12 The Campine Plateau

Koen Beerten, Roland Dreesen, Jos Janssen and Dany Van Uytven

Abstract

Once occupied by shallow and wide braided channels of the Meuse and Rhine rivers around the Early to Middle Pleistocene transition, transporting and depositing debris from southern origin, the Campine Plateau became a positive relief as the combined result of uplift, the protective role of the sedimentary cover, and presumably also base level fluctuations. The escarpments bordering the Campine Plateau are tectonic or erosional in origin, showing characteristics of both a fault footwall in a graben system, a fluvial terrace, and a pediment. The intensive post-depositional evolution is attested by numerous traces of chemical and physical weathering during (warm) interglacials and glacials respectively. The unique interplay between tectonics, climate, and geomorphological processes led to the preservation of economically valuable natural resources, such as gravel, construction sand, and glass sand. Conversely, their extraction opened new windows onto the geological and geomorphological evolution of the Campine Plateau adding to the geoheritage potential of the first Belgian national park, the National Park Hoge Kempen. In this chapter, the origin and evolution of this particular landscape is explained and illustrated by several remarkable geomorphological highlights.

Keywords

Campine plateau • Meuse terraces • Late Glacial and Holocene dunes • Tectonic control on fluvial evolution • Polygonal soils • Natural resources • Coal mining

12.1 Introduction

The Campine Plateau is an extraordinary morphological feature in northeastern Belgium, extending into the southern part of the Netherlands (Fig. 12.1a). It runs from the southeast to the northwest, from an altitude of ca. 100 m a.s.l. (TAW: Tweede Algemene Waterpassing) in the south, to ca. 30 m near the Belgian-Dutch border in the north (Fig. 12.1 b). The polygonal shape of this lowland plateau has attracted a lot of attention from geoscientists during the last 100 years, all the more since it is covered by a thick sheet of coarse fluvial deposits from the Rhine and Meuse (Paulissen 1973, 1983). As these rivers' current channels are located several to tens of km further to the east, at a much lower altitude, the Campine Plateau witnesses a unique episode in the Quaternary evolution of the region. The interplay of fault activity, uplift, weathering, fluvial incision, regressive erosion, substrate characteristics, and aeolian processes created the current shape of the plateau. The plateau can be considered a classic case of relief inversion. The resulting steep bordering slopes are in strong contrast with the otherwise flat landscape of the European sand belt.

K. Beerten (✉)
Engineered and Geosystems Analysis, Belgian Nuclear Research Centre SCK•CEN, Boeretang 200, 2400 Mol, Belgium
e-mail: kbeerten@sckcen.be

R. Dreesen
Geological Survey of Belgium, Royal Belgian Institute of Natural Sciences, Jennerstraat 13, 1000 Brussels, Belgium
e-mail: roland.dreesen@telenet.be

J. Janssen · D. Van Uytven
Werkgroep Geologie, LIKONA (Limburgse Koepel Voor Natuurstudie), Provinciaal Natuurcentrum, Craenevenne 86, 3600 Genk, Belgium

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The availability of erodible sand in the surrounding areas promoted the development of aeolian sand sheets and large inland dune areas (Fig. 12.1c) during the (Late) Weichselian when vegetation cover was sparse and the water table low. Furthermore, the poor sandy and stony soils and the deep groundwater table have made the plateau area unattractive for human occupation. Nevertheless, early agricultural activities and Mesolithic settlements seem to have existed close to wetlands (Vermeersch et al. 1974). The collection of heather sods for fertilizing soils (leading to the formation of plaggen soils), a widespread practice in the region, caused severe landscape instability from the fifteenth to the nineteenth century, as a result of which drift sand landscapes developed. Massive pine plantation during the late nineteenth century and first half of the twentieth century led to stabilization of the landscape and destruction of heathland, serving notably the coal mining industry in the Liège and Campine coal fields.

The establishment of pine plantations was a direct result of the need for wood in the galleries of underground coal mines in the nearby Liège area. Later, large amounts of pine wood were used in the Campine Plateau itself, where the geological Campine Basin hosts many coal seams at several hundreds of meters depth. The sandy and dry nature of the soils that developed on Quaternary fluvial and aeolian deposits provided an excellent habitat for pine trees. Similarly, gravel and, especially, sand extraction became widespread economic activities in the plateau region (Gullentops and Wouters 1996). Miocene and Pliocene quartz sand, coarse Rhine sand and loamy Meuse gravels are dug in huge extraction pits, the extent and depth of which can easily be detected on uncorrected digital terrain models, just as the coal spoil heaps. The deep groundwater table, as a result of uplift and river incision, enables dry extraction of some of these mineral resources, in contrast to the much more difficult wet extraction of, e.g., more valuable Meuse gravels in the present-day floodplain.

