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Ground-water Flow Systems and the Water-Resources Potential of the Dakota Aquifer in a Two-county Area in North-central Kansas

by

Alan Wade

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GROUND-WATER FLOW SYSTEMS AND THE WATER-RESOURCES POTENTIAL OF THE DAKOTA AQUIFER IN A TWO-COUNTY AREA IN NORTH-CENTRAL KANSAS

by

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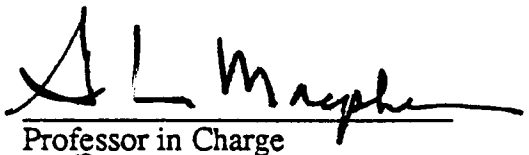
of Master of Science.



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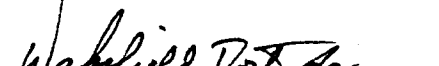

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ABSTRACT

The Dakota aquifer in Republic, Washington, and northern Cloud counties is approximately 40% sandstone and 60% mudstone, and is composed mainly of the Dakota Formation. Test holes reveal that the sandstones are largely isolated from each other by mudstone with the exception of a sheet-like, amalgamated fluvial channel sandstone at the base of the Dakota Formation, which is the most important conduit of lateral flow in the aquifer.

Field measurements of water levels, depths to water from well-drillers logs, ground-water quality and surface features of ground-water discharge show that flow systems in the aquifer are both regional and local. The regional system is driven by the topographic slope of the Great Plains toward the east. This system brings halite-solution brine, originally derived from the Permian, into the area from the west. The incised watersheds of Salt Creek in Republic County and Mill Creek in Washington County drain the regional flow system creating local flow systems which are recharged in the outcrop area. The local systems have potentiometric surfaces which are subdued forms of the local relief.

A pumping test shows that sandstones, the major conduits of horizontal flow, have a specific storage of approximately 10^{-6} ft.$^{-1}$ ($3x10^{-6}$ m^{-1}) and hydraulic conductivities as high as 76 ft./day (23 m/day), whereas mudstones, the major conduits of vertical flow, have hydraulic conductivities of approximately 10^{-3} ft./day ($3x10^{-4}$ m/day). A steady-state flow model indicates recharge to the local systems is approximately $1/4$ in./yr. (6 mm/yr.).

Ground water in local flow systems is pumped for domestic, irrigation, stock and municipal use. Simulations of a local flow system in southwestern Washington County and limited long-term monitoring data suggest that water quality and availability have not been significantly affected by pumping. Simulations of additional pumping show that it is likely the aquifer could sustain an increase in pumping in this area to double the current level without drawing in water of inferior quality. Mean drawdowns estimated for this pumping rate would be between approximately three and six feet (one and two metres).

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CHAPTER 1. INTRODUCTION

The Dakota aquifer extends across much of the Great Plains of North America. Russell (1928) claimed it was the largest and most important aquifer system in the United States and possibly the world. Many years earlier, Darton (1905) already recognized the great importance of the Dakota aquifer as a water resource. He described it as "the most widely extended and most useful" of the central Great Plains (Darton, 1905, p. 193). The Dakota aquifer remains an important source of water in the Great Plains of North America today where precipitation is limited and sporadic, evaporation rates are high, and surface water supplies are extremely limited.

The Dakota aquifer is the second most extensive aquifer system in Kansas after the High Plains aquifer (Macfarlane et al., 1990). However, until recently relatively little was known about the availability and quality of Dakota ground water. A long-term multi-agency research program to assess the water-resources potential of the Dakota aquifer in Kansas is in progress at the Kansas Geological Survey. The information from this investigation will help the state to plan the use of this valuable resource. This thesis project is an integral part of the Dakota research program.

The study area of this project is defined by the Nebraska state line on the north, the Republican River aquifer on the south and west and the limit of the lower Cretaceous rocks in outcrop on the east (Figure 1.1). Sandstones and mudstones of the Dakota Formation constitute the majority of the aquifer in this area; similar lithologies in the Kiowa Formation also comprise part of the aquifer in a few subsurface locations. The sedimentary architecture of the aquifer is complicated because the sandstone typically occurs in belt-like patterns, concentrated in irregular lenses which differ in thickness, areal extent and the degree with which they interconnect. The sandstones are the main aquifer units which yield water to wells. Although the mudstones are important for ground-water storage, they are aquitards which retard the flow of ground water. The proportion of sandstone and the degree of interconnection between sandstone bodies are therefore important factors affecting the flow of ground water in the Dakota aquifer.

There are five main objectives of this project. The first is to develop a better understanding of the geology of the aquifer in this area in order to identify distinct

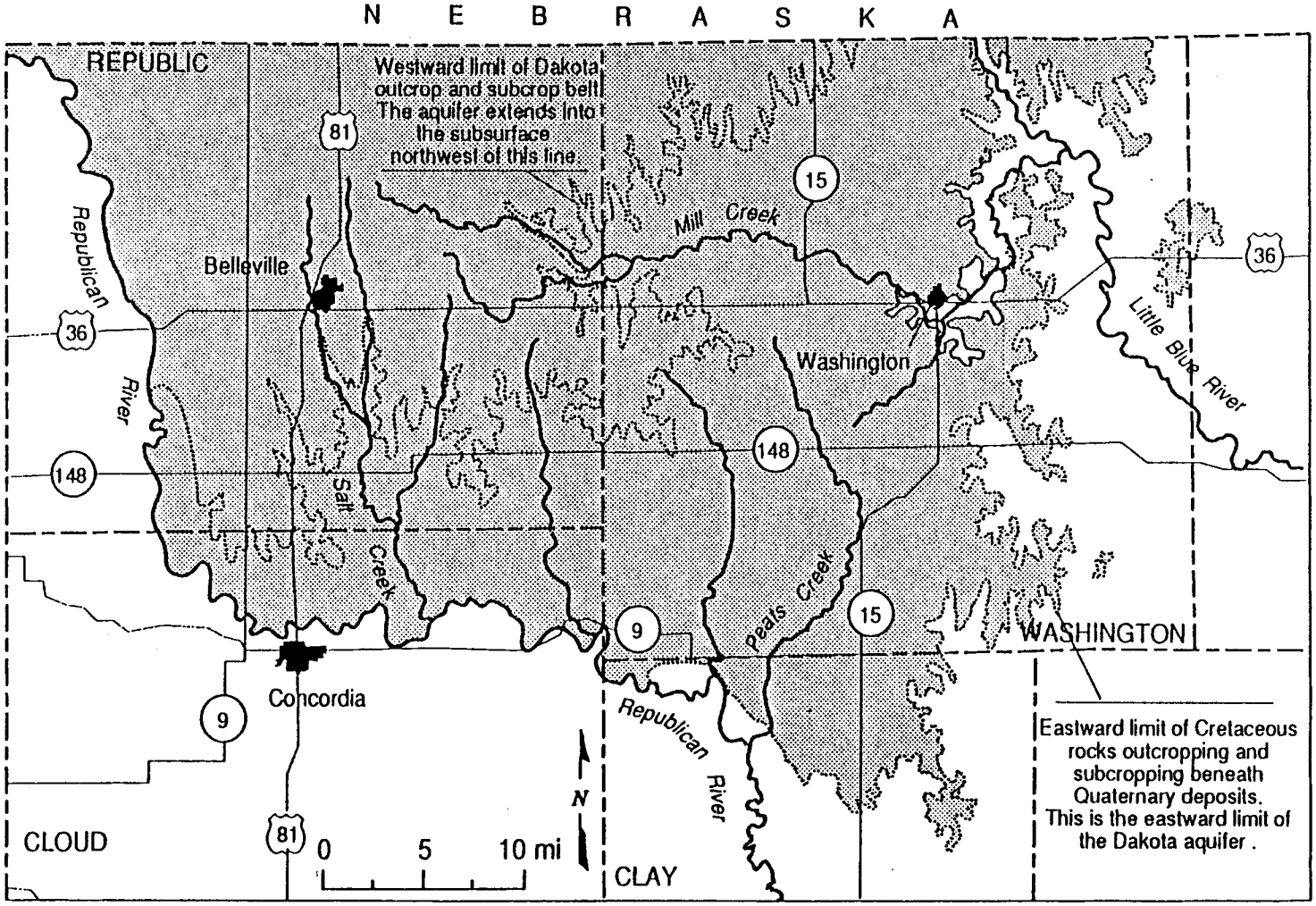


Figure 1.1. The study area is defined by the Nebraska state line on the north, the Republican river on the west and south, and the limit of the outcrop on the east. The outcrop is not mapped west and south of the Republican river.

hydrostratigraphic units. The second objective is to present recent water-level measurements from wells in the aquifer and to use these to help determine the nature of the ground-water flow systems in the aquifer. The third objective of this project is to estimate aquifer properties based on drawdowns measured during a pumping test. The fourth objective is to simulate the flow of ground water through a local flow system in the outcrop belt of the Dakota Formation in order to determine the recharge rate at the surface needed to sustain existing hydraulic heads. Finally, the fifth objective is to predict the effects of increased pumping of ground water from this local flow system.

Nine test holes were drilled to provide information concerning the subsurface geology of the aquifer. The system for identifying locations of test holes and wells in this report is described in Appendix 1. The test-hole information was used to determine the characteristics of the geologic framework of the aquifer, and to correlate sequences of rocks between different parts of the study area by their environments of deposition and stratigraphic position. The geologic data were used to define units with distinct hydraulic properties which affect the flow of ground water in the aquifer.

Hydraulic heads from wells surveyed in 1990 were used in conjunction with previous water-level measurements to map the "potentiometric surface" configuration of a sandstone at the base of the Dakota Formation. This surface was used to determine the directions of ground-water flow within the sandstone based on the principal that, ignoring anisotropy, flow direction is from locations with high hydraulic head to locations with low hydraulic head, perpendicular to lines joining points of equal head. Recharge and discharge zones were identified in the outcrop area of the formation using a combination of the water-level data, surface features and topographic information.

Hydraulic conductivity and storativity are important aquifer properties affecting the flow of ground water. These were estimated for a sandstone at the base of the aquifer by means of a pumping test using a pumping well and an observation well to measure drawdown. This is the first pumping test using an observation well that has been performed in the Dakota aquifer in Washington County.

The computer model MODFLOW (McDonald and Harbaugh, 1988) simulated steady-state ground-water flow in the sandstone at the base of the aquifer

over a 300 sq. mi. area of Washington County, which included the site of the pumping test. The recharge rate needed to sustain the potentiometric surface in this area was determined by computing synthetic hydraulic heads and minimizing the difference between these and the existing heads. The aquifer parameters used for the modelling were estimated from the pumping test. Two cross-sectional models of flow in the aquifer were also run in order to corroborate the result of the areal model. Using the recharge rate estimated in this way, the same computer model was used in a transient simulation to predict the effects of increased ground-water withdrawal on availability and quality of water.

CHAPTER 2. GEOLOGY OF THE DAKOTA AQUIFER

2.1. Previous Work

The Dakota aquifer consists of interbedded sandstones and mudstones deposited in fluvial, deltaic, and near-shore marine systems during the early Cretaceous Period (Macfarlane et al., 1991b). The Dakota Formation is the main geologic unit of the aquifer in Kansas; the Kiowa Formation and the Cheyenne Sandstone comprise the remainder of the aquifer throughout much of the state (Figure 2.1). However, in north-central Kansas the Cheyenne sandstone is not present; it thins out in the subsurface two counties west of the area of this study (Macfarlane et al., 1991a).

2.1.1. The Kiowa Formation

The Kiowa Formation was originally named the Kiowa Shale from a type section in Kiowa County by Cragin (1895). Plummer and Romary (1942) redefined the formation to include all the strata above the Cheyenne Sandstone and below the Dakota Formation (Figure 2.1). So the definition of the Kiowa Formation included some sandstone and limestone as well as marine shale. The Kiowa Formation, which is bounded by unconformities at its top and base, is upper Albian in age. It has a thickness of 300 ft. in its type area (Latta, 1946) and a maximum thickness of 380 ft. (Merriam, 1957) but is between 100 and 125 ft. thick in central Kansas. The basal unconformity in north-central Kansas represents a major gap of approximately 50 m.y. in the stratigraphic record between the age of the underlying Wellington Formation and the Kiowa Formation (Zeller, 1968). This unconformity commonly has 50 ft. and sometimes 100 ft. of local relief (Franks, 1966).

Franks (1966) defined a variety of white siltstones, lenticular sandstones, and red-mottled siltstones and mudstones at the base of the Kiowa Formation as the Longford Member, named for their type locality near Longford, Clay County (Figure 2.1). The Longford Member of the Kiowa Formation crops out in a north-south trending band along the eastern edge of the Kiowa outcrop belt in central Kansas, where it is up to 100 ft. thick (Franks, 1979). Recent test-hole drilling in Lincoln County, 45 mi. west of Longford, has revealed a 200-ft. sequence of Longford-Member strata deposited in a tidally-influenced deltaic environment, with negligible marine shale (Macfarlane et al., 1991a). The standard Kansas

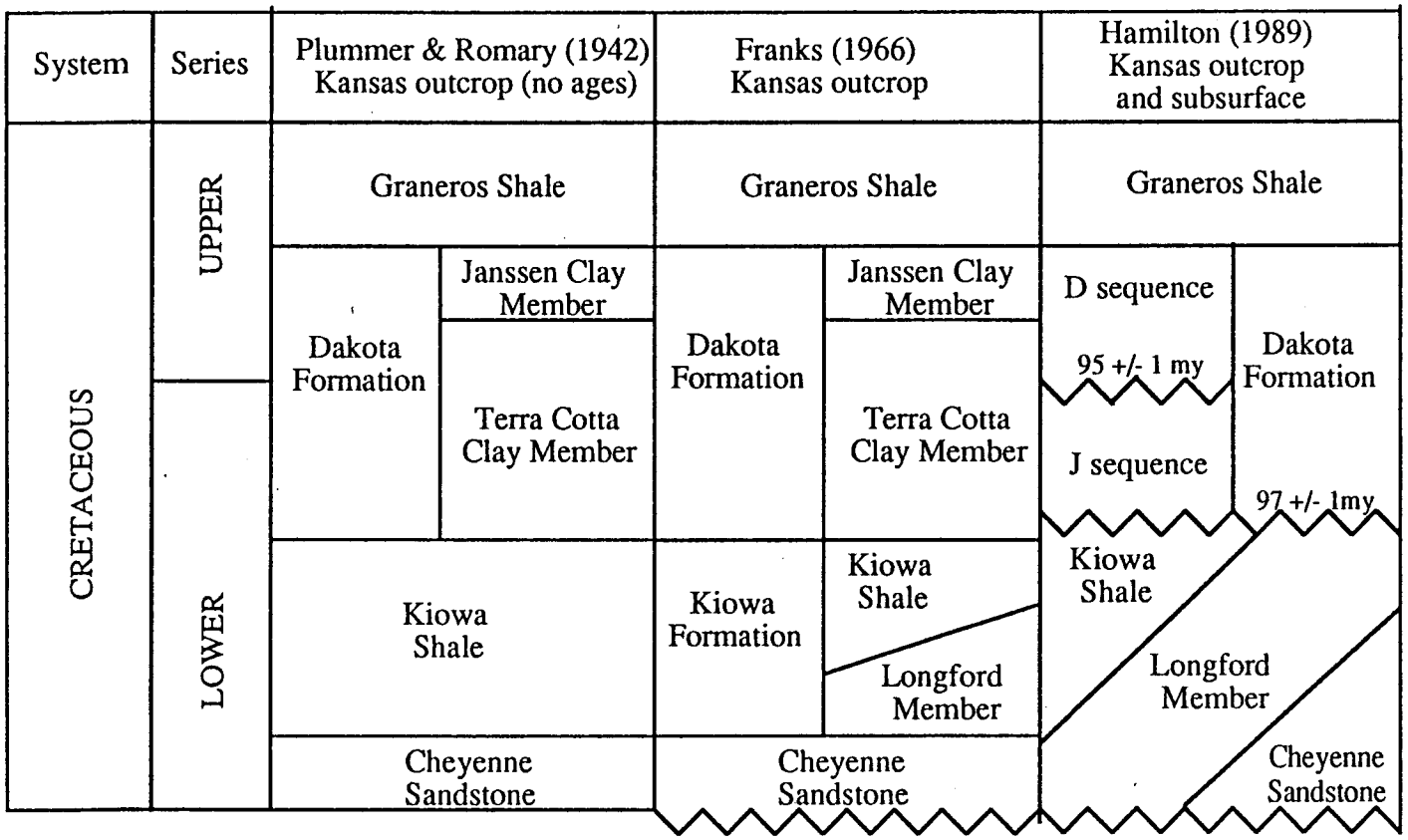


Figure 2.1. Stratigraphic nomenclature of the geologic units of the Dakota aquifer in Kansas.

hydrostratigraphic interpretation of the Kiowa Formation as a very low permeability aquitard dominated by shale does not apply in north-central Kansas.

The Kiowa Formation is not present in outcrop in Washington County. However, the Longford Member has been mapped in the northeastern corner of Clay County close to Washington County, 25 mi. northeast of its type locality (Ross et al., 1991).

2.1.2. The Dakota Formation

The Dakota Formation was first described by Meek and Hayden (1861) near Dakota City in northeastern Nebraska. In Kansas, Plummer and Romary (1942) defined the Dakota Formation to include all the non-marine and littoral clay and sandstone above the Kiowa Formation and below the Graneros Shale. They subdivided the formation into the Terra Cotta Clay Member and the overlying Janssen Clay Member (Figure 2.1) based on type sections in Ellsworth County, central Kansas. The average thickness of the Dakota Formation in Kansas is 250 ft. (Siemers, 1971) although the thickness exceeds 350 ft. in parts of western Kansas (Macfarlane et al., 1991b). The Terra Cotta Member comprises approximately the lower two-thirds of the Dakota Formation (Plummer and Romary, 1942). Both members contain siltstone, sandstone, and massive mudstone. In the original definition red-mottled mudstone was a characteristic of the Terra Cotta Clay Member only (Plummer and Romary, 1942) but Franks (1966) observed red-mottled mudstone in both members. The Janssen Clay Member is also distinguishable by the presence of lignites, and ironstone-bearing or sideritic mudstones (Siemers, 1971).

Sandstones in the Terra Cotta Clay Member are mainly fluvial channel deposits. Multistory fluvial channel sandstones up to 70 ft. thick have been described in outcrop, and conglomeratic sandstones are common at the base of the formation (Franks, 1975). The sandstone bodies are generally discontinuous and ribbon-like (Macfarlane et al., 1991b). Planar-tabular cross-bedding is common in the channel sandstones, which are fine to medium-grained.

The Janssen Clay Member also contains fluvial sandstone lenses. However, these sandstones are trough and epsilon cross-bedded and no coarser than fine-grained sand (Karl, 1976); planar-tabular cross-bedding has not been documented in these sandstones. The general differences in style of cross bedding in the two

members are thought to reflect a change from relatively high-competence, low-sinuosity streams to low-competence high-sinuosity streams as base level rose during deposition of the formation (Karl, 1971, 1976). The environment of deposition of the Janssen Clay Member was transitional from non-marine to marine. Flat-bedded, sheet-like sandstones, which are not present in the Terra Cotta Member, have been interpreted as of estuarine or deltaic origin (Siemers, 1971, Macfarlane et al., 1991b).

At the type locality, the boundary between the two clay members is placed at the top of a concretionary siderite, limonite, or hematite, and/or "quartzite" sandstone which is overlain by a bed of massive gray clay (Plummer and Romary, 1942). However, Franks (1966) noted that the contact differs in stratigraphic position from place to place and that such concretionary zones occur at various stratigraphic positions at some localities and are not present everywhere. The boundary has since been interpreted as a flooding surface above which the environment of deposition was deltaic and below which it was coastal plain (Siemers, 1971, Hamilton, 1989). This is a surface separating two progradational events, rather than a sequence boundary (Hamilton, 1989).

The base of the Terra Cotta is an unconformity (Figure 2.1) which is a result of down-cutting by streams after the retreat of the Kiowa sea (Hamilton, 1989, Macfarlane et al., 1991a). This unconformity was correlated with an unconformity at the base of the "J" sandstone in Colorado (Hamilton, 1989) where it has been dated at 97±1m.y. (Weimer, 1984). It has up to 100 ft. of local relief in north-central Kansas.

The upper boundary of the formation, with the Graneros Shale, is laterally gradational although vertically it is often abrupt. Hattin (1965) suggested that the Dakota Formation intertongues with the Graneros Shale. The contact is recognized as uneven and transitional by Franks (1966, 1975) and Siemers (1971) and as a transgressive disconformity resulting from progradational backstepping by Hamilton (1989).

Hamilton (1989) separated the Dakota Formation in Kansas into two sequences, the J and D sequences, equivalent to the J and D sandstones of Colorado (Figure 2.1). The D sequence consists of the Janssen Clay Member and the upper part of the Terra Cotta Clay Member; the J sequence consists of the lower part of the Terra Cotta Clay Member. The boundary between these sequences is defined by an

unconformity, the "D" unconformity, which was correlated from Colorado (Hamilton, 1989) where it was dated at 95 +/- 1 m.y. (Weimer, 1984). The D unconformity is generally less easy to identify than the J unconformity, located at the base of the D sequence, due to the similarity in lithologies above and below it (Macfarlane et al., 1991a). Hamilton (1989) defined the D unconformity by a seaward shift in depositional environments which he recognized in outcrops and well logs in central Kansas.

2.2. Recent Test-Hole Drilling in the Study Area

2.2.1. Introduction

In north-central Kansas there has been little drilling for hydrocarbons. Therefore, before this study subsurface information was limited to water-well drillers' logs; there were no geophysical logs or cores of the Dakota aquifer in the area. In addition, outcrops of rocks comprising the Dakota aquifer are rare and mainly limited to resistant sandstones. A test-hole drilling program was undertaken in 1989 and 1990 in order to aid in understanding the sedimentary architecture of the aquifer.

2.2.2. Method

I chose sites for nine test holes to provide a broad coverage of information along the western edge of the outcrop belt of the aquifer strata as well as in the outcrop belt itself (Figure 2.2). The legal locations of all the test holes and their lease names are listed in Table 2.1. Each hole number includes a letter indicating which county the hole was drilled in; R, W, and C, are abbreviations for Republic, Washington, and Cloud counties respectively. The holes were drilled by KGS Exploration Services using mud rotary techniques. Cores were taken from R1 and W2 (Table 2.1, Figure 2.2) and these holes were logged by a commercial wireline logging company to provide detailed information through the full thickness of the aquifer at opposite ends of the subcrop part of the study area. This information was used to determine the characteristics of the geologic framework and to infer the environments in which the sediments were deposited.

The remaining holes were logged by the KGS with a gamma-ray logger. Lithologic logs of these holes were constructed by combining information from the

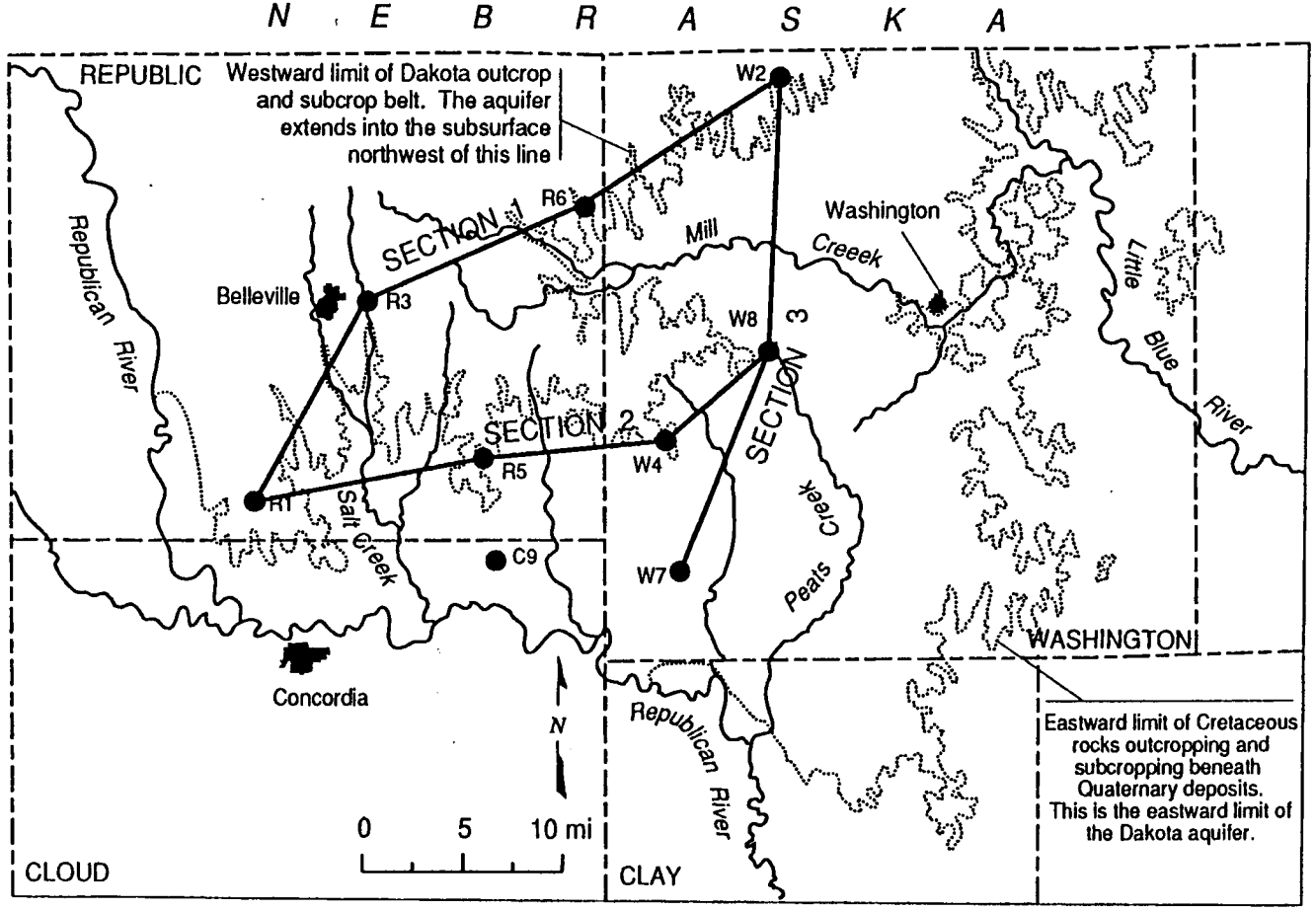


Figure 2.2. Nine test holes provide the bulk of the sedimentological data for this project. Sections 1, 2, and 3 were constructed from these test holes and are illustrated in Figures 2.12, 2.13, and 2.14. The lower limit of the Cretaceous rocks is the base of the Dakota Formation everywhere in Washington County. It is the base of the Longford Member, Kiowa Formation in much of Clay County. The outcrop is not mapped west and south of the Republican River.

gamma-ray logs, samples of drill cuttings taken every 5 ft., and observations of drilling-rig response. The gamma-ray log, a measure of the natural radiation of the rocks, clearly delineated sand-rich from clay-rich units. Clays and feldspars give a relatively high gamma-ray response whereas Dakota sandstones, which contain very little feldspar (Siemers, 1971), give a relatively low response. Rock sequences from holes R3 to C9 were correlated with those recognized in R1 and W2 with the aid of sections constructed along the lines shown on Figure 2.2.

Table 2.1. Test holes drilled for this project in Republic, Washington, and Cloud counties between September 1989 and September 1990.

Hole Number[a]	Date Drilled	Lease Name	Location
R1	9/89	Kenyon	Sec. 24, T4S, R4W
W2	11/89	Gaydusek	Sec. 10, T1S, R2E
R3	4/90	Popelka	Sec. 6, T3S, R2W
W4	4/90	Peterson	Sec. 10, T4S, R1E
R5	4/90	Benyshek	Sec. 12, T4S, R2W
R6	4/90	Cromwell	Sec. 12, T2S, R1W
W7	5/90	Leiszler	Sec. 10, T5S, R1E
W8	5/90	Nanninga	Sec. 16, T3S, R2E
C9	9/90	Feight	Sec. 7, T5S, R1W

2.2.3. Test-Hole Logs

I described and interpreted the stratigraphy of test holes R1 and W2 from the cores and geophysical logs taken from these holes (Macfarlane et al., 1991a). The core descriptions and interpretations are also reproduced in Appendices 2 and 3 of this thesis and are summarized in Figures 2.3 and 2.4. The logs of the remaining seven test holes are summarized in Figures 2.5 through 2.11. Sections 1, 2, and 3 along the lines shown in Figure 2.2 are illustrated in Figures 2.12, 2.13, and 2.14 respectively. Table 2.2 lists the thicknesses of the stratigraphic components of the aquifer in the test holes.

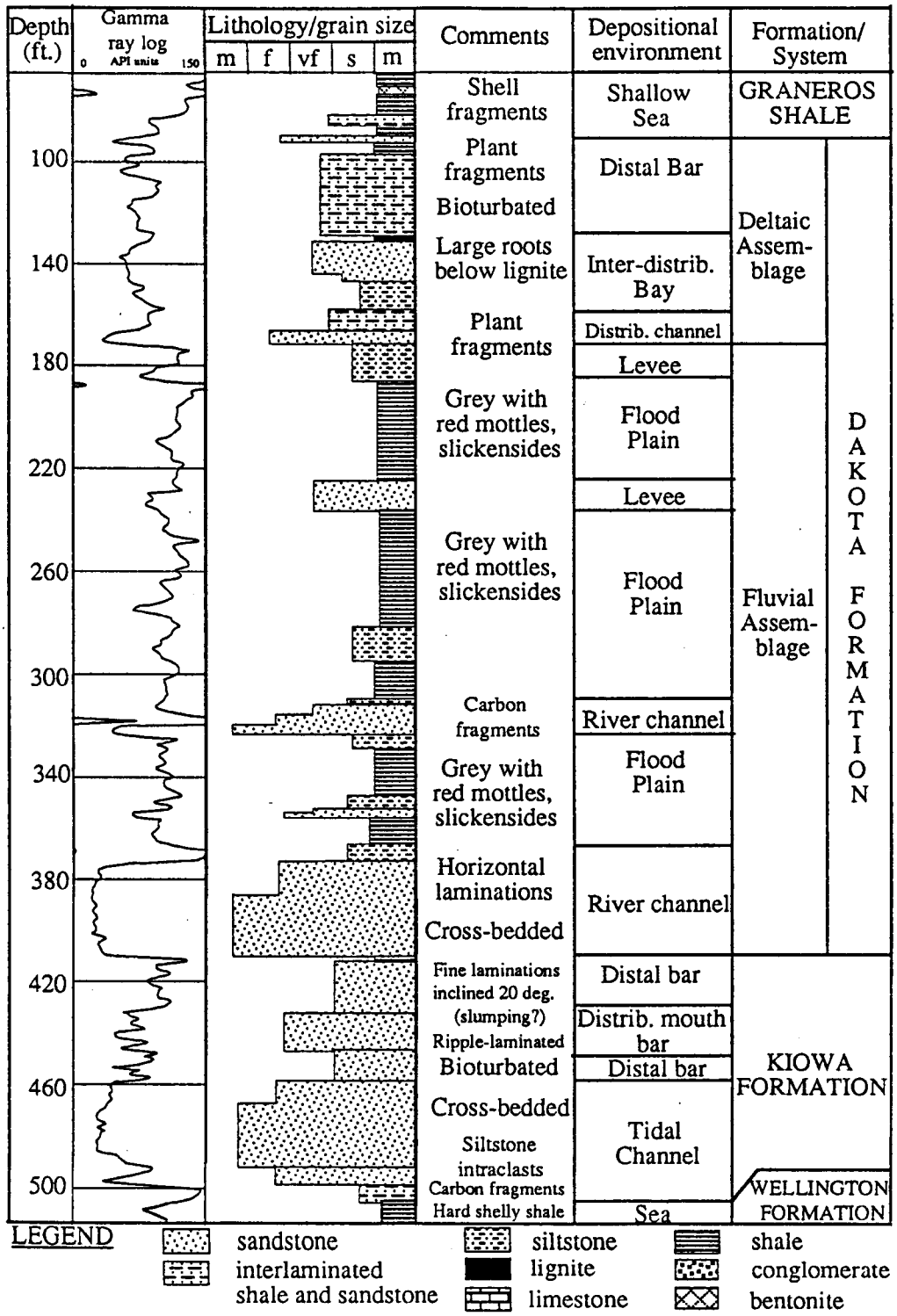


Figure 2.3. Log of test hole R1 (T4S, R4W, Section 24DD) showing lithologies and interpretations of depositional environments (Macfarlane et al., 1991a). Lithologic types are from the core description (Appendix 1). The grain-size scale ranges from mud on the right to medium-grained sand on the left. Land surface elevation = 1506ft.

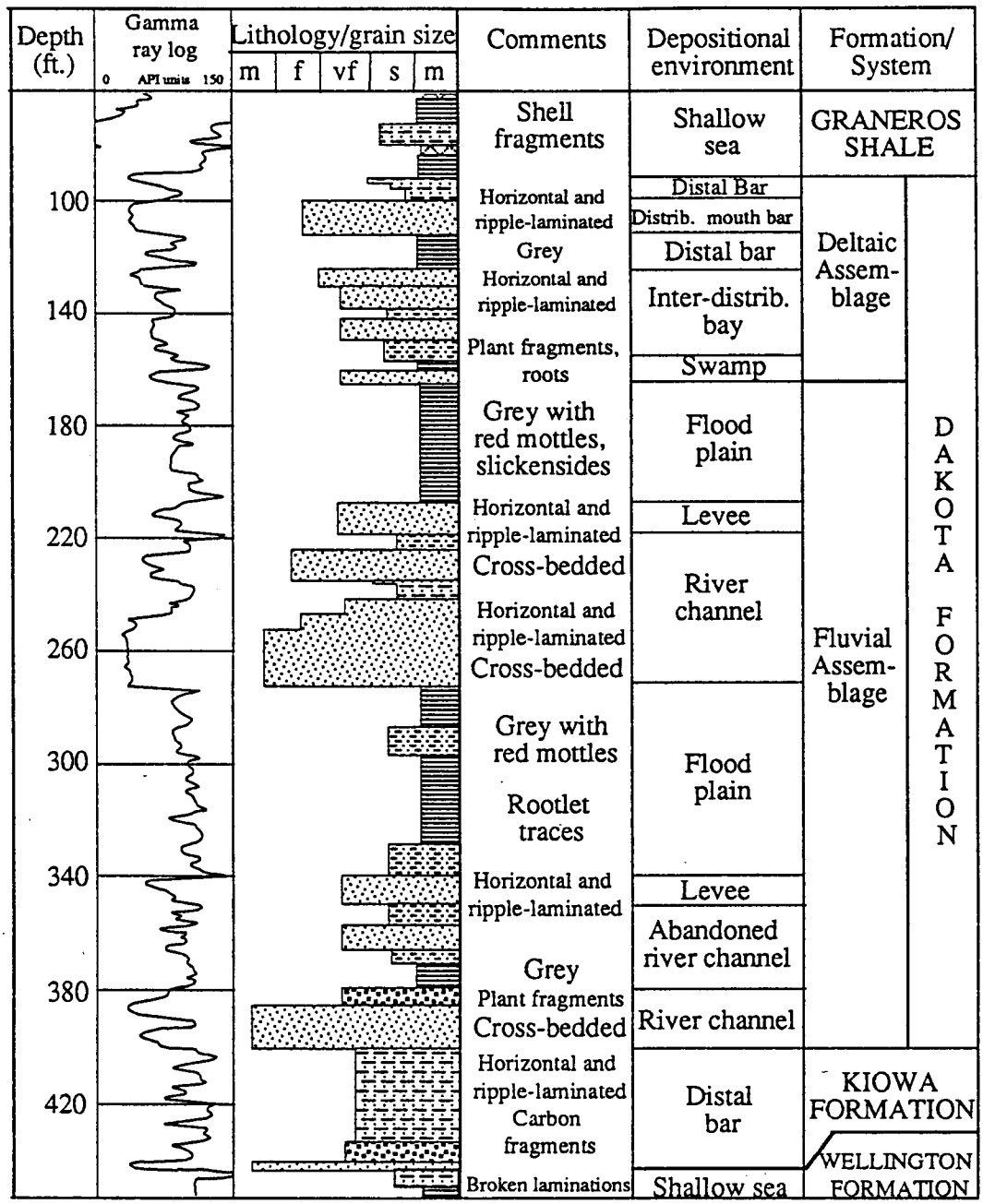


Figure 2.4. Log of test hole W2 (T1S, R2E, Section 10BCB) showing lithologies and interpretations of depositional environments (Macfarlane et al., 1991a).. Symbols for the lithologies are explained in Figure 2.3. Lithologic types and sedimentary structures are from core description (Appendix 2). The grain-size scale ranges from mud on the right to medium-grained sand on the left. Land surface elevation = 1599 ft.

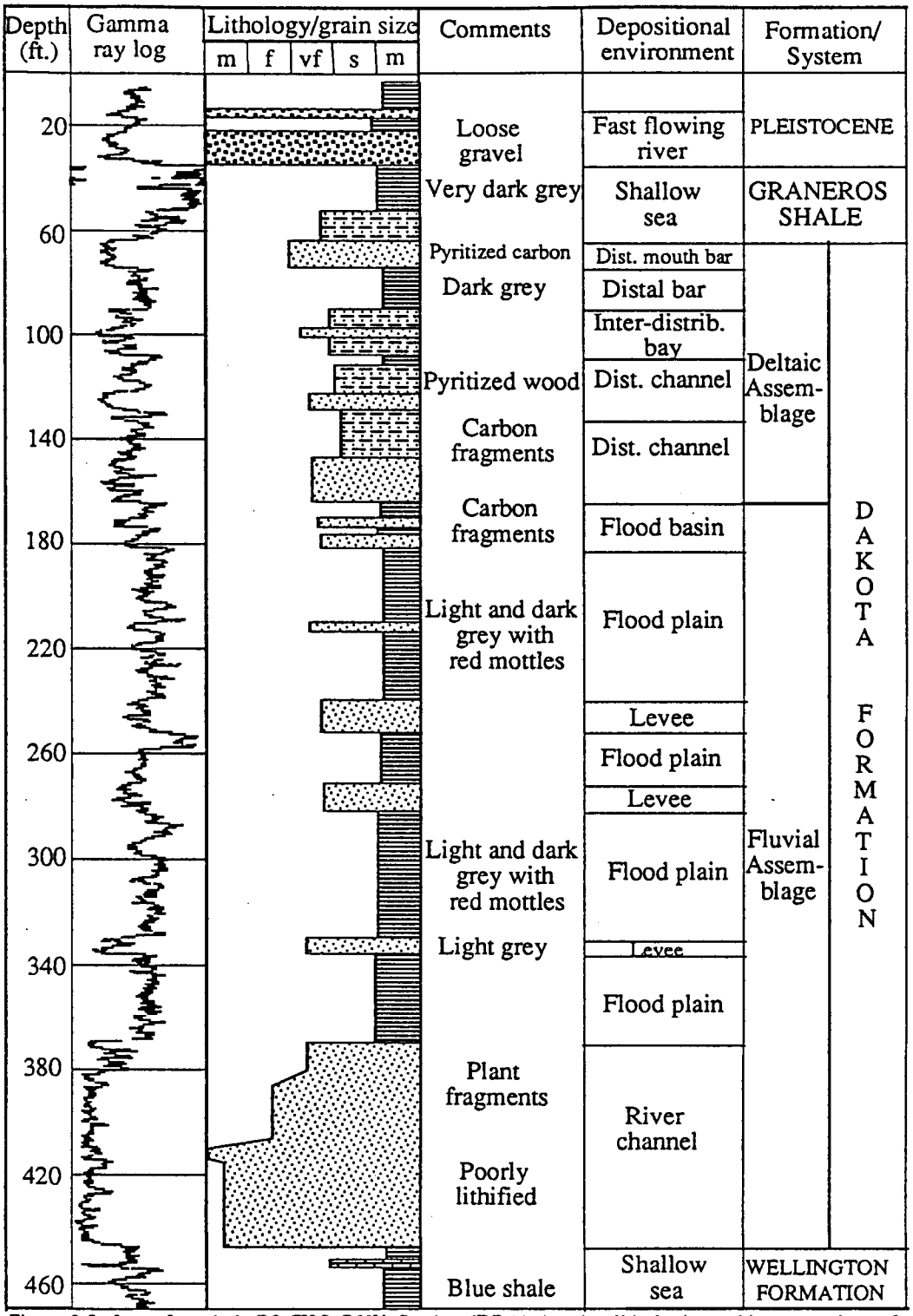


Figure 2.5. Log of test hole R3 (T3S, R2W, Section 6BBA) showing lithologies and interpretations of depositional environments. Lithologic types are from sample descriptions. The grain-size scale ranges from mud on the right to medium-grained sand on the left. Symbols for the lithologies are explained in Figure 2.3. Land surface elevation = 1479 ft.

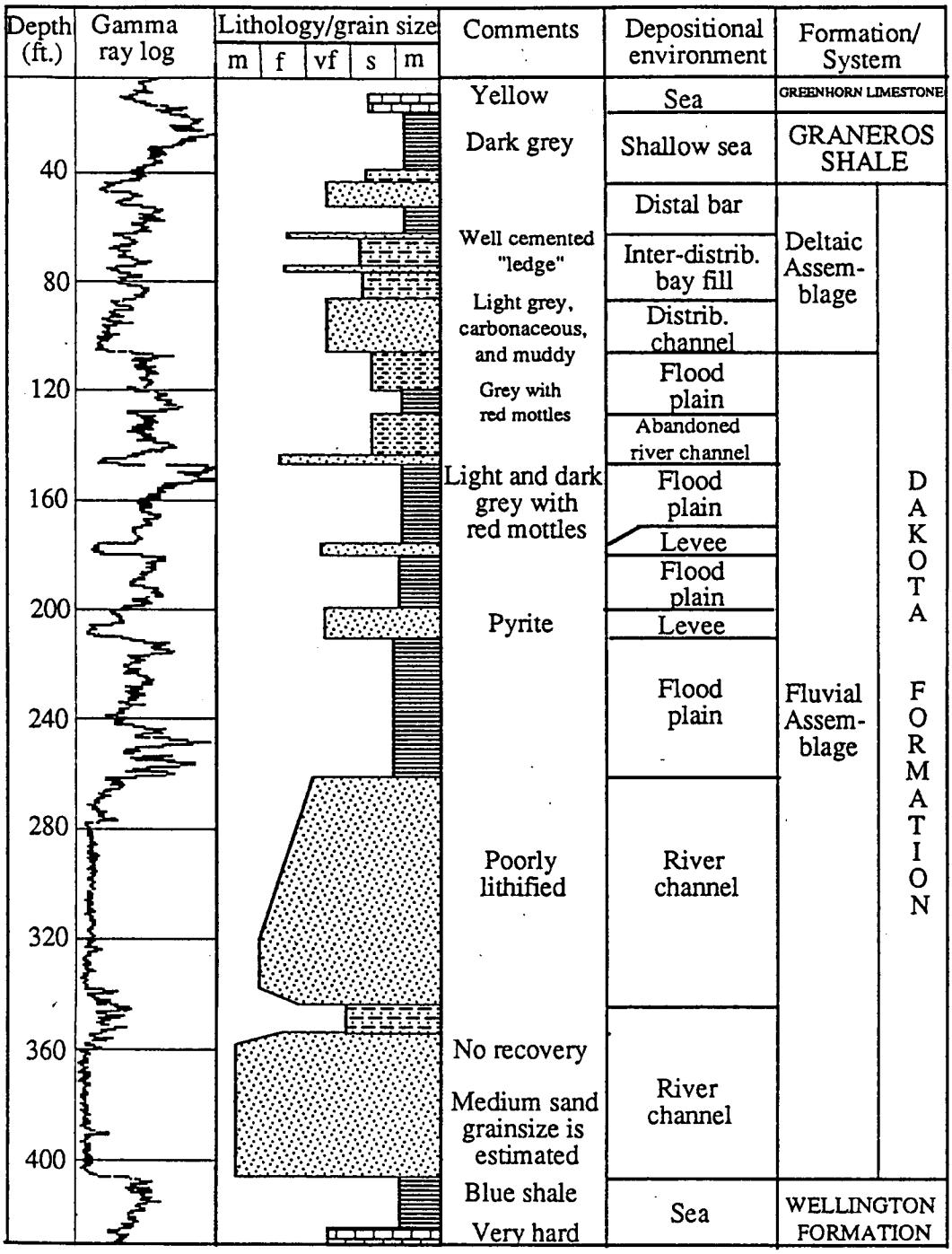


Figure 2.6. Log of test hole W4 (T4S, R1E, Section 10BBB) showing lithologies and interpretations of depositional environments. Lithologic types are from sample descriptions. The grain-size scale ranges from mud on the right to medium-grained sand on the left. Symbols for the lithologies are explained in Figure 2.3. Land surface elevation = 1588 ft.

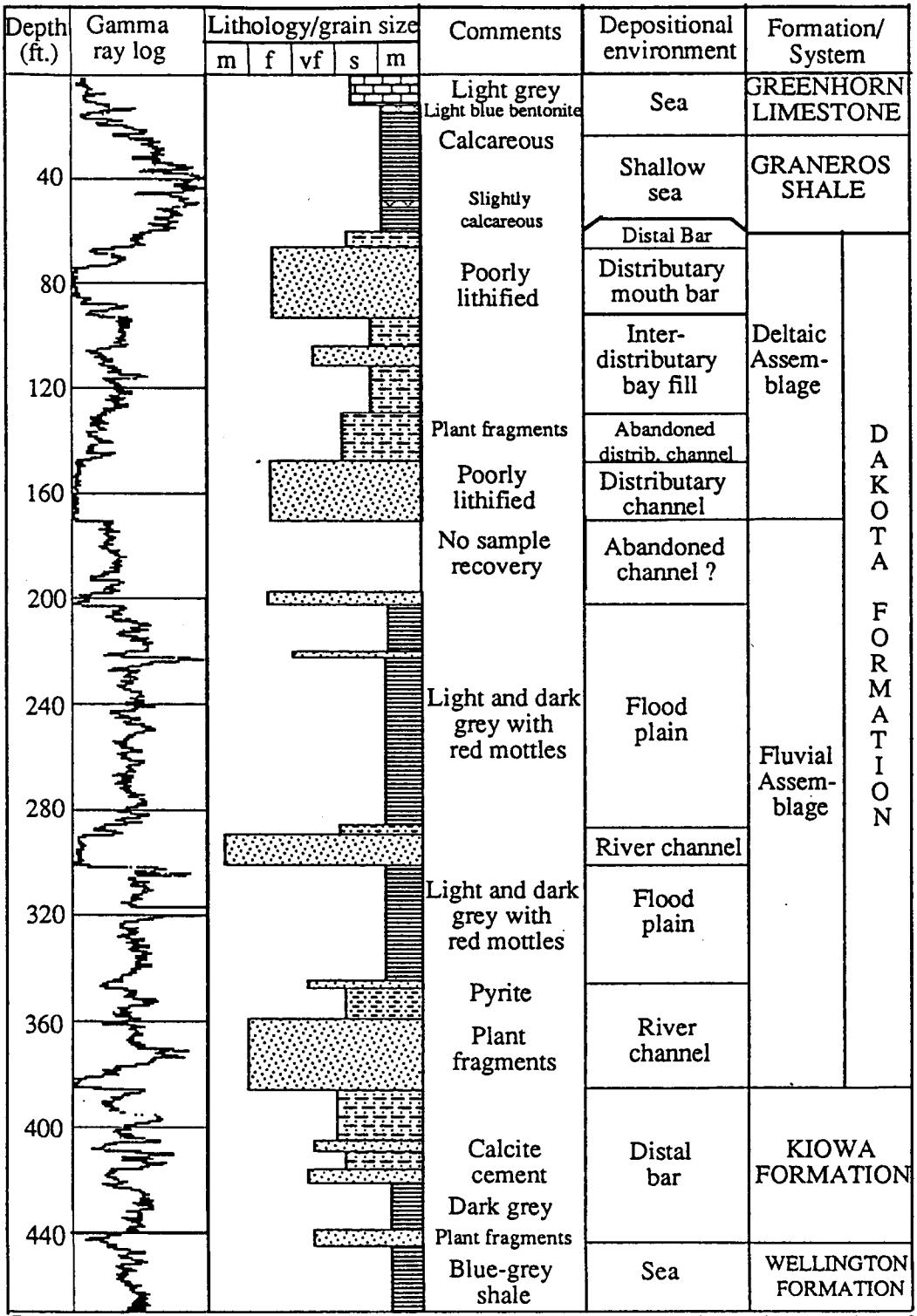


Figure 2.7. Log of test hole R5 (T4S, R2W, Section 12CDC) showing lithologies and interpretations of depositional environments. Lithologic types are from sample descriptions. The grain-size scale ranges from mud on the right to medium-grained sand on the left. Symbols for the lithologies are explained in Figure 2.3. Land surface elevation = 1546 ft.

