GEOLOGIE EN MIJNBOUW© VOLUME 58 (3), p. 295-304 1979

THE GROUNDWATER THERMAL REGIME IN THE FLEVO POLDERS AND GELDERSE VALLEI (SOUTHERN IJSSELMEER AREA, THE NETHERLANDS)[1]

S. J. DE JONG[2] & W. GEIRNAERT[3]

ABSTRACT

De Jong, S. J. & W. Geirnaert 1979 The groundwater thermal regime in the Flevo polders and Gelderse Vallei (southern IJsselmeer area, The Netherlands) – Geol. Mijnbouw 58: 295-304.

The relationship between regional groundwater-flow systems and groundwater temperatures has been studied. Temperature/depth profiles were recorded in 70 observation wells. Low temperature gradients were observed in the infiltration areas. In these areas the observed temperature/depth profiles and those calculated with the theory of one-dimensional steady-state heat flow, showed significant correlation.

Temperature/depth profiles in discharge areas showed lower shallow vertical temperature gradients than expected. This could be explained with a simple heat-budget equation of the regional groundwater-flow systems. In the area studied recharge of groundwater occurs at a lower temperature than discharge. Part of the geothermal conductive heat flow is used for warming up the relatively cold recharge water. This results in relatively low shallow temperature gradients in discharge areas.

INTRODUCTION

Hydrogeologists have recently shown growing interest in the temperature distribution in groundwater. This interest is generally focussed on subjects such as:
(1) the temperature changes associated with natural or artificial recharge of groundwater (VERRUYT, 1966; HUISMAN, 1973; NIGHTINGALE, 1975);
(2) the temperature changes associated with injection or return of cooling water (THIRRIOT & GANDU, 1975);
(3) the determination of the specific discharge of groundwater by means of temperature methods (STALLMAN, 1965; BREDEHOEFT & PAPADOPOULOS, 1965; SOREY, 1971).

So far not much attention has been paid to the study of the natural thermal regime of groundwater. A study of the groundwater thermal regime of a glacial complex in Canada was carried out by PARSONS (1969). The hypothetical and other models based on his field observations form the starting point of the present study. Parsons' study is based on the assumption that groundwater flow and heat transfer occur under steady-state conditions. According to Parsons the general effect of groundwater flow on the temperature at any given point in a groundwater basin depends on:
(1) the groundwater flux vector;
(2) the disposition of temperatures at the water table relative to areas of recharge and discharge;
(3) the disposition of groundwater-flow systems within the basin;
(4) the depth of the groundwater basin with respect to the superficial and geothermal zones;
(5) the mechanism of groundwater discharge, that is, whether groundwater is discharged directly at the surface by seepage and springs or indirectly by evapotranspiration.

The purpose of the study outlined in this paper is to investigate the groundwater thermal regime of a regional groundwater-flow system in The Netherlands. The southern part of the IJsselmeer area was selected. This area is hydrogeologically well-known because of detailed investigations of the Geological Survey of The Netherlands, the Government Institute for Water Supply and the Service of

[1]Manuscript received: 1979-04-11.
Revised manuscript received and accepted: 1979-07-09.
[2]International Water Supply Consultants IWACO B.V., P.O. Box 183, ROTTERDAM, The Netherlands.
[3]Free Reformed University, Institute of Earth Sciences, P.O. Box 7161, AMSTERDAM, The Netherlands.

