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Impact of recent climate variability on an aquifer system in north Belgium

Marc VAN CAMP, Kristine MARTENS & Kristine WALRAEVENS

Laboratory for Applied Geology and Hydrogeology, University of Gent, Krijgslaan 281, 9000 Gent, Belgium

ABSTRACT. The last decade it has been realised that climate and global change can (and very likely will affect) groundwater reserves. This will have impact on both the economical (groundwater exploitation) and ecological (ecosystems) aspects of aquifer systems. A key issue is the intrinsic variability in hydrodynamics due to natural fluctuations of meteorological conditions. In this paper the results of a modelling study are presented that reconstructs the hydrodynamic evolution of the important Neogene aquifer system in North-East Belgium, between 1833 and 2005. Boundary conditions are defined on a monthly basis. The results show that besides the yearly seasonal fluctuations also multi-year to decadal variations occur. These are especially important in the topographic higher regions like on the Campine plateau and the Campine cuesta, and can severely affect local groundwater dependent ecosystems as found in some protected areas and nature reserves. The same cyclicity is recognized in the intensity and extent of seepage of deep Neogene groundwater in the Nete basin.

KEYWORDS: modelling, hydrodynamics, Neogene, evolution

1. Introduction

Studying the impact of climate change on natural systems has become one of the main scientific challenges in recent years. Investigating the influence of natural climate variability on natural systems is a prerequisite to the understanding of system responses to external forcing stresses and helps to quantify the variability in resources. As groundwater dependent ecosystems largely depend on the availability of groundwater, they are very sensitive to groundwater changes and variations. These can be anthropogenic, as induced by groundwater pumping, which is often local. Fluctuations on a larger spatial scale may be caused by meteorological changes. Sequences of multiple dry or wet years can trigger periods with overall lower or higher groundwater levels. To quantify the potential risk of hydro-ecosystems, understanding the hydrodynamic response of the aquifer to these extreme meteorological conditions is useful.

The northern parts of the provinces of Antwerp and Limburg, in the north-east of Belgium, comprise a wealth of small streams and rivulets in the extensive Nete Basin (Fig 1), where many valuable groundwater dependent ecosystems have developed. Also the more elevated parts of the region (the Campine Plateau in the south-east and the Campine Cuesta in the north) contain many nature reserves and ecological valuable areas.

To secure the future of these valuable groundwater dependent ecosystems, an understanding of the intrinsic variations of groundwater levels and hydrodynamics in general is necessary. A first step is to quantify the changes that originate from natural variation of meteorological parameters like precipitation and potential evapotranspiration, which are the main controlling physical entities for recharge of aquifer systems. Without knowing this natural variability it is very difficult to evaluate the impact of anthropogenic influences like groundwater exploitation (pumping) or the climatic shift expected to occur by global warming. This natural variability has been investigated for the Neogene aquifer by means of a numerical groundwater model and a simulation of the hydrodynamic history during more than one and a half century, only taking into account natural conditions, excluding anthropogenic factors like pumping or changes in land use and hydrography. Spatial variation in aquifer recharge was also not considered.

2. Description of the Neogene aquifer system

2.1. Geological and hydrostratigraphical setting

The aquifer's substratum is formed by the Boom Clay (Fig 2), which is dipping in the northern and north-eastern direction. It is covered by Paleocene and Miocene sandy deposits from the Formations of Eigenbilzen, Voort, Berchem, Bolderberg, Diest and Kattendijk. These are hydraulically well connected and form a single aquifer, which is identified with HCOV("Hydrogeologische Codering van de Ondergrond van Vlaanderen", Meyus et al, 2000) code 250. The Miocene aquifer is covered by more clayey layers which belong to the Pliocene Formation of Lillo. This aquitard is identified with the HCOV code 240 and separates the underlying Miocene aquifer from the overlying Pleistocene and Pliocene aquifer (HCOV 0230).

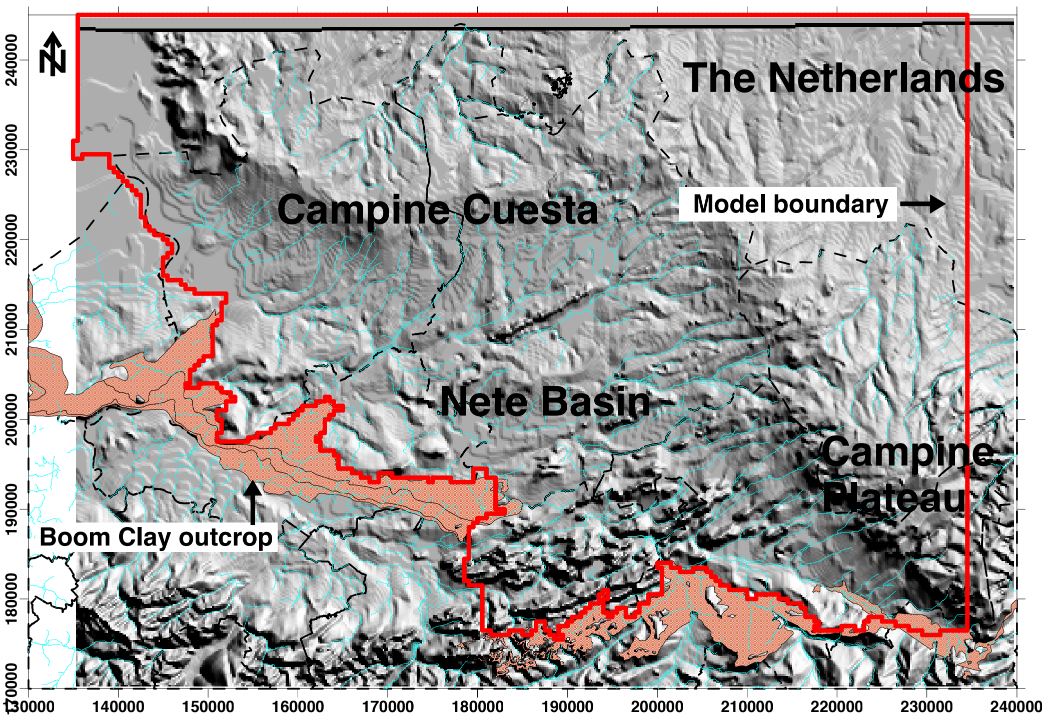

Figure 1. Location and physiography of the Neogene aquifer.

