Geohydrology of the Madison and Associated Aquifers in Parts of Montana, North Dakota, South Dakota, and Wyoming

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1273-G

Geohydrology of the Madison and Associated Aquifers in Parts of Montana, North Dakota, South Dakota, and Wyoming

By JOE S. DOWNEY

GEOLOGY AND HYDROLOGY OF THE MADISON LIMESTONE AND ASSOCIATED ROCKS IN PARTS OF MONTANA, NEBRASKA, NORTH DAKOTA, SOUTH DAKOTA, AND WYOMING

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1273-G

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON: 1984

UNITED STATES DEPARTMENT OF THE INTERIOR

WILLIAM P. CLARK, *Secretary*

GEOLOGICAL SURVEY

Dallas L. Peck, ***Director***

First printing 1984
Second printing 1984

Library of Congress Cataloging in Publication Data

Downey, Joe S.
Geohydrology of the Madison and associated aquifers in parts of Montana, North Dakota, South Dakota, and Wyoming.

(Geological Survey professional paper ; 1273-G)
Bibliography: p.
Supt. of Docs. no.: I 19.16:1273-G
1. Water, Underground—Great Plains. 2. Aquifers—Great Plains. I. Title. II. Series.
GB1019.D68 1983 551.49'0978 83-600057

For sale by the Distribution Branch, U.S. Geological Survey,
604 South Pickett Street, Alexandria, VA 22304

CONTENTS

ILLUSTRATIONS

[Plate is in pocket]

TABLES

METRIC CONVERSION TABLE

Inch-pound units in this report may be converted to metric (SI) units by using the following conversion factors:

Multily inch-pound units	*By*	*To obtain metric units*
acre-feet (acre-ft)	1,233.0	cubic meters (m^3)
cubic feet (ft^3)	0.02832	cubic meters (m^3)
cubic feet per second (ft^3/s)	0.02832	cubic meters per second (m^3/s)
cubic miles (mi^3)	4.827	cubic kilometers (km^3)
feet (ft)	0.3048	meters (m)
	30.48	centimeters (cm)
feet per second (ft/s)	0.3048	meters per second (m/s)
square feet per day (ft^2/d)	0.0929	square meters per day (m^2/d)
square feet per second (ft^2/s)	0.0929	square meters per second (m^2/s)
inches (in.)	2.540	centimeters (cm)
	25.40	millimeters (mm)
miles (mi)	1.609	kilometers (km)
pounds per square inch (lb/in^2)	0.07037	kilograms per square centimeter (kg/cm^2)
square miles (mi^2)	2.59001	square kilometers (km^2)

DEFINITIONS

Aquifer. A formation, a group of formations, or a part of a formation that contains sufficient saturated permeable material to yield appreciable quantities of water to well or springs.

Artesian well. A well in which the static water level is above the top of. an artesian or confined aquifer; a flowing artesian well is a well in which the water level is above the land surface; (*see* Confined ground water).

Confining bed. A body of relatively impermeable material adjacent to one or more aquifers; in nature, the hydraulic conductivity of a confining bed may range from about zero to some value distinctly less than that of the adjacent aquifer.

Confined gound water. Water in an aquifer that is bounded by confining beds and is under pressure significantly greater than atmospheric.

Gaining stream. A stream whose flow is being increased by the inflow of ground water from springs and seeps along its course.

Hydraulic or static head. The height above a standard datum at which the upper surface of a column of water or other liquid can be supported by static pressure at a given point (the term, which is a measure of potential, is expressed simply as "head").

Homogeneity. Identical properties everywhere in space.

Hydraulic conductivity. A measure of the ease with which a fluid will pass through a porous material, determined by the size and shape of the pore spaces in the material and their degree of interconnection as well as by the viscosity of the fluid; a term replacing "field coefficient of permeability." Hydraulic conductivity may be expressed as feet per day per square foot or cubic meters per day per square meter; hydarulic conductivity is measured at the prevailing water temperature.

Losing stream. A stream that is losing water to the ground along its course.

National Geodetic Vertical Datum of 1929 (NGVD of 1929). A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Mean Sea Level.

Porosity. The ratio of the voluime of the voids in a rock to the total volume, expressed as a decimal fraction or as a percentage. The term "effective porosity" refers to the amount of interconnected pore spaces or voids in a rock or in soil; it is expressed as a percentage of the total volume occupied by the interconnecting pores.

Potentiometric surface. The surface that represents the hydraulic or static head, such as a water table; defined as the level to which water will rise in tightly cased wells.

Specific capacity. The rate of discharge of water from a well divided by the drawdown of the water level, normally expressed as gallons per minute per foot of drawdown.

Storage coefficient. The volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in hydraulic head. In an artesian aquifer, the water derived from storage with decline in hydraulic head comes mainly from compression of the aquifer and, to a lesser extent, from expansion of the water. In an unconfined, or water-table, aquifer the water derived from the aquifer comes from gravity drainage of the voids.

Transmissivity. The rate at which water, at the prevailing temperature, is transmitted through a unit width of an aquifer under a unit hydraulic gradient. Transmissivity is normally expressed as square feet per day or square feet per second; it can be expressed as the number of cubic feet of water that will move during 1 day under a hydraulic gradient of 1 foot per foot through a vertical strip of aquifer 1 foot wide, extending the full saturated height of the aquifer.

GEOLOGY AND HYDROLOGY OF THE MADISON LIMESTONE AND ASSOCIATED ROCKS IN PARTS OF MONTANA, NEBRASKA, NORTH DAKOTA, SOUTH DAKOTA, AND WYOMING

GEOHYDROLOGY OF THE MADISON AND ASSOCIATED AQUIFERS IN PARTS OF MONTANA, NORTH DAKOTA, SOUTH DAKOTA, AND WYOMING

By Joe S. Downey

ABSTRACT

The ground-water system in the Northern Great Plains can be subdivided into at least five major regional aqufiers. Two of these are discussed in this report—the Madison aquifer of Mississippian age and the Cambrian-Ordovician aquifer, which includes the Red River Formation. These aquifers are recharged in high aras of Montana and Wyoming and in the Black Hills of South Dakota. Regional flow is generally to the east and northeast, with most of the discharge from the system occurring in eastern North Dakota and in the Canadian Province of Manitoba. High-density brine is present in both aquifers on the eastern flank of the Williston basin, and freshwater flow appears to divert to the north and south around the brine area or to leak upward into aquifers overlying the Madison. Fracture systems traverse the entire geologic section and exert a major influence on both vertical and horizontal flow.

Prior to Pleistocene glaciation, the flow system probably was similar to that which exists today. During the Pleistocene Epoch, ice sheets covered the northern and eastern parts of the study area, causing extensive changes in recharge and discharge patterns. These changes probably did not cause major shifting in the position of the high-density brines. The flow system was analyzed with the help of a three-dimensional mathematical simulation.

INTRODUCTION

The Madison Limestone is undivided in parts of the project area and is referred to as the Madison Group where subdivisions are possible. For consistency, the term Madison Limestone is mostly used throughout this report. For the past several years, the Madison Limestone and associated rocks, which constitute the Madison aquifer system, have been discussed as a possible source of water for the development of energy reserves in the Fort Union coal region and the Powder River basin of the Northern Great Plains (figs. 1 and 2). Development of these reser es will increase the demand (estimated to exceed 200,000 acre-ft per year) for the use of the region's available water resources. Surface-water sources are limited in time and space and are already fully appropriated throughout much of the region. Aquifers composed of rocks of Paleozoic age (pl. 1), which underlie most of the Northern Great Plains of northeastern Wyoming and eastern Montana, North Dakota, and South Dakota, could supply a significant percentage of the water required for the development. Two such sources of water are the Cambrian-Ordovician and the Madison aquifer systems.

In 1975, the U.S. Geological Survey was provided funds for a 5-year study of the water-supply potential of the Madison aquifer (U.S. Geological Survey 1975). A preliminary computer-based simulation of the Madison aquifer was developed to analyze regional ground-water flow in the aquifer (Konikow, 1976). The modeled area was about 63,000 mi² and encompassed approximately the area of the Powder River basin in Wyoming (fig. 2). The model was relatively simple because the amount of information available on the properties and boundaries of the aquifer system is limited.

This report is one of several reports resulting from the Madison aquifer study; it concentrates on the geohydrologic aspects of the Cambrian, Ordovician, and Mississippian rocks in the study area. Bedrock units above and below were studied for their effects on the hydrologic system existing in the Cambrian-Ordovician and Madison aquifers. Subsequent reports will discuss the hydrology of overlying aquifers in relation to the hydrology of the Cambrian-Ordovician and Madison units.

GEOGRAPHIC SETTING

This study focused on parts of the Northern Great Plains and Central Lowlands physiographic provinces

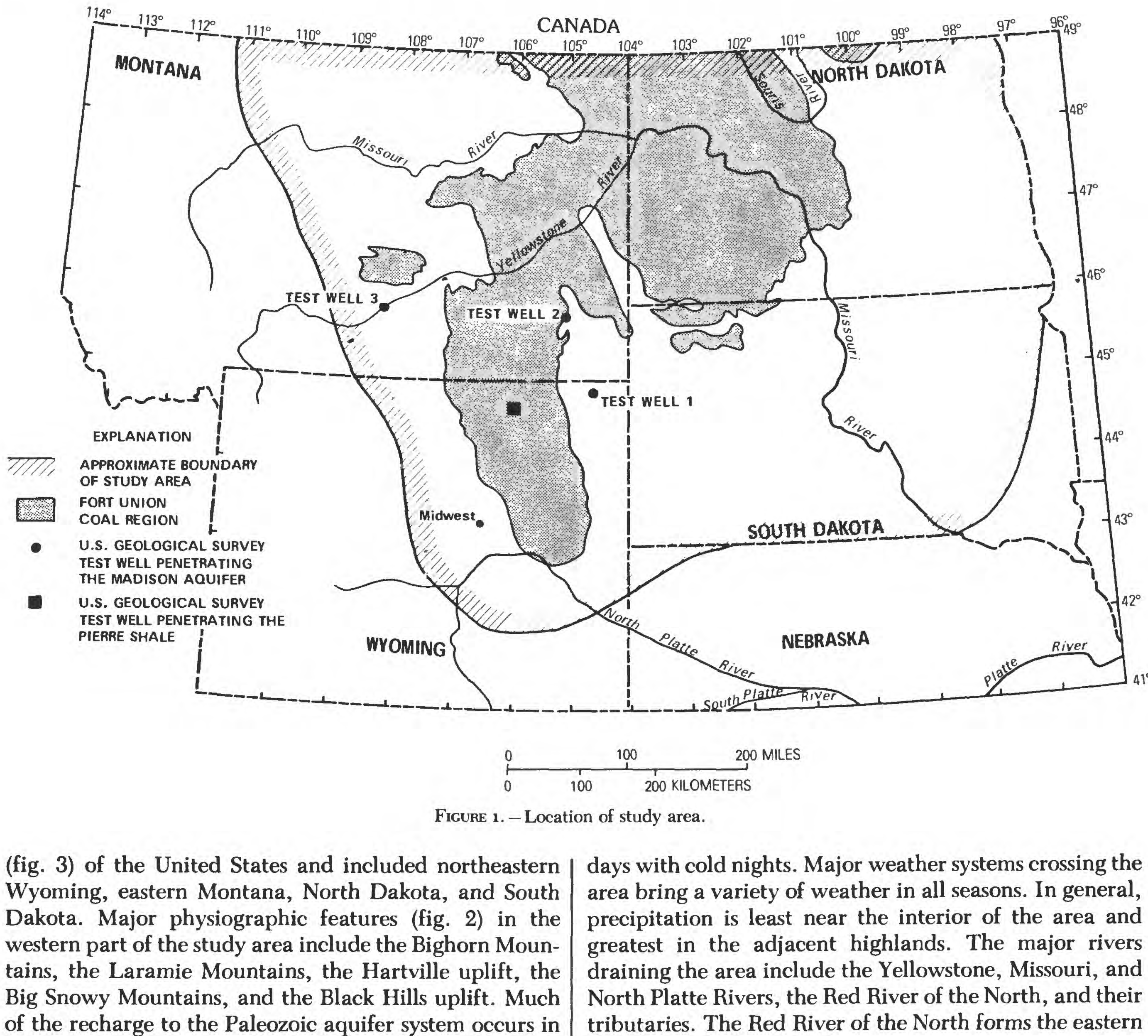

FIGURE 1.—Location of study area.

(fig. 3) of the United States and included northeastern Wyoming, eastern Montana, North Dakota, and South Dakota. Major physiographic features (fig. 2) in the western part of the study area include the Bighorn Mountains, the Laramie Mountains, the Hartville uplift, the Big Snowy Mountains, and the Black Hills uplift. Much of the recharge to the Paleozoic aquifer system occurs in high areas around these features. In the east, the study area is characterized by nearly flat plains containing the broad, flat basin formed by glacial Lake Agassiz in the northeastern part of North Dakota (fig. 2). Glacial Lake Agassiz covered an area of about 200,000 mi^2 from about 12,500 to 7,500 years ago (Flint, 1971) in the Canadian Provinces of Ontario, Manitoba, and Saskatchewan and in Minnesota, North Dakota, and the northeastern part of South Dakota (Elson, 1967). However, during most phases of Lake Agassiz, the surface area probably did not exceed 80,000 mi^2. Maximum depth exceeded 600 ft. Much of the discharge from the Paleozoic aquifer system appears to occur in the area of the Lake Agassiz basin.

The study area has a semiarid continental climate characterized by cold, snowy winters and warm summer days with cold nights. Major weather systems crossing the area bring a variety of weather in all seasons. In general, precipitation is least near the interior of the area and greatest in the adjacent highlands. The major rivers draining the area include the Yellowstone, Missouri, and North Platte Rivers, the Red River of the North, and their tributaries. The Red River of the North forms the eastern boundary of the study area.

PREVIOUS WORK

An abundance of mineral resources, including petroleum, natual gas, uranium, and coal, lies within the project area. Economic interest has encouraged a large number of reports on geologic structure, stratigraphy, mineralization, and economic geology of the area. Some of the important publications on regional geologic structure and subsurface and surface stratigraphy of the project area and surrounding regions include those by Thom (1923); Nordquist (1953); Agatston (1954); Andrichuck (1955); Sonnenberg (1956); Peterson (1957, 1966, 1972, 1978, 1981); Sandberg and Hammond (1958); Sandberg

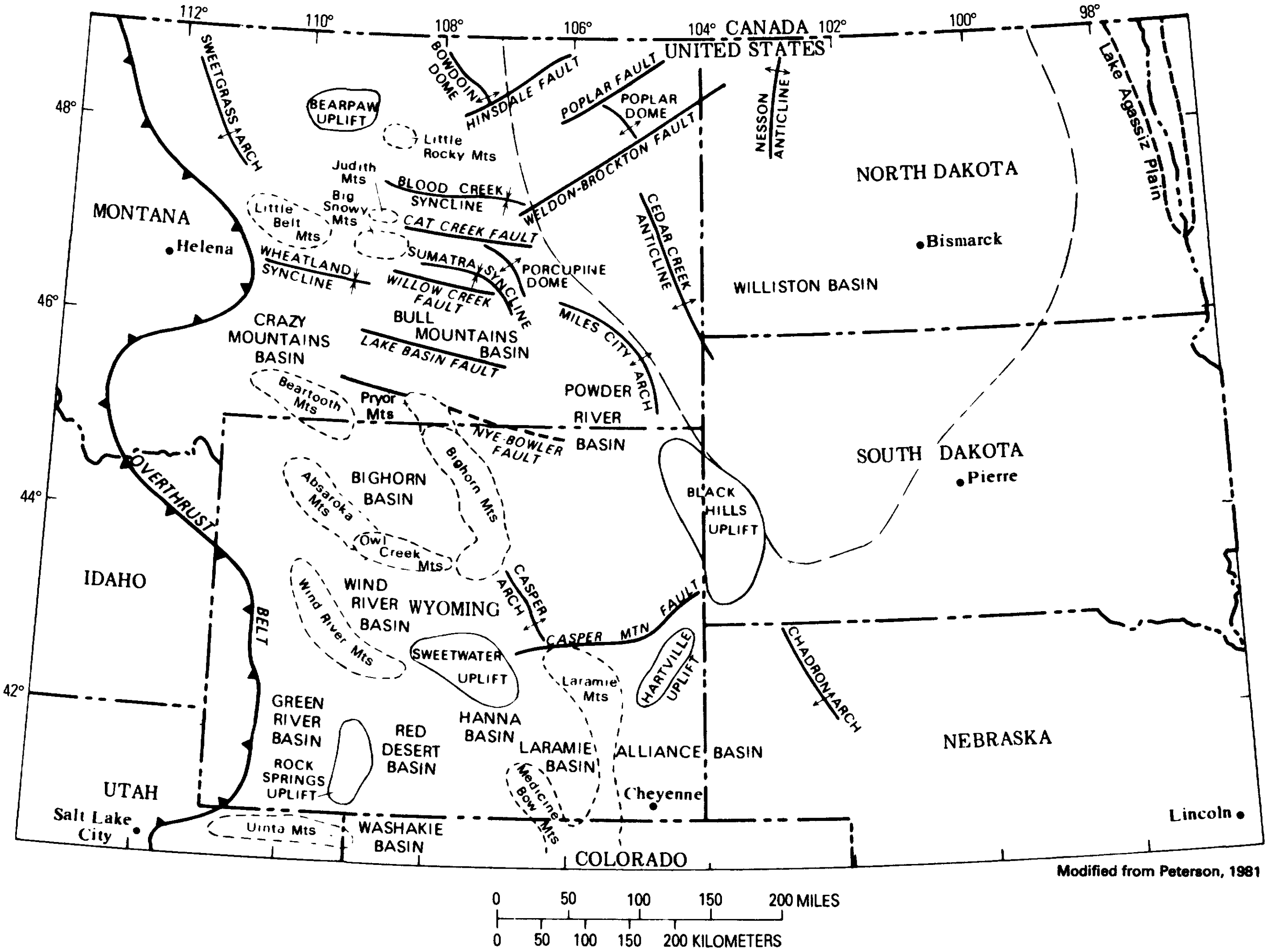

FIGURE 2. —Present-day structural and physiographic features of the Northern Great Plains.

(1961); Irwin (1965), Keefer and Van Lieu (1966); Maughan (1966); Rohl (1967); Sales (1968); Stone (1969, 1970, 1971); Keefer (1970); Thomas (1971, 1974); Sando and others (1975); Rose (1976); Sando and others (1976); Sando (1976a, 1976b); Sandberg and Poole (1977); Brown (1978); Warner (1978); and Head and others (1978).

The number of regional and subregional studies of the hydrology and geochemistry of the project area and surrounding region is limited. Some of the more important are those reported by Darton (1896); Swenson (1968); Downey (1973); Wyoming State Engineer's Office (1974); Konikow (1976); Miller (1976); Swenson and others (1976); Hopkins (1978); and Miller and Strausz (1980a, 1980b).

Historically, the geologic environment of the project area was of a stable shelf margin throughout most of Paleozoic time. Most authors agree that there were periods of transgression and regression, when seas advanced from west to east in response to mountain-building activity of the Antler orogeny to the west (Sandberg and Poole, 1977).

The style of structural deformation has been explained in terms of horizontal compression (Sales, 1968; Stone, 1969), vertical tectonics (Stearns and others, 1975), and wrench-fault tectonics (Brown, 1978). Although the mechanics of deformation have little effect on the existing hydrologic system, there is general agreement that the present-day structure has been controlled by the preexisting structural grain in the Precambrian basement and modified by the Laramide orogeny. Several authors, including Sandberg (1962); Maughan (1966); Miller (1966); Sando and others (1976); Rose (1976); and Slack (1981), have suggested that paleostructure in various parts of the project area has had a pronounced effect on the thickness and environment of deposition of sediments. Knowledge of sediment thickness and structural history is important in the estimation of various hydrologic properties of the aquifer units at locations where there are no test holes.

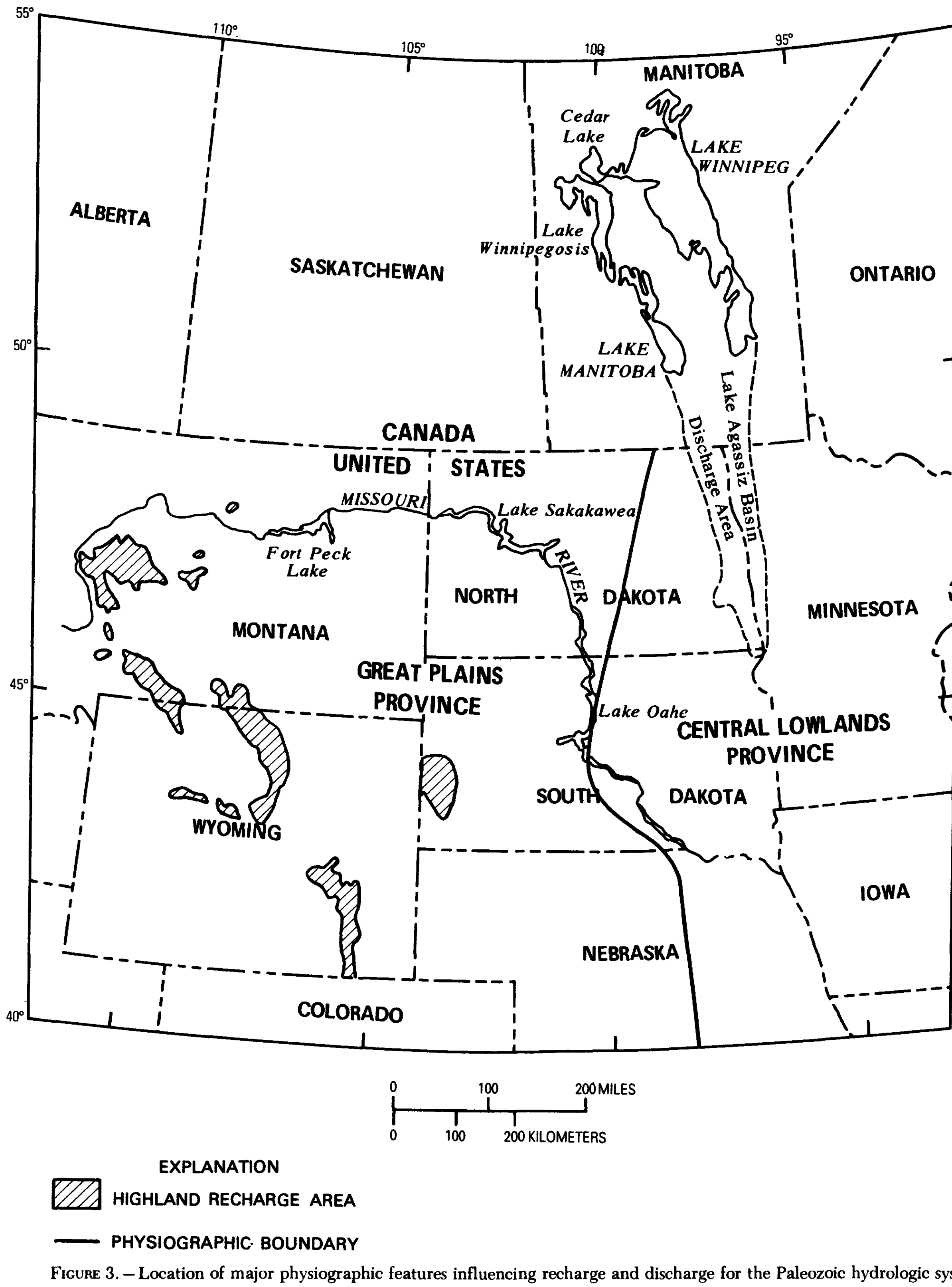

FIGURE 3.—Location of major physiographic features influencing recharge and discharge for the Paleozoic hydrologic system in the Northern Great Plains.

GEOLOGIC SETTING

During most of Paleozoic time, the study area was part of the Cordilleran platform. This broad flat area was bordered on the west by the Cordilleran miogeosyncline, oriented approximately north and south in Idaho and western Montana. Most of the detrital sediments in the synclinal trough came from the Antler orogenic belt, which probably was an island arc system to the west that had intermittent tectonism during parts of Paleozoic time. The Transcontinental arch, southeast of the study area, was low-lying and sporadically contributed minor amounts of sediment that were spread thinly by ocean currents across the platform. In general, the Cordilleran platform was a shallow-water depositional shelf on which predominantly carbonate and evaporite sediments were deposited during most of Paleozoic time.

The Black Hills uplift (fig. 2) was not a regionally significant tectonic element until Late Cretaceous time (Agnew and Tychsen, 1965) and had little effect on sedimentation during most of Paleozoic time. During the Mississippian age, the study area generally was covered by a shallow warm sea probably less than 10 ft deep (Sando, 1976b); however, there are small areas characterized by deep-water sediments. The warm sea was conducive to prolific biologic productivity, accumulation, and preservation. Shoals and reefs were common. These shallow-water areas continually changed and shifted because of the effects of geologic forces in time and space. Associated with many of the shallow areas were small reefs as well as many oolite- and crinoid-bank shoals and lagoons. Lagoons were evaporating basins in which gypsum and halite precipitated and became incorporated into the lime-rich bottom sediments. Gypsum precipitated in lagoons that received influxes of sea water, but evaporation seldom proceeded to the point of halite precipitation. Evidence of large areas with more restricted circulation or greater evaporation rates is provided by the accumulations of bedded evaporites in the Williston basin and the Central Montana trough.

The present-day structure (fig. 2) and paleogeography of the Northern Great Plains is directly related to the geologic history of the Cordilleran platform, which is a part of the stable interior of the North American continent. During geologic time, many structural features developed that affected the deposition of the various sedimentary units. Most of these structural features exist today and are important in determining the present hydrologic regime in both the Cambrian-Ordovician and Madison aquifer systems of the Northern Great Plains.

PRECAMBRIAN ROCKS

Crystalline rocks of Precambrian age form the basement in the Northern Great Plains region. Depth to the Precambrian basement from land surface ranges from outcrops in the eastern part of the study area to greater than 17,000 ft at the center of the Williston basin. Precambrian rocks also are found in the central core of the many mountain ranges in the western part of the study area.

On a regional basis, little is known about the water-bearing properties of the Precambrian rocks. Available data indicate that they contain only small amounts of water in joints and fractures. Except along major fracture systems, which appear to extend to great depths (Kanasewich, 1968), these rocks are not considered to be water-bearing. Along major fractures, Precambrian rocks may yield water coming from leakage from the overlying sedimentary sequence. Precambrian rocks in the study area, where Precambrian rocks crop out, they were taken as the eastern boundary of the hydrologic system and were represented as such in the digital-simulation model.

CAMBRIAN AND LOWER ORDOVICIAN ROCKS

Rocks of Cambrian and Early Ordovician age (fig. 4) in the Northern Great Plains consist of marine sandstone, shale, and limestone that represent the shoreward facies of a transgressive sea that occupied the area during Cambrian and Early Ordovican time (Peterson, 1981).

Several formations of Middle Cambrian to Early Ordovician age (fig. 5), including the Flathead Sandstone, Deadwood Formation, and Emerson Formation, are aquifers; however, their great depth has prevented their use as a source of water, and very little hydrologic data concerning these aquifers are available on a regional scale. Most of the data that exist are from tests performed in connection with the development of oil and gas wells.

In the subsurface, both the Deadwood and Emerson Formations consist of about 600 ft of fine- to coarse-grained quartz sandstone with dolomitic cement interbedded with dolomitic limestone and shale (Steece, 1978). Where exposed in the Black Hills of South Dakota, the Deadwood Formation is of the same lithology; however, only about 450 ft of section is present. In the Black Hills, the Deadwood lies unconformably on Precambrian schist and quartzite.

As shown in figure 5, the Deadwood Formation is absent in extreme eastern North Dakota except for a small

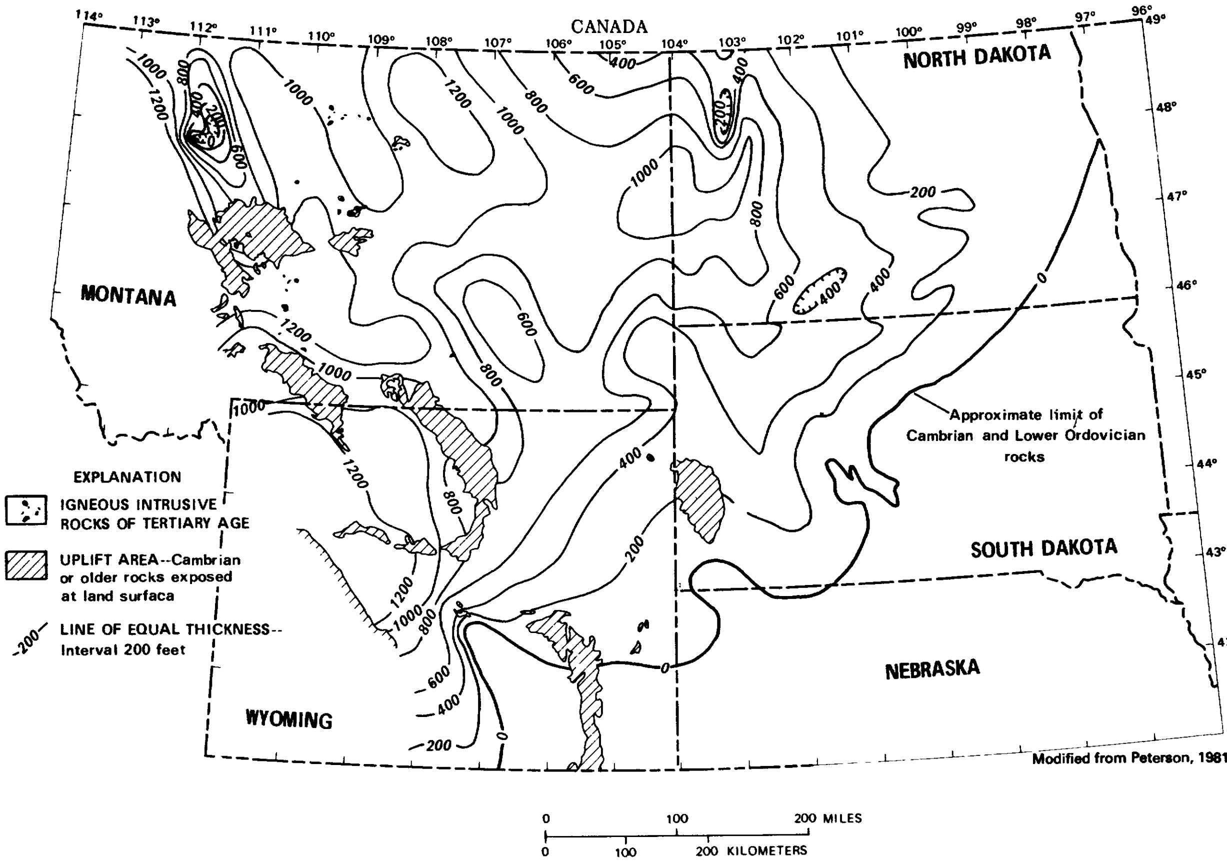

FIGURE 4. — Thickness and extent of Cambrian and Lower Ordovician rocks.