This chapter provides a state-of-the-art overview on the origin and development of the Campine Plateau, against a background of climate change, tectonic movements and human activities. The link between the current landscape and the overall evolution of the region is highlighted, as well as the geoheritage value of the landscape (see Chap. 24).

12.2 Geographical and Geological Setting

12.2.1 Location of the Campine Plateau

The Campine region is a 4000 km^2 large coversand area in NE Belgium and the southern part of the Noord Brabant province in the Netherlands. The name Campine (Kempen) is derived from Latin “campina”, which literally means open space. The location of the Campine Plateau with respect to

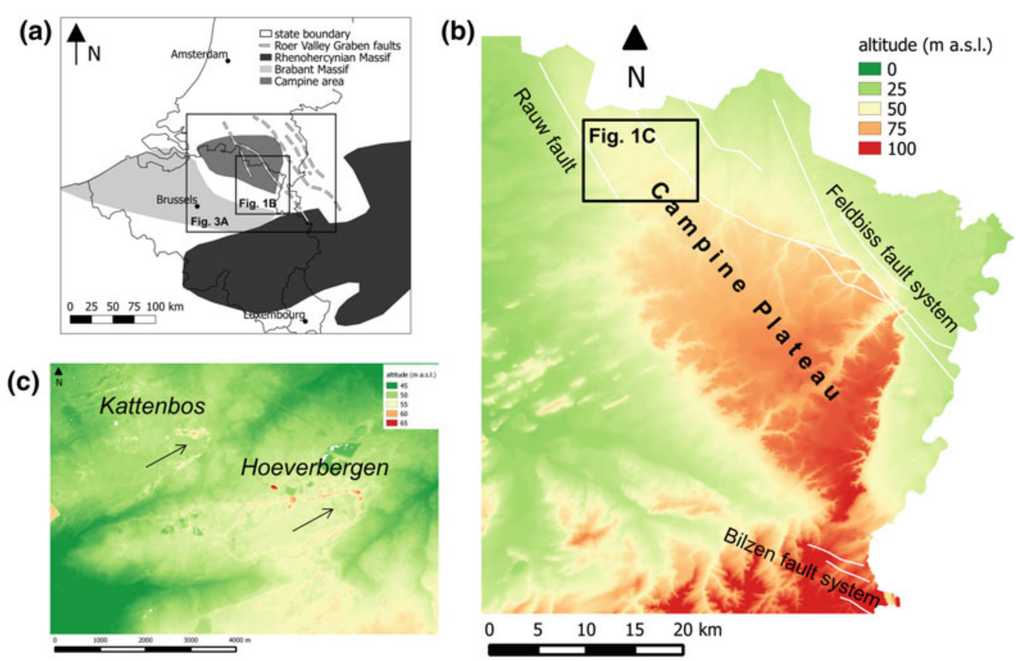

Fig. 12.1 **a** Location of the Campine area in Belgium and the southern Netherlands with indication of Roer Valley Graben faults and Caledonian/Hercynian Massifs. **b** DTM of the Campine Plateau, including Quaternary faults (*white lines*, from G3Dv2; Matthijs et al. 2013). M-G. Meeuwen-Gruitrode. **c** DTM of the parabolic dune areas near Lommel (Kattenbos-Hoeverbergen)

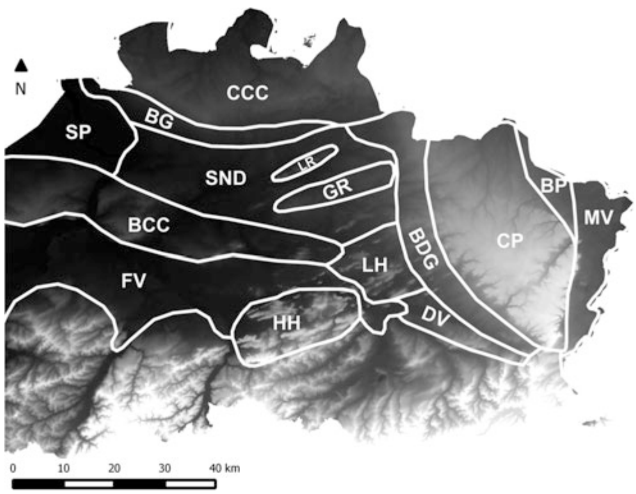

Fig. 12.2 Geomorphological units surrounding the Campine Plateau (from De Moor and Pissart 1992). *CP* Campine Plateau; *BP* Bocholt Plain; *MV* Meuse Valley; *BDG* Beringen-Diepenbeek Glacis; *SND* Schyns-Nete Depression; *LR* Lichtaart Ridge; *GR* Geel Ridge; *FV* Flemish Valley; *DV* Demer Valley; *LH* Lummen Hills; *HH* Hageland Hills; *BCC* Boom Clay Cuesta; *CCC* Campine Clay Cuesta; *SP* Scheldt Polders

other geomorphological entities in NE Belgium is shown in Fig. 12.2 (De Moor and Pissart 1992).

