Weichselian and Holocene climate history reflected in temperatures in the upper crust of the Netherlands

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Abstract

In the Netherlands the present-day thermal gradient in the shallow subsurface (i.e. the upper few 100 m), is around 20°C km^{-1}, whereas at depths between 0.5 and 3 km it is $\sim$33°C km^{-1}. This large contrast in the gradient between shallow and deeper parts of the subsurface occurs throughout the country and cannot be explained by either systematic thermal property changes with depth or the depositional setting of the region. In this paper we use a 1D thermal model for the crust and demonstrate that this observed temperature-depth trend most likely reflects a transient condition inherited from past climate change. It is shown that the prolonged cold period during the Weichselian ($\sim$110–10 kya) and the subsequent warmer conditions during the Holocene account for the increase in the thermal gradient with depth. Moreover, we demonstrate that thermal history further back in time still influences the present-day subsurface temperature. Geothermal climate-change influences on these long time scales have not been documented before for the Netherlands.

Introduction

Knowledge about the thermal state of the crust is important when aiming for a thorough reconstruction of the geological history of a region, and also when calculating maturation of hydrocarbon source rocks or for successful geothermal prospecting. Data from boreholes provide a useful but local estimate of subsurface temperatures and the thermal state of the shallow crust, and are often used as constraints when estimating heat flow (e.g. Agemar et al., 2012; Bonté et al., 2012) or lithospheric strength (e.g. Tesauro et al., 2007) at deeper levels.

Changes in the surface temperature affect the temperature at depth. The magnitude, duration and penetration depth of this effect depend on the amplitude and especially the wavelength of the variations in the surface temperature. For glacial–interglacial cycles of several hundreds of thousands of years, the penetration depth of the detectable effect extends to a few kilometres (Pollack & Huang, 2000; Bodri & Cermak, 2007; Huang et al., 2008; Majorowicz & Šafanda, 2008;

Kukkonen et al., 2011; Majorowicz & Wybraniec, 2011). The influence of these paleoclimatic variations is often neglected when heat flow at deeper crustal or lithospheric levels is estimated from temperature measurements in the shallow crust. However, the inclusion of surface temperature variations on the timescale of glacial–interglacial cycles yields heatflow corrections up to 20 mW/m^2 in Europe (e.g. Majorowicz & Šafanda, 2008; Majorowicz & Wybraniec, 2011), underlining the fact that the thermal state of the crust is a transient phenomenon.

The average present-day thermal gradient in the Netherlands is around 20°C km^{-1} in the shallow subsurface (i.e. $\sim$70–500 m) (van Dalfsen, 1981; Bense & Kooi, 2004; Karg et al., 2004; Kooi, 2008), whereas at larger depths (500 m to 3 km) it is between 30 and 35°C km^{-1} (e.g. Ramaekers, 1991; van Balen et al., 2000, 2002; Luijendijk et al., 2011; Bonté et al. 2012). This is a regional phenomenon, which may also imply that in the Netherlands the thermal field is in a transient state of heating as a response to the increase in surface temperature since

the last glacial maximum: at the end of the Weichselien (∼18 kya), yearly average surface temperatures were below –5°C.

In addition to crustal heat flow studies, borehole temperature data have been used extensively to reconstruct ground surface temperature history as a proxy for climate change. Long-term studies (i.e. up to 100 kyrs) have been carried out by, for example, Kukkonen & Šafanda (1996), Kukkonen et al. (1998), Šafanda & Rajver (2001), Kukkonen & Jõeleht (2003), Šafanda et al. (2004) and Kukkonen et al. (2011). These studies were mostly based on one or two boreholes and all focused on the Baltic Shield and the East European Platform. Data from boreholes capturing periods up to 1000 years ago have been used to make surface temperature reconstructions averaged over wider regions, especially in the Northern Hemisphere (e.g. Pollack et al. 1998; Harris & Chapman, 2005).

Deltaic areas like the Netherlands have always been avoided in studies on paleoclimatic signals in subsurface temperatures because of disturbing effects of shallow processes like groundwater flow and land use (e.g. Kooi, 2008). The Netherlands are exceptional, however, in the sense that a very large amount of data is available in which these local effects can be largely suppressed. Although this average geotherm is not sufficiently detailed to make a reconstruction of the paleo-climate from the subsurface geotherm, it allows for the investigation of whether the decrease in thermal gradient that appears in the average geotherm beneath the Netherlands is in concert with the estimated paleoclimatic signal as derived from climate proxy data.

Here we analyse the effect of changes in surface temperatures on timescales of hundreds of kyrs on the geothermal gradient in the Netherlands, both in time and as a function of depth, using a numerical model that calculates heat transfer through the crust. We compare our findings to published heat-flow data for the Netherlands.

The model

We used an implicit finite difference model that calculates changes in temperature in the crust due to conduction. The model solves the heat equation:

$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\kappa \frac{\partial T}{\partial z}\right) + \frac{A}{\rho C} \qquad (1)$$

where T is temperature (°C), t is time (s), κ is thermal diffusivity ($m^2\,s^{-1}$), A is heat production ($W\,m^{-3}$), ρ is density ($kg\,m^{-3}$) and C is specific heat ($J\,°C^{-1}\,kg^{-1}$).

Here the thermal properties are regarded as bulk properties of the crust, i.e. of rock matrix and pore fluids together. Boundary conditions are a prescribed temperature that can change with time at the surface and a constant temperature at the base of the model (base lithosphere). The start condition of the model is a steady-state geotherm resulting from the initial boundary conditions. Starting from this steady-state situation we change the surface temperature to observe the effects on the geotherm in depth and time.

Comparison with analytical solutions: the influence of Milankovitch cycles

The magnitude, duration and penetration depth of the effect of changes in the surface temperature depend on its amplitude and especially the wavelength of the temperature variations. The analytical solution for a sinusoidal variation in surface temperature, for the case without spatial variations in the thermal conductivity and in the absence of heat production, is (e.g. Furbish, 1997):

$$T(z,t,\lambda) = T_a \exp\left(-z\sqrt{\frac{\pi}{\kappa \cdot \lambda}}\right) \sin\left(\frac{2\pi}{\lambda} \cdot t - z\sqrt{\frac{\pi}{\kappa \cdot \lambda}}\right) \qquad (2)$$

where T is temperature (°C), t is time (s), z is depth (m), λ is the period of the temperature signal (s), T_a is the amplitude of the temperature change at the surface and κ is thermal diffusivity ($m^2 s^{-1}$).

