MECHANISMS, DISTRIBUTION AND FREQUENCY OF GROUNDWATER RECHARGE IN AN UPLAND AREA OF WESTERN NORTH DAKOTA

by

Bernd W. Rehm, Gerald H. Groenewold, and William M. Peterson

REPORT OF INVESTIGATION NO. 75

NORTH DAKOTA GEOLOGICAL SURVEY

Don L. Halvorson, State Geologist

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INTRODUCTION

Groundwater is essential to human development in the semi-arid region of the Northern Great Plains. Aquifers supply much of the domestic and agricultural water needs in the region and maintain the baseflow of several streams and rivers in the area. It is through the process of groundwater recharge that water levels and fluxes within the groundwater flow systems are maintained.

In North Dakota 48 percent of public water supplied and 78 percent of rural water use is obtained from groundwater (Pettyjohn, and others, 1979). Over much of the central and western portions of the state many rural and some municipal groundwater users withdraw groundwater from shallow (less than 75 m deep) flow systems whose recharge and discharge areas are within 50 km of each other. The population of this region is expected to grow during the next decades as lignite reserves in the area are developed to fuel electrical generating stations and supply feedstocks to gasification facilities. Groundwater recharge rates are one piece of data that will be required in developing adequate water resource management plans for these people.

Lignite coal underlies much of the western half of North Dakota. Surface mining for this coal will disturb large areas of the state over the next 20 to 30 years. Current mining activities disturb approximately 4.2 $km^2 \cdot yr^{-1}$ (U.S. Bureau Land Mgmt., 1978). By 1983 it is estimated that 8.5 $km^2 \cdot yr^{-1}$ will be disturbed by mining, and an additional 10.0 $km^2 \cdot yr^{-1}$ will be disturbed by projects proposed to begin operation in 1990 (U.S. Bureau Land Mgmt., 1978). The disturbance of the landscape may drastically alter the potential for groundwater recharge. If recharge rates through reclaimed landscapes are significantly less than through the currently undisturbed landscapes, water levels in shallow flow systems may decline to the point where groundwater supplies are disrupted and stream baseflows are decreased. Proposed regulations developed by the Office of Surface Mining under Section 102 (i) of the Surface Mining Control and Reclamation Act of 1977 (Pub. L. 95-89, 91 Stat. 445 (30 U.S.C. Section 1201)) recognize the potential of surface mining to change groundwater recharge rates. Section 816.38 (3) of the proposed regulations state that mine operators must undertake "...reclamation which will restore approximate premining recharge capacity through restoration of the capability of the reclaimed areas as a whole ...to transmit water to the ground water system." The essential first step in attaining this goal is to gain an understanding of how recharge occurs in undisturbed landscapes.

This paper first reviews the factors that control the movement through the unsaturated zone. The remainder of the paper then focuses on data collected from the Falkirk site in central North Dakota. Most of the study area will have been disturbed by surface mining for lignite by the year 2010. Mining operations have already

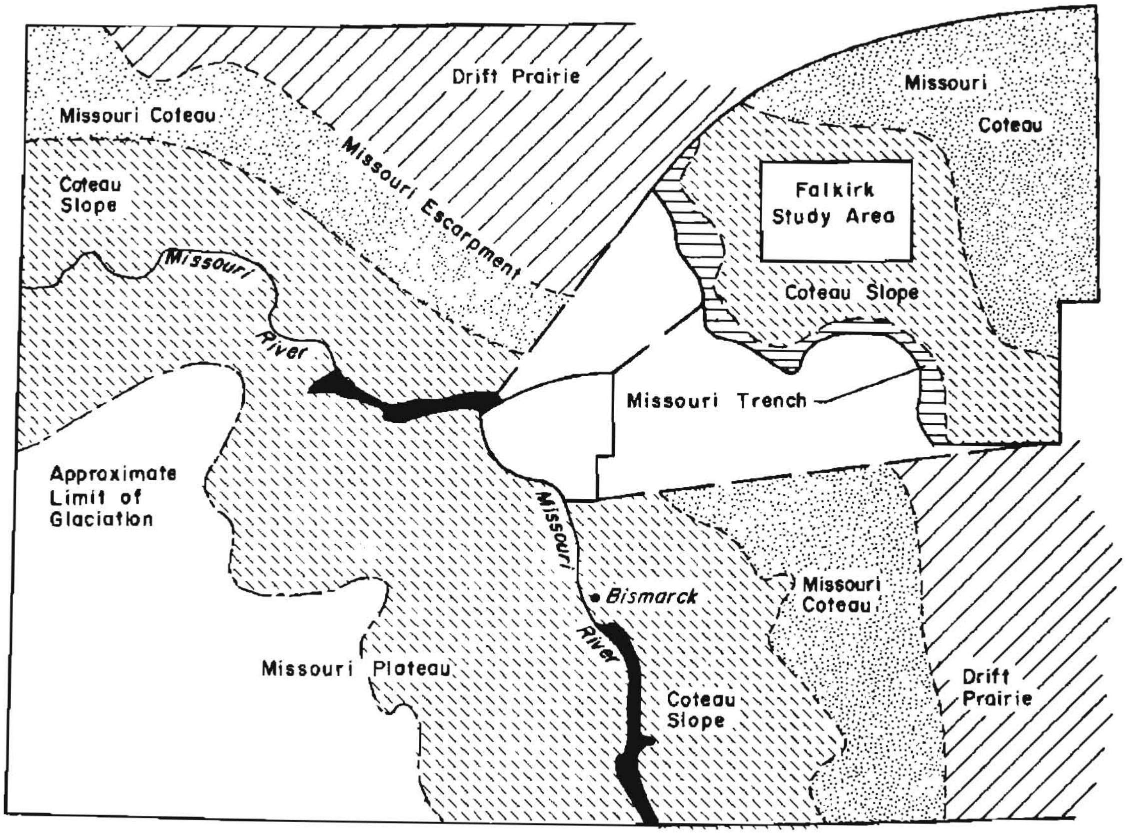


Figure 1. Location of the Falkirk study area.

begun on the southwestern edge of the site. The data reported here is being collected as part of a larger study which is designed to evaluate the impact of surface mining and reclamation on groundwater and surface water quality and quantity. This work has been supported by U.S. Environmental Protection Agency Grant No. R-803727 and Office of Water Resources Technology Grant No. A-061-NDAK. Investigations at this site are continuing under funding from U.S. Environmental Protection Agency Grant No. R-805935. The authors would also like to acknowledge the Falkirk Mining Company for their cooperation and assistance in various phases of the groundwater investigations and all those landowners in the area who allowed us access to their property.

STUDY AREA DESCRIPTION

The study area occupies approximately 150 km^2 in southeastern McLean County, North Dakota (fig. 1). The site is about 11 km to the east and north of the Missouri River and about 10 km southeast of Lake Sakakawea. The town of Underwood (population of about 1,300) occupies 0.6 km^2 in the north-central portion of the study area.

The climate in the study area is classified as a sub-humid continental climate. The site has received an average annual precipitation of 0.443 m during the period 1970 to 1979. Precipitation amounts are highly variable from year to year (table 1). June, for example, has the highest average value, 0.086 m, but during the decade

TABLE 1. Measured total monthly precipitation at the Falkirk study area. Data collected from NOAA weather station in Underwood and one to three rain gauges operated by the NDSU Soils Department. Values followed by an asterisk are an average of the NOAA data and the NDSU data.

	1970	1971	1972	1973	1974	1975	1976	1977	1978	1979	Average
Jan	1.2	2.0	1.1	0.2	0.4	1.4	1.4*	2.1*	0.4	0.5	1.1
Feb	0.8	0.4	2.1	0.4	1.6	1.0	0.9*	0.7*	1.3	3.3	1.2
Mar	2.2	2.3	2.3	1.4	1.3	3.4	1.6*	1.0*	0.8	1.9	1.8
Apr	9.7	3.6	2.4	1.8	9.1	12.8	8.6*	0.2	4.4*	2.7*	5.5
May	8.1	2.4	9.1	3.6	10.5	5.0	1.6*	5.8*	6.8*	3.1*	5.6
Jun	6.5	10.2	15.3	4.7	1.6	14.9	8.7*	6.1*	11.3*	7.0*	8.6
Jul	11.4	4.6	6.3	4.4	3.4	4.6	2.7*	9.0*	7.9*	8.5*	6.3
Aug	1.0	T	6.3	3.5	4.8	2.4	2.8*	5.2*	1.2*	6.5*	3.4
Sep	4.9	6.6	3.6	7.4	2.4	5.3	3.0*	10.2*	8.4*	3.0*	5.5
Oct	1.5	7.3	4.2	4.0	0.9	2.7	1.3*	1.9*	1.0	2.3	2.7
Nov	1.8	1.0	0.5	1.2	0.8	0.9	0.6*	2.2*	2.7	0.8*	1.3
Dec	0.5	0.6	2.3	2.1	1.1	1.7	1.1*	0.9*	1.8	1.2*	1.3
Total	49.6	41.0	55.5	34.7	37.9	56.1	34.3	45.3	48.0	40.8	44.3

the June precipitation values range from 0.016 to 0.153 metres. The average annual temperature in the area is 4°C, with July being the warmest month (average temperature of 27°C), and the coldest month is January with an average temperature of about -14°C. The length of the frost-free period ranges from 110 to 120 days. Prevailing winds in the Bismarck area are from the west-northwest except in the summer when they shift to the south-southeast and east (NOAA, 1976). Mean wind speeds vary from 16 km/h in December to 21 km/h in April with a yearly average speed of 17 km/h (NOAA, 1976).

The Falkirk study area is located on the Coteau Slope of the Great Plains physiographic province (fig. 1). The Coteau Slope is characterized by moderate local relief (generally less than 8 m) except in the vicinity of major valleys where the relief can be as great as 30 m. Drainage is well developed with small, intermittent streams draining southward and westward to the Missouri River (Bluemle, 1971). The Falkirk site is situated on the southern half of a low, upland drainage divide. Post-glacial erosion has been minor in most of the area and the landscape has remained much as it was left by the last Wisconsinan glaciation. The land surface around the town of Underwood is generally undulating to rolling. North of the town the landscape is dotted by numerous prairie pothole sloughs. East, west, and south of Underwood the land surface is characterized by longer slopes, better integrated drainage, and few sloughs. The western edge of the site is drained by small, intermittent streams that flow to the Missouri River. The southern portion of the site drains into the Weller Slough area, a complex of several large sloughs located along the trend of a broad, buried valley (fig. 2). These sloughs drain to the east and south, with the water eventually reaching the Missouri River. The eastern edge of the study area drains into Coal Lake (fig. 2). Coal Lake is situated in a closed depression in the bottom of Coal Lake Coulee, a long, narrow valley that borders the entire eastern edge of the site. The surface area of Coal Lake has varied by as much as a factor of four during the last 20 years. The slough eventually spills to the southwest, toward the Weller Slough area.

The shallow subsurface within the Falkirk study area is made up of sandy, silty, clayey, and lignitic sediments of the Sentinel Butte and Bullion Creek Formations, Fort Union Group (Paleocene) and sand, gravel, and pebble-loam of the Coleharbor Formation (Pleistocene). The Paleocene units consist of alternating beds of silt, clay, sand, and lignite. The thickness of the beds varies from less than 0.25 m to as much as 25 m. The following paragraphs briefly describe the geology and hydrogeology of the study area. A more detailed description of the site can be found in Groenewold and others (1979).

The Hagel bed (fig. 3) is the lowest stratigraphic unit which concerns this report. This unit consists of two coal beds within the study area, the upper, main Hagel bed and the lower, B bed split of the Hagel bed. The Hagel bed underlies the entire study area. The main Hagel bed is generally 3 m thick and the B bed is usually 1 m thick. The two coal beds are separated by a clay to silty clay unit that varies in thickness from less than 0.25 m to 10 m. The coal beds will be mined throughout most of the study area. Mining operations are currently under way in the southwestern portion of the study area.

The Kinneman Creek interval overlies the Hagel bed. It consists of silty sand, silt, clay, and minor lignite beds. Within the Kinneman Creek interval is the Underwood sand, a large fluvial sand unit that underlies approximately one-third of the study area (fig. 3). The Underwood sand is 24 m thick, and exposed at the ground surface, immediately south of the town of Underwood. The unit thins radially outward from this point, generally pinching out completely within 5 km.

Glacial drift of the Coleharbor Formation (Pleistocene) covers almost all of the study area. The drift covers a highly dissected pre-Pleistocene topography characterized by deep (150 m) valleys cutting through broad uplands which had a much more subdued, rolling topography. Three of these deep, pre-glacial, valleys surround the Falkirk site, Coal Lake Coulee to the east, the Weller Slough area to the south, and the former Knife River Valley to the west. The drift in the upland areas consists

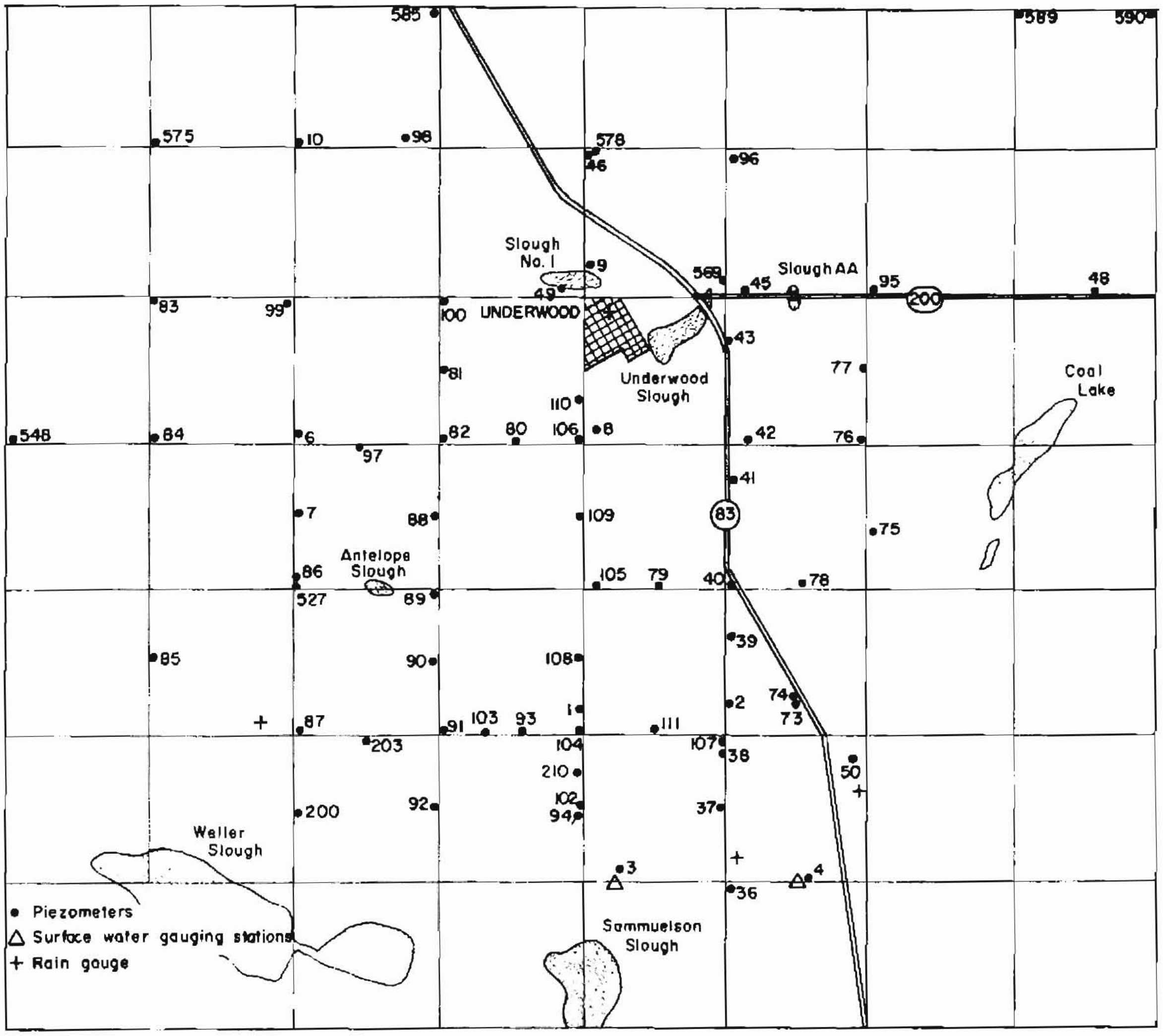


Figure 2. Location of piezometer nests, rain gauges, and surface water gauging stations.

largely of pebble-loam (till) from 3 to 15 metres thick. Glacio-fluvial sand and gravel deposits of highly variable thickness and areal extent are also found within the uplands. The deposits range from thin (0.1 to 1 metre) beds within the pebble-loam to thick (several metres) deposits that generally separate the pebble-loam from the Paleocene sediments.

Aquifers in the Paleocene sediments consist of lignite and slightly indurated sand. The mean hydraulic conductivity of these materials is 3×10^{-6} $m \cdot s^{-1}$ and 1×10^{-6} $m \cdot s^{-1}$, respectively (Rehm and others, 1980).

The clayey silt to clay units within the Paleocene sediments are aquitards. The mean hydraulic conductivity of these sediments is 3×10^{-8} $m \cdot s^{-1}$ and the range of hydraulic conductivity is from 1×10^{-5} $m \cdot s^{-1}$ to 5×10^{-11} $m \cdot s^{-1}$. The range in values results from fracturing of the sediments. The low end of the range is representative of the unfractured, intergranular permeability, and the high end is representative of highly fractured sediments.

The aquifers within the Coleharbor Formation consist of sand and gravel. These are the most permeable units in the study area with a mean hydraulic

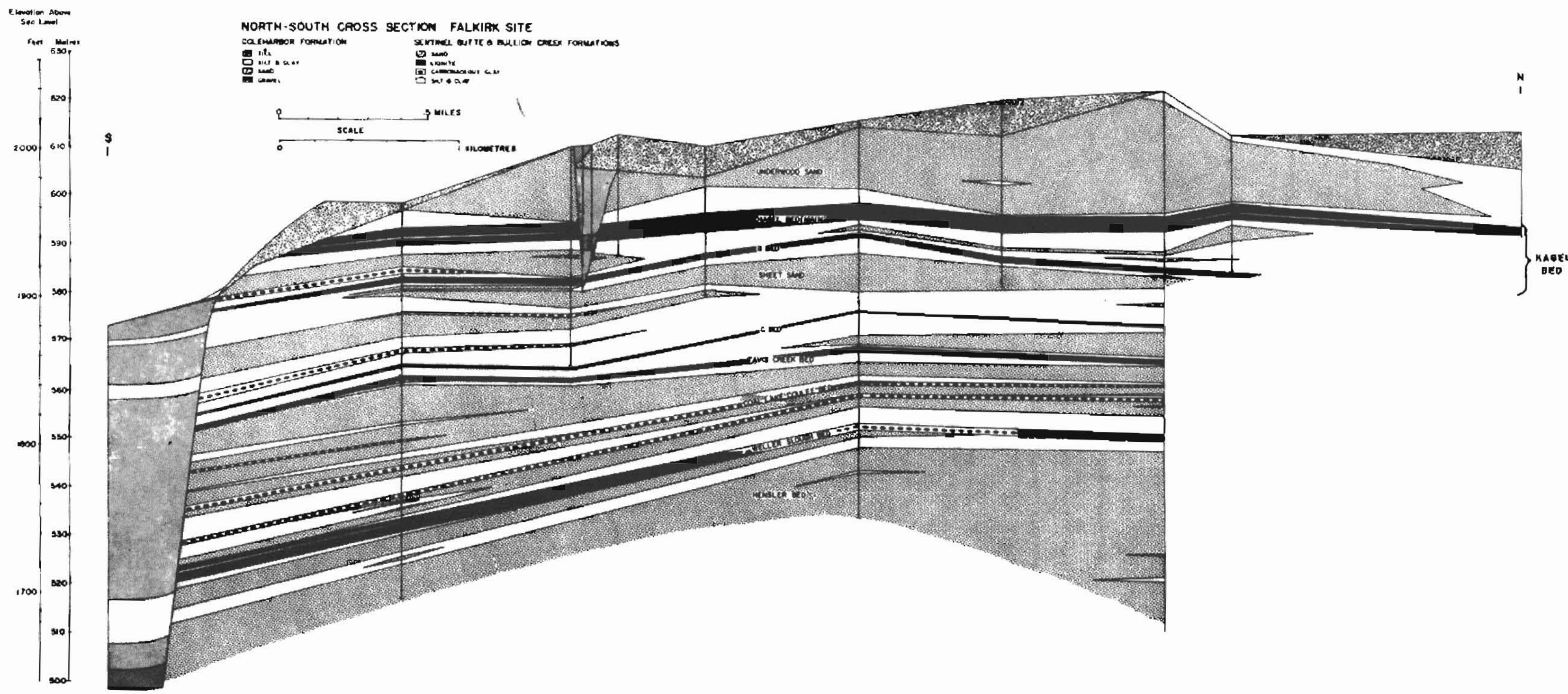


Figure 3. Geologic cross section through the Falkirk study area.

conductivity of 5 x 10^{-5} $m \cdot s^{-1}$. The very low mean hydraulic conductivity (7 x 10^{-9} $m \cdot s^{-1}$) of the pebble-loam make this unit an aquitard. The range of permeability extends over six orders of magnitude, with fractures again being the primary cause of the variability.

Groundwater flow within all the aquitards, regardless of their stratigraphic position, is predominantly vertically downward. Flow within the Underwood sand is both radially outward from an area immediately south of the town of Underwood and downward towards the Hagel bed. Within the main Hagel bed and the B bed groundwater flow is primarily horizontal, or parallel to the beds. Leakage from the base of the lignite beds does occur, with approximately 70 percent of the total inflow into the main Hagel bed being lost to downward leakage over the 150 km^2 study area (Williams, Koob, and Rechard, in prep.). The lateral flow of all the aquifers from the Hagel bed to the base of the Bullion Creek Formation discharges into the glacial drift that fills the pre-glacial valleys surrounding the study area.

MECHANISMS OF GROUNDWATER RECHARGE

Groundwater recharge is a complicated process which is influenced by a great number of variables. These can be divided into three groups: climatic variables, unsaturated zone variables, and saturated zone variables. The climatic variables include such factors as the intensity, duration, spatial, and temporal variations in precipitation, evaporation, and transpiration. The most important unsaturated zone variables are hydraulic conductivity as a function of pressure head ($K(\Psi)$), volumetric moisture content ($\theta(\Psi)$), pressure head (Ψ), and depth to the water table. Once in the saturated zone the controlling parameters are the saturated hydraulic conductivity K, porosity η, and hydraulic head ϕ.

