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Groundwater recharge in an area near Ardmore, Alberta.

By

John Horgan ©

A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree of Master of Science.

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The undersigned certify that they have read, and recommended to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled **Groundwater recharge in an area near Ardmore, Alberta** submitted by **John Horgan** in partial fulfillment of the requirements for the degree of Master of Science.

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Date: 1994 August 16

Dedication

To Judy, Keara and Colin.

Abstract

In a Boreal Parkland Transition Zone in East Central Alberta, groundwater moves dominantly downward from the water table through glacial till and outwash deposits of Quaternary age. Recharge has been estimated from the analyses of hydrographs, pumping tests and flow patterns within a 200 km^2 study area. The estimates are expressed in terms of: (i) groundwater accretion, defined as the process or rate of addition of water to the water table from the unsaturated zone, and (ii) groundwater recharge, defined as the process or rate of flow of water below and away from the water table within the saturated zone.

On the basis of hydrograph and flow system analyses, the estimated annual fluxes of groundwater recharge were 1.1×10^{-9} m/s and 1.5×10^{-9} m/s respectively, or 9 % and 12 % of the average annual precipitation (400 mm). The annual accretion flux, on the other hand, for the water year 1992/93 was estimated at 5.7×10^{-10} m/s or 5 % of the average precipitation (355 mm) for the same period. The differences in recharge and accretion fluxes are reflected in an average decline in water levels throughout the area. Estimated recharge rates vary between 3 and 18 % of the average annual precipitation within the area, whereas accretion rates vary between a net loss and 12 % of the precipitation during the 1992/93 water year. Recharge fluxes from the water table to the first major (semi-confined) aquifer, induced by pumping, are estimated to be less than the natural recharge fluxes calculated at the water table The discrepancy suggests that natural recharge rates at shallow depths are different locally from flow rates deeper in the groundwater regime.

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1. Introduction and review

1.1. Introduction

Measurements of water level fluctuations for the period 30 November, 1991 to 30 September, 1993 were collected in an area of approximately 200 km^2 near Ardmore, Alberta. Twenty-four University Observation Holes (UOH) and 4 previously existing Commercial Observation Holes (COH) were monitored: 18 with automatic recorders and 10 by hand. Two additional recorders were installed towards the end of the study.

The groundwater flow pattern was analyzed with the following information: i) the monitoring data, ii) data from 260 existing water wells, iii) previous geological studies, and iv) field mapping.

Rates and processes of groundwater recharge were examined with the available data. The recharge process has been conceptualized as a two component process involving both the addition and transport of water. The term accretion is defined as the process of water addition to the water table from the unsaturated zone, whereas recharge is reserved to describe the transport of water within the saturated zone (Sec. 5.1). Estimates were made of fluxes for each of the components separately (Ch. 6).

Three approaches have been developed in the thesis for preliminary estimates of groundwater recharge, namely those based upon the analysis of hydrographs, flow patterns, and pumping tests; all of which were suited to the Ardmore area (Ch. 5). The greatest emphasis was placed upon the hydrograph approach because it made use of data collected during the study. A comparison has been made between the results from each approach (Sec. 6.4).

1.1 Purpose and scope

The purpose of the study was to examine the processes of groundwater recharge and make preliminary estimates of recharge rates. An attempt was made to compare the estimates made by the hydrograph approach with estimates from two other approaches.

The underlying principle for the study is that the most appropriate place to study the processes and rates of recharge is at the site of its occurrence, i.e. at the water table. With careful monitoring, it is possible to identify those rises and falls in the water table that are directly related to recharge. The water level movements represent changes in storage in the saturated zone related to recharge, and can be quantified with the correct approach.

Among the other considerations for developing approaches for preliminary estimates of groundwater recharge from water table fluctuations, two were paramount: first, water level measurements in shallow observation holes are relatively simple and inexpensive to obtain and second, in the early stages of many projects, water levels are the only information available. In addition, the general lack of well established approaches for estimating groundwater recharge and the paucity of estimates from this part of Alberta were strong incentives for the study.

The field work was designed, therefore, to collect continuous water level data from a number of observation holes, some of which were installed for this study and some of which were existing. Surface manifestations of groundwater were also examined through field mapping and soil profile descriptions.

The groundwater flow pattern was described through the analysis of the monitored data, together with all other water level data from the area. Most of the other data were available from government archives and consisted of water well and test hole records for both private water supplies and commercial projects. Hydraulic parameters were derived from existing pumping test data and earlier reports.

1.3. Choice of area

Water levels in wells and observation holes from the Ardmore area were known to show seasonal fluctuations that were apparently closely related to variations in water supply at the ground surface (Union Texas of Canada Ltd. 1977, 1978; Troy, 1989). Much of this knowledge came from reports of monitoring for water supply to the Ardmore Thermal Project, owned and operated by a succession of oil companies. In one of the reports, Troy (1989) concluded that much of the groundwater flow beneath the area is recharging; in other words directed downwards away from the water table. The presence of good seasonal responses to surface water inputs, some historical records and the downward flow, made the area particularly attractive for our purposes.

Ardmore is located in the Beaver River basin as described in Section 2. Unlike larger rivers the Beaver River cannot provide a reliable supply of water. Local water supplies for municipalities, rural domestic users and industrial projects are all sourced from either lakes or from groundwater. Groundwater recharge is therefore an important issue for the region.

The volume of groundwater removed within the Ardmore study area, however, remains small and since the cessation of the above mentioned pilot project in 1978, is restricted to domestic supplies for some 300 farms and 400 villagers in the hamlets of Ardmore and Fort Kent.

On the basis of its hydrogeologic setting we were able to define suitable boundaries for the area (Fig. 1). To the north and east the base of the Beaver River valley lies some 30 m below the general height of land and represents a line sink for groundwater. It also represents a boundary of symmetry for subsurface flow with hydraulic potentials approximately equal on either side of it. To the south Muriel Creek lies in a broad shallow valley, and whereas the valley does not provide the clear groundwater divide that is present below the Beaver River, an analysis of the groundwater flow pattern indicates that flow is approximately parallel with it for most of its length. The same is true for the west boundary, a broad low lying area occupied by a ditch, north of Fort Kent.

Finally, few estimates of groundwater recharge have been made in the Parkland zone in Alberta (Sec 2.2), whereas there are studies from the zone in Saskatchewan (Freeze, 1969b; Keller et al., 1988) which may be used for comparison.

1.4. Background

It is widely accepted that groundwater forms the subsurface portion of the hydrologic cycle (Linsley et al., 1975; Todd, 1980) and that the ultimate source for most groundwater is meteoric.

For both theoretical and practical reasons the rate of groundwater replenishment from the surface is of considerable interest. Historically the main driving forces behind research have been related to groundwater supply and withdrawal rights (Meyboom, 1967b; Kitching et al., 1977; Sophocleous, 1992). More recently it has been recognized that the role of wetlands in the hydrologic cycle (Siegel, 1988; Boeye and Verheyen, 1992), soil moisture and salinity control (Tóth and Gillard, 1988; Stein and Schwartz, 1990) and the potential for groundwater contamination (Keller et al., 1988) all require calculations of natural recharge rates. Recharge rates are also an essential input parameter for numerical modeling of groundwater flow and transport (Anderson and Woessner, 1992).

Some questions arise, however, as to the importance and relevance of natural recharge rates in determining water supply sustainability. The argument is made that the real parameter of interest is the potential for higher artificial rates, induced in response to a new dynamic equilibrium, which results from exploitation (Bredehoeft et al., 1982). However, if a truly integrated resource management approach is adopted, where questions of surface moisture conditions are included in the criteria used in defining safe yield (Freeze, 1971; Domenico and Schwartz, 1990), knowledge of natural recharge rates must be required.

Natural recharge occurs at the water table that, in a recharge area, is usually located one to several meters below ground. Recharge is therefore difficult to measure without disturbing the natural processes. The recharge rates are nevertheless known to vary widely with time, location and depth (when speaking of aquifer recharge). They are affected by such diverse variables as ground cover, slope and aspect, soil permeability and moisture content, climate and the depth to the water table. Sophocleous (1992) found in the Great Bend Prairie of Central Kansas that the most influential variables affecting recharge are, in order of decreasing importance: total annual precipitation, average maximum soil-profile water storage during the spring months, average shallowest depth to the water table during the same period, and spring time rainfall rate.

A number of approaches have been developed to measure natural recharge (Simmers, 1988; Lerner, 1990). All have serious limitations and most can only be applied under suitable circumstances. They can be broadly classified as physical, chemical and numerical approaches.

The group of physical approaches is perhaps the largest and includes the following: i) direct measurements of recharge, ii) estimates using a water balance approach, iii) recharge rates derived from flux estimates either in the saturated or the unsaturated zone and iv) estimates from empirical relationships.

The chemical group involves chiefly the use of tracers, either natural or artificial and either as dated and labeled parcels or as mass balances of tracer (Lerner 1990).

Numerical modeling is not always considered a separate group but rather a set of techniques used to assist the application of other approaches Conceptually the models often fall into two subgroups, models which use a fixed water table and models which use an estimated flux as the upper boundary. In applying the second group, it becomes quickly apparent how detailed our knowledge of the spatial distribution of recharge rates must be and how inappropriate is the usual assumption of a spatially averaged rate.

The present study uses only physical approaches in the preliminary estimates and in this it follows a tradition of previous studies of recharge in North America. A direct measurement approach involving the construction of lysimeters (Kitching et al., 1977) was considered but rejected. Lysimeters are expensive to construct and in the end give accurate estimates for only a small area. For the Ardmore study several lysimeters would have been required.

The water balance approach has been applied to estimates of groundwater recharge in a number of studies on the Canadian Prairies (Meyboom 1961, 1966, and 1967b). The approach seeks to account for all the inputs and outputs to a basin, or to any defined hydrologic unit, over some time period. The governing equations usually follow the form,

Input = Output ± Change in Storage.

A solution for groundwater recharge is obtained by summing all the other sources and sinks of the unit in an equation such as the following:

$$R = P - ET - G \pm \Delta G \pm \Delta S \qquad [L^3] \qquad (1.1)$$

where R is groundwater recharge, P is precipitation, ET is evapotranspiration, G is groundwater outflow, ΔG is groundwater storage and ΔS is soil moisture storage. In this case surface runoff, changes in surface water storage and lateral movement of water have not been considered.

If the time period chosen is sufficiently long, the storage changes in the saturated and unsaturated zones will often cancel and the problem becomes a little easier. For short time periods, however, it is usually the inability to make reasonable estimates of both the storage changes and evapotranspiration which limit the applicability of the approach. Our time period was short and changes in storage in the unsaturated zone were important and obviously large. Unfortunately we did not have the capability to measure moisture changes in the unsaturated zone.

Estimates using empirical relationships in which recharge is related to some other variable, usually precipitation, often form the only approach available in the absence of good hydrogeological information (Lerner, 1990). For example, results in Table 1 indicate that recharge on the Plains Region may vary between about 1 and 63 % of the average annual precipitation. We might then develop an empirical relationship about the median value and say that recharge on the prairies is usually about 9 % of the average annual precipitation. This may be a reasonable figure to apply on a basin scale where no other information is available.

Table 1. Recharge estimates in the Northern Interior Plains Region.

Author	Location	Depth (mm/y)	% of av. annual Precipitation.
Meyboom, (1966) (1967b)	Saskatchewan Sask./Man.	10.67 2.2 - 38.1	3 0.5 - 7.5
Freeze, (1969b)	Saskatchewan	0.254 - 254	0.06 - 62.5
Tóth, (1968)	Alberta	34.3	9.0
Rhem et al., (1982)	N. Dakota	10 - 40	2.0 - 9.0
Trudel et al., (1986)	Alberta	18-22	5.5-6.3
Zebarth et al., (1989)	Saskatchewan	35	10.0
Keller et al., (1988)	Saskatchewan	0.3 - 35	0.09-10.1

Estimates of fluxes solve for recharge directly rather than as a remainder once all the other sources and sinks are accounted for. The approaches we developed for the Ardmore area belong to this group so we take time now to look at some earlier work which has application to the Ardmore study.

In a report on the hydrogeology of Three Hills Area, Alberta, Tóth (1968) used a saturated flow approach, under steady state conditions, to estimate a number of basin parameters. Building on the concept of a natural basin yield (Freeze, 1966) and assuming a complete exchange of water between recharge and discharge areas during months of ground frost, he calculated the volume of groundwater flow through a unit portion of the basin.

The natural yield of a flow system for a given time interval was expressed as (Tóth, 1968).

$$Q = S_y \frac{dh}{dt} A \qquad [L^3/T] \qquad (1.2)$$

where, Q is the flow rate, dh is the average change in water levels over the area, A, of either upward or downward flow, S_y is the average specific yield of the rock at points of water level measurement and dt is a time interval.

Providing that the seasonal changes in hydraulic head in the flow system were small in comparison to the average gradient, the winter flow rates could be extended over a full year. Assuming also that the basin contained only one flow system the total volume, converted to an average annual depth of groundwater transfer (34.3 mm/y), represented 9% of the annual precipitation in the Three Hills Area.

A comprehensive groundwater study was carried out by Freeze (1969b) on the Good Spirit Lake Drainage Basin, Saskatchewan in which he estimated recharge rates with the aid of a numerical model. Profiles were constructed through the three dimensional model and fluxes were calculated through the application of Darcy's law at each water table node. Permeability values were assumed from the geology whereas the hydraulic gradient was determined directly from the model. Using several such profiles, a map of recharge and discharge fluxes was constructed. It was then possible to calculate total basin recharge by a summation of the recharge areas multiplied by their associated fluxes. The estimates were between 1 and 5 inches (25.4-127 mm) per year on a sand-gravel plain, but were less than 0.1 and more often less than 0.01 inches (2.54 and 0.254 mm) per year on the till portion of the basin.

The generation of modeled profiles for theoretical flow regimes (Freeze and Witherspoon, 1968, Fig. 1.1), showed very clearly the degree of spatial variability in groundwater recharge. Where it can be applied, modeling may be the best method of spatially distributing point estimates of fluxes.

In the estimates discussed thus far, the authors have been primarily concerned with flow rates in the saturated zone which they presume to be sustained, on an annual basis (or longer), by water additions at the water table (Freeze and Banner, 1970b). Since water arrives at the water table from the unsaturated zone to initiate recharge, the fluxes here are of equal interest. This fact has been recognized for a number of years and has lead to the development of models which are capable of solving equations of flow under saturated or unsaturated conditions (Freeze, 1971).

In the analyses of water table fluctuations, which is the primary goal for the Ardmore study, both the rate of water addition at the water table as well as the rate of water flow away from it, must be taken into account. A study of groundwater recharge to a lignite seam in Underwood, North Dakota (Rehm et al., 1982) can be used to illustrate this point. They found that the vertical fluxes close to the water table estimated from hydraulic gradients and permeability in piezometers, were much higher than the fluxes estimated from rates of water level rises multiplied by specific yield. They point out that, "this large difference is probably the result of the aforementioned condition of having flux rates away from the water table in the saturated zone that are as large as or larger than the vertical flux rates in the unsaturated zone". They did not attempt to separate the two fluxes using hydrograph data and may, as a result, have missed some valuable information.

Trudell and coworkers (1986), studying recharge rates at the reclaimed Diplomat mine-site in east-central Alberta, also discussed the problem of distinguishing between rates of water addition from above and rates of flow below and away from the water table. They noted that deeper piezometers located more than 2 m below ground surface, responded with a more or less continuous rise following either single or multiple surface infiltration events. The relatively thick unsaturated zone tended to dampen or smooth the recharge process. In the shallow holes, however, they noted a larger amplitude of water level response to individual infiltration events. In many cases, though, a rapid rise was followed by a rapid fall in water levels which prompted the following observation: "Consequently, although there is a very large apparent recharge, a stable or long term increase in the water table elevation is unlikely to be produced, and a small amount of drying will readily remove the newly added water, causing a relatively rapid decline in the water-table to a lower elevation". These observations prompted them to exclude some of the more transient water level peaks from their estimates of cumulative recharge.

A comparative study of approaches to recharge estimates undertaken in Sweden (Johansson, 1987) lays the clearest ground for the Ardmore study. In the Swedish study, the recession method of hydrograph analysis was used. A number of authors have suggested that the rate of water level decline during winter months (winter recession) can be viewed, under certain circumstances, as being solely due to groundwater flow away from the water table in the saturated zone (Tóth, 1968). A similar concept lies behind the use of a night time rate of fall in pond levels or in very shallow water tables (White, 1932; Meyboom, 1967b; Keller, 1985; Zebarth, 1989) when evapotranspiration is assumed negligible. The concept is simply that during periods of ground frost, losses and gains to and from the saturated zone from precipitation and evapotranspiration respectively, are at a minimum and may be considered negligible. The method then uses the steepest parts of the winter recession to estimate the vertical component of groundwater flux. The added benefit of the method is that during the relatively long drain-out period, a constant specific yield may be applied (see below). In the recession method used by Johansson the rate of water level fall is added to the rate of water level rise in order to arrive at a recharge rate. This is done for a given event or some other convenient time interval (14 days for Johansson's study). The method does not, of course, answer the question of what storage term to assign to the water level rise.

Many observers have discussed the problem of assigning a suitable storage term to convert water level rises in wells to equivalent depths of water (Gillham, 1984; Trudell et al., 1986; Sophocleous, 1991). The specific yield concept defined by Bear (1979) as, the average amount or water per unit volume of soil drained from a soil column extending from the water table to the ground surface, is the most useful for our purposes. The most obvious difficulty is that to release this average amount of water takes time and varies with the hydraulic properties of the soil water system. Even if this average amount is known from experimentation, it often happens that a new pulse of infiltrated water is added to the unsaturated zone before complete drainage has occurred, so that the actual storage at the time of the addition is not known. Thus the specific yield in Equation (1.2) changes with time and position in the soil profile. If good information is not available as the soil moisture profile develops during wetting or drying, the choice of an effective specific yield can become quite arbitrary.

The problem of a storage term has lead most researchers to avoid the use of hydrographs, on their own, for estimates of groundwater recharge (Freeze and Bannner, 1970b; Johansson, 1987; Sophocleous, 1991).

Finally, for background, we indicate that many of the studies of groundwater recharge on the prairies have been conducted at the scale of a single wetland (Meyboom, 1966; Williams, 1968; Zebarth et al., 1989; Anderson and Cheng, 1993). The importance and mechanisms of recharge beneath seasonally water-filled depressions in the glaciated landscape of large parts of the Northern Plains is quite well known. Recharge is apparently focused at these locations, particularly in areas of hummocky moraine with poorly integrated drainage. Zebarth and others (1989), reported that essentially all recharge in a study area near Hafford, Saskatchewan, occurred below sloughs which occupied only 5 % by area of the landscape. The landscape around Ardmore, however, has been largely modified by post glacial erosion and drainage. The topography is, for the most part, gently undulating and we felt that depression-focused recharge might be less important here.

1.5. Statement of the problem

The currently held concepts of natural recharge generally agree that the process begins with the addition of water at the water table, and culminates in the flow of water below and away from the water table within the saturated zone (Freeze and Cherry, 1979). The water is made available at the water table by percolation from the land surface. The total volume of water that crosses the land surface, however, is usually greater than the volume available at the water table; the difference going largely to changes in soil moisture and losses to evapotranspiration.

Water flows from regions of high towards regions of low fluid potentials (Hubbert 1940). Where boundary conditions exist, as they do in many small drainage basins, water enters a flow system below areas of recharge and leaves through areas of discharge (Tóth 1962). In recharge areas, flow is directed downwards away from the water table, whereas in discharge areas the flow is towards the water table. A dynamic equilibrium must exist, between water gains and losses, for these flow systems to persist. Over long time periods, many systems appear to be in steady state since the water table is observed to occupy some mean position below land surface. As a result, the water table is often a subdued replica of the topography (Tóth 1963). These facts have lead many researchers to study the saturated zone in isolation, reasoning that if the basin is in steady state over long time periods, then the mean recharge must equal the mean discharge. This reasoning provides the basis for the concept of a natural basin yield (Freeze and Witherspoon, 1968). A further corollary is that the saturated flow rates must equal the long term average rate of water supply from the unsaturated zone. Thus from a saturated flow perspective, the mean recharge flux at the water table can be derived by dividing the recharge rate by the area of recharge for the flow system.

It is usually argued in these analyses that the ratio of the water level changes to the total head difference in all but the most local flow systems is small. Therefore, the hydraulic gradient, and thus the flow rate, will not be substantially affected by movements of the water table.

Observations of changes in the water table position over shorter time periods suggest that quite transient flow conditions actually exist. The water table changes represent changes in groundwater storage resulting from an imbalance between the rate of supply and the rate of loss of water in the system (Freeze and Banner, 1970b). This must imply that the process of recharge is a coupling of two separate components. The first component is the process of supply at the water table and the second is the process of water transport below and away from the water table. There must also be separate rates associated with each component, otherwise no imbalance could occur and from earlier statements we see that the rates of water supply to the water table and of infiltration at the land surface are not equivalent.

Examination of the changes in water table position must play an important part in our understanding of the mechanisms of recharge, and should in some cases lead to quantifiable estimates of rates. Water table fluctuations represent the cumulative effects of the different processes involved. To examine the processes, we can conceive a groundwater balance within a volume element positioned at the water table in a recharge area (Fig. 1.2). An equation representing the dynamic equilibrium controlling the storage in the saturated zone and thus the position of the water table can be written as follows

$$S_y A dh/dt = I_g - R_z - ET_g, \qquad [L^3/T] \qquad (1.3)$$

where
S_y = the specific yield of material at the interface,
dh = a change in water level,
dt = a time period,
A = the horizontal cross section area of the volume element,
I_g = the rate of water added at the water table,
R_z = the vertical component of saturated flow,
ET_g = the rate of evapotranspiration from the water table.

The two components of the recharge process are identified as I_g and R_z in Equation (1.3) and represent the water addition and flow away from the water table respectively. Following an estimate of the storage term on the left hand side of the equation, the problem addressed by the thesis is the development of approaches for estimating all the terms on the right hand side such that a solution can be found for the two recharge components.

Only those water level changes associated directly with the recharge processes may be utilized. Water level changes resulting from other causes must be separated; such as those caused by atmospheric pressure differences. The separation can be achieved with continuous monitoring in all seasons.

1.5.1. Introduction to problem solution

Following a detailed description and interpretation of the flow regime three approaches are developed that are suited to the Ardmore area. The first approach uses two methods of hydrograph separation previously developed for the estimation of evapotranspiration from diurnal water level fluctuations. The methods are extended and applied to periods up to a complete water year. A fundamental assumption is that the winter recession can be equated with the saturated flow rate during periods of ground frost. The difficulty in choosing the correct storage term is addressed largely by using measurements from relatively deep water tables which change position slowly, thus giving time for complete drainage.

In the second approach a flow system is identified which can be used to approximate the entire flow regime beneath the Ardmore area. A steady state model of the flow system is utilized, following the reasoning outlined above, and discharge through a midline section is equated to total recharge over the surface of the study area.

In the third approach an attempt is made to estimate recharge to an aquifer some 30 m below the water table using the total withdrawal rate divided by the area from within which water is being diverted to pumping.

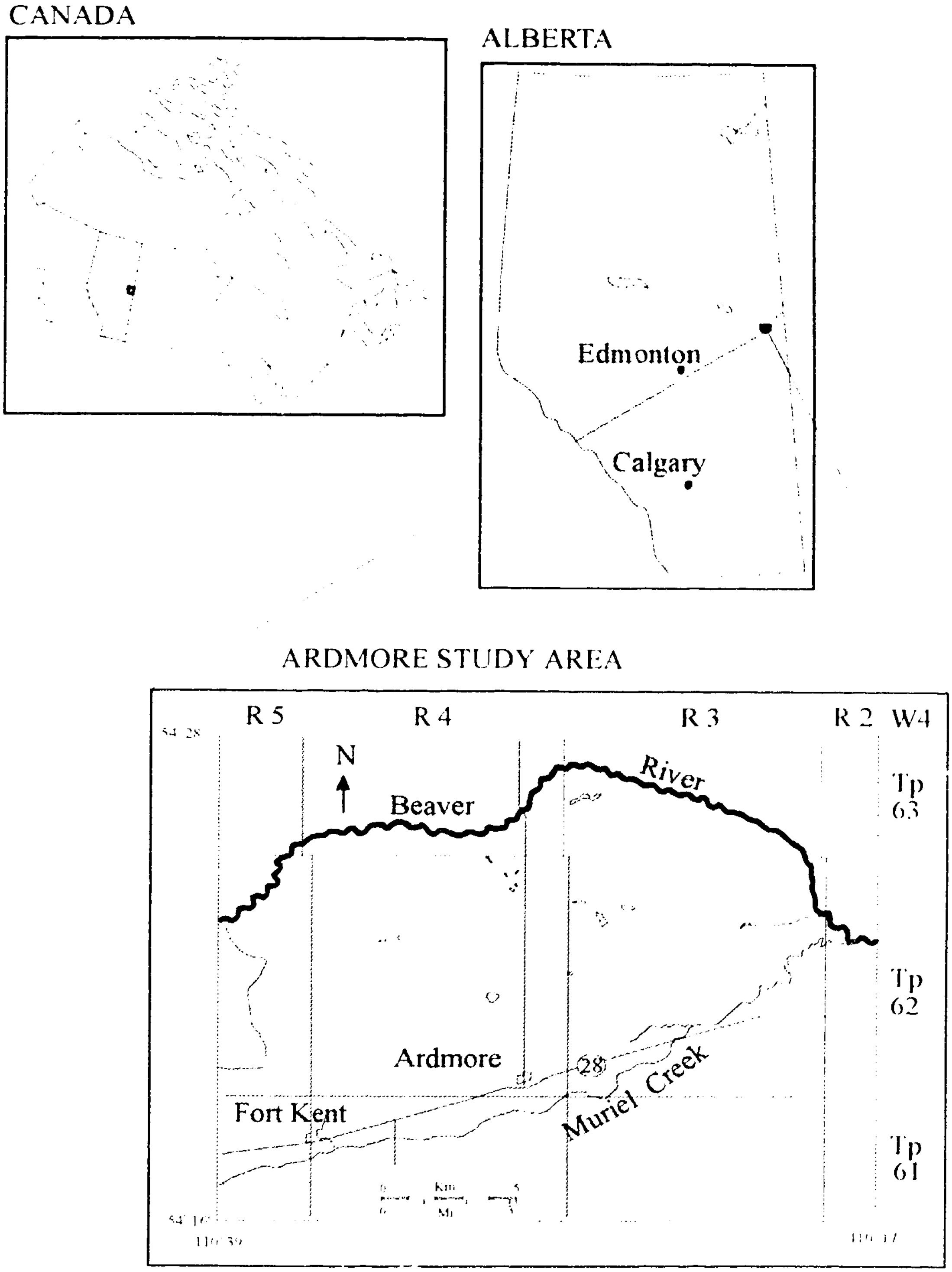


Figure 1. Study area location.

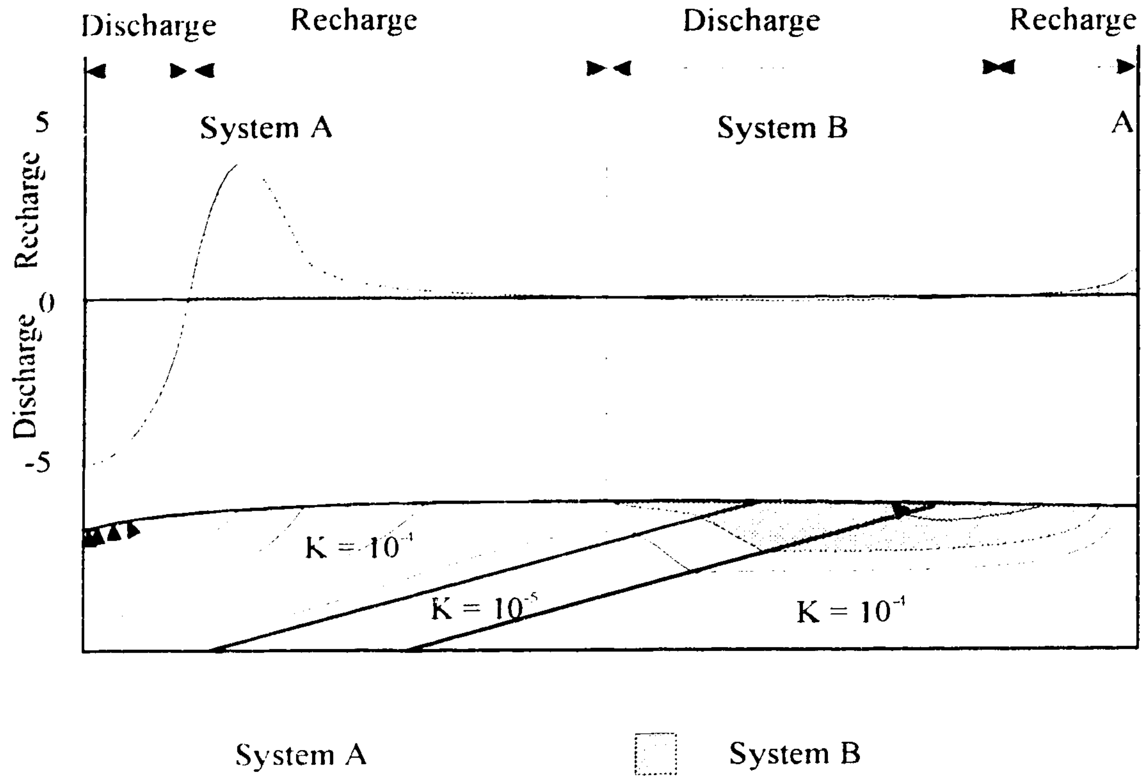


Figure 1.1. Modeled flow systems profile. Lower portion shows sketched flowlines in two hypothetical flow systems A and B. The flow regime is developed in a medium with hydraulic conductivity of 1 x 10^{-4} m/s. A bed with lower hydraulic conductivity (1 x 10^{-5} m/s) dips across the basin from right to left. The upper portion shows the calculated vertical fluxes in each flow tube; positive into the flow system. The spatial variability of recharge and discharge is clearly shown. (Modified after Freeze & Witherspoon, 1968).

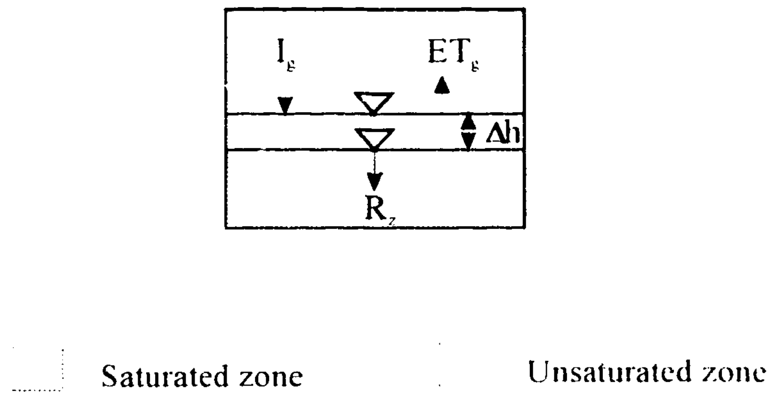


Figure 1.2. Volume element at the water table in a recharge area. The vertical water exchange is shown between the saturated and unsaturated zones; I_g is accretion, ET_g is evapotranspiration, R_z is recharge and delta h, changes in the water level (storage). (Modified after Kovács, 1981).

2. Description of the study area

2.1. Location

The area is located in east central Alberta, approximately 15 km from the Saskatchewan provincial boundary. It is covered by map sheets 73 L/7 and L/8 of the National Topographic System and mostly contained within township 62, ranges 3 and 4, west of the 4th meridian (Fig 1.0). It is bounded in the north and south by latitudes 54° 28' N and 54° 16' N and in the east and west by longitude 110° 17' W and 110° 39' W respectively. The hamlets of Fort Kent and Ardmore are located near the southern boundary on provincial highway No. 28, that runs between the towns of Bonnyville in the west and Grand Center in the east. The study area covers roughly 200 km^2 and is 290 km by road northeast of Edmonton.

2.2. Climate and Vegetation

The climate is classed as short, cool summer (Dc) Koeppen zone (Longley, 1972). In the Koeppen system, cool summers have mean temperatures of the warmest month below 22°C and less than four months with mean temperatures of 10°C or more. Mean monthly temperatures from the study area for 1992 suggest that this summer would not quite fit on the basis of the last criterion since it had 5 months with mean temperatures of 10°C or more (Fig 2a).

The frost-free period in 1992 lasted between May 25 and Aug. 23 inclusive, a total of 90 days, which is exactly equal to the long term average for this area (Chapman and Brown, 1966). The mean annual temperature was 2.0°C for the period 1977 to 1991, which is almost one degree higher than the long term average for the nineteen year period beginning in 1952 (Ozoray et al., 1980).

The average annual precipitation is slightly over 400 mm with about 70% falling as rain. The total precipitation for 1992 ranged between 355 and 358 mm in the study area (the figures are a compilation of readings taken during the summer from the study area rain gauges and from Cold Lake at all other times), whereas it was 371 mm at the Cold Lake meteorological station.

Figure 2a shows the mean monthly precipitation for the years 1977 to 1991 inclusive and the mean monthly precipitation for 1992. It can be seen that the amount of precipitation was less than the long term average throughout the summer of 1992. Also shown in the figure is the mean monthly areal evapotranspiration calculated using the complementary relationship areal evapotranspiration (CRAE) model, (Morton, 1983) averaged over the period 1954 to 1982 (Bothe et al., 1990). The greatest amount of precipitation usually falls during the growing season with the maximum occurring in July. This is also the period of maximum temperatures and evapotranspiration rates. The area is therefore sensitive to precipitation shortages at this time.

Figure 2b shows average precipitation for the sixteen years from 1977 to 1992. The data are shown in the upper curves for each complete year with the period mean. The middle and lower curves are for the periods April to October, when most precipitation falls as rain, and the period November to March, when precipitation falls mainly as snow, respectively. The upper curve shows both 1991 and 1992 were lower than average years for total precipitation. The middle curve shows below average rainfall in summer for each of the years 1990 through 1992. Snowfall on the other hand appeared to remain relatively constant.

In summary the area appears to have experienced higher temperatures and lower precipitation than normal, at least in the years just prior to and including the study period. Perhaps most importantly, rainfall through the summer months has been below average. Together with observations of drainage (Sec 2.3), the study is thought to have been undertaken during a dry cycle of the climatic regime.

The study area is included in the Subhumid Mid-Boreal ecoclimatic region (Ecoregions Working Group, 1989). This region includes a huge area of dominantly poplar forest reaching as far north as Great Slave Lake and east to Newfoundland. Ardmore is towards the southern edge of the region. From a vegetative view point the area is part of a transition zone between true Prairie Parkland in the south and Boreal forest to the north. The Parkland zone represents a large land unit which forms a band running from northern Minnesota northwest through southern Manitoba, Central Saskatchewan and Alberta (Bird, 1961). Parkland vegetation appears to alternate through time between almost complete forest cover back to grassland. Transitions probably take place quite quickly and may be related to climatic changes and forest fires. Immediately prior to widespread settlement, the Ardmore area was covered by a mixture of poplar forest and hay meadow, or marshland with willows in the wetter sites (Evans, 1920). About 80% of the land is now cleared in the west but extensive areas of jackpine woodlands are found on the sandy surficial deposits to the east. The Beaver River Valley remains largely wooded with poplar trees.

Present day land-uses in the cleared areas are cereal-crop and cattle farming. The most common cereal crops during the study were barley and canola. Approximately one third of the area is used for hay production and grazing. About one half of the hay producing area is seeded intermittently with alfalfa. Generally lower yields are obtained for cereal crops in the cultivated uplands and these areas are more commonly reserved for hay fields and pasture.

2.3. Topography and drainage

The Ardmore area is a part of the Beaver River drainage basin and is situated in a physiographic unit known as the Beaver Lowland (Andriashek and Fenton, 1989). The Beaver Lowland is described as flat to gently rolling till plain and is surrounded on almost all sides by higher ground (Fig 2.1). The study area itself is located on a slightly elevated portion of the Beaver Lowlands above the Beaver River. It has been divided into several physiographic units for ease of description (Fig 2.2). The highest elevation is reached on Ardmore Ridge at approximately 564 meters above mean sea level (m.a.s.l.). Land elevations are generally slightly higher in the west half averaging approximately 550 m.a.s.l. The average elevation in the Alexander lowlands to the east is about 535 m.a.s.l. The lowest elevation is just under 500 m.a.s.l. at the east end of Beaver River Valley. The river valley is everywhere about 30 m below the general height of land and the valley sides contain the only steep topography with slopes between 0.14 m/m (1 in 7) and 0.1 m/m (1 in 10). On the land above the river valley, slopes in the uplands rarely exceed 0.06 m/m (1 in 16) and are usually around 0.003 m/m (1 in 300), whereas slopes in the lowlands are less than this.

Long sinuous and shallow valleys, running from west to east across the area are notable topographic features on aerial-photographs. They appear to be the result of post glacial drainage and may have been formed by the Beaver river before it had located in its present valley. They often contain wet areas and sloughs (for example Tri-Lakes corridor) even though their underlying deposits are sandy.

The Beaver River rises in an upland just south of Lac La Biche. It is thus a fairly small river prone to wide seasonal and annual changes in discharge (Alberta Environment, 1983). The Beaver river drains to the Churchill River and is thus a part of the Hudson Bay drainage basin.

There is little perennial runoff from the area, although some permanent creeks are marked on the national topographic map series (Fig. 2.3). These were only observed to flow in their lowest reaches during the study, where they are greatly modified by beaver dams. Elsewhere intermittent surface water courses are mostly marked by vegetation changes. Muriel Creek, which forms the southern boundary, contained water along many reaches in the early season, but appeared to have little if any flow. The small lakes and sloughs are correctly marked as intermittent. Many of these had beaver lodges and other signs of former permanence but were dry by late summer 1992.

2.4. Geologic setting

The Bedrock in this area is Lea Park Formation; a marine mudstone of Upper Cretaceous age (Shaw and Harding, 1949). It has a conformable lower boundary with the underlying, dominantly argillaceous, Colorado group. The maximum combined thickness of this fine grained sequence in the central part of the study area is approximately 235 m below a 'bedrock high' (Amerada well, 12-25-62-4 W4). Its minimum thickness is therefore around 150 m below the deepest bedrock channel.

The bedrock has an eroded surface with a number of valleys or channels. The valleys, which are generally deeper, are thought to have been formed by drainage which pre-dated glaciation; channels on the other hand were eroded during glaciation. Two channels intersect the study area, Bronson Lake in the extreme south west corner and Big Meadow across the northern portion (Andriashek and Fenton, 1989). The remainder of the study area is underlain by a gently undulating bedrock upland which is roughly conformable with the present day topography (Fig 2.4).

The Colorado Group is underlain by rocks of the Mannville Group which are of continental margin facies. The Mannville is made up mainly of thick unconsolidated sand units which contain the Cold Lake oil sand deposit (Hitchon et al., 1989). Oil is produced within the area by enhanced methods using steam injection or, more recently, by horizontal drilling and conventional pumping. The sands are not everywhere saturated with bitumen, however, and the McMurray Formation at the base of the Mannville is almost entirely water bearing (Harrison et al., 1981).

The bedrock is covered by a sequence of Quaternary age till and sandy outwash and channel fill deposits (Fig 2.5). These have been given formation status by Andriashek and Fenton (1989). The sequence is up to 90 m thick above the north east arm of Big Meadow Channel and up to 80 m thick in Bronson Lake Channel (Fig. 2.6). Over the remainder of the study area the Quaternary sequence averages about 50 m in thickness. The distribution of the two lowermost outwash and channel fill units are determined by the bedrock topography. Sands and gravels of Empress Formation are only found in the base of Bronson Lake Channel whereas Muriel Lake Formation is more extensive, though largely absent above the bedrock upland.

2.5. Hydrogeologic setting

The Beaver River basin is about 136 km wide from north to south in the area of the Beaver Lowland (Fig 2.1). The elevation difference between the principal water divide to the south and the lowland is about 122 m. Due to the large permeability contrast between tills and channel fill deposits, much of the groundwater flow is likely to occur in the sandier units. The marine mudstone bedrock is thought to be poorly permeable and to act as a leaky barrier to flow. On the scale of the basin, the Quaternary sequence can be viewed as a thin layer in a broad, gently concave, saucer shape. A broad flat basin with an impermeable lower boundary and a relatively thin homogeneous permeable layer, can give rise to shallow cell-like local flow systems beneath an undulating topography, superimposed on a gentle regional slope (Tóth, 1963). The presence of highly permeable layers and channels, however, will tend to bypass flow beneath local cells and develop more regional systems. It is known from the preparation of potentiometric surfaces above buried channels in the Beaver River basin (Alberta Research Council, 1985), that groundwater flow does tend to become channeled in the bedrock valleys where the hydraulic gradient is along the length of the valley.

Along each side of the Beaver River valley the land rises to elevations above the remaining Beaver Lowland. There is evidence that this provides enough potential gradient away from the river to bring deeper groundwater to the surface. Alkaline soils and artesian flows in water wells occur in a strip of land north and south of the river. To the north the areas north of La Corey and around Harold Lake contain these features and are likely discharge areas (Fig 2.1). To the south manifestations of discharging groundwater can be found at Sinking Lake and in the area around Charlotte Lake. Alkaline soils are not observed in areas of shallow, local system discharge in the basin. It appears, therefore, that whereas the Beaver River is a major line sink for the basin, a considerable amount of discharge probably occurs in other parts of the Beaver Lowland as well (Ozoray et al., 1980). Ardmore is a part of the higher ground adjacent to the river and most of the groundwater recharged in the area appears to discharge in the river valley.

In 1974 a 58 m deep test hole was drilled by Alberta Environment in the Beaver River Valley about 4 km upstream from the north west corner of the Ardmore study area (Ozoray et al., 1980). The report states, "drilling encountered Quaternary sand or sand and gravel," from which we infer that recent fluvial deposits are thin or absent and that they are directly underlain by older Quaternary channel fill and that bedrock was not reached in the hole. Water was reported to have flowed from the test hole under artesian conditions, but there is no record of a measured hydraulic head. Nevertheless, the hole provides two important pieces of information concerning the regional flow in the basin; first that the Beaver River is in direct communication with bedrock channel aquifers in at least some of its reaches and second, confirmation that some groundwater discharges from these aquifers to the Beaver River.

Below the Quaternary sequence, there is potential for groundwater to leak through the poorly permeable Lea Park and Colorado marine mudstones, towards the Manville group aquifers. Potentiometric surfaces show hydraulic heads in the Quaternary aquifers (this study) are higher than in the underlying Manville (Hitchon et al., 1989 and Wickert, 1992). The hydraulic head difference is around 50 m, which results in an average gradient of about 0.25 m/m. Wickert (1992) shows a loose correlation between the potentiometric surface in the Upper Manville and surface topography. This suggests that the Manville aquifers are only partially confined below the Colorado group and indicates some direct communication with the surface. She also shows that the pore pressures in the Upper Manville are less than hydrostatic, which may indicate downward flow (Tóth, 1980).

The hydrochemistry of the area has been studied in detail by Wallick, (1984). He found the groundwater in the Quaternary aquifers to be characterized by low TDS values, usually less than 1000 mg/l; the dominant cations in wells less than 30 m deep being calcium and magnesium and the dominant anion, bicarbonate. He noted a trend towards sodium-sulphate type water with increasing well depth. He also observed a shift in composition in samples from the same wells in which sodium, magnesium and chloride were skewed in the direction of higher summer than winter concentrations. This was attributed to flushing of these mobile cations from the soil and unsaturated zone during spring and summer infiltration events.

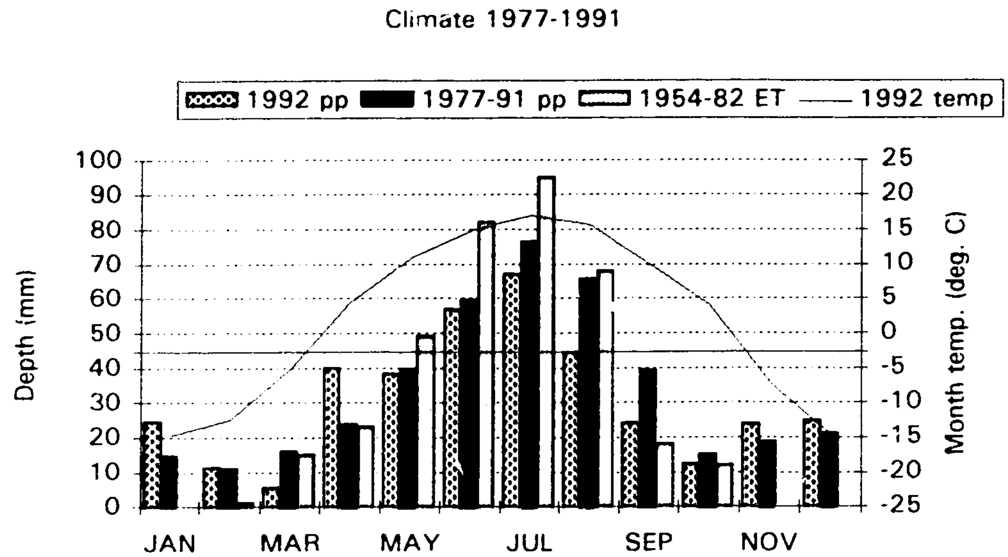


Figure 2a. Precipitation, ET and Air Temperature.

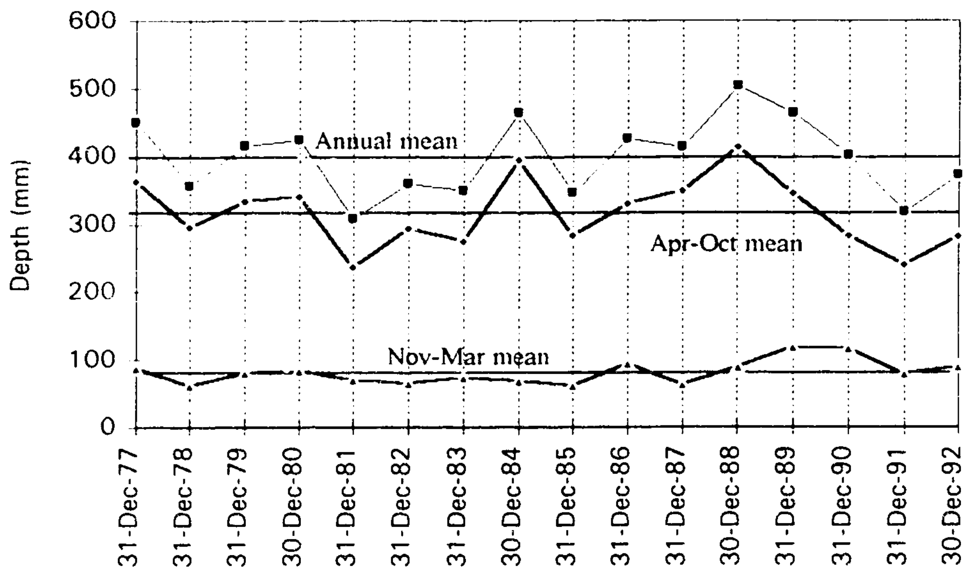
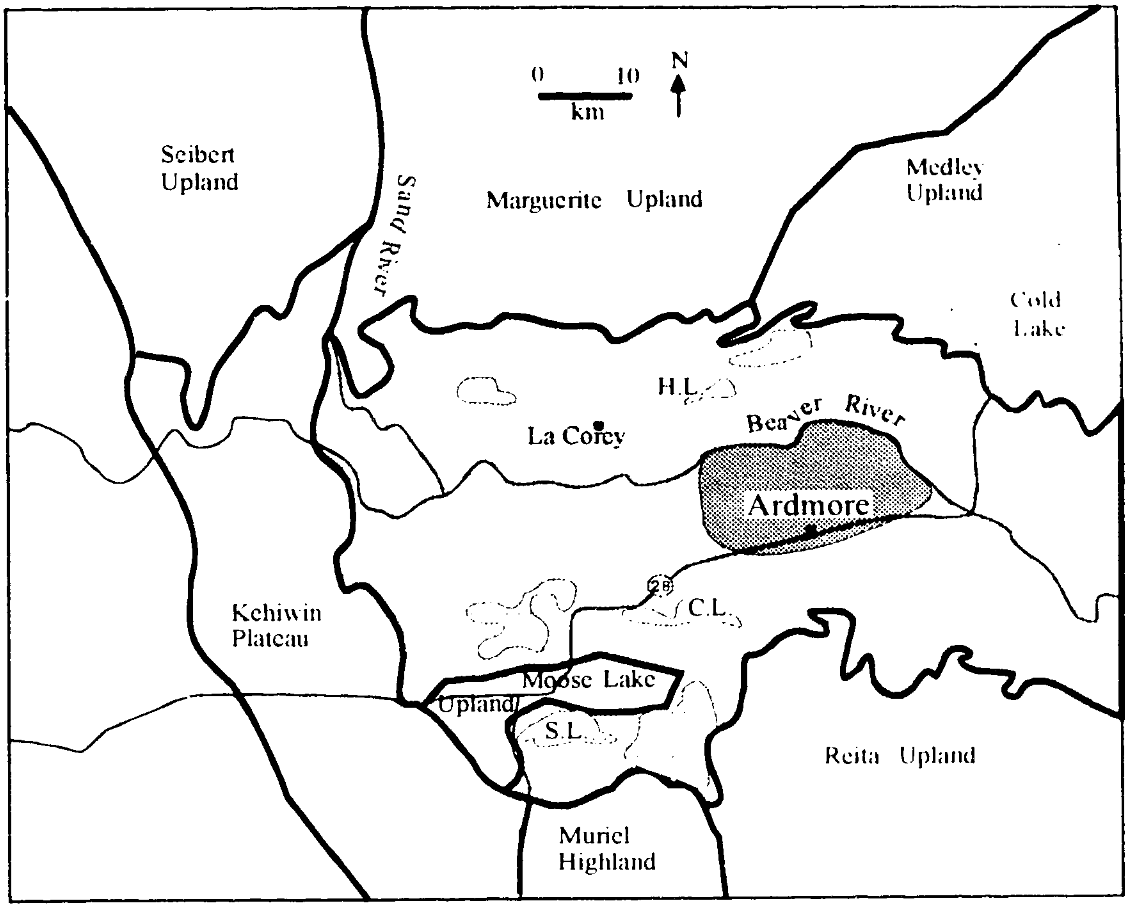


Figure 2b. Precipitation at Cold Lake Weather Station for the period 1977-1992 inclusive.
(based on data provided by Environment Canada).



Beaver Lowland | Ardmore study area | Lakes

Rivers | Roads | Physiographic Boundaries

H.L. Harold Lake | C.L. Charlotte Lake | S.L. Sinking Lake

Figure 2.1. Regional setting of the study area. (Modified from Andriashek and Fenton, 1989).

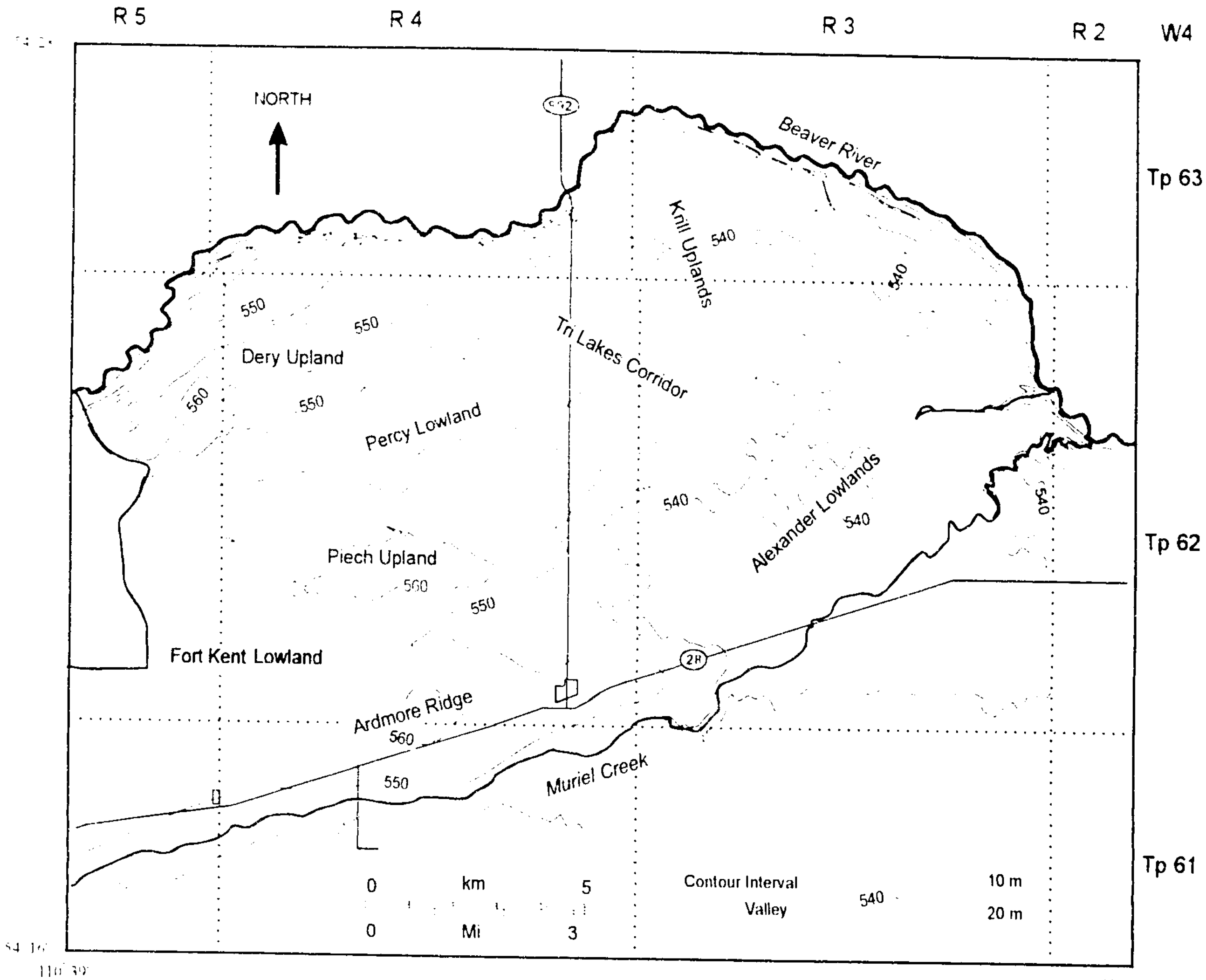


Figure 2.2. Physiographic units within the study area.

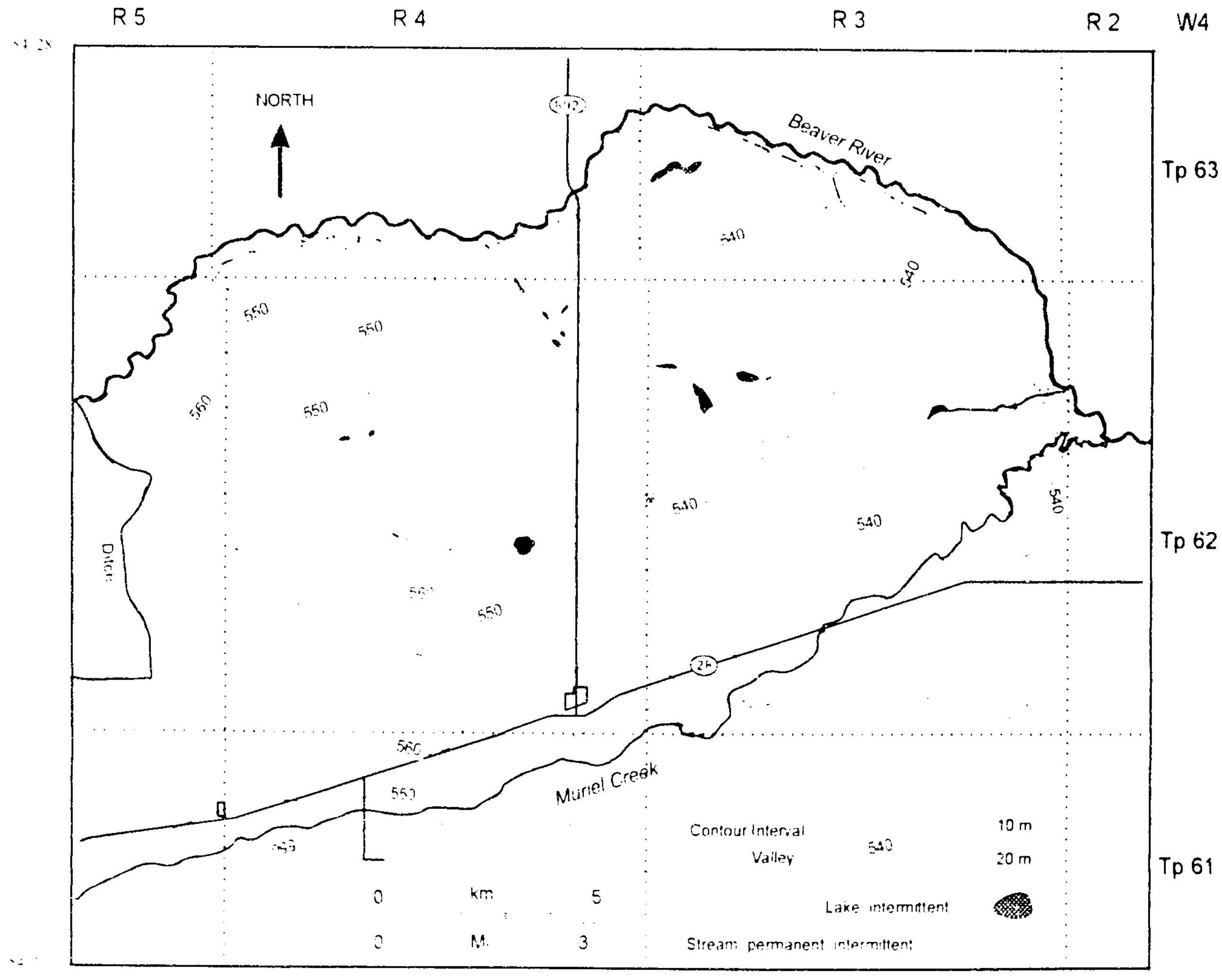


Figure 2.3. Topography and drainage.