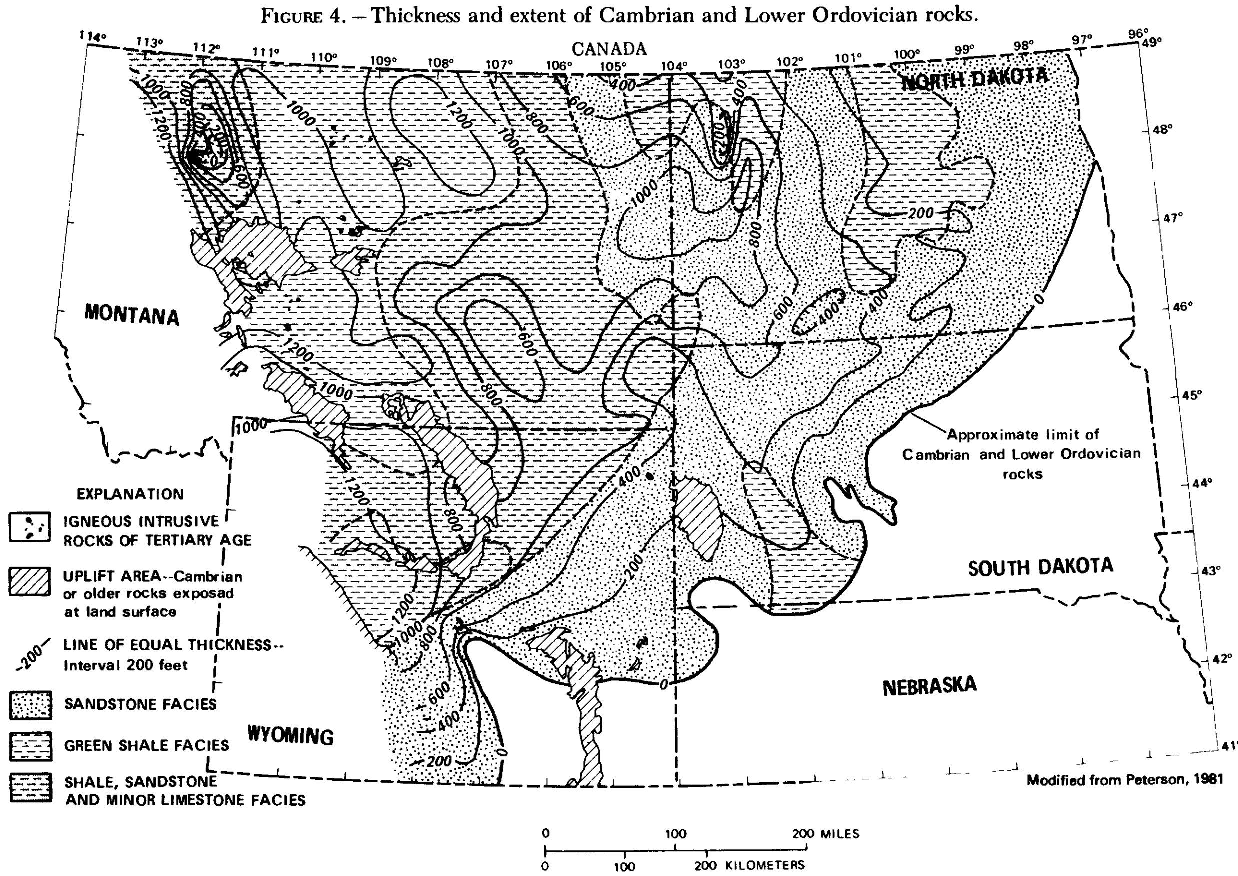

FIGURE 5. — Thickness and lithology of the Cambrian and Lower Ordovician rocks.

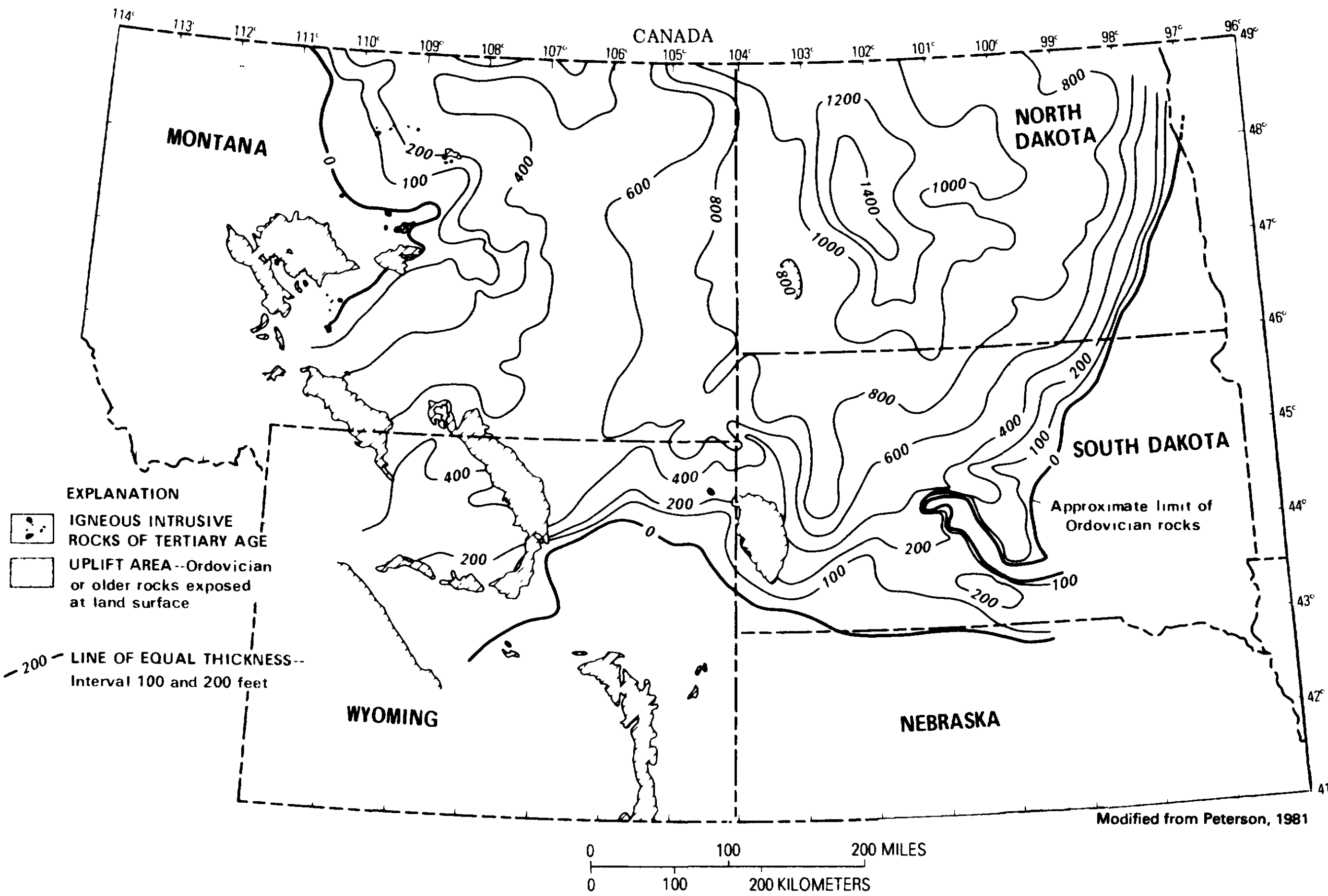

FIGURE 6.—Thickness and extent of Ordovician rocks.

area in the eastern discharge area, where a thickness of about 100 ft of quartzose sandstone is present (Ballard, 1963). Both the Deadwood and Emerson Formations are overlain by the Winnipeg Formation of Middle Ordovician age. In areas where the Deadwood or Emerson Formations are in hydrologic connection with the overlying Ordovician units, the formations along with the Flathead Sandstone have been included as a part of the Cambrian-Ordovician aquifer system.

ORDOVICIAN ROCKS

Rocks of Ordovician age (fig. 6) in the Northern Great Plains have been divided into several formations: the upper part of the Deadwood, the Winnipeg, the Red River, and Stony Mountain Formations. The Winnipeg, Red River, and Stony Mountain Formations in Montana are equivalent to the Bighorn Dolomite in the Powder River basin. In south-central Montana, the Winnipeg is absent, and only the Red River and Stony Mountain are equivalent to the Bighorn. Subsurface studies of the Ordovician rocks in the Williston basin and adjoining areas include those by Fuller (1956), Carlson and others (1960), Patterson (1961), Porter and Fuller (1964), Carlson and Anderson (1965), Ballard (1969), Foster (1972), Armstrong (1980), and Peterson (1981).

Ordovician rocks are major petroleum reservoirs in the Williston basin, and many exploratory wells that penetrate these rocks have been drilled. Ordovician rocks are not present in western Montana, southeastern Wyoming, southwestern and eastern South Dakota, and southeastern North Dakota, because of nondeposition or removal by erosion during Devonian and Early Mississippian time. Thickness increases eastward and northward from the zero line in central Montana and northeastern Wyoming to more than 1,400 ft in the central part of the Williston basin.

The Winnipeg Formation is stratigraphically equivalent to the St. Peter Sandstone that occurs in the Midwestern United States. In the western part of the study area, the formation consists of a clean, well-sorted, medium-grained porous sandstone where it is not deeply buried (Peterson, 1978). Where it is deeply buried, the

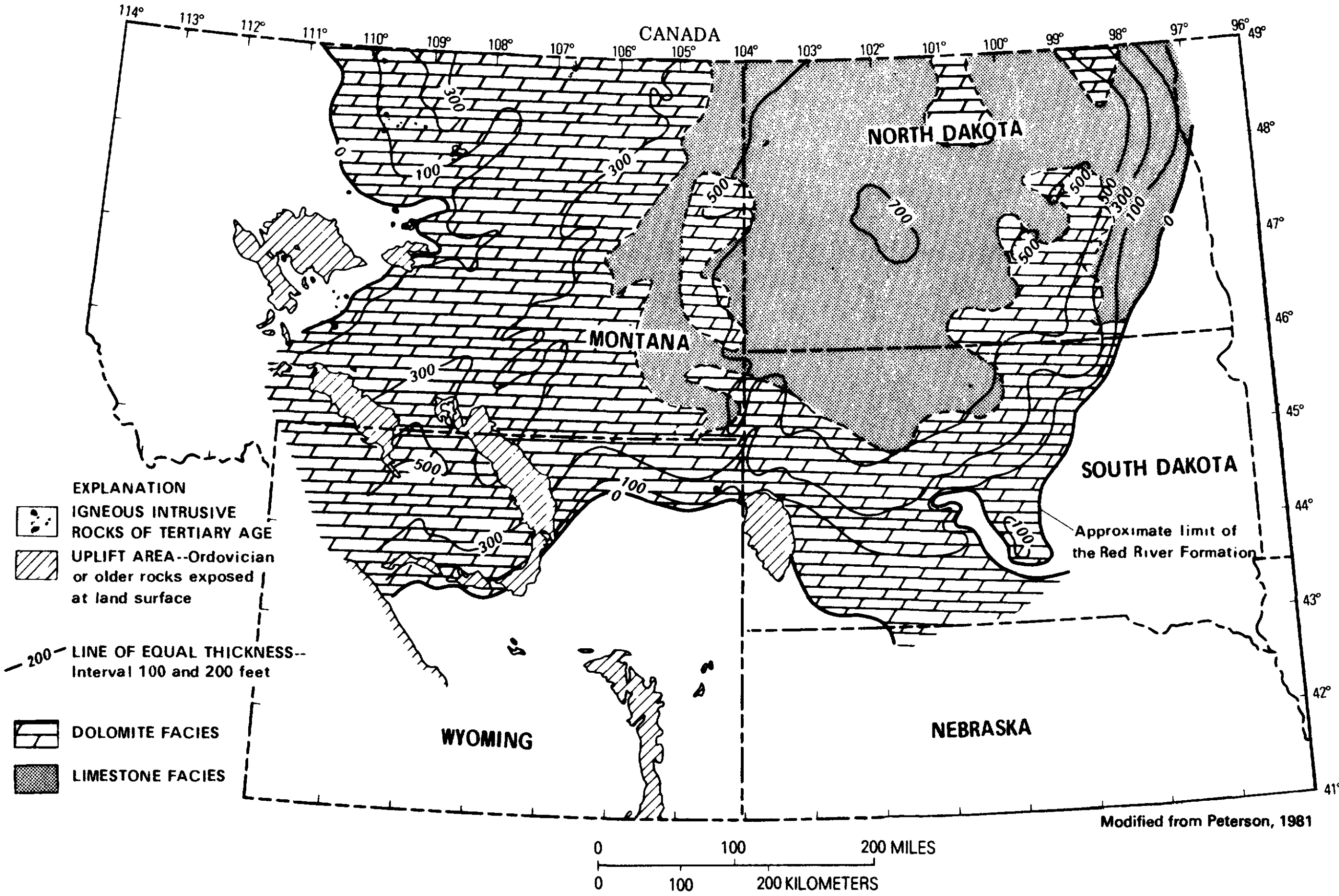

FIGURE 7. — Thickness and lithology of the Red River Formation of Late Ordovician age.

unit has minimal porosity and permeability because of silica cementation and related compaction. In the eastern discharge area of North Dakota, the Winnipeg Formation consists of a sequence of shale, sandstone, and shaly limestone ranging in thickness from 20 to about 140 ft (Armstrong, 1980). The sandstone units consist of a very fine to fine, rounded quartz grains with interbedded siltstone and shale. Armstrong (1980) reported that in Dickey and Lamoure Counties of North Dakota, near the eastern discharge area of the Cambrian-Ordovician aquifer, sands of the Winnipeg Formation (which he included in his Black Island aquifer) have a hydraulic conductivity ranging between $9.2+10^{-4}$ and $1.1+10^{-3}$ ft/s. Armstrong estimated transmissivity of the sand to be about 0.02 ft^2/s.

The Red River Formation (fig. 7), a carbonate sequence that overlies the Winnipeg Formation, extends outward past the borders of the Williston basin. The Red River Formation is more than 700 ft thick in the central part of the Williston basin and was truncated by Devonian erosion in the western part of the study area along a line extending between the central Black Hills and the southern Bighorn Mountains (Peterson, 1981). According to Porter and Fuller (1964), the Red River in the center of the Williston basin consists of a lower unit of fragmental limestone and dolomite and an upper unit with three evaporite units of fragmental limestone and anhydrite.

The Stony Mountain Formation, which conformably overlies the Red River Formation, includes a lower shaly limestone and shale interval locally known as the Stoughton Member in North Dakota. The Stoughton, which is, in general, confined to the interior of the Williston basin, is truncated on the east and south and grades into the upper part of the Bighorn Dolomite to the west (Peterson, 1981). Rocks of Late Ordovician and Early Silurian age overlying the Stony Mountain Formation in the study area are called the Stonewall Formation in Canada. The Stony Mountain Formation is composed of carbonate, shaly carbonate, and anhydrite beds and is gradational upward into the Interlake Formation (fig. 8) of Late Ordovician and Silurian age.

Both the Red River and Stony Mountain Formations were truncated by Devonian erosion around the periphery of the Williston basin. The Stony Mountain

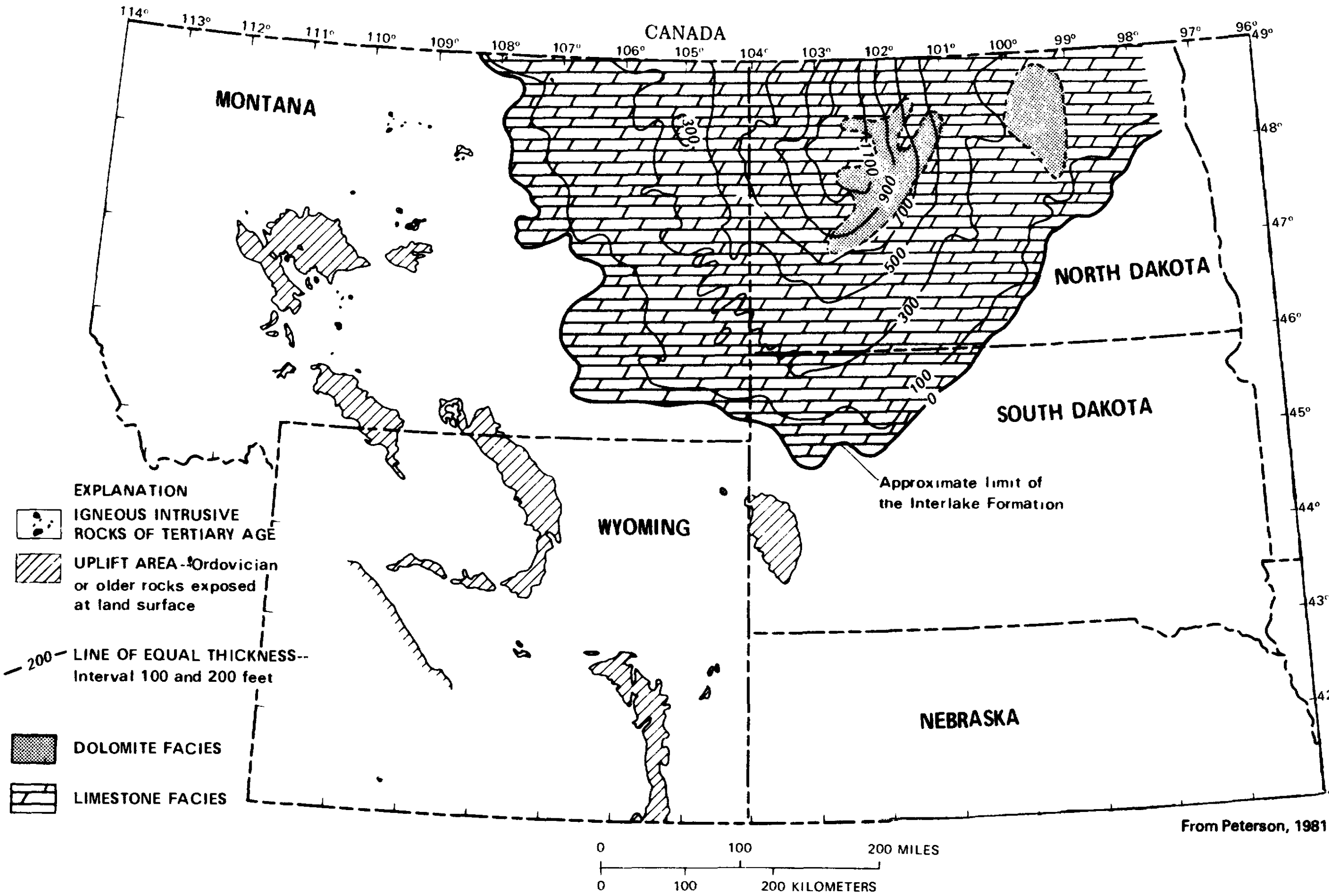

FIGURE 8.—Thickness and lithology of the Interlake Formation of Late Ordovician and Silurian age.

erosional edge is closer to the basin center than is the erosion edge of the underlying Red River Formation.

SILURIAN AND DEVONIAN ROCKS

Silurian and Devonian rocks (figs. 8 and 9) overlie the lower part of the Interlake, Red River, and Stony Mountain Formations of Late Ordovician age in most of the study area. Silurian and Devonian units consist mainly of shaly carbonates, shale, and evaporites, including Devonian halite (fig. 10), near the center of the Williston basin where the Silurian and Devonian section has a total thickness greater than 2,000 ft. The halite units in Devonian rocks extend northward into Canada about 1,200 mi, underlying the Provinces of Alberta, Saskatchewan, and Manitoba and extending into the Northwest Territories. The salt in the Prairie Formation (so-called Prairie Salt), one of the principal halite units of Devonian age in the study area, contains many structural lows occurring along the margin of the halite and locally within the halite area. These structural lows have been attributed to postdepositional solution of halite which allowed collapse of the overlying formations into the void created by salt removal (DeMille and others, 1964). Because of the fine-grained lithology and the presence of evaporites in the Silurian and Devonian units, these formations are considered to be confining beds for the underlying Cambrian-Ordovician aquifer.

MISSISSIPPIAN ROCKS

Rocks of Mississippian age (pl. 1) overlie the Devonian formations throughout the study area. The Mississippian rocks have been divided into several formations and one stratigraphic group.

The lowermost Mississippian rock unit is the Bakken Formation or the equivalent Englewood Formation (pl. 1), which overlies the Upper Devonian Three Forks Formation in parts of the area. The Bakken Formation consists of more than 100 ft of black, organic shale and siltstone and appears to be an excellent hydrologic confining bed. It is considered to be a source bed for much of the petroleum found in overlying formations. Overlying the Bakken Formation is a sequence of Mississippian rocks, mainly limestone and dolomite, that are termed the Madison Group where divided or the Madison

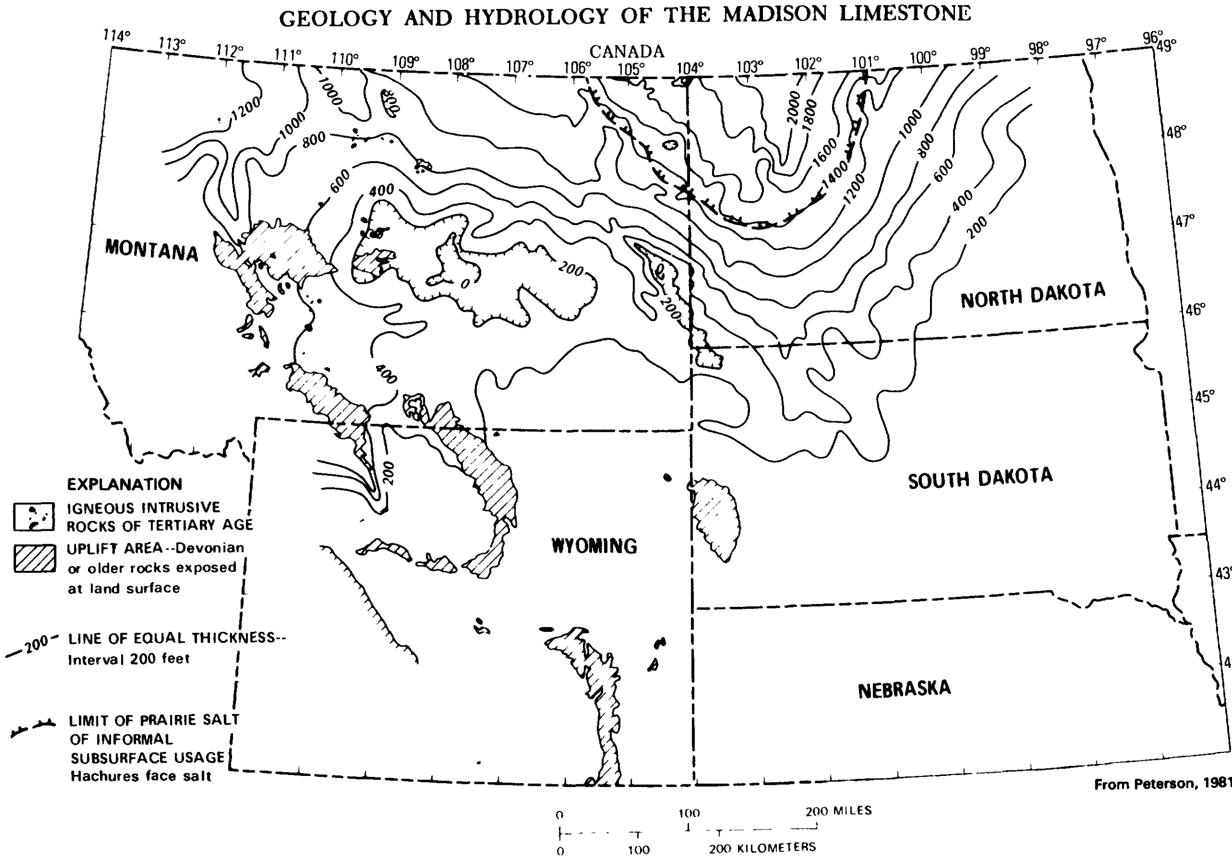

FIGURE 9. – Thickness of Devonian rocks.

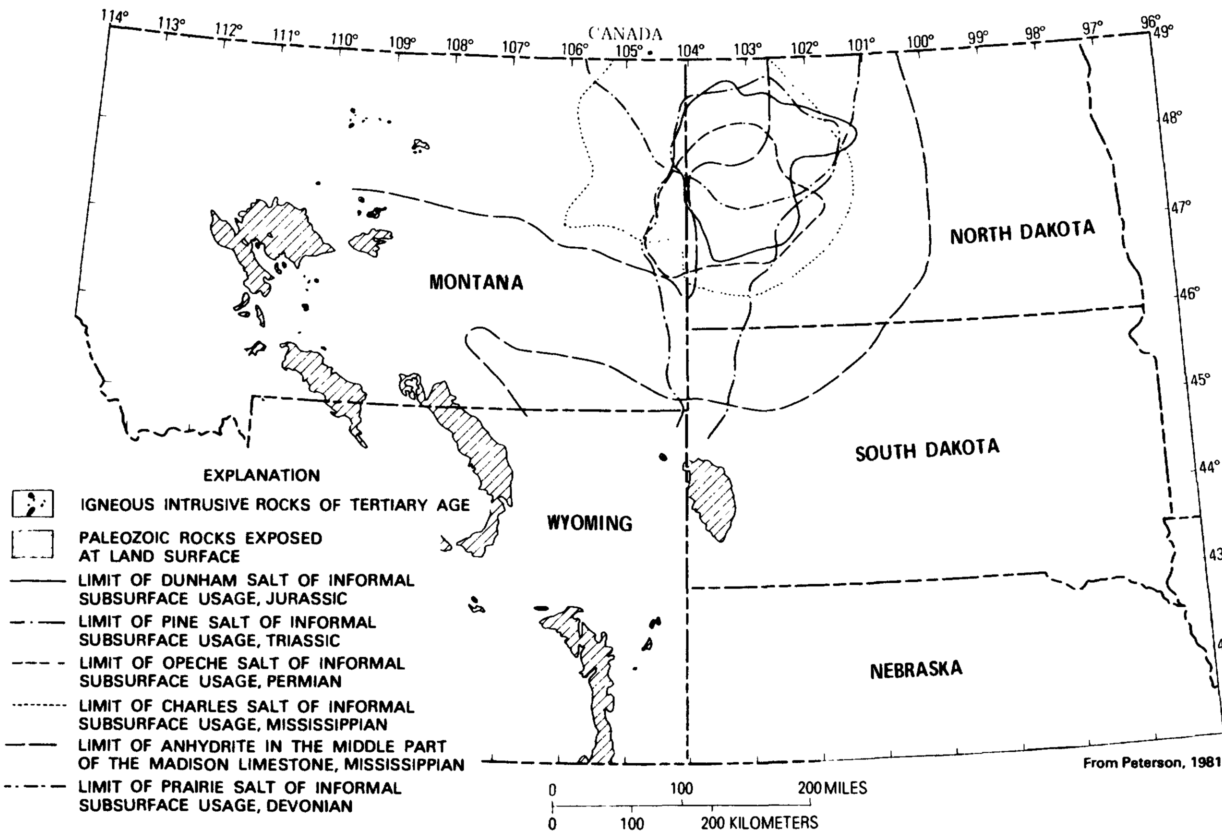

FIGURE 10. – Extent of halite and evaporite units.

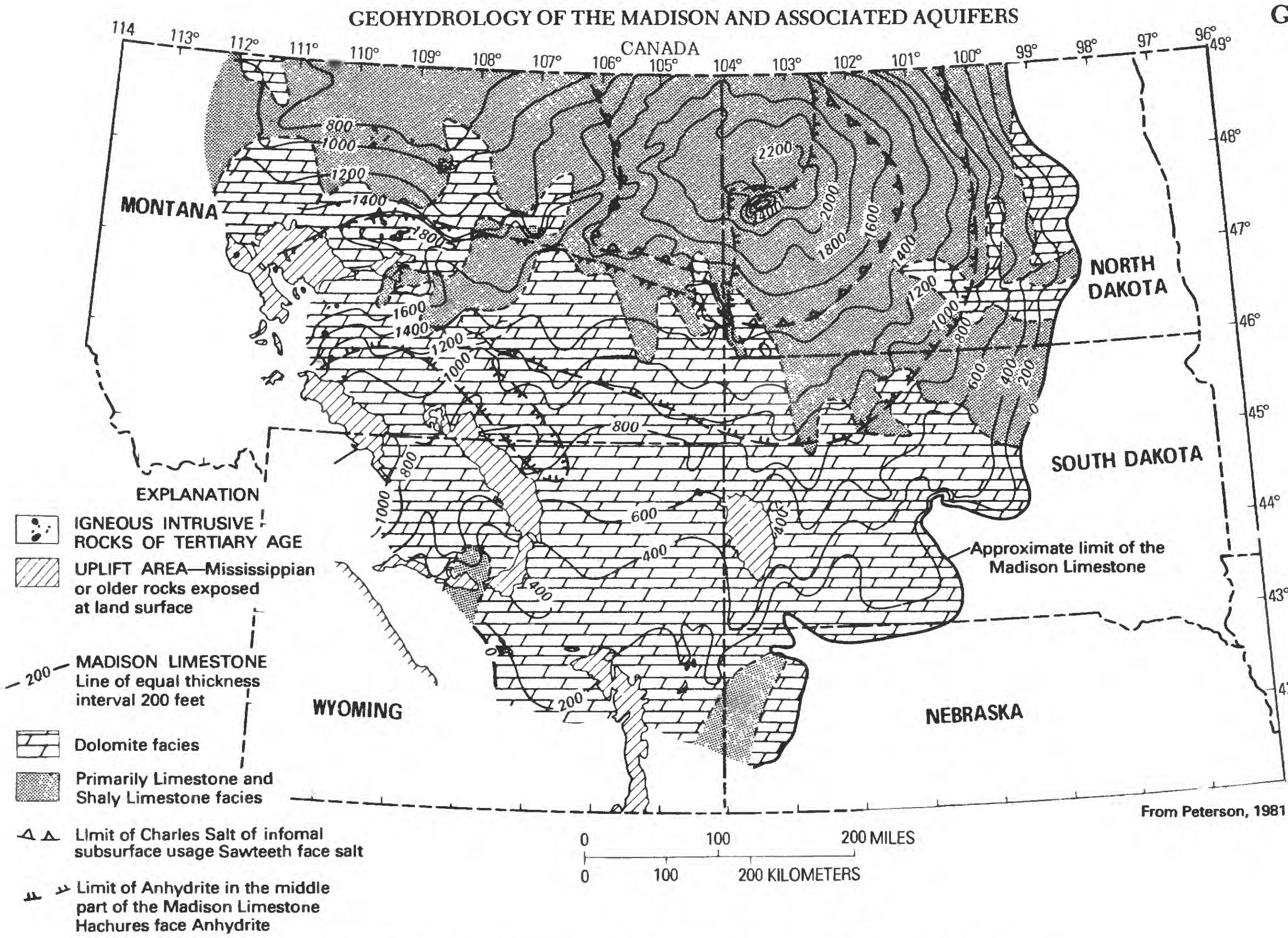

FIGURE 11. — Thickness and lithology of the Madison Limestone of Mississippian age.

Limestone where undivided.

The Madison Limestone in the study area (fig. 11) consists of a sequence of marine carbonates and evaporites deposited mainly in a warm, shallow-water environment. Depositional environments grade both laterally and vertically from shallow-marine carbonate and evaporite facies to deep-water facies. The Madison Group, from oldest to youngest, consists of the Lodgepole Limestone, the Mission Canyon Limestone, and the Charles Formation, or their stratigraphic equivalents in other parts of the study area.

The Lodgepole Limestone is predominantly a cyclic carbonate sequence consisting largely of fossiliferous to micritic dolomite and limestone units that are argillaceous and thin-bedded in most of the study area (Smith, 1972). The unit ranges from 0 to more than 900 ft in thickness and averages abut 300 ft in the study area. The Lodgepole overlies the Bakken Formation in the Williston basin, the Central Montana trough, and north-central Montana.

The Mission Canyon Limestone consists of a coarsely crystalline limestone at its base, grading upward to finer crystalline limestone and evaporite minerals near the top (D. L. Brown, U.S. Geological Survey, written commun., 1981). The formation contains one evaporite cycle and shares a second evaporite cycle with the lower part of the Charles Formation. Bedded evaporite units are absent in most of Wyoming and South Dakota, but evaporite deposits are present in southeastern Montana and northwestern North Dakota and gradualy thicken from central Montana toward their maximum thicknesses in the Williston basin. The Mission Canyon Limestone ranges from 0 to more than 650 ft in thickness and averages about 300 ft in the study area.

The Charles Formation, the uppermost unit of the Madison Group, is a marine evaporite sequence consisting of anhydrite and halite with interbedded dolomite, limestone, and argillaceous units. The Charles Formation ranges from 0 to more than 300 ft in thickness, with an average thickness of about 250 ft in the study area. Pre-Jurassic erosion has removed most of the Charles Formation in the western and southern parts of the study area. Where present, the Charles Formation is a hydrologic confining bed for the underlying aquifers.

