SVERIGES GEOLOGISKA UNDERSÖKNING
Rapporter och meddelanden nr 21

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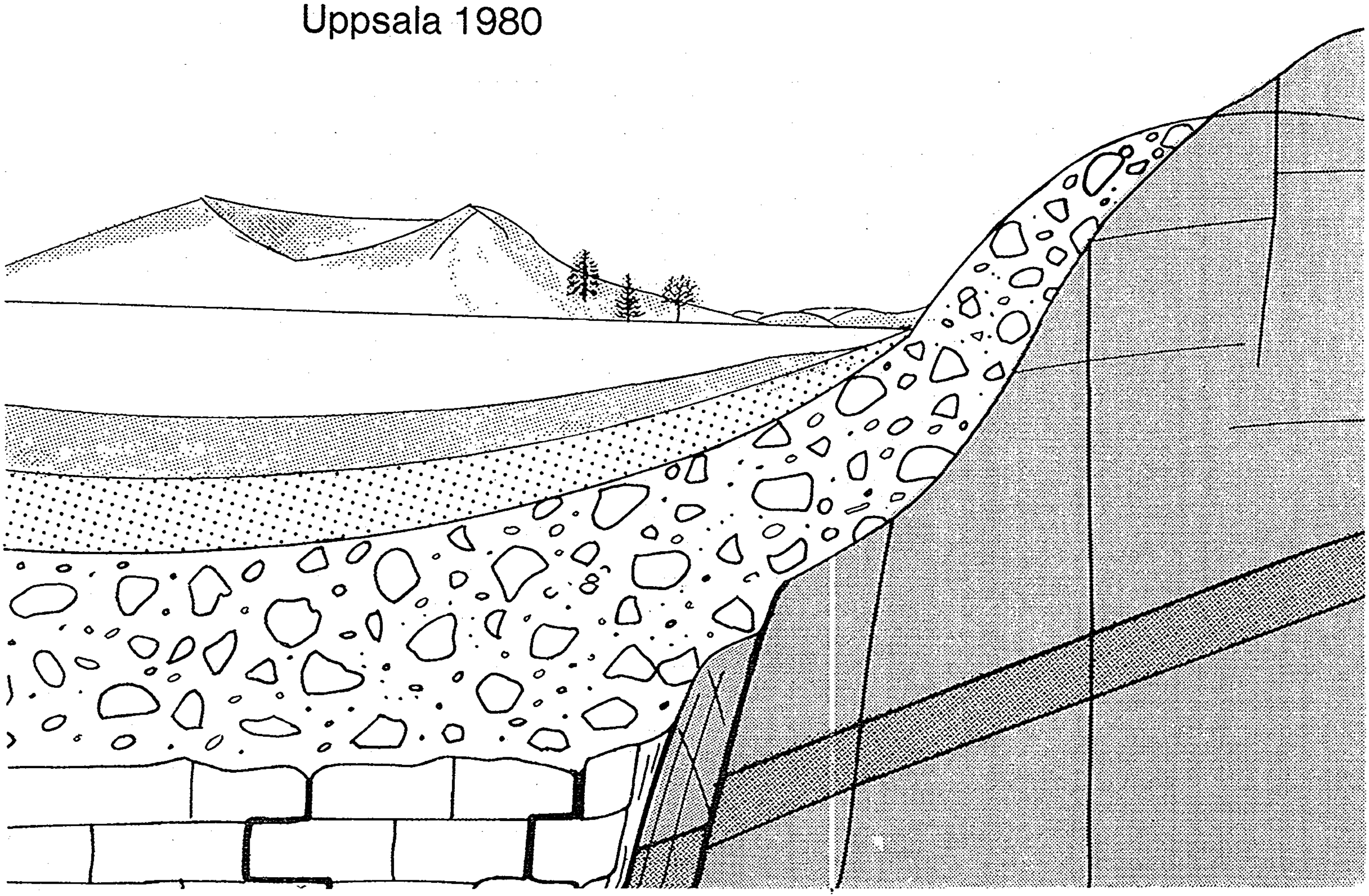
Ground-water-level fluctuations as a measure of the effective porosity and ground-water recharge

Uppsala 1980



GROUND-WATER-LEVEL FLUCTUATIONS AS A MEASURE OF THE EFFECTIVE POROSITY AND GROUND-WATER RECHARGE

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CONTENTS

1. INTRODUCTION

The ground-water recharge and effective porosity in till soils are crucial for the ground-water balance in vast areas in Sweden. As pointed out by Olsson (1980) inter alios, the soil cover often acts as an infiltration reservoir for the underlying bedrock and in hydrogeological studies of the till itself, it is obvious that these properties are of great interest. The fluctuations of the ground-water level are a characteristic feature, about which information may easily be obtained. The fluctuations in an aquifer are governed mainly by the infiltration recharge and the hydraulic properties of the formation, which makes it suitable to evaluate the level fluctuations with respect to these parameters.

The main purpose of this papaer is to discuss the ground-water fluctuations as a measure of the effective porosity of and the infiltration recharge to an aquifer. The long-term variations in ground-water level in an area at Stensele in the County of Västerbotten have been roughly examined, in order to illustrate the discussions and theories. The fluctuation analysis is not the best method of evaluating the parameters, but the data needed are usually easy to record and thereby the method may be usable for many purposes.

2. GENERAL ASPECTS OF RECHARGE

The mechanism of the recharge depends on whether the formation is exposed to the atmosphere or covered with other, more or less permeable deposits. In the first case, which will be discussed in this paper, the recharge is due to the percolation of precipitation or surface water through the unsaturated zone down to the ground-water table. The amount of the recharge to the uncovered formation is mainly determined be either the:

- infiltration or percolation capacity or the
- available precipitation

Hence the recharge in the uncovered formation depends on both the hydrometeorological and the hydrogeological conditions in the area. In terms of rises in the ground-water level, the response is also governed by the hydraulic conductivity and effective porosity of the formation.

2.1. Methods for recharge determinations

The ground-water recharge is of great importance for many purposes. This is the case in solving problems concerning water supply. It is also the case in many problems of engineering geology, for instance, in determining the steady-state, ground-water inflow into underground constructions. However, no exact method exists and many different approaches are made, for example, the following:

- The calculated value of the available precipitation (absolute precipitation reduced by the amount of evapotranspiration), as being the potential for ground-water recharge (Eriksson and Johansson 1978).
- The obtained values of the infiltration capacity of different soils (von Brömssen 1968).
- Water-balance studies of various kinds (Axelsson and Carlsson 1979, Olsson 1979).
- Direct measurements, using lysimeters (von Brömssen 1968, Soveri 1980).
- Infiltration coefficients (von Brömssen 1968).
- Determinations of the ratio between discharge and recharge areas, combined with the available precipitation (Eriksson 1977).
- Studies of the water-level response (Olsson 1979).

For very local measurements, the lysimeter method seems to be the most accurate, but such facilities do not yet exist in Sweden. I hope that lysimeters will be installed in the near future in those areas which are covered by the ground-water stations run by the Geological

Survey of Sweden. Such measurements would improve our knowledge of the phenomenon of ground-water recharge. In Finland, the Water Board now makes this kind of observation and run regular, water-level stations; at the time of writing there are more than 30 stations in operation (Soveri 1980).

On the local scale, the method based on the water-level response also seems to be useful, especially after being collated with the above-mentioned measurements with lysimeters.

For studies on the regional scale, the study of the ground-water balance and the method using the ratio between recharge and discharge areas are most suitable for determining the ground-water recharge. The main disadvantage of these methods is that they are dependent to a very high degree on the accuracy of the reliability of other hydrogeological and hydrometeorological parameters, such as the hydraulic conductivity of the formations, the hydraulic gradient, precipitation, evapotranspiration and so forth. However, in many Swedish soil types, all the available precipitation will infiltrate and thus the ratio between recharge and discharge areas will be a direct measure of the recharge.

The first-mentioned methods, available precipitation and infiltration capacity, together with the infiltration-coefficient method, only give very rough figures of the recharge. In Swedish soils (except clay and exposed bedrock), the infiltration capacity is usually much higher than the available precipitation (Olsson 1979, Eriksson 1980), which indicates that all the available precipitation may infiltrate in recharge areas. The total amount of the available precipitation usually infiltrates, exept that part which falls on discharge areas and thus remains as surface-water.

An attempt to classify these different methods on the basis of their reliability is summarized in Table 1. In this table, those methods which are fairly accurate are especially distinguished, as well as

Table 1. The reliability of different methods for the determination of ground-water recharge. Accurate methods are marked by 0 and methods which under- or over-estimate the recharge are marked by - or +.

Method	Reliability
Local scale methods	
Lysimeters	0
Water-level response	- to +
Infiltration capacity	+
Regional scale methods	
Available precipitation	+
Water balance	0
Discharge/Recharge areas	0
Infiltration coefficients	+ to -

those which under- or over-estimate the recharge.

3. WATER BALANCE OF AN AREA

As regards the ground-water balance, the recharge of the aquifer is of fundamental importance. According to Brown et al. (1972), the long-term, average discharge from a ground-water reservoir can be presumed to be in equilibrium with the long-term, average recharge. The potential for recharge to the ground-water régime in an area depends on the amount and pattern of annual precipitation in relation to the evapotranspiration and to the occurrence of surface or subsurface inflow from or to adjacent areas. The amount that actually contributes to the ground-water recharge varies seasonally and from year to year (Brown et al. 1972). Fig. 1 shows a schematic soil profile

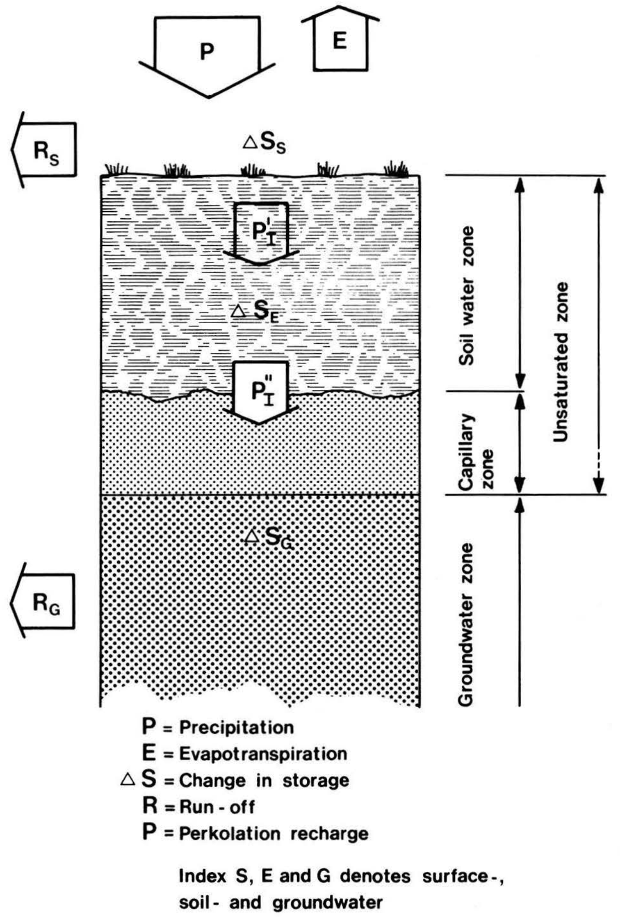

Fig. 1. Soil profile, illustrating the main hydrological processes included in and affecting the recharge.

with the division of the subsurface water.