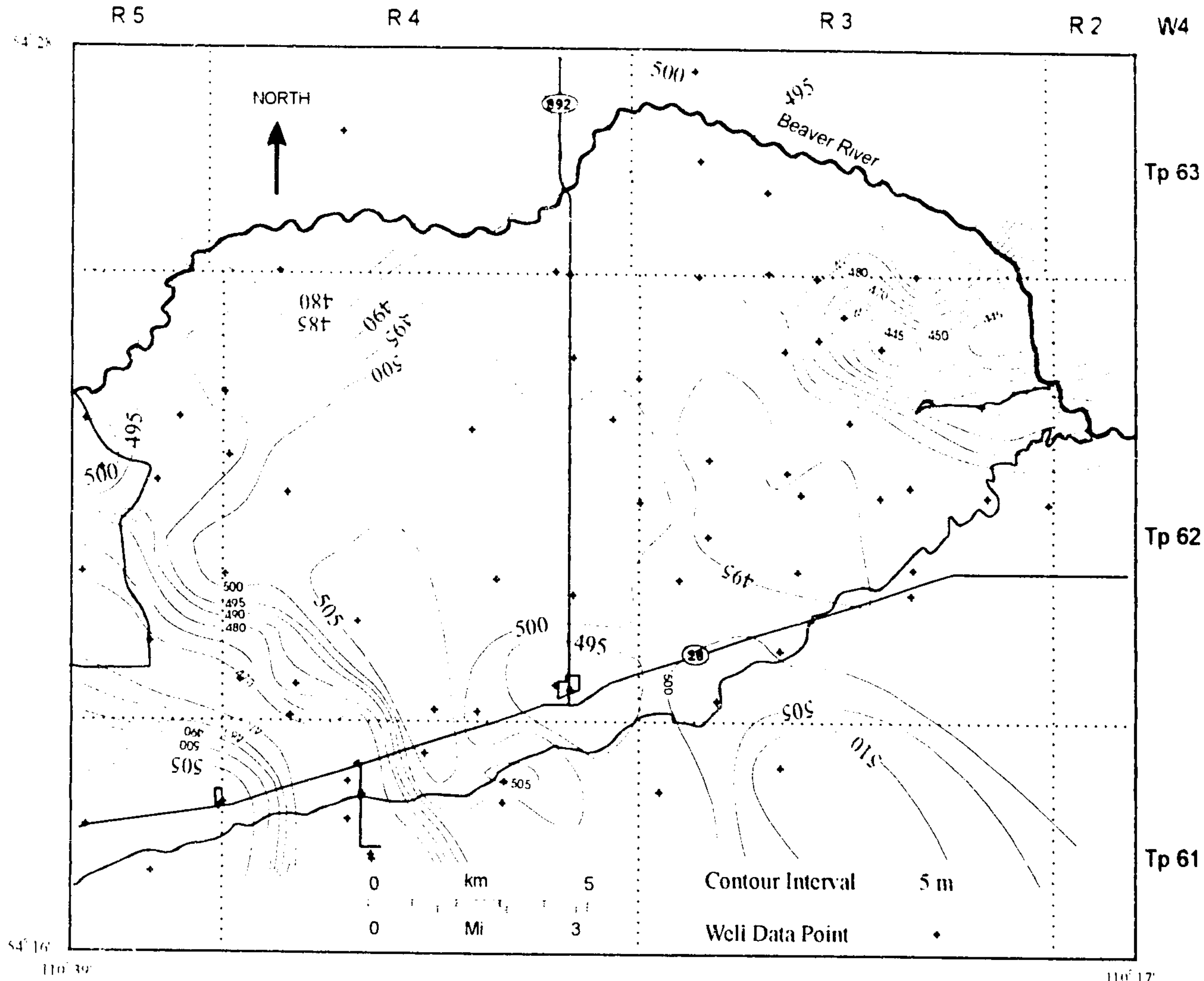


Figure 2.4. Bedrock topography.

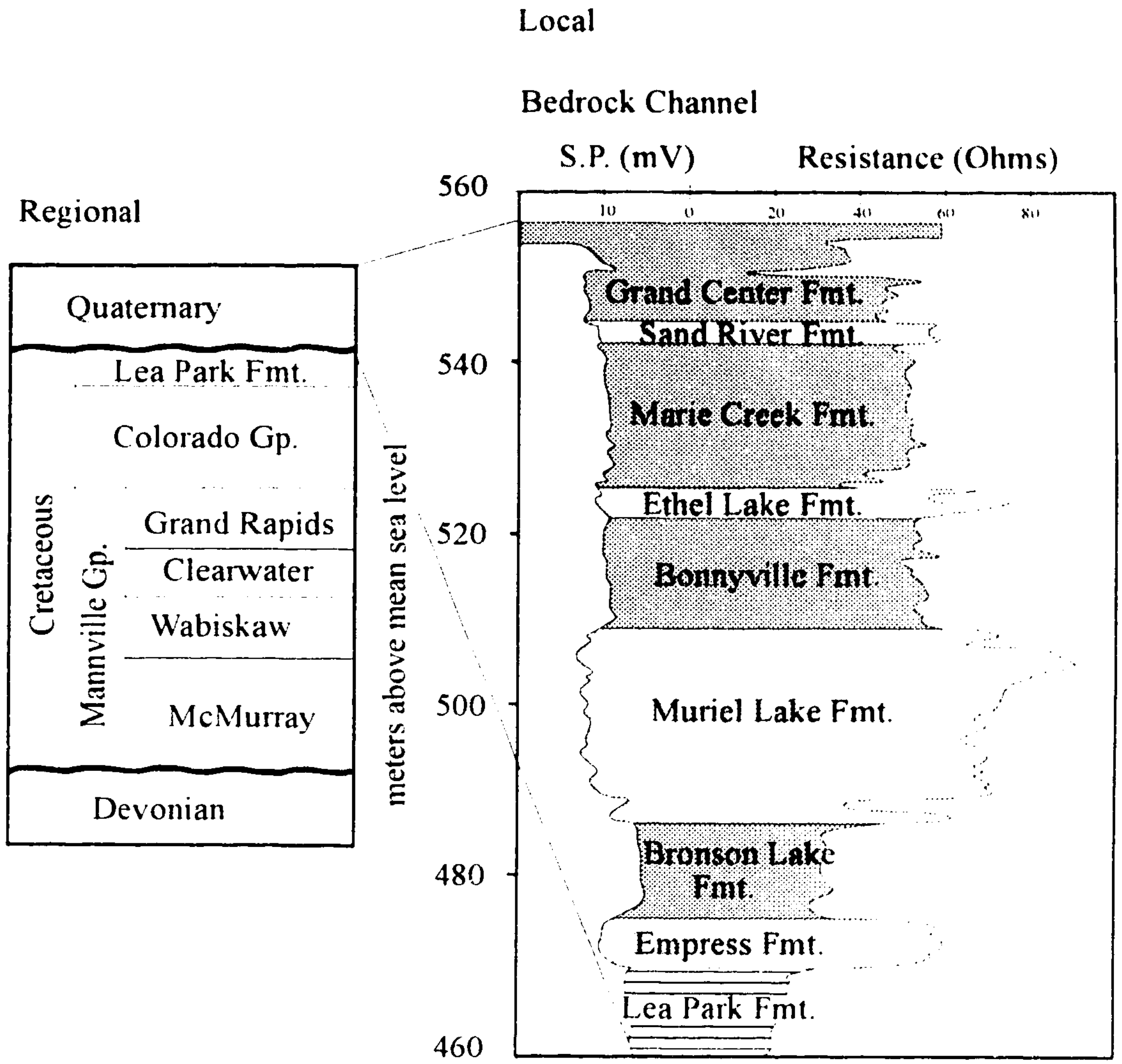


Figure 2.5. Regional and local stratigraphy. Local sequence demonstrated with geophysical logs from Weco well, LSD 4, Sec. 28, Tp. 61, R. 4, W4M, elv. 555 m.a.s.l., depth 116 m, drilled in the Bronson Lake Channel in the S. W. corner of the study area. Sandy outwash deposits, pale grey, till medium grey and marine mudstone shaded.

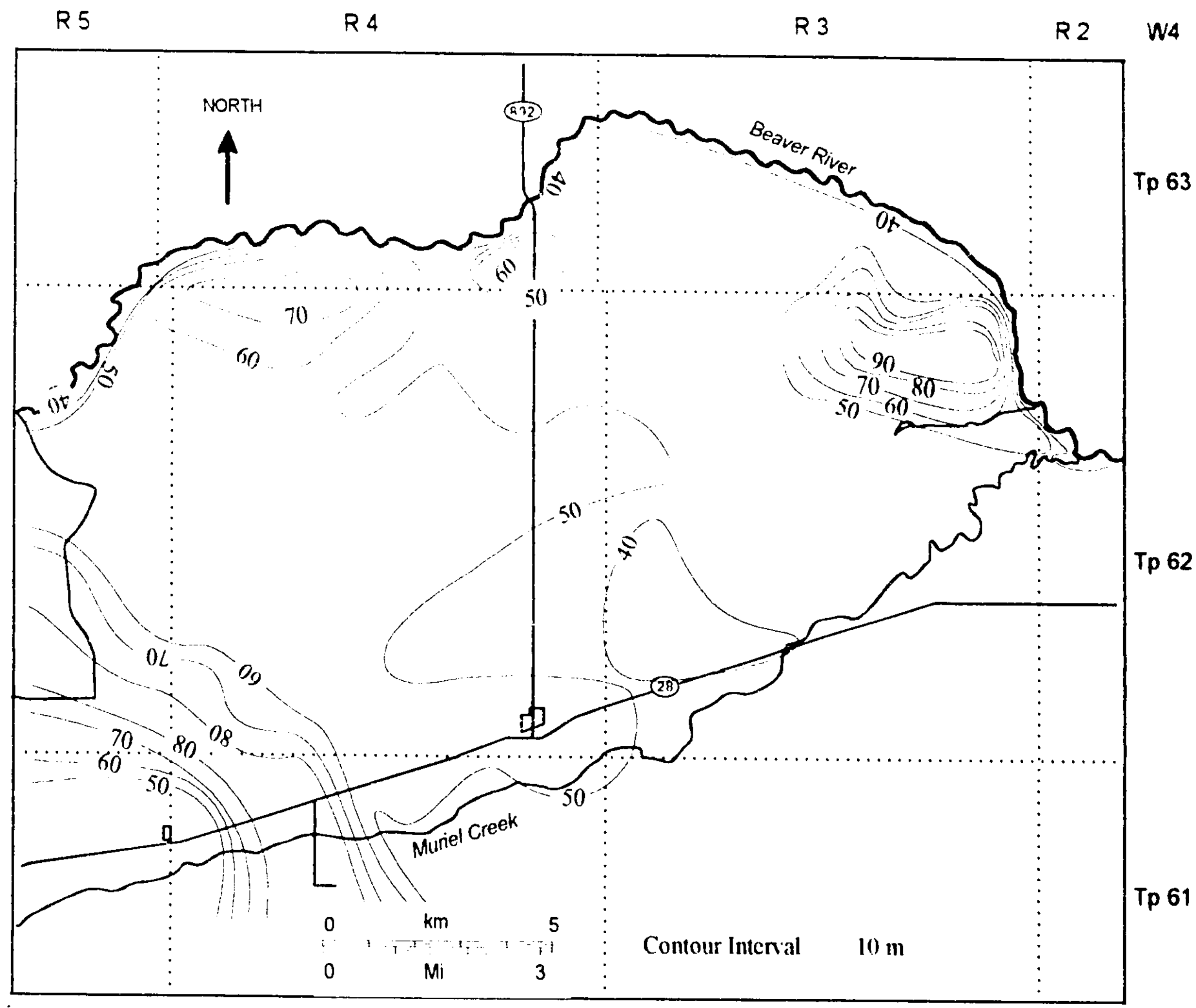


Figure 2.6. Quaternary sequence isopach.

3. Methods and Materials

The methods and materials, which provide the basic information for the recharge estimates, are discussed separately for each of the three approaches. Section 3.1 covers hydrograph and meteorological data, Section 3.2, data collection for flow pattern analysis and Section 3.3, pump testing.

3.1. Hydrographs

Data from more or less continuous monitoring of local water levels came from our own network of observation holes and from the Ardmore Thermal Project.

3.1.1. University observation holes

A total of 26, 15.24 cm (6") diameter holes were drilled with a truck mounted solid stem auger in early October, 1991. Hole number 23a could not be completed due to difficult drilling in gravels and hole number 21, located in the Beaver River Valley, could not be completed deep enough for useful water level readings. Permission to drill at site number 4 was not granted by the land owner. This left a total of 24 holes available for monitoring (Fig 3.1).

3.1.1.1. Observation hole design

The observation holes were designed to capture the full range of water table fluctuations throughout the year (Fig 3.2). The members of the drilling crew were given an estimated depth to water table, but instructed to drill until they were sure they had drilled into saturated conditions. Whereas we were anxious not to drill too far below the water table, we also wished to be sure that the holes did not go dry. A preferable method would have been to drill a number of open holes to different depths within the range of water table fluctuations at each site. This is because in regions of a flow system where the flow is anything other than static or horizontal, the true water table depth can only be measured from the level at which water stands in a shallow hole, open along its length, and penetrating the surficial deposits just deep enough to encounter standing water at the bottom (Freeze and Cherry, 1979). Since steep hydraulic gradients were not anticipated over the zone of water table fluctuation in this area, and since the depth to water table was not well known, the single well approach was considered preferable from both practical as well as economic points of view.

The observation holes were completed in unconsolidated deposits and so it was considered likely that some caving or washing out of sandy layers would occur. We therefore installed PVC casing and screens in every hole. The annular space between casing and native material was sealed only over the top 1 m or so below ground surface (Fig. 3.2). This was to prevent direct hydraulic communication between the surface and the water table down the outside of the casing. The casing itself was left standing several centimeters above ground surface and capped. The length of screen varied depending on the expected rise and fall in the water table but was usually at least 2.8 m. The screens were manufactured from 10.16 cm (4") internal diameter schedule 40 PVC pipe slotted with three vertical rows of 0.5 mm (0.02") slots. In some holes the water table fluctuated above the screen as shown in Figure 3.2b. Equilibration of pressure head through the screen was assumed to be instantaneous, so this configuration was not considered to have any effect on the water levels. It is interesting to note that most of the installations were recovered by hand (with some leverage added) at the end of the program in Oct. 1993, and that the well bores remained in good condition.

The responses in some hydrographs indicate that we were not always successful in sealing the annular space from the surface. The abrupt water level rises which occurred in March of 1993 in UOH 3, 15, and 23 (Figs 4.3 and 4.6) are thought to have been caused by leakage of meltwater directly to the hole when ponding took place above the cement plug during warm weather.

3.1.1.2 Schedule and method of measurement

The first water level readings were obtained in Dec. 1991 and again in Feb. 1992. Continuous monitoring on a daily basis could not be established until May 1, 1992. The first automatic recorders were installed in early May, 1992 and by early June we had borrowed and installed 11 F type and 7 A type Leupold and Stevens recorders for a total of 18. The remaining holes were monitored by hand, daily between May 1 and July 26, 1992 and between April 4 and June 30, 1993. They were monitored monthly at all other times until the end of the program Sept. 27, 1993 Two more Stevens F type recorders were loaned to the project in April 1993 bringing to 20 the total number of recorders.

3.1.1.3 Location and rationale

The university observation holes were drilled to sample as wide a range of landscape and substrate conditions as possible. The objective in this, together with our mapping approach, was to characterize sub-units within the area in terms of recharge rates, in order to facilitate the regionalization of our point measurements. We, therefore, located the observation holes in uplands, lowlands and in midslope positions. We also located them either close to, or away from, smaller scale depressions in all these physiographic units. The holes were also located in either sandy or in dominantly fine grained surficial deposits. Our site selection had to be approved by the individual landowners, and some compromises were reached.

3.1.2. Ardmore Thermal Project

The second largest source of water level data came from consultant reports of groundwater supply for the Ardmore Thermal Project (Fig. 3.3). These included monthly, weekly and in some cases more frequent water level records. The water was used in steam generation for enhanced oil sand recovery between the years 1975 and 1988.

The project began with two water supply wells (CWW1 and 2) completed in the Ethel Lake aquifer at 30 m below ground, and two observation holes (COH1 and 2) completed in the Sand River aquifer at about 9 m depth. These four wells were also monitored by hand during the present study between June 1992 and Sept. 1993. Continuous recorders were installed on the two supply wells in April 1993. The hydrographs from the wells matched exactly when compared on a light table, so the recorder from CWW2 was removed at the end of June and installed at UOH 9.

Three more observation holes were drilled for the project in 1978, two in the deep aquifer (COH2-25 and 5-21) and one in the shallow aquifer (COH2-33), at distances of between 1.5 and 3.5 km from the supply wells. Up to 7 private wells were monitored in the surrounding area during most of the project life. Only one of these, at some 6 km distance (not included in Fig. 3.3), was thought to be completed in the deeper aquifer.

3.1.3. Meteorology

Our main source of meteorological data was the Cold Lake weather station, located approximately 3 km from the western boundary of the study area and about 12 km from the center. Precipitation data were available throughout the year and reported as equivalent depths of water, with readings taken daily at 11:00 am. Evapotranspiration calculations from the meteorological data are published periodically by both Environment Canada and Alberta Environment. Detailed recordings of atmospheric pressures and temperatures are also available.

We recorded rainfall during the field work with three rain gauges (4 in 1993), on an east to west line across the area (Fig 3.1). We tried to maintain a rainfall reading schedule similar to the Cold Lake station. The rain gauges had apertures, (127 mm in diameter), that were funneled into glass receiving jars, contained inside overflow cans. The cans were set vertically so that the apertures were about 30 cm above ground. The depth of rain was measured by decanting the glass jar into a graduated measuring beaker which could measure accurately to 0.1 mm. The gauges were located in relatively sheltered locations and the vegetation surrounding the gauge was trimmed to about 5 cm above ground. The central location at UOH 6 had also a tipping rain gauge with a 30 cm diameter aperture and a bucket that tipped with every 0.2 mm of rain. We were therefore able to measure both the amount and the intensity of rain quite accurately.

3.2. Flow pattern data

3.2.1. Geology

Detailed stratigraphy of the Quaternary sequence underlying the Ardmore area has been completed by Andriashek and Fenton (1989). The objective of this study was to confirm their mapping and to improve, if possible, the definition of those formations which were already exploited as aquifers and those which were clearly aquitards.

Information from over 300 wells and test holes was accumulated, much of it from the government archives. Table 1 in Appendix B is a listing of all the data points by LSD location and shows the number of formation picks that were made. The information is very variable in quality. Very few holes had either detailed lithological descriptions or geophysical logs and even fewer had both together. Due to the density of wells, however, we believe that the unit mapping is sufficient for a preliminary description of the flow regime.

The sand units were traced first by constructing profiles across the area using correlations between geophysical log responses (Fig 3.4 and 3.5). These were based on the work of Gold (1978) and Andriashek and Fenton (1989). Each well with a lithological description was then compared to the general stratigraphic framework established by the profiles. If a unit described as sandy was found at the correct stratigraphic elevation, it was connected with others at the same elevation. Figure 3.6 shows an example of the datum point distribution used to define the uppermost unit (Sand River Formation).

Oil and gas logs are not very useful for shallow stratigraphy because the holes are often cased through the Quaternary deposits. They were used instead to determine the thickness of the Colorado Group and the depth to the underlying Mannville Group.

Surficial deposits were mapped using aerial photographs backed by field validation from roughly 30 soil pits, the UOH network, road cuts and other outcroppings. Lithology logs were prepared for each UOH with samples from the auger flights during drilling. They were compiled together with details of hole completion and water levels to assist in assigning storage coefficients in the analyses of hydrographs (see example Fig. 3.7).

3.2.2. Hydrogeology

3.2.2.1. Field mapping

The purpose of field mapping was to identify, if possible, from surface manifestations of groundwater flow, regions of predominantly recharging and discharging conditions following procedures originally outlined by Tóth (1966). These observations were to assist in the analysis of flow patterns and to assist in the characterization of sub-units within the area for regionalization of point estimates of fluxes.

Field mapping for surface manifestations of groundwater was conducted in August 1991, and April-May 1993, and supplemented by observations during the remainder of the program. Aerial photographs were used to plan traverses which included most of the wetland and moist depression sites. Following generally the ideas of Lissey (1968) and Leskiw (1971) the wetlands and depressions were mapped and grouped as having features which were presumed to indicate the dominant direction of groundwater flow beneath them. The area was also examined for other surface manifestations of groundwater, such as springs and seeps, soap holes and artesian conditions, again following the methods of Tóth (1966).

3.2.2.2. Soils

The importance of interactions between water and parent materials in the development of soil profiles is now well established (Eilers, 1987). Soil pits were dug at each UOH site and at intermediate positions in an effort to identify the main soil catenas of the area. Where the parent material was glacial till the pits were dug deep enough to include the carbonate horizon. The horizons were described using the Canadian System of Soil

Classification guidelines. The presence of carbonates was tested with a 10 % solution of hydrochloric acid.

Particular attention was paid to features which might indicate the dominant direction of water movement within the soil profile. These included: i) the presence or absence of gleyed horizons, and ii) eluviated and illuviated clay horizons, iii) the depth of the solum, iv) the thickness of the A and B horizons and v) the depth to calcium carbonate horizons. In the choice of these features we followed Zebarth and de Jong (1989).

3.2.2.3 Water level maps and profiles

The most widespread water level information is from private, usually domestic, water wells. Records from these wells are filed with Alberta Environment. The records normally include a non-pumping water level obtained by the driller prior to well development or testing. The reported levels in several domestic wells were checked during the study and found to be within the range of natural seasonal variation. There are approximately 260 of these data points in the study area.

Analysis of the deeper parts of the flow regime below the study area had to be based on information from the existing wells. Water level maps (represented in plan view by equipotential contours) were prepared for each of the aquifers in the Quaternary sequence. The non-pumping water levels recorded in wells at the time of their installation were used where readings were not measured during the program.

The water level in a well completed with a screen sealed to the aquifer, may be treated as a reading of the average fluid potential within the aquifer at that point. For wells completed without sealed screens, or for water levels measured before the installation of sand packs and seals, the level represents some averaged value of potentials between the depth drilled and the water table. Since there is no enforced protocol for the measuring of water levels in the drilling industry, the so called "static water levels" from drilling reports might represent either situation or some unstabilized level unrelated to the formation potentials. We simply mapped the water levels for all wells completed in each aquifer and removed the erroneous looking points (those levels which stood out as being either above or below the neighboring levels). For this reason and because the aquifers are not perfectly horizontal, we have avoided calling the maps potentiometric surfaces even though we did use them to estimate approximate hydraulic gradients in the sub-surface. Since the flow is dominantly downwards beneath the area, the maps may over estimate the fluid potentials in the deeper aquifers and thereby under estimate the vertical hydraulic gradients; they should be more accurate at shallower depths.

In cases where the holes were drilled significantly deeper than the location of well completions, more reasonable results were obtained by using the latter depth (i.e., we had fewer anomalies in the contours). From this we surmised that in most cases the hole plugs, set below the screens, were effective seals.

The first step in preparing the data for map generation was to sort the holes by completion elevation (the elevation above mean sea level at which the screens had been set) (Table 2, Appendix B). The sorting was then compared with the geology to decide within which aquifer the wells had been completed. This procedure was particularly useful for wells which did not have any lithological records. The sorted file was then sub-divided and contour maps of water level elevations were prepared for each aquifer. The software programs Surface II and Surfer were used to assist in the generation of contours by creating grids from the randomly spaced data. Cross-sections were then constructed through the contour maps, approximately parallel to the flow direction, and intersecting the groundwater divides.

The uppermost water level map prepared was an approximation of the water table; it was constructed by using water levels in the shallowest wells and observation holes. Since the number of these holes was not large, areas in between were mapped by developing a simple relation between surface elevation and depth to water in the holes. A least squares regression line was fitted through a plot of surface elevation with water level elevation (Fig 3.8). The coefficients in the equation of this line were then used to estimate the water levels from the elevation of all the other wells in the area. Since the topography is fairly smooth without rapid changes in slope this was thought to be a reasonable approach.

The outcrop elevations were used as minimums for hydraulic heads in aquifers intersected by the Beaver River valley. For Muriel Lake aquifer, which runs below the river, an arbitrary minimum value for hydraulic head was chosen at approximately 5 m above river level. The head could be much higher than this, especially if an aquitard is present below the river. The only report of a hole drilled below the valley (Sect 2.5), however, does not mention strong artesian conditions. The completion depth of the only domestic well located in the valley from our area appeared to be at about river level and was therefore presumed to be in direct communication with it.

3.3. Pumping tests

The purposes of the pumping test analyses were to estimate hydraulic parameters, and to confirm the conceptual model of the flow regime. As mentioned in Section 3.1 the Ethel Lake aquifer was exploited for water supply by the Ardmore Thermal Project.

3.3.1. Ardmore Thermal Project

Two pumping tests had been conducted for this project using the supply wells and observation holes shown in Figure 3.3. The first test, comprising a test of each well, was carried out in 1974 at the time the wells were installed. CWW1 was pumped for 14 hours and CWW2 for 24 hours. CWW1 was used as an observation hole for CWW2 during the latter test.

Transmissivity values were estimated from the pumped wells using Jacob's straight line method of solution of the Theis equation (Cooper and Jacob, 1946). The method can be used for single well tests with constant discharge in both confined and leaky aquifers without corrections for non-linear well losses, provided the conditions below are met (Kruseman and de Ridder, 1990). The early drawdown behavior in a pumped well may be influenced by well storage. These effects will show in the early portion of a drawdown curve as a steeper slope. Large storage effects will maintain the steep portion of the curve for longer time. The effects of well bore storage can be neglected, however, if the condition, (Papadopolous and Cooper, 1967)

$t > 25r_c^2/T$

is met, where t is time, r_c is the radius of the unscreened portion of the well in which the water level is changing, and T is transmissivity. In addition the influence of leakage is negligible when,

$t < cS/20$

where c is the hydraulic resistance of the aquitard as expressed by its thickness, D', divided by the vertical hydraulic conductivity, K', and S is the storativity in the aquifer.

Drawdowns in OWW1, the supply well used as an observation hole during the test of CWW2, could be analyzed with the Theis curve matching method since enough early time data were available for a unique match.

A second test had been conducted on the same wells in 1978. For this test, both wells were pumped simultaneously at approximately the same rate for just over 9 weeks, and water levels were monitored in the surrounding observation holes (Sec. 3.1 and Fig. 3.3). The most accurate information for this test came from CWW2, whereas records from CWW1 were less detailed.

The 1978 test proved difficult to analyse and in the end we were not confident enough of the results to use it quantitatively. The difficulties resulted from the following facts: i) both wells were pumped simultaneously, ii) the observation wells in the pumped aquifer were far away, iii) the records for the early time drawdowns were not available and iv) the pumping rate, including some intermittent stoppages, varied. The water level recovery data also lacked sufficiently frequent early time measurements and the pumping rate at the end of the test was uncertain. In addition to these problems, the aquifer composition and geometry appeared to be far from simple.

It was possible, from the 1978 test, to draw some qualitative conclusions about hydraulic continuity in the subsurface, however, which gave important supporting information for the conceptual model.

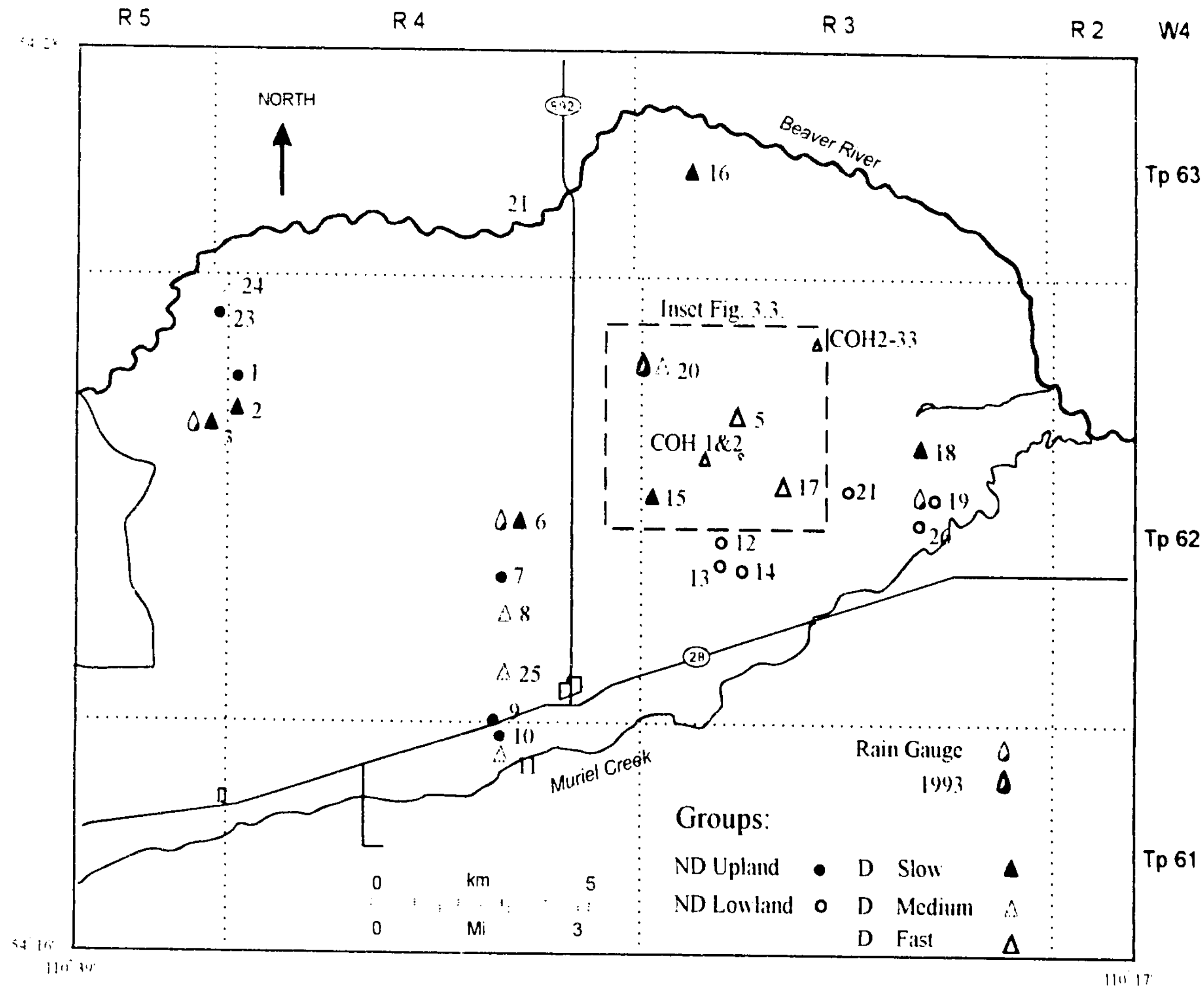


Figure 3.1. University observation hole locations. Small symbols are existing holes.

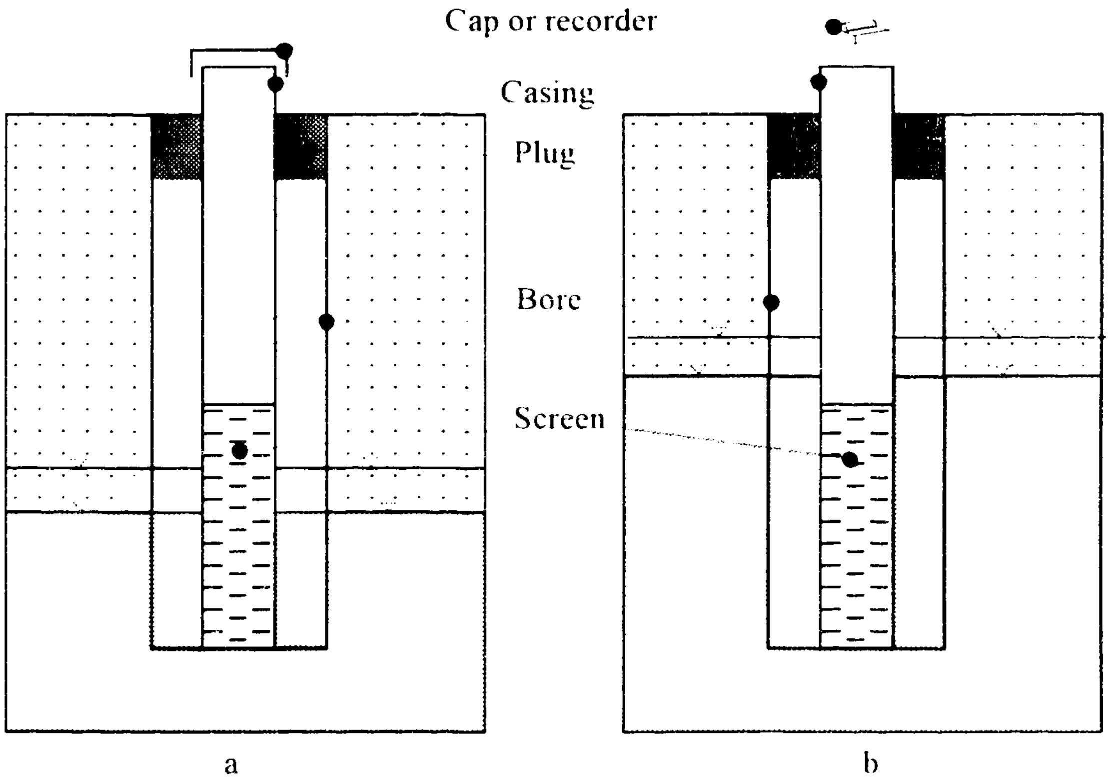


Figure 3.2. University observation hole design. Type a, has water table fluctuation entirely within screen. In type b the water table fluctuates above the screen.

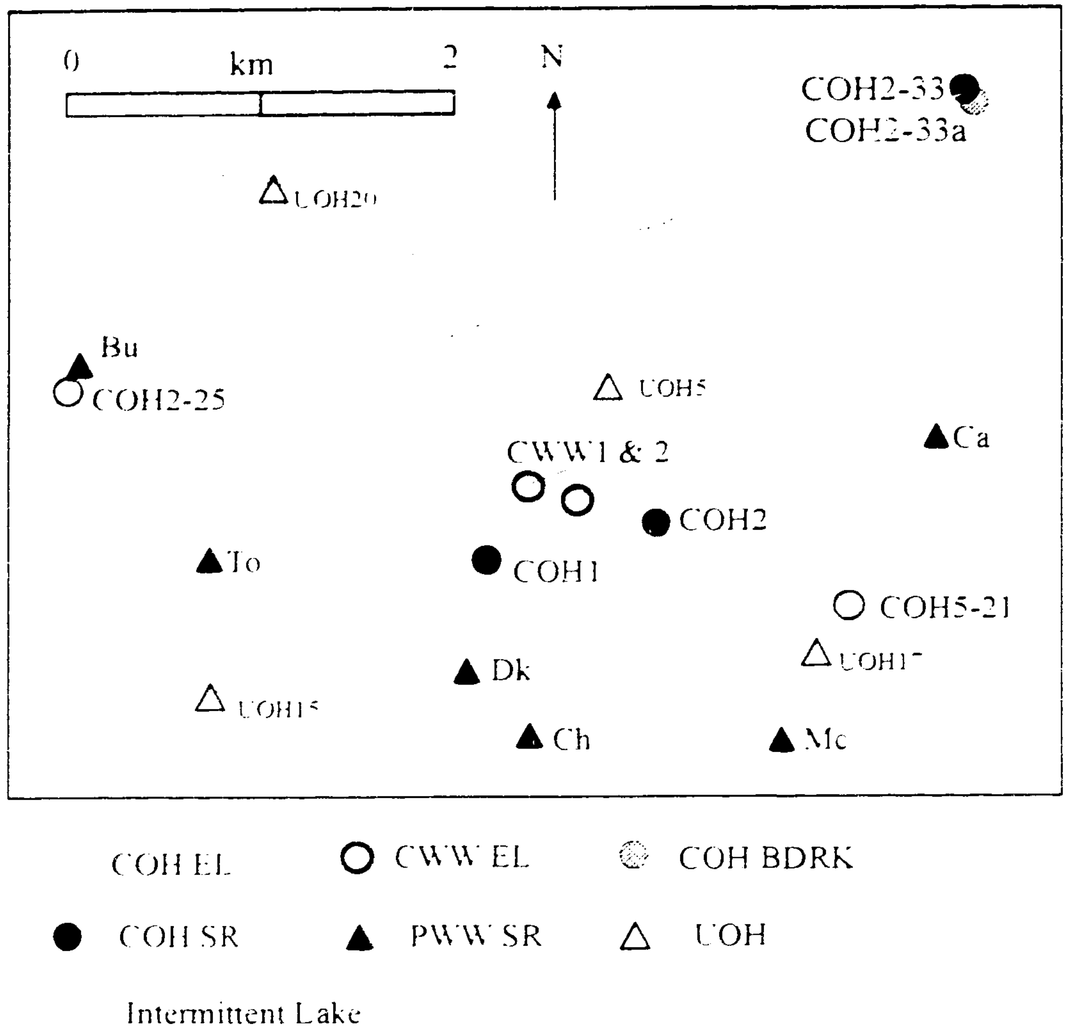


Figure 3.3. Ardmore Thermal Project: hole locations. COH commercial obervation hole. CWW commercial water well. PWW private water well. EL Ethel Lake Aquifer. SR Sand River Aquifer. BDRK bedrock and UOH University observation hole.

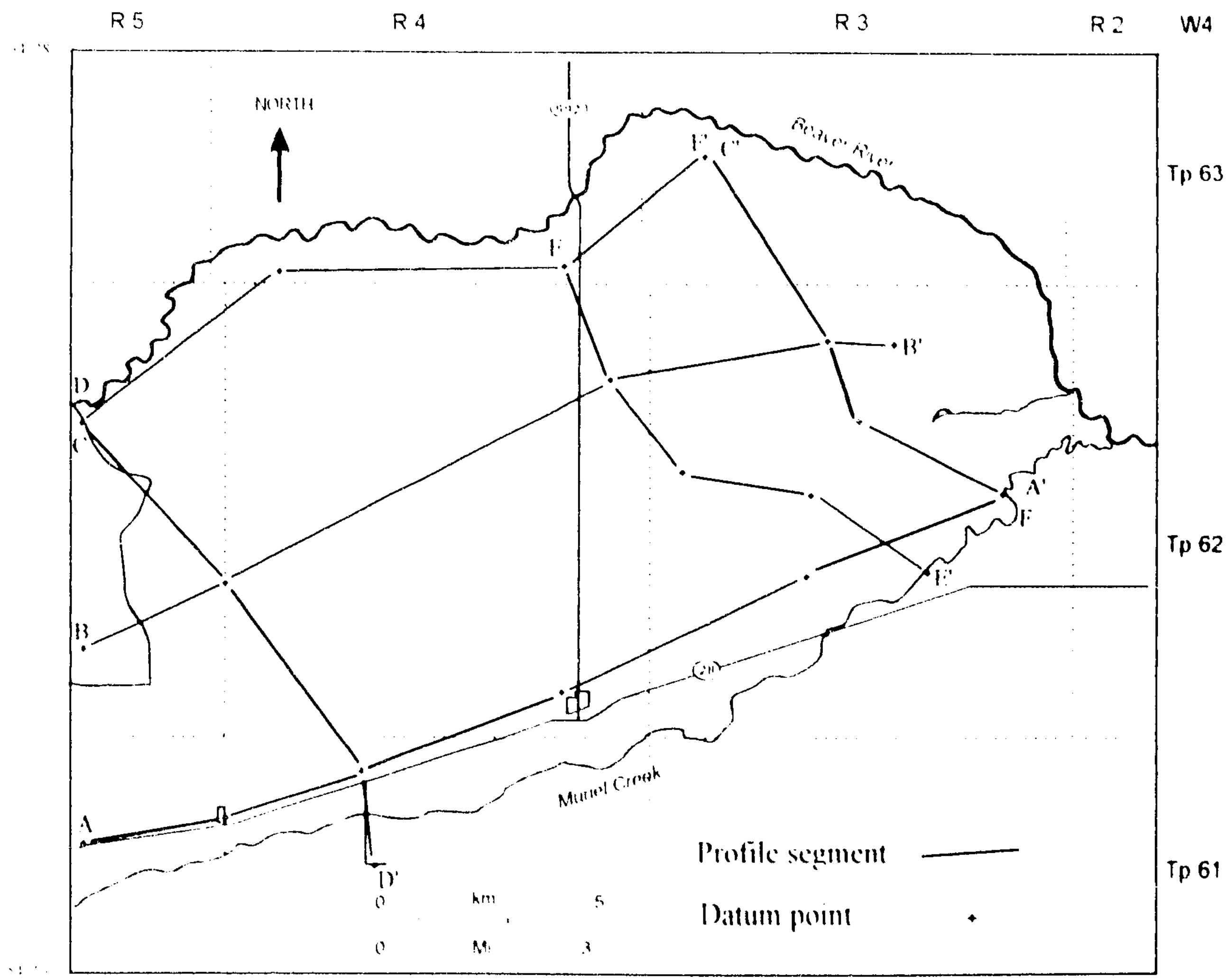


Figure 3.4 Location of main geophysical profiles.

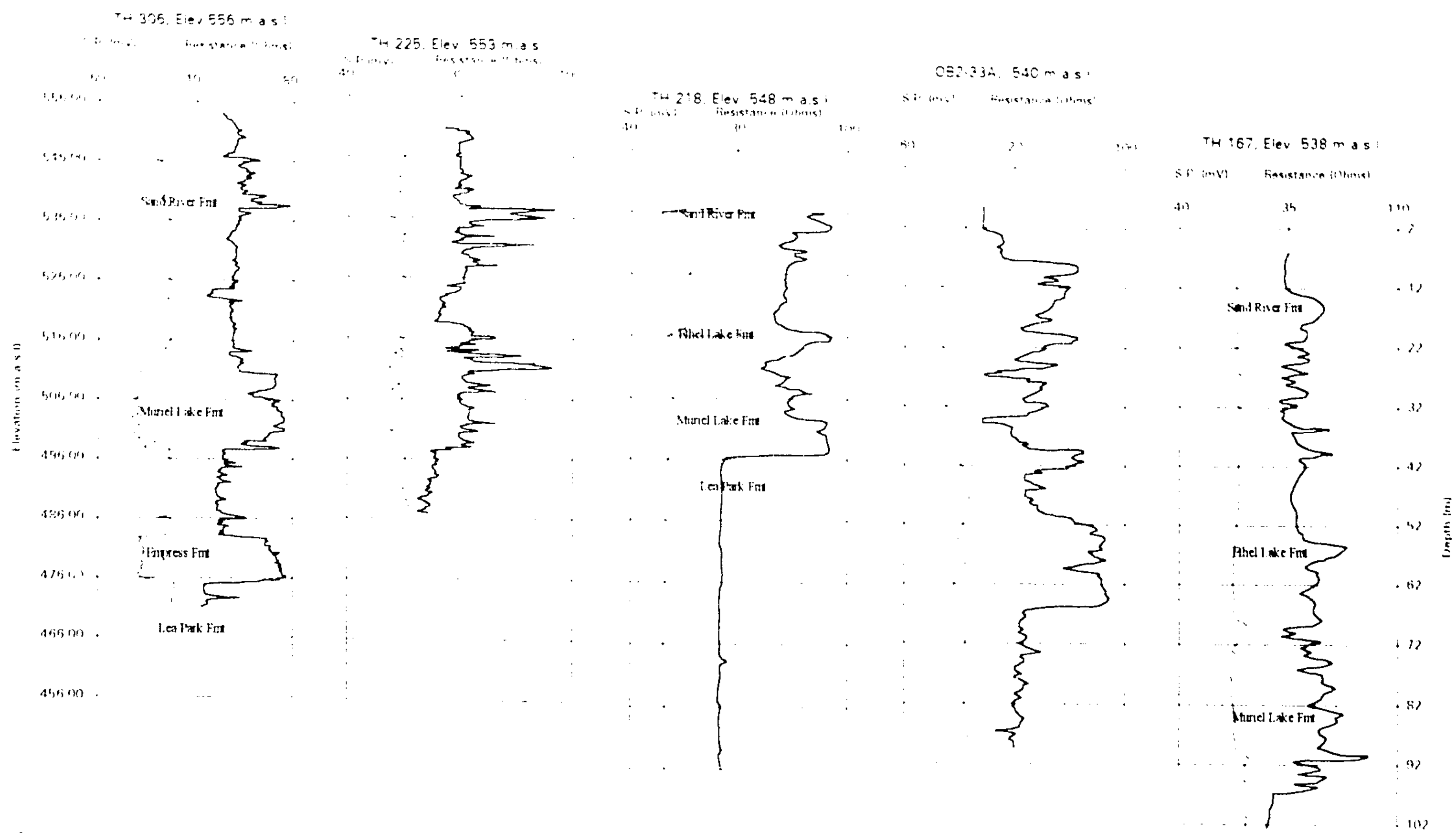


Figure 3.5. SP and Resistance profile B-B'.

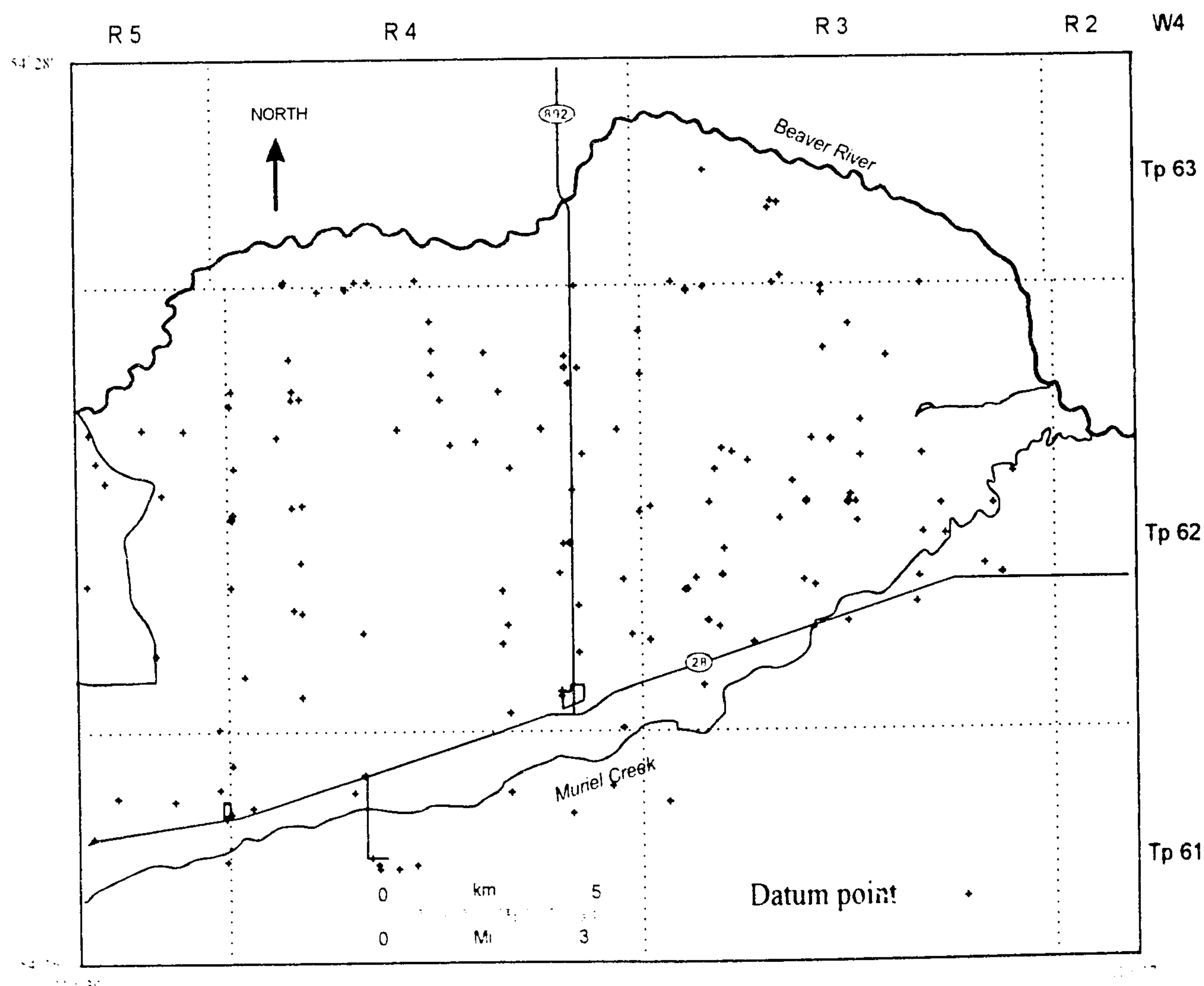


Figure 3.6. Data points in the Sand River Formation.

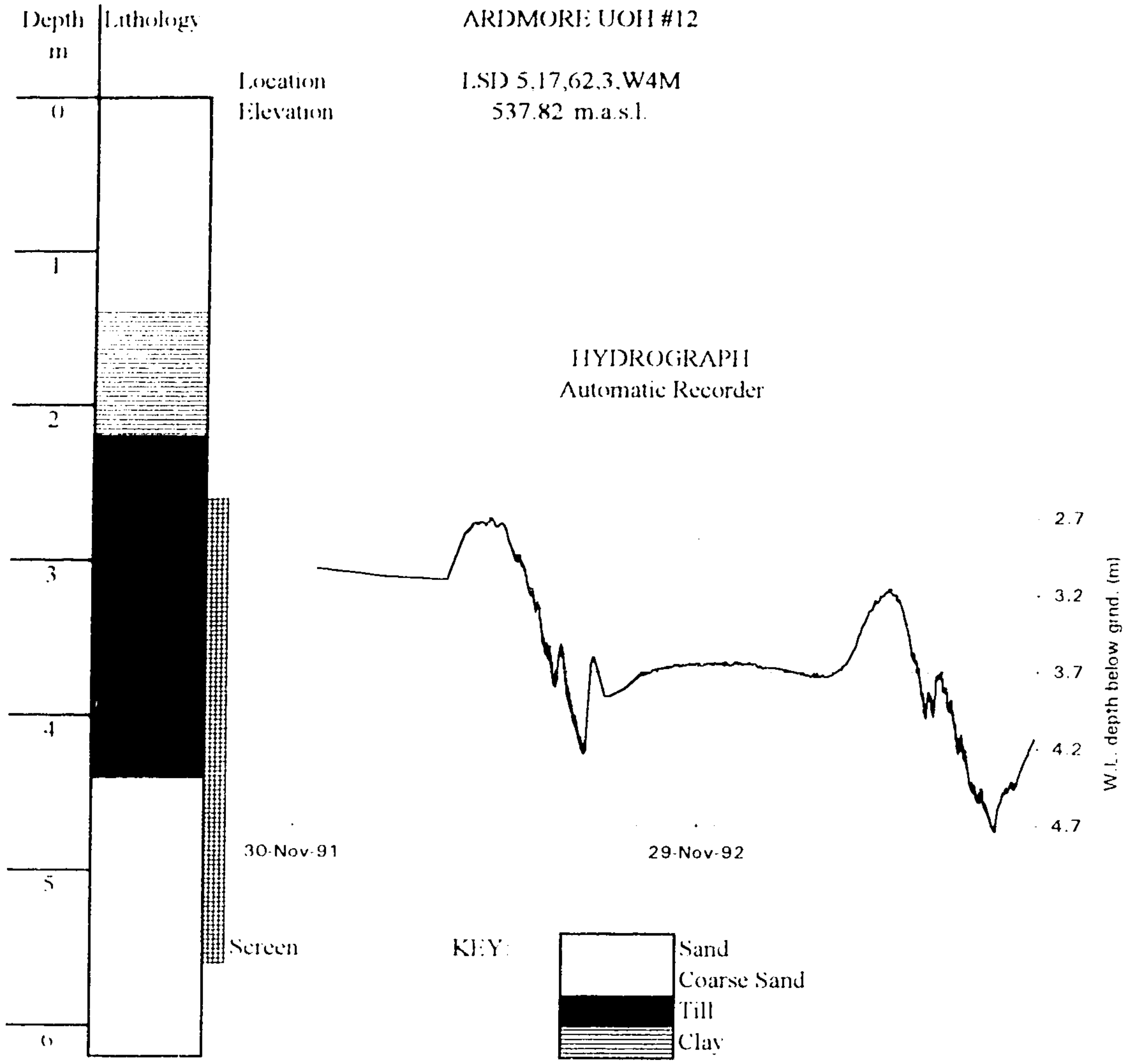


Figure 3.7. UOH 12 completion and lithology log showing zone of water level fluctuation.

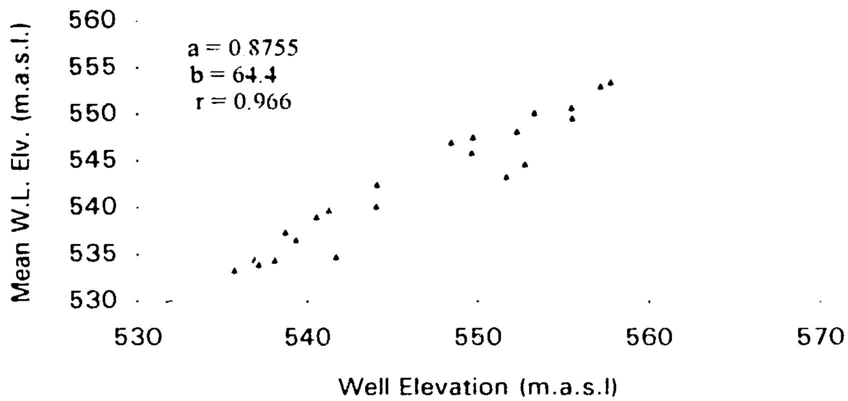


Figure 3.8. Correlation between land surface and water table elevation where mean W.L. Elv. = 0.8755(well elevation) + 64.4 and r is the correlation coefficient.

4. Results and interpretation

4.1. Hydrographs

The results of the hydrograph analyses are presented moving from the general to the particular. The first section gives an overview of the features common to all the observation holes. The next section discusses the holes by groups with similar characteristics and the last section presents some notable features from individual holes.

4.1.1. Overview

The hydrograph terms used in this section and illustrated in Figure 4.1 are in common use among hydrogeologists although formal definitions for them are not found in any textbooks of the subject.

Shallow holes (< 12 m) in the Ardmore area generally show the following pattern of water level fluctuations through the course of a water year, (the period between sequential spring minima) (see UOH9, Fig. 4.1). From a spring minimum in late March or early April, the water level rises rapidly over a period of between 2 to 3 weeks to a higher position. This period is referred to as the spring rise and occurs in response to the arrival of infiltrated water at the water table, originating at the land surface from melt-water and precipitation. Following the spring rise, the maximum water table elevation usually occurs at some time during a period when the water level is fluctuating about a relatively constant mean position, which may last for a month or so into early June. The summer is marked by the most rapid fall in water level that, in many holes, is continuous in the absence of major rainfall events, (at least 20 mm of precipitation in a 24 h period). This period is known as the summer recession and is attributed to high rates of water consumption by evapotranspiration during the growing season. The rate of fall in the water level usually slows in the late summer and autumn, or it may actually rise at this time. This event, known as the fall step, can be directly correlated with the end of the growing season and the onset of frosty conditions. The rate of water level fall usually increases again in late December or early January, which is referred to as the winter recession. We have interpreted the predominant cause of this winter recession to be saturated flow below and away from the water table in a recharging flow pattern. The rate of water level fall during the winter recession is usually much less than the rate for the equivalent summer period.

There may be intermittent and even sustained water level rises at any time during months when the ground temperatures are above freezing; under frozen conditions, however, rises can only occur where the saturated flow is upwards, towards the water table. With two notable exceptions, UOH 12 and COH 2 discussed below, we did not observe water level rises during the winter at Ardmore.

The general pattern outlined above is reported from our own data collected over a fairly short time period, but data from holes which pre-date our program show that it represents a general trend for the area over a period of several years. For example, the water level data from COH2-33 (Fig. 4.2) spans a period of roughly eleven years and the pattern is repeated in almost all of those years.

In each of our holes there was a decline in water level from the 1992 to the 1993 spring minima. The average decline for all the observation holes was 0.3 m. This indicates a general loss of groundwater storage over the period of study. The long term records from observation wells for the Ardmore Thermal Project indicate a more prolonged period of decline. Water levels were lower during our study period than at any time in the previous ten years, which includes the period of groundwater pumping for "the project". There was a marked drop in levels in 1981 (Fig. 4.2 and 4.11), a year with below average precipitation (Fig. 2.0b), from which position there was little recovery, even in the period 1986 to 1989 when precipitation was above the 16 year average.

4.1.2. Groups

The observation holes were grouped initially on the pattern of their hydrographs. These patterns were then analyzed to produce a number of quantifiable parameters (Table 4.1 and Fig. 3.1) which lead to clearer definition of the groups. Two main groups and five sub-groups resulted from the analyses (Table 4.2). Group A, were "non-depression" observation holes which were located away from any topographic depressions capable of collecting surface runoff. They included two sub-groups: those which were located towards ridge crests, upland, and those which were located in the lower slope positions, lowland.

Group B included observation holes that were deliberately located very close to a topographic depression, usually just outside the willow ring of a slough or wetland, or in the equivalent position in cleared areas. These were subdivided on the basis of their estimated recharge rates and on the general patterns of their hydrographs.

Table 4.2. Observation hole groups.

Group A. Non-Depression	Group B. Depression
1 - Upland	1 - Slow
2 - Lowland	2 - Medium
	3 - Fast.

With reference to Tables 4.1 and 4.2 we give a brief description of some of the features that define each group.

4.1.2.1. Group A, non-depression observation holes

A-1, Upland: The group contained a total of five university observation holes and had the highest mean elevation (555.3 m.a.s.l.; standard deviation (2.2 m)) and the greatest completion depth (9.8 (2.5) m) of all groups. The mean water level depth (5.0 (2.0) m) was also the lowest of all groups. The mean annual amplitude of water levels, (0.8 (0.2) m), was greater in this group than in the lowland group but was less than in many holes located near depressions.

Water levels in these holes did not rise in response to individual precipitation events, with the single exception of UOH 9 which showed uncharacteristically rapid water level rises in response to two major rainfalls in the late summer of 1993 (Fig. 4.3). A notable feature from the same year was the complete absence of a spring rise in UOH 23 and 7 and a very subdued rise in UOH 1. From this it appeared that the spring thaw and precipitation were insufficient to produce measurable accretion to the saturated zone. These same three holes, on the other hand, were all rising at the end of the program, apparently in response to the same rainfall events that caused the rises in UOH 9.

A-2, Lowland: There were six University observation holes in this group which had a mean elevation (537.9 (2.6) m.a.s.l.) approximately 17 m lower than for the upland holes and a shallower mean depth to water at approximately 3.7 (2.0) m. The mean annual amplitude of water level fluctuations (0.5 (0.4) m) and the magnitude of the spring rise (~ 0.2 m) were about half those seen in the upland holes. Before drawing conclusions from the last two statistics, however, we must take a closer look at the data.

Hydrographs from group A-2 showed a clear differentiation between materials with low specific yields on the one hand and those with high specific yields on the other. Three out of the five wells had sandy materials in the zone of water level fluctuations (UOH 19, 22 and 26). The water level fluctuations in these wells were characterized by much lower amplitudes, and the hydrographs were generally flat (Fig. 4.4). Hole UOH 14, on the other hand had a clay rich till in the zone of water level fluctuations and the hydrograph from this hole was similar to UOH 9 and 10 from the upland group.

There was a slightly greater tendency for group A-2 holes, with shallow water levels, to show responses to individual precipitation events.

The vertical component of saturated flux (Sec. 5) was quite similar for both the upland and lowland holes, indicating that, notwithstanding their different topographic settings on the scale of the study area, they were both located in a similar flow regime. Comparison of fluxes estimated for each group is discussed further in Section 5.

4.1.2.2 Group B, depression holes

A number of the university observation holes were located near depressions in both upland and lowland settings. We were interested in topographic depressions in light of the mounting evidence that depression focused recharge is important in glaciated landscapes with poorly integrated drainage (Meyboom, 1962; Lissey, 1968; Freeze and Banner, 1970; Rehm et al., 1982; Mills and Zwarich, 1986, Trudell et al., 1986, Zebarth et al., 1989).

The depression holes were divided into three groups and named in accordance with the relative fluxes of both saturated flow and accretion. Group B-1 and B-3 had the lowest and highest fluxes respectively whereas group B-2 had fluxes similar to the non-depression groups.