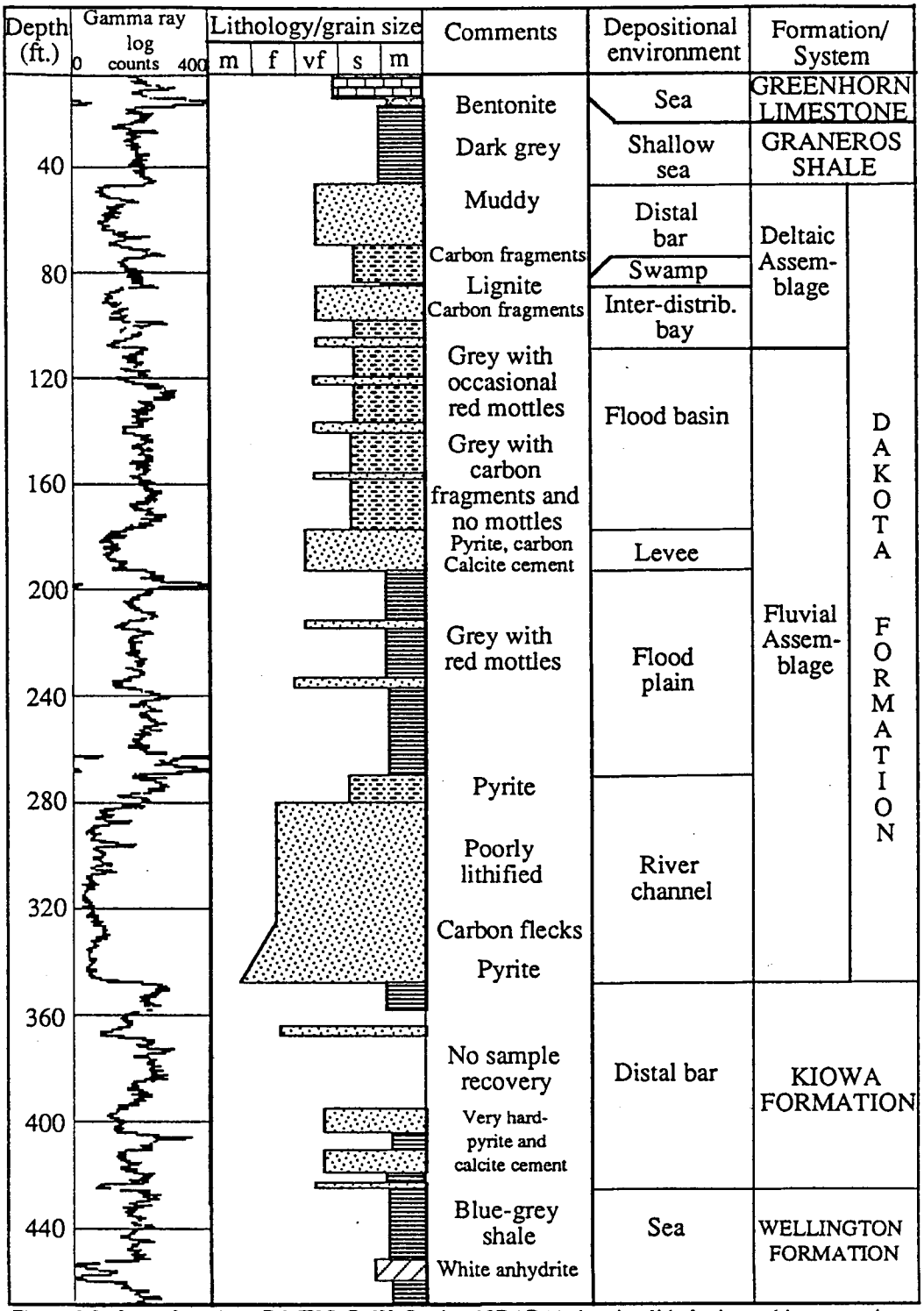


Figure 2.8. Log of test hole R6 (T2S, R1W, Section 12BABA) showing lithologies and interpretations of depositional environments. Lithologic types are from sample descriptions. The grain-size scale ranges from mud on the right to medium-grained sand on the left. Symbols for the lithologies are explained in Figure 2.3. Land surface elevation = 1505 ft.

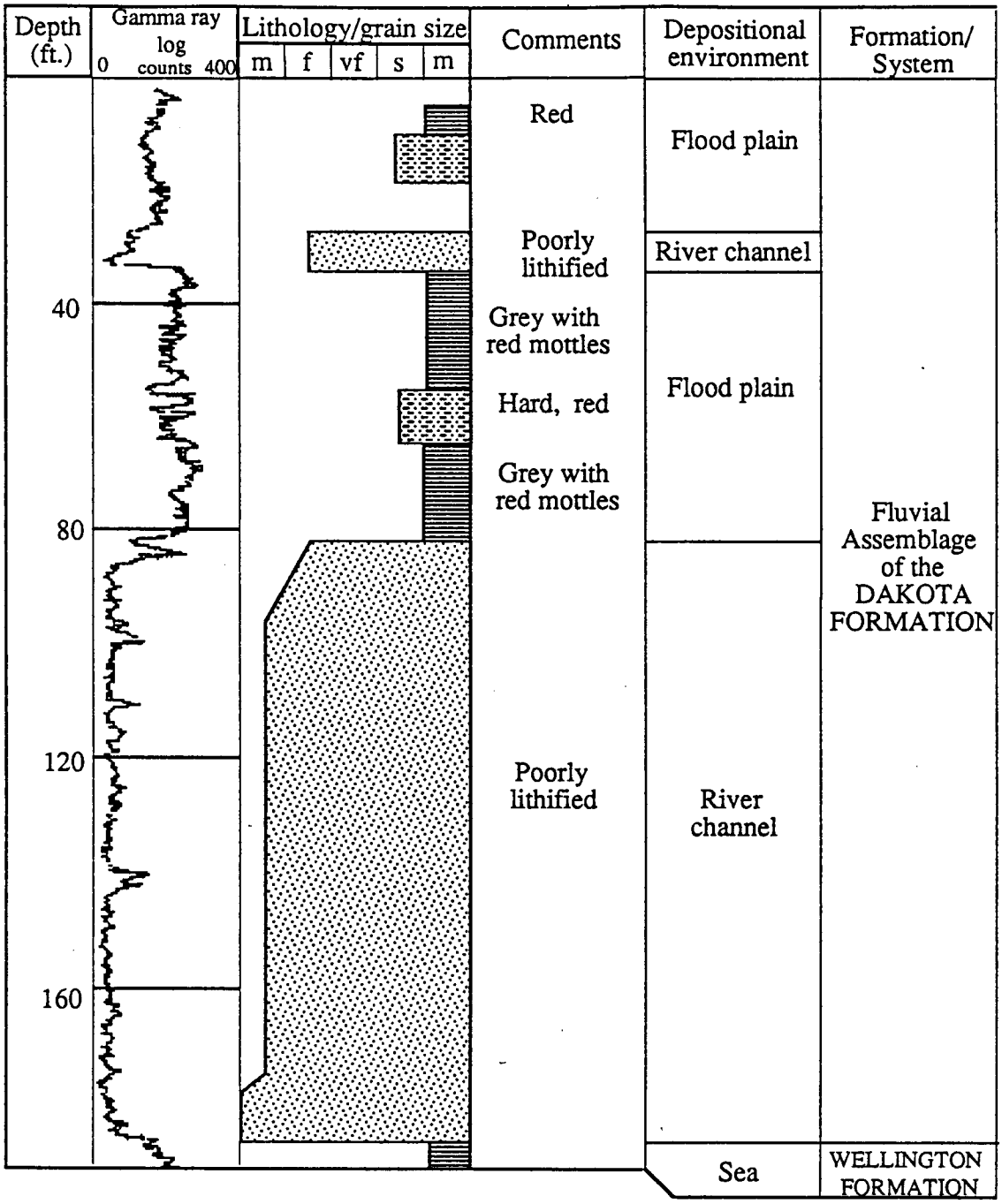


Figure 2.9. Log of test hole W7 (T5S, R1E, Section 10DDAB) showing lithologies and interpretations of depositional environments. Lithologic types are from sample descriptions. The grain-size scale ranges from mud on the right to medium-grained sand on the left. Symbols for the lithologies are explained in Figure 2.3. An observation well (O.1) was installed in this test hole to measure drawdown in a pumping test (Chapter 4). Elevation of land surface = 1367 ft.

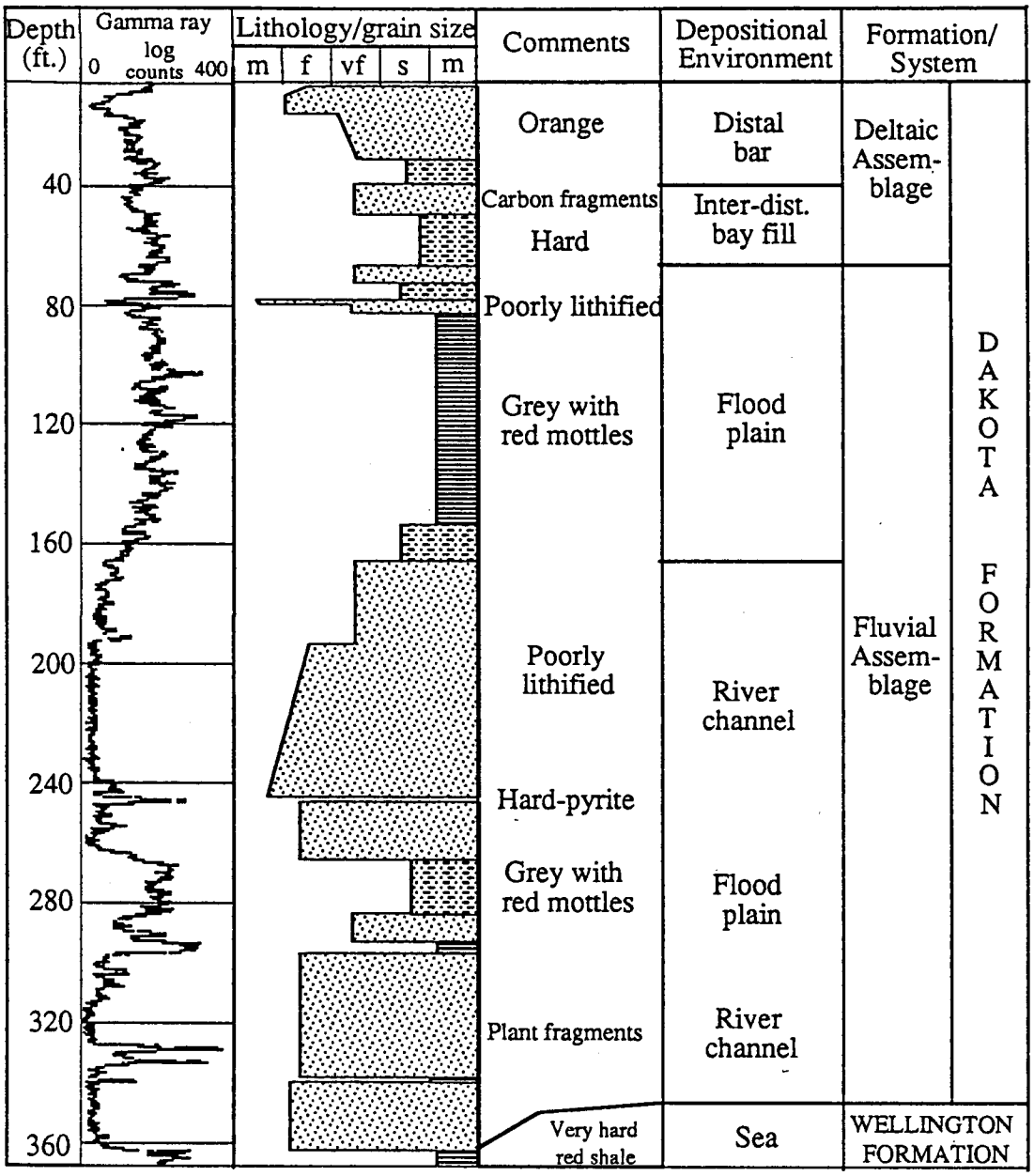


Figure 2.10. Log of test hole W8 (T3S, R2E, Section 16CCAC) showing lithologies and interpretations of depositional environments. Lithologic types are from sample descriptions. The grain-size scale ranges from mud on the right to medium-grained sand on the left. Symbols for the lithologies are explained in Figure 2.3. Elevation of land surface = 1532 ft.

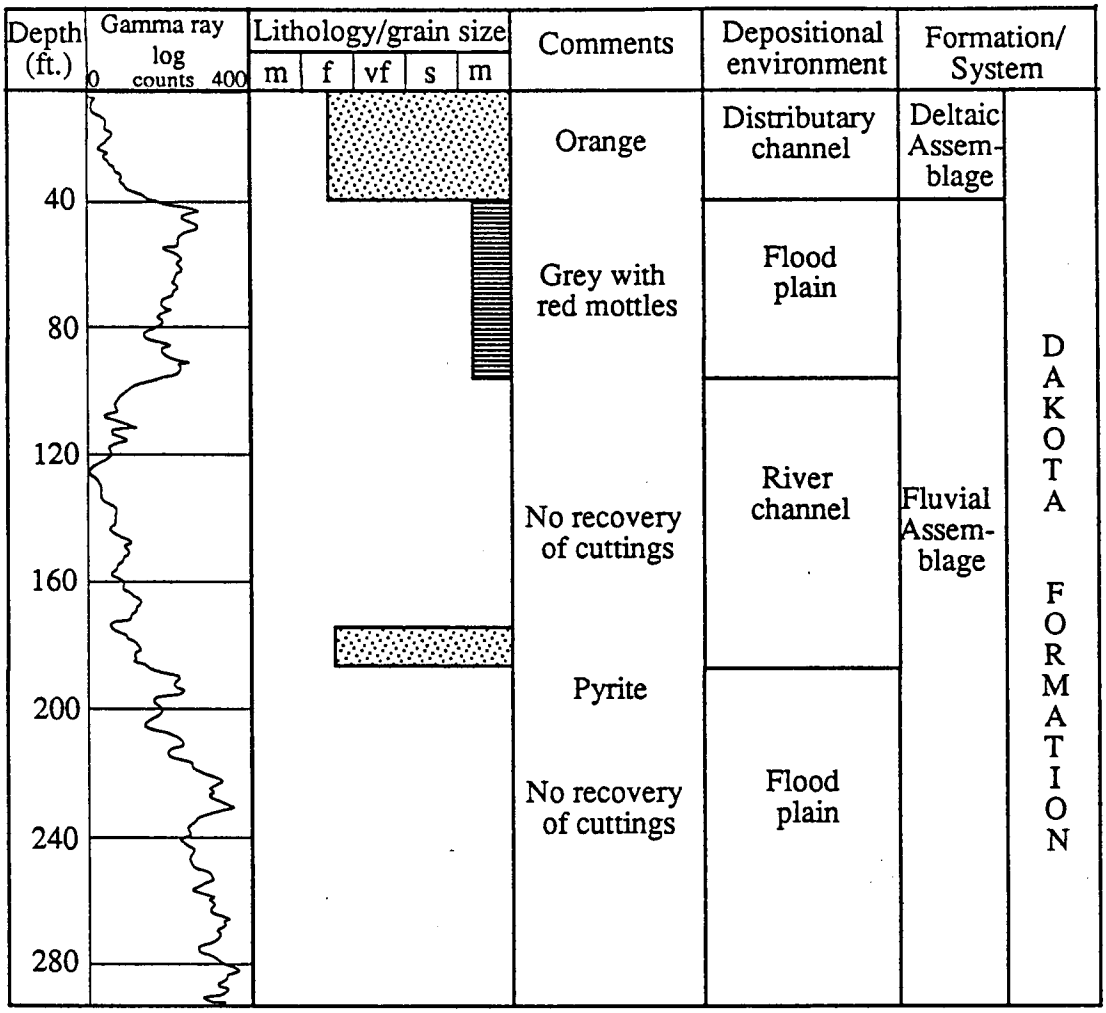


Figure 2.11. Log of test hole C9 (T5S, R1W, Section 7BBA) showing lithologies and interpretations of depositional environments. Lithologic types are from sample descriptions. The grain-size scale ranges from mud on the right to medium-grained sand on the left. Symbols for the lithologies are explained in Figure 2.3. Elevation of land surface = 1425 ft.

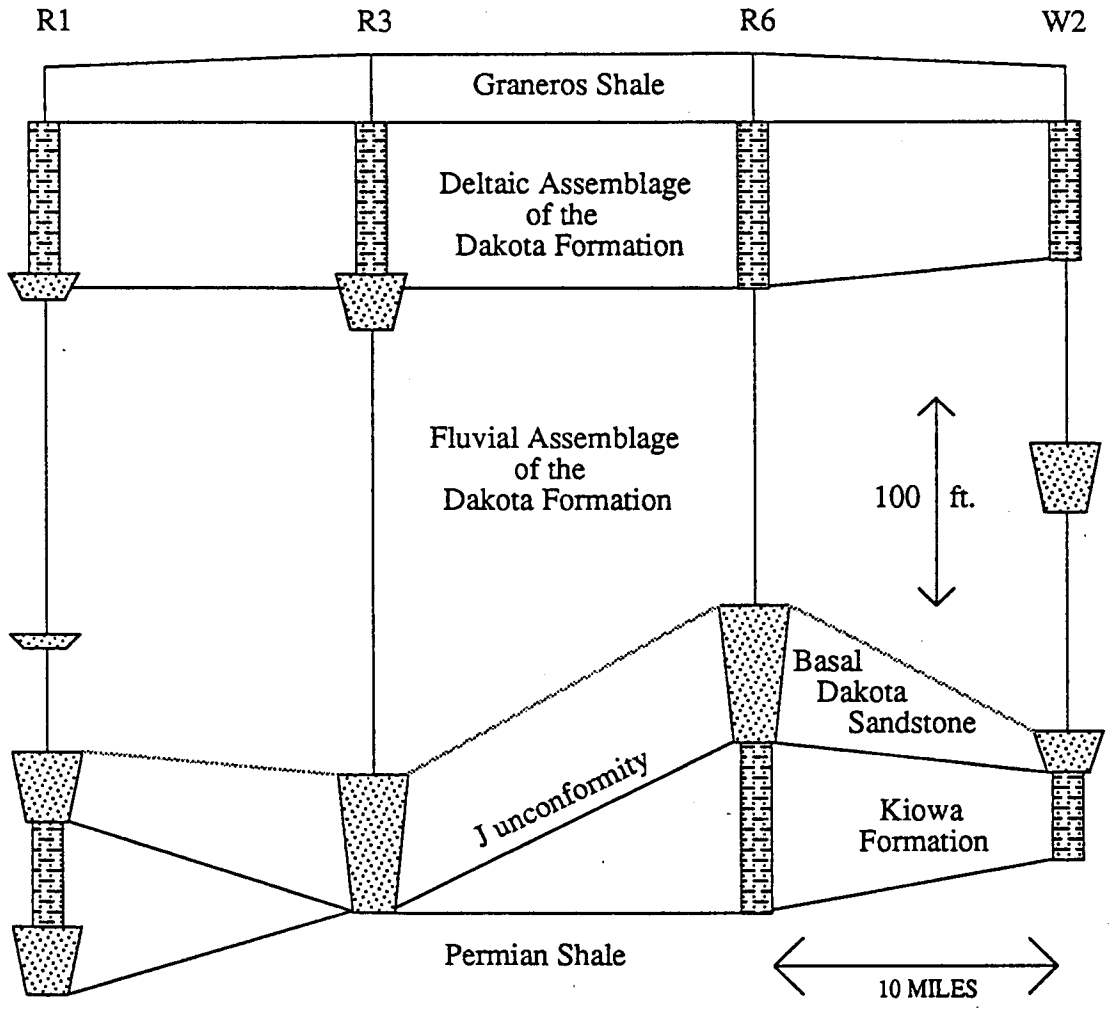


Figure 2.12. Section 1. A southwest-northeast section through the the strata of the Dakota aquifer from test hole R1 (T4S, R4W, Section 24DD) to test hole W2 (T1S, R2E, Section 10BCB). See Figure 2.2 for locations of test holes.

Channel sandstone

Interbedded sandstone and mudstone

Mudstone

The test holes are adjusted for differences in elevation above sea level so that the top of the Dakota formation is at the same level across the section. There is a difference in the thickness of the Dakota Formation of 82 ft. between R3 and R6. The basal sandstone of the Dakota Formation is present in all the test holes. Other sandstones are discontinuous. The Kiowa Formation has been completely removed by erosion on the J unconformity (Hamilton, 1989) in test hole R3.

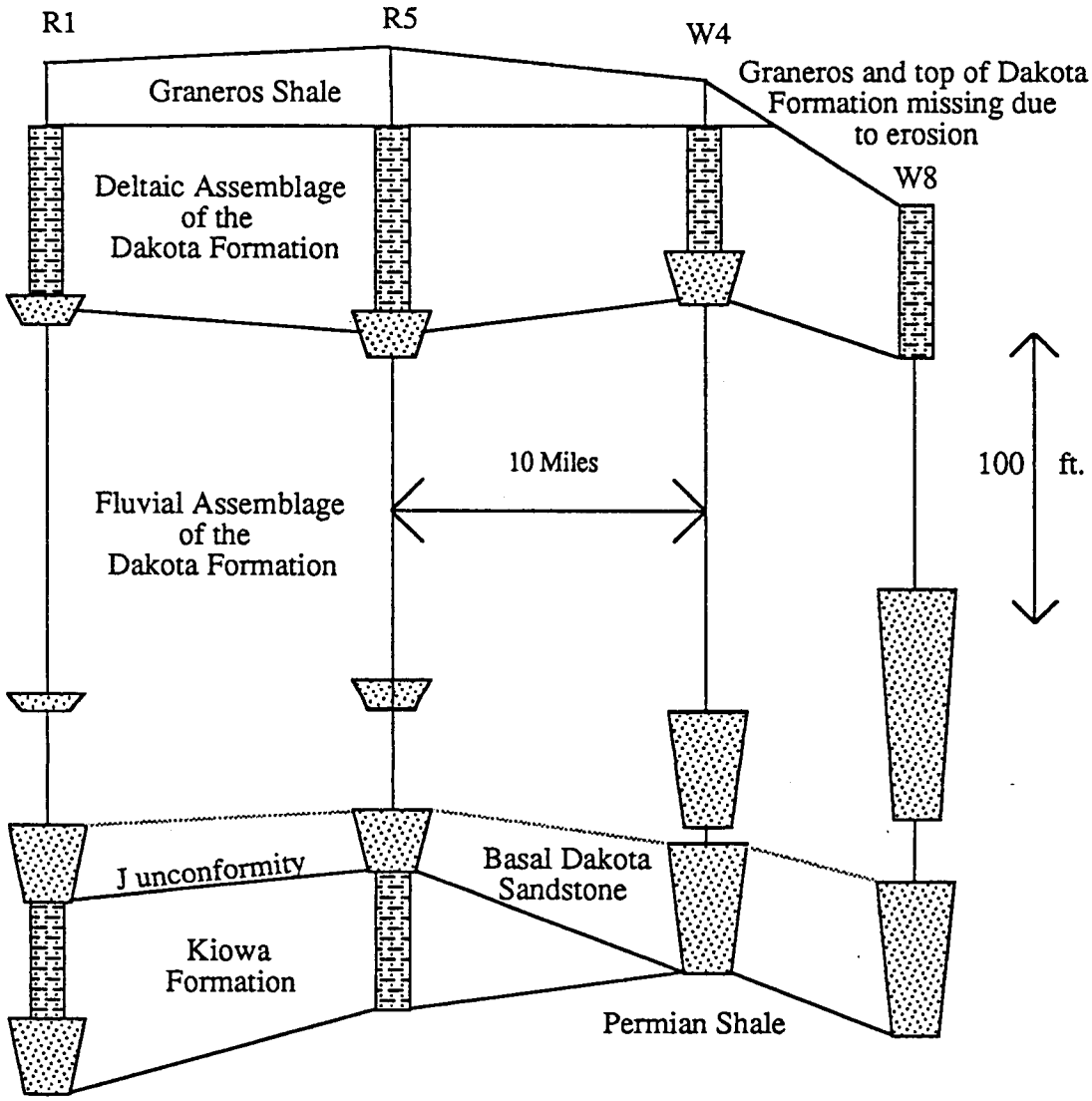


Figure 2.13. Section 2. A west-east section through the strata of the Dakota aquifer from test hole R1 (T4S, R4E, Section 24DD) to test hole W8 (T3S, R2E, Section 16CCAC). See Figure 2.2 for locations of test holes.

Channel sandstone

Interbedded sandstone and mudstone

Mudstone

The testholes are adjusted for differences in elevation above sea level so that the top of the Dakota aquifer is at the same level across the section. The Kiowa Formation has been removed by erosion on the J unconformity (Hamilton, 1989) at the eastern end of the section. The basal sandstone of the Dakota Formation is present in all the test holes.

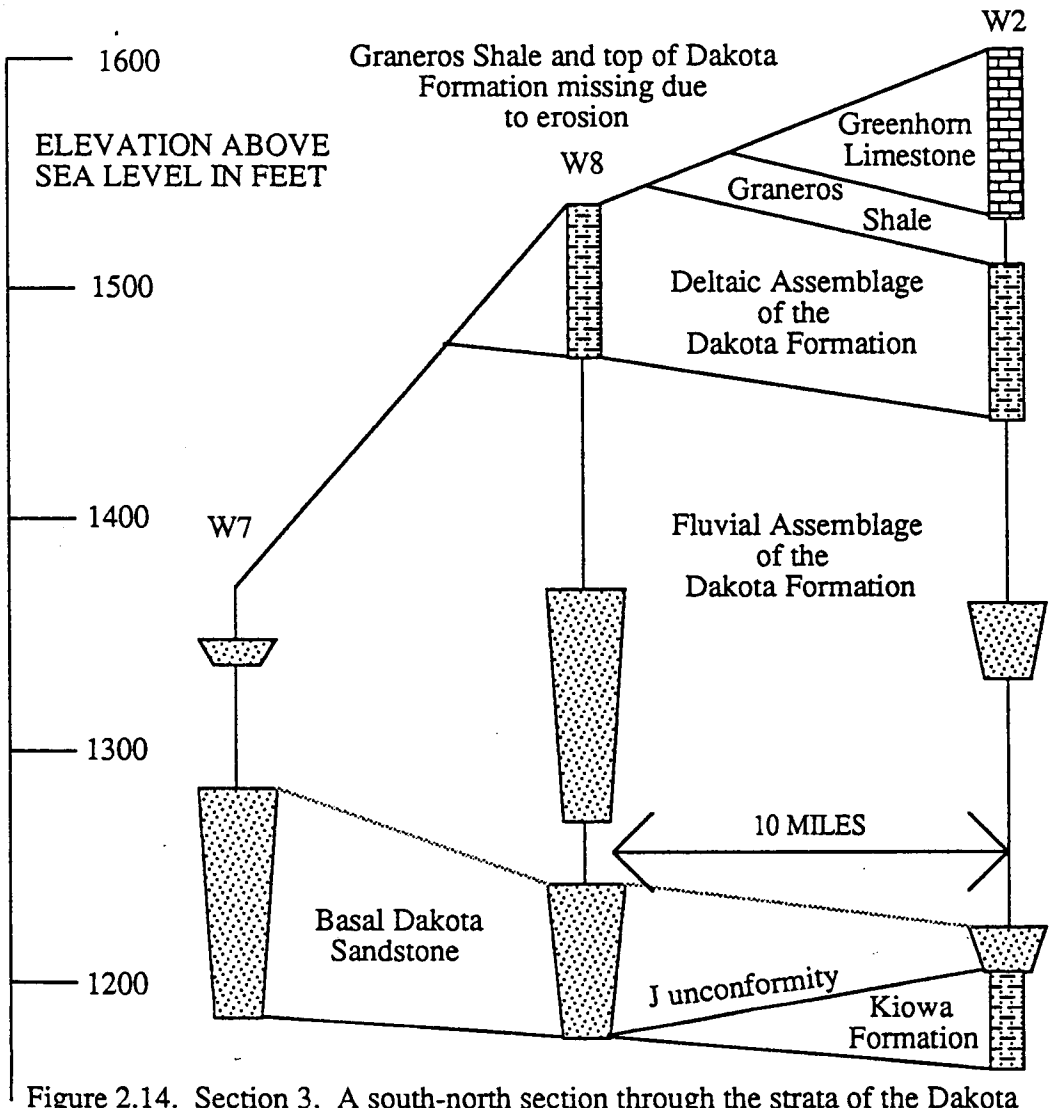


Figure 2.14. Section 3. A south-north section through the strata of the Dakota aquifer from test hole W7 (T5S, R1E, Sec. 10DDAB) to test hole W2 (T1S, R2E, Sec. 10BCB). See Figure 2.2 for locations of test holes.

Channel sandstone

Interbedded sandstone and mudstone

Mudstone

The basal sandstone ranges in thickness from 22 ft. in test hole W2 to 100 ft. in test hole W7. The Kiowa Formation has been removed by erosion on the J unconformity (Hamilton, 1989) at the southern end of the section.

Table 2.2. Thicknesses (ft.) of the major stratigraphic components of the Dakota aquifer in the test holes.

Test hole Number	Thickness of deltaic assemblage of the Dakota Formation	Thickness of fluvial assemblage of the Dakota Formation	Total Thickness of the Dakota Formation	Thickness of the Kiowa Formation	Elevation of Dakota-Graneros contact
R1	89	239	328	100	1423
W2	78	232	310	43	1509
R3	100	282	382	0	1415
W4	68	300	368	0	1550
R5	111	215	326	60	1485
R6	61	239	300	77	1457
W7	0[a]	186[a]	186[a]	0	not present
W8	66[a]	296	362[a]	0	not present
C9	40[a]	250[b]	290[b]	unknown	not present

a. This thickness has been reduced by erosion of the Dakota Formation in its outcrop belt.

b. This is not the full thickness because the test hole did not penetrate to the base of the formation.

2.2.4. Results and Interpretations

In the test holes, the aquifer consists only of the Dakota Formation and the Longford Member of the Kiowa Formation (figures 2.3 to 2.11). Cheyenne Sandstone, which constitutes part of the Dakota aquifer in much of the southern part of the state, and marine shale of the Kiowa Formation are not present. The following is a summary of observations and interpretations about each of the bedrock units penetrated by the drilling.

2.2.4.1. The Permian Wellington Formation

All the test holes except C9 (Figure 2.11) penetrated the Wellington Formation. The most distinctive feature of the Wellington Formation common to all the test holes was its great hardness compared to the Cretaceous sediments. In the cores from R1 and W2, the Wellington Formation was brecciated at the top and showed evidence of subaerial exposure (Appendices 2 and 3).

The most common lithology encountered in the Permian was bluish gray shale. Red shale and limestone were also encountered. White anhydrite was recovered in the cuttings of test hole R6; two 1-ft. beds of anhydrite at 454 ft. and 459 ft. show up well on the gamma-ray log (Figure 2.8).

2.2.4.2. The Kiowa Formation

This formation comprises only a minor proportion of the Dakota aquifer in the test holes. It consists mainly of interbedded sandstone and silty mudstone of the Longford Member which were deposited in a tidally-influenced deltaic environment. The Kiowa Formation is not present in test holes R3, W4, W7, and W8 (Figures 2.5, 2.6, 2.9 and 2.10 respectively), i.e., in western and southwestern Washington County, and central Republic County, nor is it present in outcrop in Washington County. The thickest Kiowa sediments discovered were in the southwestern corner of the study area in test hole R1 (Figure 2.3, Table 2.2). The formation is 100 ft. thick at this locality and includes 40 ft. of medium-grained, poorly-cemented sandstone, which is thought to have been deposited by tidal processes (Appendix 2). In the subsurface in other parts of Republic County the Kiowa Formation may be even thicker than this.

The Longford Member is 43 ft. thick in core hole W2 (Figure 2.4), which was drilled within 1.4 mi. of the Nebraska state line in northern Washington County (Figure 2.2). It is therefore likely that the Longford Member extends northward in the subsurface into Nebraska.

2.2.4.3. The Dakota Formation

The thickness of the Dakota Formation in test holes R1 through R6 ranged from 300 ft. to 382 ft. (Table 2.2). The top of the Dakota Formation dips to the west-northwest at approximately 10 ft./mi. Test holes W7, W8, and C9 were drilled in the outcrop area of the Dakota Formation (Figure 2.2), where the thickness of the

formation has been reduced by erosion. Test hole C9 did not penetrate to the formation base.

In the core holes (Figures 2.3 and 2.4), two distinct facies assemblages were recognized; a deltaic assemblage and a fluvial assemblage (Macfarlane et al., 1991a). A flooding surface was recognized above which sediments were deposited in the subaqueous portions of a river-dominated delta system and below which sediments were deposited by rivers in a coastal plain environment (Macfarlane et al., 1991a). This flooding surface was correlated among the other test holes.

2.2.4.3A. Fluvial Assemblage

Test holes R1 through R6 and W8 penetrate the complete fluvial assemblage. In these test holes this assemblage ranges in thickness from 215 to 300 ft (Table 2.2) and comprises approximately 75% of the full thickness of the formation.

Variegated mudstone, with some thin, very fine-grained levee and splay sandstones, constitutes approximately 60% of the fluvial assemblage in the test holes. These overbank mudstones are generally gray with red mottles which appear to follow large root traces in the cores (Appendices 2 and 3, Figures 2.3 and 2.4). There is a good exposure of mottled Dakota mudstone at the base of a cut bank of Parsons Creek, 1 mi. east of Clifton at T6S, R1E, sec. 1AAA. At this site, red mudstone is more resistant to erosion by the creek than gray mudstone. The root-shapes of the red mottles can be clearly seen where they physically stand out from the gray mudstone. I identified the upper limit of red mottling in test holes R3 through C9 by the depth at which red discoloration of the drilling mud was first encountered. Red mottles are rare close to the top of the fluvial assemblage.

The remainder of the fluvial assemblage consists mainly of fine to medium-grained sandstones (e.g. between 368 and 410 ft. in test hole R1, Figure 2.3) which commonly fine upward and contain large-scale cross beds (Appendices 2 and 3). These sandstones were deposited in river channels. Although they are normally poorly cemented in the subsurface, gray in color, and easy to drill, in outcrop the sandstones are well-cemented, orange in color, and comprise the majority of outcrops of the formation due to their resistance to erosion. In Washington County outcrops, I took fifteen measurements of dip directions of large-scale planar-tabular cross beds which ranged from 210° to 260°. Franks (1966) and Karl (1976)

reported similar dip directions in north-central Kansas and determined that the rivers which deposited these sediments drained in a west-southwesterly direction. Although the channel sandstones are generally relatively clean, channel abandonment resulted in the accumulation of poorly sorted, carbonaceous, sandy siltstone and mudstone in some fluvial channels (e.g. between 350 and 380 ft. in test hole W2, Figure 2.4).

The maximum grain size in the river-channel sandstones within the fluvial assemblage decreases upward. A fluvial channel sandstone is present on the J unconformity (Figure 2.2) at the base of the fluvial assemblage in all the test holes which penetrated to this depth. Sections 1, 2, and 3 show the persistence of this sandstone (Figures 2.12, 2.13 and 2.14 respectively). It is typically the coarsest and, in many of the test holes, the thickest sandstone of the aquifer (Figures 2.3, 2.5, 2.6, 2.7, 2.8, 2.9). It also consistently appears in drillers logs from water wells in Washington County. Hills immediately south of Mill Creek and east of Washington (Figure 1.1) are capped with this sandstone. Mudstone intraclasts are common in the basal sandstone recovered in core from test hole W2 (Appendix 3).

A laterally continuous body of sandstone seems an unlikely product of a fluvial system in which the dominant lithology is overbank mudstone. The period of erosion represented by the J unconformity (Hamilton, 1989) corresponds to a maximum in the rate of drop of the eustatic sea level. The first sediments to be deposited following this period of erosion, when the rate of reduction in the eustatic sea level slowed, were deposited by relatively high-energy, high-competence streams. They probably accumulated very slowly and underwent considerable reworking. There was therefore a high degree of interconnection between different channel sandstones and little preservation of lower-energy overbank mudstone. When the rate of eustatic sea-level rise approached a maximum, sediment accumulation accelerated, and overbank mudstones became volumetrically dominant over channel sandstones. The upward reduction in the maximum grain size of sandstones reflects the decrease in competence of rivers which deposited them as base level rose. Using sedimentation models, Cross (1987) predicted that channel sandstones would be vertically stacked and volumetrically subordinate to overbank mudstones during a period in which eustatic sea level was rising at a maximum rate. He also predicted that, in contrast, channel sandstones in a coastal plain environment would be laterally coalesced, and volumetrically dominant over

their associated overbank mudstones during a period in which the eustatic sea level was dropping.

The basal sandstone is likely to be thickest in paleovalleys of the J unconformity and at locations where stream channels later stacked on top of one another as the depositional slope lessened and overbank mudstone began to be preserved in greater volume between the channels. A comparison between test holes shows there is considerable variation in the thickness of the basal sandstone. For example, it ranges in thickness from 20 to 70 ft. in a distance of 10 mi. between test holes W2 and R6 (Figure 2.12).

The "J" unconformity at the base of the Dakota Formation is well defined in all the test holes. In the core taken from test hole W2, rotated blocks from the Kiowa Formation up to 4 ft. thick are preserved within the basal sandstone of the Dakota Formation, having slumped into a river channel which was being cut into the Kiowa Formation at this site (Appendix 3). The variation in thickness of the Dakota Formation from 300 ft. in R6 to 382 ft. in R3 (Table 2.2, Figure 2.12), 14 mi. away (Figure 2.2), is largely due to topographic relief which existed on the J unconformity surface before the Dakota Formation was deposited.

The position of the "D" unconformity (Hamilton, 1989) in these test holes is problematic. Although there is a medium-grained channel sandstone in the middle of the fluvial assemblage in some of the test holes, which could be attributed to a drop in base level and consequent rejuvenation of streams, it was not observed in all the test holes. Therefore, if this is the D unconformity, the continuity of any sandstone associated with it is limited. It is equally possible that the channel sandstones in the middle of the fluvial sequence are not all associated with an unconformity and are not connected to each other. There is no evidence in outcrop in this area of a continuous sandstone in the middle of the formation. This does not negate the existence of the D unconformity, which is not necessarily marked by a sandstone at the base of the D sequence. However, identifying the D unconformity is of questionable hydrostratigraphic value at the scale of this project.

The fluvial assemblage has many of the features of the Terra Cotta Clay Member originally defined by Plummer and Romary (1942) in outcrop in central Kansas. The flooding surface between the fluvial and deltaic assemblages is equivalent to the boundary between the Terra Cotta Clay and Janssen Clay Members.

2.2.4.3B. Deltaic Assemblage

In those test holes which penetrate it, the complete deltaic assemblage ranges in thickness from 61 ft. to 111 ft. (Table 2.2). It comprises approximately the upper 25% of the Dakota Formation.

Interlaminated mudstone and very fine to fine grained sandstone with a high organic content make up approximately 50% of the deltaic assemblage. The mudstone in this assemblage contains no red mottles. The core from R1 includes a typical, bioturbated, coarsening-upward sequence capped by lignite (Appendix 2 and Figure 2.3, 130-158 ft.) which is characteristic of the fill of an interdistributary bay in a river-dominated deltaic environment (Coleman and Prior, 1988).

Close to the base of the deltaic assemblage, fining-upward sequences rich in organic debris are typical (e.g. Appendix 2 and Figure 2.3, 158-172 ft.). These sequences were deposited in distributary channels, which are commonly abandoned and filled with fine-grained material in river-dominated deltas (Coleman and Prior, 1988). Clean sandstones deposited in active distributary channels were also encountered near the base of the deltaic assemblage (e.g. in R5, Figure 2.7).

The highest-energy conditions in a river-dominated delta are commonly on sand bars at the mouths of active distributary channels (Coleman and Prior, 1988). These distributary-mouth bars are areas of shoaling in which depositional rates are high and the sediments are constantly reworked. The cleanest, coarsest sandstones in the deltaic Dakota sediments, except for active distributary-channel sandstones, were deposited in these bars; they consist of fine-grained sand with a high organic content. Distributary-mouth-bar sandstones are likely to be more laterally extensive than the distributary-channel sandstones with which they are associated (Coleman and Prior, 1988). However, they are not very common in the test holes. The best example discovered was in hole R5 (Figure 2.7), where there is a 28 ft. sequence of well-sorted, fine-grained, distributary-mouth-bar sandstone.

The deltaic assemblage is commonly relatively sand-rich at its top and base and mud-rich in the middle (e.g. R5, Figure 2.7) due to the predominance of fining-upward distributary fill at the base and coarsening-upward delta-front deposits close to the top. The gamma-ray log in this case shows a pronounced curve through the deltaic sequence which is concave to the left.

The coarsest grain size encountered in the deltaic assemblage in the test holes is fine-grained sand. This is a consequence of the relatively low competence of the rivers supplying the sediment and the lower energy marine processes and limited marine-sediment supply.

The contact of the deltaic assemblage with the overlying Graneros Shale both in the test holes and in outcrop is an abrupt transgressive disconformity where delta-front deposits were overstepped by the Graneros sea. The abrupt nature of this contact is likely to be due to lobe switching rather than a sudden acceleration of regional sea-level rise. Although the coastline as a whole was moving landward, lobes associated with major distributaries were prograding. When a major distributary was abandoned, the supply of sediment was cut and diverted to begin forming a new lobe at another location. The old lobe was then submerged by the sea through the combined effects of sea-level rise, land subsidence and/or compaction.

The deltaic assemblage has many of the features of the Janssen Member originally defined by Plummer and Romary (1942) in outcrop in central Kansas (Figure 2.1). The high organic content and generally poor sorting are good evidence that the deltaic environment in which this assemblage was deposited was river dominated. Cross bedding and ripple laminations with a bimodal character, which are evidence of marine processes in the deltaic sediments in central Kansas (Siemers, 1971), were not observed in the area of this study.

2.2.4.4. The Graneros Shale and the Greenhorn Limestone

The Graneros Shale is a shallow marine fossiliferous gray shale which includes some slightly calcareous, very-fine, wavy-laminated sandstones and at least two bentonite beds (Appendices 1 and 2). It has a strong gamma-ray response, particularly in the bentonite beds. Plant fragments are rare. The boundary between the Dakota Formation and the Graneros Shale is easy to recognize in core (Figures 2.3 and 2.4) and from gamma-ray logs and cuttings (Figures 2.5 to 2.7).

The Greenhorn limestone overlies the Graneros Shale. It consists of interbedded gray, chalky limestone and calcareous shale as well as several bentonite beds. Its gamma-ray response is generally quite low (Figures 2.7 and 2.8).

2.3. Summary of the Hydrostratigraphy of the Dakota Aquifer in the Study Area

A hydrostratigraphic unit is defined as a portion of a formation, a formation, or a group of formations that forms a distinct hydrologic unit with respect to flow of ground water (Maxey, 1964). The Dakota aquifer was subdivided into hydrostratigraphic units on the basis of the geologic data presented above, in order to more easily analyze the flow system (Figure 2.15).

The flow of ground water through porous media is controlled by the hydraulic conductivity of the media. Poorly cemented sandstones have a hydraulic conductivity which is several orders of magnitude greater than that of mudstone (Freeze and Cherry, 1979). The primary aquifer units of the Dakota aquifer are the sandstones; these are the units which yield water to wells. Mudstones also contain water, indeed their porosity is likely to be greater than the porosity of the sandstones, and over a large area and a long time period they can transmit a significant quantity of water, but the hydraulic conductivity of the mudstone is insufficient to yield water to wells at a useful rate. The mudstones of the Dakota Formation are aquitards.

The coarsest and most poorly cemented sandstones in the aquifer are the channel sandstones of the fluvial assemblage of the Dakota Formation. These are therefore likely to have the highest hydraulic conductivities. However, most of these sandstone bodies are ribbon-shaped in plan view and isolated from each other by low-conductivity mudstone (Macfarlane et al., 1991b). An important exception to this in the area of this study is the basal sandstone of the Dakota Formation, the lateral continuity of which makes it the most important conduit for horizontal ground-water flow (Figure 2.15).

The Longford Member of the Kiowa Formation directly underlies the basal sandstone of the Dakota Formation in the subsurface throughout much of the study area (Figures 2.12 to 2.14) and, although it is unlikely to have a hydraulic conductivity as high as that of the basal sandstone due to its smaller grain size and higher mud content, it can be considered an aquifer unit rather than an aquitard. Where it occurs it effectively increases the thickness of the basal sandstone aquifer (Figure 2.15).

The fluvial assemblage above the basal sandstone is dominated by low-permeability mudstone containing isolated ribbons of channel sandstone. On the scale of the study area, this interval inhibits the flow of ground water and is

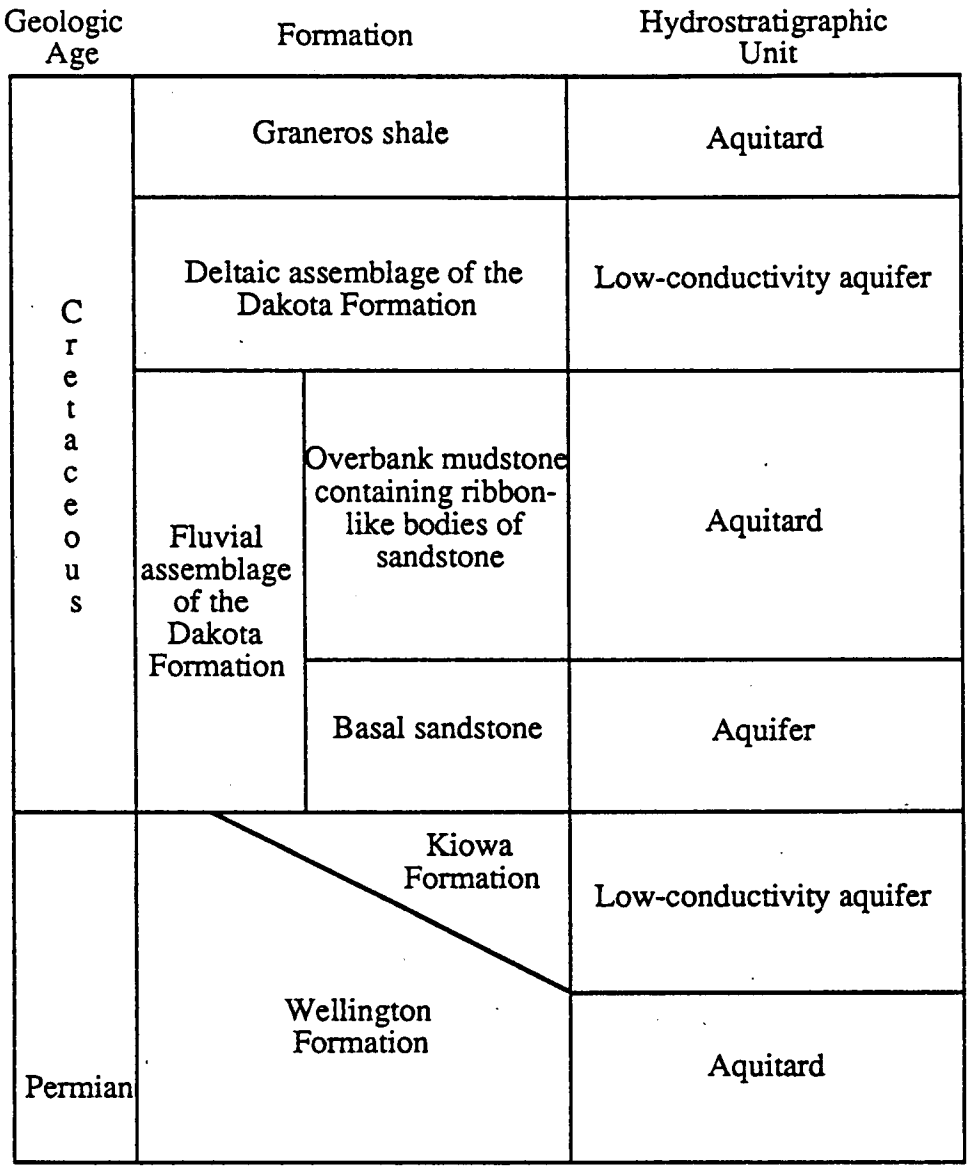


Figure 2.15. Hydrostratigraphy of the Dakota aquifer in the study area. Aquifers are units which permit appreciable amounts of water to move through them under ordinary hydraulic gradients, aquitards are sufficiently permeable to transmit water in significant quantities over a long time period and over large areas but not normally enough to yield water to wells. The aquitard in the middle of the Dakota Formation contains sandstones which yield water to wells but these sandstones are poorly interconnected. Therefore, on the scale of the study, the fluvial assemblage above the basal sandstone is classified as an aquitard because it retards flow.

therefore an aquitard (Figure 2.15), although wells screened in the sandstone bodies within this aquitard may yield significant quantities of water.

The deltaic assemblage of the Dakota Formation contains a significant proportion of sandstone. The cleanest sandstones, deposited in distributary mouth bars, are likely to have significantly greater lateral continuity than the channel sandstones typical of the fluvial assemblage. Interlaminated sandstone and mudstone is common elsewhere in the deltaic assemblage. It is therefore likely that on the scale of the study area the deltaic assemblage has a significantly greater horizontal than vertical hydraulic conductivity. Because of its sand content and degree of lateral continuity, the deltaic assemblage is an aquifer unit on the scale of the study area (Figure 2.15).

The Graneros Shale, which overlies the Dakota aquifer in the subsurface, is an aquitard which restricts the flow of ground water into and out of the aquifer. The Wellington Formation forms another aquitard immediately below the base of the Dakota aquifer (Figure 2.15).

CHAPTER 3. GEOHYDROLOGY OF THE DAKOTA AQUIFER

3.1. Introduction

The purpose of this chapter is to interpret the ground-water flow systems of the aquifer in the study area. Water-level measurements from wells, surface features of ground-water discharge, hydrogeochemical and topographic information were used to infer the directions of ground-water motion, to identify discharge zones and to differentiate between local and large-scale flow systems.

3.2. Previous Work

3.2.1. Regional Ground-Water Flow in the Dakota Aquifer

The earliest interpretation of regional flow in the Dakota aquifer was of an artesian basin in which lateral flow predominates over vertical flow (Darton, 1905). In this classic artesian aquifer scenario, the aquifer is recharged in outcrops in the high ground on the east flank of the Rocky Mountains and ground water flows eastward beneath the High Plains, discharging where the Dakota crops out in the eastern Great Plains. Darton (1905) also recognized that leakage through less permeable strata was a contributing factor to the flow system. Modelling of flow through the Dakota aquifer in South Dakota (Milly, 1978; Neuzil, 1980) has since shown that vertical flow both in and out exceeds lateral flow despite the fact that the horizontal hydraulic conductivity of the aquifer is several orders of magnitude greater than the vertical hydraulic conductivity of the confining layers. This is a result of the vast area which the confining shales cover in relation to the area of the sandstone outcrops. In large-scale studies including most of the U.S. Great Plains, Helgeson et al. (1982) and Belitz and Bredehoeft (1988) reported that hydraulic head data from drill-stem tests in the middle of this flow system are anomalously low. The reason for this under-pressuring is hydraulic isolation of the aquifer from its recharge zones (Belitz and Bredehoeft, 1988).

3.2.2. The Dakota Aquifer in Kansas

Common surface features in discharge zones of large-scale flow systems are seepages, springs, salt precipitates, relatively high mineralization of water, waterlogging, and the presence of phreatophytes (Toth, 1962). In central and north-

central Kansas, salt marshes which include all of these features of ground-water discharge have been the subject of scientific interest since the mid-nineteenth century, when the state geologist, Professor Mudge, described a salt marsh "sometimes called the Tuthill marsh" in southeast Republic County within the area of this project. Mudge (1866) measured the density of a sample of water from 4 ft. below the ground surface in the marsh and found the water to be "three times the strength of the ocean". A quarter of a century later, Hay (1891) revisited this marsh and reported a somewhat lower level of salinity than Mudge but one which was still similar to sea water.

Another well-known surface feature of ground-water discharge from the Dakota Formation now lies submerged beneath the waters of Waconda Lake, 40 mi. west-southwest of the study area of this project. Before the dam was built across the Solomon River valley near Glen Elder, in Mitchell County, the travertine cone of a major spring known as Great Spirit Spring rose approximately 35 ft. above the level of the Solomon River. The water discharging in 1954 had sodium and chloride contents of 6200 and 7700 ppm respectively (Swineford and Frye, 1955).

Macfarlane et al., (1991b) determined that halite-solution brine in the underlying Permian strata is the main source of salinity in the Dakota aquifer and that the Permian Cedar Hills Sandstone aquifer is the most likely conduit through which the brine enters the Dakota. The Cedar Hills Sandstone aquifer is hydraulically connected to the Dakota aquifer approximately 60 mi. west of the area of this project, in western Smith and Osborne counties (Macfarlane et al., Plate 10, 1990).

