Groundwater

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ABSTRACT

The subsurface of the Netherlands is, from a hydrogeological viewpoint, dominated by a regional aquifer, consisting of medium-grained Plio-Pleistocene fluvial sand with a thickness ranging from 25 to 250 m. The aquifer is at the surface in the eastern half of the country and dips below semi-confining layers of lagoonal clay and peat in the western coastal area. The groundwater table is close to the surface almost everywhere and the precipitation surplus of about 300 mm/a is discharged by a dense and largely artificial drainage system. Most of the western half of the country is below sea level and consists of a patchwork of polders, each with its own artificially controlled level of surface water and groundwater. This has resulted in a complex system of infiltration into relatively elevated polders and in discharge by diffuse upward leakage of fresh and brackish water into deep polders. The eastern half of the country comprises (i) shallow aquifers in level areas, which are drained by seasonally contracting and expanding stream systems, and (ii) deep aquifers in more elevated areas with hardly any surface drainage and a groundwater table that reacts predominantly on annual variations in rainfall. A total volume of the order of 1700 million m^3 of fresh groundwater is annually extracted in the Netherlands; about 60% of this is used for public water supply.

Keywords: Netherlands, hydrochemistry, hydrogeology, coastal aquifer, groundwater management

Introduction

Groundwater occurrence

Groundwater is present in the pores and fractures of the entire sedimentary sequence of the onshore and offshore Netherlands. The geological framework of the subsurface controls the spatial distribution of porosity and permeability, and thus the distribution, thickness and structure of the hydrostratigraphic units.

Flow in the upper groundwater zone is primarily driven by rainfall-induced and topographically controlled, potential energy. At greater depths, groundwater becomes increasingly separated from the present-day hydrological cycle. In general, there is a gradual transition between the shallow and deep zones, depending on the occurrence and character of the poorly-permeable layers in between the two zones, and the persistence of the, either natural or artificially induced, vertical flow components. The turnover time of water circulation in the upper zone is of the order of days to thousands of years (locally more than 10 000 years), whereas water in the lower zone circulates on geological time scales.

The actual hydrogeological conditions in the lower zone are strongly influenced by residual components from the complex geological history. These deep groundwater systems are mainly driven by pressure gradients, caused by large-scale (paleo-)topography and tectonic forces, and by thermo-chemical processes. Permeable formations in the deep zone consist mainly of sand, sandstone and chalk. The low-permeable units include layers of clay, claystones and particularly the evaporites of the Permian Zechstein Group and to a lesser extent the Upper Germanic Trias Group (Verweij, 2003). The deformation history and the associated deep faults and salt structures disrupt the lateral continuity of the permeable and low-permeable units.

The upper groundwater systems consist of components of different orders of magnitude of regional extent and depth of penetration, which are genetically related to the land-surface topography and the associated water-table topography (Figs 1, 2). The main elements of the topographic relief within the Dutch landscape are the Holocene coastal dunes, the remnants of glacial features of Pleistocene origin in the central-eastern part of the country, and the pre-Pleistocene uplifted parts of the southernmost Netherlands (Fig. 1). The penetration depth of the (first-order) topography-driven groundwater systems is only in the order of hundreds of metres due to the presence of low-permeable clay layers or a succession of low-permeable sediments of mainly Early Pleistocene and older age. The maximum depth is reached in the Roer Valley Graben (cf. Fig. 6), where the influence of a supra-regional flow system is observed between depths of 500 and 1000 m below surface.

There is increasing awareness that groundwater plays an important role in geological processes. Understanding of the hydrogeological conditions is important for the exploration and exploitation of natural resources such as water, oil, gas, salt, surface mineral deposits and geothermal energy, as well as for the management of the storage of energy carriers and different types of waste, and for the prediction of geohazards. This chapter mainly concerns the upper groundwater systems. For knowledge of the deep groundwater systems in relation to the geological history

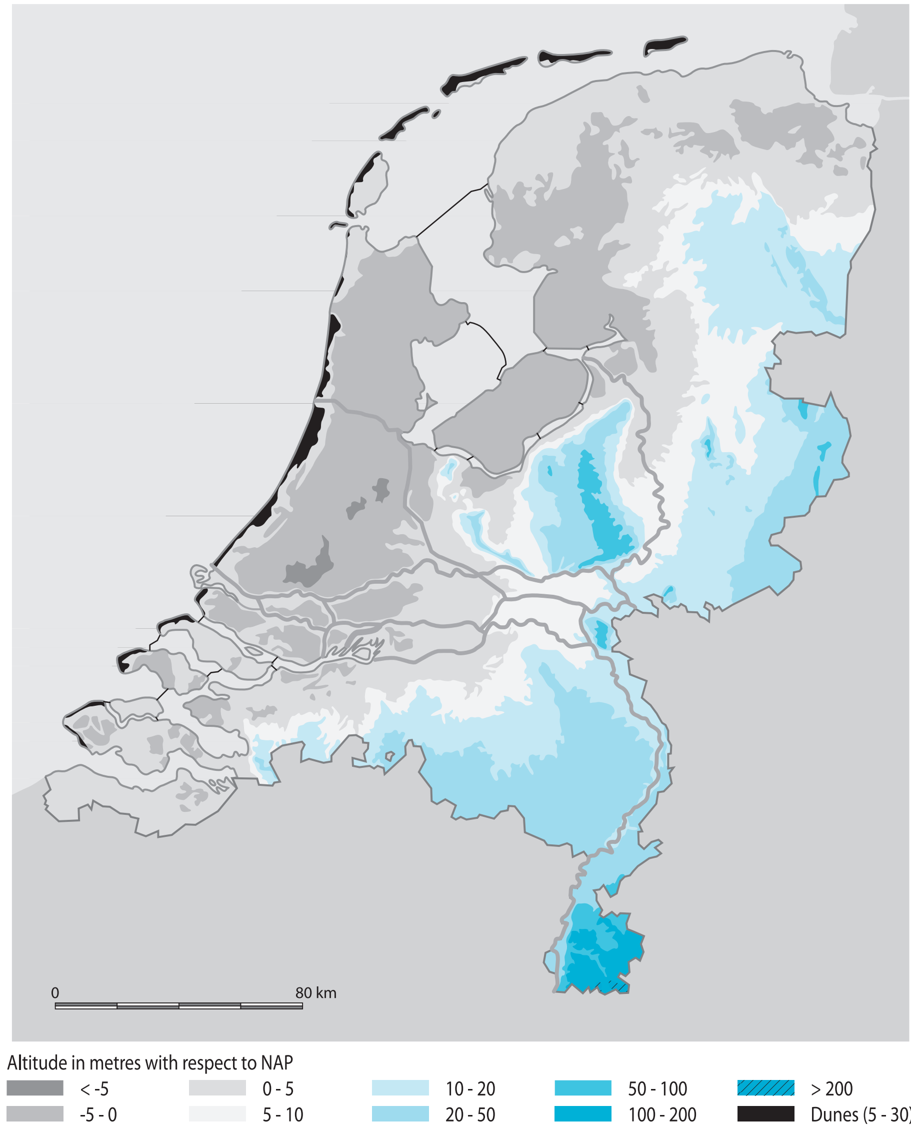


Fig. 1. Surface topography of the Netherlands; elevation relative to NAP = Normaal Amsterdams Peil = Dutch ordnance datum, approximately mean sea level (source: Dufour, 2000).

in the onshore and offshore Netherlands the reader is referred to Verweij (2003).

Development of groundwater research in the Netherlands

Groundwater research in the Netherlands is principally driven by the need to solve the specific water-related problems of coastal lowlands. These problems are in particular related to the search for suitable drinking water and the management of groundwater in an area with brackish groundwater and shallow groundwater tables in polders below sea level. Already during the founding of hydrology as a science in the 19th century, Dutch engineers and geologists, in an attempt to resolve these problems, made fundamental contributions to the concepts that at the beginning of the 20th century resulted in the basic theory of groundwater flow (De Vries, 1982, 2004).

The first systematic groundwater investigations in the Netherlands were carried out in the mid-19th century by Pieter Harting, a physician, geologist and professor of natural history. His interest in groundwater was, apart from scientific curiosity, stimulated by his concern to improve public health by the supply of good drinking water. Because of a lack of basic knowledge of hydraulics, his analysis did not really contribute to a better theoretical understanding of groundwater behaviour, but he nevertheless improved the general knowledge of the occurrence and quality of groundwater in its geological context (see Appendix). A first step in scientific groundwater exploration was made between 1839 and 1844 with the drilling of a

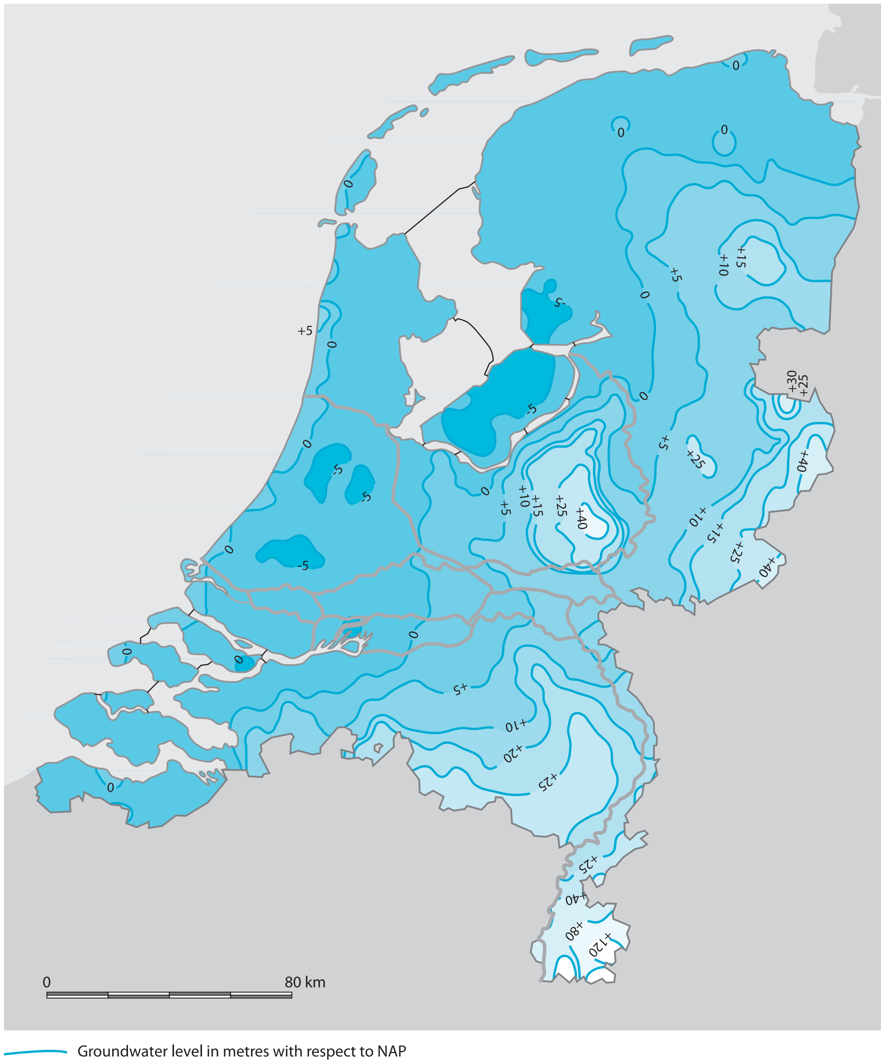


Fig. 2. Groundwater table relative to NAP, observations 9 February 1996 (source: Dufour, 2000).

well to a depth of 172 m at the Amsterdam Noordermarkt. Groundwater quality and yield of this well were disappointing, but the conscientious description and analysis of the soil samples by Harting (1852) constituted an important step towards the knowledge of the subsurface.

