

A GEOHYDROLOGIC STUDY OF EAST GELDERLAND (NETHERLANDS)

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ABSTRACT

This paper deals with the results of a geohydrological investigation carried out in the eastern part of The Netherlands (province Gelderland). In this region the demand for water for domestic and industrial use is rapidly increasing. The present demand of 34 million m³ per year is entirely met by extraction of groundwater. With a view on the agricultural interests the problem arises what effect a further increase of the groundwater extraction may have on the productivity of the drought-sensitive soils. This study was made to provide basic data for the solution of this problem. The geological investigations have given a much better insight in the suitability of the aquifers for groundwater exploitation than existed. The transmissivity of the aquifers according to field pumping tests and well logs, varies from about 300 to 10,000 m²/day. This variation is chiefly due to fluvioglacial erosion and aggradation during the Saale Ice Age (buried glacial channels). Utilizing a finite differences equation, in which the available transmissivity values and water table gradients had to be substituted, a map has been prepared showing the intensity of the net subsurface inflow. This flow rate is of practical importance because adding to it the mean precipitation and evaporation values immediately yields the mean drainage intensity. Transmissivity values and values of the drainage resistance were used for the calculation of the drawdown of the water table by extraction of groundwater from deep wells. In general, the groundwater is fresh but remarkable differences in electrical conductivity, sulphate-chloride ratio, hardness and iron content were found in the region. At some places the groundwater temperature was also measured. The differences in chemical composition and in temperature of the groundwater were found to be in fairly good agreement with the occurrence of positive and negative net subsurface inflow.

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1. INTRODUCTION

In The Netherlands the needs for domestic and industrial water are rapidly increasing. Provisional estimates show that in the next three decennia this demand will grow to a fourfold of the present demand. Now more than 60 per cent of the needs is met by extraction of groundwater from the eastern and southern higher parts of the country and from the dune areas along the coast.

Intensification of the extraction of groundwater will inevitably result in a lowering of the groundwater table. Since the growth of the crops depends on the available soil moisture, which, in its turn, depends on the depth of water table (presently 0.5 to 2 meters

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below soil surface) a further increase of groundwater extraction could cause damage to agriculture in most of the regions. It is therefore of more than academic interest to know to what extent the extraction of groundwater can be increased without doing too much harm to agriculture.

Although the population density of East Gelderland is below the country's average, the supply of high quality drinking water may become a serious problem in this region. East Gelderland is still predominantly rural and has a population of about 340,000. The present water demand is 34 million m^3 per year, of which 14 million m^3 for domestic and 20 million m^3 for industrial use. To date these quantities are entirely withdrawn from the groundwater. Recent prognoses of the population growth in the region suggest an increase to about 680,000 at the end of this century. By that time the water demand will

increase to about 136 million m^3 per year, or a fourfold of the present needs (W.O.G., 1967). It is not advisable to withdraw this quantity entirely from the groundwater reservoir because of the agricultural interests mentioned above. The Water Supply Company "Oostelijk Gelderland" has estimated that of the total demand some 98 million m^3 can be withdrawn from the groundwater, whereas the remaining 40 million m^3 will have to come from other sources.

The problem is to find the best ways and means of ensuring that all reasonable future needs for water in this region can be met. The Institute for Land and Water Management Research started an extensive study of the region with the aim to find adequate solutions to the problem. The study included a hydrogeological analysis of the region, the principal results of which will be discussed in this paper.

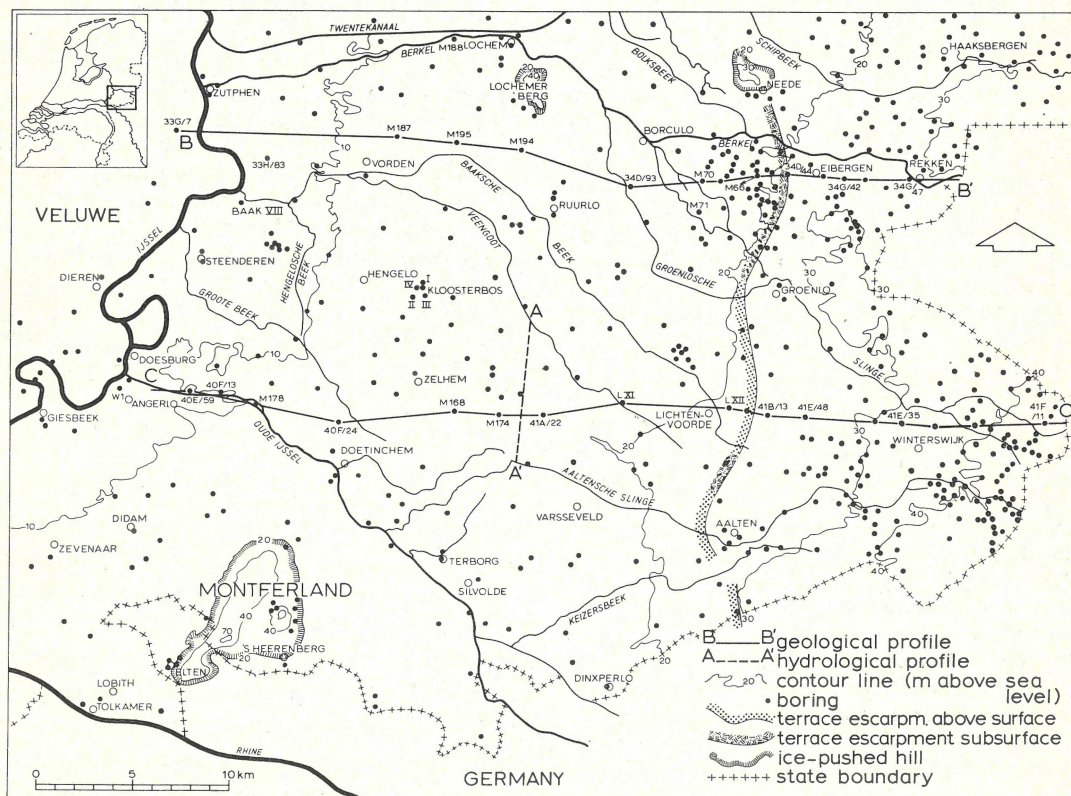


Fig. 1

Map of the eastern part of the province of Gelderland, showing major topographic features and location of borings and water wells. A - A' refers to hydrologic-section shown in fig. 2, B - B' and C - C' refer to geologic cross sections shown in fig. 3.

2. DESCRIPTION OF THE REGION

2.1. Location and size of the area

The investigated region is located between the German frontier and the river IJssel, a branch of the river Rhine. It is bounded in the North by the river Berkel and in the South by the river Oude IJssel (fig. 1). The region covers approximately 100,000 hectares; it forms a part of the drainage basin of the IJssel.

2.2. Geomorphology

From a geomorphologic viewpoint two landscapes can be distinguished: the East Netherlands plateau and the Pleistocene basin. The plateau extends from the German frontier to the line which runs from the village of Aalten in the South, via Lichtenvoorde and Groenlo towards Eibergen in the North. Here marine Tertiary clays and clayey fine sands and Mesozoic rocks are found at or close to the land surface. This surface is rather flat and is sloping from about 45 m above sea level in the east to about 25 m above sea level in the west, at the rim of the plateau.

Towards the west, the plateau passes rather suddenly into the Pleistocene basin, due to a westerly dip of the Tertiary deposits along the line Aalten-Eibergen. The transition of these two landscapes is visible in some places by a terrace escarpment of several meters high. The Pleistocene basin has been filled with predominantly coarse, gravel-bearing sands of fluvial and fluvioglacial origin. These coarse materials

are covered by a relatively thin layer of wind-blown sands.

Although in general flat, the Pleistocene basin when viewed in detail, shows minor relief features in the form of longitudinal sand ridges, a few meters high and separated from each other by minor depressions or by wide valley bottoms of small rivers. The general slope of the land surface is from east (about 25 m above sea level) to west and northwest (about 15 m above sea level). East of the line Vorden-Zelhem-Varsseveld the general slope of the land surface is about 1/5000, but west of this line it is about 1/1000. One of the higher parts of this landscape is the Lochemer berg, located in the north. It is an ice-pushed hill made up of coarse and fine sands which are Preglacial in age, locally with a veneer of glacial moraine material. Its elevation is 49 m above sea level. In the southwest and west the Pleistocene basin is bounded by the ice-pushed hills of the Montferland and the Veluwe reaching an elevation of about 100 m above sea level.

2.3. Precipitation excess and drainage

The precipitation in the area is fairly regularly distributed over the year. The average yearly precipitation, measured in 8 rain gauges over the period 1956 through 1965 was 790 mm. Potential evapotranspiration in the area is approximately 640 mm (Kramer, 1957). Since about 40 per cent of the area is used as arable land and 60 per cent as pasture,

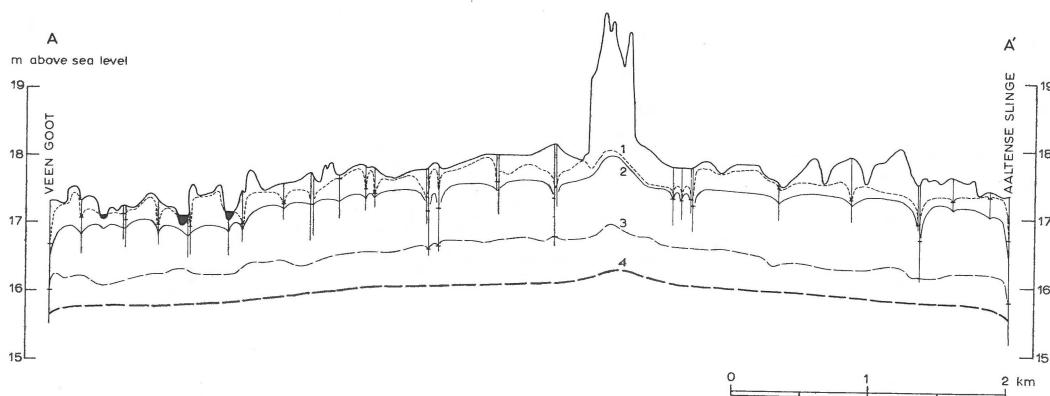


Fig. 2
Section A — A', showing the average summer and winter tables and the corresponding water levels of ditches and brooks. For location of A — A' see fig. 1. Curve 1, December 6 through 8, 1965; Curve 2, average situation during winter; Curve 3, September 10 through 20, 1965; Curve 4, average situation during summer. Curve 2 and 4 after Reuter and Kouwe, 1958.

a reduction factor of 0.75 to 0.8 can be used (M a k - k i n k, 1960; R i j t e m a, 1965). This gives an average actual evapotranspiration of about 480 mm per year.

During the summer period the precipitation is, on the average, in balance with the evapotranspiration, so that the precipitation excess of about 300 mm is restricted to the winter period. This excess water is discharged to the river IJssel by means of a dense drainage system. The quantity of 300 mm derived from meteorological data, is in fairly good agreement with the discharge measurements in the Baakse beek, carried out by the Provincial Service for Water Management. The average discharge of this rivulet over the period 1956 through 1965 was about 315 mm per year, of which 250 mm was discharged within the winter period and 65 mm in the summer period.

The groundwater table in the region is high, about 40 cm below ground surface in the winter and about 130 cm below surface in the summer period (R e u - t e r and K o u w e, 1958). Fig. 2 shows an example of the groundwater table fluctuations from which can be seen that more conduits are involved with the discharge of water in winter than in summer.

Three classes of conduits can be distinguished. Primary conduits with a spacing of about 1500 to 6000 m, such as the larger rivers, brooks and other drainage channels of major importance. Secondary conduits with a spacing of about 250 to 1000 m, as large ditches and other medium sized drainage channels. Tertiary conduits, such as small ditches, trenches, furrows and subsurface tube drains, with a spacing of the open conduits of about 50 to 200 m, and of the tube drains about 10 to 40 m.

Although they are narrowly spaced, the insufficient discharge capacity of the main conduits is the cause of inundations in the lower parts of the area occurring after excessive precipitation in winter. The eastern plateau, because of its small transmissivities, has rather high discharge peaks which contribute to the floods in the lower parts of the Pleistocene basin (B o n, 1968a and b).

3. GEOLOGIC FORMATIONS

The subsurface geological conditions of the region, though complicated, were fairly well known because in previous years a number of groundwater explora-

tion and production wells have been drilled by drinking water supply companies. In order to collect geologic and hydro-geologic information in those parts of the region where data were lacking or very scanty, an additional 130 exploration wells have been drilled for the present investigations. Samples from these wells have been lithologically described and when needed heavy-mineral analyses have been made to determine the stratigraphic position of the layers. In most of these wells a shallow and a deep filter were installed to measure the piezometric heads. Water samples were withdrawn from the deep filters to determine the chemical composition of the deep groundwater. A rather detailed description of the geological conditions of the region has been given by D e R i d d e r (1966).

Furthermore, geo-electrical surveys have been carried out in some parts of the area, for example to determine location, extent and thickness of water bearing layers and to trace buried glacial channels (C s o n k a, 1967). The results of the above mentioned investigations can be summarized as follows (see also fig. 3).

In the eastern part of the region Mesozoic rocks, chiefly composed of shale, marl, sandstone and limestone, are found at shallow depths. This area has strongly been afflicted by tectonic movements and several synclines, anticlines and horst and graben structures can be distinguished (H a r s v e l d t, 1963).

Tertiary formations, chiefly Oligocene and Miocene, are wedging out against the Mesozoic rocks. These formations are made up of marine heavy clays and clayey fine sands. At the rim of the East Netherlands plateau these marine beds dip under the Pleistocene basin where they thicken westward. During the Pliocene the sea retreated gradually from the region but evidence exists that in the central part of the basin, near the village of Hengelo, marine Lower Pliocene is present at a depth of about 40 m below sea level. Pliocene fine sands and clays of fluvial origin are found at many places in the basin, though at different depths and in varying thicknesses, indicating post-depositional erosion.

At the beginning of the Pleistocene the sea covered only the most western part of the basin, as is proved by the presence of marine Icenian at a depth of 60 m below sea level near the village of Steenderen (V a n R e e s V e l l i n g a and D e R i d d e r, 1966). Thereafter sedimentation was exclusively continental.