As well as the highly mineralized waters present in the Dakota aquifer, it has been estimated that up to 80 million acre-ft. of water containing less than 1000 mg/L of total dissolved solids (TDS) can be obtained form this aquifer in Kansas (Keene and Bayne, 1977). The aquifer is widely used in southwestern, south-central, and north-central Kansas (Macfarlane et al., 1990).

3.2.3. The Dakota Aquifer Within the Study Area of This Project

Most previous ground-water research in this study area has focused on the Republican River aquifer and the Quaternary Belleville Formation which is a large, unconfined gravel aquifer in northern Republic County along the Nebraska State Line (Fishel, 1948; Fader, 1968). However, Fishel also discussed the ground water

resources of the Dakota aquifer in Republic and northern Cloud counties. He reported saline water discharging from the Dakota Formation into streams and also noted that, as is the case elsewhere in north-central Kansas (Macfarlane et al., 1990), the salinity of ground water in the Dakota Formation increases with depth and distance into the subsurface from the outcrop belt. Fishel (1948) and Fader (1968) found that chloride concentrations of waters from wells in the Republican River alluvial aquifer north of Concordia were anomalously high because saline water from the Dakota aquifer discharges into the alluvial aquifer. Fishel (1948) recognized that local recharge allows the Dakota aquifer to yield an abundant supply of potable water in southeastern Republic and northern Cloud counties where the strata of the aquifer crop out.

Little previous work has been done concerning the ground-water resources of the Dakota aquifer in Washington County. However, Miller (1966) measured the chloride and pH of samples of water from sixty wells in northern Washington County, mostly within the outcrop belt of the Dakota Formation. In general the chloride contents were 300 mg/L or less.

3.3. Presentation of Data

3.3.1. Water-Level Database

In late winter 1990, I took 29 water-level measurements in Washington County, two in Cloud County and seven in Republic County with the help of J. Healey and T.J. Meehan of the Kansas Geological Survey (Figure 3.1, Table 3.1). At this time of year the aquifer had had time to recover from the seasonal effect of pumping during the irrigation season. We measured depths to water in domestic, irrigation, stock and municipal wells using chalked steel tape, which requires only a small well opening. I supplemented this data with depths to water taken mainly from "WWC-5" forms filled out by well drillers for the Kansas Department of Health and Environment (KDHE) (Figure 3.2, Appendix 4). Most of these water levels were documented in the 1970s and 1980s. There was no systematic difference in depths to water from wells which appear in both databases. Such a difference might be indicative of a depletion of the aquifer by pumping.

The hydraulic head at a point in an aquifer is a measure of the potential energy per unit weight of the ground water at that point (Freeze and Cherry, 1979).

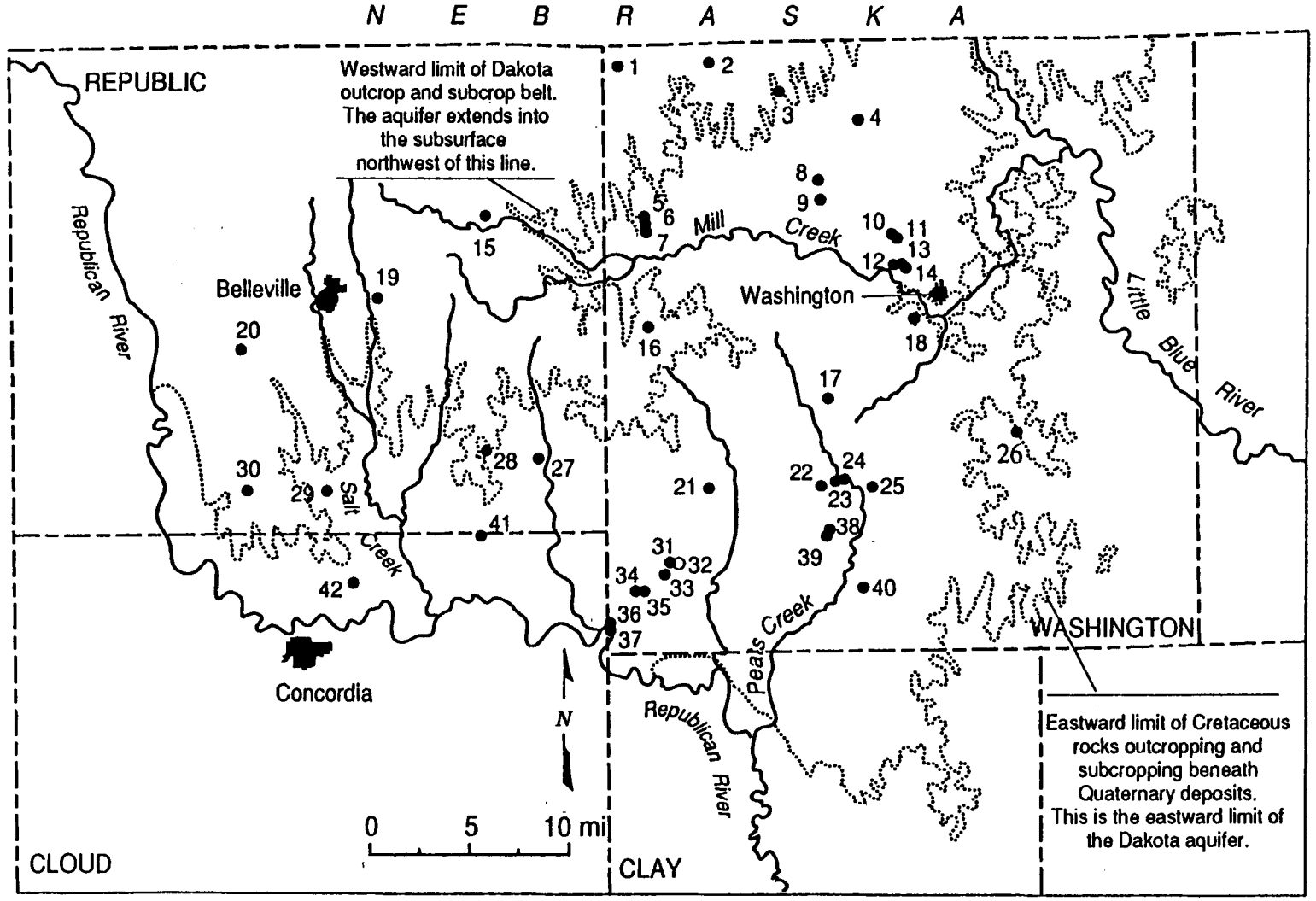


Figure 3.1. Locations of 1990 KGS water-level measurements from wells in the Dakota aquifer. Well number 32, depicted with an open circle, was used in the pumping test (Chapter 4).

Table 3.1. Water levels in wells measured by the KGS in 1990.
All the wells listed are screened in the Dakota aquifer. The locations are shown in Figure 3.1.

Number	Legal location	Farmer or municipality	Land-surface elevation above sea level (ft.)	Well depth (ft.)	Depth to water (ft.)	Hydraulic head above sea level (ft.)	Date of measurement
1	1S 1E 01DCCC	Gaydusek	1590	262	192	1398	03/21/90
2	1S 1E 07AAAC	City of Mahaska	1601	285	197	1404	03/21/90
3	1S 2E 15BCD	Huber	1516	120	74	1442	03/20/90
4	1S 3E 20CCC	Miller	1440	160	110	1330	03/21/90
5	2S 1E 21BABB	City of Haddam	1445	76	53	1392	02/15/90
6	2S 1E 21BABB	City of Haddam	1448	87	55	1393	02/15/90
7	2S 1E 21BACB	City of Haddam	1470	101	74	1396	02/15/90
8	2S 2E 01CCCB	Rosenthal	1435	200	112	1323	03/21/90
9	2S 2E 12CCD	Rosenthal	1435	80	45	1390	03/21/90
10	2S 3E 21D	Stigge	1372	147	49	1323	03/21/90
11	2S 3E 21DDBD	Stigge	1364	145	43	1321	03/21/90
12	2S 3E 33ABAA	City of Wash.	1360	120	61	1299	03/20/90
13	2S 3E 34BBB	Wash. RWD #3	1358	114	52	1306	03/20/90
14	2S 3E 34BBC	Wash. RWD #3	1362	118	61	1301	03/20/90
15	2S 1W 18BCC	Peters	1532	178	100	1432	03/28/90
16	3S 1E 16BAAB	Nat. Gas Pipe.	1590	340	203	1387	03/21/90
17	3S 2E 36C	Wilgers	1483	230	125	1358	03/22/90
18	3S 3E 10DCAB	Stigge	1330	48	21	1309	03/21/90
19	3S 2W 06ACB	Snapp	1500	138	85	1415	03/28/90
20	3S 4W 13DCD	Dooley	1528	142	89	1439	03/28/90
21	4S 1E 25BBAB	Carlson	1405	30	16	1389	03/21/90
22	4S 2E 23DDD	Herrs	1445	unknown	91	1354	03/20/90
23	4S 2E 24DAB	Herrs	1390	103	46	1344	03/20/90
24	4S 2E 24DBC	Herrs	1420	142	75	1345	03/20/90
25	4S 3E 29BAA	City of Linn	1425	124	72	1353	03/20/90

Table 3.1 (continued)

Number	Legal location	Farmer or municipality	Land-surface elevation above sea level (ft.)	Well depth (ft.)	Depth to water (ft.)	Hydraulic head above sea level (ft.)	Date of measurement
26	4S 4E 09DAA	Hoover	1390	122	40	1350	03/20/90
27	4S 1W 16ACC	City of Agenda	1413	140	70	1343	01/20/90
28	4S 2W 12DDD	Benyshek	1450	112	29	1421	03/27/90
29	4S 3W 22DDA	Parker	1420	90	52	1368	03/27/90
30	4S 4W 24DAD	Kenyon	1485	154	57	1428	03/27/90
31	5S 1E 10C	Leiszler	1370	156	77	1293	03/22/90
32	5S 1E 10DBC	Leiszler	1373	189	73	1300	03/22/90
33	5S 1E 16AAD	Condray	1340	178	34	1296	03/22/90
34	5S 1E 17DCCD	Condray	1340	162	66	1274	03/22/90
35	5S 1E 17DDCC	Condray	1340	154	62	1278	03/22/90
36	5S 1E 30CBC	Pierce	1322	103	56	1266	03/27/90
37	5S 1E 30CCC	Leiszler	1286	113	20	1266	03/27/90
38	5S 2E 01B	Wilgers	1440	135	72	1368	03/20/90
39	5S 2E 01BC	Wilgers	1450	unknown	89	1361	03/20/90
40	5S 3E 18DDD	Voelker	1380	41	26	1354	03/20/90
41	5S 2W 01BAC	Olson	1375	230	48	1327	03/27/90
42	5S 3W 13BCD	Keil	1390	66	35	1355	03/21/90

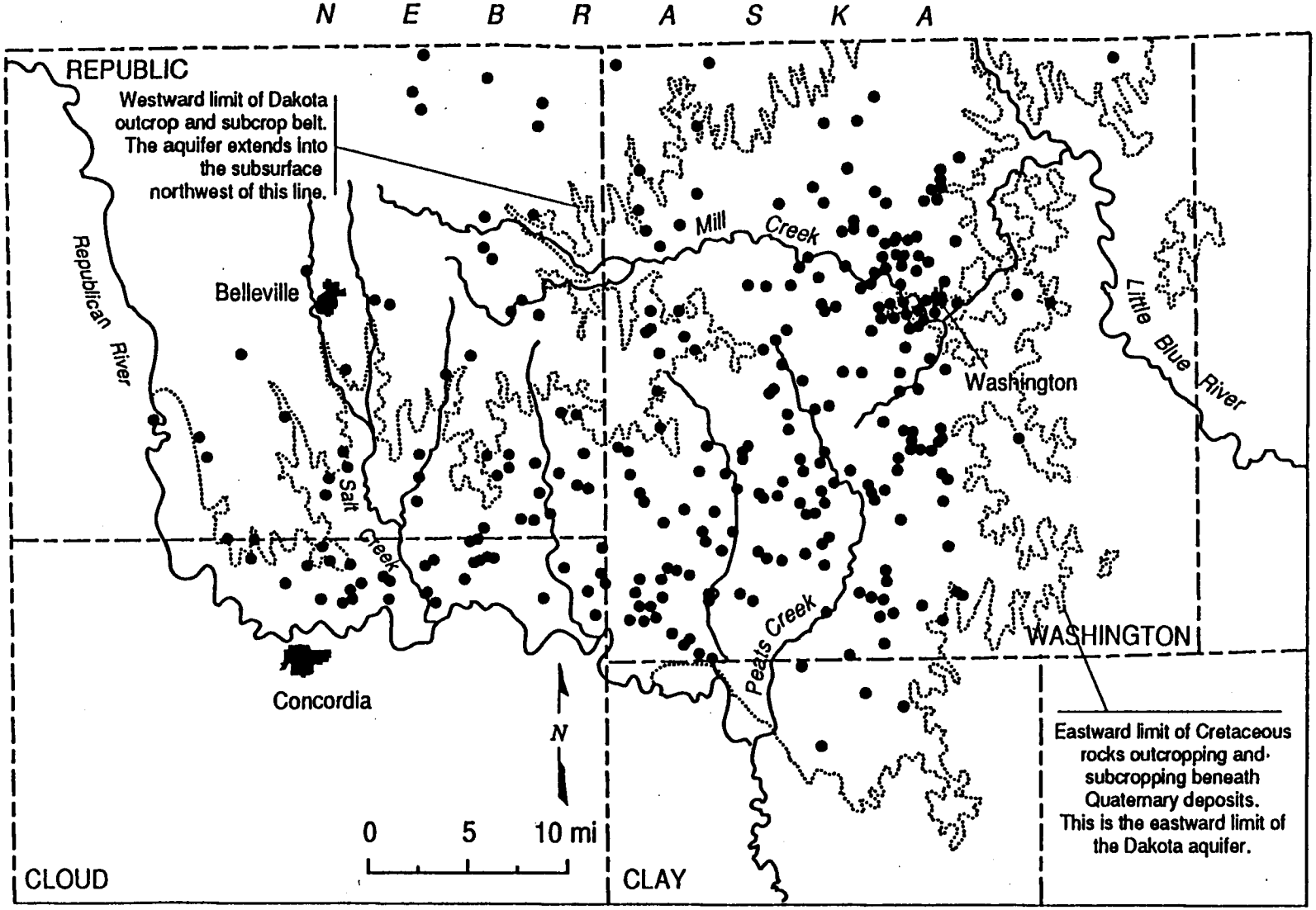


Figure 3.2. Locations of water-level measurements existing prior to the 1990 survey. These are listed in Appendix 4.

It is easily determined as the elevation of the water level in a well above a datum which is normally taken to be sea level (Figure 3.3). Hydraulic head is given by $h = z + \Psi$, where z is the elevation of the measuring point above the datum, and Ψ is the pressure head (Figure 3.3). The total energy of the water also includes a kinetic component, but this is negligible in relation to z and Ψ, the elevation and pressure components, due to the very low velocities involved in ground-water flow (Freeze and Cherry, 1979).

I determined the hydraulic head at each of the well sites by subtracting the depth to water from the elevation of the measuring point (Table 3.1, Appendix 4). The largest potential source of error in the head values is the estimate of the measuring-point elevation made from 1:24,000 $7\frac{1}{2}$ -minute topographic maps contoured in 10-ft. intervals. Errors in depth measurements were insignificant in relation to inaccuracies in the measuring-point elevations. Locations of wells measured by the KGS were field checked but, despite this, errors of up to 5 ft. in measuring-point elevation are possible because of the contour interval on the maps and the slope of the land.

Heads determined from the WWC-5 forms were generally much less accurate than the KGS water levels for the following reasons: 1. Well locations given on the forms are commonly sufficiently inaccurate that an error of up to 20 ft. is introduced in land-surface elevations and therefore hydraulic heads. Most of these locations were not field checked. 2. The heads are affected by seasonal and annual fluctuations because the measurements were made at different times of year over a period from the 1960s to the 1980s. 3. It is likely that in some cases the water level in a newly constructed well had not reached equilibrium with the aquifer water level at the time of measurement.

3.3.2. Water-Chemistry Data

Water samples from wells were not collected for this study. However, I compiled chloride (Cl) and sulfate (SO_4) data mainly from the Kansas water-chemistry database at the KGS (KWATCHEM, 1991) and from other sources (Figure 3.4, Table 3.2). As well as the quantitative information documented in Tables 3.1 and 3.2 and Appendix 4, the knowledge of local farmers was very useful in developing an understanding of the ground-water flow systems and is referred to later in this chapter.

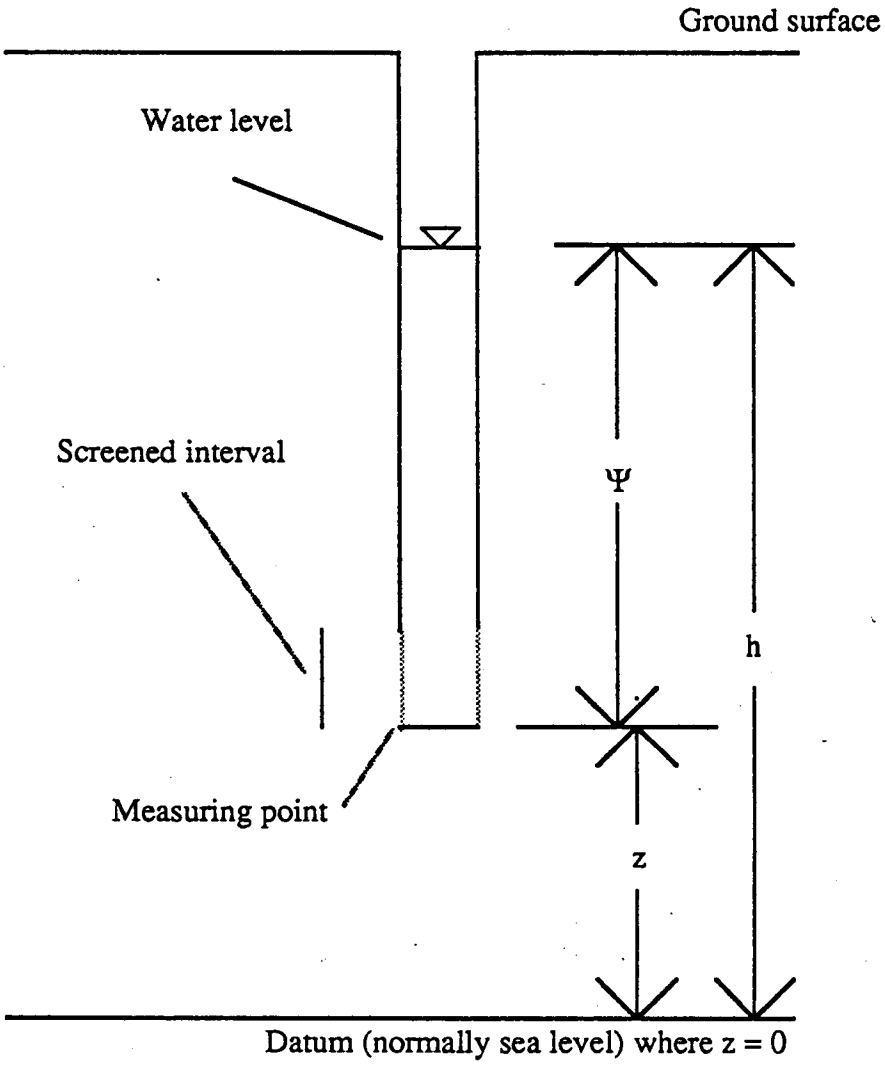


Figure 3.3. Hydraulic head in a well is the sum of the elevation head, z, and pressure head, Ψ. In most cases the elevation of the base of the well is a good approximation of the elevation head.

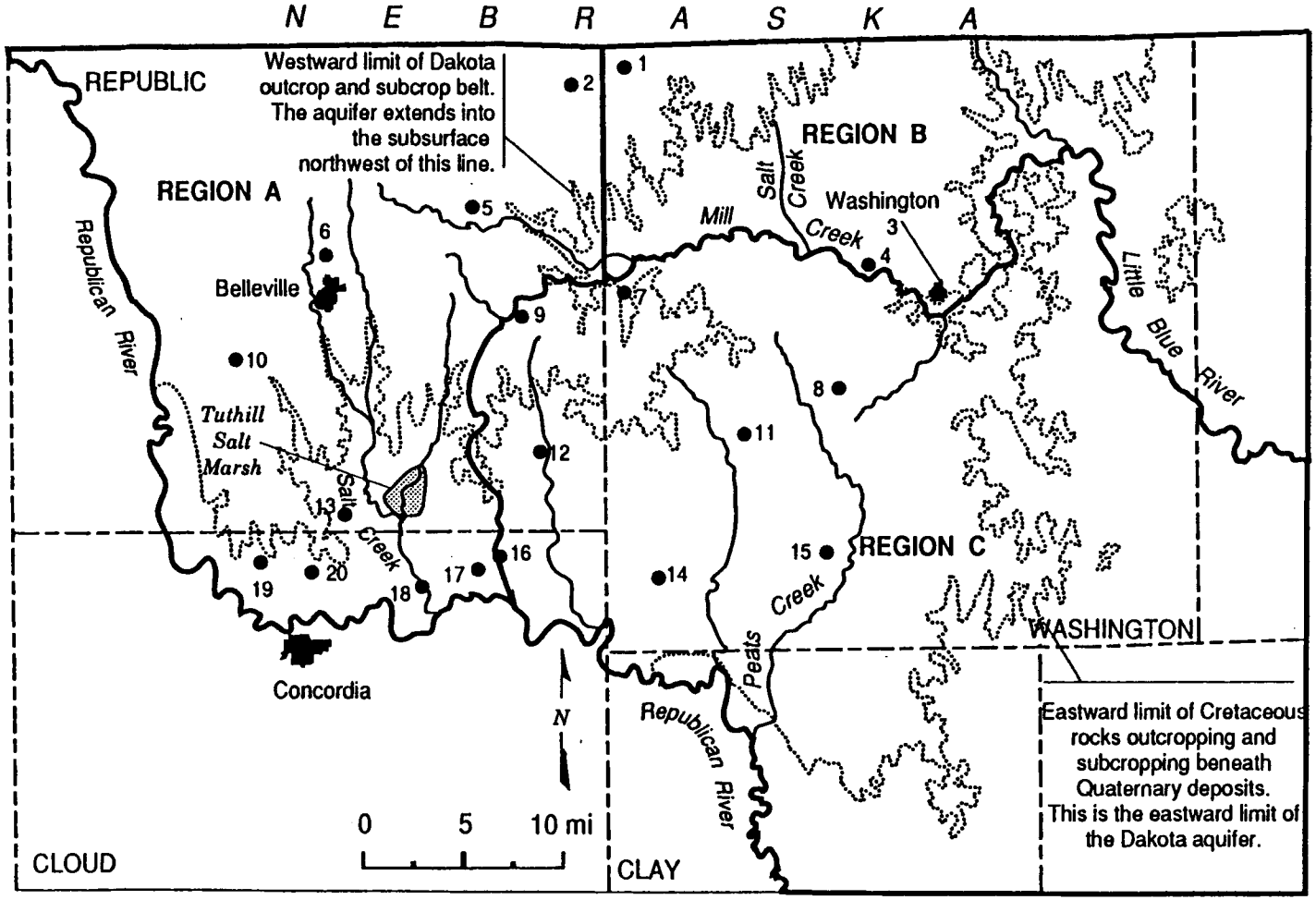


Figure 3.4. Water-sampling locations. All the samples were taken from wells screened in the Dakota aquifer except number 18, which was taken from Salt Creek. The samples are tabulated in Table 3.2. The study area was divided into 3 regions, described later in this chapter, on the basis of hydrogeochemical differences.

Table 3.2. Hydrogeochemical Data.
All the wells listed are screened in the Dakota aquifer. The locations are shown in Figure 3.4.

Number	Location	Depth (ft.), if well	Cl (mg/L)	Br/Cl	SO_4 (mg/L)	Source of data	Date of sampling
1	T1S R1E 07AA	280	141	-	115	KWATCHEM (1991)	11/22/55
2	T1S R1W 15CC	253	214	-	188	KWATCHEM (1991)	02/04/42
3	T2S R3E 21D	147	9.0	-	4.0	KWATCHEM (1991)	08/03/89
4	T2S R3E 32ABB	120	7.6	-	20	KWATCHEM (1991)	07/06/88
5	T2S R1W 18BCCB	100	75	-	233	KWATCHEM (1991)	06/05/90
6	T2S R3W 27CC	157	4000	-	540	KWATCHEM (1991)	07/30/42
7	T3S R1E 06AAA	-	3500	-	-	Miller (1966)	summer, 66
8	T3S R2E 36DDCC	181	12	-	31	KWATCHEM (1991)	06/05/90
9	T3S R1W 08C	217	26	-	143	KWATCHEM (1991)	05/23/42
10	T3S R4W 13DCDD	142	1410	-	390	KWATCHEM (1991)	06/05/90
11	T4S R2E 08DCC	125	38	-	29	KWATCHEM (1991)	07/07/81
12	T4S R1W 16ACC	100	11	-	47	KWATCHEM (1991)	06/10/87
13	T4S R3W 35AD	50	14	-	82	KWATCHEM (1991)	08/15/90
14	T5S R1E 16AAA	178	32	-	13	KWATCHEM (1991)	08/01/89
15	T5S R2E 12CBA	151	9.7	-	14	KWATCHEM (1991)	07/06/88
16	T5S R1W 07BBAB	138	13	-	16	KWATCHEM (1991)	05/03/91
17	T5S R2W 12ABBD	138	226	-	128	KWATCHEM (1991)	08/16/90
18	T5S R2W 16DDD	Salt Creek	424	2.43E-4	-	Whittemore, personal communication (1991)	02/26/91
19	T5S R3W 06CCCD	85	24	-	40	KWATCHEM (1991)	08/15/90
20	T5S R3W 15AB	341	16,000	-	-	Fader (1968)	10/18/55

3.4. Methods of Analysis

3.4.1. Correlation Between Pressure Head and Well Depth

In an ideal piezometer, which is in hydraulic connection with the aquifer only at its base, the elevation head is the elevation of the base of the piezometer and the pressure head is the height of the column of water. In a typical well, which is gravel packed for most of its depth, the elevation of the base of the well is not necessarily the elevation of the measuring point. However, in most cases the elevation of the base of the well is a good approximation of the elevation head (Figure 3.3).

A plot of pressure heads against well depths can be constructed for an aquifer and the value of the slope of the best-fit straight line can be determined. If the slope is greater than one, there is an upward hydraulic gradient and so the potential for vertical flow is upward (Fogg et al., 1983), which is a characteristic of discharge zones (Toth, 1972).

In recharge zones, hydraulic heads decrease with increasing well depth (Toth, 1972). If the slope of a plot of pressure heads against well depths is less than one, there is a downward hydraulic gradient and the potential for vertical flow is downward (Fogg et al., 1983). If there is no tendency for vertical movement through the aquifer, the slope will equal one, the slope of the hydrostatic line (Fogg et al., 1983). The degree of linear correlation between pressure heads and depths can be determined using least-squares linear regression analysis.

Pressure head was plotted against depth using well data from Table 3.1 (Figure 3.5) and Appendix 4 (Figure 3.6) to determine the slope and the correlation coefficient, r, in each case. For each well, the elevation of its base was assumed to be a good approximation of the elevation head; the pressure head was determined by subtracting the depth to water from the well depth. In both sets of data the wells are distributed throughout most of the study area, but are concentrated in the outcrop belt.

3.4.2. Generation of a Potentiometric Surface

If the water-level elevations of wells screened in a confined aquifer are plotted and contoured, the resulting contour surface of the hydraulic head of the

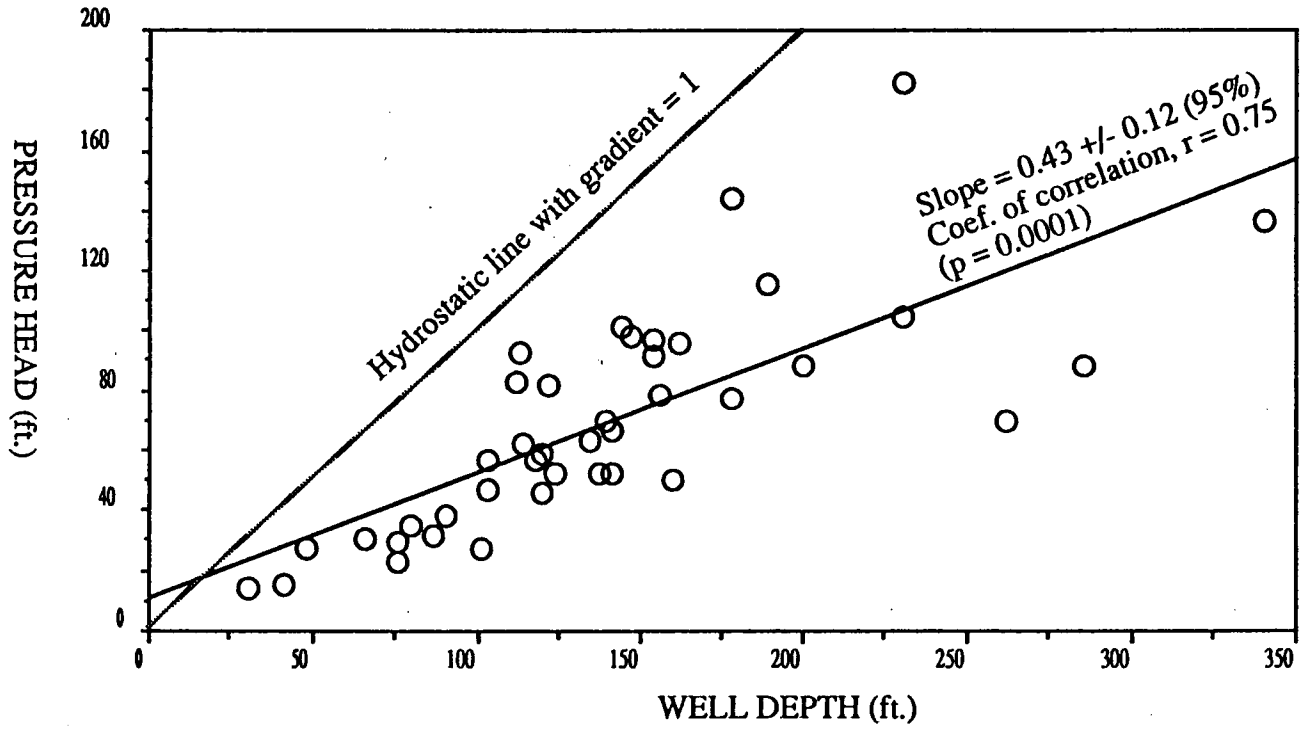


FIGURE 3.5. Relationship between pressure head and well depth for the data listed in Table 3.1. Each circle is represents a different well. The slope is less than 1, indicating a potential for downward flow. The pressure head of each well was determined by subtracting the depth to water from the depth of the well. Two wells on the subcrop side of the upper edge of the outcrop belt , in T1S R1E, both plot much below the line showing that the tendency for downward flow here is particularly strong.

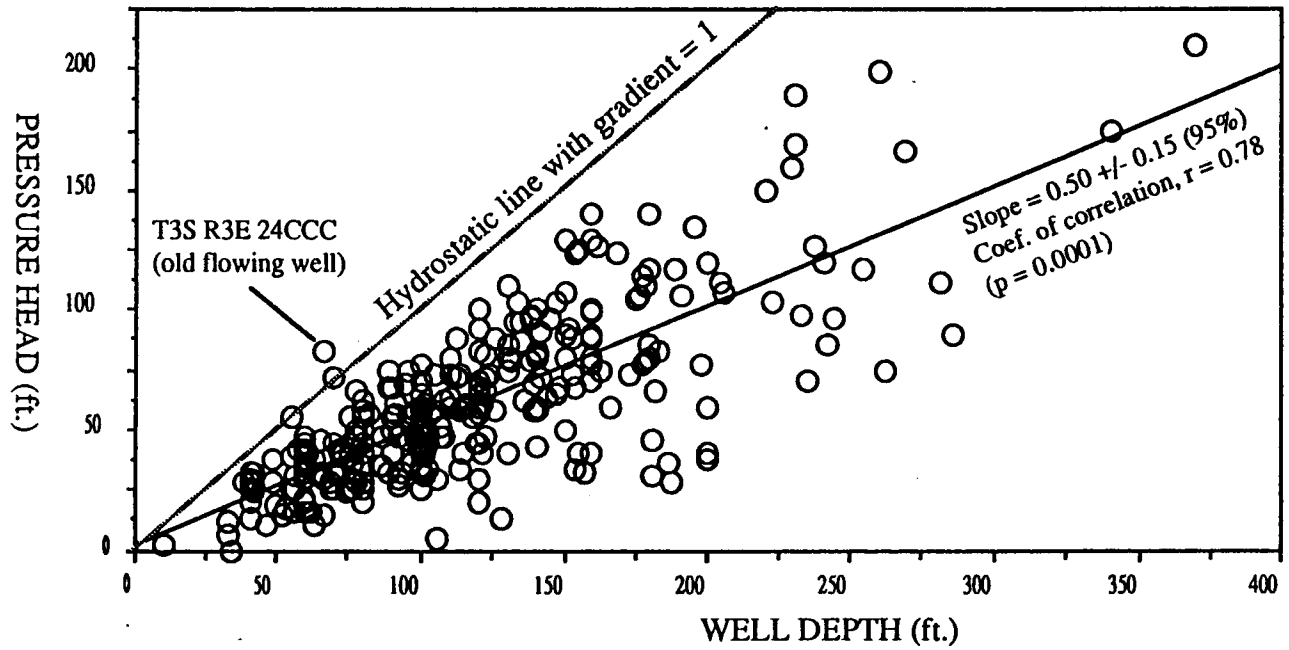


FIGURE 3.6. Relationship between pressure head and well depth for the data listed in Appendix 4. Each circle represents a different well. The slope is less than 1, indicating a potential for downward flow. The pressure head of each well was determined by subtracting the depth to water from the depth of the well.

aquifer is termed the potentiometric surface (Freeze and Cherry, 1979). Ground water in an aquifer flows from areas of high hydraulic head to areas of low hydraulic head, in a direction perpendicular to the equipotential lines, assuming the aquifer is homogeneous and isotropic. Therefore, the potentiometric surface can be a useful aid to understanding two-dimensional ground-water flow directions and to help identify zones of discharge and recharge as well as ground-water divides. The potentiometric surface is a valid concept only in horizontal aquifers with horizontal flow (Freeze and Cherry, 1979).

In the Dakota aquifer of north-central Kansas, the basal sandstone is the most important and most continuous conduit for lateral flow of ground water in the aquifer (Chapter 2). I therefore constructed a potentiometric-surface map for this sandstone (Figure 3.7) mainly using the KGS water-level data (Figure 3.1, Table 3.1) because of its greater accuracy but also using "WWC-5" water levels (Figure 3.2, Appendix 4) in areas where KGS data was limited. I referred to 1:24,000, $7\frac{1}{2}$ -minute topographic maps to help identify discharge areas by locating surface features of ground-water discharge such as the Tuthill Marsh and Mormon Spring (Figure 3.7). These maps were also useful in estimating the elevation of the potentiometric surface in low-lying discharge zones. In the lower reaches of several drainage basins, for example Mill Creek, Peats Creek and Parsons Creek, the elevation of the creek bed was used as the elevation of the potentiometric surface (Figure 3.7). The potentiometric surface was not accurate to better than ± 10 ft. due to the poor accuracy of much of the hydraulic head data.

There were problems associated with considering only the basal sandstone when constructing the potentiometric surface. In the western part of the study area the basal sandstone is deeply buried, and is therefore penetrated by few wells. Hydraulic heads in this area were inferred from a knowledge of heads higher in the aquifer combined with an estimate of vertical head gradient and/or from physiographic features at the land surface such as the Tuthill Marsh. In discharge zones there is an upward vertical component to the hydraulic gradient (Toth, 1972). The elevation of the water level in the basal sandstone below this watershed was therefore estimated to be 10 ft. higher than the elevation of Salt Creek in the vicinity of Tuthill Marsh (Figure 3.7).

The potentiometric surface was used to infer likely zones of ground-water discharge where surface features of discharge were not apparent. It was also used in

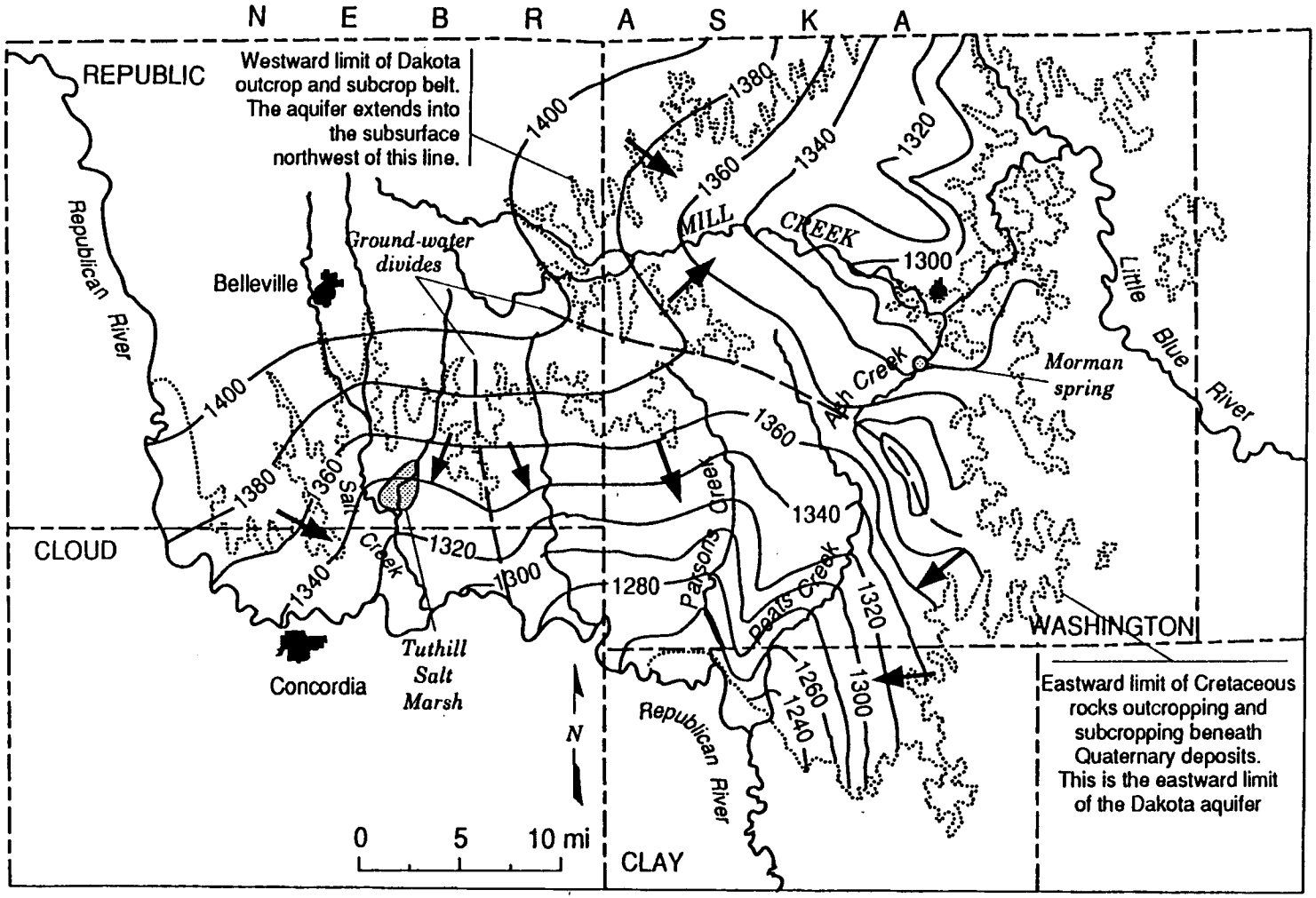


Figure 3.7. The potentiometric surface elevation of the basal sandstone aquifer of the Dakota Formation constructed from water-level measurements (Figure 3.1, Table 3.1, and Figure 3.2, Appendix 4), topographic information and surface features of ground-water discharge such as the Tuthill Salt Marsh. Contours are in feet. Arrows indicate the direction of groundwater flow.

conjunction with hydrogeochemical data to determine the nature of the flow systems in the basal sandstone and to attempt to differentiate between locally recharged flow systems and large-scale flow systems originating outside the study area. It was found that ground-water flow was best described by dividing the area into the three regions, A, B, and C based mainly on differences in hydrogeochemistry (Figure 3.4). Region A consists of most of Republic County and a small part of Cloud County, Region B is that part of Washington County north of Mill Creek and Region C is the remainder of the study area. For most of its length the boundary between Regions A and C follows the divide along the eastern edge of the Salt Creek watershed (Figure 3.4). In each region, the aquifer thins from west to east (Chapter 2). The characteristics of each of the regions are described below.

3.5. Results and Interpretations

3.5.1. Pressure Head/ Well Depth Plots

The correlation coefficient is quite low in Figures 3.5 and 3.6 reflecting scatter of the data because the wells are located at different points along flow lines. Points close to the hydrostatic line are few and the slope of the best-fit line is considerably less than one in each case. This shows that there are many more wells located in recharge areas than in discharge areas, which in turn is evidence that areas of discharge are limited in relation to areas of recharge. Therefore, the potential for vertical flow is downward throughout most of the study area. Areas of ground-water discharge occupy between 5 and 30% of the surface area of typical watersheds (Freeze and Cherry, 1979).

3.5.2. Interpretation of the Potentiometric Surface With Reference to the Hydrogeochemistry of the Aquifer

As expected, the potentiometric surface is a subdued form of the topography (Figure 3.7). The relief on the potentiometric surface is approximately 150 ft. between the uplands of west-central Washington County and the southern boundary of the aquifer at the Republican River valley. This compares to a relief on the land surface of approximately 300 ft. between these locations.

3.5.2A. Region A. Republic and Cloud Counties Northeast of the Republican River

The Dakota outcrop belt covers much of the northern part of Cloud County and extends northward into Republic County up the valleys of tributaries of the Republican River. The Dakota aquifer is deeply buried in the northwestern part of Region A. Potentiometric-surface contours were not drawn here due to the lack of data (Figure 3.7).

The potentiometric surface slopes toward the Republican River valley in the southern part of Region A (Figure 3.7). This corroborates hydrogeochemical evidence of Fishel (1948) and Fader (1968) that ground water is discharging from the Dakota into the Republican River aquifer. Where the Republican River valley runs north-south at the western boundary of the study area, there are no reports of chloride concentrations greater than 120 mg/L in the Republican River aquifer. Due to the northwesterly dip of the strata of the Dakota aquifer (Chapter 2) and the downstream slope of the river valley, the Dakota aquifer is not in good hydraulic connection with the Republican River aquifer this far upstream.

The "Tuthill marsh" (Mudge, 1866), located in the southwest corner of T4S R2W in the Salt Creek watershed, consists of salt marshes with phreatophytes, salt precipitates, and highly mineralized water (Figure 3.4). There are also records of wells with static water levels above land surface within the marsh area (T4S R2W, Sections 21 and 28, Appendix 4). These are all good evidence of ground-water discharge (Toth, 1962). The base of the basal sandstone is approximately 150 ft. below Salt Creek. Saline water may flow up through alluvium and sandstones which are hydraulically connected with the basal sandstone.

A sample of water from Salt Creek taken at T5S R2W 16DDD during low flow in February, 1991 (Figure 3.4, No. 16), had a chloride (Cl) content and bromide/chloride (Br/Cl) ratio (Table 3.2) which plot in the mixing zone between freshwater and halite-dissolution brine (D.O. Whittemore, personal communication, 1991). This is strong evidence that the salinity of the ground water discharging into the marsh was derived from dissolution of evaporites. The Permian strata which underlie the Dakota aquifer in the study area are relatively impermeable and contain small amounts of evaporites (Chapter 2) relative to Permian strata subcropping further west (Macfarlane et al., 1990). It is likely that this water entered the Dakota aquifer from Permian strata west of the study area and flowed eastward to the outcrop zone, where it mixed with locally recharged water.

Water-quality data are naturally biased in favor of the regions of usable ground water because wells which yield saline water are generally abandoned or plugged. There are therefore limited data from the basal sandstone in this region. Fader (1968) reported a sample from a deep well (sample 20, Figure 3.4, Table 3.2) which must have been screened in the basal sandstone. A water sample taken from this well in 1955 had a Cl content of 16,000 mg/L indicating that this water is not locally recharged but is part of a regional flow system. In contrast, 12 mi. east of this site, the water from the basal sandstone is sufficiently fresh to be used for irrigation.

Samples 2, 5, 6, 9, and 10 (Table 3.2) were taken from upper-Dakota wells in Republic County on the subsurface side of the upper edge of the Dakota outcrop (Figure 3.4). These had Cl concentrations ranging from 26 to 4000 mg/L and sulfate (SO_4) concentrations ranging from 143 to 540 mg/L. The SO_4 may be derived from minerals in the younger Cretaceous rocks which confine the Dakota Formation, as recharge seeps down through them (Macfarlane et al., 1990). In samples 12, 13, 16, 17, and 19, taken from the outcrop belt (Figure 3.4), Cl concentrations range from 12 to 226 mg/L and SO_4 concentrations range from 16 to 128 mg/L. The salinity of the water in the aquifer increases with depth and distance into the subsurface, as reported by Fishel (1948) and Macfarlane et al. (1990).

Sample 16 was taken from a well located on the surface water divide marking the eastern boundary of the Salt Creek watershed; sample 17 was collected 1/2 mile to the west from a well 50 ft. lower in elevation but with the same depth (Figure 3.4). The low Cl and SO_4 concentrations of sample 16 (Table 3.2) are indicative of locally recharged water. Sample 17 has much higher concentrations of these ions, which is evidence that a significant proportion of this water flowed to this point through a large-scale flow system. The significant differences in Cl and SO_4 concentrations between the two wells may reflect the influence of the surface-water divide on the ground-water flow systems. There is a ground-water divide on the potentiometric surface associated with this surface-water divide (Figure 3.7), which may effectively prevent the encroachment of saline water into Region C.

In summary, flow in the basal sandstone in Region A is dominated by a large-scale regional flow system bringing highly mineralized water into the study area from the northwest. In other sandstones, local recharge is an important source of ground water, particularly within the outcrop belt.

3.5.2B. Region B. Washington County North of Mill Creek

The Dakota Formation is exposed at the surface throughout most of this area. It is capped by younger Cretaceous sediments in the northwestern corner of the region. Strata of the aquifer thin relatively rapidly to the east, mainly due to the eastward slope of the land surface. The Little Blue River cuts through the lower part of the aquifer, so flow systems east of this river are separate from those to the west.

The potentiometric surface slopes toward the lower reaches of tributaries of Mill Creek and the Little Blue River (Figure 3.7), as well as Mill Creek itself. It is likely that ground water discharges from the aquifer in these areas. Unlike Region A, there are no physiographic features of the discharge of saline ground water in Region B. However, brackish water has been reported in the basal sandstone within 5 mi. of the town of Washington (personal communication, Doug Schwartz, local farmer). One of the tributaries of Mill Creek which drains part of this area is named Salt Creek (Figure 3.4). The creek may have derived its name from its salinity although there has been no known sampling of its water to verify this. The creek flows across township T2S R2E. Wells in this township also provide evidence that the basal sandstone contains brackish water. An irrigation well at T2S R2E 23 ADC, which was screened in the basal sandstone, is unused due to the salinity of the water (personal communication, John Stigge, local farmer). Wells located near the western edge of T2S R3E also yield water with a fairly high salt content (Personal Communication, Doug Schwartz, farmer). An old stock well (Number 8, Figure 3.1 and Table 3.1), which penetrates the basal sandstone, is unused for the same reason. However, a well 1 mi. away (Number 9, Figure 3.1 and Table 3.1), which is a shallower well screened in a different sandstone, yields potable water (personal communication, Rosenthal, landowner). The hydraulic head in well 8 is 67 ft. lower than in well 9 (Table 3.1), indicating that there is a downward vertical head gradient of approximately 0.5 ft. per ft.

Topographic relief is an important driving force for local flow systems in the basal sandstone close to the eastern limit of the contiguous outcrop belt, as well as east of the Little Blue River. Pleistocene sands and gravels blanket much of the lower part of the Dakota outcrop area, effectively creating a single unconfined

aquifer system in much of this area, particularly east of the Little Blue River (Figure 3.4).

The quality of water from shallow wells in the Dakota aquifer in Region B is generally good (Miller, 1966). There are many domestic wells screened in Dakota sandstones in addition to several municipal wells, e.g. the city wells of Haddam and Mahaska (Table 3.1 and Figure 3.1, wells 2,5,6, and 7). However, few of these wells are screened in the basal sandstone which is deeply buried at the western edge of the region.

A deep Dakota well in the town of Mahaska (T1S R1E 07AA, Table 3.2) is more than 2 mi. from the nearest point on the upper boundary of the outcrop belt. At this location the top of the aquifer is approximately 150 ft. below the land surface. A 1955 sample from the well had a chloride concentration of 141 mg/L and a sulfate concentration of 115 mg/L (KWATCHEM, 1991). The low chloride value suggests that this water is derived primarily from recharge through the thick confining layer rather than from large-scale flow from the west. There is a ground-water divide associated with the surface water divide which approximately follows the state line. Figure 3.7 shows that any saline water flowing into this region from Region A would be funneled toward Mill Creek away from the divide. The elevated heads along the state line and the relatively good quality of the ground water provide evidence of local recharge.

In summary, mineralized water in the basal sandstone in this region flows in from Region A to the west but is diluted by local recharge, particularly in the outcrop belt. At the eastern end of the outcrop belt recharge is the dominant source of water flowing into the basal sandstone. Local recharge is also the most important source of ground water in other sandstones in Region B.

3.5.2C. Region C. Eastern Republic County, Washington County South of Mill Creek, and Northern Clay County.

The Dakota Formation is exposed at the surface throughout most of this area. The basal sandstone is therefore relatively close to the surface and more accessible to local recharge than it is further west.

There is a ground-water divide on the potentiometric surface which approximately coincides with the surface-water divide between the Mill Creek and Republican River watersheds (Figure 3.7). North of this divide, water flows north

in the basal sandstone toward Mill Creek; south of the divide water flows south toward the Republican River. The basal sandstone probably receives recharge from precipitation throughout this area by slow infiltration from the surface through the confining mudstones. This recharge must be sufficient to sustain the relief on the potentiometric surface (Chapter 5).