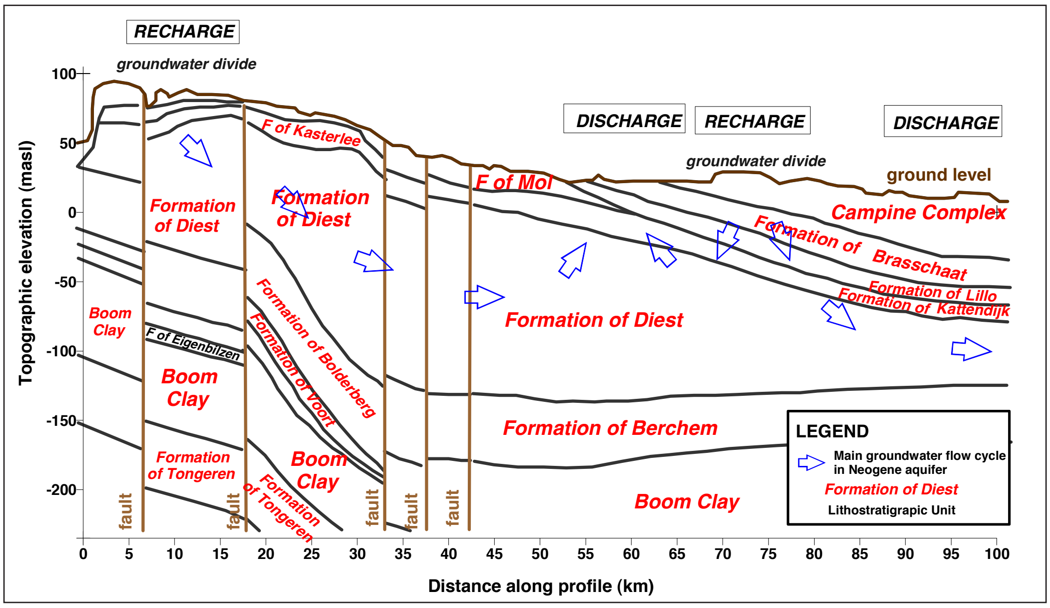

Figure 2. Cross section.

This latter includes the late Pliocene and Pleistocene sands of the Formations of Lillo, Kasterlee, Poederlee, Mol, Merksplas and Brasschaat. In most of the study region this aquifer is phreatic as it is only covered by younger Quaternary sands. In the north of the province Antwerp a very heterogeneous complex is deposited on top of the Pleistocene and Pliocene aquifer, making it semi-confined. This complex (HCOV code 220) contains sand, silt, clay and peat layers and is both vertically as laterally very heterogeneous with strongly varying textures.

2.2. Topography and hydrography

Topography (Fig 1) drops from the east to the west and to the north. The highest levels are found on the Campine Plateau (around + 70 masl) in the southeast, while the lowest levels occur in the lower Nete valley in the west (less than + 10 masl). In the north a west-east oriented ridge separates the Nete river valley from low lying polders in The Netherlands. This cuesta is formed by clay layers in the Campine Formation. The southern side of the cuesta is steeper than the northern side.

2.3. Groundwater exploitation

The Neogene aquifer is the most important aquifer system in north-east Belgium. Due to its considerable thickness, increasing to the north, and high hydraulic conductivity and accordingly high transmissivity (up to a few thousand m^2/day), it can provide high amounts of water even with relatively low drawdowns (often no more than a few meters). Over the last decades numerous pumping sites have been installed, ranging in size and capacity from local home users to drinking water companies with well fields capturing up to 10 million m^3/year. In recent years total exploitation rate in the Neogene aquifer is around 150 Mm^3/year (Coetsiers, 2007), which is nearly 12% of the estimated total aquifer recharge. Most of the water is taken from the Formation of Diest and in the northern region also from the Formation of Brasschaat, often found at 30 or 40 m depth.

2.4. Piezometric levels and hydrodynamics

Measurements in more than 500 wells were used to make a piezometric map (Fig 3) of the Neogene aquifer for the year 2003 (Coetsiers, 2007). Although most mesurements were not taken at exactly the same time,

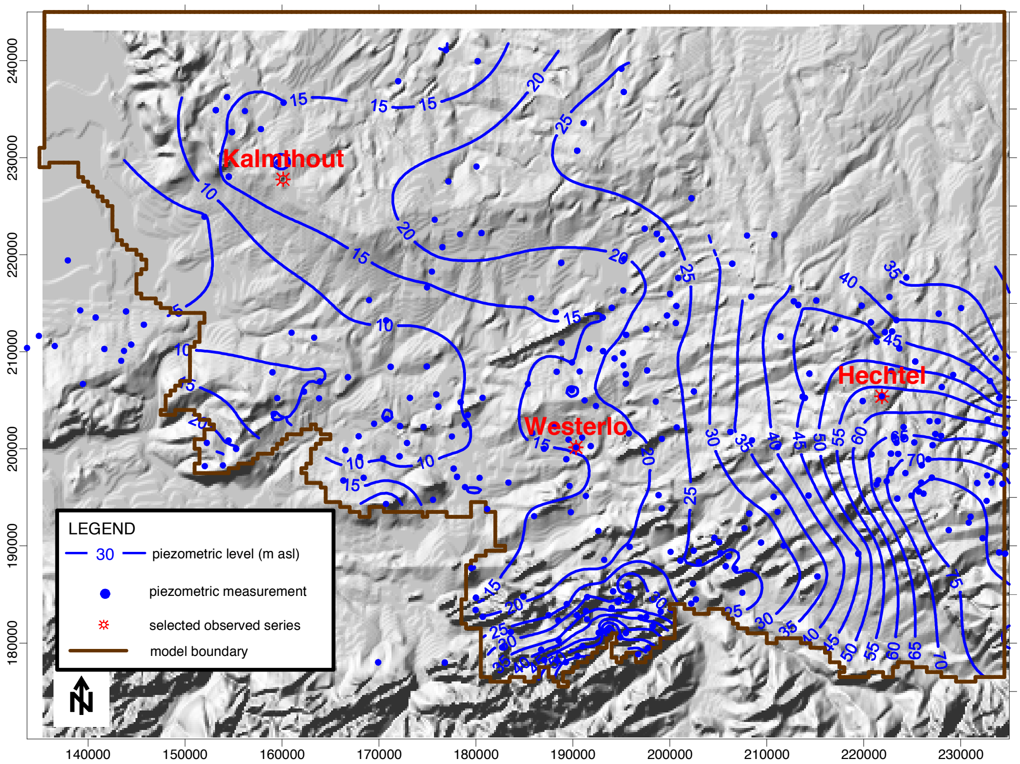

Figure 3. Piezometric map anno 2003.