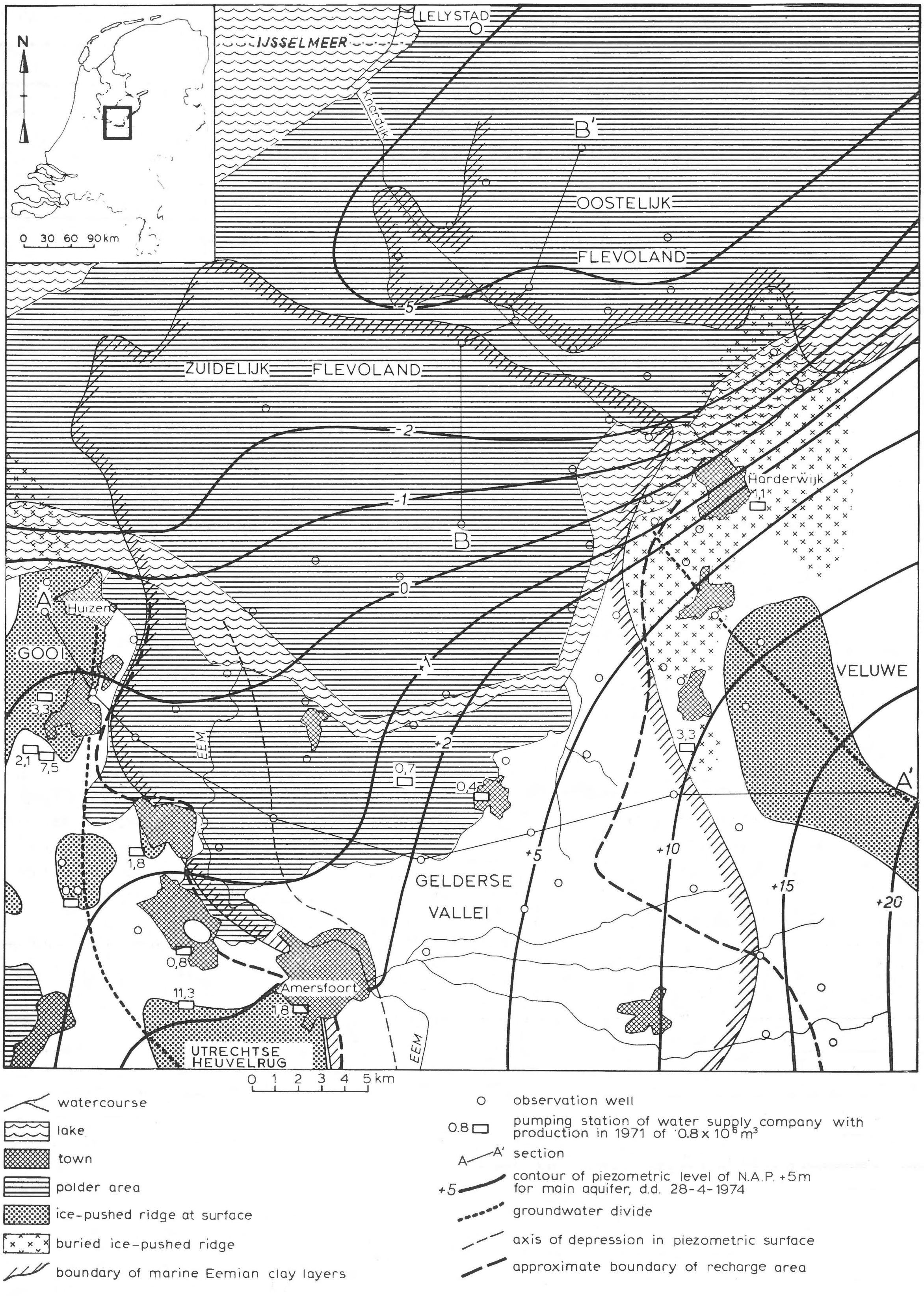

Fig. 1
Hydrogeological map of the area of study.

the IJsselmeer polders of Rijkswaterstaat (Public Works). In The Netherlands the measurement of temperature profiles in observation wells was initiated by Csonka (CSONKA, 1968, 1972). Our study is the first on the regional distribution of the temperature of groundwater and will be followed by more detailed studies like the determination of thermal characteristics of geological formations and the calculation of groundwater velocity with thermometry. The latter study is in preparation.

The field work for this investigation was carried out in 1973 and 1974. Some additional measurements were made in 1975. In total 70 temperature/depth profiles were recorded. Temperature measurements were made at intervals of 1 or 2 metres. The measurements were made in the water-filled parts of the observation wells which were provided with small-diameter PVC tubes.

Convective flow in the observation wells was occasionally observed in the upper part. This phenomenon was studied by SAMMEL (1968). Based on the results of Sammel and the experience gained in this study, it is concluded that in the area of study the thermal effects of convective heat flow in the observation wells are limited to a maximum depth of 5 to 10 m below the groundsurface. It is assumed that at depths exceeding 10 m the temperature of the water in the observation well is equal to the temperature of the surrounding ground.

For the temperature measurements several assemblies, each consisting of a thermistor connected to a Wheatstone bridge, were used. The sensitivity of the equipment was 0.01 °C. Temperature differences in a temperature/depth profile are thought to be accurate to 0.02 °C. The absolute temperature of the initial equipment may deviate ± 0.1 °C, whereas the accuracy of the equipment finally used was 0.02 °C.

HYDROGEOLOGY OF THE SOUTHERN IJSSELMEER AREA

The area of study comprises the northern part of the Gelderse Vallei, the high areas surrounding the valley, and parts of the recently reclaimed polders Zuidelijk and Oostelijk Flevoland in the IJsselmeer area (Fig. 1) (a 'polder' is an area having an artificially maintained drainage and surface water regime; in The Netherlands its elevation is generally situated near or below sea level). The reclamation of the polders of Oostelijk and Zuidelijk Flevoland was completed, respectively, in 1957 and 1968.

The hydrogeological base in the area is considered to consist of the fine-grained marine early Pleistocene deposits (Maassluis Formation, see Fig. 2). The top of this formation is found at a depth varying from 150 m in the South to 300 m below NAP (Dutch datum level, approximately corresponding with mean sea-level) in the North. Up to the Saalian glaciation a thick sequence of mainly coarse-grained fluvial sediments was deposited (predominantly Harderwijk and

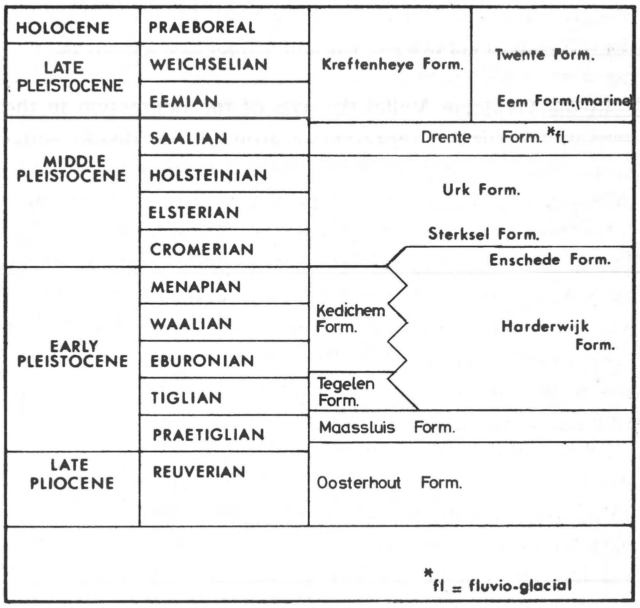

Fig. 2
Correlation chart of the Pleistocene in the central part of The Netherlands (schematically after Doppert et al, 1975).

Enschede Formations). During the Saalian glaciation an important change in sedimentation as well as in morphology took place in the area. An ice sheet invaded the area from the North. From the front of the ice sheet glacier tongues developed, scouring deep basins. Such an elongated basin was formed under the present Gelderse Vallei and Oostelijk Flevoland. The eroded material was pushed up into ridges surrounding these basins, such as the Veluwe, the Gooi hills and the Utrechtse Heuvelrug.

Since the retreat of the ice, various layers of coarse- and fine-grained sediments have been deposited in the glacial basin and the low-lying areas. Important fine-grained deposits are the glacio-lacustrine Saalian clays belonging to the Drente Formation, the marine Eemian clays and the Holocene clays. The Saalian clays are only found in the centre of the glacial basin. The thickness of the latter deposits can be more than 40 m. In the Gelderse Vallei the Eemian clays form an extensive confining layer (see Fig. 1); the Holocene clays occur mainly in and around the Flevo polders.