The Campine area is subdivided into several subregions. The Schyns-Nete Depression (SND; Fig. 12.2) in the west, also called the Campine Plain, is an erosive landform characterized by WSW–ENE trending ridges and alluvial plains that follow the strike of the Neogene subcrop (Goossens et al. 1983). The most prominent ridges are the Lichtaart Ridge (LR) and the Geel Ridge (GR). The Campine Plateau (CP) in the east is an accumulative landform, consisting of coarse fluvial deposits from the rivers Meuse and Rhine. These deposits extend towards the north into the Bocholt Plain (BP), which borders the Campine Plateau, north of the boundary faults of the Roer Valley Graben (see Chap. 13). The Beringen-Diepenbeek Glacis (BDG), a cryopediment, connects the SND with the Campine Plateau. Towards the south, the glacis merges into the Demer Valley (DV) and the Lummen Hills (LH), a continuation of the Hageland Hills (HH, see Chap. 14). In the east, the Campine Plateau is bordered by the incised Meuse Valley (MV).

12.2.2 Tectonic Framework

The Campine Plateau is situated in the Campine Basin, which constitutes the southernmost extension of the geological North Sea Basin. It coincides with a deep structural unit, referred to as the Eastern Campine Block (Fig. 12.3a). This structural unit is bounded by deep faults forming the Rauw-Beringen fault system in the (south)west (Fig. 12.3a, RaF fault), the Brabant Massif in the south, and the Feldbiss fault zone in the (north) east (Fig. 12.3a, FF fault) (Dusar and Langenaeker 1992). The thickness of the post-Oligocene sediments range from ca. 200 m on the Western Campine Block over 300 m on the Eastern Campine Block to more than 1000 m in the Roer Valley Graben. Quaternary deposits on the Eastern Campine Block are usually less than 20 m thick, while they may reach up to 100–200 m thickness in the Roer Valley Graben. The area west of the Feldbiss Fault very likely has experienced shoulder uplift during the Middle Pleistocene that is probably associated with accelerated uplift in the Ardenne (Demoulin and Hallot 2009) and/or enhanced subsidence in the Roer Valley Rift System (van Balen et al. 2000).

The morphotectonic map of the Campine Plateau shows the presence of various fault-related escarpments (Paulissen 1997). The most important one is the Bree Fault Scarp, which is the morphological expression of the Feldbiss fault (Fig. 12.3b, see also Chap. 13). To the southeast, this scarp shows a bifurcation known as the Berg Fault Scarp. Towards the northwest, the Bree Scarp splits into the Grote Brogel Fault Scarp, the Reppel Fault Scarp and the Bocholt Fault Scarp, separating the planated fault blocks north of the Campine Plateau, namely the Lommel, Reppel and Kaulille plains, and the Bocholt Plain. The western escarpment partly coincides with the Rauw-Poppel fault (Vandenberghe 1982, 1990; Gullentops and Vandenberghe 1995; Verbeeck et al. 2002), while Neogene and Quaternary displacements are reported along a fault zone inherited from the Caledonian basement directly south of the southernmost plateau tip (Gullentops and Claes 1997) (Fig. 12.1b).

12.2.3 Neogene Evolution

The Campine Block, as part of the Campine Basin, became covered by marine, estuarine, and continental sediments during the late Oligocene and Neogene (Fig. 12.4). This led to overall shoaling of the area (sea level drop) ultimately leading to the establishment of full continental conditions during the Early Pleistocene (Westerhoff et al. 2008). The Neogene sediments consist of the partly continental Lower Miocene sandy Bolderberg Formation, well-known for its high quartz content and lignite layers, the Upper Miocene glauconite-bearing marine Diest Sands, the Upper Miocene glauconitic and slightly clayey marine Kasterlee Sands, and the Pliocene estuarine Mol sands, also known for its high quartz content, and fluvial Kieseloolite Formation. Notably, the white and pure quartz sands of the Bolderberg and Mol formations are intensively quarried, and their preservation is directly related to the Pleistocene evolution of the area, as will be shown below.