Groundwater recharge is not simply the quantity of water that penetrates below the ground surface. This quantity of water, together with the processes of flow through the unsaturated zone is defined as infiltration. The downward migration of infiltrating water can be modified by evaporation, transpiration, and frost formation. Groundwater recharge refers only to that portion of the infiltrating water that reaches the saturated zone and begins to flow downward from the water table. Groundwater discharge is then defined as the removal of water from the saturated zone through the water table and the upward flow associated with that removal. The relationships between these processes are shown in figure 4. Using this conceptual model, fluctuations in the hydraulic head of the water table result whenever the flux of water to the water table (infiltration less losses to evapotranspiration and frost formation) is not equal to the rate at which water moves downward through the saturated zone. In sandy textured materials, where the saturated hydraulic conductivity is

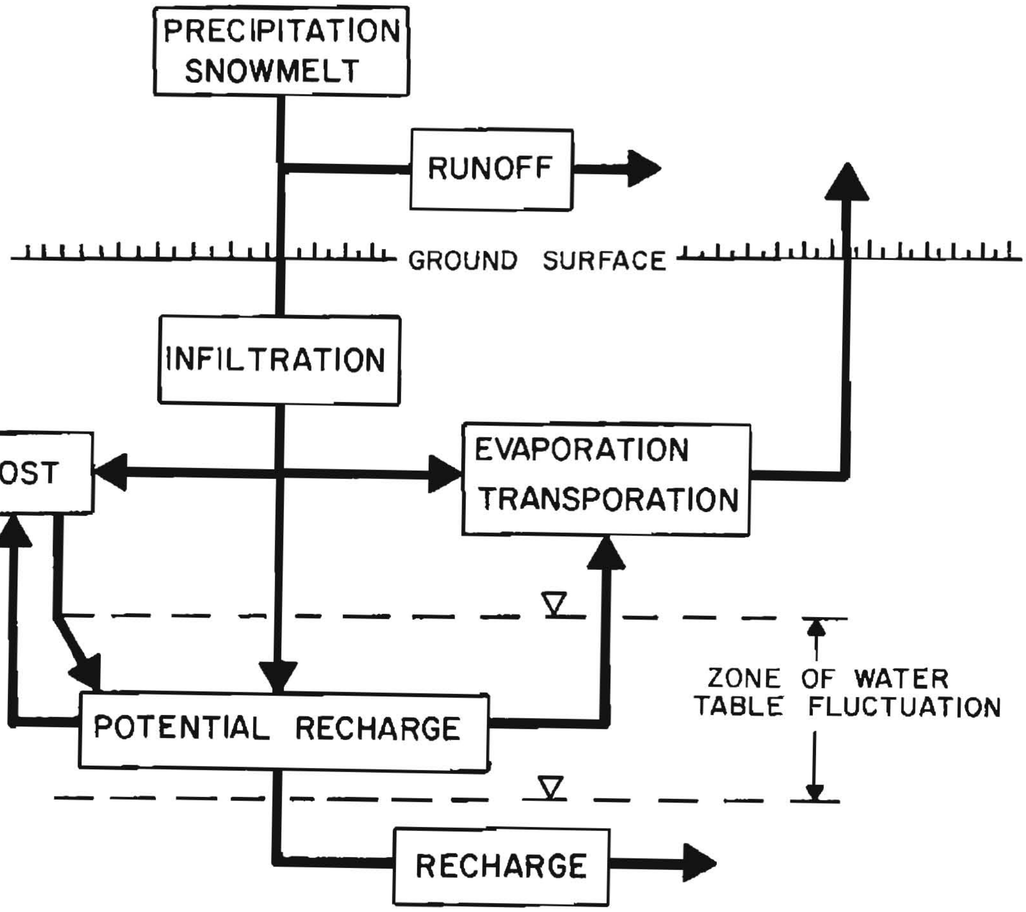


Figure 4. Potential paths of water movement from the ground surface to the water table.

high, the flux away from the water table can be large. For the water table to rise in this material the downward flux to the water table must be greater than the flux away from the water table in the saturated zone. If the infiltration rate is less than the flow from the water table, the water table will fall even though water is moving downward through the unsaturated zone.

The process of groundwater recharge begins with the infiltration of water through the unsaturated zone to the water table. A numerical modeling study by Freeze (1969) illustrates how the physical properties of the unsaturated zone can affect the rate at which water reaches the water table. The parameters that control infiltration include soil type, the rate and duration of precipitation and evapotranspiration, the allowable depth of ponding, the antecedent soil moisture conditions, and the depth to the water table.

Several generalized conclusions can be drawn from the Freeze study:

1. The ability of the unsaturated zone to transmit infiltrating water is strongly dependent on the relationships between pressure head, hydraulic conductivity, moisture content, and specific moisture capacity. Generally soils with a high hydraulic conductivity, and/or

low specific moisture capacity, and/or high moisture content over a wide range of pressure heads are most likely to transmit water to the water table.

2. If ponded water conditions are reached at the ground surface the infiltrating water will generally reach the water table if the water table is at a relatively shallow depth. Under certain moisture conditions, or for some soil types, ponding must occur before water can reach the saturated zone. The depth of ponding does little to affect the rate of movement of water through the unsaturated zone, but the surface infiltration rate does increase with increasing pond depth.
3. The potential for moving water through the unsaturated zone is more dependent upon the duration of the precipitation event than the intensity of the event. Thus, a low intensity event of long duration is more effective than a high intensity event of short duration.
4. If a soil profile has a relatively high moisture content prior to an infiltration event, there is a greater probability of the infiltrating water reaching the water table, rather than being used to make up the moisture deficit of the soil profile.
5. High evaporation rates following an infiltration event can remove much of the infiltrated water before that water reaches the water table.

The critical role of the physical properties of soil in controlling the rate at which infiltrating water reaches the water table requires that the interrelationships between hydraulic conductivity, moisture content, pressure head, and potential groundwater recharge be discussed in some detail. Freeze (1969) evaluated three types of soil: two sands and a silt-loam. The characteristic curves, ($K(\psi)$) and ($\theta(\psi)$), for the silt-loam and one of the sands are reproduced in figure 5. The second sand had characteristic curves intermediate between these first two materials.

The modeling results indicated that the ($K(\psi)$) relationship is critical to the infiltration rate. When all other variables are held constant, the silt-loam, with a saturated hydraulic conductivity of 2×10^{-6} $m \cdot s^{-1}$, showed the greatest infiltration rate. For the coarse sand, with a saturated hydraulic conductivity of 1×10^{-4} $m \cdot s^{-1}$, infiltration proceeded very slowly. The low infiltration in the more permeable material is the result of several factors. In a dry profile with a given pressure head the infiltrating water must satisfy a larger moisture deficit in the sand than for the silt-loam. In other words, the specific moisture capacity, $\frac{d\theta}{d\psi}$, is greater for the sand than for the silt-loam. In conjunction with this, the ($K(\psi)$) characteristic curves show that the hydraulic conductivity of the sand is extremely low until the moisture content of the sand approaches saturation. Given the same initial pressure heads and surface inputs of water the

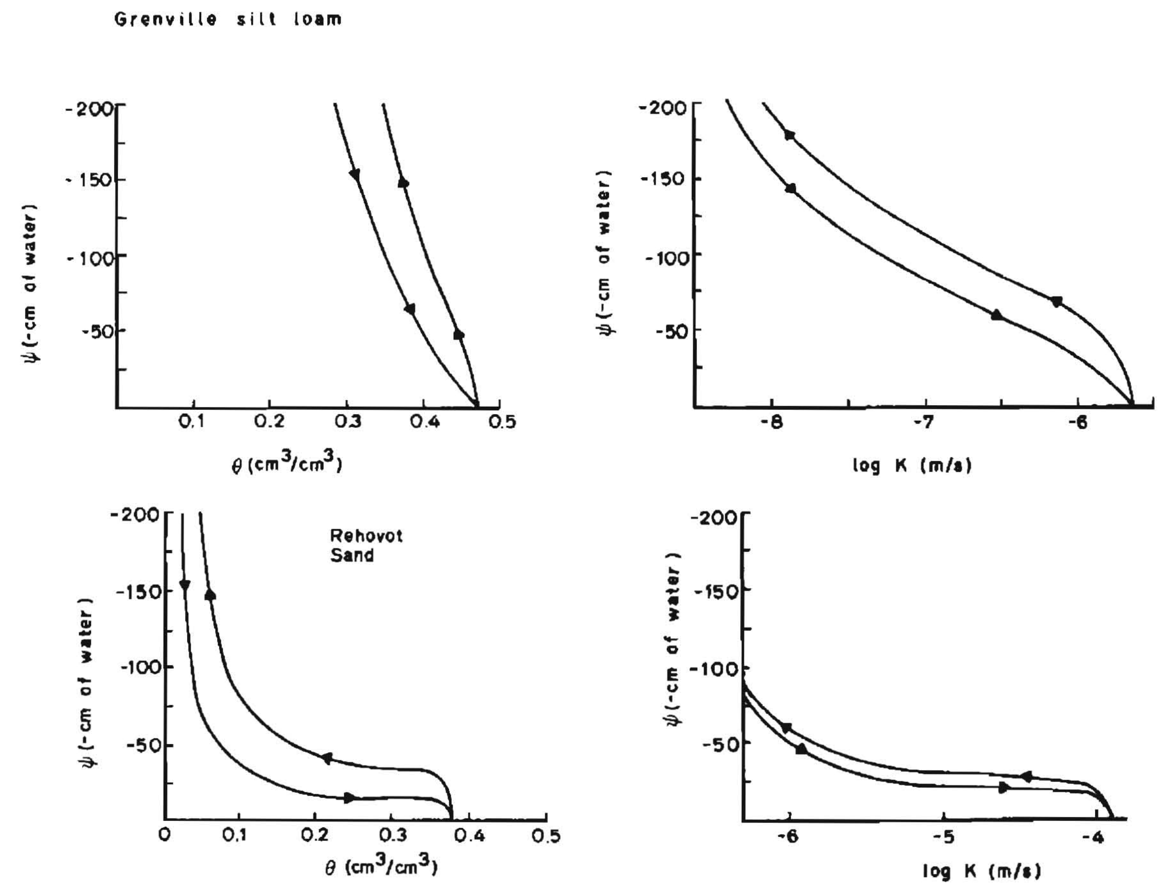


Figure 5. Measured values of hydraulic conductivity (K(Ψ)) and soil moisture (Θ(Ψ)) for Grenville silt loam and Rehouot sand (after Freeze, 1969).

actual hydraulic conductivity of a sandy material will not be significantly greater, and may even be less than the hydraulic conductivity of a loamy material.

Under simulated precipitation rates high enough to produce ponding (7.9 $cm \cdot hr^{-1}$) a saturated zone generally formed beneath the ponded area and moved downward towards the water table. The rate at which this saturated zone moved through the silt-loam was approximately 0.023 $m \cdot min^{-1}$. The rate for the intermediate sand was about 0.011 $m \cdot min^{-1}$. The coarse sand simulation never developed ponded conditions and no saturated zone formed just below the ground surface. Infiltrating water reached the water table, at a depth of 0.9 m, ahead of the downward migrating saturated zone. The water table rose 0.04 m 48 minutes after precipitation began in the intermediate sand, but no change in the water-table hydraulic head was evident before 36 minutes had elapsed. For the silt-loam the water-table rise began after 32 minutes and by 36 minutes had risen 0.03 m.

The data required to develop characteristic curves for the surficial loamy material within the Falkirk study area are not available. Cassel (1974) looked at similar, glacial, till-derived, soils in eastern North Dakota (fig. 6). These three materials have characteristic curves similar to the silt-loam evaluated by Freeze (1969). As long as

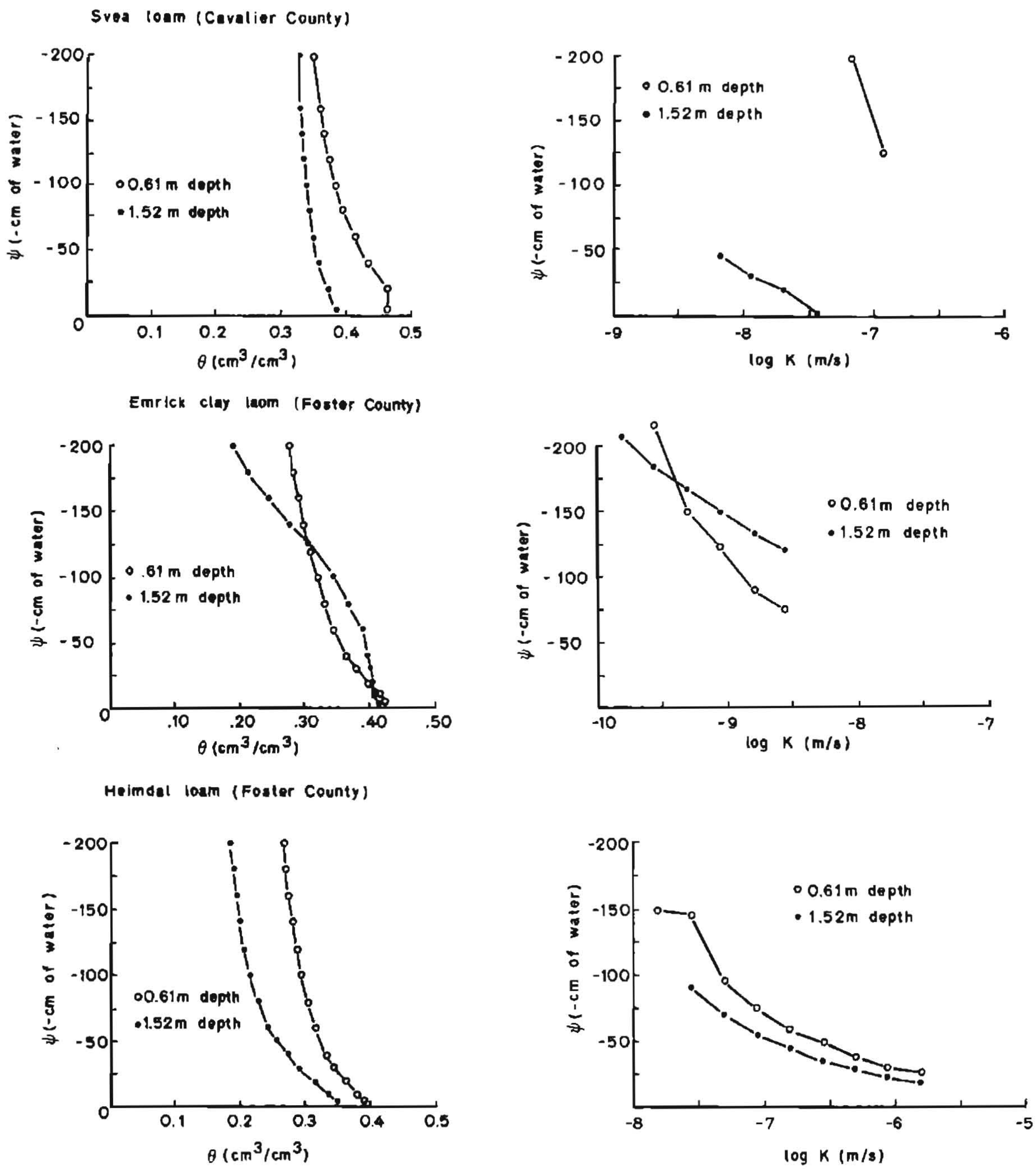


Figure 6. Measured values of hydraulic conductivity (K(Ψ)) and soil moisture (θ(Ψ)) of three loamy soils from eastern North Dakota (after Cassel, 1974).

flow through fractures is insignificant it is reasonable to assume that infiltration at the Falkirk study area will follow a pattern similar to that of the silt-loam described by Freeze.

Once the infiltrating water reaches the water table the water can move away by several routes. The water can flow downward, under saturated flow conditions, and become actual groundwater recharge. If the rate of downward saturated flow is less than the rate at which infiltrating water reaches the water table, the water table will rise in response to the infiltration event. After infiltration ceases to reach the saturated zone the water table will decline as the water continues to flow downward. Where the water table is relatively shallow, evaporation and transpiration will transport water upward into the unsaturated zone in the summer and the formation of frost can move water upward during the winter months.

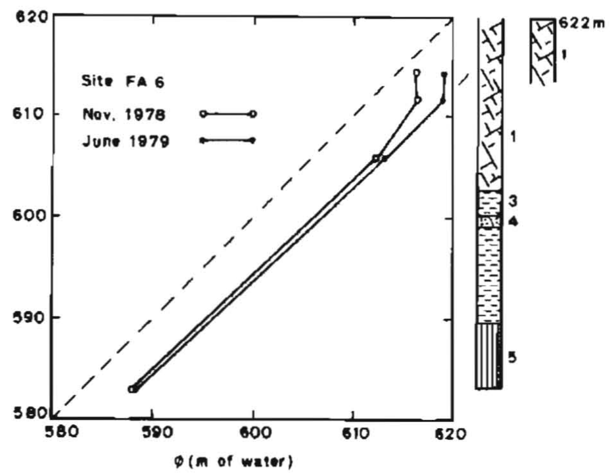
The rate of groundwater recharge, or downward flow from the water table, is controlled by the permeability of the porous medium and the hydraulic gradient below the water table. The hydraulic conductivity is invariant over the time scales involved in the determination of groundwater recharge; therefore, the gradient becomes the controlling parameter of groundwater recharge rates.

Hydraulic profiles at eight sites in the Falkirk study area that are instrumented in shallow portions of the groundwater flow system are shown in figures 7a through 7h. The figures show vertical profiles through these sites in terms of hydraulic head (ϕ). The pressure head (Ψ) is determined by determining the difference between the hydraulic head and the elevation head (Z) as indicated by the dotted line. Elevation heads were taken as the base of the screened intervals of the piezometers. Profiles are shown for the fall and spring of the year when hydraulic heads are at their lowest and highest levels, respectively.

The profiles at site Fa 6 (fig. 7a) show the greatest change in head at or near the water table over the course of a year. With increasing depth the change in head decreases from 2.5 m to less than 0.2 m. The magnitude of the gradient increases with depth, from values of -0.01 $m \cdot m^{-1}$ to -1.00 $m \cdot m^{-1}$ (negative values of gradient will indicate downward flow). The very low gradient between the uppermost piezometers indicate either stagnant or horizontal flow conditions near the water table. No additional piezometers are available to measure the horizontal gradients at the site; but the very high gradient below this zone, combined with the moderate decline in the water table and the lack of evidence of a discharge zone in the immediate vicinity of the piezometer nest, appear to rule out significant horizontal flows. Very similar hydraulic profiles exist at sites Fa 10 and Fa 95 (figs. 7b and 7f). The unifying factor at these sites is the geology. At least 30 m of low hydraulic conductivity, fine-grained materials extend from the ground surface to the top of the first aquifers.

Perched water tables can be found throughout the area. Evidence for perched water tables can be found at sites Fa 7, 38, 40, 76, 80, 86, 88, 91, and 103 (fig. 2). Profiles for two of these sites are shown in figures 7c and 7d. At site Fa 40 (fig. 7c) the hydraulic head of the second deepest piezometer is less than the elevation of the bottom of the screened interval of the piezometer, indicating a negative pressure head, or in other words, an unsaturated zone. The exact extent of the unsaturated zone, as well as the magnitude of the negative pressure head, is unknown. A pressure head value of -1.0 m of water has been used in these hydraulic profiles whenever unsaturated conditions have been

7a



7b

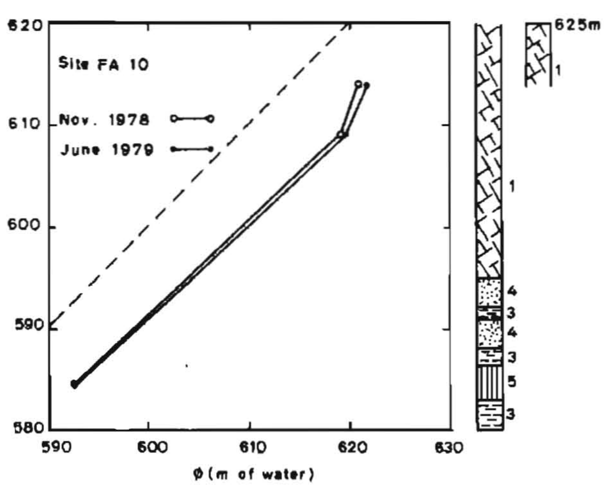
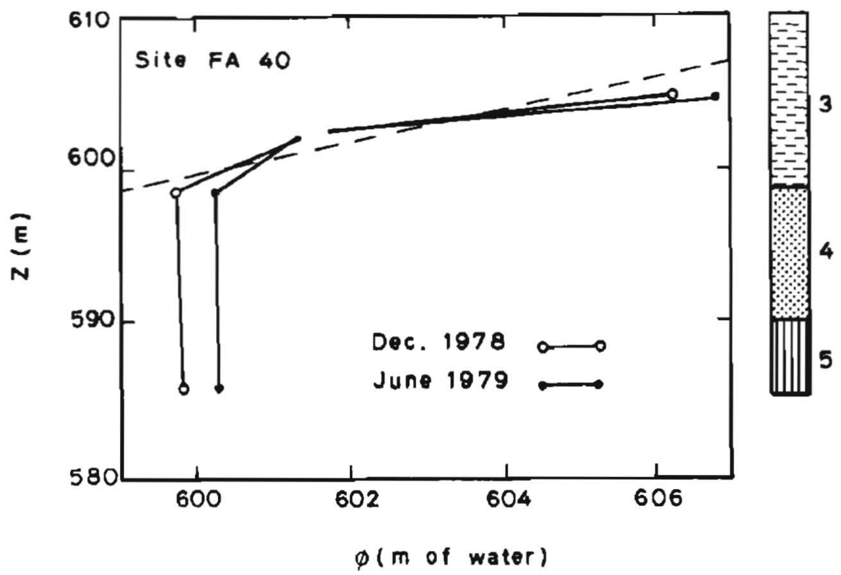
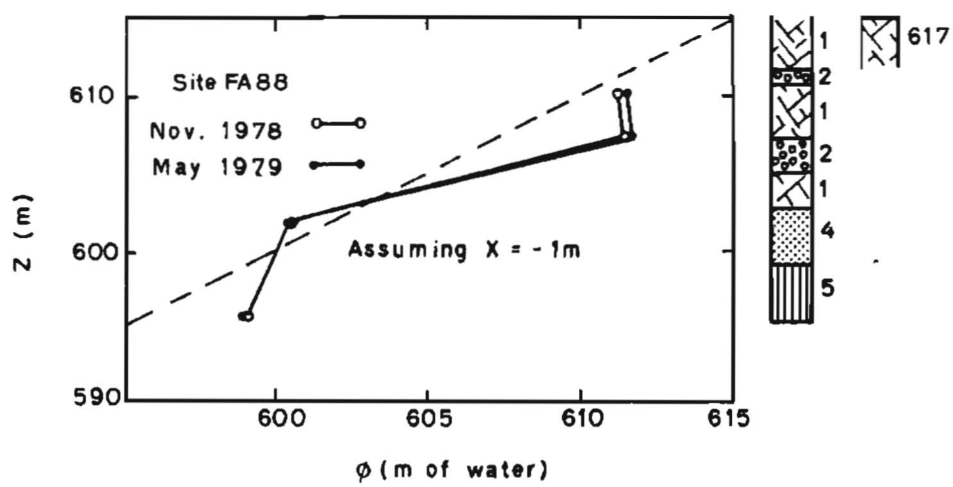


Figure 7. Measured hydraulic profiles through the saturated zone at eight piezometer nests in the Falkirk study area, where 1 is pebble-loam, 2 is sand and gravel, 3 is silty clay to clayey silt, 4 is silty fine to medium sand, and 5 is lignite. The eight graphs comprising this figure are on pages 13 through 17.