After persistent problems with polluted and brackish drinking water, the major towns in the west of the Netherlands eventually went to the groundwater reserves in the coastal dune belt for their water supply. In 1853, Amsterdam was the first to begin with the exploitation of the dune area near Haarlem, 30 km to the west. The main hydrogeological problem was that the small catchment area was surrounded by saline water, thus prone to depletion and salinization. Moreover there was no insight into the vertical extent of the fresh-water reserves, and their origin and replenishment were a matter of wild speculation. Hypotheses about the source of this water varied from local rainfall and condensation to artesian water veins originating in the remote higher grounds of the ice-pushed ridges, or even in the Ardennes in Belgium. Because of the fear of salinization, one prudently started with shallow extraction by drainage canals.

The first Dutch fundamental contribution to groundwater hydrology concerned the position of the interface between fresh and salt water below the dunes. In the 1880s, Captain Willem Badon Ghijben of the Army Corps of Engineers proposed the principle of a fresh-water lens, floating on salt water, which was recharged through local rainfall (Drabbe & Badon Ghijben, 1889). According to this principle, the thickness of the fresh-water pocket below the higher dunes along the Dutch coast was predicted to be not less than 150 m on average (Fig. 3). This remarkable hypothesis remained unnoticed until the German engineer Herzberg (1901) arrived independently at the same concept, which is now well-known as the Ghijben-Herzberg principle. Although the existence of this extensive fresh-water pocket was indeed proven by exploration drillings at the turn of the century, it took another 20 years before it was generally accepted that this fresh water originated from the limited input by local rainfall and not from an inexhaustible artesian inflow.

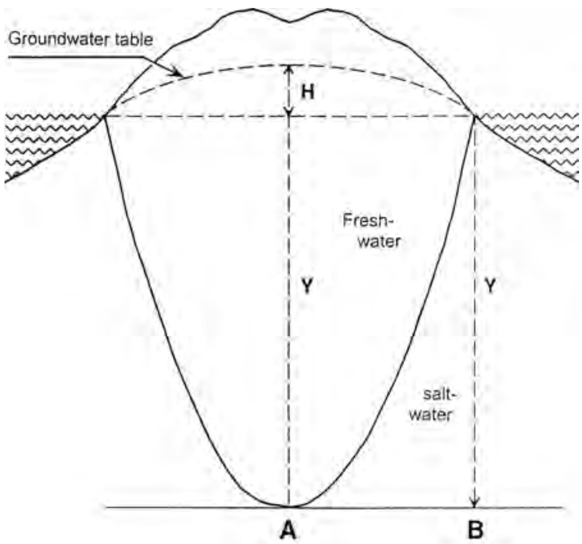


Fig. 3. The position of the fresh–salt water interface according to the Ghijben-Herzberg principle. The weight of the column of fresh water at A: $(y + H)\rho_f$, equals the weight of the column of salt water at B: $(y \cdot \rho_s)$, where ρ_f and ρ_s are the specific weights of fresh water (1 g/cm^3) and salt water (1.0238 g/cm^3) respectively, so that $y = 42H$. The average maximum height H of the water table in the Netherlands coastal dune area is in the order of 5 m above sea level.

At the turn of the century, Johan M.K. Pennink, director of the Amsterdam Municipal Water Works, warned against over-exploitation and salinization of the dune catchment. To convince the municipality, he simulated the upward movement of the salt-water boundary by overpumping, through experiments with viscous parallel-plate models. In order to cope with the growing water demand, he proposed artificial recharge of the dune area by river water, for which he developed a detailed scheme. His plans were only implemented half a century later once salinization problems had become serious. The onset of the salinization process was convincingly demonstrated by Pennink in a sound report published in 1914. Pennink experienced a long-lasting conflict with the Amsterdam municipality because he stubbornly refused to extract more water from the dunes than a percentage of the quantity he rightly assumed to be the rainfall replenishment.

Other even more essential contributions by Pennink were his field experiments to investigate the groundwater flow pattern around the drainage canals. Until then it was generally assumed that groundwater under free water-table conditions could not move in an upward direction, so that the flow to a drainage canal was limited to the depth of the canal. Pennink measured the distribution of the hydraulic head with a row of piezometers at different depths in a section perpendicular to the canals. By drawing the flow lines perpendicular to the measured equipotential surfaces, he proved the radially upward converging flow pattern near the canals, which explained the observed increase in hydraulic head with depth below the discharge areas. This downward increase in hydraulic head was pre-

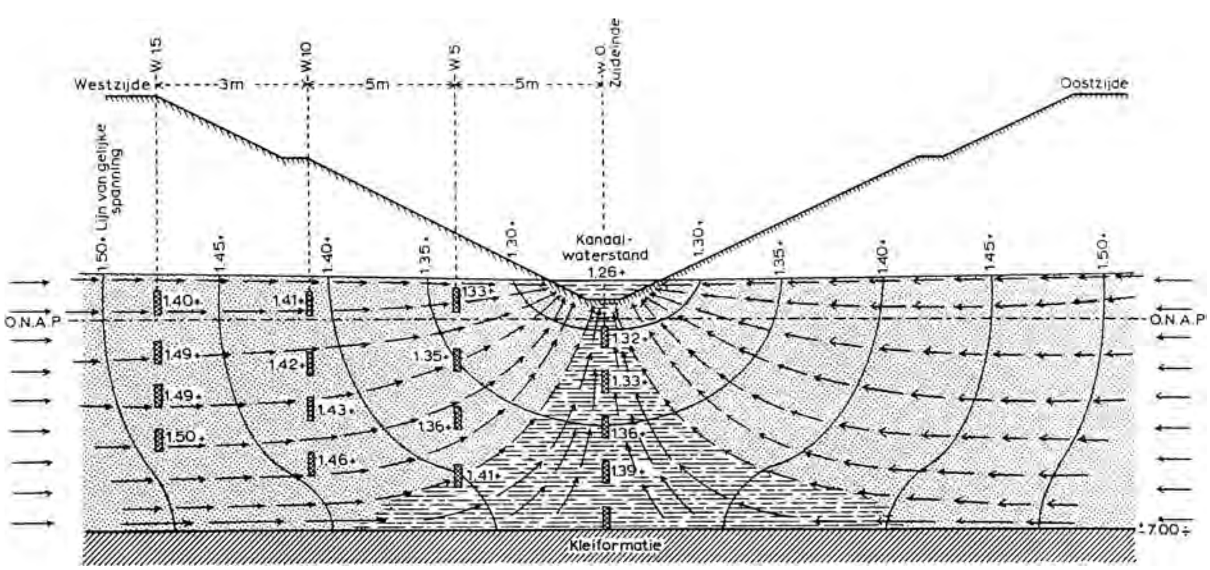


Fig. 4. Flow lines around a drainage canal based on hydraulic-head observations in piezometers; redrawn from Pennink (1905); kanaal-waterstand = water level in canal; lijn van gelijke spanning = line of equal hydraulic head; kleiformatie = clay layer.

viously used as an argument in favour of the existence of artesian water (Pennink, 1905; Fig. 4).

In fact, Pennink solved by performing field experiments, the continuity equation and made clear that groundwater flow is governed by a combination of the flow equation according to Darcy and the continuity equation according to Laplace. At about the same time, physicists like Ph. Forchheimer, C.S. Slichter and J. Boussinesq, proposed this concept on theoretical grounds. Pennink was not familiar with this theoretical approach, but he certainly was the first to prove experimentally the validity of this concept, that eventually at the beginning of the 20th century resulted in a general theory of groundwater flow.

This theory made it possible to simulate groundwater flow in a mathematical model, and to solve flow problems as boundary-value problems. The first Dutch contribution along this line was the solution of the typical lowland problem of groundwater flow in a leaky aquifer with abundant water at the surface. The derived formulas enabled the quantitative description of the complicated groundwater-flow conditions in polder areas and the prediction of the consequences of, for instance, groundwater extraction, water management and land-reclamation works. As early as 1914, J. Kooper, captain in the Army Corps of Engineers and engineer with the National Bureau for Drinking Water Supply (established in 1913), published the solution for the flow to a well in a leaky aquifer. His flow equations were subsequently elaborated by G.J. de Glee in 1930, and his formula for flow to an extraction well has become generally known as the 'De Glee formula' (Fig. 5). This solution appeared in the international literature only after World War II through the work of C.E. Jacob and M.S. Hantush in the USA.

A scientific approach to the problem of land drainage emerged in the 1930s under the leadership of S.B. Hooghoudt at the Experimental Station and Soil Science Institute in Groningen. These trail-blazing studies of Hooghoudt and his collaborators, notably L.F. Ernst, which combined theoretical analyses and plot-scale experiments, have led to a basic understanding of the processes of groundwater drainage and a sound basis for groundwater-table management (Hooghoudt, 1940; Ernst, 1962). Hooghoudt's program was relocated to the Institute for Land and Water Management Research (ICW) in Wageningen in the 1950s, which meant a shift of research focus to the influence of groundwater depth on capillary transport processes in the soil-water zone and the associated evapotranspiration and crop production.

Until the end of World War II, groundwater research and its application in water management in the Netherlands developed mainly along sectoral lines with separate solutions for problems related to public water supply, agriculture and general water management, including protection against flooding and salt encroachment. After the war, it became clear that an integrated approach was required to deal adequately with conflicting interests. This led to the founding of the Committee of Hydrological Research (CHO-TNO), under the umbrella of the Netherlands Organisation for Applied Scientific Research TNO. This Committee has played an eminent role in stimulation and coordination of hydrological research until the early 1990s. It perfectly documented the hydrological activities and achievements during this post-war period in a series of technical reports and proceedings of meetings. Another post-war initiative was the establishment of an archive for groundwater levels, which were systematically monitored in thousands of observation wells with the help of hundreds of volunteers. This work got a great impulse by a national inventory of the agricultural water condition by the Committee on Agro-hydrological Research (COLN-TNO), which was predominantly based on a survey of the groundwater-depth regime by a network of 23 000 observation wells (one per 100 ha) during the period 1952-1956 (Visser, 1958).

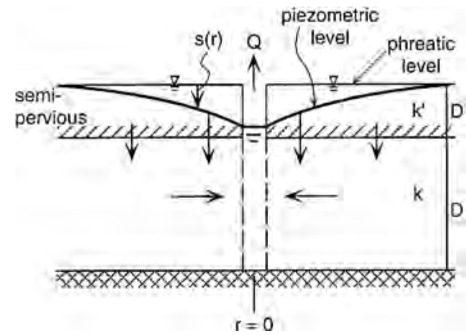


Fig. 5. Steady-state radial symmetrical flow to a well in a leaky aquifer, according to Kooper (1914) and De Glee (1930). For a small-diameter well in an infinite flow field, the solution is:
$s = \frac{Q}{2\pi T} K_0\left(\frac{r}{\lambda}\right)$,
with $\lambda = \sqrt{Tc}$, and where s is the lowering of hydraulic head at distance r from the well through a groundwater extraction Q; T and λ are transmissivity and leakage factor of the aquifer respectively; c is the vertical flow resistance, which equals D'/k', where D' and k' are the thickness and vertical hydraulic conductivity of the semi-pervious confining layer; K_0 is a modified Bessel function of zero order. This equation, and related formulas for flow around large-diameter excavations or polders, have been widely used for pumping-test analyses, evaluation of the impact of groundwater extractions, and assessments of groundwater inflow into deep polders and excavations.