Flow directions provide evidence that the most important zone of discharge from the Dakota aquifer in this area is its west-east-trending boundary with the Republican River aquifer. This alluvial aquifer cuts down to approximately 100 ft. below the current level of the river (Fishel, 1948). It is therefore likely to be in good hydraulic connection with the basal sandstone of the Dakota aquifer for several miles. Well 36 (Table 3.1) is located 1/4 mile from the Republican River and well 37 (Table 3.1) which is screened in both alluvium and Dakota basal sandstone (Figure 3.1). The elevation of the water level in the two wells (Table 3.1) and the river is approximately the same, which is evidence that there is good hydraulic connection between the two aquifers.

Ground water also discharges from the basal sandstone into the lower reaches of Peats Creek and Parsons Creek (southward-flowing tributaries of the Republican River) and at Mormon Spring (T3S R3E 23DBB) close to Ash Creek (Figure 3.7). There is a record of a flowing well close to Mormon Spring with a hydraulic head 17 ft. above land surface (Miller, 1966). The contours of hydraulic head "V" up these streams indicating that the streams are draining water from the aquifer.

Water samples from other nearby wells screened in the basal sandstone had very low Cl and SO_4 concentrations (Table 3.2 and Figure 3.4, numbers 14, 15, and 16) . The low mineralization of the water discharging from the Dakota aquifer into the Republican River aquifer along the southern edges of T5S R1E and T5S R2E is evidence that this water is part of a local flow system. There are no known ground-water salinity problems in this region except for a report of a sample from an old well near the boundary with Region A (well 7, Figure 3.4, Table 3.2) which had a chloride content of 3500 mg/L. Unfortunately there is no depth or water-level information for this well. It was probably a deep well screened in the basal sandstone. The brackish composition of the water may be explained by mixing of saline water from Region A with fresh water from above. Brackish water has not been reported east of this location in Region C. The quality of the ground water in

the basal sandstone improves toward the discharge zones of Region C and away from the main upland areas. The opposite would be expected in a large-scale flow system (Toth, 1972).

In summary, the quality of ground water in the Dakota aquifer in Region C is generally superior to the other two regions. The lack of salinity problems in Region C is probably due to a combination of two factors: 1. Saline water drains from the aquifer in Region A and is therefore prevented from occupying the basal sandstone in other tributary watersheds to the east. 2. Local recharge may be relatively high because throughout most of this area the aquifer is not confined by younger sediments.

CHAPTER 4. AQUIFER PROPERTIES DETERMINED FROM A PUMPING TEST

4.1 Introduction

To model the flow of ground water through the basal sandstone in Washington County, estimates of the transmissivity and storativity of the aquifer were needed. However, little was known about the hydraulic properties of the aquifer in this area. Pumping tests using observation wells provide a reasonably simple method of estimating hydraulic properties on a section-level scale (which is nearly within an order of magnitude of the scale of the model area used in Chapter 5). They can also be used to estimate the leakage from less permeable sediments which confine the aquifer. Prior to this study, no pumping tests using observation wells had been performed in the Dakota aquifer in Washington County.

A pumping test was performed near Clifton in August, 1990, using a high-yield irrigation well and an observation well drilled by the Kansas Geological Survey. The wells are located in T5S R1E Sec. 10 (Figures 3.1, 4.1), in the outcrop belt of the Dakota Formation where the total thickness of this stratigraphic unit is less than 200 ft. and the thickness of the basal sandstone is 100 ft. (Figure 4.2). There are several irrigation wells in this township because of the unusually thick layer of saturated, medium-grained sandstone at a shallow depth (Figure 4.1). The sandstone is confined by Dakota mudstone above and Permian shale below (Figure 2.9). The difference in hydraulic conductivity between the sandstone and its confining beds was anticipated to be many orders of magnitude, and leakage effects on drawdown during the pumping test were expected to be low.

The irrigation well was pumped at a constant rate of 592 GPM and drawdown recorded in the observation well located 270 ft. away. The drawdown was corrected by compensating for the effects of well interference, atmospheric pressure changes, and continuing recovery of the water level from a previous period of pumping. The compensated drawdown was then used to estimate the transmissivity and storativity of the aquifer and the leakage from the confining layers.

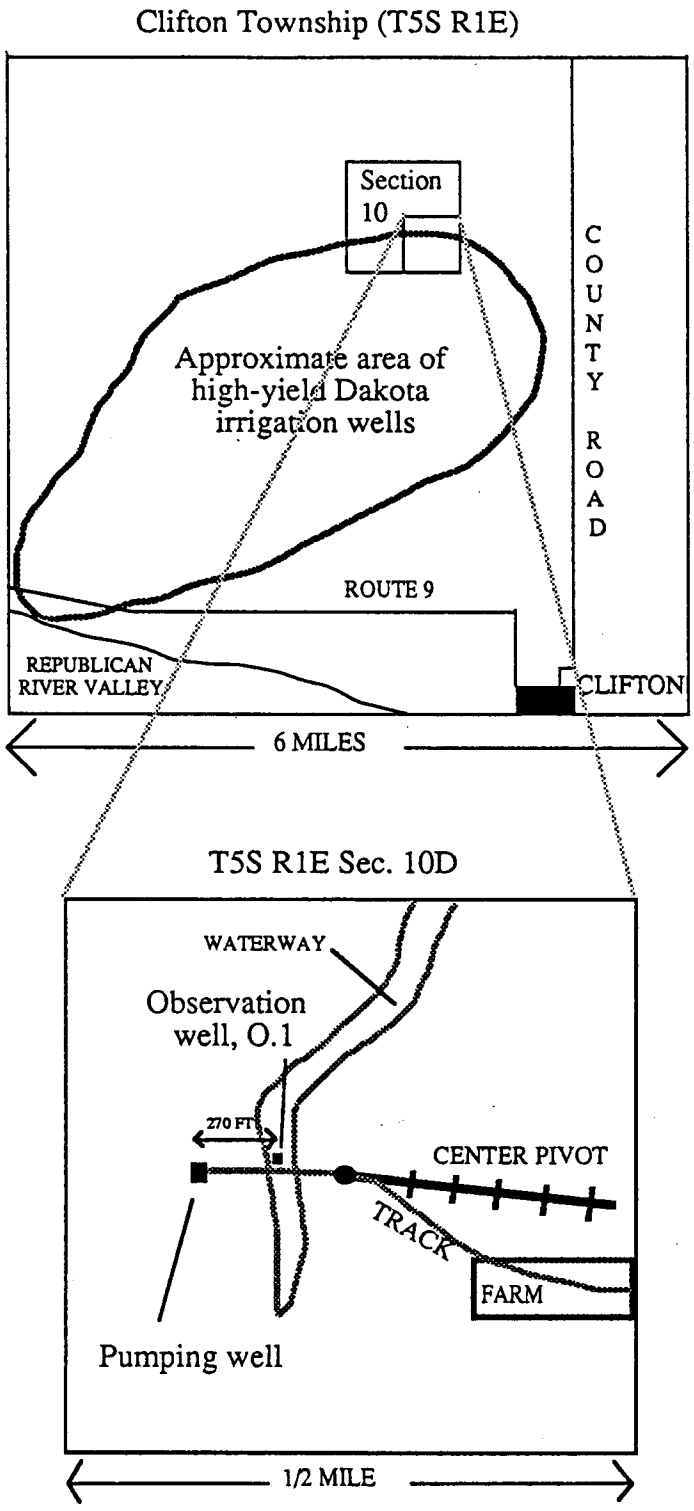


Figure 4.1. Location of pumping-test site in T5S, R1E, Sec. 10D, Washington County. The pumping well is well 32 in Figure 3.1.

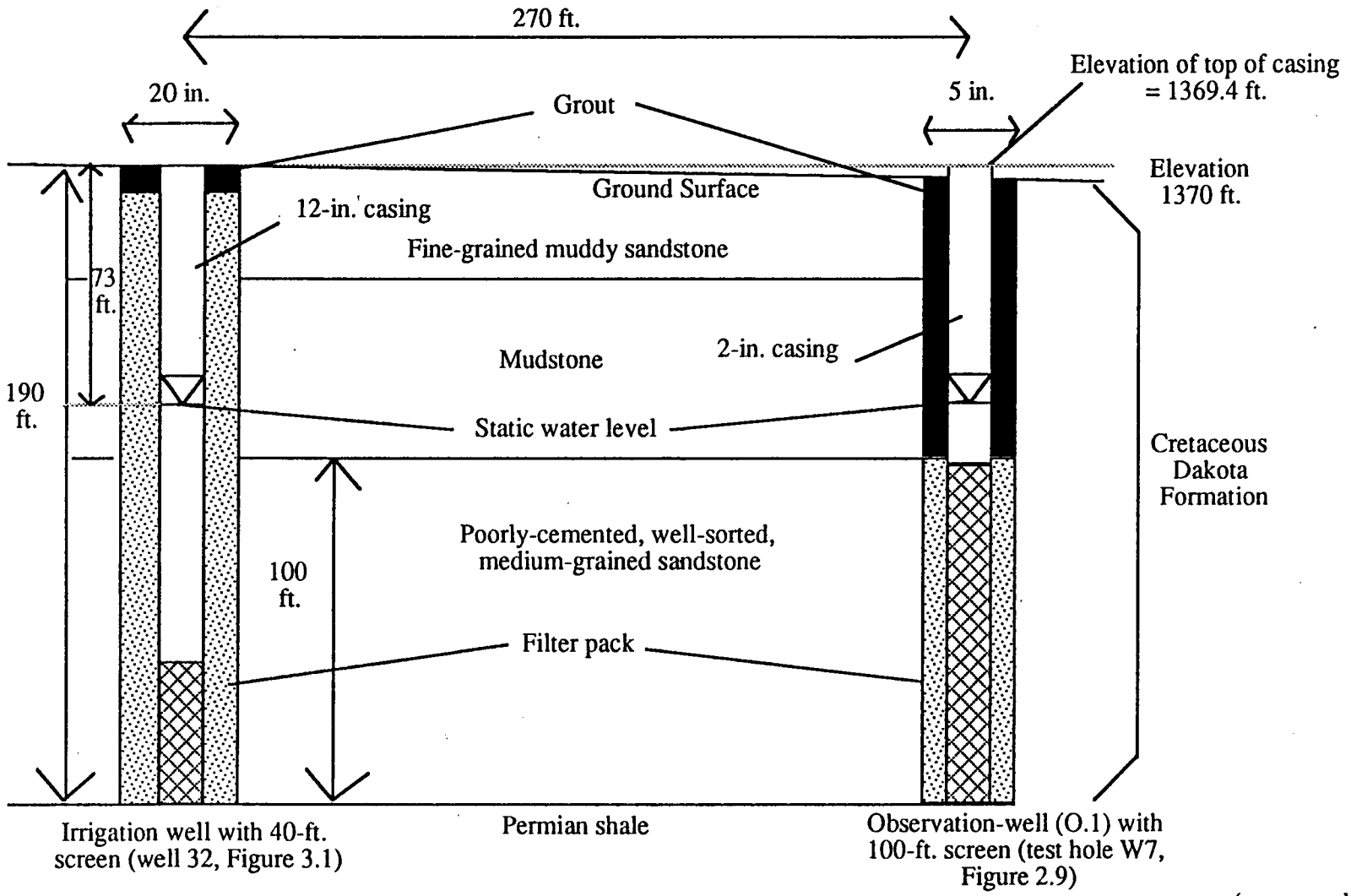


Figure 4.2. Cross section of the aquifer and the irrigation and observation wells used in the pumping test. (not to scale)

4.2 Pre-Test Preparations

4.2.1. Drilling, Construction, and Development of the Observation Well, O.1

The observation well, O.1, was drilled by the Kansas Geological Survey in May, 1990. It was located 270 ft. from the high-yield center-pivot irrigation well (well 32, Figure 3.1) on the edge of a waterway near Clifton, southwestern Washington County (Figure 4.1). O.1 was sited at the optimum distance from well 32 determined from expected drawdown based on a pre-test estimate of aquifer transmissivity. Practicalities such as accessibility, position of underground pipe and cable, and location of the waterway also influenced the siting of O.1.

A 5-in. hole was drilled to a depth of 191 ft using mud-rotary methods (test hole 7, Figure 2.9). The hole was logged using gamma ray, spontaneous potential, single-point resistivity, and caliper measuring tools. The logs were used to determine the exact depth to the top and base of the basal sandstone aquifer. Well O.1 was then constructed as shown in Figure 4.2; it was screened and gravel packed throughout the sandstone. The total length of screen used was 100 ft. Immediately above the screen and gravel pack, the borehole was sealed with 7 ft. of bentonite chips. The uppermost 75 ft. of hole was filled with a mixture of bentonite chips and shale. The hole was plugged at the top with bentonite chips and finally a steel cover was cemented into the ground over the PVC casing.

Later in the month, the well was developed using compressed air to remove accumulated fluids, mud, and cuttings. The air line was lowered to the bottom of the well and compressed air lifted water from the well at a rate of approximately 20 GPM. The well was developed for 3 hours until there was no visible trace of drilling mud in the water.

4.2.2. Monitoring Equipment

(a) On July 2, 1990, two pressure transducers were set in well O.1 at a depth of 95 ft. below the top of the casing, i.e. 22.4 ft. below the May static water-level. The transducers were designed for a pressure range of 10 psi which is equivalent to approximately 23 ft. of water. These transducers were connected to a Hermit Data Logger which was programmed to record the depth to water to the nearest .01 ft. every hour. Two transducers were used rather than one in order to insure consistency of measurements and to insure that data were not lost due to a

malfunction. The water-level monitoring station was protected with a weatherproof, insulated cover.

(b) On the same day, a barograph was placed 1/4 mile from the well in a farm building and calibrated by the atmospheric pressure measurement from the local weather station in Concordia. This instrument recorded atmospheric pressure on paper charts until mid-October. According to manufacturer's specifications, the barograph had an accuracy of ±0.15 mb and a resolution of 0.2 mb. The accuracy and resolution of this instrument were ensured by periodically comparing its reading to the local weather station reading at various pressures.

(c) On July 12, a McCrometer bolt-on saddle, propeller-driven flow meter was fitted onto the irrigation well. This device displays total gallons pumped as well as the current pumping rate. To ensure straight, laminar flow and accuracy of the meter, the manufacturer recommends a minimum of 8 in. of straight pipe downstream from the meter and 40 in. upstream. The length of straight, constant-diameter pipe leading from the well was limited so the meter was installed with 8.5 in. of straight pipe downstream and 35.5 in. upstream. Straightening vanes were therefore installed immediately upstream from the meter.

The meter was calibrated for a pipe with an internal diameter of 7.872 in. (internal radius 0.328 ft.). The actual internal diameter of the pipe, measured during installation of the meter, is 8.24 in.; its radius is 0.343 ft. Flow rate through a pipe is proportional to the square of its internal radius (IR^2). The IR^2 of the pipe is 9.6% greater than the IR^2 for which the meter was calibrated. Flow meter readings were therefore corrected by multiplying by a factor of 1.096.

4.2.3. Pre-Test Data Acquisition and Analysis

Before the pumping test began on August 7, 1990, data were collected from the monitoring equipment described above for four weeks. Well 32 was in use for most of the first 11 days of this period, until July 21. Between 04:00 hrs, July 21 and 14:36 hrs, August 7, well 32 was not pumped. From July 22 to August 6 none of the other irrigation wells in the field to the southwest were pumped either. This allowed the aquifer to recover to within 3 ft. of its pre-irrigation season level from a maximum drawdown of close to 16 ft.

The water-level data were downloaded from the datalogger directly onto a microcomputer at the Kansas Geological Survey. (Drawdown data are tabulated in

Appendix 5.) Atmospheric pressure data in millibars were entered into the microcomputer at the keyboard and converted into units of height of a column of water. These data were used

(1) to check that the transducers were set at a suitable depth and were performing accurately;

(2) to ensure that all the monitoring equipment was functioning properly;

(3) to decide on an initial pumping rate for the irrigation well which could be sustained at maximum drawdown in that well.

Depths to water and atmospheric pressure fluctuations were also used to quantify the effects of aquifer barometric efficiency, recovery from a previous period of pumping, and well interference from another irrigation well (well 31, Figure 3.1) 2200 ft. away which began pumping 29 hours before the start of the pumping test. These effects were then used to adjust the raw water-level data in order to determine fluctuations resulting from pumping only.

4.3 Methodology

Theis (1935) related drawdown in an observation well to discharge rate of water from a pumping well using aquifer properties of transmissivity and storativity:

$$s^* = \left(\frac{Q}{4\pi T}\right) W(u)$$

where W(u) is the "well function," an infinite series given by

$$W(u) = \left[-.5772 - \ln u + u - \frac{u^2}{2x2!} + \frac{u^3}{3x3!} - \frac{u^4}{4x4!} + \ldots.\right].$$

The argument, u, is given by $u = \frac{r^2 S}{4Tt}$.

In these equations, s* is the drawdown induced in the observation well at a distance, r, from the pumping well after a time, t , of pumping if the well is pumped at a constant rate, Q. T and S are the aquifer properties of transmissivity and storativity respectively.

In applying this solution, it is assumed that flow is in the range of Darcy's Law and that water is discharged instantaneously from storage in the aquifer when pumping begins. It is also assumed that the wells fully penetrate the aquifer, which has constant thickness and negligible slope and is homogeneous and isotropic.

The Theis solution does not consider the possibility of leakage of water from the confining layers above or below the aquifer. Hantush and Jacob (1955) modified the Theis solution to include consideration of leakage from a confining layer:

$$s^* = \left(\frac{Q}{4\pi T}\right) W\left(u, \frac{r}{B}\right)$$

B is the "leakage factor," given by

$$B = \sqrt{\frac{Tb'}{K'}},$$

assuming there is only one leaky confining layer, where
b' is the thickness of the confining layer, and
K' is its vertical hydraulic conductivity.

In addition to the assumptions listed above, this Hantush-Jacob solution assumes leakage through the confining layer is vertical and proportional to hydraulic gradient across the layer, the head in the deposits supplying the leakage is constant, and storage in the confining bed is negligible. The equation is valid for all values of r_s (radius of well screen), provided that the following conditions are satisfied:

$$\frac{r_s}{B} < 0.1, \text{ and } t > \left(\frac{30 r_s^2 S}{T}\right)\left[1 - \left(\frac{10 r_s}{b}\right)^2\right].$$

In order to determine T, S, and B using the Hantush-Jacob or Theis equations, drawdown in an observation well should be recorded over a period of several hours while the pumping well is discharging at a constant rate. A common method of solution facilitated by modern computing capabilities is to use non-linear regression to estimate the values of T, S, and B that produce synthetic drawdown/time data which most closely match the observed data in terms of the sum of the squared residuals. This method was used to estimate aquifer properties from the pumping-test drawdowns.

Due to the water stored in a well when pumping begins, water will not be released instantaneously from storage in the aquifer and so, at early time, the Hantush-Jacob and Theis well functions are not valid. Walton (1987, p.3) provides a formula with which to determine the critical time after which pumping well storage becomes insignificant:

$$t_s = \frac{\left[5.4E5\left(r_w^2 - r_c^2\right)\right]}{T}$$

where t_s = critical time (min.),
r_w = internal radius of well casing, (ft.),
r_c = external radius of drop pipe inside well, (ft.) and
T = transmissivity of the aquifer, (gpd/ft.).

For well 32, r_w = 0.5 ft., r_c = 0.25 ft. and T = 57000 gpd/ft. (from section 5), Therefore, using the formula above, t_s = 1.8 min. To insure no well storage effects, final values of T, S, and B were determined using only data from later than 3 min. into the pumping test (section 5).

4.4. Pumping-Test Data

The pumping test began at 14:36 hrs, August 7, 1990. Well 32 was pumped at a rate of 592 GPM, previously determined to be the normal sustained rate during long-term pumping. Initially, the pump was throttled back so as not to exceed this rate while the water level in the well was still high. Total volume pumped was read from the odometer every 5 min. for the first 30 min. of the test to determine the pumping rate as accurately as possible and to ensure it was constant. No variation in the pumping rate was detected.

The datalogger recorded depths to water to the nearest 0.01 ft. at logarithmically increasing time intervals beginning at 0.2 seconds. (Drawdown was calculated to the nearest 0.01 ft. for a time t after the onset of pumping by subtracting the initial depth from the depth at time t.) Pumping of well 32 ceased after 33 hours when drawdown was nearly 9 ft. and the time interval between depth measurements was 100 min. Eight hours later, pumping resumed at the same rate for an additional 70 hours, during which a maximum drawdown of 11.64 ft. was

reached. Recovery of the aquifer was then recorded for over 100 hours at a rate of one measurement every 100 min. for the first 50 hours, dropping to one every 127 min. for the last 50 hours.

Measured depth to water, drawdown, and compensated drawdown from this nine-day period are tabulated in Appendix 5. Drawdown was adjusted to compensate for the effects of (1) overall aquifer recovery, (2) drawdown due to well 31, and (3) atmospheric pressure changes. These corrections are described in detail in a the complete report of the pumping test (Wade, 1991). Figure 4.3 shows the fully compensated and original observed drawdown plotted against logarithmic time for the complete 9 days of data.

4.5. Potential Sources of Error in the Data

4.5.1. Flow Meter

As is often the case in pumping tests, the main source of error in this test is likely to be in the value of the pumping rate. During the first 30 min. of pumping, in which the pumping rate was determined every 5 min., there was no detectable variation in the rate. However, for the rest of the test the pumping rate was not monitored. In theory, in a homogeneous aquifer, the Jacob semilog plot (Figure 4.3) should follow a straight line after the initial curve at early time but slight deviations can be seen which may be due to small fluctuations in the pumping rate or the heterogeneities of the aquifer.

Two factors contributed to error in the flow rate measurement: one was the lack of sufficient straight pipe upstream from the meter to satisfy manufacturers recommendations. Also, a correction factor had to be introduced because the meter was calibrated for a pipe of 5% smaller I.D. The correction factor was based on field measurements made of the I.D. of the pipe when it was cut open during installation of the meter. The combined effect of these factors is that the pumping rate is not accurate to better than ± 5%, i.e. Pumping rate = 592 ± 30 GPM. This percentage error is considered in evaluating the pumping test.

4.5.2. Transducers

The accuracy of each transducer was checked at various water levels during preliminary testing in July. This was done by comparing the transducer values

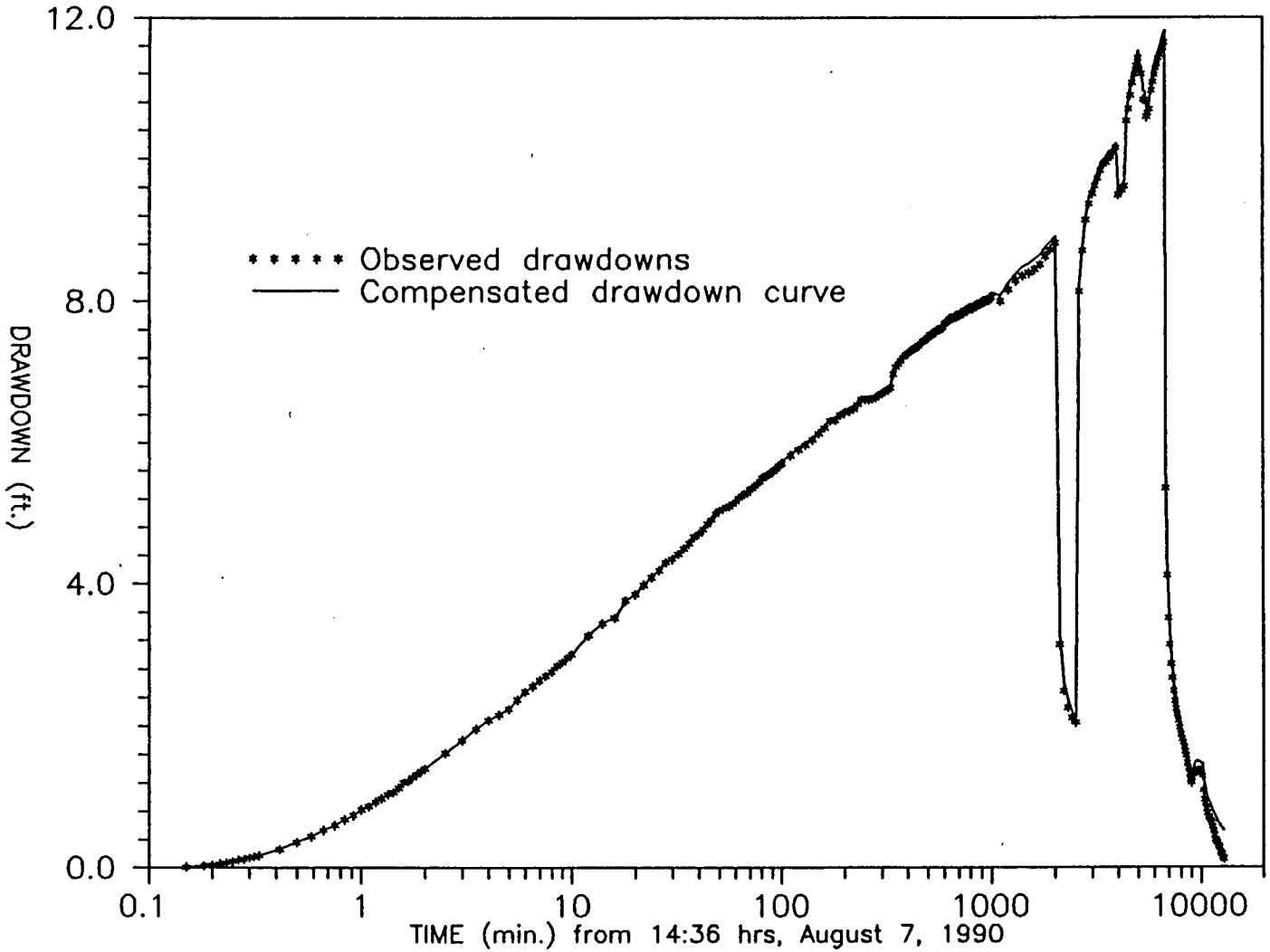


Figure 4.3. Observed and compensated drawdown over the full nine days of the pumping test. The difference is very small, particularly at early time.

displayed on the Hermit datalogger with measurements from an electric water level tape. It was found that one transducer was malfunctioning part of the time but the other one was consistently within 0.05 ft. of the electric tape measurements. At no time during the collection of water level data did the head above the transducers exceed the range for which they were designed and at no time did the water level drop below the level of the transducers. Values of depth to water used for analysis before and during the pumping test were those measured by the transducer which had proved to be consistently accurate. Transducer error is not therefore a significant factor in the determination of the aquifer properties.

4.5.3. Other Influences on Water Level

The combined effect of (1) atmospheric pressure fluctuations, (2) overall aquifer recovery since it was last pumped, and (3) well interference during the pumping test was least significant at small time (and not detectable at all before 24 min. of pumping) and most significant at large time. These factors reduced the drawdown by 0.14 ft. (1.7%) after 1600 min. of pumping and by 0.10 ft. (1.1%) after 2000 min. of pumping (Appendix 5). They had negligible effect on the estimation of T and S, for which early drawdowns are the most important. However, the leakage factor determined for the confining layer was altered by over 10% when the data were adjusted as described above. This is because leakage effects are only seen at relatively large time in a pumping test. In addition, the difference between observed and fully compensated drawdown was small partly because effects (1), (2) and (3) combined by chance to cancel each other out. Therefore, the corrections made to the drawdown are justifiable, particularly for estimating the leakage factor.

4.5.4. Vertical Flow

Well 32 is screened through the lower 40 ft. of the sandstone only (Figure 4.2). Close to a pumping well shortly after pumping begins vertical flow may be significant and this can affect the drawdown in an observation well. Well O.1 was therefore sited a large distance from well 32 and was screened throughout the sandstone making the effect of the partially penetrating pumping well negligible.

4.6. Results and Interpretations

Determinations of transmissivity, storage, the leakage factor of the confining layer, and approximate boundary conditions were made from the initial 1600-min. period of uninterrupted pumping only, for the following reasons:

1. There is a high density of data from this period; 64% of the 9 days of drawdown data were collected during the first 1600 min. (Appendix 5).
2. Well 31 was being pumped continuously through this period but the later pumping history of this well is unknown and therefore the correction for its pumping is unreliable at later times. Figure 4.3 shows significant fluctuations in drawdown during the second period of pumping of well 32 which are probably due to intermittent pumping of well 31.
3. It can be seen from Figure 4.3 and Appendix 5 that the difference between the observed and compensated drawdowns is small during this period.
4. At no time during this period did the water level in the observation well fall below the top of the screen.

SUPRPUMP (Bohling et al., 1990), a microcomputer software package incorporating the Hantush-Jacob and Theis well functions, was used to analyze the compensated drawdown/time data (see Methodology, above). SUPRPUMP uses non-linear regression to estimate the well function parameters that produce synthetic drawdown/time data which most closely match the observed data in terms of the sum of the squared residuals. The solutions of the well function parameters for various intervals of time are listed in Table 4.1.

Values of transmissivity, storativity, and leakage factor were determined using both the leaky artesian Hantush-Jacob function and the Theis well function, which does not consider leakage. Figure 4.4 shows a good fit of the compensated drawdown data points to the computer-generated Hantush-Jacob fitted curve for the period 3 to 1600 min. The Theis analysis (Figure 4.5) of the same data produced a significantly poorer fit in terms of the root mean squared residuals (Table 4.1). SUPRPUMP calculated the 95% confidence limits of the aquifer properties it determined. These were ± 1% for transmissivity, T, ± 3% for Storativity, S, and ± 9% for leakage factor, B. The program AQUITEST (Heidari and Hemmet, 1991), which also estimates aquifer parameters using non-linear regression, was used to check the results. Results obtained for T, S, and B were all within 0.5% of the SUPRPUMP results.

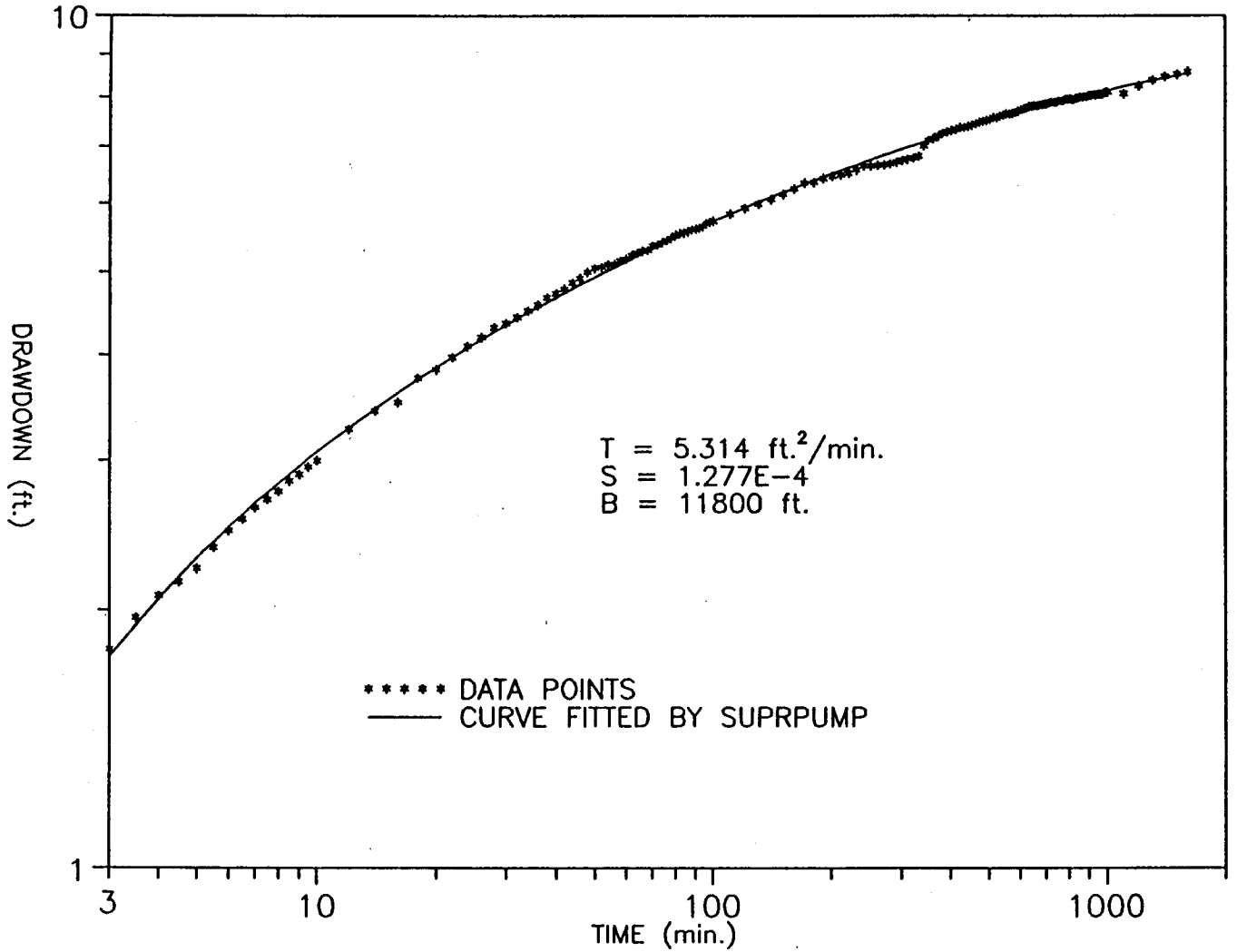


Figure 4.4. Leaky artesian analysis of compensated drawdown between 3 and 1600 minutes of pumping. The fitted curve matches the drawdown very well.

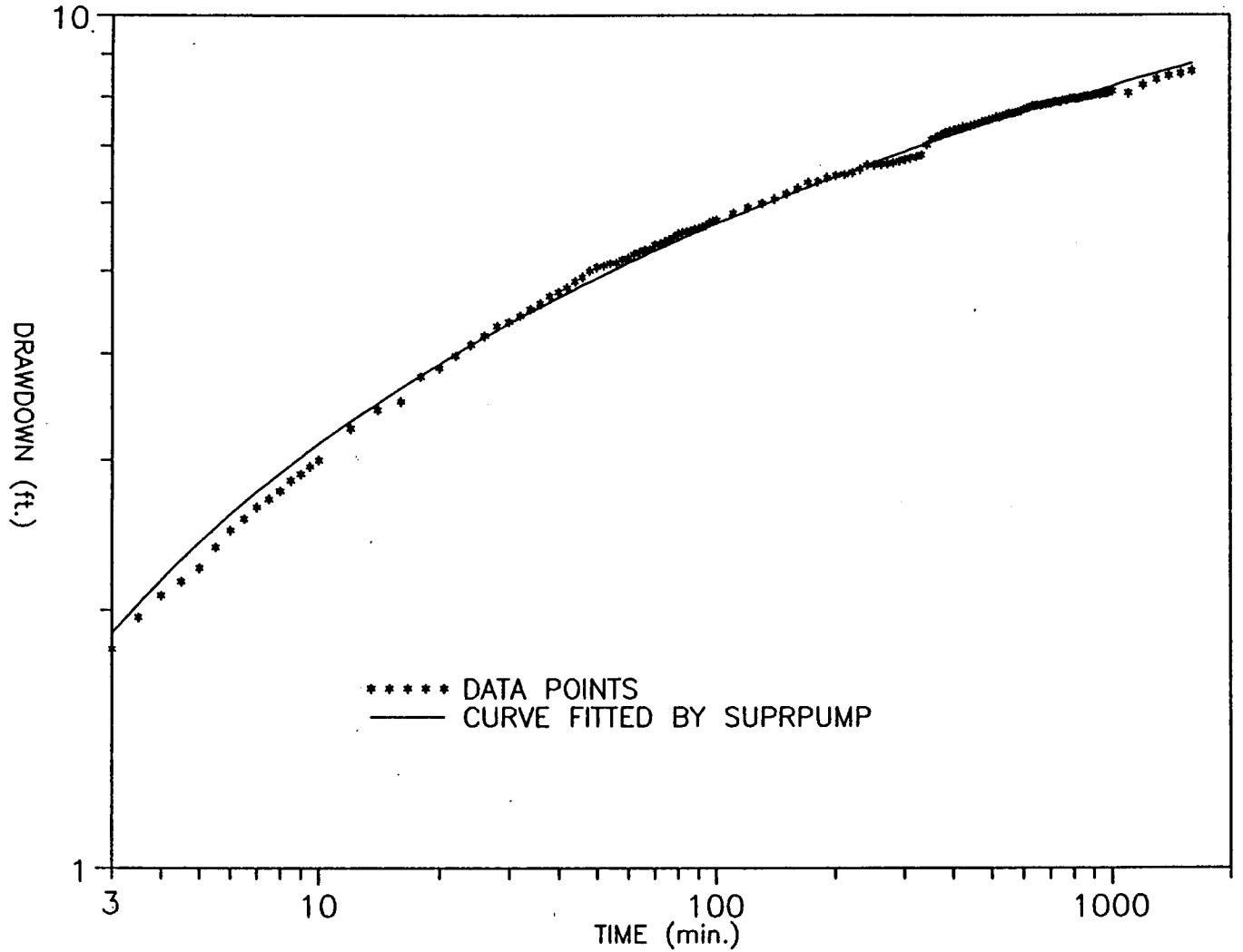


Figure 4.5. Theis analysis of compensated drawdown between 3 and 1600 minutes of pumping (leakage not considered). Fit is poor at early time.

TABLE 4.1. Values of aquifer properties determined from the pumping test. Underlined values were set as known quantities and were not permitted to vary during the simulation.

Period of data considered (min.)	Trans-missivity, T ($ft.^2$/day)	Storativity, S	Leakage factor, B (ft.)	RMS res. (ft.)	No-flow boundaries simulated
3-1600	8100	1.088E-4	Theis; B not considered	.08880	NONE
3-1000	7700	1.277E-4	11800	.06531	NONE
3-1600	7600	1.277E-4	11800	.06566	NONE
3-2000	7700	1.249E-4	13300	.06789	NONE
3-2000	7600	1.281E-4	11200	.06655	2, each 2 mi. away
6717-9000	6500	1.277E-4	22900	.06263	NONE
6717-9000	6300	1.277E-4	12400	.04261	2, each 2 mi. away

In this table, the "leakage factor", $B = (Tb'/K')^{1/2}$, and RMS res. is the "root mean squared residual" or the quadratic mean difference between observed and synthetic drawdown. It is a measure of how well the synthetic drawdown fits the observed drawdown.

Figure 4.6 shows that the computer-generated curve of Figure 4.4, when extended to early time, matches the early compensated drawdown well. Therefore well storage is probably insignificant to the results of this pumping test. Furthermore, the best fit to the first 1600 min. of data was obtained without simulating boundary conditions, suggesting that boundary effects are unimportant during that time. The best fit to the data from the interval 3–2000 min. was obtained by including parallel no-flow boundaries at a distance of approximately 2 mi. (Table 4.1). Later drawdowns are increased significantly by boundary, or aquifer thinning effects, or possibly by well interference from another pumping

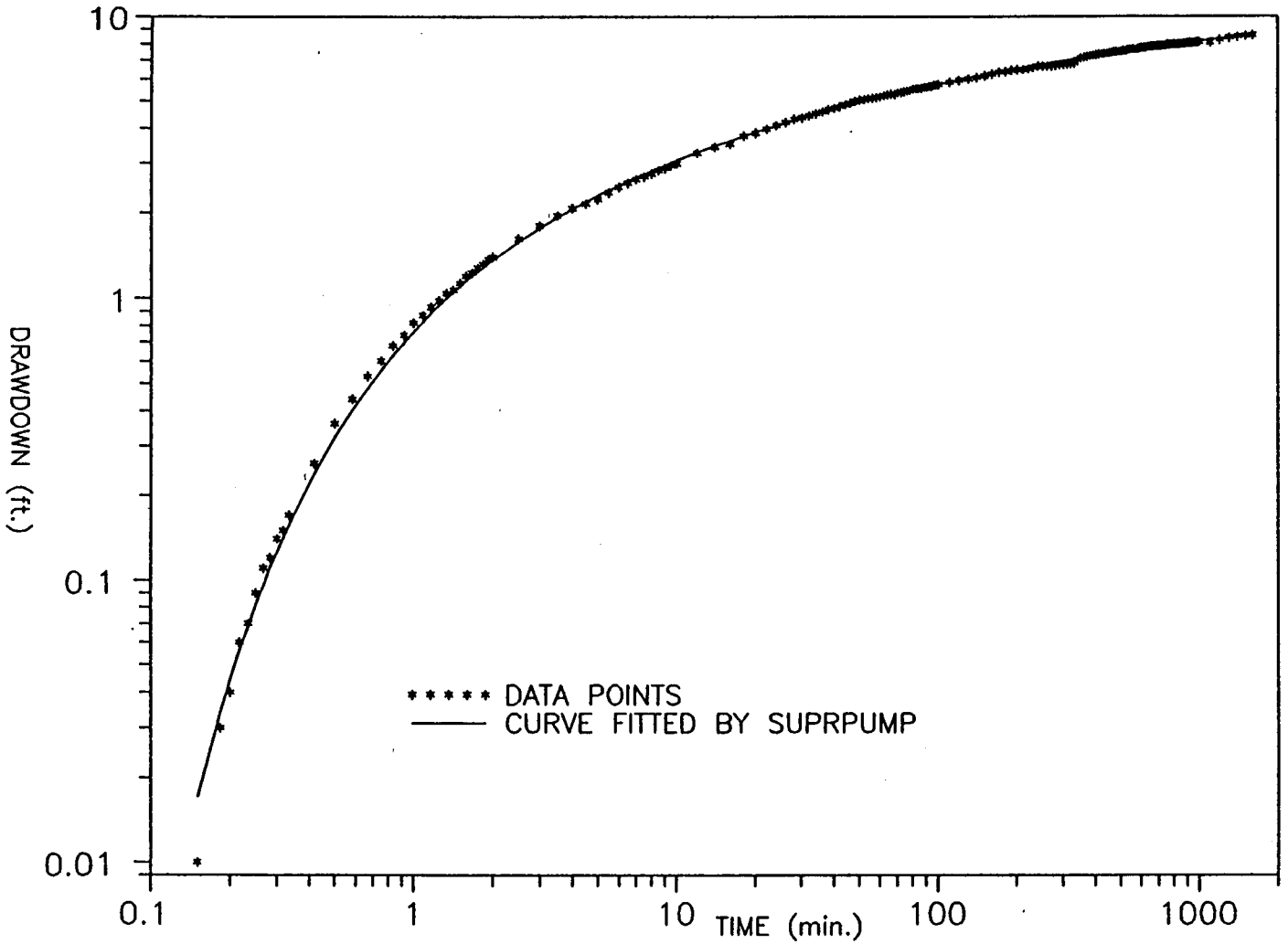


Figure 4.6. Compensated drawdown between 0.15 and 1600 minutes of pumping, and the fitted curve from Figure 4.4. Fit is excellent throughout, despite the expected effect of pumping–well storage at early time.

well. This is particularly obvious in the increase in drawdowns recorded later than 3000 min (Figure 4.3). Recovery data recorded after pumping stopped are best matched with synthetic drawdowns using a lower transmissivity than during initial pumping and also by simulating two barrier boundaries approximately 2 mi. away (Table 4.1). This is unlikely to be due to compaction of the aquifer during the pumping test because numerous irrigation wells, including wells 32 and 33, have already been pumping from this sandstone for several years.

The responses of the later drawdown and recovery are probably due to the sandstone thinning within approximately 2 mi. of the site. At this site the sandstone is exceptionally thick (Figure 2.9) and a thinning of this sandstone within a few thousand feet is therefore likely. This interpretation is supported by the fact that irrigation wells in the area are not scattered universally but are concentrated to the south and west of the site, where the yield is highest, implying that the sandstone thins toward the north and east.

The Hantush-Jacob conditions (1955) for r_s/B and t (see 4.3 Methodology) were both satisfied for all values of t at which drawdowns were measured. Hantush-Jacob (1955) analysis of the data collected between 3 and 1600 min. gave the following results:

$T = 7600 (\pm 500)$ ft.^{2}/day

$S = 1.28(\pm 0.06)\times 10^{-4}$

$B = 1.2(\pm 0.1)\times 10^4$ ft.

$S = bS_s$, where S_s is the specific storage of the aquifer (Freeze and Cherry, 1979). Therefore, $S_s = 1.28(\pm 0.10)\times 10^{-6}$ ft.$^{-1}$.

From $B = \sqrt{\frac{Tb'}{K'}}$, 'leakance', K'/b', $= T/B^2 = 5.5(\pm 0.9)\times 10^{-5}$ day^{-1}. If the mudstone was saturated to 15 ft. above the top of the sandstone, i.e. b' = 15 ft., the level of the hydraulic head in the sandstone before the irrigation season, the vertical hydraulic conductivity of the mudstone, K', would be $8.0(\pm 1.3)\times 10^{-4}$ ft./day. However, the mudstone is 47 ft. thick at the site of the pumping test (Figure 2.9) and b' may be greater than 15 ft. If it was saturated throughout, i.e. b' = 47 ft., K' would be $2.6(\pm 0.4)\times 10^{-3}$ ft./day. In reality, the saturated thickness of the mudstone is likely to be somewhere between 15 and 47 ft. and a significant proportion of leakage may be contributed by the underlying Permian shale. Therefore,

$2.6(\pm 0.4)\times 10^{-3}$ ft./day is an estimate of the maximum vertical hydraulic conductivity of the confining layer.

4.7. Conclusions

4.7.1. Aquifer Hydraulic Properties

The calculated horizontal hydraulic conductivity of the sandstone is 76 (±5) ft./day which is high for a sandstone and more typical of unconsolidated sand. The poor lithification of the Dakota sandstone at the pumping test site, as observed in the drilling of well O.1 (Figure 2.9), supports the high hydraulic conductivity. The basal sandstone in some of the other test holes (Figure 2.3 to 2.8, and 2.10) is not as coarse as in test hole W7 (Figure 2.9), at the site of the pumping test. Hydraulic conductivity can be approximately related to median grain size (Freeze and Cherry, 1979). Therefore, although this high hydraulic conductivity is likely to be within the normal range occurring in the basal sandstone, it is probably at the high end of that range, and certainly at the high end of the range in Dakota channel sandstones in general (Chapter 2). The calculated transmissivity of the sandstone is 7600 (±500) $ft.^2/day$. In addition to the high hydraulic conductivity, the exceptional thickness of the sandstone at the site of the pumping test is a major reason for the high transmissivity.

The calculated storativity of the sandstone is $1.28(\pm 0.06)\times 10^{-4}$ and the specific storage is $1.28(\pm 0.10)\times 10^{-6}\ ft^{-1}$. These results are within the normal range typical of a confined aquifer (Freeze and Cherry, 1979). Watts (1985) estimated the specific storage of the Dakota aquifer in southwestern Kansas to be $2\times 10^{-6}\ ft.^{-1}$.

Assuming there is no leakage from the Permian shale underlying the sandstone, the Hantush-Jacob leakage factor of the upper confining layer is $1.2(\pm 0.1)\times 10^{4}$ ft., its leakance is $5.5(\pm 0.9)\times 10^{-5}\ day^{-1}$, and its vertical hydraulic conductivity is between $8.0(\pm 1.3)\times 10^{-4}$ and $2.6(\pm 0.4)\times 10^{-3}$ ft./day, which is within the range typical of a clay-rich mudstone (Freeze and Cherry, 1979).

4.7.2. Thickness and Extent of Aquifer

The sandstone aquifer at the site of this pumping test is 100 ft. thick. Drawdown during pumping and recovery later than 1600 min. was affected as if no-flow boundaries were present within 2 mi. of the site. Recovery data recorded after

pumping stopped are consistent with a lower transmissivity than during initial pumping. This is probably due to the sandstone thinning away from the site of the pumping test.

CHAPTER 5. ESTIMATES OF RECHARGE RATE AND THE EFFECTS OF PUMPING IN SOUTHWESTERN WASHINGTON COUNTY

5.1. MODEL A, AN AREAL MODEL

5.1.1. Introduction

The purpose of Model A is to estimate the recharge rate to the basal sandstone aquifer in the outcrop belt of the Dakota Formation and to predict the effects of increased use of the aquifer. The ground-water flow model "MODFLOW" (McDonald and Harbaugh, 1988) was used to simulate two-dimensional flow in an areal model. In this computer program similar programming functions are grouped together in independent modules, for example a recharge module and a well-pumping module. In this way individual modules can be called up as required and the size of the program is limited to the necessary code for the specific problem.

It was hypothesized that recharge directly from the surface must be the main source of inflow to the system in order to sustain the potentiometric surface in the basal sandstone aquifer as well as the water quality. Ground-water flow was simulated to estimate the recharge rate which would sustain the potentiometric surface (Section 5.1{a}). The model was then used to determine the effects of pumping the aquifer at the present rate of water withdrawal and to predict the effects of increasing the withdrawal rate (Section 5.1{b}).

Model A was constructed to simulate ground-water flow in the basal sandstone over a 300 $mi.^2$ area of Washington and Clay counties. This area is bounded by Mill Creek to the north, the Republican River to the south, the east fork of Elk Creek to the west and Peats Creek to the east (Figures 5.1 and 5.2). It consists of most of Region C (Figure 3.7), in which the aquifer is generally unaffected by salinity problems and is consequently pumped for various uses (Chapter 3).