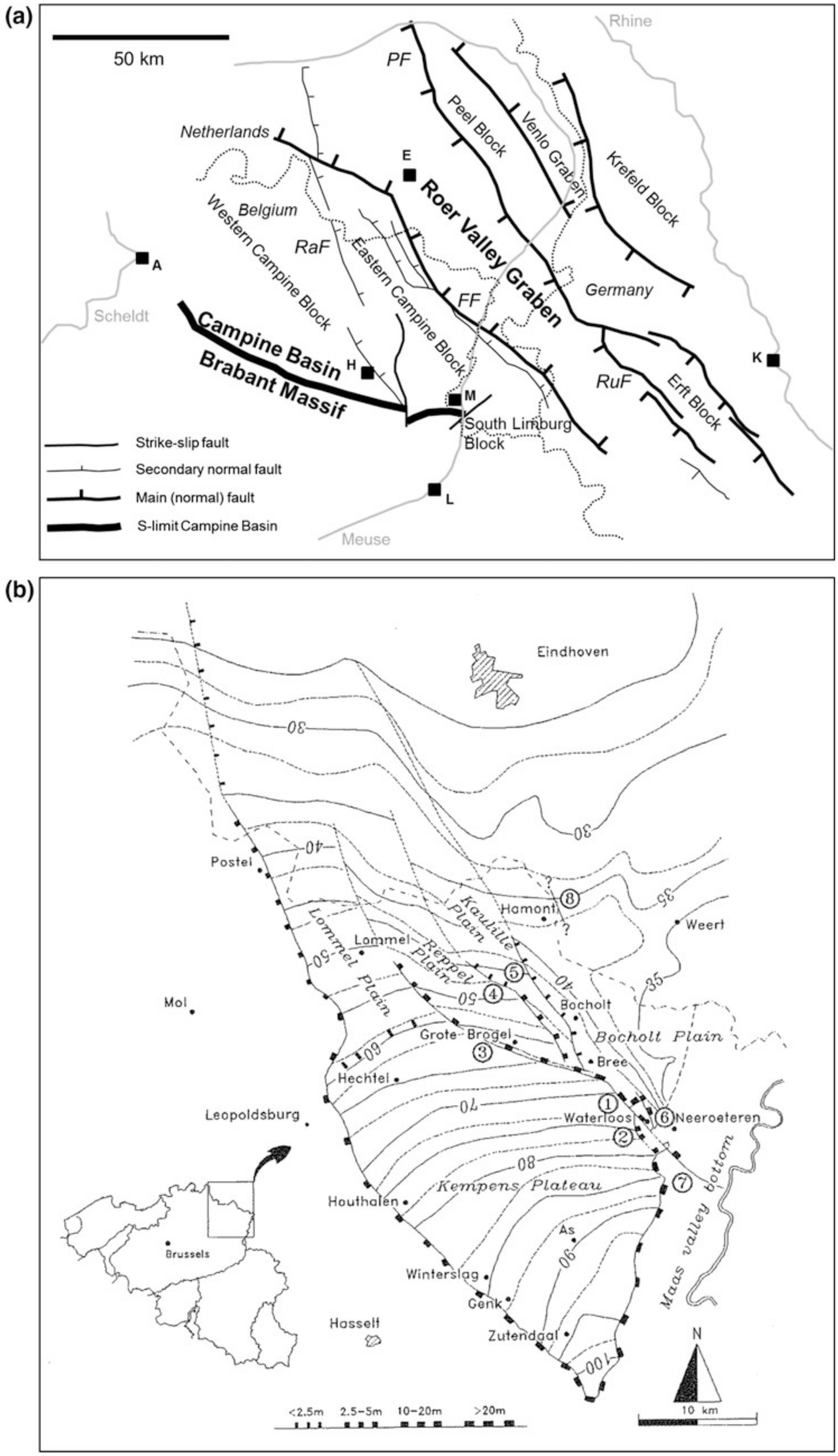

Fig. 12.3 **a** Tectonic framework of the Roer Valley Graben (modified after Geluk et al. 1994; Vanneste et al. 2001). *FF* Feldbiss Fault; *PF* Peel Boundary Fault; *RuF* Rurrand Fault; *RaF* Rauw Fault; *H* Hasselt; *M* Maastricht; *L* Liège; *K* Köln; *E* Eindhoven; *A* Antwerpen. **b** Morphotectonic map of the Campine Plateau highlighting the location of scarps, and notably the following fault scarps: *1* Bree fault scarp; *2* Berg fault scarp; *3* Grote Brogel fault scarp; *4* Reppel fault scarp; *5* Bocholt fault scarp; *6* waterloos scarp; *7* Bichterweerd Scarp; *8* Hamont Scarp. Contour lines in m TAW (Tweede Algemene Waterpassing) (from Paulissen 1997)