<u>B-1 and B-2 Slow and Medium Depression Holes:</u> These groups had similar mean water level depths of 2.7 (1.0) and 2.8 (1.6) m respectively. The depressions were located in both upland and lowland settings as can be seen from the similar mean site elevations of 545.4 (6.3) and 547.3 (6.2) m.a.s.l. respectively.

The mean annual amplitude of water level fluctuation was greatest in group B-2 approaching 2 (0.2) m, which was also the largest range for all groups. Group B-1 had the second largest mean annual amplitude of about 1 (0.6) m.

The spring rise was slightly more in the spring 1992 as compared to the spring 1993 in most of the holes (Fig. 4.5 and 4.6). In UOH 8 and 11 from group B-2, on the other hand, the opposite was true (Fig. 4.5). Both of these holes had meltwater ponding in the spring of 1993 close to the sites, which was not present in 1992. The larger spring rise was therefore attributed to increased infiltration from the meltwater.

Observation holes 15 and 16 in group B-1 showed clear water level responses to individual precipitation events in the spring of 1992, and less reliably thereafter (Fig. 4.6). The other holes in this group were not generally responsive to individual events with the exception of UOH3 in late summer 1993. Two wells in group B-2, with mean annual depths to water of less than 1.5 m, (UOH 11 and 20) responded almost as reliably as the rain gauges themselves to precipitation events, except during periods with melt-water ponding (Fig. 4.5). The remaining two holes in group B-2, with mean annual depths to water greater than 3.5 m, showed decreasing responsiveness with depth.

The factors affecting responsiveness to individual precipitation events appear to be: the duration and intensity of the event itself, the depth to the water table, and the antecedent soil moisture conditions. Moderate intensity rainfall (4-5 mm/hr) for say two to three hours, over a shallow water table (< 1.5 m), with high antecedent soil moisture conditions, will produce a measurable rise in water level in the Ardmore area. Response times in the shallow holes are between approximately 1 and 5 h. Surface ponding close to a hole appears to attenuate water level responses to individual events (recall that all the sloughs in the Ardmore area were dry except in the spring). Ground cover does not seem to be an important factor. For example, observation hole 11 is located between a partially reclaimed slough and a hay field, whereas observation hole 20 is located in a willow ring between a slough and a wood of poplars. Likewise, grain size may not be as important as the antecedent moisture content, because UOH 11 responds well to rainfall events in a sandy zone whereas the water level in UOH 20 fluctuates in a clay rich zone.

<u>B-3, fast depression observation holes:</u> The last sub-group contained two University observation holes and two commercial observation holes. They were all located in areas of sandy surficial deposits without organic soils.

The sandy deposits were 2 m or more in thickness, but were underlain by less permeable till or lacustrine clays.

Group B-3 had the shallowest mean completion depth (3.2 (0.1) m) and mean depth to water (1.3 (0.1) m) of all groups, and the lowest mean elevation (539.6 (1.3) m.a.s.l.) for depression holes. It also shared the smallest mean annual amplitude of water level fluctuations (0.46 (0.0) m) with the lowland group (Fig. 4.7).

The amplitude of the spring rise was consistent and relatively small at a mean of approximately 0.3 (0.0) m. This was about 10 cm greater than the same amplitude in the lowland group.

There were responses to individual precipitation events in the spring of 1992, and then more consistently through 1993, especially to the heavy late summer rainfalls towards the end of the monitoring period.

4.1.2.3. Summary of water level observations:

The water table in upland holes is deeper than at other locations. Individual rainfall events are less likely to produce a water level response, and most holes show little evidence of evapotranspiration from the saturated zone. Lowland holes have shallower water levels which are more likely to respond to individual precipitation events. In general lowland and depression holes have steeper summer recessions indicating that evapotranspiration rates from the saturated zone are relatively high. The less responsive hydrographs are from highly permeable sandy sites. Depression sites show the most variation and it is hard to determine from our data why some appear to show higher recharge rates than others. Substrate material permeability, hydraulic gradients and flow direction in the saturated zone are likely to be key factors.

The seasonal pattern in 1993 was different from that in 1992 in the following ways: the spring rise is generally smaller, the rate of summer recession is less and in most holes the late summer is marked by a water table rise. These features are attributed to differences in precipitation patterns between the two years. The summer of 1993 was generally wetter and two large rainfall events occurred in the late summer which were not present in 1992.

All the holes show a net drop in the water table from the annual minimum of 1992 to the same time in 1993. The difference is greatest in the upland and medium depression holes (-0.33 and -0.36 m) and smallest in the fast depression group (-0.19 m). The average drop in water table between the two years is -0.30 m for all UOH holes in the area.

Storage in the saturated zone is closely related to precipitation. The spring rise usually represents the period with greatest accretion, but as demonstrated by the upland holes, heavy rain in the summer can be equally important. Precipitation at any time of the year probably has implications for groundwater storage in that, even though no accretion may take place, losses directly from the saturated zone due to evapotranspiration are reduced. The hydrographs in general indicate that it was the lack of summer rain in 1992 rather than reduced accretion in the spring which is chiefly responsible for the overall water table decline.

All the hydrographs demonstrate that a dynamic equilibrium exists between: i) the rate of water supply at the water table, ii) the rate and direction of flow in the saturated zone, and iii) the rate of evapotranspiration. It is this dynamic equilibrium which governs the position of the water table through time. These ideas are expanded in Section 5 with the development of approaches for estimating recharge.

A key assumption in the development of the hydrograph approach to estimating recharge is that losses from the saturated zone are negligible under frozen conditions. Schneider (1961) concluded from his air temperature and water level observations in Minnesota, that the winter decline of the water table is caused partly by moisture transfers from the saturated zone to a developing frost wedge. Meyboom (1967b) noted a change in the slope of the spring rise in stream hydrographs from the Milk River, Alberta, that he also attributed to thawing of frost. We have assumed that the magnitude of this loss in the Ardmore area, is much less than observed by Meyboom. We did not observe a widespread change in slope in the spring rise that we could attribute to thaw in the frost wedge alone. Two holes with the shallowest water tables and therefore the most likely to show this effect, UOH 11 and 20, do show a step in the spring rise in 1993. This might be the result of frost thawing, but it might also be caused by seepage of ponded meltwater from the surface as the frost wedge receded (see Sec. 5.1 for further discussion).

An interesting closing note is a comparison of the Ardmore data with that of Konoplyantsev et al. (1963). These authors concluded that the ratio of the spring rise to the annual amplitude of water level fluctuations could be used to characterize about 9000 holes measured in the European part of Russia. They found that in "the zone of excessive precipitation" the ratio is always about unity. In the zone of moderate precipitation the ratio was found to lie between 0.70 and 0.85, whereas in the zone of poor precipitation it was found to vary from 0.10 to 0.50. With regard to the magnitude of the annual amplitude, they found for the three precipitation zones the following ranges: 0.91 to 1.83 m, 1.65 to 2.29 m, and 0.64 to 0.82 m, respectively (as quoted by Meyboom 1967b). If the depression holes are excluded, then according to these criteria, the Ardmore holes would lie clearly within the dry zone. If only the depression holes are considered, however, the area would lie in the zone of moderate precipitation on the basis of the ratio of spring rise to annual amplitude. This may have important implications for those areas where recharge is predominantly depression focused.

4.1.3. Individual observation hole characteristics:

Barometric efficiency:

A number of holes with automatic recorders in the Ardmore area showed, varying degrees of barometric efficiency as well as evidence of recharge. The effects of atmospheric pressure changes were not discernible in holes measured by hand, nor from holes where only weekly averages were available. Figure 4.8 shows water levels in three holes plotted together with atmospheric pressures over the period 28 April to 30 June, 1993. It is very obvious from the charts that a drop in atmospheric pressure is accompanied by a rise in water level in the observation holes. The phenomenon results where a relatively rigid high permeability layer is confined between layers of lower permeability. The pressure in the open hole is able to equilibrate quickly with changing atmospheric pressure. Water then moves rapidly from the high permeability layer as the pressure drops in the hole with respect to the formation, whereas water is forced out of the hole as the pressure rises. The barometric efficiency, BE, is defined as the ratio of the water level rise, Δh, to the atmospheric pressure change causing the rise, Δp, expressed as an equivalent depth of water (Jacob, 1940). An equation expressing the ratio is written thus (Kruseman and de Ridder, 1990),

$$BE = \gamma \Delta h / \Delta p$$

where γ is the specific weight of water.

Using this equation the barometric efficiencies in CWW1, UOH 7 and UOH 1 are approximately 0.20, 0.33 and 0.09 respectively. It is interesting to note that the shallowest hole shows the highest efficiency. UOH 7, however, did not intersect any notable sand lenses below the water table, whereas UOH 1 drilled through a 0.6 m thick sand lens at 8.5 m. The presence of barometric efficiency in a hole indicates a combination of both confinement and rigidity. The interesting result for the present study is that recharge can occur not only within a confining layer but also apparently through it.

The slight offset between peaks in the water level and troughs in the atmospheric pressure curves respectively, in the chart for UOH 1, is probably explained by the fact that this hole is the furthest from Cold Lake Weather station where the pressure changes were measured. The straight segments in UOH 7 are periods of no data which resulted from equipment problems.

In conclusion it can be seen that water level fluctuations on the order of 0.03 to 0.08 m can be caused by atmospheric pressure changes and that these can only be identified from continuous records. Where continuous records are available, the general trends in the hydrographs are easily separated (UOH 7, Fig. 4.3). We also note that CWW 1 is completed in the Ethel Lake aquifer and was the subject of the pump testing discussed in greater detail below (Sec. 4.3).

Evapotranspiration:

Many of the observation holes show a clear diurnal pattern of water level changes which were ascribed to evapotranspiration effects. Only perfectly set, Stevens F type recorders were sensitive enough to record these changes, however, so a complete inventory of wells which showed the effect is not possible. Unlike the examples given in most text books, the manifestation of the interaction between saturated flow and losses to evapotranspiration, did not follow the pattern whereby water levels recover during the night time. In our case water levels usually fell rapidly during the day and either stopped falling at night or fell at a reduced rate (Fig. 4.9a). We interpreted the night time fall to result from the downward flow of water in the saturated zone, whereas the higher day time rate of fall was caused by both saturated flow and evapotranspiration. This interpretation is similar to the conclusions reached by Keller et al. (1985) with regard to the night time rate of seepage loss from a small pond. The diurnal pattern was essentially different in UOH 12 (discussed below) and the eleven day period following that for UOH 11 is shown for comparison in Figure 4.9b. Figure 4.9a shows also the effect in UOH 11 of a rainfall event between June 8 and June 11 which resulted in an accretion rate that exceeded, for a short period, the combined rates of saturated flow and evapotranspiration.

Before discussing the diurnal evapotranspiration effects further, some more general observations concerning transpiration are as follows. In the upland holes such as UOH 1, 7 and 23 (Fig. 4.3), there was little difference in recession rates between winter and summer for the year 1992. We reasoned, therefore, that the transpiration losses at these sites were negligible from the relatively deep water table. Also, for 1992, accretion at these sites during the summer was almost completely absent presumably because of a lack of precipitation and a higher than normal soil moisture deficit. In the absence of transpiration losses and with very low levels of domestic withdrawals, the only cause of the summer recession was the same as for winter, namely, downward directed groundwater flow in the saturated zone. Thus, because other inputs and outputs were negligible for most of the year the recession rate was constant.

At lowland sites a much steeper recession curve in the summer time was apparent, which gradually flattened into the autumn. For water tables at intermediate depths, between 2 to 3 m, accretion in the summer 1992 again appeared to have been negligible. In these cases we reasoned that the summer recession was caused by both transpiration and the downward directed saturated flow. These observations lead us to an approach for estimating evapotranspiration at the water table from the difference between summer and winter rates of recession (Sec. 5.1).

UOH 12:

This observation hole is discussed separately from the others, and under the section on evapotranspiration, because it showed a unique hydrograph pattern which was interpreted to be intimately related to the latter process. Figure 4.10a shows the complete hydrograph for the period of study and Figure 4.10b is a detail showing the unusually large diurnal fluctuations with a characteristic night time recovery. In addition to the diurnal fluctuations, the hole shows a hydrograph pattern which differs generally from other observation holes in the following details: i) the annual minima occurred in late summer rather than in early spring, ii) there was a long period of water level recovery in the autumn and winter of 1992, before the usual winter recession began in early 1993, and iii) the water level showed large responses to rainfall even though it was quite deep.

Some pertinent details of the site are as follows, i) the hole was located on the edge of a small isolated poplar woodland, ii) the hole intersected a layered sequence of clay rich horizons interbedded with fluvial sandy deposits, and iii) the water level fluctuated within a till (Fig. 3.7).

The occurrence of both a winter recession and a night time recovery in summer following day time drawdown, appeared at first to be contradictory pieces of evidence. The first suggested the hole was located in a recharge area with downward flow whereas the second was indicative of discharging conditions. The explanation proposed here is that the site experienced high rates of transpiration from the deeply rooted trees, which caused a cone of depression to form under the wood with respect to water levels existing under the adjacent cultivated fields. The fields supported crops which we supposed to have roots that were less deeply penetrating and less well connected to the water table: since the crop root system (Barley) must re-grow each season it was unlikely to be as well established as that of the woodland.

Cones of depression caused by transpiration from concentrations of specific vegetation have been reported in the literature, (e.g., Troxell, 1936; Meyboom, 1962; Davis and Peck, 1986). The night time recovery was attributed to rapid lateral inflow of water from the surrounding area in response to the fluid potential differences. At UOH 12 the inflow occurred through the highly conductive sandy horizons. The cone of depression continued to develop at a fairly high rate through the summer due to the low storativity of the clay horizon and reached a minimum in late July. Its recovery was aided by rainfall events at the end of both summers, but it is theorized that the recovery would have occurred in any event once the transpiration rate declined in the cooler weather. Once an equilibrium was established with the areas outside the cone of depression (e.g., in late November 1992), the normal pattern of groundwater flow was established and the usual winter recession began.

The responsive behavior of the water table to precipitation is thought to have been caused by a capillary fringe effect (Gillham, 1984). A capillary fringe formed in the clay which intersected the overlying sandy layer. Rainwater was able to percolate rapidly through the sand and reach the top of the clay. The rate of water supply was usually fast enough to fill the low effective storage in the capillary fringe of the clay layer, resulting in a rapid rise in water level.

COH 2:

COH 2 is one of the shallow observation holes from the Ardmore Thermal Project. Its hydrograph showed a complete absence of a winter recession: a feature not found anywhere else in the study area. We have records for the hole dating from the start of the project monitoring program in 1977, and in each of the years from 1983 to 1988, the water level showed a recovery during the winter (Fig. 4.12). We also recorded a recovery in the winter of 1992/93. The occurrence of the annual minima in late summer was the same as in UOH 12.

The chart in Figure 4.12 shows that the water table elevation in COH 2 is lower than in COH 1 and that over the years the difference has increased. The explanation for these features is not clear. The hydrograph pattern of the early eighties is very similar to UOH 12, but more recently the site appears to experience more permanent discharge, because the water level continues to recover all through the winter.

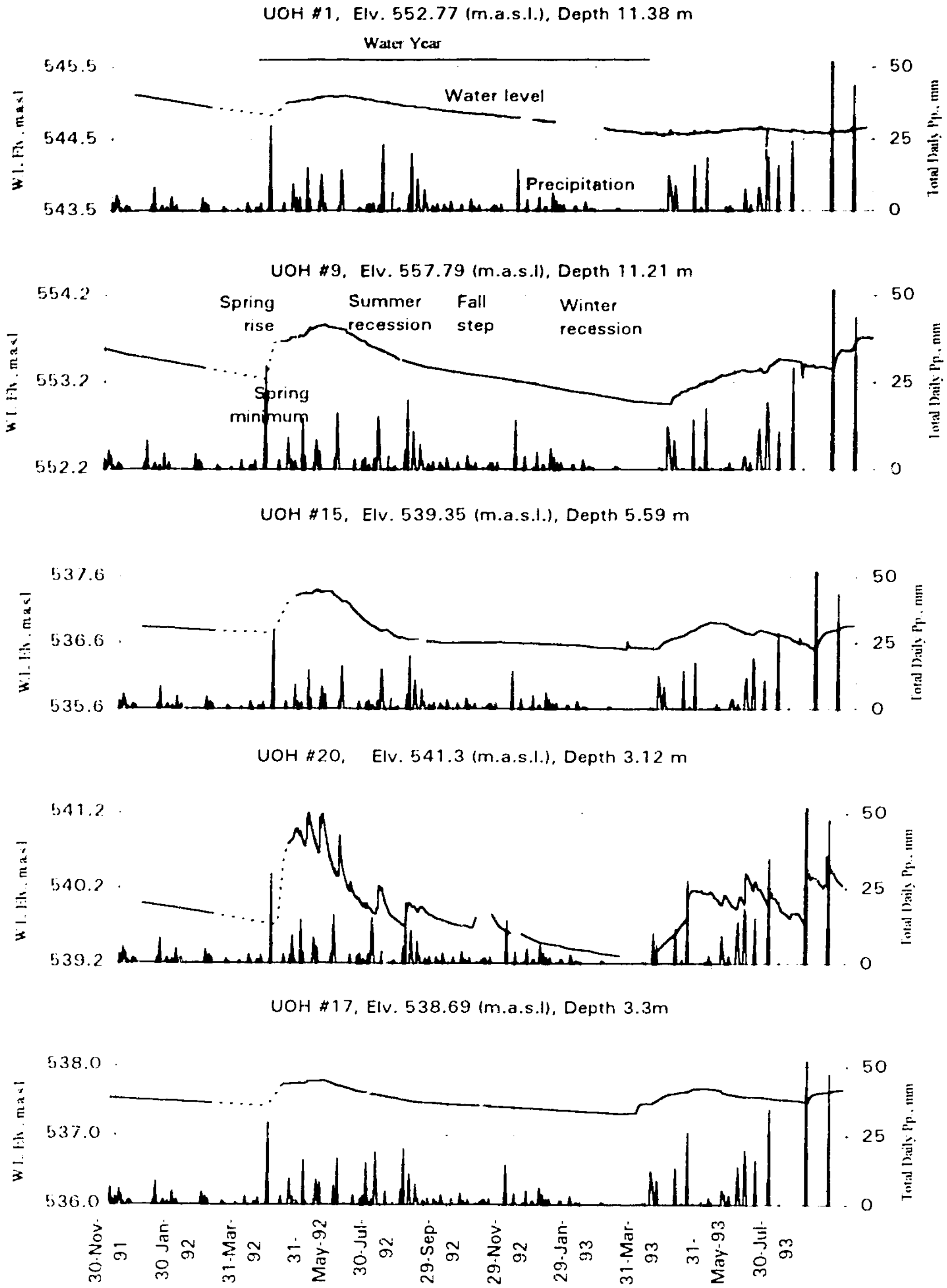


Figure 4.1. Selected hydrographs from UOH.

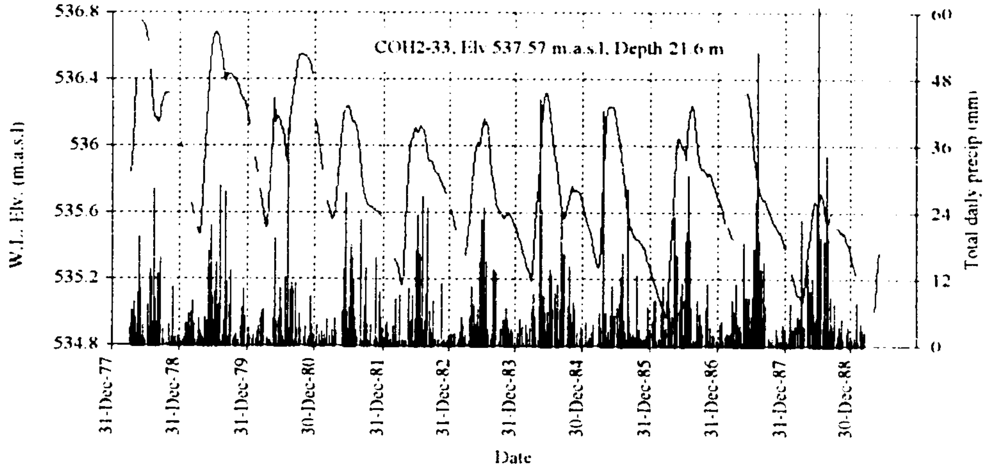


Figure 4.2. Eleven year hydrograph in COH2-33. Water level fluctuations are closely related with depth of precipitation and season. Note the constant winter recession. (Data, courtesy of Hydrogeological Consultants).

Table 4.1. Hydrograph groups summary

A Non-Depression		B Depression Holes	
A 1. Upland		B 1. Slow	B 3. Fast
A 2. Lowland		B 2. Medium	

Group	No. of Wells	Mean Depth Drilled m	Mean Base of screen Depth m	Ground Elv m.a.s.l	Mean W.L. Depth m	W.L. Depth Range m	Water year 1992/93 Date of W.L. Elv max	Water year 1992/93 Date of W.L. Elv min	Spring Rise 1992 W.L. Range m	Spring Rise 1992 Date of W.L. Elv. min	Spring Rise 1992 Date of W.L. Elv. max	Spring Rise 1993 W.L. Range m	Spring Rise 1993 Date of W.L. Elv. min	Spring Rise 1993 Date of W.L. Elv. max
A 1	5	11.62	9.75	555.32	5.00	0.79	25-Jun-92	24-Apr-93	0.46	20-Apr-92	25-Jun-92	0.38	24-Apr-93	27-Jul-93
A 2	6	6.80	5.26	537.94	3.70	0.44	31-May-92	7-Apr-93	0.18	24-Apr-92	31-May-92	0.21	7-Apr-93	10-May-93
B 1	6	7.23	5.29	545.35	2.71	0.99	29-May-92	17-Mar-93	0.80	20-Apr-92	29-May-92	0.50	17-Mar-93	24-May-93
B 2	4	6.93	6.02	547.33	2.80	1.78	25-May-92	16-Mar-93	1.42	19-Apr-92	25-May-92	1.37	16-Mar-93	28-May-93
B 3	4	5.35	3.16	539.63	1.25	0.46	2-Jun-92	22-Mar-93	0.34	17-Apr-92	2-Jun-92	0.32	22-Mar-93	14-May-93

Group	Depth to CaCO3 horizon m	Soil Order	Time period between Rainfall start and the Beginning of water level rise t (h)	Ratio of Spring Rise to Annual Amplitude 1992/92-93	Ratio of Spring Rise to Annual Amplitude 1993/92-93	Net W.L. Change Spring Min 92/93 s m	Recharge R m/s	Net Accretion Water Year 92/93 Ig m/s	Recharge % of pp R/Pav*100 P = 400mm	Net 92/93 Accretion % of pp Ig/P*100 P variable
A 1	0.76	Luvisolic		0.58	0.48	-0.33	1.11E-09	6.99E-10	9.00	5.78
A 2	0.64	Mixed	20	0.42	0.48	-0.25	1.36E-09	5.86E-10	10.83	5.07
B 1	0.34 0.12	Gleysolic	10	0.81	0.50	-0.19	6.25E-10	3.23E-10	4.93	2.85
B 2	0.22 0.03	Gleysolic	2	0.80	0.77	-0.36	1.35E-09	1.02E-09	10.68	8.11
B 3	Sandy	Gleysolic	11	0.74	0.71	-0.12	1.79E-09	1.36E-09	14.14	11.32

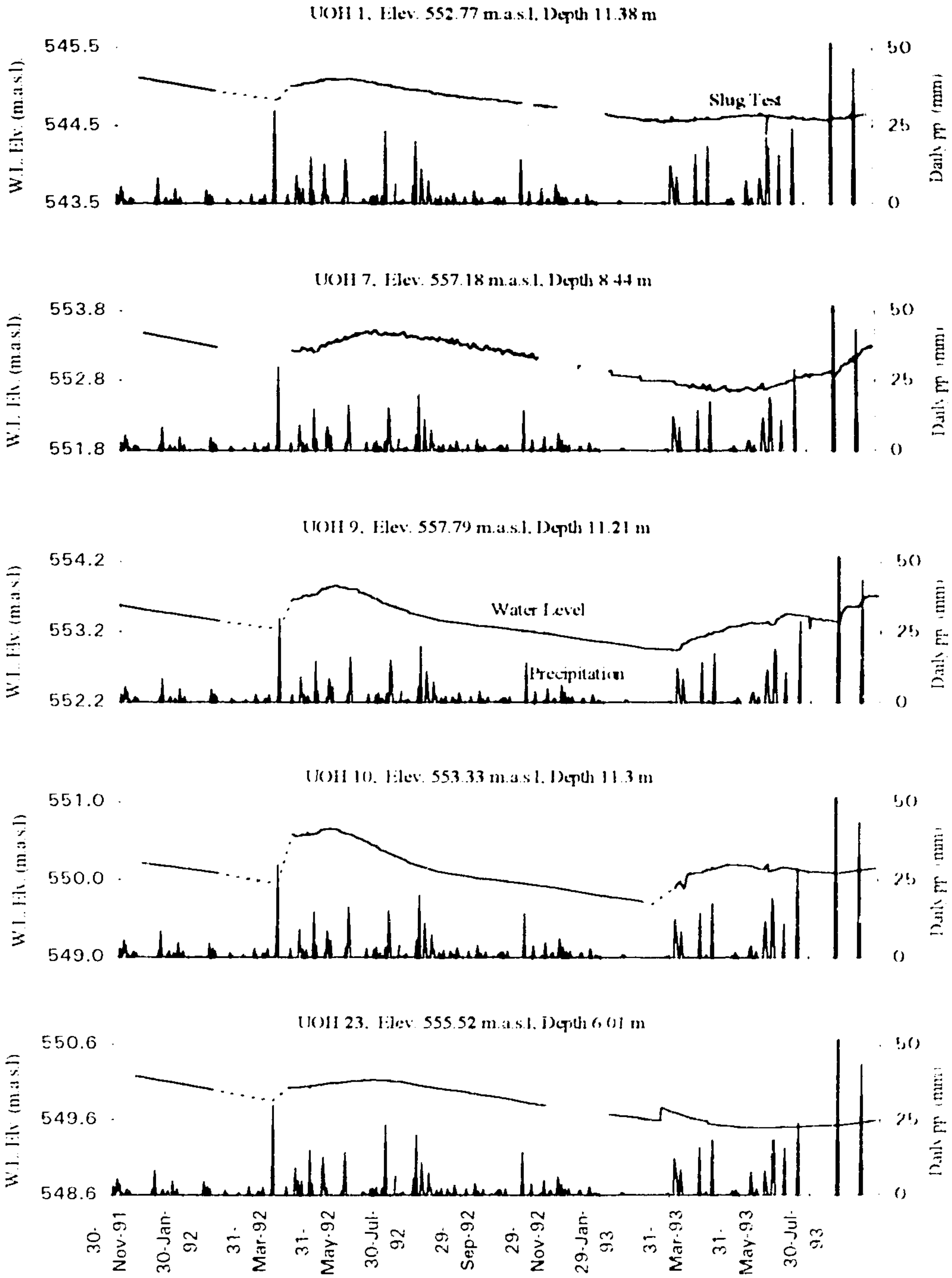


Figure 4.3. Group A1 non-depression upland holes.

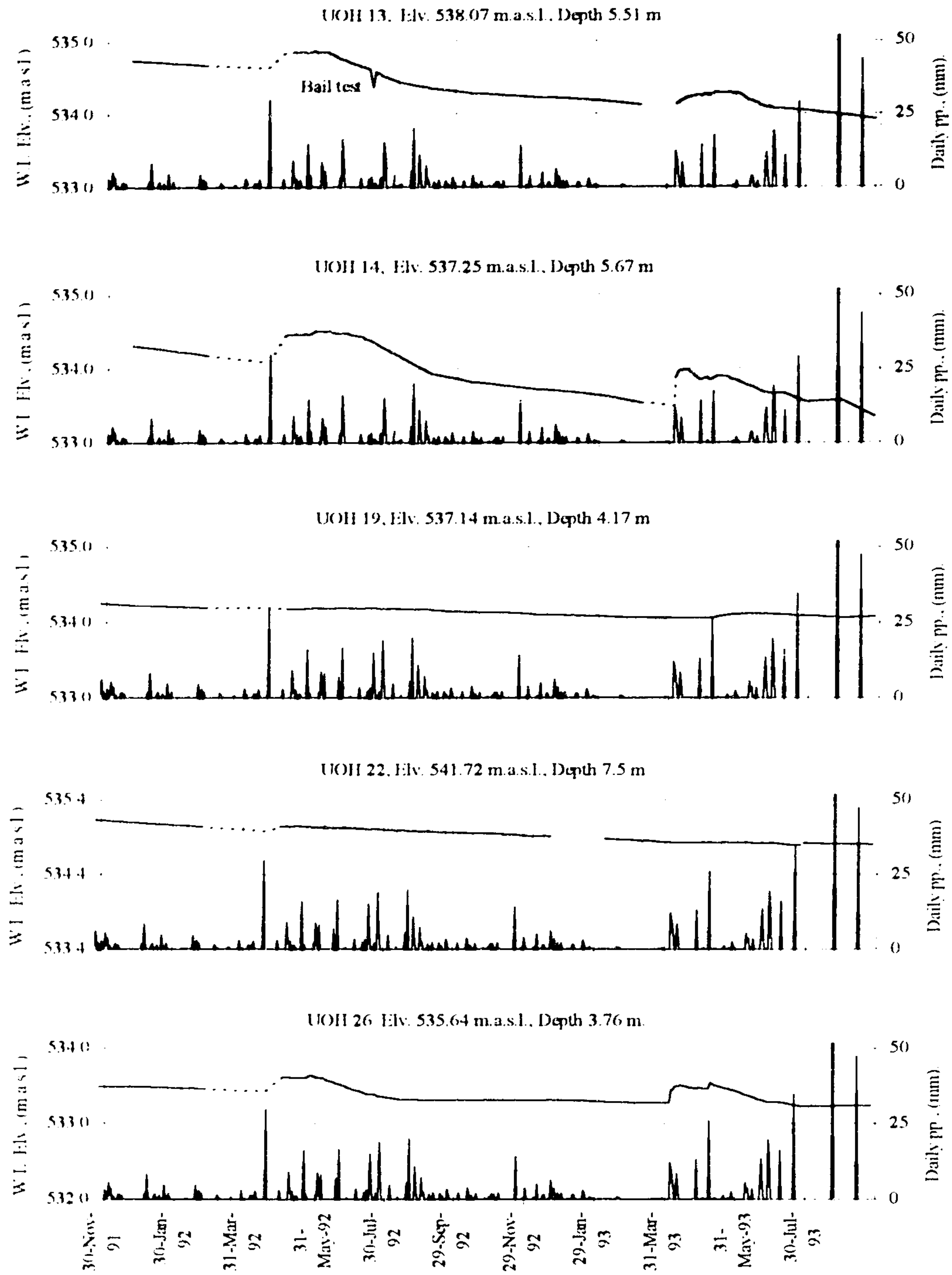


Figure 4.4. Group A2 non-depression lowland holes.

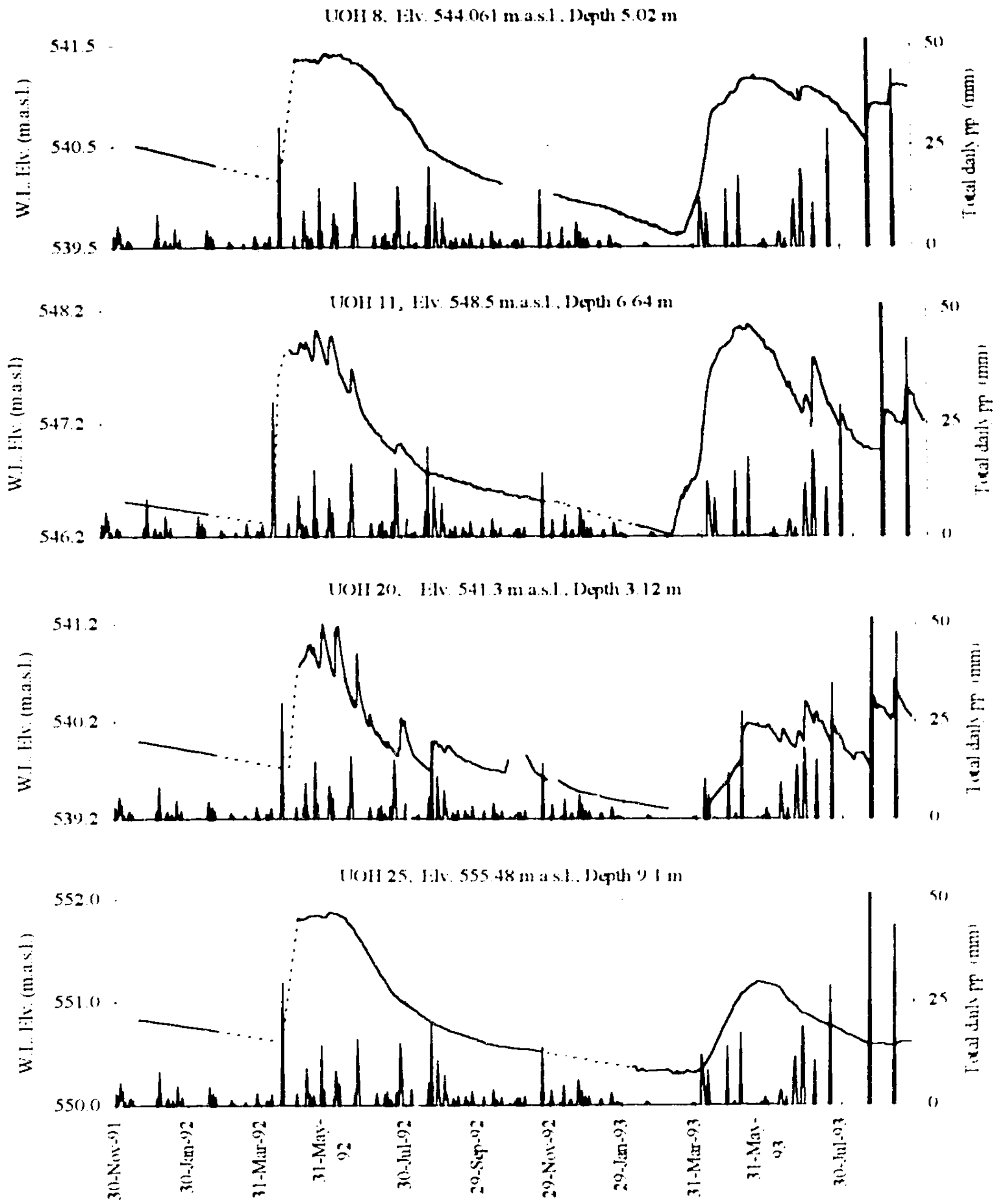


Figure 4.5, Group B2 medium depression holes.

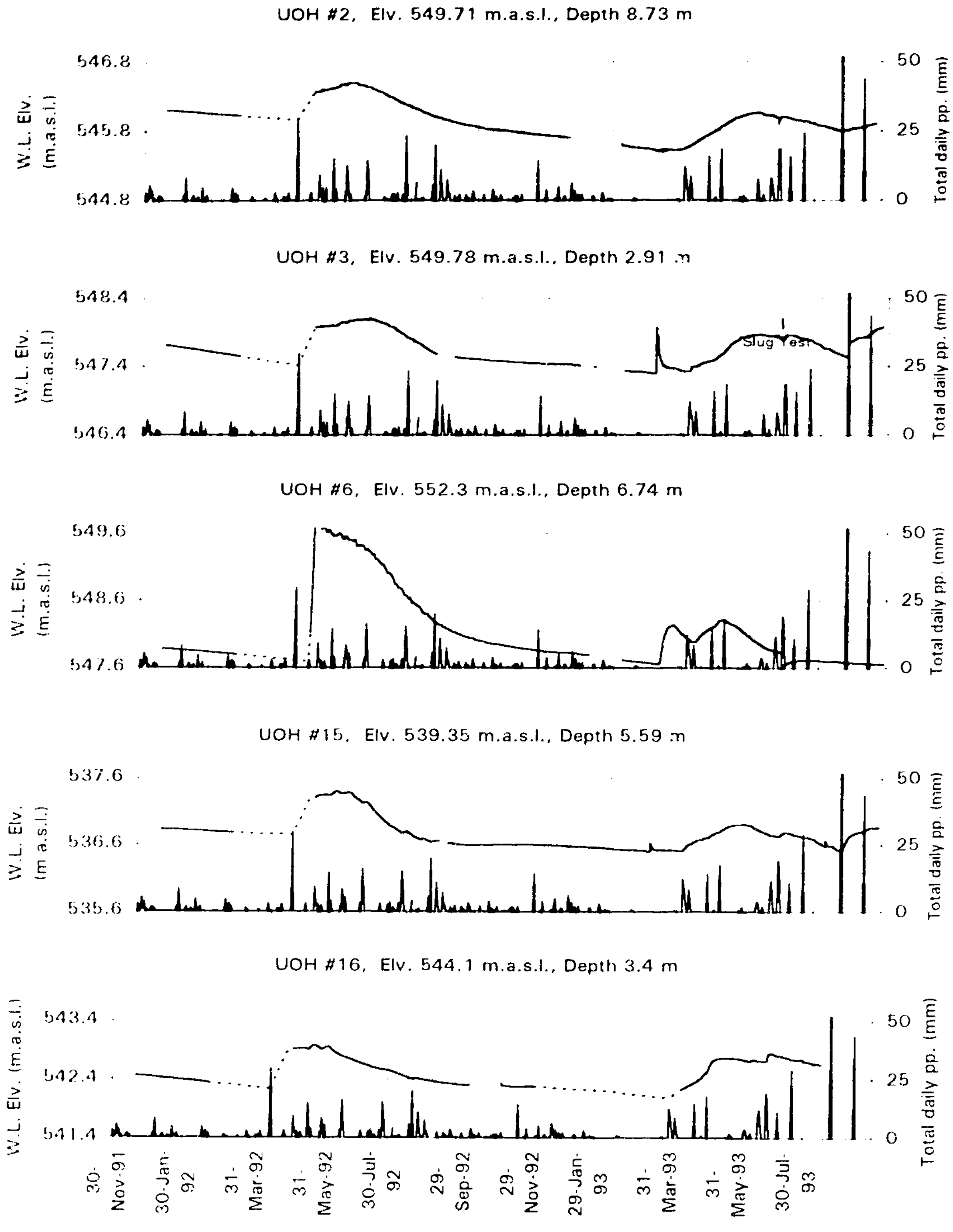


Figure 4.6. Group B-1 slow depression holes.

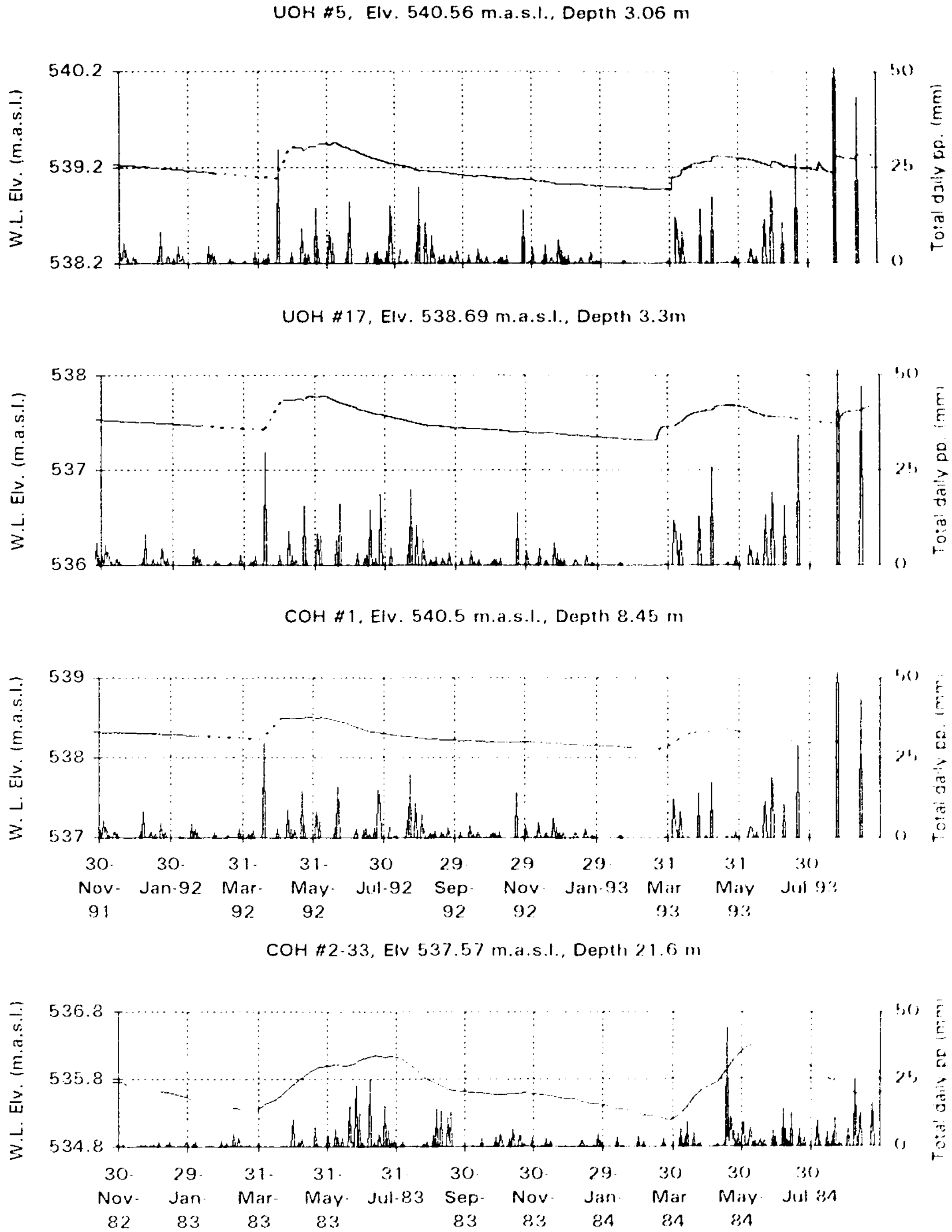


Figure 4.7. Group B-3 Fast depression holes.
The hydrograph for COH2-33 covers a different time period

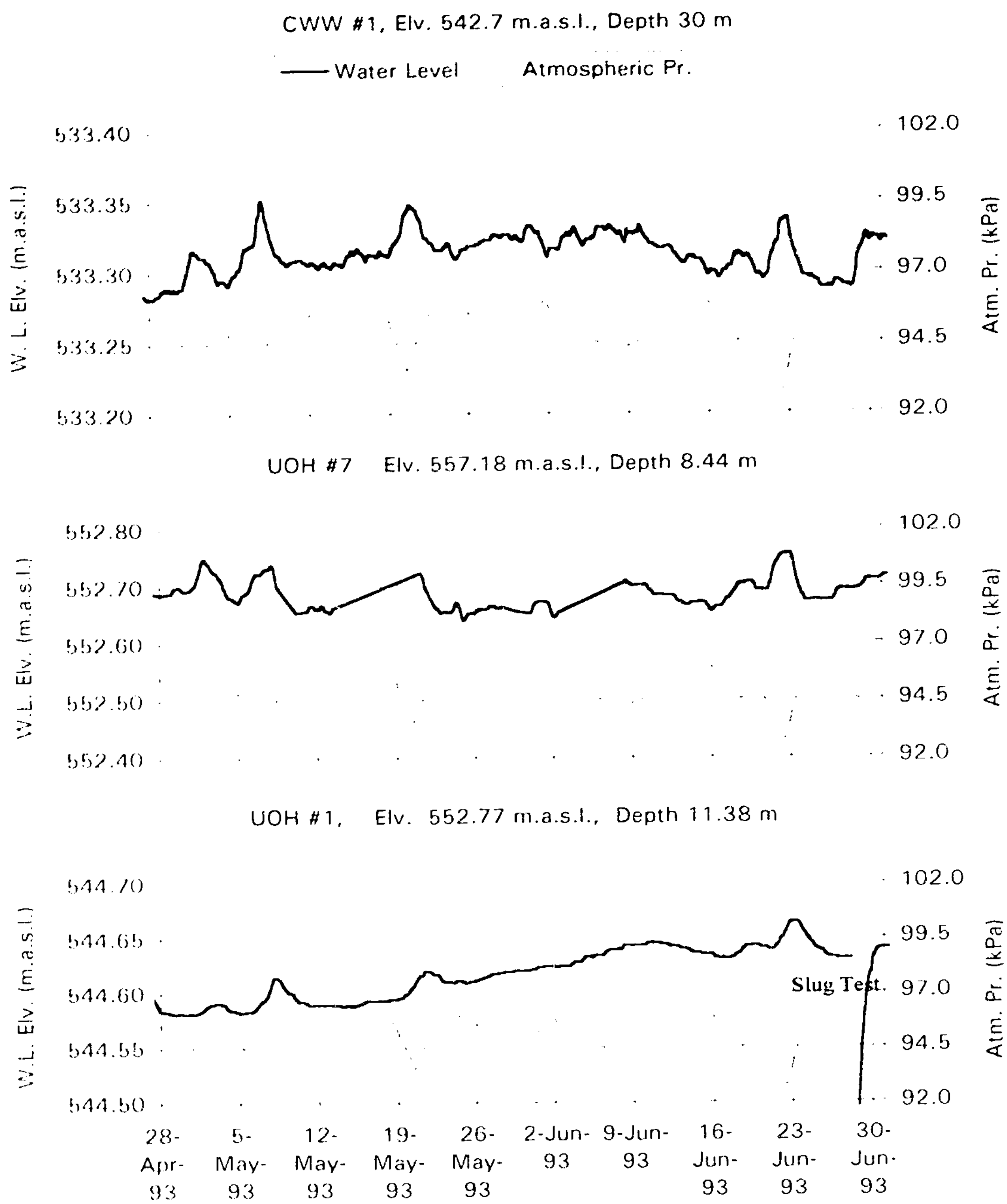


Figure 4.8. Barometric effects. Note vertical scale change in UOH 7.

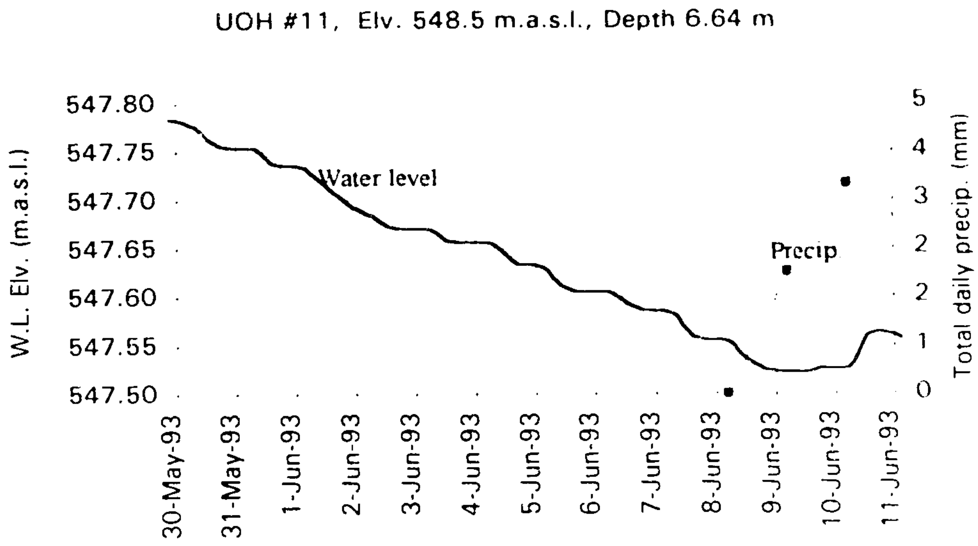


Figure 4.9a. Diurnal water level fluctutations, UOH11.

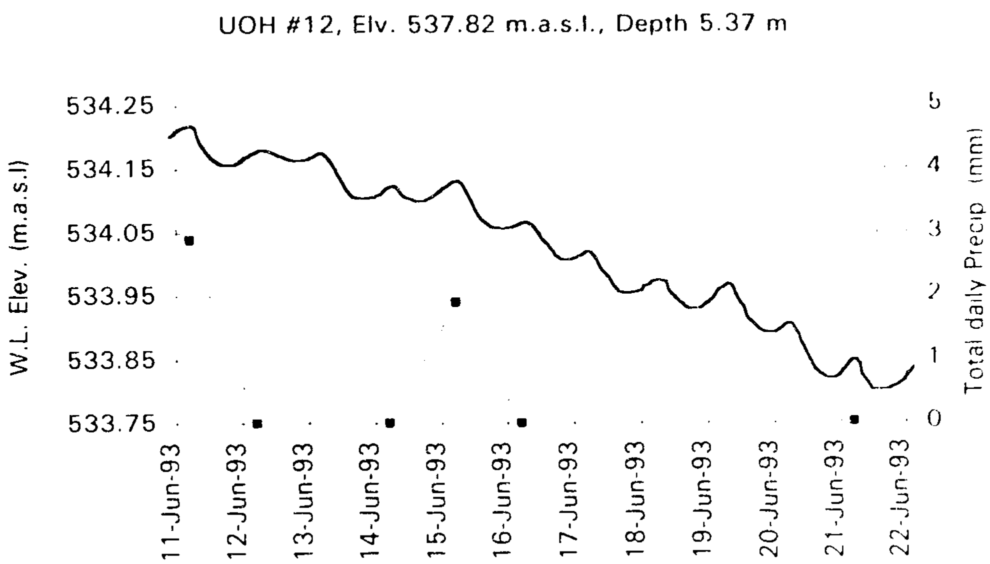


Figure 4.9b. Diurnal water level fluctuations, UOH12. Fluctuations attributed to the effects of saturated flow at night and evpotranspiration loss, together with saturated flow during the day.

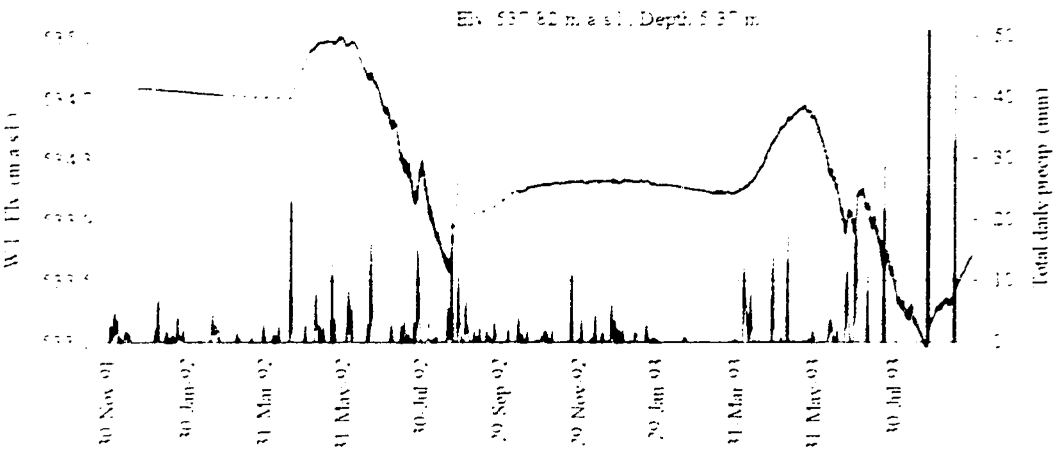


Figure 4.10a. UOH12 hydrograph.

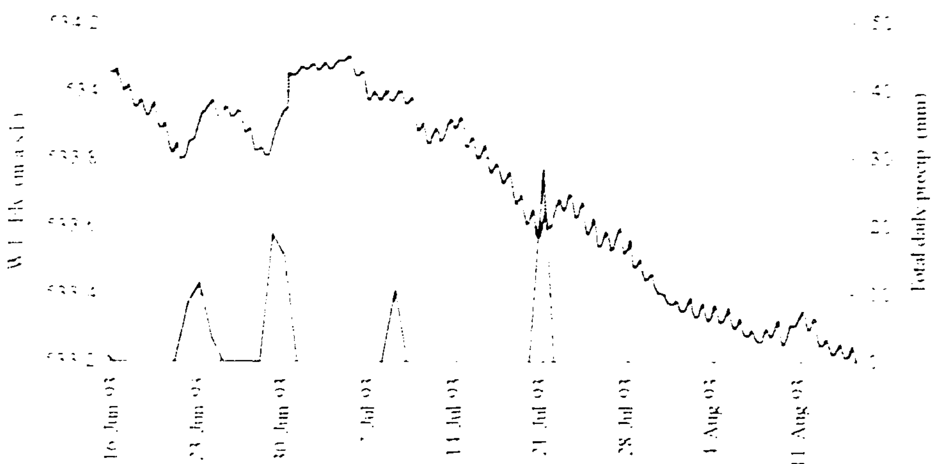


Figure 4.10b. UOH12 period with diurnal effects.

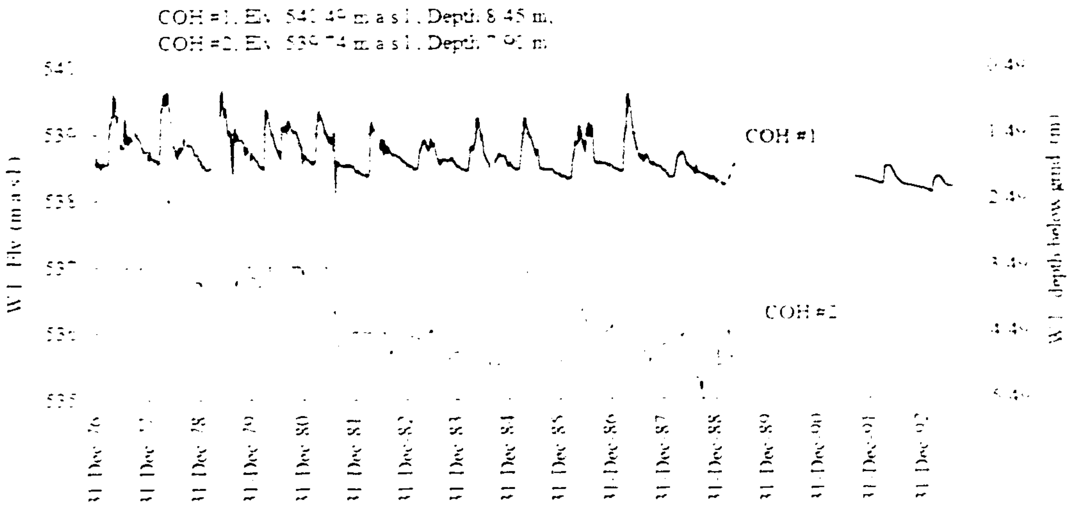


Figure 4.11. Hydrographs from COH 1 and 2. The depth scale applies only to COH 1. Water levels in COH 2 are slightly deeper. (Data, courtesy of Hydrogeological Consultants).

4.2. Flow pattern

4.2.1. Hydrosratigraphy

There are four sand rich formations present in the Quaternary sequence and each one is exploited for water supplies in different parts of the study area. They are, in ascending order, Empress, Muriel Lake, Ethel Lake and Sand River Formations (Fig. 2.5). Except where noted otherwise, they are separated from one another by much less permeable tills (also given formation status by Andriashek and Fenton, 1989) which act as aquitards. The lithostratigraphic and hydrostatigraphic units within the sequence are more or less equivalent, because of the large contrasts in hydraulic properties between the sandy formations and the tills.

Below is a brief description of the lithology, distribution and thickness of each sandy formation. The lithological descriptions are taken from Andriashek and Fenton (1989) unless specified otherwise.

Empress Formation:

The formation is well defined only in the base of Bronson Lake Channel, where the upper surface is located at an elevation between 480 and 475 m.a.s.l. (Fig. 4.12). The formation varies generally between 6 and 10 m in thickness (Fig. 4.13). In the Ardmore area the Empress Formation is represented only by unit 3, the uppermost unit (Andriashek and Fenton, 1989), and believed to be Quaternary rather than Tertiary in age. The formation is composed mainly of sands and lenticular gravel deposits which are interpreted as glaciofluvial in origin. The Empress Formation may also be present in the deep eastern end of Big Meadow Channel but it is overlain directly by Muriel Lake Formation and the two are treated as one hydrostratigraphic unit. In Bronson Lake Channel, on the other hand, the Empress Lake Formation is overlain by a dense clay rich till (called the Bronson Lake Formation) and pumping tests in this area indicate that the formation is confined.

Muriel Lake Formation:

The Muriel Lake Formation is absent over the bedrock high beneath Piech and Ardmore ridges but is fairly continuous through the remaining portions of the study area (Fig. 4.14). The upper surface is usually at approximately 500 m.a.s.l., but dips down to 480 m.a.s.l. at the east end of Big Meadow Channel. The formation thickness increases noticeably from an average of around 5 m outside the channels to 20 m in Bronson Lake and 40 m in Big Meadow Channels respectively (Fig. 4.15). The rapid thickening in Big Meadow Channel may be explained because the formation is not differentiated from the Empress Formation as previously noted.

On the western side of the study area, the Muriel Lake Formation is hard to distinguish from the overlying Ethel Lake Formation in areas outside of the bedrock channels. It appears from correlations of geophysical logs that the two formations may combine in this area where no intervening till unit is present (Fig. 3.5).

The formation is described as consisting generally of silt, sand, and sand and gravel with minor silt and clay beds all of glaciofluvial or glaciolacustrine origin.

Ethel Lake Formation:

Sandy units at elevations between approximately 515 and 520 m.a.s.l. are mapped throughout the area from drilling records, but become very thin (< 1 m) beneath Ardmore and portions of Piech ridges (Figs. 4.16 and 4.17). The units are included together as Ethel Lake Formation which is thus thought to be more or less continuous throughout the area: although in the last two places mentioned it may not yield sufficient water for pumping. The thickest development of the formation occurs in the west where it may directly overlie and be in communication with, Muriel Lake Formation. Elsewhere it is usually around 2 m thick but reaches 8 m adjacent to the east end of Big Meadow Channel. The formation is described as consisting mostly of silt and clay with minor amounts of sand and gravely sand. As we mapped it from drillers' descriptions of cuttings, however, it generally comprises sandy units. These may be quite lenticular in shape and not directly connected with each other. Below the Excel Energy site and in Big Meadow Channel on the east side of the area the formation may again come into direct contact with the underlying Muriel Lake Formation by thickening at the expense of the intervening till unit.

Sand River Formation:

Sandy units of the Sand River Formation are found at elevations between 535 and 540 m.a.s.l. (Fig. 4.18). As with the other formations it appears to dip below its usual elevation range in the east end of Big Meadow Channel. It is either discontinuous or completely absent below much of Muriel Creek Valley and portions of Ardmore Ridge. Elsewhere in the south west of the study area it is usually less than 2 m thick. The formation thickens to 6 m in the west end of Big Meadow Channel and to over 16 m at the east end. It is also thick beneath the Krill Upland where it reaches 14 m (Fig. 4.19). The formation is described as consisting of stratified sand and silt with lesser amounts of clay and gravel. The sand is generally fine to medium grained, well sorted and commonly free of pebbles.

An outcrop of clean fine to lower medium grained sand was found above the springs at the north edge of Dery Uplands at the crest of the Beaver River Valley (Fig. 4.19). It is interpreted as being Sand River Formation on the basis of its lithology and elevation. The sand was massive at first appearance, but weathered surfaces showed thin cross-beds that where hard to identify otherwise due to the uniform colour and texture of the outcrop. The sand may have been either glaciofluvial or glaciolacustrine in origin. The unit had no exposed contacts and was up to 6 m in thickness.