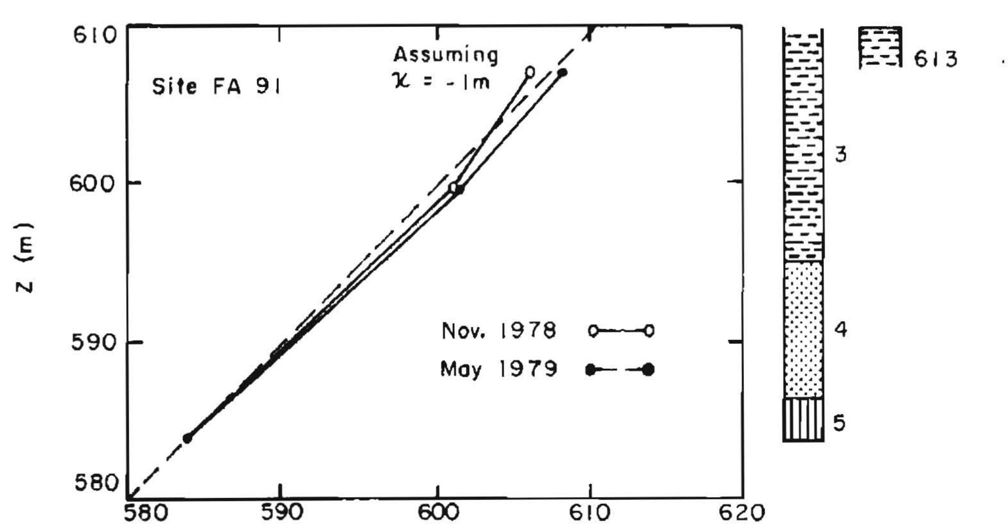
7c


610
Site FA 40
600
Z (m)
590
580
600
602
604
606
ϕ (m of water)
Dec. 1978
June 1979
3
4
5

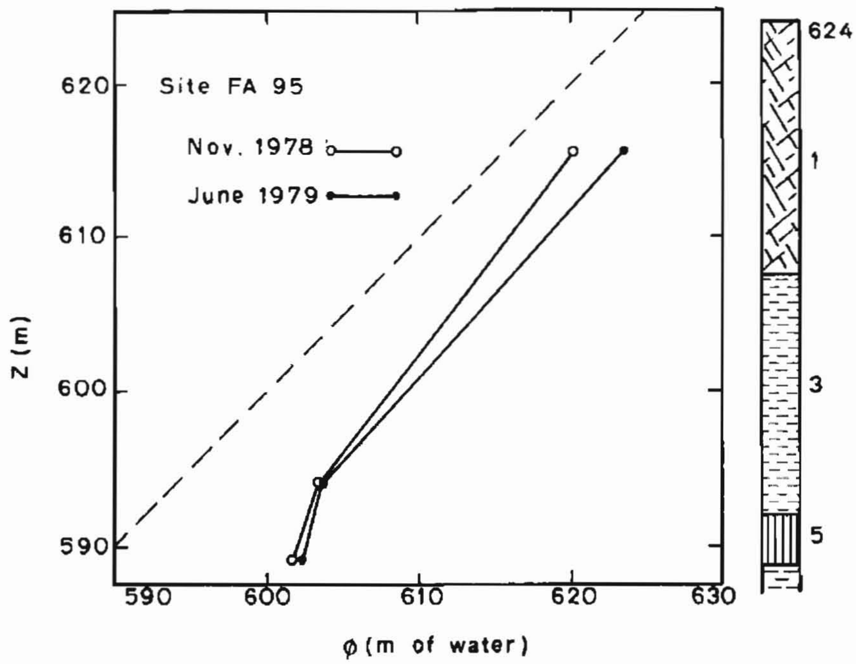
7d


Site FA88
Nov. 1978
May 1979
610
600
590
Z (m)
Assuming X = - 1m
600
605
610
615
ϕ (m of water)
1
2
1
2
1
4
5
617

7e

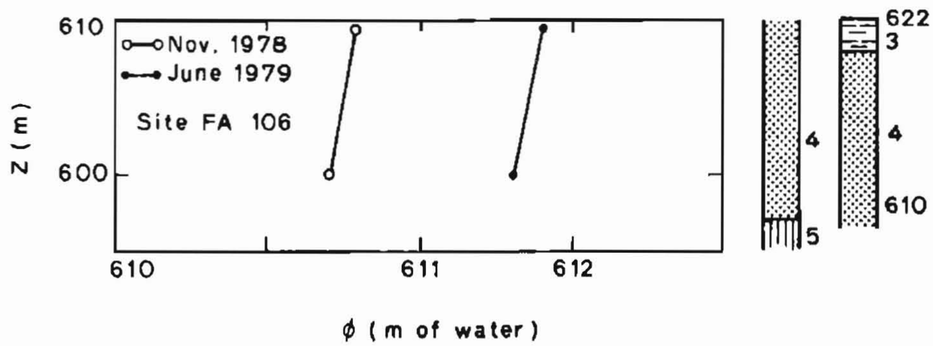

Site FA 91
Assuming
χ = -1m
Nov. 1978
May 1979
Z (m)
610
600
590
580
580
590
600
610
620
3
4
5
613

7f



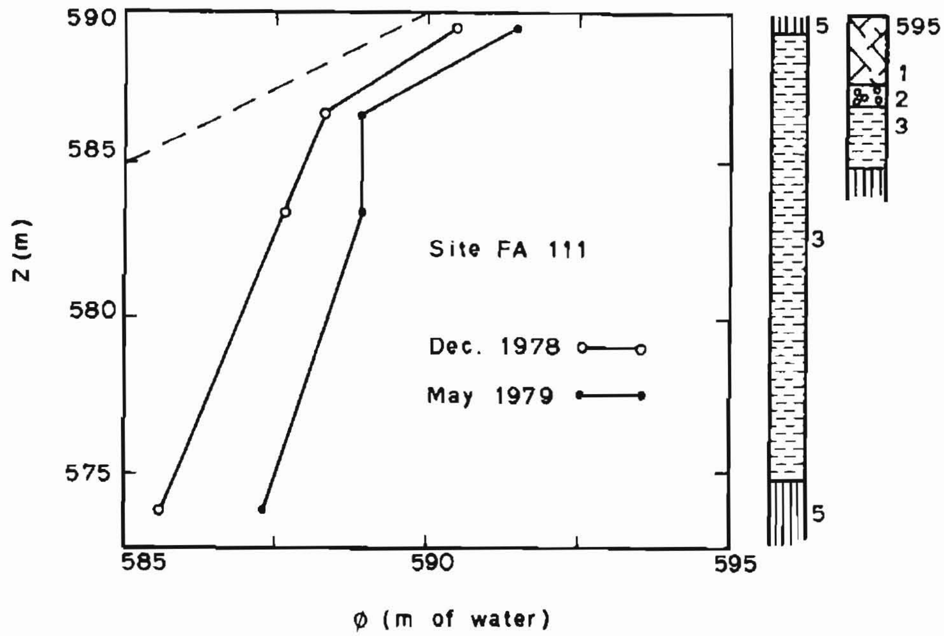
Site FA 95
Nov. 1978
June 1979
620
610
600
590
Z (m)
590
600
610
620
630
ϕ (m of water)
624
1
3
5

7g



Nov. 1978
June 1979
Site FA 106
610
600
Z (m)
610
611
612
ϕ (m of water)
4
5
622
3
4
610

7h



encountered by piezometers. If the hydraulic heads observed in the upper two piezometers were used without considering the possibility of an unsaturated zone the calculated gradient would be significantly less than -1.0 $m \cdot m^{-1}$.

It is unclear why a perched water table should exist at site Fa 40. Generally perched water tables are formed above lenses of low hydraulic conductivity materials surrounded by higher permeability materials. At site Fa 40 (fig. 7c) this sequence is reversed; the fine-grained, low permeability materials overlie sandy materials. A similar situation exists at sites Fa 38, Fa 80, and Fa 91 where pebble-loam overlies sand.

The shallowest observation wells at site Fa 88 (fig. 7d) again show very low vertical gradients, in both upward and downward directions, in the saturated zone below the water table. The lack of water in the 602 m elevation piezometer indicates a perched water table. It is probable that the perched water table includes the gravel layer between the elevations of 605 to 607 metres and that the unsaturated zone begins beneath the second till unit. At this site much of the perched water may move off laterally through the gravel which probably subcrops along the edges of a shallow valley 400 to 600 metres to the south and west of the site (fig. 2). Sites Fa 7, 76, 86, and 103 are similar to site Fa 88.

Site Fa 91 (fig. 7e) is typical of regions near the southern subcrop of the main coal bed in the study area. Quite commonly the coal bed becomes an unconfined aquifer and it is not unusual to find the coal bed completely unsaturated. The shallowest piezometer at site Fa 91 is dry for most of the year, but between April and July the pressure head at this point goes from negative values to +0.75 m of water and then back to unsaturated conditions. The middle piezometer is consistently saturated and the hydraulic head is relatively constant over the course of a year. The deepest piezometer at site Fa 91 indicates a pressure head of +0.5 to +1.0 m of water, resulting in a water table within the coal bed. It is probable that the sand above the coal is unsaturated. It is unknown at this time whether the saturated conditions displayed by the middle piezometer are a stable condition or whether the head will decline as the silt and clay drain into the underlying coal bed.

The geologic setting of site Fa 106 (fig. 7g) is typical for a small portion of the study site immediately south and east of the town of Underwood. In this area there is little or no pebble-loam or silt and clay covering the sandy textured bedrock. The hydraulic profile shows a small change in head between the highest and smallest recorded hydraulic head values, and a very small gradient on the order of -0.01 $m \cdot m^{-1}$. The very low gradient is the result of the high hydraulic conductivity of the sand (10^{-6} $m \cdot s^{-1}$) relative to the pebble-loam and silt and clay (10^{-8} $m \cdot s^{-1}$).

The final site illustrated, Fa 111, (fig. 7h) is anomalous in several respects. The greatest gradient is near the water table and decreases with depth. The greatest change in head occurs at the base of the profile rather than at the water table. Between the second and third piezometres there is actually an upward gradient, on the order of +0.10 $m \cdot m^{-1}$, during March, April, and May. The spring profile would seem to indicate that water flows toward the second piezometer from both above and below. The geology of the site would seem to make this flow pattern unlikely, for there is no evidence for a permeable zone which could transfer this water laterally out of the area.

Between April and October in North Dakota, the downward movement of water can be halted or reversed by evaporation from the ground surface and/or transpiration by vegetation. Annual free water surface evaporation is on the order of three times greater than the total annual precipitation and measured evapotranspiration is at least 1.5 times greater than precipitation. The magnitude of these numbers would appear to make the possibility of groundwater recharge unlikely. But recharge does occur and the major factor is the timing of the maximum evapotranspiration and precipitation rates.

The evaporation rate from a bare, non-vegetated soil surface is limited by the K-Ψ-θ relationships of the soil, the moisture content of the soil, the depth to the water table, and external climatic conditions such as temperature, wind speed, and humidity. Gardner and Fireman (1958) found that when the water table was within 1 m of the ground surface the evaporation rate is controlled by the climatic factors. When the water-table depth is greater than 1 m the evaporation rate is controlled by the soil properties and water-table depth.

Under constant climatic conditions the evaporation rate decreases rapidly with increasing water-table depth to a depth of 3 m. Below 3 m the rate of decline of the evaporation rate decreases. When Gardner and Fireman (1958) held the water table at a depth of 1 m, in clay and sandy loam soil columns, they measured evaporation rates of approximately 0.30 and 0.50 $cm \cdot day^{-1}$, respectively. When the depth to the water table was increased to 3 m the evaporation from both soils decreased to about 0.05 $cm \cdot day^{-1}$. With the water table at a depth of 6 m the evaporation rate only declined to approximately 0.02 $cm \cdot day^{-1}$. Shaw and Smith (1927), using a loam soil in a laboratory setting, found that evaporation rates were negligible when the water table was more than 3 m below the ground surface. Veihmeyer and Brooks (1954) obtained similar results from a field experiment using weighing lysimeters. With the water table at a depth of 0.5 m the average evaporation rate over the course of a year in southern California was 0.13 $cm \cdot day^{-1}$. When the water table was at a depth of 1.5 m the evaporation rate decreased to 0.03 $cm \cdot day^{-1}$.

The soil moisture content is an important control on the rate of evaporation from a soil surface (Green and others, 1972). Evaporation rates from a moist soil are approximately equal to the rate of evaporation from a free water surface under similar climatic conditions. As the water content decreases the remaining soil water is found in smaller and smaller pores. This decreases the evaporating surface area of the water and increases the capillary tension on the water. As a result of these processes the evaporation rate decreases. As drying continues, the evaporative surface begins to recede into soil and the continuous film of water surrounding soil grains is broken. At this point evaporation rates decrease further and moisture transfer is controlled by vapor diffusion within soil pores.

Transpiration of soil moisture by vegetation can also prevent infiltrating water from reaching the water table. Generally plants are assumed to be able to remove all water held at pressure heads greater than -1.5×10^6 Pa or approximately -150 m of water, the wilting point. At these pressure heads the water content of the soil within the root zone, the interval of greatest root development, is very low. The low water content results in low hydraulic conductivity values; therefore, the rate of movement of any infiltrating water is very small. As long as the growing season continues this slow moving water is also subject to transpiration and little infiltrating water passes through the root zone. Significant recharge in the summer months probably occurs only when very high intensity rains produced by local thunderstorms result in high enough infiltration rates to quickly raise the soil moisture content to the point where water can quickly pass through the root zone.

Transpiration rates decrease with the harvesting of crops, cutting of hay, and the first killing frost. The falling autumn temperatures also decrease the free water surface evaporation rate. As a result, the rain that falls between September and November, before the precipitation is trapped as snow, is also available for infiltration. This is illustrated by the 1977 hydrograph for observation well Fa 9-3 (fig. 64, Williams, Koob, and Rechard, in prep.) when the water table rose approximately 0.60 m in response to 0.10 m of precipitation during September. During the previous four months the water table did not respond to several rainfall events which totaled approximately 0.30 m of water.

Once the infiltrating water is below the root zone, away from the effects of transpiration, and 2 to 3 metres beneath a bare ground surface where evaporation rates are negligible, the water cannot be returned directly to the atmosphere. Limited data from the Falkirk site indicate that evapotranspiration below depths of 1.5 to 2.0 metres is insignificant (Pole, 1979). Once below the zone of evapotranspiration the movement of the infiltrating water depends upon the interrelationships of the particular soil's characteristic curves; as has been described previously. If the water table is relatively shallow the infiltrating water will quickly add to groundwater recharge. But, in the Northern Great Plains the water table is often several tens of metres below the ground surface. In these areas a single infiltration event is not sufficient to reach the water table. This water redistributes itself within the unsaturated zone above the water table, and downward movement will not resume until the next major infiltration event occurs and the water content of the unsaturated materials increases to a point at which downward movement can resume.

The third major mechanism which can affect the movement of water through the unsaturated zone is the formation of frost during the winter months. Several investigators have shown that water can move upward through the unsaturated zone under the influence of vapor pressure and temperature gradients. The degree to which a frozen soil profile can prevent or restrict infiltration has also been explored to a lesser extent. The following discussion summarizes the results presented by Willis and others (1961) and Willis and others (1964) working in western North Dakota, Benz and others (1968) in eastern North Dakota, Schneider (1961) in southern Minnesota, Sartz (1969) in Wisconsin, and Freeze and Banner (1970) in central Saskatchewan.

In the Northern Great Plains the ground freezes from the surface downward beginning in November. The downward migration of the frost continues until March or April when maximum depths of 1.5 to 2.5 metres are reached. The depth of freezing is strongly dependent on the nature of ground cover. Benz and others (1968) showed that bare soil surfaces result in freezing to a depth of 2.7 m. A straw mulch reduced the depth of freezing to 1.7 m.

The actual transfer of water within soil profiles has been documented by monitoring moisture content changes with neutron probes. Three studies (Willis and others, 1964, Benz and others, 1968, and Freeze and Banner, 1970) found water content increases in the frost zone with concurrent decreases in moisture content below the frost zone. In all three studies the water table was within 2.0 to 3.5 metres of the ground surface and fell as the depth of frost penetration increased. In the Freeze and Banner (1970) study the site was known to be located within a recharge area. The water table would have declined regardless of whether water migrated upward to the frost zone because the precipitation, in the form of snow, could not infiltrate to the water table. It is, therefore, unclear to what degree upward migration of water lowered the water table. The site investigated by Benz and others (1968) was within a discharge area; therefore, the observed water table declines of 1.5 to 2.0 metres probably resulted from water movement into the frost zone. It is not known whether the site studied by Willis and others (1964) was in a recharge or discharge area. In another North Dakota study there was little accumulation of water in the unsaturated zone (Willis and others, 1961).

Water that migrates upward into the unsaturated zone to form frost is not necessarily lost as potential groundwater recharge. As frost thaws the water once again moves downward toward the water table. In the spring the frost begins to melt both from the ground surface and the base of the frost zone, with the greater thaw rate being from the ground surface. Because of this difference in thaw rates, the last remnants of frost are commonly found in the deepest third of the frozen profile. The frost begins to thaw when air and soil temperatures rise above 0°C. Once the frost begins to thaw, melting is complete within 30 to 50 days. Coincident with, or within several days of, the start of above-freezing air temperatures and frost melting, the water-table hydraulic head also begins to rise (Freeze and Banner, 1970, Benz and others, 1968, Willis and others, 1963, Schneider, 1961 and figures 8a to 8gg).

The water-table rise precedes the general thaw of the frost profile but corresponds to the start of the spring thaw, implying that melting frost and snowmelt actually pass through the frozen soil. This implies that frozen soil is not necessarily impermeable. Storey (1955) describes four types of frozen soil structure: concrete, honeycomb, stalactite, and granular. Concrete frost consists of many thin lenses of ice and small ice crystals. It is generally very dense. Honeycomb frost has a loose porous structure which is very friable. Stalactite frost is often formed from the refreezing of a thawed honeycomb structure. It is made up of many small icicles that consist of tiny columns. The icicles are often fused into sheets or loosely bound blocks. The final frost type, granular frost, has a loose arrangement of small granules of ice scattered through the soil. Storey states that the concrete frost is relatively impermeable, but the other types do little to

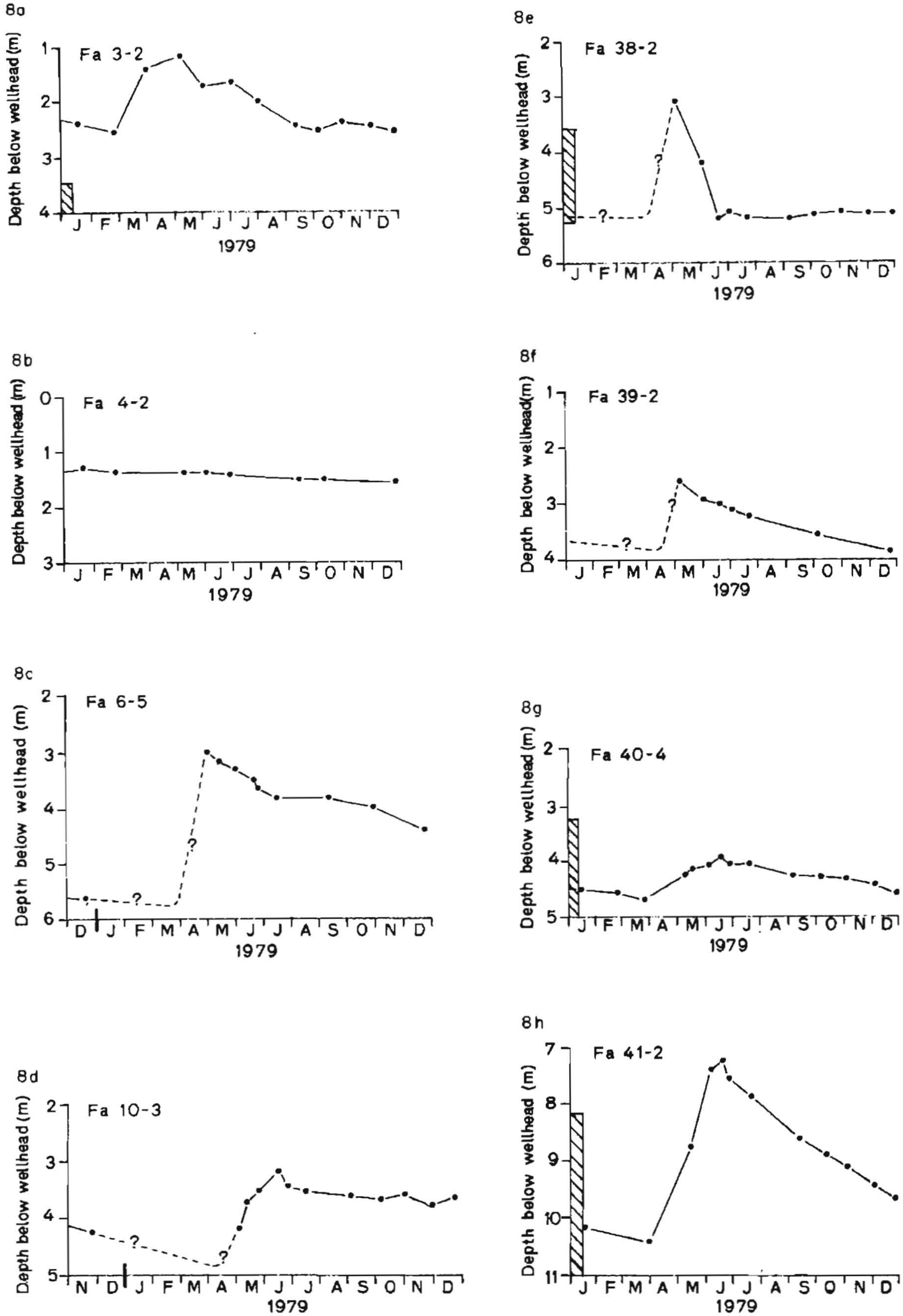
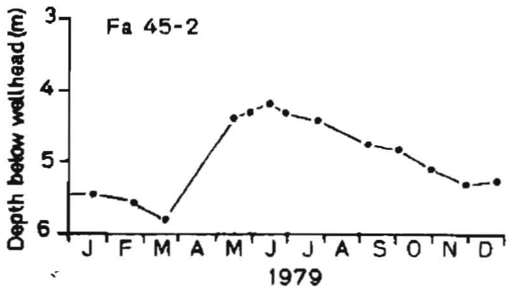


Figure 8. Hydrographs from water-table observation wells and shallow piezometers. The thirty-three graphs comprising this figure are on pages 22 through 25.

8i



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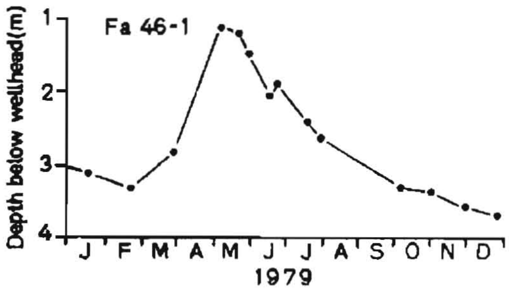

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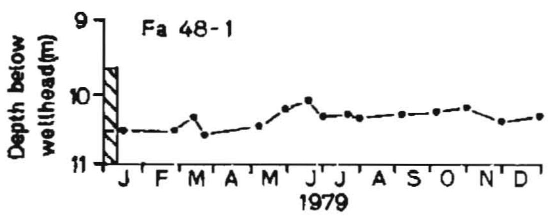

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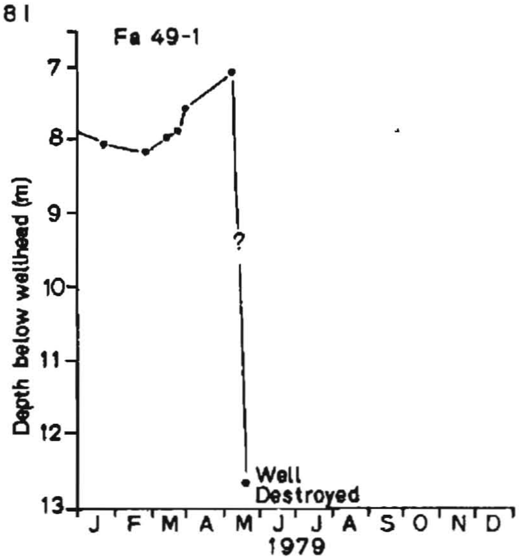

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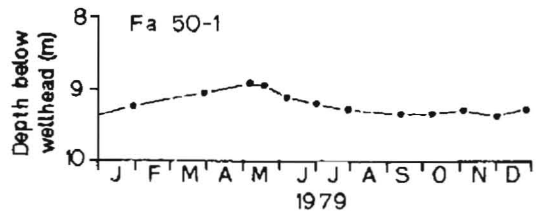

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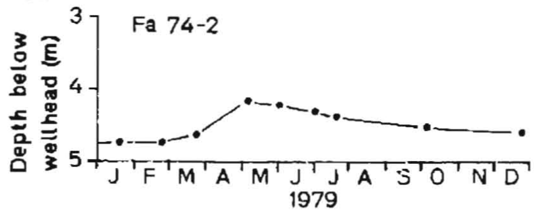

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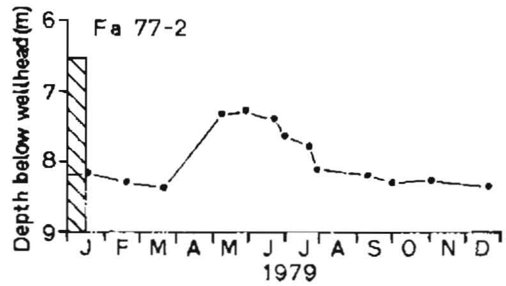

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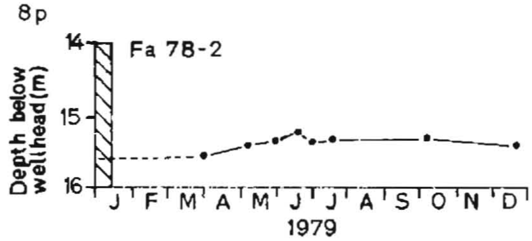