For an isolated catchment area, the following classical equation for the water balance may be stated:

$$P = E + R \pm \Delta S \qquad (1)$$

where P = precipitation
E = evapotranspiration
R = run-off
ΔS = change in water storage during the computed period

This equation may be further divided for the different, water-bearing systems:

$$P = E + R_S + R_G \pm \Delta S_S \pm \Delta S_E \pm \Delta S_G \qquad (2)$$

where R_S = surface-water run-off
R_G = ground-water run-off
ΔS_S = change in surface-water storage
ΔS_E = change in soil-water storage
ΔS_G = change in ground-water storage

If there is no surface-water run-off and no surface-water storage, all the precipitation in exess of the evapotranspiration (available precipitation) will join the subsurface-water system, in the aerated zone of soil-water and/or in the saturated zone of ground-water. Consequently, equation (2) will in this case be reduced to:

$$P = E + R_G \pm \Delta S_E \pm \Delta S_G \qquad (3)$$

As the soil-water storage has priority over the ground-water storage, the following conclusions may be drawn from equation (3):

$P - E < S_{Emax} - S_{Ea}$ The available precipitation will increase the soil-water storage by an amount equivalent to $P - E$.

$P - E > S_{Emax} - S_{Ea}$ The available precipitation will fill the soil-water storage up to its maximum, that is, by an amount equivalent to $S_{Emax} - S_{Ea}$ and, additionally, the ground-water storage will be recharged by an amount equvalent to $(P-E-(S_{Emax}-S_{Ea}))$.

These conditions show that no ground-water recharge will take place unless the soil-water storage is filled up to its maximum. The ground-water storage will only be recharged if the amount of the available precipitation is greater than the amount needed to fill the soil-water storage.

In order to obtain a positive change in the ground-water storage, it is necessary that the following condition should be fulfilled:

$$\Delta S_G > 0; \quad P - E - R_G > 0 \tag{4}$$

that is, when the amount of the available precipitation is greater than the ground-water run-off. As the increase or decrease in ground-water storage is a pure function of the variation in depth of the ground-water table, the change in level will be a measure of the recharge. However, an increase in ground-water storage only takes place when the recharge is greater than the ground-water run-off. There may still be a ground-water recharge even when the table is falling. In fact, ground-water recharge occurs when:

$$P - E > S_{Emax} - S_{Ea} \tag{5}$$

Hence, ground-water recharge may occur even if the ground-water level is falling, that is, when:

$$\Delta S_G > 0 \tag{6}$$

$$R_G > P - E - (S_{Emax} - S_{Ea}) \tag{7}$$

Consequently, it must be possible to determine the recharge by measuring the variations in ground-water level. The time-lapse of the ground-water level for different amounts of recharge is schamatically shown in Fig. 2.

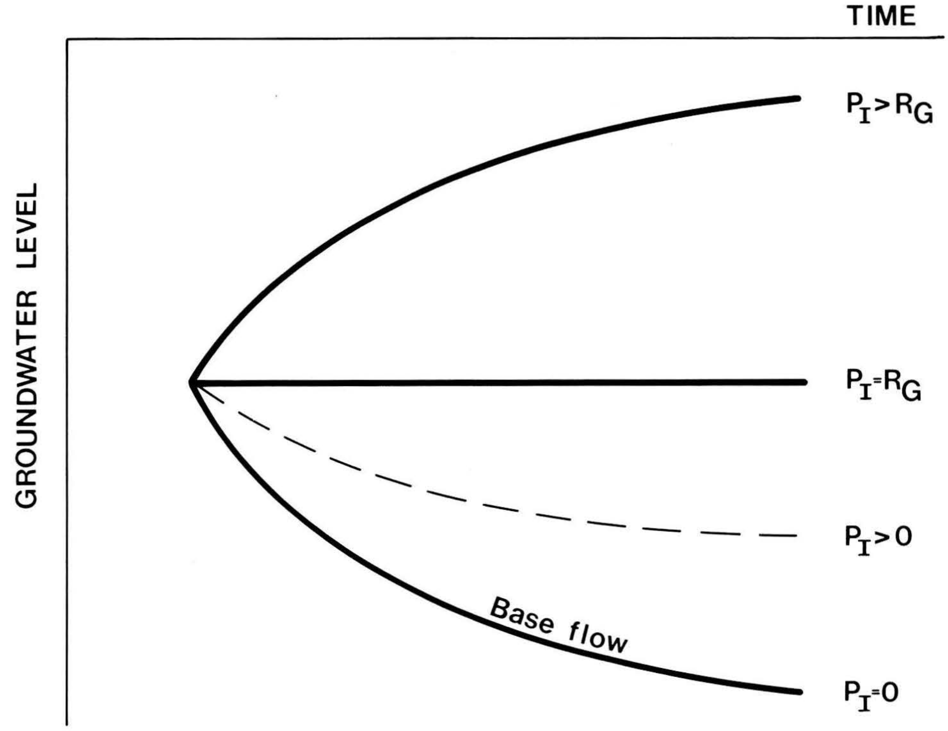

Fig. 2. Time dependent variation in ground-water level. When the recharge is equal to the discharge, the ground-water level is in equilibrium. A rising level is indicated by $P_I > R_G$ and a falling level by $P_I > 0$. Without recharge, $P_I = 0$, the variation in ground-water level is governed by the basic flow from the aquifer.

4. GROUND-WATER DYNAMICS

In order to be able to regard the variations in ground-water recharge as a function of the level fluctuations, it is necessary to study the determinants of the fluctuations. Generally speaking, the changes in head have the following form:

$$\Delta h = F(P_a, \Theta, R_G, t, f(x)) \qquad (8)$$

where Δh = change in ground-water level

$F()$ = function

P_a = available precipitation

Θ = available pore space

R_G = ground-water run-off

t = time

$f(x)$= a residual function, including other effects, such as barometric efficiency etc.

The change in ground-water level is a function of a large number of parameters, mainly the available precipitation, available porespace, ground-water run-off and time.

As stated above, all the available precipitation not needed to fill the soil-water storage will take part in the ground-water-recharge process.

On the basis of the theories presented in the literature (Todd 1959, Davies and De Wiest 1966, Raudkivi and Challander 1976 and many others) regarding the soil-water storage, the available pore space for ground-water recharge will be equal to the effective porosity of the soil. Due to the vertical, hydraulic conductivity for aerated conditions of the soil, there must be a certain time-lapse between the occurrence of the available precipitation and the response of the ground-water storage. Assuming a soil-water storage filled up to its maximum, the general opinion is that the whole amount of the available precipitation will take part in the recharge and that the pore space available will be equal to the effective porosity of the soil. The difficulty in this part of the problem will obviously be the time-lapse between the occurrence of the available precipitation and the water-level response. Depending on the aerated, vertical conductivity of the soil and the thickness of the aerated zone, the time will vary considerably. The relation between these parameters is schematically shown in Fig. 3.

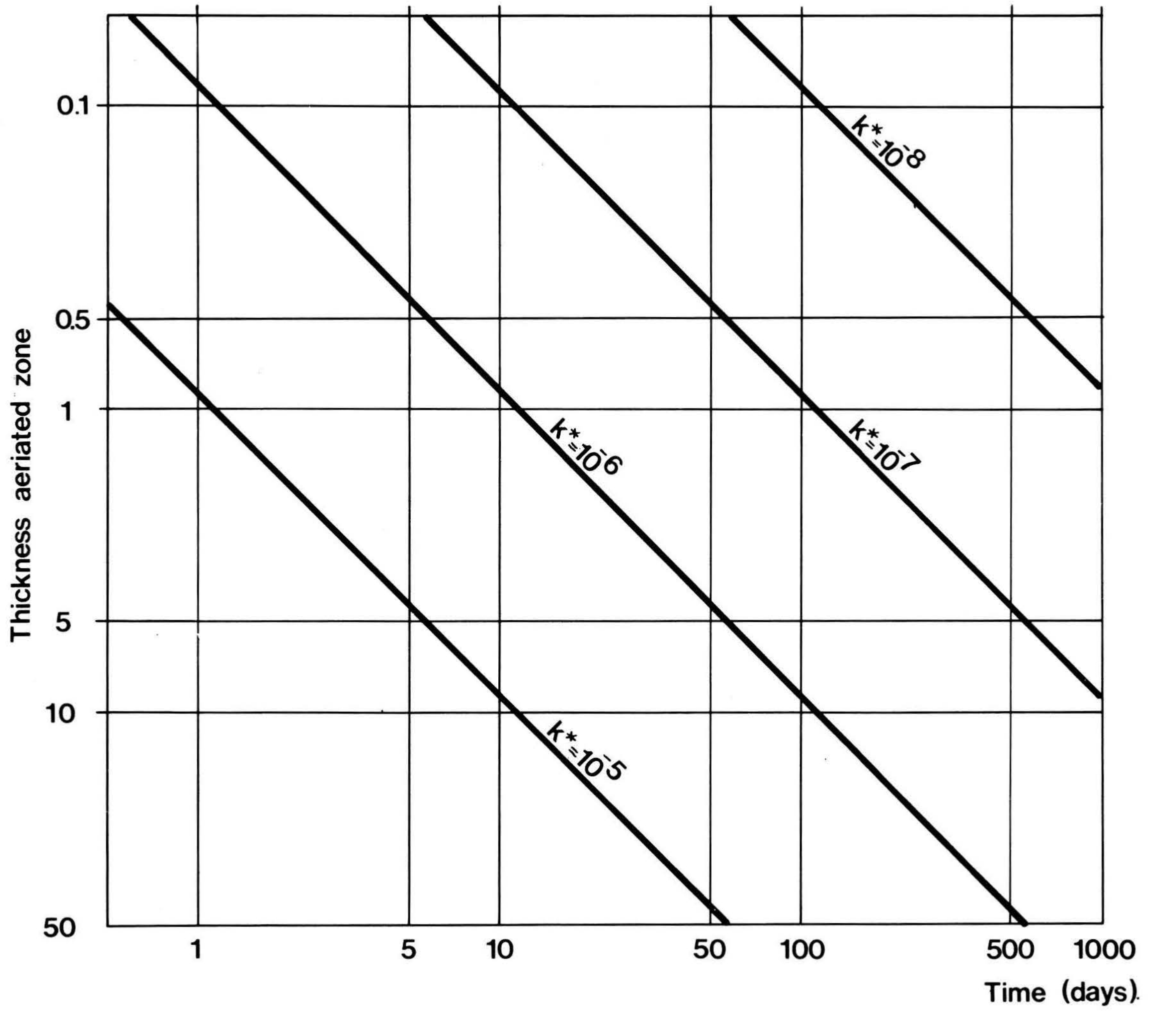

Fig. 3. Percolation time as a function of the thickness of the aerated zone for different values of the unsaturated, vertical conductivity (k^+).

5. THE PROCESS OF INFILTRATION

By studying an area of the ground on which water is available for infiltration, it is possible to establish a relation for the vertical movement of the water, The points at which water may enter the soil profile are exclusively made up of the pore system. Replacing the natural pore system by a system of parallel tubes of equal diameter makes it possible to use an approach based on the Hagen-Poiseulle law. The possibility and validity of such a substitution have been verified by Versluys (1915). According to the Hagen-Poiseulle law, the following relation is valid for a single tube saturated with water:

$$v = \frac{\gamma\, d^2}{32\, \eta}\, I \tag{9}$$

where v = the mean velocity of the water flow in the tube
γ = the density of the water
d = the diameter of the tube
η = the dynamic viscosity of the water
I = the hydraulic gradient

The water flow in a system of tubes composed of N tubes in unit area thus becomes:

$$Q = \frac{\gamma\, d^4\, \pi}{128\, \eta}\, NI \tag{10}$$

A combination of these equations with the law of Darcy will give an expression for the gross conductivity (k) of the infiltration surface and for the pore conductivity (k_p). However, during saturated conditions

$$k = \frac{\gamma\, \pi\, d^4\, N}{128\, \eta} \tag{11}$$

$$k_p = \frac{\gamma\, d^2}{32\, \eta} \tag{12}$$

Thus, the gross conductivity is in proportion to the fourth power of the pore diameter and the pore conductivity to the square of the diameter. As the infiltration takes place under the influence of gravity, the true mean velocity of the flow through the tubes will be:

$$v = \frac{\gamma\ d^2}{32\ \eta} \tag{13}$$

Assuming the percolation distance, L, the time needed for the percolation under saturated conditions is then:

$$t = \frac{32\ \eta\ L}{\gamma\ d^2} \tag{14}$$

These conditions are valid for an idealized soil profile when the water routes are composed of a certain number of circular tubes of equal diameter. However, the pore system in a natural soil is not as uniform as is indicated by these equations. Normally, the variations in pore diameter are considerable and, assuming that the pore-size distribution is equal to the grain-size distribution, it is then possible to use the grain-size distribution in order to determine the dispersion in percolation due to variations in pore size.