Table 2 in Appendix B shows the drill hole information sorted by completion elevation. It is noticeable that several wells are completed at a shallower depth than the (uppermost) Sand River Aquifer. These wells exploit sandy horizons found in the overlying till (Grand Center Formation), in the east part of the area below Piech and Ardmore Ridges. Many of these are large diameter wells (+60 cm) which do not exploit extensive aquifers.

4.2.2. Water level maps

The mean of the annual fluctutation of the water table varies between a maximum of approximately 5-8 m below ground to a minimum of just under 0.60 m. The water table reaches its maximum depth below uplands and sharply convex slopes, whereas it is shallowest below lowlands. It thus forms a subdued replica of topography (Fig. 4.20). From the water table contours it appears that flow is outwards from the study area in the shallow saturated zone. The maximum estimated horizontal gradients, excluding the Beaver River Valley, are on the order of 8 m drop in hydraulic head for 600 m of horizontal distance (1.3×10^{-2} m/m). These gradients only occur on the flanks of ridges and over distances of less than 1 km. The lowest horizontal gradients occur over the Alexander Lowlands and are on the order of 4 m drop over 4000 m (1.0×10^{-3} m/m) or less.

The water level map for the Sand River Formation shows that water level elevations are close to the water table below the Alexander Lowlands and the formation is thought to be more or less unconfined through much of this area (Fig. 4.21). Below the western portions of the study area the water level elevations are as much as 3 m below the water table under the ridges and represent maximum vertical hydraulic gradients on the order of

0.25 m/m. The equipotential contours still follow generally the present day topography, and flow is more or less outwards from the area except where it parallels the boundaries in the south and west. Horizontal hydraulic gradients are similar to those found at the water table.

In spite of the very low horizontal hydraulic gradients, potential lateral fluxes in the sandy layers are probably always higher than potential vertical fluxes across the till units. If we assume a vertical hydraulic conductivity, K_z, for the uppermost till in the shallow subsurface to be on the order of 1×10^{-8} m/s (Keller et al., 1988; Bulk K = 5×10^{-9} m/s), the vertical flux can be calculated using Darcy's law as 2.5×10^{-9} m/s from the surface to the Sand River Formation; where the flux $q = -K_z dh/dz$, and dh/dz is the vertical hydraulic gradient. The horizontal flux within the Sand River Formation, on the other hand, calculated in the same manner can be as high as 1.3×10^{-6} m/s and as low as 1×10^{-7} m/s with the hydraulic conductivity estimated at 1×10^{-4} m/s.

Water levels in the Ethel Lake Formation show the first clear departure from the influence of minor differences in surface elevation (Fig. 4.22). The contours decline in value uniformly from a high under the mid portion of Ardmore Ridge towards the Beaver River Valley. Along the southern boundary, there is not much difference in elevation between Ethel Lake and Sand River Formation water levels. The descending flow may be very weak in this area and horizontal flow along the length of Muriel Creek Valley probably dominates. There may in fact be discharging conditions in places along the west end of the valley: surface manifestations in the form of saline seeps are seen further west outside the study area. The differences in elevations between Ethel Lake Formation water levels and those of the overlying aquifer increase towards the Beaver River Valley. Below Dery Uplands the hydraulic gradient appears to approach 1.0 m/m. There is still little possibility for inflow of groundwater to the study area, since the flow directions are either outwards across the boundaries or parallel with them.

The flow pattern in Muriel Lake Formation appears to follow closely that of Ethel Lake Formation, with the exception of some inflow and outflow across the south west corner of the area (Fig. 4.23). Vertical hydraulic gradients across the till units between these formations are estimated at approximately 0.3 m/m. Flow in the underlying Empress Formation is along the length of the Bronson Lake Channel towards the north west (Fig. 4.24). There is also inflow and outflow in this formation across the south west corner of the area, but it is estimated that there is negligible net effect on the water budget.

The water level maps support the general picture of a dominantly descending flow regime beneath the area. The maps are not detailed enough to show more than a general picture, and local flow systems cannot be identified. The maps of the water table and Sand River Formations contain closures which suggest that local discharge can occur in the low lying areas, particularly below the sloughs of Percy Lowlands and in the Tri-Lakes Corridor. There may also be weak discharge below reaches of the Muriel Lake Valley. As discussed in the previous chapter, however, field mapping did not identify areas with strong perennial discharge although discharge from shallow local flow systems is harder to identify because it may not be associated with saline soils or high hydraulic gradients.

Vertical cross-sections, generated from the gridded surfaces, are thought to represent somewhat idealized and smoothed geology. One cross-section (A-A') is constructed along the length of the study area (Fig. 4.26) and three others (B-B' series) are constructed orthogonally to it (Figs. 4.27, 4.28, 4.29). The section locations are shown in Figure 4.25. An attempt has been made to represent sub-surface flow directions schematically in the sections. Where the arrows terminate in mid-section the flow becomes dominantly into or out of the section. The south end of section B1-B1' (Fig. 4.27) suggests that some discharge may occur to the Muriel Creek Valley as mentioned above. The valley is underlain by the Bronson Lake Channel in this area, however, and most of the deeper flow is probably along the valley and out of the section within the Muriel Lake and Empress Formations. The potential for local flow systems developing under more undulating topography is greatest in the western half of the study area. All the sections indicate a steepening in the hydraulic gradients into the Beaver River Valley. The hydraulic head values in the formations beneath the river are not known, as mentioned previously, so the flow patterns at the base of the valley are largely conjectural. This is an important consideration for quantitative estimates of flow within the flow systems discharging to the valley.

For the purposes of preliminary estimates of groundwater flow beneath the study area, the flow regime was approximated by a single intermediate flow system. The recharge area for the system was the whole study area above the Beaver River Valley. The discharge area was the valley itself.

4.2.3 Water level relationships with depth and elevation

In a descending flow field, fluid potentials must decrease with depth because water will only flow from areas of high potential towards areas of low potential. This statement follows from the relationship between energy and flow as given by a version of Darcy's Law as follows (Hubbert, 1940).

$$q = -K\frac{1}{g}\frac{d\phi}{dl} \qquad [L/T] \qquad (4.0)$$

where q is the saturated flux, K is a combined material parameter known as the hydraulic conductivity that takes into account properties of both the solid medium and the fluid, g is the acceleration due to gravity, ϕ is the total fluid potential energy and l is a distance over which the difference in potential energy is measured.

Hydraulic head, h, may be used as a measure of fluid potential through the relation,

$$h = \phi/g \qquad [L] \qquad (4.1)$$

which relation holds where the compressibilities of water and rock can be considered negligible. Thus the familiar form of Darcy's Law is directly applicable to our case.

$$q = -Kdh/dl \qquad [L/T] \qquad (4.2)$$

In a region of descending flow we therefore expect hydraulic head to decrease as the depth of measurement increases. A corollary to this statement is that depth to water in a hole will increase with the depth drilled. This relationship was first explained by Tóth, (1962), who found it to be a consequence of the theoretical potential distribution and boundary conditions of a modeled basin. The plot in Figure 4.30, using all the water level data available, shows that this is the relation found at Ardmore and supports the conceptual model of a dominantly descending flow direction.

If we now consider the relation between hydraulic head and elevation, it can be seen that the head in a descending flow field will decrease directly as the elevation of the point of measurement decreases; assuming that elevation is always positive upwards.

It is sometimes quite useful in a small area to examine the relationships discussed above to determine, i) whether the flow direction is generally ascending or descending and ii) to obtain preliminary estimates of the vertical hydraulic head gradients, providing the points of measurement are not too far apart in the horizontal direction. This exercise is most useful when time is not available for the construction of detailed water level maps and profiles. Some plots of these relations are presented here by way of summary and confirmation of the flow patterns outlined thus far.

To reduce the scatter seen in Figure 4.30, arithmetic averages of the water levels were calculated and assigned to each aquifer. Plots of mean water level elevation (which we considered to be equivalent to total hydraulic head) were then made with either mean elevation or depth of the aquifers as indicated by the well completion (Figs. 4.31a and b). We maintained the well completion depth and elevation on the vertical axis of the plots. This may seem intuitively correct but it is slightly confusing in terms of hydraulic head gradient. The gradient increases as the slope of the plotted curve decreases.

Following the preparation of these plots we noted differences in the estimates of hydraulic gradients between those from depth and those from elevation relationships. For example in Figure 4.31a the difference in the mean water level elevation in the shallow aquifers is quite large, whereas the depth of well completion for each is similar. This is because the uppermost aquifer is only present under uplands or ridges (previous section), and completion depths are always shallow. On the other hand, the Sand River Aquifer is well developed particularly under the Alexander Lowlands and completion depths here are also shallow.

The elevation plot eliminates the difficulties caused by undulating topography, but data from different parts of the flow regime will still be scattered.

To this end we subdivided the study area, to evaluate the two trends noted earlier, on the basis of expected similarities in flow patterns. The elevation plots showed that sub-hydrostatic conditions appeared to exist in each subdivision (Fig. 4.32).

Groundwater flow results from fluid potential differences in the subsurface. Hydrostatic conditions are those which exist where water levels in wells drilled to any depth will rise to the elevation of the water table. In other words there is no change in the value of hydraulic head with depth or elevation below the water table. The condition can result from either no flow (hence the term hydrostatic) or where flow is exactly horizontal between the point of measurement and the water table. The term sub-hydrostatic describes the condition where water from a point in the flow regime rises to elevations below that of the water table. Non-hydrostatic (dynamic) hydraulic head conditions indicate groundwater flow with a vertical component. A difference in hydrostatic and dynamic head, associated with flow, is here called the dynamic head increment.

To examine the factors controlling the hydraulic head gradient (and thus the magnitude of the dynamic head increment) we can re-arrange Equation 4.2 as,

$$q_z/K = -i_z, \qquad [L/L] \qquad (4.3)$$

where i_z is the vertical gradient dh/dz. The gradient, i_z, is observed to be directly proportional to the vertical flux, q_z, and inversely proportional to the hydraulic conductivity, K, of the medium.

A change in the slope may therefore indicate either, a change in the vertical flux, (which could result from a change in the direction of flow), or a change in the hydraulic conductivity. Since these variables are intimately related it may be difficult to determine which one dominates. An increase in the permeability, for example, will increase hydraulic conductivity, reduce the flux and change the flow direction due to refraction of flow lines.

In general, the hydraulic head gradients are constant in the study area within each subdivision, and project through an origin at the water table. This is interpreted to indicate unconfined conditions and constant vertical fluxes between aquifers, through tills of very similar hydraulic conductivities. The break in slope at about 500 m.a.s.l. in the dynamic head curve from the Muriel Creek Valley subdivision, indicates a new set of conditions between Muriel Lake and Empress Aquifers (Fig. 4.32c). The hydraulic conductivity of the intervening till (Bronson Lake Formation, Andriashek and Fenton 1989) is slightly lower than in other tills. Since the change in slope indicates a reduction in the vertical hydraulic-head gradient, the vertical flux must be considerably less through this till. This conclusion is in agreement with pumping records from the aquifer showing confinement.

It should be pointed out that the same change of slope could have resulted from an increase in hydraulic conductivity below the Muriel Lake Aquifer and a constant flux.

The fact that different gradients are observed between different subdivisions provides the explanation of the trends observed in the plots of raw data. There are at least two dominant trends, one associated with increasing gradients adjacent to the Beaver River Valley and a second with areas of weaker vertical gradients away from the river. In addition, the aquifers appear to be essentially unconfined with the exception of Empress Formation.

Finally, an equivalent freshwater head value from the Mannville Group (Hitchon et al., 1989) was included in the dynamic head plot of the Ardmore data to examine, by this method, the potential for flow across the Lea Park-Colorado Group mudstones (Fig. 4.33). The head value remains sub-hydrostatic and is also less than the value in the Empress Formation at the base of the Quaternary. Thus the potential for flow across the Lea Park-Colorado Group is confirmed.

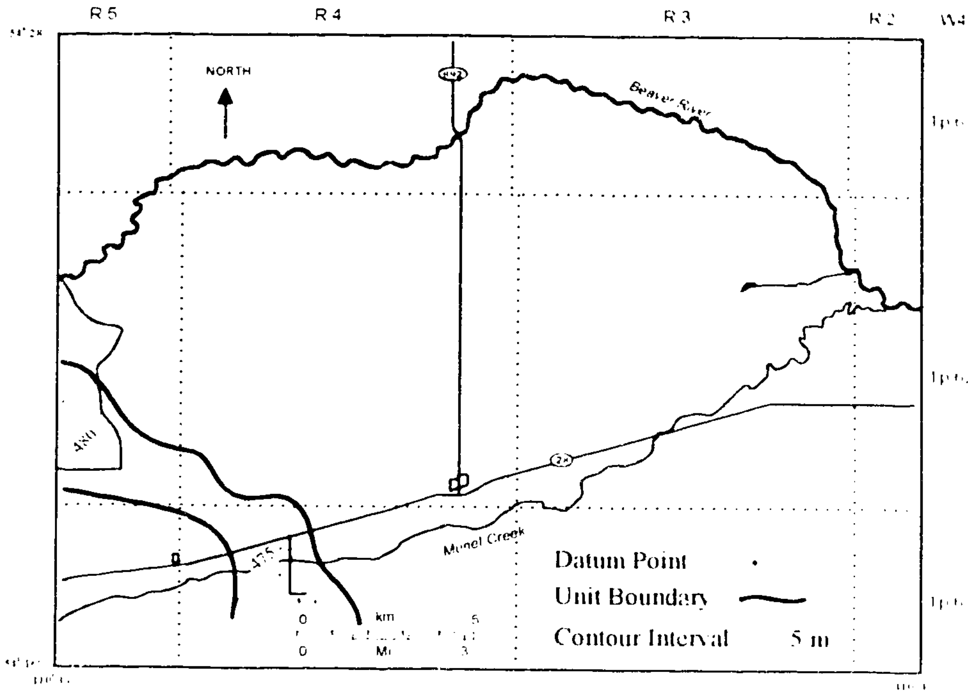


Figure 4.12. Empress formation top surface elevation.

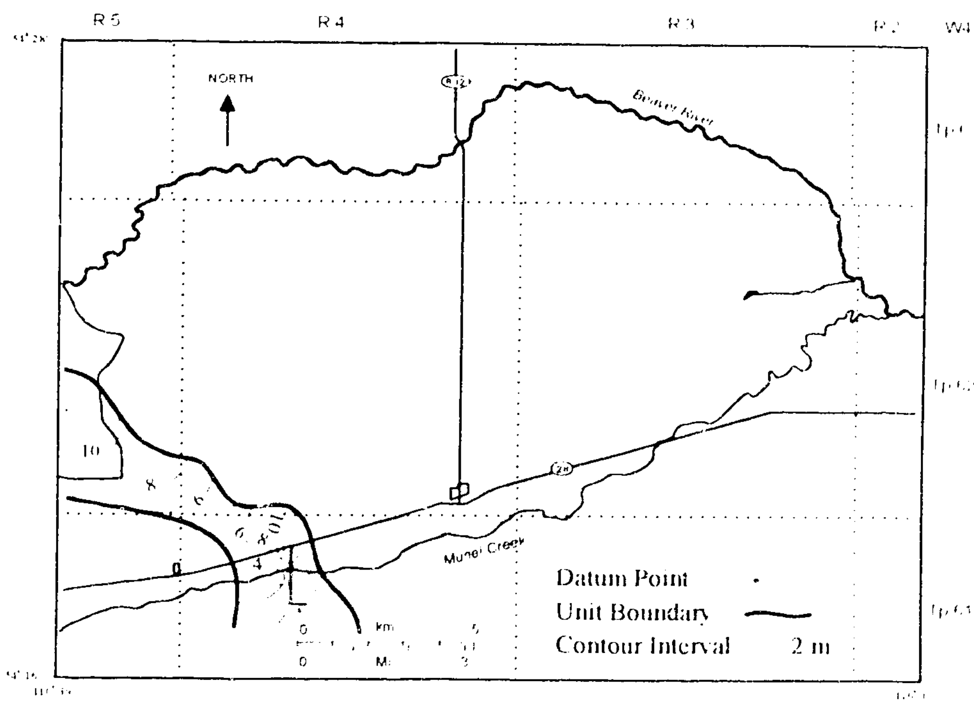


Figure 4.13. Empress formation isopach.

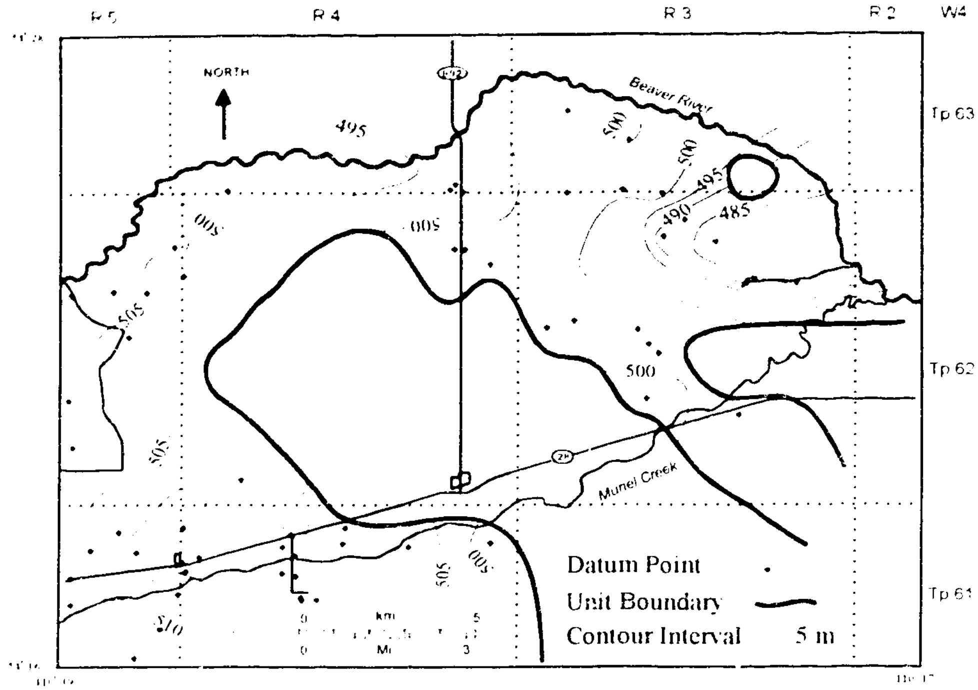


Figure 4.14. Muriel Lake formation top surface elevation.

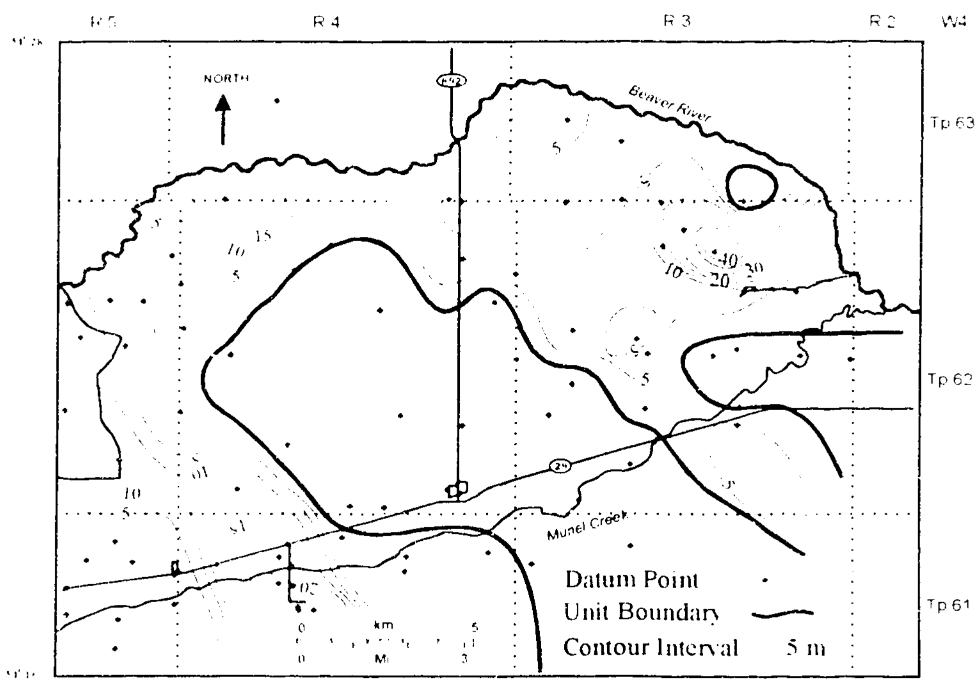


Figure 4.15. Muriel Lake formation isopach.

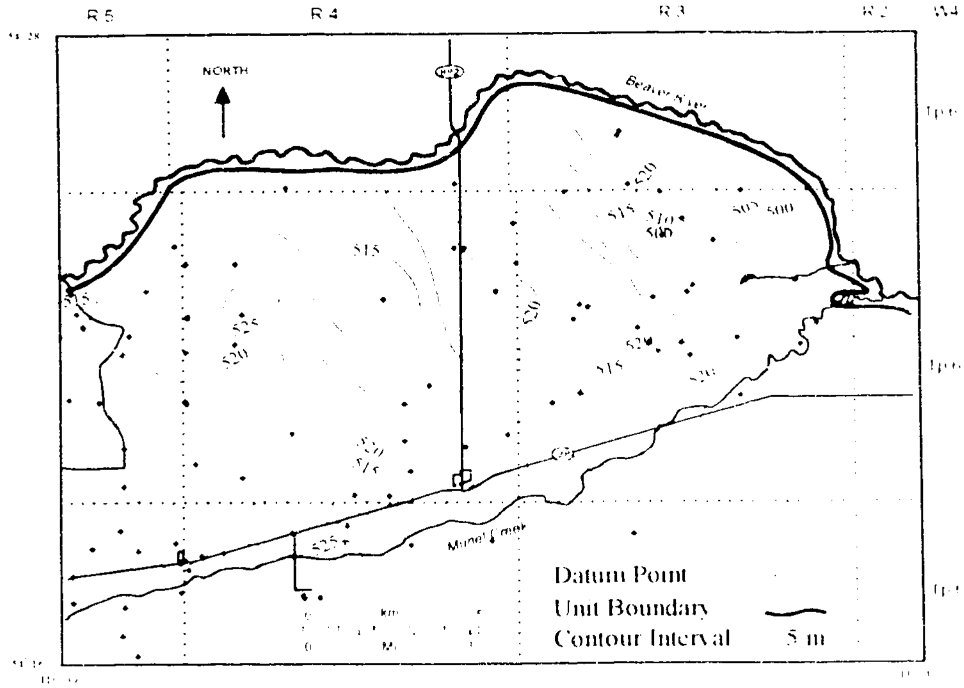


Figure 4.16. Ethel Lake formation top surface elevation.

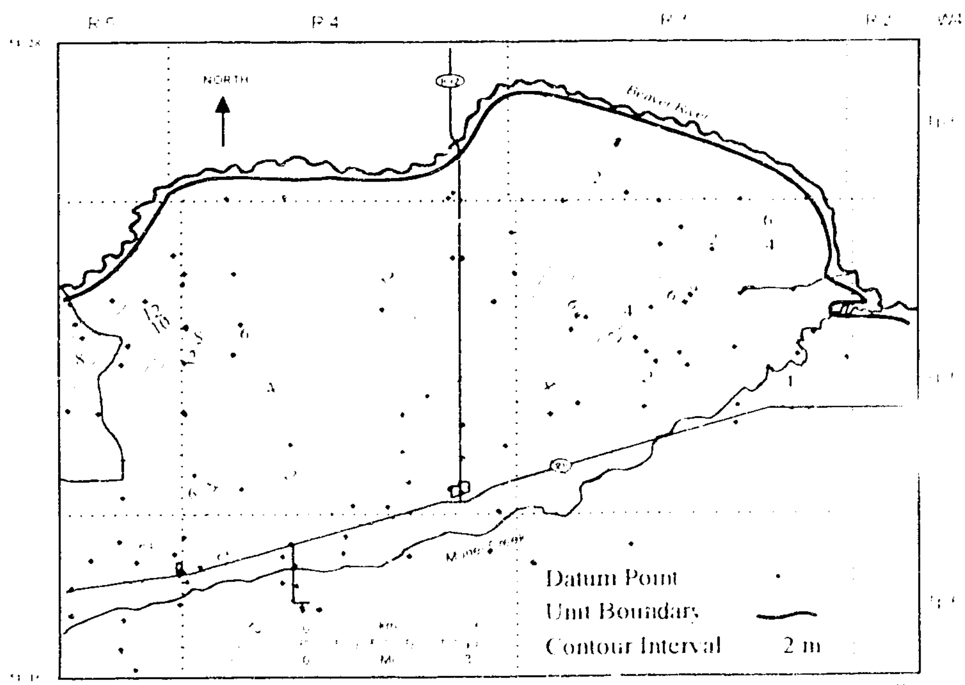


Figure 4.17. Ethel Lake formation isopach.

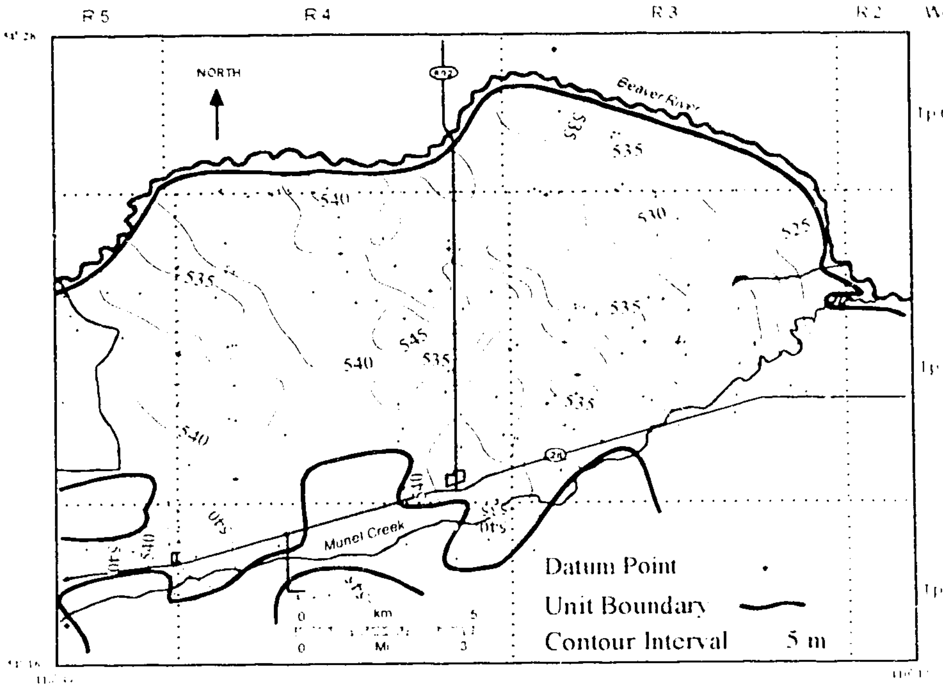


Figure 4.18. Sand River formation top surface elevation.

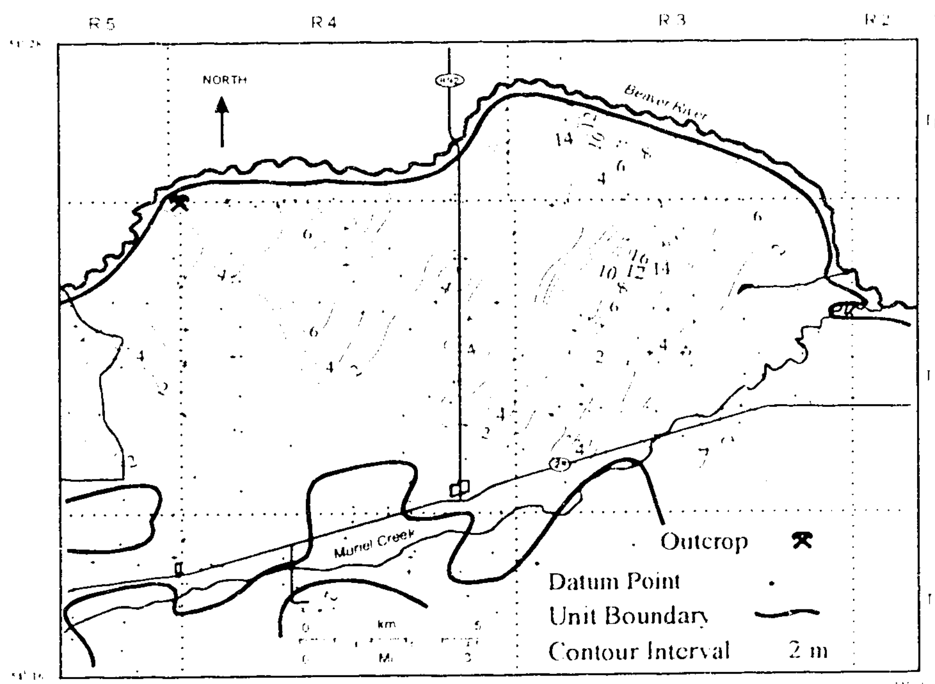


Figure 4.19. Sand River formation isopach.

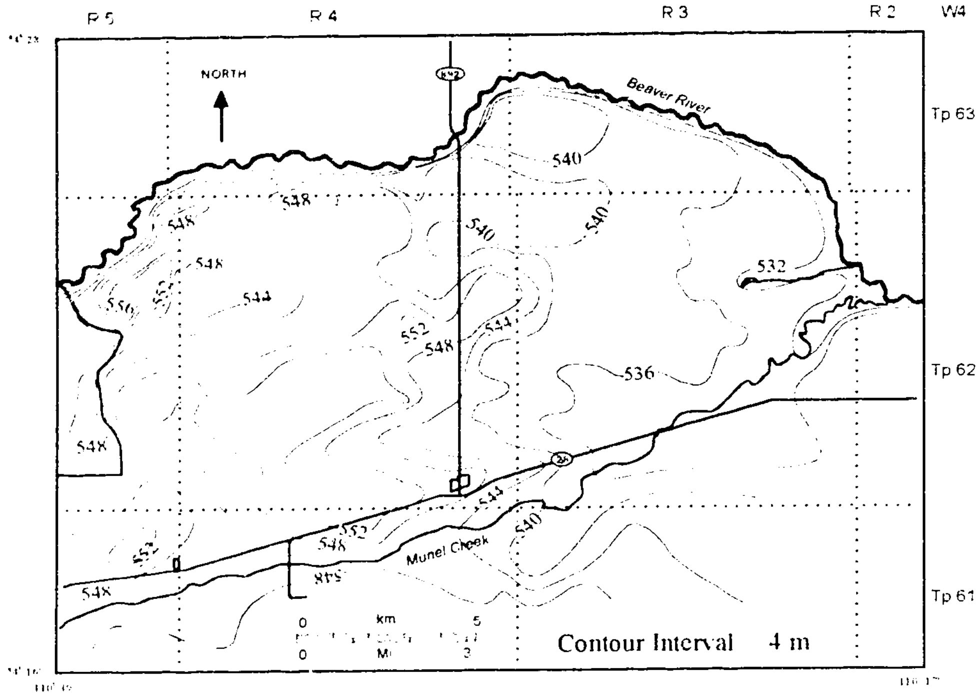


Figure 4.20. Water table elevation.

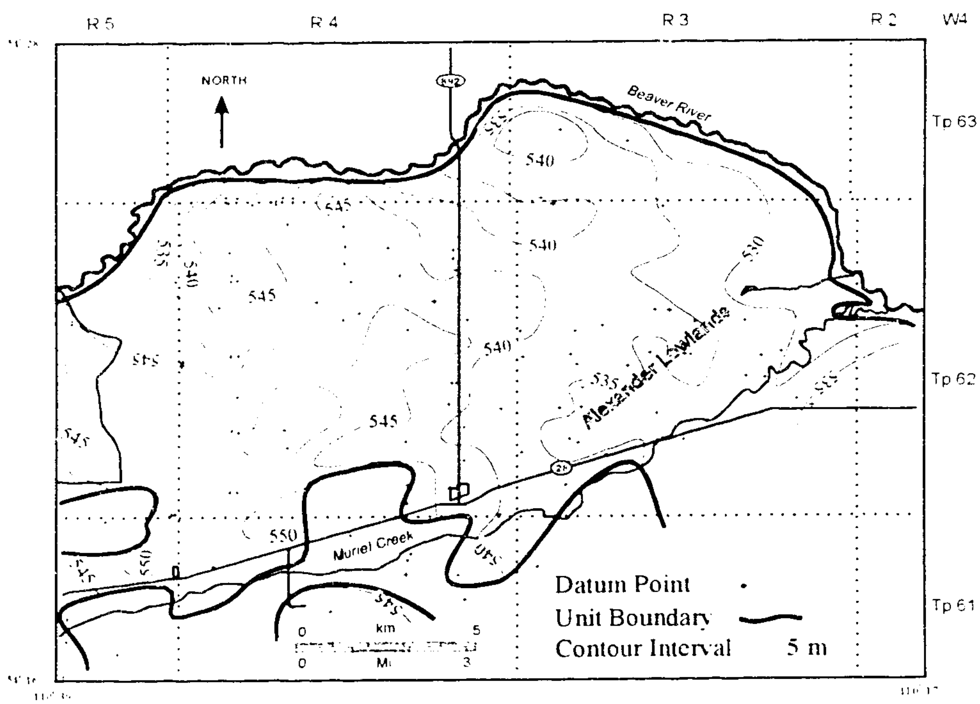


Figure 4.21. Sand River formation water level elevation.

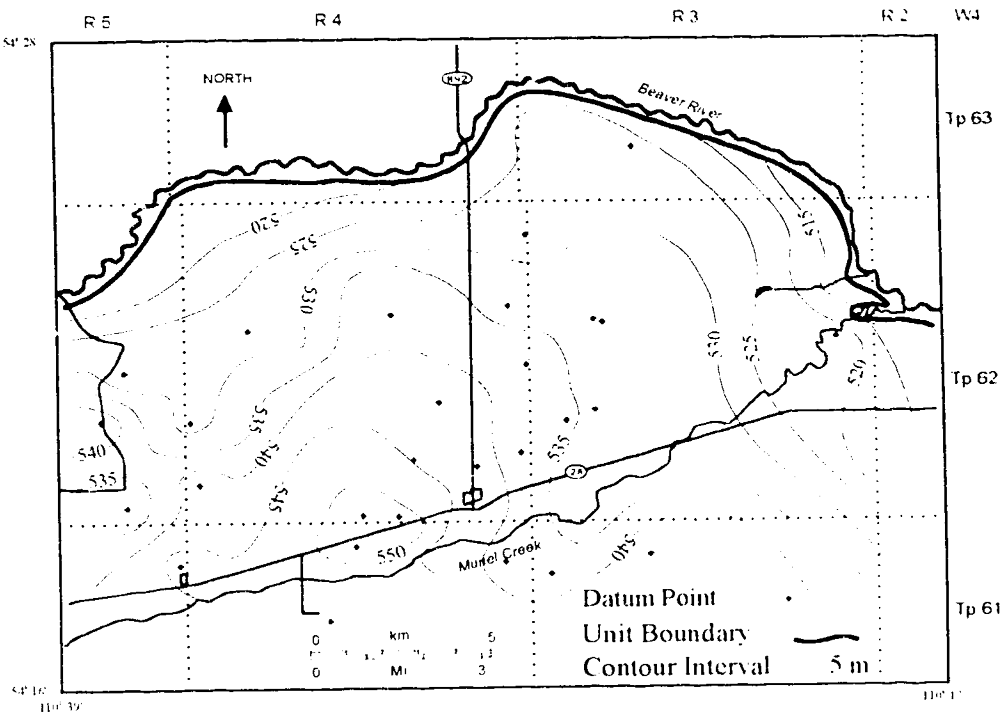


Figure 4.22. Ethel Lake formation water level elevation.

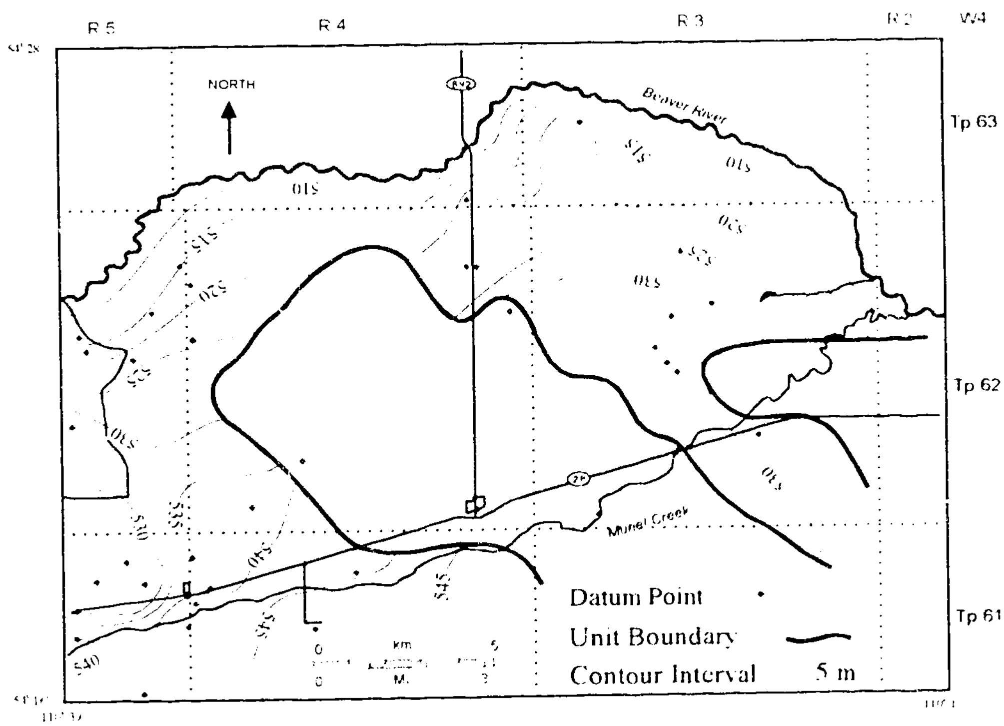


Figure 4.23. Muriel Lake formation water level elevation.

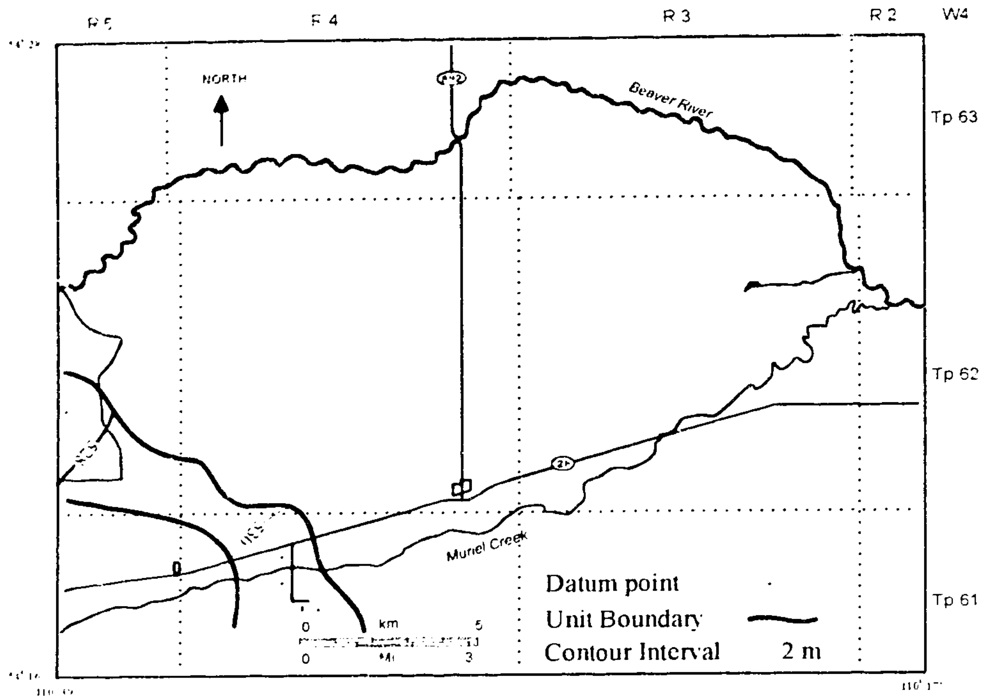


Figure 4.24 Empress formation water level elevation.

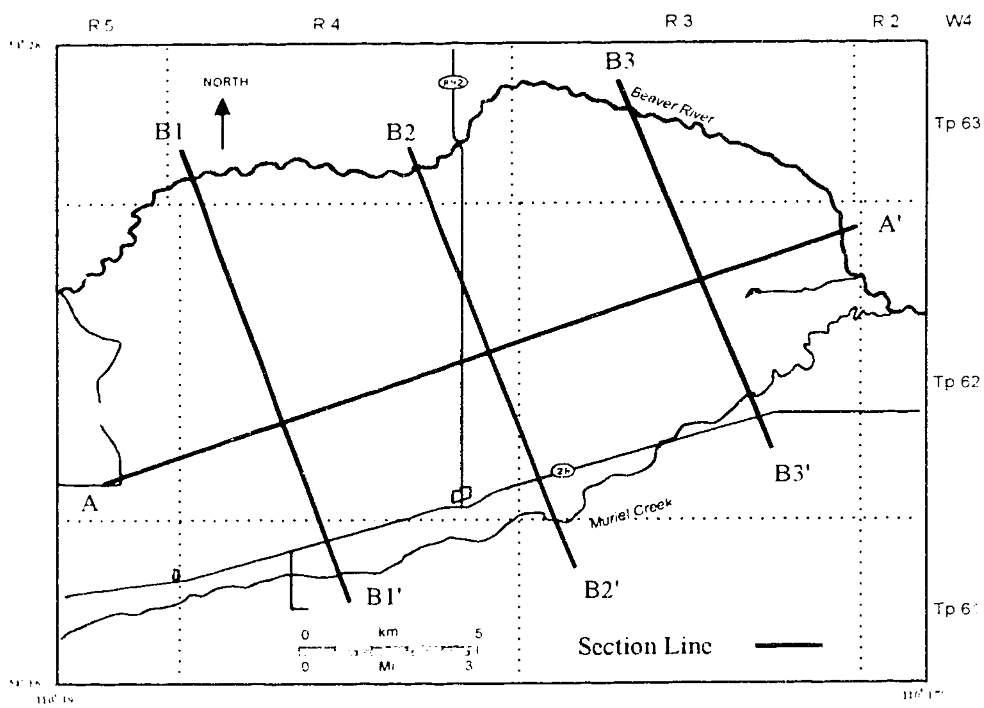


Figure 4.25. Location of cross-sections.

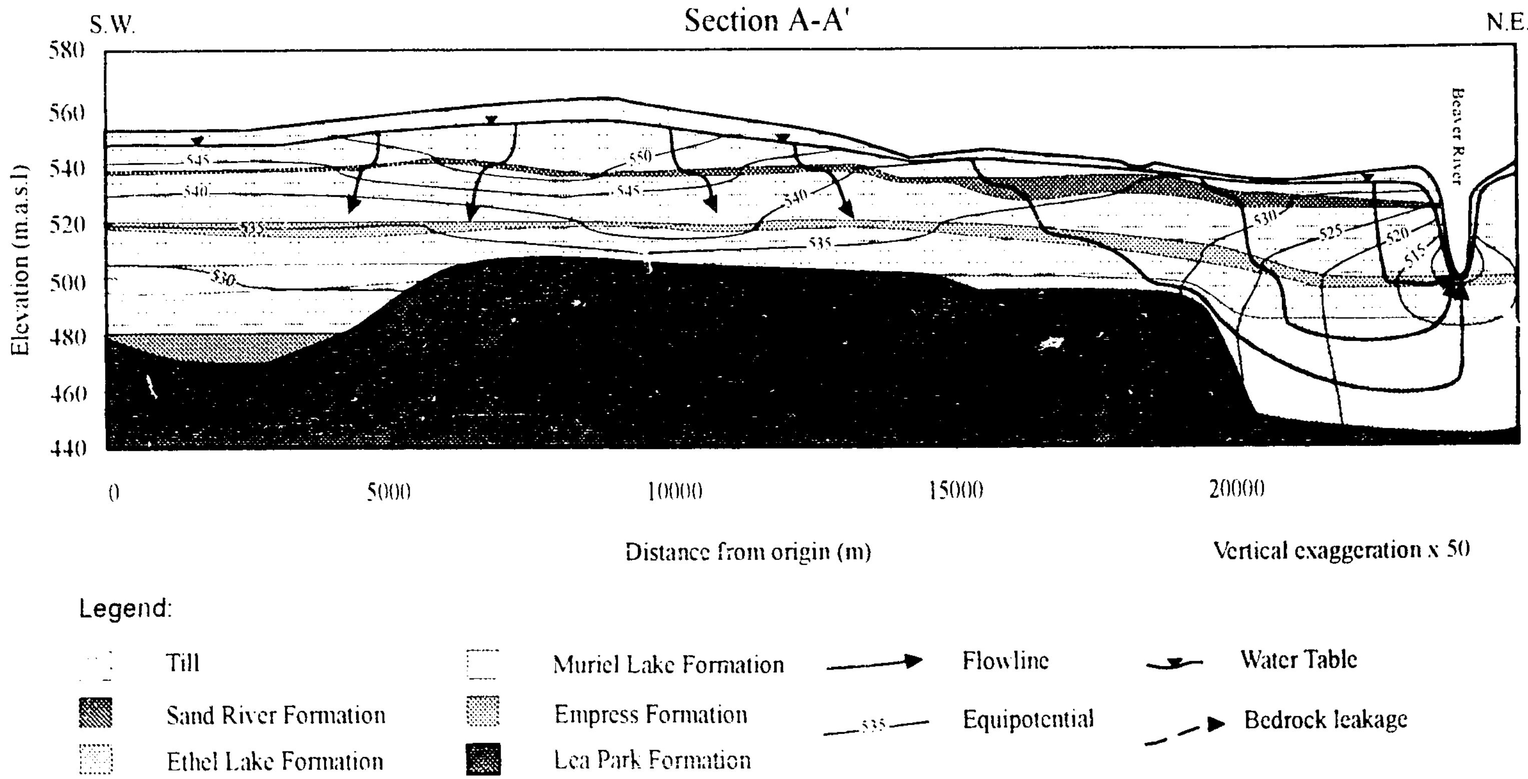


Figure 4.26. Vertical cross-section on line A-A'. Schematic projections of flow lines are sketched in to illustrate the dominant flow directions.

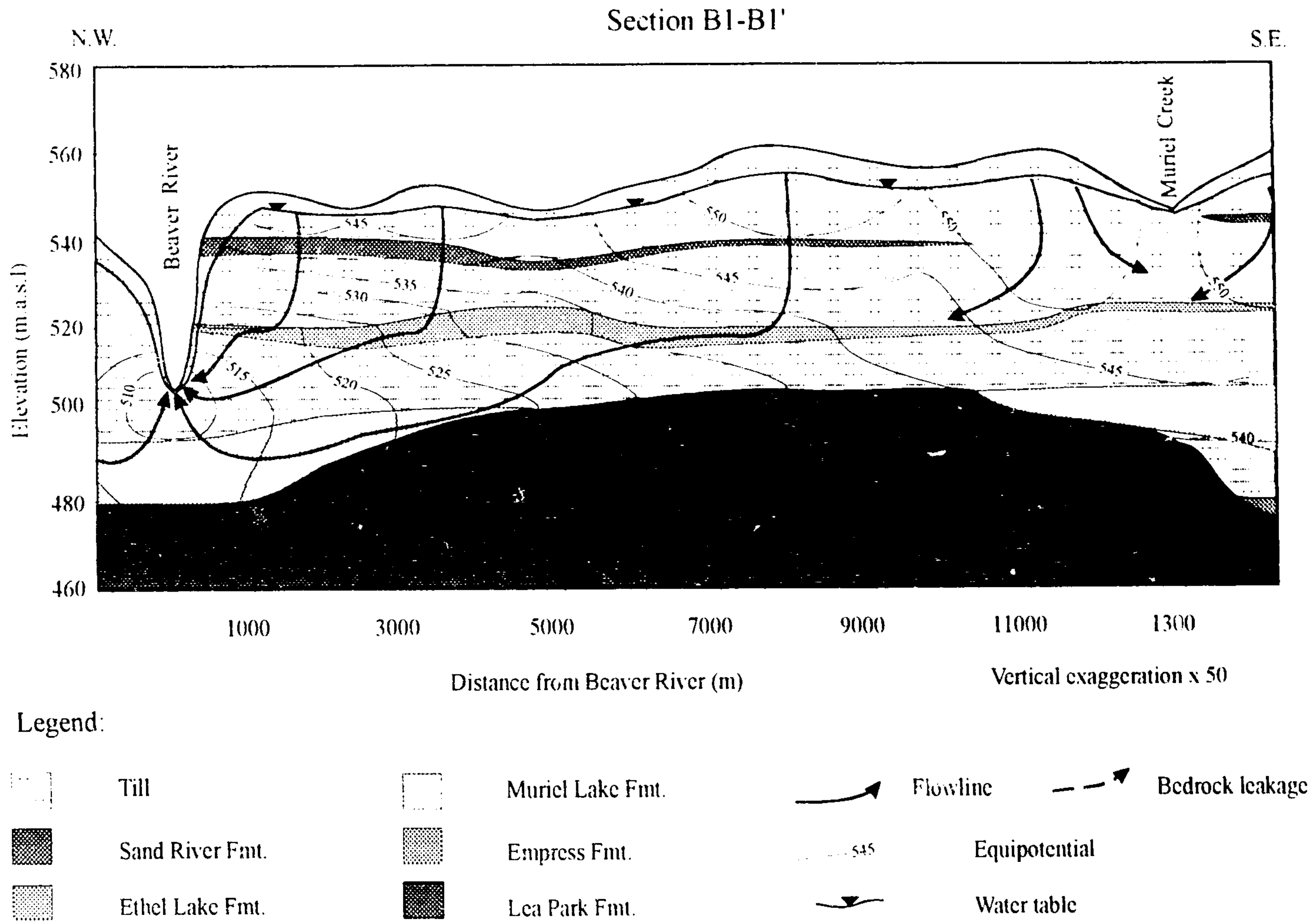


Figure 4.27. Vertical cross-section on line B1-B1'. Schematic projections of flow lines are sketched in to illustrate the dominant flow directions.

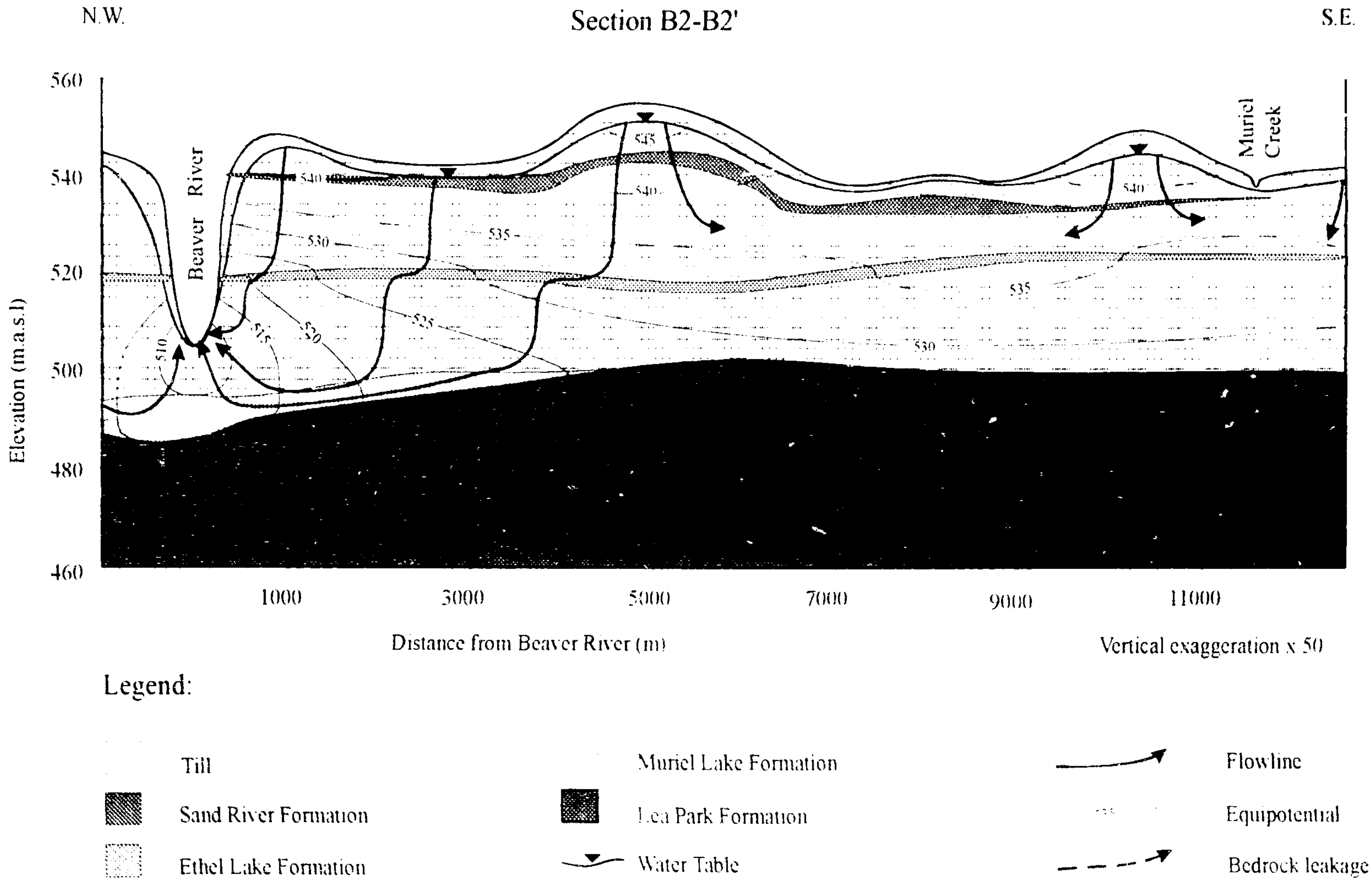


Figure 4.28. Vertical cross-section on line B2-B2'. Schematic projections of flow lines are sketched in to illustrate the dominant flow directions.

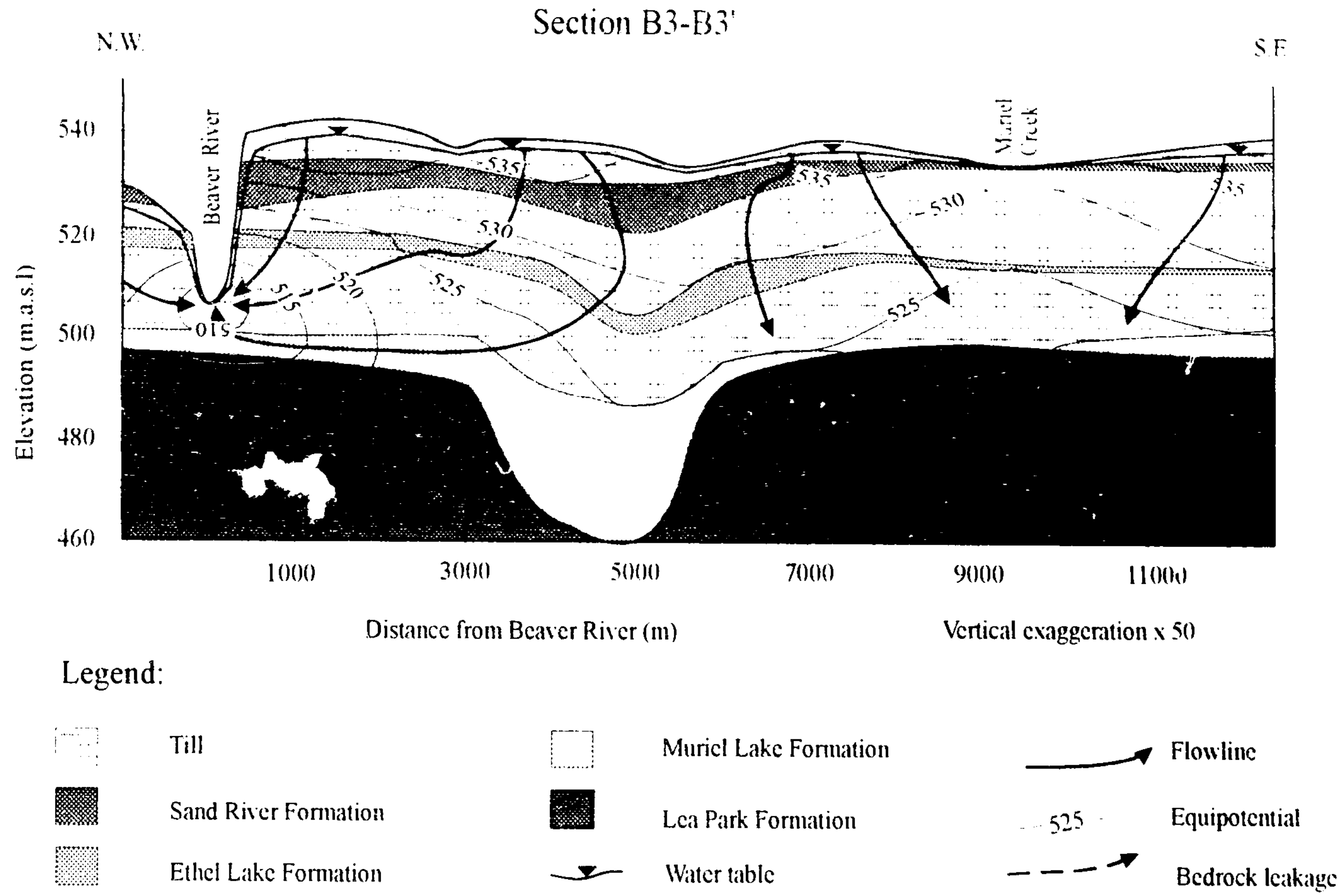


Figure 4.29. Vertical cross-section on line B3-B3'. Schematic projections of flow lines are sketched in to illustrate the dominant flow directions.

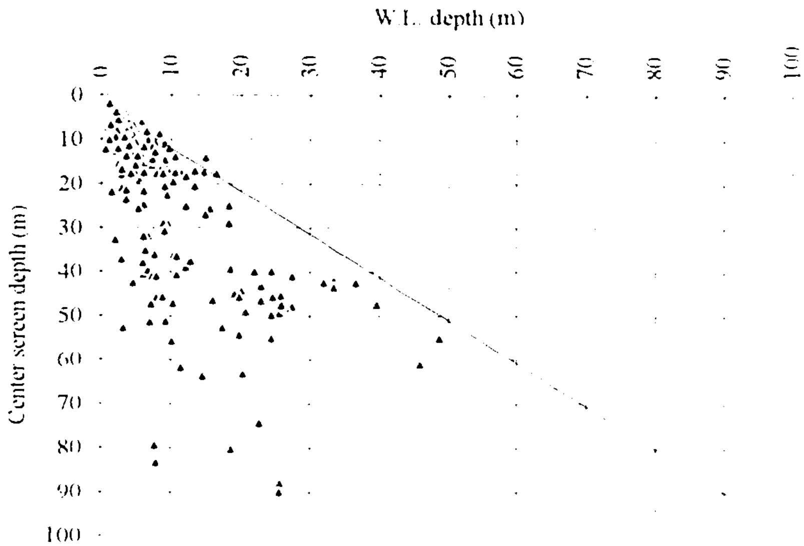


Figure 4.30. W.L. and centre of screen depths plotted together for all holes in the Ardmore area.