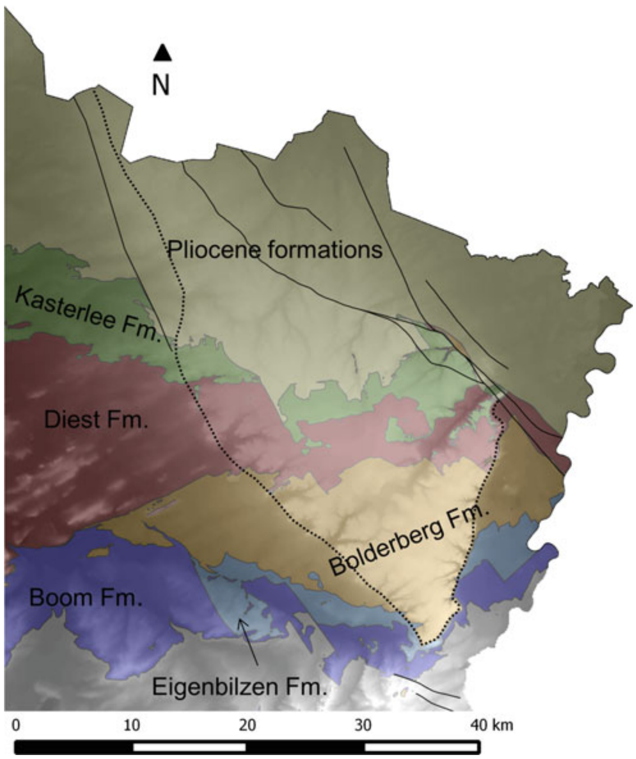

Fig. 12.4 Neogene subcrop map of the Campine Plateau and surrounding areas (from G3Dv2; Matthijs et al. 2013), superimposed on a grayscale DTM. *Dotted line* topographic outline of the plateau; *thin black lines* Quaternary faults. The Pliocene formations comprise the Mol Formation and the top of the Kieseloolite Formation, both consisting of fine to medium sands with intercalated lignite and/or clay layers. The Kasterlee Formation consists of fine glauconitic and slightly clayey sand, the Diest Formation of glauconite-rich coarse indurated (clayey) sand, the Bolderberg Formation of fine to medium pure quartz sand with lignite, the Eigenbilzen Formation of clayey fine sand, and the Boom Formation of (silty) clay

12.3 Origin and Evolution of the Campine Plateau Landscape

12.3.1 Tectonic and Climatic Controls of Early and Middle Pleistocene Fluvial Processes

Following a period of prolonged erosion during much of the Pliocene and Early Pleistocene, sedimentation on the Campine Block resumed when the hydrographic network changed completely around the Early-Middle Pleistocene transition (Juvigné and Renard 1992; Westerhoff et al. 2008). Initially, the Meuse followed an eastern course from Liège to the region north of Aachen where it merged with the Rhine (Fig. 12.5). Shifting tectonic movements along the Feldbiss and Rauw Faults and uplift of the northern margins of the Ardenne-Eifel massif, together with choking by sediments of the area where the Meuse debouched into the graben, caused the river to breach through its northern interfluve and to follow a completely different course over the Eastern Campine Block. At the same time, the Rhine shifted its course as well, flowing over the northern part of the Eastern Campine Block where it merged with the Meuse. Age control is limited, but this drastic event took place in the Early Pleistocene (Juvigné and Renard 1992; Van den Berg 1996), possibly as late as around 1 Ma ago (Westerhoff et al. 2008). Both rivers shifted their course to the east again by 0.5 Ma at the latest (Schokker et al. 2005). Three different lithostratigraphical units cover the plateau: the Lommel Sands, the Winterslag Sands, and the Zutendaal Gravels (Figs. 12.5 and 12.6). The former belongs to the Sterksel Formation while the latter two belong to the Zutendaal Formation (Gullentops et al. 2001). Further to the north, across the Feldbiss fault system, sediments in the Roer Valley Graben are essentially laid down by the Rhine (Bocholt Sands of Sterksel Formation), while the Meuse was still a tributary draining and depositing material (Zutendaal Gravels) on the eastern part of what is now the Campine Plateau (Beerten 2003). However, material with a Meuse signature is also often found within the Bocholt Sands (Beerten 2003).