The spirit of cooperation led to integrated regional surveys, which not only focussed on a more balanced water management, but also encouraged fundamental research on the interaction of surface water, groundwater, soil water and evapotranspiration. The thorough investigations in the small catchment of the Leerinkbeek in east Gelderland (Colenbrander et al., 1970) are an early example of this approach. The awareness of environmental pollution subsequently stimulated the study of hydrogeochemical aspects of groundwater. This resulted, for instance, in a textbook on hydrochemistry and groundwater pollution by Appelo & Postma (1993). To sustain the regional and integrated water management, a national groundwater-mapping, monitoring and data-management programme was established under the umbrella of TNO in the late 1960s. These groundwater surveys and analyses were based in particular on an approach in which coherent groundwater systems are considered as the basic units. The hydrogeological investigations gradually included other geotechnical aspects as well. This eventually resulted in the 1990s in a concentration of all national groundwater and geological surveys within the Netherlands Institute of Applied Geoscience (TNO-NITG).

The extensive contributions of Dutch hydrologists to the theoretical solution of groundwater-flow problems are presented in several internationally recognized textbooks, including Verruijt (1970, 1982), Kruseman & De Ridder (1970, 1990), Huisman (1972), Huisman & Olsthoorn (1983) and Bruggeman (1999).

Hydrogeological setting

Geological framework

Almost the entire territory of the Netherlands is part of the south-eastern marginal zone of the subsiding North Sea Basin. The limits of this basin are close to the south-eastern and eastern national boundaries, where they form a transition to the relatively more stable and uplifted areas of the Ardennes and the Rhenish Massif. The average elevation at the southern and eastern basin boundaries is normally of the order of 30 to 40 m, while the maximum elevation in the Netherlands of 322 m above NAP is at the southernmost tip of the country (NAP = Normaal Amsterdams Peil = Dutch ordnance datum = approximately mean sea level). Other relatively high areas are the Pleistocene ice-pushed hills in the central-eastern part of the country, notably the Utrechtse Heuvelrug and the Veluwe, with a maximum elevation of 107 m above NAP (Figs 1, 6).

The deposits participating in the present-day hydrological cycle, consist predominantly of Plio-Pleistocene, medium to coarse, fluvial sands with a thickness that increases north-westward to more than 300 m (Fig. 6; Breeuwer & Jelgersma, 1973). These sediments belong to the Upper North Sea Group and include the Echteld, Kreftenheye, Urk, Sterksel, Waalre, Beegden, Appelscha, Peize and Kieseloolite formations. The Appelscha and Peize formations were deposited by the former Eridanos River, which originated in the Scandinavian-Baltic area; the other formations have a Rhine and Meuse origin (De Mulder et al., 2003; De Gans, this volume). The lower part of the aquifer partly consists of a succession of coarse and fine, marine sediments of the Early Pleistocene Maassluis Formation and the Pliocene Oosterhout Formation.

The aquifer sands have an average permeability factor that ranges from 20 to 50 m/day, so that the average transmissivity of the subsurface is several thousands to locally more than 10 000 m^2/day (for explanation of these parameters, see Appendix). Semi-confining layers of various extent cause the Plio-Pleistocene fluvial deposits to behave as a multi-layer aquifer. On a regional scale, however, notably in the western half of the country, the aquifer can roughly be divided in an upper and a lower aquifer, separated at about 50 m depth below sea level by Middle Pleistocene clayey deposits. Glacial and fluvio-glacial clays of the Peelo (Elsterian) and Drente (Saalian) formations form extensive aquicludes in the north and north-east (Figs 7a, b). The Plio-Pleistocene aquifer is the most important source for public water supply in the east and south. Groundwater in the Pleistocene aquifer in the west is predominantly brackish due to marine influences during the Holocene.

The Pleistocene aquifer is at the surface in the east and dips under the clayey and peaty Holocene layers in the coastal area, and thus the upper aquifer shifts from predominantly phreatic in the east and south to semi-confined in the west and north. The Holocene consists mainly of fluvial, tidal-flat and estuarine deposits, accumulated under the influence of a rising sea level behind a series of coastal barriers. These deposits reach a maximum thickness of 25 m near the coast and are partly separated from the sea by a ridge of young coastal dunes, which locally reach a height of 50 m on top of the barrier. The main part of the Holocene coastal lowland is situated below sea level. The Pleistocene aquifer and the Holocene semi-confining layers, with abundant water at and near the surface, are in close communication because of vertical leakage, which is mainly induced by the artificial abstraction of water from the low polders (Figs 1, 2).

The Plio-Pleistocene aquifer lies almost everywhere on a sequence of clayey and sandy deposits of mainly marine origin, ranging in age from Oligocene and Miocene along the southern and eastern fringes, to Pliocene and Early Pleistocene in the western and central part of the basin (Fig. 7a). These low-permeable, basal sediment complexes contain predominantly brackish and saline water and form almost everywhere the lower boundary of the replenished groundwater system.

Unusual groundwater circulation occurs in the SE-NW running Roer Valley Graben system that cuts into the south-eastern basin margins (Figs 6, 8). Here the Plio-Pleistocene fluvial deposits are underlain by more than 1500 m of fine-grained sands and clays of mainly marine origin and Miocene and Late Oligocene age (Breda and Veldhoven formations). The salt–fresh water interface in the Breda Formation reaches depths of about 1000 m near the German border, which indicates that the marine sediments in the Roer Valley Graben have been desalinized by fresh groundwater inflow from the past and present outcrops of Tertiary sands in the east and south.

A series of continental sandy aquifers with intercalated lignite horizons (Ville and Inden formations) interdigitates with the Miocene marine clays. The lignite is being mined in Germany in open pits of hundreds of metres depth, just across the Netherlands-German border. Groundwater extraction to drain the excavations has caused a decline of more than 10 m in hydraulic head in some parts of the deep confined aquifers in the Dutch section of the graben, at a distance of more than 50 km from the excavations (cf. Fig. 16). Groundwater ages in this deep and supra-regional aquifer system range from 1000 years near the Dutch-German border to more than 10 000 years some 50 km to the north-west (Stuurman, 2000).

Other pre-Plio-Pleistocene aquifers are present in the uplifted area south of the Feldbiss Fault system, which forms the southern boundary of the Roer Valley Graben. Aquifers occur in Miocene fine-grained sandy layers in the northern part of this uplifted area, and in Lower Paleocene and Upper Cretaceous karstified chalk and marls further to the south. Both aquifers have a thickness in the order of 100 m. The Miocene aquifer is partly covered by Plio-Pleistocene gravel, sand and clay, whereas the northern part of the Cretaceous-Paleocene aquifer is overlain by confining Oligocene clay. The Oligocene clay dips to the north to form the lower boundary of the Miocene aquifer. The chalk aquifer is underlain by Upper Cretaceous sandstone, sand and clay of the Aken and Vaals formations (Fig. 8; Patijn, 1966; NITG, 1999).

The uplifted area, south of the Feldbiss Fault, forms a plateau, which is drained by incised branches of the river Meuse, notably the Geul in the south and Geleenbeek in the north. The exposed chalk area has a typical karst appearance with bowl-shaped depressions and dry valleys. The main draining river, the Geul with its tributary the Gulp, receives water from a number of springs, which particularly emerge in upstream areas where the river bed cuts into the interface of the chalk and the impervious clay. Several of these springs, producing tens of cubic metres per hour, have been intercepted and encased for drinking-water supply (Waterleiding Zuid-Limburg, 1941). The chalk is an important source for this supply, yielding up to 6 million m^3/a by the best producing group of wells. Large extractions take place along the Heerlerheide–Benzenrade fault system, where obstruction of groundwater flow by low-permeable fractures has resulted in concentrated vertical fluxes with associated chalk dissolution and enhanced permeability.

Groundwater at greater depths, separated from these recharged aquifers by low-permeable clayey deposits occurs in sandstone and quartzite inter-beds in Carboniferous shales and claystones. The main conduits through these low-permeable beds are related to NW-SE running fractures. In contrast, NE-SW directed fractures create barriers to groundwater flow. Breaches through these fracture aquifers during coal mining often produced temporary yields of hundreds of cubic metres of water per hour. Accumulated extraction for dewatering until the end of the mining activities in 1974 totalled 25 million m^3 (NITG, 1999).

Fig. 6. Approximate thickness in metres of that part of the Plio-Pleistocene aquifer that is involved in the present-day groundwater circulation (after De Vries, 1974); map also shows topographic names referred to in text; S and B are locations of the polders (land reclaimed from lakes) Schermer and Beemster.

There is a chemical stratification, ranging from $Ca(HCO_3)_2$ dominated water in the Cretaceous chalk, to NaCl water at depths below 500 m in the Carboniferous sediments. In between is $NaHCO_3$ water, formed by cation exchange during freshening of the original salt water (cf. section 'Process reconstruction and prediction'). Thermo-saline water with temperatures up to 50°C, ascends locally along deep fractures (Kimpe, 1963; NITG, 1999).

Most of the information on the subsurface and its

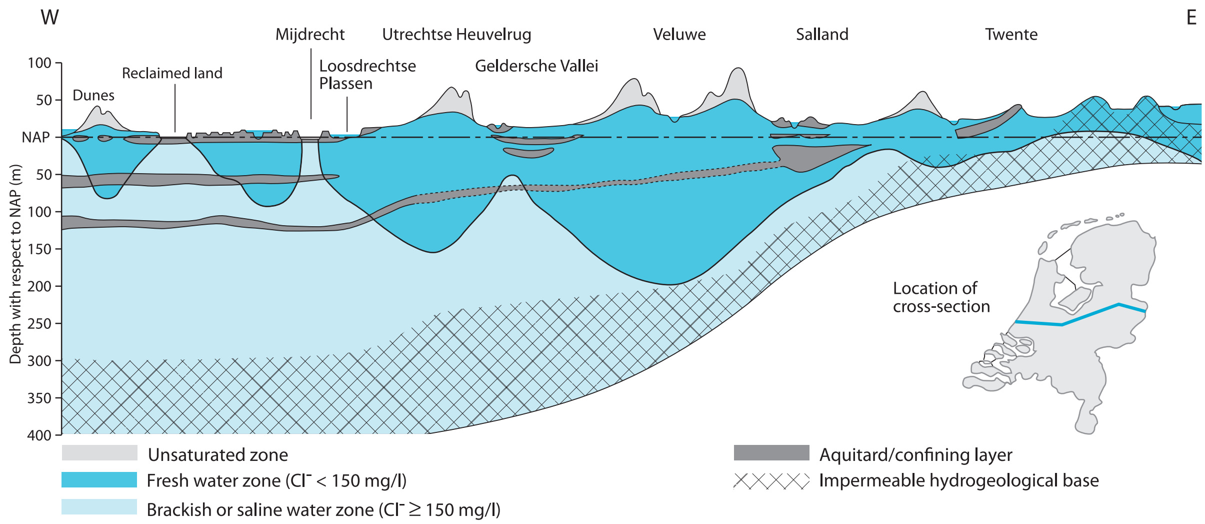


Fig. 7a. Schematic topographic-hydrogeologic east-west section showing the approximate depth of the fresh–brackish water interface (brackish > 150 mg Cl/l); length of section ca. 200 km (after Van de Ven (ed.), 1993; source Dufour, 2000).