but were rather spread over the year 2003, the resulting map gives a representative view of the piezometric surface. It captures the main hydrodynamic trends and groundwater flow cycles in the region. The highest piezometric levels are found in the southeast on the Campine Plateau, where ground level reaches around 70 masl. Piezometric levels decrease towards the east, the west and the north. Topography is the controlling factor for the main groundwater cycle. On the Campine Plateau itself the hydrographic network is sparse and most rainwater infiltrates. From here groundwater flows mainly laterally towards the low lying Nete Valley, where seepage drain it into the Nete river and contributaries. As the deep Neogene groundwater usually has high dissolved iron concentrations because of the reducing redox conditions, brown coloured precipitation in many ditches, caused by oxidation of the upwelling water, indicates the occurrence of these seepage cycles in the valleys. Under the northern region, close to the border with the Netherlands, a secondary groundwater divide is found at the topographic height of the Campine Cuesta.

A whole set of observation wells is being used now by the Flemish government (department VMM) to monitor groundwater levels, but most of these wells were installed only during the last decade. Few long time series, spanning multiple decades, exist. The oldest series started at the end of the 1970s. Some typical data sets are given in Fig 4. The location of the series are indicated on Fig 3. A first well (Fig 4a) is located in the south-east on the Campine Plateau in Hechtel. Levels here are very high (around 56 m asl) because of the high topography, and in the series the regular seasonal fluctuations (high winter and low summer levels) can clearly be seen. But the most important characteristic is the presence of a multi-year trend that is characterized by an alternation of periods with lower than normal levels, like 1990-1992 and 1996-1998, and higher levels, like 1987-1989 and 1999-2002. Winter levels in the dry periods can be lower than summer levels in the wet periods. The same pattern is found in observation wells located close to the groundwater divide on the Campine Cuesta, in the norhwestern part of the region, like the well in Kalmthout (Fig 4b). Here, summer minima show a range of up to two meters, which is more than the average difference between winter and summer. The general fluctuations in this longer series (from 1981 on) correspond very well with the behaviour of the well in Hechtel. The range of the seasonal variations depends on the specific yield of the phreatic layer, which is related to the lithology of the subsurface and can be different for these two sites. In contrast, a typical well located in the Nete Valley in Westerlo (Fig 4c), shows much less pronounced interyearly variations. Although high winter levels can still vary considerably from year to year, depending on winter rainfall, the low summer levels range much less, most of them lie between 15.2 and 15.6 m. The main reason is the density of the stream network and the proximity and good hydraulic contact of surface water that limits the lowering of the water table.

2.5. Aquifer recharge

The main source of water in the aquifer system is recharge from precipitation. Aquifer recharge is estimated using data series of the Ukkel meteorological station, obtained from the global historical climatology network (GHCN). Although Ukkel lies outside the modelled region (around 30 km from the southwest border), it is by far

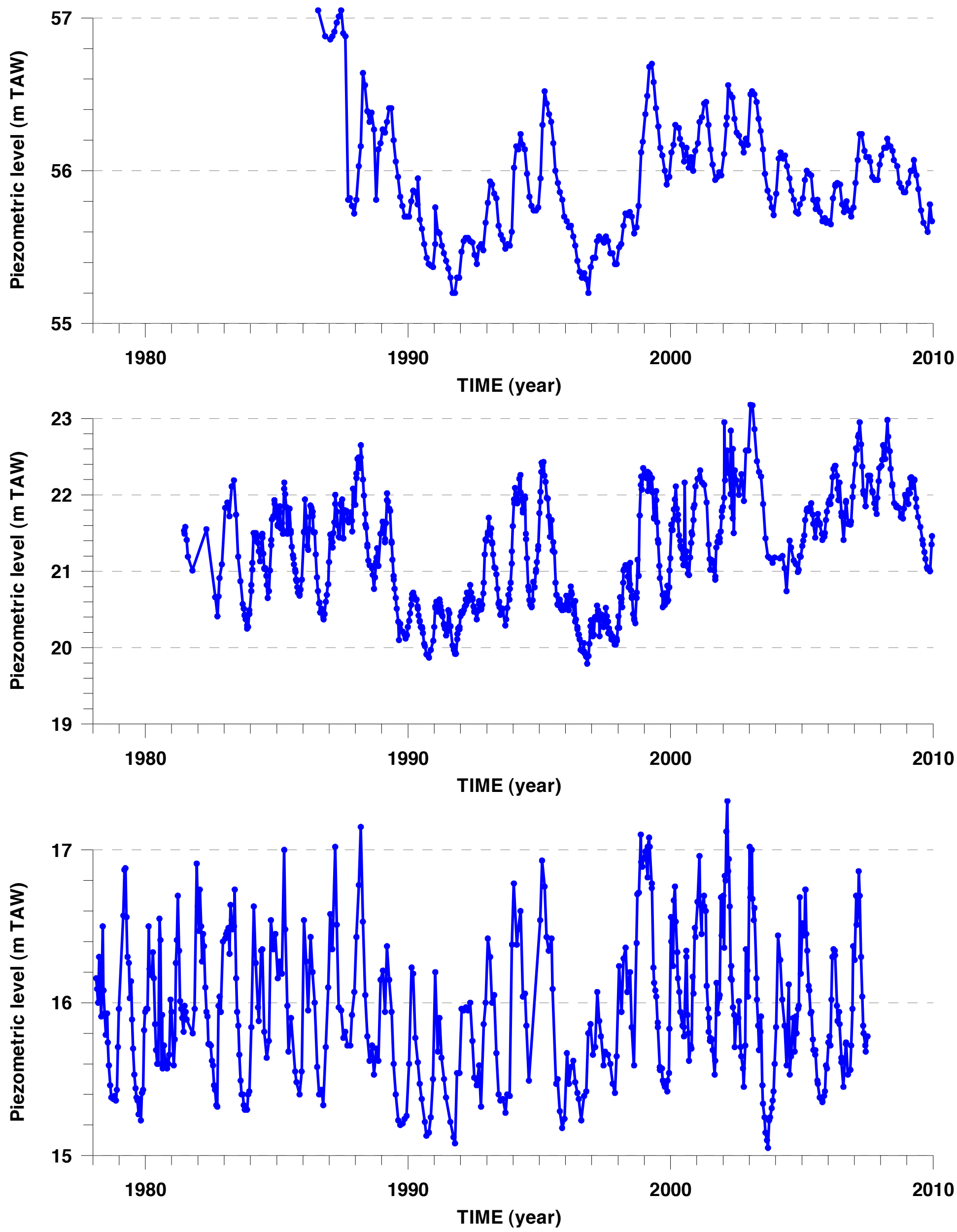

Figure 4. Selected piezometric time series.