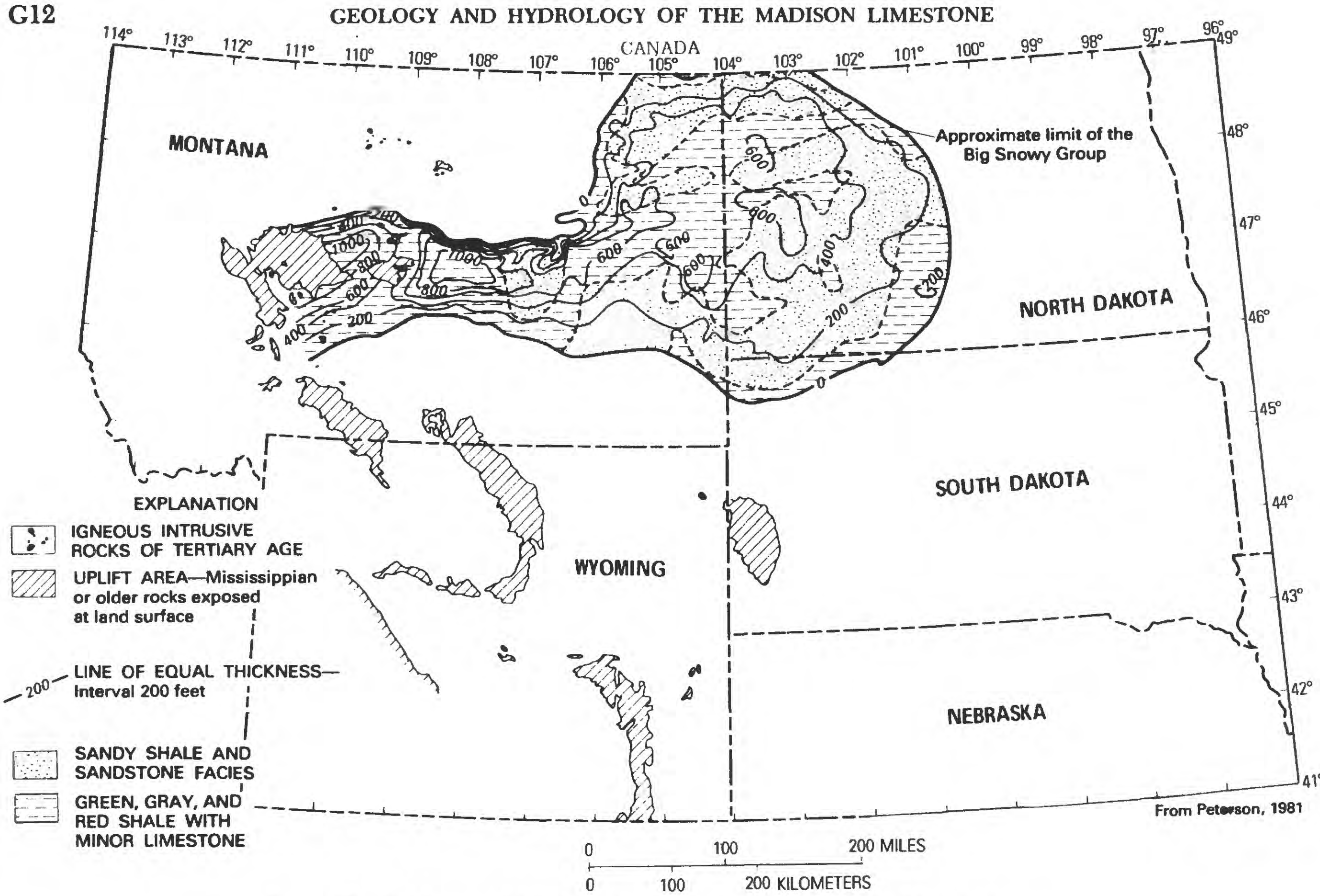

FIGURE 12. —Thickness and lithology of the Big Snowy Group of Late Mississippian age.

Because the carbonate rocks of the Madison Limestone are relatively soluble in water, the development of karst (solution) features is common. Sando (1974) describes ancient karst features, including enlarged joints, sink holes, caves, and solution breccias, that developed in the Madison Limestone in north-central Wyoming. He indicates that most of the open spaces were filled by sand and residual products reworked by a transgressive sea during Late Mississippian time. Large and extensive cave systems in outcrop areas of the Madison Limestone in the Bighorn Mountains and in the Black Hills are further evidence of the importance of the dissolution process in the development of secondary permeability in some of the units of the Madison Limestone.

Overlying the Charles Formation in parts of Montana, North Dakota, and South Dakota are rocks of Late Mississippian age belonging to the Big Snowy Group (fig. 12). The Big Snowy Group consists mainly of shale and sandstone, with minor limestone. In the hydraulic analyses made during this study, the Big Snowy Group has been combined with the Charles Formation as a confining layer.

The Madison Limestone, or its equivalents, is the major source of petroleum in the Northern Great Plains area. It is estimated that more than 90 percent of the petroleum production in North Dakota has come from reservoirs in the Madison Limestone (Carlson and Anderson, 1960). More than 40 percent of these reservoirs are of the statigraphic-trap type (Landes, 1970), underlying an unconformity on the updip flank of the Williston basin.

Many of the reservoirs in the Madison Limestone consist of porous zones in fine- to medium-grained limestone or dolomite. Studies of pressure relationships between these porous zones (Landes, 1970) indicate that in some instances they are interconnected by fractures that cut the less permeable rocks between the zones. Fracture interconnection between zones of greater permeability appears to be the major route of water flow in the Cambrian-Ordovician and Madison aquifers.

PENNSYLVANIAN AND PERMIAN ROCKS

Rocks of Pennsylvanian age (fig. 13) overlie the Mississippian units in most of the study area and consist of marine sandstone, shales, siltstone, and carbonates.

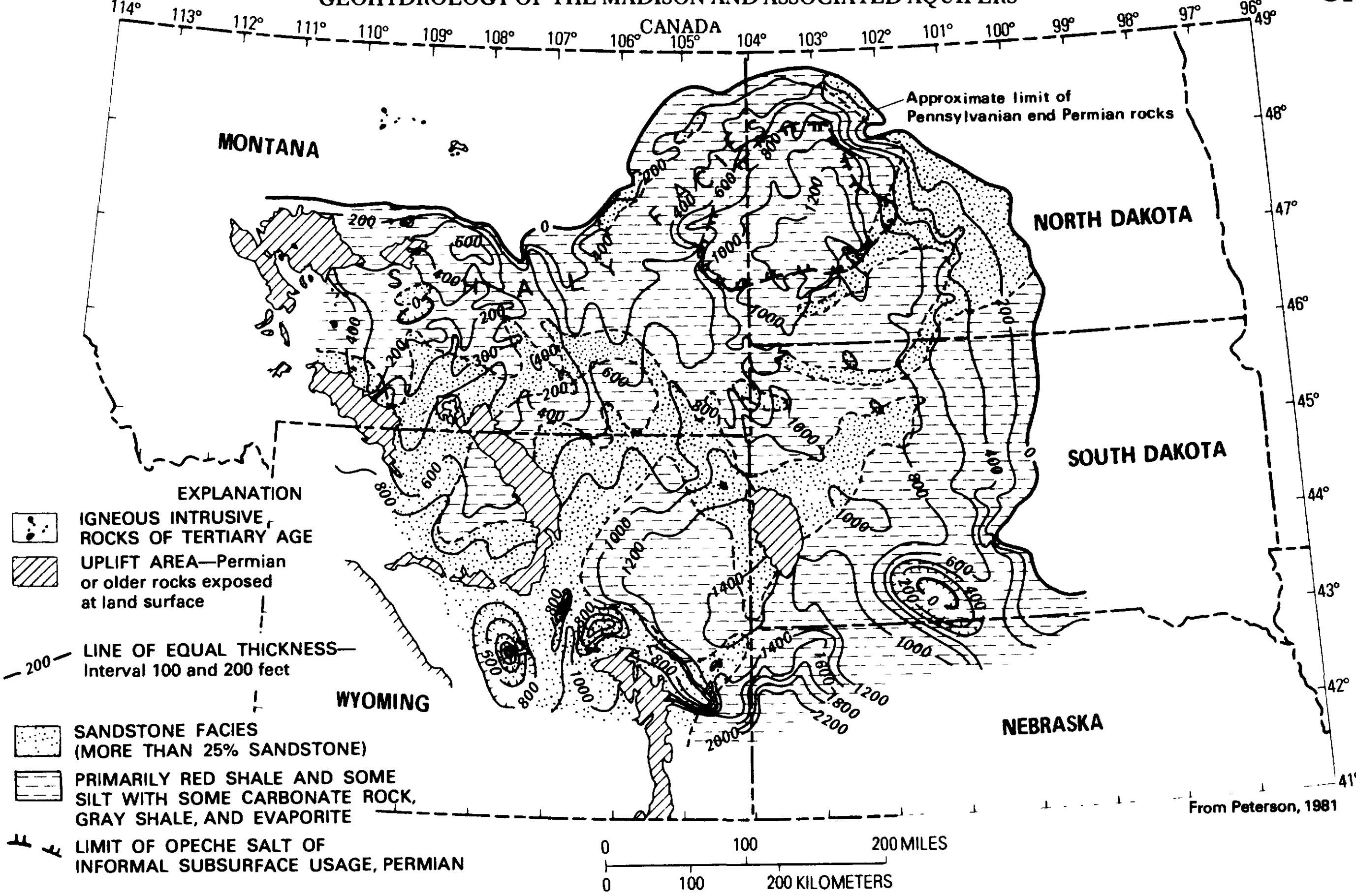

Figure 13. —Thickness and lithology of Pennsylvanian and Permian rocks.

The Pennsylvanian rocks are divided into many different formations; however, most are equivalent units.

The Tyler Formation (pl. 1) is restricted generally to the Central Montana trough and the Williston basin of Montana, but the formation extends southward and appears to be equivalent to the lower part of the Amsden Formation in south-central Montana. The Tyler also appears to be equivalent to the lower part of the Minnelusa Formation in northwestern South Dakota, Wyoming, and western North Dakota.

Middle Pennsylvanian rocks are represented by the Tensleep Sandstone in central and southern Montana and north-central Wyoming and by part of the Minnelusa Formation in the Williston basin, northeastern Wyoming, and western South Dakota. These rocks have been truncated by pre-Jurassic erosion progressively northward across central Montana, reaching a line of zero thickness (fig. 13) approximately near the axis of the Central Montana trough.

Porous sandstone units of Pennsylvanian age are present in the Tensleep Sandstone on central to north-central Wyoming and south-central Montana and in the middle part of the Minnelusa Formation in western South Dakota and along the east side of the Williston basin.

The upper part of the Minnelusa Formation in the Powder River and Williston basins and in the western part of South Dakota consists of sandstone, shale, and carbonate, with interbedded anhydrite of Early Permian age. The sandstone facies extends northward to include the southeastern part of the Williston basin. The source of the sands in the upper part of the Minnelusa is interpreted to be the reworking of earlier deposited Pennsylvanian sands derived from paleostructures to the west (D. L. Brown, U.S. Geological Survey, oral commun., 1980). Additional source areas for clastic sediments were the Siouxia uplift and the Canadian Shield on the eastern and northeastern borders of the Williston basin. Locally, Pennsylvanian rocks in the study area contain small aquifers that are in hydraulic connection with the Madison aquifer, and they provide a path for discharge of water from the Madison aquifer.

Overlying the upper part of the Minnelusa Formation are the Lower Permian Opeche Formation, the Permian Goose Egg Formation, or equivalent formations (pl. 1). The Opeche Formation is interbedded in the central part of the Williston basin with halite beds termed the Opeche

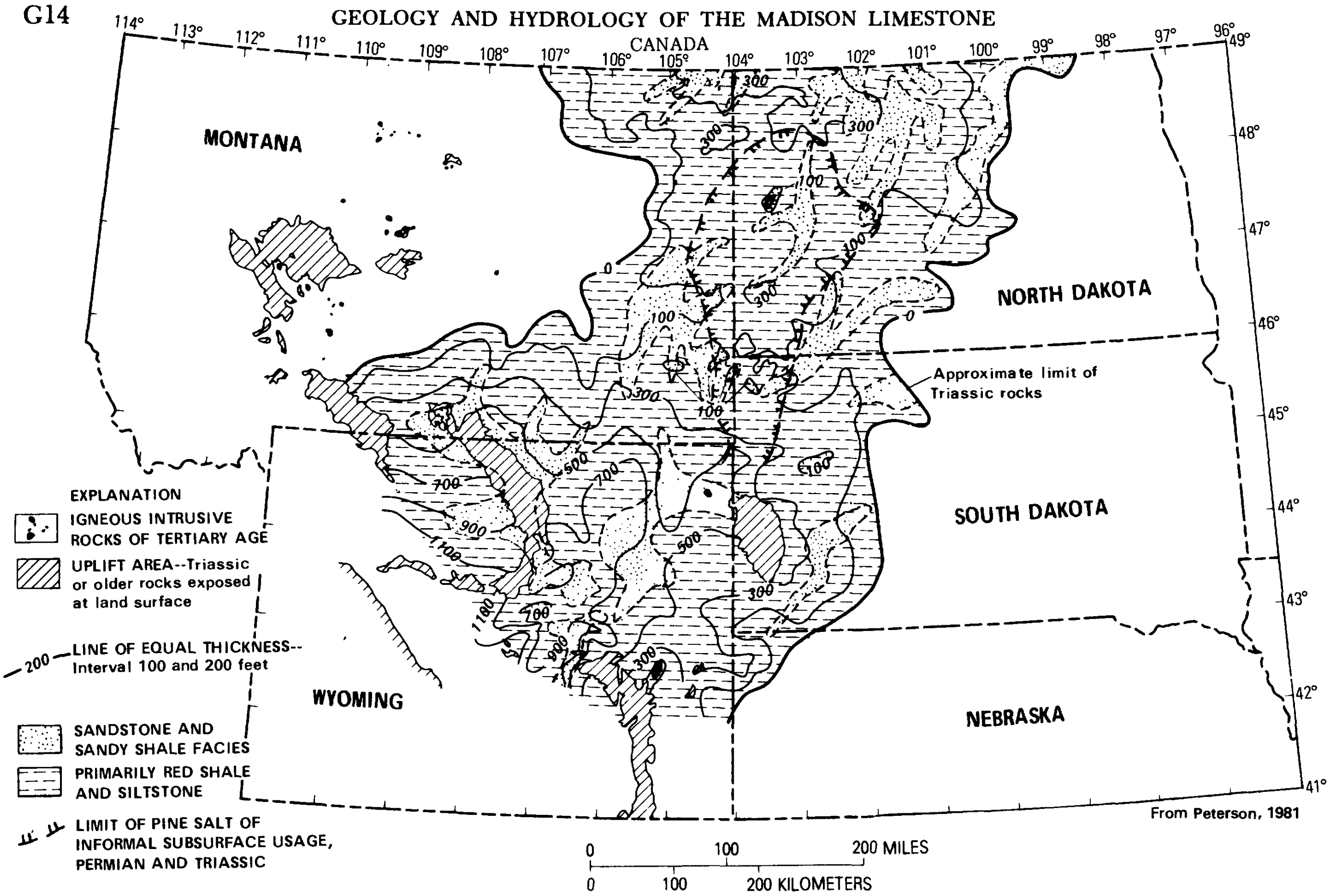

FIGURE 14.—Thickness and lithology of Triassic rocks.

salt. The Minnekahta Limestone overlies the Opeche Formation and halite units. Above the Minnekahta Limestone, the Spearfish Formation contains more than 300 ft of bedded halite, which limits the vertical flow of water through this formation.

TRIASSIC ROCKS

Rocks of Permian age in the study area are overlain by a sequence of red shale, siltstone, and evaporites belonging to the upper parts of the Goose Egg and Spearfish Formations of Triassic age (fig. 14). These formations are about 200 to 400 ft thick in the central Williston basin and thicken southwesterly to more than 900 ft in the Powder River basin.

Although shale and siltstone are the principal lithologies of the Triassic units in the study area, sandstone is present to a limited extent in the eastern part of the Williston basin as northeast-trending sandy belts probably deposited by streams flowing off the adjacent Siouxia uplift and Canadian Shield source areas to the east. Triassic beds have been truncated by pre-Middle Jurassic erosion along the southern and eastern margins of the Alberta shelf (D. L. Brown, U.S. Geological Survey, written commun., 1981) and on the east side of the Williston basin.

Rocks of Triassic age, where present in the study area, are considered to be a confining bed for flow from underlying aquifers to overlying aquifers of Cretaceous age.

JURASSIC ROCKS

Rocks of Jurassic age (fig. 15) overlie formations of Triassic age, with a pronounced disconformity. These rocks are predominantly carbonate, shale, and anhydrite. The anhydrite occurs in restricted parts of the Williston basin. The Piper Formation of Middle Jurassic age is divided into three members: a lower shale, with local nearshore sandstone in central Montana; a middle limestone; and an upper shale. In north-central Montana, the Piper Formation thins appreciably and consists chiefly of sandstone.

Total thickness of the Jurassic units in the study area range from less than 50 ft along the periphery of the Williston and Powder River basins to more than 1,000 ft north of the deepest part of the Williston basin.

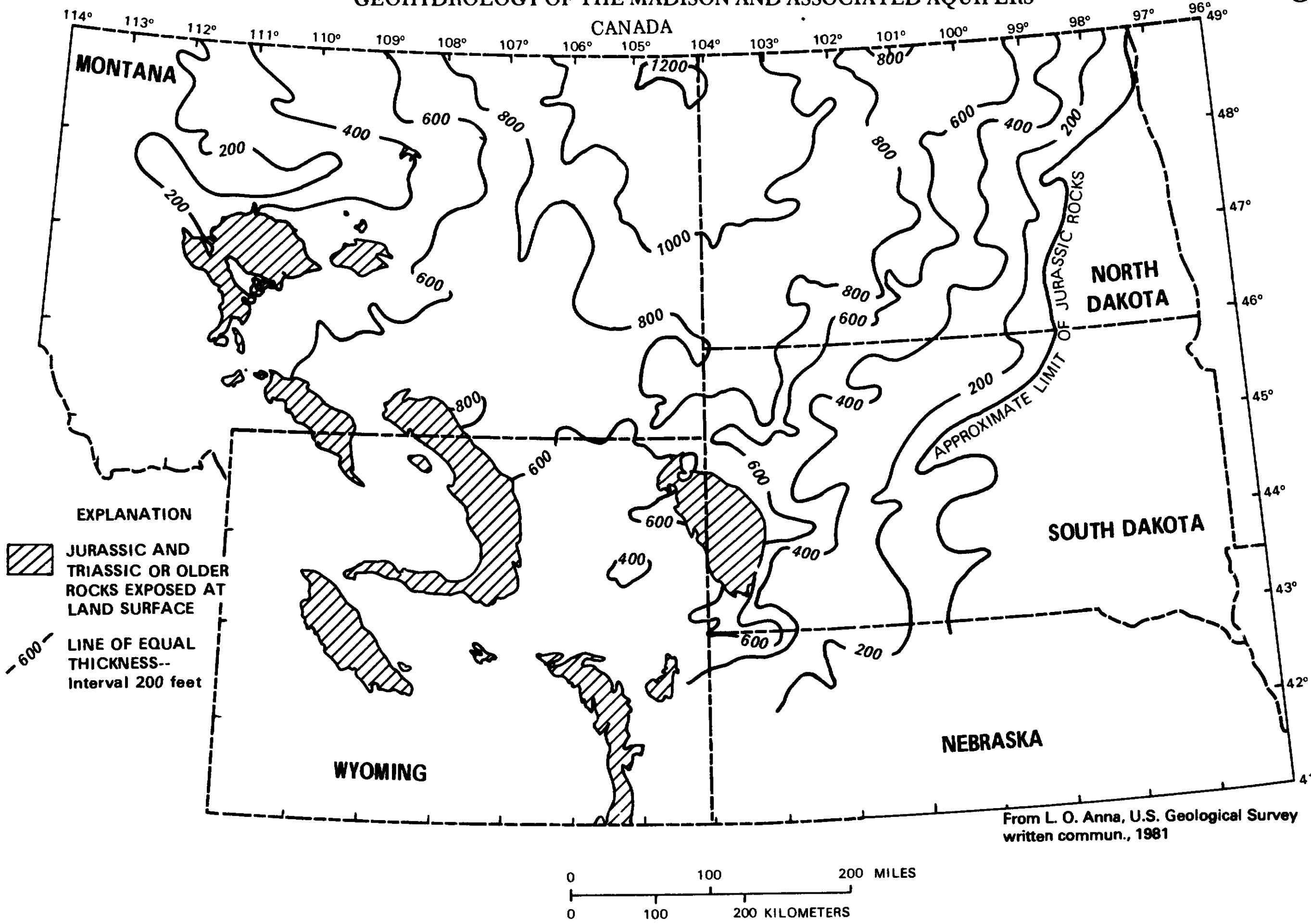

FIGURE 15. – Thickness and extent of Jurassic rocks.

GEOLOGIC STRUCTURE

Geologic structure is one of the more important factors controlling the distribution of porosity and permeability in carbonate rocks. Fracturing generally creates porosity and increases secondary permeability; both may be modified at a later time by chemical processes occurring in the aquifer as water moves through the fracture system.

Patterns of primary porosity and permeability also may be influenced during deposition as a result of structural adjustments between large basement blocks as they move in relation to each other. This block movement may create shallow-water marine environments for periods of time. Later, the same area may be lowered and a deep-water marine environment developed at the same location, resulting in a change in sedimentation rate or character. Geologic structure also can have a strong effect on permeability by influencing development of secondary porosity; faulting can influence ground-water movement by displacing or offsetting permeable beds. Some structural features that may influence ground-water flow are shown in figures 16 and 17.

Lineament patterns inferred from Landsat imagery of the study area are shown in figure 18. Where these lineament systems are associated with intensified fracturing or with increased thickness of permeable material, they indicate avenues for increased water movement (Chilingar and others, 1972), both horizontally through aquifers and vertically through confining beds. Conversely, movement normal to the directions of these features may be impeded (Konikow, 1976); this is particularly true in the case of fault zones in which permeable beds are offset.

Many of the faults and lineaments shown in figures 16, 17, and 18 are associated with present-day physiographic features that appear to affect both shallow ground-water and surface-water flow systems in the study area. An example of this is the lineament (A, fig. 18) in eastern North Dakota, which may be a control on steam-channel location, drainage, and, possibly, ground-water quality in the Devils Lake basin of North Dakota.

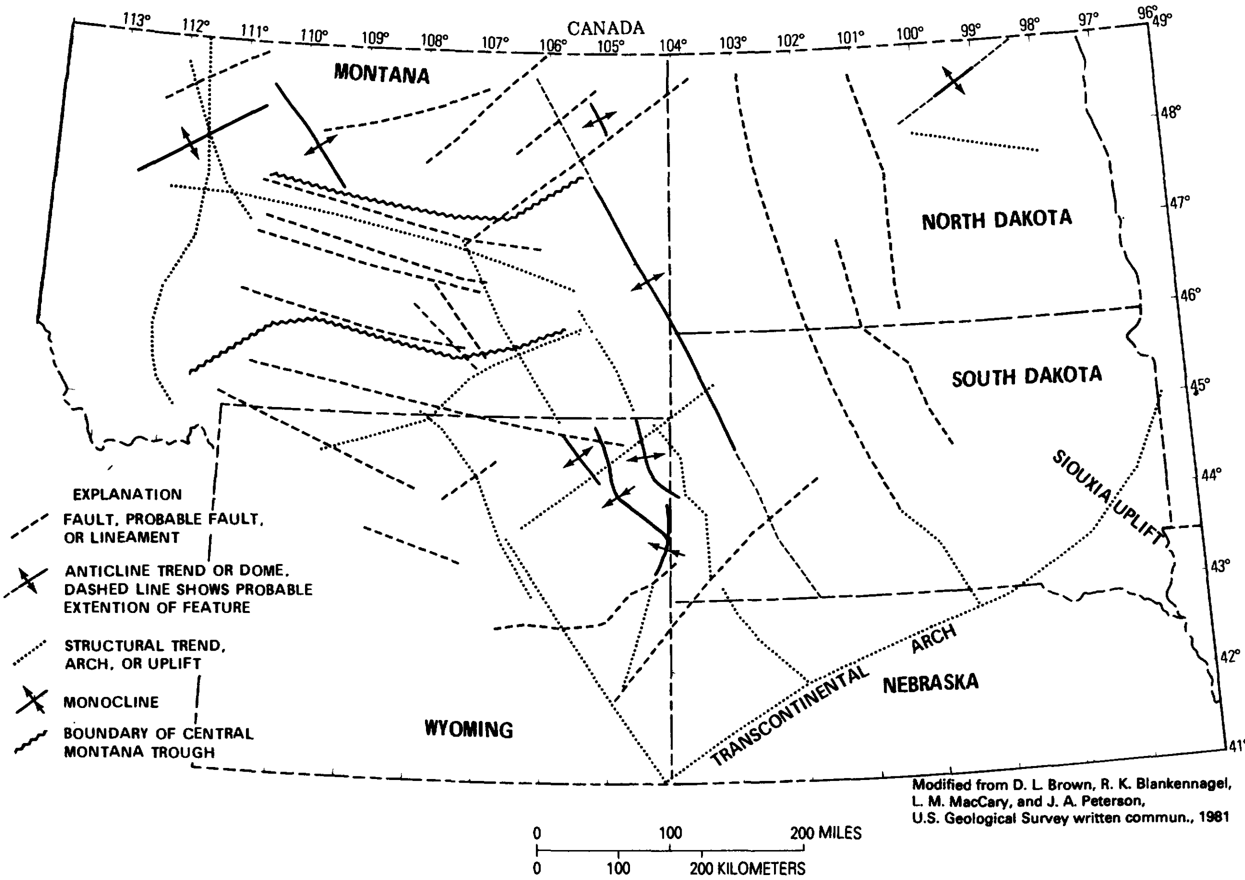

Figure 16. – Major Paleozoic structural features.

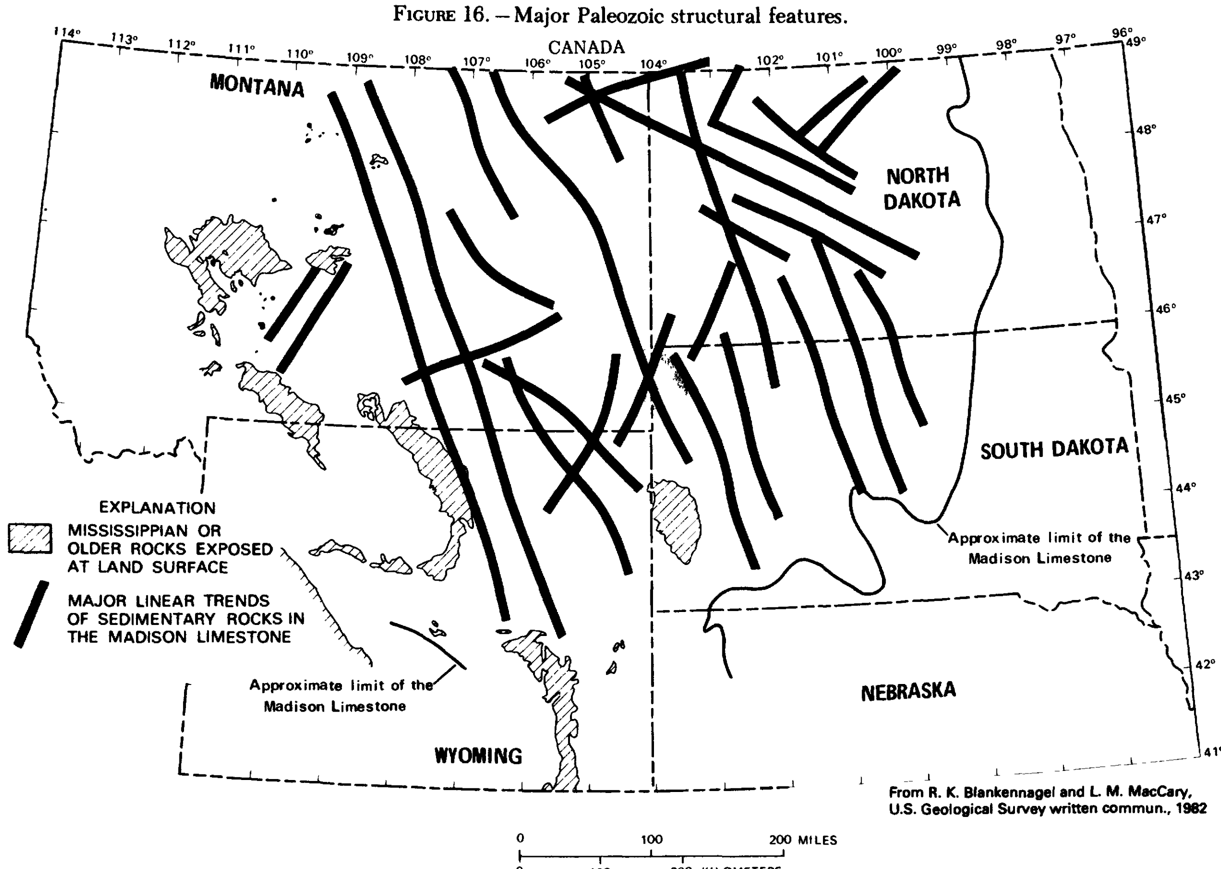

Figure 17. – Linear trends in rocks of the Madison aquifer.

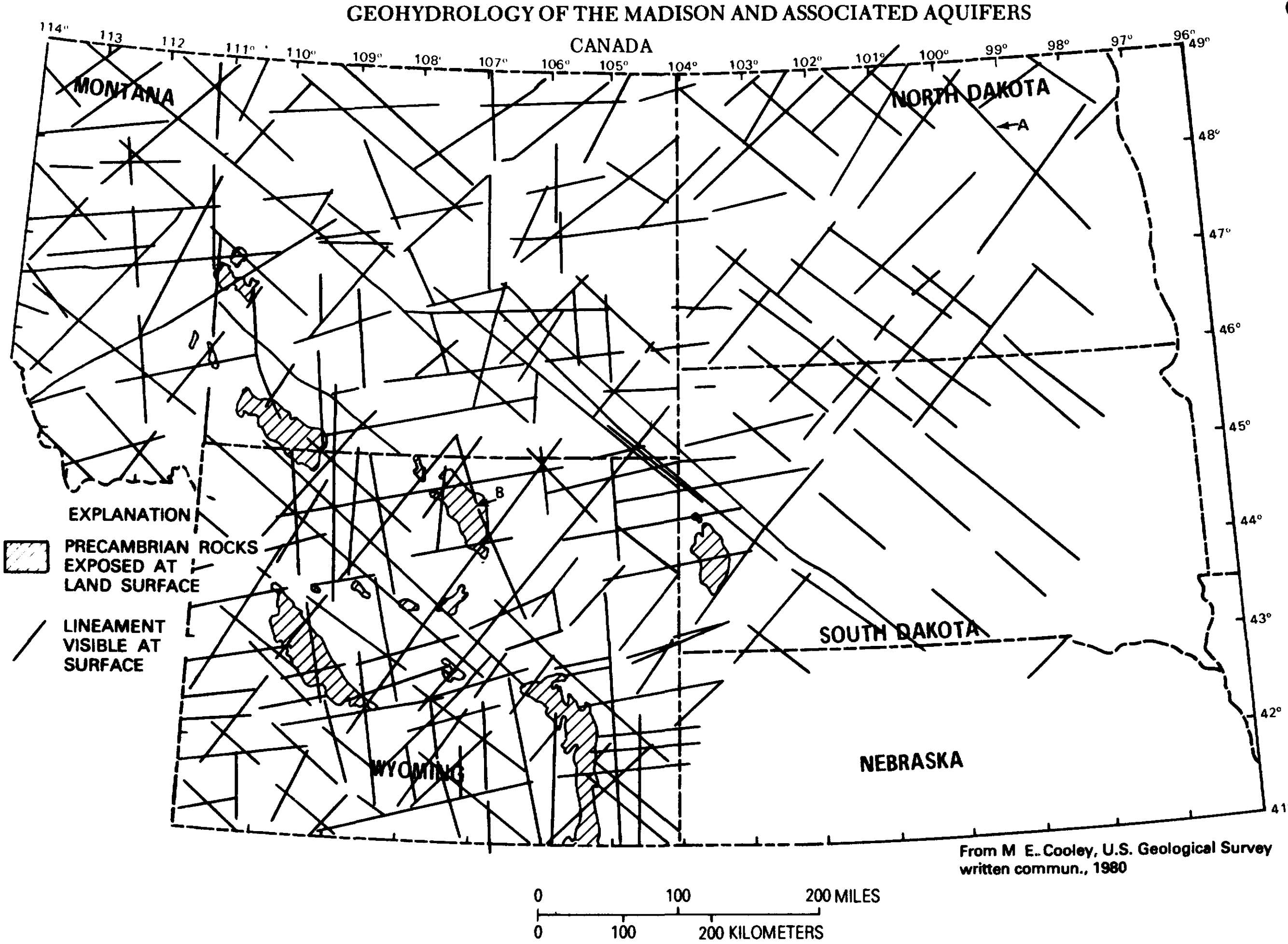

FIGURE 18. —Lineament patterns in the Northern Great Plains inferred from Landsat imagery.