The thick coarse-grained preglacial and glacial deposits form an important aquifer. On a regional scale this aquifer can be subdivided in a phreatic part in the higher areas surrounding the valley, and a semi-confined part in the lower areas. Confining layers consist here of the fine-grained sediments deposited after retreat of the ice, whereas the main aquifer consists of the preglacial deposits (see also Fig. 5).

Contours of the piezometric level of the main aquifer are presented in figure 1. The regional groundwater-flow systems are determined by recharge in the Veluwe and Gooi hills and by discharge in the Gelderse Vallei and Flevo polders. For a part of the Veluwe an average annual precipitation surplus of 360 mm, or 1.0 mm/day was determined (R.I.D., 1970). In the recharge areas generally no surface run-off takes place, as the infiltration capacity of the soil is high and a thick unsaturated zone is present.

In the Gelderse Vallei the axis of the depression in the piezometric surface represents a groundwater divide, separating the regional groundwater-flow systems associated with the Gooi hills and the more important Veluwe recharge area. Near this axis a ridge of brackish groundwater was found by geoelectrical investigations (GISCHLER, 1964). In the centre of the valley thick clay layers are present and consequently upward flow of groundwater is slow. Discharge takes place mainly along the margins of the recharge areas where a number of small streams originate. Part of the groundwater in the valley is discharged by farmers' artesian wells.

The confining layers in the area of the two Flevo polders consist mainly of Eemian and Holocene clays. In an elongated zone near the boundary of the two polders these confining layers are absent (Fig. 1).

Especially during the last decades the regional groundwater-flow systems were strongly affected by human interference. The reclamation of polders and the withdrawal of groundwater by industries and water-supply companies have changed the hydrogeological regime. This resulted among other things in a decrease of the groundwater discharge into the watercourses along the border of the recharge areas and in a significant lowering of the piezometric level.

THEORY OF THE SIMULTANEOUS FLOW OF CONVECTIVE AND CONDUCTIVE HEAT

In a homogeneous and isotropic water-bearing layer in which no heat is generated or absorbed, the equation of continuity of heat can be described by:

$$\frac{\delta \vec{q}}{\delta s} = -\varrho_g C_g \frac{\delta T}{\delta t} \qquad (1)$$

where $\frac{\delta \vec{q}}{\delta s}$ denotes the gradient of the heat flux, ϱ_g and C_g respectively the density and the specific heat of the saturated ground, t the time and T the temperature. In saturated ground, heat is transferred by conduction and convection; heat transfer by radiation may be neglected. The heat flow by conduction obeys the law of Fourier and may be expressed as:

$$\vec{q_c} = -\lambda \frac{\delta T}{\delta s} \qquad (2)$$

where $\vec{q_c}$ denotes the conductive heat flux vector, λ the thermal conductivity of the saturated ground and $\frac{\delta T}{\delta s}$ the temperature gradient. Convective heat transfer occurs when groundwater is in motion. When it is assumed that the groundwater is incompressible and that its density is constant the heat flow by convection can be written as:

$$\vec{q_w} = \varrho_w C_w \, T \, \vec{v} \qquad (3)$$

where $\vec{q_w}$ denotes the convective heat flux vector, ϱ_w and C_w respectively the density and the specific heat of the groundwater and $\vec{v}$ the specific-discharge vector. The latter is defined as the discharge per unit total area. Substitution of v from the Law of Darcy into equation (3) yields:

$$\vec{q_w} = \varrho_w C_w T \, k \frac{\delta Q}{\delta s} \qquad (4)$$

in which k denotes the hydraulic permeability of the ground and $\frac{\delta Q}{\delta s}$ the gradient of the piezometric level. As

$$\vec{q} = \vec{q_c} + \vec{q_w} \qquad (5)$$

substitution into equation (1) yields (STALLMAN, 1963):

$$\delta \frac{\left(-\lambda \frac{\delta T}{\delta s} + C_w \varrho_w k \frac{\delta Q}{\delta s}\right)}{\delta s} = -\varrho_g C_g \frac{\delta T}{\delta t}\,. \qquad (6)$$

Stallman investigated the possibilities of determining permeability by temperature methods and remarked that 'the complexity of equation (6) and the boundary conditions met in field studies preclude an analytical approach, except in those cases within which both heat and groundwater flow occur along one and the same direction'. Up to the present time two analytical studies have been published. The first deals with an analytical solution of steady-state one-dimensional flow of heat and groundwater (BREDEHOEFT & PAPADOPOULOS, 1965) and the second covers an analytical solution of steady one-dimensional fluid flow in a semi-infinite porous medium with sinusoidally varying surface temperature (STALLMAN, 1965). The solution of Bredehoeft and Papadopoulos has been applied to determine the groundwater velocity through semi-permeable layers (SOREY, 1971).

In the present paper the authors make use of some of the above-mentioned studies for the development of an equation.

The differential equation for one-dimensional flow of heat and groundwater can be derived from equation (6) and reads:

$$-\lambda \frac{\delta^2 T}{\delta z^2} + \varrho_w C_w k \frac{\delta T \frac{\delta Q}{\delta z}}{\delta z} = -\varrho_g C_g \frac{\delta T}{\delta t} \qquad (7)$$

For uniform flow of groundwater this equation can be simplified to:

$$-\lambda \frac{\delta^2 T}{\delta z^2} + \varrho_w C_w v \frac{\delta T}{\delta z} = -\varrho_g C_g \frac{\delta T}{\delta z} \qquad (8)$$

Assuming steady-state conditions this equation can be written as:

$$-\frac{\delta^2 T}{\delta z^2} + \frac{\varrho_w C_w v}{\lambda} \frac{\delta T}{\delta z} = 0 \qquad (9)$$

