Hydrogeology in the Romerike area, Southern Norway

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A hydrogeological model has been developed on the basis of field observations and data collected during the International Hydrological Decade. From water balance studies it was found that half the annual precipitation was lost due to evapotranspiration (400 mm). Nearly 60% of new groundwater percolates through the soil profile during snowmelt. The average melting rate in spring 1966 was 13.5 mm/day. Linear velocities through the unsaturated zone were between 17 and 20 cm/day. Water moves through the saturated zone with an average velocity of 10-20 cm/day, and average natural residence time is close to 30 years.

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Introduction

Comprehensive hydrological studies were carried out in Norway during the International Hydrological Decade from 1965 to 1974 (Otnes 1973, 1975). Romerike, about 40 km north of Oslo, was chosen as a representative area. A hydrogeological map of the area has been published by Østmo (1976).

Glaciofluvial deposits formed during the last deglaciation of Scandinavia are important groundwater aquifers. The purpose of this paper is to present a hydrological and hydrogeological model for an area with such deposits.

Study area

A map of the Pleistocene deposits (Fig. 1) illustrates that the aquifer is composed of glaciofluvial sand and gravel partly underlain by silty glaciomarine deposits. A marked pause in the ice retreat over the area resulted in the formation of a glaciofluvial delta which built up to the marine limit. The glaciofluvial drainage channels start where the subglacial meltwater came out of the glacier.

Due to isostatic rebound, the area has been lifted 200 m above sea level during postglacial time and, consequently, the median elevation for the area studied is close to 200 m. Podzol profiles have developed on these sandy soils, and the area is mainly covered by coniferous forest.

Fig. 2 shows the groundwater flow pattern within the subarea drained by the river Risa (Østmo 1976). The catchment is covering an

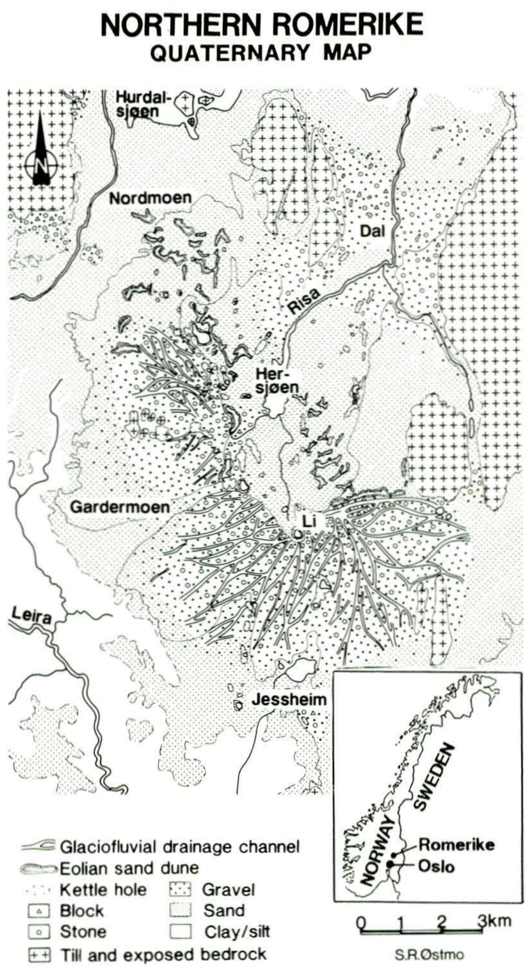

Fig. 1. Map of Pleistocene deposits within the studied area.