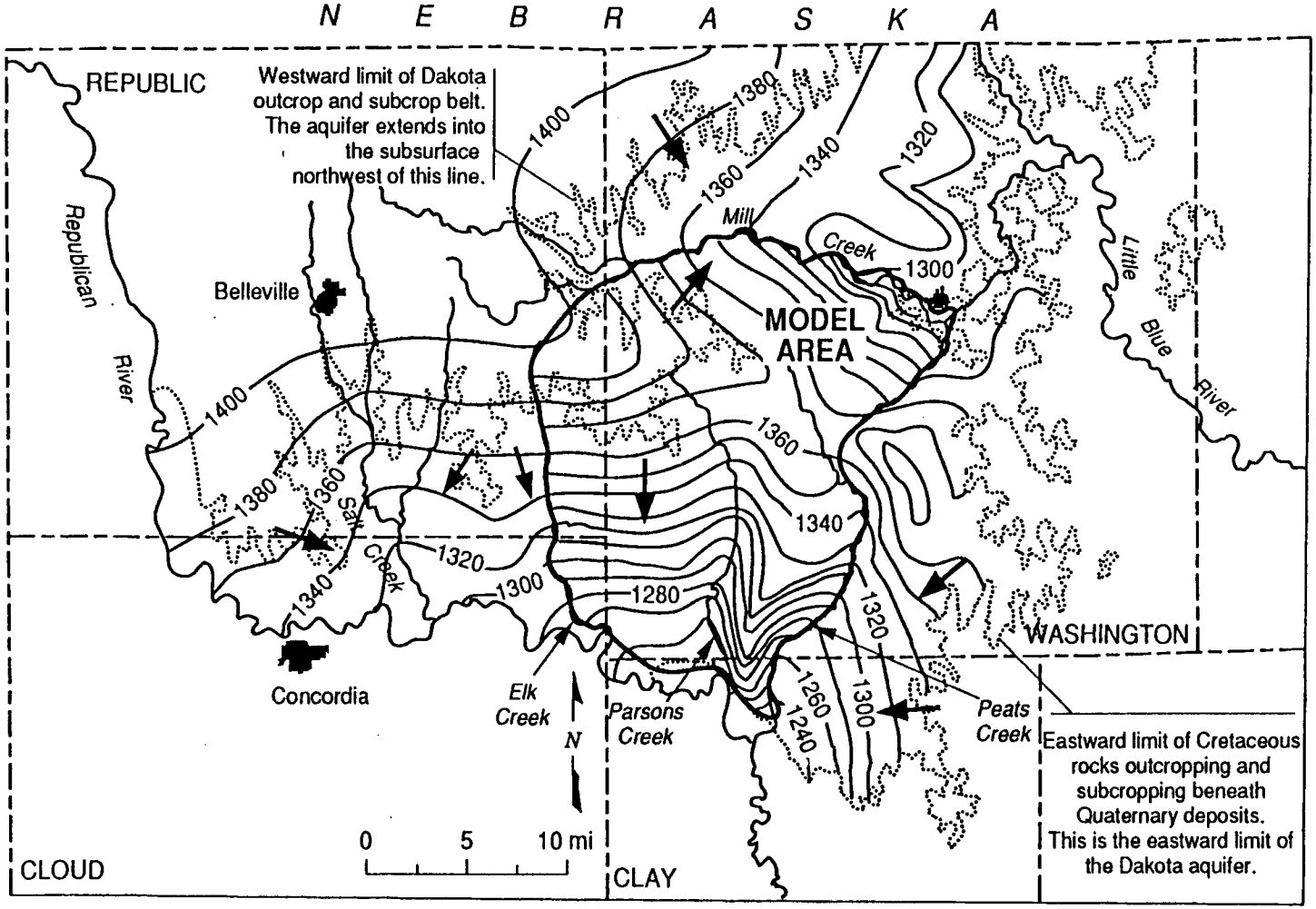


Figure 5.1. Area of model A. Contours are in feet. The potentiometric surface of the basal sandstone is contoured in 10-ft. intervals within the model area.

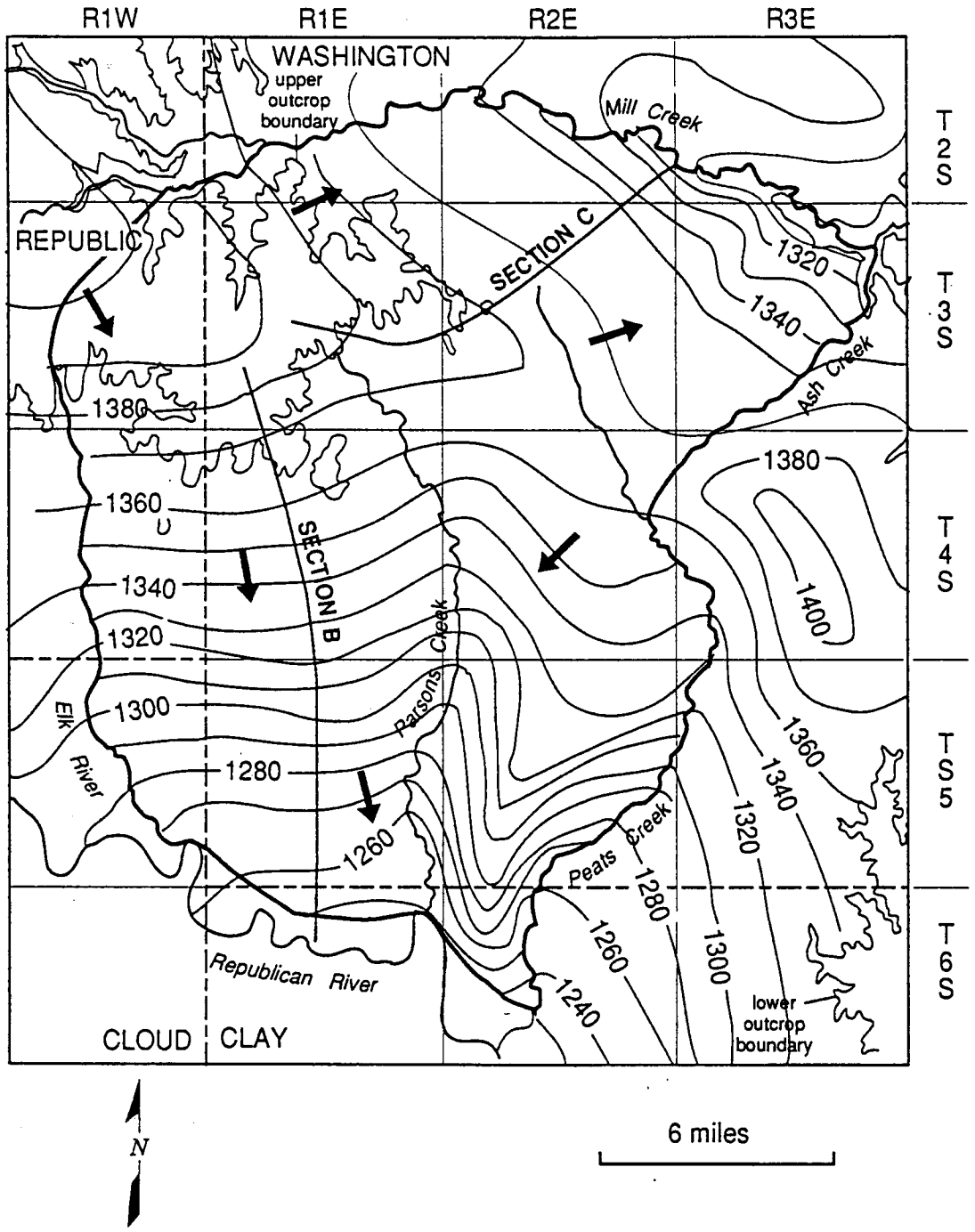


Figure 5.2. Area of Model A showing the potentiometric surface of the basal sandstone contoured in 10-ft. intervals inside the model area and 20-ft. intervals outside it. Model A boundaries are mostly parallel to the flow paths. Models B and C were used to simulate flow along Sections B and C respectively.

5.1{a}. DETERMINATION OF RECHARGE RATE (Steady-State Conditions)

5.1{a}.1. Model Setup

The basal sandstone was modelled in a steady state as a continuous, confined, one-layer system with a uniform recharge rate through the leaky confining layer. The model area was discretized into a regular grid; transmissivities were estimated and hydraulic heads were interpolated for each grid cell. Recharge rate was varied until the sum of the squared differences between computed and interpolated heads was minimized.

5.1{a}.2. Governing Equation

The model solved for the hydraulic head in a steady state using the strongly implicit procedure of finite difference iteration. For square grid cells, in a steady-state system, the finite difference equation linking hydraulic head (h) at a node i,j, distance between nodes (Δx), recharge rate (R), and transmissivity (T), is a simplified version of Poisson's equation:

$$\frac{\left(h_{i-1,j} + h_{i+1,j} + h_{i,j-1} + h_{i,j+1} - 4h_{ij}\right)}{(\Delta x)^2} = -\frac{R}{T_{ij}}$$

5.1{a}.3. Simplifying Assumptions

1. The recharge rate is uniform throughout the model area. Spatial variation in the recharge rate is likely due to the variation in the hydraulic conductivity of the confining layers. However, on the scale of the model, with a distance of 1.5 mi. between nodes, the variation is probably small.
2. The system is in a steady state. There is no evidence in the model area of significant changes in water levels since pumping of water from high-yield wells in the aquifer began during the 1970s. Computer-simulated pumping (Section 5.1{b}) shows that the current rate of water use is not sufficient to significantly affect water levels.
3. There is no leakage to or from the underlying Permian shale. There are no hydraulic-head data in this area from the Permian Wellington Formation which underlies the Dakota aquifer. Therefore, whether there is an upward or a downward hydraulic gradient between the Dakota and the Permian is unknown. However, the

hydraulic gradient is likely to be downward due to the effects of topography. Permian rocks in the study area are mainly shales which have very low hydraulic conductivity (Chapter 2). Any leakage into or out of the Permian is likely to be small.

5.1{a}.4. Model Inputs

5.1{a}.4.1. Grid

The grid consisted of 119 active cells and 33 constant-head cells, each being a square with side length 1.5 mi. (Figure 5.3). The grid-cell size was small enough to represent the locations of stream boundaries with reasonable accuracy but not too small for the density of field data originally used to generate the potentiometric surface.

5.1{a}.4.2. Boundaries

Boundaries were chosen so that in most cases they were perpendicular to the equipotential lines of the potentiometric surface (Figure 5.2). This simplified the modelling problem because a large proportion of the total length of the boundaries could be simulated with no flow into or out of the model area. Figure 5.2 shows the direction of flow in the basal sandstone to be generally parallel to the tributaries (Elk Creek, Peats Creek, Parson's Creek and Ash Creek) and toward the more major rivers (Republican River and Mill Creek). The upper reaches of the tributaries at the edges of the model area were therefore simulated as no-flow boundaries. Constant-head boundaries were used along the Republican River and the lower reaches of Mill and Ash Creeks, where ground water from the aquifer discharges into alluvial aquifers. Constant-head boundaries were also used at the northwestern corner and the eastern edge of the model area where equipotential lines are not quite perpendicular to the boundaries (Figures 5.2, 5.3) and therefore some inflow into the model area is likely.

5.1{a}.4.3. Transmissivity

From the governing equation above, the recharge-rate estimate is directly dependent on the transmissivity estimate. If T_{ij} is doubled for all nodes ij, and the

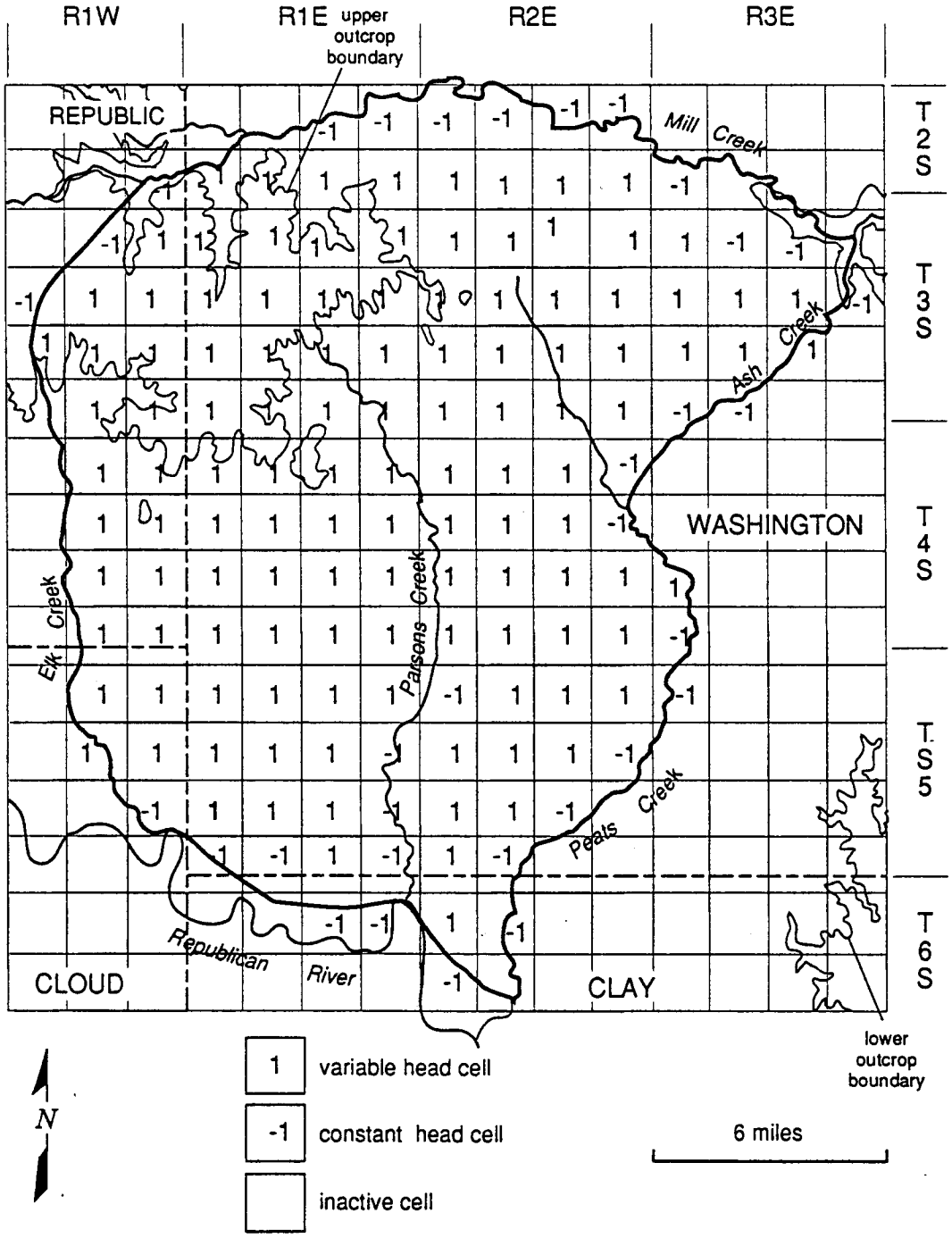


Figure 5.3. Grid used in Model A. Constant head cells are used in discharge zones and along boundaries where the potentiometric surface (Figure 5.2) shows a hydraulic gradient into the model area.

recharge estimate is also doubled, the same heads will be generated. Therefore the estimate of the transmissivity is very important.

An estimate of transmissivity (T_{ij} in the governing equation above) was needed for each node, ij, of the model. There are no historical records of pumping rates and drawdown patterns with which to calibrate T_{ij}. The pumping test (Chapter 4) provided an accurate estimate of transmissivity in T5S R1E, Sec. 10D which is located in cell 11,6 of the model grid (Table 5.1). However, no pumping tests with observation wells have been performed elsewhere in this area. There are records of performance tests of several irrigation wells in which the specific capacities of the wells were recorded on WWC-5 forms. Theis et al. (1963) developed a method to estimate transmissivity from specific capacity, duration of test, effective well radius and storage coefficient. Using this method, I obtained an estimate of transmissivity from well 32 (Figure 3.1, Table 3.1) of 4100 $ft.^2$/day which compares poorly with the estimate of 7600 $ft.^2$/day made from the pumping test using the same well (Chapter 4). For this reason, I did not use specific-capacity tests to estimate transmissivity elsewhere in the model area.

The pumping-test site is at the northeast edge of an area of high-yield irrigation wells (Figure 4.1). The mean thickness of the basal sandstone in the drillers logs of these wells is 60 ft., 60% of its value at the pumping-test site. I estimated T_{ij} to be 4600 $ft.^2$/day in cells in the area of irrigation wells, based on this thickness and assuming the hydraulic conductivity to be the same as at the nearby pumping-test site (Table 5.1).

I estimated T_{ij} throughout most of the model area to be 1500 ± 500 $ft.^2$/day, 20% of the pumping-test value, for the following reasons.

i. KGS test-hole logs and well-drillers logs throughout the area of Model A show that the mean thickness of the basal sandstone is approximately 50 ft., half its thickness at the pumping-test site. (The mean thickness of the basal sandstone in the nine KGS test holes, most of which are outside the model area, is 55 ft., which agrees with this figure.)

ii. The thickness data from water-well logs are likely to be biased because the wells are located to penetrate as thick a sandstone as possible and holes which do not penetrate a significant sandstone may be plugged.

iii. The mean hydraulic conductivity in the model area is likely to be slightly lower than the value estimated from the pumping test. The grain size of the basal

Table 5.1. The array of transmissivities, T_{ij} ($ft.^2$/day) used in Model A. The uncertainty in T_{ij} is ± 33%. The pumping-test site was located in cell 11,6.

Col. / Row	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
1	-	-	-	-	-	1500	1500	1500	1500	1500	1500	-	-	-	-
2	-	-	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	-	-	-
3	-	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	-
4	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500
5	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	-
6	-	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	-	-
7	-	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	-	-	-	-
8	-	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	-	-	-	-
9	-	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	-	-	-
10	-	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500	-	-	-
11	-	4600	4600	1500	1500	7600	1500	76	76	1500	1500	1500	-	-	-
12	-	4600	4600	1500	4600	1500	1500	76	76	76	1500	-	-	-	-
13	-	-	4600	1500	4600	1500	1500	76	76	76	-	-	-	-	-
14	-	-	-	1500	1500	4600	1500	76	76	-	-	-	-	-	-
15	-	-	-	-	-	1500	1500	76	76	-	-	-	-	-	-
16	-	-	-	-	-	-	-	1500	-	-	-	-	-	-	-

sandstone in test hole W8 (Figure 2.6), which is within the area of Model A (Figure 2.2), is significantly finer than at the site of the pumping test (Figure 2.9).

Drillers logs from the area between Parsons Creek and Peats Creek (Figure 5.2) show that the sandstone here is thin where present (less than 10 ft.). The contours on the potentiometric surface are particularly close to each other and the hydraulic gradient is therefore steep (Figure 5.2). This is evidence that the transmissivity is significantly lower here than elsewhere in the model area. In order to accomplish a better fit between observed and computed heads in this area, the transmissivity was estimated to be 76 $ft.^2$/day (Table 5.1), one hundred times lower than at the site of the pumping test.

There is an uncertainty of approximately ± 33% in the values of T_{ij} used. This is primarily due to the difficulty in quantifying the effects of ii and iii above.

5.1{a}.4.4. Interpolated Heads

A head value was interpolated at the center of each grid square in the model area from the contour map of the potentiometric surface (Figure 5.2) and from the elevations of streams taken from topographic maps. These heads are tabulated in Table 5.2. The interpolated heads are not accurate to closer than +/- 10 ft., reflecting the limited accuracy of the potentiometric surface (Chapter 3).

5.1{a}.5. Results

5.1{a}.5. 1. Recharge Rate

Using the transmissivities in Table 5.1, the sum of the squared residuals (Σ sq. res.) between computed and interpolated heads is a minimum of 6586 $ft.^2$ using a recharge rate of 0.26 ± 0.09 in./yr. (Figure 5.4). The heads computed using this recharge rate are tabulated in Table 5.3.

5.1{a}.5. 2. Volumetric Flow Budgets

The volumetric flow rates estimated in the steady-state simulation with a recharge rate of 0.26 ± 0.09 in./yr. are tabulated in Table 5.4. Volumetric flow rates are all 33% higher or lower with a respective 33% increase or decrease in recharge rate and transmissivity. However, in each case, less than 6% of the total annual flow is inflow at the western constant-head inflow boundary, between 12% and 14%

Table 5.2. Heads in the basal sandstone interpolated for the cells of Model A from the potentiometric surface contours.

Col. / Row	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
1	-	-	-	-	-	1370	1360	1350	1340	1330	1316	-	-	-	-
2	-	-	1392	1384	1376	1366	1361	1356	1352	1346	1330	1300	-	-	-
3	-	1395	1394	1391	1383	1371	1367	1363	1360	1355	1344	1330	1312	1290	-
4	1393	1393	1394	1392	1387	1380	1375	1370	1366	1361	1352	1340	1328	1324	1310
5	1388	1389	1390	1388	1383	1379	1375	1372	1369	1364	1356	1350	1343	1330	-
6	-	1378	1378	1376	1372	1368	1365	1365	1366	1366	1364	1358	1355	-	-
7	-	1365	1365	1364	1363	1361	1355	1352	1358	1362	1363	-	-	-	-
8	-	1350	1353	1354	1352	1348	1341	1337	1350	1356	1356	-	-	-	-
9	-	1333	1337	1338	1338	1335	1328	1326	1339	1350	1348	1345	-	-	-
10	-	1317	1318	1321	1322	1319	1310	1305	1329	1338	1339	1330	-	-	-
11	-	1300	1300	1302	1304	1304	1292	1292	1324	1332	1331	1315	-	-	-
12	-	1285	1282	1279	1282	1282	1275	1295	1322	1312	1305	-	-	-	-
13	-	-	1270	1267	1270	1267	1258	1295	1295	1285	-	-	-	-	-
14	-	-	-	1264	1260	1258	1247	1293	1270	-	-	-	-	-	-
15	-	-	-	-	-	1244	1235	1270	1252	-	-	-	-	-	-
16	-	-	-	-	-	-	-	1238	-	-	-	-	-	-	-

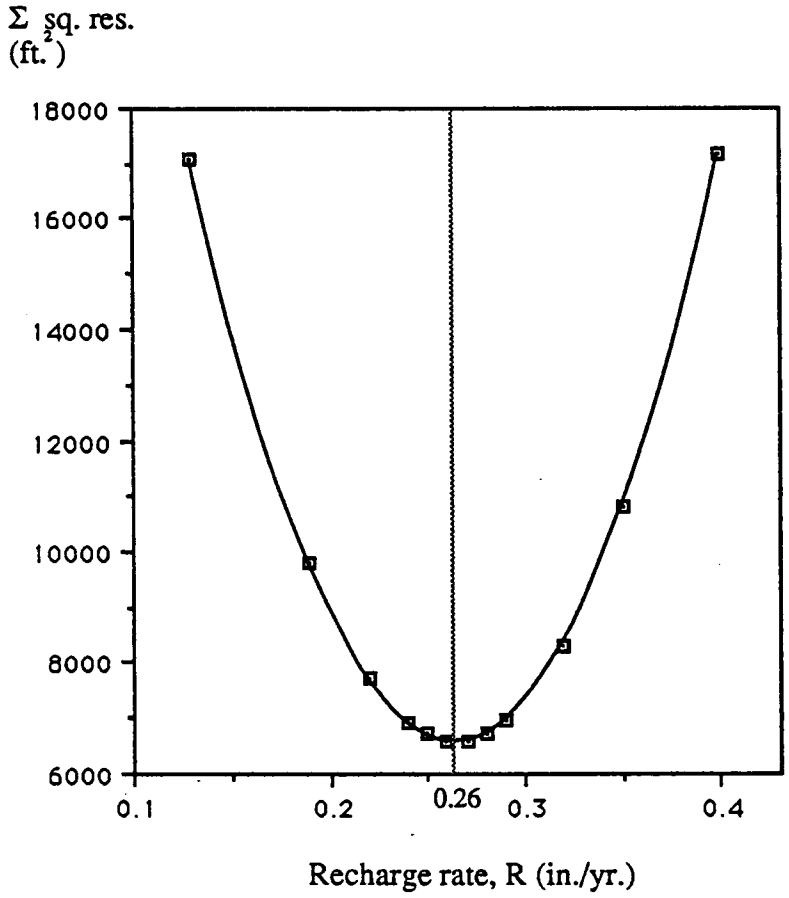


Figure 5.4. The effect of varying the recharge rate (Model A) on the sum of the squared residuals (Σ sq. res.) between heads interpolated from the potentiometric surface (Table 5.2) and computed heads (Table 5.3). The array of transmissivities used in Model A is shown in Table 5.1. Σ sq. res. is a measure of how well the computed data fit the interpolated data. The best fit is obtained when Σ sq. res. is a minimum, i.e. when R is 0.26 in./yr.

Table 5.3. Heads (ft.) computed in the steady-state simulation of Model A using a recharge rate of 0.26 in./yr.

Col. / Row	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
1	-	-	-	-	-	1370	1360	1350	1340	1330	1316	-	-	-	-
2	-	-	1392	1388	1382	1374	1365	1357	1347	1336	1322	1300	-	-	-
3	-	1395	1391	1386	1381	1375	1368	1361	1353	1344	1333	1321	1312	1290	-
4	1393	1391	1387	1384	1379	1375	1370	1364	1358	1351	1343	1335	1326	1314	1310
5	1390	1385	1382	1379	1376	1372	1369	1365	1361	1357	1352	1347	1339	1328	-
6	-	1375	1373	1372	1370	1368	1366	1364	1362	1360	1359	1358	1355	-	-
7	-	1364	1363	1362	1361	1361	1360	1360	1360	1360	1363	-	-	-	-
8	-	1350	1350	1350	1350	1351	1352	1354	1355	1355	1356	-	-	-	-
9	-	1335	1335	1336	1336	1337	1341	1346	1348	1348	1345	1339	-	-	-
10	-	1316	1316	1319	1320	1319	1325	1337	1342	1340	1335	1330	-	-	-
11	-	1302	1300	1303	1301	1302	1302	1292	1337	1332	1322	1315	-	-	-
12	-	1296	1289	1288	1286	1286	1275	1302	1330	1323	1305	-	-	-	-
13	-	-	1270	1275	1275	1268	1258	1289	1306	1285	-	-	-	-	-
14	-	-	-	1264	1260	1255	1247	1272	1270	-	-	-	-	-	-
15	-	-	-	-	-	1244	1235	1254	1252	-	-	-	-	-	-
16	-	-	-	-	-	-	-	1238	-	-	-	-	-	-	-

Table 5.4. Volumetric flow-rate estimates in Model A under steady-state conditions with no pumping considered. Increasing or decreasing the recharge rate and transmissivity by 33% creates the same set of computed heads (Table 5.3) but increases or decreases the volumetric flow rates by 33% respectively.

	1. Flow rates (acre-ft./yr.) with a recharge rate of 0.26 in./yr. and transmissivities set as in Table 5.1	2. Flow rates (acre-ft./yr.) with recharge rate and transmissivities 33% greater	3. Flow rates (acre-ft./yr.) with recharge rate and transmissivities 33% lower
Inflow from recharge	3780	5040	2520
Inflow across western constant-head bndry	250	370	180
Inflow across eastern constant-head boundary	570	860	430
Total inflow	4700	6270	3130
Outflow in Mill Creek valley discharge zone	1670	2230	1110
Outflow in Republican River valley discharge zone	3030	4040	2020
Total outflow	4700	6270	3130

is inflow at the eastern boundary, and approximately 81% is recharge from rainfall within the model area. Sixty-four percent of the discharge is along the boundary with the Republican River aquifer and the lower reaches of its tributaries, along the southern edge of the model area.

5.1{a}.6. Discussion

5.1{a}.6.1. Recharge Rate

The minimum value of Σ sq. res. (Figure 5.4) represents a standard deviation between the computed and interpolated heads of 7.4 ft. This deviation shows that the array of transmissivities used in the model (Table 5.1) is not a perfectly accurate simulation of the real transmissivity, which is likely to vary over this area more than the estimates used (Table 5.1). Adjusting the transmissivities on a cell-by-cell basis by trial and error in order to perfectly simulate the potentiometric surface would not have been a worthwhile exercise in this model, considering the limited accuracy of the heads interpolated from the potentiometric surface. However, the recognition of the relatively high transmissivity in the vicinity of the pumping-test site and the relatively low transmissivity further east is important, because it considerably reduces Σ sq. res. in comparison with using a uniform transmissivity throughout.

For the uncertainty range of ± 33% in the estimated transmissivities, the maximum recharge rate is 0.35 in./yr. and the minimum is 0.17 in./yr., assuming initial assumptions are correct. If there is significant leakage at the base of the sandstone into underlying Permian shales, the recharge rate could be higher than this range. The minimum vertical hydraulic conductivity (k') of the mudstone confining layer estimated from the pumping test is $8.0(\pm 1.3)\times 10^{-4}$ ft./day. From Darcy's Law, a hydraulic gradient of 0.1 ft./ft., well within the range possible, would be needed for this recharge rate to be transmitted to the basal sandstone through such a mudstone. Assuming k' at the pumping-test site is representative of k' elsewhere in the model area, the full range of possible recharge rates estimated using Model A could easily be transmitted through the confining layer.

I estimated the recharge rate from the baseflow of Mill Creek in order to corroborate the model estimate. Mill Creek drains an area of 326 sq. mi. upstream from a gauging station at the City of Washington. This watershed has the same climate, vegetation, soil types, soil conservation measures, and crop types as the

model area. They are in the same county, approximately the same size, and overlap to some extent. Ninety-nine percent of the watershed is underlain by the Dakota aquifer. The gauging station is located just below the base of the Dakota and is therefore well located to measure the total discharge from the aquifer, provided there is no significant evapotranspiration or water use from the creek at the time of flow measurement. Evapotranspiration and water use from the creek are at a minimum during winter so this is the best time of year to estimate the discharge rate from the formation.

During 1974 the total precipitation at Washington and Haddam, 15 mi. upstream from Washington, was only 19.6 in. and 15.5 in. respectively (NOAA, 1975). This compares with a mean annual precipitation of 29 in. (Dugan and Peckenpaugh, 1985). The flow rate of Mill Creek at the Washington gauging station reached a late-December minimum of 8.0 CFS (USGS, 1975). During this month, no significant precipitation was recorded and evapotranspiration was likely to be low. Assuming steady-state conditions, negligible flow of ground water to or from underlying Permian sediments, and no inflow from a regional flow system to the west, the aquifer must be receiving recharge at a rate of 8.0 CFS. Over the 326 sq. mi. area of the watershed, this corresponds to an annual recharge rate of 0.33 in./yr. Regional flow from outside the watershed and pumping of the aquifer both affect the accuracy of this estimate, although in opposite senses. Therefore, the estimate of recharge based on ground-water discharge to Mill Creek is likely to be a good one in terms of its order of magnitude but is not accurate to closer than ± 50%.

Dugan and Peckenpaugh (1985) studied potential recharge rates for an eight-state region in the Midwest (the central Midwest regional aquifer system). Because they were studying a large area, they used a grid size in which each cell was larger than the area of the entire model described above. However, they took into account climate, vegetation, soil type, and water use. The result they obtained for central Washington County, Kansas is a potential recharge rate of approximately 3 in./yr. This can be considered the maximum mean recharge rate possible.

The recharge-rate estimate of Model A is corroborated by these two alternate methods of recharge estimation. The estimate of Model A is an order a magnitude below the maximum rate estimated by Dugan and Peckenpaugh (1985), and comparable with the recharge rate estimated from the baseflow of Mill Creek.

5.1{a}.6.2. Volumetric Flow Budgets

A few miles beyond the western edge of the model area the basal sandstone contains saline water of the regional flow system, some of which discharges into the Salt Creek watershed (Chapter 3). The fact that less than 6% of the inflow into the model area is via the western boundary (Table 5.4) is significant. Most of this in-flowing water is also locally recharged. Therefore there is little opportunity for saline water further west to enter the model area despite the good lateral continuity of the basal sandstone. Ground water in the aquifer to the east of the model area is all locally recharged and is not affected by salinity problems. Therefore, inflow from the east is unlikely to cause any water quality problems.

5.1{b}. DETERMINATION OF THE EFFECTS OF PUMPING (Transient Conditions)

5.1{b}.1. Model Setup

Model A was set up as in Section 5.1{a} but with the additional consideration of pumping of the aquifer, with estimates of current pumping rates based on DWR appropriations. The model solved for the hydraulic head over a simulated stress period of 1000 years to ensure the system that final heads were equilibrated to the new conditions. Transmissivities and recharge rates from Section 5.1{a} were used in the transient simulations in order to estimate drawdowns and flow budgets under different conditions.

5.1{b}.2. Governing Equation

Using square grid cells, the alternating-direction implicit transient finite-difference approximation for the nth time step, linking hydraulic head (h) at a node i,j, distance between nodes (Δx), recharge rate (R), storativity (S), transmissivity (T), pumping rate (Q) and time step Δt, is

$$\frac{\left(h_{i-1,j} + h_{i+1,j} + h_{i,j-1} + h_{i,j+1} - 4h_{ij}\right)^{n+1}}{(\Delta x)^2} = \frac{S\left(h_{i,j}^{n+1} - h_{i,j}^{n}\right)}{T_{i,j}\Delta t} + \frac{Q_{i,j}}{T_{i,j}(\Delta x)^2} - \frac{R}{T_{ij}}.$$

The equation used in MODFLOW (McDonald and Harbaugh, 1988) also takes into account transmissivities in adjacent cells. The important advantage of

implicit finite-difference techniques over explicit techniques is that for each grid cell, new time-step head values from surrounding cells are used to calculate the new head value of the cell in question. This generally means that a solution is arrived at more quickly. The assumptions of isotropy, uniform recharge rate, and no leakage in Section 5.1{a}were also made in the transient simulations.

5.1{b}.3. Model Inputs

5.1{b}.3.1. Grid, Boundaries, Transmissivities, Recharge Rate

The grid, boundaries, transmissivities, and recharge rate of part {a} were also used in the transient simulations of Section 5.1{b}. The initial heads used to estimate drawdown due to pumping were the computed heads of the steady-state model (Table 5.3). Transmissivity and recharge rate were varied together by as much as ± 50% in order to investigate the sensitivity of the transient model to these parameters.

5.1{b}.3.2. Storativity

A value of storativity was entered for each cell in the model area for the transient simulations (Table 5.5). Although the Dakota Formation crops out throughout most of this model area without being "confined" by younger strata, the basal sandstone of the aquifer is effectively confined in a hydrogeological sense in all the active grid cells by Dakota mudstones.

At the site of the pumping test, the specific storage of the basal sandstone was estimated to be 1.28E-06 $ft.^{-1}$ (Chapter 4). This specific storage was assumed to be uniform in the basal sandstone throughout the model area. This is a good assumption because the degree of cementation of the basal sandstone was poor in all the test holes throughout the study area.

The thickness of the sandstone, b, varies and, therefore, so does its storativity. Thickness was estimated to be 50 ft. throughout most of the model area. However, it was set at 60 ft. in grid cells in the area of high-yield irrigation wells in T5S, R1E and 10 ft. in the low-transmissivity region to the east of Parsons Creek (Table 5.5).

Table 5.5. Values of storativity (S) used in the transient simulations. S was estimated to be 1.28E-4 in grid cell (11,5) from the pumping test. This table shows the storativities of all the grid cells within the model area expressed as a multiple of 1.28E-4.

Row Col.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
1	0	0	0	0	0	0.5	0.5	0.5	0.5	0.5	0.5	0	0	0	0
2	0	0	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0	0	0
3	0	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0
4	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5
5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0	0
6	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0	0	0	0
7	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0	0	0	0
8	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0	0	0	0
9	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0	0	0
10	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0	0	0
11	0.5	0.6	0.6	0.5	0.5	1.0	0.5	0.1	0.1	0.1	0.1	0.1	0	0	0
12	0.5	0.6	0.6	0.5	0.6	0.5	0.5	0.1	0.1	0.1	0.1	0	0	0	0
13	0	0	0.6	0.5	0.6	0.5	0.5	0.1	0.1	0.1	0	0	0	0	0
14	0	0	0	0.5	0.5	0.6	0.5	0.1	0.1	0	0	0	0	0	0
15	0	0	0	0	0	0.5	0.5	0.1	0.1	0	0	0	0	0	0
16	0	0	0	0	0	0	0	0.5	0	0	0	0	0	0	0

5.1(b).3.3. Rate of Ground-Water Withdrawal

Irrigation wells screened in the Dakota aquifer have been in use in the model area for several years. These are all screened in the basal sandstone. Table 5.6 shows the current rate of ground-water withdrawal via these wells, which was estimated from individual irrigation-well appropriations from the Division of Water Resources (DWR). The mean annual DWR estimate of water use was found for each well for which there are existing records. Total water use from a group of irrigation wells in Washington County including those in T5S, R1E, Sections 16 and 17 in Table 5.6, is significantly less than the total estimated by DWR (Appendix 6). Actual pumpage was 78% of the DWR estimate of pumpage. Therefore the water-use estimates for unmetered wells listed in Table 5.6 are 78% of the values determined directly from DWR water-use data. Mean annual water use for wells which are listed in Table A6.1, Appendix 6, and for which the pumping rate is therefore known, were determined directly from this rate and the number of hours pumped. For irrigation wells with no known record, a pumping rate of 47 acre-ft./yr. was estimated. This is the mean rate from wells which do have records.

Table 5.6. Current Rates of Ground-Water Withdrawal from High-Yield Wells

Location	Grid cell (row, column)	Number of wells	Mean annual withdrawal rate (acre-ft./yr.)	Source
T3S R2E 26	5, 11	1	47	estimated
T3S R2E 36	6, 11	1	47	estimated
T4S R2E 24	9, 11	2	81	DWR, App.1
T5S R1E 1	11, 7	1	53	DWR, App.1
T5S R1E 10	11, 6	2	136	DWR, App.1
T5S R1E 16	12, 5	1	56	DWR, App.1
T5S R1E 17	12, 5	3	124	DWR, App.1
T5S R1E 20	13, 5	1	22	DWR, App.1
T5S R1E 30	13, 4	1	47	estimated
T5S R2E 1	10, 11	2	94	DWR, App.1
T5S R1W 12	11, 3	1	47	estimated
TOTAL		16	754	estimated

In addition to the high-capacity wells listed above, there are records of 70 domestic and stock wells screened in the basal sandstone in the model area. If these were all pumped at 300 GPD, the total annual pumpage from domestic wells would be no greater than 25 acre-ft. In addition there are likely to be a similar number of low-capacity wells, such as old windmill wells, which are in use but which have no official records. The total annual pumpage from these wells is also likely to be no greater than 25 acre-ft. The total annual pumpage from all low-capacity wells in the basal sandstone was therefore estimated to be 50 acre-ft., which is approximately equivalent to the annual pumpage of one typical irrigation well.

Current rates of ground-water withdrawal were simulated using the well module of MODFLOW (McDonald and Harbaugh, 1988). The estimated ground-water withdrawal of 754 acre-ft. from high-capacity wells was applied to nodes of the relevant grid cells (Table 5.6). Withdrawals attributed to low-capacity wells were evenly distributed throughout the active nodes of the model. Pumping was assumed to be uniform throughout the year. In reality, pumping of high-capacity irrigation wells is seasonal and therefore the drawdown would vary between different seasons.

Simulations were performed using the above estimate of current pumping rate to investigate its effects on water levels and flow budgets in the aquifer. The pumping rate was then doubled throughout the model area in order to investigate the effects of increased water use.

5.1{b}.4. Results

5.1{b}.4.1. Drawdown

Drawdowns computed by Model A, with a recharge rate of 0.26 in./yr. and transmissivities set as in Table 5.1, are illustrated in Figure 5.5 and documented in Table 5.7. The maximum computed drawdowns are in the southwestern part of the model area (Figure 5.5). More than 95% of the drawdown was induced within a year of the onset of pumping in all transient simulations, but pumping was simulated for a total of 1000 years to ensure the system had equilibrated.

The effects on the computed drawdown of varying the transmissivity and recharge together by up to 50% and doubling the pumping rate are documented in

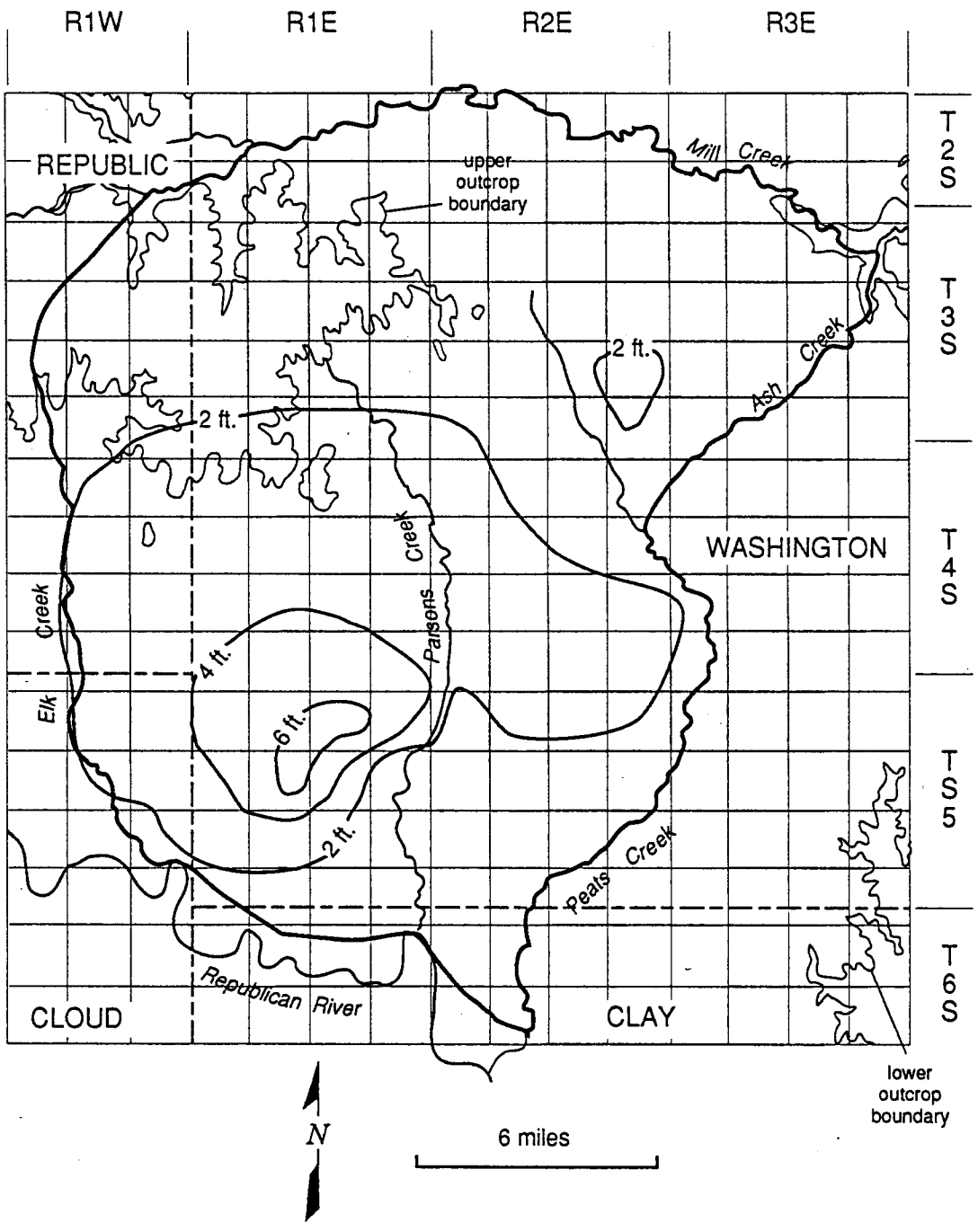


Figure 5.5. Drawdown predicted by Model A for a pumping rate of 800 acre-ft./yr., a recharge rate of 0.26 in./yr., and transmissivities set as in Table 5.1.

Table 5.8 and illustrated in Figure 5.6. At the current estimated pumping rate of 800 acre-ft./yr., mean drawdown ranges between 1.6 ft. and 3.2 ft. and maximum drawdown ranges between 5.1 ft. and 10.2 ft. if the recharge rate and transmissivity are varied by ±33% (Table 5.8). Figure 5.6 shows that lowering the transmissivity and recharge-rate estimates by 33% causes a greater change in drawdown than raising them by 33%. If the pumping rate is doubled, drawdowns are also doubled; the mean drawdown ranges between 3.2 ft. and 6.4 ft. and maximum drawdown ranges between 10.2 ft. and 20.3 ft. for the same range of recharge rate and transmissivity (Table 5.8).

Table 5.7. Drawdowns induced by pumping the aquifer at the estimated current rate of ground-water usage (ft.). Mean drawdown = 2.1 ft.

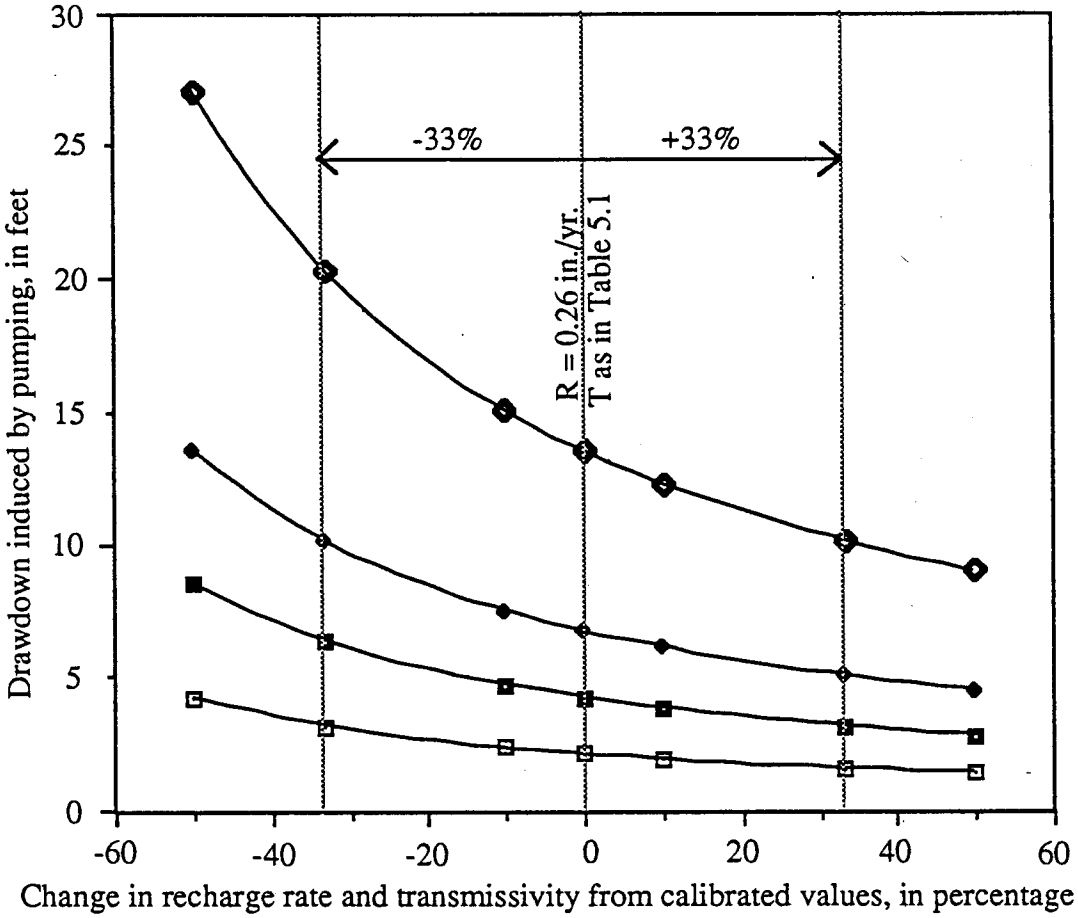
Row Col.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
2	0	0	0	0.5	0.6	0.5	0.5	0.5	0.5	0.4	0.3	0	0	0	0
3	0	0	0.4	0.7	0.9	0.9	0.9	0.9	0.9	0.8	0.7	0.4	0	0	0
4	0	0.5	0.9	1.1	1.3	1.3	1.3	1.3	1.3	1.3	1.3	0.7	0.3	0.2	0
5	.6	1.1	1.4	1.6	1.7	1.8	1.7	1.7	1.7	1.8	2.4	0.9	0.4	0.3	0
6	0	1.8	2.0	2.1	2.2	2.2	2.1	2.0	1.9	1.8	2.0	0	0	0	0
7	0	2.4	2.5	2.6	2.7	2.7	2.6	2.4	2.0	1.4	0	0	0	0	0
8	0	2.9	3.0	3.1	3.3	3.3	3.1	2.8	2.3	1.6	0	0	0	0	0
9	0	3.2	3.4	3.6	3.9	4.1	3.7	3.2	2.9	2.8	3.8	1.9	0	0	0
10	0	3.4	3.6	4.1	4.8	5.3	4.5	3.5	3.1	3.0	3.9	0	0	0	0
11	0	3.3	3.6	4.3	5.7	6.8	5.3	0	2.1	2.3	1.6	0	0	0	0
12	0	2.9	2.4	4.0	6.2	4.4	0	0.5	1.2	1.1	0	0	0	0	0
13	0	0	0	2.6	3.7	2.1	0	0.4	0.6	0	0	0	0	0	0
14	0	0	0	0	0	0.5	0	0.3	0	0	0	0	0	0	0
15	0	0	0	0	0	0	0	0.2	0	0	0	0	0	0	0
16	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0

5.1{b}.4.2. Volumetric Flow Budgets

The volumetric flow budgets estimated during pumping of the aquifer are documented in Table 5.9. With pumping simulated at the current estimated rate, and a recharge rate of 0.26 in./yr., recharge contributes approximately 77% of the total inflow into the system and wells account for 16% of the total outflow (Table 5.9, Column 1). If the pumping rate is doubled with the same recharge rate (Table

Table 5.8. Drawdowns predicted by Model A for various recharge rates, transmissivities, and pumping rates. (These are the drawdowns after the aquifer has equilibrated to the pumping.) These data are plotted in Figure 5.6. The maximum drawdowns occur in cell (11,6) which has the highest pumping rate (Table 5.6).

	Drawdown (ft.) with R = 0.26 in./yr. and T set as in Table 5.1	Drawdown (ft.) with R and T **10% greater**	Drawdown (ft.) with R and T **33% greater**	Drawdown (ft.) with R and T **50% greater**	Drawdown (ft.) with R and T **10% lower**	Drawdown (ft.) with R and T **33% lower**	Drawdown (ft.) with R and T **50% lower**
Current pumping rate	Mean = 2.1	Mean = 1.9	Mean = 1.6	Mean = 1.4	Mean = 2.4	Mean = 3.2	Mean = 4.3
(800 acre-ft./yr.)	Max. = 6.8	Max. = 6.2	Max. = 5.1	Max. = 4.5	Max. = 7.5	Max. = 10.2	Max. = 13.5
2 X current pumping rate	Mean = 4.3	Mean = 3.9	Mean = 3.2	Mean = 2.8	Mean = 4.7	Mean = 6.4	Mean = 8.5
(1600 acre-ft./yr.)	Max. = 13.5	Max. = 12.3	Max. = 10.2	Max. = 9.0	Max. = 15.0	Max. = 20.3	Max. = 27.1



- Mean drawdown in the model area due to pumping at current rate
- Maximum drawdown in the model area (cell 11,6) due to pumping at current rate
- Mean drawdown in the model area due to pumping at **double** current rate
- Maximum drawdown in the model area (cell 11,6) due to pumping at **double** current rate

Figure 5.6. The effect of the choice of Model A recharge rate and transmissivity on estimations of mean and maximum drawdown induced by pumping. Pumping is considered both at the current estimated pumping rate and at double this rate.

Table 5.9. Volumetric flow-rate estimates in Model A after the aquifer has equilibrated to a steady state, with pumping considered. Pumping rate is 800 acre-ft./yr., the estimated current rate of pumping, except in the final column, in which the pumping rate is double the estimated current rate. Volumetric flow rates are all 33% higher or lower with a respective 33% increase or decrease in recharge rate (R) and transmissivity (T).

	1. Flow rates (acre-ft./yr.) with R = 0.26 in./yr. and T set as in Table 5.1	2. Flow rates (acre-ft./yr.) with R and T **33%** greater	3. Flow rates (acre-ft./yr.) with R and T **33%** lower	4. Flow rates (acre-ft./yr.) with a R of 0.26 in./yr. and T set as in Table 5.1
Inflow from recharge	3780	5040	2520	3780
Inflow across western constant-head bndry	300	400	220	360
Inflow across eastern constant-head boundary	800	1010	590	1000
Total inflow	4890	6450	3330	5140
Outflow in Mill Creek valley discharge zone	1630	2180	1070	1600
Outflow in Republican River valley discharge zone	2450	3470	1460	1930
Discharge to wells	800	800	800	1610
Total outflow	4890	6450	3330	5140

5.9, Column 4), recharge is still the main source of inflow (74%) and 31% of the outflow is via wells (Table 5.9). The proportion of inflow derived from recharge remains the same if the recharge rate and transmissivity are both varied by ±33% but the proportion of total outflow which is discharge to wells is significantly affected (Table 5.9, Columns 2 and 3).