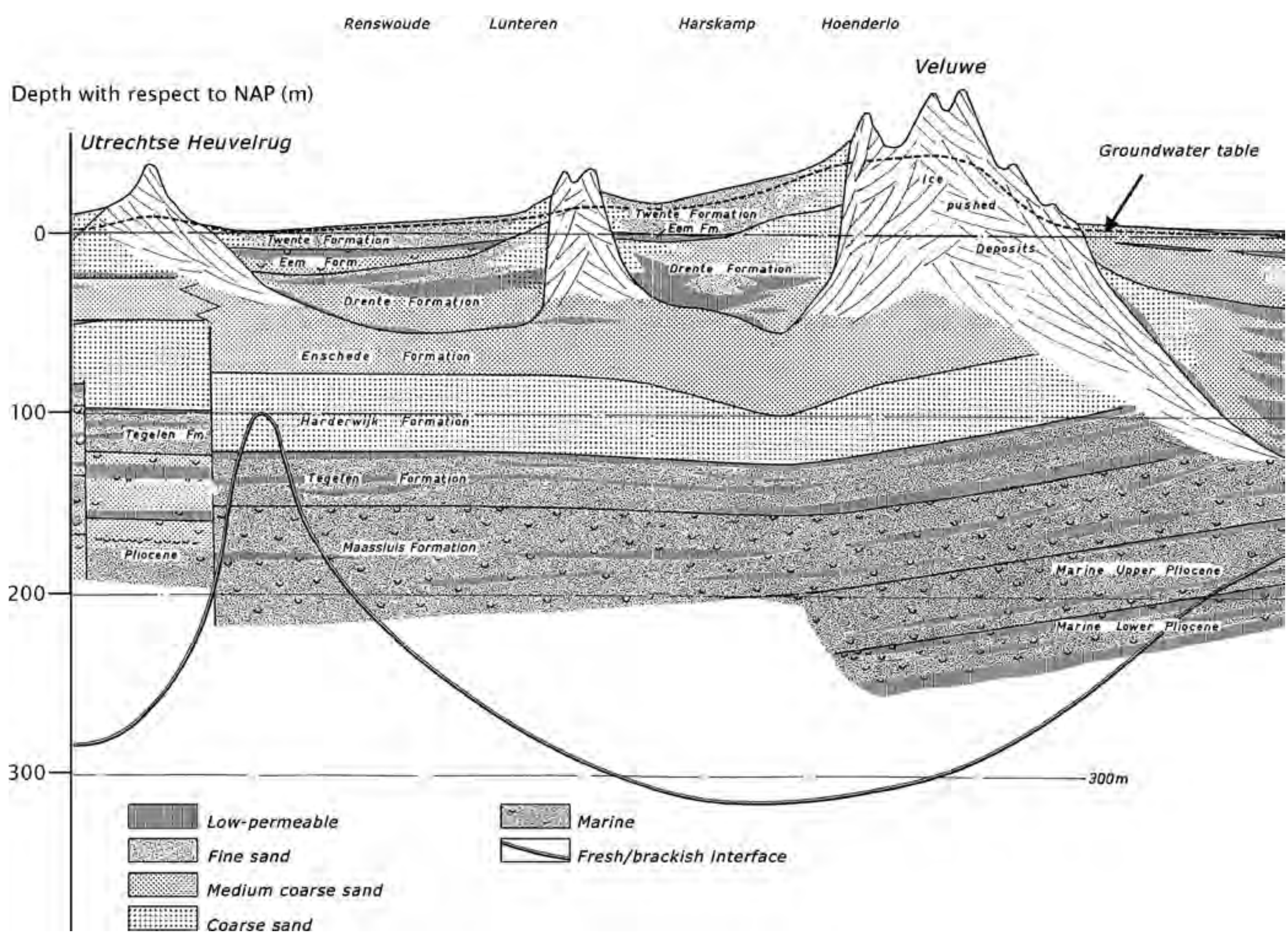


Fig. 7b. Detail section, WSW-ENE and 12 km south of Fig. 7a (after Breeuwer & Jelgersma, 1973, with stratigraphic nomenclature given by these authors). Length of section is ca. 60 km; the deeper position of the fresh–brackish interface in comparison with Fig. 7a is due to depth increase of the interface in south-eastern direction.

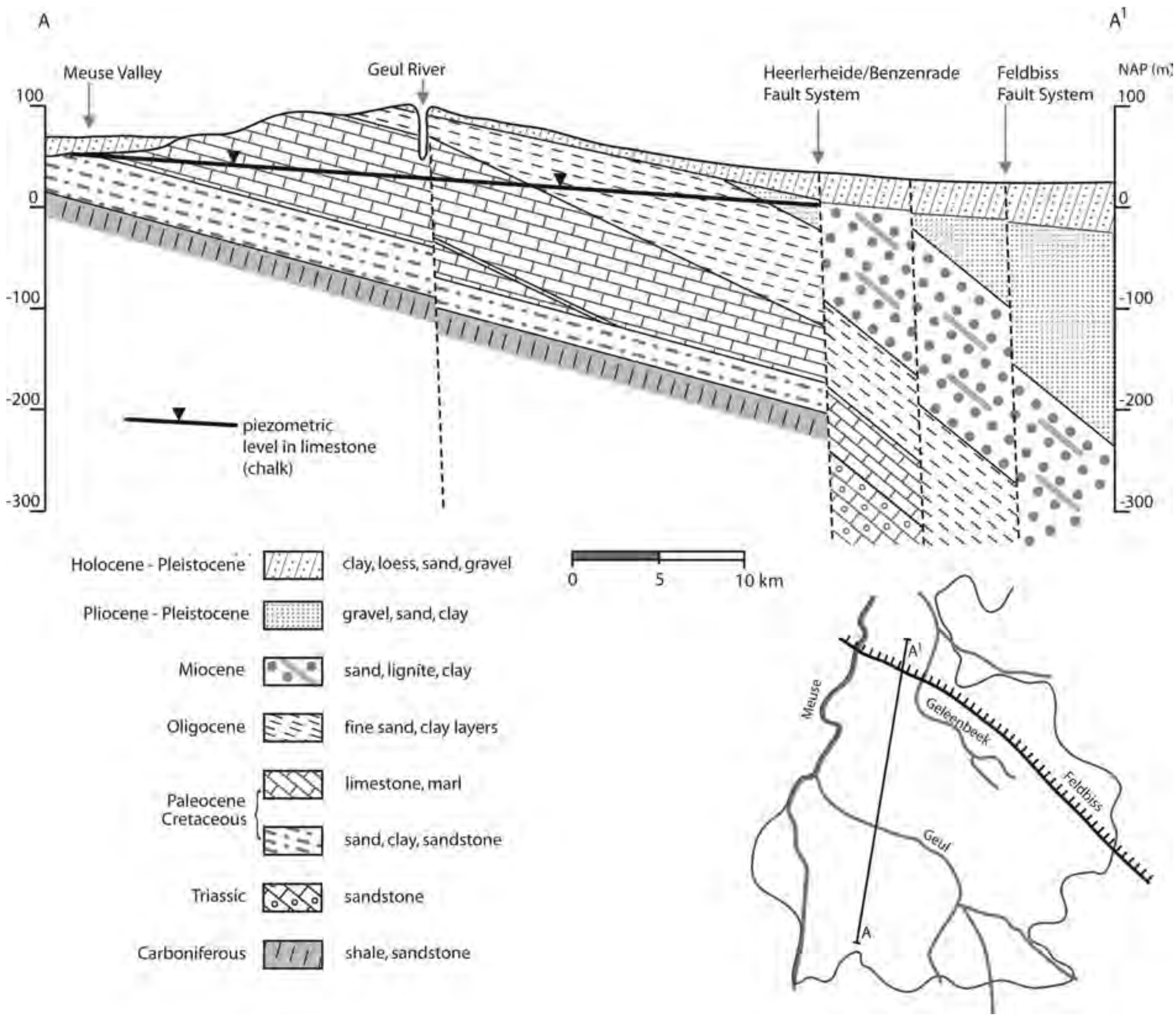


Fig. 8. Schematic hydrogeological section through the Cretaceous-Paleocene chalk aquifer in Zuid-Limburg (redrawn from Patijn, 1966). The piezometric level indicated refers exclusively to the chalk aquifer.

groundwater is contained in the archives of TNO-NITG and has been compiled in (digital) maps and databases. For a comprehensive overview of the groundwater situation in the Netherlands, the reader is referred to Dufour (1998, 2000).

Climate and groundwater recharge

Annual rainfall and evapotranspiration in the Netherlands are of the order of 800 and 500 mm respectively. Rainfall is evenly distributed over the seasons, but evapotranspiration is highest during summer, when it is more or less in balance with precipitation. Thus the average annual precipitation surplus of 250 to 350 mm is concentrated in the winter period (e.g. Meinardi, 1994). The rainfall surplus can be much higher in sandy areas with little or no vegetation; annual values of more than 600 mm were observed in bare dune sand (Stuyfzand, 1993). The lowest values, less than 150 mm, were measured in pine forest, where high interception-evaporation reduces infiltration (Stuyfzand, 1993; Gehrels, 1999). Inflow of water through the Rhine and Meuse amounts to an extra equivalent supply of 2100 mm/a (78 000 million m^3/a) on average. This river water is mainly used to combat salinization in the coastal area and to supplement public, industrial and agricultural water supply (NHV, 2004).

Almost all rainfall infiltrates the subsurface, and that part that does not return to the atmosphere by evapotranspiration, is subsequently discharged as groundwater by a dense and in many places artificial drainage system. The drainage capacity of this system normally increases with decreasing groundwater depth and reaches a maximum of the order of 15 mm/day in relatively low-lying areas with shallow groundwater tables and a dense drainage network. This applies particularly to the coastal lowlands where groundwater levels are artificially controlled by pumping. The clay and peat-containing soils in this area are strongly anisotropic due to low-permeable horizontal sediment layers. This obstructs vertical penetration of the precipitation surplus and stimulates concentrated lateral drainage through the upper metres. The marine Holocene deposits are extensively freshened, but pockets of remnant brackish water in low-permeable strata indicate the preferential flow of fresh water through relatively permeable sediments in buried tidal channels.

The groundwater table is shallow and almost everywhere less than 2 m below the surface with a maximum seasonal fluctuation of about 1 m, except for ice-pushed hills in the central and eastern Netherlands and the uplifted areas in Zuid-Limburg, where groundwater tables are much deeper. Figure 2 gives the overall pattern of the hydraulic head in the Pleistocene aquifer. Superimposed on this general pattern is the higher-order groundwater-table topography, related to the drainage system of small streams and ditches.

The Holocene coastal lowland

Hydrogeological evolution

The sea has repeatedly invaded the western part of the Netherlands during the Quaternary. Figure 9 shows the most easterly positions of the Pleistocene and Holocene coastlines and the overall present-day depth of the fresh–salt water interface. Saline groundwater is almost everywhere present in the marine Pleistocene and Tertiary deposits and has intruded a large part of the overlying fluvial Pleistocene aquifer in the coastal area. The maximum depths of this fresh–salt interface are reached below the coastal dunes and the ice-pushed hills, where the relatively elevated groundwater tables caused deep infiltration of fresh meteoric water (Figs 7, 9). Other deep fresh-water occurrences are found in the Tertiary aquifers in the south-east, where subsurface inflow from the outcrops situated further to the east and south, has flushed the marine sediments by (supra-)regional groundwater flow (cf. 'Geological framework').

Invasion by the sea took place mainly in four periods after the Early Pleistocene: Cromerian, Holsteinian, Eemian and Holocene. In most areas, salt water originated from the Pleistocene transgressions has been flushed from the coarse fluvial deposits during the last phase of the Pleistocene, the continental Weichselian, which lasted some 70 000 years. At the beginning of the Holocene a zone of brackish water below an eastward-dipping interface was probably present in the lower part of the aquifer as a result of diffusion, dispersion and compaction-driven flow from the lower marine sediments (Meinardi, 1991; Kooi & De Vries, 1998).

The Holocene sedimentation began with the development of a thick peat layer on the Pleistocene surface as a result of stagnating drainage in front of the encroaching sea. Subsequently, the peat layer became covered by low-permeable lagoonal clayey deposits, which protected the Pleistocene aquifer from salt-water intrusion by density currents, except where they became incised by deep tidal inlets. Recent transport and hydrochemical process modelling and isotope dating, make plausible that large areas may have been salinized as a result of vertical and lateral spreading by free convection of salt water from the large tidal channels (Kooi et al., 1999; Post, 2004).