HYDROLOGIC SETTING

The confined ground-water system of the Northern Great Plains includes numerous permeable horizons, many of which are discontinuous and all of which vary considerably in hydraulic properties from one location to another. During the U.S. Geological Survey investigations in this area, five major subdivisions of the aquifer system were recognized—the Cambrian-Ordovician, Madison, Pennsylvanian, Lower Cretaceous, and Upper Cretaceous aquifers. Each of these is an aggregate of permeable horizons and low-permeability, semiconfining material; each has been identified as an aquifer, primarily because vertical hydraulic-head differences within the unit tend to be smaller than those between it and the adjacent unit. To some extent, the division is arbitrary and was made to assist in analysis and discussion.

Together, these five major aquifers comprise one of the largest confined aquifer systems in the United States. The flow system extends more than 600 mi, from mountainous recharge areas in Montana, Wyoming, and South Dakota to discharge areas in the eastern Dakotas and the Canadian Province of Manitoba. The total area underlain by the aquifer system is more than 300,000 mi². In this report, attention is focused on the two lowermost aquifers—that is, the Cambrian-Ordovician and Madison aquifers. The geologic units (pl. 1 and fig. 19) comprising these aquifers and the associated semiconfining zones are as follows:

1. *Lower boundary.* Consists of crystalline rocks of Precambrian age underlying the study area.
2. *Cambrian-Ordovician aquifer* (AQ1). Consists of sandstones of Late Cambrian age and rocks of Middle to Late Ordovician age. This aquifer is composed principally of the Red River Formation of Late Ordovician age, but includes the Deadwood Formation of Middle Ordovician age, and the Stony Mountain Formation of Late Ordovician age. This aquifer underlies about 217,000 mi² in western Montana, North Dakota, South Dakota, and northestern Wyoming.
3. *Confining layer 1* (TK1). Consists of rocks of Silurian and Devonian age and the Mississippian part of the Bakken and Englewood Formations.
4. *Madison aquifer* (AQ2). Consists of rocks of Missippian age belonging to the Lodgepole and Mission

GEOHYDROLOGIC UNIT AND REPRESENTATION IN SIMULATION MODEL	SIMULATION-MODEL IDENTIFICATION
Confining unit (Charles Formation and Big Snowy Group)	TK2
Madison aquifer (Includes Lodgepole and Mission Canyon Limestones)	AQ2
Confining unit (Englewood and Bakken Formations and formations of Silurian and Devonian age)	TK1
Cambrian-Ordovician aquifer (Upper part of Deadwood, Winnepeg, Red River, and Stony Mountain Formations)	AQ1
Lower boundary Precambrian	

FIGURE 19.—Relationship between geologic units and units used in simulation model of hydrologic system.

Canyon Limestones of the Madison Group or equivalent units of the Madison Limestone. The Madison aquifer underlies about 210,000 mi² of the study area.

5. *Confining layer 2* (TK2). Consists of the Mississippian Charles Formation of the Madison Group and equivalent units of the Madison Limestone and the Big Snowy Group.

DESCRIPTION OF DIGITAL-SIMULATION MODEL

Simulation is the most versatile technique for solving problems involving large, complex, regional, ground-water flow systems such as those in the aquifers of the Northern Great Plains. Computer-based numerical simulation was used in this study to evaluate many different conceptual models of the hydrologic system. The following section briefly describes the theory, limitations, and data underlying the simulations used in this study. Geologic units and their model identifications are shown in figure 19.

THEORY

The flow of constant-density ground water in a porous medium in three dimensions may be expressed as:

$$\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial}{\partial x}\right)+\frac{\partial}{\partial y}\left(T_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(bK_{zz}\frac{\partial}{\partial z}\right) = S\frac{\partial h}{\partial t}+bW(x,y,z,t) \qquad (1)$$

where

x,y	refer to coordinate directions parallel to the bedding;
z	refers to the direction transverse to the bedding;
T_{xx}, T_{yy}	are the principal components of the transmissivity tensor (L^2T^{-1});
h	is hydraulic head (L);
b	is formation thickness (L);
K_{zz}	is the vertical permeability (LT^{-1});
S	is the storage coefficient (dimensionless);
t	is time; and
$W(x,y,z,t)$	is a source term which includes well discharge and recharge per unit surface area of the aquifer.

Equation 1 may be approximated by a finite-difference equation by applying Taylor's theorem (Pinder and Bredehkoeft, 1968). The finite-difference equation is solved at each node of a rectangular grid on a digital computer, generally using numerical methods such as the strongly implicit method (Trescott, 1975). For this study, the flow equation was solved on a digital computer using the program code developed by Trescott (1975) and modified by the addition of subroutines to calculate the effects of variable-density fluids in the aquifers.

Analysis of a complex, nonhomogeneous aquifer system is accomplished by subdividing the system into a large number of relatively small rectangular cells, which constitute a finite-difference grid. The model used in this study represented two aquifer layers, designated AQ1 and AQ2, and two semiconfining units, designated TK1 and TK2, as shown in figure 19. The horizontal finite-difference grid (fig. 20) consisted of 21 rows and 26 columns, with variable cell dimensions. Vertically, the confining units between aquifer layers were simulated using the quasi-three-dimensional approach described by Trescott (1975), in which the effects of vertical flow through the confining bed are incorporated in the vertical components of hydraulic conductivity of the adjacent aquifiers. For each cell, values of transmissivity, hydraulic head, recharge, and discharge were supplied.

Accuracy of the solution of the finite-difference equations applied to a problem in ground-water hydrology depends on a number of variables. Errors are introduced

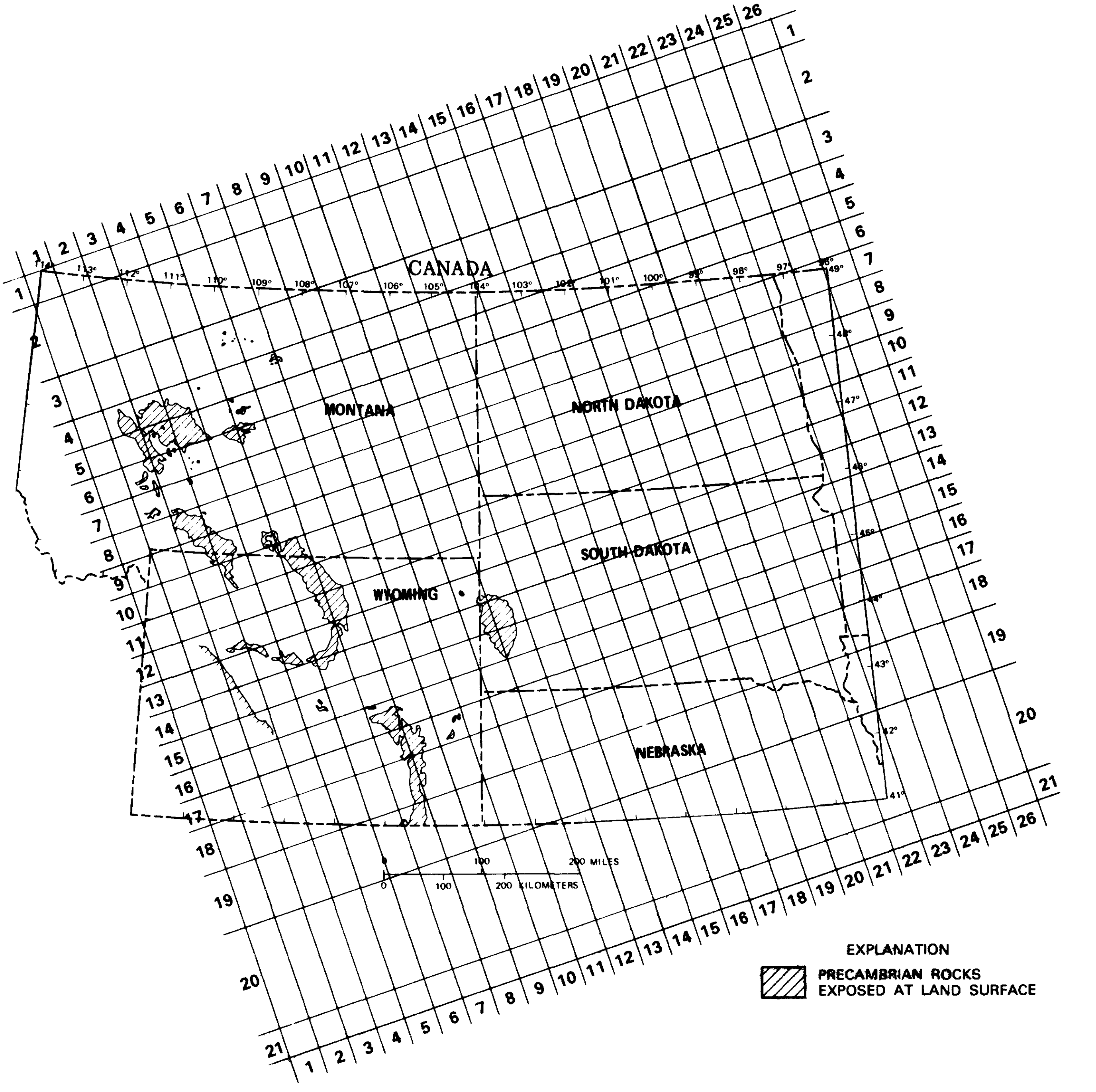

FIGURE 20. – Finite-difference grid for digital simulation of the Cambrian-Ordovician and Madison aquifers

when the continuous variation in time and space of the physical properties of a natural system are represented by discrete elements in a mathematical model. These errors can be minimized by proper design of the finite-difference grid (Pinder and Bredehoeft, 1968; Bedinger and others, 1973).

Water in the aquifer systems of the Northern Great Plains varies significantly in dissolved-solids concentration and in temperature, leading to significant variations in fluid density and viscosity from one point to another. Under undisturbed natural conditions, changes in dissolved-solids concentration or in temperature tend to occur very slowly. Thus, for purposes of studying the natural flow system, density and viscosity may be assumed to vary with position only. This assumption also can be made in studying the transient response of an aquifer to development, provided the actual movement of dissolved minerals or of thermal energy through the system, during the time interval under study, is not great enough to cause significant change in the overall density and viscosity distributions. Under the assumptions that density does not vary with time and that its distribution in space is known from onsite measurements, its effect on the flow equations can be accounted for by adding a.

function of x; y, and z to equation 1, that is,

$$\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(T_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(bK_{zz}\frac{\partial}{\partial z}\right)$$
$$=S\frac{\partial h}{\partial t}+bW(x,y,z,t)+Wps, \qquad (2)$$

where Wps is a "pseudosource" function, so termed because of its analogy to the source function W. Wps is calculated from the fluid density, which in turn is estimated from onsite data of fluid temperature, pressure, and dissolved-solids concentration; use of the pseudosource approach requires that such data be available. When formation dip is greater than 5 degrees and the pseudosource approach is used to account for hydraulic-head changes owing to variable water density, deviations from the analytically derived hydraulic-head distribution occur. Weiss (1982) found the error at 5 degrees to be about 0.03 percent. Formation dip in the Williston basin is generally less than 1 degree, and errors in the calculated hydraulic-head distribution due to the pseudosource method are small. Weiss (1982) gives a complete discussion of pseudosource terms and their use in simulating the flow of fluids of variable density, where freshwater hydraulic head is used as the field variable. Viscosity, again uder the assumption that changes with time are negligible, can be accounted for by adjustment of transmissivity and hydraulic conductivity values on the basis of onsite measurements of fluid temperature.

ASSUMPTIONS AND LIMITATIONS OF THE MODEL ANALYSIS

In using the model to analyze the predevelopment flow pattern, steady-state conditions were assumed, so that total inflow to the aquifer was taken as equal to total outflow. Thus, the value of $S\frac{\partial h}{\partial t}$ in equations 1 and 2 was always set at zero. The method of discretization is based on the assumption that within layers, which are represented as aquifers, velocity components transverse to the bedding are very small compared with those parallel to the bedding whereas between layers—that is, in the intervals represented as confining beds—components parallel to the bedding are very small compared to those transverse to the bedding. The model also incorporates the assumption that an impervious layer (corresponding to Precambrian basement material) underlies the lowest aquifer layer.

The ground-water flow equation (equation 2) does not incorporate terms to deal with the effects of chemical-osmotic potential; therefore, errors may be introduced in the final calibration results, if chemical-osmotic pressures are present in the onsite values of hydraulic head. Chemical-osmotic effects have been suggested (Hitchon, 1969) as the cause for anomalous hydraulic head and salinities in several formations in Canada north of the study area. Berry and Hanshaw (1960) noted a closed potentiometric low in the Lower Cretaceous Viking Formation of Canada, which they attribute to the effects of chemical-osmotic forces. Bredehoeft and Hanshaw (1968) suggested several chemical and physical processes that may cause anomalous pressure to occur at depth in confined aquifers.

MODEL CALIBRATION

To demonstrate that the digital model adquately simulates the hydrologic system, onsite measurements must be compared with corresponding computations of the model. The only data sets available for the project area for comparative evaluation with the model are the potentiometric surfaces based on water-level measurements for the aquifers and some general limits on the flow through the system, as indicated by the regional water balance. Although transient hydrologic effects due to Pleistocene glaciation may still be present, it was assumed that the potentiometric surfaces represented steady-state flow conditions. For comparison with the observed potentiometric surfaces, the model computed steady-state hydraulic heads from data sets containing hydrologic properties for each aquifer and confining layer and specified values of hydraulic head at certain boundary nodes.

The objective of digital simulation in this study was to improve the conceptual model of the natural flow system that existed before development of ground water or petroleum began in the area. The means of doing this was calibration of a digital model of the undisturbed flow system. The conceptual model is based on current understanding of the physical properties and nature of the aquifers and confining beds, the sources of recharge and discharge to the system, the rates and directions of flow, the variations in aquifer properties, the distribution of hydraulic potential, and the relation of the system to other aquifers. Digital simulation integrates the effects of these many factors affecting ground-water flow. If the conceptual model is sound and is accurately represented in the simulation, computed results will be consistent with field data; a difference between simulated results and field measurements indicates that some element of the conceptual model needs to be revised. In fact, maps depicting the potentiometric surfaces in parts of the study area have been reinterpreted on the basis of insights gained from the model analysis. Any adjustment of the data set used to construct the simulation model constitutes a modification of the conceptual model of the hydrologic system.

The calibration of digital-simulation models is directed toward minimizing the differences between observed and simulated steady-state potentiometric surfaces by adjusting aquifer properties and, where subject to uncertainty, specified boundary conditions. Although the number of interrelated factors affecting ground-water flow makes this a subjective procedure, the allowable adjustment of any aquifer property generally is proportional to the uncertainty of its onsite value. In this study, because hydraulic head is relatively well known, historic values of hydraulic head were adjusted very little; however, because transmissivity and vertical hydraulic conductivity are relatively unknown, many adjustments were made in these variables, and for some nodes the range of adjustment covered several orders of magnitude.

Assuming various values for aquifer properties, heads, and flows also helps to achieve another objective of the calibration procedure: To determine the sensitivity of the system to factors that affect ground-water flow. This sensitivity analysis indicates which factors control flow in the aquifers. During the calibration process, it was determined that the simulation model was sensitive to changes in leakance (a value obtained by dividing hydraulic conductivity by thickness) and relatively insensitive to changes in transmissivity. Evaluating the importance of each factor helps determine which data must be defined most accurately and which data are already adequate or require only minimal definition. Because potentiometric data were the basis of the calibration procedure, accuracy of the calibrated model was limited by the accuracy of the potentiometric data.

BOUNDARY CONDITIONS

Several different types of boundary conditions can be represented in a simulation model. These include:

1. No-flow boundary. By specifying a transmissivity equal to zero at a given node, no flow can occur across the boundary of that cell of the finite-difference grid.
2. Specified hydraulic-head boundary. When the hydraulic head in the aquifer is known, and in a transient simulation when it will not change over time, the hydraulic head in the corresponding node may be maintained at the specified value throughout the analysis.
3. Specified flux. A specified rate of withdrawal or recharge may be assigned at any of the interior notes in the model.
4. Vertical leakage. Vertical leakage into or out of the aquifer from a source bed in which the hydraulic-head distribution is known can be represented at any node by specifying the vertical hydraulic conductivity and thickness of the confining layer and the hydraulic head in the source layer. The rate of leakage is computed implicitly by the model.

In the steady-state simulations made during this study, specified hydraulic-head nodes, representing both recharge areas to the west and discharge areas to the east, were used in the active layers of the model for the Cambrian-Ordovician and Madison aquifers (figs. 21 and 22). Vertical leakage into and from the upper layer of the model was simulated by specifying hydraulic heads in an overlying source layer and assigning vertical hydraulic conductivity values and thickness values for the confining material. The confining-material hydraulic conductivities were subsequently adjusted during calibration. The specified heads used in the source layer were developed from field data on heads in Cretaceous aquifers throughout the study area.

HYDRAULIC HEAD

Field information for predevelopment hydraulic heads of the Cambrian-Ordovician and Madison aquifers is shown in figures 23 and 24. Hydraulic heads shown in these figures are actually freshwater hydraulic heads; that is, they indicate the water level that would be measured in an observation well containing water with a density of 1 g/cm^3 (gram per cubic centimeter), whether the water in the aquifer system is fresh or saline at the point of measurement. Freshwater hydraulic heads shown in figures 23 and 24 are essentially those developed by Miller and Strausz (1980a, 1980b), with some minor modifications to remove the effects of ground-water pumpage or petroleum development. These heads represent the best available synthesis of the freshwater heads that prevailed prior to the development of the formations for either of those purposes. Some of the information shown represents direct measurement of freshwater hydraulic head in the few water wells that are completed in the Paleozoic aquifers or at springs and seeps in the outcrop areas. However, most of the data consists of pressure readings from petroleum-industry drill-stem tests. These pressure data were converted to freshwater hydraulic heads by Miller and Straisz (1980a, 1980b), using standard methods.

A few of the drill-stem tests were made by U.S. Geological Survey personnel in test wells drilled specifically for the study of the Madison aquifer (fig. 1); for these tests, the freshwater hydraulic heads are estimated to be accurate to within ±50 ft. The majority of the drill-stem tests, accuracy is estimated to be ±200 ft, and the overall accuracy of the information shown in figures 23 and 24 cannot be considered better than this.

The hydraulic-head data were used for both model development and calibration: for model development, they were used to establish hydraulic heads at the specified hydraulic-head boundary nodes and in the overlying source layer for vertical leakage; for calibration, they provided the information against which simulated hydraulic heads in the interior of the model could be calibrated. Potentiometric surfaces simulated by the model at the end of the calibration process are shown

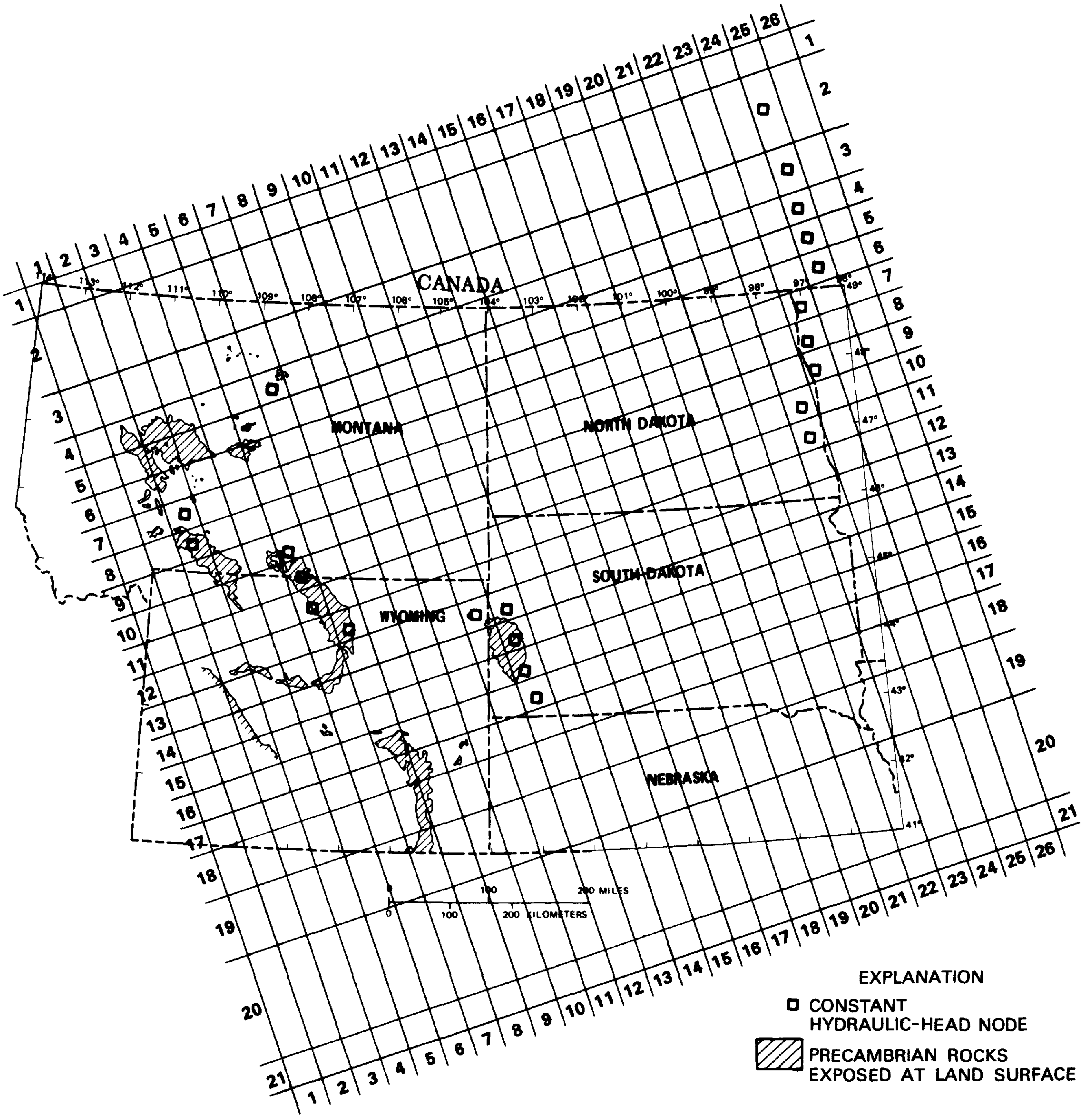

FIGURE 21.—Specified hydraulic-head nodes used in model simulation of the Cambrian-Ordovician aquifer.

in figures 25 and 26, and may be compared with the observed potentiometric surfaces shown in figures 23 and 24.

Water in the Madison and the Cambrian-Ordovician aquifers varies considerably in density, as indicated by the distribution of dissolved-solids concentrations (figs. 27 and 28). In a variable-density system of this type, gradients of freshwater hydraulic head are not directly proportional to the magnitude of flow; nor do they always indicate the actual direction of flow. However, flow direction and velocity can be calculated from freshwater hydraulic-head information, provided fluid density throughout the system is known. In this study, these calculations were made through use of a subroutine that incorporated the pseudosource term, added to the program code of Trescott (1975).

TRANSMISSIVITY

Transmissivity of an aquifer refects the rate at which ground water at the prevailing kinematic viscosity will

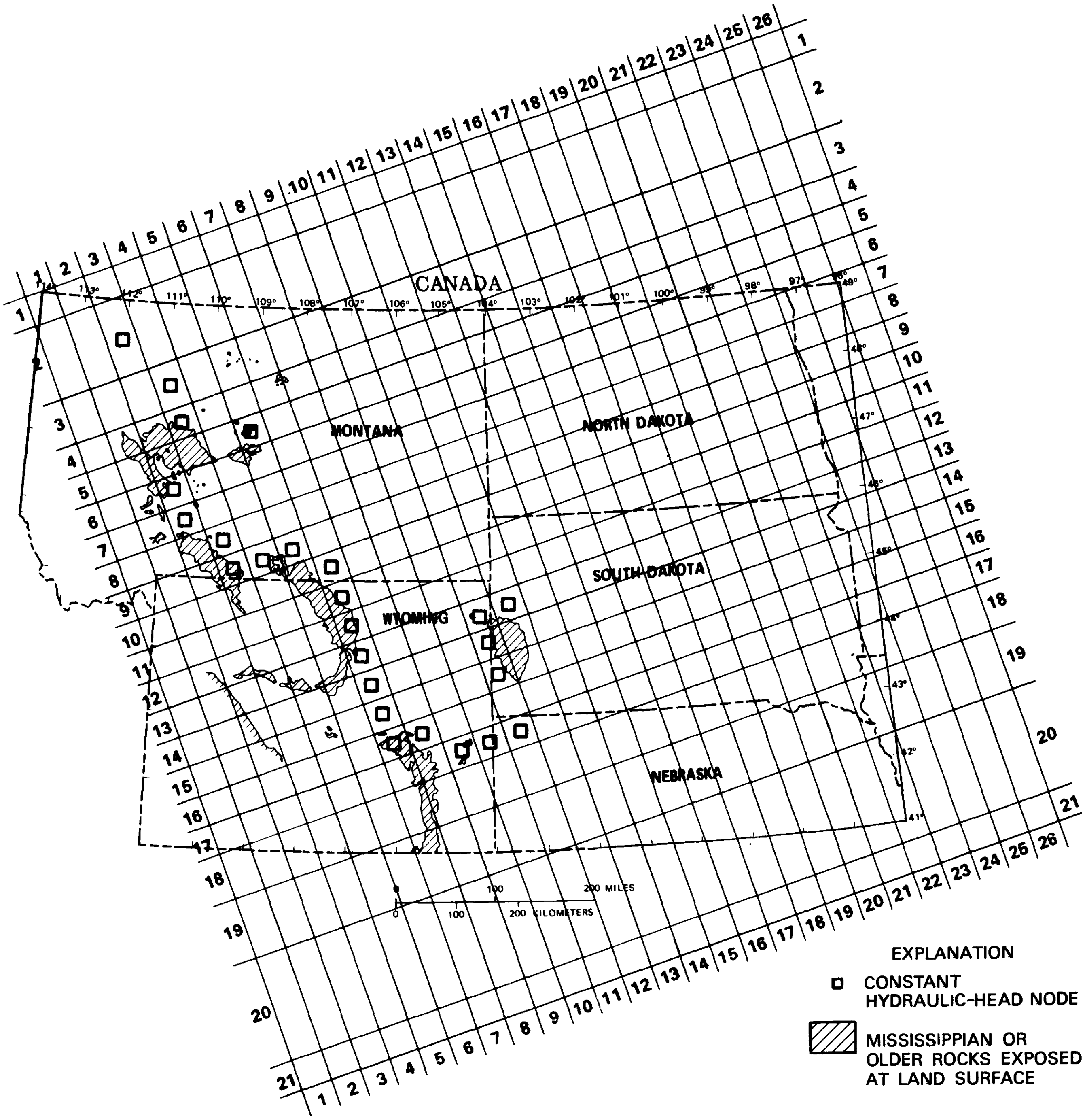

FIGURE 22. — Specified hydraulic-head nodes used in model simulation of the Madison aquifer.

flow through a unit width of the aquifer under a unit hydraulic gradient (Lohman and others, 1972). Very few data are available to describe transmissivity of either the Cambrian-Ordovician or the Madison aquifer in the study area. Miller (1976) states that data from drill-stem tests in Montana indicate that transmissivity of the Madison aquifer ranges from about 5×10^{-7} to 6×10^{-2} ft^2/s. However, Miller did not consider water viscosity changes with temperature in his calculations. The reliability of transmissivity values derived from drill-stem tests also is questionable. On the basis of an analysis of data from step-drawdown tests performed on selected perforated intervals in the Madison Limestone at Madison test well 3 (fig. 1), Blankennagel and others (1981) reported a transmissivity range from 0.37 to 0.52 ft^2/s, using an assumed storage coefficient of 2×10^{-6}. The lesser value of transmissivity appears to be associated with the upper part of the Madison Limestone at this location. The range in reported transmissivity values may be related to the significant variability of secondary porosity and permeability development along fractures and along solution features in the various limestone

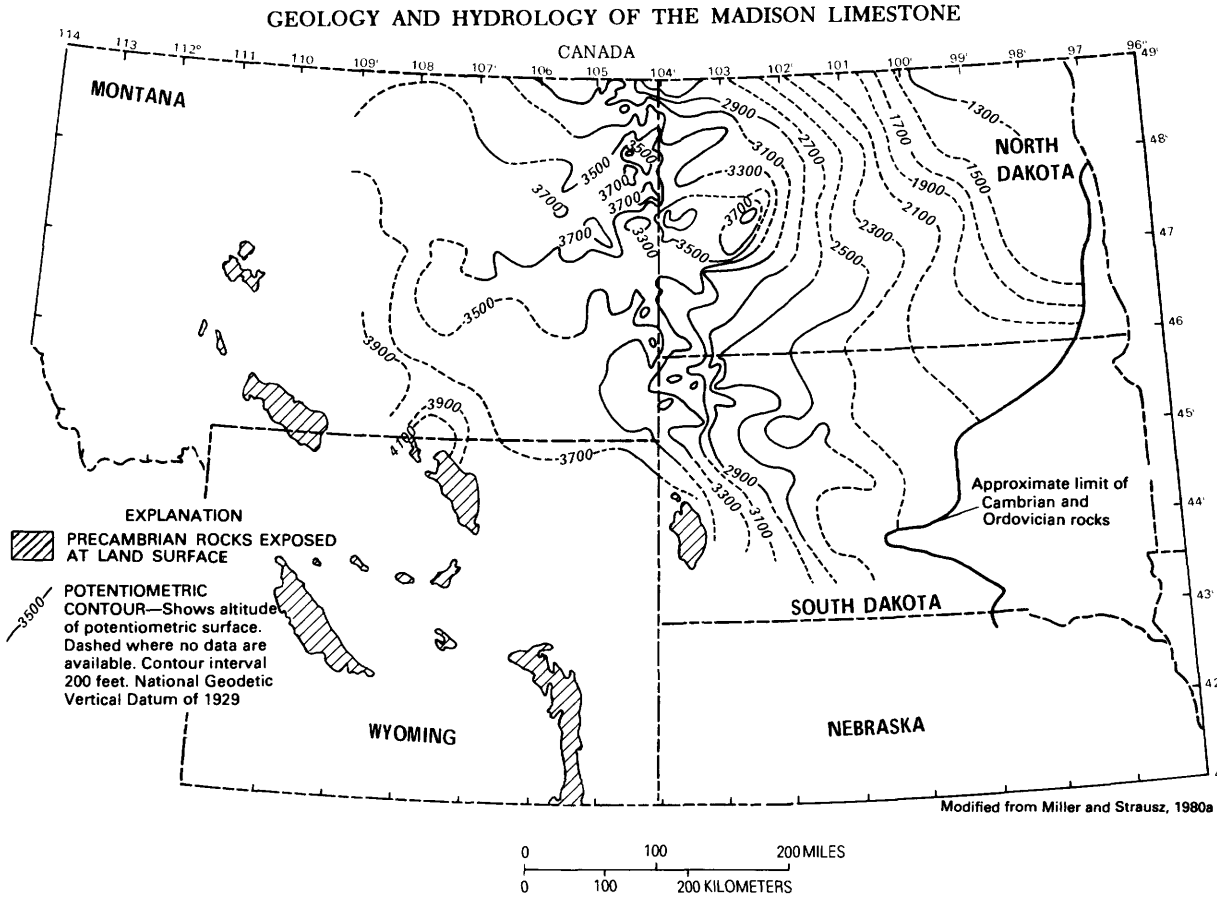

FIGURE 23. —Predevelopment potentiometric surface of the Cambrian-Ordovician aquifer.