When recharge rate and transmissivity are both increased or decreased, the constant-head boundary flow budgets increase or decrease respectively in proportion (Table 5.9, Columns 2 and 3). Inflow to the system via the western constant-head inflow boundary is not greater than 7% of the total flow budget in any of the scenarios in Table 5.9. Inflow across the eastern constant-head boundary increases from 800 acre-ft./yr. (16% of the total flow) to 1000 acre-ft./yr. (20% of the total flow) if the pumping rate is doubled (Table 5.9, Columns 1 and 4).

When the pumping rate is doubled, the discharge rate at the Republican River valley boundary of the model decreases by 520 acre-ft./yr. to 1930 acre-ft./yr. In contrast, discharge in the Mill Creek valley is hardly affected by an increase in pumping rate (Table 5.9, Columns 1 and 4). With the recharge rate and transmissivity set at 33% or more above their estimated values in Section 5.1{a}, at a pumping rate of 1600 acre-ft./yr., inflow begins to occur via constant-head cells which are discharge cells at the lower pumping rate.

5.1{b}.5. Discussion

The mean drawdowns induced by pumping the aquifer at the current estimated rate of ground-water withdrawal (800 acre-ft./yr.) were minimal for the uncertainty range of the transmissivity and recharge (Table 5.8). This is compatible with the fact that no significant drops in water level have been reported in this area. No hydrographs from within the model area are available. However, a hydrograph from an irrigation well (T5S R2W Sec. 1BA) 2 mi. west of cell (10,1) outside the model area (Figure 5.3) shows an upward trend in water level between 1970 and 1988 (Macfarlane et al., 1990). Therefore, pumping within and to the west of the model area during this period has not lowered water levels at this site.

The area of greatest drawdown in Figure 5.5 corresponds to the area of high-yield irrigation wells in southwestern Washington County (Figure 4.1). The maximum computed drawdown for the full uncertainty range of transmissivity and recharge (10.2 ft., Table 5.8) is in cell (11,6), which is subject to the highest

maximum computed drawdown for the full uncertainty range of transmissivity and recharge (10.2 ft., Table 5.8) is in cell (11,6), which is subject to the highest pumping rate (Table 5.6), and includes the site of the pumping test (Figure 4.1). Drawdown of this magnitude was not observed at this site except during the irrigation season within 270 ft. of a high-yield pumping well (Chapter 4). Therefore the recharge rate is unlikely to be as low as 0.17 in./yr., the lower limit of its uncertainty range in Section 5.1{a}. As expected, doubling the ground-water withdrawal from the system doubled the drawdowns (Table 5.8).

The quick response of the system to pumping shows it is capable of equilibrating itself to new stresses in a relatively short time. This fact, combined with the low drawdowns, means that the assumption that the system is in a steady state, which was made to simplify the estimation of the recharge rate, is reasonable.

Pumping the aquifer at up to double the current rate does not significantly increase the inflow at the western boundary (Tables 5.9 and 5.4). Therefore, with these boundary conditions there is little or no opportunity for saline water further west to enter the model area as a result of pumping. This is compatible with the fact that no significant changes in water quality have been reported in the model area. However, a hydraulic gradient is induced across the western boundary (rows 7 to 13, columns 1 and 2) by drawdowns in column 2 (Table 5.7). Drawdowns are doubled by doubling the pumping rate (Table 5.8), and consequently the hydraulic gradient across this boundary is doubled. This suggests that the no-flow boundary used along the western edge of the model under steady-state conditions may not be valid when the aquifer is pumped. With a constant-head boundary simulated instead, the western boundary transmits 15% of the total flow at the current pumping rate and 17% at double this rate, compared to 6% with no pumping. However, an increase in development of the aquifer in the model area would most likely be accompanied by an increase in development outside the model boundaries. Therefore, drawdown of the water level in the aquifer would be induced on both sides of the boundary of the model area and any hydraulic gradient across the western no-flow boundary of the model area would be small. Boundary conditions of the steady-state simulations were not changed for the transient simulations (Figure 5.2).

The proportion of total flow entering the model area at its eastern boundary increases from 12% with no pumping (Table 5.4, Column 1) to 16% when the

rate of pumping, the inflow rate across the eastern boundary estimated by Model A (between 590 and 1010 acre-ft., Table 5.9) could not be sustained. Greater drawdowns may therefore be induced in the east-central upland part of the model area than are shown in Table 5.7 and Figure 5.5. Historical water-level data would be needed to validate the model boundary condition in this area.

Most of the water pumped in the simulation of current pumping conditions can be accounted for in the interception of natural discharge from the aquifer along the Republican River valley at the southern boundary of the model. Discharge to the Republican River valley is 580 acre-ft./yr. (19%) lower when the aquifer is pumped at the current rate (Table 5.9, Column 1) than in its pre-development state (Table 5.4, Column 1). This discharge rate would be further reduced to 64% of its pre-development value if the pumping rate were doubled (Table 5.9, Column 4). In contrast, pumping has little effect on the discharge rate in the Mill Creek valley (Tables 5.4, Column 1 and Table 5.9, Columns 1 and 4). The reason for these differences is the location of the area of highest use in T5S, R1E, close to the Republican River valley. If most of the ground-water withdrawal in this area were concentrated in the central part of the model area, it is likely there would be much greater drawdown in this area, and less discharge in the Mill Creek valley. The underlying principal of ground-water development is that, by withdrawing water from an aquifer, some of the natural discharge may be made available for use (Peters, 1972). However, the interception of ground water discharging along the southern boundary of the model area reduces the recharge from the Dakota aquifer to the Republican River alluvial aquifer.

The aquifer can support the current level of pumping in the area of Model A with no detrimental effects on water availability or quality. Doubling the pumping rate in this area would be likely to significantly reduce the mean annual water level in the aquifer in the areas of highest use, but not to the extent of producing severe water-level declines in the aquifer in general, or adversely affecting its water quality.

5.2. MODELS B AND C; CROSS-SECTIONAL MODELS

5.2.1. Introduction

The purpose of Models B and C was to establish whether consideration of additional layers significantly affects the estimate of the recharge rate needed to sustain the potentiometric surface of the basal sandstone. MODFLOW (McDonald and Harbaugh, 1988) was used to simulate two-dimensional cross-sectional flow.

The upper part of the Dakota aquifer consists of many sandstones which are poorly connected to each other and separated from the basal sandstone by a mudstone aquitard (Chapter 2). Flow was simulated in a three-layer system along two sections with an upper sandstone aquifer of unknown hydraulic conductivity, an intervening mudstone aquitard, and the basal sandstone. This layered approach is a better representation of the true makeup of the aquifer than the one-layer areal model, Model A, although it is not an exact representation. Sandstones above the basal sandstone receive recharge from the surface and transmit some of this water horizontally to discharge into streams. It was hypothesized that this diversion of potential recharge away from the basal sandstone might mean that the recharge rate estimated using Model A may be too low because horizontal flow in upper sandstones and the discharge of ground water from upper flow systems are not considered by Model A.

The two cross sections along flow lines of the basal sandstone were each simulated as two-layer systems. The lines of these sections both begin in the high ground of T3S, R1E, in west-central Washington County (Figure 5.2). Section B is a north-south line which ends at the Republican River (Figure 5.2). It was selected because it is a flow line through the basal sandstone of an area in which there is a relatively high density of drillers' logs and water-level data, including the area of high-yield irrigation wells in southern Washington County (Figure 4.1). Section C follows a line to the northeast, ending at Mill Creek (Figure 5.2). It was selected because it is approximately perpendicular to Section B and located on the other side of a ground-water divide in a different region of Model A. The density of data available for Section C was considerably lower than for Section B.

Model B, which simulated flow in Section B, was used to confirm the recharge rate estimated using Model A. Model C, which simulated flow along Section C, was used to estimate the transmissivities of both aquifer layers and to

reveal variations in transmissivity in the basal sandstone. Due to the relative lack of data along Section C, this model was not used to confirm the recharge rate estimated using the areal model.

5.2.2. Model Setup

Although this is a three-layer system, it can be simulated by MODFLOW (McDonald and Harbaugh, 1988) as a two-layer system with a vertical conductance term which simulates the vertical flow through the aquitard between the two layers. This assumes no significant horizontal flow in the mudstone aquitard. This assumption does not introduce any significant errors because if the contrast between the hydraulic conductivities of the aquifers and the hydraulic conductivity of the intervening aquitard is large, flow is effectively vertical in the aquitard.

For each model, a two-dimensional, two-layer grid was set up as described below with known or interpolated heads entered for each grid cell. In Model B upper-boundary recharge rate and the transmissivity of the upper layer (Layer 1) were varied until the sum of the squared differences between computed and known (or interpolated) hydraulic heads in both layers was minimized. The transmissivity of the basal sandstone (Layer 2) was held constant. In Model C transmissivity was varied in the same way in both layers while the recharge rate was held constant: the recharge rate used in Model C was the mean of the recharge rates estimated using Model A and Model B. In both cross-sectional models, the recharge rate and the transmissivity of Layer 1 were assumed to be uniform across the section.

5.2.3. Simplifying Assumptions

The following simplifying assumptions were made for Models B and C. They are the same assumptions as were made for Model A (above).

1. The recharge rate is uniform throughout the model area.
2. The system is in a steady state.
3. There is no leakage to or from the underlying Permian shale.

5.2.4. Model Inputs

5.2.4.1. Grid

Figures 5.7 and 5.8 show the geologic and hydrologic data of Sections B and C respectively. The upper part of the Dakota Formation is not present at the southern end of Section B or the eastern end of Section C. Grids used to simulate flow in the sections are illustrated in Figures 5.9 and 5.10.

In Section B, Layer 2 was divided into 16 cells with the 8 cells of Layer 1 overlying it at the northern end of the section (Figure 5.9). The density of data along the line of section is variable. For this reason, the grid-cell length used in the simulation is smallest (0.5 mi.) in parts of the section where head data are relatively dense. The grid was set up in such a way that boundaries between grid cells in Model A coincide with boundaries in Model B (Figure 5.9). Widths of most grid cells used in Model B are therefore multiples of 0.5 mi. However, at its northern end, Section B deviates slightly from true north-south (Figure 5.2), and so for grid boundaries to line up with the plan-view grid the four most northern cells of Model B are multiples of 0.53 mi. (Tables 5.10 and 5.11).

The density and reliability of data in Section C are not as good as in Section B; only one of the water levels was measured by the KGS (Figure 5.8, Table 5.12). For Model C, Layer 2 was divided into eight cells and Layer 1 into three cells (Figure 5.10). As in Model B, the grid-cell length used in the simulation is smallest (0.5 mi.) in parts of the section where head data are relatively dense (Figure 5.10). However, the largest grid-cell length used in Model C is 4.5 mi., more than double the length of the largest cell of Model B. The grid of Model C was not set up to coincide with grid-cell boundaries in Model A, because Section C is not parallel to either of the main axes of the grid of Model A .

Layer 1 of Model B and Layer 1 of Model C were not intended to represent the same layer of sandstone. They are separate, hypothetically continuous sandstone layers. In Model C, Layer 1 was located 100 ft. higher than in Model B (Figures 5.7 and 5.8). Their true continuity or lack of it is reflected in their transmissivities estimated using the models. The dip of the aquifer in Section C (Figure 5.8) was not simulated because it does not affect the flow (Figure 5.10).

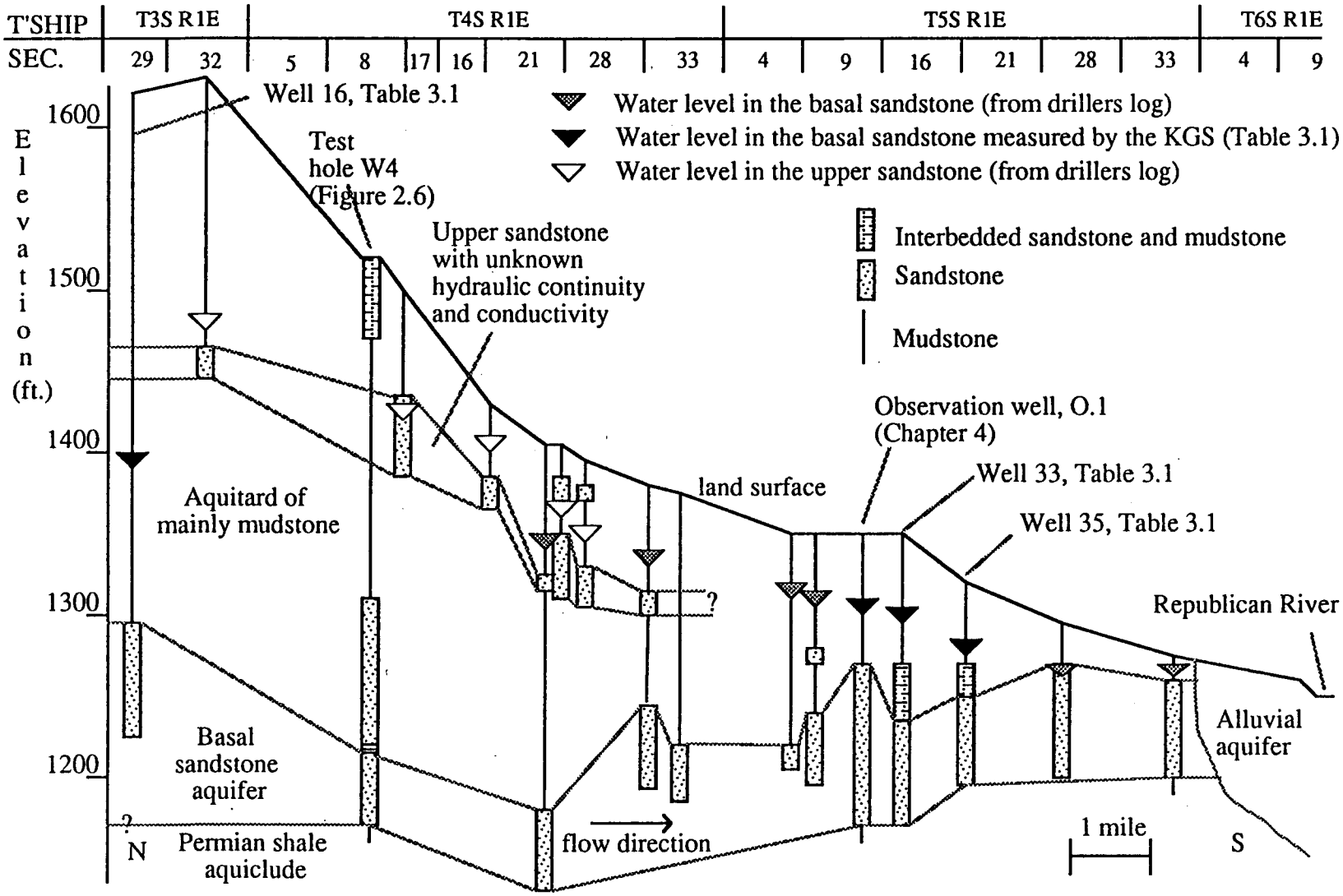


Figure 5.7. Section B, a N-S cross section along a flow line in the basal sandstone. Line of section is shown in Figure 5.2.

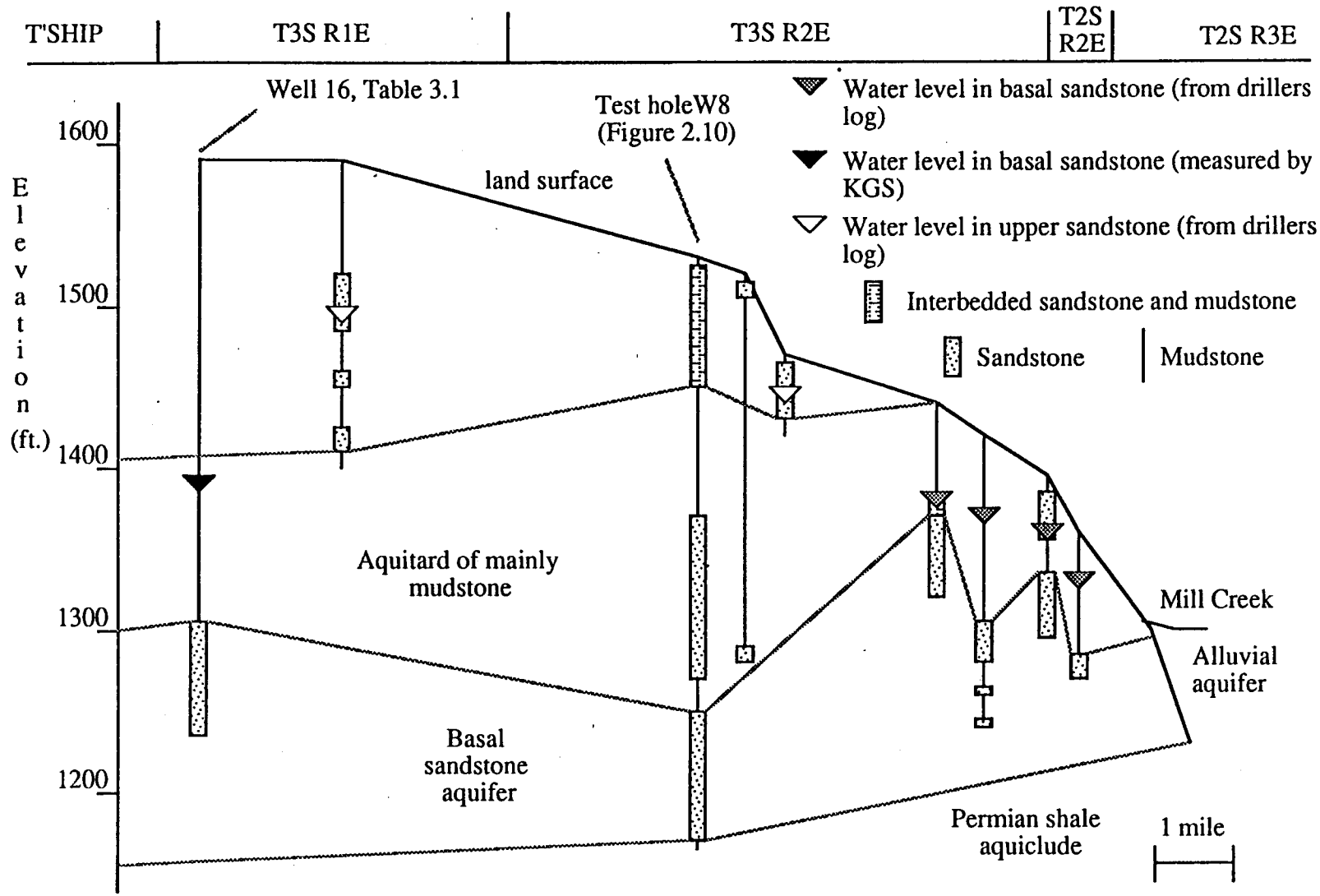


Figure 5.8. Section C. A W-E cross-section along a flow line in the basal sandstone.

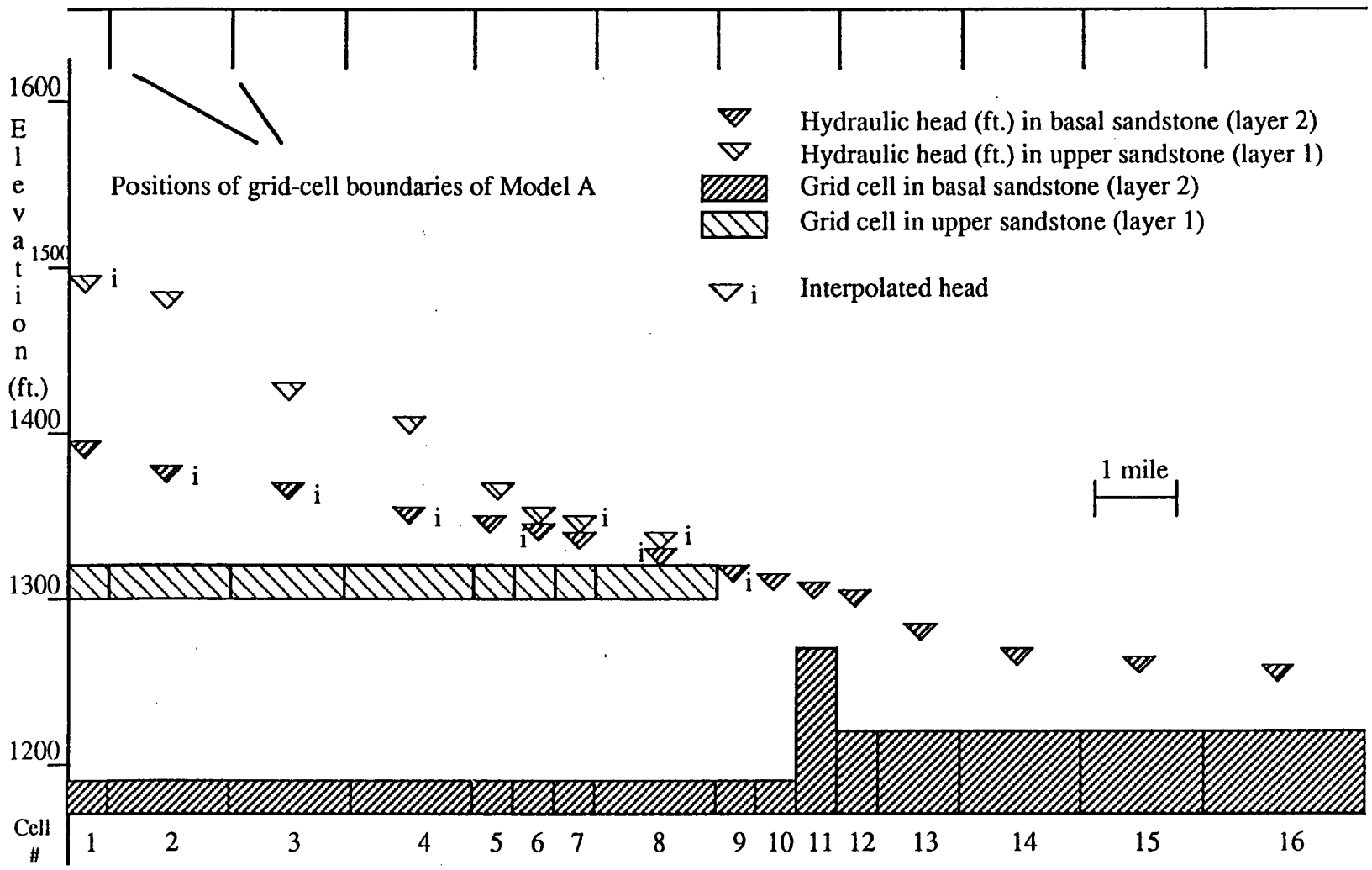


Figure 5.9. Grid used in Model B.

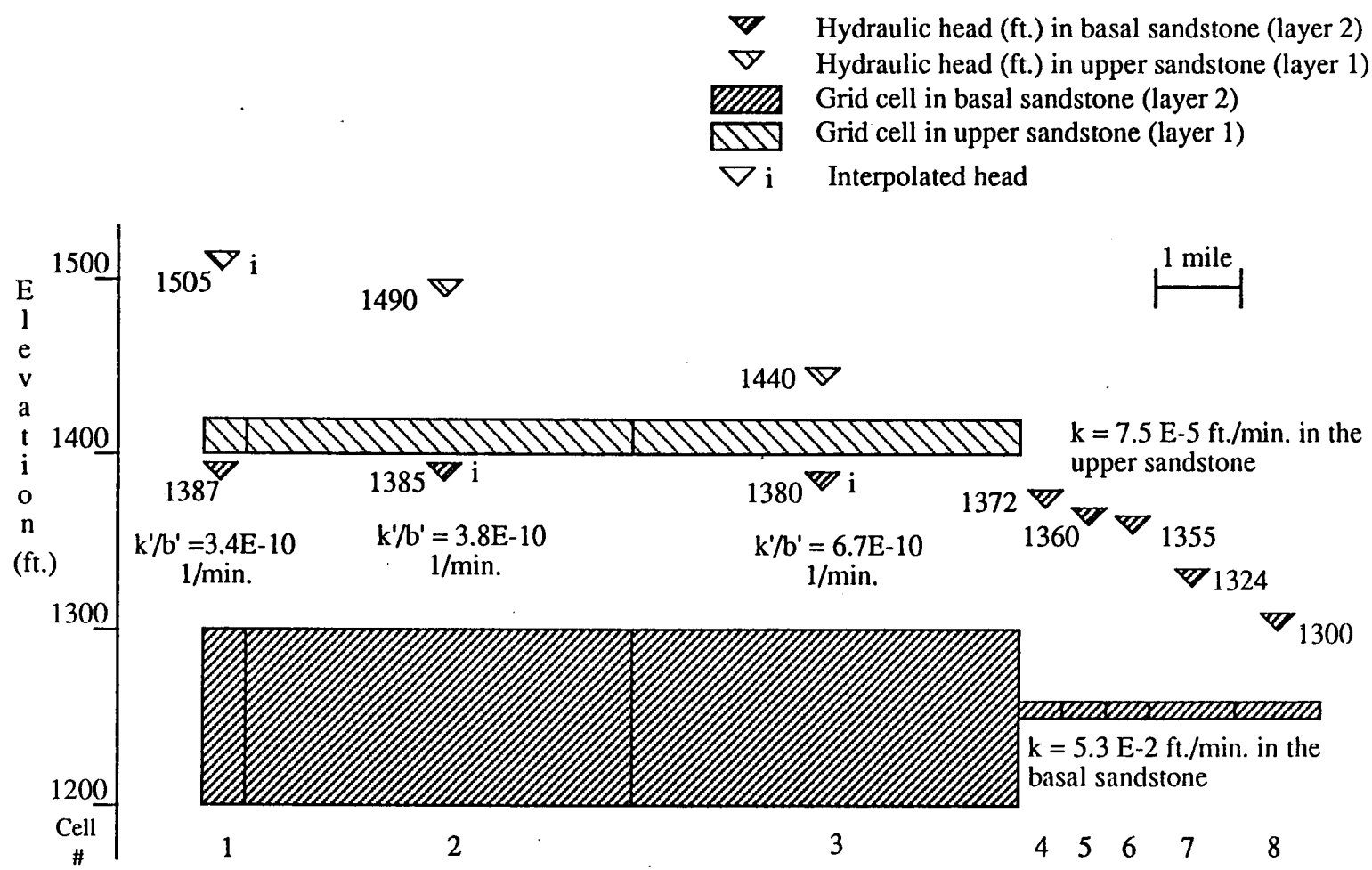


Figure 5.10. Grid used in Model C.

Table 5.10. Parameter values of the grid cells of the upper sandstone (layer 1) in Model B.

CELL #[a]	CELL TYPE[b]	CELL LENGTH (ft.)	T[c], ($ft.^2$/ day)	HYDRAULIC HEAD (ft.)	LOCATION OF HEAD MEASUREMENT	SOURCE AND DATE OF HEAD DATA	COMPUTED HEAD (ft.)	RES.[d]
1	1	5600	2.6	1485	interpolated		1479	5.7
2	1	8400	2.6	1475	T3S R1E 28DCC	DRILLER'S LOG, 1982	1476	-.91
3	1	8400	2.6	1420	T4S R1E 8CCD	DRILLER'S LOG, 1978	1426	-6.3
4	1	8400	2.6	1401	T4S R1E 20BAB	DRILLER'S LOG, 1979	1404	-2.8
5	1	2640	2.6	1360	T4S R1E 20DDC	DRILLER'S LOG, 1982	1362	-1.8
6	1	2640	2.6	1345	T4S R1E 29ADA	DRILLER'S LOG, 1983	1345	.41
7	1	2640	2.6	1340	interpolated		1336	3.8
8	2	7920	2.6	1330	interpolated		1325	4.6

a. Grid cells are numbered in increasing order from north to south.

b. Cell type 1 contains a variable head node, and cell type 2 contains a variable head node discharging to a constant head node outside the model. The constant head outside the boundary was set at 1320 ft., 1 mile south of the node at the center of cell 8.

c. Transmissivity (T) of layer 1 was not permitted to vary between cells.

d. The Residual (Res.) is the difference between the hydraulic head and the head computed by the model.

Table 5.11. Parameter values of the grid cells used to simulate the basal sandstone (layer 2) in Model B.

CELL #[a]	CELL TYPE[b]	GRID CELL LENGTH (ft.)	T[d], ($ft.^2$/day)	HYDR-AULIC HEAD (ft.)	LOCATION OF HEAD DATA	SOURCE AND DATE OF HEAD DATA	COMP UTED HEAD (ft.)	RES.[e]
1	-1	5600	1500	1387	T3S R1E 16BAAB	KGS MEASUREMENT, 1990	1387	0
2	1	8400	1500	1370	interpolated		1379	-8.9
3	1	8400	1500	1358	interpolated		1367	-8.7
4	1	8400	1500	1345	interpolated		1352	-6.8
5	1	2640	1500	1338	T4S R1E 14DCD	DRILLER'S LOG, 1979	1340	-2.3
6	1	2640	1500	1335	interpolated		1335	.5
7	1	2640	1500	1332	T4S R1E 27DDA	DRILLER'S LOG, 1988	1328	3.5
8	1	7920	1500	1320	interpolated		1316	4.2
9	1	2640	1500	1310	T4S R1E 35DDD	DRILLER'S LOG, 1979	1302	8.4
10	1	2640	1500	1305	T5S R1E 1BCB	DRILLER'S LOG, 1979	1294	11
11	1	2640	7600	1300	T5S R1E 10DCB	KGS MEASUREMENT, 1990	1290	10
12	1	2640	3800	1296	T5S R1E 16AAA	KGS MEASUREMENT, 1990	1287	8.6
13	1	5280	3800	1276	T5S R1E 17DDC	KGS MEASUREMENT, 1990	1283	-6.7
14	1	7920	3800	1262	T5S R1E 29AAD	DRILLER'S LOG, 1979	1274	-12.5
15	1	7920	3800	1258	T5S R1E 34ADA	DRILLER'S LOG, 1979	1264	-5.7
16	-1	10560	3800	1250	T6S R1E 9ACB	TOPO. MAP, 1965[c]	1250	0

a. Grid cells are numbered in increasing order from north to south.

b. Cell type 1 contains a variable head node; cell type -1 contains a constant head node.

c. The hydraulic head used in cell 16 is the elevation of the Republican River taken from a 1:24000 USGS quad sheet.

d. Transmissivities (T) are set as known quantities taken from the transmissivities used in the areal model.

e. The Residual (Res.) is the difference between the hydraulic head and the head computed by the model.

Table 5.12. Parameter values of the grid cells of both layers in Model C.

LAYER #	CELL #[a]	CELL TYPE[b]	CELL LENGTH (ft.)	T[e] (ft.^{2}/day)	V-CONT[c] (day^{-1})	HYDRAULIC HEAD (ft.)	LOCATION OF HEAD MEASUREMENT	SOURCE AND DATE OF HEAD DATA	COMPUTED HEAD (ft.)	RES.[f] (ft.)
1	1	-1	2640	17	4.90E-7	1505	interpolated		1505	0
1	2	1	23760	17	5.47E-7	1490	T3S R1E 14BCB	DRILLER'S LOG, 1978	1491	-1.4
1	3	2	23760	17	9.65E-7	1440	T3S R2E 15BAB	DRILLER'S LOG, 1977	1440	-.27
2	1	-1	2640	7600		1387	T3S R1E 16BAA	KGS , 1990	1387	0
2	2	1	23760	7600		1385	interpolated		1386	-1.2
2	3	1	23760	7600		1380	interpolated		1381	-.52
2	4	1	2640	760		1372	T3S R2E 2CDD	DRILLER'S LOG, 1988	1370	2.0
2	5	1	2640	760		1360	T3S R2E 1CCB	DRILLER'S LOG, 1969	1358	1.6
2	6	1	2640	760		1355	T2S R2E 35DAA	DRILLER'S LOG, 1980	1346	8.7
2	7	1	5280	760		1324	T2S R2E 36BBB	DRILLER'S LOG, 1988	1327	-3.4
2	8	-1	5280	760		1300	T2S R3E 31BD	TOPO. MAP, 1966[d]	1300	0

a. Grid cells are numbered in increasing order from west to east.
b. Cell type 1 contains a variable head node, -1 contains a constant head node, and cell type 2 contains a variable head node discharging to a constant head node outside the model. The constant head outside the boundary was set at 1425 ft., 2.5 mi. away from the node at the center of cell 3.
c. VCONT is a term describing the hydraulic connection of the aquitard between layers 1 and 2. It is defined as VCONT = k'/b' where k' is the vertical hydraulic conductivity of the aquitard and b' is its thickness.
d. The hydraulic head used at this location is the elevation of Mill Creek taken from a 1:24000 USGS quad sheet.
e. Transmissivity (T) was treated as an unknown in this model. The recharge rate used to estimate T was 0.25 in./yr.
f. The Residual (Res.) is the difference between the hydraulic head and the head computed by the model.

5.2.4.2. Boundaries

The nodes of the cells at each end of Layer 2 were set as constant heads in both cross-sectional models. Cell 16 of Section B (Figure 5.9, Table 5.11) simulates discharge to the Republican River alluvial aquifer; cell 8 of Section C (Figure 5.10, Table 5.12) simulates discharge to the Mill Creek alluvial aquifer. The head values used in these cells are the elevations of the respective streams at these locations, taken from 1:24000 topographic maps. In each model, cell 1 of Layer 2 is a constant head boundary so that the model simulates flow entering the layer horizontally as well as recharge penetrating from above. The head gradient from the northwest which causes horizontal inflow into both section lines can be seen in the potentiometric surface of Figure 5.2.

Horizontal flow out of Layer 1 in both sections was simulated using the head-dependent boundary module of MODFLOW (Tables 5.10 and 5.12). By setting a hypothetical constant head at a point beyond the southern limit of Layer 1 and extending the hydraulic gradient and conductance of the layer to that point, the boundary is head-dependent and recharge can be applied to it (McDonald and Harbaugh, 1988).

Section B begins near a surface-water divide between the Mill Creek and Republican River watersheds. There is unlikely to be a significant flow gradient in the direction of Section B into the northern end of Layer 1 due to the influence of topography. Therefore, grid cell 1 in Layer 1 of Model B has a variable head node with all flow into the cell coming from above (Table 5.10).

In the direction of Section C, the land surface continues to rise beyond the western end of the section and so there is likely to be a horizontal hydraulic gradient toward cell 1 of Layer 1. This grid cell was given a constant head node to simulate horizontal inflow from the west (Table 5.12). The head value of this cell was interpolated using the hydraulic gradient between the other two nodes in this layer. Therefore setting a constant head node in this cell is very similar to using the "head-dependent boundary" module in which a constant head node would be interpolated outside the model area. The only difference is that the cell is within the model so recharge into it is not simulated. However, this cell is small and has little effect on the total recharge rate.

5.2.4.3. Transmissivity

The transmissivity of Layer 1 was not known for either section although it was estimated to be several orders of magnitude lower than the transmissivity of Layer 2 due to the poor interconnection of the sandstones. The models were used to estimate this transmissivity in each section. To simplify the simulations, Layer 1 was assumed to be homogeneous within each section. In Model B, transmissivities used in the grid cells of Layer 2 were taken from those used in Model A along this line of section (Table 5.10). In Model C, transmissivities were varied to fit the computed heads to the known and interpolated heads but the recharge rate was held constant.

5.2.4.4. Hydraulic Heads

Tables 5.10, 5.11 and 5.12 show the head data which the computed data of the models were intended to duplicate. Most of the sources of these heads do not lie directly upon the lines of section but are within 3 mi. of them. Hydraulic heads were transposed into the section lines by moving them parallel to the equipotential lines. Hydraulic heads are shown transposed into the sections in Figures 5.7 and 5.8. They are accurate to ± 10 ft.

Hydraulic heads were not available for all the cells in both layers of the section; they were therefore interpolated for those cells with no data (Figures 5.9, 5.10). In Layer 1, the direction of flow of ground water, although generally in the direction of the sections, is likely to deviate from this line in places due to the influence of local relief. Some transposed heads in Layer 1 may therefore deviate from the true head at that point on the line. Of all the head data used in the two models, four heads in Layer 2 of Section B and one in Layer 2 of Section C were measured by the KGS. Head data in the basal sandstone along Section B are therefore likely to be the most reliable.

5.2.4.5. Geologic Data

Geologic data from drillers' logs were also transposed into the section lines; these were vertically adjusted in Section B to compensate for an average dip of the aquifer of 10 ft./mi. to the west. For example, geologic boundaries from a well at T4S R1E sec. 20BAB, 1 mi. to the west of Section B, were raised by 10 ft. when transposed into the section line. However, despite these adjustments, the geologic

boundaries illustrated in Figure 5.7 are not intended to accurately portray the geology of the aquifer along the line of section. Because of heterogeneity and discontinuity of sandstones in the Dakota Formation, the geology can change radically over several thousand feet, the distance over which much of the data was transposed. The geologic logs of Figures 5.7 and 5.8 are used to show the positions of the intervals from which head measurements were taken. The elevations of boundaries between sandstones and mudstones from drillers logs were not used to estimate the transmissivities of grid cells in either cross-sectional model.

5.2.4.6. Vertical Flow Through the Aquitard

In each section, the rate of vertical flow through the mudstone aquitard between the two layers was set to be the same for all cells. This is likely to be a simplification of the true situation. Variations in proportion of sandstone in the aquitard and vertical interconnection of sandstones as well as in thickness of and head difference across the aquitard all mean that the vertical flow rate through the aquitard is likely to vary between cells. However, this simplification is justified because the detailed geology of the aquitard is unknown.

5.2.5. Results

5.2.5.1. Recharge Rate

Using Model B, a recharge rate of 0.24 in./yr. was needed to minimize the difference between computed heads and interpolated heads (Figure 5.11). This is close to the recharge rate of 0.26 in./yr. determined using Model A. The computed heads for each grid cell and the difference (RES.) between observed or interpolated head data and synthetic head data are listed in Tables 5.10, 5.11, and 5.12.

5.2.5.2. Aquifer Parameters

The transmissivity of Layer 1 (T_1) in Section B was estimated to be 2.6 $ft.^2/day$ (Figure 5.12). In Section C, T_1 was estimated to be 15 $ft.^2/day$ (Figure 5.13). If Layer 1 of each section is 20 ft. thick, the calculated hydraulic conductivity of Layer 1 in Section B (K_1) is approximately 0.1 ft./day and K_1 in Section C is approximately 0.8 ft./day. Estimates of transmissivities in Layer 2 of

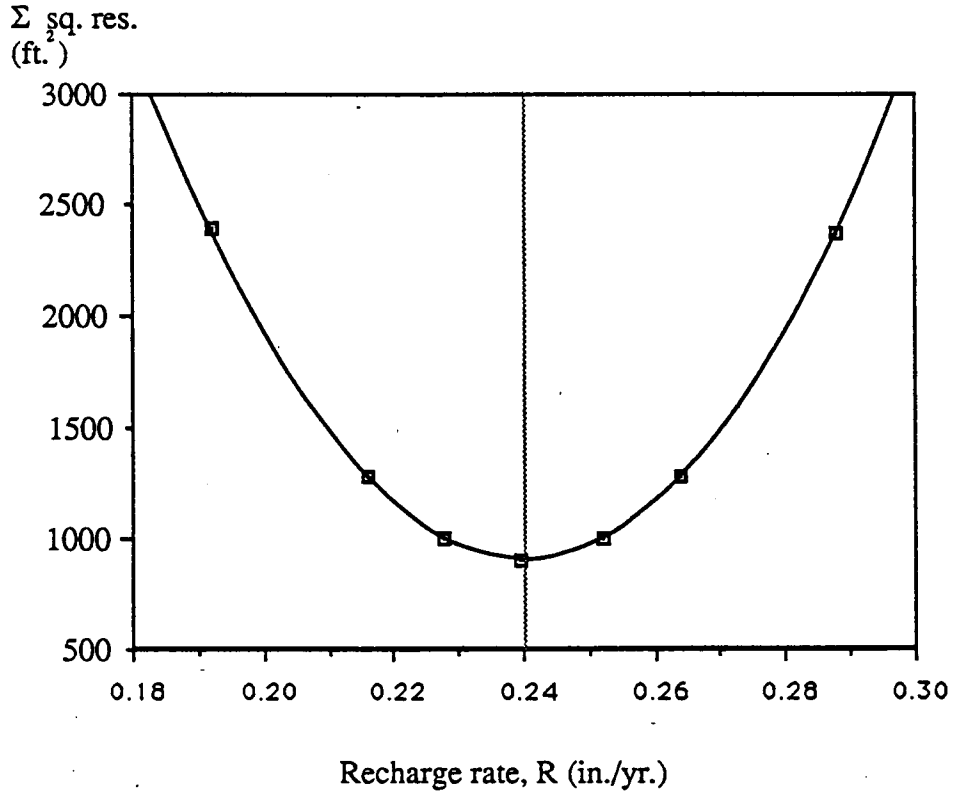


Figure 5.11. The effect of varying the recharge rate in Model B, on the sum of the squared residuals (Σ sq. res.) between interpolated heads and computed heads. Σ sq. res. is a measure of how well the computed data fit the interpolated data. The transmissivities used in Layer 2, Model B are listed in Column 4 of Table 5.11. The transmissivity of layer 1 is set at 2.6 ft./day in this sensitivity analysis. The best fit is obtained when Σ sq. res. is a minimum, i.e. when R is 0.24 ft./day. The heads computed using this

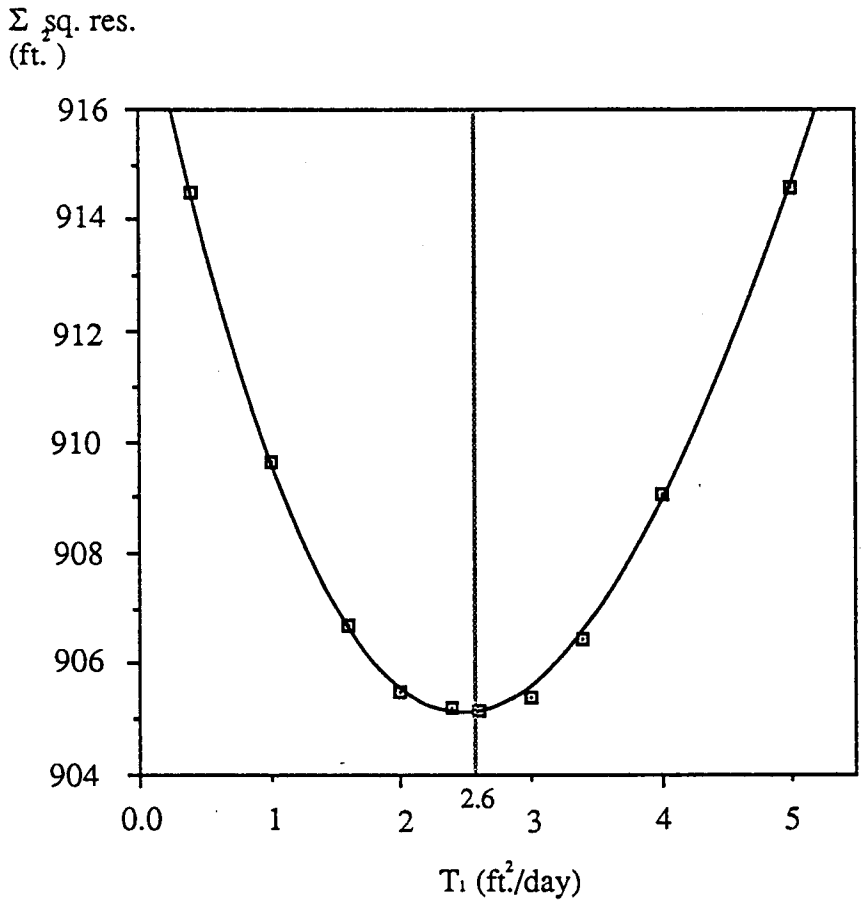


Figure 5.12. The effect of varying the transmissivity of Layer 1(T_1), Model B, on the sum of the squared residuals (Σ sq. res.) between interpolated heads and computed heads. Σ sq. res. is a measure of how well the computed data fit the interpolated data. The recharge rate is set at 0.24 in./yr. in this sensitivity analysis. The best fit is obtained when Σ sq. res. is a minimum, i.e. when T_1 is 2.6 ft.²/day. The heads computed using this transmissivity are listed in Column 8 of Tables 5.10 and 5.11.

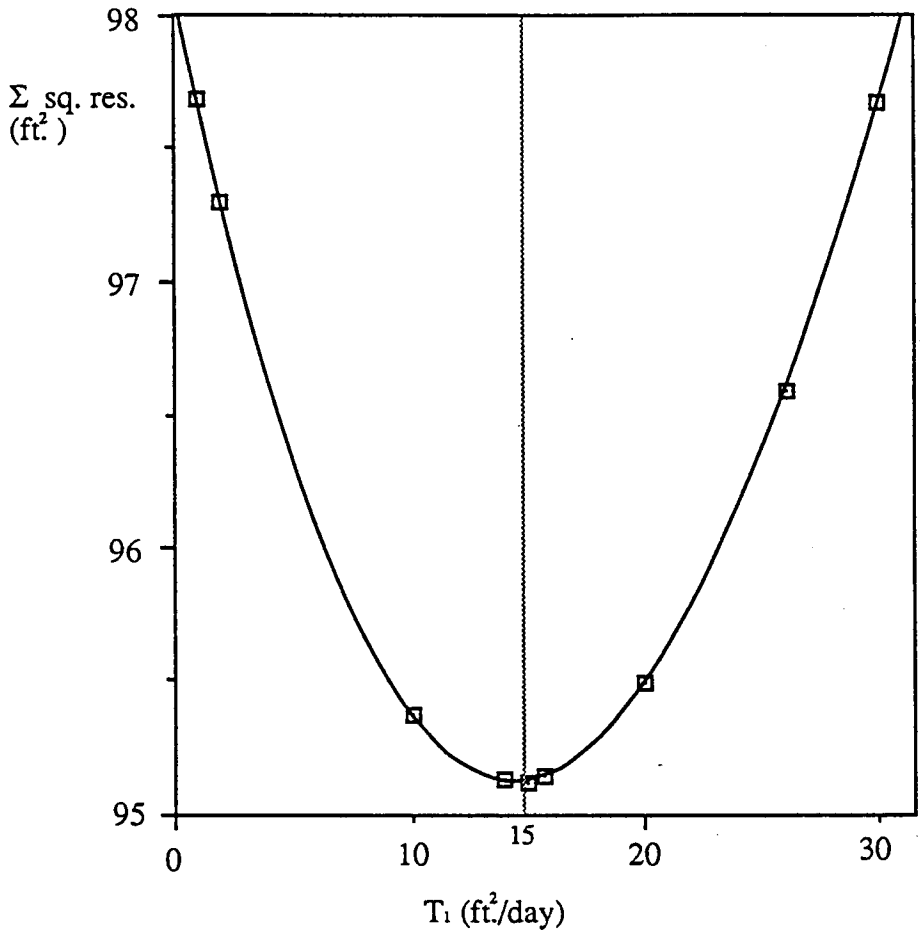


Figure 5.13. The effect of varying the transmissivity of Layer 1(T_1), Model C, on the sum of the squared residuals (Σ sq. res.) between interpolated heads and computed heads. Σ sq. res. is a measure of how well the computed data fit the interpolated data. The best fit is obtained when Σ sq. res. is a minimum, i.e. when T_1 is 15 ft.²/day. The heads computed using this transmissivity are listed in Column 10, Table 5.12.

Section C show an order-of-magnitude difference between the two ends of the section (Table 5.12).

5.2.5.3. Volumetric Flow Budgets

Volumetric flow budgets are listed below in Table 5.13.

Table 5.13. Volumetric Flow Budgets in Models B and C.

Model	Layer	Annual horizontal inflow at grid cell 1 ($ft.^3$)	Annual inflow from above ($ft.^3$)	Annual downward outflow ($ft.^3$)	Annual horizontal outflow ($ft.^3$)
B	1	0	930	930	1.0
B	2	500	1600	0	2100
C	1	7	1000	1000	8
C	2	170	1300	0	1400

Horizontal outflow from Layer 1 was less than 0.8% of the total outflow from this layer in both cross-sectional models (Table 5.13). All the remaining flow from Layer 1 leaked out of the base of the layer into the mudstone aquitard.

5.2.6. Discussion of Cross-Sectional Model Results

Neither Model B nor Model C are very sensitive to the transmissivity of Layer 1. An increase in T_1 by 25% increases Σ sq.res. by less than one in both Model B and Model C (Figures 5.12 and 5.13) whereas an increase of 25% in the recharge rate in Model B increases Σ sq.res. by over 2000 (Figure 5.11). The low sensitivity of the models to T_1 and the limited head data from this layer mean that the estimates of T_1 are not likely to be accurate to better than an order of magnitude. However, the order of magnitude of K_1 (0.1 to 0.8 ft./day) based on these transmissivities is reasonable in terms of the lithology of the layer (Freeze and Cherry, 1979, p. 29). This order of magnitude is between the hydraulic conductivities of the basal sandstone (76 ft./day) and the confining mudstone (10^{-3} ft./day) estimated from the pumping test (Chapter 4). The results of these layered

models are consistent with the fact that the aquifer is more complicated than a simple sandstone layer with a leaky mudstone confining layer.

Volumetric flow budgets (Table 5.13) show that the volume of discharge from local flow systems in sandstones in the upper part of the Dakota Formation is not significant in relation to the total volume of recharge leaking through the mudstone aquitard to the basal sandstone. In both cross-sectional models a significant proportion of the inflow into Layer 2 is horizontal inflow at the constant head node in cell 1 (Figures 5.9 and 5.10). This is consistent with the fact that there is a slope on the potentiometric surface toward the starting points of the sections (Figure 5.2). Horizontal flow through the aquifer is much more significant through the basal sandstone than through other sandstones due to its relatively high hydraulic conductivity and lateral continuity. Therefore the recharge rate estimated using Model A and Model B are very similar despite the fact that Model A simplifies the aquifer to a one-layer system.