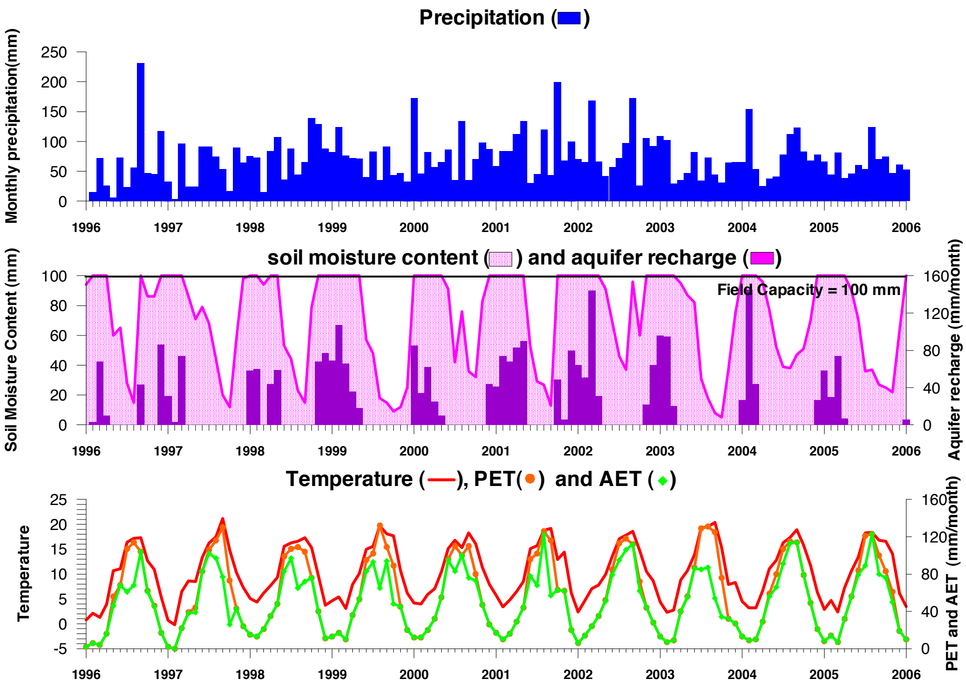

Figure 5. Result of the soil moisture water balance model for the period 1996-2005.

the longest series available in Belgium, and thus allowed for a much longer simulation period. Recharge is computed with a soil moisture balance (SMB) approach as introduced by Thornthwaite and Mather (1955) and applied by Mather (1978 and 1979), using precipitation and temperature data. Monthly PET (potential evapotranspiration in mm/day) is estimated from temperature using the Thornthwaite (1948) equation:

$$PET_i = 16\left(\frac{L_i}{12}\right)\left(\frac{N_i}{30}\right)\left(\frac{10T_a}{I}\right)^a$$

with: T_a is the average daily temperature of the month being calculated (°C), N_i is the number of days in the month I being calculated (-), L_i is the average day length of the month I being calculated (hours).

$\alpha = (6.75\ 10^{-7})\ I^3 - (7.71\ 10^{-5})\ I^2 + (1.79\ 10^{-2})\ I + 0.49$
I = heat index defined as (-):

$$I = \sum_{i=1}^{12}\left(\frac{T_{ai}}{5}\right)^{1.514}$$

This approach has the benefit that it requires only a single meteorological parameter: temperature. The calculations for the soil moisture balance are done based on monthly totals (for rainfall) and averages (for temperature). The program WATBUG (Willmott, 1977) was used for this purpose. Results of the SMB model include potential evapotranspiration, actual evapotranspiration, deficit, soil water content and aquifer recharge. The results of the SMB model for the last decade of the simulation period (1996-2005) are presented graphically as an example (Fig 5). In the upper graph the measured monthly rainfall is given. The middle graph shows the calculated aquifer recharge (in mm/month) as a bar chart and the soil moisture content (hatched graph). The maximal value at field capacity was estimated at 100 mm. The lower graph shows measured monthly averaged temperatures and the values of monthly potential (PET) and actual (AET) evapotranspiration calculated with the meteorological data of Ukkel. In Fig 5 it can easily be recognized that aquifer recharge is limited to the winter months. The main controlling factor is that PET and AET are very low in wintertime while they usually exceed precipitation in summer months. Intermonthly variation of rainfall is small but interyearly variation is large, while intermonthly variation in PET and AET values is large, but interannual variation is small.

Average recharge rate over the whole period 1833-2005 is 239 mm/year, with an average precipitation of 790 mm/year. Interyearly variation can be seen in Fig 6. A longer time cyclicity can be recognized in the 10 year simple running average, which is defined as the simple arithmetic average of the last ten values (including the year itself). Drier decades can be found around 1860-1870, 1900-1910 and 1950-1960. They apparently occur with a periodicity of 40 to 50 years. Wetter periods are interspersed with the wettest decade recently between 1980 and 1990. In the dry periods average recharge is a bit less than 200 mm/year, while the wet decades have average rates close to 300 mm, around 50% higher than de dry period average value.