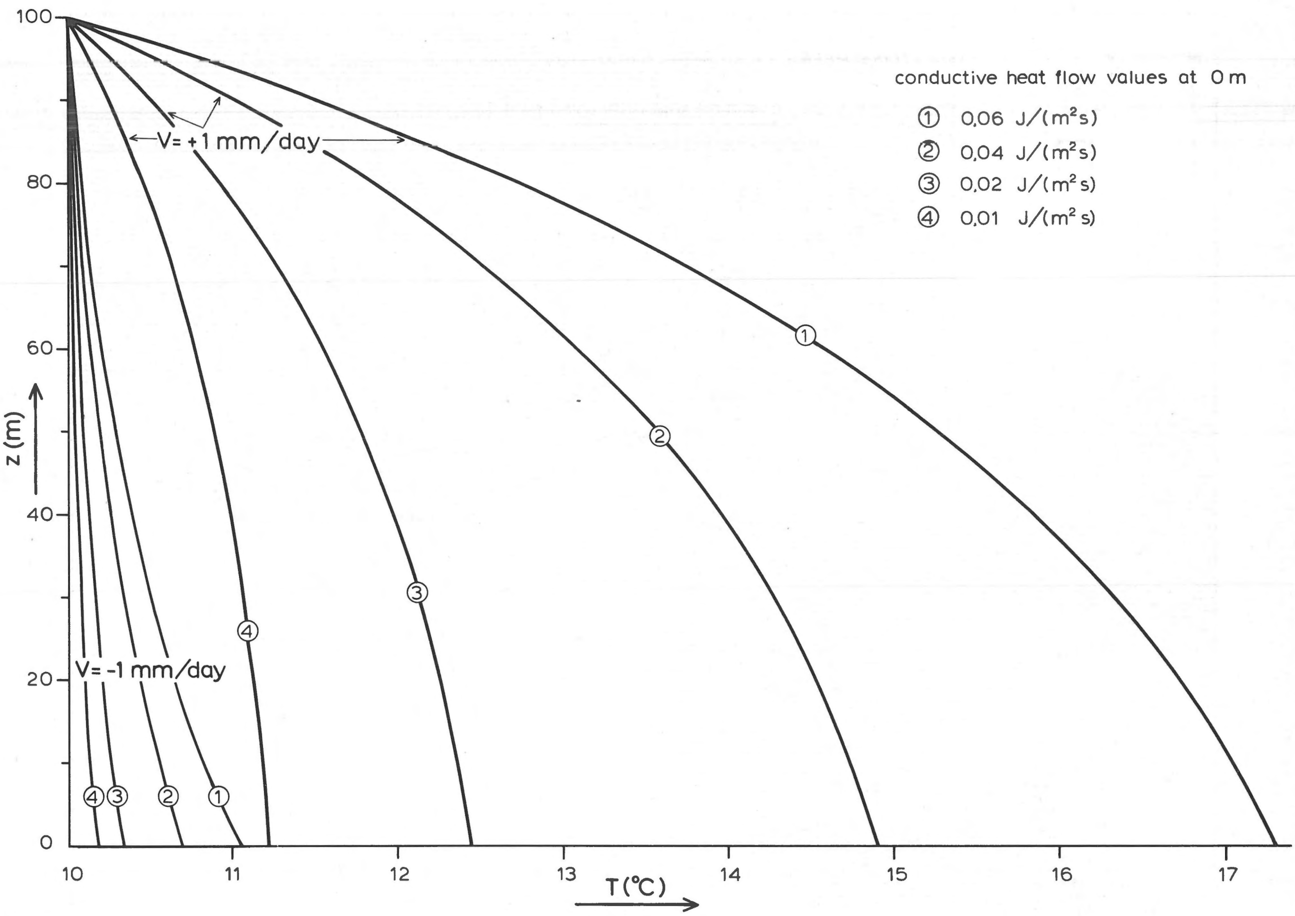

Fig. 3
Calculated temperature profiles for one-dimensional steady-state heat flow.

The general solution of this equation is:

$$T(z) = A + B \exp(\beta z) \tag{10}$$

in which: $\beta = \frac{\varrho_w C_w v}{\lambda}$

and where A and B denote constants. The solution with boundary conditions:

$$z = 0 : \lambda \left.\frac{\delta T}{\delta z}\right|_{z=0} = q^* \tag{11}$$

$$z = l : T = T_l \tag{12}$$

is: $$T(z) = T_l + \frac{q^*}{\beta\lambda} \exp(\beta l) - \frac{q^*}{\beta\lambda} \exp(\beta z) \tag{13}$$

With the aid of equation (13) temperature/depth profiles have been calculated. Values of +1.0 and −1.0 mm/day for v and values for q* equal to or smaller than the geothermal heat flow in the central parts of The Netherlands were taken. According to a map of HAENEL (1974) the value of the latter flow amounts to 0.07 J/(m²s).

Furthermore, the following values were introduced:

T_l = 10.0°C
$\varrho_w C_w$ = 4.2.10⁶ J/(°C.m³)
l = 100.0 m
λ = 2.5 J/(°C.s.m.) (average value of water-saturated quartz sands, see WOODSIDE & MESSMER, (1961).

The calculated temperature profiles are presented in figure 3. It shows that groundwater movement strongly affects the temperature distribution. The profiles are convex upward for positive values of v and convex downward for negative values of v; without groundwater flow heat is only transferred by conduction and temperature is a linear function of depth. The temperature gradients for z = 100 are small for the negative value of v; this type of temperature profile may be expected in those shallow parts of the recharge areas where predominantly steady-state vertical heat and groundwater flows take place. The temperature gradients at z = 100 are high for the positive value of v and depend strongly on the conductive heat flow for z = 0.

Let us now consider the heat budget of a two-dimensional phreatic aquifer with a simple single groundwater-flow sys-