CHAPTER 6. CONCLUSIONS

The Dakota aquifer in Republic, Washington, and northern Cloud counties is mainly composed of a fluvial assemblage of channel sandstones and overbank mudstones which are part of the Dakota Formation. Interbedded sandstones and silty mudstones of the Kiowa Formation, present in the subsurface of much of Republic, northern Cloud, and northwestern Washington counties, pinch out toward the east and do not crop out in Washington County. The sandstone with the greatest degree of lateral continuity is a fluvial channel sandstone at the base of the Dakota Formation; this is the main conduit of ground-water flow in the aquifer. The fluvial assemblage above the basal sandstone is an aquitard dominated by low-permeability mudstone but containing isolated ribbons of channel sandstone. On the scale of the study area, this interval inhibits the flow of ground water and is therefore an aquitard. The upper 25% of the Dakota Formation is composed of muddy sandstones and sandy mudstones deposited in a deltaic environment. Although the sandstones in the deltaic assemblage are likely to have greater lateral continuity than those in the underlying fluvial assemblage, the geohydrologic consequences of this greater continuity are offset by the fact that the deltaic sandstones are less permeable than the fluvial channel sandstones.

Ground water flows through a combination of regional and local systems in the aquifer. Both types of system are ultimately driven by topography, although on different scales. A regional flow system, which brings brine into the study area from the west, is driven by the regional slope of the Great Plains toward the east. The local flow systems are concentrated in the outcrop area of the aquifer, fueled directly by local recharge. The relief on the "potentiometric surfaces" in these systems is a subdued form of the local topographic relief.

Major zones of discharge from both types of flow system are in stream valleys which cut below the level of the hydraulic heads in the aquifer. Most of the discharge occurs along the contact with the alluvial aquifer in the Republican River valley and along the lower reaches of the south-flowing tributaries of the Republican River. Much of the flow in the regional flow system discharges in the watershed of one of these tributaries, named Salt Creek, in southern Republic and northern Cloud counties. East of this watershed, in southern Washington County, local flow systems predominate in the aquifer beneath south-flowing tributaries of

the Republican River. In northern Washington County, in the Mill Creek watershed, most of the discharge from the aquifer is along Mill Creek and the lower reaches of its tributaries. The aquifer contains some saline water of the regional flow system here also although it is diluted by local recharge from above.

A pumping test showed that at a site in southwest Washington County the basal sandstone has a specific storage of approximately 10^{-6} ft.$^{-1}$ and a hydraulic conductivity of 76 ft./day, whereas the mudstone layer confining the sandstone has a hydraulic conductivity of approximately 10^{-3} ft./day. The ground water in the basal sandstone at this site is recharged locally through confining layers of mudstone, as well as through limestone and shale. From a combination of areal and cross-sectional computer models, the recharge rate to the basal sandstone in a 300 sq. mi. area of southwestern Washington County was estimated to be 0.26 ± 0.09 in./yr. Although the confining layers have hydraulic conductivities several orders of magnitude lower than the basal sandstone, the areal extent of the aquifer is very large in relation to the vertical cross-sectional area of the basal sandstone. This allows the transmission of approximately 4000 acre-ft. of recharge to the basal sandstone per annum in the area simulated. Cross-sectional simulations in which an additional upper aquifer layer was considered confirmed that the volume of horizontal flow is much greater through the basal sandstone than through other sandstones. The transmissivities of these upper sandstones were estimated to be several orders of magnitude lower than the basal-sandstone transmissivity.

Water in the local flow systems is pumped from the aquifer for domestic, irrigation, stock, and municipal use. In general, water quality and availability have not been significantly affected by the pumping. Using computer simulations, it was estimated that pumping the aquifer at the current rate of ground-water withdrawal is likely to have a minimal effect on water levels and water quality in southwestern Washington County. This is compatible with the fact that no significant drops in water level have been reported in this area.

Further simulations showed that in the model area it is likely the aquifer could sustain an increase in pumping in the current areas of highest use up to double the current level without drawing in water of inferior quality. The mean drawdown estimated for this pumping rate in the model area would be between approximately 3 and 6 ft. If the increased pumping were concentrated in the current areas of highest use, drawdowns in these areas would be sufficient to significantly reduce the

maximum pumping rates of the wells and recharge of the Republican River alluvial aquifer from the Dakota aquifer would be significantly reduced. Also, drawdowns could be higher, particularly in the east-central upland part of the model area, if the inflow of water allowed for at the eastern boundary of the model area was too large. The prediction of the effects of increased use is not applicable in other parts of the study area, particularly west and north of the model area, where salinity of the ground water is already a problem.

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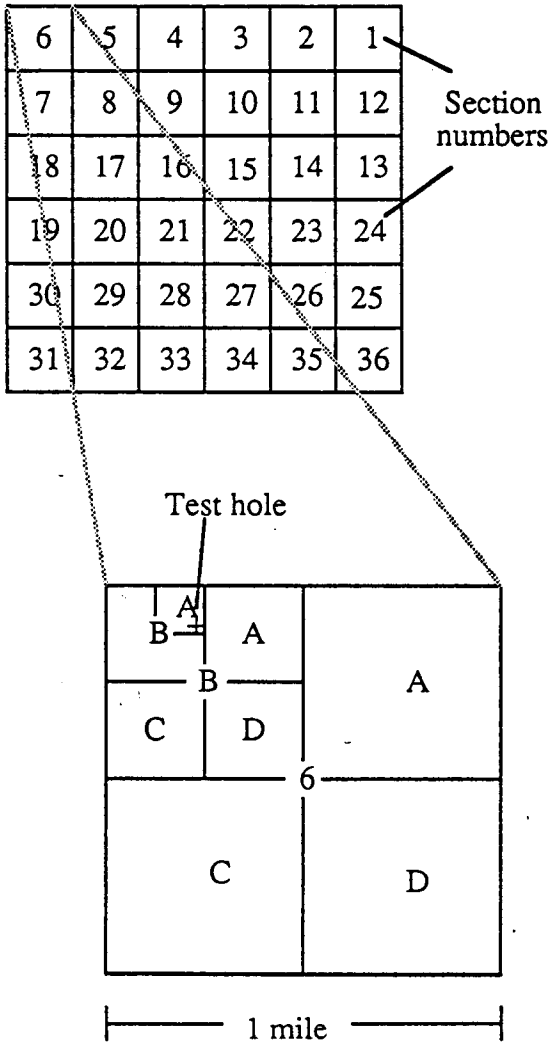
APPENDIX 1. System Used to Describe Locations in This Report

In this report, the locations of test holes, wells and other sites of interest are given by the General Land Office system of land classification. The component parts of the location are the township number, range number, section number, and one to four uppercase letters that indicatethe quarter section, quarter-quarter section, quarter-quarter-quarter section, and quarter-quarter-quarter-quarter section respectively. Sections, which are 1 sq. mi. in area, are numbered in all townships as shown below in T3S, R2W. The test hole marked with a cross is in "T3S, R2W, Sec. 6BBA", i.e. the NE1/4 of the NW1/4 of the NW1/4 of section 6, of township 3S, range 2W.

T3S, R2W



APPENDIX 2. Stratigraphy of the Core From Test Hole R1

The R1 test hole was drilled during September, 1989 in T4S, R4W, Sec. 24DD Republic County, Kansas, on land owned by Orlin Kenyon. The test hole began in the lower part of the Greenhorn Limestone (upper 58 ft.) and ended in the Permian Wellington Formation at a total depth of 537 ft. A detailed description of the core retrieved and its interpretation is given below.

58–82 ft. (Graneros Shale interval)

The dominant lithology in this interval is dark-gray shale with occasional calcareous, sandy laminations and abundant shell fragments. The sand content decreases upward and the carbonate content increases upward from the base of the interval. Hummocky cross-laminations are common in the calcareous, sandy, fossiliferous zones in the lowest 10 ft. of this facies. There are two significant bentonites; one 10 in. thick at 74 ft. and the other 3 in. thick at 60 ft. On the gamma-ray log the bentonites can be recognized from the increased levels of gamma-ray activity, and this shows up particularly well on the thorium curve produced from the spectral gamma ray detector.

This facies was deposited in the sea under generally quiet conditions below fair weather wave base but above storm wave base. Storm action is responsible for the hummocky cross-stratification. The decrease in sand content and cross-stratification and the increase in calcium carbonate content upward are due to decreasing influx of terrigenous material as the shoreline moved eastward from this site. As a result, the environment of deposition changed from shoreface to offshore marine with less restricted circulation patterns (Hattin, 1975).

82–130 ft. (uppermost Dakota Formation)

The strata in this interval consist of interlaminated dark-gray mudstone and very fine to fine grained sandstone. The organic content is low to moderate and mainly in the form of carbon flecks, although occasional small plant fragments are also present. Many of the sand laminations are lenticular. These were formed by two separate processes: migration of starved ripples and soft sediment deformation resulting from shearing along bedding planes. Evidence of migration is ripple cross-laminations within the sand lenses. Convoluted and overturned laminae within the sand lenses are evidence of deformation. Burrowing and horizontal tracks are present below 98 ft. Individual coarsening-upward sequences are

distinguishable in places, for example, in the top 9 ft. of this interval, which grades upward from a mudstone with occasional sand laminae to a calcite-cemented, rippled, fine-grained sandstone containing plant fragments. No shell fragments are found in or below this sandstone. This marks the boundary between the Graneros Shale above and the Dakota Formation below.

This sequence is a typical deposit of the distal bar of a prograding river-dominated delta (Coleman and Prior, 1988). Silt and mud settled out from suspension and variable river- or tide-related currents transported sand intermittently across the surface of the delta front forming laminations and ripples. Some soft-sediment deformation occurred as a result of loading. Marine transgression overstepping the delta front prevented the deposition of a complete coarsening-upward delta sequence above this facies.

130–158 ft.

In this interval the grain size coarsens upward from laminated siltstone through laminated, very fine grained sandstone and siltstone with abundant carbon fragments along bedding planes to very fine to fine grained sandstone lacking sedimentary structures. The top of this sequence is capped by a lignite. The laminations vary from parallel and subhorizontal to wavy and rippled and include occasional climbing ripples. Burrows are present throughout this interval except within 6 ft. of the lignite. Roots penetrate 8 ft. into the underlying sandstone from the lignite.

The lignite at the top of this interval was deposited in a swampy subaerial environment suitable for dense plant growth and with little sediment influx. The underlying sands and silts were deposited in a subaqueous environment. The sedimentary structures are typical of a distributary mouth bar sequence. Such sequences, with a capping lignite formed in a swamp or marsh environment, commonly constitute the upper part of interdistributary bay fills in river-dominated deltas (Coleman and Prior, 1988).

158–172 ft.

In this part of the core there is a fining-upward sequence grading from poorly sorted fine-grained sandstone (with mud intraclasts just above the scoured base) to poorly sorted, muddy, very fine grained sandstone and interlaminated very fine grained sandstone and silty mudstone above 167 ft. Only half the fine sandstone was recovered in the core. Abundant organic debris in the form of carbon

flecks and plant fragments is present throughout. Sedimentary structures consist of a few subhorizontal laminations throughout and occasional ripple cross-laminations in the upper 4 ft. The abundance of mica may be responsible for the general scarcity of ripples in the recovered sandstone. No large-scale bedforms are present.

The poor sorting and abundance of organic material are evidence of a relatively sudden drop in flow rate of the water transporting the sediment. The environment of deposition of this interval of the core is probably a partially abandoned distributary channel. The lower parts of such a channel are typically filled with poorly sorted sands and silts with an abundance of transported organic debris (Coleman and Prior, 1988).

172–186 ft.

Four vertically stacked, fining-upward sequences ranging from 1–6 ft., in thickness comprise the strata in this interval. The lower two sequences have a combined thickness of 2.5 ft. and consist of interlaminated very fine grained sandstone and siltstone. The upper two sequences have a combined thickness of 12 ft. and consist of light-gray massive siltstone grading up into light- and dark-gray mudstone with occasional green and red mottles.

This sequence of beds was deposited in a flood-plain environment. The laminated sandstone and siltstone in the lower 2.5 ft. was deposited on a levee by low-flow-regime currents during flood conditions.

186–310 ft.

Gray slickensided mudstone with red mottles and occasional green mottles and variable silt content comprises the bulk of this interval. Included are occasional darker gray beds and some coarsening-upward sequences of gray silty sandstone or sandy siltstone with evidence of plant colonization in the form of rootlet marks. These siltstones are not normally mottled. The finer grained zones show more red mottling than the coarser grained zones.

This facies was deposited from suspension during inundation of a flood plain by a river. The generally gray color of the sediment signifies reducing conditions, probably near or below a water table. Red mottles indicate deposition under more oxygenated conditions, possibly in topographically higher areas of the flood plain. Some mottles are in a rootlet pattern, showing that openings made by roots provided pathways for oxygen to enter the sediment.

310–324 ft.

This interval consists of orange medium-grained sandstone containing abundant wood chips that grade upward through orange and muddy gray fine-grained sandstone to very poorly sorted muddy, gray, carbon-rich, very fine grained sandstone to clean, green-mottled, very fine grained sandstone with rootlet marks. Recovery of the medium-grained sandstone during coring was poor, and its contact with underlying overbank deposits was not recovered. The only sedimentary structures are low-angle crossbedding in the orange fine-grained sandstone and faint parallel to wavy laminations in the muddy, very fine grained sandstone.

This sequence is the fill of a small channel that contained sufficient stream flow for the transport of medium-grained sandstone. The channel was then abandoned and filled with poorly sorted sediment under low-flow conditions. This was topped with levee deposits. Rootlet marks at the top reflect eventual colonization by plants. This is the highest level at which medium-grained sand was deposited. This deposit represents the last time that such high energy conditions existed during Dakota Formation deposition at this site. This was probably a function of the reduction in gradient to the sea during this period as sea level rose and the resulting reduction in competence of the streams.

324–368 ft.

This interval of the core consists mostly of gray slickensided mudstone with red mottles and occasional green mottles and variable silt content. The finer grained members of this interval contain more red mottling than the coarser grained members.

This facies was deposited from suspension during inundation of a flood plain by a river or rivers. The generally gray color of the sediment tells us that it was saturated with reduced water following deposition. Red mottles show that oxygenated water penetrated the sediment in places. Some mottles are in a rootlet pattern, showing that openings made by roots provided pathways for oxygen to enter the sediment.

368–410 ft.

Core recovery in this interval was generally poor. In the lower part of this interval medium-grained sandstone is present (a calcite-cemented sandstone occurs near the scoured base but otherwise the rock is crumbly) for approximately 24 ft. grading up into poorly indurated, fine-grained sandstone for approximately 12 ft.

This grades up through interlaminated dark-gray silty mudstone and sandstone to dark-gray organic siltstone to structureless dark-gray organic mudstone with root marks at the top. The medium-grained sand of this interval contains planar rather than trough crossbeds, which show a preferred direction of dip. Higher up, convoluted, horizontal, wavy, and small-scale trough cross-laminations are common.

This interval is a classic fining-upward channel fill. High-flow-regime conditions existed during deposition of the crossbedded medium-grained sandstone. The decreases in grain size and in scale of bedforms upward reflect a reduction in flow rate associated with shoaling, as the main axis of the flow gradually moved away from the point of deposition. The scoured base of this river channel sequence marks the base of the Dakota Formation, which is an erosional surface cut into the marginal marine sediments of the Kiowa Formation below.

410–430 ft.

The strata in this interval consist of finely laminated, gray, very fine grained, micaceous sandstone and siltstone containing little clay. Occasional burrows and plant fragments are present. Flecks of carbon are common and are concentrated in thin discontinuous laminae that are distributed throughout. The laminations are generally parallel apart from some wavy and lenticular laminations in a 2-ft.-thick coarsening-upward zone close to the top of this interval. The lack of ripple laminations may be due to the high mica content, which has an inhibiting effect on ripple formation.

Laminations are inclined at approximately 20° above 424 ft. Laminations between 424 ft. and 426 ft. are convoluted and slumped. The consistent dip of the laminations above this zone is probably a result of rotational slumping, which is often found in delta-front environments. The laminations were probably originally subhorizontal.

The limited burrowing, general lack of mud, and high carbon and mica content are evidence of a marginal marine environment consistently under the influence of low-energy currents. The plant debris is a consequence of the sediment being directly supplied from local distributaries. The environment of deposition of this facies is probably the upper distal bar of a delta.

430–457 ft.

This interval consists of two interbedded facies: (1) a gray, laminated, sometimes deformed or burrowed, muddy, very fine grained sandstone similar to that described in the above interval and (2) a ripple laminated, micaceous, fine- to medium- grained sandstone that is commonly cemented with iron and also contains some calcite cement in small quantities. Beds of facies 2 include occasional rip-up clasts near the base. Beds of facies 1 are 0.5–5 ft. thick are interbedded with up to 1.5 ft. of facies 2. The thickness of facies 2 beds generally increases with decreasing grain size from fine-medium to fine grained, whereas the thickness of beds of facies 1 generally decreases upward. This suggests that the processes responsible for deposition of 2 became more common and longer lasting but weaker with time. There is evidence of more than one direction of ripple migration in the form of crossing of adjacent sets of ripple laminae but no evidence of symmetric ripples or form-discordant or chevron internal structure; that is, there is no evidence that wave action was responsible for deposition of this facies.

This sequence is a typical deposit of the upper distal bar of a prograding river dominated delta (Coleman and Prior, 1988). Very fine grained sand, silt, and mud settled out from suspension and variable river-flood or tide-related currents transported sand intermittently across the surface of the delta-front depositing facies 2. Some soft sediment deformation occurred because of loading. A rise in relative sea level or a shift in the axis of deposition of the local distributary terminated deposition of facies 2 at 430 ft.

457–500 ft.

This interval consists mainly of medium-grained, crossbedded, and ripple laminated gray sandstone that is generally poorly cemented but contains patchy red iron oxide and calcite cements. There is an overall fining upward of the sandstone above 487 ft. from medium to fine grained. However, below 487 ft., the sandstone coarsens upward from the scoured base with abundant gray siltstone intraclasts and plant fragments in the fine-grained sandstone immediately above the scour. This basal intraclast-rich sandstone is predominantly red. It is interbedded with gray, micaceous, laminated and convoluted siltstone and very fine grained sandstone up to 487 ft. The almost continuous crossbedded sandstone of the upper 30 ft. of this interval is interrupted only by occasional mud drapes and carbonaceous lenses up to 1 in. thick.

This sequence is generally coarse grained, with an erosive base, and contains major bedforms, which is typical of the fill of an active channel. However, it has features that are not typical of a fluvial channel; in the lowest 13 ft. it coarsens upward and includes beds of relatively fine-grained, laminated silt and sand reflecting periods of lower energy. Large-scale bedforms do not become prevalent until higher in the sequence, where they are occasionally draped with mud deposited during intervals of low energy. This considerable variation in current intensity is good evidence that this interval is a tidal rather than a fluvial channel deposit.

500–507 ft.

This interval consists of interlaminated, very fine grained sandstone and mudstone that is severely deformed in the upper 2 ft. Occasional ripple and lenticular laminations and carbon flecks are present and pyrite nodules are fairly common. This interval was deposited in a subaqueous environment with intermittent currents strong enough to form ripples in very fine grained sand, such as in a tide-influenced distal bar or estuarine environment.

507–512 ft.

This part of the core consists of a hard, gray, slightly calcareous, faintly laminated mudstone with occasional shell fragments overlain by 2 ft. of brecciated, laminated, silty, very fine grained sandstone with rootlet marks at the top. The top of the Permian Wellington Formation is at 510 ft. The hard shelly shale below the top of the Permian was deposited in a marine environment. The rooted sandstone is associated with a period of subaerial deposition following erosion of the Permian sediments. Brecciation of the sandstone may have been caused by dissolution of evaporite deposits, which are common in the Wellington Formation.

APPENDIX 3. Stratigraphy of the Core From Test Hole W2

The W2 test hole was drilled during November, 1989 in T1S, R2E, Sec.10BCB, Washington County, Kansas, on land owned by Frank Gaydusek. The test hole began in the lower part of the Greenhorn Limestone (upper 62 ft.) and ended in the Permian Wellington Formation at a total depth of 534 ft. A detailed description of the core retrieved and its interpretation is given below.

62–90 ft. (Graneros Shale interval)

This interval consists of predominantly fossiliferous gray shale containing a moderate abundance of bivalves. Included in this interval is 8 ft. of interlaminated calcareous sandstone and slightly calcareous shale near the middle of the Graneros Shale with wavy and hummocky cross-laminations. Above this sandy zone the shale is slightly calcareous; below it the shale is not calcareous. There are also several beds of bentonite; in particular, two beds one at 62 ft. and the other at 81 ft. can be recognized from the increased levels of activity on the gamma-ray log. The base of this interval is marked by a hard, calcite-cemented siltstone containing cone-in-cone structures.

This sequence was deposited in a generally quiet shallow sea below fair weather wave base but above storm wave base. The sandy, hummocky cross-laminated interval was deposited by storm action. The decrease in sand content and cross stratification and the increase in calcium carbonate content upward are due to decreasing influx of terrigenous material as the shoreline moved eastward from this site. As a result, the environment of deposition changed from shoreface to offshore marine with less restricted circulation patterns (Hattin, 1975).

90–100 ft. (Upper Dakota Formation)

This section of the core consists of interlaminated sandstone and mudstone with increasing sand content and calcite cement upward from the base. Near the top of this interval wood and shell fragments are both present.

This facies was deposited in the distal bar portion of a delta front. The combination of shell and wood fragments reflects the marginal marine setting of the depositional environment, with detritus derived from both marine and terrestrial sources.

100–112 ft.

This interval consists of fine-grained horizontal and ripple-laminated micaceous, glauconitic sandstone. It includes mud clasts near the scoured base. Mud drapes up to 0.1 ft. thick and carbon flecks occur throughout along with heavily pyritized wood fragments near the top.

The small-scale sedimentary structures, grain size, and organic content are typical of a distributary mouth bar (Coleman and Prior, 1988).

112–124 ft.

This part of the core consists primarily of dark-gray mudstone with occasional thin, parallel, wavy and lenticular very fine grained sand and silt laminations with a 2-ft.-thick bed of very fine grained sandstone that was not well recovered from the core. Occasional horizontal sand-filled burrows are present in this interval.

This interval is interpreted to have been deposited in the distal bar portion of a river dominated delta front. The mudstone was deposited from suspension with occasional current action, either tide or river related, being responsible for the sandy laminations and lenses.

124–157 ft.

The predominant lithology in this interval consists of a coarsening-upward sequence of horizontal and wavy laminated, very fine grained micaceous, generally poorly sorted sandstone interbedded with siltstone. Individual coarsening-upward sequences rich in plant fragments and several feet thick can be distinguished below 140 ft. Vertical and horizontal burrows are present below 150 ft where the sequence lies conformably on top of the swamp deposits described earlier. Core recovery was poor in this interval, particularly above 137 ft.

This interval is interpreted as an interdistributary bay fill sequence consisting of splays from the local distributary. Bay fill deposits are characterized by an overall coarsening-upward sequence incorporating graded beds of laminated sand, silty sand, and silt, with burrowing more common lower in the sequence (Coleman and Prior, 1988). Subsidence or a rise in sea level resulted in submergence of the swamp deposits, creating space for sediment to accumulate beneath water level. Sea-level rise relative to the delta was sufficient to prevent the bay from filling up completely, in which case the sequence would have been capped with a lignite.

157–208 ft.

This interval is dominated by the red- (and minor green-) mottled gray mudstone with root marks. However, this grades up, above 186 ft., into structureless light-gray siltstone containing plant fragments, carbon flecks, and in situ roots. The organic content increases and the grain size decreases as this mudstone grades up into a waxy, black, organic mudstone at 157 ft. and is interbedded with rooted light-gray silty sandstone.

This part of the core was deposited on a flood plain under conditions that changed from oxidizing to reducing throughout the interval. Near the top of this section the sediments were being deposited in a marshy flood basin under reducing conditions which promoted the preservation of plant material and in which the black organic mudstone was deposited. Deposition of this mudstone in a marsh was interrupted by a splay of silty, very fine grained sandstone introduced by a flood event from a nearby river.

208–218 ft.

This part of the core consists of laminated, very fine grained sandstone and siltstone with ripple marks, scour and fill structures, slumping, sandstone dikes, inclined laminations, and general soft sediment deformation. This facies is interpreted as a levee deposit capping a channel fill sequence.

218–235 ft.

In this section the core consists of poorly indurated, trough crossbedded fine grained sandstone with occasional mud drapes grading up through rippled, fine-grained sandstone into laminated, sandy siltstone. This section is interpreted as a channel fill sequence. Upward changes in sedimentary structures and grain size reflect the change from high to low flow regime conditions.

235–272 ft.

The predominant lithology in this section consists of poorly indurated sandstone that fines upward from medium-grained sand to silty, very fine grained sand. Carbon flakes and fragments are common throughout this interval. Core recovery was poor in this section. However, large-scale crossbedding is visible in the few pieces of medium-grained sandstone recovered from the lower half of this interval. Parallel, ripple, and wavy laminations and soft sediment deformation are common in the fine-grained, very fine grained and silty sandstone near the top of this interval.

This fining-upward sequence is a classic river channel fill with high flow regimes reflected in the lower channel fill sediments (relatively coarse grain size and large-scale crossbedding) and lower flow conditions dominating higher up (very fine grain size with small-scale bedforms).

272–340 ft.

This interval is dominated by gray mudstone with red mottles and no sedimentary structures but also includes some laminated, very fine grained sandstone. Many of the mottles in the mudstone are root and rootlet shaped, that is, branching vertical tubes and fibers, which is a remnant of the plant bioturbation that removed all sedimentary structures from this facies.

This facies is interpreted as an overbank flood-plain sediment. It accumulated only during flooding of a nearby river and was thoroughly reworked by plant roots in between these flooding events. Beds of finely laminated, very fine grained sandstone within the mottled mudstone were deposited closer to the sediment source (i.e., the river) in a levee environment.

340–349 ft.

The strata in this section consist of light-gray poorly indurated, very fine to fine-grained sandstone with some wavy laminations. Core recovery was poor in this interval. The strata in this interval are interpreted to have been deposited in a levee adjacent to a river channel.

349–378 ft.

There is an overall coarsening upward of the strata in this interval. The lower 10 ft. consists of clean medium-gray, slickensided mudstone with occasional silty lenses and carbon flecks. Poorly sorted, gray, micaceous, sandy siltstone with abundant plant fragments and carbon flecks are present above. No sedimentary structures were observed in this interval.

This interval is interpreted as the fill of an abandoned river channel. The lack of red mottles and very fine grain size of the mudstone is indicative of a sheltered, reducing environment that may have been below the water table. The upward increase in grain size is due to closer proximity to a sediment source, such as an active river channel. The poor sorting of this coarser sediment shows it was dumped in a relatively quiet environment as a result of a sudden significant drop in current velocities, which is consistent with an abandoned channel.

378–400 ft.

The lithology in this interval is dominated by very poorly indurated, crossbedded, medium-grained sandstone containing rip-up clasts of siltstone and occasional plant fragments and carbon flecks. Embedded within the sandstone are large blocks up to 4 ft. thick of relatively well-lithified distal bar sediments consisting mainly of silty mudstone with thin, very fine grained sand laminations but also including a bed of ripple laminated fine-grained sandstone with abundant intraclasts (see description above of the interval 400–431 ft.).

The laminations within the distal bar sediments are inclined at an angle of 50°–60°. They could not have been deposited on a stable surface at such an angle. This high angle shows that the blocks of distal bar sediments suffered rotational slumping after lithification. Thus the base of the medium-grained sandstone is an unconformity representing a significant period of erosion. The crossbedding and relatively coarse grain size is evidence of a high flow regime typical of an active river channel. This interval was therefore deposited in a river channel cut into the underlying lithified sediments, large blocks of which collapsed into the channel and were preserved. The disconformity marks the boundary between the Dakota and Kiowa Formations.

The uppermost foot of this interval consists of conglomerate beds up to 0.5 ft. thick interbedded with clean gray mudstone beds 0.1–0.2 ft. thick. The conglomerate contains clasts of white, gray, and red mudstone and siltstone up to 0.5 in across. This reflects large variations in current intensity resulting from partial abandonment of the river channel. The intraclasts are probably reworked overbank sediments from a nearby cutbank.

400–431 ft. (Longford Member, Kiowa Formation)

Two main facies are interbedded with each other: (1) a yellow, micaceous, rippled, fine-grained sandstone facies in beds from 0.1 ft. –3 ft. thick and (2) a laminated, very fine grained sandstone and silty mudstone that makes up most of the interval. The rippled sandstone facies contains occasional to abundant plant fragments up to at least 0.2 ft. in length and carbon flecks. Ripples are unidirectional current ripples. This facies commonly contains small intraclasts (medium to very coarse sand size) of gray silty mudstone. Most beds of this facies have sharp tops and bases, but some of the beds grade down into silty mudstone. The interlaminated silty mudstone and very fine grained sandstone facies contain

laminations of sandstone that are subhorizontal, thin, wispy, and sometimes lenticular. There is little evidence of bioturbation except for some horizontal traces at 415 ft.

This interval is interpreted as having been deposited in the distal bar portion of a river-dominated delta front. The finer grained facies was mainly deposited from suspension with occasional current action, probably tidal or river related, being responsible for the sandy laminations and lenses. The fine-grained sandstone beds were deposited by longer lasting and greater currents. The moderately high organic content and small-scale crossbedding is good evidence of current action associated with floods in a nearby distributary touching bottom, each bed of fine sandstone representing an individual flood which even brought in the yellow sand and organic material and locally ripped up clasts of silty mudstone (Coleman and Prior, 1988).

431–443 ft.

In this interval beds of conglomerate 0.1–3 ft. thick are interbedded with laminated gray siltstone and silty mudstone and underlain by a poorly indurated, sugary, medium-grained sandstone, most of which was not recovered in core. The conglomerate consists of angular to subangular clasts of white, gray, and pale red siltstone or silty mudstone and gray fine-grained sandstone and plant fragments supported in a matrix of fine to medium sand with carbon flecks. There are no graded beds, and there is no evidence of imbrication. Boundaries between the beds are sharp with no gradation. The gray siltstone contains occasional thin laminations of very fine grained sand that are lenticular in places. The interlaminated siltstone and sandstone is draped directly over the uneven tops of the intraclastic conglomerates.

The poor sorting and lack of structures in the conglomerate beds are indicative of plastic flow, that is a debris flow. These debris flows may have been triggered by fracturing and subsidence of the underlying Permian sediments as a result of evaporite dissolution during transgression of the early Cretaceous sea. The gray siltstone was deposited mainly from suspension in between debris flow events. It represents the generally quiet water conditions of a lower shoreface or distal bar environment when sedimentation rates were much slower than during the debris flows. Laminations and lenses of sandstone are evidence of current action, which was probably tidal.

443–534 ft. (Wellington Formation)

This part of the core consists of finely interlaminated, silty, very fine grained calcite-cemented sand and silty mud. The laminations are generally flat and parallel, although some wavy laminations are present, indicating current or wave processes. This facies is well indurated but heavily brecciated. The core contains pieces of broken rock from fractions of an inch to several inches across. Broken laminations are common. This brecciated facies overlies a hard bluish-gray, burrowed silty shale. The contact is an irregular surface inclined at 70° to the horizontal, that is a fault plane.

These sediments are interpreted as shallow marine deposits of the Permian Wellington Formation. Much of the Wellington Formation typically consists of evaporites, although they were not observed in this core. The brecciation may be due to dissolution of evaporites by ground water that penetrated the Permian surface. Debris flow deposits within the overlying Cretaceous sediments may be related to this subsidence in the Permian deposits.

APPENDIX 4. Water-Level Data Which Were not Measured by the KGS

The locations of these data are shown in Figure 3.2. Water-use codes are D: domestic, P: public supply, I: irrigation, S: stock, T: test hole, O: observation well.

Location (Tnshp., Rng., Sec.)			Farmer/user	Water use	Land surface elev. (ft.)	Total well depth (ft.)	Depth to water (ft.)	Water level elev. (ft.)	Date of meas.
1	1E	1DCCC	GAYDUSEK, FRANK	D	1590	?	187	1403	09/09/80
1	1E	7AAAC	CITY OF MAHASKA	P	1601	285	195	1406	11/01/78
1	1E	26AAA	ARBITHOUT, Rodney	D	1500	240	157	1343	09/20/79
1	2E	14AAC	FRAGER, WILLIAM	D	1510	160	60	1450	11/18/77
1	2E	15BCD	HUBER, CHARLES	D	1516	120	100	1416	02/28/84
1	2E	23BBB	JOEBEL, EARL	D	1520	114	40	1480	05/14/76
1	2E	25BAB	FRAGER, GARY	D	1470	104	30	1440	09/10/12
1	3E	20CCC	MILLER, JERRY	D	1440	160	20	1420	04/14/76
1	3E	17DDA	LINDSLEY, LANNY	D	1480	188	160?	1320?	03/30/84
1	4E	4BA	O.WC. FARMS	D	1350	102	68	1282	05/21/86
1	4E	31CCB	STEWART, ADELBERT	D	1315	55	30	1285	10/21/76
1	5E	5DDD	SCHRAMM, MELVIN	I	1380	75	20	1360	05/06/79
1	5E	5ADA	SCHRAMM, MELVIN	I	1400	76	46	1354	08/03/89
1	5E	19DAD	GREFE, FRED	D	1330	68	40	1290	08/27/76
1	1W	7CAB	COONROD, DEAN	D	1540	160	71	1469	09/25/80
1	1W	15CCA	CITY OF NARKA	P	1580	281	169	1411	11/30/77
1	1W	28AA			1550	?	147	1403	07/14/42
1	2W	3BCC	COONROD, WESLEY	D	1562	100	68	1494	07/14/82
1	2W	16ABB	PACK, PAUL	D	1530	160	120	1410	04/02/76
1	2W	21AA			1550	?	165	1385	00/00/42
2	1E	4BCB	KUCZYNSKI, LES	D	1450	95	53	1397	10/21/87
2	1E	11ADC	COOK, KEITH	D	1440	173	100	1340	10/16/76
2	1E	14CCC	ZINGER, WARREN	D	1395	40	15	1380	12/11/80
2	1E	15DDD	ZINGER, WENDELL	D	1400	40	20	1380	02/21/81
2	1E	17ADA	ZINGER, REX	D	1440	?	40	1400	03/06/81
2	1E	21BABB	CITY OF HADDAM	P	1445	74	51	1394	09/23/87
2	1E	21DDD	WELCH, EDWIN	D	1430	89	56	1374	11/26/85
2	1E	21BACB	CITY OF HADDAM	P	1470	101	70	1400	09/22/87
2	1E	21BABB	CITY OF HADDAM	P	1448	87	52	1396	02/26/88
2	2E	11AAC	KOLMAN, JERRY	D	1420	114	80	1340	04/16/81
2	2E	12CCD	ROSENTHAL	S	1435	80	45	1390	05/10/79
2	2E	16AAA	GROVER, CECIL	D	1400	130	55	1345	09/11/79
2	2E	26BDD	N-CENT. Hugh USD#221	L	1340	78	28	1322	04/16/81
2	2E	27DDD	BAKER, PAUL	D	1350	42	18	1332	08/25/80
2	2E	32CDA	MORRE, LEO	D	1420	180	40	1380	08/09/90
2	2E	33CDC	LECUYER, BOB	D	1390	140	40	1350	10/14/80
2	2E	34DCC	LUTJEMEIER, EVERETT	D	1420	120	63	1357	12/19/79
2	2E	35DAA	SHERR, TERRY	D	1395	99	49	1355	09/11/80
2	2E	36BBB	SAWYER, JUNIOR	D	1360	110	36	1324	05/22/88
2	3E	1BCC	HENDERSON, J.L.	D	1330	60	25	1305	07/13/84
2	3E	6BDA	GORDON, Austin & Mable	D	1430	182	115	1315	10/31/78
2	3E	8AD	HENKE, HENRY	D	1425	140	97	1328	11/05/80
2	3E	9CDD	APLEY, KENNETH	D	1390	100	23	1357	02/10/77
2	3E	11ADB	IMHOFF, HENRY	D	1293	48	10	1283	03/18/76
2	3E	11CDC	HOLBROOK, DAVID	D	1325	70	38?	1287?	07/31/80
2	3E	12BBB	NEW, CLYDE	D	1340	80	40	1300	03/13/79

APPENDIX 4 (cont.)

Water-use codes are D: domestic, P: public supply, I: irrigation, S: stock, T: test hole, O: observation well.

Location (Tnshp., Rng., Sec.)			Farmer/user	Water use	Land surface elev. (ft.)	Total well depth (ft.)	Depth to water (ft.)	Water level elev. (ft.)	Date of meas.
2	3E	18DDC	DILLER, LAWRENCE	I	1375	154	30	1345	01/16/75
2	3E	18D	DILLER, LAWRENCE	T	1400	?	?	?	01/16/75
2	3E	19B	DILLER, LAWRENCE	I	1345	122	40	1305	11/23/77
2	3E	19A	DILLER, LAWRENCE	I	1390	162	35	1355	02/04/75
2	3E	20A	DILLER, LAWRENCE	I	1415	191	85	1330	10/20/75
2	3E	21DDBD	STIGGE, JOHN	I	1364	145	49	1315	08/03/89
2	3E	21D	STIGGE, JOHN	I	1372	147	43	1329	08/03/89
2	3E	21CDC	LONG, LEROY	D	1380	148	80	1300	03/28/79
2	3E	22DCB	SHUM, MARVIN	I	1370	130	50	1320	08/20/80
2	3E	22D	SHUM, MARVIN	I	1400	132	53	1347	08/20/80
2	3E	24DDC	LONG, LONZO	D	1350	80	40	1310	08/09/80
2	3E	26CDC	GRAHAM, LARRY	D	1400	114	80	1320	06/09/80
2	3E	27DAA	GRAHAM, LARRY	D	1410	85	50	1360	09/30/78
2	3E	27CBA	STIGGE, JOHN, & SONS	I	1320	75	49	1271	08/03/89
2	3E	28CBD	WILLBRANDT, Raymond	I	1360	95	20	1340	05/03/78
2	3E	30CDC	STIGGE, MARVIN	D	1320	85	45	1320	07/07/86
2	3E	32AAC	PIFER, LEE	D	1310	89	21	1289	07/08/76
2	3E	32CCC	STIGGE, MARVIN	DS	1350	122	50	1300	07/31/87
2	3E	32DCB	STIGGE, MARVIN	D	1380	120	75	1305	08/23/79
2	3E	34BBC	Wash. Cnty Rural WD#3	P	1362	118	63	1299	11/17/82
2	3E	34BBB	Wash. Cnty Rural WD#3	P	1358	114	56	1302	05/12/82
2	3E	34ADA	KEESECKER, WILLIS	D	1385	101	60	1325	12/03/75
2	3E	36C	WILGERS, CALVIN	I	1475	230	61	1414	06/01/77
2	1W	16ACA	SHARA, RICHARD	D	1525	38	10	1515	06/18/76
2	1W	18BCCA	PETERS, EUGENE	D	1532	178	100	1432	02/26/75
2	1W	19CCB	FILINGER, MRS. Frank	D	1560	181	135	1425	02/13/76
2	1W	30BD			1580	?	45	1535	07/11/42
2	3W	27CC			1538	?	125	1413	07/15/42
3	1E	3DDD	BAKER, TODD	D	1500	254	137	1363	03/18/88
3	1E	3CCC	NAT GAS PIPELINE		1513	?	?	?	08/02/89
3	1E	9ABB	KELLER, WILLIAM H.	D	1611	181	150	1461	03/19/76
3	1E	9DCC	RYSER, LEE	D	1590	370	160	1430	08/26/83
3	1E	9ABB	KELLER, WILLIAM H.	D	1600	180	60?	1540?	11/19/80
3	1E	11BBA	ALLEN, ROBERT	D	1600	130	90	1510	10/25/82
3	1E	14DCC	ALLEN, BOB	D	1555	100	30	1525	09/02/80
3	1E	14BCB	ALLEN, DEAN	D	1590	183	100	1490	05/25/78
3	1E	16DDD	ROKES, RALPH	D	1613	237	110?	1503?	04/09/82
3	1E	16BAAB	Nat. Gas Pipe. Co. USA	D	1590	340	165	1425	06/17/83
3	1E	28DDC	TAYLOR, Frank	D	1550	?	90	1460	11/08/82
3	2E	1DDC	MEYER, Harold	D	1380	101	40	1340	11/20/87
3	2E	1CCB	DURST, FRANK H.	D	1420	179	69	1369	97/06/76
3	2E	12BBA	SLATER, WILLIAM	D	1390	92	45	1345	10/22/87
3	2E	15BAB	ELDER, LESLIE	D	1470	49	30	1440	11/21/77
3	2E	16ADC	ELDER, FLOYD	D	1530	240	120	1410	09/26/78
3	2E	21BBB	WALLACE, DAVE	D	1530	200	160	1370	11/19/85
3	2E	22CBC	DITMER, LAVERYL	S	1470	155	115	1355	12/03/82

APPENDIX 4 (cont.)

Water-use codes are D: domestic, P: public supply, I: irrigation, S: stock, T: test hole, O: observation well.

Location (Tnshp., Rng., Sec.)			Farmer/user	Water use	Land surface elev. (ft.)	Total well depth (ft.)	Depth to water (ft.)	Water level elev. (ft.)	Date of meas.
3	2E	26CBB	LOHRENGEL, LARRY	D	1495	220	70	1425	11/10/76
3	2E	28DCC	HERRS, VERLIN	D	1500	160	82	1418	07/01/80
3	2E	33BAA	ELLIOT, WAYNE	D	1500	144	80	1420	01/15/75
3	2E	35DCD	COLGROVE, ELMER	D	1470	180	100	1370	02/16/77
3	2E	36C	WILGERS, CALVIN	I	1483	230	61	1422	06/01/77
3	3E	1DDB	ROLLMAN, BOB	D	1280	40	12	1269	10/29/75
3	3E	2CCA	STIGGE, JOHN	D	1300	56	25	1275	09/12/79
3	3E	4DDA	WALTHERS, Charles	D	1330	60	28	1302	06/03/78
3	3E	4DCC	Bekemeyer Enterprises	D	1350	80	30	1320	02/26/81
3	3E	5DDD	BEKEMEYER, Lester	D	1385	100	60	1325	07/14/80
3	3E	5DDD	BEKEMEYER, Lester	D	1385	110	30	1355	09/26/75
3	3E	9CBB	SCHOEN, KENNETH	D	1330	80	18	1312	12/01/77
3	3E	9ADA	TEGETHOF, Donald	D	1320	61	45	1275	07/26/79
3	3E	10DCAB	STIGGE, JOHN	S	1330	48	20	1310	07/30/80
3	3E	10DAB	STIGGE, JOHN	D	1315	40	20	1295	10/29/80
3	3E	10ADD	STIGGE, JOHN	D	1330	50	16	1314	05/01/78
3	3E	15BAA	STIGGE & SONS, INC.	S	1350	123	76	1274	07/24/79
3	3E	15CCD	MORSE, DENNIS	D	1405	120	55	1350	04/13/76
3	3E	15CCD	DAQUE, PAUL	D	1405	139	80	1325	12/19/77
3	3E	17ABA	Bekemeyer Enterprises	S	1370	90	50	1320	02/17/77
3	3E	23DBBB	MORMON SPRING		1312	?	-1	1311	
3	3E	24CCC			1345	?	-17	1352	66
3	3E	28AAD	MONTAGUE, Kenneth	D	1410	119	75	1335	01/28/83
3	3E	29BBC	KERN, BOB	D	1445	150	60	1385	05/19/83
3	3E	30BBB	BEIKMAN, DALLAS	D	1430	120	75	1355	10/29/80
3	3E	33AAA	HIESTERMAN, Ernest	D	1390	65	32	1358	08/30/78
3	3E	34AAA	HYDROC. Transport. Inc.	D	1410	100	60	1350	11/04/82
3	4E	2DCC	BRABEC, JEFF	D	1340	59	29	1320	07/12/78
3	4E	4ADD	STAMM, ERNEST	D	1410	56	40	1380	07/12/78
3	1W	5ADD	HENKE, GLENN	D	1490	102	58	1432	05/01/79
3	1W	5DCC	BRUNS, RAYMOND	D	1540	160	60	1480	04/29/80
3	1W	9AAC	KOPSA, Dennis & Julia	D	1560	154	120	1440	03/31/76
3	1W	34DD			1450	?	34	1416	07/20/42
3	1W	35DCC	KALIVODA, DON	D	1580	200	162	1418	09/09/77
3	2W	5BDC	BLACKBURN, Sylvia	D	1530	160	90	1440	04/08/77
3	2W	6ACB	SNAPP, PHIL	S	1500	?	59	1441	04/19/79
3	2W	13CDC	HUNCOVSKY, Dale	D	1530	179	100	1430	11/01/82
3	2W	23CCC	THOMPSON, William	D	1460	133	38	1422	06/10/81
3	3W	24CB			1500	?	120	1380	07/09/42
3	3W	32DDD	JENSBY, ROY L.	D	1500	108	45	1455	09/07/79
3	4W	13DCD	DOOLEY, Lawrence	D	1528	142	70	1458	08/14/80
4	1E	4DAD	Peterson, Col. Donald	D	1530	60	40	1490	11/29/82
4	1E	7DAD	GRANT, WILLARD	D	1500	140	80	1420	07/13/78
4	1E	8CCD	BREEDING, WARREN	D	1450	75	30	1420	07/20/78
4	1E	12CCA	WURTZ, HERMAN	D	1410	?	28	1382	03/31/79

APPENDIX 4 (cont.)

Water-use codes are D: domestic, P: public supply, I: irrigation, S: stock, T: test hole, O: observation well.

Location (Tnshp., Rng., Sec.)			Farmer/user	Water use	Land surface elev. (ft.)	Total well depth (ft.)	Depth to water (ft.)	Water level elev. (ft.)	Date of meas.
4	1E	14DCD	VAN CAMPEN, LARRY	S	1440	268	102	1338	06/18/79
4	1E	20DDC	Brantford Covenant Church	D	1400	98	40	1360	05/01/82
4	1E	20BAB	LUNDQUIST, ROLAND	D	1430	65	19	1411	04/13/79
4	1E	25DCC	PETERSON, DON	D	1375	94	25	1350	12/17/79
4	1E	27DDA	RUSCO, DELBERT	D	1412	200	80	1332	03/15/88
4	1E	29ADA	PEARSON, LEROY	D	1400	100	55	1345	05/04/83
4	1E	33ADD	GRANERE, HAROLD	D	1380	186	150	1230	03/07/83
4	1E	35DDD	WINTER, DENNIS	D	1380	150	70	1310	12/02/87
4	2E	3BAC	VANWINKLE, MARVIN	D	1460	160	70	1390	08/25/83
4	2E	3CDC	BISPING, NEIL	D	1453	200	140	1313	09/06/77
4	2E	8CAC	WILKENS, FRANK G.	D	1410	229	70	1340	08/13/80
4	2E	14DAA	HERRS, HARVEY	D	1425	40	10	1415	01/27/77
4	2E	14CCC	HELMS, ERNA		1485	?	43	1442	06/07/88
4	2E	14DAA	HERRS, HARVEY	D	1425	120	20	1405	09/23/82
4	2E	15ADA	BISPING, DON	D	1490	180	95	1395	12/10/78
4	2E	17BCBA	BEIKMAN, Mrs. Marie	D	1400	?	58	1342	03/30/77
4	2E	19BBA	WURTZ, HERMAN	D	1380	107	55	1325	03/17/76
4	2E	19DDB	MOORMAN, MARVIN	D	1360	90	22	1338	07/24/76
4	2E	22BCC	HERRS, LEROY	D	1440	159	60	1380	02/08/78
4	2E	23DD	HERRS, HARVEY	D	1455	139	70	1385	07/07/76
4	2E	24DBC	HERRS, HARVEY	I	1420	142	51	1369	12/15/77
4	2E	24DAB	HERRS, HARVEY	I	1390	103	46	1344	12/15/77
4	2E	25BCC	HARTMAN, Walter A.	D	1420	120	52	1368	05/30/80
4	2E	26DCD	HERRS, HARVEY	D	1390	60	20	1370	01/26/82
4	2E	27ADD	HELMS, ARNOLD	D	1460	195	60	1400	07/14/82
4	2E	28AAB	HELMS, A.& L., Farms	D	1420	160	120	1300	02/16/77
4	2E	29AAA	DETTMER, ART	D	1390	74	50	1340	08/28/78
4	2E	31CDD	BEIKMAN, ALBERT	D	1310	60	24	1286	11/03/82
4	2E	35BBA	MEYERS, DELBERT	D	1440	140	60	1380	12/26/78
4	3E	1CCC	HOLSCH, RALPH	D	1420	63	25	1395	07/07/79
4	3E	4CBC	MEYER, DELBERT	D	1455	140	80	1375	03/28/78
4	3E	10CDD	KOLLE, ELMER	D	1440	100	42	1398	10/30/81
4	3E	10BBA	DUITSMAN, BRAD	D	1440	92	62	1378	05/13/88
4	3E	10ABB	DUITSMAN, Kenneth	D	1435	100	35	1400	03/14/77
4	3E	10CAA	KOLLE, ELMER	D	1425	100	30	1395	12/29/81
4	3E	11DCC	VOELKER, LEO	D	1427	90	35	1392	09/16/75
4	3E	11DAA	VOELKER, LEO	D	1425	40	27	1398	08/24/84
4	3E	12BCC	STUNKEL, MELVIN	D	1415	89	14	1401	10/11/77
4	3E	15AAA	KOLLE, RAYMOND	D	1445	80	60	1385	10/12/83
4	3E	16DDC	SCHEELE, LARRY	D	1440	60	20	1420	04/21/87
4	3E	16ADC	SCHEELE, WILBERT	D	1445	90	40	1405	10/08/76
4	3E	18CDD	RIPPE, CLARENCE	D	1410	80	55	1355	07/24/87
4	3E	20CAD	WASH. CNTY RWD#2	P	1420	122	56	1364	10/08/85
4	3E	20CDD	WASH. CNTY RWD#2	P	1400	115	55	1345	10/14/85
4	3E	20CCC	SCHEELE, RAY	D	1380	60	30	1350	02/12/80
4	3E	20CAA	WASH. CNTY RWD#2	P	1400	116	56	1344	10/28/85

APPENDIX 4 (cont.)

Water-use codes are D: domestic, P: public supply, I: irrigation, S: stock, T: test hole, O: observation well.