Illustrative for the process of salinization through a transgression over a protecting clay layer, are the developments in the former Zuiderzee, an embayment of the North Sea, north-east of Amsterdam. This bay was separated from the sea in 1932 by a barrier dike (Afsluitdijk) and turned into a fresh-water lake (IJsselmeer), in which 2000 km^2 of polderland have subsequently been reclaimed (Figs 1, 6). The Zuiderzee came into being during medieval times by destructive encroachment of the sea through tidal inlets in this former peat area. The Holocene clayey deposits at the bottom of the sea protected the underlying Pleistocene aquifer in most of the area from salt-water intrusion by density currents, because there were no deeply incised tidal gullies opening up the aquifer from above. This is evident from Figure 10a which shows a gradual decrease of chloride content in the Holocene layer with depth, from 6000 mg/l (average chloride content of Zuiderzee) to less than 1000 mg/l at the Holocene/Pleistocene boundary. In the Pleistocene aquifer, the chloride content increases again with depth, to more than 6000 mg/l at the lower boundary of the aquifer at a depth of about 200 m (Fig. 10b).

Volker explained the Holocene profile by downward diffusion of salt from the Zuiderzee into the Holocene deposits, and attributed the Pleistocene profile to upward diffusion from the marine Lower Pleistocene into the overlying fluvial Pleistocene deposits (Volker, 1961; Volker & Van der Molen, 1991). However, deep borings in the marine Maassluis and Oosterhout formations revealed salt-water inversions, with lower chloride concentrations in the upper parts of these marine formations below a zone of higher salt content (Fig. 10c). These findings contradict the diffusion hypothesis for the Pleistocene aquifer but are consistent with salinization of this aquifer by density currents during Holocene transgressions. Age determinations of this brackish water in the lower part of the fluvial deposits also suggest the absence of important pre-Holocene components: more than 80% of 275 analysed groundwater samples from the Pleistocene aquifer in the west of the country indicate an age of less than 10 000 years (Post, 2004). Other inversions were encountered in the upper part of the Pleistocene aquifer on either side of clay lenses. This points to the delaying ef-

Fig. 9. Depth to fresh–brackish groundwater interface (isopleth for 150 mg chloride per litre; source: Dufour, 2000), and maximum eastward extension of Early Pleistocene and Holocene coastlines.

fect of low-permeable layers on the free-convection salinization process by density currents.

Different conditions prevailed north of the barrier dike, where deeply incised and shifting tidal inlets caused a complete salinization in the tidal-flat area of the Waddenzee behind the barrier islands.

Effects of land reclamation

The coastal lowlands used to form extensive peat marshes behind the coastal barrier with its dunes, and locally consisted of tidal flats and salt marshes along the tidal inlets. This situation prevailed when the first land-conversion works began at about 800 AD. The peat bogs then formed large dome-shaped areas, several metres above sea level, which were dissected by distributary branches of the main delta streams. Drainage of these areas for agricultural use caused a lowering of the ground surface by compaction and disintegration of the peat. Eventually this brought the area below sea level which made it necessary to build dikes, dams and other hydraulic structures, so that a patch-work of polders with artificially controlled water levels came into being (e.g. Van de Ven, 1993).

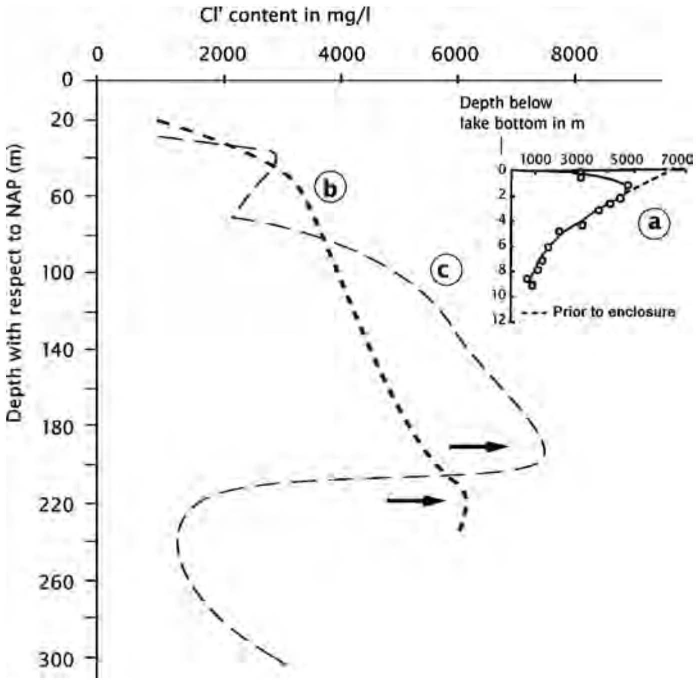


Fig. 10. Vertical chloride distribution in the area of the former Zuiderzee: (a) Holocene clay, central IJsselmeer; (b) Pleistocene sand, central IJsselmeer (after Volker, 1961); (c) Borehole 25G-32, situated east of Amsterdam at the edge of the Zuiderzee area, showing freshening of groundwater in the upper part of the marine Maassluis and Oosterhout formations below ca. 190 m. Arrows indicate clay layer at the boundary between fluvial Pleistocene and underlying marine Pleistocene and Pliocene deposits.

The reclamation of land from lakes and sea embayments was made possible by the introduction of drainage by means of windmills in the early 15^{th} century. Large areas were regained, which previously were lost by sea encroachment, river flooding and destructive digging of peat for fuel. The lowest polders are currently more than 6 m below mean sea level. The ongoing change in topography through land and water-management measures has set in motion a complicated pattern of groundwater recharge and discharge systems. Because of the artificially maintained groundwater tables in the low polders, a continuous flow of groundwater from the relatively higher regions into the lower ones occurs, causing upward seepage ('kwel') of groundwater from the Pleistocene aquifer into the polder area (Fig. 11).

The loss of water from the Pleistocene aquifer by this upward leakage is partly compensated for by subsurface inflow of fresh water from the higher grounds in the east (causing desiccation problems at the fringe of these higher areas) and by intrusion of sea water in the west (causing salinization in the lower areas). Saline groundwater brings through upward leakage a large amount of salt to the surface, which partly precipitates in the soil by evaporation during the summer. For the area between Amsterdam and The Hague, this input of chloride into soil and water may be in the order of 150 000 tonnes annually (ICW, 1976). Flushing of the soil by the precipitation surplus in winter, and flushing of the ditches and canals by river water, protect the area against complete salinization. The process of redistribution of salt and fresh groundwater takes thousands of years to reach equilibrium with new hydraulic conditions. With the acceleration in changing conditions through human interference during the last 1000 years, it is evident that the fresh–salt water interfaces are far from steady-state. Prediction of environmental problems within the framework of this long-term hydrogeological evolution is one of the priorities in groundwater research in this area.

A schematic reconstruction of the regional redistribution of fresh and salt water due to the water-management and land-reclamation works is depicted in Figure 12 (De Vries, 1981). This section is illustrative for the area between Amsterdam and The Hague, which is dominated by a central zone with deep polders, bounded by the coastal dune belt in the west and the ice-pushed ridge of the Utrechtse Heuvelrug in the east. Fresh water from these higher grounds and saline groundwater from greater depths and from the sea, discharge into the polders through upward leakage. Local groundwater flow systems, connected with the polder topography, are superimposed on these regional systems. Groundwater flow velocity calculations indicate that the landward intrusion of the sea water by lowering of the land surface during the last 1000 years, is not more than about 6 km. Thus, most of the saline groundwater originates from earlier transgressions and from old marine deposits at greater depths.

The seepage flux is mainly controlled by the hydraulic resistance of the Holocene confining layer (parameter *c* in Fig. 5), and can reach an areal average of more than 15 mm/day along the eastern margins of the area, where the remnants of Holocene peat on the Pleistocene sand form only a thin layer and where the hydraulic gradient is high. In fact, one polder, the Béthune polder (at the western fringe of the Utrechtse Heuvelrug), produces annually 30 million m^3 of fresh leakage water, providing 30% of the total drinking-water consumption of Amsterdam. The remaining 70% is extracted from the coastal dunes where the natural replenishment is supplemented by artificial recharge with water from the river Rhine.

The Holocene confining layers consist of alternating sandy channel deposits, clayey flood-plain and tidal-flat sediments and organic swamp deposits. Thorough analyses of the variation in hydraulic properties (notably the specific vertical flow resistance) in connection with the sedimentological architecture of the fluvial Rhine-Meuse Delta, and its geostatistical characteristics, were given by Bierkens (1994) and Weerts (1996).

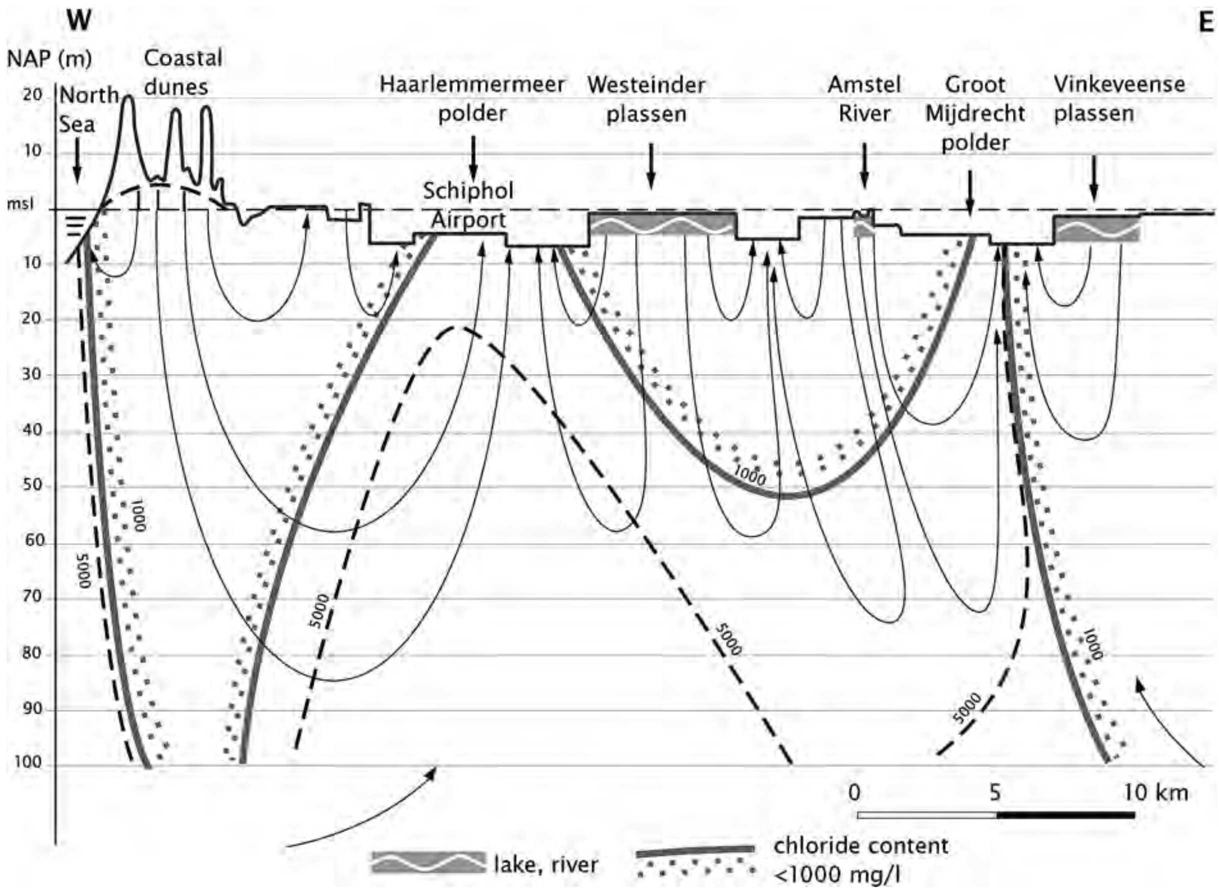


Fig. 11. Groundwater flow systems in an east-west section about 20 km south of Amsterdam, showing the strong influence of topography in this polder area (inferred from boreholes and numerical model simulation).

