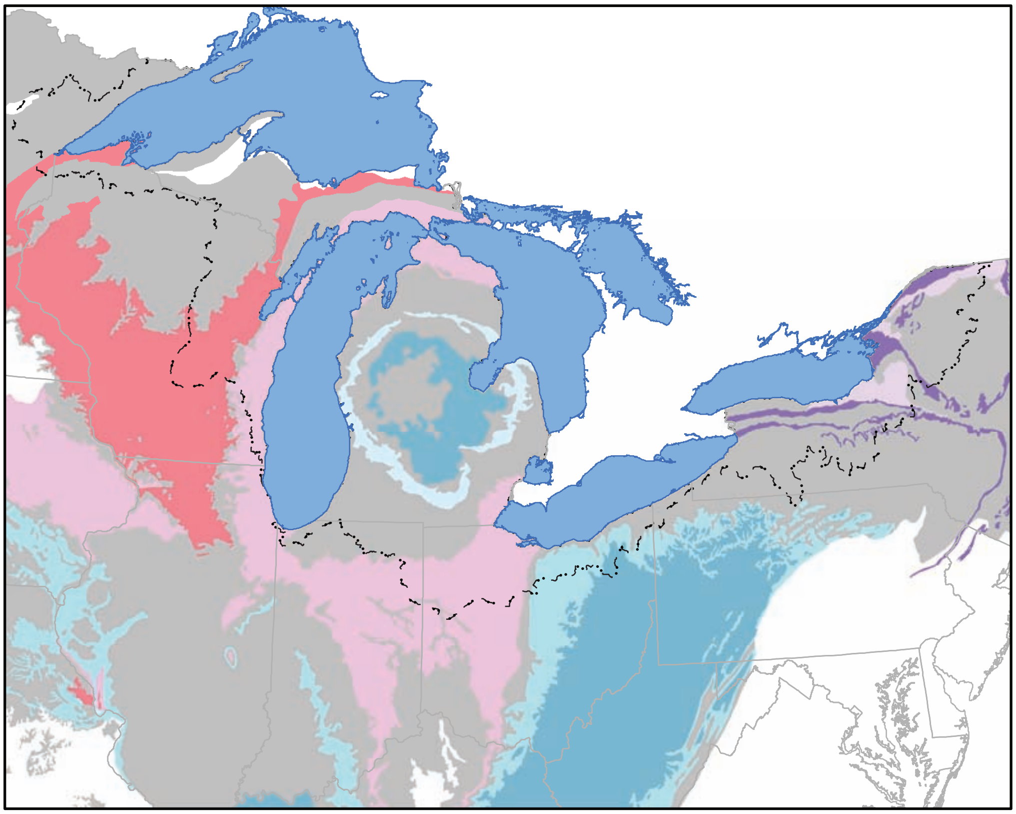
National Water Availability and Use Program

Estimate of Ground Water in Storage in the Great Lakes Basin, United States, 2006



Scientific Investigations Report 2006–5180

U.S. Department of the Interior
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By William F. Coon and Rodney A. Sheets

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U.S. Geological Survey
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Conversion Factors, Datums, and Abbreviations

Inch-Pound to International System (SI) Units		
Multiply	**By**	**To obtain**
	Length	
foot (ft)	0.3048	meter (m)
	Area	
square foot (ft^2)	0.09290	square meter (m^2)
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
gallon (gal)	3.785	liter (L)
cubic mile (mi^3)	4.166	cubic kilometer (km^3)
	Flow rate	
gallon per minute (gal/min)	0.06309	liter per second (L/s)
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Transmissivity	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

Dissolved concentrations are reported in milligrams per liter (mg/L).

Estimate of Ground Water in Storage in the Great Lakes Basin, United States, 2006

By William F. Coon and Rodney A. Sheets

Abstract

Hydrogeologic data from Regional Aquifer System Analyses (RASA) studies by the U.S. Geological Survey in the Great Lakes Basin, United States, during 1978–95, were compiled and used to estimate the total volume of water that is stored in the many aquifers of the basin. These studies focused on six regional aquifer systems: the Cambrian-Ordovician aquifer system in Wisconsin, Illinois, and Indiana; the Silurian-Devonian aquifers in Wisconsin, Michigan, Illinois, Indiana, and Ohio; the surficial aquifer system (aquifers of alluvial and glacial origin) found throughout the Great Lakes Basin; and the Pennsylvanian sandstone and carbonate-rock aquifers and the Mississippian sandstone aquifer in Michigan. Except for the surficial aquifers, all of these aquifer systems are capable of yielding substantial quantities of water and are not small aquifers with only local importance. Individual surficial aquifers, although small in comparison to the bedrock aquifers, collectively represent large potential sources of ground water and therefore have been treated as a regional system.

Summation of ground-water volumes in the many regional aquifers of the basin indicates that about 1,340 cubic miles of water is in storage; of this, about 984 cubic miles is considered freshwater (that is, water with dissolved-solids concentration less than 1,000 mg/L). These volumes should not be interpreted as available in their entirety to meet water-supply needs; complete dewatering of any aquifer is environmentally undesirable. The amount of water that is considered available on the basis of water quality and environmental, economic, and legal constraints has not been determined. The effect of heavy pumping in the Chicago, Ill., and Milwaukee, Wis., areas, which has caused the regional ground-water divide in the Cambrian-Ordovician aquifer system to shift westward, has been included in the above estimates. This shift in the ground-water divide has increased the amount of water in storage in the deep-bedrock aquifers of the Great Lakes Basin by about 36 cubic miles; however, this water is removed by wells and, after use, is mostly discharged to the Mississippi River Basin rather than to the Great Lakes Basin. The corresponding decrease in ground-water storage that has resulted from lowering of the potentiometric surface due to this heavy pumping (0.059 cubic miles) is negligible compared to the total estimated storage.

Introduction

Freshwater is vital to the economic and environmental well-being of the United States. Besides the need to maintain streamflows and lake levels for ecological purposes, freshwater is required to meet the public, domestic, commercial, and industrial needs of the American people. As developed areas expand and the competition for limited water resources intensifies in heavily populated areas and where freshwater is in limited quantities, the need for conservation, protection from pollution, and sound development of water resources becomes more important. The Nation's freshwater needs are met by withdrawals from streams, lakes, reservoirs, and ground-water systems. As water demands increase, the role of ground water to meet these demands becomes increasingly important. Ground water is the dominant source of drinking water for most rural areas, the largest source of water for irrigation and other uses in arid and most semiarid regions, and an important source of water for urban, industrial, and irrigation uses in humid areas (Heath, 1985).

Estimation of the volume of ground water that is stored in an aquifer—a water-bearing layer of rock or unconsolidated material that will yield a usable quantity of water to a well (Heath, 1983) —is an essential first step for assessing ground-water resources in a given area. Storage calculations require hydrogeologic data on the water-storage capacity and water-transmission capability of the aquifer, which are controlled by the rock type or composition of the unconsolidated materials that form the aquifer, and the amount of voids (openings), the degree of fracturing, and the interconnectedness of voids and fractures in the aquifer. Storage properties of aquifers determine the amount of water that can be released from storage. An aquifer whose upper surface is free to fluctuate under atmospheric pressure (an unconfined aquifer) can yield a greater volume of water per unit change in hydraulic head than one that is similar in all respects but is completely filled with water that is under pressure because overlying material or a rock unit restricts the movement of water (a confined aquifer). Additionally, calculations of storage are complicated by water-supply wells that can increase the area that contributes ground water to the wells and that can lower the water level and remove water from storage, at least to the point that a new equilibrium condition is attained.

In 2005, the U.S. Geological Survey (USGS) began the National Water Availability and Use Program to describe the status and trends in the availability and use of freshwater resources, which would enable informed policy decisions regarding the economic and environmental uses of the Nation's water resources (U.S. Geological Survey, 2002a). The Great Lakes Basin was selected as the pilot study area for that program (U.S. Geological Survey, 2005a). The American public and its decision makers are becoming increasingly concerned about water availability in the United States part of the Great Lakes Basin, but the ground-water resources of this basin have not been comprehensively assessed since the early 1970s (Great Lakes Basin Commission, 1975). Many conditions have changed since that assessment—competition for water has increased, new water sources and technologies have been developed, and water resources in some areas of the basin have been depleted (U.S. Geological Survey, 2005a). USGS studies of regional aquifer systems within and adjacent to the Great Lakes Basin during 1978–95 have greatly increased our knowledge of the ground-water resources of the area (Young, 1992; Westjohn and Weaver, 1998; Bugliosi, 1999; Randall, 2001) and have provided hydrogeologic data that characterize the many aquifer systems of the region.

To assess the present status of ground-water resources in the Great Lakes Basin and to provide a basis for assessing future trends in these resources, the volume of ground water that is stored in the basin needs to be estimated. The regional aquifer studies mentioned above include the hydrogeologic data on which such an estimate can be based. Therefore, as part of the Great Lakes Basin pilot study, this report (1) summarizes the hydrogeologic characteristics of the major and minor aquifers in the basin and (2) presents an estimate of the volume of ground water that is stored in these aquifers and, of this total volume and within the limitations set by water-quality criterion, the volume of freshwater. These estimates do not take into consideration any environmental, economic, or legal constraints that would limit the availability of this water for the many and varied uses in the basin. The only previous estimate of ground-water storage in the Great Lakes Basin (United States part) was "at least 1,000 mi^3" by Grannemann and others (2000); no data supporting this value were given with the estimate, however.