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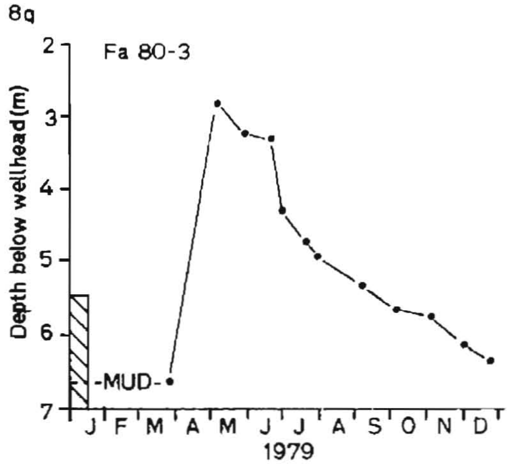
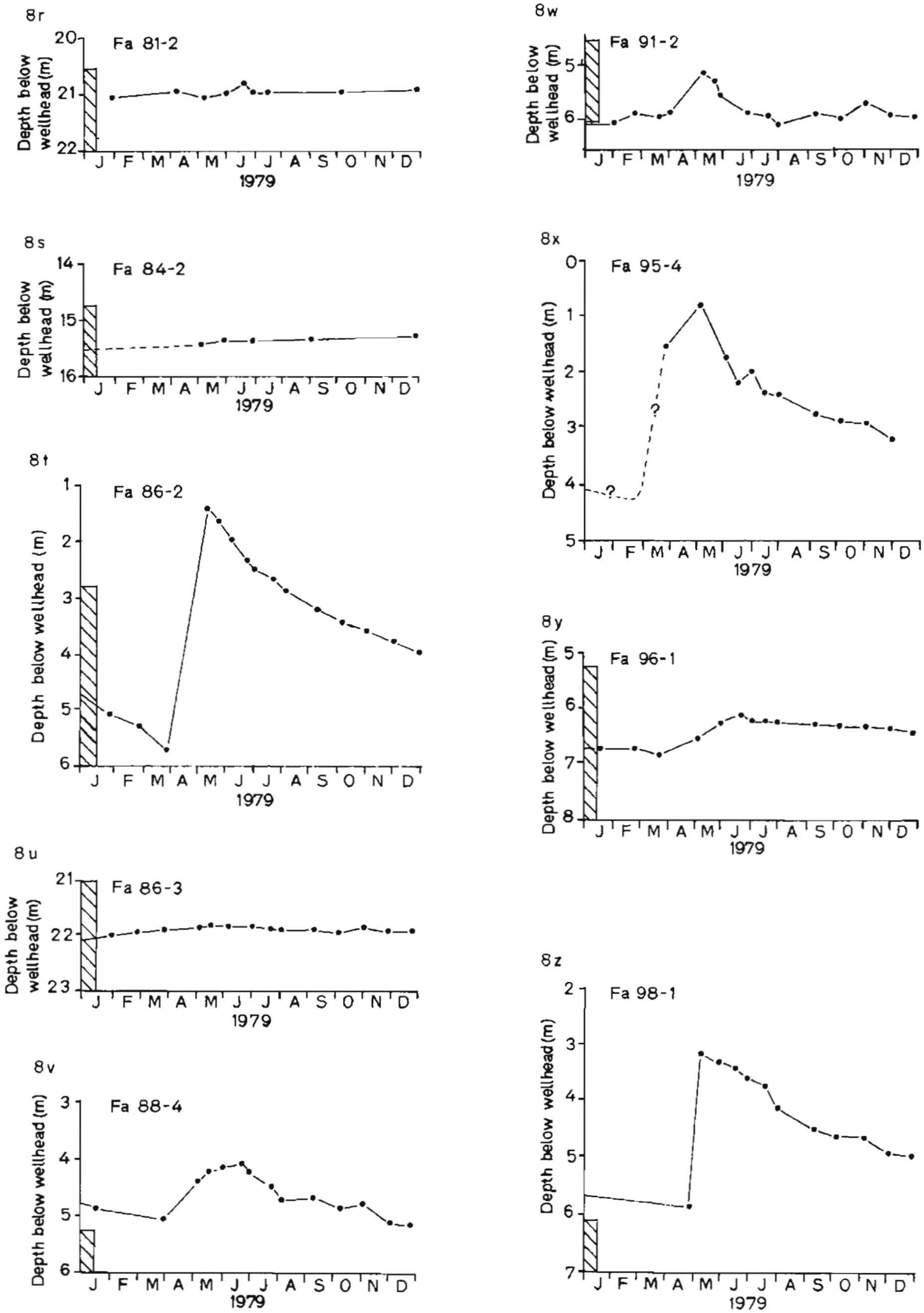
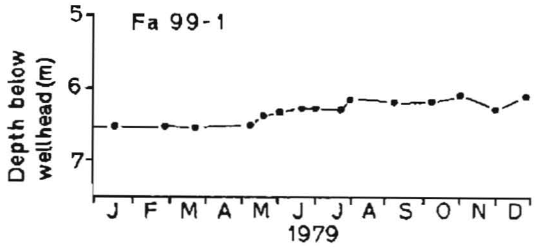



8r
Fa 81-2
Depth below wellhead (m)
1979
8s
Fa 84-2
8t
Fa 86-2
8u
Fa 86-3
8v
Fa 88-4
8w
Fa 91-2
8x
Fa 95-4
8y
Fa 96-1
8z
Fa 98-1

8aa



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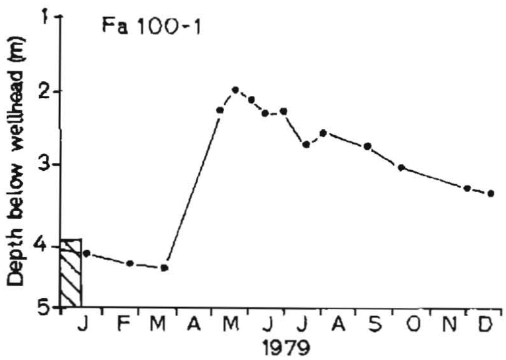

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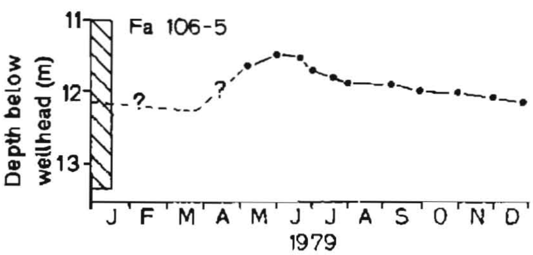

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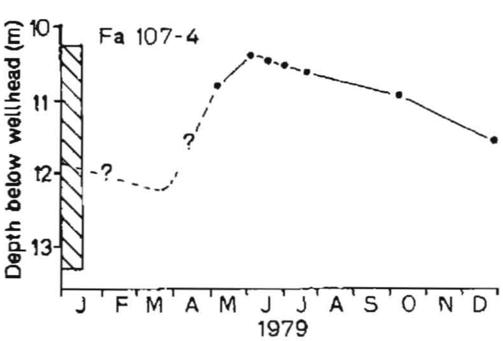

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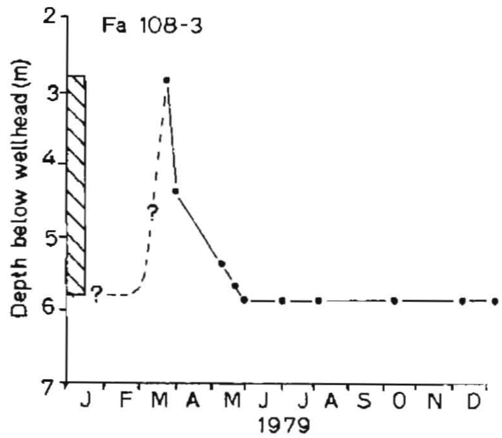

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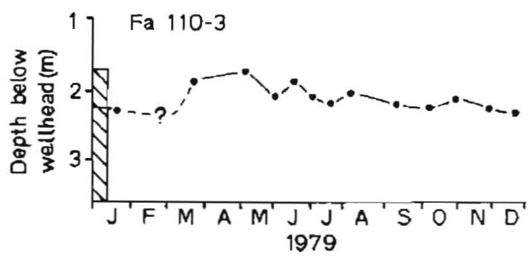

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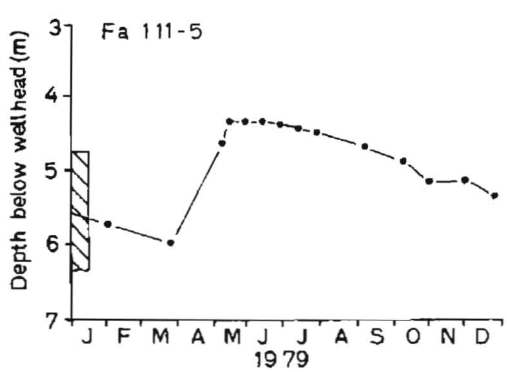


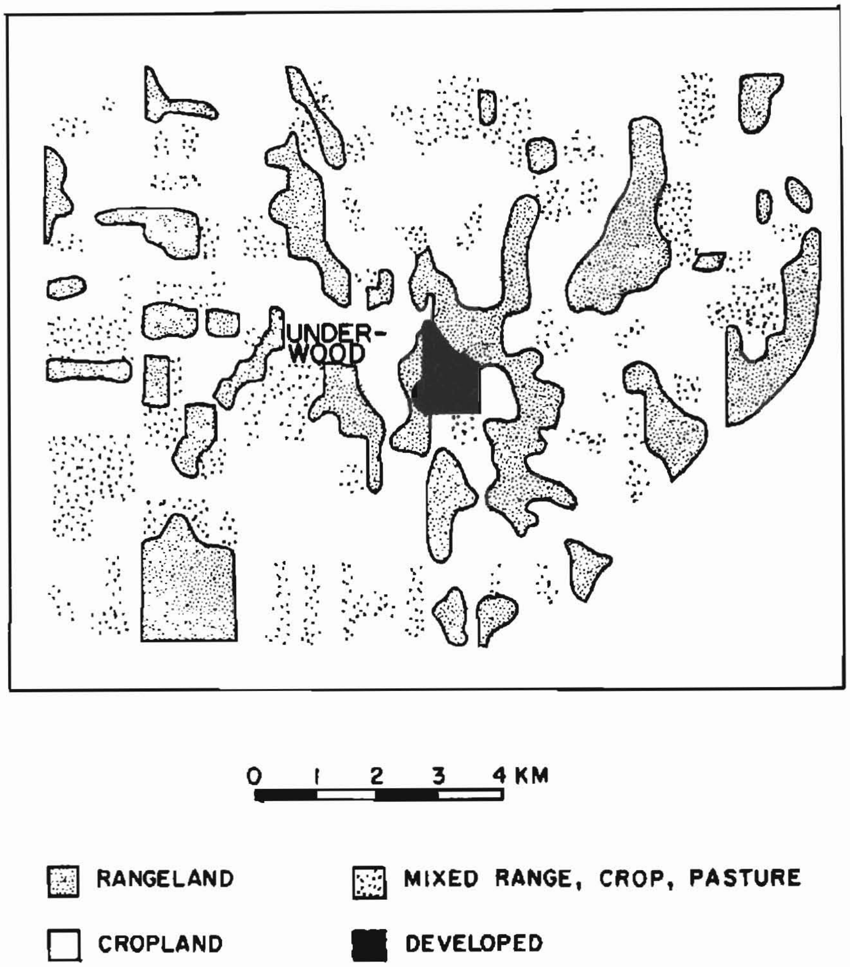


Figure 9. Land-use within the Falkirk study area.

impede the flow of water through a soil profile. In the northeastern United States, Storey (1955) observed concrete frost to occur most frequently under cultivated fields. The honeycomb and stalactite types were found most often under meadows and pastures and usually occurred during shallow freezing. In forested areas the granular frost type predominated. Observations of frost types have not been made in North Dakota. Therefore, it is unknown as to how well these frost types describe what develops during the harsher winters of the Northern Great Plains. Twenty to forty percent of the Falkirk study area is covered by some type of grassland (fig. 9). If the more permeable frost types do occur under grasslands there could be considerable opportunity of infiltration during the spring snowmelt period.

Willis and others (1964) described the following sequence of events for a spring thaw in western North Dakota. The frozen soil began to thaw by the end of the first week in April. Moisture above "field capacity" within the upper tens of centimetres may have flowed downward to the 60 to 90 centimetre depth and refroze. During the third week in April the moisture content decreased in the 60 to 90 centimetre zone and the water table started to rise; even though the frost did not completely melt until the second week of May.

Sartz (1969) also found frozen

soil to be permeable. He describes frost as being of the concrete type, but on numerous occasions during winters between 1961 and 1968 snowmelt and/or rainfall infiltrated freely through the frozen soil profiles. He also found that the amount of infiltration decreased as the frost season progressed. He assumed that the permeability of the frozen soil profile decreased as the infiltrating water formed ice crystals in the soil. During the thawing period the process of frost softening would have been accelerated by the relatively warm infiltrating water.

The preceding discussion indicates that the presence of frozen soil does not preclude infiltration and potential groundwater recharge. The major effect of cold temperatures is to tie up potential infiltration water in the immobile form of snow and ice. Once air temperatures climb above the freezing mark the snow begins to melt, the ground begins to thaw and infiltration begins. The possibility for infiltration is increased by the behavior of the snowmelt, which has been observed at the Falkirk area to percolate through the snowpack and then flow laterally over the surface of the soil. Once in contact with the soil surface it is possible for the runoff to infiltrate into the soil and eventually reach the water table. Since evapotranspiration rates are very low at this time of year the downward migration of the water will depend only on the soil properties and moisture content of the soil.

MAGNITUDE OF GROUNDWATER RECHARGE

Groundwater recharge begins with the movement of water through the unsaturated zone. Unfortunately, data on the magnitude of the infiltration rate through the unsaturated zone in the study area are almost nonexistent. Soil moisture data from several plots of disturbed and undisturbed materials from the Falkirk study area were collected by the Soils Department of NDSU (Pole, personal commun.). In plots consisting of machine compacted silt-clay-loam and undisturbed pebble-loam, approximately 0.17 m of water apparently passed below a depth of 3.66 m in a two-month period. The reliability and significance of this value is difficult to assess because the bulk of the data is from an area that has been manipulated to simulate current reclamation procedures.

Groundwater recharge does not occur until the saturated zone is reached. Approximately 200 observation wells are installed singularly or in nests over a 90 km^2 area. Thirty-three of the observation wells either intersect the water table or are screened within 2 m below the water table. Six piezometer nests are sufficiently instrumented in the shallow groundwater zone to enable the evaluation of vertical groundwater fluxes. The piezometers and water-table wells were monitored extensively during 1979 only; the calculated recharge rates, therefore, apply only to this year. In drier or wetter years, or in years when the temporal distribution of precipitation is different from that of 1979, the recharge rate at a specific site will also be different from that recorded for 1979.

An observation well in which the screened interval intersects, or is immediately below the water table, can be used to determine potential ground-water recharge (as defined previously) by monitoring changes in the hydraulic head of the water table. If the water table is at least 3 m below the ground surface, the effects of evapotranspiration and frost formation on water-table fluctuations will be negligible, and the changes in hydraulic head can be considered to indicate actual ground-water recharge. Approximately 80 percent of the water-table observation wells in the Falkirk site indicate the maximum water-table head to be greater than 3 m below the ground surface. The effects of evapotranspiration and frost can therefore be disregarded when evaluating the observation well hydrographs.

Hydrographs for the water-table observation wells for 1979 are plotted in figures 9a through 9gg. Using the procedures outlined earlier, the recharge indicated by the hydrographs is summarized in table 1. The critical parameter in calculating groundwater recharge by this method is the effective specific yield of the porous medium. Since only a limited portion of a porous medium consists of voids, and a portion of that void space is filled with water, the change in hydraulic head must be multiplied by the specific yield to determine the actual volume of water added to the porous medium. For sandy Tertiary sediment materials, the porosity has been determined to be approximately 0.40 (Williams, Koob, and Rechard, in prep.). The specific yield can be calculated from water retention--soil moisture tension relationships (Williams, Koob, and Rechard, in prep.) if it is assumed that the moisture released at a tension of 30 K Pa is approximately equal to the specific yield. The average of 19 calculations is 0.16 with a standard deviation of 0.06. This value is similar to the representative specific yield values of 0.23 for fine sand, 0.18 for loess and 0.08 for silt (Johnson, 1967). The effective specific yield or porosity of the fine-grained pebble-loam and Tertiary silty clay is more difficult to define.

The specific yield has not been determined for the fine-textured material in the study area. Representative specific yield values for pebble-silty loam to pebble-sandy loam are on the order of 0.06 to 0.16, respectively (Johnson, 1967). Clay, silt, and siltstone range from 0.03 to 0.12 (Johnson, 1967). The measured porosity of 7 samples of silty loam to clay collected from the study area averaged 45 percent (Williams, Koob, and Rechard, in prep.). This is within the same range as the representative porosity values presented by Morris and Johnson (1967) for silt (0.46) and clay (0.42). No samples of pebble-loam (till) were analyzed. Representative porosity values of pebble-loam fall between 0.30 and 0.35. Fracturing of the fine-textured material further complicates an evaluation of the effective specific yield value that should be used in the groundwater recharge calculations. The pebble-loam has been observed by the authors to be fractured within the study area and at many other locations (Williams and Farvolden, 1967, Grisak and Cherry, 1975, and Grisak and others, 1975). The fine-grained Tertiary materials have also been observed by the authors to be fractured, but to a lesser extent than the pebble-loam. Grisak (1975) has determined the fracture porosity of pebble-loam to be about 0.0002 in southeastern Manitoba. If both the intrinsic permeability and fracture spacing of a porous medium are known, the effective fracture porosity can be estimated from equations developed by Snow (1968). The permeability of the pebble-loam has been measured with single-well response tests (Hvorslev, 1951). Grisak and others found the spacings of vertical fractures in pebble-loam in the Northern Plains to be generally between 0.02 and 0.30 metres. Using this range of spacings and an average, measured, in situ permeability of $10^{-15} \cdot m^2$ (hydraulic conductivity of 10^{-8} $m \cdot s^{-1}$) yields porosity values that range from 0.0001 to 0.0008. The effective porosity was determined for two large, in situ blocks of pebble-loam in southern Sweden by Nordberg and Modig (1974). In this field experiment, the total porosity of the blocks was in the range of 18 to 32 percent, with the effective porosity calculated to be 0.074 and 0.034 for the two test blocks.

A knowledge of fracture porosity in the fine-textured material does not completely solve all the problems of determining the recharge because the pebble-loam and silty clay are dual porosity media. As the infiltrating water flows into the more permeable fractures, the hydraulic head in the fractures will increase at a greater rate than the head in the unfractured matrix. Since this matrix is not impermeable, water will flow from the fractures into the matrix. If the matrix is unsaturated, a great deal of water can flow into the matrix, rather than directly becoming groundwater recharge. Once infiltration ceases, the water will flow out through the fractures faster than through the matrix, resulting in a head reversal. Water will then flow from the matrix into the fracture. To what extent this process would affect the determination of groundwater recharge rates is not known. All that can be said is that the potential recharge calculations in fractured media are rough estimates.

A specific yield of 0.16 for the sandy material and a fracture porosity of 0.001 for fine-textured material were chosen to estimate groundwater recharge. Based on the data presented by Grisak and others (1975) and Nordberg and Modig (1974) the value of 0.001 is, at best, an order of magnitude estimate of the effective porosity. Five of the observation wells, Fa 46-1, 48-1, 86-3, 88-4, and 108-3, intersect thin layers of sand, gravel, or lignite. These wells will be omitted from the following discussion because the observed water levels may be the result of the confined pressure head of the permeable layer, rather than the actual water-table head. Another four wells, Fa 45-2, 77-2, 98-1, and 100-1, will be omitted because of the possibility that the observed water level increase was, at least partially, due to the lateral inflow of water from nearby sloughs. The observed water level changes ranged from 3.73 to 0.37 metres (table 2). During 1979, 18 of the remaining wells showed an average net increase in water-table elevation of 0.35 m and 2 wells showed no net change in hydraulic head. The remaining 5 wells had an average net water-table decline of 0.27 m.

The yearly average potential recharge, from table 2, in the sandy material is 0.17 $m^3 \cdot yr^{-1} \cdot m^{-2}$ with a range of 0.51 to 0.06 $m^3 \cdot yr^{-1} \cdot m^{-2}$. Much of the variation is due to the topographic setting of the piezometer sites and will be discussed in the following section. When a porosity of 0.001 is used to calculate recharge rates through fine-grained materials, the values average 0.0018 $m^3 \cdot yr^{-1} \cdot m^{-2}$ with a range of 0.0006 to 0.0043 $m^3 \cdot yr^{-1} \cdot m^{-2}$. Using a specific yield of 0.10 results in a hundredfold increase in the average annual recharge rate to 0.18 $m^3 \cdot yr^{-1} \cdot m^{-2}$ with a range of 0.06 to 0.43 $m^3 \cdot yr^{-1} \cdot m^{-2}$. The lower values on the order of 0.001 to 0.05 $m^3 \cdot yr^{-1} \cdot m^{-2}$ are probably the more realistic of the calculated recharge rates.

Direct measurements of recharge rates, using Darcy's law, were carried out at six piezometer nests within the Falkirk study area. Hydraulic heads were measured periodically during the course of 1979 and converted to vertical gradients between pairs of piezometers. Unfortunately, the piezometer nests were often inaccessible or buried beneath snow drifts during the winter months and head measurements could not be made. The gradient, when multiplied by the measured hydraulic conductivity at the sites, yields a downward flux of water in the saturated zone, the true groundwater recharge rate. The hydraulic conductivity of each piezometer was determined by slug testing (Hvorslev, 1951). The lower of the hydraulic conductivity values from the two piezometers at a site was used in the flux calculations. The vertical flux rates of the six sites monitored during 1979 are plotted in figure 10a through 10f. The majority of the sites had higher flux rates during the spring, but the variation in rates during the year is generally small, less than a factor of two.

The bar graphs plotted in figure 10 were derived by first plotting the flux at the time of the head measurement. The time interval between consecutive measurements was divided in half, and the measured flux was assumed to be constant between the half intervals before and after the measurement. When no data were available, the last measured flux rate (usually in November or December) was extrapolated to early April, when the head of the shallowest wells started to rise. The recharge rates were then integrated over the course of the year to yield the average yearly recharge rates listed in table 3. Recharge rates

TABLE 2. Groundwater recharge rates derived from water-table hydrographs.

Well Number	Lithology	H (m)	Recharge ($m^3 \cdot yr^{-1} \cdot m^{-2}$) Specific Yield 0.16	0.001	
6-5	pebble-loam	2.60		0.0026	
10-3	pebble-loam	1.00		0.0010	
38-2	pebble-loam	2.03		0.0020	
39-2	sand	0.93	0.15		
40-4	clay	0.70		0.0007	
41-2	sand	3.18	0.51		
45-2	pebble-loam	1.58		0.0016	
46-1	silt & clay	2.15		0.0022	*
48-1	silt & clay	0.56		0.0006	*
50-1	pebble-loam	0.68		0.0007	
74-2	silt & clay	0.56		0.0006	
77-2	pebble-loam	1.12		0.0011	
78-2	sand	0.37	0.06		
80-3	pebble-loam	3.73		0.0037	
81-2	sand	0.38	0.06		
86-2	pebble-loam	4.28		0.0043	*
88-4	pebble-loam	1.01		0.0010	*
91-2	pebble-loam	1.24		0.0012	
95-4	pebble-loam	3.32		0.0033	
96-1	pebble-loam	0.68		0.0007	*
98-1	sand	2.66	0.43		
100-1	pebble-loam	2.46		0.0025	
106-5	sand	0.80	0.13		
108-3	pebble-loam	2.98		0.0030	*
110-3	sand	0.82	0.13		

*Observed water level changes may be due to confined pressure heads.
Observed water level changes may correspond to lateral inflow of water from sloughs.