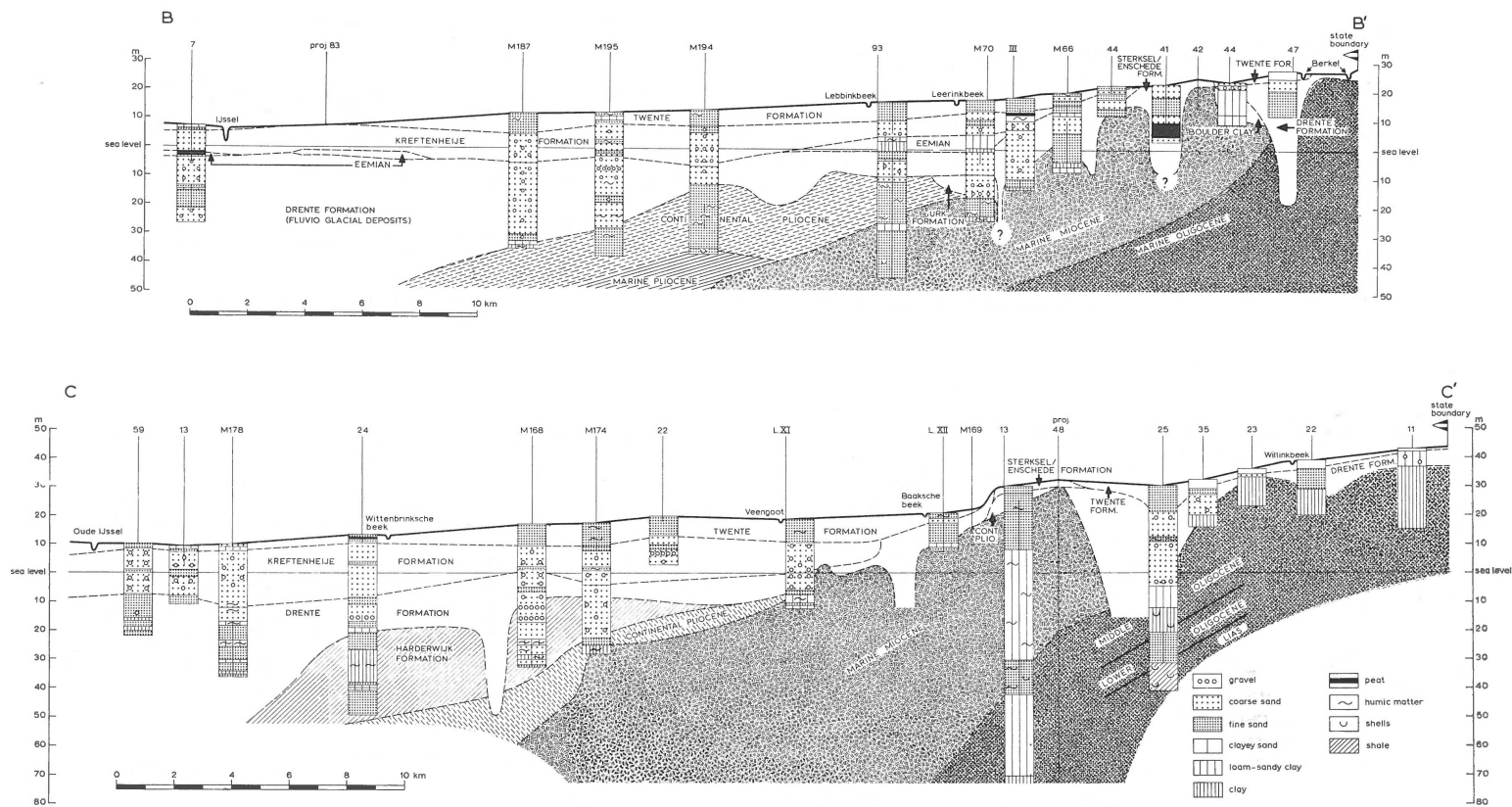


Fig. 3
Geologic cross sections B – B' and C – C'. For location of these sections see fig. 1.

During the Middle Pleistocene the river Rhine deposited the Sterksel-Enschede formation. Subsequent, climatic changes caused the sea level to drop increasing the erosion. In the Pleistocene basin most of these sediments were eroded away and even the underlying Pliocene and Miocene layers were partly removed. Relatively deep and narrow channels were formed, one of which could be traced from the village of Dinxperlo in the south to the village of Ruurlo in the north. It is filled with very coarse, gravel-bearing sands of the river Rhine which are rich in volcanic minerals, especially augite. These sands belong to the Urk formation and are probably from the Holstein Interglacial to Early Saale Glacial.

From a hydrogeological point of view the Drente formation and the Kreftenheye formation are the most important ones, because they are coarse-grained and rich in gravel thus forming the main aquifer of the region. The Drente formation (consisting of fluvioglacial deposits) was laid down during the Saalian when the entire region was covered by land ice. During the advance of the ice the initial drainage pattern was disturbed and new deep channels came into being, taking a west to southwesterly course. One of the largest buried glacial channels in the region, filled with coarse melt water deposits, can be traced from the German frontier on the eastern plateau via the towns of Winterswijk and Aalten to the village of Dinxperlo, where the bottom of the channel lies more than 45 m below sea level (De Vries and Van Rees Vellinga, in press). The thickness of the Drente formation varies from a few meters to about 70 m.

After the land ice had retreated the river Rhine did flow across the region and deposited a 10 to 20 m thick layer of gravel-bearing sand (Kreftenheye formation) in a braided river system. The Rhine did not overfill the Pleistocene basin and therefore the formation is not present on the East Netherlands plateau. Sedimentation continued in the Weichselian or last glacial period during which the land ice did not reach this region. During the Eemian (Saale-Weichsel Interglacial) fine clay-bearing sands and clay and peaty clay were deposited in local depressions. The Eem formation became known from borings in the vicinity of the villages of Borculo, Groenlo and Lichtenvoorde where it occurs at the basis of the Kreftenheye formation or intercalated within this formation.

Toward the end of the Weichselian the river Rhine took a more westerly course and under the periglacial

conditions wind was the principal agent of deposition. The Rhine deposits were covered with a layer of fine wind-blown sands (Twente formation) which varies in thickness from 0 to 12 m. These fine sands were laid down in low ridges, giving the present landscape a slightly undulating relief. Alluvial deposits, chiefly loamy sands and loam, are restricted to valley bottoms of brooks and rivers.

4. WATER-BEARING LAYERS

On lithological grounds the Kreftenheye, Drente and Urk formations can be considered to be the principal water-bearing formations of the Pleistocene basin. Since these formations are interconnected, they form one aquifer bounded at its basis by Old Pleistocene and Tertiary deposits of low permeability.

In general, the groundwater within this aquifer may be considered to occur in a phreatic state. Semi-pervious layers or lenses within this aquifer are present only in the area of Lichtenvoorde-Borculo, in front of the terrace escarpment and under the IJssel valley in the West. Therefore, the groundwater in the latter areas may show a slight artesian pressure.

The total thickness of the aquifer is shown in fig. 4. The Twente formation, which covers the aquifer and which is chiefly composed of fine sand and sandy loam, is not included in this isopach map. It can be seen that on the East Netherlands plateau the aquifer is thin, except in the buried glacial channels. In the Pleistocene basin the thickness varies from about 10 m along the terrace escarpment to more than 40 m in the western part of the region.

5. TRANSMISSIVITY OF THE AQUIFER

The transmissivity of the aquifer has an important influence on the flow of the groundwater. It is defined as the product of the average hydraulic conductivity k for horizontal flow in the aquifer and the thickness D of the aquifer. In many Anglo-Saxon publications the transmissivity is denoted by the symbol T . However, in this article it will be indicated as kD -value because this term is commonly used in The Netherlands. The transmissivity values were obtained from the data of a fairly large number of pumping tests and from the well logs of many old and new wells drilled in the region.



Fig. 4
Isopach map of the water-bearing layers, comprising the Kreftenheye, Drente and the Urk formations.

5.1. Transmissivity values estimated from well logs

It is well known that there is a relation between hydraulic conductivity and particle size of the sediment. It is a common practice to express the grain-size distribution by a single parameter for which in The Netherlands often the so-called U-figure is used. This U-figure is the surface/volume ratio of the grains, multiplied by 1/6. This multiplication means that in the case of spheres with 1 cm diameter, the U-figure

will become $\frac{1}{6} (4\pi r^2 : \frac{4}{3}\pi r^3) = \frac{1}{2r} = 1 \text{ cm}^{-1}$.

Often another definition is given (Fahmy, 1961; De Ridder and Wit, 1965) which results in the same numerical values for the U-figure with neglect, however, of the correct dimension cm^{-1} . According to various investigations the proportionality factor between k (m/day) and $1/U^2$ (cm^2), or in other words the product kU^2 , varied at a porosity of 0.40 from 31×10^3 to 71×10^3 (in units of $100 \text{ cm}^{-1} \text{ day}^{-1}$). For

the fluvial Pleistocene sands of the southern part of The Netherlands De Ridder (1967) and De Ridder and Wit (1965) found that a proportionality factor of 50×10^3 gave reasonable results when compared with those of field pumping tests.

In the investigated area the geological conditions differ from those in the southern Netherlands and another relationship between k and U can be expected, chiefly due to the occurrence of gravel layers. When the estimated kD -values were compared with those obtained from the results of pumping tests, a proportionality factor of 125×10^3 was found for the area between the river Oude IJssel and the line Lichtenvoorde-Zelhem-Doetinchem. In the remaining part of the region reasonable results were obtained when a proportionality factor of 70×10^3 was used.

In the new exploration wells a sample was taken every one or two meters of the profile. Of each sample the grain size was estimated with the aid of a binocular microscope and expressed in an U-figure.

Since the thickness of each layer was known, the transmissivity of the entire aquifer at the well site could be calculated. A similar procedure was followed to estimate the kD -values from old well logs. It should be noted, however, that these results may be considerably less accurate because descriptions of grain size in driller's logs are sometimes subject to large errors. By applying this method the transmissivity of the aquifer was estimated at more than 200 sites.

5.2. Transmissivity values derived from pumpig test data

The most reliable values of the transmissivity are obtained from the field pumping tests. For the present study such tests were performed at 15 different sites. Moreover, the results of another 10 pumping tests, carried out in the past for different purposes, were found in the Hydrogeological Archives of the Governmental Institute for Drinking Water Supply.

In table 1 the transmissivity values derived from the results of these 25 pumping tests as well as the transmissivity values estimated from grain-size data obtained from borings in the vicinity of some test sites, are given. Based on these results a transmissivity map of the entire region was compiled (fig. 5). This map can be used for a first selection of the locations for groundwater extraction. It can be seen that the East Netherlands plateau is characterized by a very low transmissivity due to the fact that the aquifer is very thin or at some places is even lacking. From the terrace escarpment to the West the transmissivity increases rapidly and reaches its highest values in the middle of the Pleistocene basin. In the most westerly part of the basin the transmissivity decreases slightly as there the aquifer contains more fine-grained material.

Besides considerable variations in aquifer thickness there are also large variations in hydraulic conductivity of the aquifer material. From the pumping tests the mean hydraulic conductivity was found to vary from 20 to more than 130 m/day.

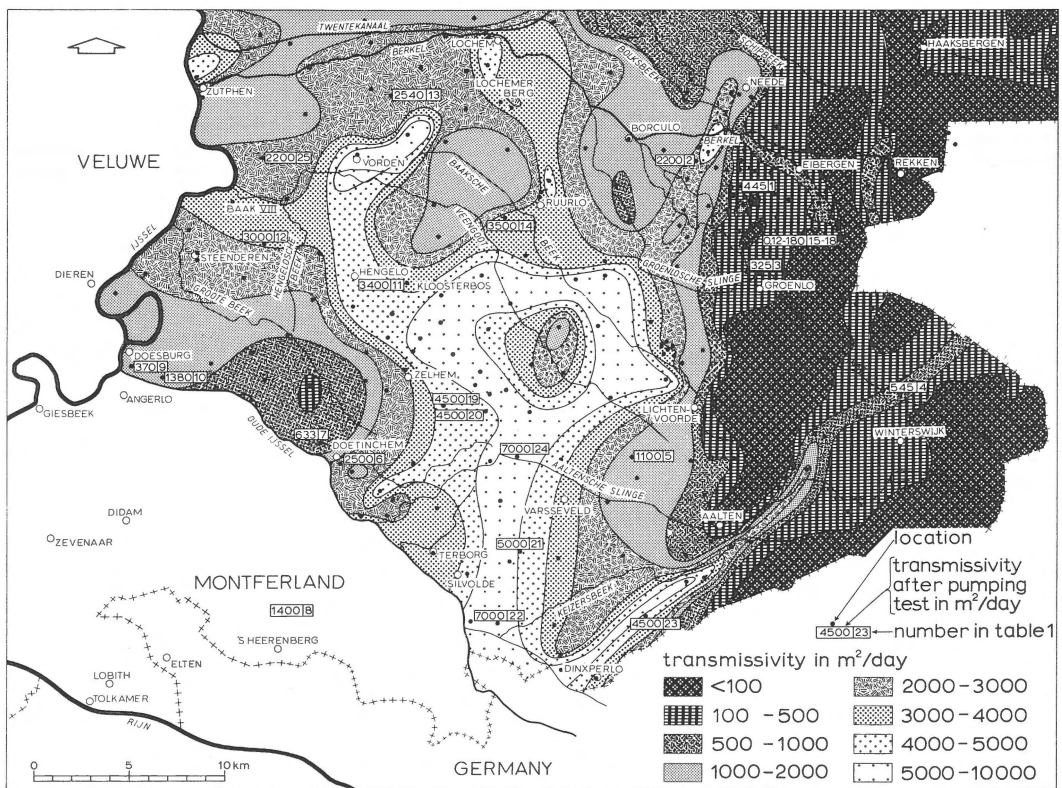


Fig. 5
Transmissivity map of the water-bearing layers.

TABLE 1

Transmissivity values derived from pumping test data and from well logs

Test number (see fig. 5)		Transmissivity in m^2/day	
		from pumping tests	from well logs
1	Eibergen, 1931	445	---
2	Haarlo M 74, upper layer	290	---
	lower layer	1940	---
3	Groenlo	325	180
4	Winterswijk, M202	545	---
5	Lichtenvoorde, 1966	1100	1250
6	Doetinchem, 1953/56	2500	---
7	Doetinchem, 1957	633	---
8	's-Heerenberg, 1962	1400	---
9	Doesburg, upper layer, 1946	370	---
10	Drempt, 1946	1380	---
11	Hengelo, 1964	3400	4400
12	Baak, 1965	3000	2400
13	Vorden, 1964	2540	2700
14	Ruurlo, 1965	3500	3950
15	Groenlo I, 1963	0.12	---
16	Groenlo II, 1963	0.05	---
17	Groenlo III, 1963	186	---
18	Groenlo IV, 1963	12	---
19	Zelhem, M168	4500	4100
20	Zelhem, M174	4500	4500
21	Varsseveld, M200	5000	3750
22	Silvolde, M165	7000	8750
23	Dinxperlo, M163	4500	5200
24	Varsseveld, M199	7000	6600
25	Vierakker, M219	2200	---

6. GROUNDWATER LEVELS

The depth to the groundwater table is in periods of drought an important factor for crop growth. Lowering of the groundwater table by pumping may cause damage to agriculture. Such lowering of the groundwater levels depends on the amount of extraction and on the groundwater flow. The direction of this flow can be detected from groundwater elevation contour maps. Multiplication of kD -value and hydraulic gradient yields the horizontal flow intensity.