An important feature of the Zutendaal Gravels is the gradual downstream fining. Near the southern tip of the plateau, the gravel is very coarse, while towards the north, fine gravel and coarse sand become the dominant size fraction. This, together with the conical shape of the plateau and the northward thinning of the gravel deposits led to the view that the plateau constitutes an alluvial fan (Pannekoek 1924). The Meuse deposits of the Campine Plateau are in terrace position relative to other gravel sheets that occur at lower altitude. They were usually linked with the term “Main Terrace” (Pannekoek 1924), according to the large areal extent, or “High Terrace” (Macar 1938), according to the elevated position with respect to the current river course. Paulissen (1973) concluded that the gravels of the Campine Plateau were deposited during the Mindel glaciation, which was correlated with the Elster glaciation in the past. Current correlation schemes correlate the Elster glaciation with Marine Isotope Stage 12, dated to ca. 0.45 Ma (Cohen and Gibbard 2011), while the originally defined Mindel glaciation is probably much older. Age control is poor, but the extensive terrace staircase that developed after deposition of the Zutendaal Gravels already hints towards a relatively old age for the Campine Plateau sediments. Each terrace has been linked with an interglacial–glacial cycle (Vandenberghe 2008), and available palaeomagnetic data suggest that the Main Terrace deposits are Middle Pleistocene in age (van den Berg 1996). This results in age estimates for the Zutendaal Gravels that range between 0.6 and 0.8 Ma, while the Winterslag Sands and Lommel Sands are thought to be slightly younger than 1 Ma (Felder and Bosch 1989; Juvigné and Renard 1992; van den Berg 1996; van Balen et al. 2000; Gullentops et al. 2001) (Fig. 12.6).

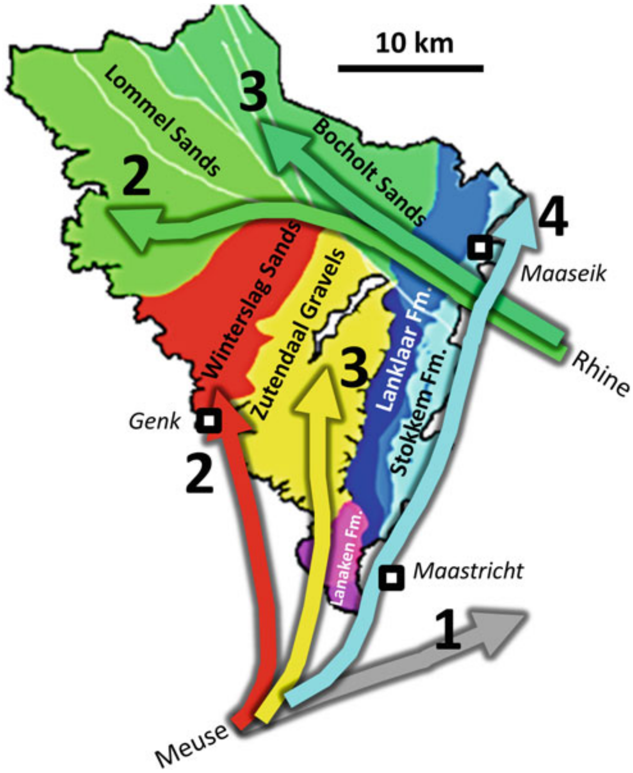

Fig. 12.5 Lithostratigraphic map of the Meuse deposits downstream of Maastricht (modified after Verstraelen 2000; Beerten 2005a, b, 2006). *Arrows* indicate flow direction of the Rhine and Meuse at the time of deposition of the different stratigraphical units. *Numbers* indicate the relative chronology of the various units (from oldest 1 to youngest 4). The entire depositional history is thought to have taken place in the timeframe between ca. 1 and 0.5 Ma (Schokker et al. 2005; Westerhoff et al. 2008). The Winterslag Sands and Zutendaal Gravels are grouped into the Zutendaal Formation (informally referred to as Main Terrace deposits from the Meuse), while the Lommel and Bocholt Sands belong to the Sterksel Formation (Gullentops et al. 2001). The Sterksel Formation consists of Rhine deposits with uptake of material from the river Meuse

12.3.2 The Aftermath: Posterior Development

12.3.2.1 Relief Inversion

The development of fluvial sequences on the Campine Block and in the Roer Valley Graben is clearly associated with the creation of accommodation space, as a result of subsidence. Wide and shallow valleys occupied by braided river channels developed during the Early to Middle Pleistocene transition. The elevated position of these fluvial sequences on top of the plateau, well above the current floodplain of the Meuse, thus needs to be interpreted as a classic case of relief inversion. The erodibility of the coarse grained deposits, ranging from coarse gravel to gravelly sand, is lower than that of the fine sandy Neogene units on which they were laid down. Regions that remained outside the sedimentation realm of the Meuse-Rhine, would thus become preferentially eroded. The western edge of the Campine Plateau can be explained in this way. Sedimentation by the Rhine-Meuse system stopped near the western edge of the Eastern Campine Block, delimited by the Rauw Fault and its southern extension (Fig. 12.3a). The topography of the area west of this fault system was subsequently lowered by regressive erosion from rivers draining the Scheldt basin. This process was accompanied by slope retreat and the development of a pediment, the BDG (Figs. 12.1, 12.2 and 12.6). It resulted in a relief difference of several tens of meters between the plateau and the area west of it.