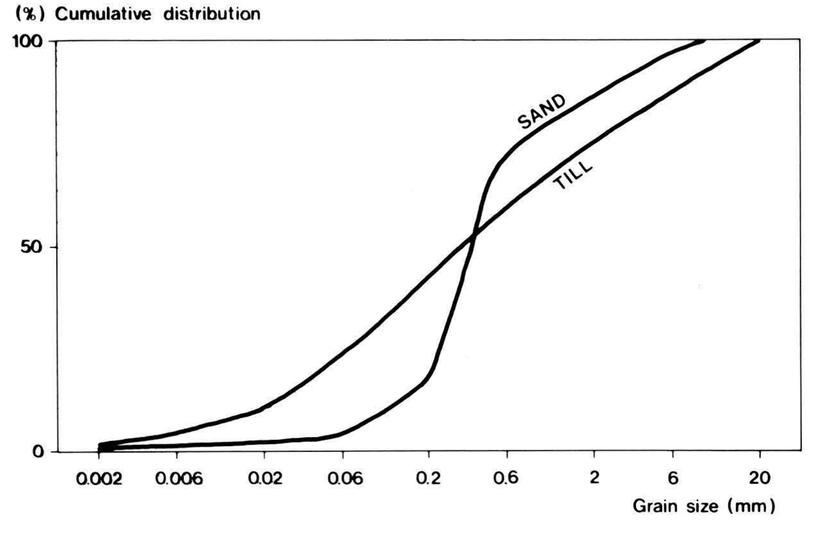

Fig. 4. Grain-size distribution of till (A) and sand (B).

Fig. 4 shows the cumulative distribution of grain sizes for two soils commonly found in Sweden. The first is a silty-sandy till and the second is a shore sand. The grain-size distributions for both soils are nearly log-normally distributed as shown in Fig. 5.

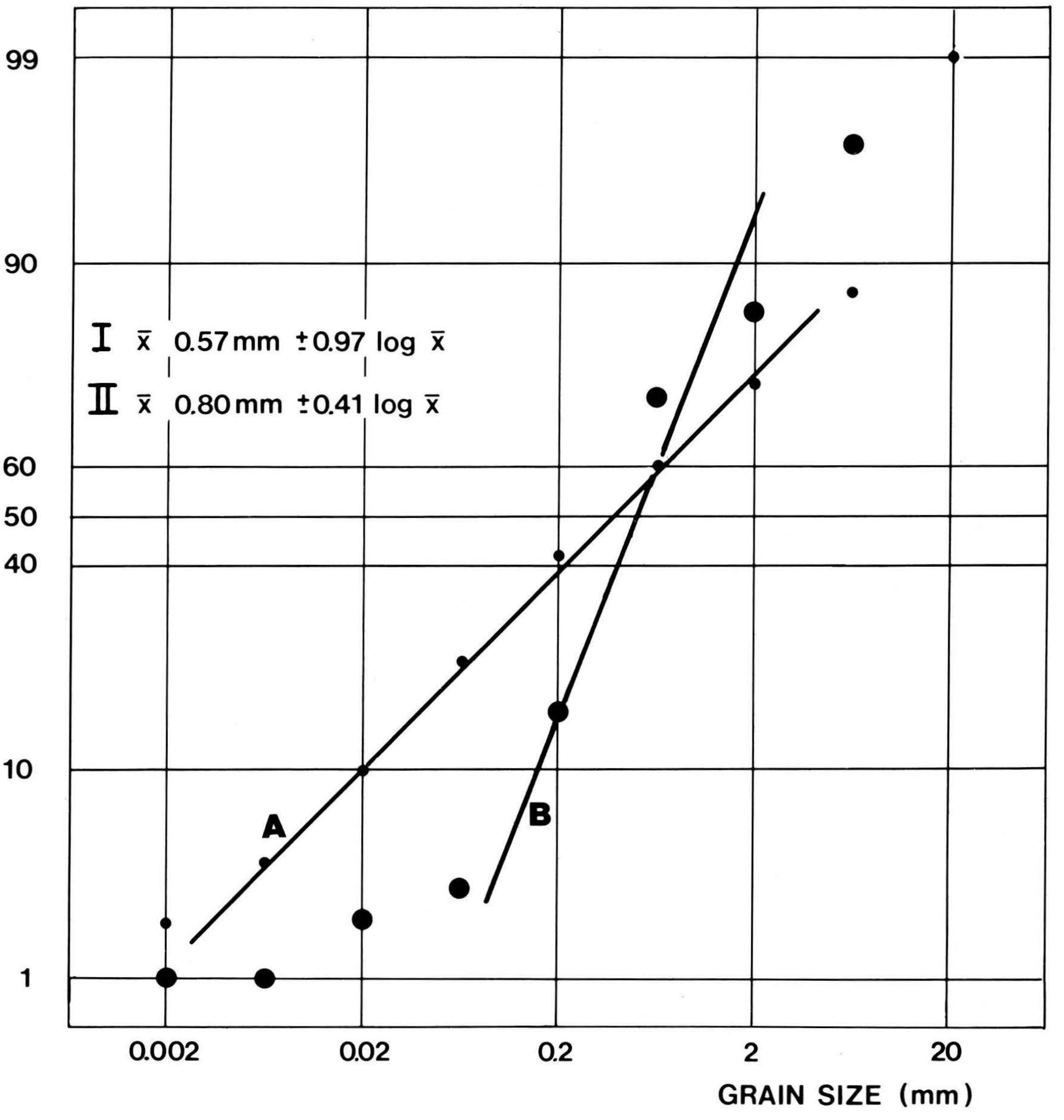

Fig. 5. Log-normal distribution of the grain sizes of till (A) and sand (B).

According to Hazen (1928), the d_{10} value for a porous medium seems to be a fairly accurate measure of the median pore-size. This is valid for saturated soils with a d_{60}/d_{10} ratio of less than 2. Neither of the soils used in this study is as much sorted as this: the till has

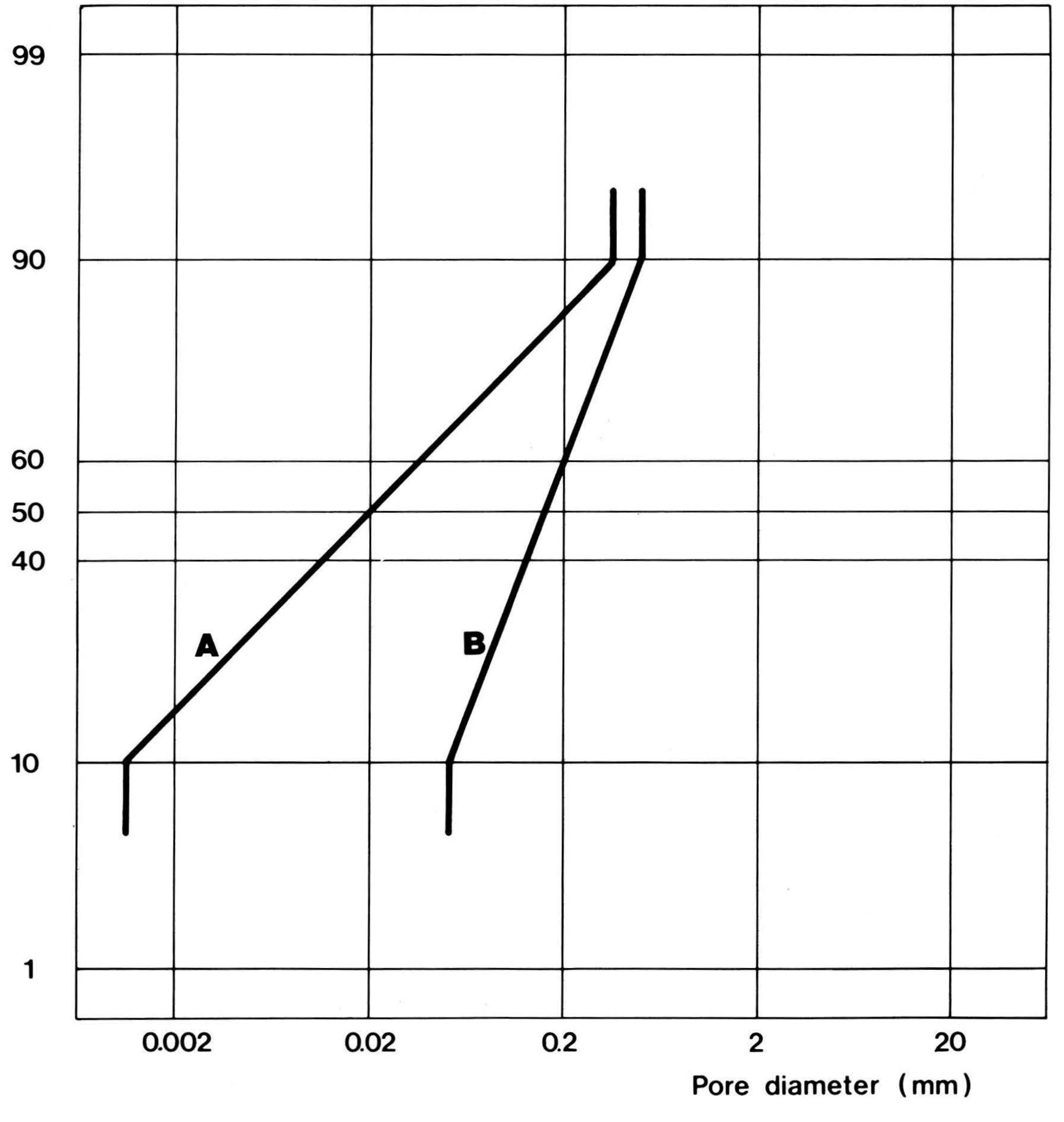

Fig. 6. Log-normal distribution of pore sizes for till (A) and sand (B).

a d_{60}/d_{10} ratio of as much as 30, while the sand has a ratio of 4. These rather low degrees of grading will probably result in the pore-size distributions not being log-normally distributed over the entire interval. For that reason, the distribution curves shown in Fig. 6 have been chosen for the pore diameters. Nevertheless, the median pore diameters for both soils are assumed to be equal to the d_{10} values of the grain-size distribution.