6.1. Piezometric surface

More than 200 deep observation wells exist in the area. Most of these wells have a filter in the main aquifer, and their water levels are measured twice a month. Since the records of April 28, 1966 were found to be the most complete, this information was used to compile a contour map of the piezometric

surface of the region outside the eastern plateau (fig. 6).

When compiling this map, minor local depressions in the piezometric surface near ditches and brooks could not be taken into account because the number of deep observation wells located near these open water courses was insufficient. As can be seen from this map the general direction of the groundwater flow is towards the rivers IJssel, Oude IJssel and Berkel. Steep hydraulic gradients of 1/160 to 1/250 are found along the terrace escarpment, whereas in the level middle part of the basin smaller gradients of 1/3000 to 1/4000 occur. West of the line Vorden-Hengelo-Varsseveld the hydraulic gradients increase to 1/700 to 1/1000, whereas in the utmost western part gradients of 1/1500 to 1/2000 are found.

This map permits to make a calculation of the inflow and outflow of groundwater across the boundaries of the area. By applying Darcy's law on the boundaries of an area of 58,000 hectares, a net

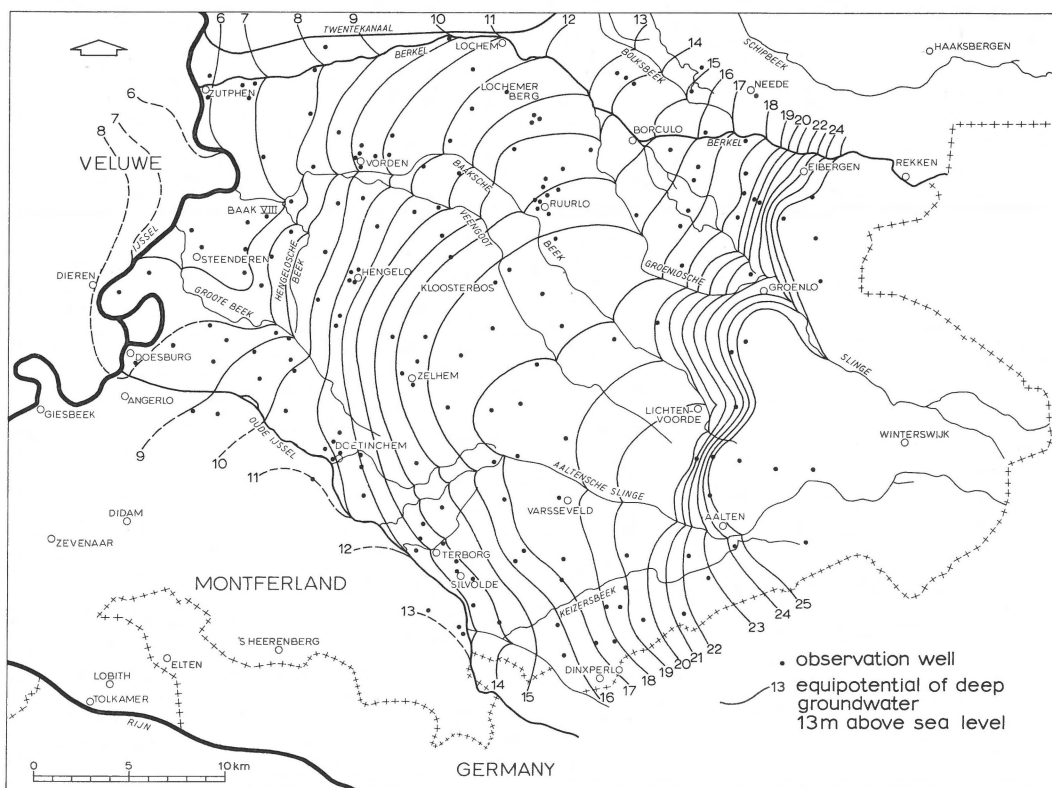


Fig. 6
Contour map of the piezometric surface on April 28, 1966.

subsurface outflow of $44,000 \text{ m}^3/\text{day}$ was found, which corresponds to 0.08 mm/day . This figure is much smaller than precipitation minus evaporation (about 300 mm/year , see section 2.3.) and in accordance with the heavy drainage by secondary and tertiary open water courses.

6.2. Phreatic surface

A contour map of the phreatic surface is presented in fig. 7. This map is based on measurements in about 750 shallow observation wells (Reuter and Kouwe, 1958; Bon, 1968b). As can be expected for a continuous aquifer under phreatic conditions, the general pattern of the contour lines looks very similar to that on the map of the deep groundwater as presented in fig. 6. The contour lines in fig. 7, however, are showing upstream bends caused by the radial groundwater outflow to the smaller open water courses.

6.3. Groundwater fluctuation

The contour maps of the groundwater depth both for average summer and average winter conditions published by Reuter and Kouwe (1958) have been used to compile a map of the difference between the summer and the winter water tables (fig. 8).

It can be seen from this map that the largest fluctuations of the water table occur chiefly in the middle part of the region, extending from Vorden in the north, via Zelhem and Varsseveld to the German frontier in the south. In this area water table fluctuations of about 1.5 m occur which can be explained by the relatively high elevation of this area and the scarcity of tertiary open water courses.

On the average, the water table fluctuations in the major part of the region are of the order of 1 m . Minor fluctuations of less than 0.5 m also occur but they are limited to areas in front of the terrace

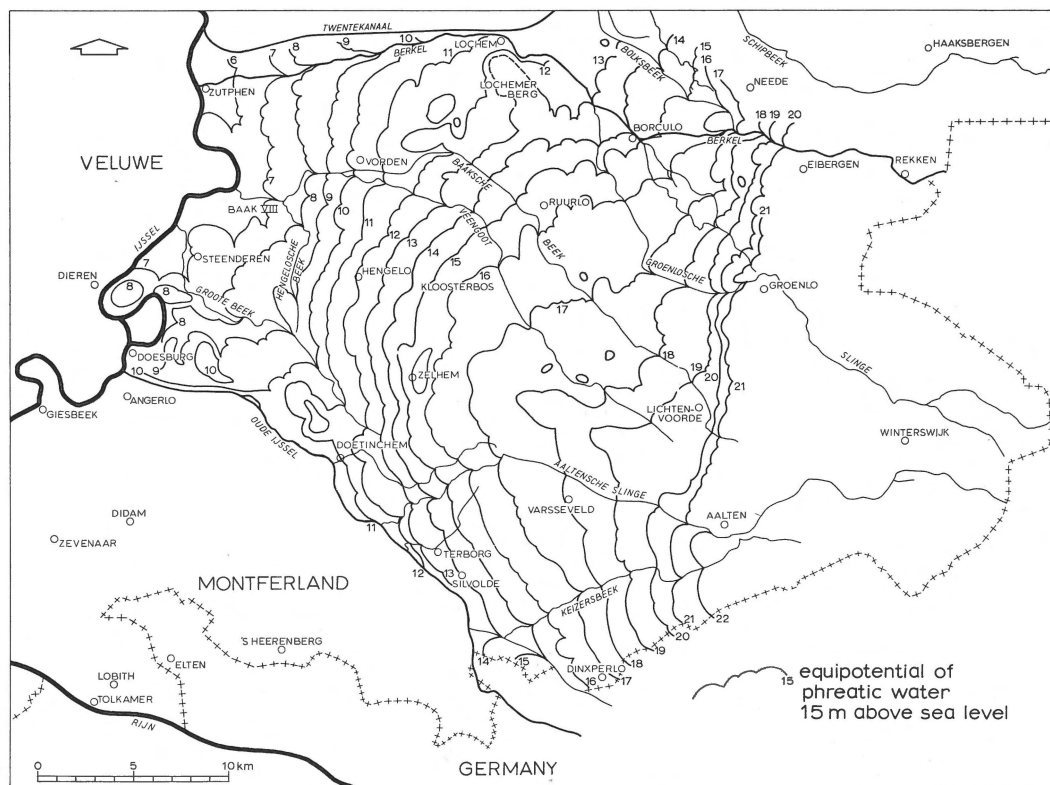


Fig. 7

Contour map of the phreatic surface. Average water table for the winter periods 1952 through 1955 (after B o n, 1968b, based on data of R e u t e r and K o u w e, 1958); 1 shallow observation well per 130 ha.

escarpment, to the transition zone of the gently sloping area west of the line Zelhem-Vorden, and to the flat IJssel valley. Finally, it should be noted that the fluctuations in the narrow strips of land bordering the rivers are strongly related to the changes in water level of these rivers. Some of these rivers, as for example the Oude IJssel, have weirs to maintain a constant water level.

6.4. Differences in hydraulic head between deep and shallow groundwater

The available contour maps could not be used to determine the difference in hydraulic head between the shallow and the deep groundwater, because these maps refer to different dates. Moreover, the accuracy is insufficient due to errors in measurement and interpolation. However, there are 58 observation wells which have a deep filter in the aquifer and

another filter in the zone in which the water table fluctuates. The differences in hydraulic head measured in the two filters of these wells on the November 7, 1968, presented in fig. 9, varied from -34 to $+34$ cm. In the middle part of the Pleistocene basin the hydraulic head of the deep water is smaller than that of the shallow groundwater, which means that a downward flow of groundwater must exist. In front of the terrace escarpment, along some of the major water courses (e.g. the river Berkel) and near the village of Baak on the other hand, the head of the deep water is larger which means that an upward flow of groundwater will be present.

The maps of figs. 8 and 9 show the same trend which can be explained by the fact that the lower areas have in general an upward flow of groundwater and small drain spacings, preventing large seasonal water table fluctuations.

Fig. 10 gives an example of the course of the water

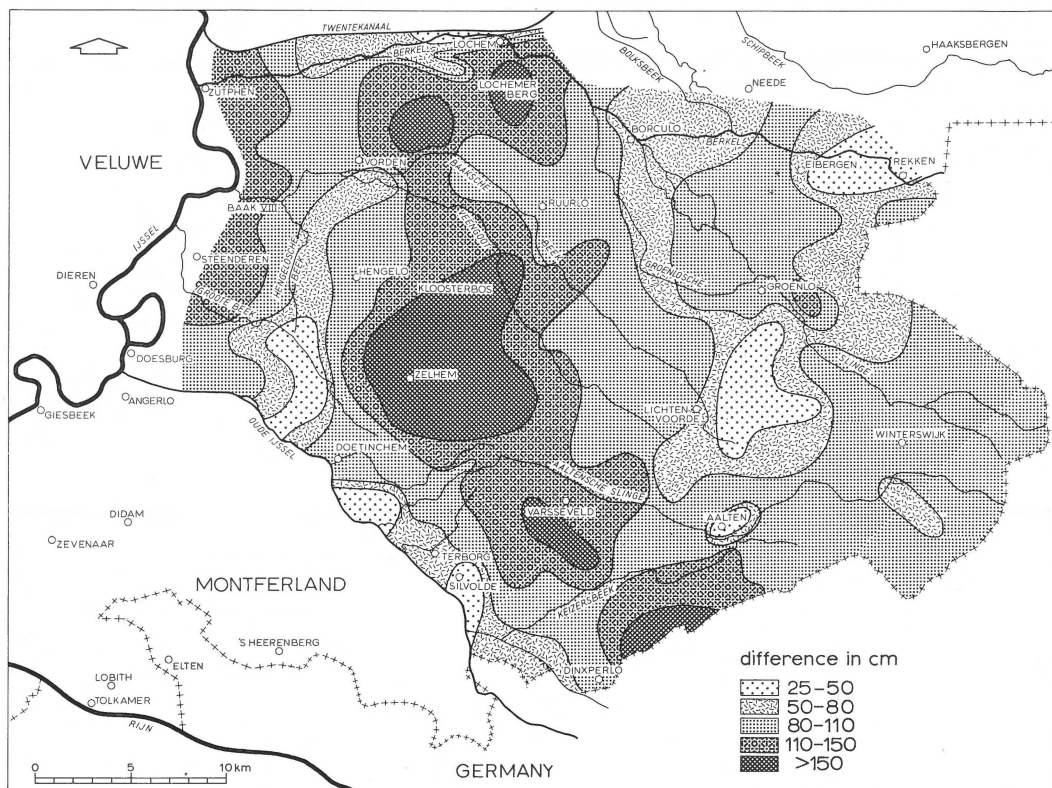


Fig. 8

Difference between the average summer and the average winter water table over the period 1952 through 1955, derived from water table maps published by Reuter and Kouwe (1958).

levels versus time in wells which are located in topographically high respectively low areas (with downward respectively upward groundwater flow). Because of the rather constant differences in hydraulic head between the deep and shallow groundwater, there is not much variation in this upward and downward flow with time, just as the intensity of the deep groundwater flow is only slightly time dependent.

Apparently there is a more or less permanent flow of groundwater towards certain regions, whereas other regions may loose considerable amounts of groundwater. An analysis of this phenomenon can provide quantitative data about the magnitude of the flows.