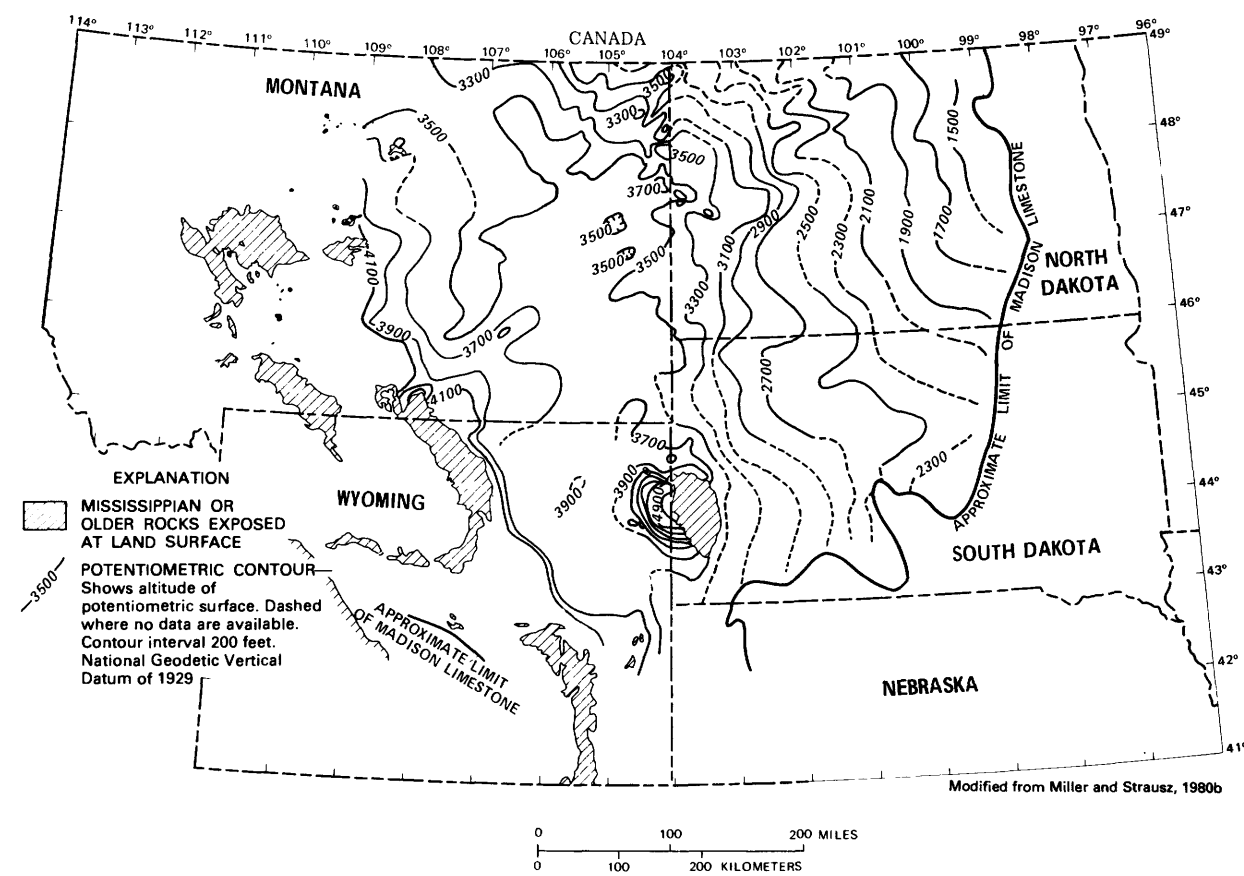

FIGURE 24. —Predevelopment potentiometric surface of the Madison aquifer.

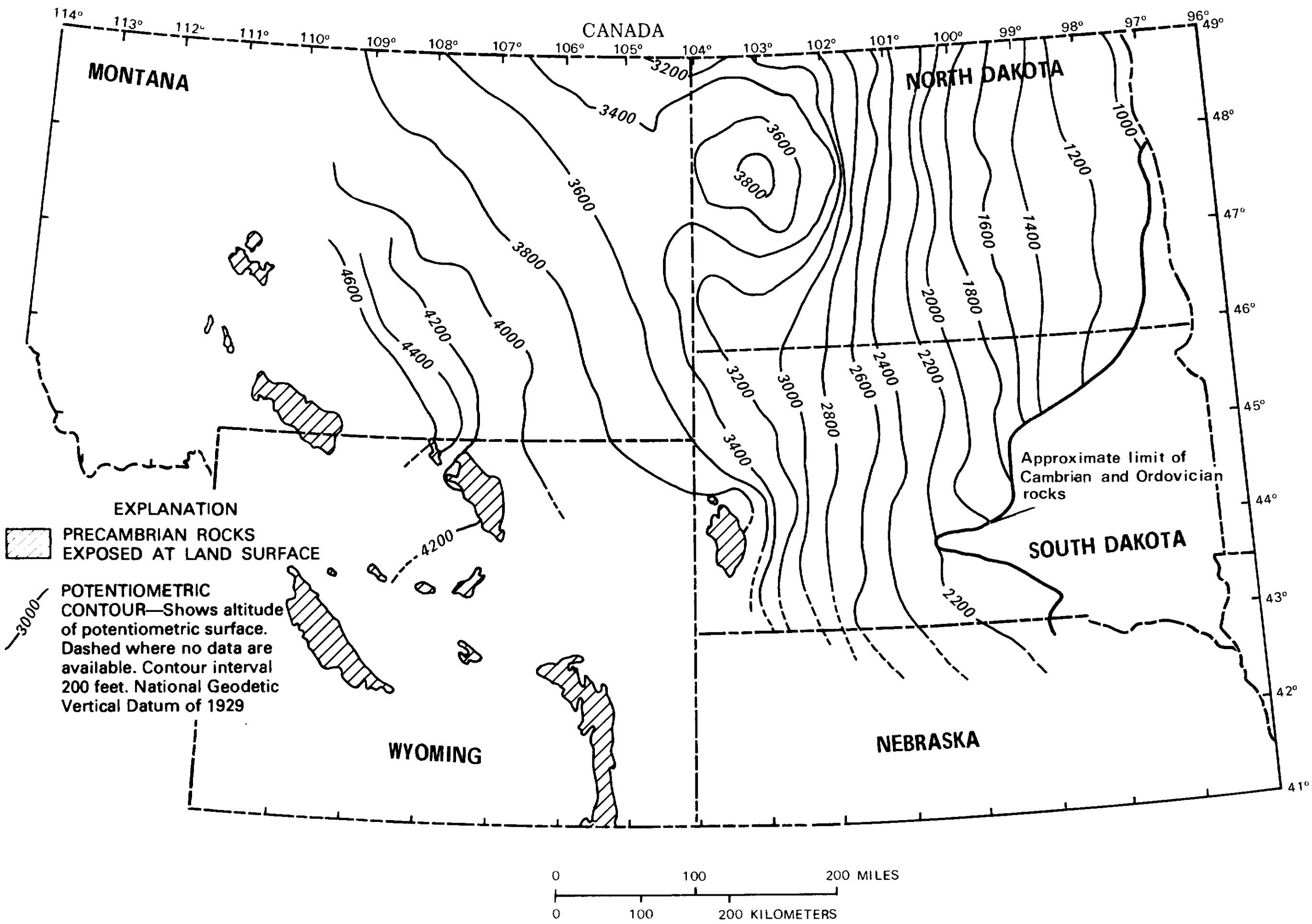

FIGURE 25. – Simulated potentiometric surface of the Cambrian-Ordovician aquifer.

formations and might indicate that point or local tests of either the Cambrian-Ordovician or the Madison aquifers do not give transmissivity values that accurately reflect either aquifer's ability to transmit water on a regional scale.

Konikow (1976) performed a flow-net analysis on the cone of depression in a well field near Midwest, Wyo., believed to yield water from the Madison aquifer. The analysis was based on the assumptions that steady-state flow exists and that no leakage is occurring from or to adjacent aquifers. Konikow's analysis indicated that an average transmissivity at this location was about 0.013 ft^2/s; however, Konikow stated that this value may be accurate only within a factor of 2 or 3 because of the lack of data about the average rate of long-term withdrawal and the exact configuration of the cone of depression. In addition, J. F. Busby, B. B. Hanshaw, and R. W. Lee (U.S. Geological Survey, written commun., 1982), in their analysis of geochemical data from the Midwest, Wyo., area, indicated that leakage from the overlying Pennsylvanian and Permian Formations into the Madison aquifer may be significant. If this is true, transmissivity of the Madison Limestone in the Midwest, Wyo., area may be somewhat smaller than 0.013 ft^2/s. Konikow used the transmissivity developed from his flow-net analysis, with adjustment at certain locations because of geologic structure, in a two-dimensional simulation of the Madison aquifer in the Powder River basin; he obtained reasonable agreement with field data during model calibration.

Final distributions of transmissivity for the Cambrian-Ordovician and Madison aquifers are shown in figures 29 and 30; they were developed as a result of the analyses carried out in this study. These transmissivity distributions were developed using many sources of data and a number of different interpretive approaches. Initially, an average hydraulic conductivity was generated by dividing Konikow's average transmissivity value by the thickness of the Madison aquifer in the area of Midwest, Wyo. This hydrauic conductivity was assumed to apply to both the Madison and the Cambrian-Ordovician aquifers, and preliminary transmissivity distributions for each aquifer were obtained by multiplying this hydraulic conductivity value by the thickness of the respective aquifer at each node of the model.

As the next step of the analysis, these initial transmissivity values were adjusted for the changes in the viscosity of water that occur with temperature (table 1). It is known that the temperature of ground water in the Cambrian-Ordovician and Madison aquifers varies from

TABLE 1.—*Adjustments to transmissivity, based on water temperature in aquifer*

Temperature range (degrees Celsius)	Median kinematic viscosity, in centistokes (square centimeters per second x 10^{-2})	Viscosity ratio[1]	Adjusted transmissivity (square feet per second)
0 - 20	1.31	0.271	0.006
20 - 40	.804	.441	.010
40 - 60	.556	.639	.015
60 - 80	.416	.853	.020
standard[2]	.355	1.00	.023
80 - 100	.328	1.08	.025
100 - 120	.269	1.32	.030
120 - 140	.227	1.56	.036

[1]This multiplication factor equals the ratio of (1) the kinematic viscosity at the temperature of the standard of reference, to (2) the kinematic viscosity at the mean temperature of a zone.

[2] All transmissivity adjustments are referenced to the standard transmissivity at 83 degrees Celsius.

about 8°C in or near outcrop areas to more than 150°C in some deeper parts of the Williston basin. Kinematic viscosity of water decreases as its temperature increases (fig. 31). Thus, hotter water can be transmitted through a given aquifer at a lower hydraulic gradient than an equal flow of cooler water. Because kinematic viscosity of water is dependent on its temperature and because transmissivity is inversely proportional to kinematic viscosity, effective transmissivity of the aquifer will vary as a function of water temperature.

Measured water temperatures in the Cambrian-Ordovician and the Madison aquifers are shown in figures 32 and 33. Lines of equal temperature in figures 32 and 33 delineate temperature zones; these zones were used in the calculation of transmissivity for each node of the model. Transmissivities thus adjusted are actually transmissivities to freshwater having a specific gravity of 1.0 and the temperature of the formation water at the point in question; these are transmissivities required when the pseudosource appoach described by Weiss (1982) is used.

These viscosity-adjusted transmissivities, based on the hydraulic conductivity calculated from Konikow's work, were then further adjusted in an attempt to account for porosity variations within the aquifers. MacCary (1981) presents maps (figs. 34 and 35) showing total thickness of rock in each aquifer having a porosity of 10 percent or greater. Using these maps, an average thickness of material having a porosity of 10 percent or more was calculated for the block of aquifer material associated with each node. A set of transmissivity factors was selected arbitrarily to correspond to the various thicknesses of porous material. These factors are shown in table 2. The viscosity-adjusted transmissivity value of each node was multiplied by the appropriate factor from table 2, according to the average thickness of porous material calculated for the node.

Transmissivity of both aquifers also was adjusted in areas where geologic data indicated the existence of major barriers to ground-water flow through the aquifer. In the Madison aquifer, for example, a zone of steep hydraulic gradient exists along the eastern flank of the Bighorn Mountains and the western side of the Black Hills (fig. 24). This steep gradient zone occurs in areas where the Madison Limestone is steeply dipping or has been offset in the subsurface by major faults (figs. 16, 17, and 18). Total offset of the aquifer would form a barrier to ground-water flow perpendicular to the faulting and would result in flow trending parallel to the barrier. Transmissivity values were, therefore, decreased to ac-

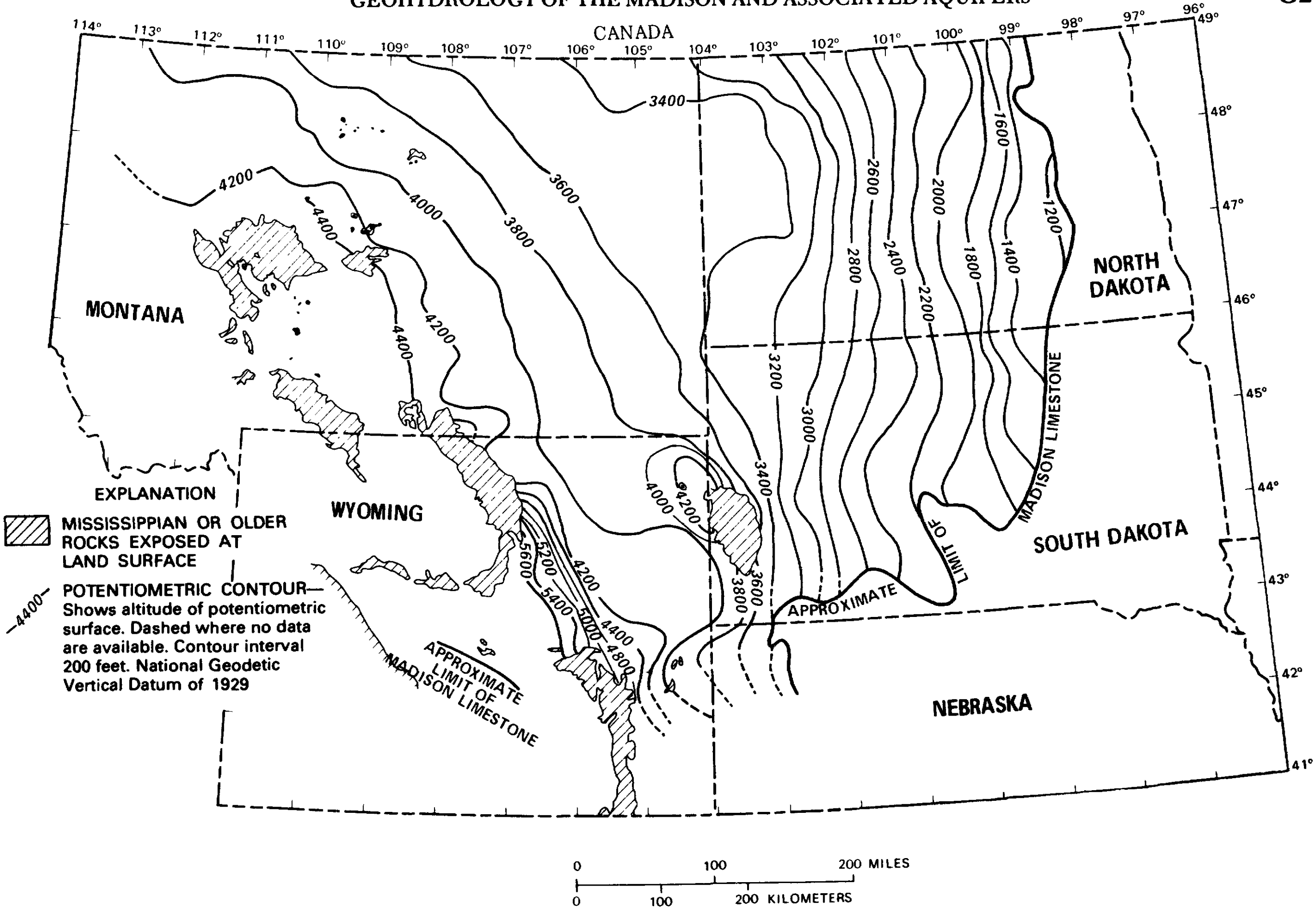

FIGURE 26.—Simulated potentiometric surface of the Madison aquifer.

Table 2.—*Transmissivity adjustment factors based on thickness of rock with porosity greater than 10 percent*

Thickness of rock with porosity greater than 10 percent (feet)	Transmissivity adjustment
0 - 10	0.2
11 - 20	.4
21 - 30	.6
31 - 50	.8
51 - 100	1.0
101 - 150	1.2
151 - 200	1.4
201 - 300	1.6
> 300	1.8

count for this barrier effect in the areas along the eastern Bighorn Mountains and the western side of the Black Hills. Similar adjustments were made in other areas where structure appears to control the flow pattern. Transmissivities thus adjusted for viscosity, porosity, and geologic structure were used as initial values in the model analysis and were further adjusted during calibration. Transmissivity distributions shown in figures 29 and 30 represent the final distributions obtained at the end of the calibration process.

VERTICAL HYDRAULIC CONDUCTIVITY

Vertical hydraulic conductivity of the aquifers and confining units is the least certain of the hydrologic properties characterizing the system. Very few core analyses from oil tests in the project area include vertical hydraulic conductivity as a part of the test results; although, in a few instances, vertical permeability to air has been determined. One test performed on core from the Lodgepole Limestone of the Madison Group in Montana indicated a vertical permeability ranging from 0.00017 to 0.05 ft/s; however, this analysis also reported oil saturation ranging from 3.6 to about 25 percent, and this may limit the applicability of the test results to the vertical flow of water through the formation.

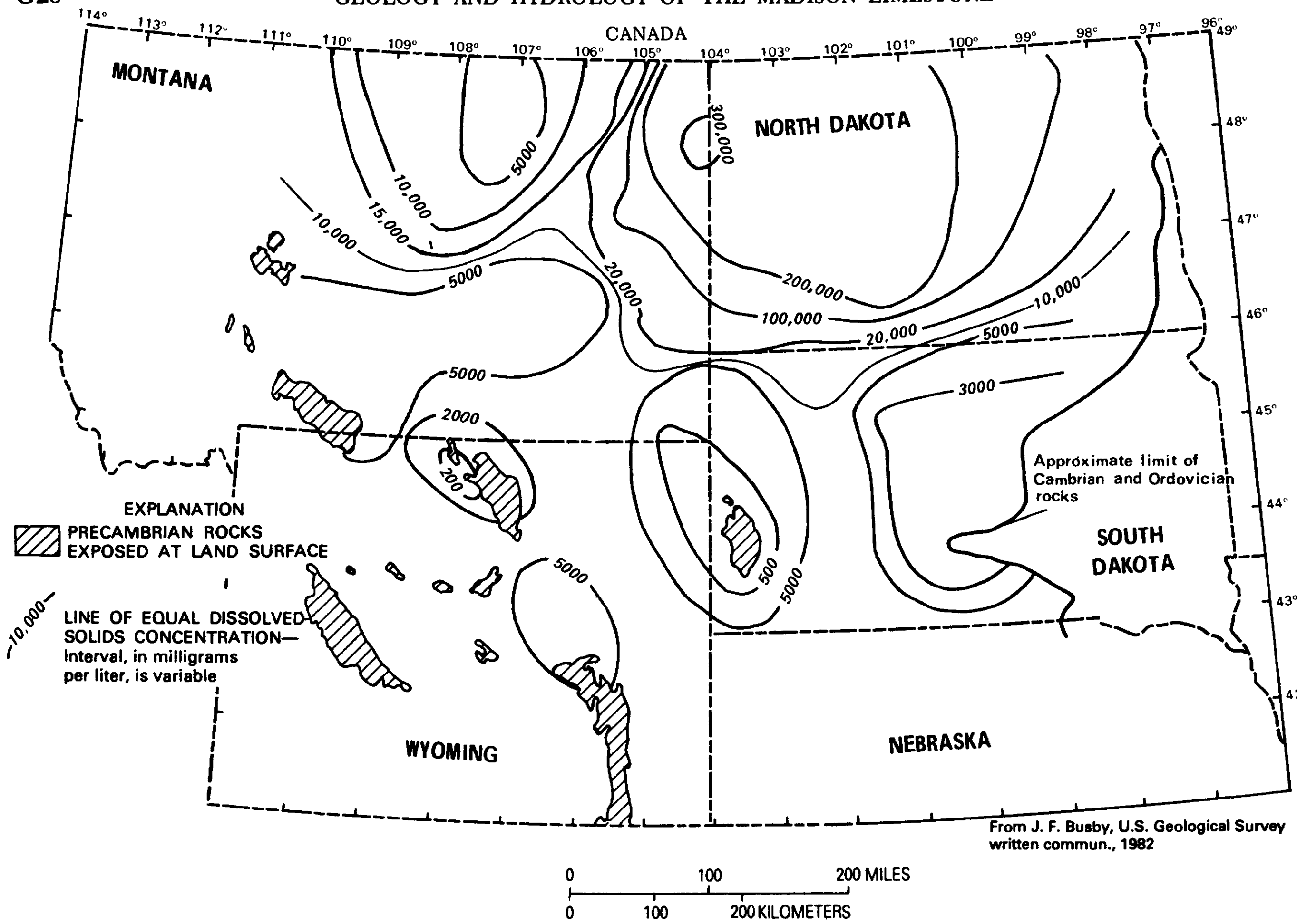

FIGURE 27.—Dissolved-solids concentration in water from the Cambrian-Ordovician aquifer.

General experience with hydrologic simulation of consolidated-rock units indicates that point values of vertical hydraulic conductivity, such as those determined in core analysis, seldom are indicative of the effective regional vertical hydraulic conductivity of confining units, that is, the hydraulic conductivity value required to describe the flow through large areal segments of those units. This is because much of the flow occurs through fractures, solution openings, and other secondary features that are not represented in isolated cores, and because it generaly is impossible to obtain sufficient core data to follow lithologic changes in the confining beds. In this study, a preliminary value of vertical hydraulic conductivity was estimated for each confining unit at each model node on the basis of lithologic and porosity maps (figs. 34 and 35), water-temperature maps (figs. 32 and 33), and information on geologic structure, particularly those aspects related to secondary permeability.

For each model node, the estimated value of vertical hydraulic conductivity was used together with confining-bed thickness to calculate a value of leakance (hydraulic conductivity divided by thickness) that was used in the preliminary model simulations. During the model-calibration process, the values of leakance for each node were adjusted within reasonable limits to assist in matching the calculated hydraulic-head values with the observed data, that is, with the hydraulic-head values from the potentiometric maps. Final values of vertical hydraulic conductivity at the end of the calibration process are shown in figures 36 and 37.

GEOHYDROLOGY

The flow system in the Cambrian-Ordovician and Madison aquifers is part of an integrated regional system that also involves several aquifer units overlying the Madison. In general, the pattern is one of recharge in high areas to the west, regional flow to the east and northeast, and discharge in the eastern Dakotas or in the Province of Manitoba in Canada. In the western segment of the area, downward flow across confining beds is the rule, and the deeper aquifers generally gain in flow. In the eastern and northeastern part of the area, upward leakage across confining beds is the rule, ultimately leading to discharge at the surface. The Williston basin exerts a major influence on the flow system, in that the aquifers reach great depths at which high temperatures

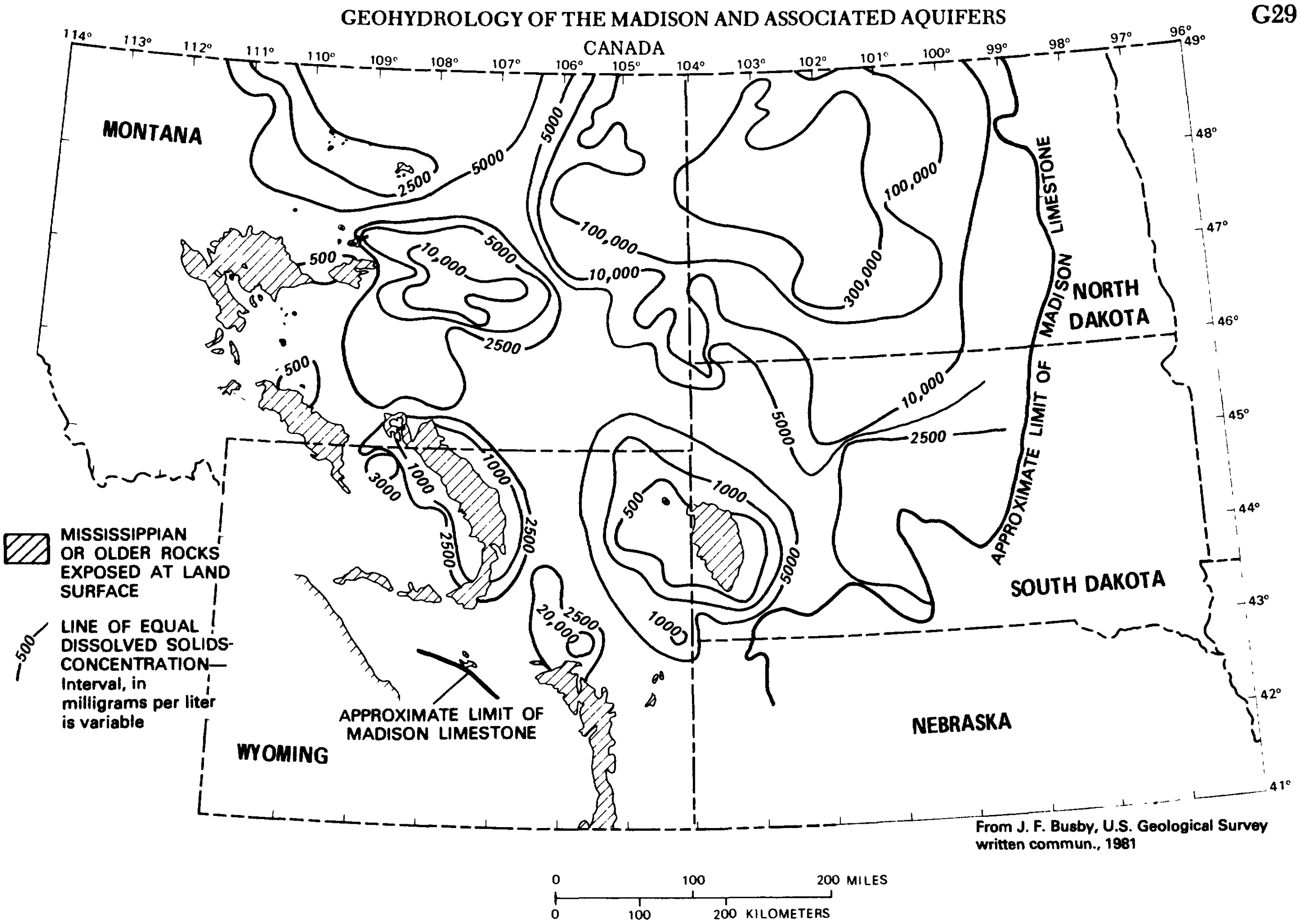

FIGURE 28. — Dissolved-solids concentration in water from the Madison aquifer.

and salinities prevail. More local geologic structures also exert an influence on the flow pattern.

Both the Cambrian-Ordovician and the Madison aquifers crop out and receive recharge in the highland areas in the western part of the study area (fig. 2). Recharge also occurs in aquifer outcrops in the Black Hills uplift area of South Dakota. Virtually all eastward-flowing streams draining these areas lose part of their flow (Swenson, 1968) as they cross the aquifer outcrop areas. Although small in quantity, recharge also is received from precipitation falling directly on the exposed rocks.

Streamflow measurements on several streams draining the east side of the Black Hills (fig. 1) indicated that as much as 10 ft³/s was being lost from the streams as they crossed the outcrop of the Madison Limestone (Swenson 1968). Prior to a program of stream-channel sealing during 1937, streamflow losses of about 100 ft³/s were reported by Powell (1940). On the basis of similar aquifer lithology and degree of weathering, it is reasonable to assume that most streams draining the western mountainous areas, such as the Bighorn Mountains, would lose similar amounts of flow as they cross aquifer outcrops of comparable area.

While available surface-water data indicate that large quantities of water enter the aquifers along the outcrop areas in the western highlands, not all of this water recharges the deep, regional flow system and moves to the eastern discharge area. A large part of the water entering each aquifer is discharged within a short distance from springs and seeps along the flanks of the mountainous areas (Swenson, 1968; Rahn and Gries, 1973; Hodson, 1974). The fraction of the total recharge that remains in the aquifer system enters the regional flow system.

The predevelopment (about 1950) regional flow regimes for the Cambrian-Ordovician and Madison aquifers, as indicated by the results of the final model simulation after completion of the calibration process, are shown in figures 38 and 39. Approximate rates of ground-water movement in the Cambrian-Ordovician and Madison aquifers are shown in figures 40 and 41. These rates of movement were calculated by using flow results taken from the simulations and assuming an average porosity of 10 percnt throughout both aquifers for purposes of calculation.

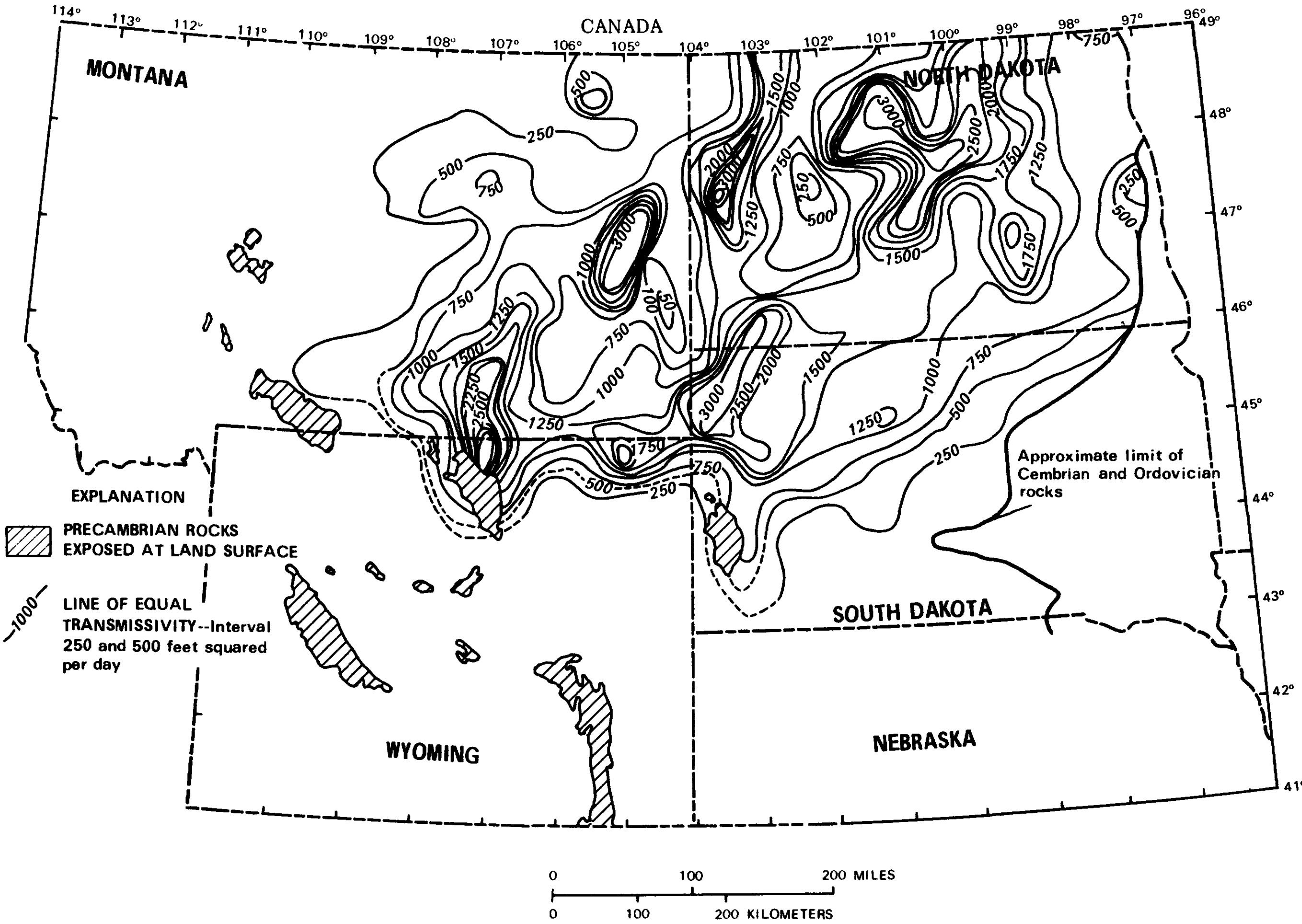

FIGURE 29. —Transmissivity distribution in the Cambrian-Ordovician aquifer.