Hydrogeology

The Great Lakes Basin, which covers nearly 300,000 mi^2, encompasses Lakes Superior, Michigan, Huron, Erie, and Ontario, and straddles the border between the United States and Canada. Eight states border the Great Lakes, parts of which represent 59 percent of the basin on the U.S. side (fig. 1); the province of Ontario represents the 41 percent of the basin that lies on the Canadian side (Neff and others, 2005). The lakes themselves occupy nearly one-third (94,000 mi^2) of the basin area (U.S. Environmental Protection Agency, 2005).

The Great Lakes Basin is underlain almost entirely by a thick sequence of sedimentary rocks. Sandstone or carbonate-rocks, where present (fig. 2), can be reliable sources of ground water. Glacial and alluvial deposits overlie the bedrock throughout much of the basin and form productive aquifers where the deposits are well sorted and composed primarily of sand and gravel (fig. 3), especially where precipitation is appreciable.

Ground water is the source of most of the water stored in the Great Lakes. Precipitation that infiltrates the soils and percolates to the water table is the main source of recharge to the aquifers. Ground water discharges directly to the Great Lakes as seepage or indirectly as base flow (dry-weather flow) in streams and rivers. Direct seepage is believed to take place near the shores (Grannemann and others, 2000) and is a relatively small component of the total water inflow to the lakes (Croley and Hunter, 1994). Base flow, after direct precipitation on the lakes, accounts for the second largest influx of water to the lakes. Using 5,735 years of daily streamflow data from 195 streams in the U.S. part of the Great Lakes Basin, Holtschlag and Nicholas (1998) estimated that the ground-water component of lake inflow from streams and rivers averages 67.3 percent of the total flow. Neff and others (2005) did a basinwide analysis of data from 959 streamflow-monitoring stations with a combined 28,784 years of daily streamflow record to estimate the ground-water component of streamflow for the United States and Canadian areas of the Great Lakes Basin. According to data reported in Neff and others (2005), the ground-water component of streamflow averages 66 percent of total streamflow for all parts of the basin.

Four studies of regional aquifer systems in the United States part of the Great Lakes Basin have been conducted by the USGS since 1975. The Great Lakes Basin contains three major aquifer systems: the Cambrian-Ordovician aquifer system (in Wisconsin, Illinois, and Indiana), the Silurian-Devonian aquifers (in Wisconsin, Michigan, Illinois, Indiana, and Ohio), and the surficial aquifer system (aquifers of alluvial and glacial origin that are found throughout the Great Lakes Basin). The basin also contains three minor aquifer systems, all in Michigan: the Pennsylvanian sandstone aquifer, the Pennsylvanian sandstone and carbonate-rock aquifer, and the Mississippian sandstone aquifer (U.S. Geological Survey, 2003). Other small aquifer systems are present in many locations in the basin but are less important as ground-water sources on a regional basis and were not included in the ground-water-storage calculations.

The hydrogeologic data generated by the USGS regional aquifer-system analyses (RASA) have increased our knowledge of the bedrock and surficial aquifers in the Great Lakes Basin and provide a basis for estimating the amount of ground water in storage in the six aquifers named above. From the total estimated storage volume, the usable quantity of ground water—that is, the amount that contains freshwater with dissolved-solids concentrations less than 1,000 mg/L—also can be computed.

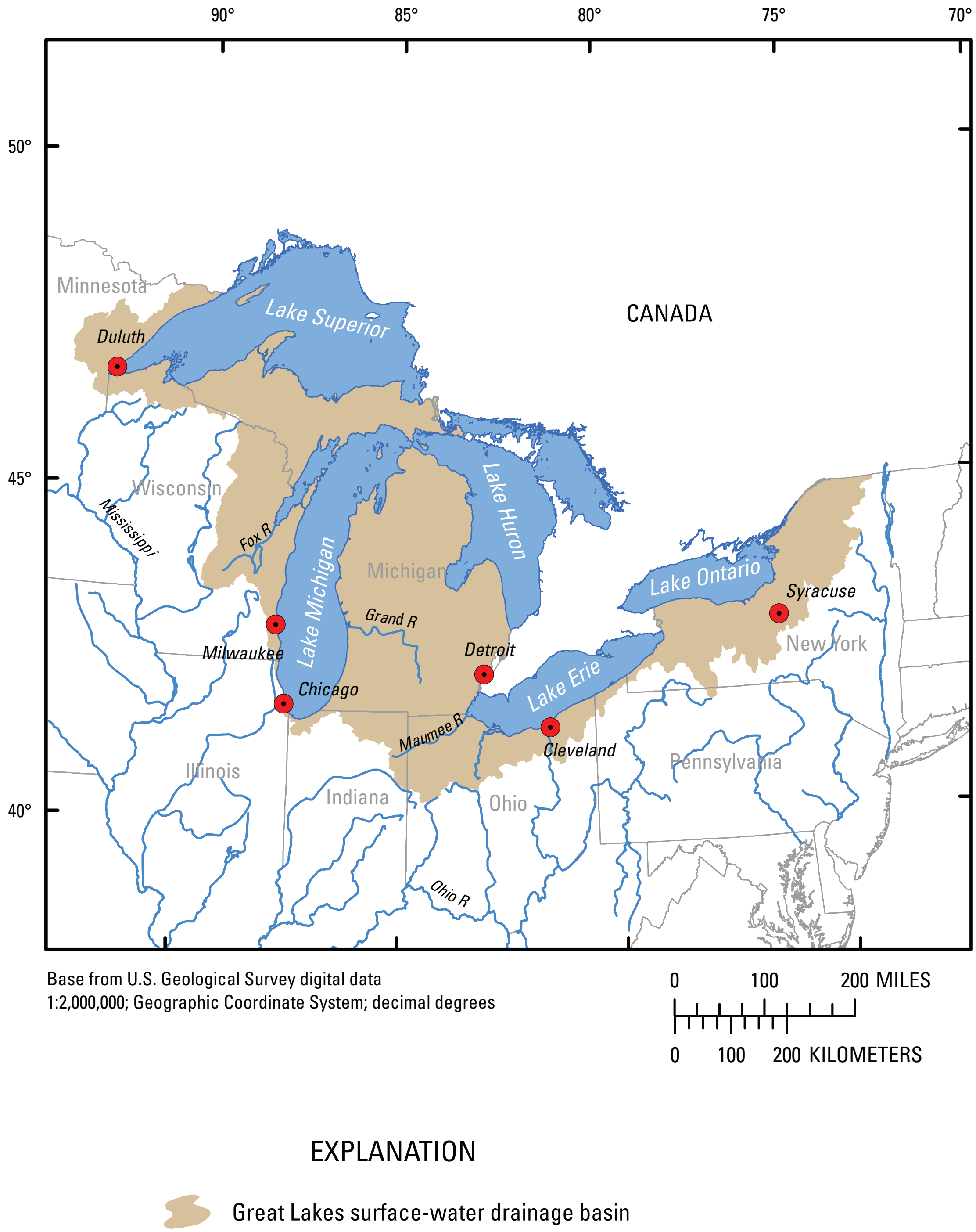

Figure 1. Drainage area of the Great Lakes Basin, United States.

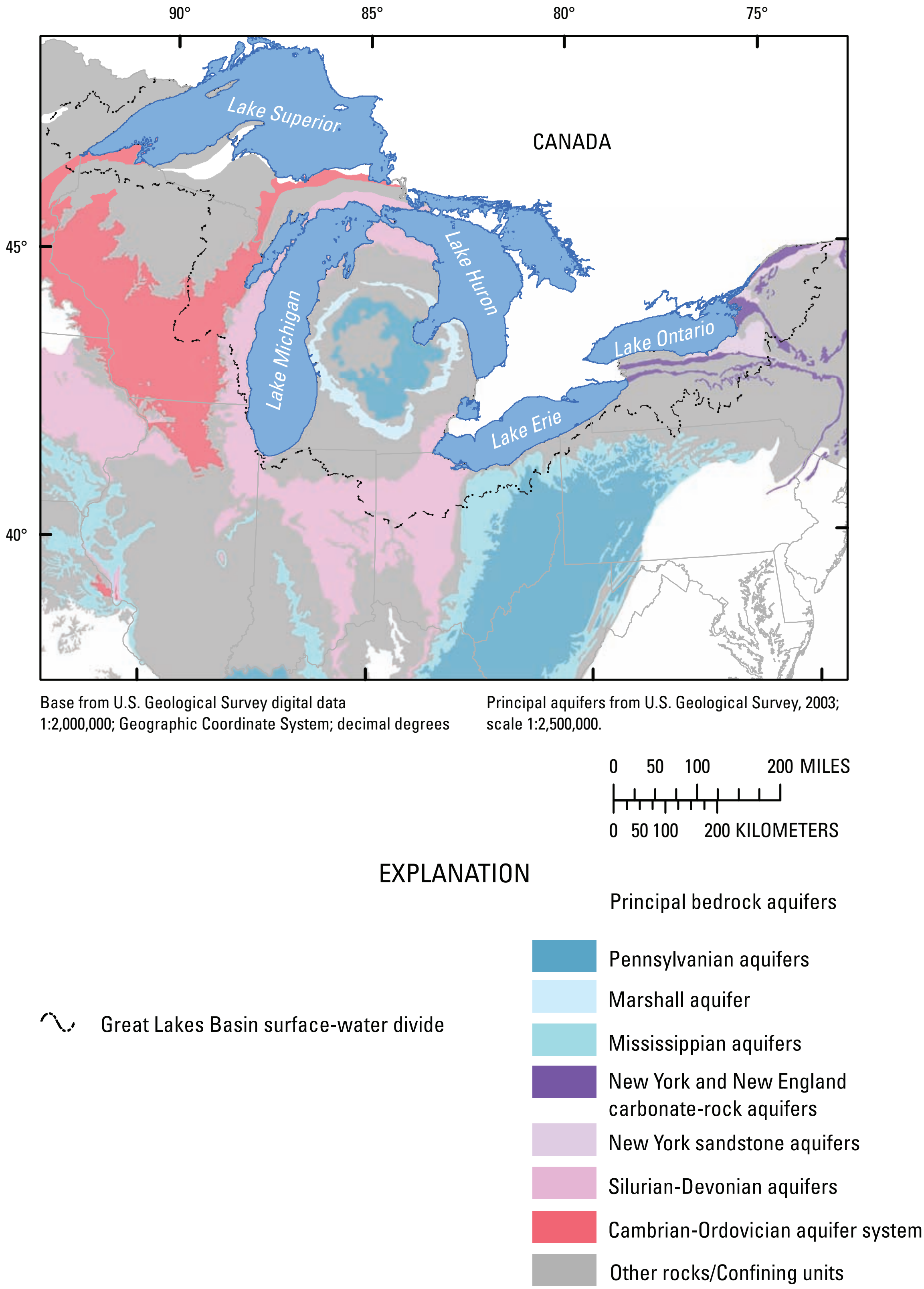

Figure 2. Regional bedrock aquifers in the Great Lakes Basin, United States.