7. GROUNDWATER FLOW ANALYSES

7.1. Water balance equation and net subsurface inflow

For the investigation of groundwater flow problems of relatively large areas, the following water balance equation is widely used. The boundaries of the area for which eq. (1) is valid are shown in fig. 11.

$$P + I_{ss} + \frac{Q_{inf}}{A} = E + \frac{Q_{dr} + Q_{dw}}{A} + \frac{\Delta W}{\Delta t} \quad (1)$$

where

- A = horizontal surface area
- P = precipitation
- E = evaporation
- I_{ss} = net subsurface inflow per unit of time

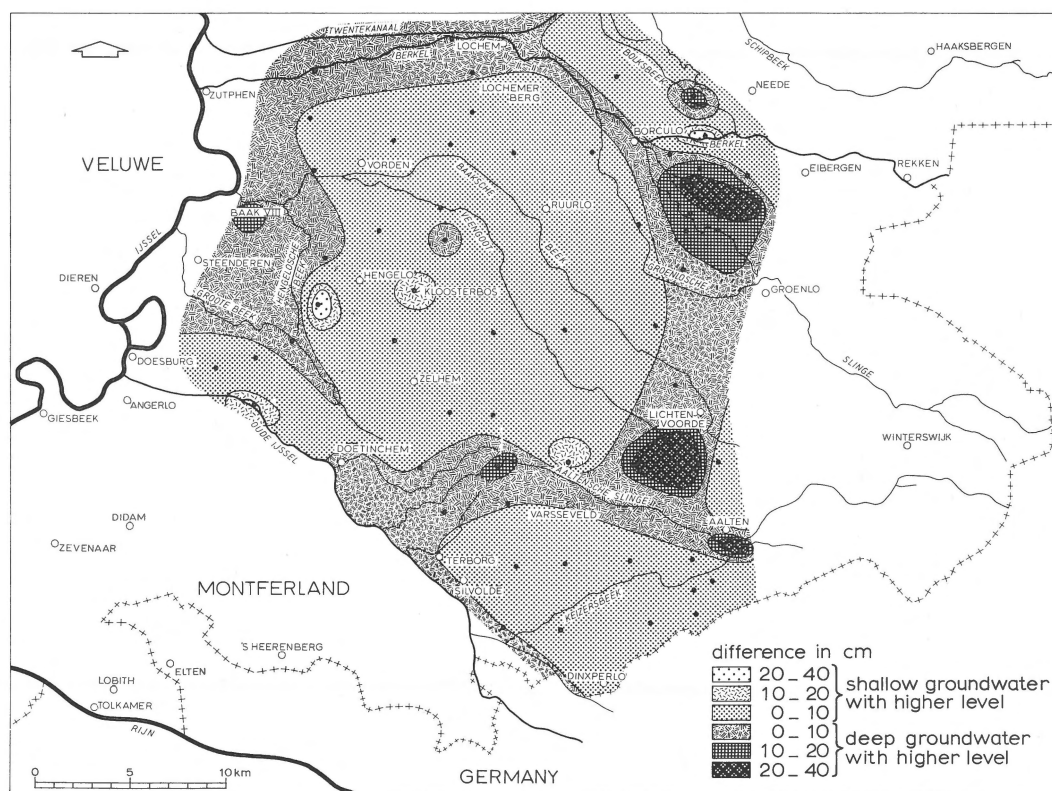


Fig. 9 Differences in hydraulic head between shallow and deep groundwater, measured on November 7, 1968 in 58 double piezometer wells with filters at 3 and 10 to 20 meter below surface.

and per unit of horizontal surface area¹⁾

Q_{inf} = influent seepage from open conduits with relatively high water levels

Q_{dr} = effluent seepage to open conduits with relatively low water levels

Q_{dw} = discharge by deep wells

$\Delta W / \Delta t$ = change in storage of water per unit of time and per unit of horizontal surface area

When both Q_{inf} and Q_{dw} are so small that they are negligible and when, moreover, only periods of long duration are considered, so the change in storage ($\Delta W / \Delta t$) is very small, the above equation can be replaced by:

$$I_{ss} = \frac{Q_{dr}}{A} - (P - E) \quad (2)$$

This formula shows clearly the practical significance of knowing the net subsurface inflow, because $A I_{ss}$ represents the difference between the amount of water which on the average has to be drained and the average supply $(P - E)A$.

It is evident that the net subsurface inflow (I_{ss}) may be positive or negative. The cause of positive or negative net subsurface inflow can always be found in differences in water level (and often also in ground surface elevation) of the area under consideration and the surrounding areas. In the Netherlands such conditions are especially known from the polder districts, where often rather large differences occur in the open water level of the drainage systems of two adjacent polders. In principle, the same phenomenon occurs in the undulating landscapes of sandy soils in the south and

¹⁾ The English language lacks an unambiguous term to go with the symbol I_{ss} , as defined in eq. (1). Net subsurface inflow is introduced as an addition to hydrologic terminology replacing various other terms presently used.

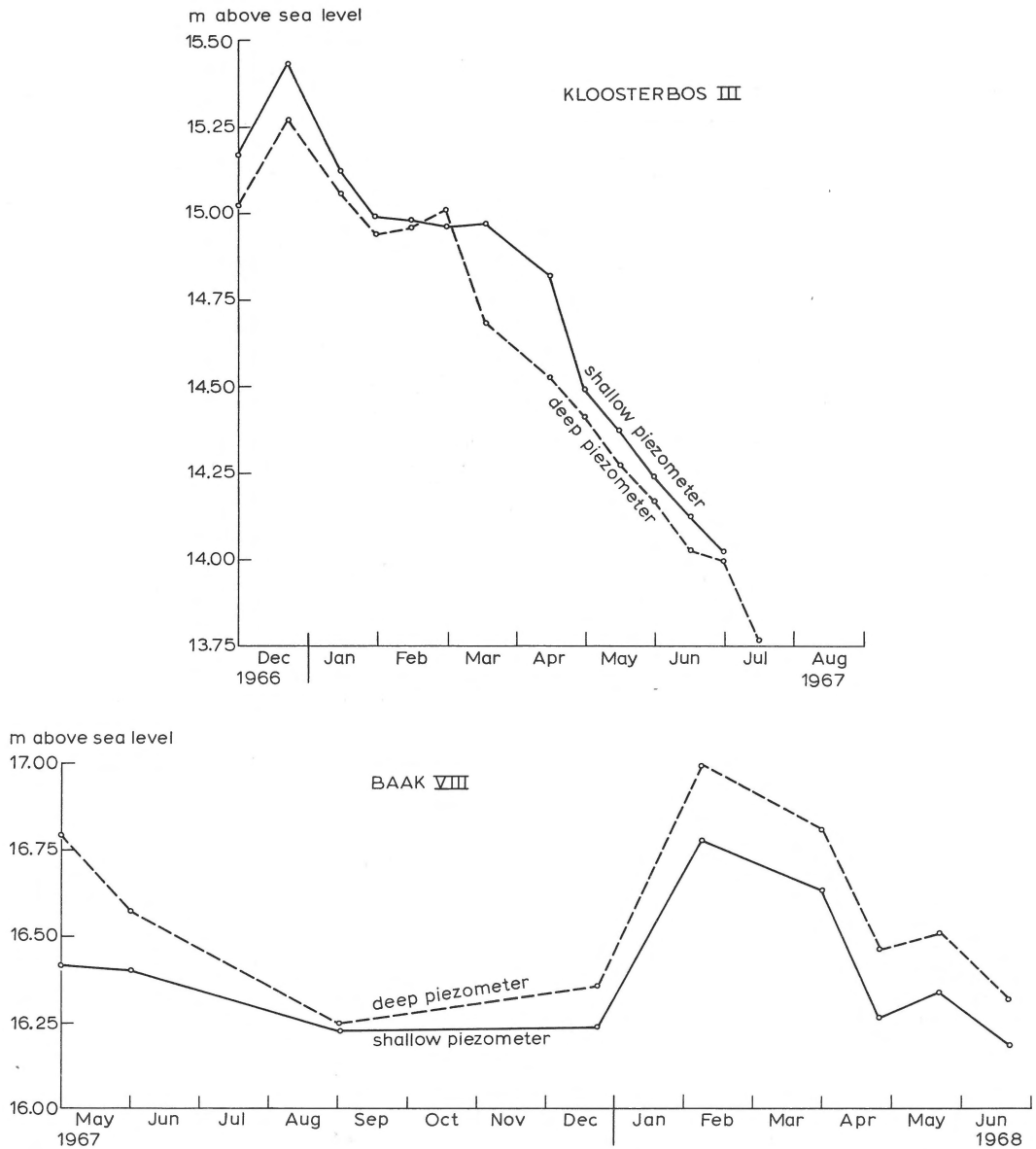


Fig. 10

Well hydrographs of the double piezometer wells Kloosterbos III and Baak VIII, showing the hydraulic heads in the shallow and the deep groundwater.

eastern parts of the country. Here the valleys of the rivers are receiving positive subsurface inflow from the higher bordering grounds (fig. 12).

The large distances over which the major differences in water level are generally found and the time independency of these differences explain the fact that I_{ss} hardly changes with time when compared

with the large and frequent variations in the drain discharge Q_{dr} caused by rain showers.

The flow of groundwater can often be considered as taking place in a horizontal aquifer or in a series of horizontal aquifers, which at the lower side are bounded by an impermeable layer. This implies that the intensity of the net subsurface inflow can be

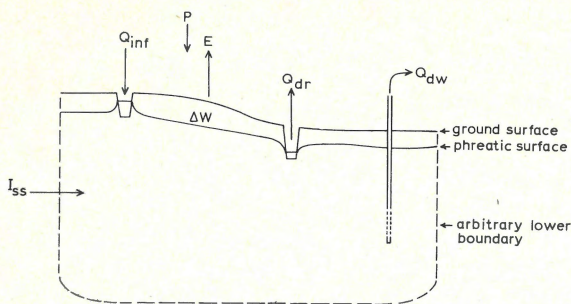


Fig. 11
Aquifer with the inflow and the outflow components contained in eq. (1).

calculated by an integration over the horizontal groundwater inflow:

$$I_{ss} = - \frac{1}{A} \oint q_{ns} ds = \frac{1}{A} \oint kD \frac{\partial h}{\partial n} ds \quad (3)$$

where

- s = length co-ordinate of the boundary of the horizontal surface area A
- h = hydraulic head
- n = direction normal to s and counted positive when directed outwards
- q_{ns} = flow component horizontal and normal to s per unit of time and per unit of length of s and taken positive when directed outwards

$$\frac{\partial h}{\partial n} = \text{hydraulic gradient normal to } s.$$

Formula (3) shows that for determination of I_{ss} only two maps are required, respectively showing contour lines of the groundwater table and of the aquifer transmissivities (kD -values).

How important the choice of the shape of the area A can be, is clearly shown when eq. (3) is replaced by:

$$I_{ss} = v_{phr} + \frac{1}{A} [\int q_{dr} ds_{dr} - \int q_{inf} ds_{inf} + \sum Q_{dw}] \quad (4)$$

where

- v_{phr} = intensity of vertical flow through the phreatic surface (positive when directed upwards)
- s_{dr} = length co-ordinate to be taken over all open conduits in area A giving some effluent seepage q_{dr}
- s_{inf} = length co-ordinate to be taken over all open conduits in area A giving some influent seepage q_{inf}

The most important conclusion to be drawn from eq. (4) is, that the boundaries of A are to be chosen in such a way that the open conduits are proportionally present in the area for which I_{ss} has to be calculated. Also, the different drain classes already mentioned in section 2.3, have to be taken into account, as the outflow depends both on size as well as on open water level of each class. Eq. (4) also shows that when the area A contains one or more pumped deep wells, I_{ss} will be rather large. The closer the boundaries of A are encircling these deep wells the more the resulting value of I_{ss} will increase.

During wet periods when the groundwater table is high, its shape is strongly affected by the smaller drains (drains of tertiary class). When, however the number of piezometers is rather small it is impossible to obtain a water table contour map which shows all these minor variations. This lack of accuracy is generally of minor importance, except for those cases where I_{ss} has to be determined for very

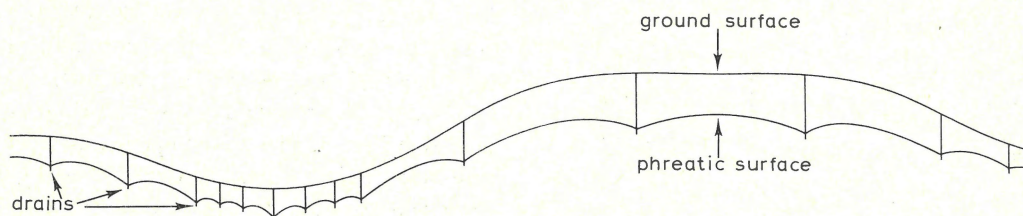


Fig. 12
In the undulating sandy areas of The Netherlands the depth to the phreatic surface and the drain spacings depend largely on the variations in the ground surface elevation (schematic profile).

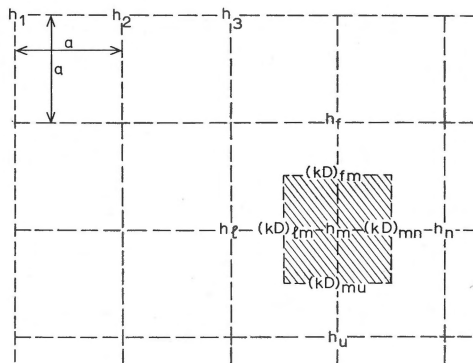


Fig. 13

Notation of hydraulic head and transmissivity used in eq. (5) for the calculation of inflow and outflow through the sides of an a^2 -area.

small areas in which a strong heterogeneity of the tertiary drains may have an appreciable influence which is not apparent on a rough water table contour map.

7.2. Finite difference method

Once a water table map of sufficient accuracy is obtained, the question arises how to perform the calculation of a large number of contour integrals or to find a simpler equation as a substitute of eq. (3). The calculation by means of contour integrals, even for a small number of sub-areas, seems less con-

venient than when a finite difference method is applied in a regular grid.

For an orthogonal grid (fig. 13), forming squares with sides a , the difference equation becomes:

$$(kD)_{fm}(h_f - h_m) + (kD)_{lm}(h_l - h_m) + (kD)_{mn}(h_n - h_m) + (kD)_{mu}(h_u - h_m) = (I_{ss})_m a^2 \quad (5)$$

Each term $(kD)_{xm}(h_x - h_m)$ in the left side of this equation can be considered as a first and fairly good approximation of the groundwater flow through one of the sides of the square shaded in fig. 13. The sum of these terms corresponds to the contour integral of eq. (3). Therefore eq. (5) can be considered to give an acceptable approximation of the average value $(I_{ss})_m$ of the net subsurface inflow for the square with nodal point m .

From eq. (4) follows immediately that when all the open conduits and pumped deep wells have a distance to the central node larger than a , the following relation is valid:

$$I_{ss} = v_{phr} \quad (6)$$

When one or more drains are separating the central node from one or more of the adjoining nodes, eq. (6) should be changed in such a way that the outflow to the drains is taken into account. For the case of closely spaced parallel drains (fig. 14) this becomes:

$$\bar{I}_{ss} = v_{phr} + \frac{\sum q_{dr}}{b} \quad (7)$$

where

$\bar{I}_{ss}$ = mean value of I_{ss} over width b
 $\sum q_{dr}/b$ = outflow to drains per unit surface area of strip b

When the nodal distance is chosen in such a way that the vertical inflow through the phreatic surface equals the outflow to the drains (e.g. area F-F in fig. 14) eq. (7) will obviously result in $\bar{I}_{ss} = 0$.