Process reconstruction and prediction

From the early 20th century onwards, Dutch geohydrologists have successfully developed mathematical models to describe and explain groundwater-flow conditions in leaky aquifers, for gravity as well as density-driven flow into polders and excavations, and for extraction wells (see section 'Development of groundwater research'). Equally important in reconstructing the paleohydrological evolution is the investigation of the regional distribution of hydrochemical water facies in combination with environmental isotopes. For example, freshening of salt water (NaCl water) and salinization of fresh water ($Ca(HCO_3)_2$ water) often leads to the following cation-exchange reactions at the clay or organic-material adsorption complex (Adc):

$$\text{Na Adc} + Ca(HCO_3)_2 \text{ water} \rightarrow \text{Ca Adc} + NaHCO_3 \text{ water},$$

$$\text{Ca Adc} + \text{NaCl water} \rightarrow \text{Na Adc} + CaCl_2 \text{ water}.$$

The occurrence of $NaHCO_3$ or $CaCl_2$ in groundwater thus testifies to the freshening and salinization processes, respectively. This occurs for example in the Frisian coastal area, where flushing by fresh water from the higher grounds produced $NaHCO_3$, whereas recent salt-water encroachment through groundwater extraction at the Noordbergum pumping station resulted in the occurrence of $CaCl_2$ (Fig. 13). The first to explain this origin of $NaHCO_3$ was the hydrologist and mining engineer Jan Versluys (1916). On the basis of this concept, Geirnaert (1973) prepared an overview of hydrochemical groundwater types in the west of the Netherlands in relation to the evolution of the distribution of fresh and salt water.

Appelo et al. (1990) recognized cation-exchange reactions to follow a chromatographic pattern along flow lines, and made use of this theoretical model to simulate paleohydrological processes. Laboratory experiments and isotope tracers were applied to calibrate the models. In this way, Beekman (1991) refined Volker's diffusion calculations (see section 'Hydrogeological evolution') and made a reconstruction of the diffusion and dispersion processes below the Zuiderzee bottom in relation to various phases of erosion and sedimentation since the first encroachment of the sea in medieval times. He further analysed the evolution of a 100-m-deep fresh groundwater occurrence in West-Friesland, west of the Zuiderzee area, and concluded from chromatographic analysis, chemical balances

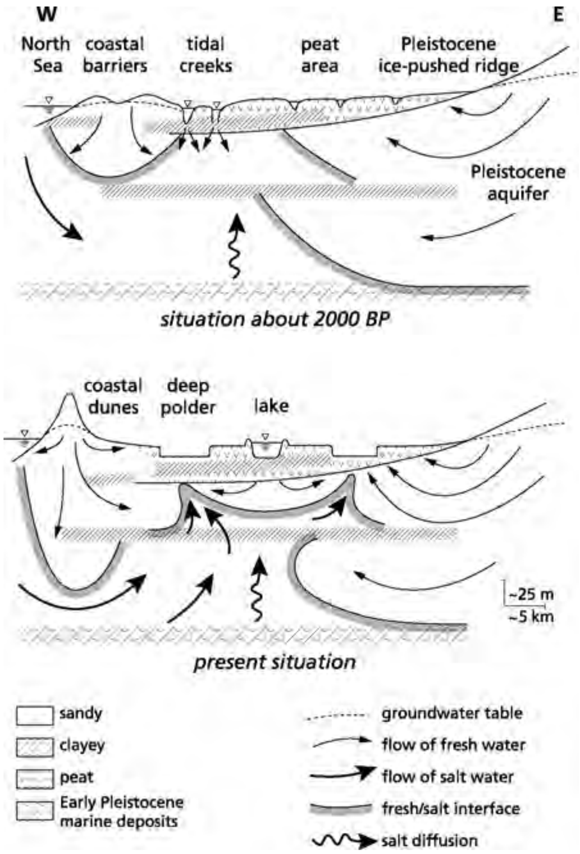


Fig. 12. Schematic 'genetic' topographic-hydrogeological sections through the area between Amsterdam and The Hague, showing the influence of land-reclamation works during the last 1000 years (after De Vries, 1981).

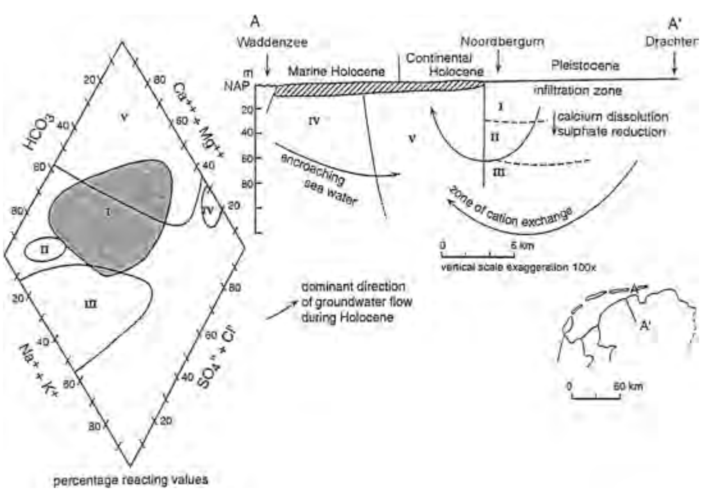


Fig. 13. Hydrochemical section across the Frisian coast and corresponding classification diagram, showing: (i) infiltration in low-carbonate soil (zone I), followed by dissolution of carbonate (II) and subsequent cation exchange with $NaHCO_3$ generation in the zone of freshening (III) and (ii) sea-water encroachment in zone IV, causing cation exchange and $CaCl_2$ generation in zone V; sea-water encroachment is caused by pumping at Noordbergum (based on data from Geirnaert (1973)).

and isotope studies an origin by local infiltration, mainly during the last 300 years. This infiltration process was initiated by the reclamation in the 17^{th} century of the deep polders Beemster and Schermer at the southern fringe of West-Friesland (cf. Fig. 6).

Another comprehensive hydrochemical-paleohydrological study was carried out by Stuyfzand (1993) in the coastal dune area. He reconstructed the evolution of the shape of the fresh-water lens and its chemical composition as caused by land-reclamation works behind the dunes as well as by groundwater extraction and artificial infiltration of river water. This study includes an advanced hydrochemical classification system, based on the origin of a water mass (hydrosome) and the characteristic hydrochemical zones (facies) within each hydrosome. A detailed chemical and transport modelling of the evolution of water artificially infiltrated from the river Rhine into the dunes, was carried out by Van Breukelen et al. (1998). This study was based on repeated sampling for chemical and isotope analysis in more than 100 mini-tubes along a distance of 1000 m over a period of 24 years. Dominant processes proved to be: (i) seasonally dependent cation-exchange, caused by the seasonally fluctuating NaCl content of Rhine water, and (ii) net dissolution of calcite from the dune sediments due to under-saturation of Rhine water and the production of CO_2 by dissimilation processes.

The application of isotopes may be illustrated by the identification of the provenance of fresh, upward seeping water in the previously mentioned Béthune polder, adjacent to a lake (Loosdrechtse Plassen) west of the Utrechtse Heuvelrug (Fig. 7a). Groundwater in the ridge of hills reflects the average Dutch rainwater ^{18}O content of –7.5‰ V-SMOW, whereas the lake water shows a value of –4‰, due to ^{18}O enrichment by evaporation. The seepage water has an intermediate content of –6‰, indicating a mixed origin (unpublished research Vrije Universiteit Amsterdam). Stuyfzand (1993) and Meinardi (1994) used various annual peaks in post-1950 nuclear-bomb-test tritium contents in rainwater as markers in vertical groundwater profiles to identify years of infiltration and to calculate the percolation flux. An example of a paleo-hydrogeological study by a combination of hydrochemical transport modelling and isotope analysis on a time scale of tens of thousands of years is the reconstruction of conditions and processes connected with rainwater infiltration and transport in the Tertiary Ledo-Paniselian coastal aquifer in Flanders south of the Dutch border (Van der Kemp et al., 2000).

Such reconstructions are a prerequisite for explaining the present situation and establishing the initial conditions to predict future developments of groundwater flow systems and redistributions of fresh and salt water by mathematical flow-simulation modelling. Oude Essink (1996) developed a comprehensive numerical computer model to simulate future shifts in the fresh–salt water interface in the coastal area, including scenarios for sea-level rise and continuing land subsidence. He predicted that, even in the case of the absence of future relative sea-level rise, it will require at least another 5000 years to reach equilibrium in the distribution of fresh and salt water.

The Pleistocene inland area

Aquifer character

In contrast to the Holocene coastal area, the Pleistocene area has remained relatively unchanged from a geological viewpoint during the last 10 000 years. The more humid conditions during the Holocene in comparison to the Weichselian, caused expansion of the stream system on sloping areas and the growth of extensive raised bogs on poorly drained flat areas. Subsequent changes were caused by human interference, including excavation of peat bogs, reclamation of dry land from marshy areas, canalization of rivers, regulation of runoff, and extraction of groundwater for public water supply.

The Plio-Pleistocene aquifer is almost everywhere filled with fresh water. Brackish groundwater is restricted to the marine, Early Pleistocene and Tertiary deposits (Figs 7, 9). The physico-chemical character of fresh groundwater normally changes from light-acidic to light-alkaline on its way from the recharge to the discharge area. Infiltrating rain water has a Na,K-sulphate/chloride character. Production of carbon dioxide and loss of oxygen by dissimilation in the upper soil layers subsequently cause the groundwater to dissolve calcareous material into bicarbonate and to reduce sulphate to sulphide. Most groundwater in the Pleistocene aquifer is therefore dominated by a Ca,Mg-bicarbonate water type. Mobilization of iron under conditions of low pH and/or low redox potential occurs particularly in marshy areas. Groundwater exfiltration in brooks in such areas has locally produced bog iron-ore deposits.

The exposed phreatic aquifers in the Pleistocene area are susceptible to contamination. Shallow groundwater is polluted diffusely by agricultural activities and atmospheric contaminants, and locally by industries and waste dumps. Nitrate concentrations often exceed the drinking-water standard of 50 ppm; concentrations up to 400 ppm have been observed at some locations as a result of manure deposition from high-intensity pig breeding. Denitrification of nitrate delivers oxygen, which in contact with pyrite (FeS_2), that is often abundant in marine sediments, can produce sulphuric acid according to the reaction:

$$2\,FeS_2 + 7\,H_2O + 15\,O \rightarrow 2\,Fe(OH)_3 + 4\,H_2SO_4.$$

This results in a low pH, which in turn can mobilize toxic metal ions.

Tritium data from the national groundwater-quality network have been used to investigate the relation between age stratification and pollution. This analysis revealed that the proportion of young (post-1950) groundwater in the upper 15 m of the sandy aquifer strongly decreases from more than 75% in the recharge areas to less than 25% in the discharge areas. The dense superficial drainage system in the low-lying discharge areas has removed a substantial part of recent water but prevents replenishment of the deeper aquifers (Broers, 2002).