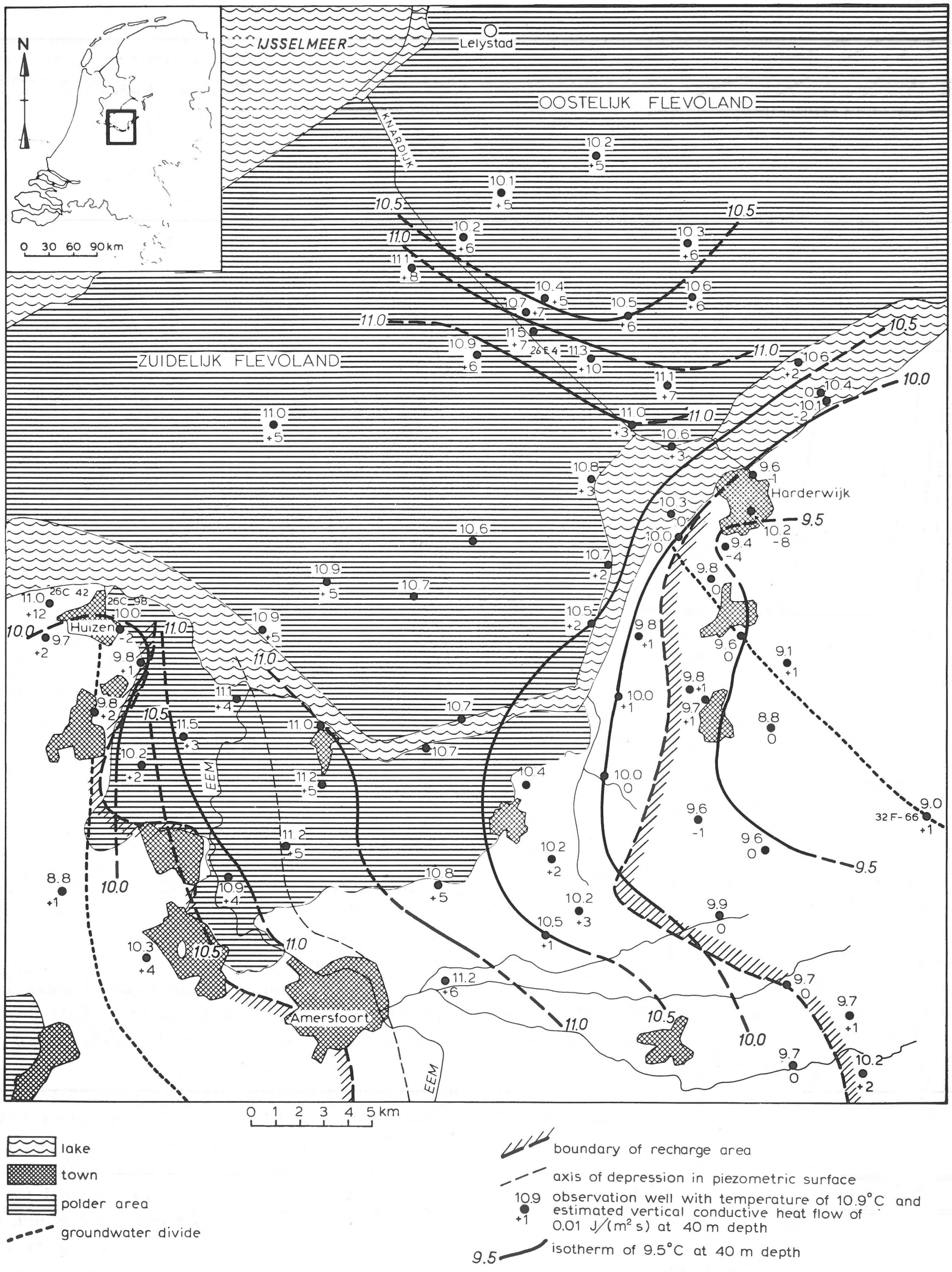

Fig. 4
Map with groundwater temperatures and estimates of vertical conductive heat flow values at 40 m below surface.

tem. It is assumed that steady-state heat and groundwater flow take place and that the aquifer is bounded by impermeable vertical walls and by an impermeable base of length L. The water table is taken as the upper thermal and hydraulic boundary. At the base of the aquifer a uniform vertical conductive heat flow is assumed, while the vertical walls are assumed to be non-conductive. In the heat budget, only the flow of heat through the upper and lower boundaries has to be taken into account.

The heat budget equation can be written as:

$$Q_{c,l} + Q_{c,u} + Q_{w,u} = 0 \tag{14}$$

in which $Q_{c,l}$ denotes the total inflow of conductive heat through the lower boundary and $Q_{c,u}$ and $Q_{w,u}$ the respective inflow of conductive and convective heat at the upper boundary.

In order to determine the influence on the heat balance of variations of the temperature of the water table it will be assumed that both recharge and discharge of 1 mm/day takes place over a distance of ½ L. The temperature of the water-table in the discharge and recharge areas is constant, the temperature in the discharge areas being 1.0°C higher than that in the recharge areas. Elaboration of the values given above yields:

$$Q_{w,u} = -2{,}4.10^{-2}.L \ [J/(m.s)]$$

For the uniform conductive heat flow at the base of the aquifer, the stated geothermal heat flow in the central part of The Netherlands is taken. It follows that:

$$Q_{c,l} = 7{,}0.10^{-2}.L \ [J/(m.s)]$$

Elaboration of equation (14) yields:

$$\frac{Q_{c,u}}{Q_{c,l}} = -0.66$$

It appears that in this example the heat budget is sensitive to variations of the temperature of the water table relative to areas of recharge and discharge; 34% of the geothermal heat is used for the warming-up of the cold recharge water.

THE GROUNDWATER THERMAL REGIME IN THE SOUTHERN IJSSELMEER AREA

Introduction

Since the depth of a large number of observation wells is about 50 m, only the thermal regime of the upper part of the aquifer could be investigated in detail. The seasonal temperature variation appeared to be perceptible down to a depth of approximately 20 m. For a presentation of the aerial distribution of groundwater temperature and of estimated vertical conductive heat flow a depth of 40 m below the surface was chosen; this depth is attained by the majority of the observation wells and lies well below the zone influenced by the seasonal temperature variation. The groundwater temperatures at this depth were read from the temperature/depth profiles, while the estimates of the conductive heat flow were calculated with the law of Fourier (equation (2)); the temperature gradients were determined from the profiles and for the thermal conductivity a theoretical value of water-saturated quartz sands of 2.5 J/(°C.m.s) was taken. If at the 40 m depth no sand was present, the temperature gradient was extrapolated to this depth from the depth where sand prevailed. As the z-direction is taken positive upwards, a negative temperature gradient gives a positive (upward) conductive heat flow value.