When surface drains are absent, but an important discharge by a number of pumped deep wells exists, a term equivalent to the last term of eq. (7) has to be added to take the outflow of these wells into account:

$$\bar{I}_{ss} = v_{phr} + \frac{\sum Q_{dw}}{\pi r^2} \quad (8)$$

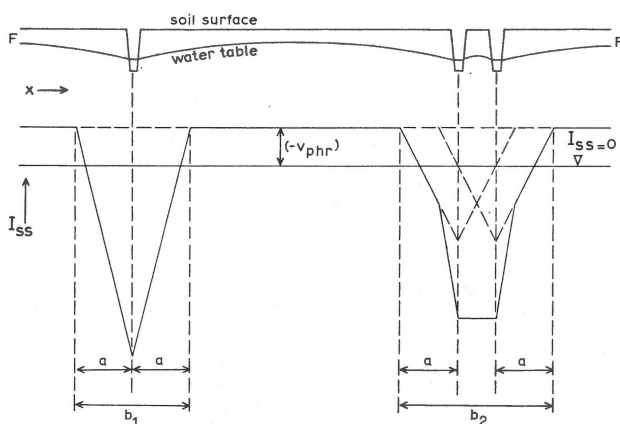


Fig. 14

Example of I_{ss} -values between parallel drainage channels when an orthogonal grid is used with one set of lines perpendicular and the other parallel to the drains, and with continual shifting in order to obtain I_{ss} -values for each value of x .

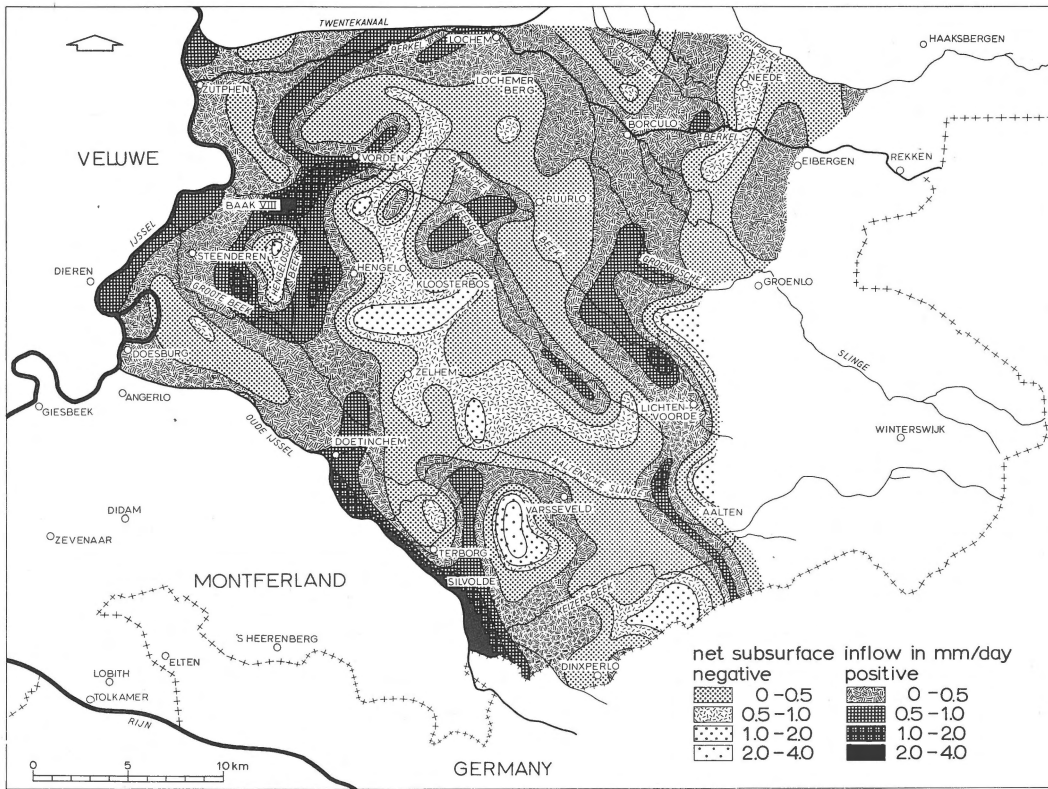


Fig. 15
Map showing the distribution of the net subsurface inflow or outflow. Inflow and outflow calculated by applying eq. (5) for $a = 2000$ m and transmissivity and hydraulic head values as given in figs. 5 and 6.

where

$$\bar{I}_{ss} = \text{mean value of } I_{ss} \text{ for the surface area inside a circle with radius } r$$

$$\Sigma Q_{dw}/\pi r^2 = \text{outflow from pumped wells per unit of the same area}$$

7.3. Application

To analyze the groundwater flow of the Pleistocene basin, an orthogonal grid with squares having sides a of 2000 meters was utilized. In order to determine the water level elevation at the nodal points this grid was superimposed on the piezometric contour map for April 28, 1966 (fig. 6). The transmissivity at midpoint between the nodes was found in a similar way by superimposing the grid on the transmissivity map of fig. 5. The intensity of the net subsurface inflow at each nodal point was then found by applying eq. (5). The results of these calculations are presented in fig. 15.

This map shows that areas of positive net subsurface inflow alternate with areas of negative net subsurface inflow. The intensity of the negative net subsurface inflow is in good agreement with the average precipitation of the previous winter months. The areas of positive net subsurface inflow are chiefly found in an elongate zone in front of the terrace escarpment between the villages of Groenlo and Aalten, along the major drainage channels as the rivers Oude IJssel, Berkel and IJssel, and west of the relatively high sandy area near the village of Hengelo in the middle of the basin. These areas are characterized by relatively high water tables, and relatively small drain distances.

The areas of negative net subsurface inflow are found on the rather high-lying terrace deposits of the eastern plateau and on the high sandy area of Hengelo in the middle of the basin, areas with lower water tables and larger drain distances. In the area bordering the lower course of the river Oude IJssel, the net subsurface inflow is not positive as would

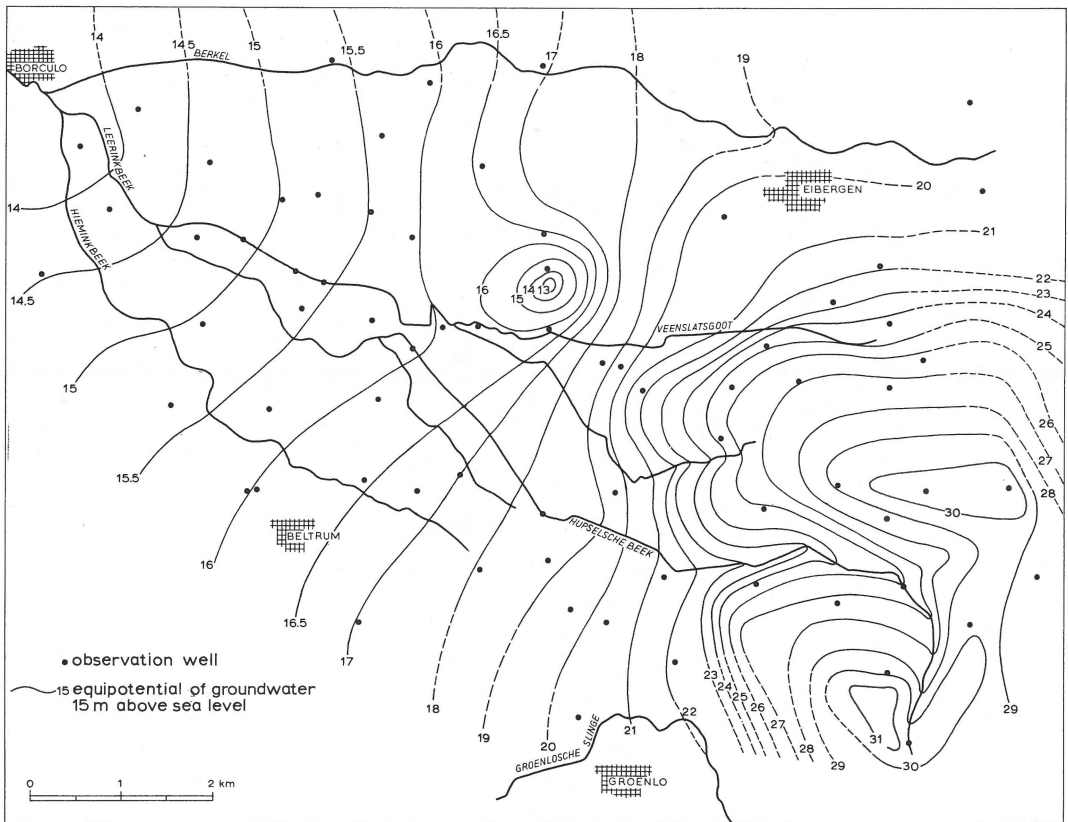


Fig. 16

Contour map of the piezometric surface in the drainage area of the Leerinkbeek on August 28, 1964 (after Colenbrander, 1968).

be expected but negative which is due to the high water level of the river caused by a weir near the town of Doesburg.

A more detailed groundwater flow analysis has been made of a small portion of the region, namely the 5000 ha large drainage area of the Leerinkbeek, between the villages of Eibergen, Groenlo and Borculo. A fairly accurate contour map of the piezometric surface in the summer period (August 28, 1964), based on measurements in more than 200 observation wells (fig. 16) has been compiled by Colenbrander (1968). This area has been analyzed in the same way as mentioned above by using an orthogonal grid, but now having squares with sides a of 500 m. The result of the calculations is presented in fig. 17.

On this map the pumping station for the drinking water supply, located to the west of the town of Eibergen, shows up clearly as an area of positive net

subsurface inflow (artificial drainage by pumping). This pumping station has an annual groundwater production of about $1.5 \times 10^6 \text{ m}^3$, causing a fairly deep depression cone. The areas of negative net subsurface inflow are better defined on this map than on the map of fig. 15, not only as a more accurate piezometric contour map was used but also because the distance between the nodal points was considerably smaller. The accuracy of analyses of this kind largely depends on the accuracy of the piezometric map and the transmissivity map.

The optimum site of a pumping station is not necessarily there where a large positive net subsurface inflow occurs, although areas with positive net subsurface inflow often have a dense drainage system which reduces the drawdown by deep well pumping. The shape of the cone of drawdown of the groundwater table by pumping depends on the amount of water withdrawn per unit of time, on

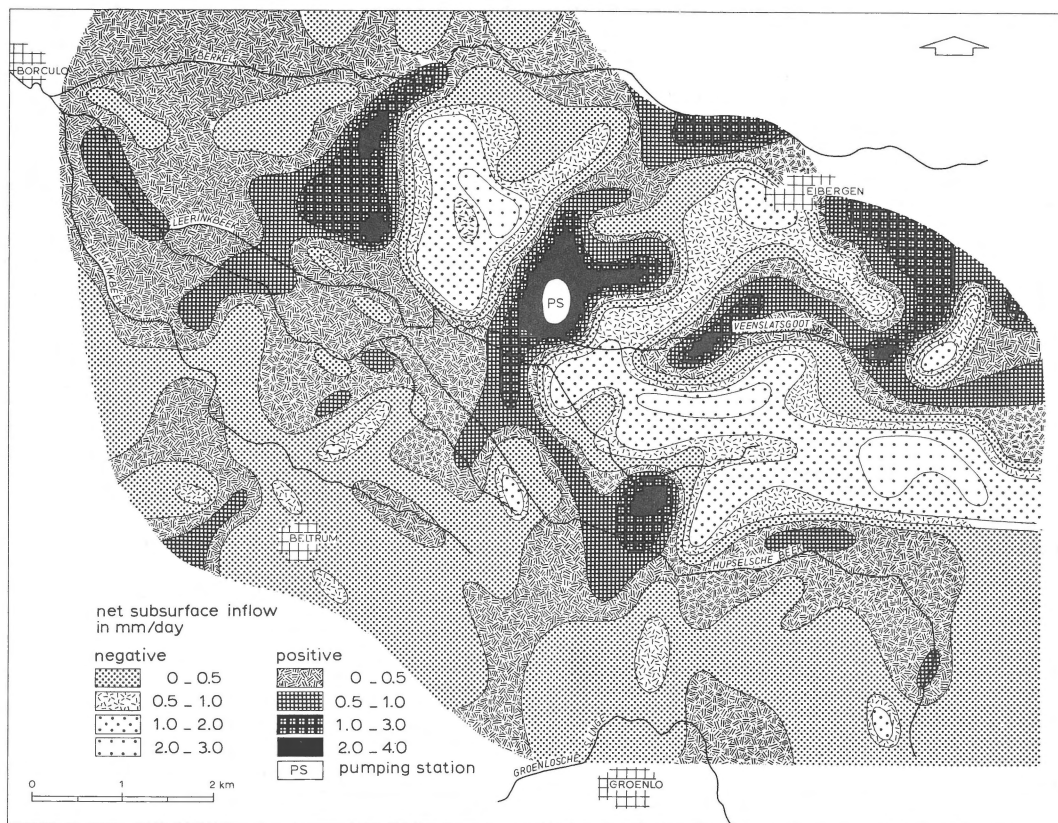


Fig. 17

Map showing the distribution of the net subsurface inflow or outflow in the drainage area of the river Leerinkbeek. Inflow and outflow calculated by applying eq. (5) for $a = 500$ m. Transmissivity and hydraulic head values as given in figs. 5 and 17.

the transmissivity of the aquifer, and on the drainage resistance (see section 8).

An adequate agricultural criterion cannot be based on the magnitude of the groundwater decline alone. Other factors, for example the distribution of arable land, grassland, forests and waste lands, the damage suffered by each crop per unit decline of the groundwater table and the present depth of the groundwater table are equally important. Other non-agricultural factors, for example transport distance from extraction site to sites of actual use or the chemical composition of the groundwater (see section 9), may be more important than the largest possible positive net subsurface inflow.

8. DRAWDOWN OF THE WATER TABLE AROUND PUMPED DEEP WELLS

As any withdrawal of groundwater results in a decline of the groundwater table, the choice of the site of the pumping station has an impact on the future agricultural use of the area. In the present case calculation of the drawdown is complicated by the presence of various kinds of drainage channels.