The effect of regressive erosion in the SND can clearly be observed from the DTM (Fig. 12.1), where the headwaters of the Kleine and Grote Nete can be followed onto the plateau. Enhanced regressive erosion in the Nete basin is probably triggered by erosional events around 450 ka in the southern North Sea basin when the English Channel was opened as a result of a proglacial lake burst (Gupta et al. 2007; Toucanne et al. 2009).

Several uplift records indicate that the southern part of the Campine Plateau experienced around 50 m of uplift since the last 0.7 Ma (van Balen et al. 2000). Around this time, the Meuse was gradually shifting its course to the east again, and started developing the lower part of its extensive terrace staircase (Figs. 12.5 and 12.6). The southern tip of the plateau shows traces of palaeo-meanders that are preserved in the landscape when they laterally carved and incised into older Meuse deposits. An example is the Lanaken terrace (Lanaken Formation), consisting of a relatively thin gravel sheet ca. 10–20 m below the Campine Plateau surface (Paulissen 1973). Subsequent downcutting-infilling cycles were responsible for the development of the Caberg-Pietersem terrace (Lanaken Formation) and Eisden-Lanklaar, Maasmechelen and Geistingen terraces (Lanklaar Formation) (Figs. 12.5, 12.6 and 12.7). The vertical offset between the Campine Plateau Main Terrace and the Saalian Eisden-Lanklaar terrace ranges up to 50 m as a result of fluvial erosion.

The northeastern edge of the Campine Plateau is tectonic in origin. It is known as the Bree Fault Scarp, which is the morphological expression of the Feldbiss fault system (Vanneste et al. 2001; see also Chap. 13). The diffuse continuation of the Meuse deposits north of the Bree escarpment severely complicates their correlation with Rhine deposits in the Roer Valley Graben (Beerten 2003). Nevertheless, it is likely that the Feldbiss Fault caused a cumulative vertical displacement of ca. 25–40 m since deposition of the Zutendaal Gravels (Beerten et al. 1999; Vanneste et al. 2001).

12.3.2.2 Weathering, Soil Formation, and Periglacial Deformation

Several glacial–interglacial cycles occurred after deposition of the Rhine-Meuse sediments on top of the Campine Plateau, leaving sufficient time for post-depositional processes