Using this equation we can compare the effects of the three major climatic (Milankovitch) cycles, with periods of 23, 42 and 100 kyr (e.g. Hays et al., 1976). The equation shows that:

(a) The decrease in the amplitude of the temperature variation with depth is larger when the frequency of the temperature cycle is higher: in that case the temperature at depth has less time to adapt to the changes. As can be calculated from eqn 2, when assuming a thermal diffusivity κ of 10^{-6} $m^2 s^{-1}$, the amplitude reduces by 63% every kilometre for the 100 kyr cycle, whereas for the 42 kyr cycle the amplitude reduction is 79% per kilometre and for the 23 kyr cycle it is 88% per kilometre.

(b) There is a phase change between the variation in surface temperature and the variations at depth. For the 100 kyr cycle, the response has a delay of 15.9 kyr per kilometre extra depth. At a depth of 3.1 km, the oscillation has thus been reversed, implying that a minimum surface temperature coincides with a maximum temperature at a depth of 3.1 km and vice versa. For the 42 kyr cycle, the delay of the response is 10.3 kyr per kilometre extra depth and the reversal is at a depth of 2 km, and for the 23 kyr cycle, the phase delay is 7.6 kyr km^{-1} and the reversal occurs already at a depth of 1.5 km.

These findings are in accordance with the numerical solutions for these cyclic surface temperature changes, which are depicted for the 23 kyr and 42 kyr period in Fig. 1. In addition, Fig. 2 shows both the modelling result and the analytical solution in one graph for a surface temperature signal with a scaled amplitude of 1, a thermal diffusivity κ of 10^{-6} $m^2 s^{-1}$ and a period of 100 kyr.

Ultimately, the combination of the reduction in amplitude and the delay of the phase of the oscillation with depth results in variations in the thermal gradient (Fig. 3). The figure shows that the shallow thermal gradient is reduced when the surface temperature is increasing, and rises when the surface temperature is decreasing. When the surface temperature is at its
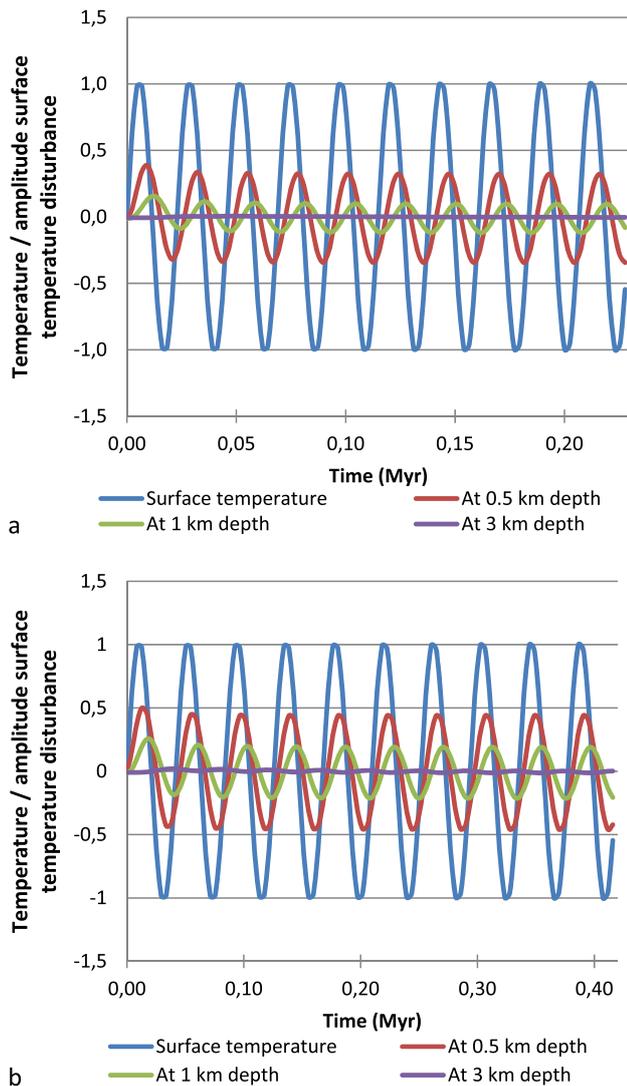
Fig. 1. *The relative response to cyclic surface temperature variations (a) with a period of 23 kyr and (b) with a period of 42 kyr. The amplitudes of the cycles are scaled to 1.*

minimum, the shallow gradient is at its maximum, and vice versa. Moreover, as demonstrated by Fig. 3, the geotherm does not reach its steady-state configuration at any moment during the oscillations, indicating that a transient temperature field is not an exception but the rule.

In reality, Milankovich cycles act simultaneously, causing the thermal phases to offset or amplify each other. Furthermore, other processes also have an influence on the climate, therefore, in order to investigate the effect of the climate on the subsurface temperature in the Netherlands, we used climate reconstructions which were available for up to 130 kya, and we tested the effect of several end member scenarios for the surface temperature history before 130 kya.

The thermal state of the Netherlands: variations with depth

The average geothermal gradient in the Netherlands at the depth interval from 0.5 to 3 km is between 31 and 33°C km^{-1}