Both the former and the future depth of the phreatic surface will show seasonal fluctuations. However, the assumption that the difference in the two situations is nearly time independent seems to be justified as far as there is not a continuous

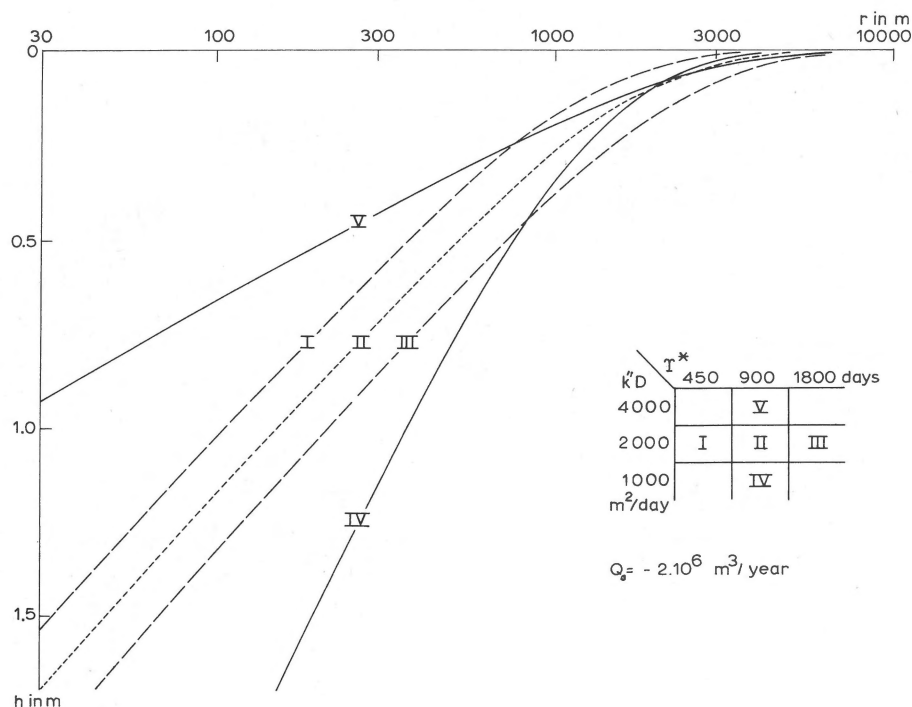


Fig. 18

Decline of the hydraulic head (h'') according eq. (11) for a deep well with constant discharge (Q_0) from a homogeneous aquifer with transmissivity kD and differential drainage resistance γ^* .

lowering of the water table by excessive pumping. Theoretical considerations indicate that the errors introduced are reasonably small. This is also supported by the available information on the effects of the pumping station "Olden Eibergen" and another pumping station west of the province of Gelderland. Also in these cases water is pumped from a phreatic aquifer. The following differential equation can therefore be used (Ernst, 1967a):

$$kD \left[\frac{d^2 h''}{dr^2} + \frac{1}{r} \frac{dh''}{dr} \right] = - [\tilde{U}' - \tilde{U}'''(r)] \quad (9)$$

where

- $h''(r)$ = time independent decline of the hydraulic head h at distance r of the pumped well
- $\tilde{U}'$ = annual mean value of the discharge of the drains before pumping (independent of r)
- $\tilde{U}'''(r)$ = annual mean value of the discharge of the drains during pumping with constant

intensity Q_0 (equilibrium state; sufficiently approached after one or two years of constant pumping)

A very simple solution can be obtained by assuming that the following substitution with constant γ^* is of sufficient accuracy:

$$- [\tilde{U}' - \tilde{U}'''(r)] \approx [\tilde{h}''' - \tilde{h}'] \left[\frac{d\tilde{U}}{dh} \right]_{\tilde{h}} = \frac{h''(r)}{\gamma^*} \quad (10)$$

where

- $\tilde{h}'$ = annual mean value of the hydraulic head before pumping
- $\tilde{h}'''$ = annual mean value of the hydraulic head during pumping (equilibrium state)
- $\tilde{h}$ = annual mean value of the hydraulic head during pumping in the most important range of r , e.g. between 500 and 2000 m (equilibrium state)
- γ^* = differential drainage resistance $\tilde{dh}/d\tilde{U}$

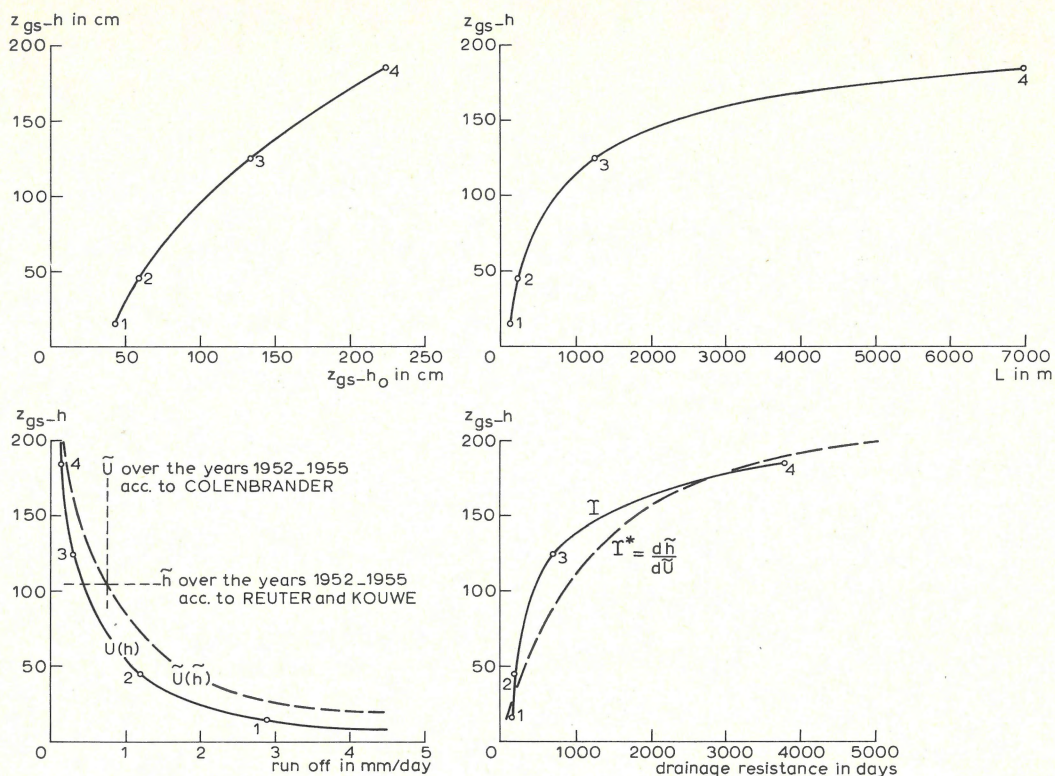


Fig 19

Relations between groundwater level h , level of open water h_0 , drain spacing L , runoff U and drainage resistances Γ and Γ^* . Drainage resistances and runoff values calculated by means of eq. (14) and (13); other quantities derived from fig. 2. z_{gs} = elevation of the ground surface; 1, 2, 3 and 4 refer to curves in fig. 2; curves 2 and 4 are mean values over 1952 through 1955.

Substitution of eq. (10) in eq. (9) and integration of the latter equation gives:

$$h''(r) = \frac{-Q_0}{2\pi kD} K_0 \left[\frac{r}{\sqrt{kD\Gamma^*}} \right] \quad (11)$$

where

Q_0 = constant discharge of the well (h'' is negative, Q_0 is positive for discharging walls)
 K_0 = modified Bessel-function of the second kind and zero order

Fig. 18 based on eq. (11) shows that for small values of r the transmissivity kD is of more influence than the differential drainage resistance Γ^* , while for larger values of r the drainage resistance is of more influence.

The result of eq. (11) is only valid when other wells, large rivers or canals are situated at infinite

(very large) distances from the well in question. In the case of finite distances the interference may be of importance. Because of the assumed linearity of the resultant differential equation, addition is allowed of two or more solutions eq. (11) and in this way the interference can very easily be taken into account. From fig. 18 it follows that in the investigated region a not quite negligible interference effect may be expected up to distances of about 5 km and perhaps even more.

If there is a linear relationship between U and h , then Γ^* is equal to the common drainage resistance Γ multiplied by a coefficient α slightly smaller than 1. By using the definition of the common drainage resistance:

$$\Gamma = \frac{h_m - h_0}{U} \quad (12)$$

where:

- h_m = highest level of the phreatic surface (mid-way) between the drains
 h_0 = water level in open drains or tube drains

and remembering that $\tilde{h}$ is an average value irrespective of the distance to the drains, it can be shown immediately that:

$$\Upsilon = \frac{\tilde{h} - \tilde{h}_0}{\alpha \tilde{U}(\tilde{h})} = \frac{\Upsilon^*}{\alpha} \quad (13)$$

A linear relation $U(h)$ is not generally valid, especially not when different drain classes are present (primary, secondary drains, etc., see section 2.3. and E r n s t, 1956). This can also be shown by using the information contained in fig. 2. The four situations shown in that figure were used to obtain graphical relations between the hydraulic head h (viz. $z_{gs}-h$) and some other variable parameters like drain spacing, runoff, drainage resistance, etc. (fig. 19).

The Υ -values were obtained by using the following formula (E r n s t, 1956; 1962):

$$\Upsilon = \frac{L^2}{8kD} + L \Omega_{rad} \quad (14)$$

where:

- L = spacing between parallel drains
 Ω_{rad} = radial flow resistance

For a determination of the radial flow resistances not much information was available. Because the top layers are of a much smaller permeability than the deeper layers (according measurements in some disturbed samples $k_1 \approx 1.5$ m/day and $k_2 \approx 30$ m/day), the following logarithmic formula can be used (E r n s t, 1956; 1962):

$$\Omega_{rad} = \frac{1}{\pi k_1} \ln \frac{4D_1}{B_{wp}} \quad (15)$$

where:

- D_1 = thickness of the top layer with permeability k
 B_{wp} = wet perimeter

For the primary drains with $B_{wp} = 10$ m it was found with eq. (15) that $\Omega_{rad} = 0.3$ day/m. For the smallest drains with $B_{wp} = 0.6$ m it was found like-

wise that $\Omega_{rad} = 0.9$ day/m¹). Substitution of these values in eq. (14) with $kD = 3000$ m²/day and the proper values of L yielded Υ -values as shown in fig. 19. The corresponding U -values were found by means of eq. (12). Finally the Υ^* -values were obtained by determining the gradients in the $\tilde{U}(\tilde{h})$ -graph (fig. 19).

For the whole of East Gelderland there was too little information available for a similar calculation of Υ^* . It seemed worthwhile, however, to use eq. (13) in order to obtain a map giving the common drainage resistance of the area (fig. 20).

A recent analysis (B l o e m e n, 1967) of the fluctuation of the phreatic surface in the region over the years 1954 through 1957 (R e u t e r and K o u w e, 1958), gave sufficient information to find the corresponding differential resistance values for 21 observation wells. For high groundwater tables this resulted in Υ^* -values between 110 and 1400 days and for low groundwater tables in values between 400 and 4000 days. The large regional variations in these results agree with the influence of drain spacing as shown in fig. 19. Calculation of mean Υ^* - and Υ -values over these 21 spots resulted in $\Upsilon^*/\Upsilon \approx 1.4$.

An approximately linear relation $U(h)$ as used for the derivation of eq. (11), implies moreover the validity of the following simple expression:

$$V = Q_0 \Upsilon^* \quad (16)$$

where:

$$V = \text{volume of the depression cone} = - \int_0^\infty 2\pi r h'' dr$$

Eq. (16) was applied to the depression cone of the pumping station "Olden Eibergen" (fig. 16). Since $V = 8 \times 10^6$ m³ and $Q_0 = 4700$ m³/day, it follows that $\Upsilon^* = 1700$ days. This result is quite satisfactory in comparison with the Υ -value of 1330 days which has been found for average conditions at the same place as indicated in fig. 20b west of Eibergen.

It should be remembered that a lowering of the mean depth of the groundwater table involves in

¹) During the writing of this paper the results of a small number of field measurements became available, showing Ω -values about half as large as those used for fig. 19. Although this will give much larger values for the runoff U at high water tables, the Υ^* -values to be used in eq. (11) will be influenced by such errors only to a small degree.

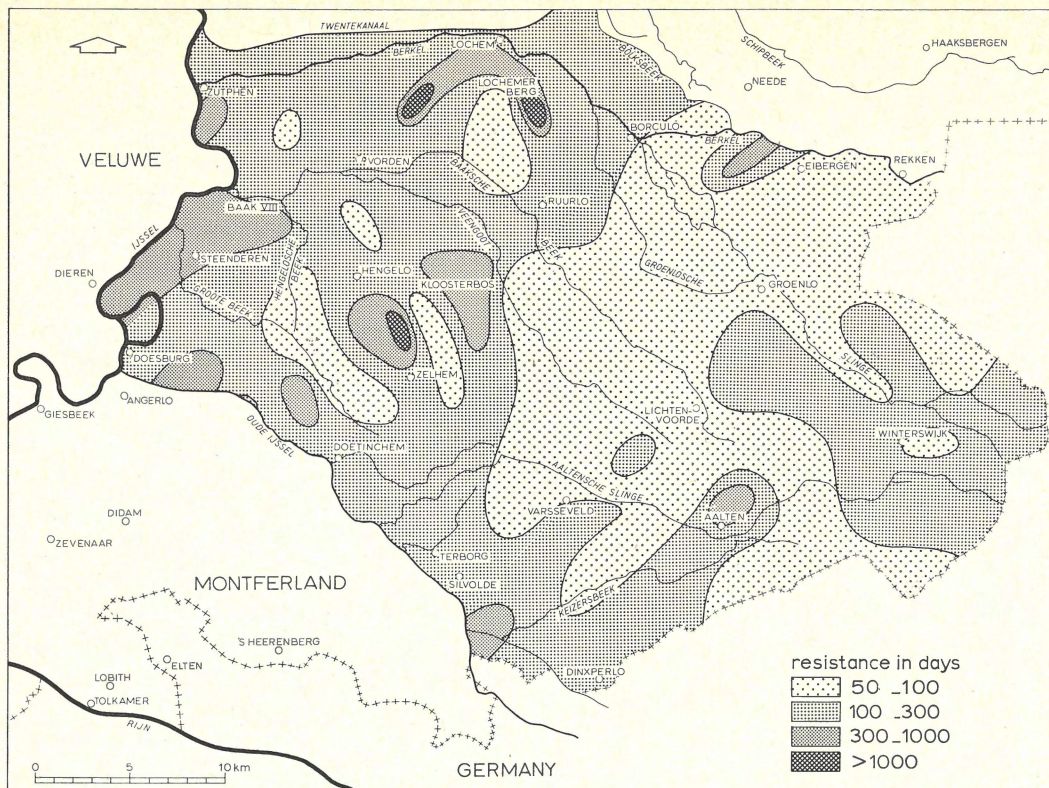


Fig. 20a
Drainage resistances in East Gelderland for a high water table (winter situation with drainage by all classes of drains; average value of Ω_{rad} in western, middle and eastern part respectively 1.5; 0.75 and 0.3 day/m.