On the basis of these hypothetical distribution curves, the relative time needed for the percolation of a specific amount of water has been calculated and the results are shown in Fig. 7. As these figures show, the percolation rate is fairly high in the till. The time needed for 80 per cent of the available water to reach the ground-water level at a certain depth is 12.6 time units for the sand and 33.1 time units for the till. Accordingly, for 100-per-cent percolation, the times

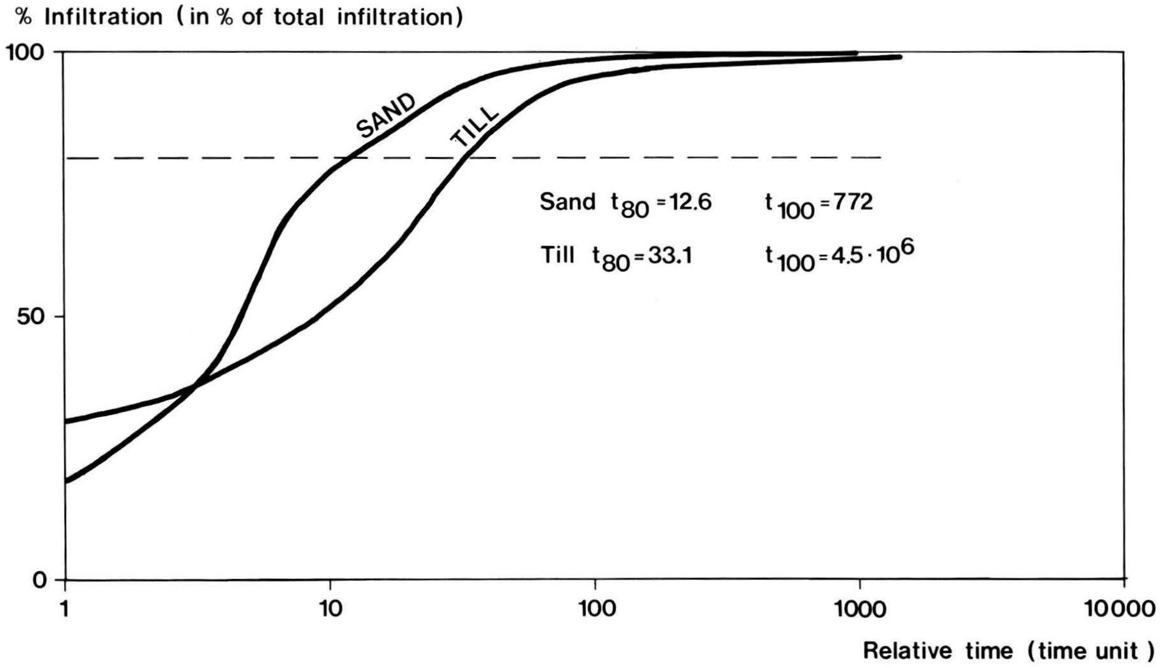

Fig. 7. Relative time needed for the percolation of a unit of water through a unit-thick, aerated zone for sand and till.

are 772 and 4.5 10^6 time units respectively.

However, due to the presence of a few tubes of large diameter in the till, the recharge to the aquifer will be faster in the initial stage of the percolation. On account of the distribution of the pore diameter, single pores of large diameter are of great importance for the percolation process. A pore with a diameter twice as large as another can conduct 16 times as much water in unit time. The ungraded structure of the till soil does not influence the arrival time of the water, which will be about the same for both soil types, as they have about the same, maximum, pore diameter. The effect of the grading will mainly influence the dispersion and then take the form of a pronounced decay in the poorly sorted till soil.

On comparing equations (11) and (12), the difference is seen to be in proportion to the cross-sectional areas of the tubes, which may be regarded as a measure of the two-dimensional porosity. In reality, the difference between the gross and the pore conductivity is due to the effective porosity of the soil. Hence, the pore and gross conductivities are related to each other as follows:

$$k = \theta\, k_p \qquad (15)$$

where k = gross conductivity
θ = effective porosity
k_p = pore conductivity

The difference between equations (11) and (12) will be:

$$k = N \frac{\pi d^2}{4} k_p \qquad (16)$$

where N = the number of pores in unit area
d = the diameter of the pores

Equation (16) is, however, not a true description of the effective porosity of a natural soil, as the distribution of the pore diameter, compaction, etc. affect the effective porosity to a very high degree.

Fig. 8 shows the percolation time as a function of the pore diameter and the depth to the ground-water table. Using the d_{10} values for the different soil types as the median pore diameter, the arrival times will be as indicated in the figure. This gives 3 min for the sand and 3.5 hr for the till soil. These times are however, too short, due mainly to the following circumstances:

- The pore system in which the percolation takes place does both have pores of equal diameter.
- The percolation rate in a single pore will be governed by the narrowest sections of the pore tube.
- The pore system is usually not saturated with water, which

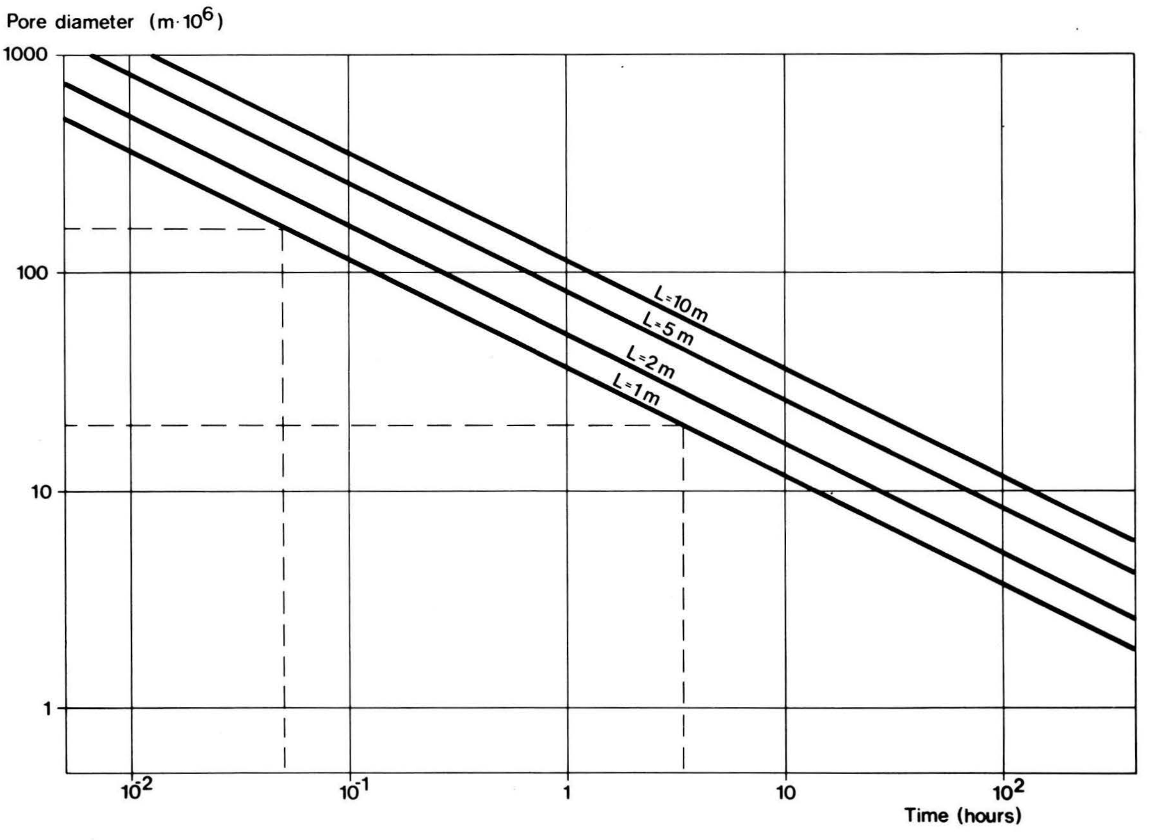

Fig. 8. Percolation time versus pore diameter for different values of the thickness of the aerated zone. The time needed for percolation in sand and till with median pores as in Fig. 6.

is a presumption for the validity of the Hagen-Poiseulle law. As pointed out by Raudkivi and Challander (1976), Todd (1959) and others, the conductivity depends to a very high degree on the soil moisture, with a much lower conductivity when the air content is high.

- For unsorted soils, no verified relation exists between the grain-size and the pore-diameter distributions. Thus, Hazen´s relation is not valid.

The effect of these conditions is that the hydraulic conductivity of the soil under aerated conditions is lower than in a saturated soil. Thus, the times obtained from calculations based on the Hagen-Poiseulle law will be too short, compared with the actual conditions at the percolation. However, the calculations give a rough idea of the effects of the pore-size distribution, time-lapse and decay.

The degree of saturation is much more important for a coarse, graded soil than for a unsorted soil, because of the differences in effective porosity. This condition may be illustrated as shown in Fig. 9.

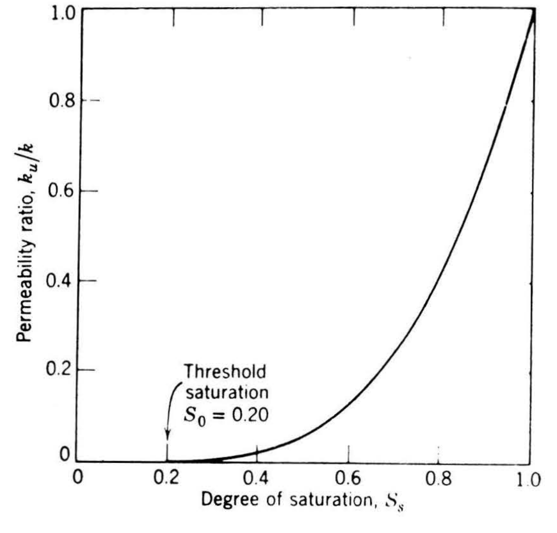

Fig. 9. The effect of the degree of saturation on the unsaturated conductivity (from Todd 1959).

Thus, percolation through the aerated zone in graded formations is usually very slow, while in ungraded formations it may be much faster and more distinct. These conditions have been pointed out by Andersen and Sevel (1974), Aneblom and Persson (1979) and others.

6. VARIATIONS IN GROUND-WATER LEVEL

5.1. General conditions

For more than 10 years, the ground-water levels in different geological environments have been recorded at a great number of measuring stations. These stations are included in the National Swedish Groundwater Network and the stations are run by the Geological Survey of Sweden (Nordberg and Persson 1974). On the basis of these long-term observations, Nordberg and Persson (1976) have distinguished four different regions for the annual ground-water fluctuations. These regions are shown in Fig. 10. In each of the regions, the annual

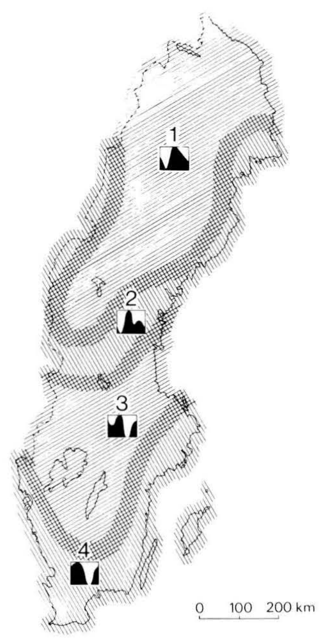

Fig. 10. Schematic, yearly variation in ground-water level (Nordberg and Persson 1976).

variations shows different characteristic patterns. In southern Sweden (Region 4), the ground-water storage reaches its maximum filling during the winter, due to very low evapotranspiration and consequently a great amount of available precipitation. The lowest ground-water level occurs during late summer or early autumn. In northern Sweden (Region 1), the maximum ground-water level occurs in connection with the snowmelting in spring, while the lowest level usually occurs just before the start of the snow-melting. These conditions show that the main recharge period occurs at the snow-melting due to the existence of large amounts of water in combination with low evapotranspiration. In spite of the usually high, summer precipitation, the ground-water storage is generally being lowered due to high evapotranspiration. Between these two regions, another two also exist with the annual patterns shown in Fig. 10.

The maximum annual periods of recharge (the maximum increase in ground-water levels) occur at different seasons for the different regions. Generally, these periods occur as follows:

Region	Primary maximum	Secondary maximum
1	Snow-melting	—
2	Snow-melting	Autumn
3	Snow-melting	Winter
4	Winter	—

Thus, in neither of the regions will the ground-water level be at its maximum during the summer. The reason for this is that the evapotranspiration in summer is usually so high as to exceed the precipitation. In Sweden, the main occasion for ground-water recharge occurs during the snow-melting while the other seasons are of less importance.