Figure 4 shows that in the Gelderse Vallei both the groundwater temperature and the estimated conductive heat flow are related to the two regional groundwater-flow systems. The temperature increases from approximately 9.0 °C in the centres of the recharge areas to 11.0-11.5 °C in a zone along the axis of depression in the piezometric surface. This is also illustrated in the section across the Gelderse Vallei (Fig. 5). The estimated vertical heat flow increases gradually from 0.00-0.01 J/(m^2.s) in the recharge areas to 0.05-0.06 J/(m^2.s) in the valley; here they approach the regional geothermal heat flow of 0.06-0.07 J/(m^2.s). It is interesting to note that the heat-flow values show more variation than the groundwater temperatures.

In the Flevo polders, excluding a zone along the Knardijk, the groundwater temperature and the estimated conductive heat flow are more or less constant; the conductive heat flow equals the geothermal heat flow and the groundwater temperature varies from 10.0 to 11.0 °C. Along the Knardijk groundwater temperatures above 11.0 °C and high heat-flow values of up to 0.10 J/(m^2.s) were found (see Fig. 4). This zone is known to be an important discharge area.

Temperature of the water table

The water table is generally situated in the zone influenced by the seasonal temperature variation. This variation is superimposed on a more or less steady temperature field. The average annual temperature of the groundwater table is an important value and can be estimated by extrapolation of the deeper temperatures which are independent of seasonal influences (see figure 6).

The vegetation influences the temperature of shallow groundwater (PLUHOWSKI & KANTROWITZ, 1963). In the zone under study the recharge areas are usually wooded, while in the discharge areas arable land is generally found. The average annual temperature of the groundwater table in the recharge areas amounts to 9.0-10.0 °C, and in the discharge areas to 9.5-10.5 °C. The lowest temperatures of the groundwater table, viz. 8.8 °C, were encountered in the ice-pushed ridges (see Fig. 5). This low temperature is probably caused by convective heat flow in the unsaturated zone, which can reach a thickness of 30 m. As the effective precipitation falls mainly in the winter, the percolation of the cold

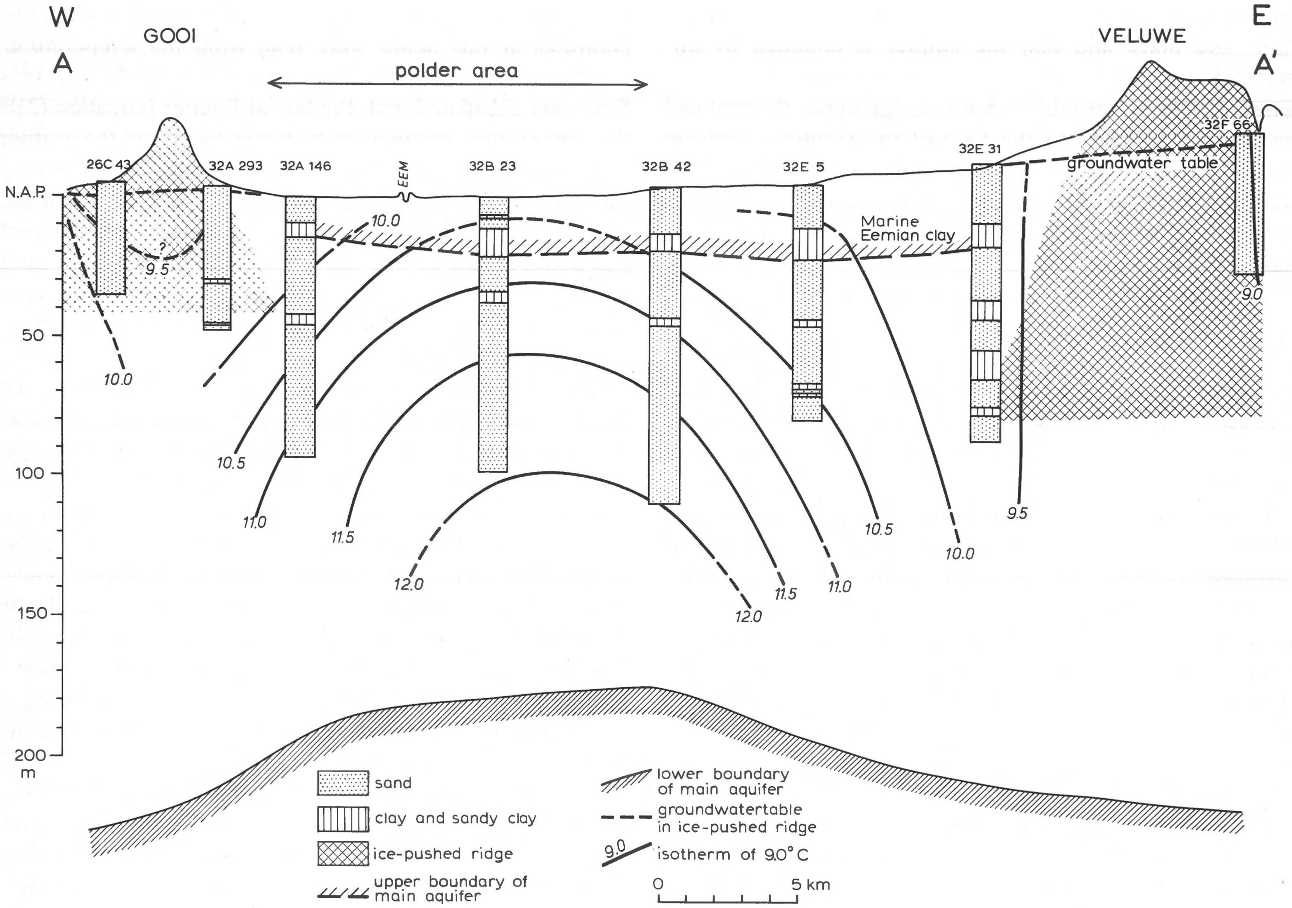

Fig. 5
Section AA' over the Gelderse Vallei (for location see fig. 1).

recharge water through the unsaturated zone results in a low temperature of the groundwater table. Due to the low thermal conductivity of unsaturated quartz sands, the unsaturated zone acts as an insulating layer for conductive heat flow.