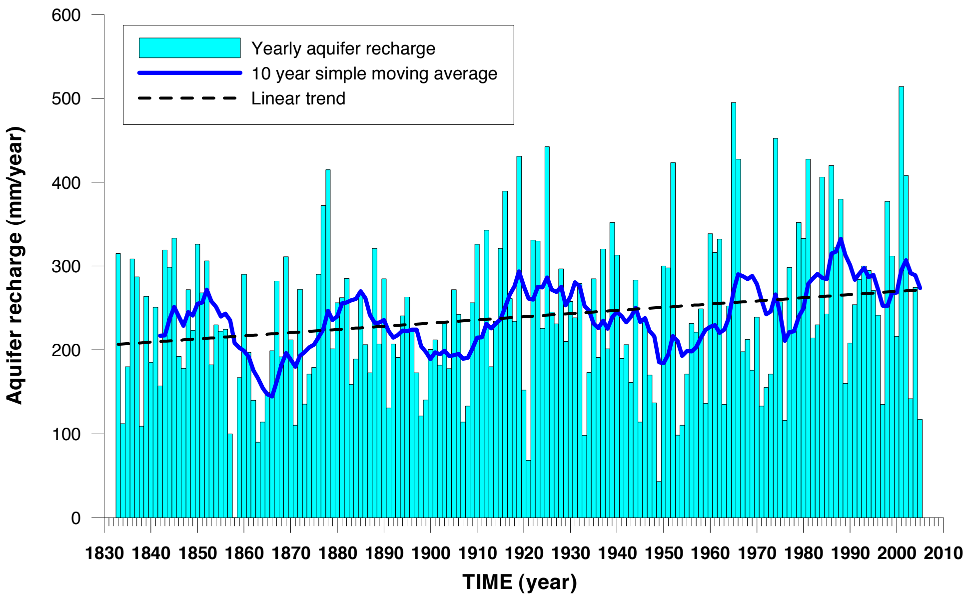

Figure 6. Calculated yearly aquifer recharge 1833-2005, 10 year simple moving average and linear trendline.

A linear correlation analysis shows a slight increasing trend with an increase of ca 3.8 mm a decade, but the coefficient of determination (r squared) is only 0.04.

3. Groundwater flow model

The groundwater flow model used for the simulations is based on the model conceptualized and implemented by Coetsiers (2007). It uses the MODFLOW simulator (Harbaugh and Mc Donald , 1996). The model domain is discretised using a rectangular grid with a uniform spacing of 500 m. The whole modelled region includes an area of 100 km by 75 km, subdivided in 200 columns and 150 rows. The south border (Fig 1) lies at the outcrop of the underlying Boom Clay (Formation of Rupel), which is considered here as the basement of the model. Flow through this basement is considered negligible compared to the fluxes in the overlying aquifer. The south border is therefore a no flow boundary. The west boundary of the model is located at the Scheldt river. As the Neogene aquifer is phreatic there, the river can be considered as a hydraulic boundary. The river is incorporated in the model with constants head cells. The east model border lies at or near to a flowline as can be seen on the piezometric map (Fig 3). This border is treated as a no flow boundary. The north border lies in The Netherlands and its position is chosen rather arbitrarily. At this border constant heads were entered into the model. The levels were estimated based on the topographic elevations, derived from a digital terrain model (DTM). Hydraulic properties are based on results of pumping tests that were done in the past in some of the exploitation well fields. Average hydraulic conductivity for the lower part of the Neogene aquifer is 10.1 m/d, for the upper part 15.3 m/d. The cover sands have an average value of 12.9 m/d (Meyus et al., 2004)

The main groundwater source is recharge from rainfall. The model uses a uniform distribution of aquifer recharge, with a long term average of 239 mm/year. It should be kept in mind that spatial variations in recharge pattern will occur, related to differences in soil texture, land use and vegetation, but these are smoothed by the use of a single uniform value. First a steady state run with the long term average recharge of 239 mm/year was done. The calculated piezometric levels were used as initial heads for a transient simulation starting in 1833 with monthly stress periods. The first years of this simulation has to seen as a spinup period.In the transient simulation run, recharge is defined with monthly stress periods. These were calculated with the soil moisture balance model (see 2.5)

The main groundwater sink mechanisms are groundwater discharge into the river system and groundwater flow over the outer model boundaries. Groundwater exploitation is not considered in the simulation 1833-2005 as the objective was to investigate natural variations in the aquifer hydrodynamics. Cross border flow is only relevant over the north border as the other borders are no flow boundaries. The south border is located at the northern outcrop of the Boom Clay. Inflow is considered here as negligible. The eastern border is chosen along a flowline and is therefore by definition a no flow boundary. Along the western border, groundwater is drained towards the Scheldt river, which is included as a river. Baseflow to the river system has been included in the model with a diffuse approach. In the Nete Valley the number of rivers, streams and small ditches is too large to consider them as individual river entities. Instead in each model cell a drainage level is defined, which corresponds with the average bottom level that the river reaches. This bottom level is derived from a DTM, not from in situ field measurements. In each model cell, also outside the Nete valley, the possible drainage is defined using the MODFLOW DRAIN module. If calculated water table is lower than the predefined drain level, the discharge flux is zero:

$WT <= LDRN \rightarrow Q_{DRN} = 0$

With WT = calculated water table level in a cell, LDRN = predefined drainage level in a cell, and Q_{DRN} = amount of water drained from the cell.

If the water table is above the defined drainage level, the discharge flux is proportional to the head difference:

$WT > LDRN \rightarrow Q_{DRN} = CF*(WT-LDRN)$

The constant CF (units: m^2/day) defines the hydraulic contact between the aquifer and the drain mechanism.

By combination of the recharge and drainage modules of the MODFLOW model, the extension of the recharge and discharge areas can be calculated. From the model's output file, the discharge flux of each model cell in the phreatic layer can be retrieved. If the discharge rate is zero, only recharge takes places in this cell and it belongs to the "recharge" area. If there exists a discharge flux, the model cell belongs to the "discharge" area, although it should be realised that also in these cells rainwater is entering the aquifer system. By accounting for the total extent of the discharge area, where groundwater discharges into the river network, total and average seepage fluxes can be calculated.

First a steady state run with exploitation data and boundary conditions for 2003 was performed (Fig 7), to check the ability of the model to reproduce the recently measured piezometric distribution (Fig 3). The correspondence of this calibration was sufficient to accept the entered parameterisation. Model calculated baseflow in the subbasin of the river "Kleine Nete" was compared with estimation of baseflow derived from measured river flow rates at a monitoring station in Grobbendonk (Fig 8) (HIC, 2004) . As an approximation, the minimum daily flowrate in each month was considered as constituted largely of baseflow alone. Deviations between modelled and measured values for the year 2003 were on the order of ca 10%, which is probably well within the limits of accuracy of the simple approximation of river baseflow.