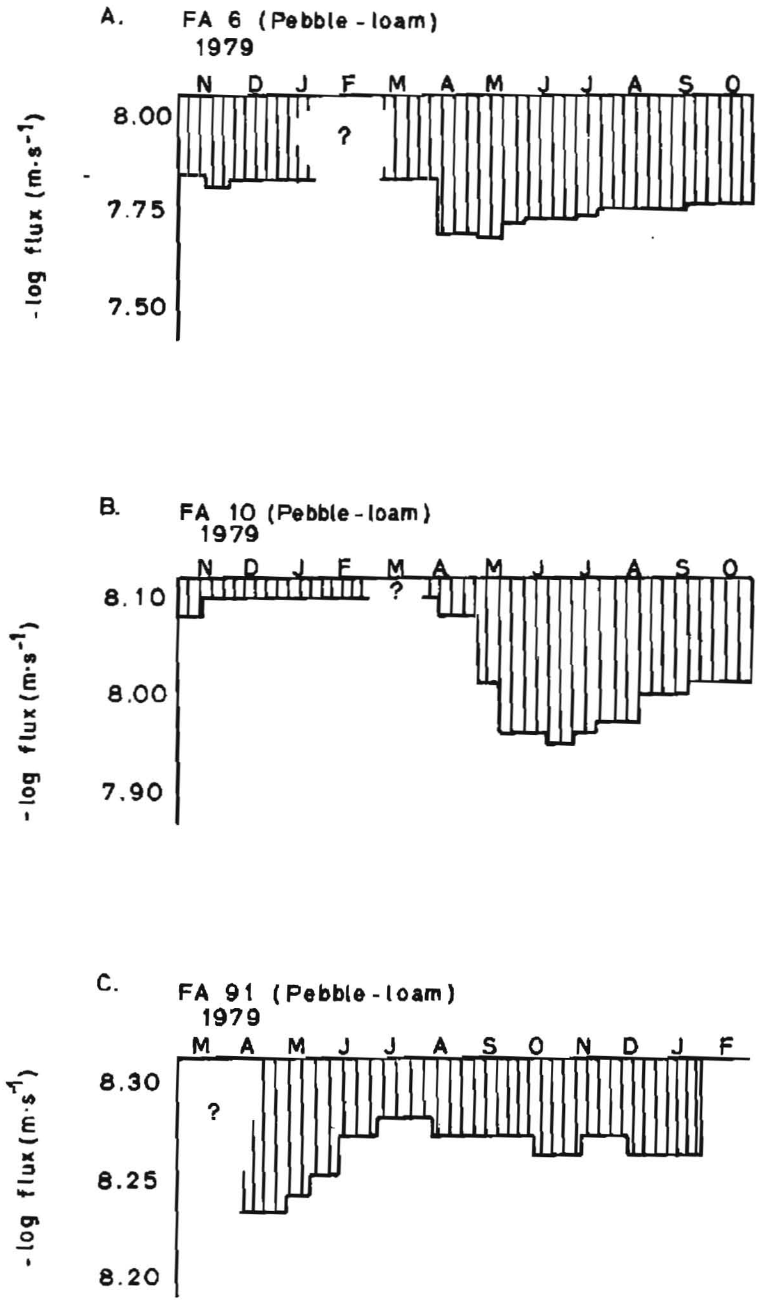
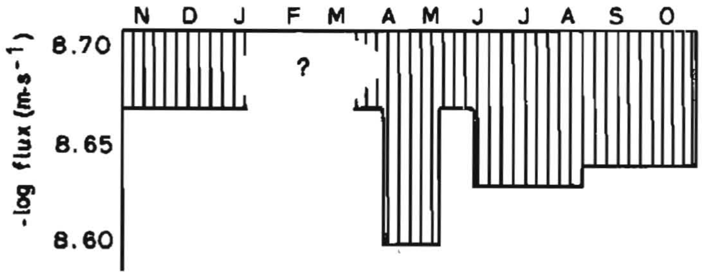
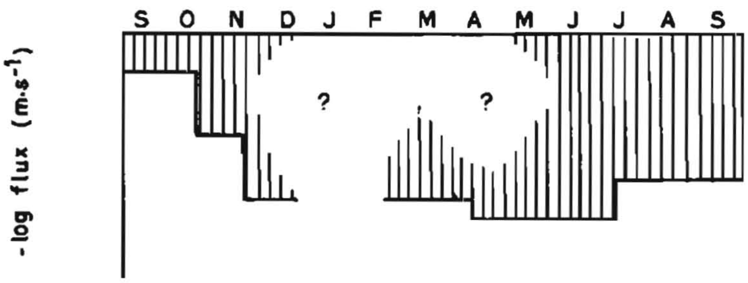


Figure 10. Vertical flux of water in the saturated zone at six piezometer nests in the Falkirk study area. The six graphs comprising this figure are on pages 32 and 33.

D. FA 95 (Pebble-loam/bdrk silt & clay)



E. FA 106 (bdrk sand)



F. FA 111 (bdrk silt & clay)

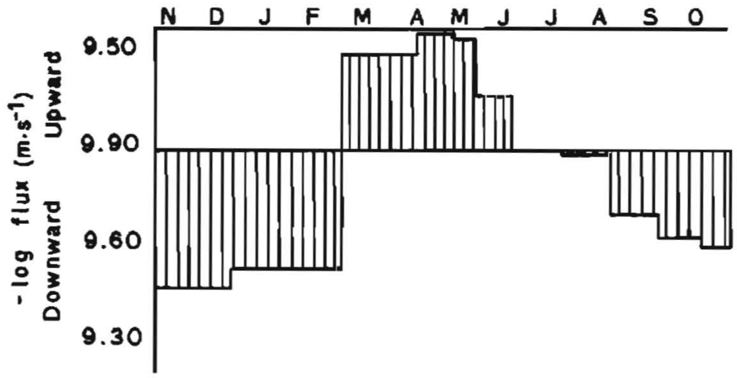


TABLE 3. Groundwater recharge rates derived from vertical flux calculations.

Site	Hydraulic Conductivity ($m \cdot s^{-1}$)	Range of Gradient ($m \cdot m^{-1}$)	Average Yearly Recharge ($m^3 \cdot yr^{-1} \cdot m^{-2}$)
Fa 6	2.2×10^{-8}	-0.74 to -1.06	0.603
Fa 9	3.2×10^{-8}	-0.58	0.587
Fa 10	7.6×10^{-9}	-0.31 to -0.44	0.086
Fa 91	7.1×10^{-9}	-0.76 to -0.84	0.115
Fa 95	2.8×10^{-9}	-0.78 to -0.92	0.072
Fa 106	8.0×10^{-6}	-0.006 to -0.011	0.711
Fa 111	2.6×10^{-9}	-0.18 to +0.15	0.006

from site Fa 9 are also included from Williams, Koob, and Rechard (in prep.).

Site Fa 111 (fig. 10f) is the only site in the study area that shows both upward and downward gradients during the course of the year. It is possible that the deeper of the pair of piezometers used to calculate the gradient intersects a series of fractures that extend to the ground surface or the shallower, saturated coal bed (fig. 9h), while the shallower piezometer does not intersect any fractures. During the spring the hydraulic head in the fractures could increase at a greater rate than in the unfractured material, resulting in an apparent gradient reversal. This phenomenon has also been observed and analyzed in more detail by Williams and Farvolden (1965).

The largest downward flux, 0.711 $m^3 \cdot yr^{-1} \cdot m^{-2}$, was found at site Fa 106 (table 3) where all but the uppermost 1.5 m of the unsaturated and shallow saturated zones consists of sandy bedrock. This site is in a roadside ditch, but no ponding of precipitation or snowmelt has ever been observed.

The recharge rate indicated by the water-table rise in piezometer 106-5 is only 0.13 $m^3 \cdot yr^{-1} \cdot m^{-2}$ (table 2). Of these two numbers, the higher value may be more accurate because in a sandy, permeable medium the ability of the medium to move water away from the water table, through the saturated zone, may approach or even possibly exceed the rate at which water is transmitted through the unsaturated zone to the water table.

Piezometer nests Fa 6, 10, 91, and 95 are indicative of vertical fluxes through pebble-loam and the fine-textured bedrock. The rate calculated for site Fa 6 is nearly one order of magnitude greater than the other three sites. There is no reason to believe that the value is due to faulty piezometer installation. The water-table elevation at site Fa 6 is much higher than 0.8 km to the east and south or 1.6 km to the west. The hydraulic conductivity of the pebble-loam is also 3 to 10 times greater than at the other three sites. Thus, the flux determined for site Fa 6 is apparently correct, but does not appear to be representative of the study area as a whole. In the Falkirk area, an average value of 0.091 $m^3 \cdot yr^{-1} \cdot m^{-2}$ is probably more representative of vertical flux through fine-grained materials where water tables are relatively shallow in the study area. The average vertical flux rate is 50 times larger than the average recharge rate calculated from water-table charges when a fracture porosity of 0.001 is used, but only half of that is calculated when a specific yield of 0.10 is used. The difference between these values is the result of the uncertainty in the magnitude of the parameters used to calculate the recharge rates. The largest uncertainty in the vertical flux calculations lies in the value of hydraulic conductivity while the uncertainty in the water-table response evaluations is in the effective specific yield or porosity values used in the calculation. Site Fa 9 was within, or on the edge of, a slough throughout 1977; therefore, the vertical flux of 0.587 $m^3 \cdot yr^{-1} \cdot m^{-2}$ is representative of areas where sloughs are situated over fine-grained materials.

AREAL DISTRIBUTION OF RECHARGE

Two major site variables are important to the areal distribution of groundwater recharge. The first concerns the effects of topographic variability. The second concerns the lithologic variability within the study area. The first variable is most evident in areas of rolling topography where drainage systems are generally non-integrated. On hilltops and slopes, precipitation and snowmelt flow overland to be concentrated in depressions. Depressions, and associated catchment areas, can vary greatly in size, from 1 m^2 hollows in the soil surface that pond to depths of several centimetres to 100 m^2 ephemeral sloughs with ponding depths of several decimetres to a metre. In the ponded areas infiltration will occur, with the magnitude of the infiltration depending on the characteristic curves of the soil, depth to the water table, and the length of ponding. Depending on the characteristic curves of the soil, the evapotranspiration rates from the soil, and the antecedent soil moisture conditions in the area of slopes and flat settings, little or no infiltration will take place.

The significance of depression-focused infiltration cannot be adequately defined with the existing instrumentation within the study area. The collection of water in depressions is evident during spring snowmelt periods and following heavy summer and autumn rainfalls. The major unanswered question is how much greater the magnitude of yearly infiltration is in the vicinity of small depressions and ephemeral sloughs than in the intervening, non-ponding areas. A limited amount of soil moisture content data from neutron probe measurements is available for the Falkirk area (Pole, personal commun.). Two access tubes located in the vicinity of shallow depressions indicated that water moved through the entire upper 2.75 m of the soil following very large rainfall events or long wet periods. Two additional access tubes situated on sloping ground showed little water movement below 2.75 m, during the same period of time. Above a depth of 1.52 m the soil moisture content responded to summer rainfalls, but the water content between 1.52 and 2.75 metres remained relatively constant. One access tube located on a ridge top showed the water content of the upper 2.75 m of the soil responded only to the unusually high precipitation and snowmelt of the spring of 1978. An expansion of the access tube network and continued monitoring will be required to quantify the magnitude of depression-focused recharge.

The second important variable, lithologic variability, is shown by the fence diagram of figure 11. Along the northern edge and western third of the area fine-textured materials predominate. The central portion of the study area is predominantly sand. The lithology has several direct and indirect effects on recharge rates within the area. The direct effects include the ability of both the unsaturated and saturated to transmit water which has already been discussed in detail.

The depth to the water table is an important variable in determining the rate at which infiltrating water reaches the water table. Four observation wells (Fa 78-2, 81-2, 84-2, 86-2) intersect the water table at depths between 15 and 22 metres. All four of the hydrographs for these wells (fig. 9) are quite flat and show very little response to the spring recharge events seen in many of the shallower piezometers. To a large extent, the depth to the water table is controlled by the rate at which water moves down to the water table, through the unsaturated zone, and the rate at which the water moves away from the water table through the saturated zone. These flow rates are, in turn, controlled by the lithology both above and below the water table.

When the water-table configuration in figure 12 is compared to the stratigraphic fence diagram of figure 11 the area of shallow water table generally corresponds to an area of predominantly fine-grained materials along the northern edge of the diagram. In these areas the average yearly recharge, based on nested piezometers and observed water-table rises, ranges from 0.09 to 0.002 $m^3 \cdot yr^{-1} \cdot m^{-2}$ (table 4), depending on the assumed value of effective porosity. In the area of piezometer nest Fa 106, the water table also remains near the ground surface because the sand is covered by only 3 to 0 metres of fine-grained materials. Areas of exposed or thinly covered sand have recharge rates on the order of 0.7 to 0.1 $m^3 \cdot yr^{-1} \cdot m^{-2}$. The area between piezometer nests Fa 106 and 45 also contains a large slough. Recharge rates under two sloughs situated over fine-textured materials

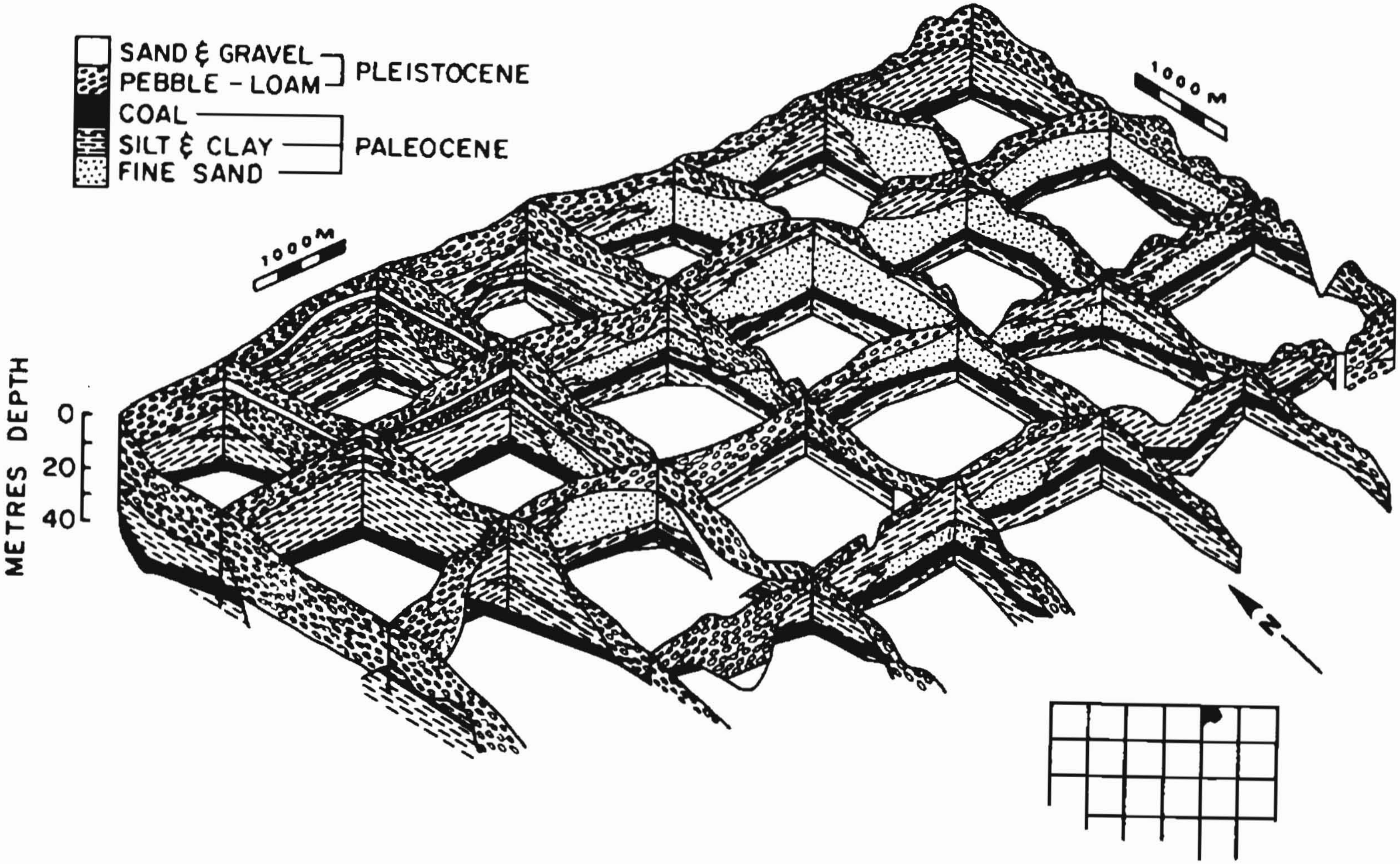


Figure 11. Stratigraphic fence diagram of the Falkirk study area.

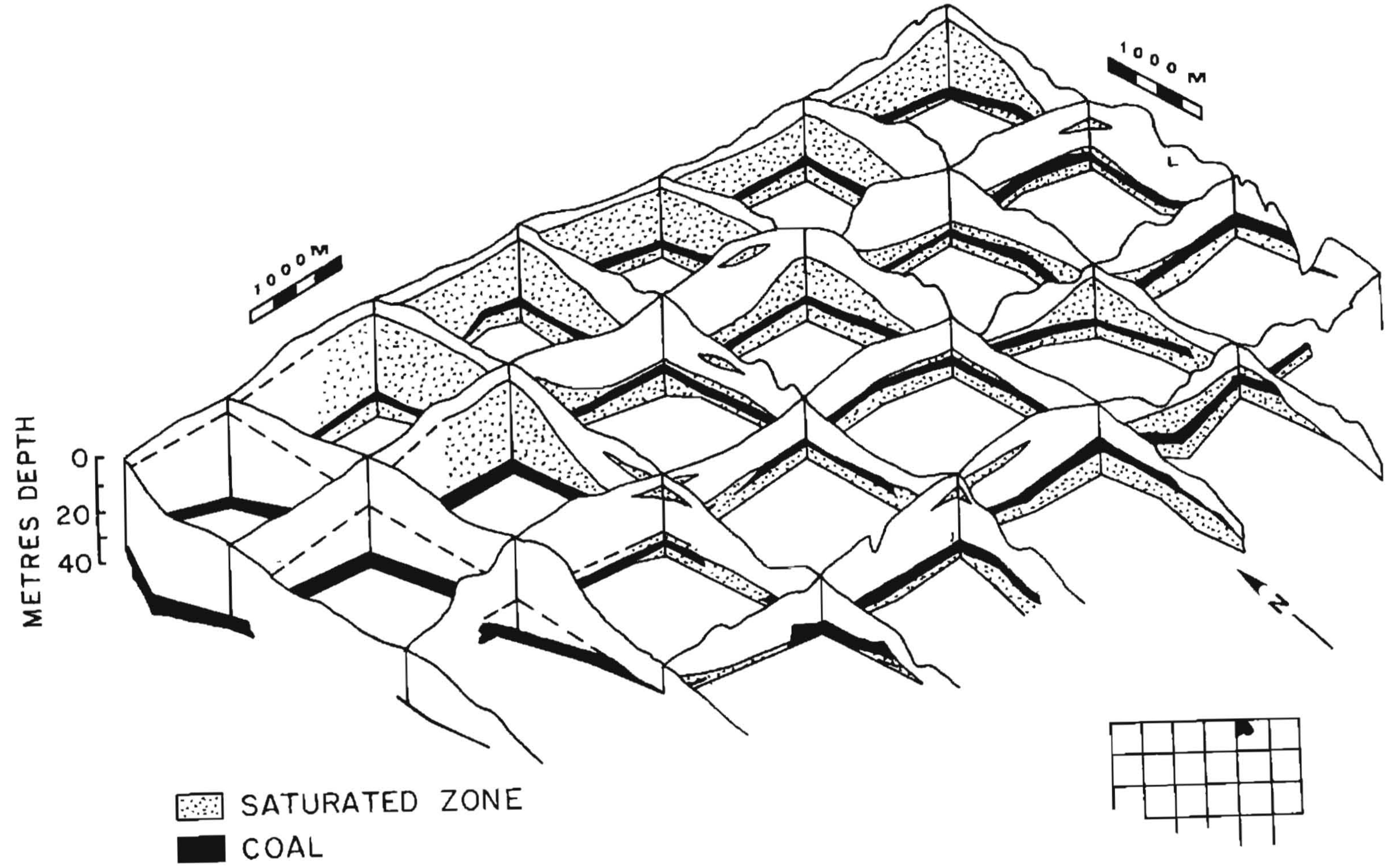


Figure 12. Configuration of the water table within the Falkirk study area.

TABLE 4. Groundwater recharge rates through four generalized hydrogeologic - stratigraphic settings.

Setting	Recharge Rate ($m^3 \cdot yr^{-1} \cdot m^{-2}$)	Area of Occurence (m^2)	Volume of Recharge ($x10^5 m^3 \cdot yr^{-1}$)
Sloughs	0.59 to 0.51	6.8×10^5	4.0 to 3.5
Exposed Sand	0.71 to 0.13	8.0×10^5	5.7 to 1.0
Fine-grained Materials Covering Sand	0.06	4.0×10^6	2.4
Predominantly Fine-grained Materials	0.091 to 0.0018	2.5×10^7	23 to 0.45
	Total	3.0×10^7	35 to 7.4

are on the order of 0.7 and 0.6 $m^3 \cdot yr^{-1} \cdot m^{-2}$ while one ephemeral slough situated in a predominantly sandy area (Fa 41-2) indicated recharge to be on the order of 0.5 $m^3 \cdot yr^{-1} \cdot m^{-2}$. These high recharge rates keep the water table near the ground surface even though the material is predominantly sand.

In the vicinity of piezometer nest Fa 40 and in between nests Fa 38 and 104 the water table is also very shallow. The water-table configuration at site Fa 40 is probably more the result of coal stratigraphy than areal variations in recharge rates. The elevation of the coal bed is 5 to 10 metres lower at site Fa 40 than to the north, east, and west. The coal bed is saturated and is separated from the sand by a thin and possibly discontinuous bed of clay. Therefore, the water table is forced to correspond to the elevation of the coal bed. A similar mechanism is probably the major influence in controlling the water-table elevation between sites Fa 38 and 104 (fig. 12). The magnitude of recharge that may be contributed from the runoff that flows down the valley between these sites is not known, but may be significant.

Sand covered by fine-textured Tertiary materials and/or pebble-loam is found extensively in the southeastern and central portion of the study area (fig. 11). In much of this area, the water table is just above the coal bed. The water table response of two observation wells (Fa 81-2 and 78-2) indicates recharge rates on the order of 0.06 $m^3 \cdot yr^{-1} \cdot m^{-2}$ (table 2), a rate that is somewhat higher than that for the finer grained material alone. The water table does not approach the ground surface because the sand and coal have a high enough hydraulic conductivity to allow recharge water to flow laterally from the site, preventing the formation of a shallow water table. The most permeable (10^{-4} $m \cdot s^{-1}$) portion of coal bed in the Falkirk study area is found along the eastern edge of the study area. The coal, in effect, acts as a drain in this part of the study area.

The water table is also quite deep

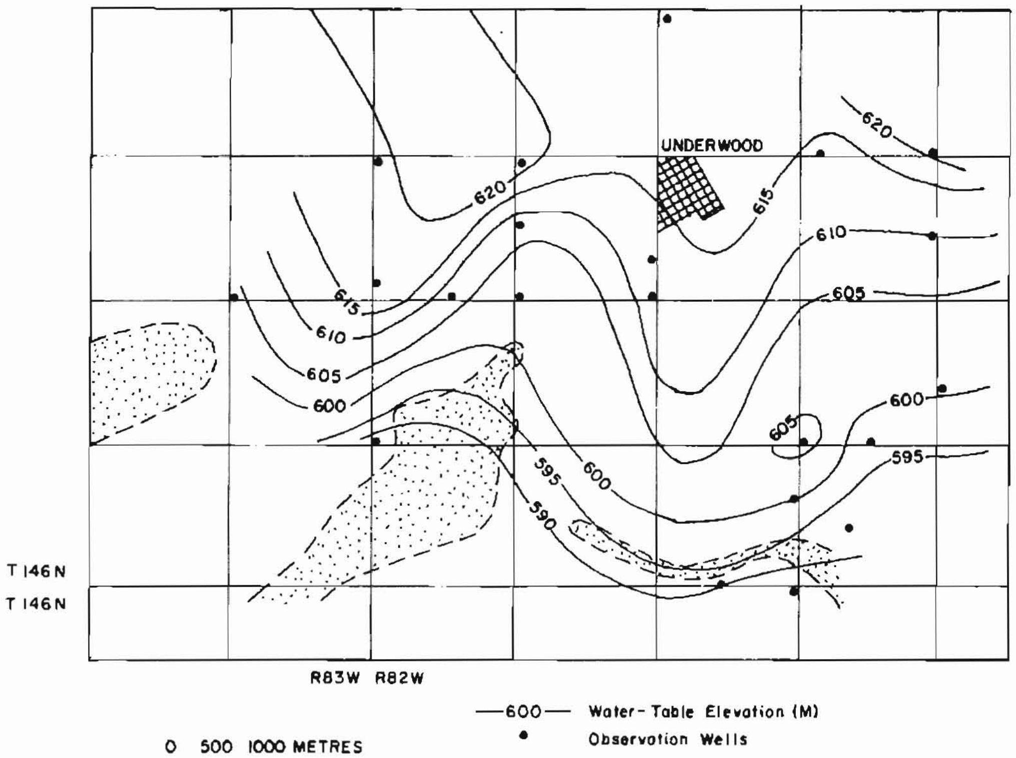


Figure 13. Water-table map of the Falkirk study area.

in the area bounded by sites Fa 82, 86, 89, 91, and 87 (fig. 12). The sand and gravel deposits buried in the area act as drains which lower the water table. This can be seen in the water-table map (fig. 13) where the surface contours of the water table bend around the gravel zones. This is especially evident with the central sand and gravel body.