(Ramaekers, 1991). Regional values have been established, for example for the Roer Valley Graben, where the temperatures at depths of 2500 m below the surface range around an average of 98°C, implying an average geothermal gradient of 35°C km^{-1} up to this depth (Luijendijk et al., 2011, Fig. 4), and for the West Netherlands Basin, where numerical modelling based on well data resulted in average geothermal gradients of 30± 2°C km^{-1} in the upper 5 km of this region (van Balen et al., 2000). The most recent analysis of the temperature of the entire onshore Dutch underground, based on a dataset of 1241 measurements in 437 wells, yielded an average thermal gradient of 31.3°C km^{-1} (Bonté et al. 2012, Fig. 5), combined with a surface temperature of 10°C (Bonté et al., 2012; see also van Dalfsen, 1983). Interestingly, a much lower geothermal gradient is documented in the shallow (<200 m depth) subsurface in the central part of the Netherlands (Kooi, 2008), in the southeastern part of the Netherlands (Bense & Kooi, 2004), and in the Dutch and German parts of the Lower Rhine Embayment (Karg et al., 2004).
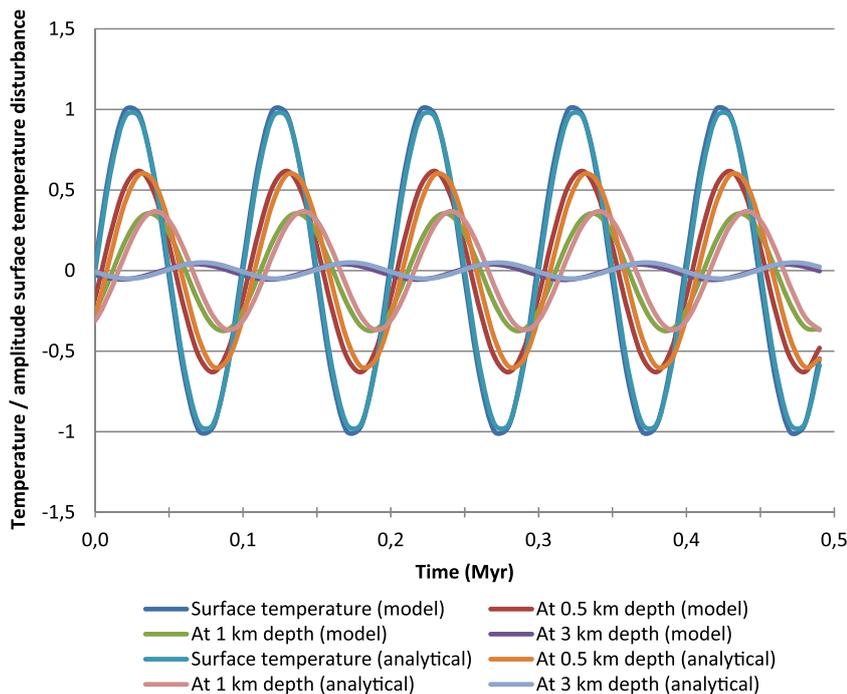
Fig. 2. The relative response to cyclic surface temperature variations with a period of 100 kyr and an amplitude scaled to 1, calculated both analytically and numerically.

Kooi (2008) compared repeated borehole temperature measurements in the central parts of the Netherlands that were recorded some three decades apart. He found gradients around 20°C km^{-1} to depths of about 200 m in measurements from the late 1970s. Similar gradients to depths of 400 m are common throughout the Netherlands (van Dalfsen, 1981). In 2006, i.e. 30 years later, warming and reduced or even negative temperature gradients were observed at depths shallower than 70 m, whereas temperatures and gradients were found to be unaltered at depths larger than 70 m. The shallow subsurface warming was found to be consistent with documented increased surface air temperatures in the Netherlands since about 1980. This climate warming is too recent to have influenced temperatures at depths in excess of 70 m, explaining the relatively stable gradients of around 20°C km^{-1}.

Below, we investigate whether the relatively low gradient of 20°C km^{-1} in the depth interval of 70 to 400 m, when compared to the interval from 0.5 to 3 km, can be explained by the temperature increase at the surface in the more distant past.

Input of the model

The existence of thick sediment layers in the Netherlands has a pronounced influence on its subsurface temperature structure (Bonté et al., 2012): sediments act as an insulating layer for the crust. The sediment layering shows large spatial variations in the Netherlands, mostly due to structures and different subsidence rates. Since we use an average geotherm to constrain our models, and since our aim is to isolate the influence of the surface temperature variations from other

effects, we choose to ignore this sediment blanketing effect. As shown by Bonté et al. (2012), this entails that the temperatures as observed in boreholes in the upper few kilometres of the subsurface can only be explained by combining an extremely low value for the thickness of the lithosphere with an unrealistic high value for the heat production in the upper crust.

The average thickness of the lithosphere beneath the Netherlands is hard to estimate due to uncertainties in composition and temperature at larger depths. Values in the literature vary from 90 km (Cloetingh et al., 2010) to more than 150 km (Goes et al., 2000). In order to obtain present-day values for the geothermal gradient that are comparable with those measured in the Netherlands, we had to adopt a rather thin lithosphere of 90 km and a heat production of 3.6 μW/m^3 in the upper 10 km of the crust. The resulting steady-state geotherm is shown in Fig. 6. Its gradient decreases with depth due to the effect of the heat production, from 32.2°C km^{-1} at the surface, via 28.7°C km^{-1} at a depth of 3 km, to 22.1°C km^{-1} at a depth of 8 km.

Starting from this geotherm, we have applied the surface temperature proposed by Luijendijk et al. (2011, see Fig. 7), to represent the thermal evolution in the Netherlands during the past 130 kyr. This surface temperature history was estimated from climate proxy data such as pollen, fossil beetles (Coleoptera), plant macroremains and periglacial indicators in northwest and central Europe (Zagwijn, 1996; Huijzer & Vandenberghe, 1998; Caspers & Freund, 2001), and in the northeastern part of the Netherlands (van Gijssel, 1995).
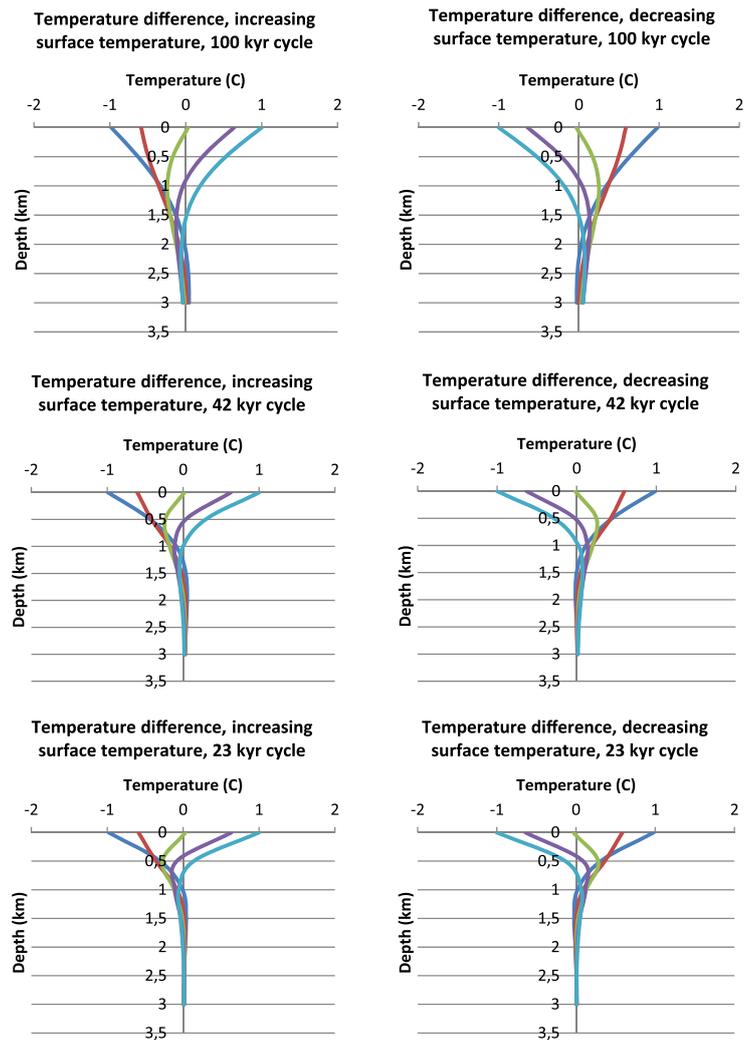
Fig. 3. *The downward propagation of changes in surface tempera-*
ture, with a dimensionless (scaled) amplitude of 1 and periods of
100 kyr (upper panels), 42 kyr (central panels) and 23 kyr (lower
panels), separated in phases of increasing surface temperature
(left panels) and decreasing surface temperatures (right panels).
Dark blue, red, green, purple and light blue curves represent suc-
cessive moments, with time steps of 10 kyr (for the 100 kyr cyle),
4.2 kyr (for the 42 kyr cycle) or 2.3 kyr (for the 23 kyr cycle).