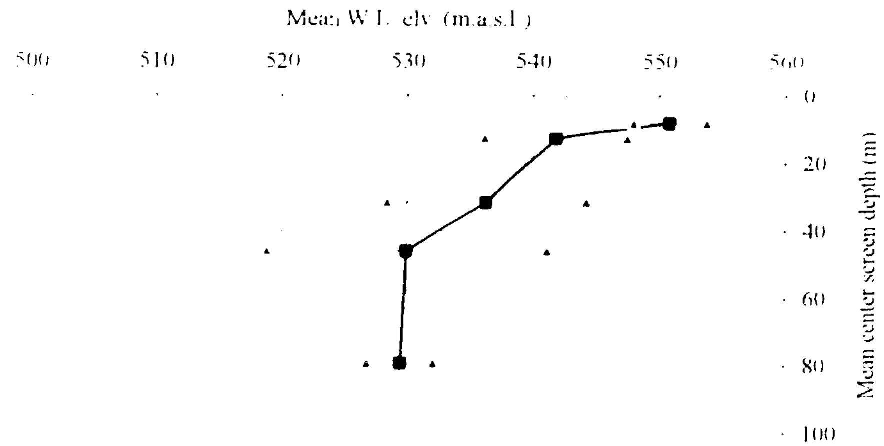


Figure 4.31a. Water level elevation and centre screen depths.

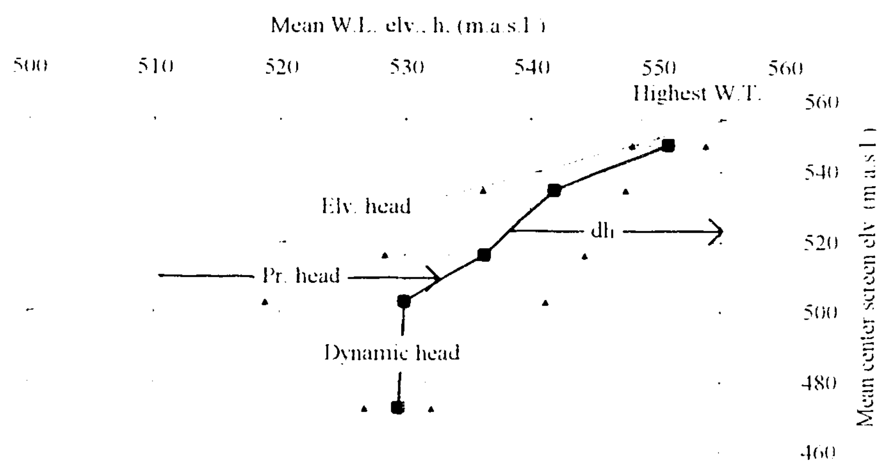


Figure 4.31b. Water level and centre screen elevations.
Triangles show one standard deviation from the mean (squares). The lower plot can be used quantitatively as shown; dh is the dynamic head increment.

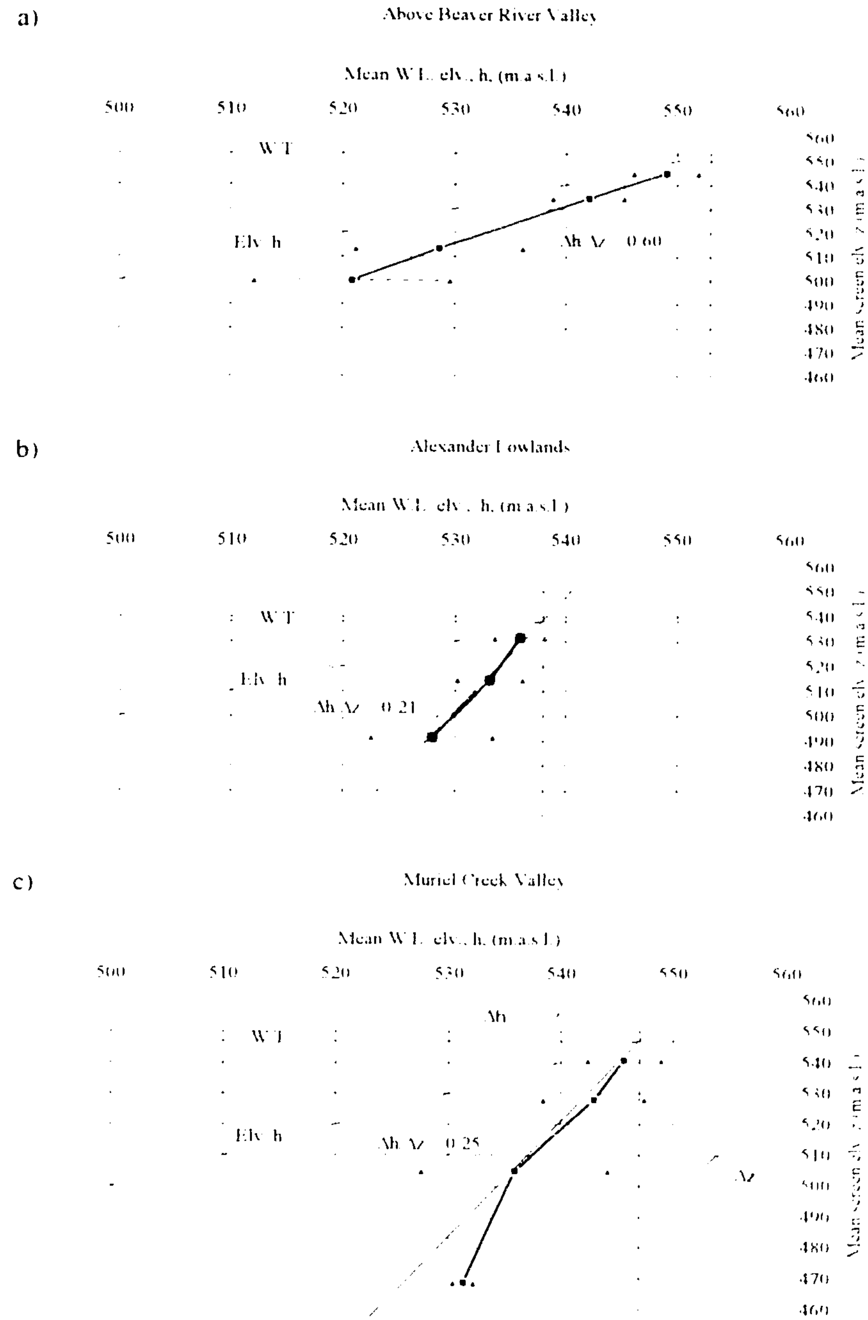


Figure 4.32a, b, and c. Water level and centre screen elevations in sub-areas. Estimates of hydraulic head gradients are shown. Largest gradients occur close to the river valley.

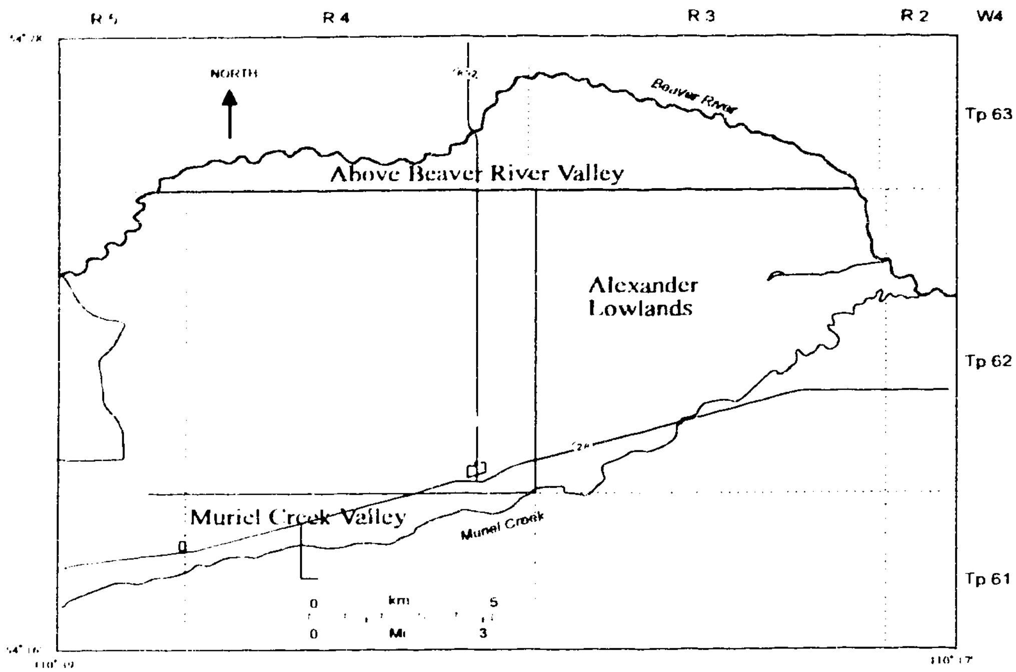


Figure 4.32d. Map of sub-areas.

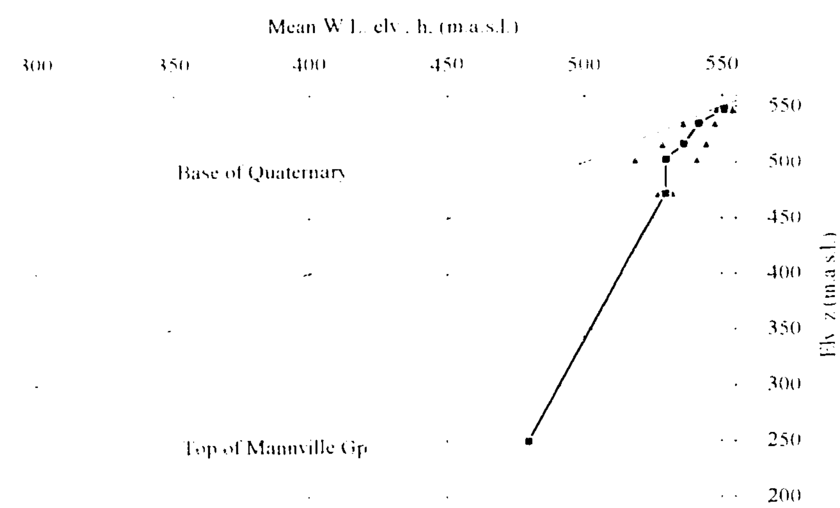


Figure 4.33. Water level and formation elevations.

4.2.4. Soils and mapping

4.2.4.1. Mapping

It was possible to distinguish by mapping two basic types of depressions in the Ardmore area: those which had clearly recharging conditions and those which were suspected of having at least some discharge component ((S) Fig. 4.34). The second type had many features in common with the slow recharge sloughs of Leskiw (1971). This in itself was a significant result indicating that any discharge within the area was most likely to be from local flow systems and was perhaps also intermittent. The features we defined for each depression type are listed below.

Depressions where recharging groundwater conditions are assumed to dominate have the following characteristics:
i) An upland setting in which they are quite variable in shape, sometimes long and sinuous.
ii) They do not contain open water except in the spring and early summer. They will support a person approaching on foot.
iii) The vegetation is only poorly zoned, and the whole area is usually covered by patchy willow and alder shrubs. A continuous ring of willows is always present around the outer edge. The ground cover is a mixture of sedges and grasses with many broadleaf species.
iv) The organic layer is only a few centimeters thick, and the soil parent material is usually a sandy diamicton. The soils at these sites are orthic luvic and fera gleysols.
v) Anthills are commonly present. The hills are built several decimeters above the ground surface and are usually composed of sandy material.
vi) They are quite readily reclaimed, at least temporarily, for hay and grazing.

Depressions with suspected groundwater discharge at some time in the year, had the following features:
i) A lowland setting, commonly forming part of complex depressions (those with more than one core area containing ponded water) which are frequently long and sinuous,
ii) Usually a core zone of open water until late summer or throughout the year and cannot be approached on foot beyond the innermost ring of vegetation.
iii) The vegetation grows in clearly defined zones around the open water. The zones usually occur in the following order, an inner cattail zone, an intermediate sedge zone and an outer willow ring, which may be followed by balsam poplar depending on the amount of clearing. No

halophytic species were found.
iv) There is always an organic layer below the core zone which is usually greater than 0.5 m in thickness. Soils beneath the depression are orthic to humic gleysols and organic where the organic layers are thick enough. The parent soil material is variable, but sandy deposits dominate in the long sinuous type. Very thin encrustations of evaporitic salts are sometimes observed especially where livestock has turned-up clods of earth. The organic layers contain numerous bivalve and gastropod shells.
v) They support a great variety of water-fowl and other birds. Beaver lodges and pond bottom tracks are common.
vi) The land is not readily reclaimable and persists within areas of cultivation.

All the intermittent lakes in the Ardmore study area were dry by the end of 1992; and whereas some contained ponded water from spring runoff early in 1993, they were mostly dry through the remainder of the study period. Patches of standing water did appear in Krill Lake, adjacent to UOH 16, following summer rain in late June, 1993 (Fig. 4.34). The surficial deposits surrounding the lake are sand and the ponding did not result from runoff. It was clear from daily observations that, as the water level in UOH 16 rose in early July 1993, the areas of open water in the lake spread. These events show that the lake is sustained by groundwater at least some of the time, and that the recharge relationship between the two is probably quite complex. UOH 16 is included in the slow depression group (Sec. 4.1).

In addition to the depressions, a number of springs and seeps were located around the north and east boundaries of the study area (Fig. 4.34). From their elevation they were interpreted to indicate discharge from the Sand River aquifer along its outcrop edge. The outcrop is everywhere covered with alluvial deposits with the exception of the outcrop above Dery Springs (Sec. 4.2).

4.2.4.2 Soils

The pedogenic processes that form soil layers, which in turn can be identified and described as a soil profile, are governed by three primary factors; i) the climate, ii) the biota, and iii) the parent material of an area. Pedogenesis begins with the weathering of the upper soil layers and continues as a complex process of decomposition and transportation of both weathered and non-weathered products deeper into the ground. One of the major climatic variables governing these processes is precipitation. The rate, total volume and depth of penetration of infiltrated water, originating at the ground surface, influences in varying degrees the depth to which pedogenic processes are active, and thus the type, thickness, and distribution of the soil layers. The depth and the magnitude and direction of the net water fluxes through the soil are in turn strongly influenced by the groundwater flow pattern.

Soil characteristics which reflect the position of the water table, and the net movement of infiltrated water, are the key factors of interest to the hydrogeologist. A useful approach in examining the relations between soils and the water regime, is to describe soils which are linked by the inherited properties of the parent material, but which show differences in acquired properties related to drainage. These groups of hydrologically linked soils are known as catenas.

Typical Soil Catena.

A typical soil catena for the Ardmore area is shown along a line running between UOH Sites 11 and 8. The line starts at the base of the Muriel Creek valley, and continues due north through a midslope position at site 10 to the crest of Ardmore ridge at Site 9 (Fig. 4.35 and Fig. 4.36). The line then continues down the north slope of the ridge through the midslope at Site 8b and ends in the thalweg of Fort Kent lowlands at Site 8.

In general, the lower slope positions have shallower soil profiles and thinner A and B horizons. They show gleying (g) and usually do not have well developed eluviated horizons (e). The soils in the upper slope positions have deeper profiles and thicker A horizons underlain by B horizons with a well developed blocky structure (t). Where the parent material is till, the depth to the carbonate layer is much shallower in the lower than in the upper slope positions. Often the surface horizons will react to dilute solutions of hydrochloric acid at the base of slopes whereas the reactive layer is usually more than 0.5 m deep at the ridge crest.

Unsaturated flow of water provides the mechanism for a dominantly physical process which flushes out the finer particles in the upper layers and deposits them in lower horizons. This leads to the development, in alkaline soils, of the blocky B horizon structure (t) discussed below. The water also dissolves the more soluble compounds and precipitates them in lower horizons in which the net movement of water is no longer downwards. The calcium carbonate of calcite, for example, is less soluble than the calcium sulphate of gypsum and more mobile species involving sodium or magnesium. Soluble compounds are dissolved in the order of their solubility in a solution and are precipitated in the reverse order. Since the tills are fairly rich in carbonate bearing rocks, such as dolomite, the absence of carbonate is taken as indicating prolonged flushing and dissolution by relatively fresh water.

The position of the water table and its associated capillary fringe dictates the thickness of the unsaturated zone and thus to a large extent the depth of soil development. The annual amplitude of the water table fluctuations (the phreatic belt) determines the zone in which reactions can alternate between those which occur in oxidizing and those in reducing conditions. This alternation results in the iron staining and mottling known as gley.

Site 11, in the soil catena outlined above, is indicative of a shallow water table and the presence of gleying indicates water-logged (reducing) conditions through portions of the year. Though the net movement of groundwater is downward at the site, as indicated by the winter recession of UOH 11 hydrograph, frequent intermittent contact with the capillary fringe is enough to replenish carbonates in the upper soil profile. The absence of more soluble salts tends to confirm the net downward movement of water.

The profile at Site 8 on the other hand shows a more complicated relationship. The presence of a weakly developed solonetzic B horizon (nj) suggests that the site was at one time an area of groundwater discharge. Conditions appear to have reverted to net downward flow, however, and the horizon enriched in soluble salts (s) is now much deeper: below 1.5 m. Carbonates (k) have either not yet been flushed from the profile or are replenished frequently by a fluctuating water table.

Soil profiles from the midslope and ridge crest positions are typical of those with net downward movement of water. They almost always contain an eluviated horizon from which the more mobile clays have been removed. The B horizon below is enriched in clay through a largely mechanical process called lessivage. Percolating water carrying suspended clays is absorbed towards the dry interior of soil peds and deposits the clays in a filter cake on the outside. The clay forms coatings, in this manner, on surfaces and fractures to define the blocky structure characteristic of the luvisolic B horizon. Thus, even though the clay content in the B horizon is enriched, the resulting soil structure is suspected to be relatively permeable to water.

Carbonates have been completely flushed from the upper soil horizons in the mid and upper slope positions. They are normally found precipitated in the C horizon as coatings on the sides of root holes and cracks. The presence of the blocky B horizon structure combined with a dense network of open root holes to below the base of the soil pits (≈ 2 m), obviously served to increase the bulk permeability of these near surface materials.

Soil sequence in the Ardmore Lowland:

This sequence is selected as being representative of the broad lowlands on the east side of the study area. The sequence is illustrated schematically in Figure 4.37. It begins at Site 14 on lacustrine, clay-rich deposits and continues northwards to Site 12 where the clays are thinning above fluvial sands. Site 5b in the sequence represents the height of land and is the only soil without gleying in the profile. From Site 5b there is a very gentle slope northwards to a broad linear depression at Site 5. Since the last three sites are all located on sandy parent materials, the sequence in this part is a catena where the profiles are reflecting slight changes in the drainage.

Orthic Gleysols are the most common soil type across the lowland, but they tend towards the Luvisolic subgroup. This means that they often contain an incipient eluviated horizon which is obviously better developed on clay rich parent materials and hard to distinguish in sandy soils. The presence of eluviation and of B horizons usually more than 0.2 m thick suggests a net downward movement of water through the profiles. The gleying indicates seasonal water logging which may have occurred more commonly in the past especially in the early season following the spring thaw. The near surface is probably drained more efficiently by road ditches in modern times although some standing water was seen in open fields in the spring. The gleying could thus be the product of perched saturated conditions for short periods. Gleying in the sandy soils, outside obvious depressions where the occurrence of perching is less likely, indicates that the water table rises close to the land surface fairly frequently.

The Lowlands contain widespread deposits of fluvial and lacustrine sands (Fig. 4.38). Brunisols tend to develop in the upper slope positions on these soils. They tend to have rather thin B horizons, but this is not thought to indicate a reduction in downward flow. In fact the underlying C horizons are streaked with vertical mottles to a depth usually greater than 1 m. Below this, there is often an iron rich horizon similar to a fera layer. The fera layer appears to be the location at which the iron compounds are being concentrated and is most likely the location of the mean position of the top of the capillary fringe. The B horizon is likely to be less well developed in any case in these more acidic sandy soils. The dominant vegetation is a Jack Pine forest with a poorly developed understory. There is, therefore, less production of organic matter and a high rate of decomposition in the leaf-litter.

In summary the soils in the extended lowland areas also indicate net downward movement of water through the soil profiles. This is the expected condition above a dominantly recharging groundwater flow regime.

Soil profiles within and adjacent to depressions:

Two soil sequences are selected to show the degree of variation observed in soils associated with depressions. The first sequence (Fig 4.39) is from the Fort Kent Lowlands and the second from the Alexander Lowlands (Fig 4.40).

The Fort Kent lowland depression has been cleared for hay production, whereas the Alexander depression is a permanent wetland. The Fort Kent depression contained ponded water only in the early spring following snow-melt. The Alexander depression which normally contains open water, dried up in September 1991 and no ponded water was observed after that time.

The Alexander depression is thought to be typical of many slow recharge depressions with core zones containing semi-permanent ponded water. The slough had apparently not been completely dry since the 1930s (pers. comm. Mr. Butler, landowner). The water table had dropped to approximately 1 m below ground level in the slough by the spring of 1991 and is estimated to have risen to within 0.5 m of the surface following the spring rise in the same year (UOH 15). The water level then fell rapidly through the summer 1992 such that there was almost no net accretion (see Sec. 5.1) to the groundwater by the spring of 1993.

Soil pit 15b was dug in the core zone of the slough normally occupied by water. The organic layers extended to 0.64 m below ground. The organic material was not very reactive to dilute HCl which indicated a low carbonate content, but the horizons contained numerous freshwater gastropod and bivalve shells which of course reacted strongly. The mineral layers became very strongly reactive and appeared to be enriched in carbonate below 1.3 m. The water table recovered to 1.16 m below ground during the excavation. The mineral layers were gleyed without mottling which indicates permanently saturated and reducing conditions.

Pit 15 was dug in the willow ring adjacent to UOH 15. Thin organic layers were developed above an eluviated A horizon. A well developed B horizon was present in the profile which was moderately rich in carbonates.

The soil at Pit 15c, dug outside the willow ring on land cleared of poplars, also showed eluviation but carbonates were present through most of the profile.

Several differences were observed between the soils of the Fort Kent and Alexander depressions. Site 101 was dug in the center of the Fort Kent depression. The organic horizon was relatively thin at 0.17 m and was underlain by thick A and B horizons. The shoveling arm ran out of energy in heavy clay at 1.18 m without having reached layers showing any reaction to HCl, so that the profile appeared to be completely flushed of carbonates. The iron staining and mottling, especially in the lower half of the B horizon, indicated alternating saturated and unsaturated conditions and suggested a large amplitude of water level fluctuations.

The second profile, 102, in the Fort Kent soil sequence was measured in a trench dug just outside the former position of a willow ring. There was a complete absence of gleying in this soil. A well developed eluviated A horizon was found above a well-structured B horizon. The carbonate layer was located just below the B Horizon at 0.76 m.

Site 103 is located approximately 500 m north and upslope from the depression at site 101. The slope is very gentle with a gradient estimated at less than 0.003 m/m. The soil profile is very similar to that at the edge of the depression but is developed to a slightly greater depth. The calcium carbonate horizon at this site was found at 0.9 m.

The depressions at Fort Kent and Alexander Lowlands were interpreted as fast and slow recharge sloughs respectively of the freshwater type (Lissey 1968). In the case of the slow recharge slough, the water table does not usually recede far below the base of the slough, and the general level of the water table is normally close to the land surface all around it. In the fast recharge slough the water table appears to rise and fall quite rapidly and to recede well below the base of the slough following periods of low to average rainfall and high transpiration. The Fort Kent depression is probably the site of focused recharge from a dissipating water table mound, of the type first outlined by Meyboom, (1966). The

Alexander site on the other hand is thought to represent very inefficient groundwater recharge with a high percentage of accreted water lost to evapotranspiration and a much smaller percentage remaining permanently in the saturated zone (see hydrograph UOH 15 Fig. 4.6).

Summary

The soil profiles greatly increased our confidence that the general conditions of downward flow were widespread even in the near surface layers. One of our concerns, that the water table monitoring was not sensitive enough to show the details of shallow local flow systems, appeared to be unfounded. Local systems might have resulted in a large amount of groundwater discharge that we did not take into account in the flow system approach to recharge estimates. The soils, however, indicated that even the sloughs with semi-permanent ponding also had dominantly recharging flow conditions and were not locations of groundwater discharge in the sense of upward flow.

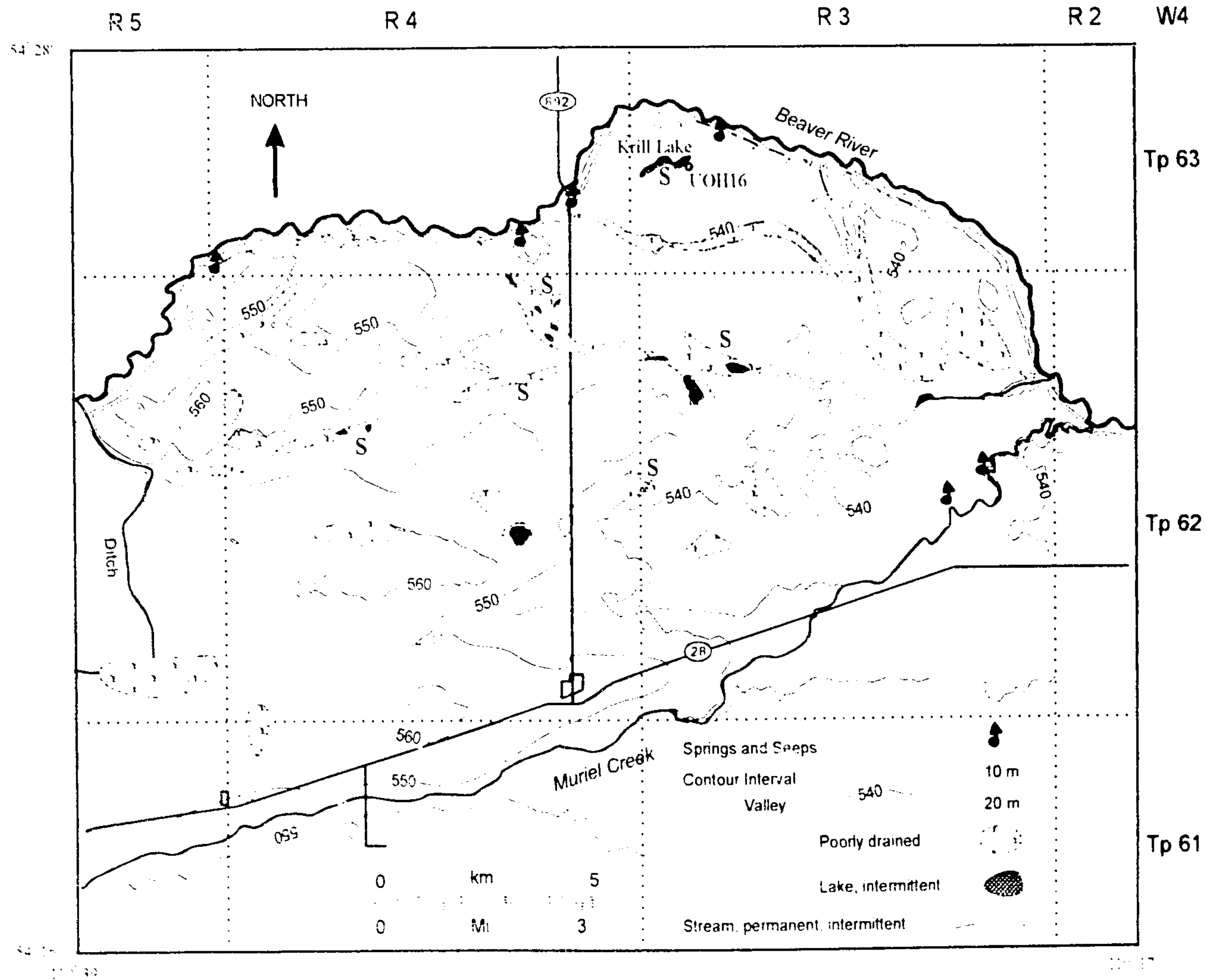


Figure 4.34. Mapping of wetlands, springs and seeps. S wetlands are slow recharge types.

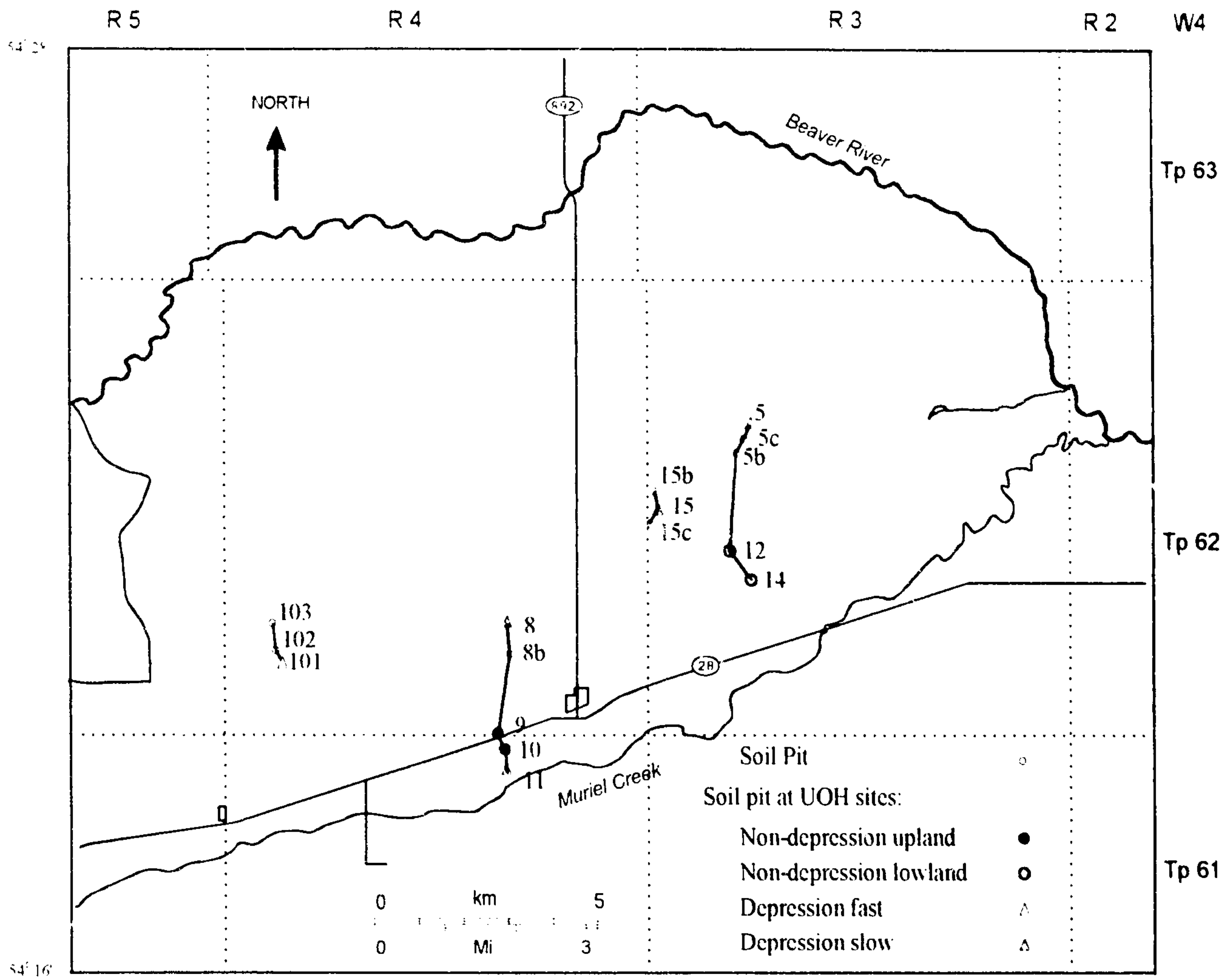


Figure 4.35. Location of selected soil profiles and sequences.
Note: no UOH at site 101.

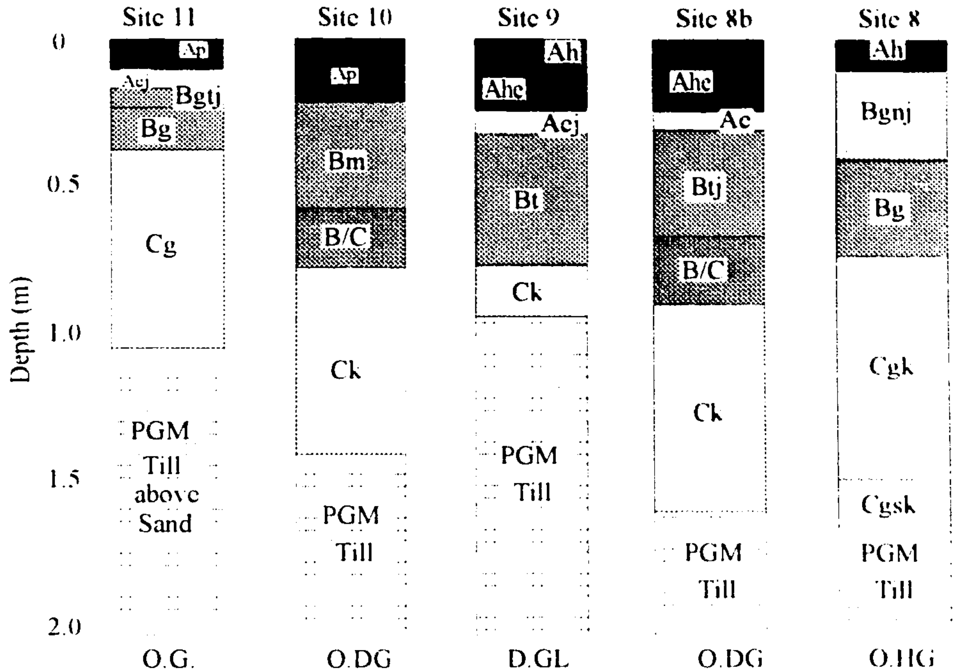


Figure 4.36. A typical soil catena. O.DG. Orthic Dark Gray. D.GL. Dark Gray Luvisol, O.HG. Orthic Humic Gleysol (see Fig. 4.35 for locations).

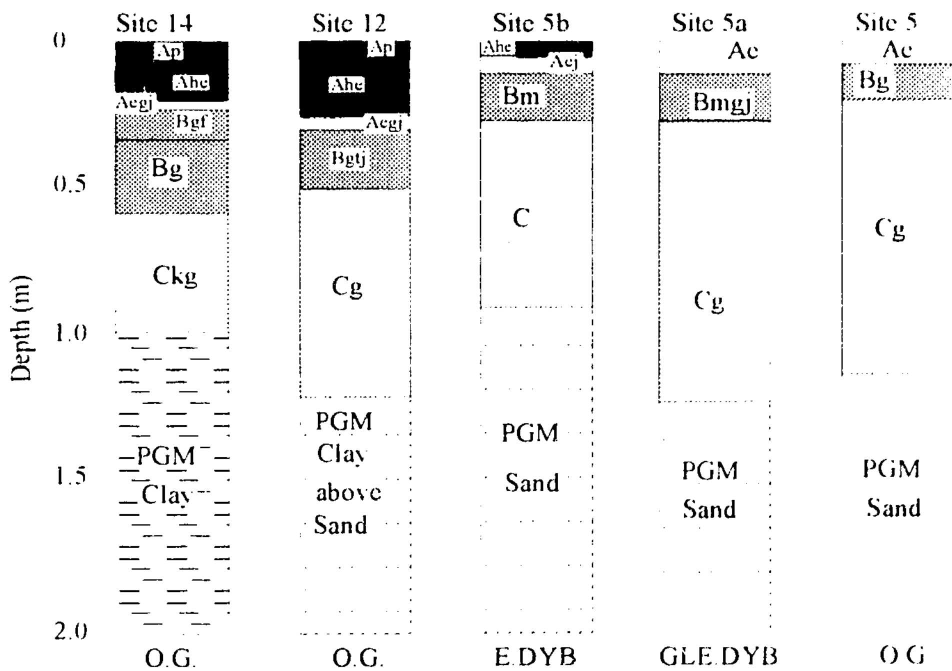


Figure 4.37. A lowland soil sequence. O.G. Orthic Gleysol, E.DYB. Eluviated Dystric Brunisol, GLE.DYB, Gleyed Eluviated Dystric Brunisol.

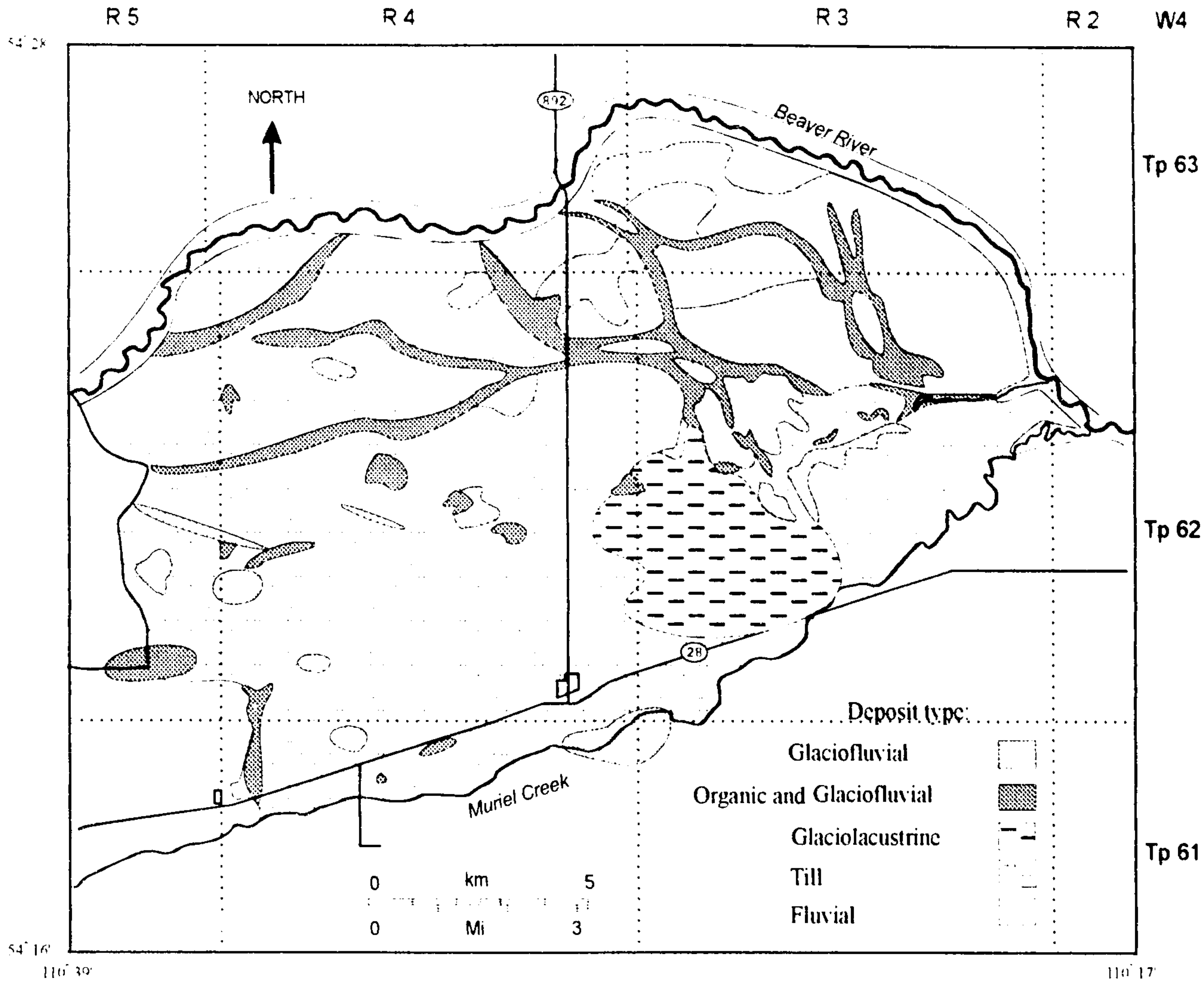


Figure 4.38. Surficial geology.

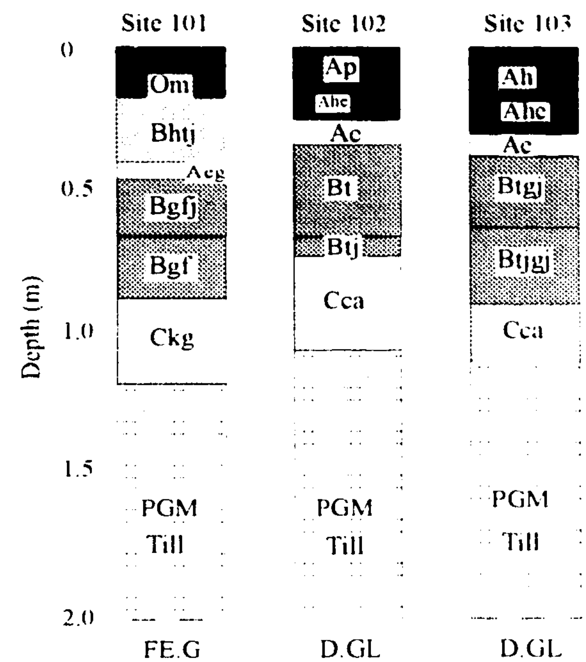


Figure 4.39. Dry depression soil sequence. FE.G. Fera Gleysol, D.GL. Dark Gray Luvisol (see Fig. 4.35 for locations).

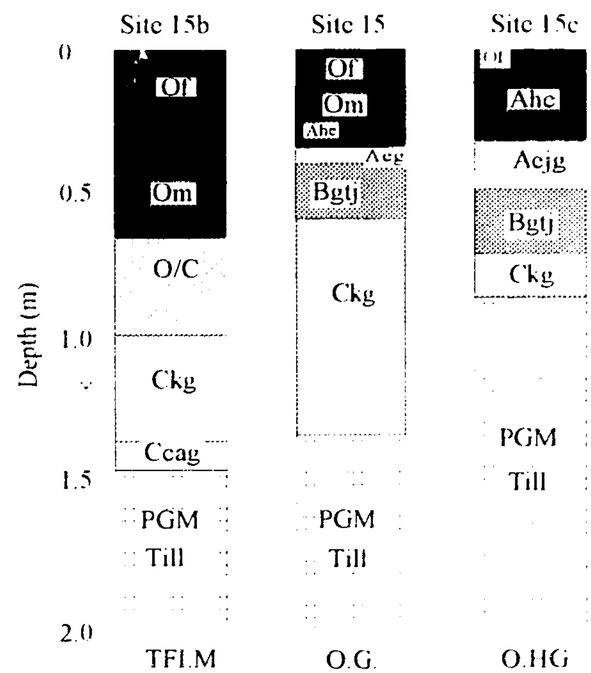


Figure 4.40. Wet depression soil sequence. TFI.M. Terric Fibric Mesisol. O.G. Orthic Gleysol, O.HG. Orthic Humic Gleysol.

4.3. Pumping tests

In the first part of this section the results of parameter estimates are given using data from the Ardmore Thermal Project (Sec. 3.3). In the second part we examine the implications of the responses to pumping, for the conceptual model of the flow regime. Whereas the pump test data were supplied by Hydrogeological Consultants, the aquifer parameters and interpretations presented below are from this study.

4.3.1. 1974 pump test

The pumping rate during the 1974 tests of both wells was reported to be 606.2 m^3/d. Transmissivity values were calculated, from the drawdown and recovery curves in the pumped wells, using Jacob's (Cooper and Jacob, 1946) straight line method (Fig. 4.41). Storage in the well appeared small because of the rapid early drawdowns. The test suggested in Section 3.3, $t > 25r_c^2/T$, was applied for confirmation that we could consider well storage effects as negligible with the following result:

$t > 25(0.089)/294 = 7.57 \times 10^{-3}$ d $\approx$ 11 min.

Also the effects of leakage can be considered negligible when $t < cS/20$ (Sec. 3.3). The result of this calculation was,

$t < (1.4 \times 10^5)(3.54 \times 10^{-4})/20 = 0.25$ d = 357 min.

There appeared, therefore, to be a large enough period of time in the drawdown data to apply the Jacob method with reasonably accurate results.

A value of transmissivity was also calculated from the drawdown data in OWW1, which was used as an observation hole during pumping in CWW2. In this case the Theis (1935) curve matching method of solution was used (Fig. 4.42). The early drawdown data from OWW1 followed the type curve for approximately the first 370 min. The data curve then rose above the type curve suggesting that the cone of depression had reached a barrier. Since the drawdown became larger than that predicted by the type curve, the barrier was of the negative kind. This meant that the barrier reduced the transmissivity of the aquifer. As the cone of depression spreads away from the pumped well and samples a greater volume of the surrounding rocks, negative barriers can be the result of several factors: facies changes within the aquifer, aquifer thinning, a contact with a less permeable unit or some combination of these.

The estimates of transmissivity from drawdown and recovery curves in the pumped wells were similar suggesting that disturbances, such as turbulence and surging water levels, caused by the pumping were minimal (Appen. A, Table 3). Nevertheless we used the estimates from the recovery curve in CWW1 and the drawdown in OWW1 to determine average aquifer parameters. The average transmissivity value prior to any barrier effects was 3.4×10^{-3} m^2/s. Storativity estimated from the Theis curve fitting method in OWW1 was 3.27×10^{-4}.

4.3.2. 1978 pump test

The pumping rates, during the 1978 test, in CWW1 and 2 were reported as 340.4 and 412.4 m^3/d respectively.

Analysis of the 1978 pump test was a little more complicated. Both CWW1 and 2 were pumped simultaneously and the nearest observation hole, COH5-21, completed in the pumped aquifer was located approximately 1500 m to the south east (Fig. 3.3).

The early portions of the drawdown curve in CWW2 were the most suitable for analysis (Fig. 4.43), although in the records we were able to obtain, the first reading was taken 30 min after the start of pumping. An initial estimate of transmissivity was made using the Jacob method following which, the data were adjusted for interference effects from CWW1 and the transmissivity recalculated. The drawdown curve steepened with time at approximately 200 min, in a similar manner to the 1974 data from OWW1, and again at 2000 min. The transmissivity estimates from the early portions of the curve were much lower than in 1974 at approximately 1×10^{-3} m^2/s. In addition the specific capacity of CWW2 had dropped from 68 $m^3/d/m$ in 1974 to 48 $m^3/d/m$ in 1978 which suggested a loss in well efficiency over the four year period. Due to the uncertainties discussed in Section 3.3 we were not confident of the accuracy of these results.

The early drawdown history in both pumping tests suggested the existence of negative barriers which had the effect of reducing transmissivity. At later times in the 1978 test, however, the drawdown in both pumping wells appeared to stabilize after about 250 h (Fig. 4.43). Drawdowns in the two observation wells completed in the pumped aquifer also stabilized with time indicating that withdrawals from the aquifer were being matched by recharge (Fig. 4.44).

We attempted to use the pumping test data to estimate the vertical hydraulic conductivity in the overlying aquitard, with a view to estimating potential natural fluxes across the unit, using undisturbed hydraulic gradients. Unfortunately, as discussed in Section 3.3, the data were not suitable for this level of analysis.

4.3.3. Support of the conceptual model

The more important result of the pump test for our purposes, was the information it gave regarding hydraulic communication. The conceptual model outlined in Section 4.2 has dominantly descending groundwater flow in an essentially unconfined flow regime. For this to be possible, there must be hydraulic communication between aquifers through the less permeable tills. The analyses of the 1978 pump test help to confirm the model in two fundamental ways. First, the test showed leakage across a till aquitard and second it showed good hydraulic communication through a 50 m thick sequence of tills and outwash deposits over an area of approximately 50 km^2. It was also obvious that the aquifers were not a simple single layered unit of constant thickness. The trend of declining transmissivity with time and distance indicates either a heterogeneous single unit or the sampling of multiple units with some hydraulic continuity between them.

Leakage through the aquitard was demonstrated by the stabilized drawdowns in the pumped aquifer. The main source of recharge appeared to be from the overlying Sand River aquifer from responses to the pumping measured in shallow observation wells, completed above the till aquitard (Fig. 4.45 and 4.46). The same observation wells also responded to precipitation events, as shown in the figures, indicating that the upper aquifer was being recharged in turn from meteoric sources. A distance drawdown graph for the shallow observation wells was constructed to show the reduced response to pumping with distance from the pumped wells and to estimate the area of influence (Fig. 4.47). The radius of the area estimated from the graph is on the order of 5000 m which corresponds to an area of 78.5 km^2.

The early drawdown data are characteristic of outwash aquifers containing both rapid changes in thickness and in composition. Spontaineous potential and resistance logs from a nearby hole (TH132, Fig. 4.48) suggested that the Ethel Lake and the underlying Muriel Lake aquifers might be in communication with each other. Also the variation in the completion elevations between CWW1 at 513 m.a.s.l. and COH5-21 at 500 m.a.s.l. suggested that the Ethel Lake aquifer might represent a zone with interconnecting sand bodies rather than a single homogeneous unit.

Figure 4.48 shows a series of lithological and electrical logs along a section line drawn from S.E. to N.W. through the area.

Hydraulic communication is also demonstrated over extended periods during normal pumping schedules. Figure 4.49 shows the hydrographs from the two observation holes and CWW1 used in the 1978 test for a period between 1982 and 1989. Also shown is a fourth observation hole, COH2-33a, completed below the pumped aquifer and located approximately 3.5 km to the north east. The pattern of water level fluctuations in the pumped well is matched by similar patterns in the observation holes throughout the period.

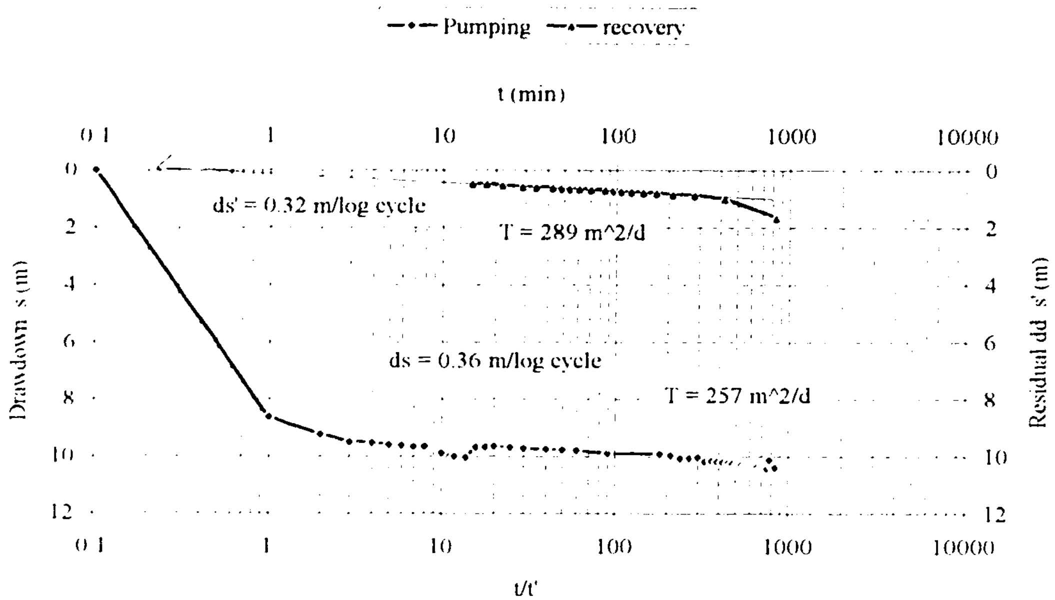


Figure 4.41a. Drawdown and Recovery in CWW1, 1974 pumping test. (Data courtesy of Hydrogeological Consultants).

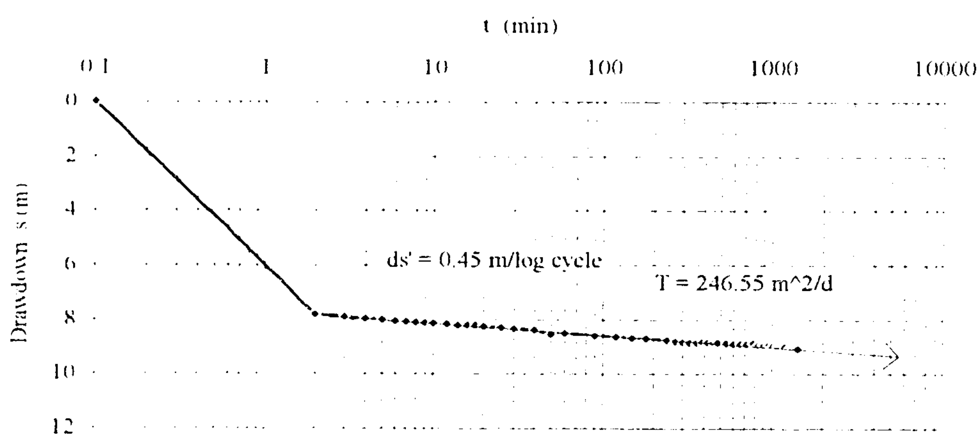


Figure 4.41b. Drawdown in CWW2, 1974 pumping test. (Data courtesy of Hydrogeological Consultants).

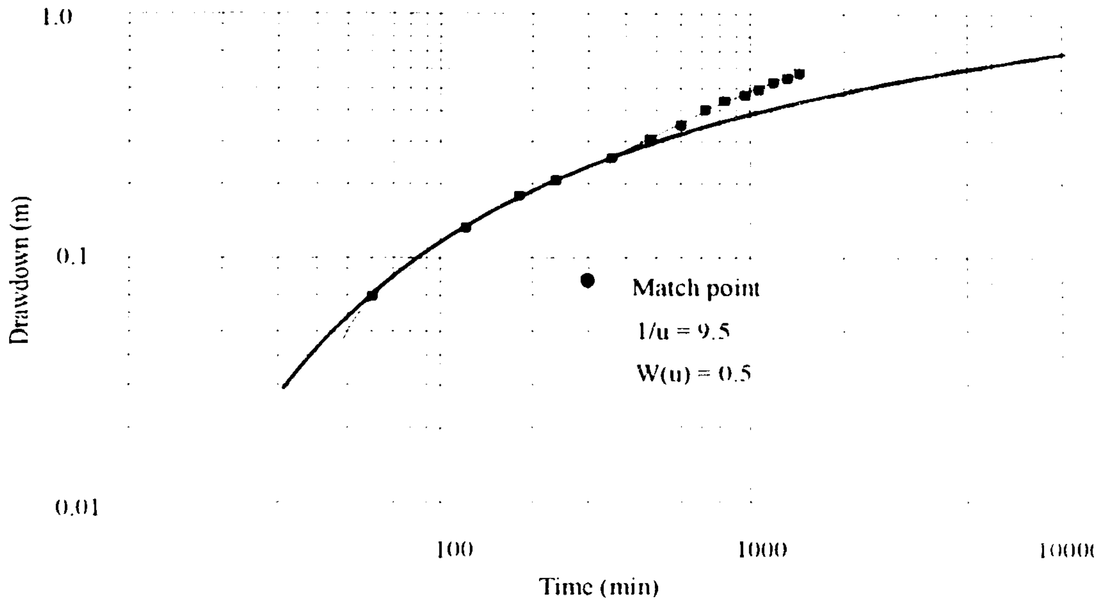


Figure 4.42. Drawdown in OWW1, 1974 pumping test of CWW2. Shows match point information for early times and departure from Theis type curve. (Data courtesy of Hydrogeological Consultants).

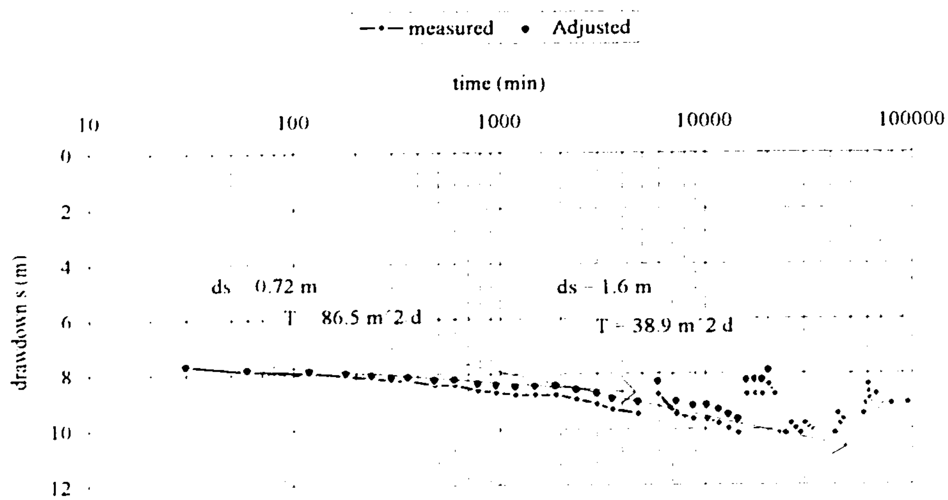


Figure 4.43. Drawdown in CWW2, 1978 pumping test (time; log scale). (Data courtesy of Hydrogeological Consultants).

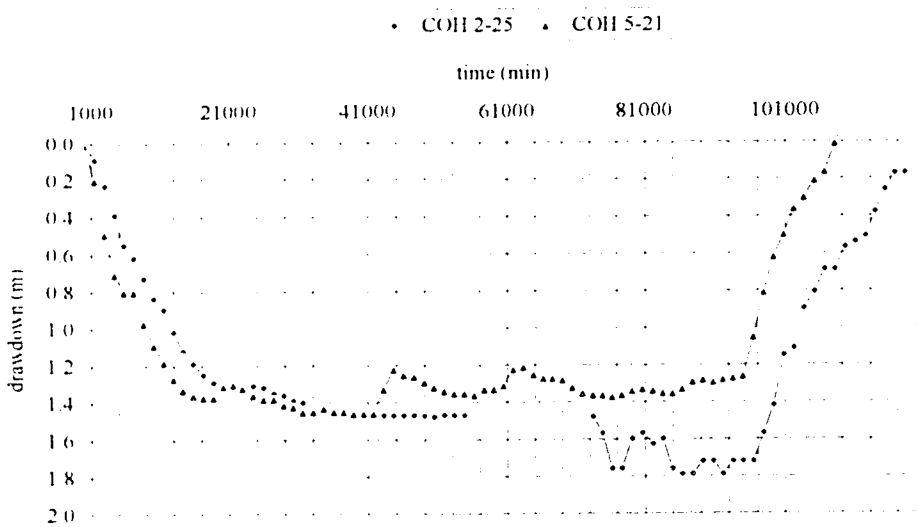


Figure 4.44. Drawdown in COH2-25 and 5-21, 1978 pumping test (time; linear scale). (Data courtesy of Hydrogeological Consultants).