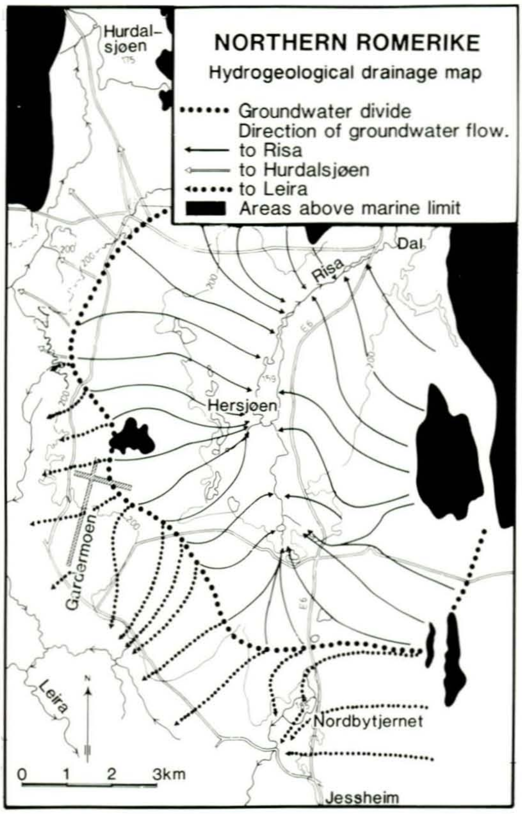

Fig. 2. Groundwater flow pattern within the studied area (Østmo 1976).

area of 55.1 km^2, and the distance from groundwater divides to the lake or river varies between 2.5 and 4 km.

Hydrological budget

Infiltration

The infiltration capacity of the sandy surface layers is high, and overland flow is normally absent or very low. This is the case even during snowmelt in the spring with frozen soils. During the summer the infiltration rate, measured with a double-ring infiltrometer, varied between 4 and 5 cm/min under saturated conditions (R. Sørensen pers. comm. 1985). Lemmelä & Tattari (1988) measured the infiltration capacity on similar soils in Finland. They found that the infiltration capacity, even during the frozen period in the spring, exceeded the rate of snowmelt by more than two orders of magnitude. The infiltration rate during the frost period was half the rate measured in the summer.

Hydrological measurements and calculations

The storage equation, where we balance inflow, outflow and storage, can be written as:

$$P = E + Q + D + dH$$

where P = precipitation, E = evapotranspiration, Q = flow out of the catchment, D = subsurface drainage and dH = change in storage. The quantities were measured for periods of 9 or 10 years.

Precipitation (P)

Fig. 3 shows the annual precipitation at 4 stations (series A to D) during the years 1966-1974 (Otnes 1973, 1975). Romerike is a flat catchment subjected to fairly uniform precipitation. The precipitation gauges are protected from wind action by the surrounding forest. As a result there is good correlation between the quantities determined at the various stations. Tollan (1970) found very good agreement between precipitation measured in a precipitation gauge and snow accumulated on a snow pillow during the winter.

The arithmetic average for the annual precipitation is: P = 794 mm (rain: 508 mm, rain and snow: 34 mm, snow: 250 mm). The amount of snow varied between 114 and 375 mm during the Hydrological Decade.

Flow out of the catchment (Q)

Lake Hersjøen and the river Risa (Fig. 2) are mainly fed by groundwater and precipitation falling directly on the lake. The discharge is almost constant through the year. The average value during the period 1967-1974 was 0.85 m^3 sec^{-1}, and average annual discharge was 26.8 x 10^6 m^3. This is equivalent to: Q = 486 mm (catchment area with lake Hersjøen = 55.1 km^2).

Change in storage (dH)

Changes in the groundwater level were recorded during the period 1968-74. According to Østmo (1986) the average value for the catch-

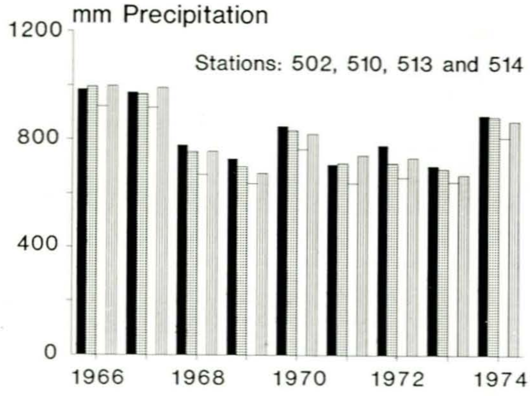

Fig. 3. Annual precipitation at 4 stations during 1966—1974.

ment during this period (6.5 years) was 600 mm (precipitation) and dH = + 92 mm.