Results

Fig. 8 shows the calculated temperature evolution with time
at depths of 0.5, 1.0 and 2.0 km. In Fig. 8a we started with
a steady-state geotherm with a surface temperature of –6.5°C
and calculated the effect of the surface heating that had
occurred since 20 kya (see Fig. 7). In this model the surface
gradient decreases from 32.2°C km^{-1} at 20 kya to 14.4°C km^{-1} at
present.

Fig. 8b shows the thermal evolution for the entire 130 kyr
period of climate changes as depicted in Fig. 7. Again, the influence
of the last cooling event is still clearly visible in the calculated pres-
ent-day geotherm, although the effect is damped: The present-day
geotherm resulting from this model run has a surface gradient of
20.6°C km^{-1}. Since the model started from a surface temperature
of 12°C, which cooled and heated up again in a period of ~100
kyr, a new steady state during the cold period was never reached,
therefore when the surface temperature increased to its current
value, the thermal field could recover much faster than in the

scenario with instantaneous cooling starting from a steady-state
geotherm with a low surface temperature at 20 kya.

The present-day geotherm resulting from this model run has
a surface gradient of 20.6°C km^{-1}, a gradient of 32.1°C km^{-1} at
a depth of 3 km and a gradient of 21.9°C km^{-1} at a depth of
8 km. These values are comparable to the gradients found by
Kooi (2008) for the upper 200 m, and by Bonté et al. (2012)
and Luijendijk et al. (2011) for the upper 3 km. Although the
absolute values are dependent on the start geotherm, which
is not very well known, we can conclude that the surface
temperature history in the Netherlands has been able to cause
a significant increase in the gradient with depth.

Our model results show that the increase in the thermal
gradient with depth in the Netherlands can be explained as
a remnant from the relatively cold period that lasted until
18 kya. Furthermore, it demonstrates that the thermal history
before 18 kya still has a profound influence on the subsurface
thermal field: the climate warming that started 18 kya inter-
rupted a period of 100 kyr in which subsurface temperatures
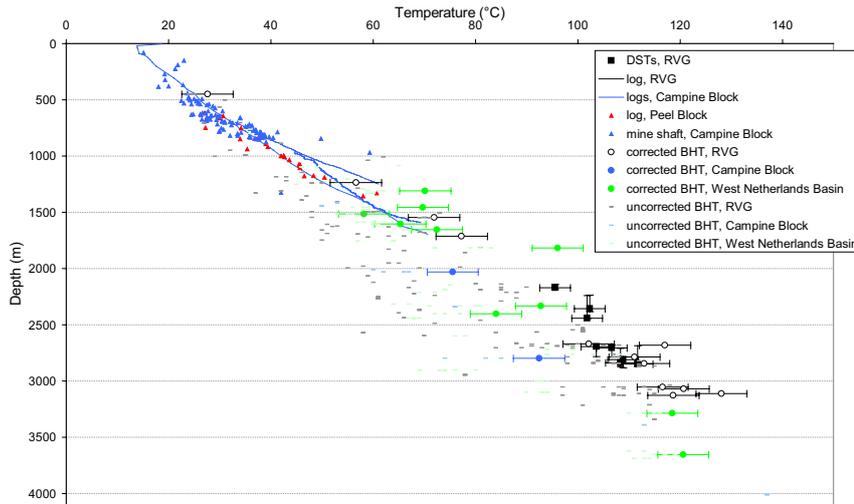
Fig. 4. Compilation of temperature data and their uncertainty range in the Roer Valley Graben Area, after Luijendijk et al. (2011).
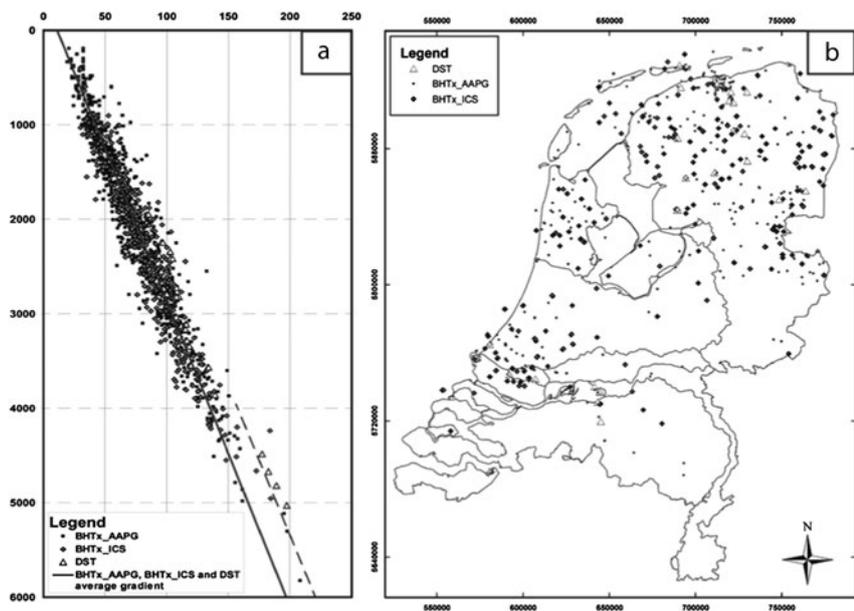
Fig. 5. Average geotherm in the Netherlands based on compilation of available borehole data, corrected for thermal perturbations by drilling (adopted from Bonté et al, 2012). (a) Temperature as a function of depth. (b) Location of the used drillholes.
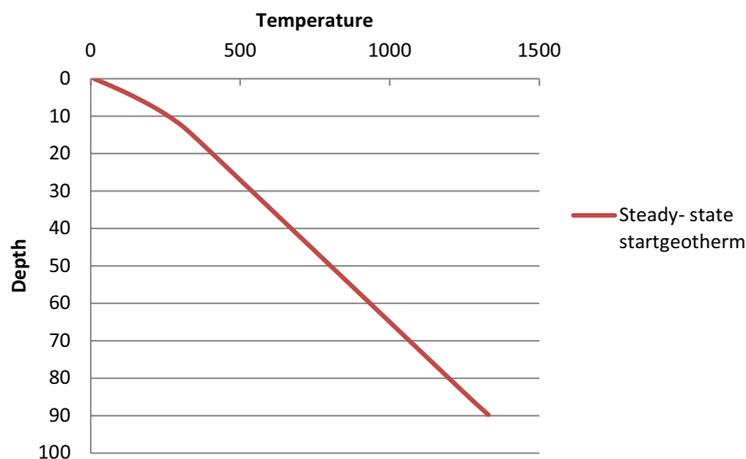
Fig. 6. Temperature with depth at the start of the modelling (i.e. 130 kya).