Base from U.S. Geological Survey digital data
1:2,000,000; Geographic Coordinate System; decimal degrees

Aquifers from U.S. Geological Survey, 2002b;
scale 1:2,500,000.

0 50 100 200 MILES

0 50 100 200 KILOMETERS

EXPLANATION

Surficial Aquifer System

Southern limit of Wisconsinan glaciation (U.S. Geological Survey, 2005c)

Great Lakes Basin surface-water divide

Figure 3. Surficial aquifer system in the Great Lakes Basin, United States.

Major Aquifer Systems

The **Cambrian-Ordovician aquifer system** (fig. 2) underlies an area of about 161,000 mi^2 in parts of Minnesota, Iowa, Missouri, Wisconsin, Illinois, and Indiana (Young, 1992). The part of this system within the Great Lakes Basin is in eastern Wisconsin, northeastern Illinois, and northwestern Indiana. It primarily comprises three units: the St. Peter-Prairie du Chien-Jordan aquifer, the Ironton-Galesville aquifer, and the Mount Simon aquifer. These units are confined by several formations, the main component of which is the Maquoketa Shale (Young, 1992).

The *St. Peter-Prairie du Chien-Jordan aquifer* consists of sandstone and dolomite. The part of this aquifer within the Great Lakes Basin ranges from 0 to 500 ft in thickness (Young, 1992) and has storage coefficients of 7.5×10^{-5} to 1.2×10^{-3} (Young and Siegel, 1992). Transmissivity varies greatly, depending mainly on the degree of fracturing and solution enlargement of the dolomite, and commonly ranges from 500 to 5,000 ft^2/d (Young, 1992). Hydraulic conductivity values used in a RASA ground-water model of the Chicago-Milwaukee area ranged from 1.7 to 5.4 ft/d (Young and Siegel, 1992).

The *Ironton-Galesville Sandstone aquifer* is the most productive unit of the Cambrian-Ordovician system of northeastern Illinois and southeastern Wisconsin (Young, 1992). This unit ranges from 0 to more than 200 ft in thickness but generally is 50 to 150 ft thick within the Great Lakes Basin (Young, 1992). Storage coefficient ranges from 1.0×10^{-6} to 7.5×10^{-5}, and hydraulic conductivity ranges from 1.0 to 31 ft/d (with a median 8.4 ft/d) (Young and Siegel, 1992). The RASA ground-water models of this aquifer used narrower ranges of these values—5×10^{-5} to 9×10^{-4} for storage coefficient, and 2.6 to 8.6 ft/d for hydraulic conductivity (Mandle and Kontis, 1992). Reported transmissivity values range from 71 to 11,000 ft^2/d, but most range from the upper hundreds to about 3,000 ft^2/d (Young, 1992).

The *Mount Simon Sandstone aquifer* is the most extensive aquifer in the Cambrian-Ordovician aquifer system (Young, 1992). Its thickness exceeds 2,000 ft in northeastern Illinois but decreases to 100 ft near and north of Milwaukee, Wis. (Young, 1992). Ground-water yields are generally not commensurate with the aquifer's great thickness because only the upper few hundred feet of the aquifer, if penetrated at all, can be tapped for water supply; water in the lower part of the aquifer is saline near Lake Michigan (Young, 1992). Storage coefficient for this unit ranges from 1×10^{-6} to 1×10^{-2} (Young and Siegel, 1992), but the RASA models used a smaller range of 5×10^{-5} to 9×10^{-4} (Mandle and Kontis, 1992). Hydraulic conductivity ranges from 0.027 to 23 ft/d (Young and Siegel, 1992); the RASA models used a much narrower range—from 0.5 to 8.6 ft/d (Mandle and Kontis, 1992). Reported transmissivity of the Mount Simon aquifer ranges from 270 to 9,400 ft^2/d (Young, 1992).

Because the Cambrian-Ordovician aquifer system is entirely confined, except for a small area in southeastern Wisconsin (J.T. Krohelski, U.S. Geological Survey, written commun., 2006) and in relatively small areas where the aquifer has been dewatered, specific yields have not been calculated for individual aquifers nor for the system as a whole. Porosity values of the various stratigraphic units in eastern Wisconsin, northeastern Illinois, and western Michigan have been measured in 36 boreholes; the average porosity of the sandstone units is 15.2 percent (Carlson, 2000). Specific yields are typically somewhat lower than porosities, and an estimated specific yield of 5 percent has been applied to these units in ground-water-flow models of southeastern Wisconsin (Feinstein and others, 2005a); this value was not based on measured data, however (D.T. Feinstein, U.S. Geological Survey, written commun., 2006).

The **Silurian-Devonian aquifers** (fig. 2) underlie parts of Wisconsin, Michigan, Illinois, Indiana, and Ohio, although the part that lies in Indiana is outside the Great Lakes drainage area. Water occurs primarily in fractures, bedding planes, and other openings, and most of the active flow within fractures is within the upper 100 ft of the aquifers (Bugliosi, 1999). The RASA study boundaries of these aquifers in the Indiana-Ohio-Michigan area were partly defined on the basis of dissolved-solids concentration, which is representative of the degree of salinity and varies greatly in this area—from less than 500 to greater than 1,000 mg/L (Bugliosi, 1999). Areas where dissolved-solids concentrations in the Silurian-Devonian aquifers in southern Michigan and northeastern Ohio exceed 10,000 mg/L were excluded from the RASA study of this area and from the ground-water-storage estimates presented in this report.

The Silurian-Devonian aquifers in eastern Wisconsin and northeastern Illinois range from 0 to more than 1,000 ft in thickness but generally are 200 to 500 ft thick in their subcrop areas (Young, 1992). In this area, the aquifers are confined by Pennsylvanian, Mississippian, and Devonian rock units (Young, 1992). Aquifer thickness ranges from about 200 to 1,400 ft across northern Michigan (Mandle and Kontis, 1992) and from 200 to more than 2,500 ft elsewhere in the Great Lakes Basin, primarily northwestern Ohio and southeastern Michigan (Casey, 1996), although the average thickness in this area is about 900 ft. In northern Indiana and Ohio, these aquifers are semiconfined by surficial deposits—mostly till—or confined by overlying shales (Bugliosi, 1999). Storage coefficient ranges from 9×10^{-5} to 4.8×10^{-3} in the Wisconsin-Illinois area (Young and Siegel, 1992) and from 1×10^{-5} to 5×10^{-2} in the Indiana-Ohio-Michigan area (Joseph and Eberts, 1994). The median storage coefficient for these aquifers in Ohio and Indiana is 1×10^{-4} (Joseph and Eberts, 1994, table 7). Specific yield ranges from 0.017 to 0.03 in northeastern Illinois (Prickett and others, 1964; Sasman and others, 1981) and from 0.01 to 0.05 in Ohio (Bloyd, 1974).

Hydraulic conductivity of the Silurian-Devonian aquifers in Wisconsin and Illinois ranges from 0.068 to 40 ft/d; the RASA models used values of 0.17 to 7.9 ft/d (Young and Siegel, 1992). Hydraulic conductivity in Indiana and Ohio ranges from 0.0016 to 12 ft/d (Joseph and Eberts, 1994). Reported transmissivities in Wisconsin and Illinois range from 67 to 360,000 ft^2/d (Young, 1992), but most values range from about 670 to 2,000 ft^2/d (Young and Siegel, 1992). Transmissivity values obtained from aquifer tests in Indiana and Ohio ranged from 70 to 52,000 ft^2/d (Joseph and Eberts, 1994). Transmissivity is low where the aquifers are confined or have few fractures or solution openings and high in areas with large solution openings or extensive fractures (Young and Siegel, 1992).

The **surficial aquifer system** (fig. 3) consists of discontinuous sand-and-gravel valley-fill aquifers of glacial and alluvial origin that are found throughout the Great Lakes Basin. Although individual surficial aquifers do not cover large areas similar to those covered by the Cambrian-Ordovician and Silurian-Devonian bedrock aquifers, their ubiquity has a regional effect on ground-water resources and, therefore, allows them to collectively be treated as a regional system. To be considered a surficial aquifer, the glacial deposits must have adequate thickness and capacity to transmit water to meet local water demands at a sustained level. The presence, areal extent, and thickness of these deposits vary widely across the basin.