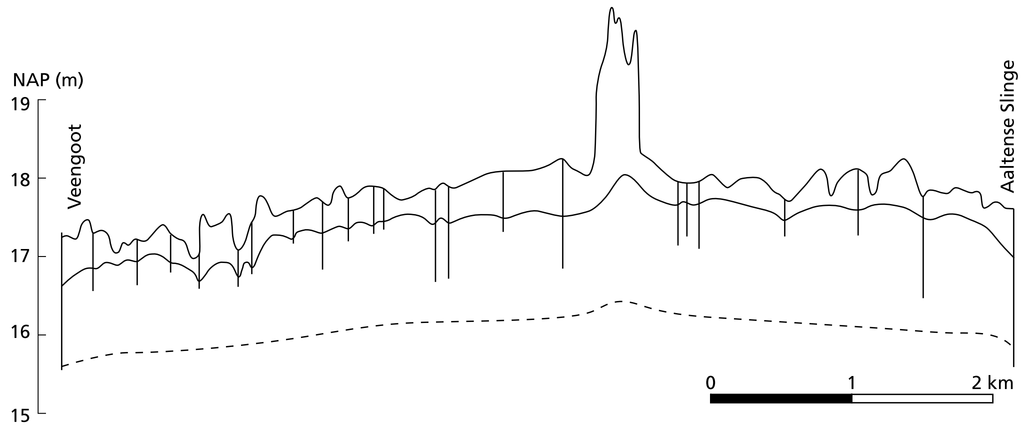
Groundwater abstraction from deep aquifers causes salinization and termination of groundwater seepage in valleys with valuable groundwater-dependent ecosystems. Shallow nutrient-rich groundwater thus displaces pristine groundwater, and rare oligothropic ecosystems are accordingly replaced by ordinary euthrophic communities (Stuurman, 2000). For an overview of anthropogenic influences on groundwater quality and the associated geochemical and microbial processes, reference may be made to Griffioen et al. (2003).

Local and regional groundwater and stream systems in level areas

The Pleistocene deposits outside the ice-pushed hills form a slightly undulating north- and westward sloping topography with an overall 1 : 2500 gradient (Fig. 1). Fluvial coarse-grained strata are covered by a layer of peri-glacial and fluvio-eolian, fine-grained sandy and loamy deposits with a maximum thickness of 10 m. Groundwater in the undulating areas is normally part of a hierarchy of nested groundwater flow systems. It is characterized by deep regional systems between the higher-order topographic elements, and by shallow, local or intermediate systems, which are driven by lower-order topographic elements like brook valleys and small streams and ditches (Engelen & Kloosterman, 1996).

The area is drained by a hierarchical stream system that expands and contracts with the seasonal fluctuation of the groundwater table (Fig. 14). This fluctuation thus regulates the number of channels that participate in the drainage process, so that the drainage density increases with increasing precipitation surplus. The maximum groundwater discharge capacity of a drainage system ranges between the average annual rainfall surplus of about 1 mm/day for the relatively higher areas with the deepest groundwater tables (> 5 m below surface), to 12 mm/day for the relatively low and flat areas with shallow groundwater tables (< 0.5 m below surface). This connection between decreasing groundwater depth and increasing discharge rate is due to the reduction in storage capacity with decreasing groundwater depth as well as to a reduction of the hydraulic-head gradient under low and flat topographic conditions.

De Vries (1974, 1994) characterized this topography-



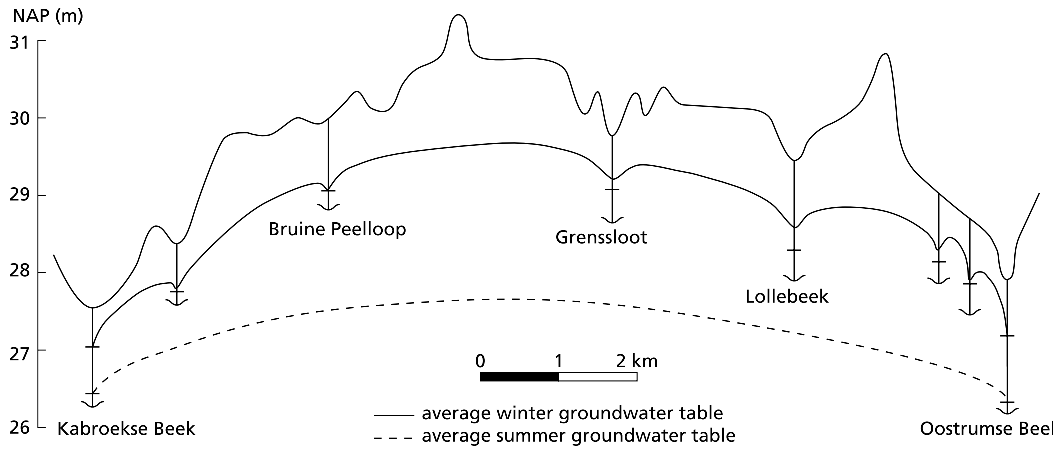


Fig. 14. Hydrological-topographic sections through two small stream systems in the Netherlands Pleistocene area, eastern Gelderland (above) and northern Limburg (below), indicating: (i) higher stream density with decreasing groundwater depth, and (ii) seasonal expansion and contraction of the discharge-contributing stream system, regulated by the fluctuating water-table depth (after De Vries, 1994).

controlled drainage system as an expression of the continuum of groundwater and surface water in which streams are initiated by groundwater seepage and sapping erosion at the intersection of groundwater table and topography in such way that the stream spacing corresponds to the required discharge capacity.

An interesting phenomenon in the area of the Roer Valley Graben is the influence of faults on groundwater flow in the Pleistocene sandy aquifer. Smearing and mineral deposition on the fault planes have created high resistance to horizontal flow. The result is a relatively steep hydraulic gradient across the fault with shallow, obstructed groundwater on the elevated block at the upstream side of the groundwater flow system. These unusual, wet soils on higher grounds are locally known as 'wijstgronden' (Bense, 2004). Exfiltration of groundwater upstream of the fault and associated sapping erosion have locally influenced the stream network.

Regional groundwater systems in higher areas

Groundwater in the more elevated areas, like the ice-pushed ridges, constitutes regional groundwater recharge systems with deeper groundwater tables. These areas are characterized by the absence of a shallow, lower-order drainage system. Such regional systems are only moderately susceptible to rainfall events and even seasonal effects, because the storage capacity is large enough to accommodate seasonal precipitation-surplus differences. Seasonal groundwater-level fluctuations are thus moderate because of the buffering effect of the thick soil–water retention zone. Main fluctuations are caused by the long-term imbalance between recharge and discharge in years of exceptionally high or low precipitation as well as by long-term climatic fluctuations (Fig. 15). Time lags between recharge and reaction of the groundwater table are of the order of one year because of (i) the time it takes the soil–water pressure wave to propagate through the soil–water zone (about 1 month per 10 m), and (ii) the delay

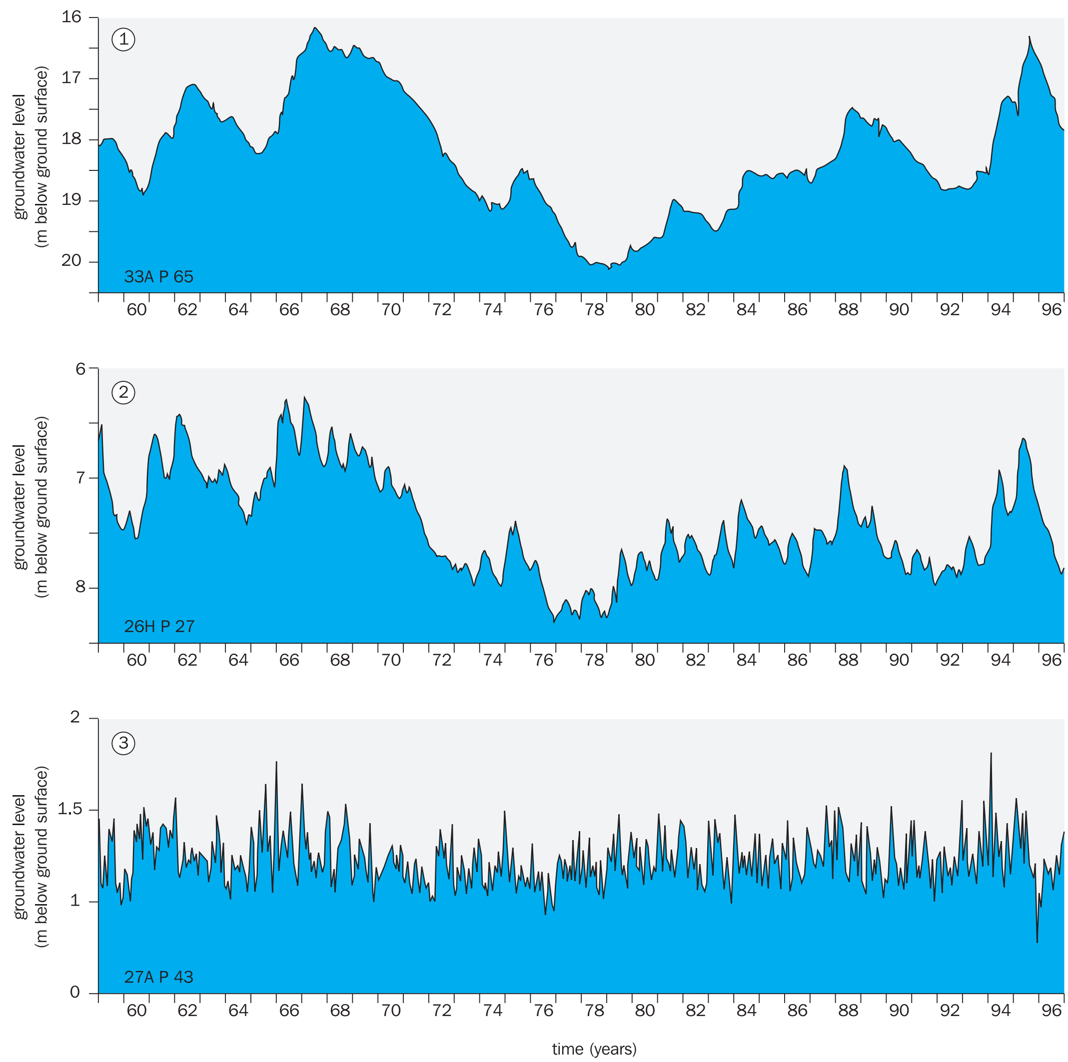


Fig. 15. Characteristics of groundwater-level time-series for three localities in the Veluwe ice-pushed ridge. This figure illustrates the buffering effect of groundwater depth (thickness of the unsaturated zone) on the impact of short-term, seasonal and annual rainfall fluctuations on the groundwater level. The absence of annual fluctuations in the shallow aquifer is due to the more or less artificially maintained regional groundwater table in this low-lying area as well as to the continuous subsurface inflow at the fringe of the ridge (after Van Bracht, 2001).

in the groundwater-level oscillation itself, due to the high drainage resistance of these large groundwater systems (Gehrels, 1999; De Vries, 2000). Conditions are different on the chalk plateau of Zuid-Limburg. The karstic chalk is characterized by a spatially irregular, but often fast response of the groundwater table to rainfall, and by strong groundwater-level fluctuations of sometimes more than 10 m (Jongmans & Van Rummelen, 1935).

Gehrels et al. (1994) carried out a time-series analysis of groundwater-level fluctuations in the Veluwe and could distinguish the impact of the reclamation in 1956 of the polder Oostelijk Flevoland (540 km^2) in the adjacent IJsselmeer from climate-induced fluctuations. Gehrels determined that it took about 25 years for the reclamation-induced head decline to arrive at the centre of the Veluwe, over a distance of 25 km. An interesting aspect of the groundwater recharge process in the Veluwe area is that the overall ^{18}O content of the groundwater (–7.8‰ V-SMOW) closely reflects the average annual content of local rainwater, although the rainfall surplus of about 350 mm originates mainly in the winter season. This winter rainfall has an average ^{18}O content of precipitation that is about 1‰ below the annual average. The explanation is a strong dispersion of seasonal rainwater in the root zone of shrubs and trees (Gehrels, 1999).