were generally decreasing. During that time period, the thermal field was still in the process of adapting to the change from
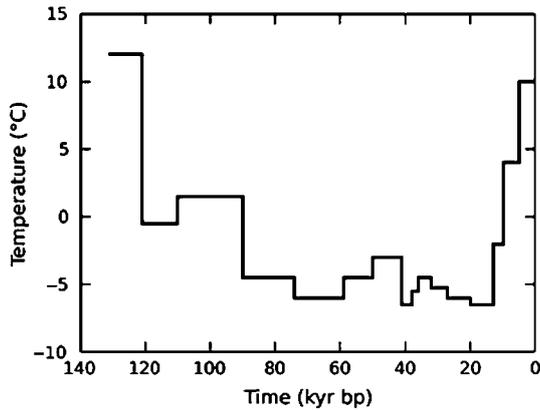
Fig. 7. *Surface temperature history used for the simulation of temperature–depth profiles by Luijendijk et al. (2011). Based on climate proxy data published by Caspers & Freund (2001), Van Gijssel (1995), Huijzer & Vandenberghe (1998), and Zagwijn (1996).*

warm surface temperatures of the Eemian to the glacial conditions of the Weichselian, without reaching equilibrium.

Going even further back in time, a simulation starting with the low surface temperature of the preceding Saalian glacial (from ∼238 to 128 kya) yields a predicted present-day surface gradient of 15.8 °C km^{-1}, a gradient at 3 km depth of 29.6°C km^{-1} and a gradient at a depth of 8 km of 21.9°C km^{-1} (Fig. 9).

Since the air temperature during the Oostermeer (∼243 to 238 kya) and the Saalien (i.e. during the last climate cycle before the Eemian) probably fluctuated between the same values as during the Eemian and Weichselien (e.g. Cheddadi et al., 2005), the scenarios depicted in Fig. 8b and in Fig. 9 can be considered as endmembers: one assuming a surface temperature of 12°C and the other surface temperature of –5°C for the entire period before 130 kya.