We cannot determine the subsurface drainage, but we have assumed that it is small compared with the measured flow out of the catchment: D = O.

The evapotranspiration is found by solving the storage equation:

E = P-Q-dH = 794 - 486 + 92 = 400 mm.

This value for the annual evapotranspiration is close to that calculated by Høiland et al. (1952) and Johannesen (1970).

The results of these measurements and calculations are presented in Fig. 4. The amount of water which percolates through the soil profile and continues to the groundwater reservoir is 394 mm.

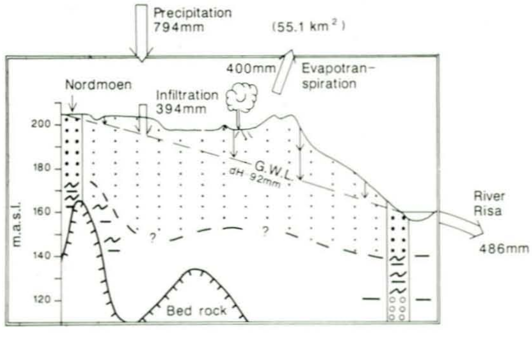

Fig. 4. Hydrological model for northern Romerike (See legend Fig. 8).

Evapotranspiration through the year

Evaporation during the summer was determined by pan evaporation (Otnes 1973, 1975). The average values for the years 1968 to 1974 are given in Table 1.

The amount for these 5 months (413 mm) is more than the value calculated for the whole year from the storage equation. This is expected since evaporation from the small amount of water in a pan will be faster than evapotranspiration in the field.

Table 1. Average values for pan evaporation (1968-1974) in mm precipitation.

Mai	June	July	Aug	Sept
83	103	94	83	50

We have assumed that the real evapotranspiration follows the same pattern as the pan evaporation and evapotranspiration in other areas (Aslyng 1966), but the total amount is 400 mm.

Fig. 5 illustrates the distribution of precipitation and evapotranspiration through the year. The difference between these two bar diagrams gives the quantities percolating to the groundwater aquifer.

Snowmelt and groundwater renewal

Fig. 5 shows that a major part of the precipitation is accumulated as snow during winter. The average value (water equivalents) for 1968-1974 was 286 mm, deposited during November to March. The evaporation during the snowmelting period is close to 50 mm (Fig. 5). This means that 236 mm (or 60%) of new groundwater percolates the soil profile during snowmelt. This is very similar to what was observed in Finland (Lemmelä & Tattari 1988).

The snowmelt is normally due to increased air temperature, combined with precipitation as rain. The accumulated snow in this forested area disappears normally during a period of 3 to 5 weeks. Observations made during spring 1966 are used to illustrate the connection between snowmelt, percolation and the rise of the groundwater level (Fig. 6). The maximum amount of accumulated snow that year (148 cm) was equivalent to 375 mm of water (Fig. 6). Rain, sleet and warm weather started the snowmelting at the beginning of April, but the