A RASA study of glacial (stratified-drift) deposits throughout the northeastern United States (Kontis and others, 2004) provided detailed information on surficial aquifers in the RASA study area, as well as a basis for transferring glacial-deposits' characteristics from the northeastern United States to other areas of the Great Lakes Basin where similar data were missing. Kontis and others (2004) identified more than 5,000 sand-and-gravel aquifers that were created by or in glacial meltwater. These surficial aquifers, which were defined as those with saturated thicknesses of at least 10 ft and(or) the potential for sustained yields of at least 10 gal/min to wells, underlie 12.6 percent (15,400 mi^2) of the glaciated Northeast (Kontis and others, 2004). More than 80 percent of these aquifers are valley-fill aquifers with a saturated thickness of 10 to 150 ft in valleys from about 1,000 to 10,000 ft wide (Kontis and others, 2004). The porosity of these deposits ranges from 27 to 45 percent, with a median value of about 38 percent (Kontis and others, 2004). Laboratory measurements of specific yield (or the long-term unconfined storage coefficient) range from 16 to 47 percent, with a median value of 33 percent; values obtained from pumping tests range from 3 to 13 percent—much lower than those obtained by laboratory methods (Kontis and others, 2004). Storage coefficients reported for confined surficial aquifers in the glaciated Northeast range from 10^{-4} to 10^{-2} (Kontis and others, 2004). Horizontal hydraulic conductivity typically ranges from about 50 to 500 ft/d but can be much greater, whereas vertical hydraulic conductivity is far lower than horizontal conductivity—commonly only a tenth as large (Kontis and others, 2004). Transmissivities for these surficial aquifers typically range from 1,000 to 50,000 ft^2/d (Kontis and others, 2004).

Surficial aquifers are generally absent where Precambrian crystalline rocks are at or near the land surface (Young and Siegel, 1992), as in the parts of the basin that lie in Minnesota, northern Wisconsin, and the western part of the Upper Peninsula of Michigan, but are estimated to cover about 40 percent of eastern Wisconsin and northeastern Illinois (on the basis of a surficial-aquifer map presented in Olcott (1992)). Where surficial aquifers exist in Wisconsin and adjacent states, aquifer thickness generally ranges from 50 to 100 ft, but in buried bedrock valleys the thickness commonly ranges from 200 to 400 ft (Young and Siegel, 1992). These aquifers, which are potential sources of water in Wisconsin and Illinois, typically were not developed in the past because deep wells that mostly tap the Cambrian-Ordovician aquifer system described previously have yielded a reliable and abundant water supply (Young, 1992). Degradation of the quality of water extracted from the Cambrian-Ordovician system in recent years, however, has resulted in increased withdrawals from the surficial aquifers, at least in southeastern Wisconsin (J.T. Krohelski, U.S. Geological Survey, written commun., 2006). The specific yield for these aquifers has been estimated at 0.15 (Feinstein and others, 2005a).

Surficial deposits are absent in some parts of the Lower Peninsula of Michigan but form the largest reservoir of fresh ground water in other parts of the State (Westjohn and Weaver, 1998). These deposits range in thickness from 0 to 400 ft in the eastern and southern parts of the Lower Peninsula where sand and gravel make up from 25 to 75 percent of the deposits (Westjohn and others, 1994). Aquifer tests indicate that these "thin" deposits are typically partly or fully confined; the median storage coefficient is 8×10^{-4} (C.J. Hoard, U.S. Geological Survey, written commun., 2005). The thickness of the surficial deposits increases northwestward across the Lower Peninsula from 400 ft to more than 1,000 ft (Olcott, 1992), and the percentage of sand and gravel increases to more than 75 percent in the northwestern quarter of the Lower Peninsula (Westjohn and others, 1994). These "thick" deposits are assumed to be mostly unconfined; sparse data in this area prevents verification of this assumption, however (C.J. Hoard, U.S. Geological Survey, written commun., 2005). Despite the ubiquitous presence of sand-and-gravel deposits, visual inspection of a surficial-aquifer map in Olcott (1992) indicates that only about 50 percent of the thin-deposits area and about 60 percent of the thick-deposits area are covered by surficial aquifers. Freshwater is found in most of the surficial deposits of the Lower Peninsula, except within a 1,600-mi^2 area of the Saginaw Lowlands, where deposits are thin (less than 200 ft thick) and have a low percentage of sand and gravel (less than 25 percent) and where saline water is common (Westjohn and Weaver, 1996b).

Thickness of the surficial deposits in the Great Lakes Basin of Indiana and Ohio ranges from less than 100 ft to more than 400 ft (Casey, 1996). About 90 percent of northwestern Indiana was estimated to be covered by surficial aquifers (Olcott, 1995a). Elsewhere in northern Indiana and Ohio, about 13 percent of the area—similar to the Northeast United States RASA value (Kontis and others, 2004)—was estimated to be covered by surficial aquifers; this determination was based on comparison of surficial-aquifer maps of Indiana-Ohio (Olcott, 1995a) and the northeastern United States (Olcott, 1995b). Storage coefficients for confined conditions in these states range from 2×10^{-5} to 0.3; the median value is 6×10^{-4} (Joseph and Eberts, 1994, tables 1 and 2). Specific yields for unconfined conditions range from 1.6×10^{-3} to 0.38; the median value is 0.13 (Joseph and Eberts, 1994, table 1). Horizontal hydraulic conductivity in this area ranges from 0.33 to 1,000 ft/d, and transmissivity ranges from 300 to 69,700 ft^2/d (Joseph and Eberts, 1994,). Similar to the bedrock aquifer in this area, dissolved-solids concentrations in ground water vary widely—from less than 500 to greater than 1,000 mg/L (Bugliosi, 1999).

Minor Aquifer Systems

The **Pennsylvanian and Mississippian aquifers in the Lower Peninsula of Michigan** (fig. 2) consist of three bedrock aquifers—the Saginaw aquifer (Pennsylvanian sandstone), the Parma-Bayport aquifer (Pennsylvanian sandstone and carbonate), and the Marshall aquifer (Mississippian sandstone). The total areal extent of these three aquifers is about 22,000 mi^2 (Westjohn and Weaver, 1996c). Saline ground water underlies freshwater-bearing aquifers everywhere in the Lower Peninsula of Michigan (Westjohn and Weaver, 1998). Therefore, the usable supply of freshwater (defined as water that has a maximum dissolved-solids concentration of 1,000 mg/L) is limited to those parts of each aquifer that underlie, and are in direct hydraulic connection with, glacial deposits. Elsewhere, the remaining parts of each aquifer are overlain by confining rock layers, dissolved-solids concentrations exceed 1,000 mg/L, and saline water or brine is present. Most aquifer tests indicate that these aquifers are either partly or fully confined; the median storage coefficient is 3×10^{-4} (C.J. Hoard, U.S. Geological Survey, written commun., 2005).

The *Saginaw aquifer* encompasses about 10,400 mi^2 (Westjohn and Weaver, 1998) and about 5,000 mi^2 contains freshwater (Westjohn and Weaver, 1996b). Aquifer thickness ranges from 100 to 370 ft (Westjohn and Weaver, 1996a), and porosity ranges from 3 to 34 percent; the median porosity is 20 percent (Westjohn and others, 1990). Transmissivity ranges from 130 to 2,700 ft^2/d, and vertical hydraulic conductivity ranges from 0.0001 to 55 ft/d (Westjohn and Weaver, 1996a). Hydrogeologic characteristics of this sandstone unit are matrix controlled; that is, the hydraulic values are generally a function of type and degree of cementation (Westjohn and Weaver, 1998).

The *Parma-Bayport aquifer* encompasses about 11,000 mi^2 (Westjohn and Weaver, 1998), of which 8,700 mi^2 contains freshwater and 2,300 mi^2 contains brine (Westjohn and Weaver, 1996b). Aquifer thickness generally ranges from 100 to 150 ft (Westjohn and Weaver, 1996a). Porosity of the upper part of the sandstone component of this sandstone and carbonate-rock aquifer ranges from 25 to 35 percent and decreases as the sandstone becomes increasingly consolidated with depth and as the carbonate-rock component increases; porosity of the lower part of the aquifer ranges from 2 to 25 percent (Westjohn and Weaver, 1996a). In general, porosity and hydraulic conductivity values for this aquifer are similar to those reported for the Saginaw aquifer (Westjohn and Weaver, 1996a).

The *Marshall aquifer* encompasses about 22,000 mi^2, of which about 12,000 mi^2 contains freshwater and about 10,000 mi^2 contains brine (Westjohn and Weaver, 1996c). Aquifer thickness typically ranges from 75 to 175 ft but exceeds 200 ft in the northwestern part of the aquifer (Westjohn and Weaver, 1996c). Porosity ranges from 16 to 25 percent; the median porosity is 21 percent (Westjohn and others, 1990). Transmissivity of highly fractured sandstone can range from 3,000 to 29,000 ft^2/d (Westjohn and Weaver, 1998) and hydraulic conductivity from 150 to 550 ft/d (Westjohn and Weaver, 1996c). In contrast, transmissivity of well-cemented, unfractured sandstone can range from 7 to 50 ft^2/d and hydraulic conductivity from 0.2 to 1.8 ft/d (Westjohn and Weaver, 1998).