High shallow-groundwater temperatures were measured in observation wells temporarily placed in the lakes between the Flevo polders and the old mainland near Harderwijk (for location, see figure 4). The average annual temperature at the lake bottom varied from 10.2 to 10.5 °C. The higher temperatures in these lakes were found in the shallow parts near the old mainland, whereas the lower temperatures were encountered in the deeper parts near the Flevo polders.

Vertical conductive heat flow

In the centres of the recharge areas small temperature gradients were observed (see Fig. 4). An example is observation well 32F66 (Fig. 6 A). This phenomenon can be fully explained by the effect of the downward groundwater flow present there. In these areas predominantly vertical steady-state heat and groundwater flows take place.

The specific discharge is taken to be the stated precipitation surplus of 1.0 mm/day. For these conditions calculated temperature versus depth profiles have been presented in figure 3. Comparison of the observed and calculated profiles show that the downward groundwater flow of 1.0 mm/day explains the small temperature gradients. It is assumed that the downward convective heat flow is also dominant in the other parts of the recharge areas.

Negative temperature gradients were not only encountered in the zone influenced by the seasonal temperature variation, but also at greater depths. An example is a profile measured in an observation well in the town of Huizen (Fig. 6B). The temperature of the groundwater table there is estimated to be higher than 12.0 °C. Underneath the city conductive heat is transferred from the warm groundwater table to the mainly horizontally flowing groundwater originating from the recharge area of Het Gooi. This heat transfer is perceptible down to a depth of approximately 40 m below the surface of the ground. Below this depth temperature is nearly a linear function of depth. Similar profiles were measured in the town of Harderwijk, where also high temperatures of the groundwater table were encountered.

Less pronounced negative temperature gradients were

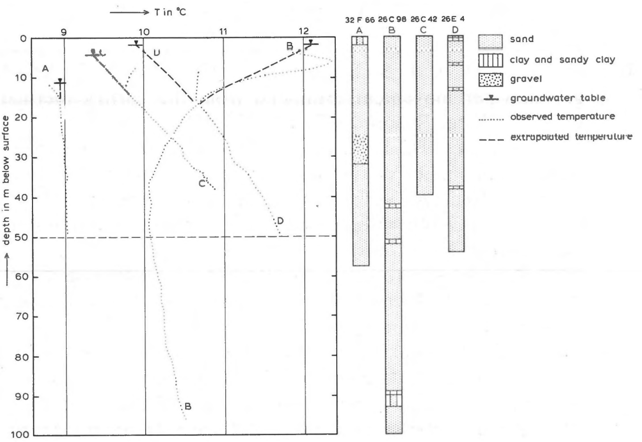

Fig. 6
Representative temperature profiles from observation wells (for locations see fig. 3).

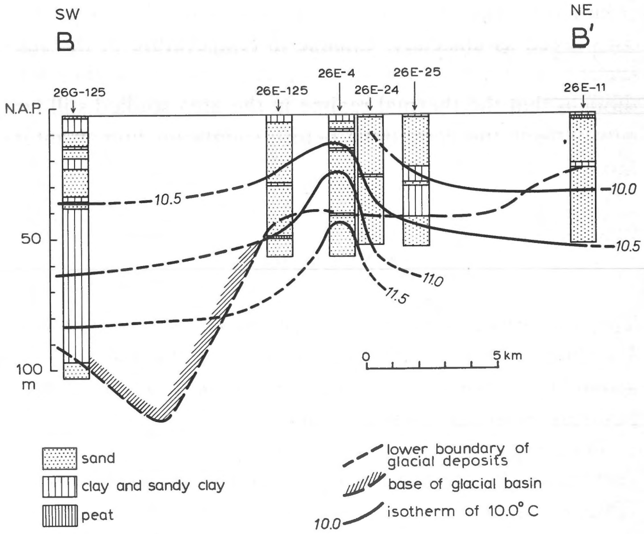

Fig. 7
Section BB' over the Knardijk (for location see fig. 1).

found at places where horizontal flow of groundwater predominates and where at the surface abrupt changes from wood to grassland and from land to lake water were present. The latter were observed in the observation wells temporarily placed in the lakes near Harderwijk.

In the major part of the Flevo polders and also in the Gelderse Vallei along the axis of depression in the piezometrice surface the temperature distribution seems to be little influenced by groundwater movement; hence the conductive heat-flow values approximate the regional heat-flow value. In this part of the Gelderse Vallei the horizontal flow of groundwater is slow, and also the upward movement is sluggish, due to the existence of thick clay layers.

In the IJsselmeer area the hydraulic gradients were probably very small before the reclamations. As temperature changes in the subsurface are slow, the present subsurface temperature distribution is thought to reflect the hydraulic conditions prevailing before the reclamation. Discharge in the zone along the Knardijk existed before the reclamation (CLAESSEN, 1972). This zone is characterized by high groundwater temperatures and heat-flow values (see Fig. 7). Temperature/depth profile D in fig. 6 was measured in this zone. This profile is slightly convex upwards. For steady-state one-dimensional flow of heat and groundwater this is an indication of upward seepage. The shallow-groundwater flow, however, is strongly influenced by the pattern of drainage ditches and is therefore not one-dimensional.

The highest heat-flow value at 40 m depth was calculated for an observation well northwest of the town of Huizen (see Fig. 6 C). This observation well is located near a local discharge area in an ancient sand quarry. This profile indicates that at least part of the discharge water in this area is derived from the deeper part of the aquifer.

It is remarkable that the estimated heat flow at 40 m depth seldom exceeds the geothermal heat flow. Also in the major discharge areas in the Gelderse Vallei, which are located along the margins of the infiltration areas, conductive heat flow values are well below the value of the regional geothermal heat flow.

Heat budget of the regional groundwater-flow systems

The previously presented steady-state heat-budget equation states that in a regional groundwater-flow system the geothermal inflow of heat at the lower boundary equals the sum of conductive and convective outflow of heat at the upper boundary.

Figure 4 presents estimated conductive heat-flow values at 40 m depth. As the temperature gradients above 40 m depth are generally constant, the estimated conductive heat flow at 40 m depth approximates to the conductive heat flow at the upper boundary, which is the groundwater table.