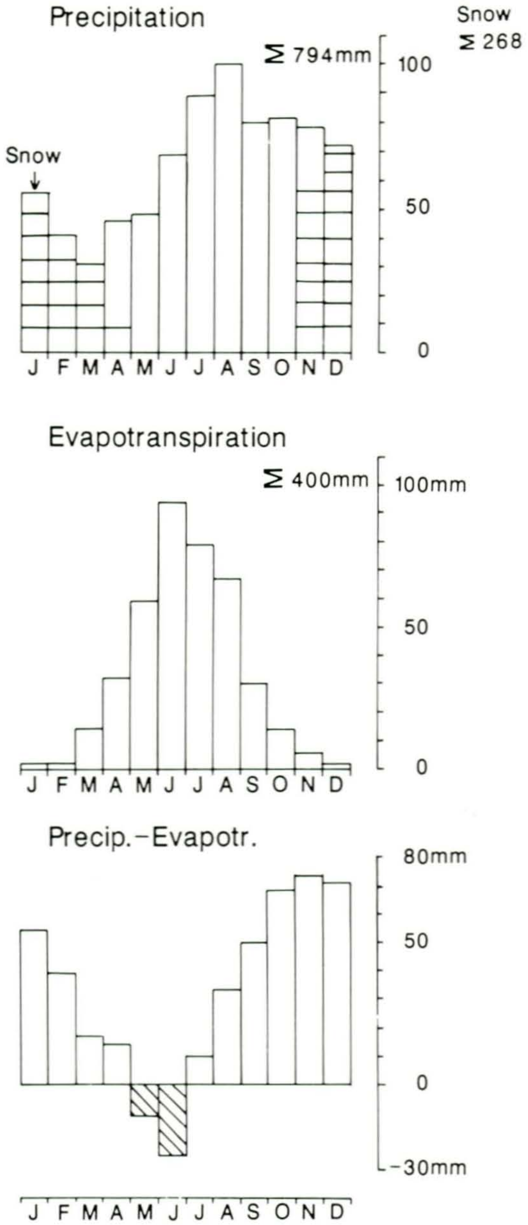

Fig. 5. Annual water budget for the period 1966—1974.

process stopped partly due to one week with cold weather. After this, the temperature increased and the snow melted rapidly. The snow had totally disappeared by May 15. During the melting period there was 40 mm of rain while the evapotranspiration was close to 50 mm (Fig. 5). This means that about 365 mm of water infiltrated during this period.

Changes in the groundwater level during the same period were observed by Bjor & Huse (1988). The measurements were taken at Nordmoen close to the water divide (Fig. 4). The groundwater level is supposed to change evenly in this fairly flat area with homogeneous sandy deposits (Fig. 4). The groundwater slowly sank during late winter by 4.12 mm/day until it started rising 90 mm/day. The groundwater level rose 129 cm during 23 days. If no water had been added it would have been lowered 9 cm (4.12 mm x 23 days). Consequently, the infiltration of meltwater raised the groundwater level 138 cm (Fig. 6).

The volume filled with water corresponds to a porosity (n_f) of:

$$n_f = (365/1380) \times 100 = 26\%$$

The efficient porosity of these soils (pF = 0.1) is close to 37% (Stuanes & Sveistrup

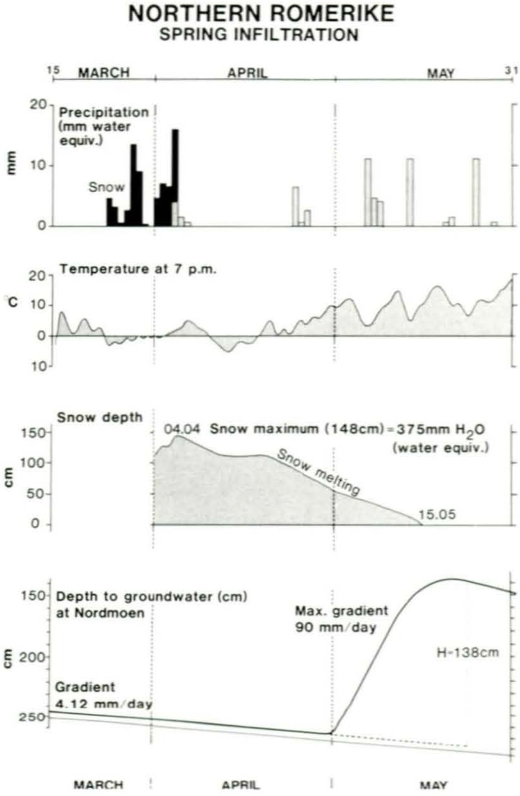

Fig. 6. Infiltration of groundwater during snowmelt in 1966.