Our results show that the continuous change in surface temperature excludes equilibrium geothermal conditions, and also that the present-day geotherm is in a transient state, for
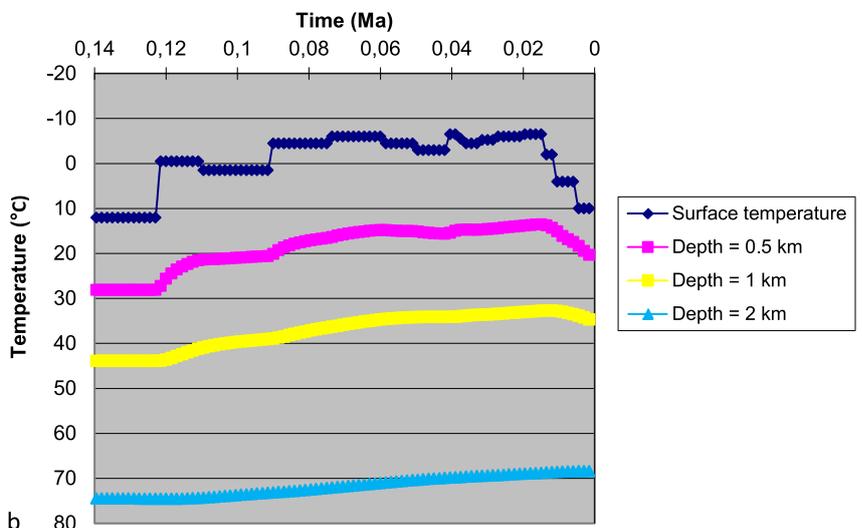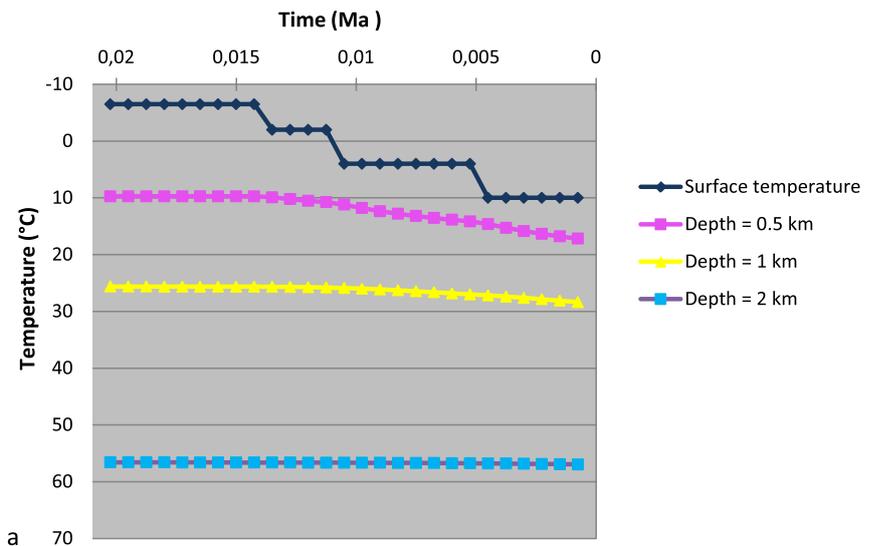
Fig. 8. *(a) Temperature as a function of time at depths of 0, 0.5, 1.0 and 2.0 km. Results for the scenario based on estimated surface temperatures in the Netherlands during the past 20 kyr. (b) Temperature as a function of time at depths of 0, 0.5, 1.0 and 2.0 km. Results for the scenario based on estimated surface temperatures in the Netherlands during the past 130 kyr.*
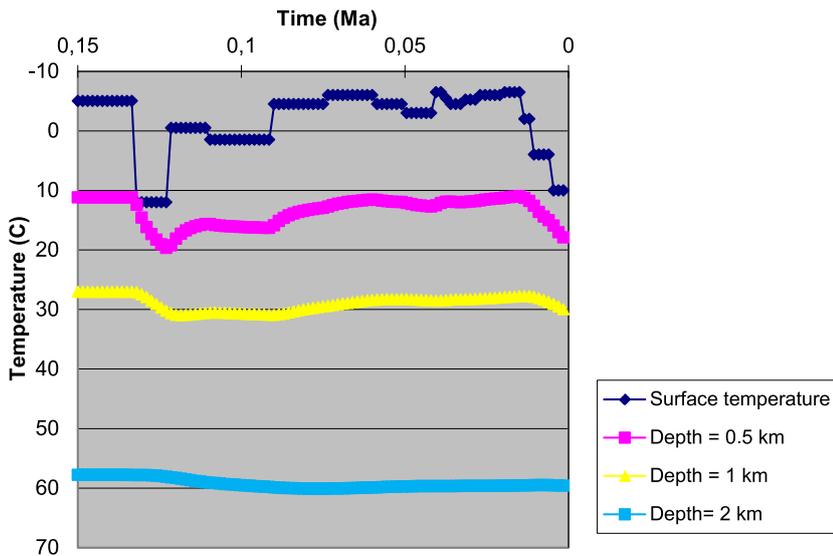
Fig. 9. Temperature as a function of time at depths of 0, 0.5, 1.0 and 2.0 km. Results for the scenario based on estimated surface temperatures in the Netherlands during the past 130 kyr, extended with a cold period before 130 kya.
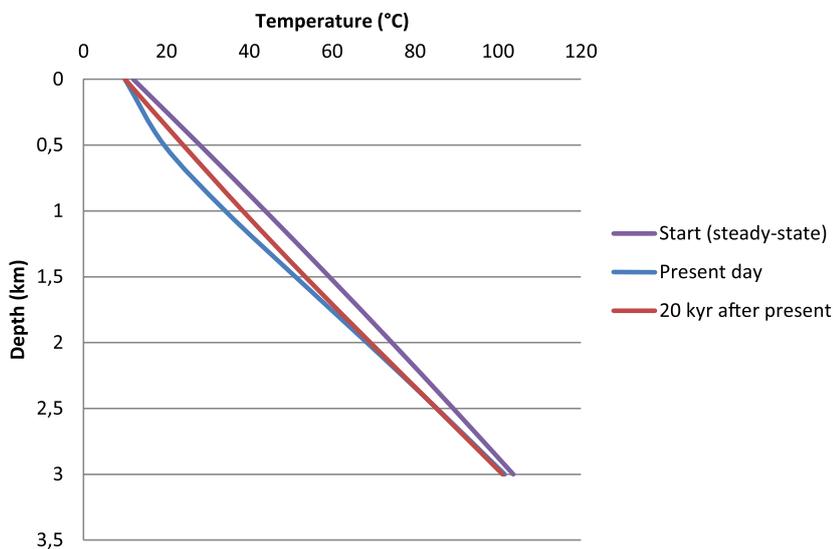
Fig. 10. Geotherms in the upper 3 km for the scenario that started at 130 kya. The purple line is the steady-state geotherm at the start of the model run, with a surface temperature of 12°C, the blue line is the modelled present-day geotherm resulting from the surface temperature history of Luijendijk et al. (2011, see Fig. 8), with a present-day surface temperature of 10°C, and the red line is the geotherm 20 kyr from now, assuming that the surface temperature remains 10°C.

example when we extend the modelled period to calculate the future evolution of subsurface temperatures, based on a no future climate change scenario (i.e. applying a constant surface temperature), the model predicts a further heating of the shallow subsurface (Fig. 10): the geotherm at 20 kyr from now shows a surface gradient of 28.1°C km⁻¹.

Discussion

We have shown that, in the Netherlands, the changes in the surface temperature of the past 130 kyr affected the temperature at shallow crustal depths. The magnitude of this effect decreases with depth. We found changes in temperature of up to 10°C at a depth of 1 km, but of just a few degrees centigrade (<3) at a depth of 3 km.

Other processes causing changes in the subsurface temperature field with time are (1) crustal deformation, sedimentation and erosion, (2) the variation of thermal parameters, both in time and in space, (3) groundwater flow and (4) latent heat by melting and/or freezing.

(1) As shown previously (Grasemann & Mancktelow, 1993; ter Voorde & Bertotti, 1994; Luijendijk et al., 2011), temperature changes due to crustal deformation by faulting in the upper crust are only detectable at short distances (< 10 km) from the fault. The effect is dependent on the rate of faulting. For fault rates up to several millimetres per year the influence of faulting is in the same order of magnitude as the changes due to the surface temperature variations. However, when the rate of faulting is in the order of tens of millimetres per year the resulting temperature changes increase to values of > 50°C. For these cases, the influence of the surface temperature on the subsurface thermal field is insignificant compared to the effect of faulting.

(2) Dependent on the rate of sedimentation, the thermal response to sediment infill of a basin may be a decrease in temperature (due to the low temperature of the sediments) or an increase

(due to their insulating effect): if the sediment conductivity is 50% lower than the basement conductivity, the turning point between cooling and heating is at a sedimentation rate of 1.8 mm/yr (ter Voorde & Bertotti, 1994). When sedimentation has ceased, and other parameters are constant with time, the thermal field will return to a steady-state situation in which thermal gradients are inversely proportional to the conductivity. Common variations in thermal conductivity between the basement and sediments are up to a maximum of a factor 1.5, although the presence of salt, for example, may cause larger variations. Depending on the thickness and conductivity of the sediment layer, this effect may dominate the effect of the surface temperature variations.