Location (Tnshp., Rng., Sec.)			Farmer/user	Water use	Land surface elev. (ft.)	Total well depth (ft.)	Depth to water (ft.)	Water level elev. (ft.)	Date of meas.
4	3E	21CCD	HERRS, HARVEY	D	1410	91	30	1380	09/13/77
4	3E	21CCD	HERRS, HARVEY	I	1410	78	30	1380	02/12/77
4	3E	24BBB	OHLDE, BOB	D	1440	100	58	1382	08/03/82
4	3E	24B	OHLDE, BOB & Norma	I	1420	113	40	1380	09/06/75
4	3E	25BCC	SINGULAR, DON	D	1440	100	53	1387	08/03/82
4	3E	29BAA	CITY OF LINN	P	1425	124	50	1375	05/11/77
4	3E	29ACC	CITY OF LINN	P	1400	110	50	1350	05/08/77
4	3E	33ADC	DITMER, LAVERN	D	1430	90	34	1396	07/14/76
4	4E	9DAA	HOOVER, RICHARD	D	1390	122	50	1340	11/02/78
4	1W	8CCD	ANDERSON, STEVE	D	1410	175	70	1340	09/10/81
4	1W	12CCC	OSTLUND, ART	D	1480	120	60	1420	12/18/79
4	1W	15DDC	PACKARD, BILL	D	1400	100	50	1350	09/23/78
4	1W	16ACCA	CITY OF AGENDA	P	1412	140	82	1330	10/25/83
4	1W	17CBA	KIEFFER, JEROME	D	1410	80	40	1370	09/29/75
4	1W	18DCC	ANDERSON, JIM	D	1420	180	63	1357	11/05/80
4	1W	21DDC	THOMPSON, Gerald	D	1380	70	45	1335	04/07/84
4	1W	23DBAA	KOCH, JIM		1340	?	37	1303	08/01/89
4	1W	24CB			1420	?	82	1338	07/16/42
4	1W	32AAA	LEGERE, JOHN	D	1375	160	30	1345	05/06/81
4	1W	33BAD	KIEFFER, HERMAN	D	1340	150	20	1320	11/18/80
4	1W	34BAA	KIEFFER, HERMAN	D	1358	130	20	1338	11/13/84
4	2W	9DC			1400	?	38	1362	07/13/42
4	2W	12DDD	BENYSHEK, TERRY	S	1450	112	40	1410	11/17/79
4	2W	21ABB	BORCHARDT, R.D.	D	1350	55	-1	1351	06/14/79
4	2W	28BAD	BORCHARDT, Retta O.	D	1350	?	-2	1352	04/23/85
4	2W	36DAA			1400	?	101	1299	07/01/66
4	3W	11DA	TROST, ELDON	D	1440	100	65	1375	01/28/76
4	3W	13BCC	MCGINNIS, GLEN	D	1430	100	40	1390	09/13/77
4	3W	14CC			1440	?	61	1379	08/31/42
4	3W	22DDA	SHERWOOD, Kenneth	D	1420	90	50	1370	07/25/78
4	4W	5BAA	BARCLAY, PETE	D	1415	42	10	1405	05/20/78
4	4W	10ABB	HERMAN, Clarence	D	1470	96	50	1420	03/25/81
4	4W	14BB			1465	?	68	1397	06/06/42
4	4W	24DAD	KENYON, ORLIN	D	1485	?	?	?	
5	1E	1BCB	WURTZ, GARY	D	1350	130	45	1305	08/10/78
5	1E	1DDB	DAMMAN, THALE	I	1340	138	42	1298	05/24/78
5	1E	10C	KIEFFER, DOUGLAS	I	1370	176	70	1300	12/28/77
5	1E	10DBC	KIEFFER, DOUGLAS	I	1373	189	72	1301	11/20/75
5	1E	10DDD	KIEFFER, DOUGLAS	D	1380	115	74	1306	10/14/83
5	1E	13CCC	MAKIN BEACON, Inc.	D	1310	98	50	1260	05/15/76
5	1E	13CCB	MAKIN BEACON, Inc.	S	1317	116	55	1262	03/20/76
5	1E	16AAD	CONDRAY, ROBERT	I	1340	178	64	1276	08/01/89
5	1E	17A	NELSON, NORMAN	I	1340	152	60	1280	06/18/80
5	1E	17DDCC	CONDRAY, ROBERT	I	1340	154	86	1254	08/01/89

APPENDIX 4 (cont.)

Water-use codes are D: domestic, P: public supply, I: irrigation, S: stock, T: test hole, O: observation well.

Location (Tnshp., Rng., Sec.)			Farmer/user	Water use	Land surface elev. (ft.)	Total well depth (ft.)	Depth to water (ft.)	Water level elev. (ft.)	Date of meas.
5	1E	17DCCD	CONDRAY, ROBERT	I	1340	163	88	1252	08/01/89
5	1E	20D	NELSON, NORMAN	I	1350	153	65	1285	06/29/76
5	1E	20ADA			1325	?	46	1279	03/09/77
5	1E	21CDD	BUNER, WILLIAM C.	D	1310	120	37	1273	09/22/78
5	1E	21AAA	CONDRAY, KENT	D	1325	134	30	1295	09/07/84
5	1E	26CCD	REED, LEO	D	1350	100	60	1290	04/30/77
5	1E	27CDC	LYON, LAURENCE	D	1340	121	80	1260	06/18/85
5	1E	29BAB	BOHM, WILBERT	D	1355	166	106	1249	07/13/76
5	1E	29AAD	VALLEY VIEW RANCH	D	1320	121	60	1260	08/16/82
5	1E	30CCC	TOBYNE, Lowell & Lois	I	1286	112	23	1263	02/13/76
5	1E	31BBA	TOBYNE, LOWELL	D	1285	136	73	1212	04/21/81
5	1E	34ADA	BENTEMAN, ARLAN	D	1320	108	60	1260	07/08/80
5	1E	34ADA	BENTEMAN, ARLAN	D	1320	107	60	1260	07/07/80
5	1E	34ADA	BENTEMAN, ARLAN	D	1320	100	30	1290	06/02/82
5	1E	35DBA	VAN BEEK, EARL	D	1330	80	36	1294	09/22/86
5	1E	35DBD	BESWICK, RALPH	D	1330	80	52	1278	10/03/77
5	1E	35DBD	TOLL, ROY	D	1330	99	55	1275	03/30/78
5	1E	35DBA	WATSON, MAURKE	D	1330	97	60	1270	03/29/77
5	1E	36DDD	JACKSON, LEE	D	1300	?	15	1285	11/30/78
5	1E	36ABB	BURK, CLARENCE	D	1270	40	?	?	10/20/78
5	1E	36CCD	CHAPUT, DENNIS	D	1280	76	50	1230	04/11/79
5	2E	1B	WILGERS, WENDELL	I	1440	135	48	1392	04/07/79
5	2E	1BCC	WILGERS, WENDELL	D	1440	153	80	1360	06/15/79
5	2E	5DDC	BARGMAN, MARYLIN	D	1350	73	32	1318	03/10/81
5	2E	9BBD	BEIKMAN, RAYMOND	D	1345	72	30	1315	07/13/76
5	2E	9AD	STANKEL, HAROLD	D	1410	60	25	1385	10/08/85
5	2E	11BBA	DAMMAN, THURL	D	1370	78	50	1320	09/07/77
5	2E	12CBB	TIEMEYER, ROLAND	D	1430	97	40	1390	04/07/77
5	2E	17CCC	LONG, FRANCIS	D	1350	96	40	1310	09/15/78
5	2E	20BADB	LONG, FRANCIS	D	1360	155	30	1330	10/13/75
5	2E	20BAD	LONG, FRANCIS	D	1370	100	45	1325	11/09/84
5	2E	24CCA	LOHMEYER, ARNOLD	D	1325	53	36	1291	03/10/77
5	3E	1CDB	WORNER, RAYMOND	D	1450	140	60	1390	07/31/78
5	3E	9CCD	OHLDE, RAYMOND	D	1415	64	33	1382	02/14/76
5	3E	16BCD	ST. JOHNS CHURCH	D	1400	129	50	1350	12/17/76
5	3E	18DDD	BEARBAUM, LEWIS	D	1380	78	12	1368	08/18/76
5	3E	20BAA	MEYER, ORVAL	D	1360	60	40	1320	04/05/79
5	3E	21BCB	BOTT, DELWYN	D	1400	92	65	1335	12/21/78
5	3E	21CDD	BOTT, HERMAN	D	1360	72	32	1328	12/27/78
5	3E	22DAD	BOTT, HERMAN	D	1390	100	75	1315	04/29/78
5	3E	24AAB	OHLDE, ERNEST	D	1400	56	30	1370	07/15/81
5	3E	24AAD	OHLDE, LEONARD	D	1420	60	15	1405	10/08/80
5	3E	25BCC	LOHMEYER, ARNOLD	D	1330	60	18	1312	09/07/77
5	3E	29AAB	BOTT, RONALD	D	1410	71	40	1370	09/02/80
5	3E	31CDC	GILBERT, WILLIAM	D	1390	61	20	1370	10/29/77
5	3E	33BCC	BESWICK, ARNOLD	D	1400	46	35	1365	09/15/77

APPENDIX 4 (cont.)

Water-use codes are D: domestic, P: public supply, I: irrigation, S: stock, T: test hole, O: observation well.

Location (Tnshp., Rng., Sec.)	Farmer/user	Water use	Land surface elev. (ft.)	Total well depth (ft.)	Depth to water (ft.)	Water level elev. (ft.)	Date of meas.
5 1W 1DA	DANIELSON, A. F.	S	1386	?	35	1351	09-31-54
5 1W 7BBA			1430	?	53	1377	08/01/54
5 1W 10DAA	LAMBERT BROTHERS	D	1340	141	60	1280	08/18/80
5 1W 12D	COX, Daryl & Barbara	I	1395	223	120	1285	07/01/77
5 1W 13ADD			1378	205	93	1284	09/01/62
5 1W 13CC	RESCO, ANNA	S	1336	?	43	1293	08-31-54
5 1W 24DCC	RESCO, MORRIS	D	1360	150	100	1260	04/09/82
5 2W 1DDD			1390	62	46	1343	08/01/68
5 2W 1ABA			1390	206	98	1291	08/01/68
5 2W 1BAC	PALMER, MRS. Mary	I	1375	?	40	1335	03/04/88
5 2W 7DD	JANSON, J. A.	S	1433	?	96	1337	08-31-54
5 2W 9ADA	PRICE, GERALD	D	1372	105	46	1326	08/29/80
5 2W 10BBA	ELKSTROM, L. B.	S	1357	?	29	1328	08-31-54
5 2W 12ABA	FEIGHT, MIKE	D	1368	80	22	1346	12/29/86
5 2W 12ABC	FEIGHT, GENE	D	1360	138	40	1320	04/17/76
5 2W 14AAA	OLSON, ED	D	1360	80	40	1320	09/28/76
5 2W 15CBC	DECKER, J. W.	S	1359	?	52	1307	08-31-54
5 2W 17BBB	MUNGER, CLAIR	D	1420	93	60	1360	08/10/75
5 2W 17CCC			1356	?	35	1321	04/14/54
5 2W 22BAA	NELSON, SCOTT	D	1325	58	30	1295	01/18/83
5 3W 3AAA	CURTIS, DON	D	1500	135	40	1460	06/21/77
5 3W 3A	MCGINNIS, GARY	D	1540	106	76	1464	08/16/76
5 3W 6CCD			1423	?	34	1389	08/01/42
5 3W 6BAA	BARLEEN, Leonard	D	1450	87	40	1410	02/27/75
5 3W 10BCB	MORGAN, W. E.	S	1435	?	63	1372	08-31-54
5 3W 11BBB	LEWIS, DON	D	1455	233	135	1320	08/26/78
5 3W 12BC	BRAY, LELAND	D	1400	100	65	1335	01/27/76
5 3W 13ABA	KEIL, WILLIS	I	1380	75	36	1344	04/13/77
5 3W 13BCD	KEIL, WILLIS	I	1390	66	36	1354	06/30/75
5 3W 13C	KEIL, WILLIS	I	1410	76	40	1370	10/15/76
5 3W 14DDD			1384	?	52	1332	08/01/54
5 3W 15DC	ERLEKSON, DAVID	D	1380	60	39	1341	11/08/77
5 3W 17AA	USGS	O	1411	?	54	1357	06-14-54

APPENDIX 5. Observed and Compensated Drawdown in the Observation Well During the Pumping Test (CHAPTER 4)

Details of corrections made to compensate for aquifer recovery, well interference, and atmospheric pressure fluctuations are documented in Wade (1991).
t = time from 14:36 hrs, August 7, 1990,
d = depth to water measured by transducer from top of casing, and recorded on the datalogger,
s = drawdown calculated directly from recorded depth to water,
s' = drawdown compensated for interference from another irrigation well,
s" = drawdown compensated for aquifer recovery and well interference,
s*= drawdown compensated for aquifer recovery, well interference, and atmospheric pressure changes, i.e. the "compensated drawdown."

t (min.)	d (ft.)	s (ft.)	s' (ft.)	s" (ft.)	s*(ft.)
0.000	77.18	0.00	0.00	0.00	0.00
0.003	77.18	0.00	0.00	0.00	0.00
0.007	77.18	0.00	0.00	0.00	0.00
0.010	77.18	0.00	0.00	0.00	0.00
0.013	77.18	0.00	0.00	0.00	0.00
0.017	77.18	0.00	0.00	0.00	0.00
0.020	77.18	0.00	0.00	0.00	0.00
0.023	77.18	0.00	0.00	0.00	0.00
0.027	77.18	0.00	0.00	0.00	0.00
0.030	77.18	0.00	0.00	0.00	0.00
0.033	77.18	0.00	0.00	0.00	0.00
0.050	77.18	0.00	0.00	0.00	0.00
0.067	77.18	0.00	0.00	0.00	0.00
0.083	77.18	0.00	0.00	0.00	0.00
0.10	77.18	0.00	0.00	0.00	0.00
0.12	77.19	0.01	0.01	0.01	0.01
0.13	77.19	0.01	0.01	0.01	0.01
0.15	77.19	0.01	0.01	0.01	0.01
0.17	77.21	0.03	0.03	0.03	0.03
0.18	77.21	0.03	0.03	0.03	0.03
0.20	77.22	0.04	0.04	0.04	0.04
0.22	77.24	0.06	0.06	0.06	0.06
0.23	77.25	0.07	0.07	0.07	0.07
0.25	77.27	0.09	0.09	0.09	0.09
0.27	77.29	0.11	0.11	0.11	0.11
0.28	77.30	0.12	0.12	0.12	0.12
0.30	77.32	0.14	0.14	0.14	0.14
0.32	77.33	0.15	0.15	0.15	0.15
0.33	77.35	0.17	0.17	0.17	0.17
0.42	77.44	0.26	0.26	0.26	0.26
0.50	77.54	0.36	0.36	0.36	0.36
0.58	77.62	0.44	0.44	0.44	0.44
0.67	77.71	0.53	0.53	0.53	0.53
0.75	77.78	0.60	0.60	0.60	0.60
0.83	77.86	0.68	0.68	0.68	0.68

0.92	77.92	0.74	0.74	0.74	0.74
1.00	78.00	0.82	0.82	0.82	0.82
1.08	78.05	0.87	0.87	0.87	0.87
1.17	78.11	0.93	0.93	0.93	0.93
1.25	78.16	0.98	0.98	0.98	0.98
1.33	78.22	1.04	1.04	1.04	1.04
1.42	78.25	1.07	1.07	1.07	1.07
1.50	78.31	1.13	1.13	1.13	1.13
1.58	78.38	1.20	1.20	1.20	1.20
1.67	78.41	1.23	1.23	1.23	1.23
1.75	78.46	1.28	1.28	1.28	1.28
1.83	78.50	1.32	1.32	1.32	1.32
1.92	78.55	1.37	1.37	1.37	1.37
2.00	78.58	1.40	1.40	1.40	1.40
2.50	78.80	1.62	1.62	1.62	1.62
3.00	78.98	1.80	1.80	1.80	1.80
3.50	79.14	1.96	1.96	1.96	1.96
4.00	79.26	2.08	2.08	2.08	2.08
4.50	79.34	2.16	2.16	2.16	2.16
5.00	79.42	2.24	2.24	2.24	2.24
5.50	79.55	2.37	2.37	2.37	2.37
6.00	79.66	2.48	2.48	2.48	2.48
6.50	79.74	2.56	2.56	2.56	2.56
7.00	79.82	2.64	2.64	2.64	2.64
7.50	79.88	2.70	2.70	2.70	2.70
8.00	79.94	2.76	2.76	2.76	2.76
8.50	80.02	2.84	2.84	2.84	2.84
9.00	80.07	2.89	2.89	2.89	2.89
9.50	80.13	2.95	2.95	2.95	2.95
10.0	80.18	3.00	3.00	3.00	3.00
12.0	80.45	3.27	3.27	3.27	3.27
14.0	80.62	3.44	3.44	3.44	3.44
16.0	80.70	3.52	3.52	3.52	3.52
18.0	80.94	3.76	3.76	3.76	3.76
20.0	81.03	3.85	3.85	3.85	3.85
22.0	81.16	3.98	3.98	3.98	3.98
24.0	81.27	4.09	4.09	4.09	4.10
26.0	81.37	4.19	4.19	4.19	4.20
28.0	81.48	4.30	4.30	4.30	4.31
30.0	81.53	4.35	4.34	4.35	4.36
32.0	81.60	4.42	4.41	4.42	4.43
34.0	81.68	4.50	4.49	4.50	4.51
36.0	81.75	4.57	4.56	4.57	4.58
38.0	81.84	4.66	4.65	4.66	4.67
40.0	81.89	4.71	4.70	4.71	4.72
42.0	81.95	4.77	4.76	4.77	4.78
44.0	82.03	4.85	4.84	4.85	4.86
46.0	82.09	4.91	4.90	4.91	4.92
48.0	82.17	4.99	4.98	4.99	5.00
50.0	82.22	5.04	5.03	5.03	5.05
52.0	82.24	5.06	5.05	5.05	5.07

54.0	82.27	5.09	5.08	5.08	5.10
56.0	82.28	5.10	5.09	5.09	5.11
58.0	82.32	5.14	5.13	5.13	5.15
60.0	82.35	5.17	5.16	5.16	5.18
62.0	82.40	5.22	5.21	5.21	5.23
64.0	82.43	5.25	5.24	5.24	5.26
66.0	82.46	5.28	5.27	5.27	5.29
68.0	82.47	5.29	5.28	5.28	5.30
70.0	82.52	5.34	5.33	5.33	5.36
72.0	82.54	5.36	5.35	5.35	5.38
74.0	82.57	5.39	5.38	5.38	5.41
76.0	82.60	5.42	5.41	5.41	5.44
78.0	82.63	5.45	5.44	5.44	5.47
80.0	82.68	5.50	5.49	5.49	5.52
82.0	82.70	5.52	5.51	5.51	5.54
84.0	82.71	5.53	5.52	5.52	5.55
86.0	82.73	5.55	5.54	5.54	5.57
88.0	82.76	5.58	5.57	5.57	5.60
90.0	82.77	5.59	5.57	5.58	5.61
92.0	82.79	5.61	5.59	5.60	5.63
94.0	82.81	5.63	5.61	5.62	5.65
96.0	82.85	5.67	5.65	5.66	5.69
98.0	82.87	5.69	5.67	5.68	5.71
100	82.89	5.71	5.69	5.70	5.73
110	83.00	5.82	5.80	5.81	5.84
120	83.08	5.90	5.88	5.89	5.93
130	83.15	5.97	5.95	5.96	6.00
140	83.22	6.04	6.02	6.03	6.07
150	83.31	6.13	6.10	6.12	6.16
160	83.39	6.21	6.18	6.19	6.25
170	83.49	6.31	6.28	6.29	6.35
180	83.50	6.32	6.29	6.30	6.36
190	83.57	6.39	6.36	6.37	6.43
200	83.61	6.43	6.40	6.41	6.47
210	83.63	6.45	6.42	6.43	6.49
220	83.66	6.48	6.44	6.46	6.52
230	83.72	6.54	6.50	6.52	6.58
240	83.79	6.61	6.57	6.59	6.65
250	83.79	6.61	6.57	6.59	6.65
260	83.79	6.61	6.57	6.59	6.66
270	83.80	6.62	6.58	6.59	6.67
280	83.82	6.64	6.59	6.61	6.69
290	83.85	6.67	6.62	6.64	6.72
300	83.88	6.70	6.65	6.67	6.75
310	83.90	6.72	6.67	6.69	6.77
320	83.93	6.75	6.70	6.72	6.80
330	83.96	6.78	6.73	6.75	6.83
340	84.15	6.97	6.92	6.94	7.02
350	84.25	7.07	7.01	7.04	7.12
360	84.29	7.11	7.05	7.08	7.16
370	84.34	7.16	7.10	7.13	7.21

380	84.39	7.21	7.15	7.18	7.26
390	84.42	7.24	7.18	7.21	7.29
400	84.44	7.26	7.20	7.22	7.31
410	84.47	7.29	7.23	7.25	7.34
420	84.50	7.32	7.25	7.28	7.37
430	84.51	7.33	7.26	7.29	7.38
440	84.53	7.35	7.28	7.31	7.40
450	84.55	7.37	7.30	7.33	7.42
460	84.58	7.40	7.33	7.36	7.45
470	84.61	7.43	7.36	7.39	7.48
480	84.63	7.45	7.38	7.41	7.50
490	84.64	7.46	7.38	7.42	7.51
500	84.67	7.49	7.41	7.45	7.54
510	84.70	7.52	7.44	7.48	7.57
520	84.70	7.52	7.44	7.48	7.57
530	84.72	7.54	7.46	7.50	7.60
540	84.75	7.57	7.49	7.53	7.63
550	84.77	7.59	7.51	7.54	7.65
560	84.77	7.59	7.51	7.54	7.65
570	84.78	7.60	7.51	7.55	7.66
580	84.80	7.62	7.53	7.57	7.68
590	84.82	7.64	7.55	7.59	7.70
600	84.86	7.68	7.59	7.63	7.74
610	84.88	7.70	7.61	7.65	7.76
620	84.89	7.71	7.62	7.66	7.77
630	84.93	7.75	7.66	7.70	7.81
640	84.93	7.75	7.65	7.70	7.81
650	84.94	7.76	7.66	7.71	7.82
660	84.94	7.76	7.66	7.71	7.82
670	84.96	7.78	7.68	7.73	7.84
680	84.96	7.78	7.68	7.73	7.85
690	84.97	7.79	7.69	7.74	7.86
700	84.97	7.79	7.69	7.74	7.86
710	84.99	7.81	7.71	7.75	7.88
720	85.01	7.83	7.72	7.77	7.90
730	85.01	7.83	7.72	7.77	7.90
740	85.01	7.83	7.72	7.77	7.90
750	85.04	7.86	7.75	7.80	7.93
760	85.04	7.86	7.75	7.80	7.93
770	85.05	7.87	7.76	7.81	7.94
780	85.07	7.89	7.78	7.83	7.96
790	85.07	7.89	7.78	7.83	7.96
800	85.08	7.90	7.79	7.84	7.97
810	85.07	7.89	7.77	7.83	7.96
820	85.07	7.89	7.77	7.83	7.96
830	85.10	7.92	7.80	7.86	7.99
840	85.10	7.92	7.80	7.86	7.99
850	85.12	7.94	7.82	7.88	8.01
860	85.12	7.94	7.82	7.88	8.01
870	85.13	7.95	7.83	7.89	8.02
880	85.13	7.95	7.83	7.89	8.02

890	85.15	7.97	7.84	7.91	8.04
900	85.16	7.98	7.85	7.91	8.05
910	85.16	7.98	7.85	7.91	8.05
920	85.18	8.00	7.87	7.93	8.07
930	85.16	7.98	7.85	7.91	8.05
940	85.18	8.00	7.87	7.93	8.07
950	85.19	8.01	7.88	7.94	8.08
960	85.19	8.01	7.88	7.94	8.08
970	85.19	8.01	7.88	7.94	8.08
980	85.21	8.03	7.89	7.96	8.10
990	85.23	8.05	7.91	7.98	8.12
1000	85.23	8.05	7.91	7.98	8.12
1100	85.18	8.00	7.85	7.93	8.08
1200	85.34	8.16	8.00	8.08	8.26
1300	85.46	8.28	8.11	8.20	8.39
1400	85.53	8.35	8.17	8.27	8.47
1500	85.57	8.39	8.20	8.30	8.52
1600	85.62	8.44	8.24	8.35	8.58
1700	85.69	8.51	8.30	8.42	8.65
1800	85.80	8.62	8.40	8.53	8.76
1900	85.89	8.71	8.49	8.61	8.83
2000	85.99	8.81	8.58	8.71	8.91
2100	80.32	3.14	2.90	3.04	3.24
2200	79.67	2.49	2.24	2.39	2.59
2300	79.44	2.26	2.00	2.16	2.36
2400	79.29	2.11	1.85	2.01	2.20
2500	79.23	2.05	1.78	1.95	2.12
2600	85.31	8.13	7.85	8.02	8.19
2700	85.88	8.70	8.42	8.59	8.75
2800	86.32	9.14	8.85	9.03	9.21
2900	86.54	9.36	9.06	9.25	9.44
3000	86.68	9.50	9.20	9.39	9.59
3100	86.79	9.61	9.30	9.50	9.72
3200	86.90	9.72	9.40	9.61	9.82
3300	87.01	9.83	9.51	9.72	9.95
3400	87.09	9.91	9.58	9.80	9.99
3500	87.12	9.94	9.60	9.83	10.02
3600	87.19	10.01	9.67	9.90	10.10
3700	87.20	10.02	9.67	9.91	10.10
3800	87.27	10.09	9.74	9.98	10.17
3900	87.33	10.15	9.79	10.05	10.22
4000	86.67	9.49	9.13	9.39	9.56
4100	86.73	9.55	9.18	9.45	9.62
4200	86.73	9.55	9.18	9.45	9.63
4300	86.79	9.61	9.23	9.51	9.69
4400	87.71	10.53	10.15	10.43	10.62
4500	87.88	10.70	10.31	10.60	10.82
4600	88.07	10.89	10.50	10.79	11.03
4700	88.25	11.07	10.67	10.97	11.20
4800	88.36	11.18	10.78	11.08	11.31
4900	88.49	11.31	10.90	11.21	11.44

5000	88.60	11.42	11.01	11.33	11.52
5100	88.44	11.26	10.84	11.17	11.36
5200	88.37	11.19	10.77	11.10	11.30
5300	88.01	10.83	10.41	10.74	10.95
5400	88.01	10.83	10.40	10.74	10.95
5500	87.77	10.59	10.16	10.50	10.70
5600	87.82	10.64	10.20	10.55	10.73
5700	87.88	10.70	10.26	10.62	10.79
5800	88.14	10.96	10.51	10.88	11.07
5900	88.26	11.08	10.63	11.00	11.22
6000	88.36	11.18	10.73	11.10	11.33
6100	88.45	11.27	10.81	11.19	11.42
6200	88.52	11.34	10.88	11.26	11.48
6300	88.60	11.42	10.95	11.35	11.55
6400	88.64	11.46	10.99	11.39	11.60
6500	88.69	11.51	11.04	11.44	11.67
6600	88.75	11.57	11.09	11.50	11.74
6700	88.82	11.64	11.16	11.57	11.81
6800	82.54	5.36	4.88	5.30	5.53
6900	81.30	4.12	3.63	4.06	4.30
7000	80.70	3.52	3.03	3.46	3.66
7100	80.32	3.14	2.65	3.08	3.27
7200	80.05	2.87	2.37	2.81	3.01
7300	79.85	2.67	2.17	2.62	2.81
7400	79.67	2.49	1.99	2.44	2.63
7500	79.53	2.35	1.84	2.30	2.49
7600	79.42	2.24	1.73	2.19	2.35
7700	79.34	2.16	1.65	2.11	2.25
7800	79.25	2.07	1.55	2.03	2.15
7900	79.15	1.97	1.45	1.93	2.05
8000	79.06	1.88	1.36	1.84	1.97
8100	78.99	1.81	1.28	1.77	1.89
8200	78.95	1.77	1.24	1.73	1.83
8300	78.88	1.70	1.17	1.67	1.73
8400	78.82	1.64	1.10	1.61	1.66
8500	78.76	1.58	1.04	1.55	1.61
8600	78.66	1.48	0.94	1.45	1.54
8700	78.58	1.40	0.86	1.38	1.48
8800	78.50	1.32	0.77	1.30	1.42
8900	78.42	1.24	0.69	1.22	1.36
9000	78.38	1.20	0.65	1.18	1.34
9100	78.44	1.26	0.70	1.25	1.39
9200	78.50	1.32	0.76	1.31	1.44
9300	78.54	1.36	0.80	1.35	1.49
9400	78.54	1.36	0.80	1.35	1.50
9500	78.52	1.34	0.77	1.33	1.49
9600	78.54	1.36	0.79	1.36	1.52
9700	78.55	1.37	0.80	1.37	1.51
9800	78.55	1.37	0.79	1.37	1.50
9900	78.55	1.37	0.79	1.37	1.49
10000	78.52	1.34	0.76	1.35	1.47

10127	78.47	1.29	0.71	1.30	1.47
10254	78.27	1.09	0.50	1.10	1.30
10381	78.14	0.96	0.37	0.98	1.19
10508	78.09	0.91	0.32	0.93	1.13
10635	78.01	0.83	0.23	0.85	1.04
10762	77.95	0.77	0.17	0.79	0.99
10889	77.90	0.72	0.12	0.75	0.95
11016	77.87	0.69	0.08	0.72	0.92
11143	77.84	0.66	0.05	0.69	0.91
11270	77.81	0.63	0.02	0.67	0.85
11397	77.76	0.58	-0.03	0.62	0.81
11524	77.70	0.52	-0.10	0.56	0.77
11651	77.63	0.45	-0.17	0.50	0.74
11778	77.57	0.39	-0.23	0.44	0.71
11905	77.55	0.37	-0.26	0.42	0.69
12032	77.52	0.34	-0.29	0.40	0.65
12159	77.49	0.31	-0.32	0.37	0.64
12286	77.46	0.28	-0.35	0.34	0.62
12413	77.40	0.22	-0.42	0.29	0.58
12540	77.38	0.20	-0.44	0.27	0.57
12667	77.36	0.18	-0.46	0.25	0.56
12794	77.32	0.14	-0.51	0.22	0.53
12921	77.30	0.12	-0.53	0.20	0.52

APPENDIX 6. Pumping Rates of Irrigation Wells in Washington County

Black (Macfarlane et al, 1990) test-pumped seven irrigation wells in Washington County as part of a statewide study of Dakota-well energy use. The pumping rates of the wells were measured using meters attached by the test crew. Table 2 shows these rates and the rates estimated by the Division of Water Resources (DWR) for appropriation purposes and the ratio of the former to the latter is shown in the right-hand column. This comparison shows that, for these seven high-capacity wells in Washington County, the mean ratio between the actual pumping rate and the rate used by the DWR to estimate total ground-water withdrawals is 0.78. This may be due to the fact that pumps and wells rarely operate with the efficiency they had immediately following their installation. It is likely that this discrepancy is also true of other wells. Therefore, to estimate the rate of ground-water use in Washington County for modelling purposes, DWR estimates were multiplied by a factor of 0.78.

Table A6.1. The measured pumping rate (Macfarlane et al., 1990) and the estimated pumping rate (Div. of Water Resources) of seven Dakota irrigation wells in Washington County.

Owner	Location	Measured pumping rate, r, (GPM)	Pumping Rate Estimated for DWR Water Appropriations , R (GPM)	r/R
Schramm	1S 5E 05ADA	278	200	1.39
Stigge	2S 3E 21D	411	488	0.84
Stigge	2S 3E 21DDB	411	600	0.68
Stigge	2S 3E 27CBA	313	330	0.95
Condray	5S 1E 16AAA	648	800	0.81
Condray	5S 1E 17DDC	365	650	0.56
Condray	5S 1E 17DCC	470	650	0.72
TOTAL		2896	3718	0.78

Appendix B. Chemical and Hydrogeological Data for Sampling Sites

ID	Township	Range	Section	1/4 Section	Well Depth (m)*	Top of Screen (m)*	Well Age (yrs)	Depth Water (m)*	Application Distance (m)*	Clay above Screen (m)*	Clay Upper 3.3 m*	Clay Upper 8.3 m*	NO3-N mg/L	Atrazine ppb	Deethyl-atrazine ppb	Deisopropyl-atrazine ppb	Soil Suface Texture Estimate	Geographic Location	Sampling Date	pH	Temp °C	Specific Conductance µmhos/cm
1	T12S	R20E	Sec 35	SE NW	25.9	19.8	25	6.7	30.5	6.7	2.1	6.7	.	<0.10	<0.10	<0.10	4	N	11/10/93	7.3	14.5	830
1B	T12S	R20E	Sec 35	SE NW	36.6	19.8	25	6.7	30.5	6.7	2.1	6.7	0.5	<0.10	<0.10	<0.10	4	N	12/20/94	7.1	15.4	790
2	T14S	R6E	Sec 34	NW NW SE NE	22.9	30.5	61	12.8	0.0	2.0	1.8	2.0	10.7	0.37	0.23	<0.10	2	N	07/14/94	7.3	16	900
3	T15S	R6E	Sec 2	NW NW NW	18.3	18.9	4	20.7	15.2	0.6	0.6	0.6	8.6	<0.10	<0.10	<0.10	2	N	07/14/94	7.3	18.1	750
4	T14S	R6E	Sec 36	SE SW SE	28.4	8.2	4	8.5	7.6	1.5	1.5	1.5	15.1	<0.10	<0.10	<0.10	2	N	07/14/94	7.2	16.4	860
6	T10S	R8E	Sec 9	NE NE NW	24.4	10.1	8	8.2	12.2	2.1	2.1	2.1	1.8	<0.10	<0.10	<0.10	2	N	08/02/94	7	19.7	780
7	T27S	R18W	Sec 36	NE SW	18.3	18.3	15	5.5	402.4	4.0	0.9	4.0	8.1	<0.10	<0.10	<0.10	8	S	11/02/94	7.3	16.9	380
8	T27S	R17W	Sec 34	NW	22.9	12.2	20	7.6	201.2	4.0	0.6	2.7	8.9	<0.10	<0.10	<0.10	8	S	11/02/94	7.2	16.9	300
9	T28S	R09W	Sec 31	SE	39.0	16.8	11	16.8	4.6	10.4	3.0	6.4	3.6	<0.10	<0.10	<0.10	6	S	11/02/94	7.4	17.3	470
10	T27S	R10W	Sec 5	NE SE	23.2	20.7	20	5.8	45.7	4.0	0.0	0.9	9.8	<0.10	<0.10	<0.10	6	S	11/03/94	7	16.3	310
11	T10S	R8E	Sec 8	NE NE SE	23.2	15.5	14	5.5	12.2	5.5	1.5	4.0	0.5	0.37	<0.10	<0.10	4	N	04/27/94	7.5	15.6	1040
11B	T10S	R8E	Sec 8	NE NE SE	19.8	15.5	14	5.5	12.2	5.5	1.5	4.0	0.2	1.1	0.59	0.3	4	N	12/20/94	7.4	15.6	650
12	T10S	R8E	Sec 9	NW NW SW	20.7	13.7	9	5.5	21.3	2.4	2.4	2.4	0.5	0.51	0.22	0.1	4	N	04/27/94	7.6	14.2	470
13	T10S	R8E	Sec 9	NE NE SW	20.7	12.5	9	5.8	15.2	4.3	2.4	4.3	0.5	1.3	0.45	0.18	4	N	04/27/94	7.5	12.6	560
13B	T10S	R8E	Sec 9	NE NE SW	20.7	12.5	9	5.8	15.2	4.3	2.4	4.3	0.2	0.87	0.32	0.15	4	N	12/20/94	7.3	16.5	630
14	T10S	R8E	Sec 9	NE NW SW	20.7	13.1	6	5.6	12.2	8.5	2.4	7.0	0.5	0.31	0.13	<0.10	4	N	05/05/94	7.5	12.5	560
14B	T10S	R8E	Sec 9	NE NW SW	21.6	13.1	6	5.6	12.2	8.5	2.4	7.0	0.3	1.0	0.59	0.28	4	N	12/20/94	7.4	14.2	530
15	T10S	R8E	Sec 9	NE NE NW SW	21.6	12.8	6	5.8	12.2	0.0	0.0	0.0	0.7	0.48	0.21	<0.10	4	N	05/05/94	7.3	15.7	660
15B	T10S	R8E	Sec 9	NE NE NW SW	19.7	12.8	6	5.8	12.2	0.0	0.0	0.0	0.2	0.64	0.28	0.13	4	N	12/20/94	7.2	16.7	650
16	T11S	R5E	Sec 35	NE NE NW	10.7	12.8	16	5.2	30.5	2.4	2.4	2.4	0.2	0.74	0.23	0.12	4	S	05/05/94	7.6	12.5	680
17	T22S	R1W	Sec 16	SW SW SW	10.4	5.8	15	5.8	22.9	4.9	2.1	4.9	6.4	5.1	0.61	0.32	2	S	06/14/94	7.1	17.8	960
18	T26S	R7W	Sec 13	SE SE SE	19.5	6.4	2	5.4	1.5	3.0	1.8	3.0	6	0.22	1.1	0.19	6	S	06/14/94	6.8	18.5	240
19	T24S	R6W	Sec 30	SW	19.8	6.1	50	3.4	22.9	1.8	1.8	1.8	10.6	<0.10	<0.10	<0.10	6	S	06/14/94	7.6	17.6	850
20	T26S	R7W	Sec 3	NW NW NW	6.1	15.9	2	10.4	22.9	1.2	0.6	1.2	11.6	<0.10	<0.10	<0.10	6	S	06/15/94	7.4	18.1	240
21	T26S	R7W	Sec 8	SE SE SE	19.5	2.1	2	3.3	3.0	3.4	1.8	3.4	16.8	1.1	0.72	<0.10	6	S	06/15/94	7.6	16.1	720
22	T26S	R7W	Sec 15	NE NE NE	17.7	15.5	2	16.0	18.3	1.5	0.3	0.3	19.1	<0.10	<0.10	<0.10	6	S	06/15/94	7.2	18.7	320
23	T13S	R2E	Sec 29	NE NE NW	17.7	11.6	4	5.8	15.2	3.4	1.5	3.4	0.7	<0.10	<0.10	<0.10	4	N	06/21/94	7.2	17.8	790
23B	T13S	R2E	Sec 29	NE NE NW	24.4	11.6	4	5.8	15.2	3.4	1.5	3.4	0.9	<0.10	<0.10	<0.10	4	N	12/20/94	7.1	15.5	800
24	T24S	R13W	Sec 16	SW SE	24.4	16.8	unknown	6.1	76.2	1.2	1.2	1.2	10.7	<0.10	<0.10	<0.10	6	S	06/21/94	7.4	22	530
25	T26S	R11W	Sec 30	SE NE	42.7	19.8	63	9.5	182.9	7.3	1.2	5.5	11.7	0.22	0.18	<0.10	6	S	06/21/94	7.4	16.8	570
26	T03S	R16E	Sec 17	NE NW NE	19.5	12.2	17	15.2	22.9	13.4	1.8	6.4	20.2	<0.10	<0.10	<0.10	2	N	08/02/94	7.3	17	630
27	T11S	R15E	Sec 20	NE NW SE	22.9	11.0	23	5.2	6.1	0.0	0.0	0.0	0.7	0.13	0.4	<0.10	4	N	08/02/94	7.2	17	640
28	T11S	R13E	Sec 1	SE NW SW	13.1	14.3	23	6.1	15.2	7.3	1.2	5.8	1.4	0.1	0.2	<0.10	1	N	08/02/94	7.3	16.7	640

Sample IDs with B indicate a duplicate sample. * Multiply number by 3.28 ft/m to obtain value in feet.

ID	Township	Range	Section	1/4 Section	Well Depth (m)*	Top of Screen (m)*	Well Age (yrs)	Depth Water (m)*	Application Distance (m)*	Clay above Screen (m)*	Clay Upper 3.3 m*	Clay Upper 8.3 m*	NO3-N mg/L	Atrazine ppb	Deethyl-atrazine ppb	Deisopropyl-atrazine ppb	Soil Suface Texture Estimate	Geographic Location	Sampling Date	pH	Temp °C	Specific Conductance µmhos/cm
29	T24S	R4W	Sec 10	SE NE NE	9.8	9.1	4	3.8	3.0	1.5	1.5	1.5	2.9	<0.10	0.12	<0.10	6	S	08/09/94	7.3	18.3	1360
30	T23S	R5W	Sec 31	SW SW SE	13.7	6.7	5	4.3	18.3	2.1	2.1	2.1	2.8	<0.10	<0.10	<0.10	6	S	08/09/94	7.8	18	1860
31	T25S	R08W	Sec 1	SW SW	11.3	7.6	47	2.1	402.4	1.5	1.5	1.5	7	<0.10	<0.10	<0.10	6	S	08/09/94	7.2	20.2	520
32	T26S	R7W	Sec 3	SW SW NW	25.9	6.7	13	3.1	122.0	0.0	0.0	0.0	18.1	<0.10	<0.10	<0.10	6	S	08/10/94	7	18.2	350
33	T29S	R11W	Sec 26	C SE	15.2	19.8	18	11.6	45.7	7.6	1.2	5.8	9.8	<0.10	<0.10	<0.10	8	S	08/10/94	7.2	18.2	480
34	T29S	R1E	Sec 22	NW NW NW	27.7	7.6	4	4.0	61.0	0.0	0.0	0.0	0.1	<0.10	<0.10	<0.10	3	S	08/10/94	7.3	18.6	870
35	T26S	R14W	Sec 12	NW SW	19.5	25.6	3	12.7	61.0	12.8	1.2	4.3	2.6	<0.10	<0.10	<0.10	8	S	08/23/94	7.4	18.9	390
36	T26S	R13W	Sec 9	NW NW	20.1	11.6	15	8.8	91.5	7.6	0.0	4.6	10	<0.10	<0.10	<0.10	8	S	08/23/94	7.1	18.5	510
37	T26S	R15W	Sec 2	NW SW	36.6	14.0	14	7.0	30.5	1.5	0.0	1.5	16.8	0.22	0.57	<0.10	8	S	08/23/94	7.3	17.5	590
38	T29S	R12W	Sec 22	NE NW	24.4	30.5	9	18.3	30.5	7.6	0.0	7.6	2	<0.10	<0.10	<0.10	7	S	08/24/94	7.8	17.8	340
39	T27S	R15W	Sec 10	NE SE	23.2	18.3	8	10.7	76.2	4.6	0.0	3.4	6.3	<0.10	<0.10	<0.10	8	S	08/24/94	7.3	17.8	270
40	T26S	R16W	Sec 2	SE SE	18.3	20.1	10	5.5	91.5	7.3	3.0	4.9	2.1	<0.10	<0.10	<0.10	8	S	08/24/94	7.6	19.7	390
41	T07S	R2E	Sec 27	SW SW	17.1	12.2	5	6.4	30.5	9.1	3.0	7.6	0.1	<0.10	<0.10	<0.10	8	N	09/07/94	7.1	16.6	1140
42	T07S	R2E	Sec 25	NW NW	17.4	11.0	10	5.5	30.5	4.9	2.1	4.6	13.7	0.28	1.1	<0.10	4	N	09/07/94	7.1	15.9	1170
43	T08S	R2E	Sec 18	SW NW	27.4	11.3	10	3.0	3.0	9.8	1.8	6.1	30	<0.10	<0.10	<0.10	2	N	09/07/94	7	18.7	1980
44	T21S	R12W	Sec 35	SW SW	24.4	21.3	12	5.5	45.7	21.3	3.0	7.6	11	<0.10	<0.10	<0.10	8	S	10/10/94	7.4	17.2	520
45	T21S	R12W	Sec 18	SE SE	29.3	18.3	10	7.9	45.7	13.7	3.0	7.6	8.1	<0.10	<0.10	<0.10	8	S	10/10/94	7.7	16.3	650
46	T21S	R12W	Sec 26	SE SW	25.9	24.1	13	7.0	61.0	21.6	0.0	4.0	40.2	<0.10	<0.10	<0.10	8	S	10/10/94	7.4	17.8	1000
47	T26S	R16W	Sec 4	NE NW	19.5	19.8	12	8.5	76.2	2.1	0.9	0.9	15.3	<0.10	0.19	<0.10	6	S	10/11/94	7.7	16	530
48	T25S	R16W	Sec 27	SW	30.5	16.5	11	8.5	76.2	5.8	0.0	0.0	9.3	<0.10	<0.10	<0.10	6	S	10/11/94	7.6	17.8	550
49	T27S	R17W	Sec 15	NE SE	24.4	24.4	5	10.7	45.7	7.0	1.2	5.8	13.1	<0.10	0.29	<0.10	8	S	10/11/94	7.7	17.1	390
50	T27S	R17W	Sec 21	SE NE	23.2	18.3	4	9.5	9.1	6.4	2.4	6.1	4.3	<0.10	<0.10	<0.10	8	S	10/11/94	7.6	17.1	290
51	T24S	R17W	Sec 24	NW NW	24.4	17.1	11	10.1	45.7	2.7	0.3	2.7	8.5	<0.10	<0.10	<0.10	8	S	10/11/94	7.7	16.6	530
52	T22S	R12W	Sec 31	NE NE	31.1	18.3	21	14.9	14.9	14.6	2.1	6.7	6.5	<0.10	<0.10	<0.10	4	S	10/12/94	7.7	16.1	500
53	T21S	R12W	Sec 26	NE NE	25.9	25.0	13	7.0	804.9	15.9	0.6	5.2	4.4	<0.10	<0.10	<0.10	8	S	10/12/94	7.7	18	660
54	T21S	R12W	Sec 26	NE NE	28.0	19.8	27	11.0	807.9	11.6	3.0	7.6	3.7	<0.10	<0.10	<0.10	8	S	10/12/94	7.6	17.2	1000
55	T23S	R1W	Sec 19	NE SE SE	15.2	24.7	0.92	14.3	30.5	13.7	0.9	2.1	9.1	<0.10	<0.10	<0.10	2	S	11/02/94	6.6	16.4	310
56	T24S	R3W	Sec 22	SW SE	29.9	9.1	24	4.3	61.0	0.0	0.0	0.0	0.2	<0.10	<0.10	<0.10	5	S	11/02/94	6.6	17.1	1090
57	T26S	R6W	Sec 18	NE SE	36.6	23.2	0.33	8.2	7.6	1.8	1.2	1.2	8.2	<0.10	<0.10	<0.10	6	S	11/02/94	7.2	17	320
58	T23S	R2W	Sec 30	SE	22.0	30.5	28	12.8	402.4	27.1	1.5	4.0	17.3	<0.10	0.18	<0.10	2	S	11/09/94	6.8	16.6	550
59	T24S	R3W	Sec 16	SE	18.3	14.0	14	3.7	15.2	4.6	0.0	1.5	0.1	<0.10	<0.10	<0.10	5	S	11/09/94	7	16	1320
60	T24S	R3W	Sec 30	SE	33.5	12.2	11	3.0	152.4	2.1	0.6	0.6	12	<0.10	<0.10	<0.10	5	S	11/09/94	7.1	15.6	1020
61	T20S	R4W	Sec 1	NW	22.0	30.5	18	13.1	42.7	21.6	2.1	6.4	9.9	0.48	0.23	<0.10	4	S	11/09/94	7	15.9	950
62	T20S	R1W	Sec 15	SW SE	10.7	15.9	32	5.2	402.4	6.1	1.8	6.1	2.7	<0.10	<0.10	<0.10	2	S	11/09/94	7.1	16.2	360
63	T17S	R5W	Sec 26	NW	12.8	7.6	10	7.6	804.9	2.7	1.2	2.7	5.6	<0.10	<0.10	<0.10	3	S	11/09/94	6.9	15.8	960

Sample IDs with B indicate a duplicate sample. * Multiply number by 3.28 ft/m to obtain value in feet.

ID	Township	Range	Section	1/4 Section	Well Depth (m)*	Top of Screen (m)*	Well Age (yrs)	Depth Water (m)*	Application Distance (m)*	Clay above Screen (m)*	Clay Upper 3.3 m*	Clay Upper 8.3 m*	NO3-N mg/L	Atrazine ppb	Deethyl-atrazine ppb	Deisopropyl-atrazine ppb	Soil Suface Texture Estimate	Geographic Location	Sampling Date	pH	Temp °C	Specific Conductance µmhos/cm
64	T5S	R2W	Sec 30	SW	13.7	9.8	18	4.6	22.9	0.0	0.0	0.0	2.5	<0.10	0.23	<0.10	6	N	11/30/94	7.2	14.6	1100
65	T5S	R2W	Sec 20	NE NE	18.3	6.1	21	6.1	22.9	0.0	0.0	0.0	4.3	<0.10	<0.10	<0.10	6	N	11/30/94	7.3	14.3	700
66	T5S	R2W	Sec 32	SW NE	29.9	15.2	18	7.3	30.5	0.0	0.0	0.0	16.2	<0.10	<0.10	<0.10	6	N	11/30/94	7.5	12.7	1030
67	T5S	R2E	Sec 12	SW SW	23.8	23.5	14	12.2	6.1	5.8	2.1	2.1	10.2	<0.10	<0.10	<0.10	4	N	11/30/94	7.8	14.5	340
68	T5S	R2E	Sec 11	NW NW	42.1	17.7	17	15.5	45.7	2.1	2.1	2.1	32.1	3.3	0.51	0.11	4	N	11/30/94	6.8	14.7	860
69	T5S	R2E	Sec 25	NW SW	58.5	39.0	3	26.5	15.2	38.1	2.1	6.7	5.8	<0.10	<0.10	<0.10	4	N	11/30/94	6.9	14.6	780
70	T26S	R25W	Sec 25	SW NE	23.2	46.3	42	20.7	0.0	20.7	3.0	6.4	6.3	<0.10	0.11	<0.10	4	S	12/12/94	7.4	18	910
71	T27S	R10W	Sec 1	NE SE	24.4	17.1	21	19.8	122.0	5.5	0.6	0.6	2.7	0.34	0.6	<0.10	6	S	11/03/94	7.1	16.6	180
72	T27S	R8W	Sec 16	SE NE	24.7	18.3	28	8.2	0.0	7.0	1.8	6.4	27.8	<0.10	0.39	<0.10	8	S	11/03/94	7.3	17.4	560
73	T7S	R3E	Sec 29	SW SW	36.6	18.6	21	15.2	201.2	0.0	0.0	0.0	0.3	<0.10	<0.10	<0.10	2	N	12/02/94	7	15.9	off scale
74	T8S	R3E	Sec 22	NW NW	49.4	30.5	10	12.2	91.5	10.4	0.0	5.2	1.9	<0.10	<0.10	<0.10	4	N	12/02/94	7.1	15.8	780
75	T26S	R25W	Sec 35	NW SW	55.8	37.2	22	11.9	0.0	23.2	1.2	2.1	4.9	<0.10	<0.10	<0.10	3	S	12/12/94	7.5	15.8	940
76	T27S	R24W	Sec 6	NW SW	19.8	43.6	13	21.3	30.5	20.1	2.7	4.6	3.9	<0.10	<0.10	<0.10	8	S	12/12/94	7.7	16.5	450
78	T9S	R18W	Sec 34	NE NE	14.3	13.7	13	11.9	0.0	5.2	1.5	4.0	11.4	0.63	<0.10	<0.10	1	R	12/13/94	7.5	14.2	off scale
79	T9S	R18W	Sec 34	NW NE	19.8	8.8	13	7.9	45.7	7.3	0.9	4.0	18.6	<0.10	<0.10	<0.10	1	R	12/13/94	7.6	14.6	1490
80	T9S	R18W	Sec 34	NE SE	65.0	13.7	10	13.7	402.4	7.6	2.1	6.7	6.4	<0.10	<0.10	<0.10	1	R	12/13/94	7.4	15.3	1940

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