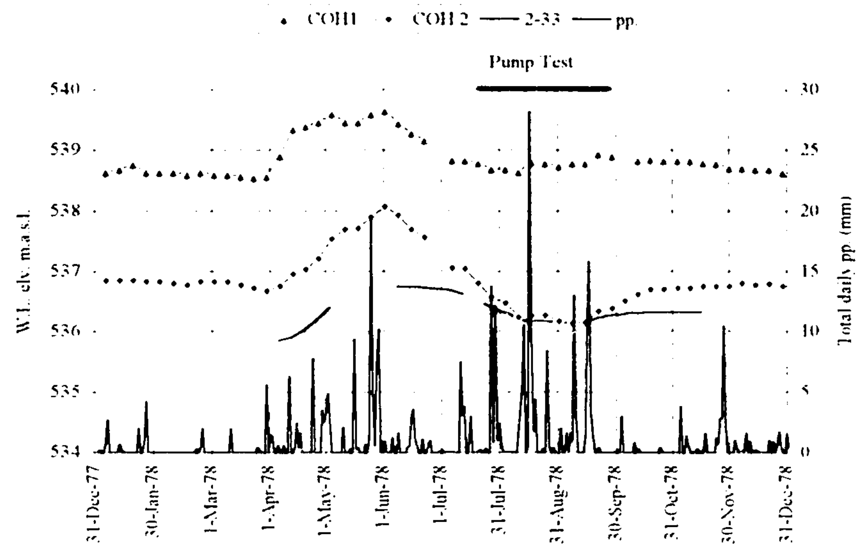


Figure 4.45. Hydrographs from commercial observation holes (COH) in Sand River aquifer. (Data courtesy of Hydrogeological Consultants).

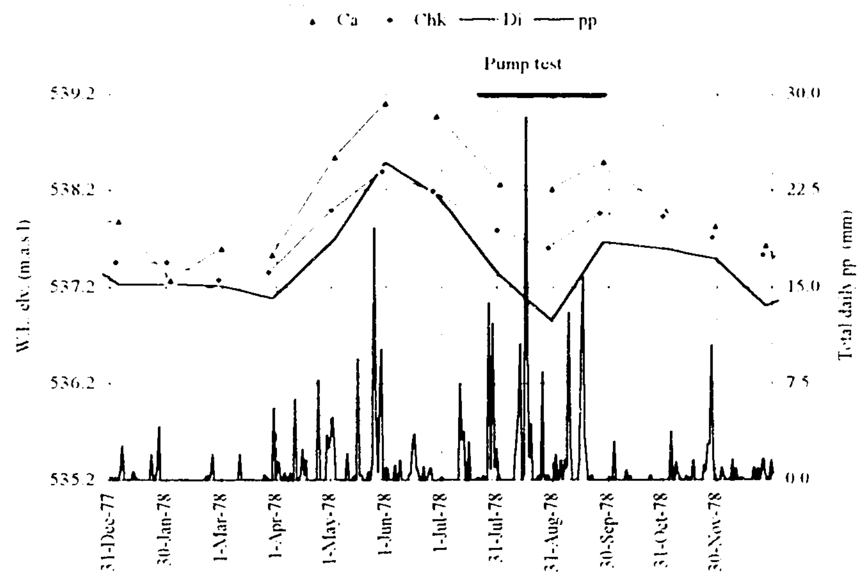


Figure 4.46. Hydrographs from private wells completed in the Sand River aquifer. (Data courtesy of Hydrogeological Consultants).

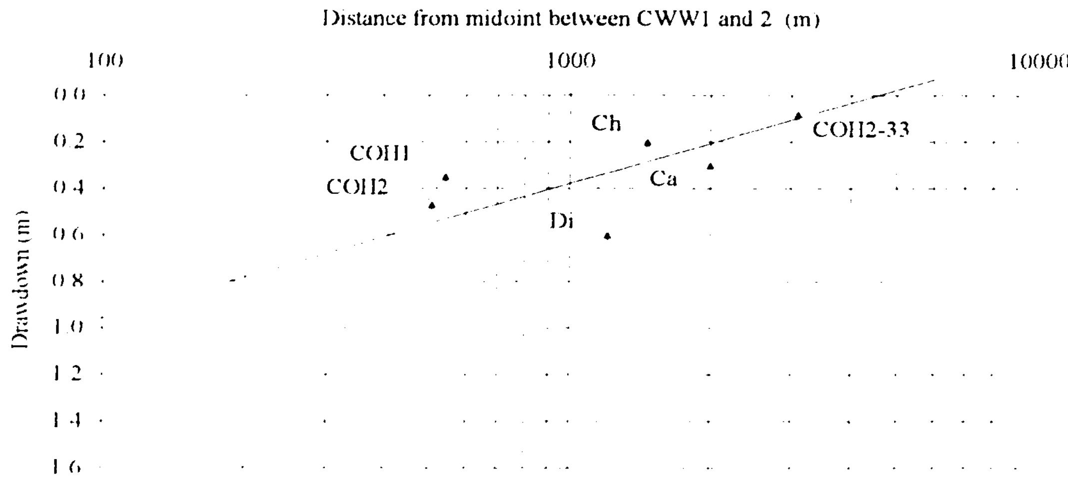


Figure 4.47. Drawdown in the Sand River aquifer, 1978 pumping test. (Data courtesy of Hydrogeological Consultants).

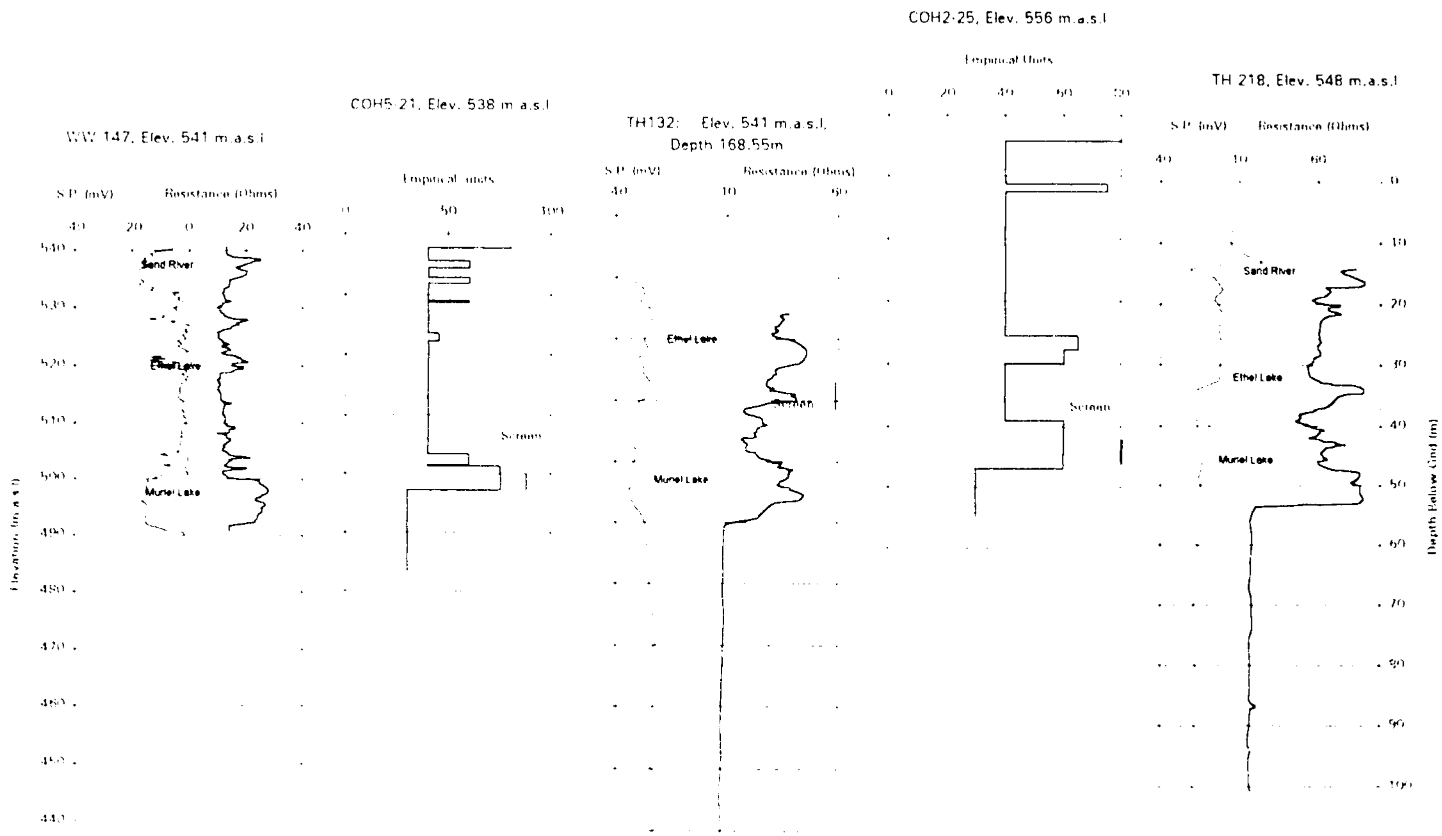


Figure 4.48. SP Resistance and lithology profile through the Ardmore Thermal Project site. Screen location in TH132 is the equivalent elevation in CWW2

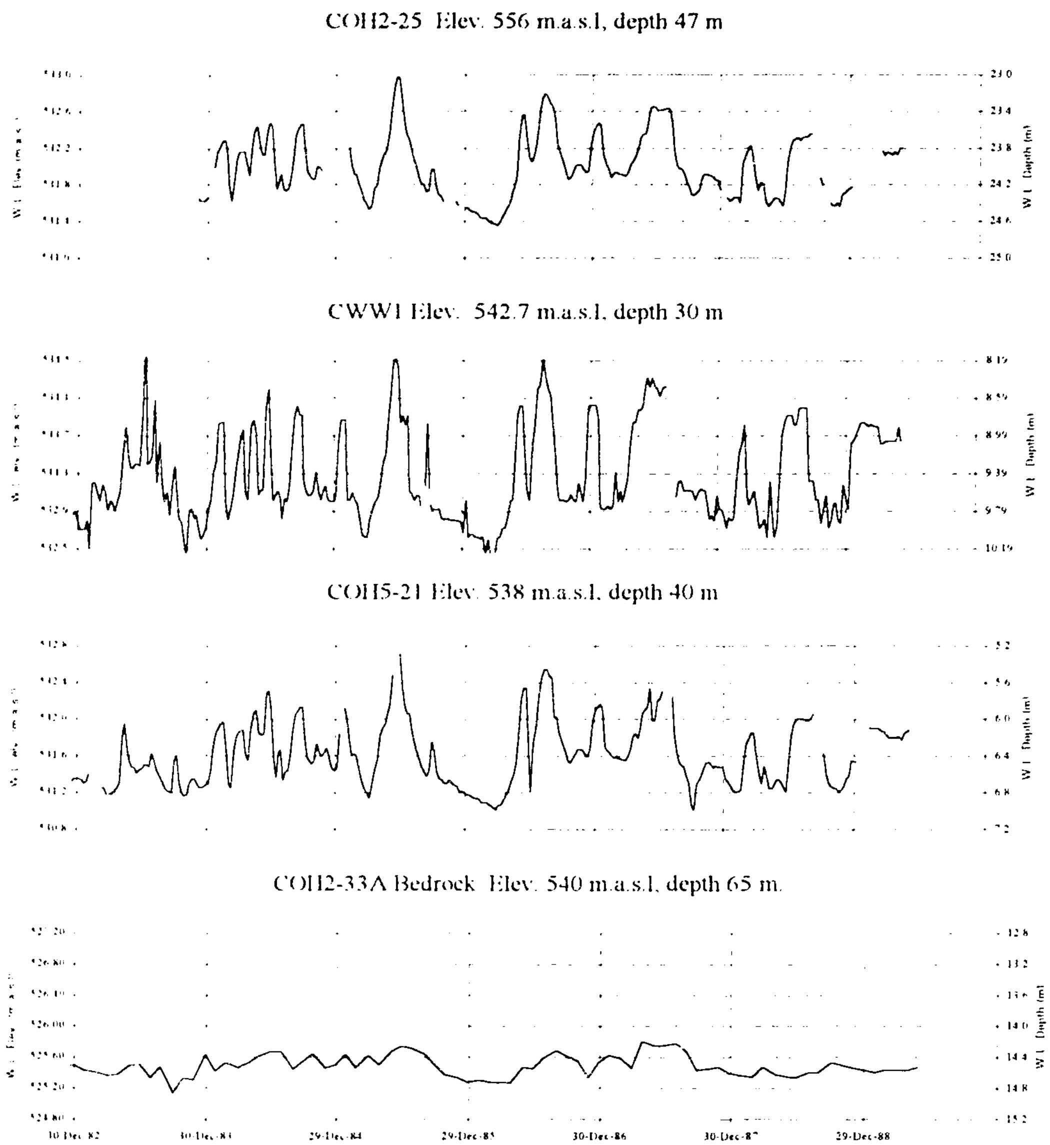


Figure 4.49. Water level responses to pumping, 1978 pumping test. (Data courtesy of Hydrogeological Consultants).

5. Development of approaches for preliminary recharge estimates

Three approaches were developed from the data interpretation which appeared to be suited to the Ardmore area: the hydrograph, flow system and pump test approaches. This section describes each approach individually.

5.1. Hydrograph Approach

A hydrograph curve for an observation hole, is a graphical representation of water level position through time. The water level position is the cumulative result of rates of inflow minus rates of outflow to and from the saturated zone respectively.

We assumed that in shallow holes the water level was a good approximation of the water table. The water level in holes drilled below its actual position, represented an averaged value of hydraulic head over the intersected saturated interval (Sec. 3.1). In the worst cases in which the holes were drilled several meters below the water table, we estimated that the error between the measured and the actual level should not be more than 1.5 m, based on the hydraulic head gradients.

5.1.1. Theoretical background

To begin the development of an approach for recharge estimation using shallow-hole hydrographs we return to the concept of an volume element introduced in Section 1.5 (Fig. 5.1a). The volume element is placed in a recharge area where the saturated flow is downward and away from the water table. The volume includes the unsaturated zone and a portion of the saturated zone below the water table. The upper boundary is the land surface. We assume for simplicity that flow is vertical in the unsaturated zone and that the sum of any horizontal inputs and outputs to the volume, in the saturated zone, is zero.

Water enters the volume during an accretion event (defined below) through the land surface at a rate known as the infiltration rate (I_s). Assuming that the actual rate of infiltration does not match or exceed the maximum infiltration rate for the soil, the water flows by processes of unsaturated flow generally towards the water table. A number of water sinks are present in the unsaturated zone: for example, some of the water may be used to bring the soil moisture to field capacity and some may be lost to transpiration during the growing season. During an accretion event, however, some water will reach the water table at the rate known here as accretion (I_g), and be added to the saturated zone. Water will flow away from the water table at the recharge rate (R), within the saturated zone.

Some water is lost again directly from the saturated zone in a recharge area, mainly through the combined processes of evaporation and transpiration (ET_g). These losses, together with losses from the unsaturated zone, leave the volume element through the upper surface as exfiltration (ET_s).

In the interpretation of observation hole hydrographs, we were concerned only with the exchange of water between the saturated and unsaturated zones at the water table. In view of the flow regime prevailing in the Ardmore area we confined our attention to processes occurring in a recharging flow system. We therefore considered a smaller volume element (Fig. 5.1b), and wrote an expression for the water balance for some time period, Δt, with the stipulation of steady state conditions,

$$I_g\Delta t = (R + ET_g)\Delta t \qquad [L^3] \qquad (5.1)$$

where the rates are defined above.

The condition of steady state described by Equation 5.1 would appear as a horizontal line on the hydrograph; in other words there would be no change in the position of the water table with time. The fact that the water table is a subdued replica of the topographic surface is strong evidence that over long time periods a form of steady state must indeed prevail. The mean position of the water table over long periods is in fact the result of a dynamic equilibrium in which the rates of water addition are matched by those of water losses.

The hydrographs collected in the Ardmore area showed that over shorter time periods the position of the water table is constantly changing. This demonstrated that in practice, transient conditions are normal at the water table. The change in elevation of the water table thus represented a change in storage, ΔS, so for transient conditions we wrote the water balance equation to include this change,

$$I_g\Delta t = (R + ET_g \pm \Delta S)\Delta t \qquad [L^3] \qquad (5.2)$$

where all the terms in the equation are rates, $[L^3/T]$.

We now wished to relate the change in the water table position, Δh, to the change in storage. This was done by multiplying the length Δh and the cross-sectional area of the volume element, A, with a storage coefficient as follows.

$$S_y \Delta h A = \Delta S \Delta t \qquad [L^3] \qquad (5.3)$$

where the coefficient, S_y, was the specific yield of the material in which the water table was fluctuating.

Equation 5.2 was rewritten with the storage changes expressed in terms of changes in the position of the water table, and solved for storage,

$$(I_g - R - ET_g)\Delta t = \pm S_y \Delta h A. \qquad [L^3] \qquad (5.4)$$

Using the hydrograph we could estimate a rate of change in storage from the slope of the hydrograph curve over a given time interval, $\Delta h/\Delta t$. It was thus convenient to perform our interpretations using fluxes rather than volumes. We converted the terms of Equation 5.4 to fluxes by expressing them as rates per unit surface area. The water balance equation for the vertical exchange of water at the water table then appeared as,

$$I_g - R_z - ET_g = \pm S_y \Delta h/\Delta t. \qquad [L/T] \qquad (5.5)$$

which is the basic equation we used to estimate fluxes from hydrographs.

5.1.2. Estimates based on hydrograph interpretation

The hydrograph approach for the estimation of any of the terms on the left hand side of Equation 5.5 is to choose periods when one or more of them are either constant or equal to zero. We selected the period of winter ground frost to estimate the saturated recharge flux by assuming that accretion and evapotranspiration were negligible under frozen conditions (see discussion below). Equation 5.5 then becomes,

$$R_z = S_y \Delta h/\Delta t. \qquad [L/T] \qquad (5.6)$$

Based on the flow regime described for Ardmore, we made the further assumption that the change in water table position did not substantially alter the hydraulic gradient in the flow systems, and that the recharge flux remained approximately constant throughout the year. That a more or less constant gradients can exist under a fluctuating water table throughout the year, was demonstrated by Freeze and Banner (1970) at their experimental site near Calgary, Alberta.

The evapotranspiration flux was estimated during periods of negligible accretion and constant recharge with,

$$ET_g = S_y \Delta h/\Delta t - R_r \qquad [L/T] \qquad (5.7)$$

and likewise the accretion flux was estimated during periods of negligible evapotranspiration,

$$I_g = S_y \Delta h/\Delta t + R_r. \qquad [L/T] \qquad (5.8)$$

or during periods when the evapotranspiration had already been estimated.

$$I_g = S_y \Delta h/\Delta t + R_r + ET_g. \qquad [L/T] \qquad (5.9)$$

5.1.3. Frost effect

The assumption of negligible I_g and ET_g under frozen conditions is probably accepted by most researchers in this field. There is, however, much speculation regarding the transfer of moisture within the unsaturated zone during freezing and thawing (Sec. 4.1). This implies that estimates of the recharge flux directly from the winter recession may be too high since a portion of the water level fall cannot be attributed to downward flow. Two studies in Alberta appear to suggest that the loss of water during winter to a developing frost wedge is not important from relatively deep water tables (> 2.0 m). Freeze and Banner (1970), working on an experimental plot in Calgary, showed that, whereas moisture gains could be measured in the zone of freezing, during winter, at the expense of soil moisture losses from lower in the profile, there was no clear link between these losses and the declining water table. Soil moisture gains in the zone of freezing were also recorded in winter by Maclean (1974) above a water table at approximately 2 m depth. In this case, he estimated that the water gain represented approximately half the equivalent depth of water loss from the saturated zone. Holes with deeper water tables in his study area near Vegreville, however, showed no similar transfers.

We looked for evidence of water transfers to and from a developing frost wedge in our own data. It appears to take at least a month for the frost wedge to thaw completely in spring (Freeze and Banner, 1970; Maclean, 1974). Provided that most of the frost-moisture returns to the water table, we might expect to see a relatively slow and constant rise in the water table during the period of thaw. The period would presumably start when the mean air temperature at the surface begins to rise above freezing. At Ardmore this was approximately the end of March 1993. We were able to detect frozen conditions within the first 0.5 m of soil with a hand auger, at most sites, until the middle of April. Wetland sites remained frozen to this depth through most of May. The winter recession in some depression holes ended in early March and the hydrograph curve flattened prior to spring rise. UOH 2 for example has a flat portion in the hydrograph curve between March 16 and April 1, 1993 (Fig. 4.6). A gradual change in the water table position, such as this, could easily be attributed to ground frost thawing but its occurrence may be a little early. On the other hand, the large and relatively rapid rises in late March and early April, before much of the frost had thawed, are interpreted by us to be the result of meltwater penetration of the frozen zone along unfrozen pathways (Startz, 1969).

The results of a literature search indicated that there has been little recent attention, outside of the regions of continuous permafrost, given to the transfer of water from the water table to freezing layers within the unsaturated zone. Many studies have focused on the potential for infiltration into frozen ground (Kane and Stein, 1983b; Granger et al., 1984; and Burn, 1991) but none of these are concerned with upward migration from the water table. Soil moisture profiles from the paper by Kane and Stein show little evidence of moisture movement upwards in the profile below the frozen layers. The position of the water table is not shown but it is inferred from the profiles, which extend to a depth of almost 2 m, to be much deeper than 2 m.

The combination of relatively low permeability and deep water tables at most of our sites was assumed to ensure in negligible water transfers from the saturated zone to a developing frost wedge in winter. The volume of transfer is anyway probably smaller than the errors involved in the choice of an effective specific yield.

As previously stated, in some instances, transfers of water to and from the developing frost wedge may result in an overestimate of fluxes. The difference between the overestimated values may remain approximately the same, however, and our estimates of net accretion from the Troxell method, should be the most accurate (see below).

5.1.4. Methods of hydrograph solution

The approaches we developed are based on two methods of hydrograph analyses originally used to estimate evapotranspiration rates from diurnal water table fluctuations. In this section we describe how we have adapted these methods to estimate all rates of inflow and outflow at the water table, based on the relationships developed above, and extended them over longer periods of up to a water year.

The first method is credited to White (1932) who observed that the constant rate of night time rise between the hours of approximately midnight and 4 a.m. should be directly related to the rate of groundwater discharge (discharge is defined here as the rate of saturated flow below and towards the water table). His reasoning was that during this time, losses to evapotranspiration should be negligible.

To demonstrate the method, we used a portion of the hydrograph from UOH 12 (see Section 4.1 for discussion) over the period July 18 to July 20, 1992 (Fig. 5.2). Letting r equal the hourly rate of rise of the water table between midnight and 4 a.m. July 19, as shown in the upper curve, and s the net fall or rise of the water table during the subsequent 24 hour period, then the equation

$$H = S_y(24r \pm s) \qquad [L] \qquad (5.10)$$

can be used to give a good approximation of the diurnal depth of water lost to evapotranspiration. The result can also be expressed as an average daily flux.

A second method of estimating the same evapotranspiration loss was developed by Troxell (1936). He noted that if the slope of the night time rise on the hydrograph represented the saturated flow rate, then the flat portions of the curves must represent periods when the rates of discharge and evapotranspiration were equal. Likewise the inflection point during the daytime fall represented the peak period of evapotranspiration; at which time it greatly exceeded the discharge rate. To estimate these rates he plotted the slope, $\Delta h/\Delta t$, of the hydrograph curve for hourly intervals over the 24 h period, as shown in the lower portion of Figure 5.2 (note: 2 h intervals shown in the figure). The depth of evapotranspiration was then estimated at any time as the difference between the slope of the hydrograph and the maximum night time rate. The area between the curves was used to approximate the total diurnal evapotranspiration as computed by the White method above. We also noted that the total discharge could be computed from the area between the night time maximum and zero slope curves as shown on Figure 5.2.

To demonstrate our application of the methods to longer time periods, we have applied them to the interpretation of a schematic hydrograph shown in Figure 5.3. The hydrograph shown in the upper portion of Figure 5.3

represents a complete water year with a recharging groundwater flow regime typical of the Ardmore area. In this case the slope of the winter recession (letter a, Fig. 5.3) is used as the constant rate, in place of the night time rate for the diurnal case.

To apply the White method over the period of a water year we projected the slope of the winter recession from the spring minimum of one year below the spring minimum of the following year. The White formula is then written as,

$$H = - S_y(Yr - s) \qquad [L] \qquad (5.11)$$

where H is the net depth of accretion to the groundwater over the water year, Y is the length of the water year, r is the rate of winter recession (negative slope) and s is the difference in height between the spring minima. The result, H, net accretion over the period, is discussed in more detail below.

The method can only be applied if the recessions for the two winters have approximately equal and constant slopes; the underlying assumptions are i) that the material in which the water table is fluctuating is homogeneous and ii) that there is no change in the hydraulic gradient. This method should result in a reasonable estimate since we assumed that water flowed into the system at a constant saturated flow rate (recharge) even though rates of supply and depletion vary at the water table.

The application of the Troxell method to a water year is more involved and becomes quite laborious for complicated hydrographs. Returning to Figure 5.3, we note again that the slope of the winter recession is taken to represent the rate of recharge (letter a, Fig. 5.3). Where the slope of the hydrograph flattens at the beginning and end of the spring rise (letters b and d, Fig. 5.3), recharge is exactly balanced by accretion, assuming that evapotranspiration at the water table is negligible in this part of the growing season. The inflection point in the slope of the spring rise (letter c, Fig. 5.3) represents the time for this period when the accretion rate is at a maximum. Likewise, the inflection point in the summer recession is the time of maximum evapotranspiration (letter e, Fig. 5.3).

The hydrograph is divided into suitable time intervals and the slopes are estimated over each interval. The slopes are then plotted on a second graph shown in the lower portion of Figure 5.3.

The slope of the winter recession plots on the negative half of the slope graph (letter a, Fig. 5.3) and can be extended across the graph because the saturated flow is thought to be approximately constant with time. This then represents the constant recharge rate.

The Troxell method is now applied to the estimation of depths as follows. The depth of accretion, I_g, is estimated at any time from the distance between the recharge curve and the slope of the hydrograph during periods when evapotranspiration is negligible. This is the graphical solution to Equation 5.8. The solution to Equation 5.9 is given by the distance h in Figure 5.3 where the accretion depth is estimated during periods of evapotranspiration.

The depth of evapotranspiration, ET_g, at the water table is calculated during periods with no accretion. It is then the difference in slope between the summer and winter recessions as indicated in Figure 5.3 which is the solution to Equation 5.7. For the intermediate periods at the beginning and end of the summer, when the evapotranspiration flux is less and overlapped by periods of obvious accretion, it is estimated by sketching the trend of evapotranspiration on the slope graph (letter g, Fig. 5.3).

The total accretion, I_{gT}, recharge, R_T, and evapotranspiration, ET_{gT}, in any given time period can be estimated from the slope graph by calculating the areas between the curves.

Net Accretion:

Of particular interest is the accretion that is converted to recharge over the period of a water year, here called net accretion. This is the water that is added permanently to the saturated zone and in effect, sustains the saturated flow. It is the difference between total accretion, I_{gT}, and total evapotranspiration, ET_{gT}, at the water table, represented by,

$$I_{gnet} = I_{gT} - ET_{gT} \qquad [L/T] \qquad (5.12)$$

and can be compared directly with the result of the White method.

A characteristic specific yield is applied in the White method, during several months of frozen conditions, to a relatively deep water table which is falling slowly. These are the most appropriate conditions for the application of a characteristic specific yield (see below). During the shorter periods of estimation by the Troxell method an effective specific yield might be expected to show some variation for which adjustments may or may not be applied. The comparison of a net accretion from the two methods, therefore, provides a form of verification of the results.

In an area such as Ardmore, which experiences very little runoff, it is worth noting that the average precipitation less the net accretion, for long periods, gives an estimate of total evapotranspiration from both surface and subsurface sources assuming that the soil moisture storage remains more or less constant (Kovács, 1981).

The net accretion for a water year can be compared to the annual recharge rate. If it is less, as was the case at Ardmore for the period of study, it indicates a net loss from groundwater storage and will be accompanied by an overall drop in the water table position. The average net drop in the water table at Ardmore between 1992 and 1993 was 0.3 m.

5.1.5. Definitions of recharge and accretion

The hydrograph approach lead to the need for specific definitions of the terms in the water balance equations. From Equations 5.1 and 5.2 it is clear that the terms accretion (I_g) and recharge (R) are not equal. We therefore define the two terms as follows: i) accretion is the rate or process of supply of water at the water table from the unsaturated zone, and ii) recharge is the rate or process of flow of water below and away from the water table within the saturated zone. In our view, when the word recharge is used to describe the replenishment of groundwater at the water table it in fact describes a two component process: the first is the process of loading a flow system at the water table, accretion, and the second is the process of transport within the flow system, recharge. Neither the processes nor the rates of the two components are equivalent.

5.1.6. The application of a storage term

A serious difficulty arises in the application of the hydrograph approach over the use of a meaningful storage term. Conceptually the correct term to use is the specific yield.

Specific yield is defined as the volume of water released from storage per unit surface area per unit decline in the water table (Freeze and Cherry, 1979). Freeze and Cherry suggest that the best way to visualize the specific yield is with reference to the saturated-unsaturated interaction it represents. They use a figure similar to Figure 5.4 which shows the water-table position and the vertical profile of soil moisture at two times, t_1 and t_2. The shaded area represents the volume of water released from storage in a column of unit cross section. The water-table drop can be represented as a unit decline so that the shaded area represents the specific yield.

For a homogeneous medium, the two controlling factors for an effective specific yield are the shape of the soil moisture profile (antecedent moisture conditions) and the time allowed for drainage. Highly permeable materials such as sand will release water from storage more quickly than low permeability materials such as clay. Experiments show, however, that even in clean well sorted sands complete drainage does not occur for several days (Johnson et al., 1963).

The shape of the soil moisture profile is dictated by the previous wetting and drying history. Its effect can again be best examined with the use of a figure representing the saturated-unsaturated boundary. Two end member cases are shown in Figure 5.5. In the first case the water table is very close to the land surface and small fluctuations occur in a fully saturated profile. Here the moisture content is equal to the porosity, n; the saturated part above the water table being held under tension. Under these circumstances there is no change in storage with time as the water table drops a unit distance and the specific yield is zero. The surface layers remain saturated, but under increasing suction. In the second case the water table declines under a constant soil moisture profile, the soil in the upper part now representing field capacity, c_{fc}, (Fig. 5.5). In this case the specific yield is calculated as the difference between the initial moisture content, c_0, and field capacity. The resulting constant value of specific yield is a materiel parameter characteristic of the soil type and occurs where the water table is deep and falling fairly slowly (Childs, 1969).

It was with these considerations in mind that we applied the characteristic specific yield to the winter recession for estimates of the recharge rate, and to the late intervals of prolonged dry periods for estimates of losses to evapotranspiration.

There are many possible conditions between the end members of Figure 5.5 that require the use of an effective specific yield. Gillham (1984) has discussed at length the effects on water table responses when the lower part of the moisture profile (capillary fringe) intersects the land surface. This is the condition close to the first case discussed above where the effective specific yield is very small. This can lead to large and rapid fluctuations in the water table which appear disproportionate to the volume of infiltrated water. Furthermore, if the characteristic value of specific yield was used, it would result in a large over-estimate of the volume of infiltrated water.

In summary, if the portion of the soil moisture curve above field capacity intersects the land surface we can expect a highly variable value for the effective specific yield. It should also be pointed out that the quantity of water released from storage will not be exactly the same as the quantity taken into storage as the water table rises. This is because of hysteretic effects in the soil in response to wetting and drying and as a result of the entrapment of air by rising water levels.

The method of estimating accretion from the area between slope curves allows the assignment of new values for effective specific yield for each time interval. In a hole such as UOH 11 where the water table was very shallow in the early season, an estimate of the specific yield was attempted using the equation of Walton (1962),

$$S_y = (P - ET - R_z)/\Delta h. \qquad [L/L] \qquad (5.13)$$

It was assumed for the application of this equation that all evapotranspiration was occurring directly from the shallow water table (and therefore no change in soil moisture storage) and that runoff was negligible. We used published values of areal ET (Bothe and Abraham, 1993) estimated by the CRAE method (Morton, 1983) and our own measurements and estimates of P and R_z respectively.

5.1.7. Practical considerations

There is a delay in the water level response caused by the hole geometry and the hydraulic conductivities of the enclosing materials (Horslev, 1951). Although these delays can affect the flux estimates by the hydrograph method, we did not attempt to make corrections. We assumed that, whereas the time lag produced a phase shift in the water table fluctuations it did not substantially change the slope of the curves or the values of their maxima and minima. This assumption seemed reasonable for most of the period of observation, because the time between accretion events was usually long.

In applying the White method to a water year, we sometimes took the average of the slopes of winter recessions if they were different, but we usually relied on the slope for the 92/93 winter because we had better records for this period.

5.2. Flow system approach

The hydrograph approach already incorporates a flow system approach in a fundamental way. The interpretation of the winter recession and resulting estimate of recharge, is based on the assessment that the area is in a dominantly descending flow regime.

In the present section we wish to develop an approach which may help to support the validity of the recharge rates estimated from the hydrographs. The estimate can also be considered independent of the hydrograph approach since it is based on the choice of a value of hydraulic conductivity rather than of specific yield. We start from another result of the flow pattern analysis which has indicated that the flow regime can be approximated by a single intermediate system with most of the recharge within the study area and discharge along the northern boundary to the Beaver River Valley. Our approach is then to attempt an estimate of the total discharge, D, to the valley for comparison with total recharge, R, over the surface of the study area computed with the hydrograph generated recharge fluxes. In this approach we are assuming that inflows to the area in the subsurface are either negligible or matched by outflows as in the Bronson Lake Channel. We are also assuming that discharges from local systems within the area, which are too small to be identified with the methods we applied, are either negligible or accounted for by the cumulative nature of the hydrograph itself; in other words, that the discharge in local systems is a part of the estimated ET_g.

We have noted that some loss can occur through the poorly permeable bedrock and an estimate of this loss must be added to the valley discharge. Artificial water withdrawals must also be accounted for so that the total water balance equation for the area appears as,

$$R = D + W + R_{col} \qquad [L^3] \qquad (5.2.1)$$

where R and D are defined above, W is local withdrawals and R_{col} is leakage through the poorly permeable bedrock.

Discharge to the Beaver River Valley was approximated by estimating flow through vertical slices of each cross-section (Sec.4.2) multiplied by the length of valley measured to half the distances between them. The slices were placed very close to the valley wall and thus approximated the position of a midline representing the boundary through which water is exchanged between recharge and discharge areas (Tóth, 1962).

The total discharge, D, from the area is the sum of discharge estimated along each cross-section segment of the valley as follows,

$$D = \Sigma(D_1 l_1 + D_2 l_2 + \ldots + D_n l_n)\Delta t \qquad [L^3] \qquad (5.2.2)$$

where D_i, is the discharge through a midline of unit width and l_i, is the length of valley assigned to each cross section.

Hydraulic conductivity values obtained from local reports and estimates from local pump tests (Tokarsky, 1975; Schwartz, 1987; this study), were assigned to each hydrostratigraphic unit. The horizontal hydraulic gradients across the midline slice in each unit were estimated from the potentiometric surfaces (Sec. 4.2). Flow, D_i, through a midline section of unit width was computed by summing the products of the hydraulic conductivities, the unit thicknesses and the horizontal hydraulic gradients, using Darcy's Law,

$$Di = \sum_{1}^{n} \left(K_1 m_1 \frac{\Delta h_1}{\Delta x_1} + K_2 m_2 \frac{\Delta h_2}{\Delta x_2} + \ldots + K_n m_n \frac{\Delta h_n}{\Delta x_n} \right) \qquad [L^3/T] \qquad (5.2.3)$$

where m_i and K_i are unit thickness and hydraulic conductivity respectively and Δh_i, the change in hydraulic head over the horizontal distance Δx_i in each unit.

For the estimate of local withdrawals we observed that the only local source of domestic water is from the Quaternary aquifers. The estimate was thus made by determining the number of inhabitants and average number of livestock in the area and multiplying by water usage statistics released by Alberta Agriculture.

The vertical hydraulic gradient across the Colorado Group was derived from Figure 4.33. Leakage through the bedrock was then estimated from the product of an assigned hydraulic conductivity, the vertical gradient and the horizontal surface area.

5.3. Pump test approach

The 1978 pump test data from the Excel Energy Site showed that drawdowns in the pumping and observation wells in the exploited aquifer stabilized after a period of time (Fig 4.44). When no further drawdowns are observed in an aquifer, it is assumed that the aquifer is being recharged and that losses from aquifer storage become negligible. The aquifer is acting simply as a conduit to flow (Kruseman and de Ridder, 1990).

The approach described in this section gives an estimate of the recharge flux across the aquitard induced by pumping. The assumption in the approach is that recharge to the aquifer is entirely by leakage through the overlying aquitard. The estimate can be made, following Walton (1962), from the quotient of the total withdrawal rate and the area of influence. The withdrawal rate was calculated as the sum of the mean pumping rates in both wells over the period of the 1978 pump test. The radius of the area of influence was estimated from the distance drawdown curves generated from the observation wells in the overlying Sand River aquifer (Fig. 4.47). The equation to estimate the vertical recharge flux, R_z, is then,

$$R_z = Q/A_e \qquad L/T \qquad (5.3.1)$$

where Q is the pumping rate and Ae the effective area, (i.e., the area within which vertical leakage is being diverted to pumping).

The induced recharge flux so calculated is not evenly distributed over the area of influence, but increases with the increasing hydraulic gradients towards the centers of pumping. The result from Equation 5.3.1 is a normalized flux which may be greater or less than the value existing at any point within the area.

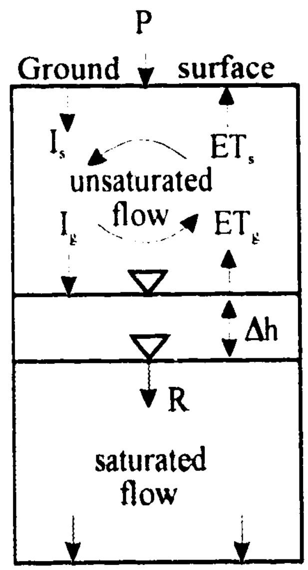


Figure 5.1a. Volume element in a recharge area showing water movement in portions of the saturated and unsaturated zones. The letters refer to rates of inputs and outputs where P = precipitation, I_s infiltration, I_g = accretion, R = recharge, ET_g = evapotranspiration at the water table and ET_s = exfiltration. Δh is the water table fluctuation. (Modified after Kovăcs, 1981).

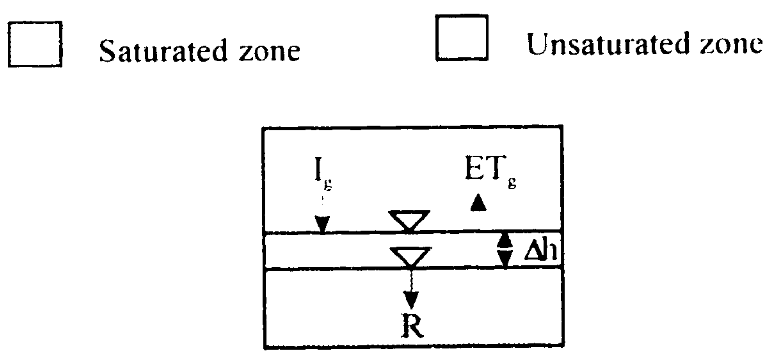


Figure 5.1 b. Volume element at the water table in a recharge area showing water exchange between the saturated and unsaturated zones. Letters are defined above. (Modified after Kovăcs, 1981).

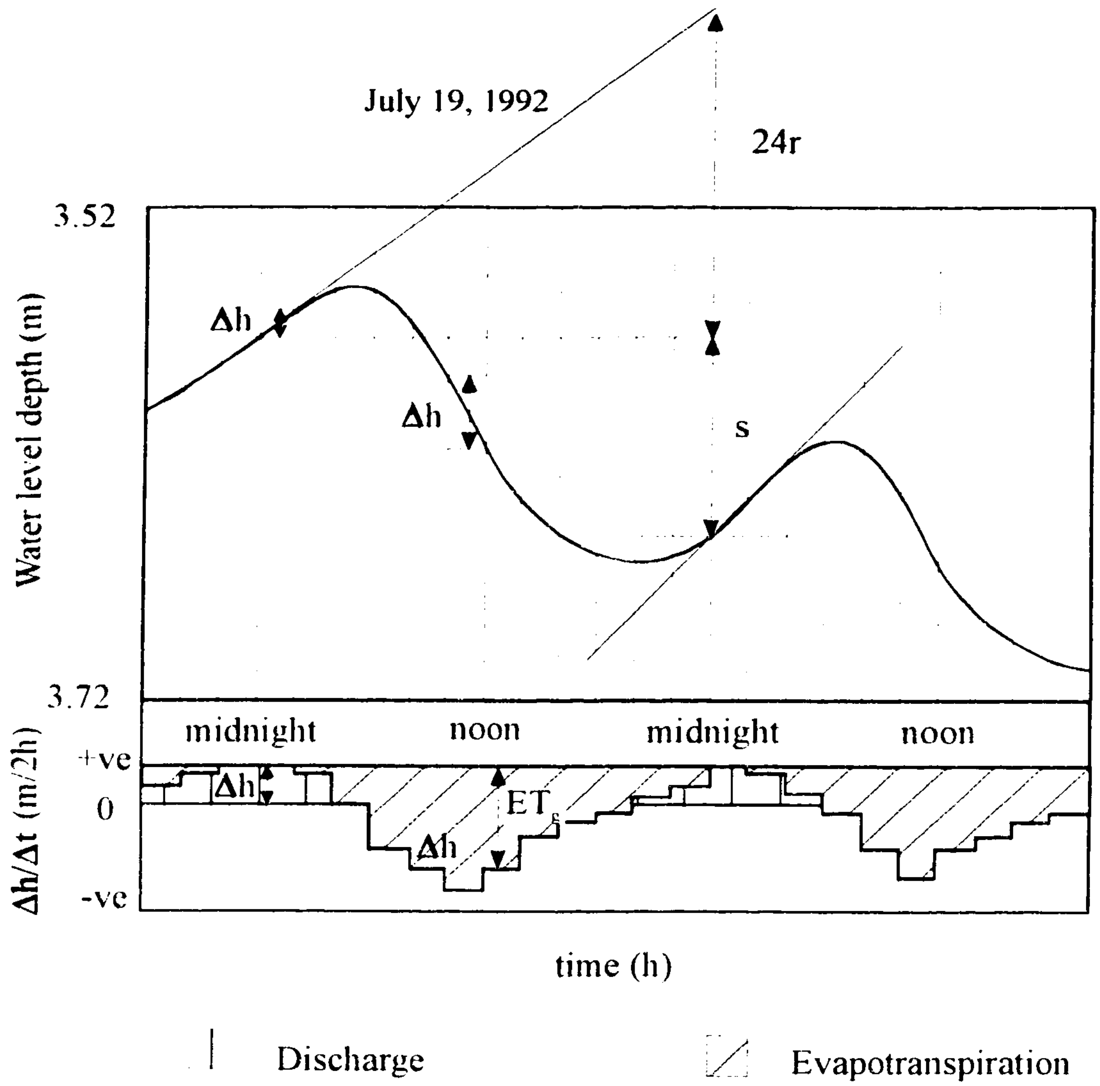


Figure 5.2. White and Troxell methods applied to diurnal fluctuations. The upper portion shows diurnal fluctuations in the water table in UOH12 with an illustration of the White method of ET_g estimation. Depth of water loss, H = $S_y(24r + s)$. The lower portion is a plot of the hydrograph slope averaged over 2 h intervals with time. The Troxell method estimates the depth of water loss to ET_g at any time fom the distance between the curves. The total depth of water lost and gained is estimated from the shaded areas between the curves.

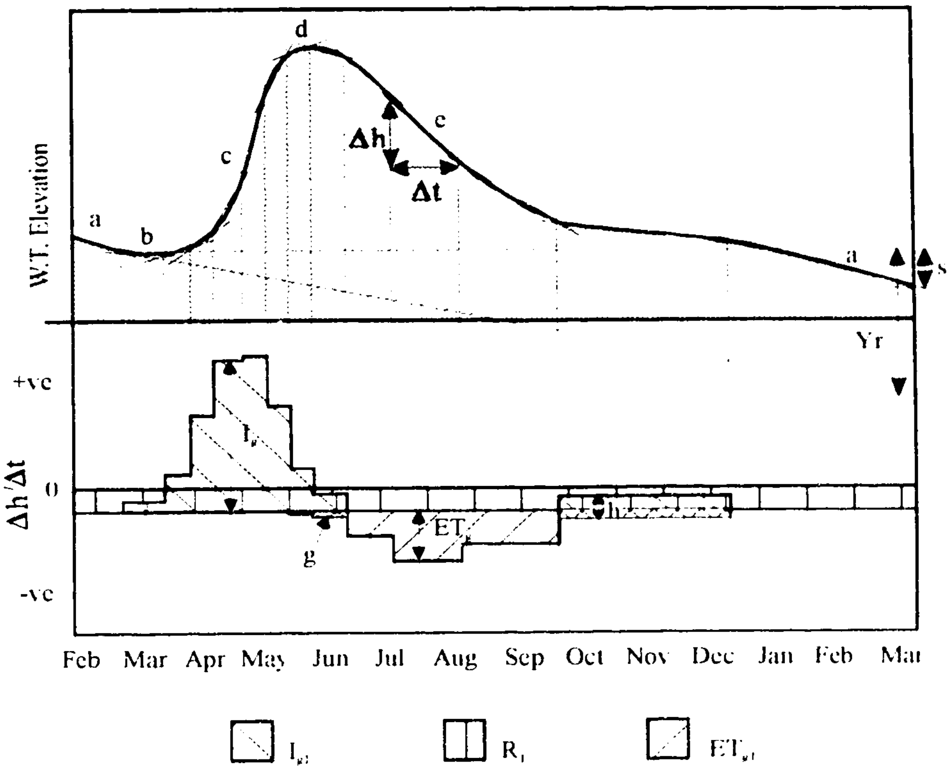


Figure 5.3. White and Troxell methods applied to a water year (schematic). Total annual depths are calculated from the shaded areas.

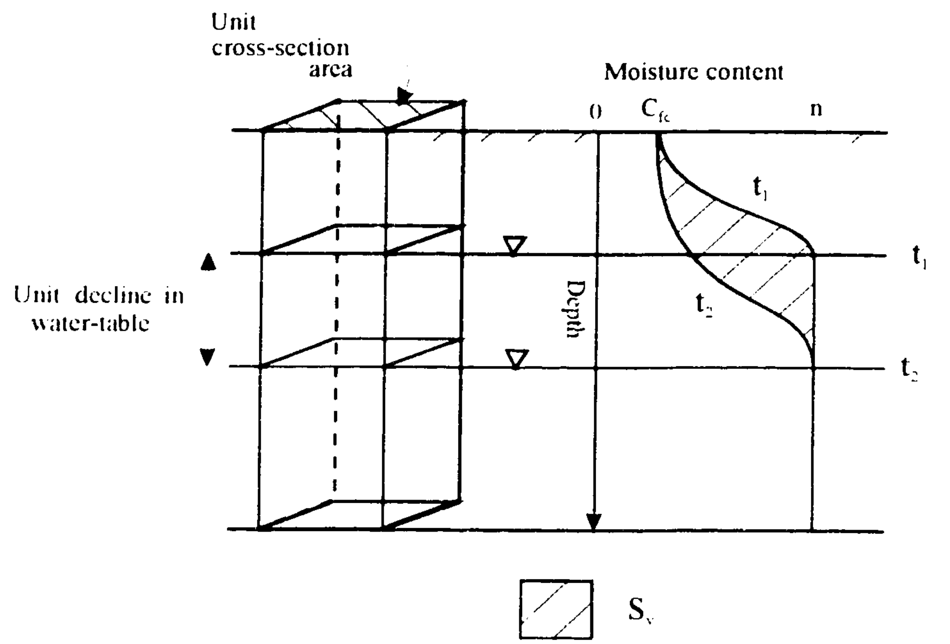


Figure 5.4. The concept of specific yield; n is porosity, C_{fc} is soil moisture at field capacity. The water released from storage for a unit decline in the water table is represented by the shaded area between the two moisture profile curves at times t_1 and t_2. (Modified from Freeze and Cherry, 1979).

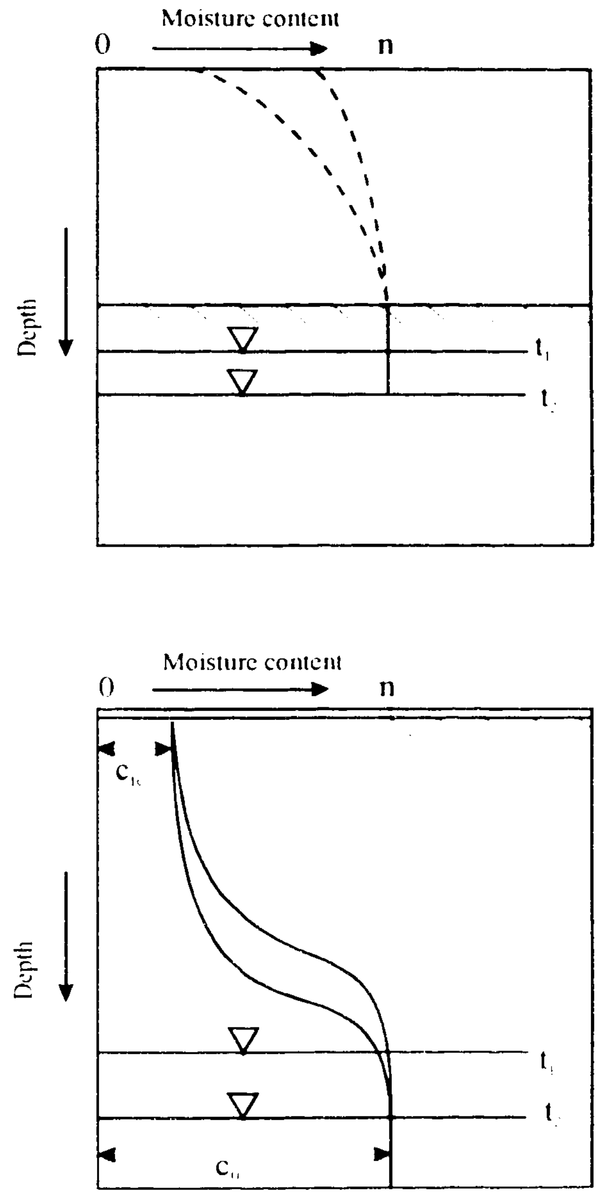


Figure 5.5. Soil moisture profiles. Upper diagram illustrates the case of zero specific yield for a decline in water table. Lower diagram illustrates characteristic specific yield from drainage below a constant moisture profile; the upper parts of the profiles are at field capacity whereas the lower parts are fully saturated (Modified after Childs, 1969).

6. Preliminary estimates

Preliminary estimates of fluxes at the water table have been calculated by each of the three approaches: Section 6.1 presents the estimates by the hydrograph approach, Section 6.2 the estimates by flow system analyses and Section 6.3 those from pump testing. Using the methods developed in the hydrograph approach, it is possible to estimate each of the fundamental fluxes active at the water table (Sec. 5.1) and to examine them over a variety of time intervals. For these reasons and because the emphasis of the study was on water table observations, the estimates using the hydrograph approach are the most extensive

6.1. Hydrograph Estimates

Preliminary estimates of fluxes using the hydrograph approach are presented in order to exemplify the spatial and temporal variability seen at Ardmore. We show spatial variability on an annual and an event basis, and temporal variability on an event and a diurnal basis. Estimates have been made of each of the following fluxes: i) recharge, ii) accretion (total and net for each period) and iii) evapotranspiration at the water table. In Section 6.1.5 an estimate is made of the total recharge over the study area for comparison with the result of the flow system approach.

6.1.1. Spatial variability, annual basis

Tables 6.1 and 6.2 give a summary, by groups, of the results from hydrograph analyses over the period of the 1992/93 water year. They are given as percentages of precipitation and as fluxes in the two tables respectively.

The mean recharge flux is slightly less in the upland and approximately equal in the lowland and medium depression sites but the variances about the means increase towards the depressions. Recharge is greater and less in the fast and slow depressions respectively. The reason the mean recharge flux appears to be slightly higher in the lowlands, as compared to the uplands, may be a result of the Schneider frost effect, (Meyboom, 1967b) as discussed in Section 5.1. The loss of groundwater during winter to a developing frost wedge, may be greater in the lowlands because the mean depth to the water table is less. This would cause an apparent increase in the recharge flux.

Table 6.1

Equivalent depths** expressed as a percentage of precipitation for the 1992/93 water year (355 mm).

Sites	Recharge % of pp.	Net Accretion % of pp.	Total Accretion % of pp.	Total ET_g* % of pp.
Upland	9.0 (0.4)	5.8 (0.3)	6.8 (2.0)	1.1 (2.0)
Lowland	10.8 (2.8)	5.2 (0.8)	15.8 (6.6)	10.6 (7.4)
Depressions				
Slow	4.9 (2.0)	2.9 (2.5)	9.6 (4.2)	6.7 (5.3)
Medium	10.7 (5.2)	8.1 (5.9)	29.5 (15.3)	21.4 (16.7)
Fast	14.1 (1.9)	11.3 (3.7)	29.4	18.1

* Evapotranspiration at the water table.
**= Measured depth x S_y.
(2.0) = Standard deviation.

Table 6.2

Average fluxes for the 1992/93 water year.

Sites	Recharge 10^{-9} m/s	Net Accretion 10^{-9} m/s	Total Accretion 10^{-9} m/s	Total ETg* 10^{-9} m/s
Upland	1.1 (0.08)	0.7 (0.05)	0.8 (0.3)	0.1 (0.3)
Lowland	1.4 (0.4)	0.6 (0.1)	1.9 (1.2)	1.3 (1.2)
Depressions				
Slow	0.6 (0.2)	0.3 (0.2)	1.2 (0.5)	0.9 (0.7)
Medium	1.4 (0.7)	1.1 (0.7)	3.8 (2.0)	2.7 (2.0)
Fast	1.8 (0.2)	1.3 (0.4)	3.5	2.2

The net accretion (Equation. 5.12) for the specified water year is slightly greater in the upland than in the lowland sites and varies widely in the depressions. The reason for a higher accretion in upland sites may be related to the seasonal pattern of precipitation (Sec. 2.2). Precipitation occurring in the early spring rather than later in the growing season is more likely to reach a deep water table because the vegetation is dormant. If the water table is beyond the reach of shallow rooted cereal crops later in the summer, water accretion in the spring will remain as net accretion; in lowland sites on the other hand, the spring accretion is available to plants later in the growing season from a shallower water table and may be lost through transpiration. The fact that accretion is everywhere less than the recharge reflects the average decline in the water table of about 0.3 m over the time frame of the study.

With the exception of the slow group, total accretion is greatest in depressions and least on the uplands. The same pattern is seen in total evapotranspiration at the water table; it shows a more rapid decline, however, as the uplands are approached.

The reason why the slow group of depressions appeared to receive less total accretion and ET_g than other depression sites is not immediately obvious. The average depth to the water table, for example, is similar to the medium group (Table 4.1). It is easy to understand that ET_{gT} would be lower from a dried wetland in which the phreatophytic vegetation has died-back, as was the case in the summer of 1992, but this occurred at all the sites. We also need to explain why the slow depressions received less accretion. It is possible that they did not receive as much runoff and meltwater in the spring as other depression sites. The most likely explanation, however, is that the material in the unsaturated zone is less permeable at the slow depression sites. Infiltration and accretion are both slower and more water is lost to evapotranspiration in the unsaturated zone than directly from the water table.

Generally there is a correlation between the estimated fluxes and the depth to the water table. Figure 6.1 is a plot, with depth on the ordinate axis, of the values I_{gT}, ET_{gT} and I_{gnet} estimated in the university observation holes (Appendix B, Table 3). Data points from individual holes representing values of total accretion are plotted on the positive side of the abscissa, whereas points of total evapotranspiration are plotted on the negative side; points of net accretion plot in between. Lines and shaded areas are sketched on the graph to indicate interpreted trends. The lines are constructed to approximate the mean position through each of the three groups of data. The shaded areas show the trends for total accretion and evapotranspiration. Several interesting features are demonstrated by the graph:

i) There is generally an inverse relationship between water table depth and both I_{gT} and ET_{gT}.
ii) Very little water is lost to evapotranspiration from a water table located below 5 m depth.
iii) Increases in I_{gT} appear to be offset by increases in ET_{gT} above 5 m depth.
iv) Accretion still occurs to deep water tables at a more or less constant rate below 5 m to a depth of 8 m: which is about the deepest position for the water table at Ardmore.
v) The constant rate in iv) is surprisingly similar to the I_{gnet} occurring above 5 m.
vi) The minimum depth on the plot represents the mean minimum depth to the water table for the Ardmore area.

In addition to the above points we can infer a number of other relationships from such a diagram. For example, the horizontal dimensions of the geometric shape described by the shading, must be related to the climate of an area but would also be strongly influenced by the direction of flow in the saturated zone. The shape in a discharge area would presumably be quite different even for the same climatic zone since, in the absence of springs and seeps, losses to ET_g would consume both discharge and accretion. In addition, it would be expected that the net accretion curve would cross to the negative side of the diagram and record a net loss of water from the saturated zone.

The vertical dimensions of the geometric shape in Figure 6.1, on the other hand, must be governed by the hydraulic properties of the geologic materials. Fine-grained materials might restrict both the accretion flux and the depth of root penetration, thereby restricting also evapotranspiration.

The higher rates of both ET_{gT} and I_{gT} in depression sites with shallow water tables reflects the inverse relationship with depth. Since both total evapotranspiration and total accretion increase together, the net accretion remains approximately constant. This can also be seen in a comparison between upland and lowland sites and goes a long way to explaining the mechanism which maintains the water table as a subdued replica of topography. In other words, although more water is able to accrete to the saturated zone in areas with shallow water tables, much of this is rapidly lost again to evapotranspiration; the smaller volumes of water which penetrate to the deeper water tables in upland sites, however, remain permanently in the saturated zone beyond the reach of evapotranspirative mechanisms (see Kovács, 1981 p194, for a more complete discussion with very similar conclusions).

Important note:

The geometric shape given in Figure 6.1 is characteristic of a particular combination of environmental and hydrogeologic parameters found within an area. The preparation of such diagrams may become a useful visual representation which may be used to characterize, and then classify, different hydrogeologic areas.

6.1.2. Spatial variability, event basis

An event is defined, for the purposes of unambiguous comparison between holes, as a period of rain which produces a measurable water level rise in an observation hole. It must be stated that a recharge event does not by definition lead to a water level rise; it may, for example, result only in a decline in the rate of water level fall. The accounting period is then taken between the trough and peak of the rise. The Troxell method of analysis was used to estimate rates for holes with continuous recorders whereas the White method was applied to holes measured by hand. This was because the Troxell method lends itself readily to "spreadsheet" manipulation for continuous records and the White method is better suited to readings taken on a less regular time schedule.

Spatial variability on an event basis is more pronounced than on an annual basis: for example, many of the upland sites responded only to the spring recharge, whereas some of the depression sites responded to almost every rainfall. Statistics for an event in May 1992 are shown in Tables 6.3 and 6.4. Individual holes have been selected for this purpose because the variability between sites within each group is quite high. It is clear from the need to show more details of the site characteristics at the event time-scale, that more variables are involved in controlling the input/output processes than topographic position alone. We have added depth to the water table, discussed above, as well as the dominant surficial deposits at the site. The material in which the water table is fluctuating together with the moisture content are the chief factors controlling the effective specific yield (Sec. 5.1).

The 18 mm rainfall event did not produce a measurable rise in UOH 9 which is one of the more responsive holes among the upland sites. In other settings, however, the event resulted in high net accretion in the range 10 to 40% of precipitation. This is considerably higher than the rate averaged over a water year. Net accretion is expected to be much higher for an event accounting period, because of the intermittent nature of the process. Whereas this event produced widespread measurable responses in the

Table 6.3

Equivalent depths expressed as percentage of precipitation for a May 1992 rainfall event (18 mm).