Rates of recharge for selected recharge areas are shown in figures 38 and 39; these rates represent infiltration that enters both aquifers in areas where they are close to or at land surface and infiltration that remains within the regional flow system. Rates of discharge for selected areas also are shown. Discharge from the Madison aquifer is to adjacent and overlying formations, particularly along the eastern subcrop of the Madison aquifer. Discharge from the Cambrian-Ordovician aquifer is to adjacent aquifers, to surficial aquifers, and thence to lakes and springs in eastern North Dakota; discharge from the Cambrian-Ordovician aquifer in Canada is through springs and seeps where the aquifer crops out. Rates of recharge shown in figures 38 and 39 represent water that actually enters the regional flow system in the indicated areas, as determined from the simulation results. Shallow infiltration that discharges locally from springs and seeps is not included in these figures.

The Madison aquifer does not crop out in the eastern part of the study area. Formations comprising the aquifer terminate in the subsurface, in the pattern shown in figure 42, and are overlain by younger rocks consisting mostly of shales. Thus, discharge from the Madison aquifer in this area consists of vertical leakage through these overlying units and eastward flow into the Cambrian-Ordovician aquifer; this discharge, although stratigraphically deeper than the Madison, is carried upward past the subsurface termination of the Madison by the regional dip. This flow from the Madison aquifer to the Cambrian-Ordovician aquifer is indicated as a recharge of 20 ft^3/s to the Cambrian-Ordovician aquifer in figure 38, and as a loss of 20 ft^3/s from the Madison aquifer in figure 39. As indicated in figure 39, groundwater discharge is concentated along the eastern, rather than the northern, limits of the Madison aquifer. This is in agreement with the observation that, in the Canadian Provinces of Saskatchewan and Manitoba, stratigraphic unconformities between Paleozoic aquifers and overlying confining units have resulted in conditions favorable to accumulations of oil and gas in stratigraphic traps (McCabe, 1963). As a corollary to this, conditions in this area must be generally unfavorable for discharge of ground water from the Madison aquifer by vertical leakage.

Recharge and discharge rates shown in figures 38 and 39 do not account for the entire steady-state flow through the Cambrian-Ordovician and Madison aquifers.

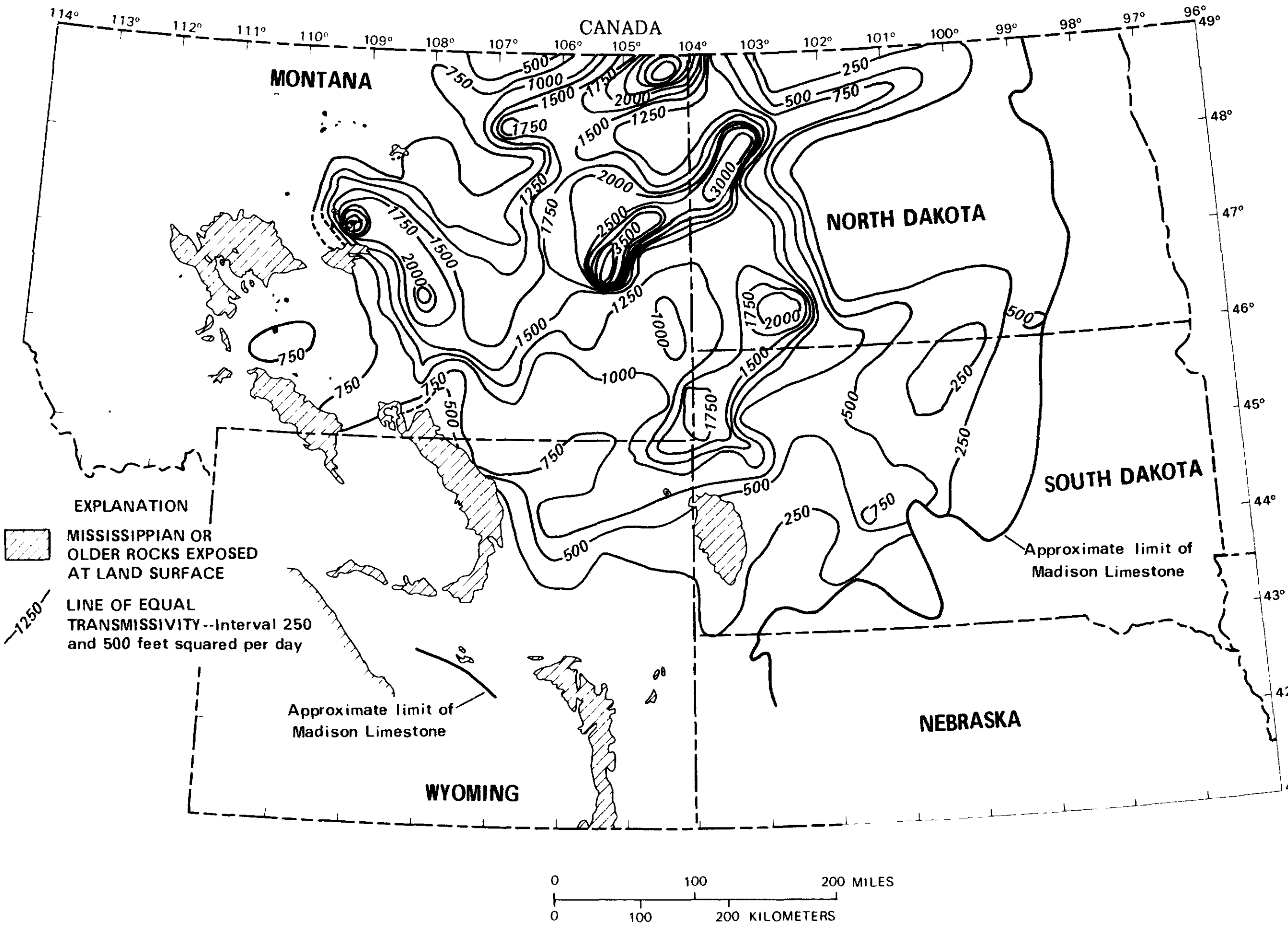

FIGURE 30. — Transmissivity distribution in the Madison aquifer.

Discharge occurs from both aquifers northward from the Big Snowy Mountains and associated highland areas in Montana into the Canadian Province of Saskatchewan. The Weldon-Brockton fault zone and other structural features in northestern Montana (fig. 2) appear to be channels for water movement toward discharge areas in Canada north of the Williston basin.

Geologic structure appears to be an important control (Weimer and others, 1982) in determining the rate and direction of water movement in the aquifer systems in the study area. For example, the Casper Mountain fault (fig. 2) appears to prevent ground-water flow (fig. 38) to the south from the Powder River basin, and the major fault system bounding the Bighorn Mountains on the east (B, fig. 18) limits recharge to the Powder River basin (Konikow, 1976). Recharge from the Bighorn Mountains appears to be channeled by geologic structure associated with a major lineament zone (B, fig. 18); south of the Cedar Creek anticline, to join the flow system from the Black Hills recharge area (fig. 38). This flow system continues along the southern part of the Williston basin northeastward to the discharge area in northeastern North Dakota and eastern Manitoba. The Weldon-Brockton fault zone (fig. 2) appears to be a major channel for water movement from the big Snowy Mountains and associated highland areas in Montana to discharge areas in Canada north of the Williston basin.

Both the Cambrian-Ordovician and the Madison aquifers exhibit areas of very low ground-water flow on the eastern flank of the Williston basin (figs. 40 and 41) coinciding with areas of high dissolved-solids concentration and, therefore, high fluid density (fig. 27). Three hypotheses were considered in developing an explanation of the hydrologic flow system in and near the areas of high-density fluid. The first hypothesis is that the brine is static and that the hydrologic situation is similar to that described by Hubbert (1969), in which freshwater flowing through a synclinal structure comes into contact with static, dense brine along a sharp, fluid interface. Hubbert shows that the body of saline water under these conditions does not lie uniformly in the deepest of the outflow flank. That is, while the inflow flank is occupied entirely by moving freshwater, the outflow flank contains static brine in the lower part of the aquifer and moving freshwater above this lower part. For the Paleozoic aquifers of the Williston basin, Hubbert's analysis would

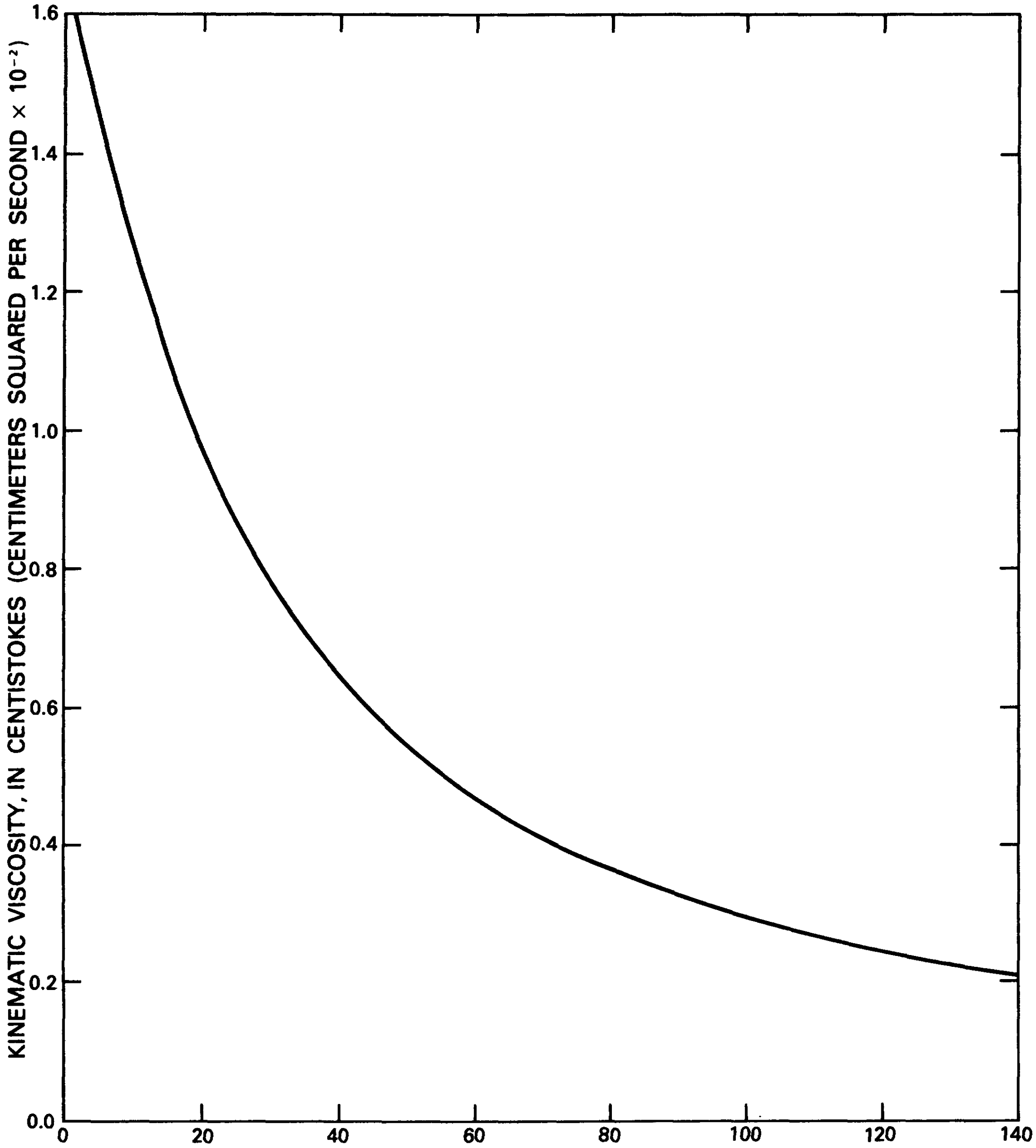

FIGURE 31. — Relationship between viscosity and temperature of water.

be modified to allow flow around, as well as above, the dense brine, reflecting the fact that the structure is actually a basin rather than a simple syncline. Moreover, the flow above the brine would presumably occur through upward leakage into units overlying the Madison, rather than as a flow in the upper part of the Madison itself.

In any case, however, simulation results show small but consistent flow velocities generally directed to the east and northeast, through the high-density brines, in both the Cambrian-Ordovician and Madison aquifers. This suggests that a small component of the regional flow actually moves directly across the Williston basin from west to east, through the high-salinity areas. Thus, the second hypothesis regarding the brine is that it actually represents a very slow-moving segment of the regional flow system. This intepretation allows an explanation for the origin of the brine: solution of salt from halite beds as the water moves through the basin, as described, by Grossman (1968) and Milner (1956) for corresponding areas in Canada. The process of solution is enhanced by increasing water temperature with depth; the highest salinities are found in regions of high temperature. Reduction in salinity in updip areas on the eastern flank of the basin presumably is due at least in part to precipitation of halite associated with cooler temperature, although dilution by fresher water undoubtedly is also a factor. To the extent that precipitation of halite occurs, it should result in a very gradual reduction in permeability, over geologic time intervals, in the

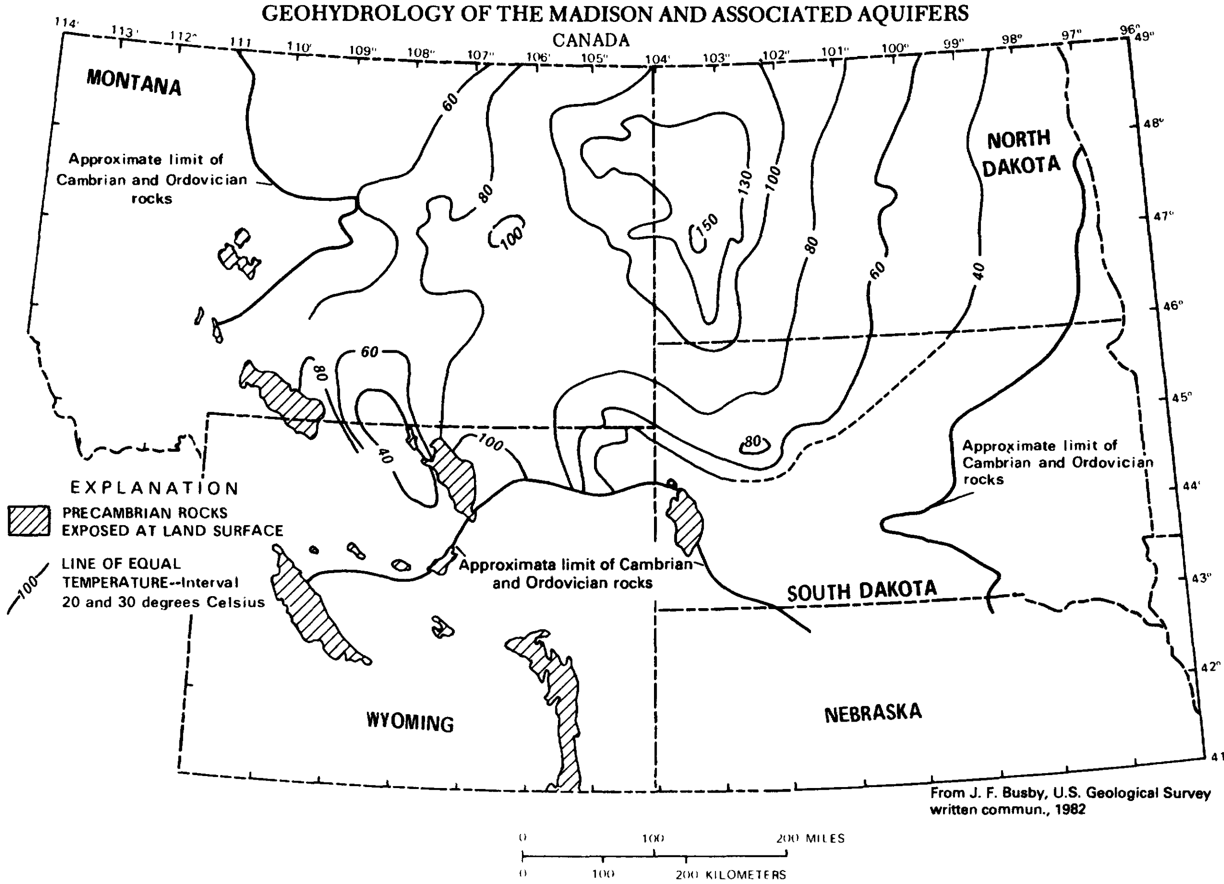

FIGURE 32. – Water temperture in the Cambrian-Ordovician aquifer.

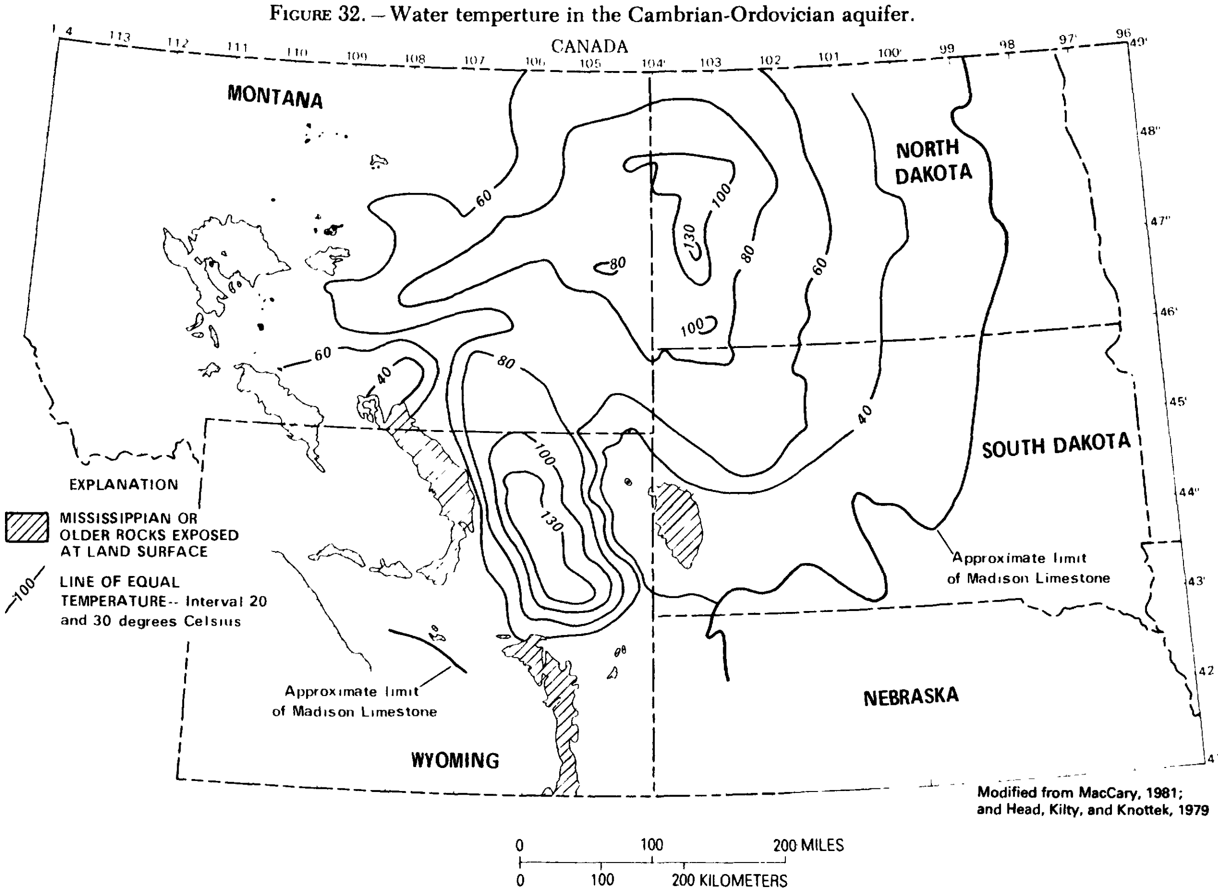

FIGURE 33. – Water temperature in the Madison aquifer.

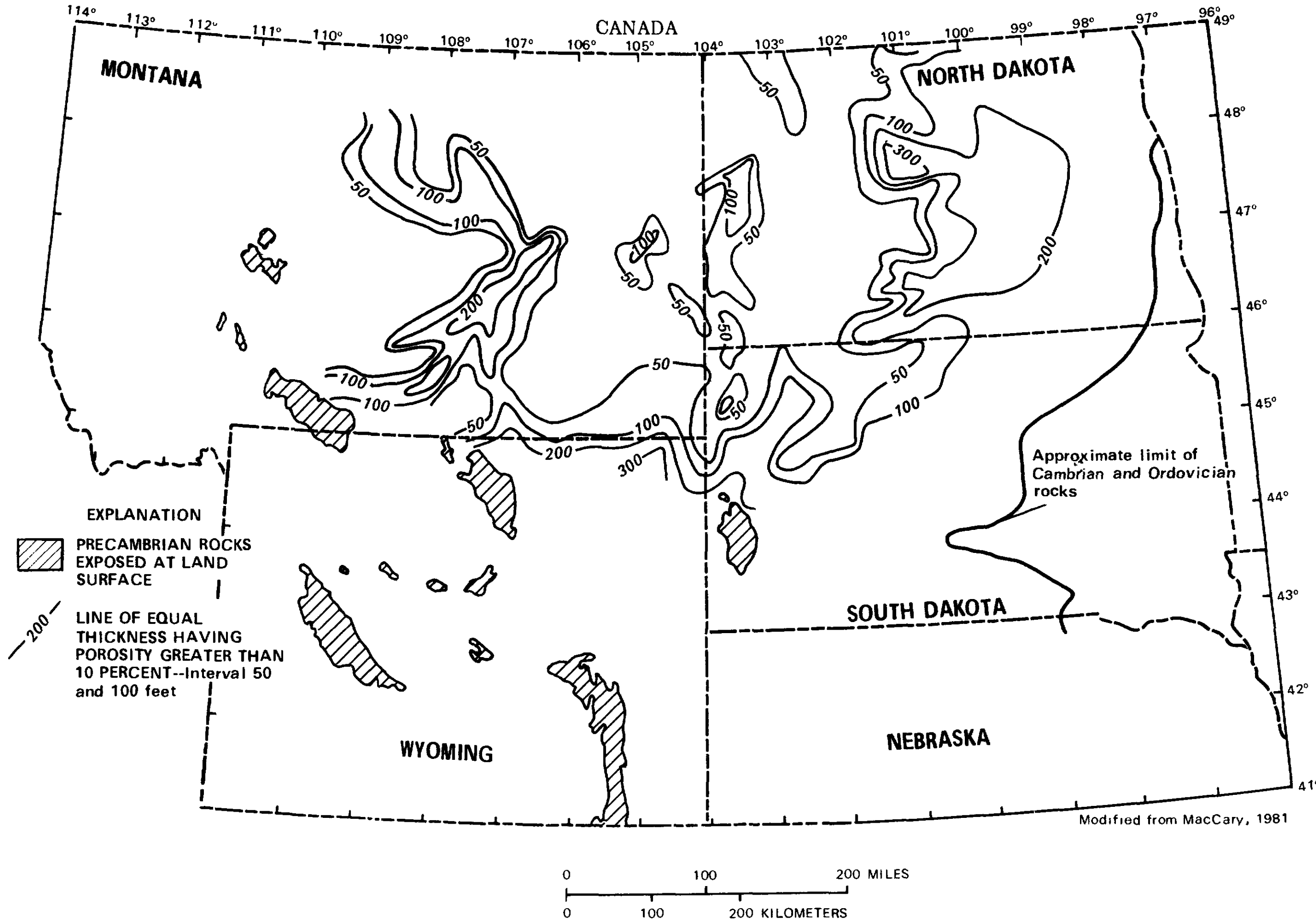

FIGURE 34.—Porosity distribution in the Cambrian-Ordovician aquifer.

areas where precipitation occurs. Finally, with regard to the second hypothesis, it should be noted that even though some flow exists, the situation still seems to approximate that described in the first hypothesis (to some extent). Velocities of flow are very low relative to those elsewhere in the system, and most of the flow of fresher water appears to be deflected around the brine to the north or south or through the confining beds into aquifers overlying the Madison aquifer. Thus, both the hydraulics and the density distribution seem to be fairly close to what would be observed in a system of totally static brine conforming to Hubbert's (1969) analysis.

The third hypothesis regarding the brine is that it is in motion, but that its movement represents an attempt of the system to adjust to changes in recharge and discharge with the end of Pleistocene glaciation. These changes associated with glaciation are discussed further in subsequent paragraphs. Here we simply note that if the brine were in a static configuration, of the type described by Hubbert, during Pleistocene glaciation, that configuration could not be at equilibrium with the new boundary conditions imposed with the retreat of the ice sheets. Thus, the brine would have been set in motion at the end of glaciation, seeking a new equilibrium configuration compatible with the new recharge and discharge patterns; this readjustment could still be in progress. Such a process could be contributing to some extent to the apparent movement of the brine; however, velocities computed from simulation results appear to conform more to the interpretation of a simple flow across the basin than to the interpretation of adjustment to changes.

In summary, the second hypothesis, that the brine represents a sluggish segment of the regional flow pattern across the basin, seems to agree best with simulation results and with observed field data. Origin of the brine appears to have been the solution of halite; as the density of the brine has increased, undoubtedly it has had an increasing influnce on the flow pattern, causing fresher water to divert around it to the north and south or above it into formations overlying the Madison aquifer. Although the second hypothesis thus seems most acceptable, elements of the other hypotheses also are probably reflected in the field situation. The present configuration of the brine on the outflow side of the basin, and its generally low velocity, approximate the static brine situation described by Hubbert, even though the brine is

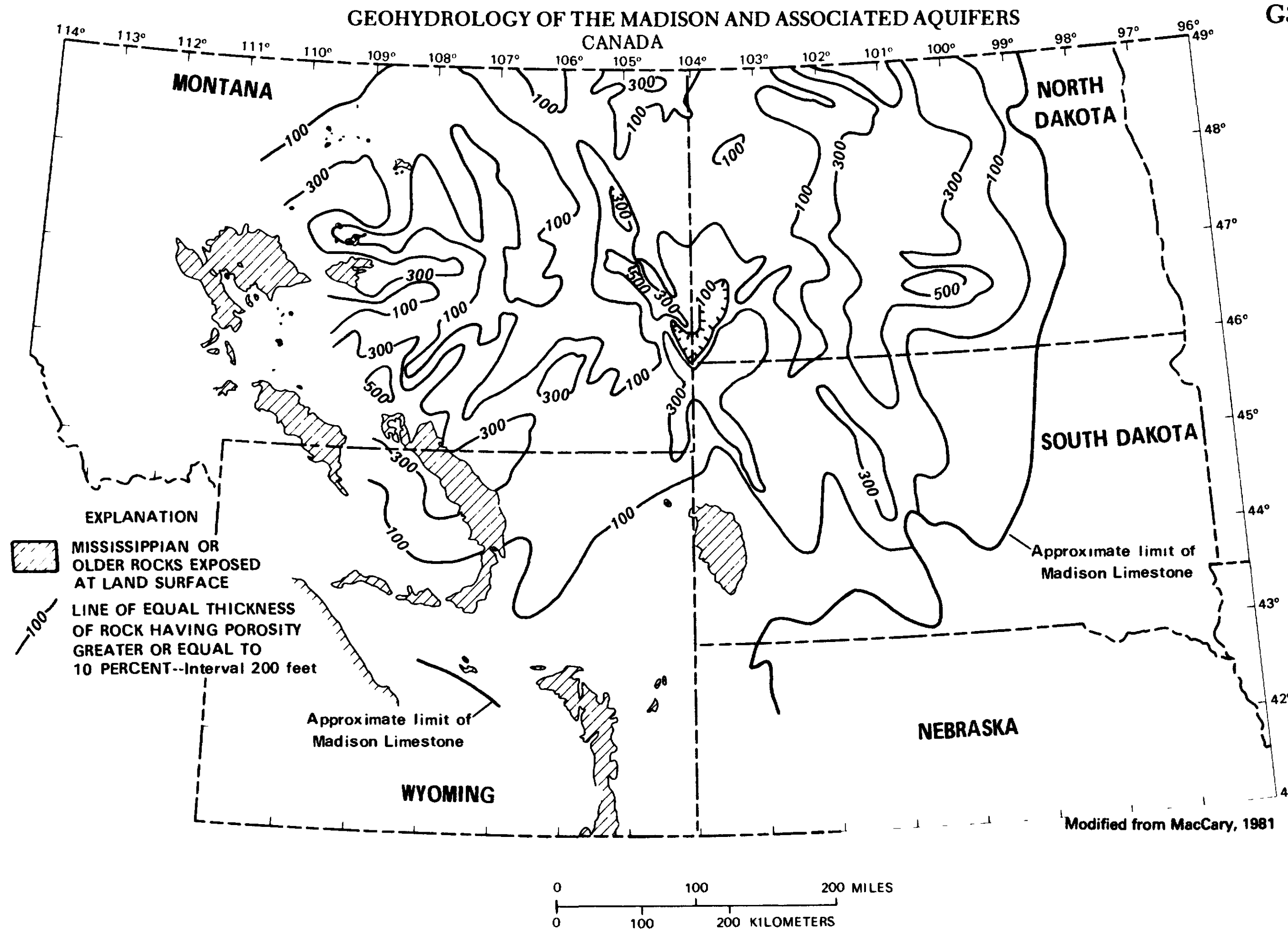

FIGURE 35. — Porosity distribution in the Madison aquifer.

not totally static; distribution of the saline water may still be shifting in response to changes in recharge and discharge at the end of the Pleistocene.

Saline water also is found in other parts of the project area — for example, in deeper parts of the Powder River basin — although not at the concentrations encountered in the Williston basin. Similar processes presumably control its distribution and movement.

The Cambrian-Ordovician aquifer extends farther to the north and east than the Madison does, and it crops out in the Province of Manitoba, Canada. The aquifer apparently discharges to a number of saline lakes in eastern North Dakota and through saline springs and seeps in Canada.