6.2. Ground-water-level response

Gottschalk and Nordberg (1977) have presented a mathematical model for the ground-water-level response due to infiltration.This model was calculated on the basis of a number of records from measuring stations included in the Ground-water Network. The results of their work show that large, glaciofluvial eskers, with, thick unsaturated zones, have a small and low response, whereas till has a large and rapid response. These results are in full agreement with the discussion given in the former section of the present paper. However, the input values used in the study of Gottschalk and Nordberg have clearly affected the results. At stations where the ground-water loggings were made at intervals of a few days the model shows a peak response after only a few days. At other stations, with recording intervals of a week or fortnight, the peak response occurs after some weeks. This condition is shown in Table 2.

Table 2. Observation interval and response lag for the stations utilized by Gottschalk and Nordberg (1977).

Station	Observation interval t_1 (days)	Response decay t_2 (days)	t_2/t_1
Kristianstad	7	7	1
Emmaboda	14	98	7
Djurarpsdalen	14	14	1
Sätrabrunn	7	21 - 60	3 - 8
Tärnsjö	7	14	2
Hälla	14	14 - 45	1 - 3
Lappträsket	1	1 - 4	1 - 4
Abisko	1	45	45

The extremely slow response for the Abisko station is explained by low conductivity and the great depth to the ground-water table (5 - 8 m). This deep-sited ground-water level indicates that the till may be underlain by some conductive formation.

It seems reasonable to assume that the long response times reported by Gottshalk and Nordberg may be an effect of the observation period used, as this interval is the shortest time within which any respons may be observed. Consequently, the peak response may in reality occur more quickly than at the intervals reported. Such a quicker response is in agreement with the discussion above in the present paper. The reason for such a quick response is that the coarser pores will conduct the water rapidly, causing a fast peak response.

In determining the time-lag between the occurrence of the available precipitation and the response of the ground-water level, the recharge due to the snow-melting would be of the greatest significance. This response would be most clear-cut in the northernmost parts of the country, where the main recharge occasion occurs at the snow-melting in spring. Thus, all of the accumulated precipitation in the snow cover will be available almost immediately. Fig. 11 shows the relation between the daily mean temperature, the the decreasing snow cover and the response of the ground-water level at the Abisko station (Nordberg and Persson 1979). This figure indicates that the melted snow creates a rise in ground-water level about 2 days after the first day in spring when the temperature rises above 0^{0} C. However, it is obvious that the snow-melting must have started earlier during the daytime, when the temperature was also above zero. The response in the ground-water level is probably an effect of this melting and consequently the time-lag between the occurrence of the available precipitation and the rise in water level would be longer than the 2 days indicated, probably between 2 and 20 days. This time-lapse seems reasonable when we compare the hydraulic properties of the soil cover at Abisko with those reported in Fig. 3.

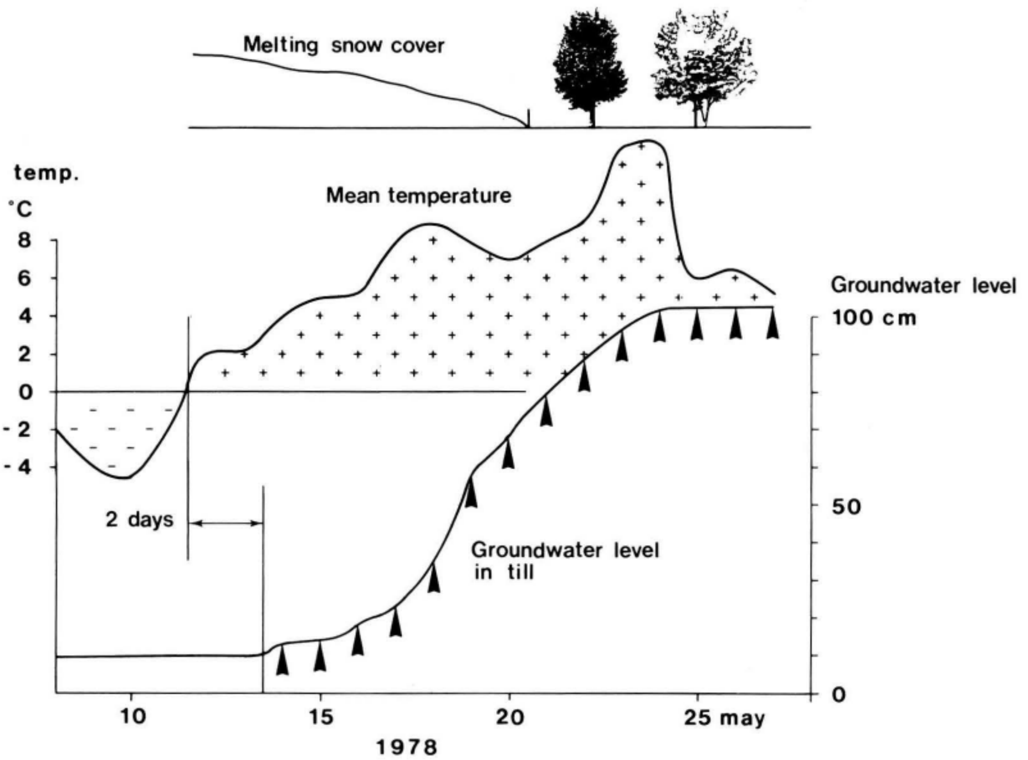

Fig. 11. The effect of the snow-melting on the ground-water level obtained at Abisko (Nordberg and Persson 1979).

7. GROUND-WATER RECHARGE IN A TILL AQUIFER

To illustrate this study, an area in northern Sweden has been chosen. In this area, Stensele in Västerbotten County, the Geological Survey is running a ground-water station with measurements made both in till and in glaciofluvial sand. Recordings of various hydrogeological data are continuosly made at this station. The ground-water levels in till and in sand, the duration and thickness of the snow cover and the conditions of soil-freezing are some of the parameters which are recorded. Continuous recordings have been made since 1971 and the station has been described by Nordberg and Persson (1974). In order to utilize these recordings to determine the ground-water recharge, the following parameters are needed:

- Ground-water-level fluctuations
- Available precipitation
- Effective porosity of the aquifer

The ground-water fluctuations are directly obtained from the records and the available precipitation may be calculated from meteorological data. The effective porosity and the recharge may be determined by an iteration method based on the ground-water fluctuations and the available precipitation. In the following pages, the determination of the parameters needed will be shown.

7.1. Available precipitation

For the available precipitation, the actual precipitation and evapotranspiration are crucial, as the available precipitation is made up of the difference between these parameters. While the precipitation may be directly measured, it is much harder to determin the evapotranspiration,but several empirical, analytical and direct-measuring methods exist. In this study, the evapotranspiration is determined by a water-balance study of the soil-water storage based on Bodyko´s method of combination (Bodyko 1948,Landberg and Ryttar 1978, Carlsson and Falk 1968, and others). The use of this method produces directly a figure of the available precipitation and the method requires the following data:

- Precipitation
- Potential evapotranspiration
- Soil-water content available to plants

In order to determine the potential evapotranspiration, the Thornthwaite method has been used; this is an empirical relation based on the monthly average temperature. The more accurate Penman method would normally have been used, but this method calls for comprehensive meteorological data, which would complicate the determinations.

The Thornthwaite method gives a fairly accurate result, as will be seen in Table 3, which presents a comparison between the Penman and the Thornthwaite methods. Both methods yield similar values of the potential evapotranspiration throughout the year, yet the monthly values are different. It seems as if the Thornthwaite method underestimates the evapotranspiration during spring and summer, while overestimating it during autumn. However, it has been judged to give sufficiently accurate values for the present purpose.

Table 3. Comparison between values of the potential evapotranspiration obtained via Thornthwaite's and Penman's formulae from the climatological data from Stensele for the period 1931 - 1960.

Method	J	F	M	A	M	J	J	A	S	O	N	D	Year
Penman	0	0	9	34	68	93	86	50	23	0	0	0	363
Thornthwaite	0	0	0	0	44	71	86	76	51	12	0	0	340

The third of the above-mentioned requirements is constituted by the soil-water content above the field capacity. Errors caused by variations in this parameters are, however, small. Calculations show that the difference in evapotranspiration is less than 10 per cent for a soil-water content ranging from 50 to 250 mm (cf. Eriksson and Johansson 1978, and Olsson 1979). A comparison between obtained evapotranspiration values calculated for 50 mm and 250 m soil-water content is shown in Fig.12. For the present study, a soil-water content of 250 mm has been used, which overestimates the evapotranspiration and underestimates the available precipitation. The following equations were used (Carlsson and Falk 1978), in order to determine the actual evapotranspiration and available precipitation:

$$E = E_p \frac{S}{S_a}$$

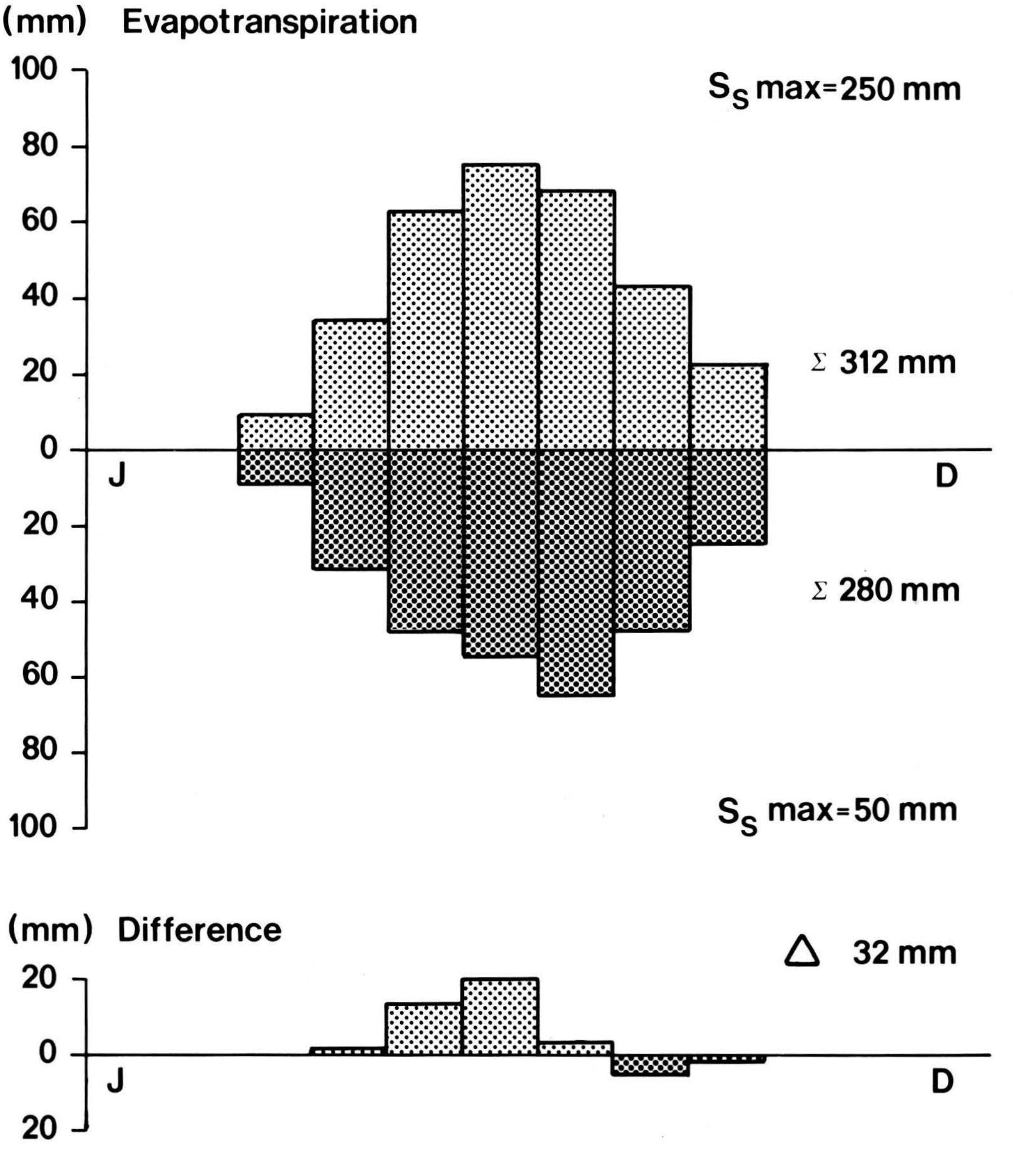

Fig. 12. A comparison of the effects of different values of the soil-water content (S_{smax}) on the evapotranspiration.