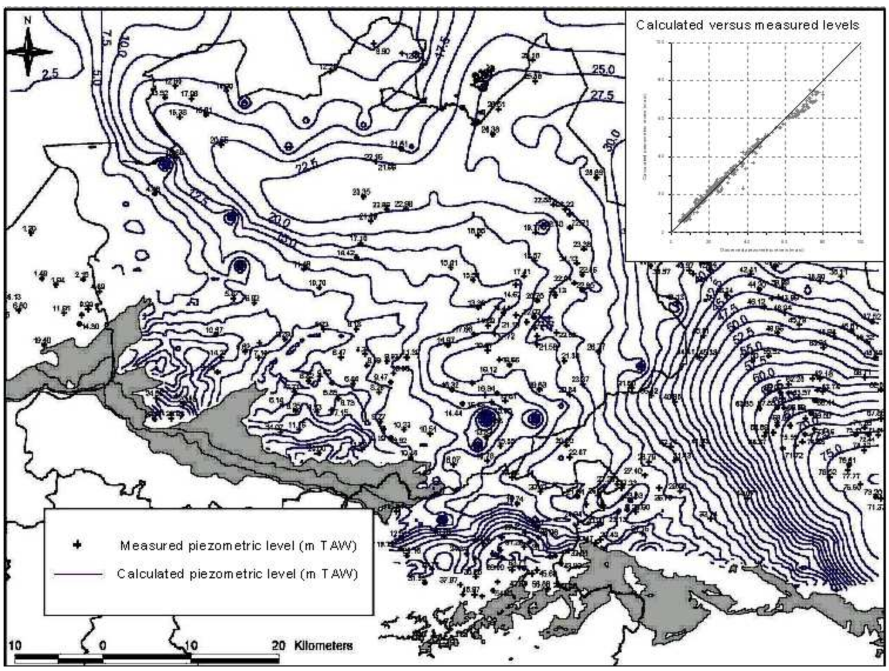

Figure 7. Results of the calibration run.

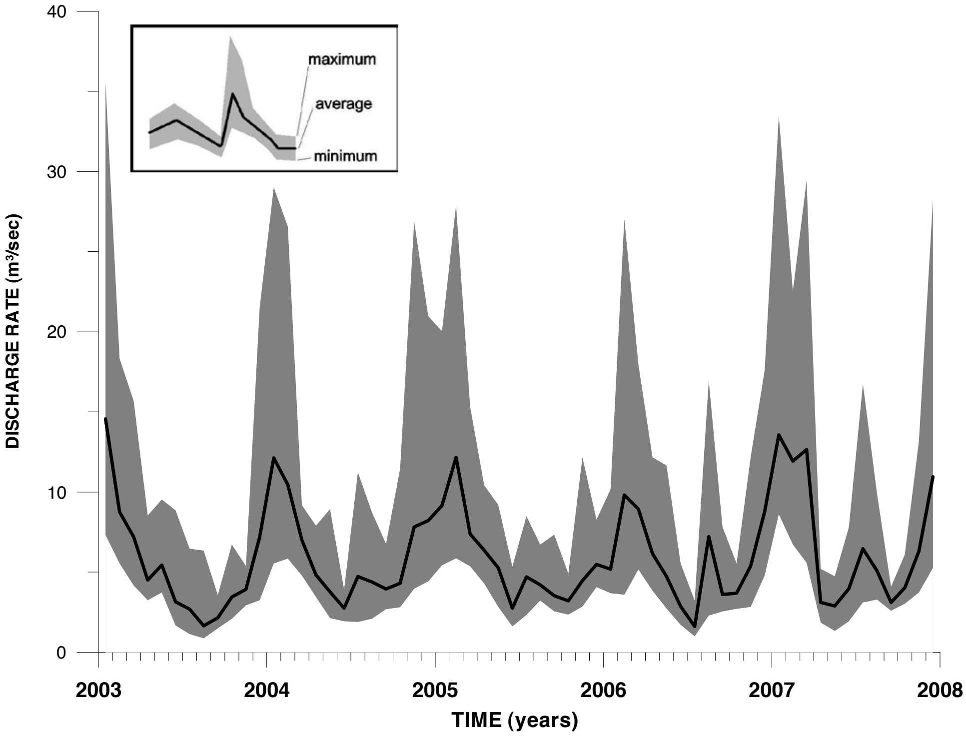

Figure 8. Monthly minimum, maximum and average daily flow rate on the Kleine Nete river in Grobbendonk.

The final model run uses a transient flow regime with stress period length of one month. Each stress period is divided into 5 time steps. Initial heads for the run were obtained from the steady state model run with the average recharge value of 239 mm/year.

4. Simulation results

The 1833-2005 model run simulates the situation without groundwater extraction, representing the natural, pristine state of the aquifer. Output of the model can be represented in the form of graphs and maps. Results indicate that, besides the yearly seasonal variation in the main groundwater flow cycle, also a longer multi-year cyclicity occurs. This is mainly in the regions where the aquifer is being recharged. Both simulated and measured series of the last two or three decades show this behaviour. The model outputs monthly distributions of hydraulic heads, fluxes from or to defined boundary conditions and flow rates inside the model domain (between model cells and layers).

The long term evolution of piezometric levels can be illustrated by time series graphs of the calculated hydraulic head at two selected locations (Fig 9). The levels are plotted as deviations (in m) from the average of the total series. The main trend is indicated by the two year running average which attenuates seasonal fluctuations. In the upper graph the evolution on the Campine Plateau is given. Here the main variations in level occur on a multi-year time scale and the yearly seasonal fluctuations are less important than the longer cyclicity. The three main dry periods of 1860-1870, 1900-1910 and 1950-1960 are easily identifiable with levels that are 1 to 2 meters below average. Since the early 1980s, levels were almost always above the long term average of the whole series. The lower graph shows a representative location in the Nete valley basin. In this series, variation can be largely attributed to the seasonal cycle exclusively.