In the Netherlands, the influence of the sediments causes large but local variations in the subsurface temperature (Bonté et al., 2012). For example, thick upper Carboniferous (Silesian) deposits, mainly composed of shale, have a strong insulation effect in the province Zuid-Holland, whereas evaporitic (high conductive) sediments in the Zechstein layers in the northern part of the country cause a relative warm shallow (<1 km) subsurface and lower temperatures at depths around 2 km.

(3) Groundwater flow can have a substantial effect on the subsurface temperature, even in a relatively flat region like the Netherlands: Luijendijk (2012) showed that topography-driven groundwater flow in the Roer Valley graben, despite its relatively low relief, caused locally more than 25°C of cooling at a depth of 1.5 km. The thermal response of sediments on changes in surface temperature also depends on the direction and velocity of groundwater flow (e.g. Goto et al., 2005).

Although these three effects are able to dominate the influence of the varying surface temperature, they are all local effects, whereas the climatic signal will cause a change in the geothermal gradient on a regional scale. Only in large scale, deep reaching systems does the groundwater flow have a regional effect on the subsurface temperature distribution. In the Netherlands, such deep groundwater movements occurred during the glacials and associated low sea levels, for example during the Elsterian and Middle Saalien glaciations (Verweij, 2003).

Interpretation of local borehole data can be further complicated by the fact that ground surface temperature can vary spatially over both short and large distances depending on surface conditions (e.g. vegetation, moisture availability, snow cover characteristics), which determine the local surface energy balance (e.g. Chisholm & Chapman, 1992). Although magnitudes of these variations appear to be generally less than 2°C in the Netherlands (Kooi, 2008), they can affect temperatures to kilometres depth and create significant variations in thermal gradients at shallow (i.e. <200 m) depth intervals.

For freezing or thawing of ice-rich permafrost, latent heating moderates the response of the subsurface temperatures to changes in surface temperature. When the surface temperature increases and thawing occurs, this will take energy, implying that temperatures at larger depth will increase less than without thawing. Where there is a decrease in surface temperature, and this is accompanied by freezing, energy is released and the decrease in subsurface temperature is reduced (see Kitover et al., 2013). This mechanism dims the effects of the varying surface temperature.

Other effects that influence the subsurface temperature are surface conditions, for example solar insolation and vegetation effects (e.g. Chisholm & Chapman, 1992). In the Netherlands, changes in surface temperature due to variations in the surface energy balance, outside the built environment, are ~1.5–2°C (e.g. Kooi, 2008).

Conclusions

We used a numerical model to study the effect of past climate changes on the thermal field in the upper crust. The effect of periodic changes in surface temperature due to climatic cycles extends to depths of 2–3 km: the higher the frequency of the changes in surface temperature, the lower the amplitude of changes in temperature at depth. Because of these cycles, the upper few kilometres of the subsurface thermal field are prevalently in a transient state.

The surface temperature history of the past 130,000 years in the Netherlands has resulted in a decreased geothermal gradient in the upper kilometres of the Dutch subsurface, and a slightly increased gradient at depths between 1 and 3 km. In addition, the climatic history before the Eemian is still visible in the surface thermal gradient of the Netherlands. Our results demonstrate that the upper 3 km of the Dutch crust are currently in a stage of heating.

Acknowledgements

References

Agemar, T., Schellschmidt, R. & Schulz, R., 2012. Subsurface temperature distribution in Germany. Geothermics 44: 65-77.

Bense, V.F. & Kooi, H., 2004. Temporal and spatial variations of shallow subsurface temperature as a record of lateral variations in groundwater flow. Journal of Geophysical Research 109, B04103, doi:10.1029/2003JB002782. Available at http://onlinelibrary.wiley.com/doi/10.1029/2003JB002782/full, last accessed 31 March 2014.

Bodri, L. & Cermak, V., 2007. Borehole Climatology: a new method of how to reconstruct climate. Elsevier (Oxford), 352 pp.

Bonté, D., Van Wees, J.D. & Verweij, H., 2012. Subsurface temperature of the on-shore Netherlands: new temperature dataset and modelling. Netherlands Journal of Geosciences 91(4): 491-515.

Caspers, G. & Freund, H., 2001. Vegetation and climate in the Early- and Pleni-Weichselian in northern Central Europe. Journal of Quaternary Science 16(1): 31-48.

Cheddadi, R., De Beaulieu, J.L., Jouzel, J., Andrieu-Ponel, V., Laurent, J.M., Reille, M., Raynaud, D. & Bar-Hen, A., 2005. Similarity of vegetation dynamics during interglacial periods. Proceedings of the National Academy of the USA 102(39): 13939-13943.

Chisholm, T. & Chapman, D.S., 1992. Climate change inferred from analysis of borehole temperatures, an example from Western Utah. Journal of Geophysical Research 97(B10): 14155-14175.

Cloetingh, S, van Wees, J.D., Ziegler, P. A., Lenkey, L., Beekman, F., Tesauro, M., Förster, A., Norden, B., Kaban, M., Hardebol, N., Bonté, D., Genter, A., Guillou-Frottier, L., Ter Voorde, M., Sokoutis, D., Willingshofer, E., Cornu, T. & Worum, G., 2010. Lithosphere tectonics and thermo-mechanical properties: An integrated modelling approach for Enhanced Geothermal Systems exploration in Europe. Earth-Science Reviews 102(3–4): 159-206.

Furbish, D., 1997. Fluid physics in geology. An introduction to fluid motions on Earth's surface and within its crust. Oxford University Press (New York and Oxford).

Goes, S., Govers, R. & Vacher, P., 2000. Shallow mantle temperatures under Europe from P and S wave tomography Journal of Geophysical Research 105(B5): 11153-11169.

Goto, S., Yamano, M. & Kinoshita, M., 2005. Thermal response of sediment with vertical fluid flow to periodic temperature variation at the surface. Journal of Geophysical Research 110(B01106): abstract.

Grasemann, B. & Mancktelow, N.S., 1993. Two-dimensional thermal modelling of normal faulting: the Simplon Fault Zone, Central Alps, Switzerland. Tectonophysics 225: 155-165.