The **Mississippian and Upper Devonian aquifers in northeastern Ohio** consist of rock units within the Cuyahoga Group (Lower Mississippian), Cussewago and Berea Sandstones (Upper Devonian; Pashin and others, 1995). Aquifers in these units are also present to a much lesser extent in northwestern Pennsylvania (fig. 2). Altogether, Mississippian and Upper Devonian units underlie about 7,790 mi^2 of the Great Lakes Basin. The Cuyahoga Group has the largest areal extent (4,650 mi^2) and is the thickest of these units (up to 170 ft), but it is sometimes considered a leaky confining unit on a regional basis because few of the sandstone or shale units with the Cuyahoga Group are highly permeable (Eberts and others, 1990). Of the aquifers in the remaining units, as much as 25 percent produce water considered to be non-potable, with dissolved-solids concentrations greater than 1,000 mg/L (Rau, 1969; Eberts and others, 1990; Barton and Wright, 1997). In fact, in parts of northeastern Ohio, the Cussewago Sandstone and Berea Sandstone are also tapped for oil and gas production (Barton and Wright, 1997). For this reason, the Mississippian and Upper Devonian aquifers in Ohio were excluded from any of the RASA studies (Bugliosi, 1999), and because the total storage in these aquifers is estimated to be small compared to the total storage in the Great Lakes Basin, these aquifers also were excluded from the ground-water-storage estimates presented in this report.

Estimate of Ground-Water Storage

The estimate of the volume of water stored in the aquifers of the Great Lakes Basin, described herein, was based mainly on data from RASA reports, which focused on regional aquifer systems that are capable of yielding substantial quantities of water. The many bedrock aquifers that are small or incapable of meeting water needs beyond those of domestic or small-community uses were excluded from the storage computation. Among these small bedrock aquifers are the Jacobsville

Table 1. Hydrogeologic characteristics of the regional aquifer systems in the United States part of the Great Lakes Basin.

[mi^2, square miles; ft, feet; S, southern; E, eastern; N, northern; NE, northeastern; NW, northwestern; <, less than; >, greater than]

Aquifer	Area (mi^2)	Area of saline water (mi^2)	Aquifer thickness (ft)	Porosity (percent)	Specific yield	Storage coefficient	References
Major aquifers							
Cambrian-Ordovician aquifer system							
St. Peter-Prairie du Chien- Jordan aquifer	24,951		0–500		0.05	7.5×10^{-5} – 1.2×10^{-3}	1, 2, 3
Ironton-Galesville aquifer	33,116		50–150		0.05	5.0×10^{-5} – 9.0×10^{-4}	1, 2, 3, 4
Mount Simon aquifer	33,116		[a]100–2,000		0.05	5.0×10^{-5} – 9.0×10^{-4}	1, 2, 3, 4
Silurian-Devonian aquifer system							
Parts of Ohio and S. Michigan only	8,616	([b])	200–1,600		0.01–0.05	1.0×10^{-5} – 5.0×10^{-2}	5, 6
E. Wisconsin and N.E. Illinois	9,695		200–500		0.017–0.03	9.0×10^{-5} – 4.8×10^{-3}	1, 2
N. Michigan	2,913		200–1,400			[c] 9.0×10^{-5} – 4.8×10^{-3}	4
Surficial aquifer system							
New York and Pennsylvania	[d]2,730		10–150	27 –45	0.16–0.47	1.0×10^{-4} – 1.0×10^{-2}	7
Minnesota and N. Wisconsin	([e])						2
E. Wisconsin and N.E. Illinois	[f]12,680		50–400		0.15	([g])	2, 3, 8, 9
N.E. Indiana	[d]395		200–400		0.0016–0.38	2×10^{-5} – 3×10^{-1}	6, 9
N. Ohio	[d]1,500		100–400		0.0016–0.38	2×10^{-5} – 3×10^{-1}	5, 6
N.W. Indiana	[h]514		100–200		0.0016–0.38	2×10^{-5} – 3×10^{-1}	6, 9
Michigan, Lower Peninsula:							
Deposits generally < 400 feet thick	[i]13,950	[j]800	50–400		([g])	8.0×10^{-4}	10, 12, 14
Deposits generally > 400 feet thick	[i]7,500		400–1,000		([g])	8.0×10^{-4}	8, 12
Minor aquifers (in the central Lower Peninsula of Michigan)							
Pennsylvanian sandstone							
Saginaw aquifer	10,400	5,400	100–370	3–34		3.0×10^{-4}	10, 11, 12, 14
Pennsylvanian sandstone and carbonate rock							
Parma-Bayport aquifer	11,000	2,300	100–150	2–35		3.0×10^{-4}	10, 11, 12, 14
Mississippian sandstone							
Marshall aquifer	22,000	10,000	75–175	16–25		3.0×10^{-4}	10, 11, 13, 14

[a] Water in the Mount Simon sandstone is generally saline below the upper few hundred feet in eastern Wisconsin and northeastern Illinois (Young, 1992).

[b] Not included in storage estimate.

[c] Same values as for eastern Wisconsin.

[d] Thirteen percent of total area is considered surfical aquifer system; that is, sand-and-gravel glacial deposits with a saturated thickness of at least 10 ft and(or) the potential for sustained yields of at least 10 gallons per minute to wells (Kontis and others, 2004).

[e] Surfical aquifer system is generally absent where Precambrian crystalline rocks are at or near the land surface (Young and Siegel, 1992).

[f] Forty percent of total area is assumed to be covered by surfical aquifer system, as defined in footnote d.

[g] Values for the stratified-drift aquifers in New York and Pennsylvania (from Kontis and others, 2004).

[h] Ninety percent of total area is assumed to be covered by surfical aquifer system, as defined in footnote d.

[i] Data presented by Westjohn and Weaver (1996b) indicate that the area of Michigan in which “thin” glacial deposits (deposits generally less than 400 ft thick) occur represents about 69 percent (about 27,900 mi^2) of the Lower Peninsula; the remaining 31 percent (about 12,500 mi^2) is assumed to be covered by “thick” glacial deposits (generally greater than 400 ft thick). Of these total areas, 50 percent of the thin-deposits area and 60 percent of the thick-deposits area are assumed to be covered by surfical aquifer system as defined in footnote d.

[j] Ground water in 1,600 mi^2 of the area covered by thin glacial deposits (the Saginaw Lowlands) is saline. Fifty percent of this area is assumed to consist of surfical aquifer system, as explained in footnote i.

REFERENCES: (1) Young, 1992. (2) Young and Siegel, 1992. (3) Feinstein and others, 2005a. (4) Mandle and Kontis, 1992. (5) Casey, 1996. (6) Joseph and Eberts, 1994. (7) Kontis and others, 2004. (8) Olcott, 1992. (9) Lloyd and Lyke, 1995. (10) Westjohn and Weaver, 1998. (11) Westjohn and Weaver, 1996a. (12) C.J. Hoard, U.S. Geological Survey, written commun., 2005. (13) Westjohn and Weaver, 1996. (14) Westjohn and Weaver, 1996c.

Table 2. Estimation of ground water stored in the regional aquifer systems in the United States part of the Great Lakes Basin.

[ft, feet; mi^2, square miles; mi^3, cubic miles; S, southern; E, eastern; N, northern; NE, northeastern; NW, northwestern; <, less than; >, greater than]

Aquifer	Area of aquifer (mi^2)		Average thickness of aquifer (ft)	Hydraulic head (ft)	Specific yield	Storage coefficient	Storage (mi^3)	
	Total	With fresh-water					Total	Freshwater
Major aquifers								
Cambrian-Ordovician aquifer system								
St. Peter-Prairie du Chien- Jordan aquifer	24,951		250	750	0.05	[a]3.0×10^{-4}	60.13	60.13
Ironton-Galesville aquifer	33,116		100	600	.05	[a]2.12×10^{-4}	32.16	32.16
Mount Simon aquifer:								
Entire aquifer	33,116		1,050	1550	.05	[a]2.12×10^{-4}	331.34	
Freshwater part		33,116	[b]300	800	.05	[a]2.12×10^{-4}		95.14
Silurian-Devonian aquifer system								
Parts of Ohio and S. Michigan only	8,616		900	1000	.03	[c]1.0×10^{-4}	44.22	44.22
E. Wisconsin and N.E. Illinois	9,695		350	450	.02	[a]6.6×10^{-4}	11.47	11.47
N. Michigan	2,913		800	900	.03	[a]6.6×10^{-4}	13.57	13.57
Surficial aquifer system								
New York and Pennsylvania:								
Unconfined	[d]1,365		80	80	.33	[a]1.0×10^{-3}	6.84	6.84
Confined	[d]1,365		80	180	.33	[a]1.0×10^{-3}	6.87	6.87
E. Wisconsin and N.E. Illinois:								
Unconfined	[e]6,340		225	225	.15	[f]1.0×10^{-3}	20.09	20.09
Confined	[e]6,340		225	325	.15	[f]1.0×10^{-3}	20.14	20.14
N.E. Indiana and N. Ohio								
Unconfined	[d]947		250	250	.13	[c]6.0×10^{-4}	5.86	5.86
Confined	[d]947		250	350	.13	[c]6.0×10^{-4}	5.87	5.87
N.W. Indiana	[g]514		150	150	.13	[c]6.0×10^{-4}	1.91	1.91
Michigan, Lower Peninsula								
Deposits generally > 400 ft thick and assumed to be mostly unconfined	[h]7,500		700	700	[f].33	[c]8.0×10^{-4}	328.92	328.92
Deposits generally < 400 ft thick and assumed to be mostly confined	[i]13,950	[i]13,150	225	325	[f].33	[c]8.0×10^{-4}	196.86	185.57
Minor aquifers (in the central Lower Peninsula of Michigan)								
Pennsylvanian sandstone								
Saginaw aquifer:								
Confined	5,400		235	435	[j].2	[c]3.0×10^{-4}	48.20	
Partly confined	5,000	5,000	235	235	[j].2	[c]3.0×10^{-4}	44.57	44.57
Pennsylvanian sandstone and carbonate rock								
Parma-Bayport aquifer:								
Confined	2,300		125	325	[j].2	[c]3.0×10^{-4}	10.93	
Partly confined	8,700	8,700	125	125	[j].2	[c]3.0×10^{-4}	41.26	41.26
Mississippian sandstone								
Marshall aquifer:								
Confined	10,000		125	325	[j].21	[c]3.0×10^{-4}	49.90	
Partly confined	12,000	12,000	125	125	[j].21	[c]3.0×10^{-4}	59.74	59.74
TOTAL							1,340.86	984.34

[a] Average of log-transformed range of values.