Where E = actual evapotranspiration

E_p = potential evapotranspiration

S = actual soil-water content

S_a = maximum soil-water content available to plants

The term S was defined as:

$$S = S_o + \frac{\Delta S}{2} \tag{18}$$

where S_o = the water content at the start of the calculation period

ΔS = change in water content during the calculation period

and ΔS is defined as:

$$S = \frac{P\, S_a - E_p\, S_o}{S_a + E_p/2} \tag{19}$$

where P = precipitation

The calculations in this paper are based on a period of one month and are for the period 1971 — 1978.

The precipitation used is that measured at the climatological station at Stensele and the precipitation through the winter was regarded as stored until the first month in spring when the temperature rose above 0 °C. This, however, is a very rough approximation, as the melting of the snow cover is long drawn out and melting seasons may also occur even during the winter. Fig. 13 shows the recorded values of temperature and precipitation and the calculated values of the soil water content, evapotranspiration and available precipitation.

7.2. Infiltration capacities of till soils

In the literature, there are several determinations of infiltration capacities obtained by infiltration tests in till soils. Most of the results show infiltration rates of 1 to 100 mm/min (Eriksson 1977). To take an example, von Brömssen (1968) made measurements which resulted in stationary infiltration rates of 3 - 11 mm/min. In the Juktan area in the vicinity of Stensele, a few simple tests of the infiltration capacities of different till deposits were made (Olsson 1979). For these measurements, a double-ring infiltrometer was used and the tests showed on the average, infiltration rates of

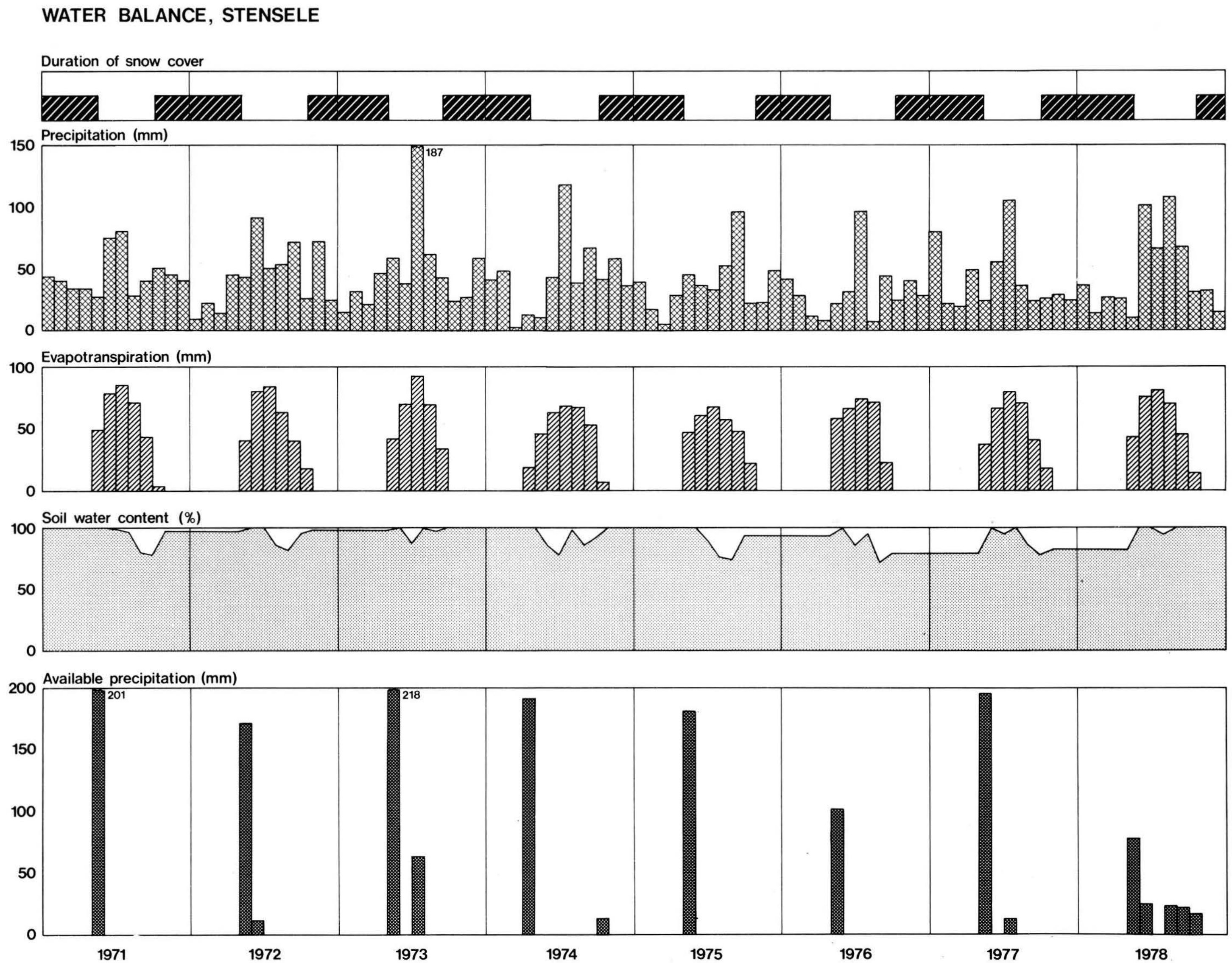

Fig. 13. Water balance of the Stensele area.

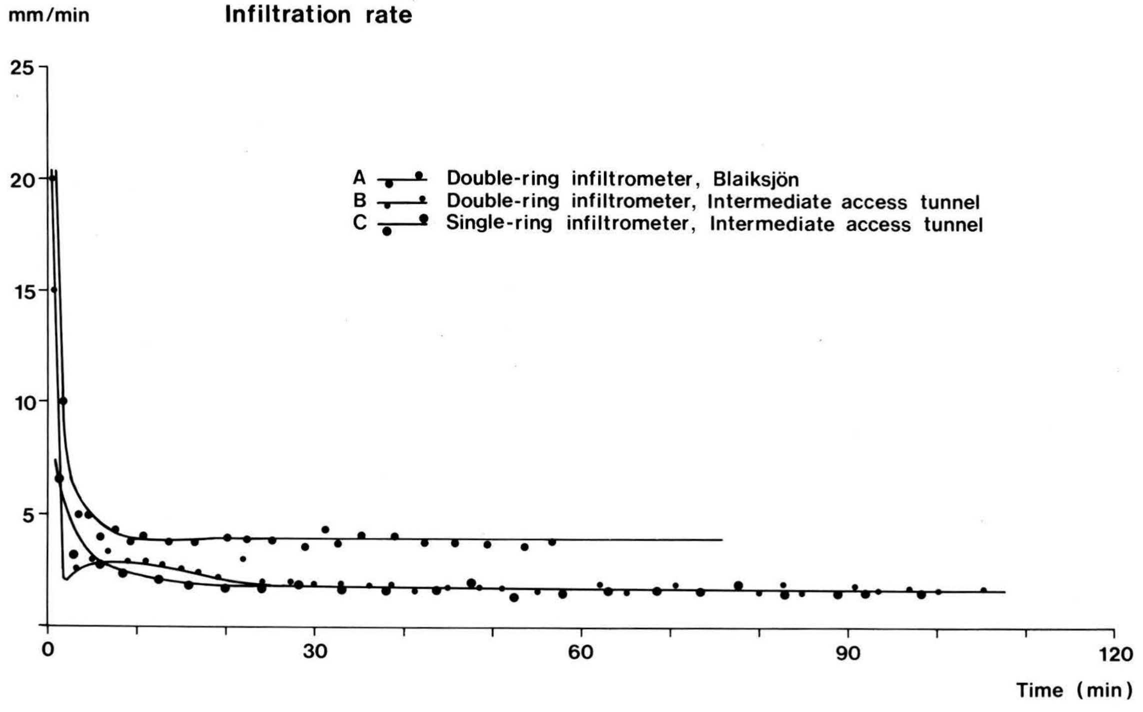

Fig. 14. Infiltration rate versus time in some till soils in the Juktan area (Olsson 1979).

3 — 5 mm/min, i.e. of the same order of magnitude as those of von Brömssen and those reported by Eriksson. The infiltration rates versus time for the Juktan tests are shown in Fig. 14.

All these measurements show that the infiltration capacity is normally much higher than the prevailing rain intensities. As Eriksson (1977) pointed out, these conditions indicate that all precipitation may infiltrate immediately. Consequently, surface-run-off is rare in till soils.

7.3. Effective porosity and ground-water recharge

Fig.15 shows the recordings from the ground-water station at Stensele and the recorded parameters are the ground-water level, precipitation, snow cover and frost penetration. The measurements have been going on