UOH	W.T. depth m	Material	Date May, 92	Net I_g %	Total I_g %	Total ET_g %
9 Gp1	3.95	Till	N/A	0	0	0
26 Gp2	2.00	Sand*	22 - 26	46.1	64.7	18.6
15 Gp3	2.00	Till	21 - 27	13.9	21.0	7.1
11 Gp4	0.70	Sand*	20 - 25	38.4	67.6	29.2
5 Gp5	1.20	Sand*	21 - 24	29.8	35.6	5.8

* The specific yield used for the deeper sandy sites was 0.2, whereas 0.033 was estimated in UOH 11 using equation 6.1.

Table 6.4

Fluxes averaged over the same event as Table 6.3.

UOH	W.T. depth m	Material	Date May, 92	Net I_g 10^{-9} m/s	Total I_g 10^{-9} m/s	Total ET_g 10^{-9} m/s
9 Gp1	3.95	Till	N/A	0	0	0
26 Gp2	2.00	Sand*	22 - 26	21.6	30.3	8.7
15 Gp3	2.00	Till	21 - 27	4.8	7.3	2.5
11 Gp4	0.70	Sand	20 - 25	16.4	28.9	12.5
5 Gp5	1.20	Sand*	21 - 24	20.7	24.7	4.0

lowland sites, events of a similar size occurring later in the summer did not. The reason for this is thought to be a combination of higher antecedent moisture content and a shallower water table position in the early season.

The correlation between depth and total accretion discussed earlier is not as clear in the sandy sites. Sandy soils may promote deeply penetrating root systems which give relatively constant rates of ET_g from a wide range of depths. In general, however, the water level rises in UOH 5 and 26 are not as well defined as for the other holes and their estimates are probably less accurate.

It is interesting to note from the dates shown in the tables that water levels started to respond earlier in UOH 15 than in 26, even though the water levels were at the same depth. The material of the former site is till as compared to sand in the latter and therefore the opposite result might have been expected. A combination of preferred pathways in close contact with a thick capillary fringe in the till is the mechanism suggested here to explain the apparent discrepancy. We do not have the soil moisture data to support this, but open fractures were often observed in soil pits extending from the surface to zones with higher moisture contents. These fractures may have a tendency to close or fill-in during rainfall events as the moisture content rises and thus the event is prolonged in the finer grained deposits as compared to the sandy sites.

6.1.3. Temporal variability, event basis

Temporal variations in recharge occur on many scales. It is already obvious from the data presented above that the average fluxes increase from water year to event based estimates in the same hole. It is also obvious that fluxes vary considerably from year to year but our study period was not long enough to make a reasonable analysis of the variations at this scale.

In attempting to illustrate temporal variation at the event scale we have to point out that this is difficult to accomplish with any degree of quantitative accuracy using the hydrograph technique alone. To exemplify this aspect we show results of event based recharge for UOH 15 and 11 in late July 1992 (Tables 6.5 and 6.6). These results may be compared directly with those of Table 6.3 and 6.4 noting that the rainfall in July was greater by approximately 40 % and that the water table has dropped from the May position by 0.6 m and 0.8 m in UOH 15 and 11 respectively. In addition, a period of roughly one month with only very light rainfall preceded the July event, whereas the soil moisture conditions in May were much wetter.

Table 6.5

Equivalent depths expressed as a percentage of precipitation for a July 1992 rainfall event (25 mm)

UOH	W.T. depth m	S_y	Date July	Net I_g %	Total I_g %	Total ET_g %
15 Gp3	2.60	0.045	26 - 31	0.6	7.7	7.1
11 Gp4	1.50	0.17	25 - 31	44.9	69.5	24.6
		0.10	25 - 31	26.5	40.9	14.4

Table 6.6

Fluxes average over the same event as Table 6.5.

UOH	W.T. depth m	S_y	Date July	Net I_g 10^{-9} m/s	Total I_g 10^{-9} m/s	Total ET_g 10^{-9} m/s
15	2.60	0.045	26 - 31	0.4	4.5	4.1
11a	1.50	0.17	25 - 31	24.0	37.1	13.1
11b		0.10	25 - 31	14.1	21.8	7.7

For UOH 15 we in fact assumed that soil moisture conditions near the relatively deep water table had not changed significantly, and no adjustment was made to the specific yield. In UOH 11, on the other hand, we present two sets of results (11a and 11b) that assume much drier soil conditions in July. In the first set, we assumed that the effective specific yield increased from 0.033 to 0.17. Equation 5.13 was used to estimate the new effective specific yield, but it is considered to be an over estimate since we could not account for changes in soil moisture storage. In the second set of figures, therefore, we chose an arbitrary but lower value of 0.1 for the effective specific yield.

The results in UOH 15 indicate that the drop in the water level and the drier soil conditions near the surface, have lowered substantially the percentage of rainfall accreting at the water table. In UOH 11, on the other hand, two different conclusions can be reached depending on which effective specific yield value is considered to be correct. From the second set of results using the lower value, our conclusions would be the same as for UOH 15. The higher choice of effective specific yield, however, resulted in flux values very similar to the May event.

We know that the high specific yield value is probably inaccurate and we can say that the result using 0.1 looks intuitively more correct, but we cannot say which is closer to reality. It seems likely from the diurnal estimates below, however, that the true value of effective specific yield is closer to 0.1 than 0.17. We also note that the ratio of each input and output estimated did not change with the different choices of an effective specific yield. The ratio of I_{gT} to ET_{gT} increased slightly between May and July and this is the expected result for a declining water table (Fig. 6.1). A valuable feature of the Troxell method is that it could readily incorporate independently measured parameters, such as effective specific yield from soil moisture measurements, if they were available.

6.1.4. Temporal variability, diurnal basis

Results of temporal variation are shown in this section by analyses of diurnal fluctuations in UOH 11 and 12 for three different periods in the summer (Table 6.7 and 6.8).

The night time fall in UOH 11, resulting from groundwater recharge, is not measurable over such a short period and was therefore treated as negligible. In UOH 12 on the other hand, the night time rise is rapid and easily measured. The increasing discharge flux through the summer, demonstrated by Table 6.8, suggests that hydraulic gradients increase as the cone of depression below the woodland deepens in response to increasing rates of transpiration (Sec. 4.1).

Both holes show an increase in evapotranspiration through the summer. The increase in average ET_g flux in UOH 11 is much less than in UOH 12, however. This suggests an early build-up in the transpiration rate from a relatively shallow water table at the former site. In this estimate we used the 0.1 effective specific yield value in late summer at UOH 11. We reasoned that the transpiration rates were unlikely to show a large increase at this site because the sedges and grasses were cropped by cows all through the summer. By comparison, the transpiration at site UOH 12 (woodland) in early June is only between one third and one quarter that in July.

Table 6.7

UOH 11. ET_g fluxes estimated from diurnal fluctuations.

W.T. depth (m)	Date	S_y	Recharge/ discharge 10^{-9} m/s	ET_g Av. daily 10^{-9} m/s	ET_g Max. daily 10^{-9} m/s
0.5	May. 93 13 - 16	0.033	0	4.8	27.5
1.0	June. 93 2 - 8	0.033	0	9.6	27.5
1.7	Aug., 92 10 - 17	0.1	0	11.!	55.6

Table 6.8

UOH 12. ET_g fluxes estimated from diurnal fluctuations.

W.T. depth (m)	Date	S_y	Discharge 10^{-9} m/s	ET_g Av. daily 10^{-9} m/s	ET_g Max. daily 10^{-9} m/s
2.8	Jun., 92 12 - 16	0.03	0.5	25.0	44.4
3.7	July, 92 18 - 20	0.033	54.2	88.5	182.0
4.1	Aug., 92 15 - 18	0.1	66.7	96.5	172.0

6.1.5. Estimates of areal recharge

An estimate of the total annual volume of recharge was attempted for the Ardmore area. The arithmetic mean flux, calculated in groups of holes, were distributed over sub-regions of the study area based on the depth to the water table, soil types, surface mapping and surficial geology (Figure 6.2). The sum of the products of each area multiplied by the respective recharge fluxes was approximately 8.7 x 10^6 m^3 per year. The same estimate using net accretion for the 1992/93 water year was 5.2 x 10^6 m^3 per year.

6.1.6. Summary of hydrograph estimates

It became apparent from parameter and flux estimates that distinct quantitative boundaries did not exist between the subgroups. The differences between individual sites were almost as great as the differences between subgroups. This should not be surprising since there were no sharp breaks in the landscape. The rather smooth changes in water table responses that resulted from topography were, however, broken by quite rapid changes in surficial deposits. Changes in lithology of the surficial deposits created differences in water level responses to precipitation as great as those that resulted from changes in topographic position. Depression sites appeared to show less variation in the fast and slow subgroups than in the medium subgroup, but over all they were more variable than non-depression sites. The differences appeared to be closely related to permeability in the subsurface, and of course to the local groundwater flow direction, although our monitoring program did not allow us to say much about this aspect. It would be a simple conclusion to say that we did not have enough holes to characterize the area and whereas this may be true, it may also be true that more observations would simply yield more subgroups.

We believe that the preliminary flux estimates made in this study indicate the ranges occurring within the Ardmore area. The detailed distribution of the fluxes is beyond the scope of this study. To illustrate this point, the estimate of total areal recharge using the arithmetic mean recharge flux for all the observation holes is approximately 8.0 x 10^6 m^3/y which is within the range of error in our estimates using regionalized flux estimates.

The arithmetic means for recharge and net accretion fluxes in the university observation holes were approximately 1.1 x 10^{-9} m/s and 5.7 x 10^{-10} m/s respectively. The recharge flux represented 9 % of the mean annual precipitation for the last 16 years (400 mm), whereas the net accretion flux was 5 % of the precipitation for the 1992/93 water year (355 mm).

On an individual site basis, recharge was estimated to vary between 3 and 18 % of the mean annual precipitation. Net accretion varied between a net loss at some sites and about 12 % of the precipitation for the 1992/93 water year.

Total accretion and total evapotranspiration varied widely between sites and between individual events and seasons. At sites with shallow water tables and permeable soils, total accretion was estimated to be as much as 70% of the rainfall during an event. Groundwater losses at these sites to evapotranspiration were usually also high, however, with day time fluxes estimated as high as 2 x 10^{-7} m/s. Some sites with deep water tables, on the other hand, did not appear to experience any groundwater loss to evapotranspiration so that total and net accretion became equivalent.

Temporal variability in accretion was related to the season and the depth to the water table. ET_g increased to a maximum in late July and early August. The increasing rates were sustained mainly by losses from groundwater storage in the summer of 1992 and the only net accretion occurred in the spring and early summer. Losses to ET_g in 1993, however, were largely replenished by summer precipitation and net accretion occurred in July and August at many sites. The processes of accretion and ET_g appeared to be dormant during periods of ground frost and the rates of each were negligible.

Estimates of rates of accretion and evapotranspiration are dependent in part on the time scale applied, whereas recharge rates appear to be more constant in the Ardmore flow regime. Maximum rates of accretion and evapotranspiration must be estimated on the event and hourly time scales respectively at sites with shallow water tables.

6.2. Flow system estimates

Recharge is sustained by accretion, but recharge fluxes are controlled by hydraulic parameters within the saturated zone. Recharge flux estimates are thus equally well derived from knowledge of the flow system. The other fluxes (I_g and ET_g) at the water table, however, cannot be measured directly with this approach.

Total discharge to the Beaver River Valley was estimated to be between 4 and 16 x 10^6 m^3/y. The estimate was found to be most sensitive to the choice of hydraulic conductivity (K) in the aquifers. The range in estimate was created by assuming that the error in the choice of K in the aquifers was between one half and two times the original value chosen.

Total leakage through the poorly permeable bedrock was estimated, using the same error approximation, to be between 0.2 and 1.6 x 10^6 m^3/y. The average leakage flux through the bedrock was estimated to range between 2.5 x 10^{-11} m/s and 2.5 x 10^{-10} m/s. Reporting the Valley discharge as a flux is less meaningful because the flux varies between units and with vertical position in each cross-section. The maximum horizontal flux estimated was 1.7 x 10^{-6} m/s in the Muriel Lake formation at the base of cross-section B3.

The estimated rural population of the Ardmore study area in 1992 was 600 based on the number of dwellings multiplied by 2. The average number of cattle was estimated at 1000 based on local observations and verbal communications with farmers and government officials. Using an average of 500 l/day and 70 l/day for humans and cattle respectively, the total annual water withdrawal was estimated at 0.1 x 10^6 m^3/y. The water use in Ardmore and Fort Kent was not included in the total because the water source for both hamlets is a single supply well completed in the Muriel Lake aquifer in Bronson Lake Channel. Flow in the channel is across the south west corner of the study area and was not believed to affect the water balance substantially (Sec 4.2).

The total water losses from the area were thus estimated to range between approximately 4.5 and 18.0 x 10^6 m^3/y. On the assumption that these losses were sustained by recharge within the study area, they were also an estimate of the average annual total recharge. It is obvious even from these approximations that discharge to the Beaver River Valley is the dominant sink. The combined volumes from bedrock leakage and withdrawals are less than the possible errors in its estimate.

We assumed that over long time periods the conceptual flow system beneath the area was in steady state and that recharge and discharge were approximately equal (Sec. 1.4). An average recharge flux can therefore be estimated by dividing the total discharge, leakage and water withdrawals by the recharge area. The resulting recharge flux at Ardmore was 1.5 x 10^{-9} m/s which can be compared to the mean recharge flux estimated by the hydrograph approach of 1.1 x 10^{-9} m/s.

6.3. Pump test estimates

The estimated average induced recharge flux beneath the area influenced by pumping was approximately 1.3 x 10^{-10} m/s. The estimate was calculated using Equation 5.3.1 and the following input values:

area of influence = $\pi(4500)^2 \approx 6.4 \times 10^7\ m^2$.
total average combined pumping rate from both source wells 684.4 m^3/d.

The recharge flux was then calculated as,

$684.4/(6.4 \times 10^7) \approx 1.1 \times 10^{-5}$ m/d $\approx 1.3 \times 10^{-10}$ m/s.

The natural recharge flux at the water table, calculated from the winter recession of local hydrographs, was approximately 1.3×10^{-9} m/s which is an order of magnitude greater than the induced rates estimated across the till aquitard. This result was unexpected because the conceptual model suggested that vertical fluxes at the water table should be similar in magnitude to vertical fluxes deeper in the flow regime below the Alexander Lowlands (see Fig. 4.26). The induced rates were actually expected to be greater than the natural rates since they were the result of artificially increased hydraulic gradients.

The significance of this apparent discrepancy cannot be evaluated because of the uncertainties involved with both the pumping test and the hydrograph approaches. The conceptual model is undoubtedly oversimplified and variations of this magnitude in the fluxes with changes in depth are quite possible: especially when one considers that only the vertical component was measured. Fluxes in the shallower portions of flow regimes are usually higher than in the deeper portions as a result of higher hydraulic head gradients, often generated in local flow systems. We did not find evidence, however, of permanent local discharge in the Ardmore area; but temporary discharge in local systems during the early season, when the water table is elevated, remains a possibility. High permeability layers with large horizontal flow components can also account for variable vertical fluxes with depth, but our conceptual model may not be detailed enough to demonstrate this.

Finally, as discussed earlier, the frost effect could have contributed to an overestimate of fluxes at the water table by the hydrograph approach. We do not believe, however, that the resulting error could be as high as one order of magnitude

6.4. Comparison of approaches

The areal estimate of annual recharge using flux values generated with the hydrograph approach falls within the range estimated by the flow system approach. The error estimate of between one half and two times the original value chosen, could equally apply to the choice of hydraulic conductivity or specific yield. Providing the errors in both estimates are in the same direction the results remain close, if the errors are in opposite directions then the estimates remain almost within one order of magnitude.

The net accretion for the area is estimated to range between 2.3 and 9.4 x 10^6 m^3/y. The total water withdrawals estimated in the Ardmore area are on the order of 0.1 x 10^6 m^3/y or a maximum of about 6% of the net accretion. The effect on the average position of the water table due to local groundwater withdrawals is therefore likely to be small. The precipitation for the 1992/93 water year represented about 89% of the average annual value. The net accretion for the same period, however, was about 50% of the recharge rate. It appears that accretion, although proportional to precipitation, declines at a much faster rate. Therefore changes in precipitation have a large effect on shallow water levels.

The results of the pump test approach are inconclusive. They suggest that recharge fluxes are higher in shallow rather than in deep parts of the flow regime. This seems intuitively correct, and if the approach was applied to tests designed specifically for the purpose of estimating recharge it could supply useful information. The question of how recharge is distributed to deeper portions of a flow regime is important and methods should be devised to answer it.

Only the hydrograph approach allows estimates to be made of total accretion and evapotranspiration, and therefore of the essentially unsaturated fluxes. The others deal exclusively with fluxes within the saturated zone, i.e. recharge. They cannot estimate the rates of water made available at the water table and lost again to other processes.

The estimates made by the hydrograph approach are very sensitive to the choice of a value of effective specific yield. Efforts to measure this parameter on a continuous basis should be made. The temporal and spatial variation in specific yield could be easily combined with the Troxell method of analysis in the application of a hybrid approach as recommended by Sophocleous (1991).

The question of the magnitude of fluxes above and away from the water table in the unsaturated zone under frozen conditions remains unanswered. We were forced to make the assumption that they were negligible from relatively deep water tables in poorly permeable materials. The sandy sites with shallow water tables such as UOH 5 and 17 may be the most likely to have overestimated recharge fluxes due to winter-time losses to a developing frost wedge.

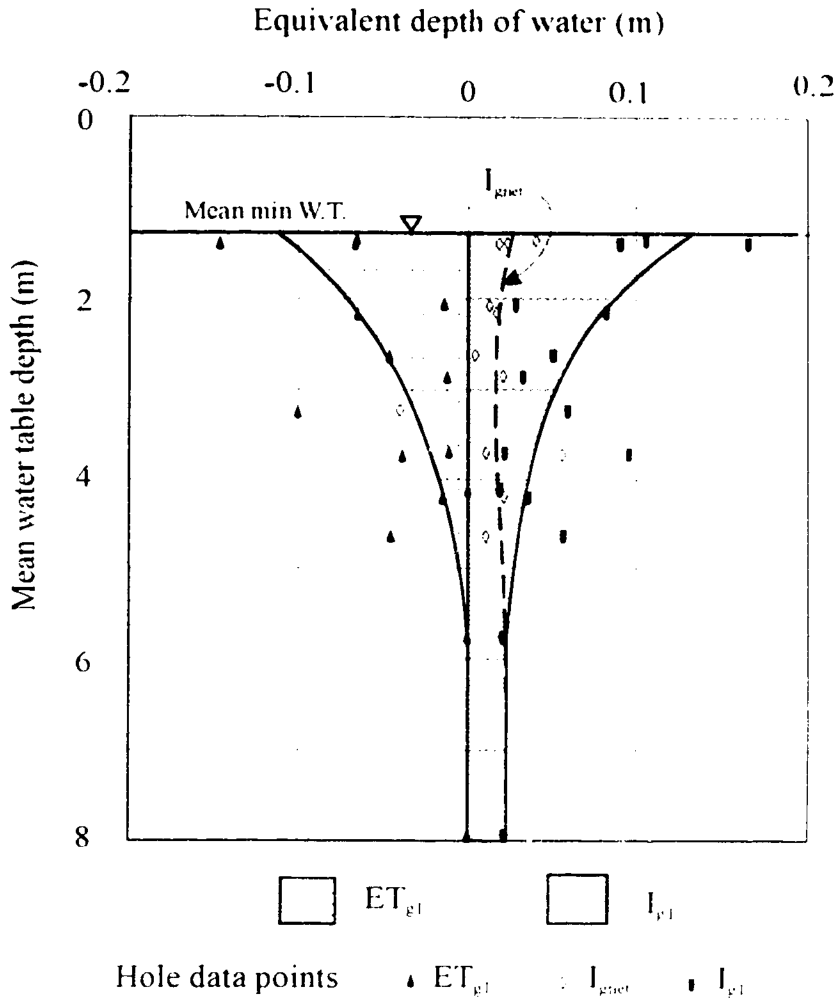


Figure 6.1. Plot of points in each hole for which total ET_g and I_g have been estimated by the Troxell method. Shaded areas represent general trends.

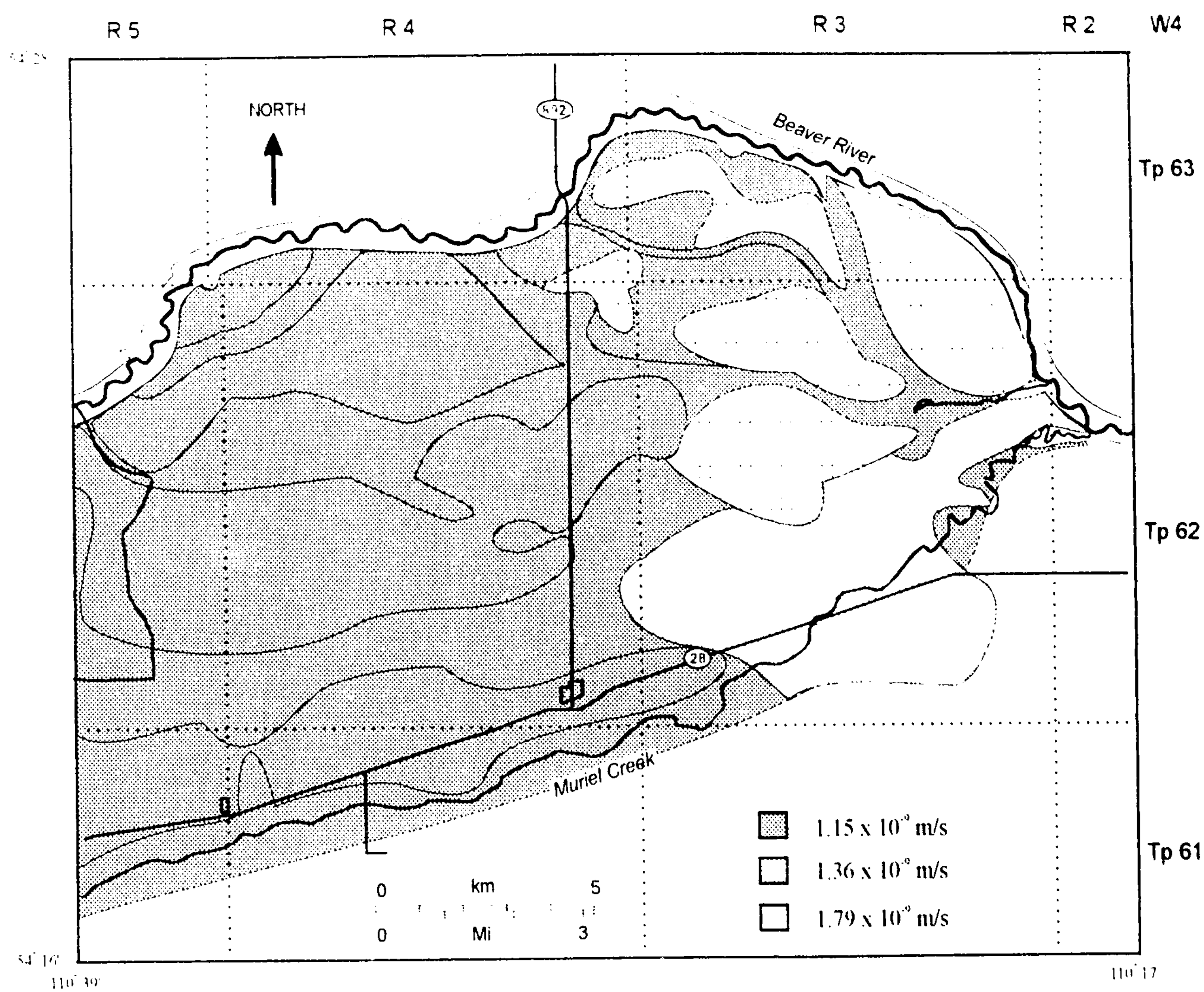


Figure 6.2. Areal distribution of recharge fluxes.

7. Summary and Conclusions

The chapter comprises three sections; Section 7.1 lists in point form the main project findings, Section 7.2 gives a separate paragraph to each conclusion and is itself concluded by a brief discussion of two major implications for the study, and Section 7.3 lists the recommendations which result from the work and its main scientific contributions.

7.1. Summary

The main project findings are summarized as follows:

i). The study area is underlain by a sequence of till and outwash deposits overlying poorly permeable bedrock. The tills function hydraulically as leaky aquitards and the sandy outwash deposits as aquifers. There is hydraulic continuity throughout the Quaternary sequence in spite of the presence of till.

ii). The fluid potential gradients beneath the area are dominantly upwards and consequently, groundwater flow in the till aquitards is dominantly downwards. Flow in the more permeable aquifers, however, is thought to have large horizontal components under much of the study area.

iii). The main source of recharge is precipitation within the area and net subsurface inflows are thought to be negligible.

iv). Discharge from the area is largely to the Beaver River Valley although the potential exists for a small volume of leakage, downwards, through the poorly permeable bedrock.

v). We have identified that the position of the water table is governed by a dynamic equilibrium of three fundamental fluxes, accretion, recharge and evapotranspiration at the water table. Infiltrated water which has percolated to the water table is called here accretion and is defined as follows; the process or rate of addition of water to the water table from the unsaturated zone. The water added to the water table is divided into that which remains as part of the saturated zone and that which is lost to the combined processes of evapotranspiration. The water which becomes a part of the saturated zone is called here recharge and is defined as, the process or rate of flow of water below and away from the water table within the saturated zone.

vi). Hydrographs in the area show that accretion from precipitation can occur everywhere even to relatively deep water tables.

vii). The hydrographs can be broadly classified using multiple parameters into groups of depression and non-depression types.

viii). The groupings reflect the spatial and temporal variability of accretion and recharge.

ix). Total accretion fluxes show very large temporal variations and spatial variations which are strongly influenced by the depth to the water table.

x). Recharge rates appear to be constant over time but show spatial variations which are most pronounced between depression sites.

xi). Accretion does not normally occur during winter months.

7.2. Conclusions

Recharge is sustained in the Ardmore area by intermittent accretion originating at the land surface as precipitation and as captured runoff or meltwater.

Accretion that is added permanently to the saturated zone, and thus contributes to recharge, is known here as net accretion. It may be estimated as the difference between total accretion and total evapotranspiration at the water table. Over long time periods net accretion and recharge rates must be equal in areas where the mean position (the average for all seasons averaged over a number of years) of the water table is relatively constant.

Provided accretion events continue to take place, the recharge rate is governed by the fluid potential field and permeability of materials in the subsurface. Consideration of the saturated flow system, therefore, gives the best estimates of long term recharge rates because the rate and direction of flow in the saturated zone (at the water table) controls the rate of entry of water made available at the water table. For this reason it is inferred by this study that water which accretes at the water table in a discharge area cannot become net accretion.

Accretion shows great variation both temporally and spatially because of the wide range of variables which affect the process. Chief among these at Ardmore appear to be the availability of water at the land surface, depth to the water table and soil moisture conditions in the unsaturated zone.

Hydrograph interpretation is best accomplished by examining the balance of fluxes at the water table. When this is done, it is clear that recharge is a two component process involving accretion at the water table and transport within the saturated zone.

Hydrographs can be used to make preliminary estimates of recharge, accretion and evapotranspiration from the water table over different time scales. The estimates probably become less accurate as the time period becomes shorter.

The most important additional parameter required for the hydrograph approach is the effective specific yield. For this reason, the late part of recession curves are generally more useful than water level rises because they represent more complete draining.

Continuous hydrographs are essential for separating natural recharge processes from other effects, such as atmospheric pressure changes or pumping.

The recharge flux estimated by the hydrograph approach is the vertical component of flow or rate per unit surface area.

The natural saturated flow rate may not be significant for withdrawal calculations (Bredehoeft et al., 1982), but it seems intuitively reasonable that accretion rates must be. A new potential field imposed artificially by pumping has little chance of increasing accretion. If the water table is lowered, for example, with the intention to capture accretion otherwise lost to evapotranspiration, we can show that the volume of accretion available to deeper water tables naturally decreases. In addition, the ultimate source of accretion in the Ardmore area, namely precipitation, remains beyond the influence of pumping.

Accretion rates are perhaps most important for estimates of transport in the unsaturated zone and are therefore most applicable to surface contamination problems and consideration of soil moisture budgets for agricultural applications.

Recharge rates on the other hand, are directly related to transport in the saturated zone and are more applicable, for example, to studies of deep waste disposal, certain ore deposits and petroleum migration.

The estimates of fluxes at Ardmore are in good agreement with those of Keller et al. (1988) at the Dalmeny site near Saskatoon, Saskatchewan and Freeze (1969b) in similar hydrogeologic environments.

Practical implications:

Water management in the Ardmore area must take into consideration the fact that, with the exception of pumping directly from aquifers in the Bronson Lake Channel, groundwater water supplies elsewhere beneath the area are sustained directly by precipitation. In addition, the aquifers are essentially unconfined, which implies that lowering water levels in the aquifers will eventually result in a lowering of the water table. This phenomenon will be particularly noticeable in years with lower than average precipitation. Lowering the water table will tend to reduce soil moisture and the water levels in lakes and wetlands. Projects which require high rates of groundwater withdrawals should be encouraged to obtain groundwater from the discharge area beneath the Beaver River Valley. This is because more of the naturally discharging, and thus otherwise lost, water is captured by wells located in discharge areas than by wells in recharge areas (Ophori and Tóth, 1990). In addition, pumping in these areas is able to make use of natural accretion which may now remain as net accretion due to the modified flow regime. The effects of groundwater withdrawals on the local water table will be less in the valley than in the surrounding recharge areas, and the effects on the total river discharge may be negligible.

The rapid response of the water table to precipitation events is a clear indication of the good communication between the surface and the saturated zone. This is an important consideration when planning land use practices in the area. Large concentrations of livestock (feedlots), inappropriate application of fertilizers and herbicides, landfill sites and water treatment facilities, all have a great potential to pollute the groundwater in this area. In each of these examples the accretion flux rather than the recharge flux, at the water table, is of most interest in predicting contaminant transport. We have estimated maximum accretion fluxes as high as 1.3×10^{-5} m/s which are equivalent to more than one meter per day.

7.3 Recommendations and scientific contribution

Flux estimates using the hydrograph approach should be tested against estimates made by other approaches in controlled experiments. An attempt should be made to establish the conditions under which the hydrograph approach can be useful (water table depth, frequency of precipitation events, duration of ground frost and so on). The Dalmeny test site in Saskatchewan may be one possible place for comparative experiments because the saturated vertical fluxes have already been accurately measured (Keller et al., 1988).

Efforts to estimate water transfers in winter from the saturated zone to the frost wedge should be continued. The differences in hydrograph estimates and the known saturated fluxes during winter, for example, should be equal to the flux of water lost to a developing frost wedge.

Scientific contribution:

The most important scientific contribution of the study is a clarification of the concepts and processes of groundwater recharge. To assist in the clarification and understanding of the recharge process we have suggested terminologies, with associated definitions, for two fundamental fluxes at the water table; namely accretion and recharge.

The development of approaches for preliminary flux estimates at the water table using water table hydrographs should be useful in other geographic areas and may have application to other fields of research: such as hydrology, forestry and botanical research especially of wetlands. By increasing the range of application of two hydrograph analysis techniques, originally designed to estimate evapotranspiration from diurnal fluctuations, we have extended the scientific contribution of earlier researchers.

The study represents the first attempt at estimating natural rates of groundwater recharge in the Cold Lake-Beaver River Basin; an area heavily dependent on groundwater supplies for which appropriate uses are being currently debated. In conducting the study we have obtained a unique data base of actual field observations. The study has also resulted in a detailed hydrogeological model of the Ardmore area.

Finally the conceptual model of the recharge processes was developed with the ideas from two fundamentally different approaches to the study of groundwater; the civil engineering approach of Kovács and the traditional hydrogeological approach of Tóth. The linking of fundamental concepts of groundwater flow (Tóth 1962 and 1963) with the concepts of a water balance at the water table (Kovács, 1959b) is a progressive step in demonstrating the relationships between recharge processes active in the unsaturated zone with those active in the saturated zone.

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Appendix A

Sample Calculations:

Example of preliminary estimates of fluxes using the hydrograph approach. Estimates made in UOH 9.

With reference to figure A1 the following estimates can be made by the hydrograph approach.

1) Recharge from the slope of the winter recession,

$R = S_y r = 0.045*0.0022 = 9.9 \times 10^{-5}$ m/d $= 1.146 \times 10^{-9}$ m/s

where S_y is the specific yield of the material of the winter recession and r is the slope, $\Delta h/\Delta t = 0.2/91$ m/d.

2) Net Accretion (I_{gnet}) by the White method,

Number of days in the water year, Y, = 322 d
Specific yield for winter recession, S_y, = 0.045
Slope of the winter recession, r, = 0.2/91 m/d
Net difference in W.T. beginning and end of year, s, = 0.22 m

$I_{gnet} = S_y(Yr - s) = 0.045((322*0.0022) - 0.22) =$
$= 0.022$ m/yr_w $= 7.91 \times 10^{-10}$ m/s

where yr_w denotes a water year.

Troxell methods:

3) Total Accretion, I_{gT}

The periods of accretion are selected from the slope graph in figure A1. The differences in slopes, length of each interval and the depth of accretion are compiled in a table (table A1). The total accretion flux is calculated from the product of the total depth and a specific yield, divided by the length of the period, in this case a full water year. Fluxes can also be estimated over any shorter period. A constant or specific yield may be applied to the whole period, or varied for each time interval.

Table A1, worksheet for estimating I_{gT}

Period	difference in slopes I_g m/d	Number of Days d	depth m
1	0.02849	12	0.34186
2	0.00776	40	0.30652
3	0.00182	18	0.03276
6	0.00159	20	0.0318
7	0.00128	73	0.09344
Total		163	**0.806382**

Applying a constant Sy because there is no material change and a deep water table, the total accretion flux estimated is,

$I_{gT} = S_y I_g H/Y = 0.045*0.806382/1 = 0.0363\ m/yr_w = 1.3 \times 10^{-9}\ m/s$

where $I_g H$ is the total depth of accretion and Y is the water year.

4) <u>Total evapotranspiration</u> ET_{gT}.

The same approach is used to estimate evapotranspiration fluxes at the water table.

Table A2, worksheet for estimating ET_{gT}.

Period	difference in slopes ET_g m/d	Number of Days d	depth m
2	0.00031	40	0.012245
3	0.00281	18	0.05058
4	0.00463	46	0.21298
5	0.00137	30	0.0411
6	0.00031	20	0.0062
Total		154	**0.323105**

Constant S_y used as above.

$ET_{g_T} = S_y ET_g H/Y = 0.045*0.323105/1 = 0.0145\ m/yr_w = 5.2 \times 10^{-10}\ m/s$

where ET_gH is the total depth of evapotranspiration.

5) <u>Net accretion</u> is the difference between total accretion and total ET_g.

$I_{g_{net}} = I_{g_T} - ET_{g_T} = 0.0363 - 0.0145 = 0.0218\ m/yr_w = 7.8 \times 10^{-10}\ m/s.$

The result is quite comparable to the estimate by the White method. This case, with no variation in S_y through the year, was normal at Ardmore.

Aquifer parameter calculations from the 1974 Ardmore Thermal Project pumping test data.

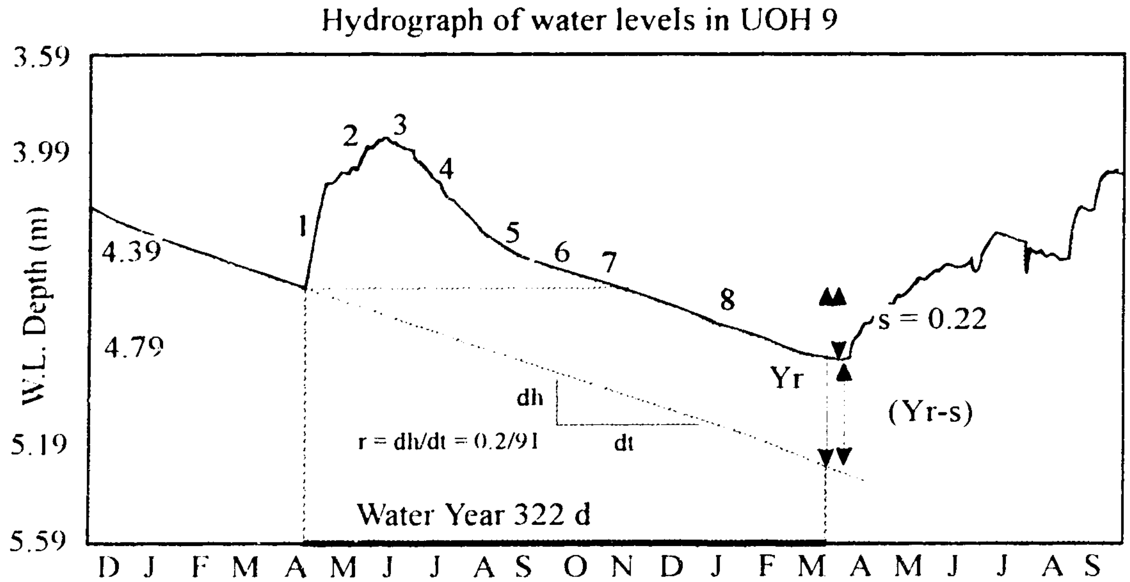
Table A3

Transmissivity:

Well	Method	Equation	Calculation	Result
CWW1	Jacob straight line	$T = \frac{2.3Q}{4\pi\Delta s}$	$= \frac{2.3(505.37)}{4\pi(0.36)}$	256.94 m^2/d
CWW1	Jacob recovery		$= \frac{2.3(505.37)}{4\pi(0.32)}$	289.05 m^2/d
CWW2	Jacob		$= \frac{2.3(606.18)}{4\pi(0.45)}$	246.55 m^2/d
OWW1	Theis	$T = \frac{Q}{4\pi s}W(u)$	$= \frac{606.18}{4\pi(0.07)}(0.5)$	344.56 m^2/d
	Jacob not applicable	$t \geq \frac{r^2 S}{4T(0.01)}$	$= \frac{292^2(3.54 \times 10^{-4})}{4(344.56)0.01}$	2.19 d 3153.6 min

Storativity:

OWW1	Theis	$S = \frac{4Tt}{r^2}u$	$= \frac{4(344.56)0.208}{(292)^2}(0.1)$	$= 3.54 \times 10^{-4}$



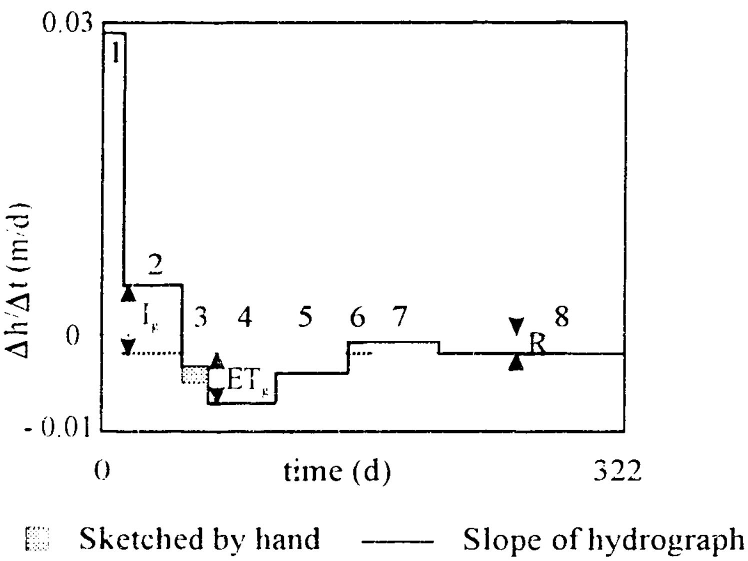


Figure A1. Flux estimates by hydrograph analyses (see text for details).

Appendix B, Table 1, Data points for geologic interpretations.

TWP	RG	SC	LS	WELL ID	UTM E	UTM N	ELEV.	YR	DEPTH	Bedrk Elv	EMP Elv	Brn Lk Elv	Mur Lk Elv	Bvlle Elv	Eth Lk Elv	Mar Ck Elv	Snd Rv Elv.
					mE	mN	m.a.s.l	19'	m	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l
61	3	30	1	WW602	37020.00	16610.00	[illegible]	89	27.13							543.42	544.03
61	3	31	3	TH657	35990.60	18176.67	542.00	91	27.74	496.89							
61	3	31	3	WW658	36049.07	18179.52	542.00	79	24.08							546.77	547.08
61	3	31	3	WW661	36036.47	18154.09	542.00	84	25.60								
61	3	33	5	WW601	38820.00	18750.00	548.00	86	26.21	512.03				519.34	520.56		
61	4	25	2	WW735	34865.00	16675.00	553.00	79	24.08								
61	4	25	5	WW736	34005.00	17180.00	552.00		0.00								
61	4	26	13	WW732	32315.00	17925.00	550.50		0.00								
61	4	26	13	WW733	32320.00	17925.00	550.50	78	23.77	495.33							
61	4	26	16	WW734	33805.00	17920.00	549.00	78	23.77							543.42	544.64
61	4	28	2	WW727	30150.00	16780.00	555.00	80	24.38							542.50	545.86
61	4	28	3	WW729	29280.00	16700.00	557.00	78	23.77	465.56	472.09	485.68	507.01	517.81	520.25	543.11	544.03
61	4	28	3	WW730	29720.00	16710.00	557.00	79	24.08		472.09	486.29	506.40	514.16	521.17	543.42	545.25
61	4	28	3	WW731	29760.00	16720.00	557.00	78	23.77								
61	4	28	4	WECO	29253.98	16785.70	555.00	75	22.86	469.05	475.14	486.12	508.67	522.08	525.13	541.59	544.64
61	4	28	5	WW729	29100.00	16940.00	553.00	74	22.56							545.25	547.08
61	4	28	12	WW726	29105.00	17435.00	551.00	92	28.04			488.52	508.33				
61	4	29	1	OB725	28695.00	17525.00	550.00	78	23.77	469.23	473.50	485.69	507.02				
61	4	30	12	FK SCHL	25797.09	17851.25	550.00		0.00	507.94			509.77	526.23	526.84	539.64	540.55
61	4	30	12	WW702	25920.00	17550.00	549.50		0.00				509.88	522.07	522.68		
61	4	30	14	WW724	26305.00	17990.00	550.00	79	24.08				507.94	525.01	527.44	541.47	543.29
61	4	31	2	WW719	26770.00	18125.00	550.50		0.00								
61	4	31	2	WW720	26585.00	18135.00	550.00		0.00								
61	4	31	2	TH721	26922.00	18085.00	550.50	78	23.77			482.83	501.43	510.88	516.06		
61	4	31	12	WW701	25830.00	18915.00	554.00	88	26.82				505.23			540.28	543.33
61	4	31	12	WW722	25831.00	18915.00	554.00		0.00								
61	4	31	12	WW723	25830.00	18916.00	554.00	81	24.69							539.67	540.28
61	4	32	1	WW716	23925.00	18235.00	550.00		0.00								
61	4	32	1	WW717	28920.00	18230.00	550.00	35	10.67								
61	4	32	1	WW718	2?695.00	18335.00	551.00	85	25.91	467.18	470.84	486.99	507.11			533.02	533.93
61	4	32	8	TH3	28931.97	18731.24	554.00	75	22.86	465.61	476.58	487.55	504.32	524.13	525.65	540.28	541.20
61	4	33	1	WW708	30444.00	18455.00	552.00	64	19.51								
61	4	33	1	WW709	30445.00	18455.00	552.00	74	22.56								
61	4	[illegible]	1	WW710	30445.00	18456.00	552.00	78	23.77								
61	4	[illegible]	1	WW711	30444.00	18456.00	552.00	84	25.60				505.06	524.26	524.57		
61	4	33	4	WW715	29050.00	18065.00	549.00	85	25.91	468.23	476.76	488.95	505.41				
61	4	33	4	WW715a	29050.00	18065.00	549.00	85	25.91	468.23	477.37	489.87	506.02				
61	4	33	9	WW714	30500.00	18955.00	558.00	86	26.21	505.57			506.18	526.00	526.61		
61	4	34	9	UOH 11	32196.83	18960.48	548.50	91	7.60								
61	4	34	12	WW712	30725.00	19145.00	559.00	86	26.21								

Appendix B, Table 1, Data points for geologic interpretations.

TWP	RG	SC	LS	WELL ID	UTM E mE	UTM N mN	ELEV m.a.s.l	YR 19'	DEPTH m	Bedrk Elv m.a.s.l	EMP Elv m.a.s.l	Brn Lk Elv m.a.s.l	Mur Lk Elv m.a.s.l	Bvlle Elv m.a.s.l	Eth Lk Elv m.a.s.l	Mar Ck Elv m.a.s.l	Snd Rv Elv m.a.s.l
61	4	34	13	WW713	30695.00	19465.00	562.00	82	24.99								
61	4	34	16	UOH 10	32207.43	19373.86	553.33	91	11.30								
61	4	34	15	WW707	31465.00	19240.00	557.00	81	24.69								
61	4	35	4	WW705	32360.00	18365.00	549.00	35	10.67								
61	4	35	4	WW737	32360.00	18370.00	549.00	86	26.21								
61	4	35	4	TH738	32360.00	18375.00	549.00	92	28.04	0.00			509.02	528.27	530.10	541.38	542.29
61	4	35	4	WW739	32362.00	18375.00	547.42	92	28.04					530.35	530.66	539.80	540.10
61	4	35	5	WW706	32450.00	18550.00	548.00	67	20.42								
61	4	36	7	WW703	34730.00	18515.00	545.00	85	25.91				499.28	522.14	523.66	540.43	541.95
61	4	36	7	WW704	34731.00	18516.00	545.00	88	26.82					529.76	530.37		
61	5	23	8	WW1	24023.31	15468.12	554.00	77	23.47			508.28	511.33	522.00	523.52		
61	5	23	8	WW2	24023.21	15452.88	554.00	87	26.52								
61	5	23	12	WW3	22593.28	15820.33	549.00	83	25.30							531.93	532.24
61	5	23	16	TH15	24100.00	16360.00	549.00	84	25.60	509.38				523.40	524.01		
61	5	24	4	WW4	24474.42	14854.02	554.00	79	24.08			497.31	500.36	518.34	522.30		
61	5	25	8	WW5	25701.18	16843.81	549.00	83	25.30			511.20	512.73	519.43	520.65	537.42	537.72
61	5	25	16	WW19	25715.00	17885.00	549.50	55	16.76				509.88				
61	5	25	16	TH1	25682.04	17760.43	549.00	75	22.86	504.19		507.24	509.38	524.62	526.44	536.50	537.42
61	5	26	4	WW6	22595.44	16506.08	549.00	78	23.77								
61	5	26	4	WW16	22595.44	16500.00	549.00	89	27.13			507.85	510.90	520.96	521.57		
61	5	26	12	WW7	22580.76	17331.60	553.00	48	14.63								
61	5	26	12	TH17	22585.00	17331.60	553.00	89	27.13	0.00				524.35	526.18	538.67	539.28
61	5	26	12	WW18	22585.00	17330.00	553.00	89	27.13	0.00			510.63	525.87	526.18		
61	5	35	8	WW8	23967.06	18759.45	554.00	82	24.99			503.40	507.06	530.84	533.58		
61	5	35	3	WW9	23149.03	18192.25	554.00	88	26.82				510.11	521.08	522.30	535.71	536.32
61	5	35	3	WW15	23151.00	18192.00	554.00	81	24.69	0.00		504.01	505.84	522.00	522.61	542.42	543.03
61	5	35	15	WW10	23638.92	19505.21	552.00		0.00								
61	5	36	1	WW11	25546.48	18385.68	555.00	70	21.34					531.84	532.14	540.06	540.37
61	5	36	1	WW12	25546.29	18357.75	555.00		0.00								
61	5	36	4	WW13	24486.76	18132.11	555.00	81	24.69			506.23	509.89	522.08	524.52	543.72	545.25
61	5	36	4	WW14	24481.50	18106.74	555.00		0.00								
62	3	4	13	RCA30	38791.45	21271.70	536.40	76	39.50	499.90				522.40		532.90	
62	3	5	4	WW102	37326.70	20157.53	540.00	79	24.08	501.60						533.90	
62	3	6	16	WW103	36863.98	20702.86	542.00	89	27.13							535.90	536.51
62	3	7	8	WW105	36967.99	22116.28	538.00	85	25.91							527.64	528.86
62	3	7	8	WW108	36966.93	22111.26	538.00	85	25.91							524.59	525.81
62	3	7	4	WW104	35603.44	21690.14	540.50	82	24.99							532.27	534.40
62	3	7	15	WW107	36480.72	22790.86	539.50	88	26.82					516.64	518.77	530.36	533.40
62	3	7	15	TH106	36419.74	22782.91	539.50	88	26.82	496.83						531.58	531.88

Appendix B, Table 1, Data points for geologic interpretations.

TWP	RG	SC	LS	WELL ID	UTM E. mE	UTM N. mN	ELEV. m.a.s.l	YR 19'	DEPTH m	Bedrk Elv m.a.s.l	EMP Elv m.a.s.l	Brn Lk Elv m.a.s.l	Mur Lk Elv m.a.s.l	Bvlle Elv m.a.s.l	Eth Lk Elv m.a.s.l	Mar Ck Elv m.a.s.l	Snd Rv Elv. m.a.s.l
62	3	8	2	WW110	38029.44	21637.79	538.00	84	25.60							532.51	533.12
62	3	8	5	WW109	37229.41	21983.36	538.00	76	23.16							526.11	533.43
62	3	8	14	WW111	37525.00	22874.79	538.00	0	0.00								
62	3	9	8	WW112	40244.88	22106.85	536.50	72	21.95							529.03	533.76
62	3	9	14	WW113	39463.12	22886.33	536.50	86	26.21							527.36	532.54
62	3	10	9	WW114	41852.91	22531.95	539.00	69	21.03	494.80			503.34			529.86	531.38
62	3	13	4	WW115	43854.48	23154.68	539.50	72	21.95							532.79	534.62
62	3	13	4	WW116	43851.87	23169.88	539.50	83	25.30							530.36	534.01
62	3	13	16	TH117	45081.87	24564.70	540.00	90	27.43	495.50						530.86	
62	3	14	1	WW119	43431.25	23367.94	538.00	83	25.30							528.86	533.43
62	3	14	12	UOH 26	41990.99	24049.85	535.64	91	5.20							531.04	534.44
62	3	14	11	WW118	42517.66	24017.33	535.00	69	21.03							529.82	531.34
62	3	15	1	TH121	41903.85	23075.15	534.68	66	20.12	497.49				511.52	512.73	525.54	527.67
62	3	15	13	WW120	40448.57	24297.32	540.50	77	23.47					522.21	525.87	535.32	538.06
62	3	16	3	TH123	39198.13	22998.99	536.57	66	20.12	493.59		495.42	497.56			526.82	530.47
62	3	16	15	WW122	39535.03	24411.36	539.50	81	24.69				501.40	507.50	509.63	531.88	
62	3	17	3	UOH 14	37883.36	23006.78	537.25	91	6.70								
62	3	17	4	WW124	37298.00	23117.63	539.00	86	26.21					508.52	514.62	529.86	536.87
62	3	17	4	UOH 13	37300.66	23092.27	538.07	91	6.70							532.17	534.57
62	3	17	5	UOH 12	37336.13	23680.98	537.82	91	6.10							532.02	533.52
62	3	17	13	Ch	37165.12	24382.65	540.00	0	0.00								
62	3	17	16	Mc	38632.51	24352.95	540.50	0	0.00							534.40	535.32
62	3	18	2	WW130	36670.32	23035.44	539.50	87	26.52							531.58	535.84
62	3	18	7	TH127	37080.00	23745.00	538.00	66	20.12	493.50							
62	3	18	13	TH128	35481.37	24494.83	539.34	66	20.12	503.37							
62	3	19	1	Di	36980.60	24698.68	541.00	76	23.16							533.08	534.90
62	3	19	4	UOH 15	35600.36	24607.11	539.35	91	6.70							533.85	534.45
62	3	19	7	TH132	37087.08	25414.58	541.00	66	20.12	490.71			499.55	513.87	517.53		
62	3	19	12	To	35530.61	25410.78	548.00		0.00								
62	3	20	11	OBS2	37881.80	25591.65	539.74	75	22.86							533.03	535.47
62	3	20	12	OBS1	37107.08	25404.58	540.49	75	22.86							534.09	536.22
62	3	20	13	WSW2	37270.25	25869.72	542.42		0.00					512.86	518.34	523.83	532.97
62	3	20	14	WSW1	37512.19	25787.40	542.70		0.00					513.13	519.84	524.11	536.30
62	3	21	1	WW141	40232.95	24732.43	542.00	78	23.77							530.42	536.82
62	3	21	1	WW142	40243.25	24704.59	542.00	66	20.12							531.64	536.21
62	3	21	1	WW146	40223.01	24681.64	542.00	83	25.30					509.39	513.35	527.37	531.64
62	3	21	1	UOH 22	40283.22	24864.64	541.72	91	10.70							532.62	535.62
62	3	21	3	WW139	39258.51	24742.60	540.50	80	24.38							530.44	534.40
62	3	21	4	UOH 17	38744.31	24931.96	538.69	91	6.10								
62	3	21	3	WW147	39243.81	24702.39	541.00	86	26.21	491.62			500.16	519.66	522.10	532.77	540.09
62	3	21	5	OBS 21	38921.32	25153.67	539.00	78	23.77	497.16			503.25	522.15	523.37	528.55	532.82

Appendix B, Table 1, Data points for geologic interpretations.

TWP	RG	SC	LS	WELL ID	UTM E mE	UTM N mN	ELEV m a s l	YR 19	DEPTH m	Bedrk Elv m a s l	EMP Elv m a s l	Brn Lk Elv m a s l	Mur Lk Elv m a s l	Bvlle Elv m a s l	Eth Lk Elv m a s l	Mar Ck Elv m a s l	Snd Rv Elv m a s l
62	3	21	14	Ca	39377.48	26064.45	540.00	75	22.86								
62	3	21	14	WW144	39369.74	26089.77	540.00	84	25.60					508.39	513.57	529.33	536.95
62	3	21	15	WW138	39812.41	26054.26	541.00	82	24.99							531.86	536.43
62	3	21	15	WW143	39817.38	26079.66	541.00	86	26.21							533.38	538.56
62	3	22	1	TH151	41806.65	24875.86	540.00	90	27.43	498.85				515.92	517.14		
62	3	22	11	TH150	41115.00	24655.00	538.50	66	20.12	497.05							
62	3	22	4	WW149	40411.08	24710.62	542.00	79	24.08							530.42	534.38
62	3	22	12	WW148	40518.47	25705.69	540.50	80	24.38							532.27	536.23
62	3	23	2	WW152	42997.37	24806.54	535.00	80	24.38								
62	3	23	4	UOH 19	42400.18	24681.36	537.14	91	4.60								535.94
62	3	23	12	UOH 18	41957.79	25759.60	536.87	91	6.10							532.27	533.47
62	3	24	4	TH154	43633.86	24683.31	533.00	66	20.12	495.51				507.09	512.27	524.47	526.90
62	3	24	4	WW153	44098.81	25368.46	532.00	85	25.91					515.54	516.15	523.16	523.77
62	3	26	8	WW155	43482.35	26676.62	530.40	90	27.43	473.71		482.24	484.68	502.97	506.02		
62	3	27	4	TH157	40368.40	26268.01	537.85	66	20.12	499.75			Log off scal	511.64	515.60		
62	3	27	4	WW156	40517.47	26484.51	537.50	86	26.21					505.80	515.55	525.92	531.40
62	3	28	5	TH158	38838.05	27809.18	540.00	66	20.12	498.85			Log off scale				
62	3	29	3	UOH 5	37628.76	26434.91	540.56	91	4.60								
62	3	30	14	UOH 20	35864.18	27540.84	541.30	91	3.34								
62	3	31	16	WW159	36420.57	29319.65	545.59	78	23.77							535.23	537.36
62	3	31	16	TH161	36814.43	29375.22	546.00	80	24.38	496.01		500.28	502.72	511.80	513.08	535.48	537.77
62	3	31	15	WW160	36430.90	29284.20	545.59	86	26.21							534.92	535.23
62	3	33	2	OB2-33A	39635.49	28049.72	540.00	66	20.12	472.03			486.05	495.19	498.24	510.74	529.03
62	3	33	2	OB2-33	39623.07	27986.23	540.00	66	21.60							nr	528.11
62	3	33	8	TH164	40228.27	28563.07	538.00	78	23.77	465.15			486.79	513.62	519.71	525.20	526.11
62	3	33	15	TH163	39581.08	29391.38	540.50	78	23.77	493.56		497.83	503.92	515.20	517.94	525.26	528.31
62	3	33	15	WW162	39591.91	29246.84	541.02	85	25.91							528.31	531.57
62	3	34	2	TH167	41111.33	27888.31	538.09	66	20.12	442.08			484.75	498.47	499.99	521.33	525.90
62	3	35	13	TH168	41920.65	29458.32	538.50	78	23.77	497.69				500.40	508.02	528.64	532.40
62	4	1	2	WW202	34988.21	19773.74	546.00	80	24.38							537.77	538.08
62	4	1	2	WW201	34978.42	19780.16	546.00	84	25.60							527.71	529.24
62	4	1	5	ARDSCH	33893.24	20354.76	554.00	71	21.64	493.65							
62	4	1	12	WW203	34007.05	20558.32	554.00	70	21.34								
62	4	2	1	WW205A	33536.62	20568.81	553.00	78	23.77							531.97	532.27
62	4	2	5	WW273a	32327.52	20103.77	558.00	84	25.60							541.85	542.76
62	4	2	8	TH2	33534.41	20454.59	554.00	75	22.86	492.43			493.95			532.36	533.88
62	4	2	12	WW274	32328.38	20662.15	556.00	89	27.13								
62	4	2	12	TH274	32328.00	20662.00	556.00	87	26.52					524.30	524.61		
62	4	2	9	WW205	33568.78	20565.47	553.00	69	21.03								
62	4	3	1	UOH 9	32000.66	19698.68	557.79	91	12.10								

Appendix B, Table 1, Data points for geologic interpretations.