[b] Water in the Mount Simon sandstone is generally saline below the upper few hundred feet in eastern Wisconsin and northeastern Illinois (Young, 1992). Used 300 ft for thickness of freshwater part of aquifer.

[c] Median value.

[d] Total aquifer area is 13 percent of the surface-water drainage area (from Kontis and others, 2004). This area was divided equally between unconfined and confined aquifers.

[e] Forty percent of total area is considered surfical aquifer system. This area was divided equally between unconfined and confined aquifers. Of the area affected by the shift in the ground-water divide, 13 percent of the estimated ground water in the surficial aquifers is assumed to recharge the Great Lakes Basin bedrock aquifers and to flow toward the pumping centers in the Chicago-Milwaukee area (U.S. Geological Survey, 2004). The remainder of the water discharges to local streams that drain to basins outside of the Great Lakes Basin.

[f] Used values for the surfical aquifer system in the glaciated northeastern United States (Kontis and others, 2004).

[g] Ninety percent of total area is considered surfical aquifer system; all of which was assumed to be unconfined.

[h] Sixty percent of total area is considered surfical aquifer system.

[i] Fifty percent of total area is considered surfical aquifer system.

[j] Porosity value.

sandstone aquifer in northern Michigan (Olcott, 1992) and the carbonate-rock and sandstone aquifers in western and central New York (Olcott, 1995b). Many individual surficial aquifers fit this exclusionary definition but were collectively treated as components of a regional surficial aquifer system, so their contributions were included in the ground-water-storage estimates. The parts of the Silurian-Devonian carbonate-rock aquifer in northern Indiana, northwestern and eastern Ohio, and southern Michigan that were excluded from the Midwestern Basins and Arches RASA study on the basis of dissolved-solids concentrations greater than 10,000 mg/L (Bugliosi, 1999), also were excluded from the storage calculations.

Methods of Computation

Ground-water storage volume is a function of storativity, the volume of water released from (or taken into) storage per unit surface area of aquifer per unit decline (or rise) in hydraulic head (Freeze and Cherry, 1979). In a confined aquifer, storativity, also referred to as the storage coefficient, is the product of specific storage and aquifer thickness. Storage volume in a confined aquifer is solely a function of the compressibility of pore water and the mineral matrix of the aquifer. In an unconfined aquifer, where storativity can be several orders of magnitude larger than if the aquifer is confined, the amount of water stored owing to the compressibility of water and the aquifer is negligible. Therefore, in an unconfined aquifer, storativity essentially equals specific yield and reflects the volume of water that will drain from an aquifer under the influence of gravity.

The volume of ground water (V) stored in each of the Great Lakes aquifer systems was computed as the product of storativity, aquifer area, and the change in hydraulic head (Fetter, 1980):

$$V = S\,A\,\Delta h \qquad (1)$$

where

S is storativity (or storage coefficient), dimensionless,
A is area underlain by an aquifer, in square feet, and
Δh is change in hydraulic head, in feet.

As mentioned above, storativity is defined differently on the basis of whether the aquifer is confined or not. For a confined aquifer, and until a confined aquifer becomes unconfined as a result of drawdown of the potentiometric surface below the top of the aquifer, storativity is defined as follows (Fetter, 1980):

$$S = b\,S_s \qquad (2)$$

where

b is aquifer thickness, in feet, and
S_s is specific storage, per foot.

Combining equations 1 and 2, the volume of water stored in a confined aquifer becomes

$$V = (b\,S_s)\,A\,\Delta h \qquad (3)$$

For an unconfined aquifer, storativity is defined as follows (Fetter, 1980):

$$S = S_y + h\,S_s \qquad (4)$$

where

S_y is specific yield, dimensionless,
h is hydraulic head, which in the case of an unconfined aquifer, equals the saturated thickness of the aquifer, in feet, and
S_s is as previously defined.

Combining equations 1 and 4, the volume of water stored in an unconfined aquifer becomes

$$V = (S_y + h\,S_s)\,A\,\Delta h \qquad (5)$$

To compute the total storage available in an aquifer, Δh must be replaced with different terms depending on whether confined or unconfined conditions prevail. In a confined aquifer, Δh would equal the height of the potentiometric surface above the bottom of the aquifer (*H*). In an unconfined aquifer, Δh would equal the saturated thickness (*h*), and the maximum *h* could be approximated by the aquifer thickness (*b*), as would be the case when an aquifer changes from a confined to an unconfined condition. Computation of total storage is the sum of “expansion water”—water released due to compression of the aquifer solids and expansion of water that accompanies a decline in head under both confined and unconfined conditions—and (2) “gravitational water”—water released due to gravitational drainage under unconfined conditions. In a confined aquifer, gravitational drainage will not occur until the potentiometric surface drops below the top of the aquifer and the aquifer becomes unconfined; that is, an aquifer whose upper surface is a water table, free to fluctuate under atmospheric pressure. Therefore, the total storage volume (V_t) of a confined aquifer is the sum of expansion water when the potentiometric surface is above the top of the aquifer and “gravitational plus expansion” water when the potentiometric surface is below the top of the aquifer:

$$\begin{aligned} V_t &= [(b\,S_s)\,A\,(H-b)] + [(S_y + b\,S_s)\,A\,b] \\ &= [(b\,S_s)\,A\,(H-b)] + [S_y\,A\,b] + [b\,S_s\,A\,b] \\ &= [(b\,S_s)\,A\,(H-b+b)] + [S_y\,A\,b] \\ &= [b\,S_s\,A\,H] + [S_y\,A\,b]. \end{aligned} \qquad (6)$$

The total storage volume (V_t) of an unconfined aquifer is the sum of gravitational and expansion water in the saturated thickness (*h*) of the aquifer. Although gravitational water dominates the water released from storage in an unconfined aquifer, the expansion-water component is included in the following equation so as to be consistent with equation 6:

$$V_t = (S_y + h\,S_s)\,A\,h = [S_y\,A\,h] + [S_s\,A\,h^2]. \qquad (7)$$

Assumptions and Limitations Inherent in Storage Estimate

The hydrogeologic characteristics of the many aquifers included in estimation of the volume of water stored in the Great Lakes Basin vary widely between aquifers and within a given aquifer (table 1). Therefore the total ground-water storage for each aquifer was calculated from average, median, or other representative values of these characteristics—thickness, porosity, specific yield, and storage coefficient—for the parts of each aquifer that lie within the Great Lakes Basin (table 2). Assumptions and transference of data from one area to another were necessary to provide reasonable parameter values for a given aquifer in an area that was not studied and to allow inclusion of all the principal aquifers in the storage estimate. Many of the RASA reports discuss the development of ground-water models and present the parameter values used therein; these model values generally constrained the range of parameter values that were deemed most reasonable for the specific region and, where given, were preferentially used in the calculations of storage.

Hydraulic head can vary greatly across an aquifer; therefore, representative values of the height of the potentiometric surface above the top of the confined aquifers in the study area were estimated from generalized hydrogeologic sections presented in the RASA reports. These values were estimated as the average distance from the top of the aquifer to the land surface, and are considered gross approximations in that large discrepancies can be noted at any specific point in a particular aquifer; nonetheless, they provide a basis for simplifying a calculation that otherwise would be unmanageable. The following representative values were added to the average thickness of their respective aquifers and were used in the calculation of ground-water storage in the confined aquifers of the Great Lakes Basin: 500 ft for the Cambrian-Ordovician aquifer system (Young, 1992), 100 ft for the Silurian-Devonian (Bugliosi, 1999) and the surficial aquifer systems (T.S. Miller, U.S. Geological Survey, oral commun., 2005), and 200 ft for the minor aquifer systems of Michigan (Westjohn and Weaver, 1998). To compute the maximum volume of water stored in an unconfined aquifer, the aquifer thickness (b) was used as an approximation of the saturated thickness (h). In reality, the saturated thickness of an unconfined aquifer would be some value less than the total thickness of the aquifer.

Specific-yield values are applicable to unconfined conditions; however, these values were required to compute the total storage volume for confined aquifers in the case where pumping might lower the potentiometric surface below the top of an aquifer, thus creating an unconfined condition. The possibility of this actually occurring on a widespread scale in any of the bedrock aquifers included in this study is unanticipated. Median values of specific yield and storage coefficients for a given aquifer were used when available. Alternatively, averages of the maximum and minimum values in a range of published values were used. Because storage-coefficient values can span several orders of magnitude, averaging was performed on log-transformed values. For the minor bedrock aquifers in the Lower Peninsula of Michigan, where specific yields were unavailable, porosities (which are typically slightly larger than specific yields) were substituted in the storage calculations.