The areally distributed average groundwater recharge to the shallowest coal bed in the Falkirk study area was approximated utilizing a simple numerical model (Williams, Koob, and Rechard, in prep.). The model was used to evaluate groundwater fluxes through the coal bed. By assuming that the groundwater flow system was a steady state system, the vertical, downward flux to the coal bed can be calculated by simple mass balance equations. The vertical flux out of the base of the coal bed was determined from the average hydraulic conductivity of the fine-textured materials below the coal bed and the average gradient between the shallow coal bed and the next deepest coal bed. The vertical flux out of the coal bed was assumed to be constant over the modeled portion of the study area (fig. 14). The coal bed itself was divided into small cells bounded by two stream lines and two equipotential lines. The lateral flows into and out of each cell were determined from the average hydraulic conductivity of the coal within each cell and the average horizontal gradients across the upgradient and downgradient boundaries of each cell. The difference between the two outflow

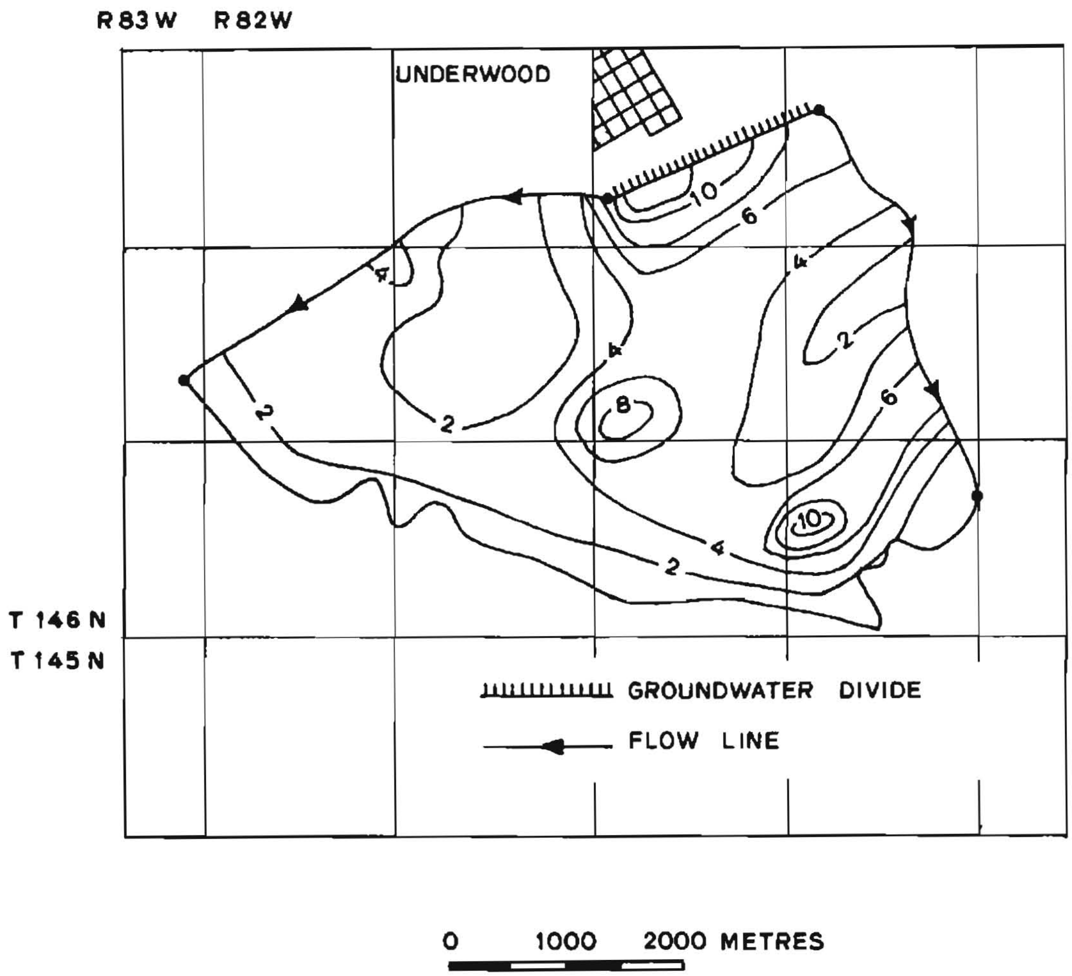


Figure 14. Spatial distribution of groundwater recharge rates ($cm \cdot yr^{-1}$) derived from the flow cell model.

fluxes and the one inflow flux then equals the vertical flux into each cell. If it is assumed that all the flow through the sediments above the coal is vertically downward, the vertical flux into a cell would be equal to the recharge rate above that cell. This assumption is probably reasonable for the fine-textured materials, but many of the sand units above the coal have appreciable horizontal gradients and fluxes. The results of the modeling effort are shown in figure 14.

Based on this model, the largest values of long-term average recharge were calculated to occur just south of the town of Underwood, where both sand is exposed at the ground surface and a large slough is situated. The data collected during 1979 and presented in this report suggest that the recharge rate is 2 to 8 times greater in this area than those predicted by the model. This difference is probably due to two factors. The lateral flow of groundwater through the sand, which increases the amount of recharge needed to maintain a steady state, was not taken into account by the model. The numbers generated by the model are long-term average values that would maintain the observed flow system in a steady state. The data collected during 1979, on the other

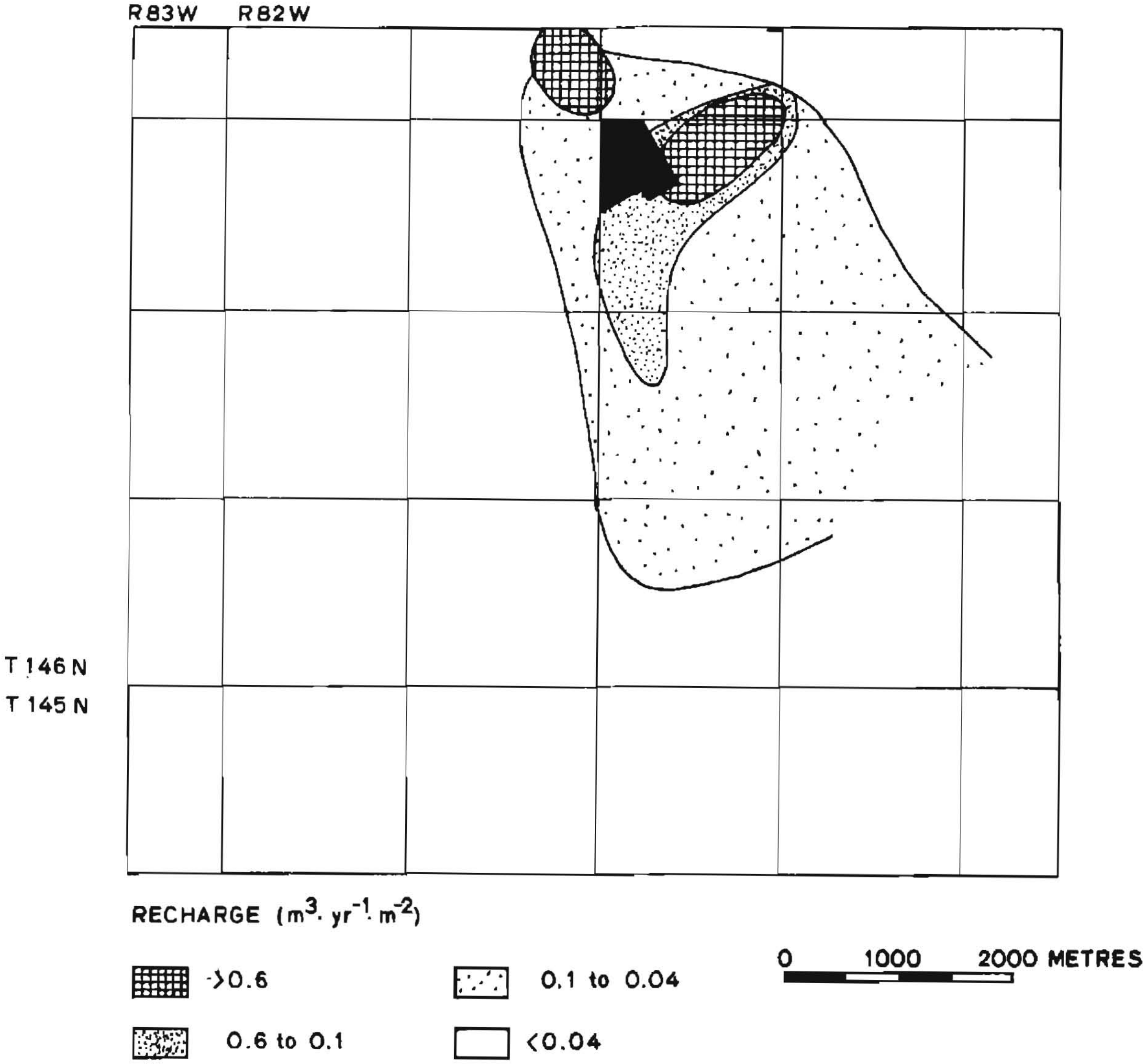


Figure 15. Spatial distribution of groundwater recharge rates based on field data.

hand, are applicable only to that year. The high recharge value in the southeastern portion of the flow system (fig. 14) may be related to a glacial meltwater valley (fig. 13) that could transmit water into the coal. The cause of the high value in the center of the model area is not known and may be due to the assumptions used in the model. In general, the model results in recharge rates smaller than, but within the same order of magnitude as, the recharge rates presented here.

The areal distribution of recharge shown in figure 15 is the result of mapping the generalized distribution of the four hydrogeologic settings listed in table 4. The distributions shown in figures 14 and 15 are roughly similar. Both show the highest recharge rates immediately southeast of the town of Underwood. An area of intermediate recharge rates is found to the east of a line running straight south from the town. The area west of the north-south line has the lowest recharge rates in both distributions.

The approximate correlation between the two methods of determining the areal distribution lend some support to the division of the area in the four hydrogeologic settings of sloughs, exposed sand, sand covered by fine-grained materials, and predominantly fine-grained materials. The difference in magnitude between the two maps is the result of several factors. The most important factor is the uncertainty involved with the assumptions used in each of the methods. Of secondary importance is that the numerical model approximates the long-term (tens of years) average recharge rate required to maintain a steady state flow system, while magnitude of groundwater recharge reported for 1979 is a function of the quantity and distribution of precipitation and evapotranspiration rates over the course of that year only.

A nonuniform distribution of groundwater recharge rates is probably typical at this local scale of investigation. A study of the Good Spirit Lake drainage area in southeastern Saskatchewan (Freeze, 1969) indicates the distribution of recharge rates (fig. 16) is as complex as that shown in figure 15 for the Falkirk study area. The range of groundwater recharge is also similar to those found in the Falkirk study area. The map in figure 16 was developed using a three-dimensional finite difference groundwater flow model. Both sites are similar in terms of climate, geology, physiography, and land use. The changes of recharge rates in the Good Spirit Lake are due primarily to changes in lithology across the site. The portion of the study area with recharge rates less than 0.0025 $m^3 \cdot yr^{-1} \cdot m^{-2}$ coincides with areas that are predominantly pebble-loam. Values greater than 0.0025 $m^3 \cdot yr^{-1} \cdot m^{-2}$ correspond to sand-covered areas, while values greater than 0.025 $m^3 \cdot yr^{-1} \cdot m^{-2}$ corresponds to a gravel outwash plain.

The variability of groundwater recharge outlined in table 4 and figure 15 are significant on a local scale. The recharge rates from individual settings can be generalized to an average areal groundwater recharge rate by multiplying the area occupied by a given setting by the average recharge rate for that setting. This has been done in columns two and three of table 4. The average areal recharge rate over the study area ranges from 0.12 to 0.025 $m^3 \cdot yr^{-1} \cdot m^{-2}$. The large range of values results from the uncertainty in the recharge rate through the fractured, fine-grained materials. The uncertainty is amplified by the large proportion (83 percent) of the study area that is covered by these materials. If the fractured media recharge rate of 0.0018 $m^3 \cdot yr^{-1} \cdot m^{-2}$ is assumed to be correct to within plus or minus one order of magnitude the range of the average areal recharge rate decreases to 0.042 to 0.025 $m^3 \cdot yr^{-1} \cdot m^{-2}$.

The magnitude of the recharge is for 1979 only. Variations in precipitation patterns, temperature, and evapotranspiration from year to year will change the magnitude of groundwater recharge. The distribution of precipitation during 1979 was somewhat atypical for the decade of the 1970s. Exceptionally heavy snowfall during the winter of 1978-1979 resulted in

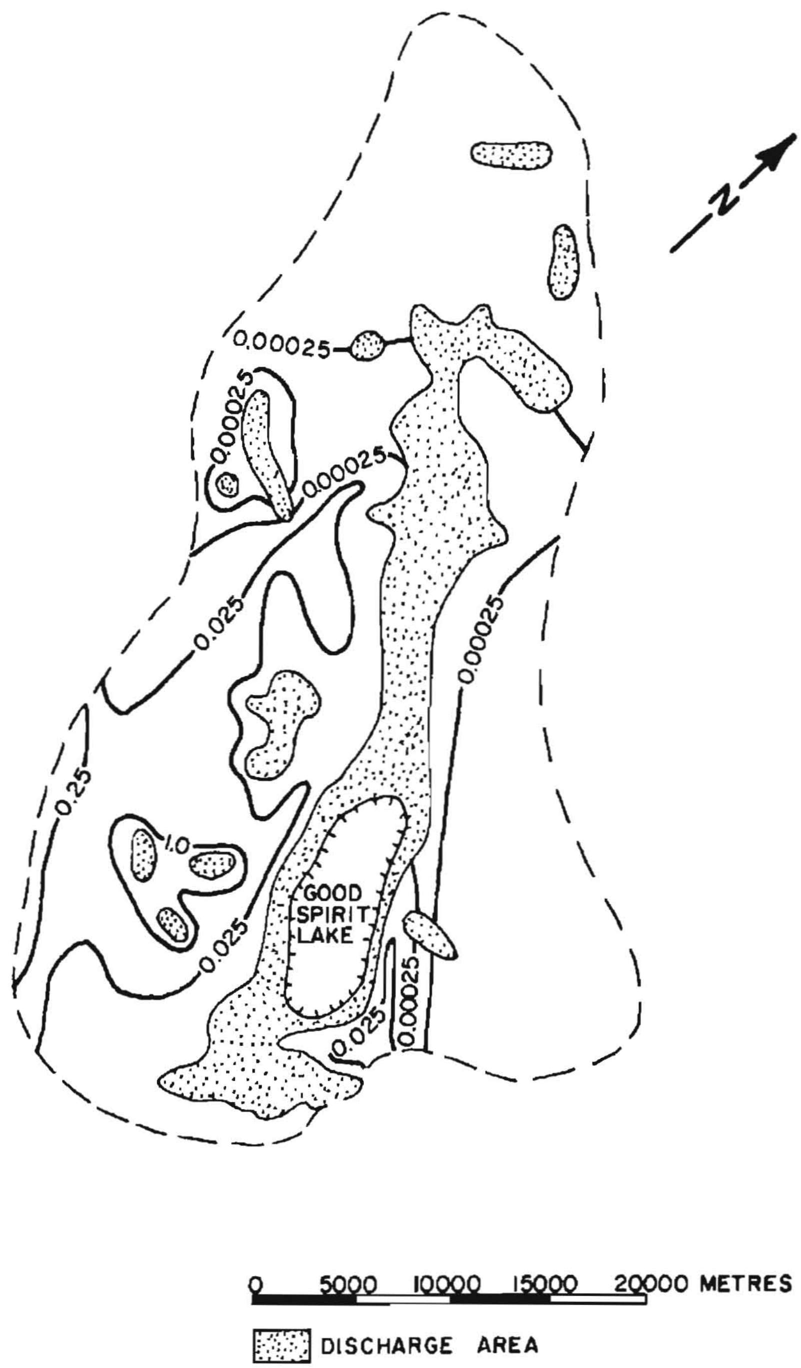


Figure 16. Spatial distribution of groundwater recharge rates ($cm \cdot yr^{-1}$) within the Good Spirit Lake drainage basin (after Freeze, 1969).

exceptional runoff in the spring of 1979. Thus, groundwater recharge during the spring of 1979 was probably much greater than was typical for the decade of the 1970s.

Groundwater recharge studies have been conducted in the Canadian portion of the Northern Great Plains by Meyboom (1966, 1967) and Freeze (1969). Climate land-use and geology are similar to the study area described here. Within the Arm River Valley, Meyboom (1966) calculated groundwater recharge to be 0.001 $m^3 \cdot yr^{-1} \cdot m^{-2}$. In a second study, that looked at groundwater at numerous sites in southeastern Saskatchewan and southwestern Manitoba, Meyboom (1967) found recharge to vary from 0.0023 to 0.038 $m^3 \cdot yr^{-1} \cdot m^{-2}$. From data presented by Freeze (1969, fig. 8) an average areal recharge rate over the Good Spirit Lake drainage basin is approximately 0.012 $m^3 \cdot yr^{-1} \cdot m^{-2}$. The average areal recharge are all quite similar even though the values were derived from sites with differing geology and topography, at different times during the last 30 years and with different methods. The similarity suggests that over large areas of the Northern Great Plains, under current climatic conditions, average groundwater recharge rates are on the order of 0.01 to 0.04 $m^3 \cdot yr^{-1} \cdot m^{-2}$.

TEMPORAL DISTRIBUTION OF GROUNDWATER RECHARGE

Groundwater recharge does not occur at a constant rate over the course of a year or from year to year. Recharge will occur when soil moisture conditions conducive to unsaturated flow coincide with periods of substantial rainfall or snowmelt. These two variables are not completely independent of each other, for the soil moisture conditions at any instant are a function of the availability of water at the ground surface prior to that instance. The availability of water is, in turn, a function of precipitation and evapotranspiration. The soil moisture conditions required for infiltrating water to reach the water table have already been outlined in previous sections of this report. The temporal distribution of precipitation will, therefore, be used as the independent variable in the following discussion of the temporal distribution of groundwater recharge.

Precipitation data have been collected within the study area for a number of years by the National Oceanic and Atmospheric Administration, at their Underwood weather station. Since 1976 additional data have been available from one to as many as three precipitation gauges operated by the Soils Department of the North Dakota State University. The data collected from 1970 to 1979 are summarized in table 1. The mean yearly total precipitation for the decade was 0.443 m with a range of 0.341 to 0.561 metres. Only four years of the ten had total precipitation values within ±0.050 m of the ten-year mean.

Generally 40 percent of the total yearly precipitation falls during the summer (June, July, and August).

Winter (November, December, January, February, and March) is usually the driest season, registering only 15 percent of the average yearly precipitation. Spring and autumn have intermediate values of 25 and 18 percent of the yearly precipitation, respectively. Seasonal precipitation values can vary dramatically from year to year for the spring, summer, and fall. Variations of a factor of two between successive years is not atypical for the 1970s. With the exception of the winter of 1978-1979, winter precipitation is reasonably consistent from year to year, ranging from 0.047 to 0.078 metres. The winter of 1978-1979 experienced an exceptionally heavy winter snowfall equivalent to 0.112 m of water, 180 percent above the average of the other nine years of the decade.

Winter precipitation in North Dakota is not available for groundwater recharge until the last few days of March when mean daily air temperatures begin to rise above the freezing mark. The snowmelt period generally extends for 2 to 3 weeks near the middle or end of April. The water available for groundwater recharge during the spring months of April and May is, therefore, the sum of the spring precipitation and the accumulated winter snowfall. This combined value has made up as much as 70 percent of the total yearly precipitation to as little as 30 percent of a year's precipitation. The average combined total for the decade is 0.172 m or about 40 percent of the average annual precipitation. The spring value is the same as that for the average summer precipitation of 0.183 m. The winter of 1978-1979 (immediately preceding the recharge events presented in this report) was extremely wet, with 0.112 m of water accumulated as snow.

The total evapotranspiration for the April through August period in 1979 was 0.890 m. Only 0.484 m of precipitation was recorded from September 1978 through August 1979, just slightly more than the evapotranspiration losses during the summer of 1979. The total evaporation from a Class A Evaporation Pan (0.941 m), which is proportional to the evaporation from the surface of a body of water, was approximately equal to the measured evapotranspiration. Evapotranspiration rates were less than 0.002 $m \cdot d^{-1}$ until the middle of June and pan evaporation rates were below 0.005 $m \cdot d^{-1}$ until the middle of May. Before these water losses became significant, 0.139 m of accumulated winter snow and rainfall were available for infiltration before the first of May.

From the preceding discussion it would seem that both the spring and summer present equal opportunity for groundwater recharge to occur. But one major factor, evapotranspiration, limits the amount of precipitation that will be available for groundwater recharge during the summer months. Evapotranspiration has been measured at the Falkirk study area by North Dakota State University personnel using weighing lysimeters planted with wheat. The measured evapotranspiration and the evaporation from a Class A Evaporation Pan are plotted in figure 17 for 1978 and 1979 (Mike Pole,

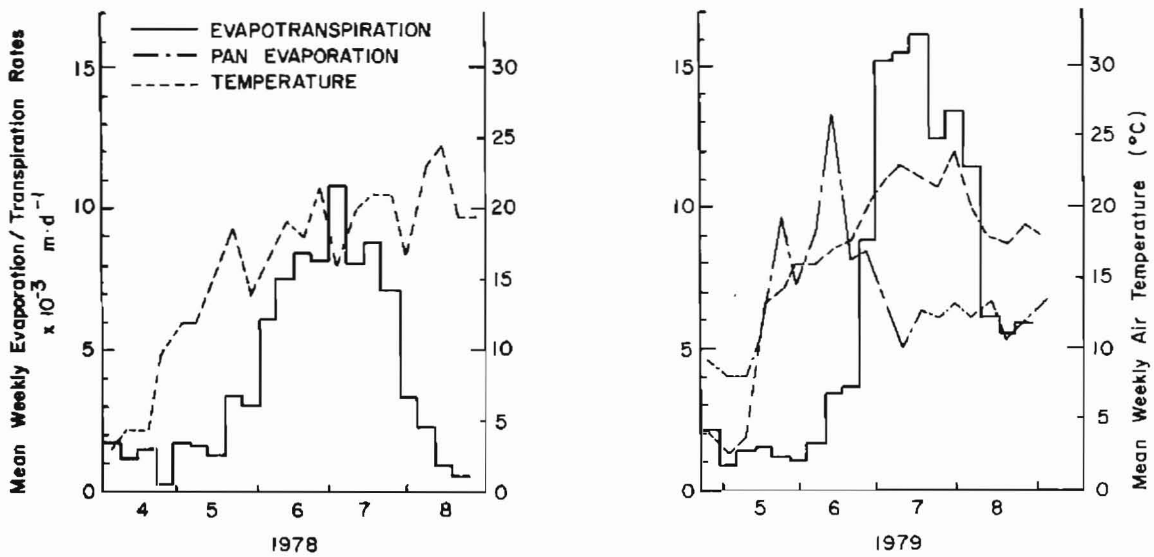


Figure 17. Measured evapotranspiration and pan evaporation rates within the Falkirk study area.

personal commun.). The values shown are approximate values because mechanical failures of the lysimeters resulted in a lack of data for several days every month. Mean weekly evapotranspiration rates were used in place of the missing values in calculating the total evapotranspiration.

The total evapotranspiration from April 1978 through August 1978 was 0.605 m which is approximately equal to the 0.595 m of precipitation that fell from September 1977 through August 1978. The equality of these figures would seem to indicate that there is no water available for groundwater recharge. During the spring, however, when 0.168 m of spring rainfall and accumulated snowmelt are available as potential groundwater recharge, the evapotranspiration rate is very low (fig. 17).

Hydrographs for the water-table wells (fig. 8a-8gg) show that the spring of 1979 was a period of large scale groundwater recharge. Of the 31 monitored wells, 25 showed varying degrees of water-table rises by the end of April to the middle of May. Two deep water-table wells, Fa 78-2 and Fa 81-2, did not respond to the spring infiltration until late May to June. An additional three observation wells showed no response at all during 1979.