Figure 4 also shows that the outflow of conductive heat at the upper boundary is low compared with the inflow of geothermal heat at the lower boundary. According to the heat-budget equation the outflow of convective heat at the upper boundary should consequently be positive. It has been shown that the temperature of the water table in the recharge areas is low compared with the temperature of the water table in the discharge areas; this difference results in an important outflow of convective heat at the upper boundary.

It should be noted that the variables introduced in the numerical example of the heat budget equation in the theory are of the same size as those found in the Gelderse Vallei and that therefore the steady-state heat budget for the studied area is more or less balanced.

The groundwater-flow pattern in the studied area has been significantly altered by human activities. The largest influences are the reclamation of the Flevo polders and the extraction of groundwater by industries and water-supply companies. The subsurface heat flow should therefore be considered as unsteady. Change in temperature of the subsurface is, however, a slow process and it is therefore thought that the thermal regime in the area studied still largely reflects the conditions before significant human interference.

CONCLUSIONS

The subsurface temperature distribution in the Southern IJsselmeer area is strongly dependent on the regional groundwater-flow systems and on the variation of the temperature of the groundwater table.

Shallow-groundwater (< 40 m) temperatures are mainly determined by the average temperature of the groundwater table, which may vary approximately 1.5 °C. The variation in deeper-groundwater temperatures is also strongly dependent on the regional groundwater-flow systems and amounts to approximately 3 °C at 100 m depth and might increase to 5 °C at 200 m depth. The latter variation is thought to influence the direction and magnitude of conductive heat flow below the base of the main aquifer.

In the area studied average temperatures of the groundwater table in the discharge areas are higher than those in the recharge areas. The total outflow of conductive heat at the surface is small, as part of the geothermal heat flow is used to warm up the cold recharge water. This principle can probably be applied to most of the regional groundwater-flow systems associated with the higher parts of The Netherlands. This is caused in the first place by low shallow-groundwater temperatures prevailing in the infiltration areas, either by the presence of woodlands, or by the considerable thickness of the unsaturated zone or by both. In the second place high shallow-groundwater temperatures are expected in the discharge areas; among other reasons because part of the discharge areas consist of surface water, which generally shows a higher average temperature than the land surface.

Especially during the last decades the regional groundwater-flow systems and their thermal boundary conditions have been considerably changed by human interference. The thermal regime of the groundwater-flow systems in the area studied is therefore unsteady. The possibilities of determining the natural specific discharge of groundwater in the main aquifer with temperature methods are therefore limited. Change in temperature of the subsurface is, however, a slow process and the distribution of deeper-groundwater temperatures still partly reflects the distribution before human interference.

REFERENCES

Bredehoeft, J. D. & I. S. Papadopoulos 1965 Rates of vertical groundwater movement estimated from the earth's thermal profile – Water Resources Res. 1: 325-328.

Claessen, F. A. M. 1972 De geohydrologische gesteldheid van het gebied rondom het Veluwemeer tussen Harderwijk en Nijkerk – Dienst Zuiderzeewerken B 72-25 (internal report).

Csonka, J. 1968 Geothermisch onderzoek – Dienst Grondwaterverkenning TNO (Delft).

—— 1972 Rapport inzake een geo-elektrisch en geothermisch onderzoek in de Gelderse Vallei – Dienst Grondwaterverkenning T.N.O. (Delft) (unpublished).

Doppert, J. W. Chr. et al. 1975 Formaties van het Kwartair en Boven-Tertiair in Nederland. In: W. H. Zagwijn & C. J. van Staalduinen (eds.): Toelichting bij geologische overzichtskaarten van Nederland – Rijks geol. Dienst (Haarlem): 11-56.

Gischler, C. E. 1964 Hydrogeologisch en geo-electrisch onderzoek in de omgeving van Amersfoort (unpublished).

Haenel, R. 1974 Heat flow measurements in northern Italy and heat flow maps of Europe – J. Geophys. 40: 367-380.

Huisman, L. 1973 Temperatuurveranderingen bij kunstmatige infiltratie – H_2O 6: 653-656.

Meinardi, C. R. 1976 Geohydrologische gegevens van Zuidelijk Flevoland en de Gelderse Vallei – R.I.D. Meded. 78-4.

Nightingale, H. I. 1975 Ground-water recharge from thermometry – Ground Water 13: 340-344.

Parsons, M. L. 1969 Groundwater movement and subsurface temperatures in a glacial complex, Cochrane district, Ontario – Thesis Univ. Michigan.

Pluhowski, E. J. & I. H. Kantrowitz 1963 Influence of land-surface conditions on groundwater temperatures in Southwestern Suffolk county, Long Island, N. Y. – U.S. Geol. Survey Prof. Paper 475-B: 186-188.

R.I.D. 1970 Nota betreffende de waterbalans van een ruimte in het noordwestelijke deel van de Veluwe (internal report).

Sammel, E. A. 1968 Convective flow and its effect on temperature logging in small-diameter wells – Geophysics 33: 1004-1012.

Sorey, M. L. 1971 Measurements of vertical groundwater velocity from temperature profiles in wells – Water Resources Res. 7: 964-970.

Stallman, R. W. 1963 Notes on the use of temperature data for computing ground-water velocity – In: R. Bentall: Methods of collecting and interpreting groundwater data – U.S. Geol. Survey Water-Supply Paper 1544-H: 36-46.

—— 1965 Steady one dimensional fluid flow in a semi-infinite porous medium with sinusoidal temperature – J. Geophys. Res. 70: 2821-2829.

Thirriot, C. & R. Gandu 1975 Le doublet hydrothermique dans les nappes souterraines – Bull. B.R.G.M. (2), 111, 1-1975: 19-24.

Verruyt, A. 1966 On the dampening of seasonal fluctuations of the temperature of infiltrated groundwater – De Ingenieur G: 43-48.

Visser, W. A. 1978 Early subsurface temperature measurements in The Netherlands – Geol. Mijnbouw 57: 1-10.

Woodside, W. & J. H. Messmer 1961 Thermal conductivity of porous media 1, unconsolidated sands – J. Appl. Phys. 32: 1688-1699.