Calculation of ground-water storage in surficial aquifers was complicated by the wide range in aquifer thickness, extent, and composition and by lack of data in many parts of the Great Lakes Basin. The problem of missing data was resolved by assuming that the surficial-aquifer characteristics documented by the RASA study of the northeastern United States (Randall, 2001; Kontis and others, 2004) would be comparable, and therefore transferable, to surficial aquifers elsewhere in the Great Lakes Basin. Only about 13 percent of the glaciated northeastern United States is covered by deposits that are, or could be considered, potential aquifers, and more than 80 percent of these are valley-fill aquifers (Randall, 2001; Kontis and others, 2004). Therefore, the values found for valley-fill stratified-drift aquifers in the glaciated Northeast were applied to the sand-and-gravel surficial aquifers found elsewhere in the Great Lakes Basin unless otherwise stipulated in a published report. The areal extent of surficial aquifers was estimated from surficial-aquifer maps in Olcott (1992; 1995a) in comparison with a surficial-aquifer map of northeastern United States (Olcott, 1995b), in which surficial aquifers had been estimated to cover about 13 percent of the land area. The coverage of surficial aquifers in northeastern Indiana and northern Ohio appeared to be similar to that in the northeastern United States; therefore the percentage of coverage was estimated as 13 percent in these areas. Elsewhere in the Great Lakes Basin, surficial aquifers were assumed to cover 40 percent of eastern Wisconsin and northeastern Illinois; 90 percent of northwestern Indiana; and 60 and 50 percent of the thick-deposits and thin-deposits areas, respectively, of the Lower Peninsula of Michigan.

No estimate of the percentage of surficial aquifers that are confined (as opposed to unconfined) is available. This distinction strongly affects the calculation of ground-water storage because the amount of water that can be released from an unconfined aquifer (with a median specific yield of 0.33 in the glaciated Northeast) is at least two orders of magnitude greater than that which can be released from a confined aquifer (with an average storage coefficient of 0.001 in the glaciated Northeast). Therefore, the estimated aquifer area in parts of the basin that were judged to contain a mix of unconfined and confined surficial aquifers was divided equally (50 percent each) between unconfined and confined conditions.

The confining units—that is, those rock layers (typically shales) or glacial deposits (typically till or lacustrine silt and clay) that impede the movement of water and overlie an aquifer—can also be potential sources of water. The hydrogeologic characteristics of confining units are much different from aquifers in that their hydraulic conductivity or rate of water transmission is very low. Although present, the water in the confining units is difficult to remove and is therefore considered unavailable; these water volumes were not included in the storage calculations.

The aquifer areas were either obtained directly from a RASA report or calculated from the summation of subbasin drainage areas within a GIS (Geographic Information System) database of the Great Lakes Basin (U.S. Geological Survey, 2005b). The boundaries of the aquifers were defined by the current known or assumed boundary of the Great Lakes Basin ground-water divide (Sheets and Simonson, 2006). In the absence of detailed information, the ground-water divide was assumed to approximate the surface-water divide.

In southeastern Wisconsin and northeastern Illinois, the ground-water divide of the deep Cambrian-Ordovician sandstone aquifer system is many miles west of the surface-water divide. In addition, recent studies have documented that heavy pumping has caused the cones of depression that surround the pumping centers in Chicago and Milwaukee to coalesce and to shift the divide westward 10 to 20 mi (Young, 1992; U.S. Geological Survey, 2004; Feinstein and others, 2005b; Sheets and others, 2005; fig. 4). This relocation of the ground-water divide has increased the area of the Cambrian-Ordovician aquifer system that currently contributes water to the pumping centers by about 3,220 mi^2 (Sheets and Simonson, 2006). This water is captured within the area defined by the Great Lakes deep ground-water divide and therefore is included in the storage estimate. Based on a regional ground-water-flow model of southeastern Wisconsin (Feinstein and others, 2005b), however, this water actually does not contribute to the Great Lakes Basin because it is removed from the deep-bedrock aquifers by wells, most of which are located in the Mississippi River Basin. Assuming that well water, after use, is returned to surface-water bodies in the same basin in which a given well is located, about 5 percent of the pumped water is discharged to the Great Lakes Basin, whereas about 95 percent is discharged to the Mississippi River Basin (D.T. Feinstein, U.S. Geological Survey, written commun., 2006).

The shift in the deep-aquifer ground-water divide in southeastern Wisconsin and northeastern Illinois has little effect on the shallow aquifers, except where drawdown of the potentiometric surface of the deep aquifer induces downward movement of water from the shallow aquifers. The surficial glacial deposits that might be affected by this shift were estimated to cover about 40 percent of a 2,630-mi^2 area (Young and others, 1989) or about 1,052 mi^2. The estimate of the ground water stored in the surficial aquifers of this area was further complicated, not just by movement of the deep ground-water divide due to heavy pumping, but by the presence or absence of the confining rock layer, the Maquoketa Shale. One-hundred percent of the ground water stored in the surficial aquifers that lie within the area defined by the Great Lakes Basin surface-water divide was included in the storage estimate. Where surficial aquifers overlie the Maquoketa Shale between the surface-water divide and the pre-development deep-ground-water divide (fig. 4)–which approximates the location of the western edge of the Maquoketa Shale (D.T. Feinstein, U.S. Geological Survey, written commun., 2006)–about 1.9 percent of the recharge to the surficial aquifers leaks through the Maquoketa Shale to the deep Cambrian-Ordovician aquifer system (U.S. Geological Survey, 2004). West of the Maquoketa Shale, between the pre-development and current deep-ground-water divides, about 12.8 percent of the recharge to the surficial aquifers leaks through to the deep aquifer system (U.S. Geological Survey, 2004). These percentages, rounded to 2 and 13 percent, respectively, were assumed to be reasonable approximations of the volume of ground water stored in the surficial aquifers that would supply the deep aquifers and were used for this purpose to calculate the contributions of these surficial aquifers to the Great Lakes Basin storage volume. The remaining water stored in these surficial aquifers discharges to local streams that drain to adjacent basins and was not included in the storage estimate.

Storage Volume

The total volume of ground water stored in the many aquifer systems in the Great Lakes Basin was estimated to be about 1,340 mi^3 (table 2), which is greater than the earlier estimate of 1,000 mi^3 by Grannemann and others (2000) and more than the volume of Lake Michigan (1,180 mi^3 at low-water datum; U.S. Environmental Protection Agency, 2005). Of this total, about 984 mi^3 is considered freshwater; that is, water with dissolved-solids concentrations less than 1,000 mg/L. Several parts of certain aquifers contain water that is saline (dissolved-solids concentrations exceeding 1,000 mg/L); these include the surficial deposits in a 1,600-mi^2 area of the Saginaw Lowlands in Michigan, below the upper few hundred feet of the Mount Simon Sandstone aquifer in Wisconsin and northeastern Illinois, and the confined parts of the sandstone aquifers in the Lower Peninsula of Michigan. The storage volumes of each of these aquifer parts, which were included in the total storage estimate, were not included in the freshwater estimate.

The movement of the ground-water divide along the western side of Lake Michigan has increased the volume of ground water stored within the Great Lakes Basin by about 36 mi^3. The major component of this increase—almost 90 percent—is the estimated ground water stored in the Mount Simon Sandstone aquifer; water below the upper few hundred feet is considered impotable, however. Therefore, of the total increase in storage, only about 13 mi^3 —including 9.2 mi^3 contained in the upper part of the Mount Simon aquifer—is considered freshwater. Heavy pumping in the Chicago-Milwaukee area has lowered the potentiometric surface and decreased the quantity of ground water stored in the confined Cambrian-Ordovician aquifer system. In southeastern Wisconsin this decrease in storage is reported to be about 65.4 billion gallons or 0.059 mi^3 (U.S. Geological Survey, 2004) and represents less than 0.005 percent of the total estimated storage. Drawdown in the deep aquifers can cause an increase in downward leakage from shallow bedrock and surficial aquifers, however this downward leakage in Wisconsin is inhibited by the