The observation wells began to show groundwater recharge within several days to two weeks after the mean daily air temperature was greater than 0°C. Mean daily air temperatures dipped below the freezing mark on November 10, 1978 and stayed below freezing almost continuously until April 10, 1979 when mean daily temperatures again returned to values greater than 0°C. Only four days during the third week of March broke the continuous sub-zero temperatures of the winter season. The first water level readings recorded after the start of the April 10 thaw were taken on May 4. By this time 25 observation wells had shown water level elevation increases over either the March 26 water level measurements or extrapolations of winter water level declines. The initiation of groundwater recharge within days of the start of above freezing temperatures has also been observed in Minnesota by Schneider (1961), in Manitoba by Ackerman (1973), and in Wisconsin by Drescher (1955). It should be remembered that spring is not the only time when groundwater levels respond to precipitation or snowmelt events. During 1977 water levels responded to snowmelt and precipitation in April. Water levels declined throughout the summer of 1977 as in 1979. But, in August and September of 1977, 0.15 m of rain fell when the average rainfall for this two-month period during the 1970s was only 0.09 m. The 67 percent increase in rainfall resulted in groundwater recharge during September and October. Preliminary evaluation of data collected during 1980 shows an entirely different pattern of groundwater recharge. The winter of 1979-1980 and the spring of 1980 were unusually dry; therefore, there was only a small to zero response of groundwater levels to the limited snowmelt and rainfall. The water levels again declined all during the summer. As in 1977, there was a larger than average amount of rainfall during the end of August and early in September and again, as in 1977, there was a response by the groundwater to the precipitation. The response was smaller than that seen during the spring of 1979, and more than half the monitored water-table observations showed no response to the precipitation. A longer period of observation is needed to define the variability in the temporal distribution of groundwater recharge.

The concentrations of certain environmental isotopes of oxygen and hydrogen in groundwater can be used to supply an independent check on the season during which water entered a groundwater flow system. The isotopes of use in the evaluation of groundwater recharge are oxygen -18 (^{18}O) and deuterium (D). The isotope data are expressed as isotope ratios relative to a standard by the definition

$$\delta = \left[\frac{\left(\frac{^{18}O}{^{16}O}\right)\text{sample}}{\left(\frac{^{18}O}{^{16}O}\right)\text{standard}} - 1 \right] \times 1000$$

where the ratios are expressed as delta (δ) units in parts per thousand (per mille or $^{o}/oo$). The standard used in this expression is Standard Mean Ocean Water (SMOW). The standard is an approximation of the isotopic composition of the world's oceans. The isotope ratios are measured with a mass spectrometer. Further details on the physics and measurement of isotopes can be found in Hoefs (1973).

Based on the SMOW standard the $\delta\,^{18}O$ values of water within the hydrologic cycle generally ranges from 0 to -25 per mille. Colder climatic conditions tend to produce lighter, more negative, $\delta\,^{18}O$ precipitation values. The lowest $\delta\,^{18}O$ values are generally found in snow. The $\delta\,^{18}O$ values for individual rainfall events can vary depending on the source of the vapor mass from which the rain is produced and on the temperature of the air in which the rain is condensed. The relationship between the climatic conditions under which precipitation occurs and the $\delta^{18}O$ value of that precipitation has been defined by the expression

$$\delta^{18}O = 0.695\ T_a - 13.6$$

where T_a is the mean annual temperature (Dansgaard, 1964).

The deuterium content of precipitation correlates linearly with the ^{18}O content of the precipitation. The relationship

$$\delta D\ (^o/oo) = 8\,\delta^{18}O\ (^o/oo) + 10$$

defined by Craig (1961) is the best fit equation from precipitation samples collected from all over the world. The line defined by this equation is generally referred to as the meteoric water line.

The evaporation of meteoric water enriches the water in the heavy isotopes of both D and ^{18}O, making the δ-values of both isotopes less negative. The rates of enrichment of the two isotopes differ, such that the slope of the line defining the enrichment caused by evaporation is less than that of the meteoric water line. This difference in the rates of enrichment makes it possible to differentiate water that has undergone evaporation from non-evaporated, meteoric water.

The ^{18}O content of the water that infiltrates to a water table is a function of the ^{18}O content of the precipitation and the degree to which the water is evaporated before it reaches depths at which evapotranspiration is no longer effective. Once the water can no longer be evaporated the ^{18}O content of the water remains unchanged by chemical or biological processes except in unusual circumstances, such as the high temperature rock-water interactions found in geothermal areas (Craig, 1963). Oxygen -18 can, therefore, be considered a conservative, non-reactive chemical species that reflects climatic conditions at the time precipitation developed and the degree to which the water was evaporated prior to entering a groundwater flow system. Under favorable conditions this information can be used in the evaluation of groundwater recharge.

As water moves through the unsaturated zone and across the water table, mixing of water from several recharge events smooths the isotopic compositions of the individual events. The result is that the isotopic composition of the groundwater corresponds to the mean isotopic composition of the infiltrating water. The concentration of ^{18}O and D in the groundwater is, therefore, related to the composition of the precipitation at the time of recharge. If groundwater recharge originates primarily from the infiltration of spring snowmelt, the recharge water would be expected to have relatively light $\delta^{18}O$ values and the ratio of D to ^{18}O would plot on the meteoric water line. If, on the other hand, the predominant source of groundwater recharge is seepage from sloughs, lakes, or streams the recharge water will have a less negative value for ^{18}O and D, and the ratio of

D to ^{18}O will reflect the degree to which that surface-water body has been evaporated.

Stable isotope analyses of water samples from the Falkirk area were carried out by Dr. Robert Koob of North Dakota State University. The samples included precipitation, spring snowmelt, slough water, water from the unsaturated zone, and groundwater collected from the water table to depths of 50 m. These data are listed in tables 5 and 6 and are summarized as a $\delta^{18}O$ - δD plot in figure 18.

Samples of snow, snowmelt, and rain were collected within the study area. The $\delta^{18}O$ values range from -26.26 $^{o}/oo$ for a snow sample to -4.49 $^{o}/oo$ from an August thunderstorm while δD values range from -227 $^{o}/oo$ to -42 $^{o}/oo$ (table 5). Snow and snowmelt samples collected during March and April have a mean $\delta^{18}O$ value of -21.13 $^{o}/oo$ with a range of -26.26 to -16.76 per mille. Rainfall has much heavier $\delta^{18}O$ values. Two summer convective storms in May and August had ^{18}O contents of -4.91 and -4.49 per mille, respectively. A two-day frontal storm sampled in September had slightly lighter values of -8.62 and -7.64 per mille. Three snow samples had a mean δD value of -192 $^{o}/oo$ while the single rain sample had a value of -42 $^{o}/oo$. Precipitation samples with both $\delta^{18}O$ and δD fall near the world meteoric water line.

Samples of slough water were collected periodically from April through October during 1980 and 1978 (fig. 18). Oxygen -18 contents were measured for both years but deuterium was analyzed for the 1980 samples only. The majority (15) of the slough samples plotted on figure 18 were collected from three sloughs in the Falkirk study area, one permanent and two semi-permanent sloughs. Three samples from two ephemeral sloughs at the Falkirk site are also included. Seven additional samples were collected from four sloughs 50 km to the south and west of the study area. All but one sample collected from Falkirk during March plot just below and to the right of the meteoric water line, indicating slight evaporative enrichment. Only one sample collected west of the study area was very enriched, with a $\delta^{18}O$ value of +1.7 $^{o}/oo$ and a δD value of -42 $^{o}/oo$. The slough waters show only minor evaporative enrichment because of the large and relatively steady input of isotopically heavy precipitation during the months of maximum evaporation rates--June, July, and August.

The sloughs followed different patterns of ^{18}O enrichment during 1978 and 1980. During 1978 there was a steady enrichment from -18.5 $^{o}/oo$ to -5.5 $^{o}/oo$. In 1980 the rate of isotope enrichment was much greater during March and April. From April through July of 1980 the sloughs maintained a 3 to 6 per mille enrichment in ^{18}O over the same period of 1978. By October of 1980 the sloughs were 3 $^{o}/oo$ less enriched in ^{18}O than during October of 1978.

Climatically, 1978 and 1980 were quite different. April of 1980 had a mean temperature 4°C warmer than April of 1978. This corresponds to the

TABLE 5. Stable isotope data from precipitation and surface water within the Falkirk study area.

Sampling Point	Date	$\delta^{18}O$ (°/oo)	δD (°/oo)	Sampling Point	Date	$\delta^{18}O$ (°/oo)	δD (°/oo)
Rain	05-78	- 4.91		Sammuelson	06-21-77	- 7.29	
	08-78	- 4.49	- 42	Slough	04-21-78	-11.68	
	09-78	- 8.62			05-17-78	-10.70	
	09-78	- 7.64			06-13-78	-10.86	
Snow	04-78	-16.76			08-23-78	- 9.47	
	04-78	-18.64			09-18-78	- 6.57	
	04-78	-19.34			04-15-80	- 7.79	- 74
	03-79	-25.93	-206		05-09-80	- 6.68	- 74
	03-79	-26.25	-195		06-04-80	-10.51	- 82
	03-80	-21.41	-156		07-02-80	- 7.22	- 84
Snowmelt	03-78	-18.89			10-29-80	- 7.28	- 64
	03-78	-18.05		Slough AA	05-17-78	-14.26	
	03-79		-134		08-23-78	- 7.24	
	03-80	-20.57			09-18-78	- 7.02	
	03-80	-25.47			04-15-80	- 9.42	- 93
					06-04-80	- 6.55	- 73
Underwood	04-21-78	-15.52			05-09-80	- 4.14	- 48
Slough	05-17-78	-10.20			07-08-80	- 3.22	- 34
	09-18-78	- 7.85	- 51		10-29-80	- 7.97	- 61
	10-26-78	- 5.73		Antelope	04-14-80	-17.81	-132
	05-04-79		-140	Slough	05-09-80	- 9.88	- 67
	06-02-79		-107	Slough No. 41	04-21-78	-14.73	
	03-28-80	-17.21	-139		03-28-80	-21.91	-164
	05-09-80	- 8.94	- 71				
	06-04-80	- 9.34	- 58				
	07-02-80	- 2.96	- 38				
	10-29-80	-10-47	- 89				
Slough No. 1	04-21-78	-19.10					
	05-17-78	-14.62					
	08-23-78	- 6.28					
	09-18-78	- 4.69	- 50				
	10-26-78	- 5.24					
	06-01-79	-13.41	-116				

TABLE 6. Stable Isotope data from subsurface water within the Falkirk study area.

Sampling Point	Stratigraphic Position*	Date	$\delta\,^{18}O(^{o}/oo)$	$\delta\,D(^{o}/oo)$
Soil-water Samplers				
48-2	1	08-79	-15.47	-114
50-1	1	08-79		-114
		04-80	-17.19	-125
50-2	1	04-80	-14.93	-117
50-3	1	04-80	-17.94	-111
50-4	1	04-80	-15.43	-117
86-1	1	08-79		-120
		04-80	-19.96	-117
86-2	1	08-79		-122
		04-80	-19.23	-131
86-3	1	04-80	-17.41	-127
86-4	1	04-80	-16.48	-128
86-5	1	08-79	-17.42	-120
		04-80	-16.96	-125
86-6	1	08-79		-117
		04-80	-19.51	-123
Wells and Piezometers				
1-1	3	05-17-78	-15.76	
2-1	3	05-18-78	-14.84	
2-2	3	05-18-78	-15.22	
3-2	1	05-17-78	-19.93	
4-2	1	05-17-78	-14.50	
6-2	3	05-18-78	-13.60	
6-3	2	05-18-78	-18.38	
6-4	2	06-06-78	-18.54	
6-5	1	06-05-78	-18.22	-126
		03-27-80	-20.15	-135
		05-01-80	-19.93	-135
		07-08-80	-18.13	-125
		07-29-80	-19.86	-134
7-3	1	06-07-78	-13.68	
8-1	2	07-07-77	-12.43	
8-2	2	07-07-77	-13.14	
9-2	2	06-06-80	-13.20	-107
10-1	3	09-08-78	- 6.84	- 73
10-2	2	06-06-78	-13.28	
		09-08-78	-13.24	
		06-06-80	-16.50	-105
10-3	1	06-06-78	-13.91	
		09-08-78	-13.11	- 91
		08-06-80	-14.18	-107
36-1	3	05-25-78	-17.62	
39-1	3	07-29-80	-18.07	-122
39-2	1	04-30-80	-14.46	-119
		07-29-80	-15.62	-120
		10-29-80	-16.17	-123
40-1	3	05-25-78	-17.11	
40-2	2	05-31-78	-15.63	
40-3	2	05-31-78	-11.96	- 85

*1 = Soil water samplers and water-table observation wells - all above the Hagel bed.
2 = Piezometer above the Hagel bed.
3 = Piezometer within or below the Hagel bed.

TABLE 6. Stable Isotope data from subsurface water within the Falkirk study area.--Continued

Sampling Point	Stratigraphic Position*	Date	$\delta^{18}O$(°/oo)	δ D(°/oo)
40-4	1	10-31-78	-12.54	
41-1	3	07-21-77	-15.66	
		06-06-78	-17.46	
		04-30-80	-16.88	- 96
		07-08-80	-16.09	-123
41-2	1	10-31-78	-10.39	- 68
		03-28-80	-18.64	-133
		04-30-80	-14.48	-104
42-1	3	07-06-77	-15.73	
		05-31-78	-13.87	
43-1	3	07-26-77	- 8.49	
		06-06-78	-10.55	- 96
		04-30-80	-16.88	
43-2	1	04-30-80	-14.48	
45-1	3	05-31-78	-12.70	-118
45-3	3	07-07-80	-13.06	-115
46-1	1	11-01-78	-14.49	-117
		05-01-80		-119
		07-29-80	-15.60	-122
		10-29-80	-15.78	-129
50-1	1	10-31-78	-16.70	-123
		04-15-80	-18.59	-125
		07-29-80	-18.57	-135
		10-29-80	-16.96	-123
73-1	2	06-01-78	-15.61	
74-1	2	06-01-78	-16.12	
		06-06-80		-131
75-1	3	08-12-77	-11.38	
		06-06-80	-16.49	-115
76-1	3	07-28-77	-12.68	
		07-22-77	-16.76	
77-1	3	08-16-77	-15.14	
78-1	3	08-12-77	-15.28	
80-1	3	06-01-78	-17.24	
		07-08-80	-18.40	-143
80-3	1	06-01-78	-16.69	-125
81-1	3	07-11-77	-15.41	
		07-08-80	-16.06	-115
81-2	1	07-08-80	-14.47	-112
82-1	3	07-20-77	-14.95	
		06-01-78	-14.59	
		07-08-80	-15.11	-125
82-3	2	06-01-80	-14.96	
83-1	3	08-16-77	-12.35	
86-1	3	07-29-80	-14.38	-115
86-2	1	09-08-78	-16.45	
		03-26-80	-20.19	-132
		04-14-80	-18.36	-131
		07-29-80	-17.36	-134
88-1	3	07-20-77	-13.59	

*1 = Soil water samplers and water-table observation wells - all above the Hagel bed.
2 = Piezometer above the Hagel bed.
3 = Piezometer within or below the Hagel bed.

TABLE 6. Stable Isotope data from subsurface water within the Falkirk study area.--Continued

Sampling Point	Stratigraphic Position*	Date	$\delta^{18}O(^o/oo)$	$\delta D(^o/oo)$
		07-08-80	-17.60	-133
88-3	2	05-24-78	-14.27	
		03-27-80	-14.21	-112
		07-08-80	-14.89	-116
89-1	3	07-18-77	-15.56	
91-1	3	06-07-78	-15.29	
91-3	1	06-07-78	-15.81	
94-1	3	07-15-77	-15.86	
		06-07-78	-16.13	
95-1	3	11-01-78	-14.37	
95-2	3	11-01-78	-14.16	
95-3	2	11-01-78	- 4.37	- 52
95-4	1	11-01-78	-12.25	- 89
		04-30-80		-164
		07-07-80		-100
		07-29-80	- 9.38	- 76
		10-29-80	-13.09	- 83
98-1	1	11-01-78	-14.72	-121
		03-27-80	-15.17	-120
		04-30-80	-15.60	-121
		07-29-80	-14.68	-114
99-1	2	09-08-78	-14.59	
		07-07-80	-16.34	-114
		10-14-80	-14.24	-114
99-2	3	07-07-80	-14.90	-104
		10-14-80	-15.12	-117
99-3	3	07-07-80	-15.05	-121
		10-14-80	-13.30	-102
99-4	1	07-07-80	-12.50	- 84
		10-14-80	-12.87	- 86
100-1	1	09-08-78	-16.52	
		03-27-80	-16.22	-112
		07-08-80	-16.31	-119
		07-29-80	-15.35	-104
		10-29-80	-15.24	-108
100-2	3	07-08-80	-15.48	-116
100-3	3	07-08-80	-13.92	-107
100-4	2	07-08-80	-13.34	- 98
101-1	3	08-11-77	-16.80	
101-2	3	08-11-77	-14.23	
102-1	3	07-13-77	-13.45	
102-2	3	07-13-77	-16.29	
103-1	3	06-07-78	-14.67	
104-1	3	08-09-77	-15.91	
105-6	3	10-29-80	-16.57	-137
106-2	3	06-08-78	-14.88	
106-3	3	08-01-77	-15.40	
106-4	2	08-02-77	-15.68	
106-5	1	04-30-80	-18.84	-127
		07-28-80	-17.66	-133

*1 = Soil water samplers and water-table observation wells - all above the Hagel bed.
2 = Piezometer above the Hagel bed.
3 = Piezometer within or below the Hagel bed.

TABLE 6. Stable Isotope data from subsurface water within the Falkirk study area.--Continued

Sampling Point	Stratigraphic Position*	Date	$\delta^{18}O$ (°/oo)	δD (°/oo)
106-6	3	08-02-77	-15.74	
107-1	3	06-16-76	-12.38	
107-2	3	06-01-78	-15.02	
108-1	3	06-02-78	-18.06	
108-2	3	06-02-78	-16.96	
109-1	3	08-16-77	-17.12	
		05-26-78	-16.40	
109-2	3	08-17-77	-16.78	
109-3	3	07-18-77	-14.72	
		06-02-78	-15.67	
		07-29-80	- 9.59	- 78
110-2	3	06-06-78	-18.98	
		07-29-80	-12.36	-114
111-1	3	04-25-78	-16.87	
111-2	3	04-25-78	-16.60	
111-3	2	04-25-78	-15.72	
111-4	2	04-25-78	-15.00	
111-5	1	04-25-78	-17.12	

*1 = Soil water samplers and water-table observation wells - all above the Hagel bed.
2 = Piezometer above the Hagel bed.
3 = Piezometer within or below the Hagel bed.

period during which the rate of isotope enrichment was greatest in 1980. During May, June, and July, mean monthly temperatures were about the same in both 1978 and 1980 and the rates of ^{18}O enrichment were approximately equal. During August, September, and October of 1980, the mean monthly temperature was 3°C less and there was 0.20 m more precipitation than in 1978. These climatic conditions resulted in the decline of $\delta^{18}O$ values during 1980 while $\delta^{18}O$ values continued to increase during 1978. Generally, 1978 was much more typical of seasonal climatic changes for the decade of the 1970s than was 1980. This implies that the isotopic composition of 1980 surface waters plotted on figure 18 is not typical of sloughs in the study area. The major discrepancy between the data collected during 1980 and a more typical year occurs in the spring when sloughs are probably much lighter, isotopically, than the 1980 data would suggest. If isotopically light, springtime surface waters were included on figure 18 the surface water field would extend down the meteoric water line to $\delta^{18}O$ values on the order of -15 to -18 per mille. This field would then overlap with many of the groundwater samples.

The bulk of the shallow groundwater samples (from above the Hagel Lignite bed) and deep groundwater samples from the Hagel bed and deeper confined aquifers all plot along the meteoric water line, midway between the snow and snowmelt and slough

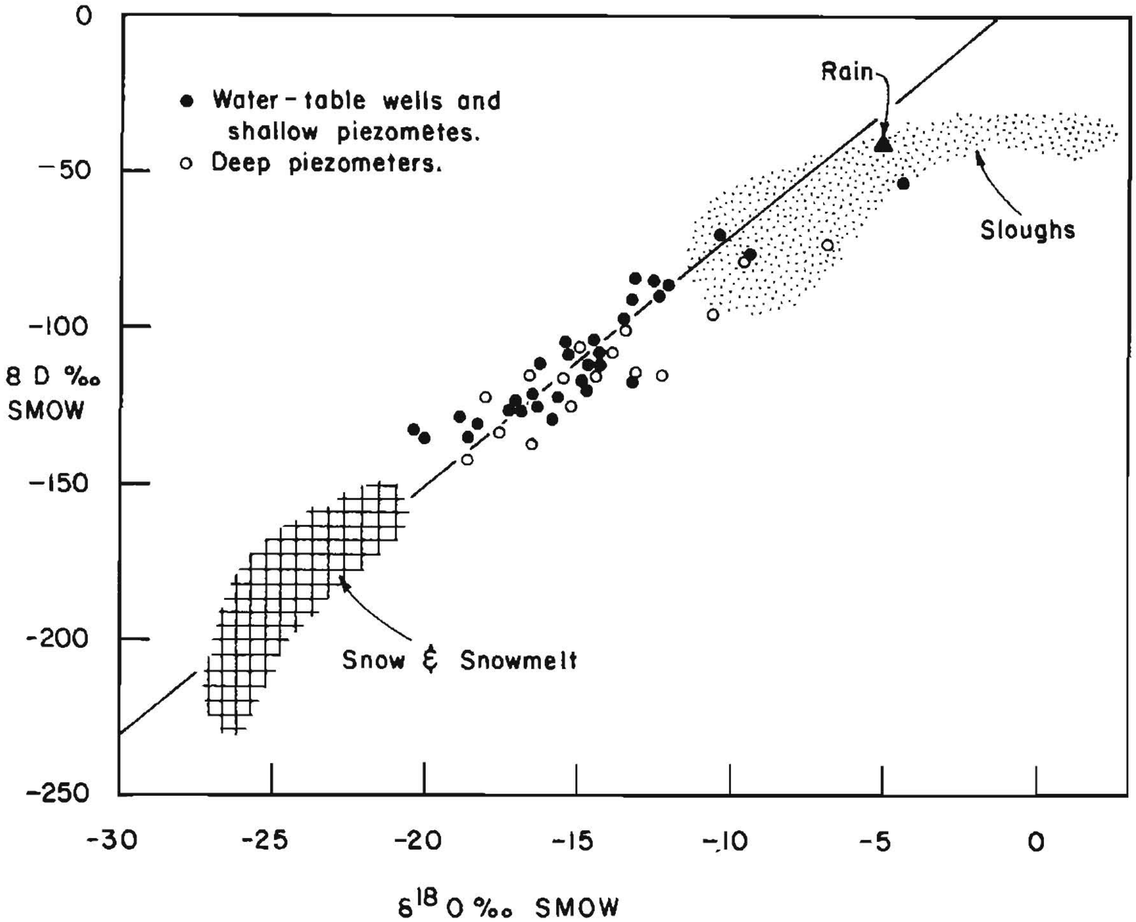


Figure 18. Stable isotope composition of precipitation, surface water, and groundwater within the Falkirk study area.

water samples (fig. 18). Water from the unsaturated zone, collected with pressure-vacuum soil-water samplers (Wood, 1973) to depths of 12 m, plot with the isotopically light groundwater samples. The $\delta^{18}O$ - δD plot shows groundwater to be generally nonevaporated with an isotopic composition that tends toward what would be a typical spring surface water. Those groundwater samples that plot farthest to the right of and up the meteoric water line are all from wells and piezometers that are near the large permanent sloughs that are situated over sandy bedrock (fig. 3). In this area significant quantities of summer and autumn rains and slightly evaporated slough water enter the groundwater flow system. As the water moves toward the discharge areas on the edges of the study area, infiltrating water, with an isotopic composition that approaches spring snowmelt, contributes greater proportions of the total groundwater flux, resulting in groundwater with an isotopic composition between snowmelt and spring slough water. The similarity between the $\delta^{18}O$ and δD values of subsurface water and spring snowmelts and sloughs suggests that the bulk of groundwater recharge occurs from March through May.