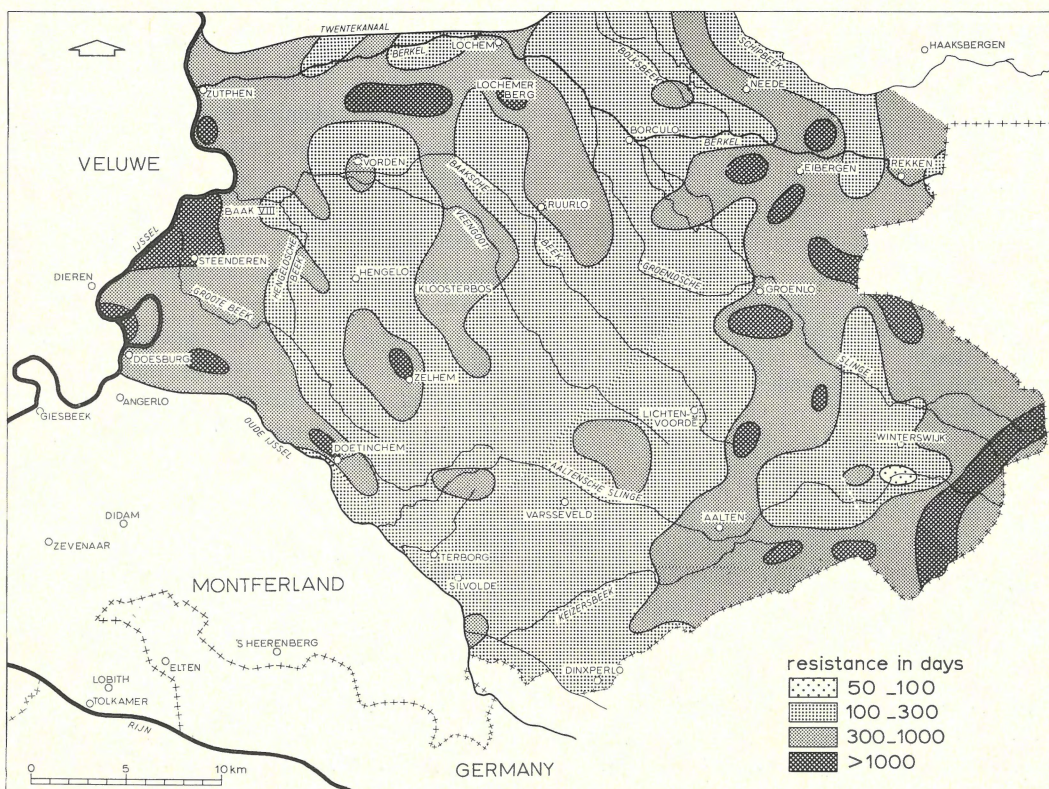


Fig. 20b
Drainage resistances in East Gelderland for an intermediate water table (no drainage by tertiary drains; average value of Ω_{rad} in western, middle and eastern part respectively 1.0; 0.5 and 0.2 day/m.

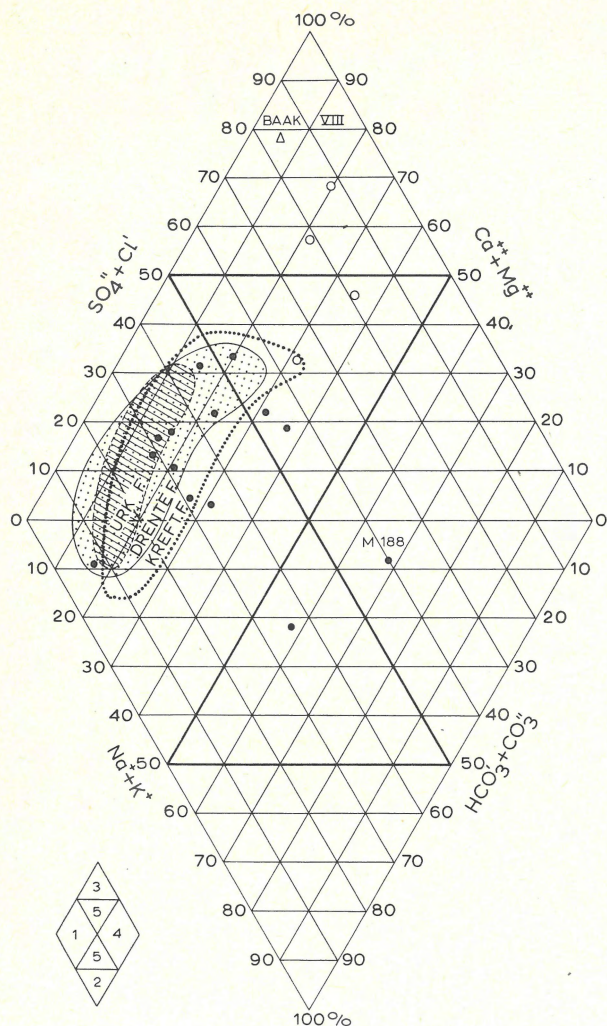


Fig. 21

Classification of 147 groundwater samples present in the various geological formations, according to the Piper diagram. Kreftenheye formation, 29 samples; and Δ Drente formation, 90 samples; --- Urk formation; 10 samples; $\bullet$ Pre-Saalian formations, 14 samples; $\circ$ Sterksel-Enschede formation, 4 samples.

general a decrease of the evapotranspiration of the crops by a reduction in available water in the topsoil and a reduction in capillary rise from the deeper layers. The discharge of water by deep wells is withdrawn for about 80 to 90 per cent from the runoff $\bar{U}$ and for 10 to 20 per cent from the evapotranspiration $\bar{E}$ (Ernst, 1967b), both $\bar{U}$ and $\bar{E}$ to be considered as averages over time and area ($\bar{U} \approx 250$ mm/year; $\bar{E} \approx 500$ mm/year). This lessen-

ing of evaporation was neglected in the preceding calculations. As a consequence slightly overestimated values were obtained of the water table decline. This also means that the best value for a resistance in eq. (11) must be somewhat closer to Υ than suggested by the differences between Υ and Υ^* as shown in fig. 19.

9. CHEMICAL CHARACTERISTICS OF THE GROUNDWATER

9.1. Groundwater classification

For a chemical classification of the groundwater in the investigated region use was made of the so-called Piper diagram (Piper, 1944). In this diagram the overall chemical character of the water, as represented by the relationship between the $\text{Na} + \text{K}$, $\text{Ca} + \text{Mg}$, $\text{CO}_3 + \text{HCO}_3$, and $\text{Cl} + \text{SO}_4$ ions, is indicated by a single point (fig. 21).

In this figure nearly all 147 samples plot in field 1 of this diagram. It can be concluded that the water confined within these formations belongs to the same type, namely the calcium bicarbonate type. Of the 119 samples withdrawn from the Drente formation and the Kreftenheye formation only 5 plotted in field 5 because the sulphate plus chloride percentage exceeds 50 per cent. This relatively high percentage could perhaps be explained by the fact that these samples were collected in the vicinity of the East Netherlands plateau, which is made up of marine sediments. The two samples from the Sterksel-Enschede formation in field 3 of the diagram is probably also due to a marine influence because the formation is thin and located directly on top of marine Tertiary formations.

In general, the water in the ice-pushed hills (Pre-Saalian deposits) belong also to the calcium-bicarbonate type. However, it is not clear why some of the samples from these deposits plot in the fields 4 and 5. It is also not fully understood why the sample withdrawn from the Drente formation in the well Baak VIII plots in field 3 of the diagram, unless the influence of the marine Icenian is assumed, which underlies the Drente formation at this site.

9.2. Electrical conductivity

The electrical conductivity was determined of about 200 groundwater samples, chiefly withdrawn from the Drente formation, the Kreftenheye forma-

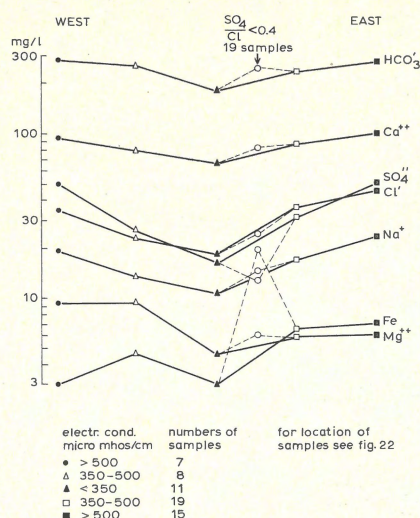


Fig. 23

Chemical West-East section of the groundwater, showing the mean chemical composition of the groundwater samples from the sites given in fig. 22. For the o-points see text on sulphate-chloride ratio.

It appears that the hardness of the groundwater in the ice-pushed hills (Lochemerberg, Montferland and Veluwe) and in the relatively high-lying middle part of the basin is low (less than 10°D). Hard groundwater (more than 15°D) is found in the areas between Ruurlo and Lichtenvoorde, Aalten and Varsseveld, and in a narrow zone bordering the river IJssel and extending eastward to Vorden. In general areas of low hardness correspond with areas of negative net subsurface inflow (see fig. 15) and areas of hard water with areas of positive net subsurface inflow.

9.5. Sulphate-chloride ratio

As can be seen from fig. 23 sulphate and chloride content varies from about 20 mg/l to about 50 mg/l. In general, the sulphate content in the eastern and the western part of the basin is higher than the chloride content, whereas in the middle the sulphate content is lower than the chloride content.

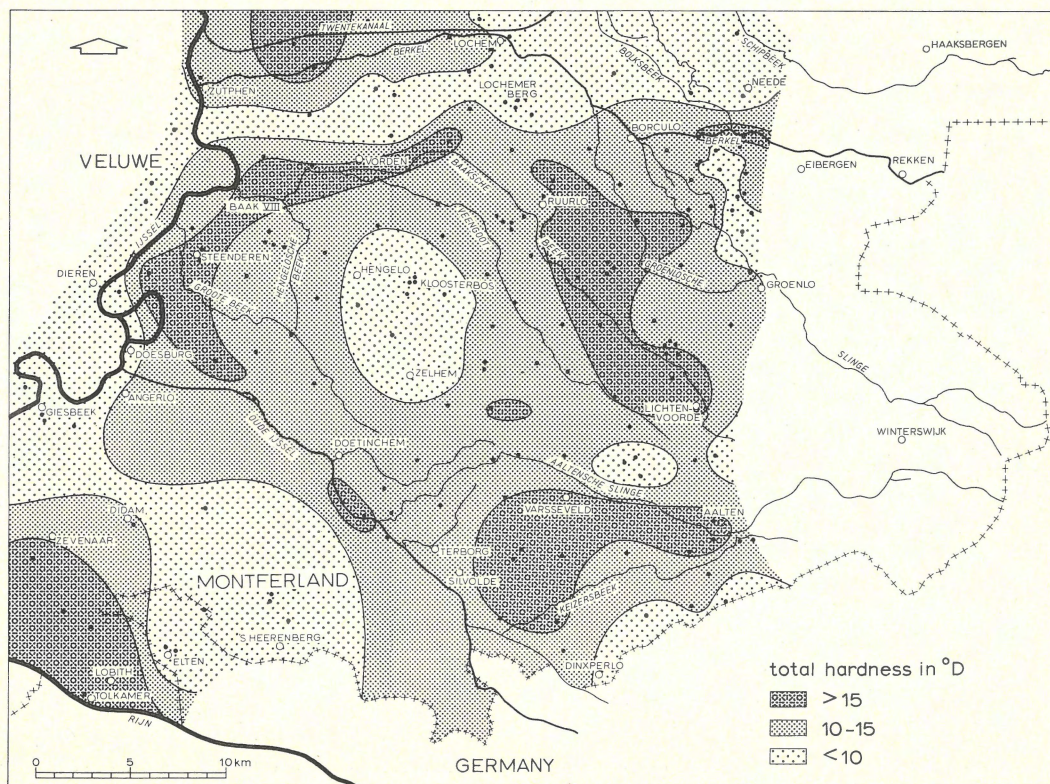


Fig. 24

Hardness (expressed in German degrees) of the groundwater mainly within the Drente and Kreftenheye formations and in the ice-pushed hills of the Lochemerberg and Montferland.

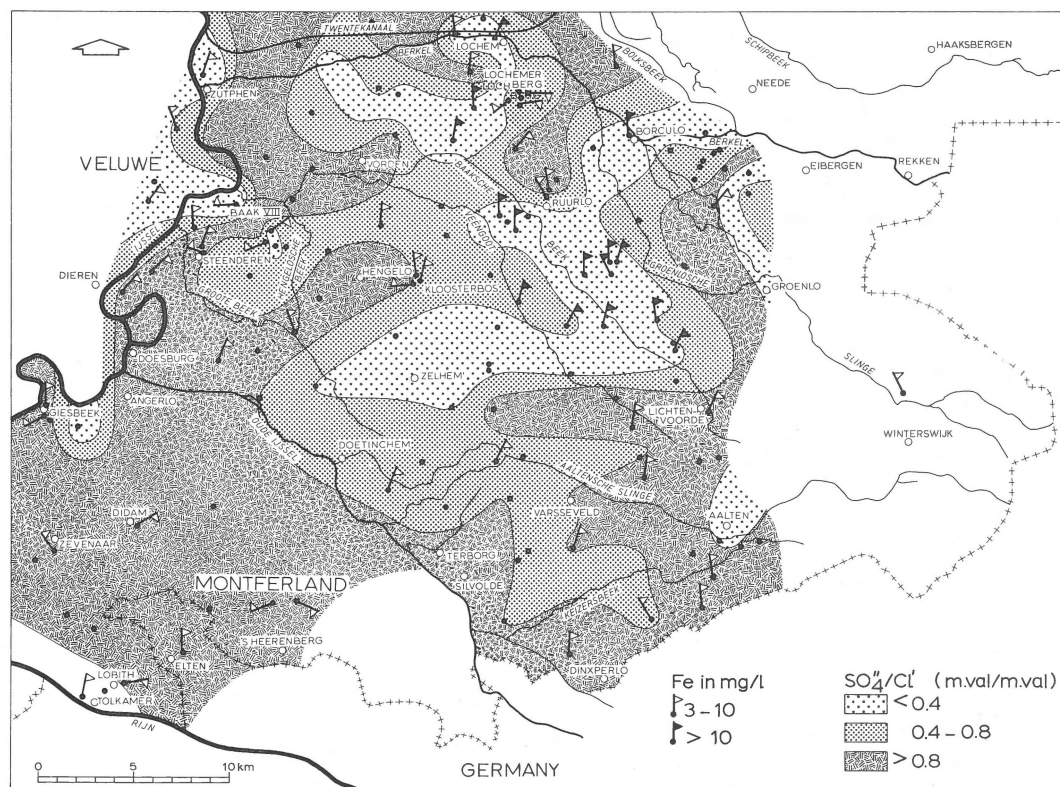


Fig. 25

Sulphate-chloride ratio, and contents of the groundwater mainly within the Drente and Kreftenheye formations and in the ice-pushed hills of the Lochemerberg and Montferland.

Of further interest is the areal distribution of the ratio of sulphate and chloride contents in mval/l of the groundwater within the most important water-bearing formations (fig. 25). This map is based on the results of about 140 analyses. Low sulphate-chloride ratios of less than 0.4 are found in the area between Borculo, Lichtenvoorde and Zelhem, and in the area south and southwest of Lochem. High ratios of more than 0.8 are found in front of the terrace escarpment and in the western part of the basin. In order to show whether the low sulphate-chloride ratio in the Borculo-Zelhem area is caused by either a low sulphate or a high chloride concentration, the mean chemical composition of the groundwater from this area is indicated with o-points in the chemical section given in fig. 23. It appears that this water mainly deviates from the groundwater in the neighbourhood by its relatively low sulphate concentration of about 12 mg/l (and high iron content).