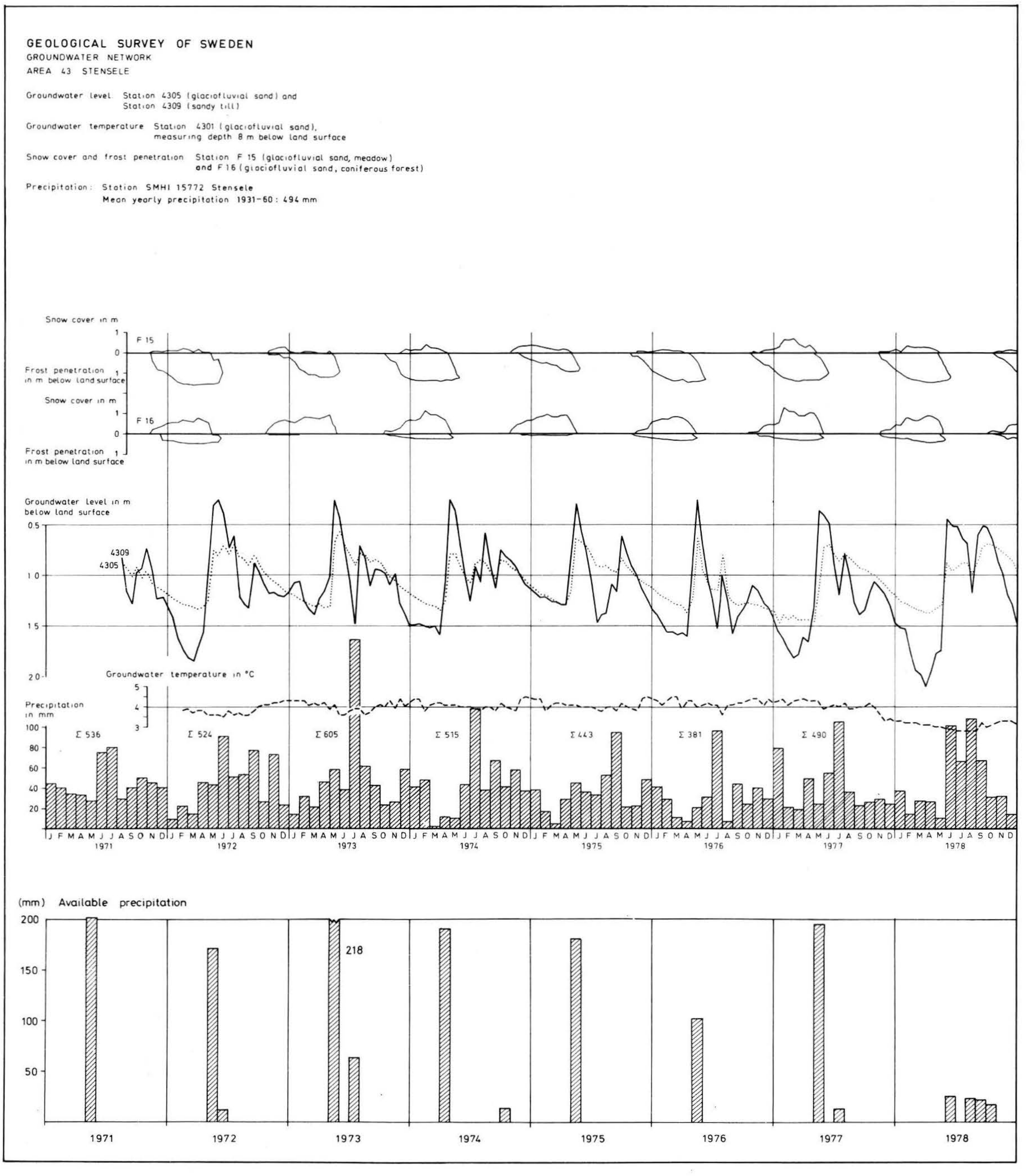

Fig. 15. Recordings of the ground-water-level fluctuation, precipitation, available precipitation, snow cover and frost penetration at Stensele during the period 1971 — 1978. Modified from the Geological Survey of Sweden.

since 1971 and the recordings in Fig. 15 represent the period 1971 — 1978. In this figure, the calculated values of the available precipitation as the potential for ground-water recharge are also included.

As regards the effective porosity and recharge, the significance of these measurements is the response of the ground-water system due to a specific amount of precipitation. The response in the aquifer varies greatly, mainly due to the precipitation, the soil-water conditions and the evapotranspiration. The intensity and duration of the precipitation have an additional importance for the aquifer response (Bruce and Clark 1966).

Variations in resposne are mainly influenced by the hydraulic properties of the aquifer (hydraulic conductivity and effective porosity). In an extensive formation with high hydraulic conductivity and porosity, the response will be leveled due to the fast pressure release within the aquifer. In a till formation with lower conductivity and small porosity, the pressure release will be much slower and, hence, a more direct-acting response due to the precipitation will be obtained (cf. Brown et al. 1972).

From the recordings at Stensele, it is possible to observe the following conditions as regards the ground-water recharge to the till aquifer:

- The main ground-water recharge occurs during the snowmelting. During the observed period 1971 — 1978, the aquifer reaches its lowest, annual, ground-water level just before the start of the melting period and its highest during the melting period.
- No considerable recharge is obtained during the summer, in spite of high precipitation rates. However, recharge has occurred during a few summer periods, due to extremely high precipitation.
- Usually, a rise in the ground-water level occurs in connection

with precipitation in autumn, during the months of August and September.

- Usually, no recharge takes place during the winter (October/ November — March). On the contrary, the aquifer is gradually being drained.

The potential for recharge is constituted by the available precipitation. This precipitation will be divided between any of the main hydrological systems: surface run-off, soil-water storage, and ground-water storage and run-off. Surface run-off may occur when the amount of the available precipitation is higher than the infiltration capacity of the ground, and ground-water recharge only occurs when the soil is saturated with water, i,e, the soil-water content is higher than the field capacity.

As Fig. 15 shows, water is available for recharge and run-off mainly during April, September and October. The fluctuations in ground-water level also indicate that the main part of the recharge usually takes place in these months. Hence, the recorded hydrological and hydrometeorological conditions during the recharge period may be used in order to evaluate the amount of recharge and the effective porosity based on the aquifer response.

The most important occasion ofground-water recharge is during the melting of the stored snow. Depending on the lapse of the melting, a varying part of the water will contribute to the recharge. If the melting is slow, the major part or all of the water may infiltrate. But if the melting is fast, the available amount of water may be too large, compared with the infiltration capacity of the ground. Hence, a great part of the melt-water will be discharged as surface run-off. These conditions indicate that the ground-water response may be different, due to varying conditions in the course of the melting.

During the second period with available precipitation — August to

October — the ground-water recharge is directly dependent on the amount of available water. In this case, the infiltration capacity may hardly be exceeded and, consequently, the response of the aquifer will probably be more clear-cut.

According to Brown et al. (1972), the ground-water balance may be defined by the following relation:

$$\mu \, \Delta H = \Delta Q \quad \Delta t + P \, \Delta t \tag{20}$$

where μ = the specific yield of the soil
ΔH = the rise in ground-water level
ΔQ = the difference in ground-water recharge and discharge from and to adjacent areas
Δt = computation interval of time
P = rate of ground-water recharge from above

However, if the recharge from and the discharge to adjacent areas are in equilibrium and if an individual recharge occasion is computed, the equation will become:

$$P = \mu \, \Delta H \tag{21}$$

On each recharge occasion, the amount of the recharge is limited by either the amount of available precipitation or by the infiltration capacity of the ground. For a given soil profile, the specific yield varies with time, water level, temperature and barometric pressure, as pointed out by Bear et al. (1967). But, as the specific yield in an unconfined aquifer is equal to the effective porosity, the rise in ground-water level will also be a function of the effective porosity of the soil, provided that the recharge occurs instantaneously. If the recharge is prolonged for some time, the response may also be influenced by the discharge from the aquifer.

According to equation (21), the rise in ground-water level is in direct proportion to the ground-water recharge and inversely proportinal to the effective porosity. This gives us an equation with two unknown parameters and one known. But as the available precipitation is also known, it will be possible to solve the equation with respect to both the effective porosity and recharge.

For the determinations, individual rises in the ground-water level were analysed with respect to the available precipitation. The calculations were made in two separate steps:

1. Determination of the effective porosity
2. Determination of the recharge

In the first step, the monthly amount of available precipitation on each recharge occasion was attributed to the obtained rises in ground-water level, which resulted in a number of varying values of the effective porosity. Of these different values, the smallest may be presumed to represent the actual soil conditions, while the higher values were assumed to be apparently increased on the account of prolonged recharge and consequent ground-water discharge. The obtained value of the effective porosity was then used in the second step, in which the recharge was determined. The recharge during the year was determined as the sum of the individual recharge occasions. In Appendix 1, the calculations are described in more detail.

The mean values obtained for the ground-water recharge as percentages of the available precipitation are shown in Table 4. The obtained values of the effective porosity of the till aquifer is also shown. Generally, the part of the available precipitation which is contributed to the recharge is smaller during the snow-melting than during the autumn, which may indicate a greater surface run-off or losses during snow-melting during the winter. The variation in recharge from year to year is also greater in the values based on the snow-melting.

Totally, the ground-water recharge shown in Table 4 amounts to about 115 mm/year as an average for the period 1971 — 1978 and this is

Table 4. Ground-water recharge and effective porosity of the till aquifer at Stensele.

Season	Recharge (Per cent of available precipitation) Mean value	Standard deviation	Effective porosity (per cent)
Spring	60	± 20	4
Autumn	70	± 15	4

about 60 - 70 per cent of the total available precipitation in the Stensele area and about 20 - 30 per cent of the total annual precipitation. The effective porosity has been determined to be about 4 per cent. Both these values seem reasonable. Perhaps the recharge values, especially during the autumn, should in reality reach 100 per cent to correspond to the infiltration rates.

The magnitude of the effective porosity obtained for Stensele agrees with other values reported in the literature from other till deposits. Nordberg and Modig (1974) report values for the effective porosity of about 5 per cent and for the total porosity of 25 per cent, and Lundin (1977) reports a value of 2.3 per cent.

8. DISCUSSION AND CONCLUSIONS

The main disadvantage of the method presented is that the calculations are based on monthly values. The response due to the precipitation is usually much faster than a month (Gottschalk and Nordberg 1977), which indicates that both recharge and discharge of the aquifer may occur during any specific month. Fig. 16 illustrates these conditions. This figure shows a month with a number of precipitation occasions, some of which cause a response in the ground-water level, while others do not. In this particular month, the groundwater level

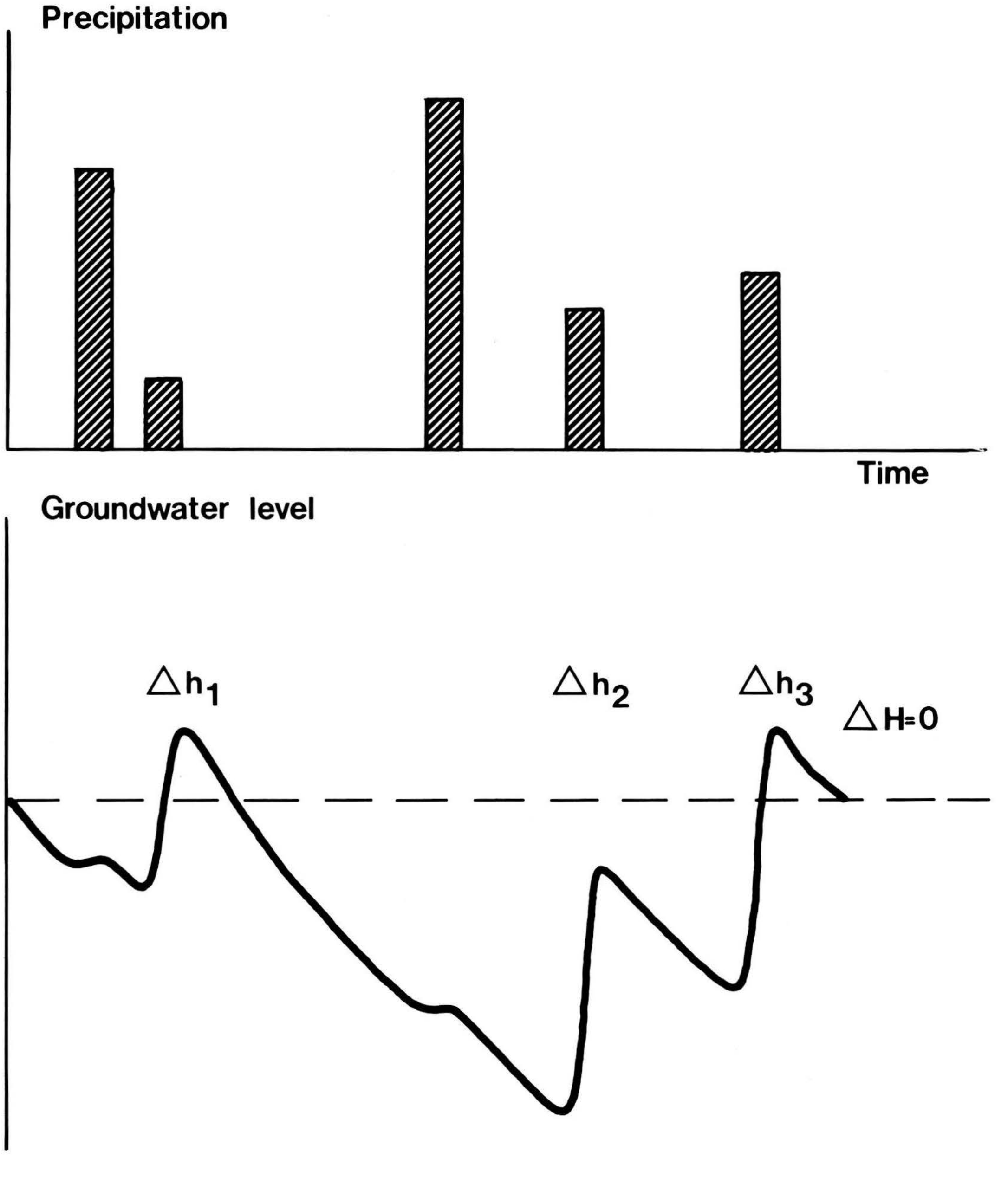

Fig. 16. Schematic illustration of the importance of the short-term variations in the response.

is equal at its start and at its end. Thus, the monthly response in the ground-water level is 0, which obviously is not correct. The real rise during the month is in fact the sum of each individual rise, but during the month the discharge is in equilibrium with the recharge. Consequently, in order to obtain more reliable results, the calculations should be based on continuous measurements or on measurements made at maximum intervals of one or a few days.