It already has been noted that recharge and discharge patterns may have been significantly different during glacial periods in the Pleistocene than at present. Such variation would have a number of implications in terms of the regional flow pattern. The existence of water at the base of continental ice sheets has been suggested by many studies (Robin, 1955; Gow and others, 1960; McGinnis, 1968; Weertman, 1972). The water would have resulted from melting at the base of ice due to geothermal and frictional heat. McGinnis (1968) estimated that the heat available to a temperate ice sheet from these sources could produce about 0.32 ft^3 of meltwater per square foot of surface area of the ice sheet per year. This water might, at least in some areas, be under high pressure from the weight of overlying ice; in any case, widespread areal recharge to the ground-water system must have occurred in the areas covered by the continental ice sheets. The configuration of ice over the Northern Great Plains varied continuously during the Pleistocene (Flint, 1971), and the resulting distribution of recharge similarly would have varied continuously. High regions to the west that form the present recharge areas also would presumably have been recharge areas prior to glaciation and also during both glacial and interglacial periods of the Pleistocene. However, during periods of strong gacial advance, recharge also would have occurred over the northern and eastern parts of the study area. The regional flow of ground water under these conditions must have been to the south, toward the only available ground-water discharge areas. With successive retreats and advances of the continental ice, directions of regional ground-water flow would have varied accordingly, generating a complex history of fluid movement. However, solution of halite in the deeper parts of the

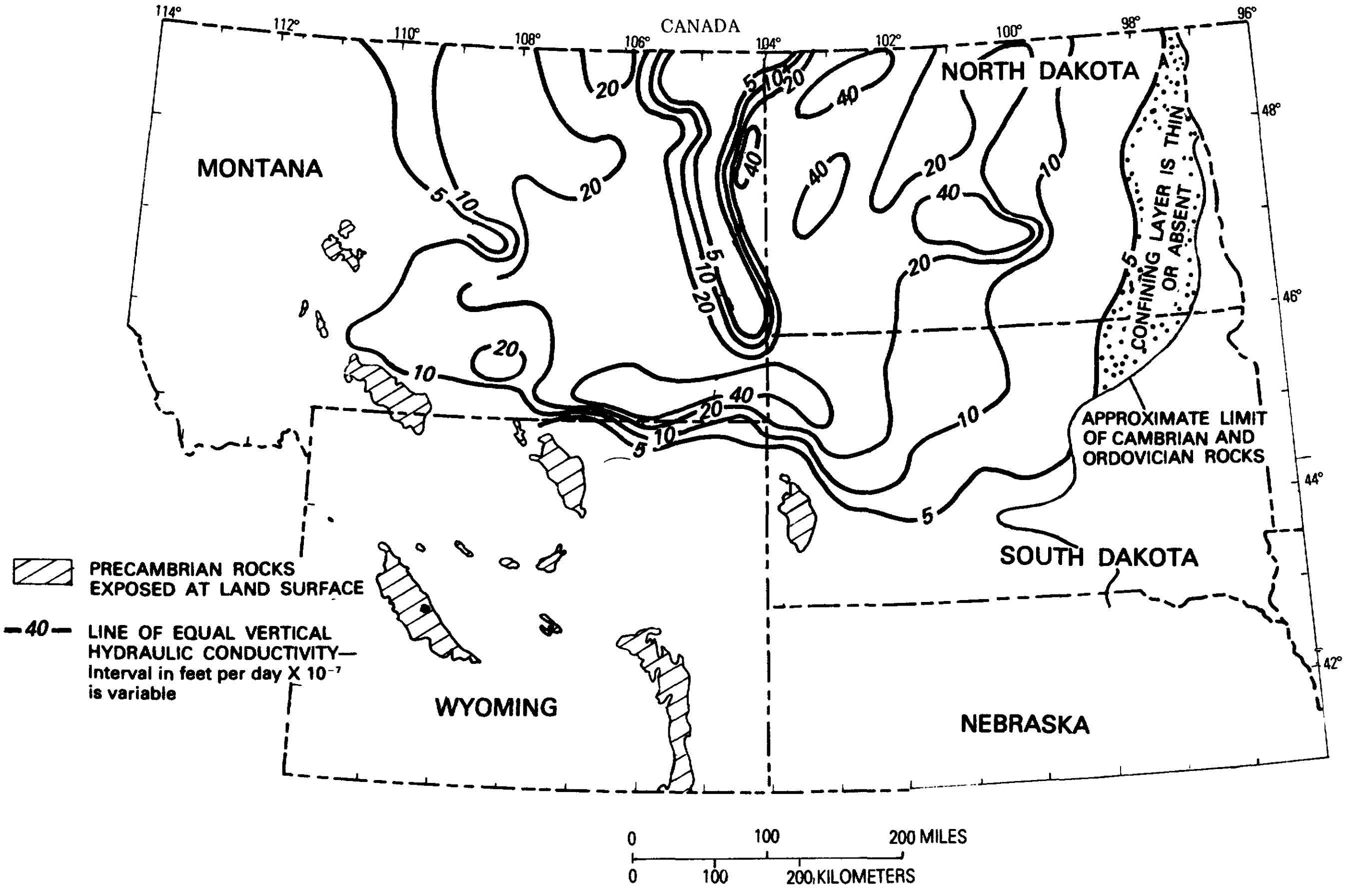

Figure 36. —Simulated vertical hydraulic conductivity of confining layer 1 overlying the Cambrian-Ordovician aquifer.

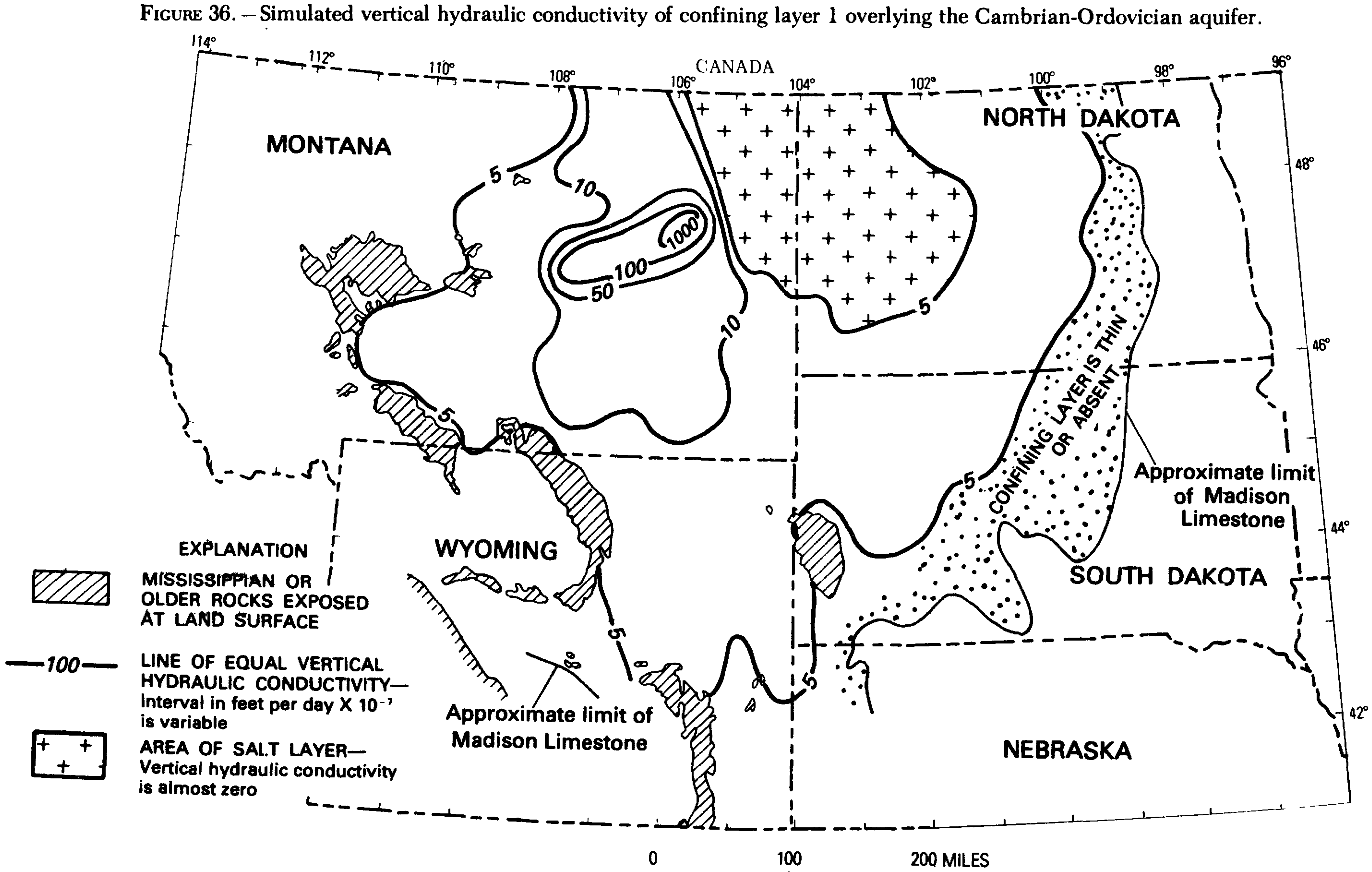

Figure 37. —Simulated vertical hydraulic conductivity of confining layer 2 overlying the Madison aquifer.

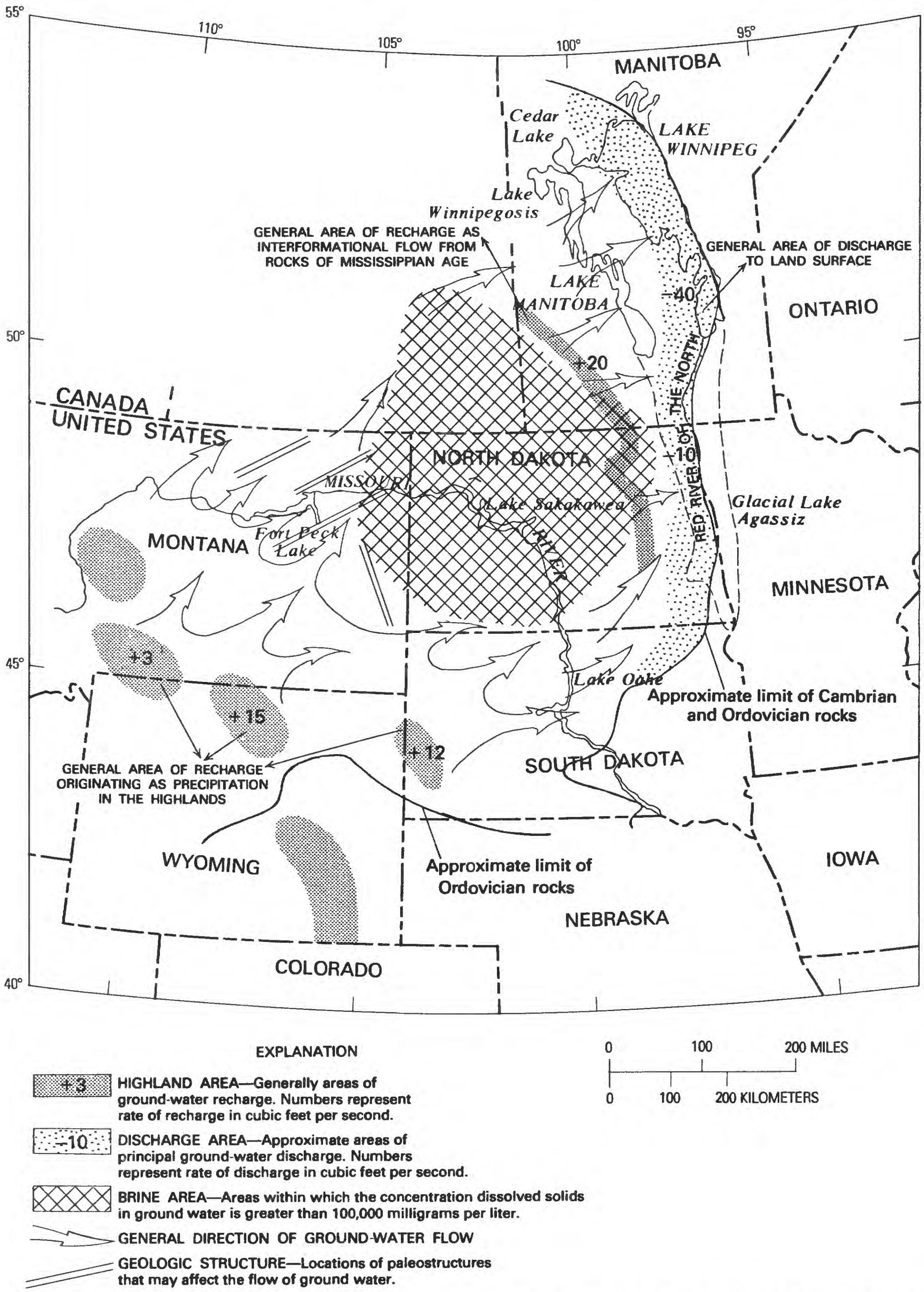

FIGURE 38.—Predevelopment flow directions, recharge areas, and discharge areas in the Cambrian-Ordovician aquifer (AQ1) as determined through simulation.

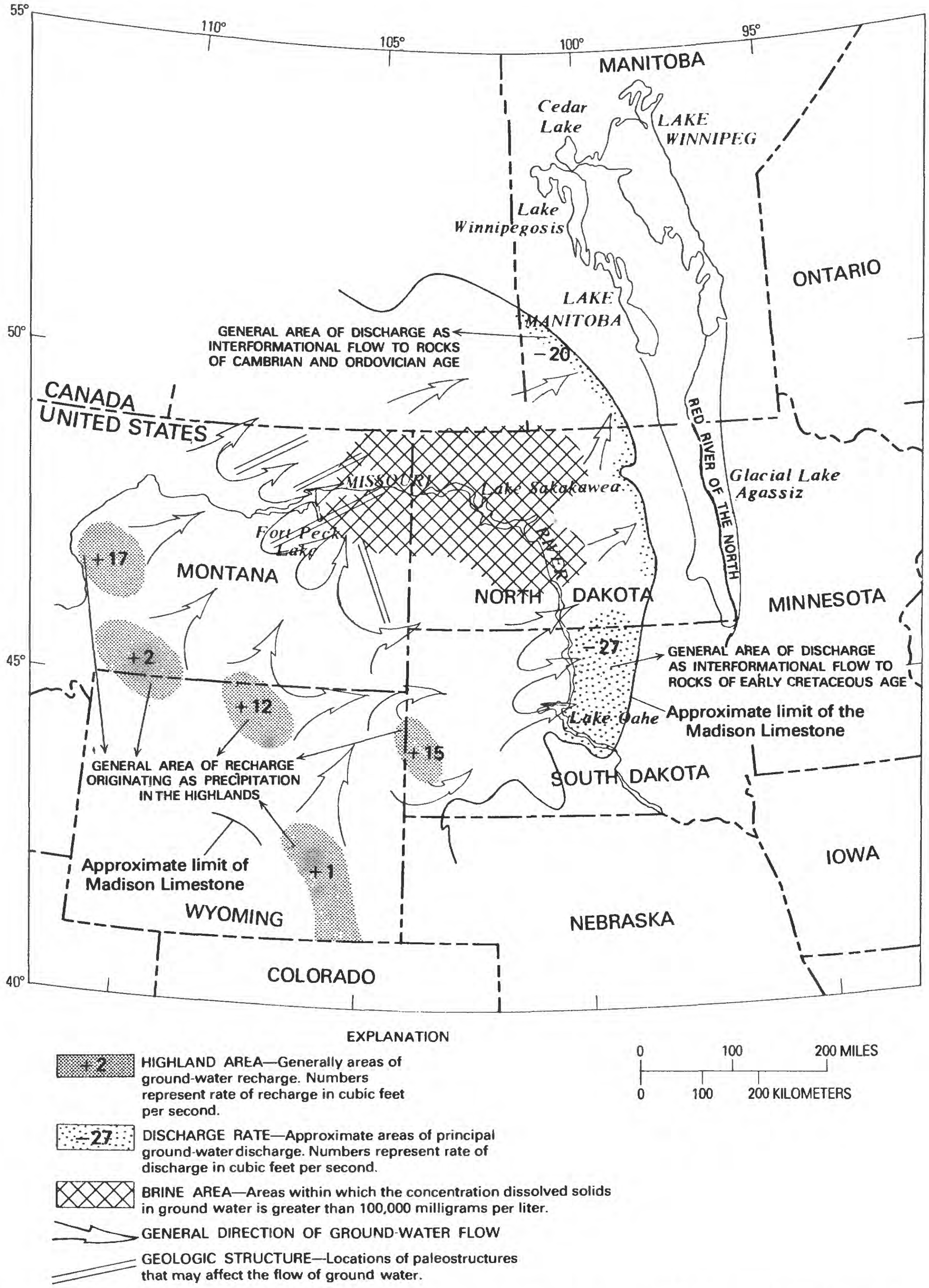

FIGURE 39.—Predevelopment flow directions, recharge areas, and discharge areas in the Madison aquifer (AQ2) as determined through simulation.

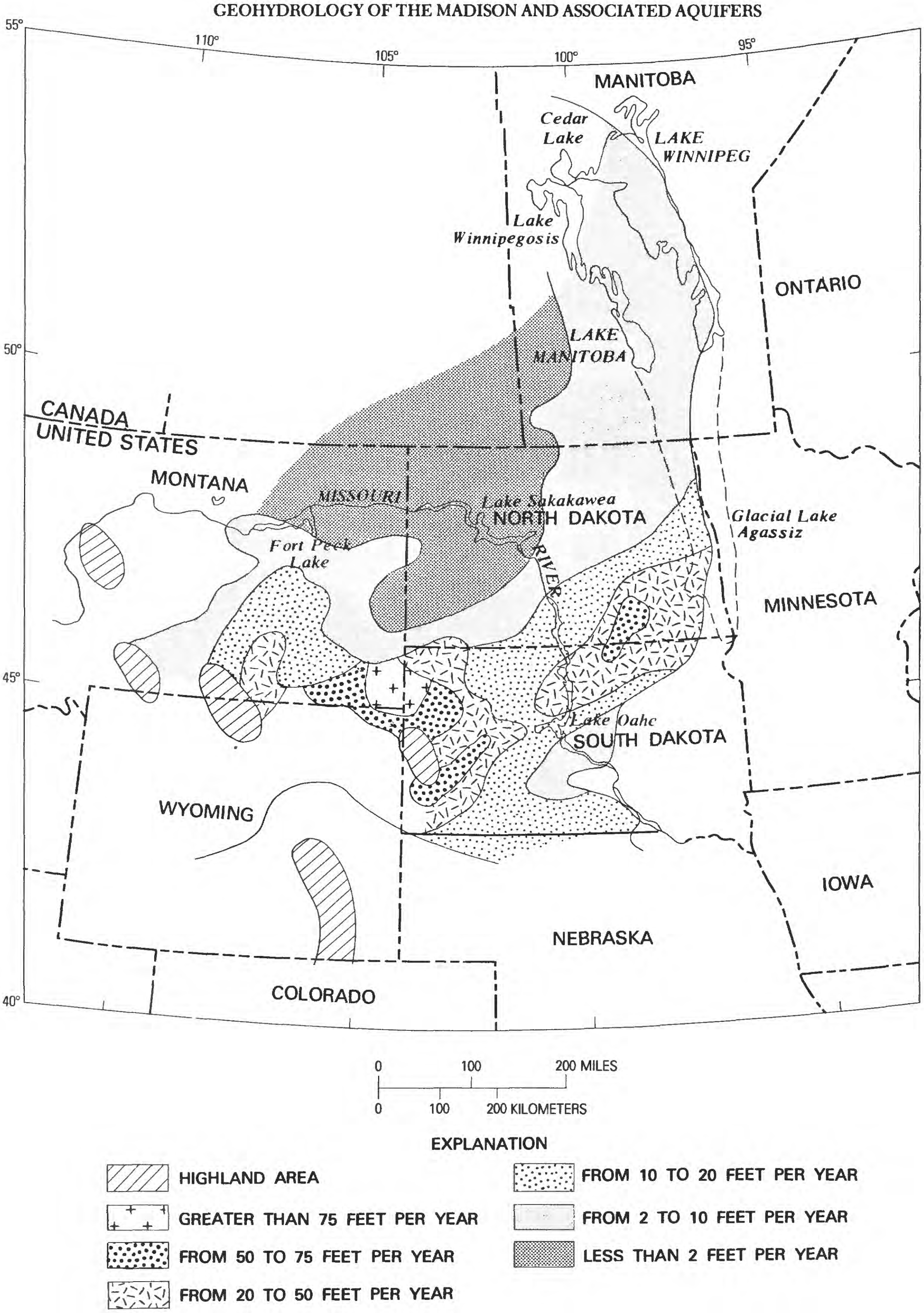

FIGURE 40. — Rates of movement of ground water in the Cambrian-Ordovician aquifer (AQ1).

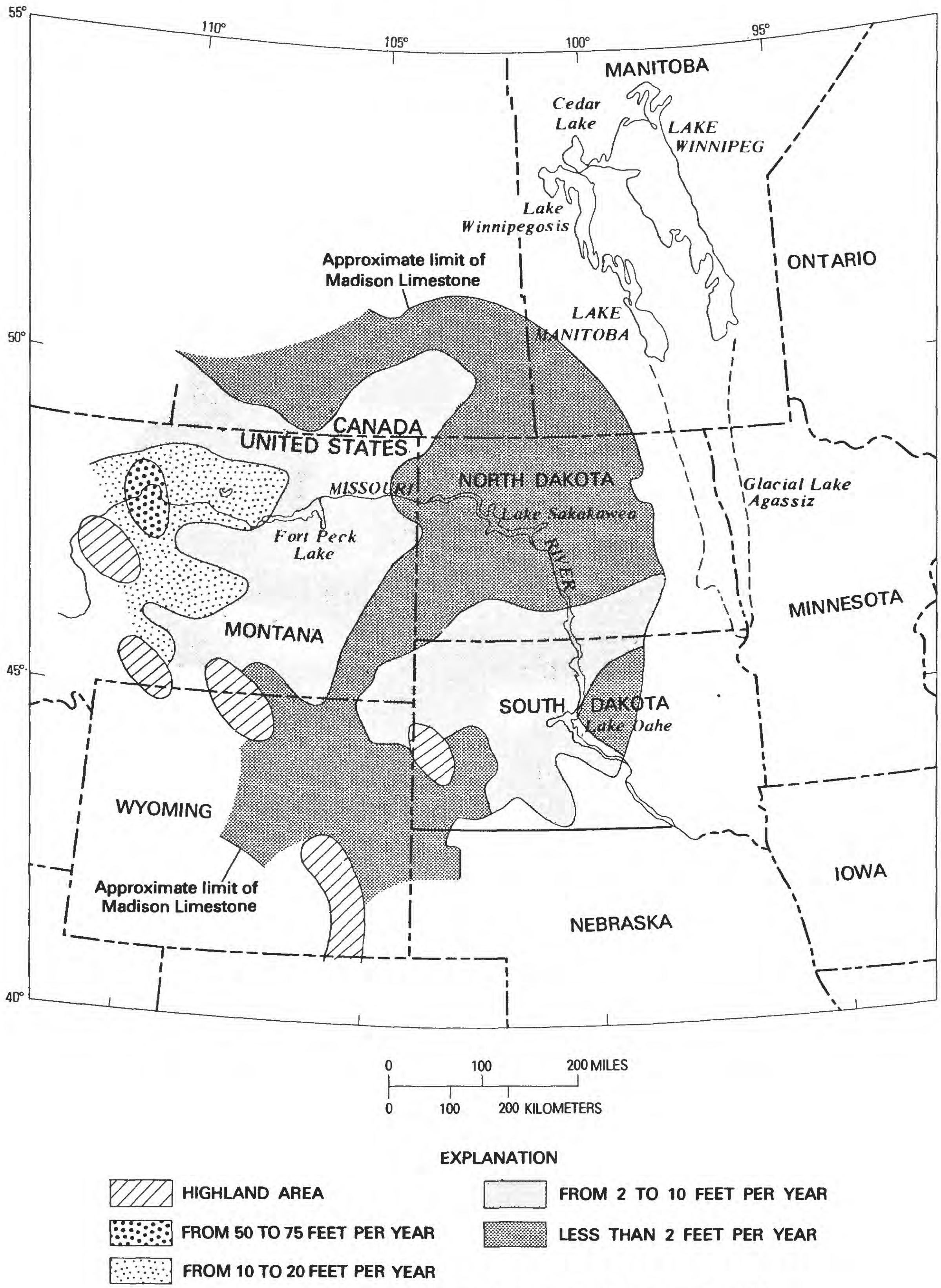

FIGURE 41. —Rates of movement of ground water in the Madison aquifer (AQ2).

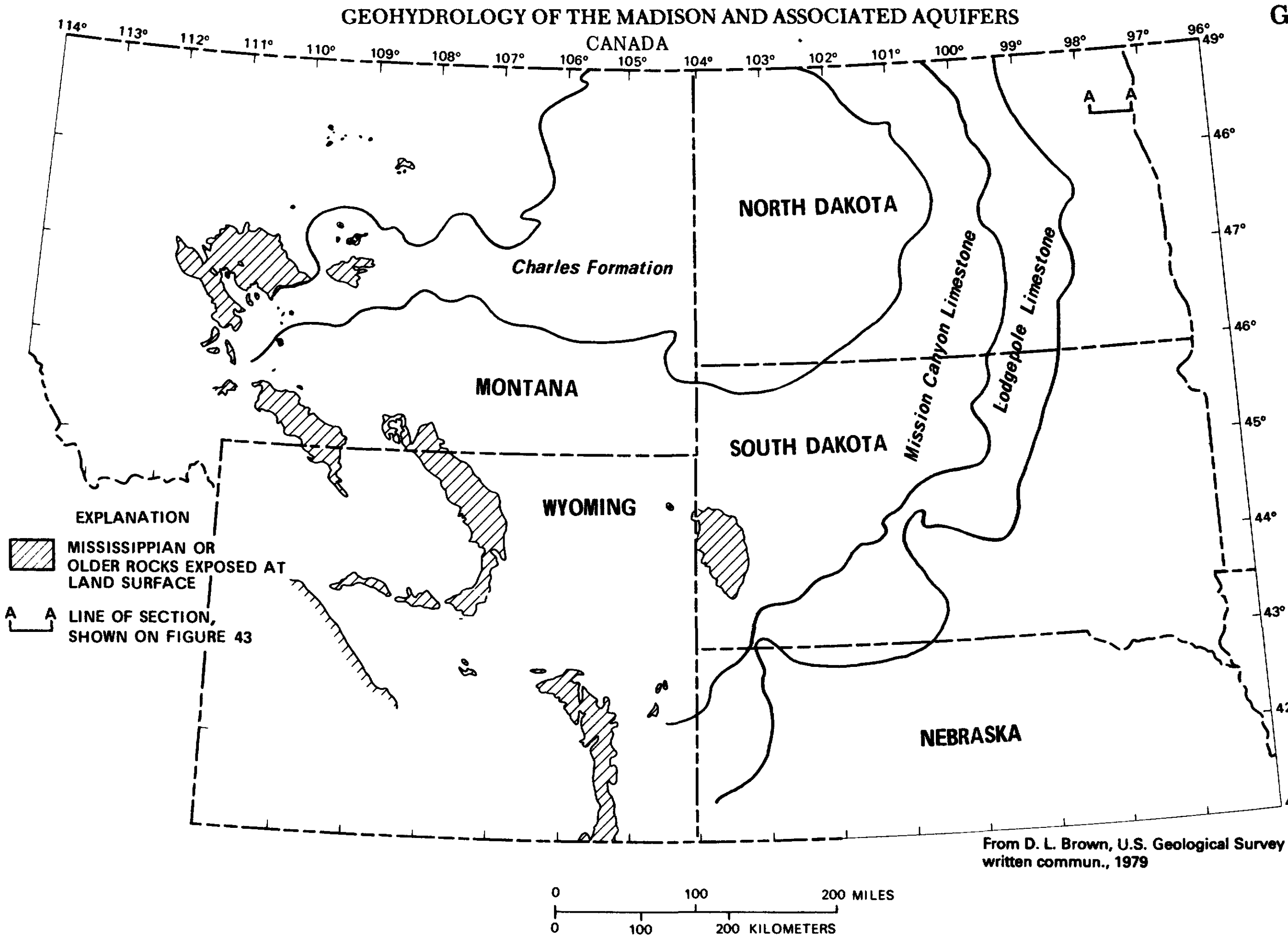

FIGURE 42. — Subcrop limits of the various formations of the Madison aquifer.

Williston basin must have occurred throughout the history of the flow system, and velocities of movement in the areas of brine accumulation must have remained small regardless of the direction of flow. Trial calculations show that repeated changes in flow direction probably would not have resulted in major displacement of the saline water but, rather, in relatively minor shifting of its position. Thus, in terms of the deep brines, the glacial changes probably represented transient disturbances but not major redistributions.

In summary, the flow pattern prior to glaciation presumably as similar to that at present, sustained by recharge in high areas to the west. Glaciation produced repeatedly varying patterns of flow, but, in general, it does not seem that the altered patterns prevailed over long enough periods to cause major changes in distribution of high-salinity brines.

A generalized hydrogeologic cross section across eastern Walsh County, N. Dak., is shown in figure 43. The location is near the eastern terminus of the Cambrian-Ordovician aquifer, and figure 43 illustrates the general relationship of the Cambrian-Ordovician aquifer to shallow ground-water systems and surface-water bodies in this area. As noted previously, part of the discharge from the Cambrian-Ordovician aquifer is apparently to a number of saline lakes in eastern North Dakota. These lakes lie in the eastern discharge area of the Cambrian-Ordovician aquifer (fig. 43) and are associated with depressions that overlie deposits of fine sand and gravel. Origin of these lake depressions has been attributed to artesian water discharging from deep, regional flow systems (Laird, 1944). A supporting hypothesis (Downey, 1969, p. 12) relates lake-depression formation to alternations of recharge and discharge associated with Pleistocene glaciation of the area. During glacial advances, meltwater at the base of the ice sheet (Gow and others, 1968; McGinnis, 1968) presumably was forced into the bedrock aquifers by hydrostatic pressure. Upon deglaciation, hydrostatic pressure would have been released, allowing large quantities of water to move rapidly out of the bedrock. This rapid movement of water through overlying materials may have resulted in erosion of the lake sediment, forming the depressions in which the lakes exist today.

The fact that these lakes now function as drains for regional ground-water flow is supported by both geologic evidence and water chemistry. Test drilling indicates that fairly thick deposits of glacial sand and gravel

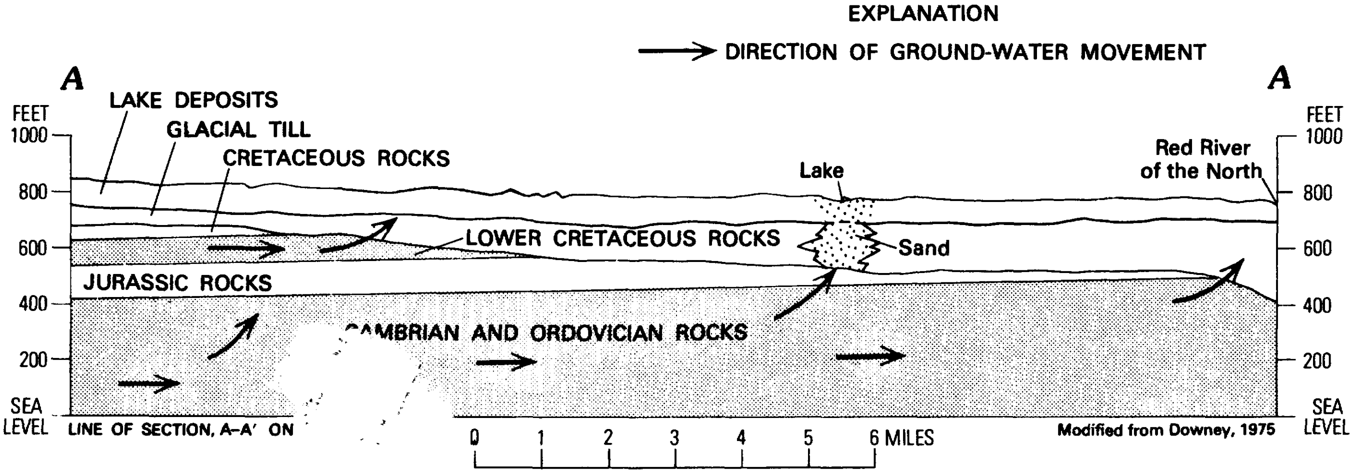

FIGURE 43.—Generalized hyd- section across eastern Walsh County, N. Dak., showing discharge area of the Cambrian-Ordovician aquifer.

underlie the depressions a... ar raulically connected with underlying bedrock (Dov ey, 1973). Chemical analyses of water samples taken from the test holes and lakes (Downey, 1971) indicate a chemical similarity to water from the Cambrian-Ordovician aquifer. These analyses indicate that water is able to move upward from the bedrock aquifers through the glacial sand and gravel deposits to discharge points at the bottom of the lakes.