Harris, R.N. & Chapman, D.S., 2005. Borehole temperatures and tree rings: Seasonality and estimates of extratropical Northern Hemispheric warming. Journal of Geophysical Research 110(F04003).

Hays, J.D., Imbrie, J. & Shackleton, N.J., 1976. Variations in the Earth's orbit: Pacemaker of the Ice-Ages. Science 194(4270): 1121-1132.

Huang, S.P., Pollack, H.N. & Shen, P.-Y., 2008. A late Quaternary climate reconstruction based on borehole heat flux data, borehole temperature data, and the instrumental record, Geophysical Research Letters 35(L13703).

Huijzer, B. & Vandenberghe, J., 1998. Climatic reconstruction of the Weichselian Pleniglacial in northwestern and Central Europe. Journal of Quaternary Science 13(5): 391-417.

Karg, H., Bücker, C. & Schellschmidt, R., 2004. New subsurface temperature maps for the Tertiairy Lower Rhine Basin and the adjacent Variscan Basement – Germany, The Netherlands, Belgium. Netherlands Journal of Geosciences 83(2): 135-146.

Kitover, D., van Balen, R.T, Roche, D., Vandenberghe, J. & Renssen, H., 2013. New estimates of permafrost evolution during the last 21k years in Eurasia using numerical modelling Permafrost and Periglacial Processes 24: 286-303.

Kooi, H., 2008. Spatial variability in subsurface warming over the last three decades; insight from repeated borehole temperature measurements in The Netherlands. Earth and Planetary Science Letters 270: 86-94.

Kukkonen, I, & Jõeleht, A., 2003. Weichselian temperatures from geothermal heat flow data. Journal of Geophysical Research 108(B3): 2163.

Kukkonen, I.T. & Šafanda, J., 1996. Palaeoclimate and structure: the most important factors controlling subsurface temperatures in crystalline rocks. A case history from Outokumpo, eastern Finland. Geophysical Journal International 126: 101-112.

Kukkonen, I.T., Gosnold, W.D. & Šafanda, J., 1998. Anomalously low heat flow density in eastern Karelia, Baltic Shield: a possible palaeoclimatic signature. Tectonophysics 291: 235-249.

Kukkonen, I.T., Rath, V., Kivekäs, L., Šafanda, J. & Čermak, V., 2011. Geothermal studies of the Outokumpu Deep Drill Hole, Finland: Vertical variation in heat flow and palaeoclimatic implications. Physics of the Earth and Planetary Interiors 188(1–2): 9-25.

Luijendijk, E., 2012. The role of fluid flow in the thermal history of sedimentary basins – Inferences from thermochronology and numerical modeling in the Roer Valley Graben, southern Netherlands. PhD thesis, Vrije Universiteit Amsterdam (Amsterdam).

Luijendijk, E., ter Voorde, M., van Balen, R., Verweij, H. & Simmelink, E., 2011. Thermal state of the Roer Valley Graben, part of the European Cenozoic Rift System. Basin Research 23: 65-82.

Majorowicz, J. & Šafanda, J., 2008. Heat flow variation with depth in Poland: evidence from temperature logs in 2.9-km-deep well Torun-1. International Journal of Earth Sciences 97(2): 307-315.

Majorowicz, J. & Wybraniec, S., 2011. New terrestrial heat flow map of Europe after regional paleoclimatic correction application International Journal of Earth Sciences (Geol Rundsch) 100: 881-887.

Pollack, H.N. & Huang, S.P., 2000. Climate reconstruction from subsurface temperatures Annual Review of Earth and Planetary Sciences 28: 339-365.

Pollack, H.N., Huang, S. & Shen, P-Y., 1998. Climate change record in subsurface temperatures: A global perspective. Science 282: 279-281.

Ramaeckers, J.J.F., 1991. The Netherlands. In: Hurtig, E., Cermak, V., Haenel, R. Zui, V., (eds): Geothermal Atlas of Europe. Hermann Haack Verlagsgesellschaft mbH (Gotha).

Šafanda, J. & Rajver, D., 2001. Signature of the last ice age in the present subsurface temperatures in the Czech Republic and Slovenia. Global and Planetary Change 29: 241-257.

Šafanda, J., Szewczyk, J. & Majorowicz, J., 2004. Geothermal evidence of very low glacial temperatures on a rim of the Fennoscandian ice sheet. Geophysical Research Letters 31: L07211.

ter Voorde, M. & Bertotti, G., 1994. Thermal effects of normal faulting during rifted basin formation. 1. A finite difference model. Tectonophysics 240: 133-144.

Tesauro, M., Kaban, M.K., Cloetingh, S.A.P.L., Hardebol, N.J. & Beekman, F., 2007. 3D strength and gravity anomalies of the European lithosphere. Earth and Planetary Science Letters 263: 56-73.

van Balen, R.T., van Bergen, F., de Leeuw, C., Pagnier, H., Simmelink, E., van Wees, J.D. & Verweij, J.M., 2000. Modelling the hydrocarbon generation and migration in the West Netherlands Basin, the Netherlands. Netherlands Journal of Geosciences 79(1): 29-44.

van Balen, R.T., Verweij, J.M., van Wees, J.D., Simmelink, E., van Bergen, F. & Pagnier, H., 2002. Deep subsurface temperatures in the Roer Valley Graben and the Peelblock, the Netherlands – new results. Netherlands Journal of Geosciences 81(1): 19-26.

van Dalfsen, W., 1981. Geothermal investigation in shallow observation wells. Project G/A 9 – Contracts 073-76. The shallow subsurface temperature field in the Netherlands, Delft. Groundwater survey TNO, Report OS 81–05.

van Dalfsen, W., 1983. Het ondiepe ondergrondse temperatuurveld in Nederland. Report no. OS 83–31. Dienst Grondwaterverkenning TNO (Delft), 52 pp.

van Gijssel, K., 1995. A hydrogeological and paleoenvironmental data set for large-scale groundwater flow simulations in the northeastern Netherlands. Mededelingen Rijks Geologische Dienst 52: 105-134.

Verweij, H.M., 2003. Fluid flow systems analysis on geological timescales in onshore and offshore Netherlands. PhD thesis ,Vrije Universiteit Amsterdam (Amsterdam). Available at http://www.nlog.nl/resources/Publicaties/Thesis_Verweij2003.pdf, last accessed 31 March 2014.

Zagwijn, W.H., 1996. An analysis of Eemian climate in Western and Central Europe. Quaternary Science Reviews 15(5–6): 451-469.