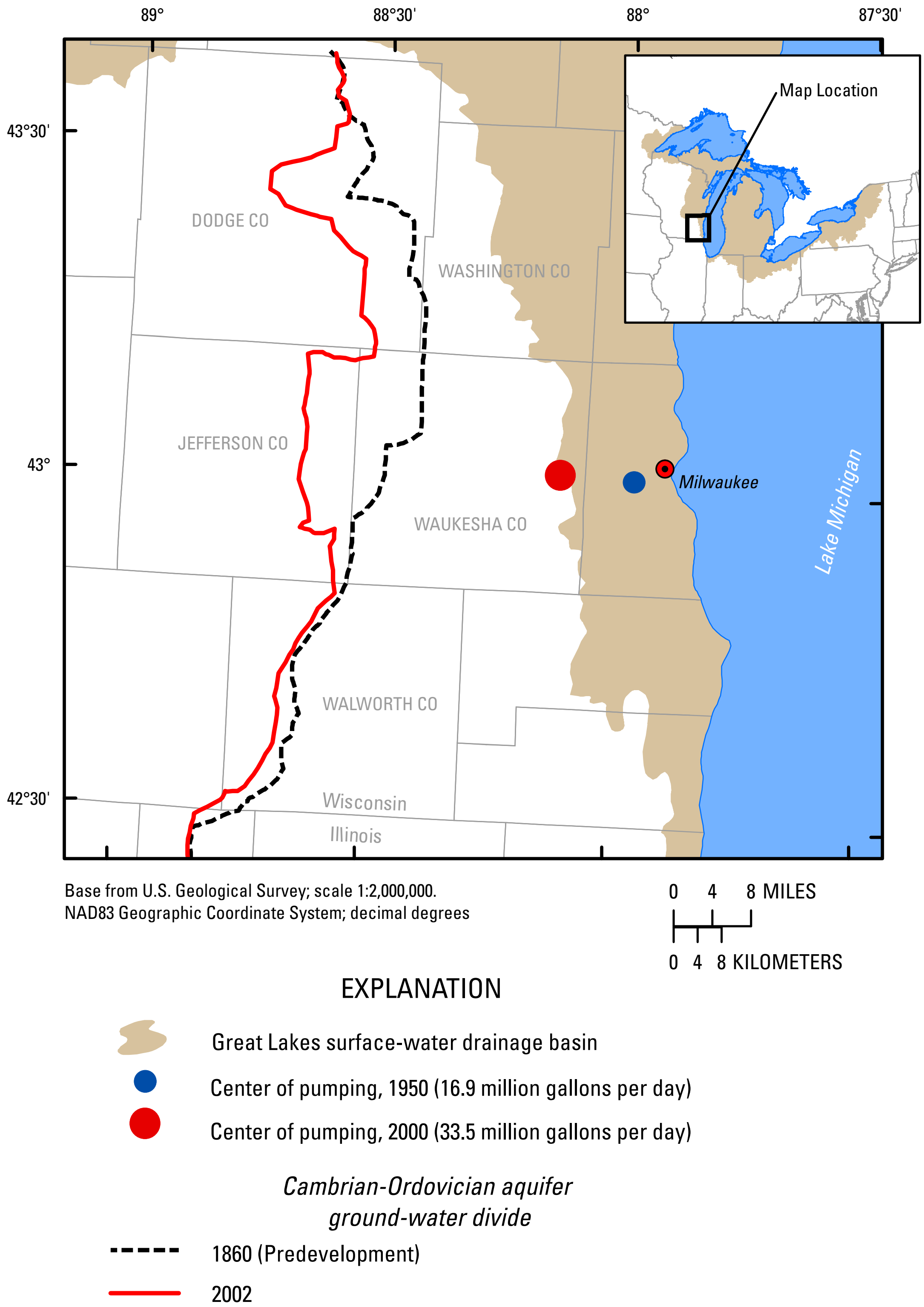

Figure 4. Movement of simulated ground-water divide in the Cambrian-Ordovician aquifer in response to intensive pumping in the Milwaukee, Wisconsin area (modified from Feinstein and others, 2005b).

Maquoketa Shale confining unit and most of the leakage is derived not from storage (drainage of pores in the overlying unconfined aquifers) but through capture of water that would otherwise flow toward and into streams and lakes. Therefore, the loss of storage in the shallow aquifers through the pumping of the deep aquifers is considered to be negligible and is not represented in the 0.059-mi^3 loss from Cambrian-Ordovician aquifer system (D.T. Feinstein, U.S. Geological Survey, Milwaukee, Wis., written commun., 2005).

Uncertainty in Storage Estimate

The computed storages are estimated values on a regional basis and should not be interpreted to mean that this quantity of water is available in its entirety to meet water-supply needs. To do so would require the complete dewatering of an aquifer, an action that is environmentally undesirable. The amount of water that is considered available on the basis of water quantity and quality and on the basis of environmental, economic, and legal constraints has not been determined.

The original estimate of total ground-water storage—1,340 mi^3—was based on average, median, or other representative values of the hydrogeologic parameters that were included in the ground-water storage formulas (eqs. 6 and 7). As a measure of the uncertainty in the total storage that might have arisen from selection of these values, the storage was recalculated with low and high values that bracketed the respective ranges of aquifer thickness, hydraulic head, specific yield, and storage coefficient, as listed in table 1. For each recalculation, only one parameter was adjusted and in only one direction—either lower or higher; however, this adjustment was done for all aquifers in the basin. The only exception to this rule concerned aquifer thickness; for this parameter, a decrease or increase required a comparable adjustment to the hydraulic-head value as well. Because hydraulic heads were estimated to begin with, the assessment of uncertainty associated with this parameter was performed on hydraulic-head values that were 50 percent lower and 50 percent higher than their original values. For parameters for which a range of values was not given in the reference documents, the original parameter value was retained. The eight recalculated storage values (table 3) ranged from 464 to 2,217 mi^3, both values of which pertained to changes in aquifer thicknesses and differed from the original estimate by 65 percent in both directions. Changes in specific yields produced the next largest deviation from the original estimate—about 38 percent. The remaining estimates departed from the original estimate by less than 14 percent.

Table 3. Uncertainty in the estimate of ground-water storage in the United States part of the Great Lakes Basin.

[Values are in cubic miles. The original estimate of storage—1,340 cubic miles—was based on average, median, or other representative values of the hydrogeologic parameters that are included in the ground-water storage formulas. (See equations 6 and 7.) The following estimates are based on low and high values of the indicated hydrogeologic parameter (table 1).]

Adjusted hydrogeologic parameter	Storage estimate	
	Low value	High value
Aquifer thickness (including required adjustment of hydraulic-head values)	464	2,217
Hydraulic head	1,156	1,345
Specific yield	830	1,819
Storage coefficient	1,337	1,480

Summary

Hydrogeologic data from Regional Aquifer System Analyses (RASA) studies that were conducted by the U.S. Geological Survey in the Great Lakes Basin, United States, during 1978–95 were compiled and used to estimate the volume of water stored in the many aquifers of the basin. Storage estimates focused on six regional aquifer systems. Three are major aquifer systems: the Cambrian-Ordovician aquifer system (in Wisconsin, Illinois, and Indiana), the Silurian-Devonian aquifers (in Wisconsin, Michigan, Illinois, Indiana, and Ohio), and the surficial aquifer system (aquifers of alluvial and glacial origin that are found throughout the Great Lakes Basin). Three are minor aquifer systems: the Pennsylvanian sandstone aquifer, the Pennsylvanian sandstone and carbonate rock aquifer, and the Mississippian sandstone aquifer (all in Michigan) (U.S. Geological Survey, 2003). Parts of these aquifers that were excluded from the RASA studies due to poor water quality and other small aquifer systems throughout the basin that are less important as ground-water sources on a regional basis were not included in the storage estimates.

Summation of ground-water volumes in the many regional aquifers of the basin indicates that about 1,340 mi^3 of water is in storage; of this, about 984 mi^3 is considered freshwater (that is, water with dissolved-solids concentration less than 1,000 mg/L). These volumes should not be interpreted as available in their entirety to meet water-supply needs; complete dewatering of any aquifer is environmentally undesirable. The amount of water that is considered available on the basis of water quality and environmental, economic, and legal constraints has not been determined. Heavy pumping in the Chicago and Milwaukee areas has caused the ground-water divide in the Cambrian-Ordovician aquifer system to shift westward and thereby has increased the area of the deep-bedrock aquifer that can potentially contribute water to the Great Lakes Basin. This increased contribution—about 36 mi^3, of which about 13 mi^3 is considered freshwater—was included in the ground-water-storage estimates above. The corresponding decrease in ground-water storage that has resulted from lowering of the potentiometric surface due to this heavy pumping (0.059 mi^3) is less than 0.005 percent of the total estimated storage.

Uncertainty in the storage estimates was assessed by recalculating the storage with low and high values of selected parameters—aquifer thickness, hydraulic head, specific yield, and storage coefficient. The recalculated storages differed from the original value by 65 percent when aquifer thicknesses were varied by each aquifer's respective range in thickness. The differences between the original and recalculated storages were less than 40 percent when values for the other parameters were varied.

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Glossary

Definitions adapted from Freeze and Cherry, 1979; Heath, 1983; Miller, 2004; or Kontis and others, 2004.

Confined aquifer (artesian aquifer) – an aquifer that is completely filled with water under pressure and that is overlain by material that restricts the movement of water.

Hydraulic gradient – change in hydraulic head per unit distance.

Hydraulic head – the height of the free surface of a body of water above a given point beneath the surface; composed of elevation head and pressure head.

Hydraulic conductivity – the capacity of a geologic material or deposit to transmit water. It is expressed as the volume of water that will move in a unit time under a unit hydraulic gradient through a unit area.

Potentiometric surface – a surface that represents the total head in an aquifer; that is, it represents the height above a datum plane at which the water level stands in tightly cased wells that penetrate an aquifer.

Specific storage – the volume of water released from or taken into storage in a unit volume of aquifer per unit change in hydraulic head owing to the compressibility of pore water and the mineral matrix of the aquifer.

Specific yield – the volume of water that will drain from a unit volume of an unconfined aquifer under the influence of gravity. It is the long-term unconfined storage coefficient.

Storativity or storage coefficient – the volume of water released from or taken into storage per unit surface area of aquifer per unit change in hydraulic head. The storage coefficient is the product of specific storage and aquifer thickness for a confined aquifer, and it essentially equals specific yield for an unconfined aquifer.

Transmissivity – the rate at which a volume of water is transmitted through a unit width of an aquifer under a unit hydraulic gradient. It equals the hydraulic conductivity multiplied by the aquifer thickness.

Unconfined aquifer – an aquifer whose upper surface is a water table free to fluctuate under atmospheric pressure.

Valley-fill aquifer – proglacial valley that has been partly filled with glacial deposits that typically consist of 10 to 150 feet of saturated sand and gravel and may be underlain, bordered, or in part overlain by fine-grained deposits and is always bordered on two sides and underlain by relatively impermeable bedrock.

Water table – the water level in the saturated zone at which the pressure head is equal to the atmospheric pressure.

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