Vertical leakage through confining beds is an integral feature of the flow pattern in the bedrock-aquifer systems in the Northern Great Plains. Leakage between aquifers may be detected by geochemical methods, although the lack of geochemical data for many areas limits the application of this method. An area of significant leakage inferred on the basis of similarities in water chemistry is shown in figure 44 (J. F. Busby, U.S. Geological Survey, written commun., 1982). Leakage occurs both through the confining-bed matrix and along fractures in the confining bed associated with lineament zones (Weimer and others, 1982). Confining beds are not present everywhere, having been removed by erosion or, in some areas, never having been deposited. In those areas where the confining bed is absent, extensive hydraulic connection exists between aquifers, and leakage tends to be high; where the confining bed is thick and unfractured, leakage is minimal. Leakage along fractures is dependent on the degree of fracturing and interconnection between fractures. Work done in connection with the flow of oil in petroleum reservoirs (Chilingar and others, 1972) demonstrates that a fracture with a width of about 0.02 in. could account for more than 90 percent of the total fluid flow through a dense rock, such as limestone. Thick halite units, such as those in the Charles Formation, are considered in this study to be impermeable to the vertical flow of water.

SUMMARY

Rocks of Paleozoic age have been grouped for purposes of this study into two major aquifers underlying the Northern Great Plains region—the Cambrian-Ordovician aquifer and the Madison aquifer—and into semiconfining units above each of these aquifers. These units are actually part of an integrated system of aquifers and aquitards that also includes a number of units overyling the Madison aquifer; however, the emphasis in this report is on the Cambrian-Ordovician and Madison aquifers. These aquifers crop out and receive recharge in highland areas in the eastern parts of Montana and Wyoming; recharge also occurs in the aquifer outcrops in the Black Hills uplift area of South Dakota. From these recharge areas, water generally moves eastward and northeastward to discharge areas located in the Lake Agassiz basin in northeastern North Dakota and in the Canadian Provinces of Manitoba and southern Saskatchewan. Water also is recharged or discharged by leakage from the aquifers to adjoining units throughout the entire area; leakage is greatest at locations where confining beds are traversed by major fault systems or where confining beds are thin or absent.

Conceptually, the hydrologic system considered in this study incorporates Precambrian rocks as a lower boundary; rocks of Cambrian and Ordovician age are grouped into the lower major aquifer, the Cambrian-Ordovician aquifer; the principal formation of this group is the Red River Formation. Rocks of Silurian and Devonian age and the Mississippian part of the Bakken and Englewood Formations, which overlie the Cambrian-Ordovician aquifer, are considered to act as confining units above this lower system. The second major aquifer, the Madison aquifer, consists of the Madison Limestone or

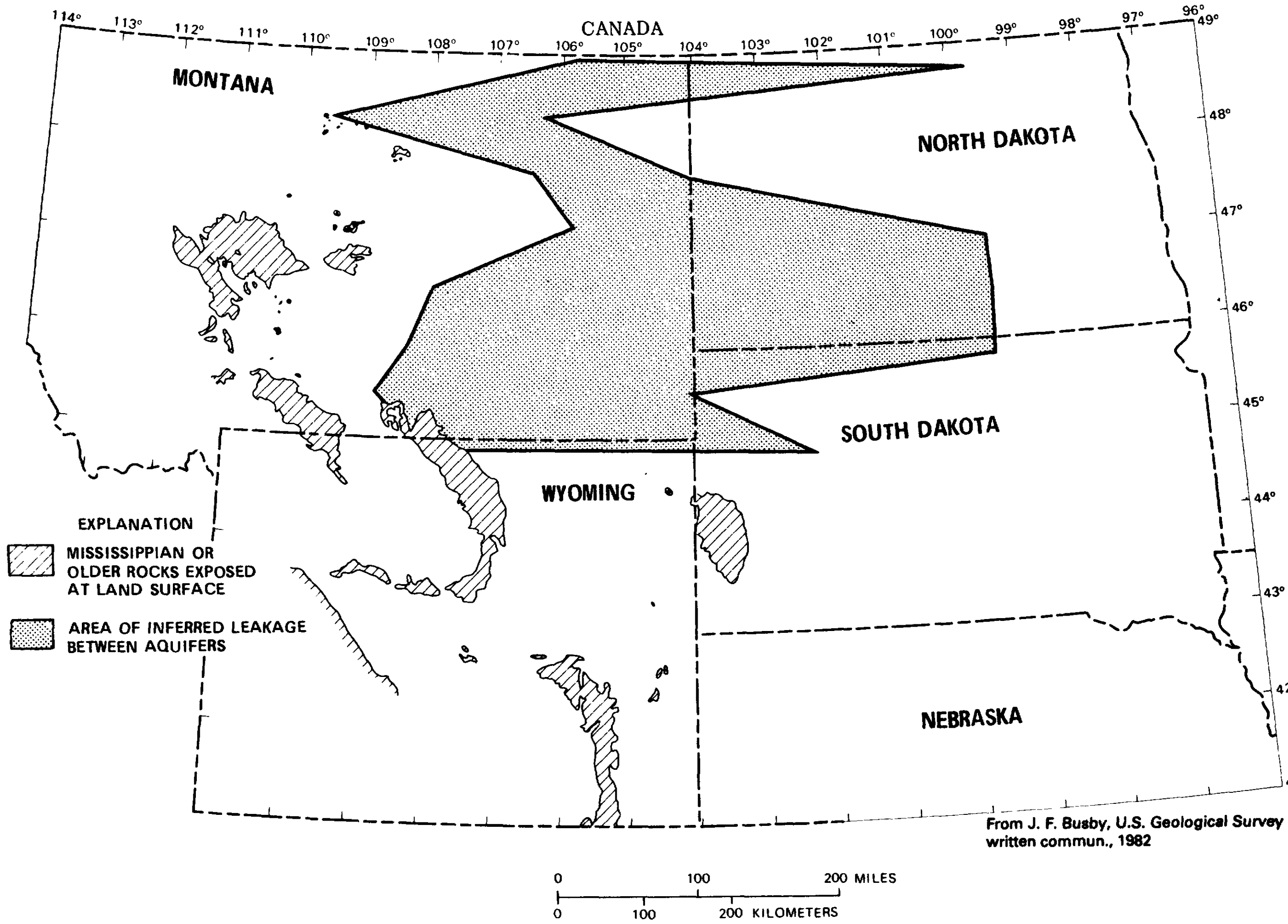

FIGURE 44.—Area of inferred leakage from the Cambrian-Ordovician aquifer to the Madison aquifer based on geochemical data.

the Lodgepole and Mission Canyon Limestones of the Madison Group of Mississippian age. These formations are composed mainly of limestone or dolomite, but also contain minor anhydrite or gypsum beds. The Mississippian Charles Formation, the uppermost part of the Madison Group, consisting mainly of halite and anhydrite beds, and the Big Snowy Group have been included in the confining unit, which also includes shales and siltstone of Permian, Triassic, and Jurassic age. The upper boundary of the hydrologic system considered in this study consists of the Lower Cretaceous rocks, mainly sandstone, which directly overlie rocks of Jurassic age.

Geologic structure is an important factor in determining the rate and direction of water movement in the bedrock-aquifer system of the Northern Great Plains. The Williston basin exerts a major influence, carrying the aquifers to great depths below the surface, where high-density brines have accumulated. Smaller structures also exert an influence. For example, the Casper Mountain fault appears to prevent flow to the south from the Powder River basin, and the major fault system bounding the Bighorn Mountains on the east limits flow from the recharge area to the Powder River basin. Recharge from the Bighorn Mountains appears to be channeled by geologic structure associated with a major lineament zone; it moves northeastward across the northern part of the Powder River basin, south of the Cedar Creek anticline, to join the flow system from the Black Hills recharge area. This flow system continues around the southern part of the Williston basin and then turns northeastward to discharge in northeastern North Dakota and eastern Manitoba. The Weldon-Brockton fault zone appears to be a major channel for water movement from the Big Snowy Mountains and associated highland areas in Montana to discharge areas in Canada north of the Williston basin.

Vertical leakage across confining units is an integral feature of the regional flow system. Vertical leakage occurs both through the rock matrix of the confining beds and along fractures associated with major lineament zones. In the western recharge area, leakage tends to be downward, In the areas of discharge, to the east, it tends to be upward.

Water quality and temperature are major considerations in the analysis of hydrologic systems existing in the Paleozoic aquifers of the Northern Great Plains.

Dissolved-solids concentrations in water from both aquifers range from less than 1,000 to greater than 300,000 mg/L; temperatures range from about 8° to greater than 300°C. Water temperature has a significant effect on viscosity of water; temperature changes alone result in large changes in transmissivity along a flow path. Both dissolved-solids content and temperature influence density, which, in turn, has a major influence on magnitude and direction of flow.

The solution of halite and other minerals in the water moving through the system has resulted in the accumulation of very dense brine in the deeper parts of the Williston basin. Because of the nature of the flow system in the Cambrian-Ordovician and Madison aquifers, the brine has been forced up the east limb of the Williston basin toward the eastern discharge area. The major part of the freshwater flow through the basin is deflected to the north and south around the brine or through the confining beds into aquifers overlying the Madison aquifer.

On the basis of available data, it appears that the brine is moving very slowly eastward and northeastward toward discharge areas in northeastern North Dakota and in Manitoba. As it moves updip on the eastern flank of the Williston basin into regions of cooler formation temperature, precipitation of halite may be occurring, causing a very gradual deterioration in permeability over periods of geologic time.

Prior to Pleistocene glaciation, the flow system probably was similar to that which exists at present, with recharge in high areas to the west and discharge to the east and northeast of the Williston basin. Glaciation caused repeated changes in recharge and discharge patterns, with corresponding alterations in flow direction. However, the period of these changes, relative to the residence time of fluid in the deeper parts of the basin, probably was short, so that major changes in the distribution of the deep brines probably has not occurred.

SELECTED REFERENCES

Agatston, R. S., 1954, Pennsylvanian and Lower Permian of northern and eastern Wyoming: Bulletin of the American Association of Petroleum Geologists, v. 38, no. 4, p. 508–583.

Agnew, A. F., and Tychsen, . C., 1965, A guide to the stratigraphy of South Dakota: South Dakota State Geological Survey Bulletin 14, 195 p.

Andrichuck, J. M., 1955, Mississippian Madison Group stratigraphy and sedimentation in Wyoming and southern Montana: Bulletin of the American Association of Petroleum Geologists, v. 39, no. 11, p. 2170–2210.

Armstrong, C. A., 1980, Ground-water resources of Dickey and LaMoure Counties, North Dakota: North Dakota State Water Commission County Ground Water Studies no. 28, p. III, 61 p.

Ballard, F. V., 1963, Structural and stratigraphic relationships in Paleozoic rocks of eastern North Dakota: North Dakota Geological Survey Bulletin 40, 42 p.

Ballard, W. W., 1969, Red River of northwest Montana and northwest North Dakota: Montana Geological Society Annual Conference, 20th [Eastern Montana Symposium], Billings, Mont. October 1969, Proceedings, p. 15–24.

Bedinger, M. S., Red, J. E., and Griffin, J. D., 1973, Digital-computer programs for analysis of ground-water flow: U.S. Geological Survey open-file report, 85 p.

Berry, F. A. F., and Hanshaw, B. B., 1960, Geologic field evidence suggesting membrane properties of shales: International Geological Congress, 21st, Copenhagen, 1960, Proceedings [abs.], p. 209.

Blankennagel, R. K., Howells, L. W., and Miller, W. R., 1981, Completion and testing of Madison Limestone test well 3, NW¼SE¼ sec. 35, T. 2 N., R. 27 E., Yellowstone County, Montana: U.S. Geological Survey Open-File Report 81–528, 91 p.

Blankennagel, R. K., Howells, L. W., Miller, W. R., and Hansen, C. V., 1979, Preliminary data for Madison Limestone test well 3, NW¼SE¼ sec. 35, T. 2 N., R. 27 E., Yellowstone County, Montana: U.S. Geological Survey Open-File Report 79–745, 201 p.

Blankennagel, R. K., Miller, W. R., Brown, D. L., and Cushing, E. M., 1977, Report on preliminary data for Madison Limestone test well 1, NE¼SE¼ section 15, T. 57 N., R. 65 W., Crook County, Wyoming: U.S. Geological Survey Open-File Report 77–164, 110 p.

Bredehoeft, J. D., and Hanshaw, B. B., 1968, On the maintenance of anomalous fluid pressures—I, Thick sedimentary sequences: Geological Society of America Bulletin, v. 79, no. 9, p. 1097–1106.

Brown, D. L., 1978, Wrench-style deformational patterns associated with a meridional stress axis recognized in Paleozoic rocks in parts of Montana, South Dakota, and Wyoming: Montana Geological Society Annual Conference, 24th [Williston Basin Symposium], Billings, Mont., September 1978, Guidebook, p. 17–31.

Brown, D. L., Blankennagel, R. K., Busby, J. F., and Lee, R. W., 1977, Preliminary data for Madison Limestone test well 2, SE¼SE¼ sec. 18, T. 1 N., R. 54 E., Custer County, Montana: U.S. Geological Survey Open-File Report 77–863, 135 p.

Carlson, C. G., and Anderson, S. B., 1960, Stratigraphy of the Winnipeg and Deadwood Formations in North Dakota: North Dakota Geological Survey Bulletin 35, 149 p.

_____1965, Sedimentary and tectonic history of North Dakota part of Wiliston basin: Bulletin of the American Association of Petroleum Geologists, v. 49, no. 11, p. 1833–1846.

Carlson, C. G., Bakken, W. E., and Kume, J. B., 1960, Subsurface geology and development of petroleum in North Dakota: North Dakota Geological Survey Bulletin 34, 20 p.

Chilingar, G. V., Mannon, R. W., and Rieke, H. H., III, 1972 Oil and gas production from carbonate rocks: New York, American Elsevier Scientific Publishing Company, 408 p.

Darton, N. H., 1896, Preliminary report on artesian waters of a portion of the Dakotas: U.S. Geological Survey 17th Annual Report, p. II, p. 603–694.

DeMille, G., Shouldice, J. R., and Nelson, H.W., 1964, Collapse structures related to evaporites of the Prairie Formation, Saskatchewan: Geological Society of America Bulletin, v. 75, no. 4, p. 307–316.

Downey, J. S., 1969, *in* Pederson, D. T., and Reid, J. R., eds., Geology of northeastern North Dakota: North Dakota Geological Survey Miscellaneous Series 39, p. 12.

_____1971, Ground-water resources of Walsh County, northeastern North Dakota: U.S. Geological Survey Hydrologic Investigations Atlas HA–431.

_____1973, Ground-water resources, Nelson and Walsh Counties, North Dakota: North Dakota Geological Survey Bulletin 57, p. III, and North Dakota State Water Commission County Ground Water Studies 17, p. III, 67 p.

Earlougher, R. C., 1977, Advances in well test analysis: Dallas Tex., Society of the Petroleum Engineers of the American Institute of Mining, Metallurgical, and Petroleum Engineers, p. 172.

Elson, J. A., 1967, Geology of glacial Lake Agassiz, *in* Mayer-Oakes, W. J., ed., Life, land and water—Conference on Environmental Studies of the Glacial Lake Region, Winnipeg, 1966, Proceedings: Winnipeg, Manitoba, Canada, University of Manitoba Press, p. 414.

Flint, R. F., 1971, Glacial and Quaternary geology: New York, John Wiley, 892 p.

Foster, N. H., 1972, Ordovician system, *in* Geologic atlas of the Rocky Mountain region: Denver, Rocky Mountain Association of Geologists, p. 76–85.

Fuller, J. G. C. M., 1956, Mississippian rocks and oil-fields in southeastern Saskatchewan: Province of Saskatchewan, Department of Mineral Resources, 72 p.

Gow, A. J., Veda, H. T., and Garfield, D. E., 1968, Antarctic ice sheet—Preliminary results of first core hole to bedrock: Science, v. 161, no. 3845, p. 11011–1013.

Grossman, I. G., 1968, Origin of the sodium sulfate deposits of the northern Great Plains of Canada and the United States, *in* Geological Survey Research 1968: U.S. Geological Survey Professional Paper 600–B, p. B104–B109.

Hanshaw, B. B., and Bredehoeft, J. D., 1968, On the maintenance of anomalous fluid pressures—II, Source layer at depth: Geological Society of America Bulletin, v. 79, no. 9. p. 1107–1122.

Head, W. J., Kilty, K. T., and Knottek, R. K., 1978, Maps showing formation temperatures and configurations of the Minnelusa Formation and the Madison Limestone, Powder River basin, Wyoming, Montana, and adjacent areas: U.S. Geological Survey Open-File Report 78–905, scale 1:1,000,000, 2 sheets.

Hitchon, Brian, 1969, Fluid flow in the western Canada sedimentary basin—2, Effect of geology: Water Resources Research, v. 5, no. 2, p. 469–480.

Hodson, W. G., 1974, Records of water wells, springs, oil- and gas-test holes, and chemical analyses of water for the Madison Limestone and equivalent rocks in the Powder River basin and adjacent areas, northeastern Wyoming: Wyoming State Engineer Report, 27 p.

Hopkins, W. B., 1978, Deep aquifers of eastern Montana Montana Geological Society Annual Conference, 24th [Williston Basin Symposium], Billings, Mont., September 1978, Guidebook, p. 392–407.

Hubbert, M. K., 1969, The theory of ground-water motion and related papers: New York, Hafner, 310 p.

Irwin, M. L., 1965, General theory of epeiric clear water sedimentation: Bulletin of the American Association of Petroleum Geologists, v. 49, no. 4, p. 445-459.

Kanasewich, E. R., 1968, Precambrian rift—Genesis of strata-bound ore deposits: Science, v. 161, p. 1002–1005.

Keefer, W. R., 1970, Structural geology of the Wind River Basin, Wyo.: U. S. Geological Survey Professional Paper 495–D, p. D1–D35.

Keefer, W. R., and Van Lieu, J. A., 1966, Paleozoic formations in the Wind River Basin, Wyo.: U. S. Geological Survey Professional Paper 495–B, p. B1–B60.

Keene, J. R., 1973, Ground-water resources of the western half of Fall River County, South Dakota: South Dakota State Geological Survey Report of Investigations 109, 82 p.

Konikow, L. F., 1976, Prelimnary digital model of ground-water flow in the Madison Group, Powder River basin and adjacent areas, Wyoming, Montana, South Dakota, North Dakota, and Nebraska: U.S. Geological Survey Water-Resources Investigations 63–75, 44 p.

Laird, W. M., 1944, The geology and ground water resources of the Emerado quadrangle: North Dakota Geological Survey Bulletin 17, 35 p.

Landes, K. K., 1970, Petroleum geology of the United States: New York John Wiley, 571 p.

Lohman, S. W., and others, 1972, Definitions of selected ground-water terms—Revisions and conceptual refinements: U.S. Geological Survey Water-Supply Paper 1988, 21 p.

MacCary, L. M., 1981, Apparent water resistivity, porosity, and ground-water temperature of the Madison Limestone and underlying rocks in parts of Montana, Nebraska, North Dakota, South Dakota, and Wyoming: U.S. Geological Survey Open-File Report 81–629, 43 p.

MacCary, L. M., Cushing, E. M., and Brown, D. L., 1981, Potentially favorable areas for large-yield wells in the Red River Formation and Madison Limestone in parts of Montana, North Dakota, South Dakota, Wyoming, and Nebraska: U.S. Geological Survey Open-File Report 81–220, 35 p.

McCabe, H. R., 1963, Mississippian oil fields of southwestern Manitoba: Province of Manitoba, Department of Mines and Natural Resources, 50 p.

McGinnis, L. D., 1968, Glaciation as a possible cause of mineral deposition: Economic Geology, v. 63, p. 390–400.

Maughan. E. K., 1966, Environment of deposition of Permian salt in Williston and Alliance basins, *in* Rau, J. L., ed., Symposium on Salt, Geology, Geochemistry, Mining, 2d, Cleveland, Ohio, 1966, Proceedings: Cleveland, Ohio, Northern Ohio Geological Society, v. 1, p. 35–47.

______1967, Eastern Wyoming, eastern Montana, and the Dakotas, Chap. G, *in* McKee, E. D., and others, eds, Paleotectonic investigations of the Permian system in the United States: U.S. Geological Survey Professional Paper 515–G, p. G125–G152.

Meyboom, Peter, 1966, Groundwater studies in the Assiniboine River drainage basin: Geological Survey of Canada Bulletin 139, 65 p.

Miller, D. N., Jr., 1966, Diagenesis of sedimentary rocks: Wyoming Geological Association Annual Conference, 12th [Symposium on Recently Developed Geologic Principles and Sedimentation of the Permo-Pennsylvanian of the Rocky Mountains], Casper, Wyo., August 31–September 2, 1966, Proceedings, p. 17–21.

Miller, W. R., 1976, Water in carbonate rocks of the Madison Group in southeastern Montana—A preliminary evaluation: U.S. Geological Survey Water-Supply Paper 2043, 51 p.

Miller, W. R., and Strausz, S. A., 1980a, Preliminary map showing freshwater heads for the Red River Formation, Bighorn Dolomite, and equivalent rocks of Ordovician age in the northern Great Plains of Montana, North Dakota, South Dakota, and Wyoming: U.S. Geological Survey Open-File Map 80–730, scale 1:1,000,000, 1 sheet.

____1980b, Preliminary map showing freshwater heads for the Mission Canyon and Lodgepole Limestones and equivalent rocks of Mississippian age in the northern Great Plains of Montana, North Dakota, South Dakota, and Wyoming: U.S. Geological Survey Open-File Map 80–729, scale 1:1,000,000, 1 sheet.

Milner, R. L., 1956, Effects of salt solution in Saskatchewan, *in* International Williston Basin Symposium, 1st, Bismarck, N. Dak., October 1956, Proceedings [abs.]: Bismarck, N. Dak., North Dakota Geological Society, p. 111.

Murphy, W. C., 1965, The interpretation and calculation of formation characteristics from formation test data: Duncan, Okla., Halliburton Services Division of Halliburton Company, 19 p.

Nordquist, J. W., 1953, Mississippian stratigraphy of northern Montana: Billings Geological Society Annual Field Conference, 4th, September 1953, Guidebook, p. 68–82.

Patterson, J. R., 1961, Ordovician stratigraphy and corrlations in North America: Bulletin of the American Association of Petroleum Geologists, v. 45, no. 6, 1363–1377.

Peterson, J. A., 1957, Marine Jurassic of the northern rocky Mountains and Williston basin: Bulletin of the American Association of Petroleum Geologists, v. 41, no. 3, p. 339–440.

______1966, Sedimentary history of the Sweetgrass arch: Billings Geological Society Annual Field Conference, 17th, 'August 1966, Guidebook, p. 112–133.

______1972, Jurassic system, *in* Geologic atlas of the Rocky Mountain region: Denver, Rocky Mountain Association of Geologists, p. 177–189.

______1978, Subsurface geology and porosity distribution, Madison Limestone and underlying formations, Powder River basin, northeastern Wyoming and southeastern Montana and adjacent areas: U.S. Geological Survey Open-File Report 78-783, 32 p.

______1981, Stragigraphy and sedimentary facies of the Madison Limestone and associated rocks in parts of Montana, North Dakota, South Dakota, Wyoming, and Nebraska: U.S. Geological Survey Open-File Report 81-642, 92 p.

Pinder, G. F., and Bredehoeft, J. D., 1968, Application tion of digital computer for aquifer evaluation: Water Resources Research, v. 4, no. 5, p. 1069–1093.

Porter, J. W., and Fuller, J. G. C. M., 1964, Ordovician-Silurian, *in* Geological history of western Canada: Calgary, Alberta, Canada, Alberta Society of Petroleum Geologists, p. 34–42.

Powell, B. F., 1940, Construction history and technical details of Sheridan Dam: Black Hills Engineer, v. 6, p. 180–196.

Rahn, P. H., and Gries, J. P., 1973, Large springs in the Black Hills, South Dakota and Wyoming: South Dakota State Geological Survey Report of Investigations 107, 46 p.

Robin, G. de Q., 1955, The movement and temperature distribution on glaciers and ice sheets: Journal of Glaciology, v, 2, p. 523–532.

Roehl, P. O., 1967, Stoney Mountain (Ordovician) and Interlake (Silurian) facies analogs of recent low-energy marine and subaerial carbonates, Bahamas: Bulletin of the American Association of Petroleum Geologists, v. 51, no. 10, p. 1979–2032.

Rose, P. R., 1976, Mississippian carbonate shelf margins, western U.S.: U.S. Geological Survey Journal of Research, v. 4, no. 4, p. 449–466.

Sales, J. K., 1968, Crustal mechanics of Cordilleran foreland deformation—A regional and scale-model approach: Bulletin of the American Association of Petroleum Geologists, v. 52, no. 10, p. 2106–2044.

Sandberg, C. A., 1961, Distribution and thickness of Devonian rocks in Williston basin and in central Montana and north-central Wyoming: U.S. Geological Survey Bulletin 1112-D, p. 105–127.

______1962, Geology of the Williston basin, North Dakota, Montana, and South Dakota, with reference to subsurface disposal of radioactive wastes: U.S. Geological Survey Open-File Report TEI-809, 148 p.

Sandberg, C. A., and Hammond, C. R., 1958, Devonian system in Williston basin and central Montana: Bulletin of the American Association of Petroleum Geologists, v. 42, no. 10, p. 2293–2334.

Sandberg, C. A., and Poole, F. G., 1977, Conodont biostratigraphy and depositional complexes of UpperDevonian cratonic-platform and continental shelfrocks in the western United States, *in* Murphy,M. A., and others, eds., Western North America; Devonian:University of California at RiversideCampus, Museum Contributions 4, p. 144–182.

Sando, W. J., 1974, Ancient solution phenomena in the Madison Limestone (Mississippian) of north-centralWyoming: U.S. Geological Survey Journal of Research, v. 2, no. 2, p. 133–141.

______1976a, Madison Limestone, east flank of Bighorn Mountains, Wyoming: Wyoming Geological Association Annual Field Conference, 28th, 1976, Guidebook, p. 45–52.

______1976b, Mississippian history of the northern Rocky Mountain region: U.S. Geological Survey ofResearch, v. 4, no. 3, p. 317–338.

Sando, W. J., Dutro, J. T., Jr., Sandberg, C. A., and Manet, B. L., 1976, Revision of Mississippian stratigraphy, eastern Idaho and northeastern Utah: U.S. Geological Survey Journal of Research, v. 4, no. 4, p. 467–479.

Sando, W. J., MacKenzie, G. M., Jr., and Dutro, J. T., Jr., 1975, Stratigraphy and geologic history of the Amsden Formation (Mississippian and Pennsylvanian) of Wyoming: U.S. Geological Survey Professional Paper 848-A, p. A1-A893.

Slack, P. B., 1981, Paleotectonics and hydrocarbon accumulation, Powder River basin, Wyoming: Bulletin of the American Association of Petroleum Geologists, v, 65, no. 4, p. 730–743.

Smith, D. L., 1972, Depositional cycles of the Lodgepole Formation (Mississippian) in central Montana: Montana Geological Society Annual Field Conference, 21st, September 1972, Guidebook, p. 27-36.

Sonnenberg, F. P., 1956, Tectonic patterns of Central Montana: Billings Geological Society Annual Field Conference, 7th, 1956, Guidebook, p. 73–81.

Stearns, D. W., Sacrison, W. R., and Hanson, R. C., 1975, Structural history of southwestern Wyoming as evidenced from outcrop and seismic, *in* Bolyard, D. W., ed., Symposium on Deep Drilling Frontiers in the Central Rocky Mountains, Steamboat Springs, Colorado, 1975, Proceedings: Denver, Rocky Mountain Association of Geologists, p. 9–20.

Steece, F. V., 1978, Deadwood Formation in the Williston basin: Montana Geological Society Annual Conference, 24th [Williston basin Symposium], Billings, Mont., September 1978, Guidebook, p. 63–69.

Stone, D. S., 1969, Wrench faulting and Rocky Mountain tectonics: Wyoming Geological Association Earth Science Bulletin, v. 2, no. 2, p. 27–41.

______1970, Principal horizontal stress in the central Rocky Mountains versus California: The Mountain Geologist, v. 7, no. 2, p. 69–82.

______1971, Tectonic sketch map of the central Rocky Mountains: Wyoming Geological Association Annual Field Conference, 23d, 1971, Guidebook, scale 1 in. = 30 mi, 1 sheet.

Swenson, F. A., 1968, New theory of recharge in the artesian basin of the Dakotas: Geological Society of America Bulletin, v. 79, no. 1, p. 163–182.

Swenson, F. A., Miller, W. R., Hodson, W. G., and Visher, F. N., 1976, Map showing the configuration and thickness and potentiometric surface and water quality in the Madison Group, Powder River basin, Wyoming and Montana: U.S. Geological Survey Miscellaneous Investigations Series I–847–C.

Thom, W. T., Jr., 1923, The relation of deep-seated faults to the surface structural features of central Montana: Bulletin of the American Association of Petroleum Geologists, v. 7, no. 1, p. 1–13.

Thomas, G. E., 1971, Continental plate tectonics, southwest Wyoming, *in* Renfro, A. R., ed., Symposium on Wyoming Tectonics and Their Economic Significance, Casper, Wyo., September 1971, Guidebook: Capser, Wyo., Wyoming Geological Association, p. 103–123.

______1974, Lineament-block tectonics—Williston-Blood Creed basin: Bulletin of the American Association of Petroleum Geologists, v. 58, no. 7, p. 1305–1322.

Trescott, P. C., 1975, Documentation of finite-difference model for simulation of three-dimensional ground-water flow: U.S. Geological Survey Open-File Report 75–438, 30 p. plus appendix.

Trescott, P. C., and Larson, S. P., 1976, Supplement to U.S. Geological Survey Open-File Report 75–438, Documentation of finite-difference model for simulation of three-dimensional ground-water flow: U.S. Geological Survey Open-File Report 76-591, 17 p.

U.S. Geological Survey, 1975, Plan of study of the hydrology of the Madison Limestone and associated rocks in parts of Montana, Nebraska, North Dakota, and Wyoming: U.S. Geological Survey Open-File Report 75-631, 44 p.

Warner, L. A., 1978, The Colorado lineament—A middle Precambrian wrench fault system: Geological Society of America Bulletin, v. 89, no. 2, p. 161-171.

Weertman, Johannes, 1972, General theory of water flow at the base of a glacier or ice sheet: Reviews of Geophysics and Space Physics, v. 10, no. 1, p. 287-533.

Weimer, R. J., Emme, J. J., Farmer, C. L., Anna L. O., Davis, T. L., and Kidney, R. L., 1982, Tectonic influences of sedimentation, Early Cretaceous, east flank Powder River basin, Wyoming and South Dakota: Colorado School of Mines Quarterly, v. 77, no. 4, 61 p.

Weiss, Emanuel, 1982, A model for the simulation of flow of variable-density ground water in three dimensions under steady-state conditions: U.S. Geological Survey Open-File Report 82-352, 70 p.

Wyoming State Engineer's Office, Underground water supply in the Madison Limestone: Cheyenne, Wyo., 117 p.

______1976, Investigation of recharge to groundwater reservoirs of northeastern Wyoming (the Powder River basin): Washington, D. C., Old West Regional Commission, 111 p.