TWP	RG	SC	LS	WELL ID	UTM E mE	UTM N mN	ELEV. m.a.s.l	YR 19'	DEPTH m	Bedrk Elv m.a.s.l	EMP Elv m.a.s.l	Brn Lk Elv m.a.s.l	Mur Lk Elv m.a.s.l	Bvlle Elv m.a.s.l	Eth Lk Elv m.a.s.l	Mar Ck Elv m.a.s.l	Snd Rv Elv m.a.s.l
62	4	3	2	WW217	31715.22	19887.69	558.00	88	26.82	495.82				511.37	513.80		
62	4	3	4	UOH 25	32185.81	20709.71	555.48	91	9.10								
62	4	3	4	WW218	30707.84	19918.46	561.00	80	24.38	509.18				509.18	512.23		
62	4	5	5	WW219	27467.05	20441.43	555.00	81	24.69								
62	4	5	5	WW220	27474.76	20416.07	555.00	90	27.43				503.79	523.00	523.61	534.88	535.80
62	4	5	5	TH220A	27474.00	20416.00	555.00	90	27.43	488.00			500.14	523.30	524.52		
62	4	6	8	WW222	27243.11	20427.68	555.00	81	24.69								
62	4	6	12	WW221	26114.14	20843.92	552.00	78	23.77					513.90	521.52	536.76	537.37
62	4	7	13	TH225	25794.73	22795.34	556.50	77	23.47	501.03				514.44	516.88	539.13	540.96
62	4	7	13	WW226	25843.36	22716.83	557.00	85	25.91					514.94	517.68		
62	4	7	9	WW223	27300.72	22327.26	559.00	79	24.08								
62	4	7	16	WW224	27280.41	22309.39	559.00	84	25.60							542.24	542.54
62	4	8	1	WW227	28898.09	21806.40	558.00	80	24.38	506.49				518.07	520.20	535.75	537.58
62	4	8	1	WW227a	28898.00	21800.00	558.00		21.00								
62	4	8	1	WW227b	28910.00	21800.00	558.00		0.00								
62	4	8	14	WW229	28044.96	22805.47	559.50	80	24.38								
62	4	8	9	WW228	28789.56	22334.01	559.50	80	24.38								
62	4	8	12	WW230	27468.97	22244.23	559.50	83	25.30							544.26	544.87
62	4	10	1	WW231	32150.07	21592.84	547.00	81	24.69					523.23	525.05	536.94	538.16
62	4	10	5	UOH 8	32273.93	22000.00	544.06	91	7.60						nr		537.06
62	4	10	16	UOH 7	32181.93	22834.25	557.18	91	12.20								
62	4	10	16	WW232	32138.93	22755.36	557.00	85	25.91	503.66				517.68	518.90	533.84	534.44
62	4	10	16	Piech O	32138.00	22755.00	556.50		12.10								
62	4	10	16	Piech N	32138.00	22765.36	556.50		16.43								
62	4	12	8	WW209	35185.63	21804.73	541.00	86	26.21					525.15	525.76	534.90	536.43
62	4	12	4	WW208	33940.93	21418.34	547.50	87	26.52							532.87	534.09
62	4	12	4	WW208A	33940.00	21418.00	547.50	84	25.60					523.73	524.64	533.78	534.39
62	4	12	12	TH7A	33932.00	22441.00	543.00	89	27.13	502.46						534.47	538.12
62	4	12	12	WW7B	33933.00	22441.20	543.00	89	27.13							530.50	530.81
62	4	12	12	WW207	33932.81	22441.11	543.00	79	24.08							534.16	535.99
62	4	13	2	WW210	34976.18	23004.18	542.00	78	23.77							534.38	538.95
62	4	13	16	WW211	35353.17	24495.68	540.30	85	25.91							530.55	531.16
62	4	14	1	WW214	33495.87	23147.29	550.50	85	25.91							533.74	536.78
62	4	14	3	WW215	32874.79	23329.81	553.00	88	26.82					516.42	519.47		
62	4	14	12	UOH 6	32582.89	23980.91	552.31	91	9.10								
62	4	14	9	WW212	33717.69	23808.24	549.50	85	25.91							531.82	532.74
62	4	14	9	WW213	33582.95	23792.41	550.00	86	26.21							532.32	533.54
62	4	15	12	WW233	30723.75	23850.83	559.00	85	25.91								
62	4	16	4	WW234	29090.24	23023.46	561.00	80	24.38								
62	4	17	4	WW235	27440.02	23346.18	555.00	84	25.60							539.46	540.98
62	4	18	13	WW236	25907.69	24306.58	552.00	79	24.08					525.73	526.70	534.93	538.28

Appendix B, Table 1, Data points for geologic interpretations.

TWP	RG	SC	LS	WELL ID	UTM E mE	UTM N mN	ELEV m a s l	YR 19	DEPTH m	Bedrk Elv m a s l	EMP Elv m a s l	Brn Lk Elv m a s l	Mur Lk Elv m a s l	Bvlle Elv m a s l	Eth Lk Elv m a s l	Mar Ck Elv m a s l	Snd Rv Elv m a s l
62	4	18	13	WW238	25825.57	24286.34	552.00	83	25.30							535.85	538.59
62	4	18	13	WW237	25832.86	24395.59	552.00	86	26.21							536.15	538.28
62	4	18	13	WW240	25825.49	24311.74	552.00	88	26.82							536.15	538.59
62	4	18	13	WW239	25853.21	24400.76	552.00	88	26.82							535.85	537.98
62	4	19	1	TH242	27232.54	24559.18	551.00	80	24.38			506.50		514.73	517.17	528.14	529.66
62	4	19	1	WW241	27232.45	24587.12	551.00	80	24.38	502.23				512.60	519.30		
62	4	19	12	WW284	25855.00	25439.64	549.00		0.00								
62	4	19	12	WW246	25857.51	25439.64	549.00	79	24.08					506.94	517.30		
62	4	19	12	TH247	25862.78	25383.78	549.00	88	26.82	494.36				509.68	516.39	530.41	531.93
62	4	19	12	WW245	25857.61	25409.16	549.00	88	26.82					506.94	517.30		
62	4	19	15	WW244	26868.76	26081.51	550.50	89	27.13							530.38	532.21
62	4	20	4	WW243	27464.03	24605.91	551.00	80	24.38							538.50	539.42
62	4	20	12	WW248	27435.73	25504.84	546.00	84	25.60					521.01	527.71		
62	4	21	16	WW250	30450.85	25723.48	549.50	0	0.00								
62	4	22	13	WW251	30920.94	25923.55	549.00	88	26.82							538.64	542.90
62	4	22	15	WW249	31518.58	25999.74	550.00	89	27.13	501.23				511.60	513.42	540.86	543.29
62	4	23	1	WW216	33777.83	24960.94	553.00	76	23.16							540.81	545.38
62	4	23	12	WW252	32308.90	25429.32	558.00	88	26.82							543.67	543.98
62	4	24	13	WW273	34006.96	25733.64	553.00	82	24.99							542.33	544.47
62	4	25	2	WW277	34824.17	26275.29	556.00	78	23.77	502.96				520.03	524.30	541.37	542.59
62	4	25	2	Bu	34824.00	26285.29	556.00		7.42								
62	4	25	2	Bu	34826.78	26252.46	556.00	83	25.30					521.25	523.39		
62	4	25	9	TH218	35428.00	27160.00	548.33	66	20.12	496.27			502.61	512.97	516.33		
62	4	25	13	WW276	33892.00	27609.00	542.00	89	27.13	495.37			501.77	516.09	517.92	538.65	539.87
62	4	25	13	WW274b	33892.03	27609.27	542.00	86	26.21					517.01	517.62	531.64	532.86
62	4	25	16	WW275	35354.66	27463.01	542.00	84	25.60							532.25	532.86
62	4	26	2	WW279	33041.38	26298.19	547.00	78	23.77							536.03	541.51
62	4	26	1	WW280	33046.53	26277.90	547.00	82	24.99							532.67	533.28
62	4	26	16	WW278	33676.89	27263.09	545.00	80	24.38							535.86	541.95
62	4	26	16	WW225	33589.32	27628.33	542.00	79	24.08							535.29	535.60
62	4	26	16	WW281	33597.00	27613.13	542.00	85	25.91				501.77	517.92	521.88	531.33	535.90
62	4	27	5	WW254	30678.77	26902.69	549.00	86	26.21							536.81	542.60
62	4	27	9	WW253	32057.15	27086.21	545.00	83	25.30							534.03	534.64
62	4	28	3	WW255	29683.29	26256.02	548.00	79	24.08							538.25	544.95
62	4	28	16	WW256	30481.15	27455.47	549.00	95	25.91							536.81	540.77
62	4	29	4	WW257	27405.81	26911.82	551.00	78	23.77							540.94	541.86
62	4	30	8	WW260	27207.32	26890.67	551.00	85	25.91							533.32	534.85
62	4	30	4	UOH 2	25795.64	26465.60	549.71	91	12.00								
62	4	30	5	WW261	25754.00	26747.40	550.00	85	25.91	496.00			501.23			530.19	533.24
62	4	30	12	UOH 1	25923.31	27146.94	552.77	91	12.20								
62	4	30	13	WW258	25803.00	27070.00	553.00	81	24.69					508.80	516.42	534.71	537.76

Appendix B, Table 1, Data points for geologic interpretations.

TWP	RG	SC	LS	WELL ID	UTM E mE	UTM N mN	ELEV m.a.s.l	YR 19'	DEPTH m	Bedrk Elv m.a.s.l	EMP Elv m.a.s.l	Brn Lk Elv m.a.s.l	Mur Lk Elv m.a.s.l	Bvlle Elv m.a.s.l	Eth Lk Elv m.a.s.l	Mar Ck Elv m.a.s.l	Snd Rv Elv m.a.s.l
62	4	30	13	WW262	25803.88	27070.23	553.00	88	26.82					508.80	516.42	536.54	538.37
62	4	30	16	WW259	27232.19	27073.66	551.00	83	25.30					515.64	522.04	536.98	545.21
62	4	31	1	WW264	27156.12	27782.04	549.50	84	25.60							540.36	543.71
62	4	31	6	WW263	27110.63	29275.57	550.50	89	27.13								
62	4	32	13	WW265	27818.50	29237.76	549.00	85	25.91							541.08	545.04
62	4	32	16	WW269	28464.90	29291.16	552.00	73	22.25							544.99	548.34
62	4	32	9	WW266	28490.34	29296.35	552.00	77	23.47							540.42	542.86
62	4	32	15	WW267	28833.40	28662.71	551.00	84	25.60								
62	4	32	16	WW268	28513.26	29293.90	552.0[illegible]	84	25.60							545.90	548.04
62	4	33	1	WW270	30484.63	27978.63	550.50	79	24.08							534.04	535.56
62	4	33	16	WW271	30436.93	28595.57	549.00	89	27.13							531.93	539.25
62	[illegible]	34	2	WW272	31708.57	27940.50	550.50	78	23.77							540.14	540.75
62	4	35	1	WW282	33586.06	27872.05	542.00	83	25.30							533.47	534.38
62	4	36	8	WW283	35331.56	28407.30	547.00	85	25.91					516.52	519.57	537.25	537.86
62	4	36	13	TH231	33812.99	29388.81	547.00	77	23.47	493.05			496.10			538.77	540.29
62	5	1	1	WW301	25646.69	19752.38	553.50	82	24.99								
62	5	1	1	WW302	25644.25	19724.43	553.50	82	24.99								
62	5	1	1	WW303	25519.57	19698.45	553.50	88	26.82							536.74	538.26
62	5	2	8	WW304	24059.27	20118.18	552.00	73	22.25					512.38	512.99		
62	5	9	16	TH185			548.80	77	23.47	490.89			505.52	514.97	518.62	532.95	536.61
62	5	10	16	TH184a	22453.50	22820.50	551.28	77	23.47				508.91	514.09	518.36	534.52	539.09
62	5	10	16	TH184b	22453.50	22820.50	551.28	77	23.47	468.98	481.48	495.20	508.91	514.09	518.36	534.52	539.09
62	5	11	4	TH306	24055.00	21300.00	551.80	77	23.47	471.03	478.34	493.58	50[illegible]	518.88	519.19	534.43	536.56
62	5	11	15	WW305	23340.13	22743.65	554.00	84	25.60					514.38	515.90		
62	5	12	8	WW307	25578.44	23546.14	559.00	88	26.82								
62	5	13	4	TH308	24253.73	22950.96	559.30	68	20.73								
62	5	14	16	WW309	23992.31	24235.11	552.00	85	25.91					512.38	516.03		
62	5	23	6	WW310	22873.94	25078.73	551.00	88	26.82	501.01				506.50	516.25	533.63	536.37
62	5	23	12	WW312	22654.90	25529.71	550.50	80	24.38								
62	5	23	12	WW313	22662.39	25509.35	550.50	85	25.91					507.22	512.10	534.65	535.26
62	5	24	4	WW3[illegible]	24179.86	24820.19	551.[illegible]	82	24.99	492.00		493.70	5[illegible].45	508.92	516.56	533.32	539.11
62	5	24	15	[illegible]	25302.85	26095.17	549.78	91	6.10								
62	5	25	3	[illegible]	24693.84	28214.52	562.[illegible]	81	24.69	493.42		500.43	504.09	504.09	5[illegible]	531.22	532.74
62	5	25	8	[illegible]	25618.66	26966.23	553.[illegible]	[illegible]	24.69								
62	5	25	16	[illegible]	25483.79	[illegible]	55[illegible]	[illegible]	26.82			502.84	5[illegible].89	513.8[illegible]	518.08		
62	5	26	1	WW[illegible]	23733.60	26239.39	[illegible]	[illegible]	[illegible]			50[illegible].99	[illegible]			534.22	534.82
62	5	26	4	WW[illegible]	[illegible]	[illegible]	549.[illegible]	[illegible]	[illegible]								
62	5	26	4	TH319	22498.74	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]		[illegible]	[illegible]	514.95	517.69	[illegible]	539.03
62	5	36	9	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]								
62	5	36	9	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]								

Appendix B, Table 1, Data points for geologic interpretations.

TWP	RG	SC	LS	WELL ID	UTM E. mE	UTM N. mN	ELEV. m.a.s.l	YR 19'	DEPTH m	Bedrk Elv m.a.s.l	EMP Elv m.a.s.l	Brn Lk Elv m.a.s.l	Mur Lk Elv m.a.s.l	Bvlle Elv m.a.s.l	Eth Lk Elv m.a.s.l	Mar Ck Elv m.a.s.l	Snd Rv Elv. m.a.s.l
62	5	3	5	TH330	20920.00	20195.00	554.00	77	23.47	479.02						542.72	543.33
63	3	4	4	WW402	38636.55	29621.78	543.00	78	23.77					521.05	521.66	527.46	527.76
63	3	4	4	TH401	38443.97	29470.99	543.00	80	24.38	491.79		496.37	500.33			526.85	531.42
63	3	6	2	WW403	36074.02	29485.04	543.00	84	25.60							531.42	535.08
63	3	7	8	WW409	36621.40	31568.24	547.00	0	0.00								
63	3	7	8	UOH 16	36500.88	31790.74	544.10	91	3.10								
63	3	8	1	WW404	38337.33	31114.15	543.00	85	25.91					514.96	521.66	527.76	536.90
63	3	8	13	Kr	36830.00	31924.57	546.00	92	46.41								
63	3	8	13	Kr	36825.74	31924.57	546.00	77	23.47	496.62			501.80			519.48	534.42
63	3	9	4	TH407	38405.33	31259.15	543.15	78	23.77	493.16		496.82	499.87	517.55	521.81	524.86	535.53
63	3	9	4	WW408	38558.04	31232.14	543.15	78	23.77							528.52	535.53
63	3	18	9	TH410	36680.00	33840.00	551.00	88	26.82	500.71				525.09	526.62	529.05	535.76
63	4	1	4	WW501	33608.83	29619.06	547.00	84	25.60				499.45	520.18	521.09		
63	4	2	1	TH502	33461.80	29453.42	547.00	80	24.38	492.75			497.62				
63	4	2	13	WW510	32130.00	30695.00	510.54		8.07								
63	4	2	15	UOH 21	32842.25	30779.08	507.00	91	3.10								
63	4	4	1	WW503	30095.72	29490.74	550.50	78	23.77							539.53	540.75
63	4	4	4	WW505	28693.49	29452.11	550.50	81	24.69					516.36	516.97	532.82	538.31
63	4	4	4	WW504	28988.74	29453.29	551.00	86	26.21							530.58	538.50
63	4	5	3	WW506	27400.29	29459.64	549.00	72	21.95								
63	4	5	4	TH508	27003.32	29394.54	550.50	80	24.38	475.52			493.20			535.87	541.66
63	4	5	4	WW507	27036.21	29460.72	550.50	86	26.21							534.04	540.14
63	4	8	16	WW509	28440.00	32460.00	541.00	89	27.13	480.00							
					Total number		304.00		Mean	464.72	475.42	496.31	503.48	516.59	519.76	534.08	536.93

Appendix B. Table 2. Wells sorted by completion elevation.

TWP	RG	SC	LS	WELL ID	UTM E.	UTM N.	ELEV (m)	YR	DEEP (m)	Completion	W.L Elv.	Bedrk Elv	EMP Elv	Brn Lk Elv	Mud Lk Elv	Bvile Elv	Eth Lk Elv	Mar Ck Elv	Snd Rv Elv
					mE	mN	m.a.s.l	19	m	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l
62	4	7	9	WW223	27300.72	22327.26	559.00	79	24.08	552.29	555.95								
62	4	10	16	UOH 7	32181.93	22834.25	557.18	91	12.20	551.39	553.20								
62	4	15	12	WW233	30723.75	23850.83	559.00	85	25.91	551.38	555.34								
62	4	16	4	WW234	29090.24	23023.46	561.00	80	24.38	551.25	554.90								
62	5	36	9	UOH 23	25501.15	28560.25	555.52	91	12.20	549.42	549.77								
62	4	3	1	UOH 9	32000.66	19698.68	557.79	91	12.10	549.26	553.61								
62	4	8	14	WW229	28044.96	22805.47	559.50	80	24.38	549.14	553.71								
62	4	3	4	UOH 25	32185.81	20709.71	555.48	91	9.10	547.86	550.85								
62	4	25	2	Bu	34824.00	26285.29	556.00		7.42	547.80	550.20								
62	4	14	12	UOH 6	32582.89	23980.91	552.31	91	9.10	547.13	548.29								
62	4	1	12	WW203	34007.05	20558.32	554.00	70	21.34	546.99	551.71								
62	5	24	15	UOH 3	25302.85	26095.17	549.78	91	6.10	546.73	547.70								
61	5	26	12	WW7	22580.76	17331.60	553.00	48	14.63	546.29	550.26								
61	5	36	4	WW14	24481.50	18106.74	555.00		0.00	545.55	553.17								
61	4	32	1	WW716	28925.00	18235.00	550.00		0.00	545.43	547.56								
61	4	31	12	WW722	25831.00	18915.00	554.00		0.00	545.31	549.43								
61	5	23	8	WW2	24023.21	15452.88	554.00	87	26.52	545.01	549.12								
61	4	34	13	WW713	30695.00	19465.00	562.00	82	24.99	544.63	547.37								
61	4	25	2	WW735	34865.00	16675.00	553.00	79	24.08	544.47	548.43								
62	4	10	16	Piech O	32138.00	22755.00	556.50		12.10	544.31	546.59								
61	4	28	2	WW727	30150.00	16780.00	555.00	80	24.38	544.18	548.90							542.50	545.86
61	4	28	5	WW729	29100.00	16940.00	553.00	74	22.56	544.16	546.90							545.25	547.08
61	4	34	12	WW712	30725.00	19145.00	558.00	86	26.21	543.98	551.29								
62	4	30	12	UOH 1	25923.31	27146.94	552.77	91	12.20	543.93	544.82								
62	5	1	1	WW301	25646.69	19752.38	553.50	82	24.99	543.75	549.54								
61	4	34	16	UOH 10	32207.43	19373.86	553.33	91	11.30	543.57	550.33								
62	5	13	8	WW307	25578.44	23546.14	559.00	88	26.82	543.46	556.56								
62	4	5	5	WW219	27467.05	20441.43	555.00	81	24.69	543.42	551.95								
62	4	8	12	WW230	27468.97	22244.23	559.50	83	25.30	543.35	552.18							544.26	544.87
62	5	36	9	UOH 24	25660.46	28908.97	551.69	91	8.61	543.08	543.44								
62	4	8	9	WW228	28789.56	22334.01	559.50	80	24.38	543.04	552.79								
61	4	25	5	WW736	34005.00	17180.00	552.00		0.00	542.86	545.90								
62	4	30	4	UOH 2	25795.64	26465.60	549.71	91	12.00	542.78	546.00								
61	3	30	1	WW602	37020.00	16610.00	554.00	89	27.13	542.72	549.12							543.42	541.03
61	4	32	1	WW717	28920.00	18230.00	550.00	35	10.67	542.68	549.09								
62	4	2	9	WW205	33568.78	20565.47	553.00	69	21.03	542.64	548.12								
62	4	32	16	WW269	28464.90	29291.16	552.00	73	22.25	542.55	546.51							544.99	548.34
62	5	1	1	WW302	25644.23	19724.43	552.50	82	24.99	542.53	547.40								
63	3	7	8	UOH 16	36500.88	31790.74	544.10	91	3.10	542.48	542.60								
62	4	2	5	WW273a	32327.52	20103.77	558.00	84	25.60	542.15	550.38							541.85	542.76
61	4	33	1	WW708	30444.00	18455.00	552.00	64	19.51	541.94	544.99								
61	4	34	9	UOH 11	32186.83	18960.48	548.50	91	7.60	541.79	547.11								
62	4	21	16	WW250	30450.85	25723.48	549.50	0	0.00	541.58	544.01								
61	5	35	15	WW10	23638.92	19505.21	552.00		0.00	541.33	544.38								
62	4	7	16	WW224	27280.41	22309.39	559.00	84	25.60	541.32	542.54							542.[illegible]	542.54
62	4	32	13	WW265	27818.50	29237.76	549.00	85	25.91	541.08	546.26							541.[illegible]	545.04
62	4	31	1	WW264	27156.12	27782.04	549.50	84	25.60	540.97	546.76							540.36	543.71
62	4	34	2	WW272	31708.57	27940.50	550.50	78	23.77	540.14	545.93							540.14	540.75

Appendix B. Table 2. Wells sorted by completion elevation.

TWP	RG	SC	LS	WELL ID	UTM E.	UTM N.	ELEV. (m)	YR	DEEP (m)	Completion	W L. Elv.	Bedrk Elv	EMP Elv	Brn Lk Elv	Mur Lk Elv	Bvlle Elv	[illegible] L. Elv	[illegible] Ck Elv	Snd Rv Elv.
					mE	mN	m.a.s.l	19'	m	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l
62	4	10	16	Piech N	32138.00	22765.36	556.50		16.43	540.04	546.52								
62	4	23	12	WW252	32308.90	25429.32	558.00	88	26.82	540.02	549.77							543.67	543.98
61	4	31	12	WW723	25830.00	18916.00	554.00	81	24.69	539.98	539.06							539.67	540.28
61	4	34	15	WW707	31465.00	19240.00	557.00	81	24.69	539.93	543.59								
63	4	5	3	WW506	27400.29	29459.64	549.00	72	21.95	539.55	546.56								
62	3	30	14	UOH 20	35864.18	27540.84	541.30	91	3.34	539.54	539.86								
63	4	4	1	WW503	30095.72	29490.74	550.50	78	23.77	539.53	547.45							539.53	540.75
62	4	6	8	WW222	272[illegible].11	20427.68	555.00	81	24.69	539.46	545.55								
62	4	17	4	WW235	27440.02	23346.18	555.00	84	25.60	539.46	549.21							539.46	540.98
62	4	10	16	WW232	32138.93	22755.36	557.00	85	25.91	539.32	544.81	503.66				517.68	518.90	533.84	534.44
62	3	29	3	UOH 5	37628.76	26434.91	540.56	91	4.60	538.75	539.21								
62	4	10	5	UOH 8	32273.93	22000.00	544.06	91	7.60	538.57	540.33							nr	537.06
62	4	20	4	WW243	27464.03	24605.91	551.00	80	24.38	538.50	547.95							538.50	539.42
62	4	29	4	WW257	27405.81	26911.82	551.00	78	23.77	538.20	547.95							540.94	541.86
62	4	30	16	WW259	27232.19	27073.66	551.00	83	25.30	538.20	548.87					515.64	522.04	536.98	545.21
61	4	33	1	WW709	30445.00	18455.00	552.00	74	22.56	538.06	541.33								
61	4	26	16	WW734	33805.00	17920.00	549.00	78	23.77	538.03	544.43							543.42	544.64
61	4	35	4	WW705	32360.00	18365.00	549.00	35	10.67	538.03	539.86								
62	3	19	12	To	35530.61	25410.78	548.00		0.00	537.94	541.80								
62	4	26	2	WW279	33041.38	26298.19	547.00	78	23.77	537.86	543.34							536.03	541.51
62	4	23	1	WW216	33777.83	24960.94	553.00	76	23.16	537.76	546.90							540.81	545.38
62	4	32	9	WW266	28490.34	29296.35	552.00	77	23.47	537.67	544.99							540.42	542.86
62	4	22	13	WW251	30920.94	25923.55	549.00	88	26.82	537.42	544.73							538.64	542.90
62	4	28	3	WW255	29683.29	26256.02	548.00	79	24.08	537.33	541.90							538.25	544.95
63	3	7	8	WW409	36621.40	31568.24	547.00	0	0.00	537.25	543.34								
61	4	26	13	WW732	32315.00	17925.00	550.50		0.00	537.24	544.86								
62	4	8	1	WW227a	28898.00	21800.00	558.00		21.00	536.97	552.00								
62	3	21	4	UOH 17	38744.31	24931.96	538.69	91	6.10	536.86	537.55								
62	5	1	1	WW303	25519.57	19698.45	553.50	88	26.82	536.74	542.83							536.74	538.26
62	4	18	13	WW238	25825.57	24286.34	552.00	83	25.30	536.61	547.89							535.85	538.59
62	3	6	16	WW103	36863.98	20702.86	542.00	89	27.13	536.21	539.26							535.90	536.51
62	4	2	12	WW274	32328.38	20662.15	556.00	89	27.13	536.19	550.51								
62	4	8	1	WW227b	28910.00	21800.00	558.00		0.00	536.05									
61	4	35	4	WW739	32362.00	18375.00	547.42	92	28.04	535.99	544.37					530.35	530.66	539.80	540.10
61	5	23	12	WW3	22593.28	15820.33	549.00	83	25.30	535.89	544.43							531.93	532.24
62	5	23	12	WW312	22654.90	25529.71	550.50	80	24.38	535.87	545.62								
62	4	26	16	WW278	33676.89	27263.09	545.00	80	24.38	535.86	540.43							535.86	541.95
62	4	18	13	WW236	25807.69	24305.58	552.00	79	24.08	535.69	548.39					525.79	526.70	534.93	538.28
62	4	18	13	WW237	25832.86	24395.59	552.00	86	26.21	535.54	547.73							536.15	538.28
62	4	24	13	WW273	34005.96	25733.64	553.00	82	24.99	535.32	541.42							542.33	544.47
62	4	26	16	WW225	33589.32	27628.33	542.00	79	24.08	535.29	537.43							535.29	535.60
62	3	31	16	WW159	36420.57	29319.65	545.59	78	23.77	535.23	543.76							535.23	537.36
61	4	35	4	WW737	32360.00	18370.00	549.00	86	26.21	534.83	539.86								
62	3	17	13	Ch	37155.12	24382.65	540.00	0	0.00	534.82	537.10								
63	4	5	4	WW507	27036.21	29460.72	550.50	86	26.21	534.65	543.18							534.04	540.14
62	4	18	13	WW240	25825.49	24311.74	552.00	88	26.82	534.63	547.73							536.15	538.59
61	5	26	4	WW6	22595.44	16506.08	549.00	78	23.77	534.37	539.86								
62	4	18	13	WW239	25853.21	24400.76	552.00	88	26.82	534.32	547.43							535.85	537.98

Appendix B. Table 2. Wells sorted by completion elevation.

TWP	RG	SC	LS	WELL ID	UTM E	UTM N	ELEV. (m)	YR	DEEP (m)	Completion	W.L. Elv	Bedrk Elv	EMP Elv	Brn Lk Elv	Mur Lk Elv	Svlle Elv	Eth Lk Elv	Mar Ck Elv	Snd Rv Elv
					mE	mN	m.a.s.l	19[illegible]	m	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l
62	3	31	15	WW160	36430.90	29284.20	545.59	86	25.21	534.31	541.93							534.92	535.23
61	4	35	5	WW706	32450.00	18550.00	548.00	67	20.42	534.28	543.43								
62	3	19	4	UOH 15	35600.36	246[illegible].11	539.35	91	6.70	534.16	536.73							533.85	534.45
62	4	12	12	WW207	33932.81	22441.11	543.00	79	24.08	534.16	539.95							534.16	535.99
62	3	21	15	WW143	39817.38	26079.66	541.00	86	25.21	534.14	537.95							533.38	538.56
62	3	21	1	UOH 22	40283.22	24864.64	541.72	91	10.70	534.10	534.94							532.62	535.62
61	3	31	3	WW658	36049.07	18179.52	542.00	79	24.08	533.92	535.90							546.77	547.08
62	3	17	4	UOH 13	37300.66	23092.27	538.07	91	6.70	533.90	534.57							532.17	534.57
62	3	17	5	UOH 12	37336.13	23680.98	537.82	91	6.10	533.86	534.33							532.02	533.52
63	4	4	4	WW505	28693.49	29452.11	550.50	81	24.69	533.74	545.93					516.36	516.97	532.82	538.31
62	3	23	12	UOH 18	41957.79	25759.60	536.87	91	6.10	533.64	534.61							532.27	533.47
62	4	35	1	WW282	33586.06	27872.05	542.00	83	25.30	533.47	538.65							533.47	534.38
62	4	27	5	WW254	30678.77	26902.69	549.00	86	26.21	533.46	544.12							536.81	542.60
62	3	7	4	WW104	35603.44	21690.14	540.50	82	24.99	533.34	537.15							532.27	534.40
62	4	30	8	WW260	27207.32	26890.67	551.00	85	25.91	533.32	546.12							533.32	534.85
62	4	32	15	WW267	28833.40	28662.71	551.00	84	25.60	533.17	542.77								
62	4	28	16	WW256	30481.15	27455.47	549.00	85	25.91	533.15	544.12							536.81	540.77
62	3	17	16	Mc	38632.51	24352.95	540.50	0	0.00	532.88	536.20							534.40	535.32
62	3	23	4	UOH 19	42400.18	24681.36	537.14	91	4.60	532.87	534.12								535.94
62	4	33	16	WW271	30436.93	28595.57	549.00	89	27.13	532.85	540.47							531.93	539.25
62	4	27	9	WW253	32057.15	27086.21	545.00	83	25.30	532.81	541.95							534.03	534.64
62	3	17	3	UOH 14	37883.36	23006.78	537.25	91	6.70	532.68	534.40								
62	4	14	1	WW214	33495.87	23147.29	550.50	85	25.91	532.52	541.36							533.74	536.78
62	3	8	14	WW111	37525.00	22874.79	538.00	0	0.00	532.51	535.56								
61	4	31	2	WW719	26770.00	18125.00	550.50		0.00	532.21	545.93								
62	3	21	14	Ca	39377.48	26064.45	540.00	75	22.86	532.08	537.56								
62	4	14	9	WW213	33582.95	23792.41	550.00	86	26.21	532.02	546.95							532.32	533.54
61	5	36	1	WW11	25546.48	18385.68	555.00	70	21.34	531.99	548.90					531.84	532.14	540.06	540.37
62	3	14	12	UOH 26	41990.99	24049.85	535.64	91	5.20	531.99	533.49							531.04	534.44
62	4	2	1	WW205A	33536.62	20568.81	553.00	78	23.77	531.97	540.20							531.97	532.27
62	4	14	9	WW212	33717.69	23808.24	549.50	85	25.91	531.82	538.83							531.82	532.74
62	3	20	11	OBS2	37881.80	25591.65	539.74	75	22.86	531.81	537.45							533.03	535.47
62	3	22	12	WW148	40518.47	25705.69	540.50	80	24.38	531.66	535.93							532.27	535.23
62	4	12	4	WW208	33940.93	21418.34	547.50	87	26.52	531.65	544.45							532.87	534.09
61	4	33	1	WW710	30445.00	18456.00	552.00	78	23.77	531.43	545.42								
62	4	32	16	WW268	28513.26	29293.90	552.00	84	25.60	531.43	538.59							545.90	548.04
63	4	4	4	WW504	28988.74	29453.29	551.00	86	26.21	531.34	540.33							530.58	538.50
62	3	21	1	WW142	40243.25	24704.59	542.00	66	20.12	531.33	537.12							531.64	536.21
62	5	26	4	WW318	22577.76	26154.97	549.00	83	25.30	531.32	541.08								
62	4	19	15	WW244	26868.76	26081.51	550.50	89	27.13	531.30	545.01							530.38	532.21
62	3	20	12	OBS1	37107.08	25404.58	540.49	75	22.86	531.04	538.40							534.09	536.22
62	4	33	1	WW270	30484.63	27978.63	550.50	79	24.08	530.99	540.14							530.38	535.56
62	3	21	3	WW139	39258.51	24742.60	540.50	80	24.38	530.44	533.84							530.44	534.40
62	3	22	4	W[illegible]149	40411.08	24710.62	542.00	79	24.08	530.42	534.68							530.42	534.38
62	3	13	4	WW115	43854.48	23154.68	539.50	72	21.95	530.36	[illegible]36.15							532.79	534.62
61	5	23	8	WW1	24023.31	15468.12	554.00	77	23.47	530.23	547.90			508.28	511.33	522.00	523.52		
61	4	36	7	WW704	34731.00	18516.00	545.00	88	26.82	530.06	537.08					529.76	530.37		
62	3	18	2	WW130	36670.32	23035.44	539.50	87	26.52	530.05	535.84							531.58	535.84

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TWP	RG	SC	LS	WELL ID	UTM E.	UTM N.	ELEV. (m)	YR	DEEP (m)	Completion	W.L. Elv.	Bedrk Elv	EMP Elv	Brn Lk Elv	Mur Lk Elv	Bvlle Elv	Eth Lk Elv	Mar Ck Elv	Snd Rv Elv.
					mE	mN	m.a.s.l	19'	m	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l
62	4	26	1	WW280	33046.53	26277.90	547.00	82	24.99	529.63	537.25							532.67	533.28
61	4	31	2	WW720	26585.00	18135.00	550.00		0.00	529.43	543.90								
62	3	21	15	WW138	39812.41	26054.26	541.00	82	24.99	529.42	536.43							531.86	536.43
63	3	6	2	WW403	36074.02	29485.04	543.00	84	25.60	529.28	538.43							531.42	535.08
63	3	9	4	WW408	38558.04	31232.14	543.15	78	23.77	529.28	532.48							528.52	535.53
62	3	33	15	WW162	39591.91	29246.84	541.02	85	25.91	529.13	531.27							528.31	531.57
62	3	9	8	WW112	40244.88	22106.85	536.50	72	21.95	528.58	532.84							529.03	533.76
62	4	31	16	WW263	27110.63	29275.57	550.50	89	27.13	528.55	548.98								
62	4	1	2	WW201	34978.42	19780.16	546.00	84	25.60	528.47	535.33							527.71	529.24
62	4	13	2	WW210	34976.18	23004.16	542.00	78	23.77	528.28	536.82							534.38	538.95
61	4	26	13	WW733	32320.00	17925.00	550.50	78	23.77	527.94	541.05	495.33							
62	3	7	8	WW105	36967.99	22116.28	538.00	85	25.91	527.94	536.78							527.64	528.86
62	4	1	2	WW202	34988.21	19773.74	546.00	80	24.38	527.71	533.81							537.77	538.08
62	3	21	1	WW141	40232.95	24732.43	542.00	78	23.77	527.67	534.68							530.42	536.82
62	3	13	4	WW116	43851.87	23169.88	539.50	83	25.30	527.61	533.40							530.36	534.01
62	5	25	8	WW315	25618.66	26866.23	553.00	81	24.69	527.40	537.46								
62	3	19	1	Di	36980.60	24698.68	541.00	76	23.16	527.28	537.34							533.08	534.90
62	3	14	1	WW119	43431.25	23367.94	538.00	83	25.30	526.42	531.90							528.86	533.43
62	3	8	5	WW109	37229.41	21983.36	538.00	76	23.16	526.11	535.56							526.11	533.43
62	4	19	1	WW241	27232.45	24587.12	551.00	80	24.38	525.85	538.81	502.23				512.60	519.30		
62	3	8	2	WW110	38029.44	21637.79	538.00	84	25.60	525.81	537.39							532.51	533.12
62	3	14	11	WW118	42517.66	24017.33	535.00	69	21.03	525.55	531.34							529.82	531.34
62	3	23	2	WW152	42997.37	24806.54	535.00	80	24.38	525.55	531.65								
62	4	12	12	WW7B	33933.00	22441.20	543.00	89	27.13	525.32	533.86							530.50	530.81
62	3	9	14	WW113	39463.12	22886.33	536.50	86	26.21	525.22	532.54							527.36	532.54
62	3	7	8	WW108	36966.93	22111.26	538.00	85	25.91	525.20	530.38							524.59	525.81
62	4	25	16	WW275	35354.66	27463.01	542.00	84	25.60	524.63	531.33							532.25	532.86
62	4	12	4	WW208A	33940.00	21418.00	547.50	84	25.60	524.03	543.84					523.73	524.64	533.78	534.39
62	4	12	8	WW209	35185.63	21804.73	541.00	86	26.21	523.63	534.90					525.15	525.76	534.90	536.43
62	4	13	16	WW211	35353.17	24495.68	540.30	85	25.91	523.54	537.25							530.55	531.16
62	3	15	13	WW120	40448.57	24297.32	540.50	77	23.47	522.82	535.23					522.21	525.87	535.32	538.06
62	4	10	1	WW231	32150.07	21592.84	547.00	81	24.69	522.31	540.90					523.23	525.05	536.94	538.16
61	3	31	3	WW661	36036.47	18154.09	542.00	84	25.60	521.43	532.85								
63	3	4	4	WW402	33636.55	29621.78	543.00	78	23.77	521.36	536.90					521.05	521.66	527.46	527.73
62	4	20	12	WW248	27435.73	25504.84	546.00	84	25.60	521.01	527.71					521.01	527.71		
61	4	28	3	WW731	29760.00	16720.00	557.00	78	23.77	520.73	546.61								
62	3	33	2	OB2-33	39623.07	27986.23	540.00	64	21.60	519.00	537.00							nr	528.11
62	3	7	15	WW107	36480.72	22790.85	539.50	88	26.82	518.16	535.84					516.64	518.77	530.36	533.40
62	4	36	8	WW283	35331.56	28407.30	547.00	85	25.91	518.04	528.71					516.52	519.57	537.25	537.86
61	4	33	9	WW714	30500.00	18955.00	558.00	86	26.21	517.46	551.90	505.57			506.19	526.00	526.61		
62	4	14	3	WW215	32874.79	23329.81	553.00	88	26.82	517.03	545.38					516.42	519.47		
63	3	8	1	WW404	38337.33	31114.15	543.00	85	25.91	516.03	528.06					514.95	521.66	527.76	536.90
62	4	7	13	WW226	25843.36	22716.83	557.00	85	25.91	516.00	529.57					514.94	517.68		
62	4	6	12	WW221	26114.11	20843.92	552.00	78	23.77	515.58	541.03					513.90	521.52	536.75	537.37
62	5	11	15	WW305	23340.13	22743.65	554.00	84	25.60	515.14	541.81					514.38	515.90		
62	3	24	4	WW153	44098.81	25369.46	532.00	85	25.91	514.32	523.16					515.54	516.15	523.16	523.77
62	3	20	14	WS1	37512.19	25787.40	542.72		0.00	513.74	533.10					513.13	519.84	524.11	535.30
62	3	20	13	WS2	37270.25	25869.72	542.42		0.00	513.47	533.52					512.85	518.34	523.83	532.37

Appendix B. Table 2. Wells sorted by completion elevation.

TWP	RG	SC	LS	WELL ID	UTM E mE	UTM N mN	ELEV (m) m.a.s.l	YR 19'	DEEP (m) m	Completion m.a.s.l	W.L. Elv m.a.s.l	Bedrk Elv m.a.s.l	EMF Elv m.a.s.l	Brn Lk Elv m.a.s.l	Mur Lk Elv m.a.s.l	Bvle Elv m.a.s.l	Eth Lk Elv m.a.s.l	Mar Ck Elv m.a.s.l	Snd Rv Elv m.a.s.l
62	3	17	4	WW124	37298.00	23117.63	539.00	86	26.21	513.40	533.65					508.52	514.62	529.86	536.6[illegible]
61	3	33	5	WW601	38820.00	18750.00	548.00	85	26.21	512.95	541.60	512.03				519.34	520.56		
62	5	14	16	WW309	23992.31	24235.11	552.00	85	25.91	512.83	527.62					512.38	516.03		
62	5	2	8	WW304	24059.27	20118.18	552.00	73	22.25	512.68	533.41					512.38	512.99		
62	4	22	15	WW249	31518.58	25999.74	550.00	89	27.13	512.36	537.20	501.23				511.60	513.42	540.86	543.29
62	4	3	2	WW217	31715.22	19887.69	558.00	88	26.82	512.13	550.08	495.82				511.37	513.80		
61	5	25	8	WW5	25701.18	16843.81	549.00	83	25.30	511.97	545.95			511.20	512.73	519.43	520.65	537.42	537.72
61	5	25	16	WW19	25715.00	17885.00	549.50	55	16.76	511.70					509.88				
62	4	30	13	WW262	25803.88	27070.23	553.00	88	26.82	511.09	519.47					508.60	516.42	536.54	538.37
62	4	3	4	WW218	30707.84	19918.46	561.00	80	24.38	510.25	551.86	509.18				509.18	512.23		
62	3	21	1	WW146	40223.01	24681.64	542.00	83	25.30	510.15	535.29					509.39	513.05	527.37	531.64
61	5	26	12	WW18	22585.00	17330.00	553.00	89	27.13	509.72	530.14	0.00			510.63	525.87	526.18		
62	4	25	2	WW277	34824.17	26275.29	556.00	78	23.77	509.67	532.96	502.96				520.03	524.30	541.37	542.59
62	4	30	13	WW258	25803.00	27070.00	553.00	81	24.69	509.57	519.47					508.80	516.42	534.71	537.76
61	5	26	4	WW16	22595.44	16500.00	549.00	89	27.13	509.38	542.29			507.85	510.90	520.96	521.57		
62	4	19	12	WW245	25857.61	25409.16	549.00	88	26.82	509.22	524.62					506.94	517.30		
62	4	19	12	WW246	25857.51	25439.64	549.00	79	24.08	509.07	527.05					506.94	517.30		
62	4	25	2	Bu	34826.78	26252.46	556.00	83	25.30	508.91	533.06					521.25	523.39		
61	5	35	3	WW9	23149.03	18192.25	554.00	88	26.82	508.74	529.62				510.11	521.08	522.30	535.71	536.32
61	4	30	12	WW702	25920.00	17550.00	549.50		0.00	508.66	542.19				509.86	522.07	522.68		
62	5	23	6	WW310	22873.94	25078.73	551.00	88	26.82	508.48	519.00	501.01				506.50	516.25	533.63	536.37
61	5	25	16	TH1	25682.04	17760.43	549.00	75	22.86	508.31	538.09	504.19		507.24	509.38	524.62	526.44	536.50	537.42
62	4	1	5	ARDSCH	33893.24	20354.76	554.00	71	21.64	508.28	545.16	493.65							
62	3	21	14	WW144	39369.74	26089.77	540.00	84	25.60	508.15	533.90					508.39	513.57	529.33	536.95
62	5	23	12	WW313	22662.39	25509.35	550.50	85	25.91	507.98	513.92					507.22	512.10	534.65	535.26
61	4	28	4	WECO	29253.98	16785.70	555.00	75	22.86	507.76	547.84	[illegible]69.05	475.14	486.12	508.67	522.08	525.13	541.59	544.64
62	3	5	4	WW102	37326.70	20157.53	540.00	79	24.08	507.39	537.87	501.60						533.90	
61	5	36	4	WW13	24486.76	18132.11	555.00	81	24.69	506.99	527.57			506.23	509.89	522.08	524.52	543.72	545.25
62	5	10	13	TH184a	22453.50	22820.50	551.28	77	23.47	506.93	531.16				508.91	514.09	518.36	534.52	539.09
62	3	27	4	WW156	40517.47	26484.51	537.50	86	26.21	506.87	528.36					505.80	515.55	525.92	531.40
62	4	8	1	WW227	28898.09	21806.40	558.00	80	24.38	506.79	548.76	[illegible]06.49				518.07	520.20	535.75	537.58
63	4	2	15	UOH 21	32842.25	30779.08	507.00	91	3.10	505.05	505.01								
61	5	35	8	WW8	23967.06	18759.45	554.00	82	24.99	504.93	533.27			503.40	507.06	530.84	533.58		
61	5	35	3	WW15	23151.00	18192.00	554.00	81	24.69	504.62	528.40	0.00		50[illegible]	505.84	522.00	522.61	542.42	543.03
61	4	30	14	WW724	26305.00	17990.00	550.00	79	24.08	503.98	539.64				5[illegible].94	525.01	527.44	541.47	543.29
62	5	25	16	WW316	25483.79	27627.79	551.00	88	26.82	503.60	511.38			502.84	505.89	513.81	518.08		
63	4	2	13	WW510	32130.00	30695.00	510.54		8.07	502.47	504.04								
62	5	26	1	WW317	22733.60	26238.39	565.00	85	25.91	501.75	544.58			500.99	510.14			534.22	534.82
61	4	31	12	WW701	25830.00	18915.00	554.00	88	26.82	501.27	536.64				505.23			540.28	543.33
62	5	24	4	WW311	24178.86	24820.19	551.00	82	24.99	501.17	526.62	490.00		493.70	503.45	508.02	516.56	533.32	539.11
62	5	25	3	WW314	24693.84	26214.52	562.00	81	24.69	501.04	515.98	493.42		500.43	504.09	504.09	518.11	531.22	532.74
62	4	5	5	WW220	27474.76	20416.07	555.00	90	27.43	500.75	535.19				503.79	523.00	523.61	534.88	535.80
61	4	33	1	WW711	30444.00	18456.00	552.00	84	25.60	500.64	544.99				505.06	524.26	524.57		
63	3	8	13	Kr	36825.74	31924.57	546.00	77	23.47	500.58	520.24	496.62			501.80			519.48	534.42
63	3	8	13	Kr	36830.00	31924.57	546.00	92	46.41	500.28	521.40								
62	3	21	5	OB5-21	38921.32	25153.67	538.00	78	23.77	500.20	532.00	497.16			503.25	522.15	523.37	528.55	532.82
61	5	36	1	WW12	25546.29	18357.75	555.00		0.00	499.83	506.23								
63	4	1	4	WW501	33608.83	29619.06	547.00	84	25.60	498.84	520.48				499.45	520.18	521.09		

Appendix B. Table 2. Wells sorted by completion elevation.

TWP	RG	SC	LS	WELL ID	UTM E.	UTM N.	ELEV. (m)	YR	DEEP (m)	Completion	W. L. Elv.	Bedrk Elv	EMP Elv	Brn Lk Elv	Mur Lk Elv	Bvlle Elv	Eth Lk Elv	Mar Ck Elv	Snd Rv Elv.
					mE	mN	m.a.s.l	19'	m	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l	m.a.s.l
61	5	24	4	WW4	24474.42	14854.02	554.00	79	24.08	498.83	529.62			497.31	500.36	518.34	522.30		
62	3	16	15	WW122	39535.03	24411.36	539.50	81	24.69	498.66	531.58				501.40	507.50	509.63	531.88	
61	4	32	8	TH3	28931.97	18731.24	554.00	75	22.86	498.22	543.79	465.61	476.58	487.55	504.32	524.13	525.65	540.28	541.20
61	4	36	7	WW703	34730.00	18515.00	545.00	85	25.91	497.45	519.09				499.28	522.14	523.66	540.43	541.95
62	4	30	5	WW261	25754.00	26747.40	550.00	85	25.91	497.27	546.74	496.00			501.23			530.19	533.24
62	4	25	13	WW276	33892.00	27609.00	542.00	89	27.13	497.04	522.92	495.37			501.77	516.09	517.92	538.65	539.87
61	4	31	2	TH721	26922.00	18085.00	550.50	78	23.77	496.70				482.83	501.43	510.88	516.06		
62	3	10	9	WW114	41852.91	22531.95	539.00	69	21.03	496.48	534.35	494.80			503.34			529.86	531.38
62	4	25	13	WW274b	33892.03	27609.27	542.00	86	26.21	496.28	522.19					517.01	517.62	531.64	532.86
62	4	26	16	WW281	33597.00	27613.13	542.00	85	25.91	495.37	519.14				501.77	517.92	521.88	531.33	535.90
62	3	21	3	WW147	39243.81	24702.39	541.00	86	26.21	493.91	530.64	491.62			500.16	519.66	522.10	532.77	540.09
61	4	28	12	WW726	29105.00	17435.00	551.00	92	28.04	489.28	539.42			488.52	508.33				
62	3	26	8	WW155	43482.35	26676.62	530.40	90	27.43	483.92	514.40	473.71		482.24	484.68	502.97	506.02		
62	5	10	16	TH184b	22453.50	22820.50	551.28	77	23.47	476.91	528.58	468.98	481.48	495.20	508.91	514.09	518.36	534.52	529.09
62	3	33	2	OB2-33A	39635.49	28049.72	540.00	66	20.12	476.30	525.40	472.03			486.05	495.19	498.24	510.74	529.03
61	4	33	4	WW715a	29050.00	18065.00	549.00	85	25.91	470.36	541.41	468.23	477.37	489.87	506.02				
61	4	29	1	OB725	26695.00	17525.00	550.00	78	23.77	469.84	531.35	469.23	473.50	485.69	507.02				
61	4	33	4	WW715	29050.00	18065.00	549.00	85	25.91	469.75	541.26	468.23	476.76	488.95	505.41				
61	4	28	3	WW730	29720.00	16710.00	557.00	79	24.08	469.07	531.27		472.09	486.29	506.40	514.16	521.17	543.42	545.25
61	4	32	1	WW718	28695.00	18335.00	551.00	85	25.91	467.79	543.04	467.18	470.84	486.99	507.11			533.02	533.93
61	4	28	3	WW729	29280.00	16700.00	557.00	78	23.77	467.08	531.43	465.56	472.09	485.68	507.01	517.81	520.25	543.11	544.03
61	3	31	3	TH657	35990.60	18176.67	542.00	91	27.74			496.89							
61	4	30	12	FK SCHL	25797.[illegible]	17851.25	550.00		0.00			507.94			509.77	526.23	525.84	539.64	540.55
61	4	35	4	TH738	32360.00	18375.00	549.00	92	28.04			0.00			509.02	528.27	530.10	541.38	542.29
61	5	23	16	TH15	24100.00	16360.00	549.00	84	25.60			509.38				523.40	524.01		
61	5	26	12	TH17	22585.00	17331.60	553.00	89	27.13			0.00				524.35	526.18	538.67	539.28
62	3	4	13	RCA30	38791.45	21271.70	536.40	76	39.50		536.40	499.90				522.40		532.90	
62	3	7	15	TH106	36419.74	22782.91	539.50	68	26.82			496.83						531.58	531.88
62	3	13	16	TH117	45081.87	24564.70	540.00	90	27.43			495.50						530.85	
62	3	15	1	TH121	41903.85	23075.15	534.68	66	20.12			497.49				5[illegible].52	512.73	525.54	527.[illegible]
62	3	16	3	TH123	39198.13	22998.99	536.57	65	20.12			493.59		495.42	497.56			526.82	530.4[illegible]
62	3	18	7	TH127	37080.00	23745.00	538.00	66	20.12			493.50							
62	3	18	13	TH128	35481.37	24494.83	539.34	66	20.12			503.37							
62	3	19	7	TH132	37087.08	25414.58	541.00	66	20.12			495.[illegible]			499.5[illegible]	513.87	517.53		
62	3	22	1	TH151	41805.65	24875.86	540.00	90	27.43			495.65				515.92	517.14		
62	3	22	11	TH150	41115.00	24655.00	538.60	66	20.12			497.05							
62	3	24	4	TH154	43633.86	24683.31	533.00	66	20.12			495.61				507.09	512.27	524.47	526.90
62	3	27	4	TH157	40368.40	26268.01	537.85	66	20.12			499.75			Log off scale	511.64	515.60		
62	3	28	5	TH158	38838.05	27609.18	540.00	66	20.12			495.85			Log off scale				
62	3	31	16	TH161	35814.43	29375.22	546.00	80	24.38			495.[illegible]		[illegible]	502.72	511.86	513.08	535.46	537.[illegible]
62	3	33	8	TH164	40228.[illegible]	28563.[illegible]	538.00	78	23.77			488.[illegible]			486.79	[illegible].62	[illegible]	525.2[illegible]	52[illegible]
62	3	33	15	TH163	39561.09	29391.36	540.50	78	23.77			491.66		497.93	5[illegible].92	515.20	517.94	526.26	52[illegible]
62	3	34	2	TH167	4[illegible].33	27836.31	538.79	66	20.12			44[illegible]			[illegible].75	[illegible]	499.99	52[illegible]	52[illegible]
62	3	35	13	TH168	41920.65	29459.32	538.50	78	[illegible]			497.[illegible]				[illegible]	50[illegible].02	528.64	53[illegible]

Appendix B. Table 2. Wells sorted by completion elevation.

TWP	RG	SC	LS	WELL ID	UTM E mE	UTM N mN	ELEV (m) m.a.s.l	YR 19'	DEEP (m) m	Completion m.a.s.l	W.L. Elv m.a.s.l	Bedrk El. m.a.s.l	EMP El. m.a.s.l	Bm Lk El. m.a.s.l	Mur Lk El. m.a.s.l	Bve El. m.a.s.l	Eth Lk El. m.a.s.l	Mar Ck El. m.a.s.l	Snd Rv El. m.a.s.l
62	4	2	8	TH2	33534.41	20454.59	554.00	75	22.86			492.43			493.9[illegible]			[illegible]	[illegible]
62	4	2	12	TH274	32328.00	20562.00	556.00	87	26.52							524.30	524.61		
62	4	5	5	TH220A	27474.00	20416.00	555.00	90	27.43			488.00			502.14	523.30	524.52		
62	4	7	13	TH225	25794.73	22795.34	556.50	77	23.47			501.03				514.44	516.58	539.[illegible]	540.96
62	4	12	12	TH7A	33932.00	22441.00	543.00	89	27.13			502.46						534.4[illegible]	538.12
62	4	19	1	TH242	27232.54	24559.18	551.00	80	24.38					506.50		514.73	517.17	528.14	529.66
62	4	19	12	WW284	25855.00	25439.64	549.00		0.00		533.42								
62	4	19	12	TH247	25862.78	25383.78	549.00	88	26.82			494.35				509.68	516.39	530.41	531.93
62	4	25	9	TH218	35428.00	27160.00	548.33	66	20.12			496.27			502.61	512.97	516.33		
62	4	36	13	TH231	33812.99	29388.81	547.00	77	23.47			493.05			496.10			538.77	540.29
62	5	9	15	TH185			548.80	77	23.47			490.89			505.52	514.97	518.62	532.95	536.61
62	5	11	4	TH306	24055.00	21300.00	551.80	77	23.47			471.03	478.34	493.58	506.99	518.88	519.19	534.43	536.56
62	5	13	4	TH308	24253.73	22950.96	559.30	68	20.73										
62	5	26	4	TH319	22498.74	26135.13	550.00	72	21.95			499.71		505.80	508.55	514.95	517.69	538.42	539.03
62	5	3	5	TH330	20920.00	201[illegible].00	554.00	77	23.47			479.02						542.72	543.33
63	3	4	4	TH401	38443.97	29470.99	543.00	80	24.38			491.79		496.37	500.33			526.85	531.42
63	3	9	4	TH407	38405.33	31259.15	543.15	78	23.77			493.16		496.82	499.87	517.55	521.81	524.86	535.53
63	3	18	9	TH410	36680.00	33840.00	551.00	88	26.82			500.71				525.09	526.62	529.05	535.76
63	4	2	1	TH502	33461.80	29453.42	547.00	80	24.38			492.75			497.62				
63	4	5	4	TH508	27003.32	29394.54	550.50	80	24.38			475.52			493.20			535.87	541.66
63	4	8	16	WW509	28440.00	32460.00	541.00	89	27.13		522.10	480.00							

Appendix B, Table 3. Flux summary; data base for Figure 6.1.

Based on 1992/93 water year

UOH	Depth to W.L. m	Recharge Calculated Depth m	pp ratio pp/d	S_y	Equiv depth m	flux m/s	% of pp	Accretion Measured Depth m	pp ratio pp/d	Equiv depth m	flux m/s	% of pp	ET Depth Measured m	pp ratio pp/d	Equiv depth m	flux m/s	% of pp	Net Accr. A-ET Equiv. d m	m/s	% of pp
1	7.94	-0.73	0.55	0.045	-0.03	1.08E-09	8.19	0.49	0.73	0.02	7.20E-10	6.17	0		0			0.022	7.20E-10	6.17
2	3.71	-0.7194	0.56	0.03	-0.02	7.62E-10	5.40	0.75	0.47	0.02	7.93E-10	6.33	-0.369	0.96	-0.01	-3.91E-10	3.12	0.011	4.03E-10	3.22
3	2.08	-0.60	0.67	0.03	-0.02	6.53E-10	4.48	0.95	0.37	0.03	1.04E-09	8.03	-0.4957	0.72	-0.01	-5.41E-10	4.19	0.014	4.97E-10	3.84
5	1.35	0.3243	1.23	0.2	-0.06	2.18E-09	16.22	0.53	0.68	0.11	3.53E-09	29.40	-0.3235	1.11	-0.06	-2.17E-09	18.07	0.041	1.36E-09	11.32
7	4.10	-1.05	0.38	0.03	-0.03	1.01E-09	7.88	0.64	0.56	0.02	6.22E-10	5.40	0		0			0.019	6.22E-10	5.40
8	3.73	-1.64	0.24	0.045	-0.07	2.57E-09	18.41	2.11	0.17	0.09	3.30E-09	26.46	-0.8605	0.42	-0.04	-1.35E-09	10.82	0.056	1.95E-09	15.65
9	4.20	-0.70	0.57	0.045	-0.03	1.14E-09	7.93	0.81	0.44	0.04	1.31E-09	10.14	-0.32311	1.11	-0.01	-5.23E-10	4.06	0.022	7.83E-1[illegible]	6.07
11	1.38	-1.07	0.37	0.03	-0.03	1.16E-09	3.05	3.00	0.12	0.09	3.24E-09	25.15	-2.18898	0.16	-0.07	-2.36E-09	18.34	0.024	8.76E-10	6.81
11a	1.38			Variable						0.165	5.92E-09	46.02			-0.14531	-5.22E-09	40.59	0.019	6.99E-10	5.43
12	3.24	0.51		0.03	0.02	5.44E-10		1.99	0.18	0.06	2.10E-09	16.66	-3.31266	0.11	-0.10	-3.51E-09	27.76	-0.040	-1.40E-09	-11.10
14	2.85	-1.05	0.38	0.03	-0.03	1.04E-09	7.86	1.09	0.33	0.03	1.08E-09	9.17	-0.3852	0.93	-0.01	-3.81E-10	3.23	0.021	7.65E-10	5.94
15	2.62	-0.32	1.23	0.045	-0.01	5.21E-10	3.65	1.14	0.31	0.05	1.83E-09	14.34	-1.02195	0.35	-0.05	-1.64E-09	12.85	0.005	1.92E-10	1.50
23	5.75	-0.77	0.52	0.045	-0.03	1.20E-09	8.71	0.47	0.75	0.02	7.34E-10	5.98	-0.01371	25.90	0.00	-2.13E-11	0.17	0.021	7.13E-10	5.81
25	4.63	-0.60	0.67	0.045	-0.03	9.51E-10	6.70	1.26	0.28	0.06	2.02E-09	15.88	-1.00545	0.35	-0.05	-1.61E-09	12.64	0.012	4.12E-10	3.24
26	2.15	-0.19	2.07	0.20	-0.04	1.27E-09	9.64	0.41	0.89	0.08	2.71E-09	22.36	-0.32192	1.14	-0.06	-2.12E-09	17.98	0.018	5.88E-10	5.00