1979). The conclusion must be that the soils were partly saturated due to subsnow melting prior to the observed rapid melting in April and May.

Percolation velocity

It is difficult to calculate the linear velocity of the water flux through the unsaturated zone since we cannot observe exactly when the infiltration started. On the basis of air temperature, we assumed that rapid infiltration started on April 18 and ended on May 15 (27 days). The groundwater level started to rise May 1, and maximum height was reached on May 23. In the beginning there was 265 cm from the surface to groundwater level while at the end there was 135 cm. These data give linear velocities through the unsaturated zone of 20 and 17 cm/day. The average water-flux during a melting period of 27 days, when 365 mm of water infiltrates the soil, will be 13.5 mm/day.

Tollan (1970) used snow pillows to determine melting rates in this area. He found that, during the most intense week of snowmelt in 1969, the daily average was 14 mm of water with a maximum of 30 mm. Lemmelä & Tattari (1986) found that the maximum daily melting varied from 8 to 30 mm at the same latitude in Finland. We have assumed that 13.5 mm of water passed through a 20 cm thick layer during 24 hours (the linear velocity was 20 cm/day). A layer of this thickness, with an effective porosity of 0.37, could contain 74 mm of mobile water. This means that the degree of saturation (average S-value) has been close to 20% during infiltration and percolation of new groundwater. From this we can calculate the hydraulic conductivity at 20% saturation ($k_{0.2}$):

$$V = k_{0.2} \times (i/S)$$

The value of $k_{0.2} = 4.6 \times 10^{-7}$ m/sec if we assume a gradient of -1.

Water movement through these deposits has also been determined with neutronmeter measurements (B. Rognerud pers. comm. 1987). The measurements were carried out after rainfall in summertime. A typical velocity of the 'wetting front' movement was 10 cm/day.

A lower rate of infiltration, compared to the snowmelt period, seems reasonable since the degree of saturation is lower during the summer. The hydraulic conductivity rapidly decreases even for a small decrease in the degree of saturation (Bouma 1977).

Change of groundwater level during periods with no infiltration

During the periods January-March 1968 and 1988 the groundwater level (Fig. 6) was lowered by even rates of 4.12 and 3.75 mm/day (Bjor & Huse 1988, L. Kirkhusmo pers.comm. 1989).

The discharge with river Risa is 0.85 m^3/sec, which means 73,500 m^3/day. The efficient porosity of these sediments is close to 37% and the catchment area is 55.1 km^2. A discharge of 73,500 m^3/day would lead to an average groundwater lowering of 3.6 mm/day. This agreement between the measured and the calculated lowering is good, since we have assumed similar conditions for the whole catchment area.

Exchange of water in the unsaturated zone

The thickness of the unsaturated zone varies between 1 and 30 m (Fig. 4) and the position of the groundwater level varies during the year.

The ability to store water in the upper part of the unsaturated zone is illustrated in Fig. 7. Samples were collected shortly after snowmelt in 1986. A total of 98.4 mm of water was stored in the upper 50 cm at field capacity. Below this 'soil profile' the deposits contained 56 mm/m (4.1%). This is close to the water contents determined in the laboratory at pF = 0.1 (Stuanes & Sveistrup 1979).

The average annual amount of water percolating from the surface to the groundwater reservoir (394 mm, Fig. 4) is equivalent to the amount stored in a 6 m deep unsaturated zone at field capacity.

At Nordmoen, where the depth to groundwater level varies between 1.5 and 2.5 m, the water in the unsaturated zone will be renewed between one and two times each year. In most of the catchment (Fig. 4) the average residence time for water in the unsaturated zone will be several years.

Residence time in the saturated zone

The average theoretical residence time (t_r) can be calculated:

$$t_r = Q_w/A_w$$

where A_w is the annual amount of new groundwater and Q_w is the total amount of water in the reservoir. The average value of A_w is 394 mm (Fig. 4).