Since a decrease in sulphate content often results

from sulphate reduction it could be expected that the sulphate content decreased with depth. For this reason the sulphate and chloride concentrations were examined in relation to depth below ground surface. The available analyses were divided according two depth classes respectively 5 to 15 m and deeper than

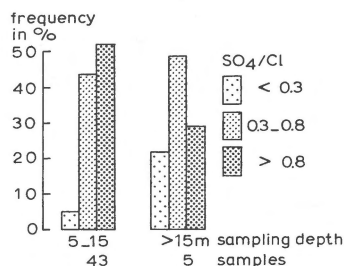


Fig. 26

Frequency distribution of the sulphate-chloride ratio classes according groundwater sampling depths.

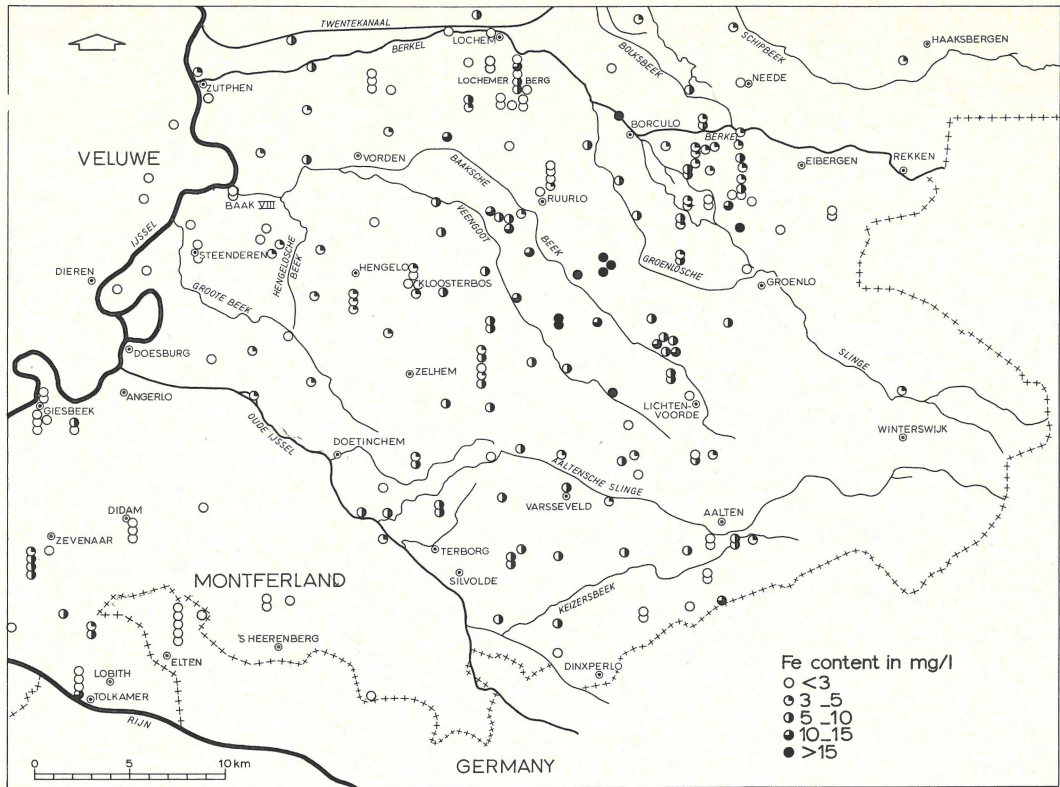


Fig. 27

Iron content of the groundwater including that in the preglacial deposits of the ice-pushed hills of the Lochemerberg and Montferland.

15 m below ground surface. Fig. 26 shows the frequency distribution of the sulphate-chloride ratios in these depth classes. The deep groundwater has a fairly normal frequency distribution, whereas the groundwater in the higher layers has a top in the highest sulphate-chloride ratio class.

Reduction of sulphate to sulphide takes only place when the redox potential of the groundwater is sufficiently low. Such low redox potentials can be expected in those areas where the water-bearing layers are rich in organic material and the water is poor in oxygen. Since sulphate reduction seems to be limited to certain areas, especially the area between Borculo, Lichtenvoorde and Zelhem (see fig. 25), the explanation of this phenomenon can probably be sought in the presence of the Eem Formation. As was mentioned earlier, the Eem Formation contains in general much organic matter and even layers of peat. Its presence within the water-bearing formations of the Borculo-Lichtenvoorde area was demonstrated in

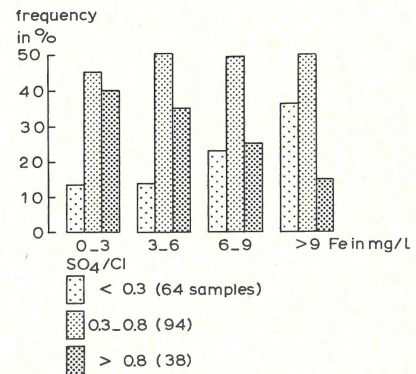


Fig. 28

Frequency distribution of the sulphate-chloride ratio classes according the iron content classes. Water with a low sulphate-chloride ratio tends to have a high iron content and vice versa.

several borings. The low sulphate content of the groundwater in this area is probably due to these humic Eem deposits.

9.6. Iron content

To study the iron content of the groundwater in the region, use could be made of the results of some 240 water analyses. The iron content of the water varies from 0 to 20 mg/l. The areal distribution of the iron content is presented in fig. 27. This map shows that water with iron contents of 10 mg/l and higher is chiefly found in the area between the villages of Borculo and Lichtenvoorde, whereas the water with an iron content of less than 3 mg/l occurs in the ice-pushed hills, in a zone bordering the river IJssel and at several places in the east and south of the region.

A comparison of this distribution pattern with the areal distribution of the sulphate-chloride ratio as given in fig. 25, leads to the conclusion that high iron contents tend to go with low sulphate contents, and low iron contents with high sulphate contents (De Vries, 1969).

Of 196 water samples analyses of both sulphate, chloride and iron concentrations were available. The frequency distribution of the three sulphate-chloride ratio classes over the iron content classes is shown in fig. 28.

It can be seen from this figure that the lowest sulphate-chloride ratio class shows a frequency top at the highest iron class, whereas the highest sulphate-chloride ratio class shows a top at the lowest iron class. Medium high sulphate-chloride ratios occur with nearly the same frequency in all iron classes. It may be concluded that a high iron content is related to a low sulphate content, and vice versa. This may be explained by a chemical reduction which reduces the sulphate to sulphide and the almost insoluble ferri-iron to the soluble ferro-iron.

In the foregoing section it was shown that the sulphate-chloride ratio is generally decreasing with depth. If chemical reduction is the true cause of this phenomenon it can be expected that the iron content should increase with depth. Table 2 shows the sulphate-chloride ratio and the iron content of the samples collected from shallow and deep filters installed in the same borings.

TABLE 2

Sulphate-chloride ratio and iron content of the groundwater at different depths (de Vries, 1969)

Boring	Filterdepth in m	SO ₄ /Cl (mval/mval)	Fe(mg/l)
M 61	11.29–12.29	1.42	5.7
	28.75–29.75	0.51	2.2
M 68	8.86–9.86	0.30	4.1
	36.06–37.06	0.33	7.2
M 70	8.88–9.88	0.49	6.0
	26.69–27.69	0.55	5.2
M 71	10.46–11.46	0.49	3.5
	41.34–42.34	0.91	3.7
M 72	11.94–12.94	1.14	5.4
	28.56–29.56	1.56	3.7
M 73	9.29–10.29	1.33	4.0
	23.06–24.06	0.82	6.2
M 74	6.30–10.30	0.43	4.1
	24.20–32.20	0.00	3.8
M 76	13.73–14.73	1.32	7.2
	25.03–26.03	0.27	3.6
M 139	16.92–17.92	0.30	6.4
	28.90–31.50	0.27	7.6
M 140	20.94–21.94	0.46	16.5
	37.00–38.00	0.13	18.0
M 150	12.70–13.70	1.20	2.7
	29.00–30.00	1.19	6.2
M 153	14.00–15.00	0.89	4.5
	41.03–42.03	0.73	5.8
M 154	20.92–21.92	0.88	7.8
	38.00–39.00	0.67	8.8
M 158	15.87–16.87	1.19	6.5
	35.70–36.70	0.57	7.1
M 166	14.00–15.00	1.55	5.9
	28.00–29.00	0.67	6.3
Median	shallow filters	0.98	5.1
	deep filters	0.61	5.6

10. VERTICAL TEMPERATURE COURSE

It can be expected that the temperature of the groundwater in areas with negative net subsurface inflow will be different from that in areas with positive net subsurface inflow. In order to check whether the observed differences in groundwater circulation are reflected in the vertical temperature course, the temperature of the groundwater was measured at various depths in a number of wells

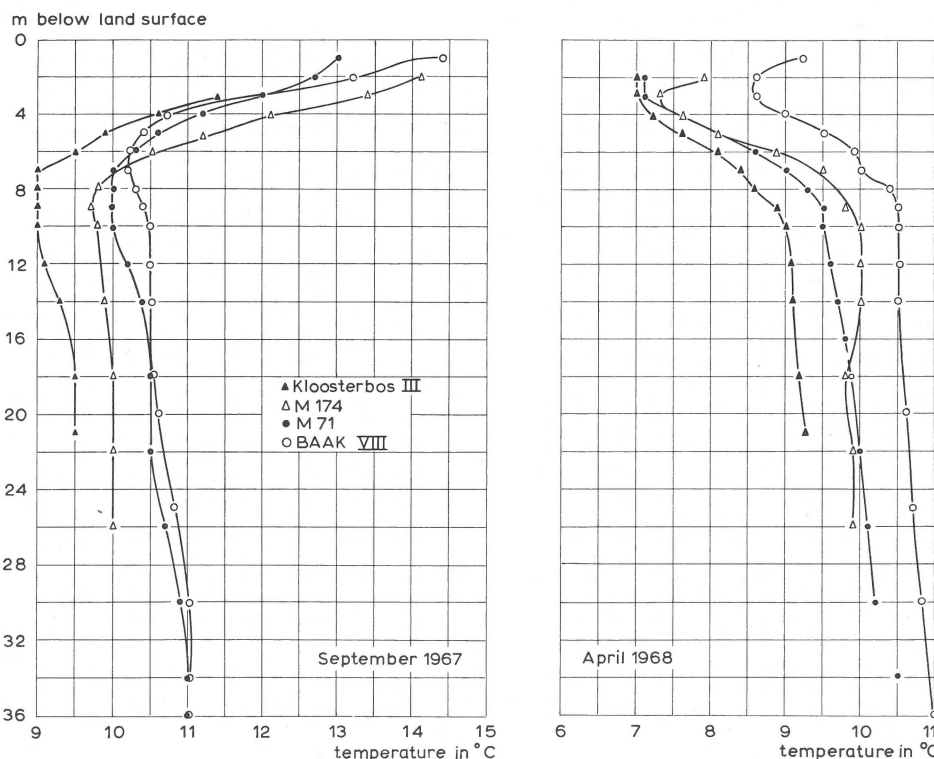


Fig. 29

Temperature of the groundwater at various depths in the wells Kloosterbos III, Baak VIII, M 71 and M 174.

located in the eastern, middle and western part of the basin. To this aim a thermocouple, by which measurements could be made with an accuracy of 0.1°C , was lowered into the wells.

The results of the measurements in 4 wells, taken in September 1967 and April 1968, are presented in fig. 29. The wells Kloosterbos III and M 174 are located in an area of negative net subsurface inflow to the north and southeast of the village of Zelhem. The wells Baak VIII and M 71 are located in areas of positive net subsurface inflow respectively between the river IJssel and the village of Hengelo and between the villages of Groenlo and Borculo.

The temperature-depth diagrams show that, at depths where the seasonal fluctuations are negligible, the groundwater in the areas of positive net subsurface inflow is 1° to 2°C warmer than that in the areas of negative net subsurface inflow.

In the area where the discharge balances the excess precipitation, the temperature of the groundwater at a depth of 30 m below ground surface is 10°C . This is in agreement with the average annual temperature of

9°C of the superficial groundwater and a geo-thermal depth gradient of 1°C .

11. REMARKS ON FUTURE GROUNDWATER EXTRACTION

Geological investigations have shown, the presence of a continuous 10 to 40 m thick water table aquifer, chiefly made up of coarse, fluvioglacial and post glacial deposits of the river Rhine. This aquifer is wedging out against the much less permeable Tertiary formations which occur near the surface in the most eastern part of the region.

For locating future pumping plants the map of aquifer transmissivities may be used. It may seem obvious that the areas of high transmissivity (order of magnitude 3000 to 10,000 m^2/day) are the most appropriate for groundwater extraction. However, from an agricultural point of view those areas are often not very suitable for that purpose. This because of the rather large seasonal fluctuations of the

groundwater table involving that during dry summers the groundwater table often drops to undesired low levels.

By applying a method of finite differences a detailed analysis of the groundwater movement within the aquifer was made. In this way areas of positive and negative net subsurface inflow could be delineated. From an agricultural point of view the areas of positive net subsurface inflow can be considered as better suited for future groundwater extraction. Here the water tables are shallow, do not change much during the course of the year and most of the excess water is flowing off to the main rivers by a dense network of drains. However, any groundwater extraction in areas of positive net subsurface will also cause the water table to decline although the extent of the depression cone may be restricted by the numerous open water courses. The small pumping station near Eibergen located in an area of positive net subsurface inflow provides evidence for this statement. A quantitative analysis of the agricultural repercussions of a decline of the groundwater table is recommended before any decision is taken on future pumping sites.

For the calculation of the possible drawdown caused by a pumped well a formula has been derived which gives solutions of sufficient accuracy. The presence of secondary and tertiary drainage channels in the area can be taken into account by means of the differential drainage resistance. For this purpose the given map of common drainage resistance values, being of the same order as the differential drainage resistances, can be used.

The results of the geochemical studies demonstrate that, generally speaking, the groundwaters within areas of positive net subsurface inflow are, because of their longer flow path, more mineralized than those in areas of negative net subsurface inflow. They have a relatively high electrical conductivity (500 to 600 micromhos/cm), a high total hardness (exceeding 15°D), and a high sulphate-chloride ratio (exceeding 0.8), rendering them less suitable for domestic use. The best groundwater qualities are generally found in areas of negative net subsurface inflow.

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