A seasonal trend is evident for both oxygen-18 and deuterium samples collected from above the Hagel bed. The samples collected during the period March to June from this portion of the landscape are significantly lighter (at the 0.01 and 0.05 significance level) than the samples from above the Hagel bed collected from July to November. Early (March to June) samples average -17.76 $^{o}/oo$ ^{18}O content while the samples collected from July to November have an average $\delta^{18}O$ value of -14.26 $^{o}/oo$. The early samples analyzed for D average -122 $^{o}/oo$ while the summer/autumn values for D average -98 $^{o}/oo$ (table 6).

Eight water-table wells and three piezometers were sampled for ^{18}O in more than one month, though the samples were not all collected during one year. Nine of these sample sets showed an increase in the $\delta^{18}O$ value of the groundwater from spring to autumn. The enrichment ranged from 1 to 13 per mille with a median value of 3 $^{o}/oo$. The remaining two sample sets showed a slight, less than 0.30 $^{o}/oo$ decrease in ^{18}O content over the course of the summer.

Although a seasonal enrichment in D is evident from the bulk sample, the same trend is not evident from wells and piezometers that were sampled during several different seasons. Of the water-table wells for which seasonal D analyses are available, only one (Fa 95-4) shows a significant enrichment (75 $^{o}/oo$) from the spring to the fall (table 6). The values from this well, as well as those from Fa 6-5 and Fa 10-3, plot along the meteoric water line (fig. 18). This implies that isotopically heavy summer rainfall reached the shallow water table at these three sites. Two wells (Fa 46-1 and Fa 86-2) have large $\delta^{18}O$ shifts of 2.5 to 5.5 per mille with negligible δD shifts of 2 and 5 per mille. This usually indicates mineral-water interactions, but the nature of possible chemical reactions at these sites are not known. The final site (Fa 98-1) showed no change in either ^{18}O or D.

Combining the surface water isotope data with the information supplied by the hydrographs of the water-table wells (fig. 10) does much to explain the observed patterns of isotope contents in the subsurface waters. The snowmelt of March, with its very negative δ values does not contribute significantly to groundwater recharge. By mid-April, when the water-table hydrographs have begun to rise, the surface water isotope values rise to the range of values found in the unsaturated zone and the water-table samples collected between March and June. By the end of May to June, when the water table is beginning its usual summer decline, the δ values of precipitation and surface water are generally above that found in most of the groundwater samples analyzed to date. Some of this summer water, as evidenced by the increasingly heavier isotope ratios of the shallower wells and piezometers, may be reaching the water table, possibly from depression focused or slough recharge. The semi-permanent and permanent sloughs in the area probably contribute to groundwater recharge from at least April to December. The extent to which the isotopically heavy, evaporated slough water contributes to groundwater recharge is difficult to assess because of the small number of piezometers in the immediate vicinity of a slough and the limited isotope sampling over the life of these piezometers. Two piezometer sites near sloughs, Fa 98 and Fa 106, indicate seasonal enrichment of ^{18}O while piezometers Fa 43-1, Fa 45-1, and Fa 110-1 and 2 have slightly evaporated waters as evidenced by the $\delta^{18}O - \delta D$ plot (fig. 18). The overall pattern of ^{18}O variation in groundwater indicates, however, that the contribution of evaporated water from permanent sloughs, due to their limited areal extent, plays a less significant role in the regional groundwater flow system than water that enters the landscape in the spring from the numerous small depressions and ephemeral sloughs that are located over the entire Falkirk area. The major role played by the evaporated slough water may be to increase the δ values of groundwater by mixing with the isotopically light spring recharge water, which enters the groundwater flow system from throughout the landscape as well as from permanent sloughs.

SUMMARY

Groundwater recharge is a complex process that is difficult to quantify. It is a function of the spatial and temporal distribution of precipitation, topography-runoff relationships, and the unsaturated and saturated hydraulic properties of a spatially heterogeneous geologic environment. Estimates of groundwater recharge presented here are based on a network of piezometers and water-table wells within the Falkirk study area. These data were supplemented with observations on evapotranspiration rates, soil moisture-content changes, and stable isotope concentrations.

Groundwater recharge takes place throughout the landscape at widely varying rates (table 4). Semi-permanent to permanent sloughs generally have the highest recharge rates, but their restricted areal extent limits the overall contribution of sloughs to the groundwater flow system to between 10 to as much as 50 percent of the total annual recharge. The role played by small, several-square-metre depressions and ephemeral sloughs is more difficult to define because of a lack of data. Limited observations of soil-moisture content indicate that water will move below the root zone near shallow depressions following short, heavy rainfalls, long, light to moderate rainfalls, or spring snowmelt.

The large range of values for the contribution of sloughs results from the uncertainty of the recharge rates through the fractured, fine-textured materials that cover over 80 percent of the study area. Estimates of groundwater recharge through these materials vary by two orders of magnitude depending on whether flow takes place exclusively through the fracture network or exclusively through intergranular pore spaces.

The average areal groundwater recharge rate through the 150 km^2 study area is on the order of 0.12 $m^3 \cdot yr^{-1} \cdot m^{-2}$ to 0.025 $m^3 \cdot yr^{-1} \cdot m^{-2}$. These recharge rates are for 1979 only. The spring of 1979 was the wettest spring of the decade of the 1970s, so the recharge rates are probably greater than the long-term, average groundwater recharge rate. The 1979 Falkirk groundwater recharge rates generally fall on the high end of the range of recharge values that have previously been reported from the Northern Great Plains.

The spatial distribution of recharge is largely dependent upon two factors--the size of surface depressions and underlying lithology. Groundwater recharge rates are very high in the vicinity of the town of Underwood. This area is characterized by several, large, permanent sloughs and is underlain by sandy bedrock exposed at the ground surface. Lower recharge rates in areas of large sloughs correspond to areas underlain by predominantly fine-textured materials. Depression-focused recharge, where small depressions and ephemeral sloughs collect runoff from spring snowmelt and heavy rainfall events was not evaluated in the course of this project.

Based on an evaluation of water-table hydrograph records, the majority of groundwater recharge occurs during the months of March and April. During these months the snow that has accumulated over a 4- to 5-month period, as well as the precipitation that falls, is available for infiltration. Evapotranspiration rates during March and April are negligible to very small. Much of this infiltrating water therefore reaches the water table or gets below the plant root zone where it cannot be recycled to the atmosphere. A limited amount of water passes below the root zone following heavy summer and autumn rains, but the water table was not seen to respond to any summer or autumn rains during 1979. This observation is apparently the result of water reaching the water table at a slower rate than the rate at which water flowed away from the water table through the saturated zone.

Stable isotope data indicate that much of the water at or above the water table during the spring resembles snowmelt or slough waters. Late summer and autumn water-table samples (not necessarily collected during the same year) are enriched in both ^{18}O and D relative to the spring samples. This enrichment does not appear to be the result of evaporated water entering the flow system, but rather of mixing of summer rains with the spring snowmelt. Isotope analyses from deeper in the flow system are very limited but suggest that slightly evaporated water may be entering the groundwater flow system in the vicinity of the sloughs around the town of Underwood.

Several areas require further research. The movement of water through the unsaturated zone is one of these areas. Of particular relevance are the questions of movement through fractures versus movement through the unfractured porous matrix and the dynamics of water movement through thick, 3- to 25-metre, unsaturated zones. Another area requiring further work is the quantification of the role of depression-focused recharge in small depressions and Class 1 ephemeral sloughs. A more complete analysis of deuterium contents in groundwater samples is also needed to clarify the contribution of summer and autumn rains and slough water to the groundwater flow system.

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APPENDIX

APPENDIX

Piezometer No.: Fa 3-2
145-082-04 CDC
Wellhead Elevation: 577.51 m
Ground Elevation: 576.77 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	2.44	575.07
2-21	2.56	574.95
3-29	1.46	576.05
5- 3	1.25	576.26
5-31	1.77	575.74
6-26	1.73	575.78
7-31	2.00	575.51
9- 7	2.44	575.07
10- 3	2.51	575.00
10-31	2.42	575.09
11-29	2.45	575.06
12-27	2.51	575.00

Piezometer No.: Fa 4-2
145-082-03 CDD
Wellhead Elevation: 583.37 m
Ground Elevation: 582.23 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	1.31	582.06
2-21	1.41	581.96
3-22	frozen	
5- 3	1.41	581.96
6-26	1.42	581.95
9- 7	1.50	581.87
10- 3	1.52	581.85
12-27	1.54	581.83

Piezometer No.: Fa 6-5
146-083-24 DDD
Wellhead Elevation: 621.94 m
Ground Elevation: 621.66 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	inaccessible	
2-21	inaccessible	
3-28	inaccessible	
5- 3	3.00	618.94
5-15	3.17	618.77
5-30	3.35	618.59
6-19	3.52	618.42
6-28	3.65	618.29
7-18	3.82	618.12
7-31	3.91	618.03
9- 7	3.84	618.10
10- 3	3.97	617.97
10-31	4.02	617.92
11-29	4.27	617.67
12-26	4.46	617.48

Piezometer No.: Fa 10-3
146-082-07 CCC
Wellhead Elevation: 625.46 m
Ground Elevation: 624.84 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	inaccessible	
2-21	inaccessible	
3-28	inaccessible	
5- 3	4.21	620.81
5-15	3.81	621.21
5-30	3.58	621.44
6-19	3.43	621.59
6-28	3.49	621.53
7-18	3.55	621.47
7-31	3.55	621.47
9- 7	3.68	621.34
10- 2	3.69	621.33
10-31	3.66	621.36
11-29	3.85	621.17
12-26	3.80	621.22

Piezometer No.: Fa 38-2
145-082-04 AAA
Wellhead Elevation: 600.10 m
Ground Elevation: 599.50 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	inaccessible	
2-21	inaccessible	
3-29	inaccessible	
5- 3	3.12	596.98
5-30	4.18	595.92
6-19	dry	
6-26	dry	
7-18	5.14	594.96
7-31	5.14	594.96
9- 7	5.14	594.96
10- 2	5.15	594.95
10-31	5.09	595.01
11-29	5.12	594.98
12-27	5.12	594.98

Piezometer No.: Fa 39-2
146-082-34 BCC
Wellhead Elevation: 603.52 m
Ground Elevation: 602.94 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	inaccessible	
2-21	inaccessible	
3-29	inaccessible	
5- 4	2.68	600.84
5-30	2.94	600.58
6-19	3.02	600.50
6-28	3.12	600.40
7-18	3.22	600.30
10- 2	3.52	600.00
12-27	3.83	599.69

Piezometer No.: Fa 40-4
146-082-27 CCC
Wellhead Elevation: 610.48 m
Ground Elevation: 610.20 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	4.49	605.99
2-21	4.56	605.92
3-22	4.65	605.83
5- 4	4.26	606.22
5-15	4.12	606.36
5-30	4.05	606.43
6-19	3.95	606.53
6-28	3.99	606.49
7-18	4.03	606.45
7-31	4.13	606.35
9- 7	4.25	606.23
10- 2	4.27	606.21
10-31	4.30	606.18
11-29	4.46	606.02
12-27	4.52	605.96

Piezometer No.: Fa 41-2
146-082-27 BBC
Wellhead Elevation: 611.83 m
Ground Elevation: 611.15 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	10.20	601.63
2-21	inaccessible	
3-22	inaccessible	
3-29	10.46	601.37
5- 4	inaccessible	
5-14	8.69	603.14
6- 2	7.37	604.46
6-19	7.28	604.55
6-28	7.53	604.30
7-18	7.86	603.97
7-31	8.12	603.71
9- 7	8.59	603.24
10- 2	8.83	603.00
10-31	9.09	602.74
11-29	9.45	602.38
12-26	9.65	602.18

Piezometer No.: Fa 45-2
146-082-15 CCC
Wellhead Elevation: 619.24 m
Ground Elevation: 618.86 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	5.50	613.74
2-21	5.61	613.63
3-22	5.80	613.44
5-15	4.40	614.84
5-30	4.32	614.92
6-19	4.22	615.02
6-28	4.33	614.91
7-18	4.46	614.78
7-31	4.58	614.66
9- 8	4.79	614.45
10- 2	4.83	614.41
10-30	5.02	614.22
11-29	5.30	613.94
12-26	5.26	613.98

Piezometer No.: Fa 46-1
146-082-16 BBB
Wellhead Elevation: 620.37 m
Ground Elevation: 619.76 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	3.06	617.31
2-21	3.25	617.12
3-29	2.74	617.63
5- 3	1.11	619.26
5-15	1.18	619.19
5-30	1.49	618.88
6-19	2.01	618.36
6-27	1.91	618.46
7-18	2.39	617.98
7-31	2.62	617.75
10- 2	3.31	617.06
10-31	3.43	616.94
11-29	3.58	616.79
12-26	3.63	616.74

Piezometer No.: Fa 48-1
146-082-13 CDD
Wellhead Elevation: 610.15 m
Ground Elevation: 609.27 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	10.49	599.66
2-21	10.50	599.65
3-12	10.41	599.74
3-22	10.55	599.60
5- 3	10.47	599.68
5-31	10.25	599.90
6-19	10.17	599.98
6-29	10.30	599.85
7-18	10.30	599.85
7-31	10.34	599.81
9- 8	10.30	599.85
10-2	10.28	599.87
10-31	10.25	599.90
11-29	10.39	599.76
12-26	10.39	599.76

Piezometer No.: 49-1
146-082-17 DDC
Wellhead Elevation: 615.05 m
Ground Elevation: 614.12 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	8.06	606.99
2-21	8.20	606.85
3-12	7.99	607.06
3-22	7.93	607.12
3-29	7.67	607.38
5- 3	7.07	607.98
5-15	12.63	602.42
6- 1	12.89	602.16
6-19	13.01	602.04
6-28	12.94	602.11
Destroyed		

Piezometer No.: Fa 50-1
145-082-03 ABD
Wellhead Elevation: 595.64 m
Ground Elevation: 594.87 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-30	9.33	586.31
2-21	inaccessible	
3-29	9.00	586.64
5- 4	8.90	586.74
5-15	8.94	586.70
5-31	9.06	586.58
6-19	9.14	586.50
6-28	9.16	586.48
7-18	9.30	586.34
7-31	9.31	586.33
9- 8	9.36	586.28
10- 2	9.38	586.26
10-31	9.28	586.36
11-29	9.35	586.29
12-27	9.33	586.31

Piezometer No.: Fa 74-2
146-082-34 CDD
Wellhead Elevation: 596.74 m
Ground Elevation: 596.30 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	4.77	591.97
2-21	4.77	591.97
3-22	4.69	592.05
5- 4	4.21	592.53
5-30	4.30	592.44
6-19	4.34	592.40
6-28	4.37	592.37
7-18	4.43	592.31
10- 2	4.53	592.21
12-27	4.61	592.13

Piezometer No.: Fa 77-2
146-082-22 ADD
Wellhead Elevation: 618.66 m
Ground Elevation: 618.10 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	8.18	610.48
2-21	8.29	610.37
3-22	8.40	610.26
5- 3	7.31	611.35
5-30	7.28	611.38
6-19	7.43	611.23
6-28	7.59	611.07
7-28	7.81	610.85
7-31	8.06	610.60
9- 8	8.28	610.48
10- 2	8.24	610.42
10-30	8.26	610.40
11-29	dry	
12-26	8.41	610.25

Piezometer No.: Fa 78-2
146-082-27 DCC
Wellhead Elevation: 611.67 m
Ground Elevation: 610.78 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	inaccessible	
2-21	inaccessible	
3-28	15.56	596.11
5- 3	15.46	596.21
5-30	15.41	596.26
6-19	15.29	596.38
6-28	15.38	596.29
7-18	15.39	596.28
10- 2	15.35	596.32
12-26	15.42	596.25

Piezometer No.: Fa 80-3
146-082-20 CDD
Wellhead Elevation: 632.40 m
Ground Elevation: 631.60 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	dry	
2-21	dry	
3-22	dry	
3-29	frozen shut	
5- 4	2.87	629.53
5-30	3.35	629.05
6-28	4.30	628.01
7-18	4.74	627.66
7-31	4.94	627.46
9- 7	5.37	627.03
10- 3	5.63	626.77
10-31	5.74	626.66
11-29	6.07	626.33
12-26	6.35	626.05

Piezometer No.: Fa 81-2
146-082-20 CBB
Wellhead Elevation: 627.23 m
Ground Elevation: 626.53 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-30	21.03	606.20
2-21	inaccessible	
3-28	20.95	606.28
5- 3	21.03	606.20
5-30	20.99	606.24
6-19	20.83	606.40
6-28	20.93	606.30
7-18	20.97	606.26
10- 3	20.96	606.27
12-26	20.94	606.29

Piezometer No.: Fa 84-2
146-083-24 BBB
Wellhead Elevation: 619.53 m
Ground Elevation: 618.84 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	inaccessible	
2-21	inaccessible	
5- 3	15.49	604.04
5-30	15.47	604.06
6-28	15.47	604.06
10- 2	15.43	604.09
12-26	15.40	604.12

Piezometer No.: Fa 86-2
146-082-30 CCC
Wellhead Elevation: 609.29 m
Ground Elevation: 608.79 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-29	5.04	604.25
2-21	inaccessible	
2-26	5.25	604.04
3-28	muddy bottom	
5- 3	1.44	607.85
5-15	1.59	607.70
6- 1	1.98	607.31
6-19	2.30	606.99
6-28	2.42	606.87
7-18	2.65	606.64
7-31	2.85	606.44
9- 7	3.18	606.11
10- 3	3.38	605.91
10-31	3.49	605.80
11-29	3.74	605.55
12-27	3.94	605.35

Piezometer No.: Fa 86-3
146-082-30 CCC
Wellhead Elevation: 609.38 m
Ground Elevation: 608.72 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-29	22.00	587.38
2-21	inaccessible	
2-26	21.92	587.46
3-28	21.87	587.51
5- 3	21.87	587.51
5-15	21.85	587.53
6- 1	21.85	587.53
6-19	21.85	587.53
6-28	21.88	587.50
7-18	21.90	587.48
7-31	21.89	587.49
9- 7	21.93	587.45
10- 3	21.95	587.43
10-31	21.89	587.49
11-29	21.94	587.44
12-27	21.94	587.44

Piezometer No.: Fa 88-4
146-082-30 DAA
Wellhead Elevation: 617.05 m
Ground Elevation: 616.70 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	4.87	612.18
2-21	inaccessible	
3-28	5.04	612.01
5- 3	4.37	612.68
5-15	4.27	612.78
5-30	4.24	612.81
6-19	4.19	612.86
6-28	4.32	612.73
7-18	4.47	612.58
7-31	4.60	612.45
9- 7	4.57	612.48
10- 3	4.86	612.19
10-31	4.73	612.32
11-29	5.04	612.01
12-26	5.06	611.99

Piezometer No.: Fa 91-2
146-082-32 CCC
Wellhead Elevation: 613.20 m
Ground Elevation: 612.80 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	dry	607.17
1-30	dry	607.15
2-21	5.91	607.29
3-22	5.94	607.26
3-29	5.93	607.27
5- 3	5.21	607.99
5-17	5.35	607.85
5-30	5.52	607.68
6-26	5.86	607.34
7-18	5.91	607.29
7-31	dry	607.16
9- 7	5.92	607.28
10- 3	5.94	607.26
10-31	5.69	607.51
11-29	5.90	607.30
12-27	5.91	607.29

Piezometer No.: Fa 95-4
146-082-15 DDD
Wellhead Elevation: 624.22 m
Ground Elevation: 624.07 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	inaccessible	
2-21	inaccessible	
3-22	inaccessible	
3-29	1.55	622.67
5- 3	0.80	623.42
6- 2	1.77	622.45
6-19	2.26	621.96
7-18	2.45	621.77
7-31	2.49	621.73
9- 8	2.80	621.42
10- 2	2.94	621.28
10-30	2.85*	621.22
11-29	3.13*	620.94
12-26	1.47*	622.60

*Measured from ground level

Piezometer No.: Fa 96-1
146-082-15 BBB
Wellhead Elevation: 632.71 m
Ground Elevation: 632.23 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	6.72	625.99
2-21	6.76	625.95
3-22	6.85	625.86
5- 3	6.61	626.10
5-30	6.32	629.39
6-19	6.17	626.54
6-28	6.20	626.51
7-18	6.23	626.48
7-31	6.24	626.47
9- 8	6.30	626.41
10- 2	6.33	626.38
10-30	6.36	626.35
11-29	6.46	626.25
12-26	6.50	626.21

Piezometer No.: Fa 98-1
146-082-07 DDD
Wellhead Elevation: 624.05 m
Ground Elevation: 623.49 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-30	inaccessible	
2-21	inaccessible	
3-28	5.79	618.26
5- 3	3.14	620.91
5-30	3.29	620.76
6-19	3.45	620.60
6-28	3.61	620.44
7-18	3.89	620.16
7-31	4.12	619.93
9- 7	4.48	619.57
10- 2	4.58	619.47
10-31	4.67	619.38
11-29	4.95	619.10
12-26	4.98	619.07

Piezometer No.: Fa 99-1
146-083-24 AAA
Wellhead Elevation: 622.66 m
Ground Elevation: 621.94 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	6.54	616.12
2-21	6.50	616.16
3-22	6.52	616.14
5- 3	6.52	616.14
5-15	6.45	616.21
5-30	6.40	616.26
6-19	6.36	616.30
6-28	6.33	616.33
7-18	6.32	616.34
7-31	6.27	616.39
9- 7	6.24	616.42
10- 2	6.23	616.43
10-31	6.16	616.50
11-29	6.28	616.38
12-26	6.13	616.53

Piezometer No.: Fa 100-1
146-082-20 BBB
Wellhead Elevation: 625.98 m
Ground Elevation: 625.29 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	4.07	621.91
2-21	4.20	621.78
3-22	4.34	621.64
5- 3	2.24	623.74
5-15	1.98	624.00
5-30	2.09	623.89
6-19	2.33	623.65
6-28	2.30	623.68
7-18	2.59	623.39
7-31	2.53	623.45
9- 7	2.73	623.25
10- 2	2.98	623.00
11-29	3.27	622.71
12-26	3.35	622.63

Piezometer No.: Fa 106-5
146-082-21 CCC
Wellhead Elevation: 622.89 m
Ground Elevation: 622.08 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	inaccessible	
2-21	inaccessible	
3-28	inaccessible	
5- 4	11.67	611.22
6- 1	11.47	611.42
6-19	11.53	611.36
6-28	11.65	611.24
7-18	11.75	611.14
7-31	11.81	611.08
9- 7	11.90	610.99
10- 2	11.98	610.91
10-31	11.98	610.91
11-30	12.05	610.84
12-26	12.11	610.78

Piezometer No.: Fa 107-4
145-082-04 AAA
Wellhead Elevation: 600.07 m
Ground Elevation: 599.34 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-30	inaccessible	
2-21	inaccessible	
3-29	inaccessible	
5- 3	10.81	589.26
5-30	10.45	589.62
6-19	10.47	589.60
6-28	10.51	589.56
7-18	10.58	589.49
10- 2	10.95	589.12
12-27	11.51	588.56

Piezometer No.: Fa 108-3
146-082-32 ADD
Wellhead Elevation: 610.44 m
Ground Elevation: 609.79 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	dry	604.62
2-21	dry	
3-22	2.84	607.60
3-29	4.34	606.10
5- 4	5.35	605.09
5-17	5.70	604.74
6- 1	dry	
6-19	dry	
6-28	dry	
7-31	dry	604.49
9- 7	dry	
10- 2	dry	
10-31	dry	
11-30	dry	
12-28	dry	604.50

Piezometer No.: Fa 110-3
146-082-20 DAD
Wellhead Elevation: 613.1 m
Ground Elevation: 612.4 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-17	2.29	610.81
2-21	inaccessible	
3-22	1.89	611.21
5- 4	1.73	611.37
6- 1	1.98	611.12
6-19	1.88	611.22
6-29	2.00	611.10
7-18	2.08	611.02
7-31	1.98	611.12
9- 7	2.08	611.02
10- 2	2.11	610.99
10-31	2.05	611.05
11-29	2.24	610.86
12-28	2.26	610.84

Piezometer No.: Fa 111-5
146-082-33 DCC
Wellhead Elevation: 595.80 m
Ground Elevation: 595.40 m

Date 1979	Depth below wellhead (m)	Water level elevation (m)
1-30	5.72	590.08
2-21	inaccessible	
3-27	5.96	589.84
5- 3	4.58	591.22
5-15	4.33	591.47
5-30	4.30	591.50
6-19	4.29	591.51
6-28	4.31	591.49
7-18	4.39	591.41
7-31	4.45	591.35
9- 7	4.65	591.15
10- 2	4.77	591.03
10-31	5.07	590.73
11-29	5.08	590.72
12-27	5.41	590.39