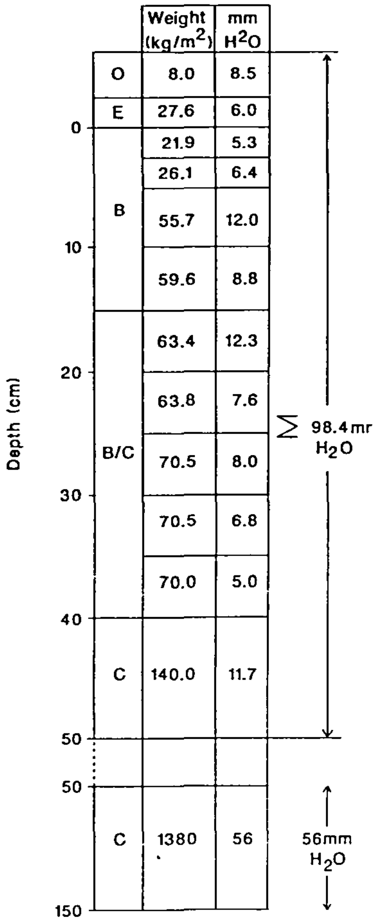

Fig. 7. Water contents in the unsaturated zone at field capacity (pF = 0.1). The left column shows the soil horizons.

Several geophysical methods have been applied to determine the depths to groundwater level and to bedrock. The amount of sediments within this drainage area has been calculated at Norsk Hydro a.s (Østmo, unpublished data). The total amount of sediments between the surface and bedrock for the whole drainage area is 4.03×10^9 m^3. This means that the average thickness of sediments, covering 55.1 km^2, is close to 73 m. The amount above groundwater level is 0.68×10^9 m^3, while the amount below is 3.35×10^9 m^3.

With a total porosity near 40% the sediments below groundwater level will contain 1.34×10^9 m^3 of water, which is equivalent to 24,300 mm of precipitation. As pointed out previously this aquifer is composed mainly of glaciofluvial gravel and sand underlain by silty (glaciomarine) deposits (Fig. 8). The geophysical methods are not well suited to differentiate between water-saturated sand/gravel and more fine-grained deposits. By using data from drillings combined with geophysical data we have found that the amounts of coarse-grained (sand, gravel) and fine-grained deposits are about equal. The amount of water in the coarse-grained water-saturated deposits will then be equivalent to about 12,000 mm. Even if there are some uncertainties in our calculations, we conclude that the average residence time in the coarse-grained deposits, with the major part of the hydraulic flow, is close to 30 years, while it is much longer in the fine-grained deposits.

Saturated flow through the aquifer

The following equation describes flow through saturated sediments:

$$V = k/n \times i$$

where V = filter velocity (m/day), i = gradient (m/m), n = porosity (decimal fraction) and k = hydraulic conductivity (m/day).

The gradient of the water surface, normal to equipotential lines, was determined from the hydrogeological map (Østmo 1976). The gradient varies between 0.12 and 0.002, but in the major part of the area the values are between 0.01 and 0.007. For the efficient porosity we have used a value of 0.36. With this porosity and a gradient of 0.01 the flow equation will be:

$$V = k \times 0.028 \text{ (m/day)}$$

The hydraulic conductivity for porous and well sorted sediments can either be calculated from particle-size data or it can be measured in the field. Jenssen (1986) has done field measurements for saturated flow through similar sediments. The relationship between the 10% value on the particle-size distribution curve (d_{10}) and the hydraulic conductivity is shown in Fig. 9. The coarse sand at Nordmoen (Fig. 8) has an average d_{10}-value of 0.09 mm while the value for the most fine-grained sediment below is close to d_{10}=0.003 mm.