Another source of errors is the difficulties in the determinations of the available precipitation, which is based on uncertain estimations of the evapotranspiration and the soil-water storage. However, the differences due to this condition are probably less than 30 per cent. The error caused by the soil water storage amounts to less than 10 per cent and probably the error caused by the evapotranspiration amounts to about 20 per cent.

The results of the calculations show that the recharge amounts to about 60 -70 per cent of the available precipitation. Assuming, as indicated by the infiltration rates, that the infiltration capacity is greater than the normal rain intensities, the actual recharge should in fact be 100 per cent at least for the autumn values. This shows that the maximum error of the method should be about 30 - 40 per cent. In addition to this error, there is that of the available precipitation. A reasonable value of the total error should then be about 50 - 60 per cent. The effect of these errors are that the recharge will be underestimated and the effective porosity overestimated. Consequently, more accurate values would be 115 - 170 mm/year for the recharge and 2 - 4 per cent for the effective porosity.

The effective porosity of a formation is not constant throughout the aquifer, but varies considerably with depth. Lundin (1977) found that the total porosity (very likely also the effective porosity) decreases with depth, from about 60 per cent in the upper 10 cm to 20 per cent at a depth of 1 m. These measurements were, however, made in another till deposit and thus the values are not valid for Stensele. Nevertheless, it is probable that a similar pattern also exists at Stensele. According to Lundin, the conductivity also decreases rapidly with depth. He reports horizontal, ground-water velocities in the saturated zone of 0.18 m/day aand 0.06 m/day at depths of 0.4 m and 1 m respectively. These velocities occurred during about equal hydraulic gradients.

In spite of these uncertainties, the method may be useful for determinations of the effective porosity in till soils and also for studies of the ground-water recharge. It is possible to improve the method by calibrating it with other measurements, such as neutron measurements for the porosity and lysimeter tests for the recharge.

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APPENDIX 1

ANALYSIS OF GROUND-WATER-LEVEL FLUCTUATIONS

As stated in the main paper. the rise in the ground-water level due to a specific amount of precipitation is in proportion to this amount (recharge) and inversely proportional to the effective porosity, that is to say,

$$\Delta H = \frac{P}{\theta_e} \tag{1.1}$$

The validity of this equation presumes that the recharge is instantaneous. If not, an apparently increased value of the effective porosity will be obtained, because the quantity of the discharge during the recharge period carries great weight. The total rise in the ground-water level is in this case governed by the actual recharge minus the discharge. The estimations based on the total recharge will therefore yield too high a value of the effective porosity.

Assuming that the total available precipitation from a given rainfall will contribute to the ground-water recharge, a fictive value of the effective porosity may be determined. The effective porosity is fictive in the sense that the recharge is not in fact instantaneous, but a small quantity of water will invariably be discharged during the recharge period. In other words, the obtained value is composed of the effective porosity plus an unknown discharge. The assumption as to instantaneous recharge and the contribution of the total available precipitation will probably hold good occasionally. Consequently, it is possible to obtain a fair approximation of the effective porosity by analysing a great number of recharge occasions. The lowest value obtained for the effective porosity will be approximately equal to the real effective porosity of the aquifer.

Fig. 1.1. shows the variations in ground-water level in an aquifer

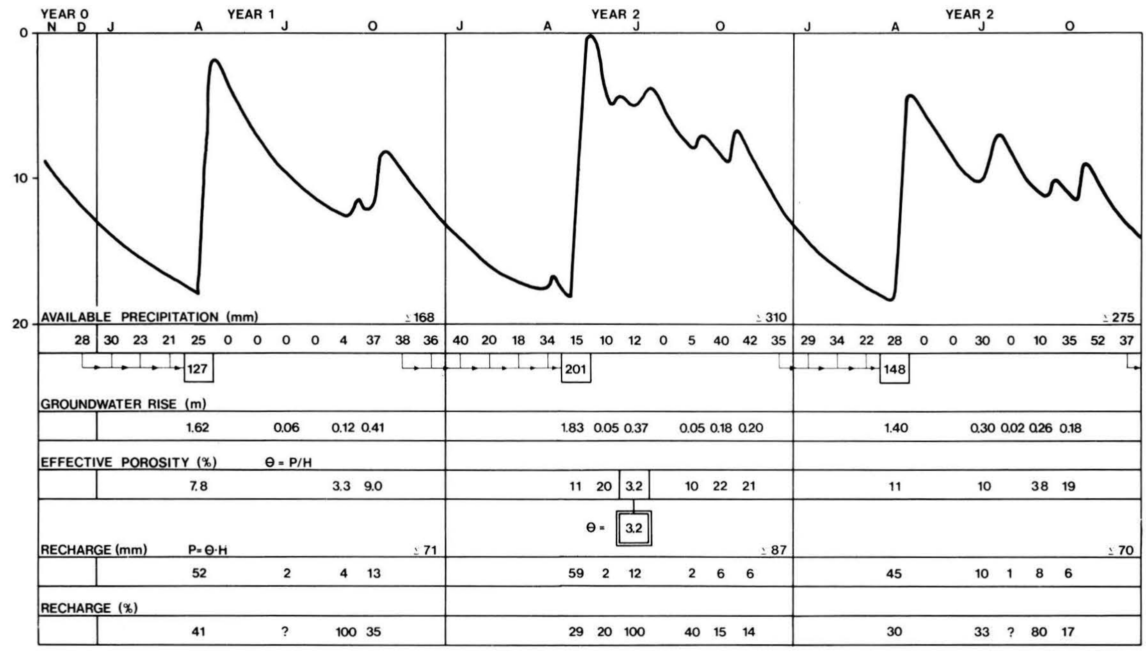

Fig. 1.1. Schematic illustration of the calculation method.

during some years. Additionally, the calculated values of the available precipitation are also included. It should be noted that the values presented in Fig. 1.1 are not authentic but hypothetical and are given only for the purpose of describing the approximate method used for the determinations.

From the figure, it can be seen that the available precipitation is present during the winter (stored as snow, but released in April — May) and autumn. The month in which the snow melts has the highest available precipitation and during this month the main filling of the ground-water storage takes place. Over the years, the aquifer in this example was recharged on 15 different occasions recorded by the rises in the ground-water level.

In the diagram, the magnitude of each recharge is included. On the basis of the given values of the available precipitation and the ground-water-level rise, the effective porosity can be determined by using equation (1.1), assuming that all the available precipitation infiltrates. This determination will result in 15 different values of the effective porosity, one for each recharge occasion.

In the example, the effective porosity ranges from 3.2 per cent up to 22 per cent. The highests of these porosities are probably a result of long-lapsing recharge and simultaneously occurring discharge. Surface-water run-off is another process which creates such high values of the porosity. However, the lowest values obtained represent a porosity which is probably fairly close to the real effective porosity of the soil. In accordance with the principles of this method the value obtained should be a maximum value and, to obtain an accurate value, it is necessary to analyse a great number of recharge occasions.

After the fixing of the most reliable value of the effective porosity, the recharge on each occasion may be determined by equation (1.1).

In the example in Fig. 1.1, the recharge on each occasion is given both in millimetres per year and as a percentage of the available precipitation. The range of the recharge is from 14 per cent up to 100 per cent. For each year, it varies between 25 and 40 per cent. Compared with the results in the main papaer, these values are much lower.

For the long-lapsing recharge, the rise in ground-water level should be measured as indicated in Fig. 1.2. and not as the difference between the initial low level and the peak rise. This type of determination of the rise results in too high a value of the porosity and too low a value of the recharge. Another main disadvantage of the present method is that the calculation on a monthly basis does not take account of the short-term fluctuations (one or a few days). Their effect also leads to the porosity being overestimated and the recharge underestimated.

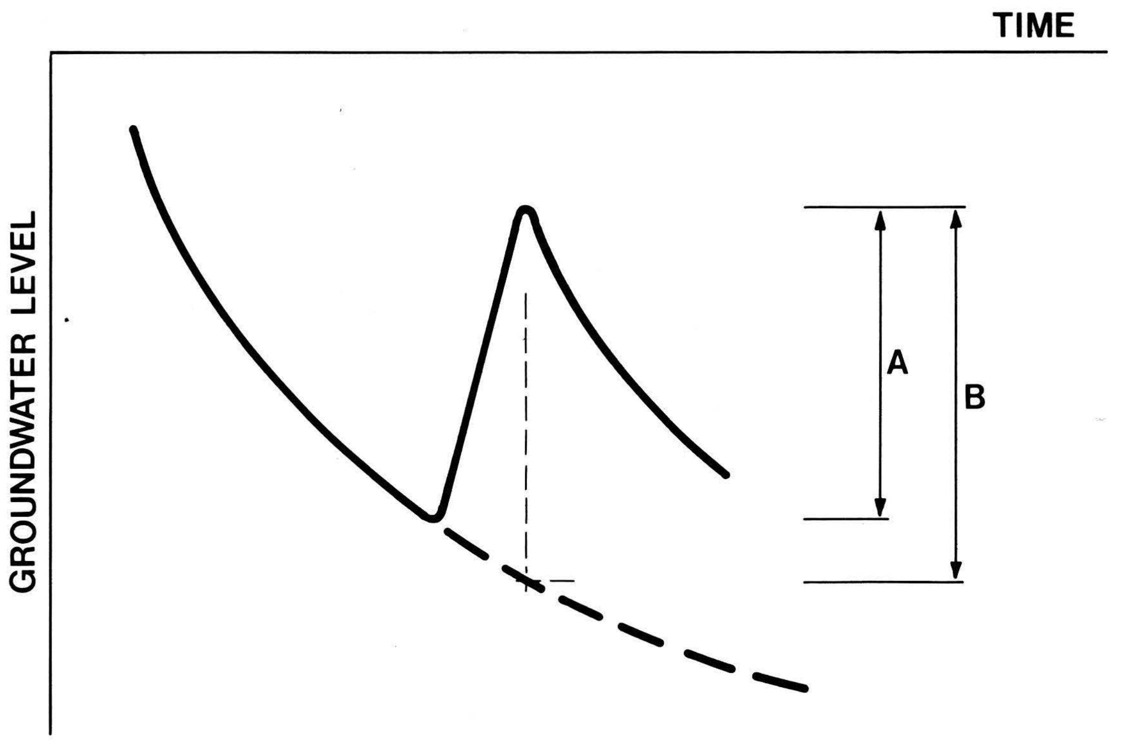

Fig. 1.2. The effect of the duration of the recharge on the response. The ongoing discharge during the recharge period makes it necessary to consider this effect. A. Incorrectly determined rise. B. Correctly determined rise.

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ISSN 0349-2176
Offsetcenter ab, Uppsala 1980
ISBN 91-7158-213-4