As can be seen on the time graphs, the maximum range of variation over the years is a few meters, between 3 and 4 meters on the Campine Plateau. This is only a small fraction of the total range in

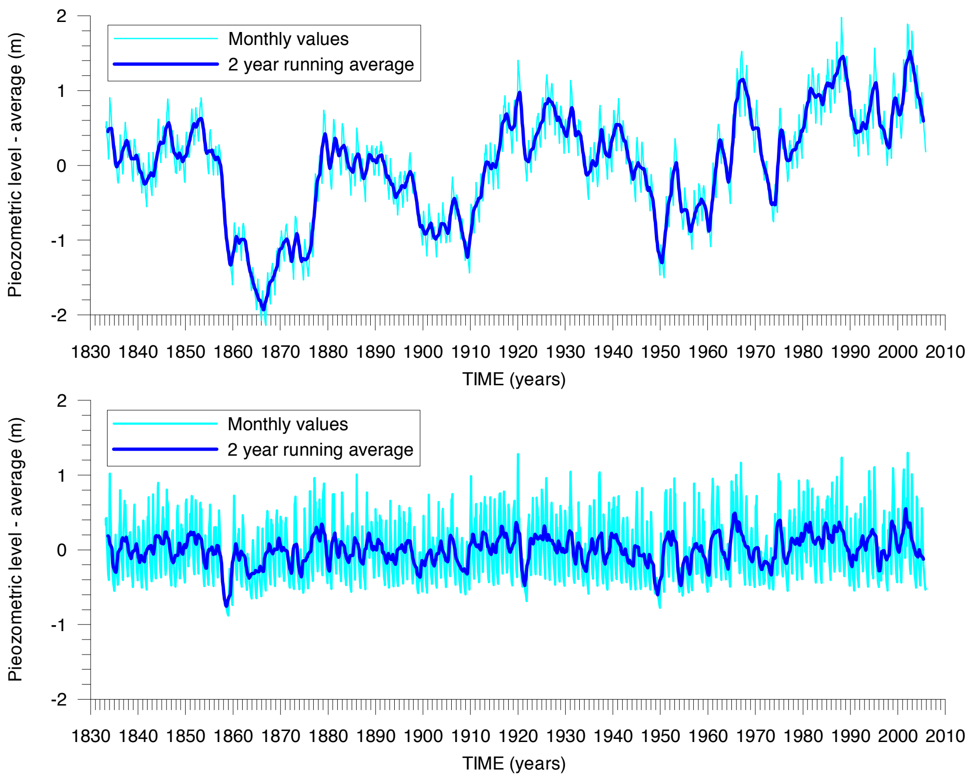

Figure 9. Calculated evolution of piezometric levels at 2 selected locations: on the Campine plateau (upper graph) and in the Nete basin (lower graph).

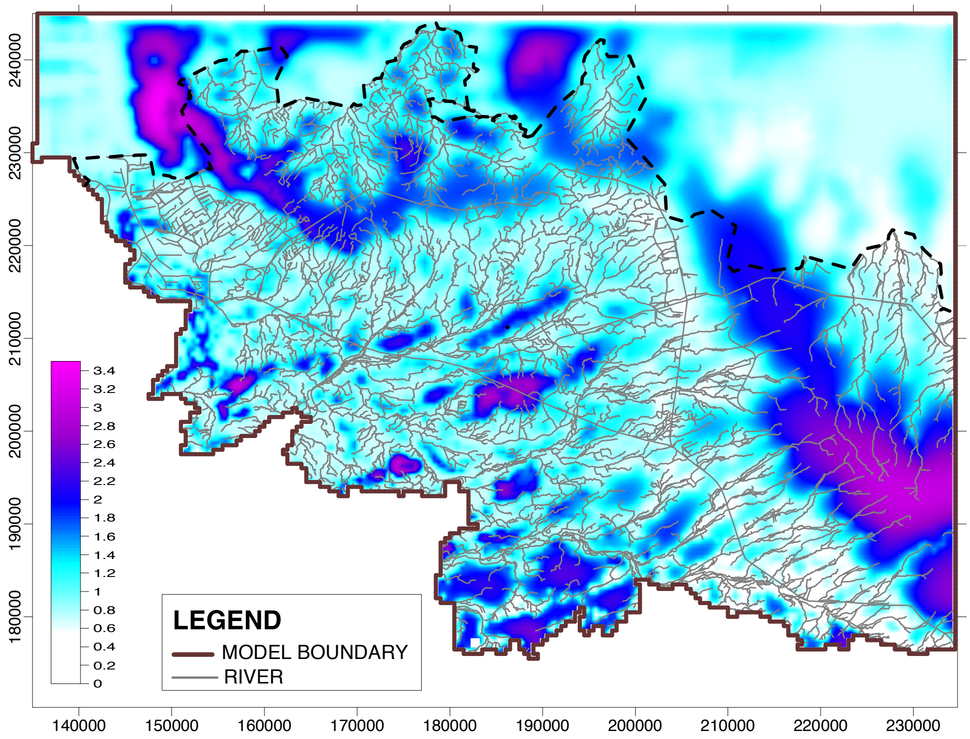

Figure 10. Difference in piezometric level of the winter 2002 maximum compared to the long term average levels.

piezometric levels in the whole aquifer system, as there is at least 50 m difference beween the Campine Plateau and the Scheldt river in the west. Consequently, the general flow pattern will not fundamentally change between seasons or between extremely dry and wet years. Deviations from the long term average can be greater on the Campine Plateau and Campine Cuesta as can be seen on the map of Fig 10. For the wet winter situation of January 2002, the piezometric levels are compared with the average levels of the whole simulated period. They can be as much as 3 m above normal.

As a result of the variations in hydraulic heads, the intensity of the groundwater cycles will vary over time. This is especially important for the upward groundwater flow in the Nete Valley, where deep Neogene groundwater is upwelling and is drained away by the river network. Here both size of the region as the intensity of the discharge flow will vary with time (Fig 11). Total calculated discharge into the river shows a strong seasonal behavior with low summer values, nearly always between 15 and 20 Mm^3/month. There is however a rather small interyearly variation on the summer minimum baseflow. Maximum baseflow appears in wintertime, but varies from year to year over a wide range between around 40 and 140 Mm^3/month. Interannual variability is much larger than for the summer lows. The same conclusion is valid for the extension of the discharge region; in summer time discharge has the smallest extension, between 1000 and 1500 km^2, and is limited to the main river valleys. Many small ditches can dry up in a dry summer period. In winter time, the discharge region's size can triple to more than 3000 km^2. The average discharge flux intensity can be calculated by dividing the total baseflow for each month by the extension over which it occurs. A cross-plot of the baseflow and the average discharge intensity as function of the extension of the discharge region (Fig 12) shows that the relation between total baseflow and extension of the seepage region is non-linear, but rather exponential. This means that, as its areal size increases, also the intensity will increase. In dry summer months, average seepage flux is around 1 mm/day, but it increases to 2.5 mm/day in wet winter months. The exponential fits indicated in Fig 12 have a coefficient of determination (r squared) of 0.89 for the relationship extension-discharge intensity and 0.98 for the relation extension-baseflow.