The saturated hydraulic conductivity for the sand is close to 5 m/day, while for the fine-grained sediment it is close to 0.06 m/day. From this we get:

$$V_{sand} = 5 \times 0.028 = 0.14 \text{ m/day}$$
$$V_{silt} = 0.06 \times 0.028 = 0.0015 \text{ m/day}$$

We can draw the conclusion that water un-

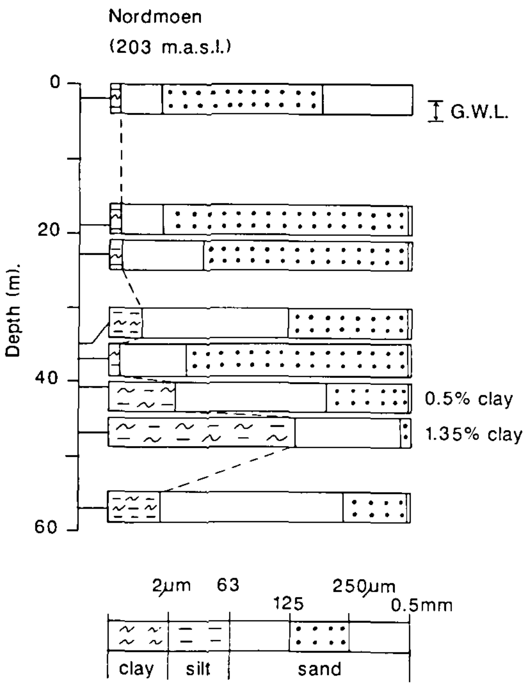

Fig. 8. Particle-size distribution of Pleistocene deposits at Nordmoen.

der the same gradient flows 50-100 times faster through the sand than through the underlying clay-containing silt. We have already assumed that the flow through sand is much faster than through silt when we calculated an average residence time of 30 years.

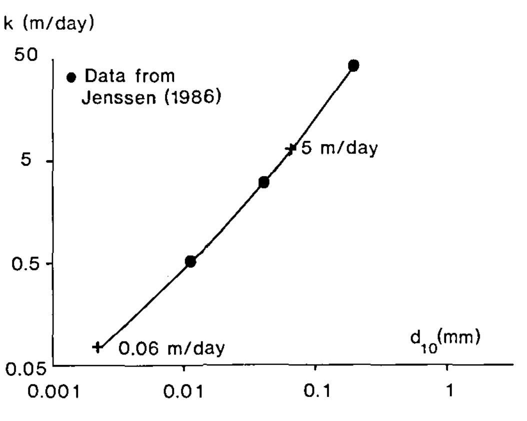

Fig. 9. Relationship between hydraulic conductivity and particle-size (d_{10}-value in mm).

A water molecule which moves with a filter velocity of 0.14 m/day will in 30 years move 1500 m. The distance from the groundwater divide to the nearest lake or river varies between 1 and 4 km (Fig. 2). The agreement between calculated average travelling distance (1500 m) and the expected travelling distance (Fig. 2) is reasonable when we remember the assumptions made during the calculations of an average residence time of 30 years and a filter velocity through sand of 0.14 m/day.

Summary and conclusions

The aquifer studied is a glaciofluvial deposit with thick sand and gravel underlain by glaciomarine silt and clay. A hydrogeological model shows that 50% of the annual precipitation is lost due to evapotranspiration. Close to 60% of new groundwater is formed during a 3-5 week long snowmelting period in spring. The unsaturated zone varies between 1 and 30 m. The amount of new groundwater infiltrating during one year is equivalent to the amount adsorbed at field capacity in a 6 m-deep unsaturated zone.

The average melting rate in 1966 was 13.5 mm (water) per day.

Water moves through the unsaturated zone during snowmelt with filter velocities close to 0.2 m/day and the degree of saturation was close to 20%. Lower velocities (0.1 m/day) were found during the summer. This is probably due to a lower degree of saturation, which will reduce the hydraulic conductivity.The major part of groundwater flow is through the upper sandy part of the aquifer. Average residence time for the groundwater in these coarse deposits is close to 30 years, with a filter velocity in the range 0.1-0.2 m/day for flow in the saturated zone.

Acknowledgements

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