Another large-scale and supra-regional groundwater system occurs in the deep confined aquifer systems of the Roer Valley Graben (see section 'Geological frame-

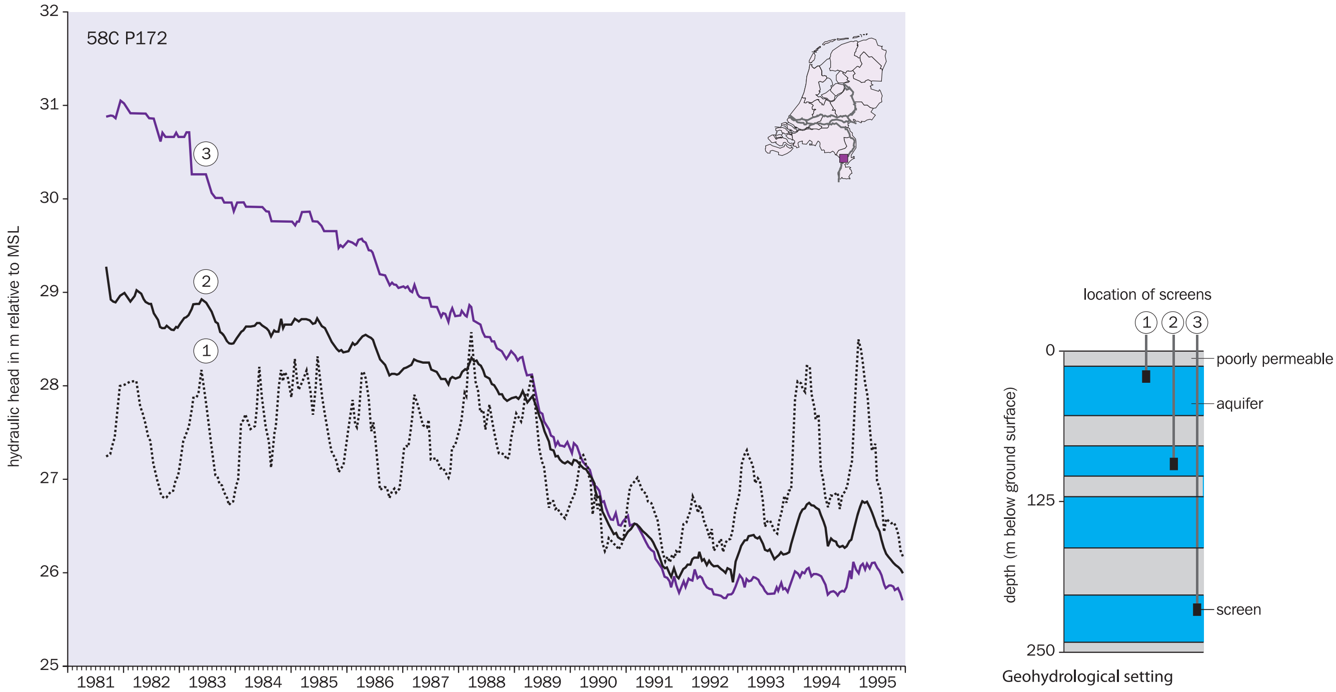


Fig. 16. Hydrographs for three screens in observation well 58C-P172 in the Roer Valley Graben in central Limburg. Groundwater extraction from the deepest aquifer in nearby Germany, caused the flow direction to change from upward seepage to downward infiltration; MSL is mean sea level (after Van Bracht, 2001).

work'). Figure 16 depicts the influence of groundwater extractions in the German lignite mines on the hydraulic head of the deep, confined Tertiary aquifers in the Dutch section of the graben. Accumulated extractions are in the order of 100 million m³ of water per lignite excavation. This cross-border decline in hydraulic head has only propagated through the confined elastic aquifers; the upper aquifers are unaffected. This caused the groundwater system to change from a condition with upward seepage into a situation with downward infiltration. The loss of seepage of old deep water with its specific chemical composition and high pH, has locally affected the natural vegetation in brook valleys in this area (Stuurman, 2000; Griffioen et al., 2003).

Relicts of paleo-conditions were encountered in the Early Pleistocene sediments beneath the ice-pushed ridges and their surrounding areas. For instance, the groundwater chemistry in borehole 32G-137 in the Eemvallei, 10 km SE of Amersfoort, shows the influence of a freshening in the upper part of the marine Maassluis Formation, with a downward increase of chloride to more than 6000 mg/l at a depth of 285 m. The ^{18}O content of this brackish water is between –8.5 and –9.5‰ (corrected for the seawater component), suggesting cold conditions during infiltration. Since ^{14}C age indications of this water are in the order of 10 000 to 20 000 years, it is plausible that the formation was partly flushed during the Weichselian by deep regional subsurface inflow from the adjacent ice-pushed hills (Meinardi, 1975). Borehole 25G-132 at the northern fringe of these hills (Fig. 10c) probably shows fresh-water relicts of the same process.

Large infiltration areas like the Veluwe and the Utrechtse Heuvelrug show relatively low subsurface temperatures, due to deep penetration of relatively cold rain water. The subsurface temperature in the Veluwe is less than 10°C at a depth of 125 m, whereas the average temperature in the surrounding areas is generally in the order of 13°C. Relatively high temperature anomalies are also observed in connection with groundwater ascending from greater depths along fractures, and in relation with rocks of high heat-conductivity, notably updoming evaporites in the northern Netherlands (Dufour, 1998).

Groundwater use and legislation

Groundwater exploitation

The total annual volume of groundwater extracted by water-supply companies for public and industrial use is in the order of 1000 million m³, of which about 18% originates from artificially induced recharge of river water, mainly into the dune area. Industry itself produces ca. 300 million m³, including water for cooling purposes (excluding cooling for power plants). Agriculture extracts 300 to 400 million m^3 from private wells for irrigation during dry summers. Most groundwater for the drinking-water supply is extracted from the Pleistocene area; it equals about 15% of the total precipitation surplus that this area receives. Considering that only part of this surplus recharges the deep aquifers and taking into account other restrictions, it is estimated that the potentially extractable resource is about twice the present extraction. Thus there is no overall over-exploitation, but nevertheless groundwater extraction has locally caused a drawdown of the groundwater table. This local drawdown together with the regional influence of intensified land drainage and reduced infiltration in built-up areas, has resulted in an overall lowering of the groundwater table in the Pleistocene area in the order of decimetres over the last 50 years (Dufour, 1998). Damage to valuable wetlands was one of the effects.

Groundwater production in the Holocene area is mainly restricted to the flood plains of the larger rivers and to induced river infiltration along stream beds. In addition, artificially infiltrated river water in the coastal dunes is an important source of drinking water for the cities in the coastal region. A special problem in this region is the necessity to maintain a shallow water table in urban areas, to protect the often centuries-old wooden foundation piles from decay.

Legislation

No special regulations with respect to groundwater had been developed until the mid- 20th century. Before that time every landowner could sink a well and extract water as long as no damage was done to the property of others. Limitations for the exploitation of groundwater were thus set by private law. This posed a problem to water supply- companies who could never be sure of continuity. Therefore the Grondwaterwet Waterleidingbedrijven (Groundwater Act Water Supply Companies) was enacted in 1954. From then on drinking-water companies needed a licence for any abstraction, specifying conditions and damage compensation; groundwater abstractions by others, however, were not authorized under this law. In 1981 an overall Grondwater Wet (Groundwater Act) became effective for all abstractions and activities related to infiltration and groundwater recharge. According to this law, the provinces are the authorities responsible for permission, registration and reporting. Quality aspects are mainly related to protection of recharge areas. Other groundwater-quality issues are dealt with in the Wet Bodembescherming (Soil Protection Act) of 1987, which includes regulations for prevention of subsurface pollution and for remediation of contaminated soils. Within this legislation, the provinces are obliged to set up and maintain a groundwater monitoring and management plan (Dufour, 1998, 2000).

Appendix

Some theoretical groundwater flow concepts (Fig. 17)

Quantitative groundwater hydrology was founded by Henry Darcy in 1856 with the results of his experiments on groundwater percolation through a vertical cylinder filled with sand. Figure 17 depicts this experiment with free outflow under atmospheric pressure. Darcy concluded that the flow rate per square metre of surface perpendicular to the flow path was proportional to the loss of hydraulic head Δh, inversely proportional to the distance e between the upper and lower flow boundary, and proportional to a coefficient k, depending on the nature of the sand. This principle that holds for all porous media is known as *Darcy's Law*; in formula:

$$q = k\Delta h/e,$$

with $\Delta h = h_1 - h_2$, where q is flow rate ($m^3/day/m^2$); h_1 and h_2 are *hydraulic head* at the upper and lower boundary of the sand layer respectively (m); k is *permeability factor* or *hydraulic conductivity* (m/day) in vertical direction; e is length of flow-line path (m). Hydraulic head is elevation head plus pressure head; assuming atmospheric pressure (zero pressure) at the lower boundary and the elevation at this boundary as arbitrary zero reference level, then the hy-

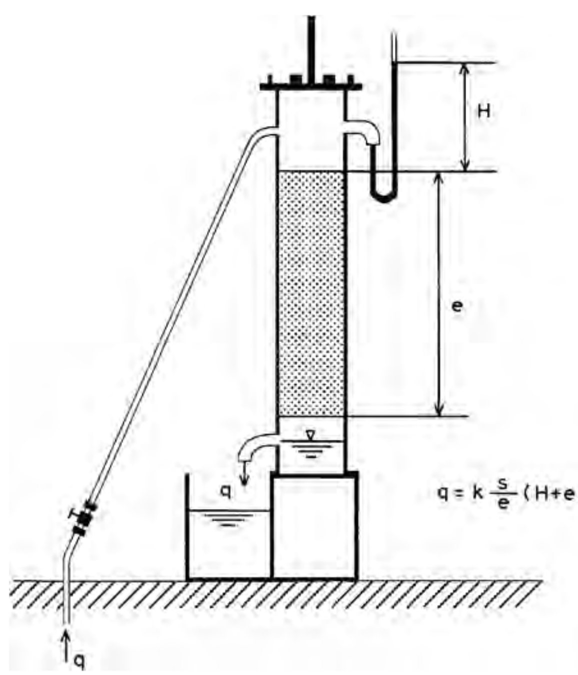


Fig. 17. Principle of percolation experiments according to Harting (1852) and Darcy (1856); see Appendix for explanation. s = surface area of cylinder.

draulic head $h_1 = H + e$ and hydraulic head $h_2 = 0$, where H is the hydraulic pressure at the upper boundary, thus:

$$q = k(H + e)/e.$$

For horizontal flow through an aquifer, the total flow rate per metre width is proportional to the product of the permeability factor k in horizontal direction and the thickness D of the aquifer. This product kD (m^2/day) is termed *transmissivity*, and often indicated with symbol T (see also Fig. 5). For vertical flow (leakage) through a semi-confining layer, the flow rate per m^2 is proportional to the hydraulic head difference between the upper and lower boundary of the confining layer, proportional to k' in vertical direction, and inversely proportional to the thickness D' of the confining layer, thus:

$$q = \frac{k' \Delta h}{D'}.$$

D'/k' is often termed *vertical flow resistance c* (day) of the confining layer (see also Fig. 5). Pieter Harting (1852) carried out similar experiments, but wrongly left out the elevation-head term e in h_1 (see 'Development of groundwater research in the Netherlands').

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