5. Conclusions

The natural variations of the main groundwater cycle characteristics in the Neogene aquifer system in north-east Belgium have been

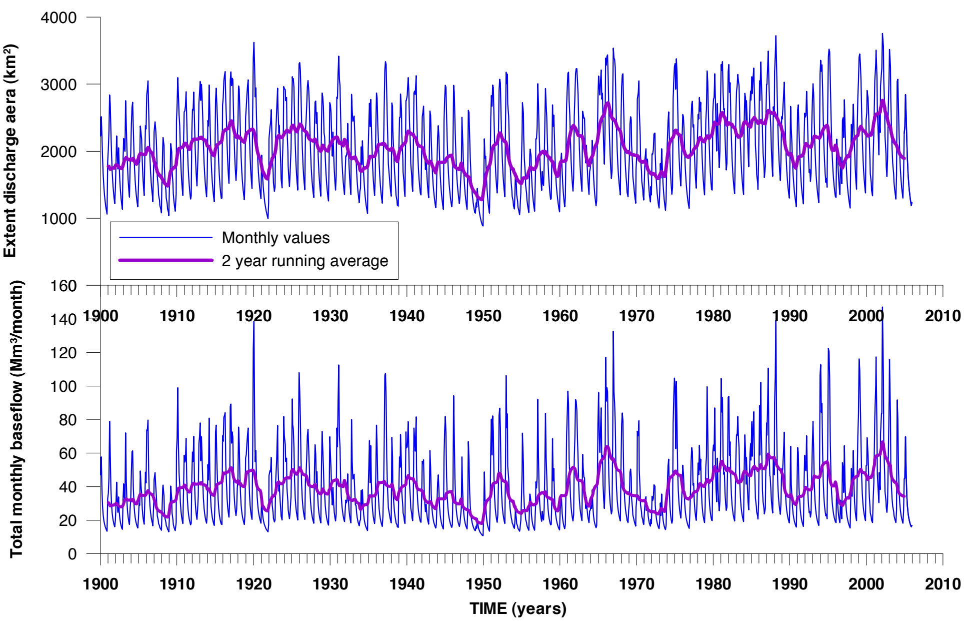

Figure 11. Evolution of the monthly baseflow and the extent of the discharge area.

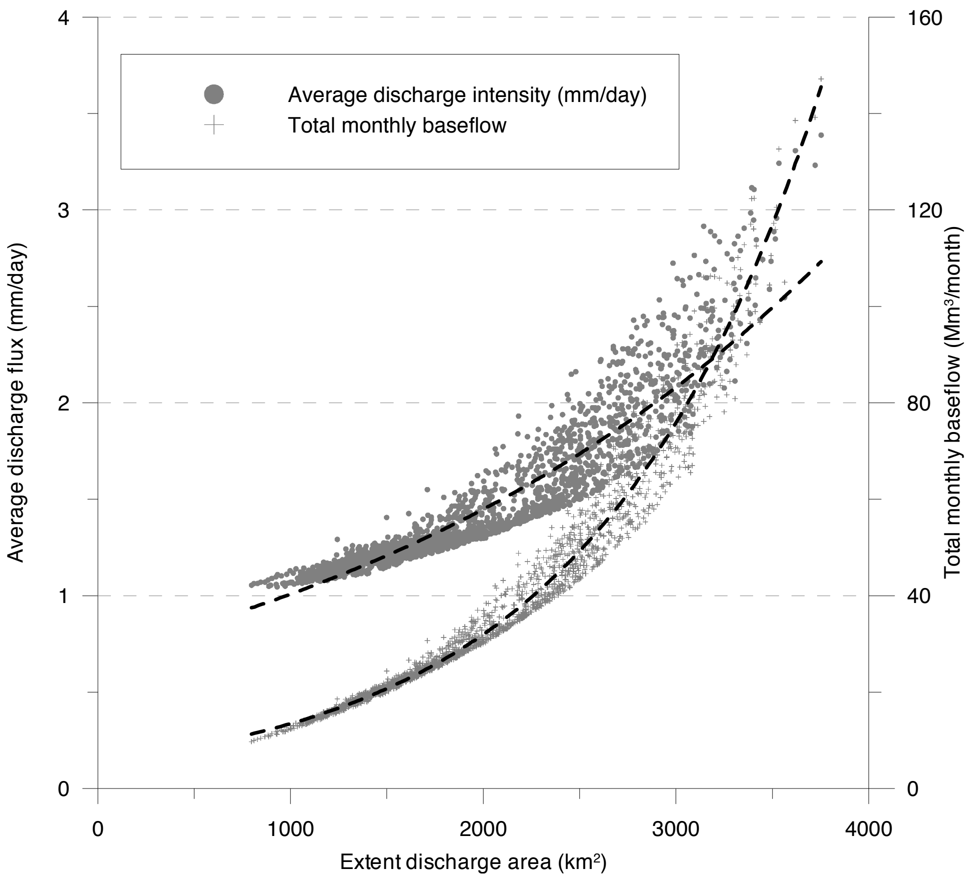

Figure 12. Relation between the model calculated monthly baseflow and the extent of the discharge area.

investigated using a numerical model. Using the longest available meteorological series in Belgium, from the Ukkel station, a transient flow simulation of a 173 year long period from 1833 till 2005 was performed. Aquifer recharge is calculated on a monthly basis with a soil moisture balance model and shows the presence of three drier decades which occur with intervals of 40 to 50 years: in 1850-1860, in 1900-1910 and in 1950-1960. The model simulation uses monthly stress periods with changing but spatially uniform recharge rates and uses a combination of the recharge and drainage capabilities of the simulator to calculate extension and intensity of the discharge component of the main groundwater cycle. The model results indicate the existence of a multi-year cyclicity in the piezometric levels which is more pronounced in the topographic higher areas where most of the aquifer recharge takes place. The multi-year variations are here more important than the yearly seasonal winter-summer cycle. Unfortunately, no very long observation series of piezometric levels exist, to compare with model results. The oldest series started back near the end of the 1970s. Discharge of deep Neogene groundwater in the Nete Valley shows a strong seasonal dependency for intensity and areal extent, and the interannual variation of peak winter maxima is much larger than for summer lows which are more constant.

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