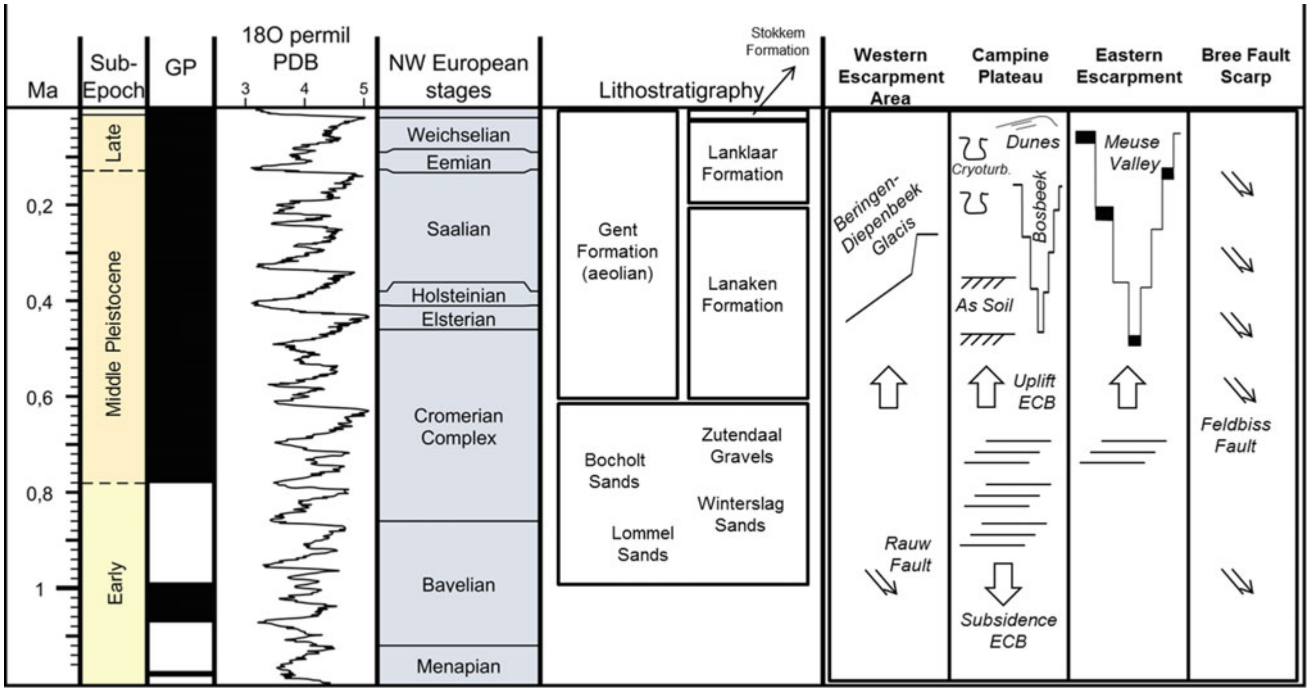

Fig. 12.6 Lithostratigraphic table of the units depicted in Figs. 12.5 and 12.7 and the associated geomorphological, climatic and tectonic development of the area before, during, and after deposition of the Main Terrace deposits (created with TimeScaleCreator-PUBLIC-6.4-21Feb2015). GP refers to the direction of the palaeomagnetic field (*black* normal; *white* reversed). The Early to Middle Pleistocene transition coincides with the Brunhes-Matuyama geomagnetic boundary (~0.78 Ma). *ECB* Eastern Campine Block (see Fig. 12.3a). *Arrows up* refer to the onset of uplift in the region, *arrows down* to subsidence; phases of normal faulting are indicated by *doubled oblique arrows*. *As soil* interglacial weathering soil in Zutendaal Gravels. According to Gullentops et al. (2001), the Meuse deposits on top of the Campine Plateau are younger than the Rhine deposits. This opinion has recently been questioned in the Quaternary subcommission of the National Commission for Stratigraphy. The lateral relationship between the different units is not fully understood yet

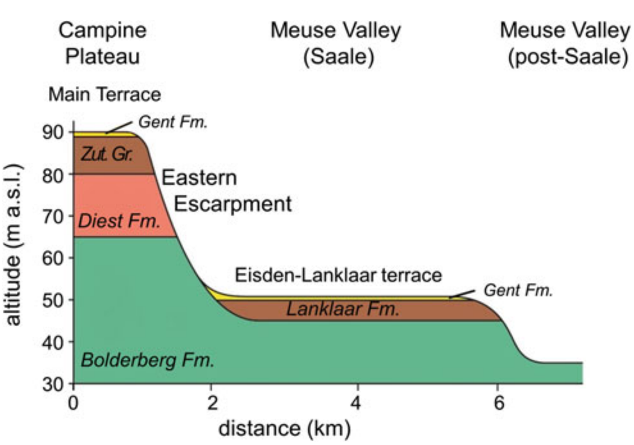

Fig. 12.7 Development of the plateau's eastern escarpment through fluvial erosion and terrace formation. This snapshot shows the situation at the end of the Saalian glaciation or the beginning of the Eemian interglacial (modified after Janssen and Dreesen, 2010). Note the thin aeolian sand layer (Gent Formation) on top of the fluvial terraces

such as weathering, soil development, and periglacial deformation. At some places, the typical red-brown interglacial weathering soil (As Soil; e.g., Gullentops et al. 1993) developed deeply into the gravel deposits (Fig. 12.8). Sufficient precipitation, a deep groundwater table, and the availability of mobile iron are necessary boundary conditions for the development of this soil. Limonite and goethite concretions often give the gravel a cemented appearance, further contributing to its resistance to erosion. This soil is very well developed in the eastern part of the Campine Plateau that more or less coincides with the Zutendaal Gravel outcrop zone (Figs. 12.4 and 12.5). Interestingly, the As Soil is not preserved in the western and northwestern part of the Campine Plateau, or it did not develop there. Several factors might account for this discrepancy. First, the availability of iron could be larger in the Zutendaal Gravels than in other fluvial units of the Campine Plateau, because of the initial presence of now dissolved iron-rich limestones (hematite oolitic ironstones) from Upper Devonian rocks in the south of the Namur syncline, and/or Jurassic minette-type oolitic ironstones from the Lorraine area. Second, the Zutendaal Gravels are more erosion-resistant than the more sandy Winterslag and Lommel Sands, such that the soil may have been eroded from these sandy units. Third, southeast-northwest tilting of the plateau resulted in deep fluvial incision by the Meuse, causing the groundwater table to be significantly lower along the southeastern edge where the Zutendaal Gravels are exposed, in comparison with the sandy facies to the northwest. Finally, the most plausible explanation involves intercalation or superposition of fine-grained sediments, which readily yielded iron in, or on top of, the gravels (M. Dusar, written communication). Where the availability of iron was low, Holocene soil formation processes transformed the top layers of the plateau into a podzol soil, with the typical albic and spodic horizons. In places where the in situ fluvial deposits are overblown by aeolian sand, soil formation took place in the latter. The plateau top layers are heavily deformed by periglacial processes during successive glacials, leading to the development of polygonal soil patterns and cryoturbations (see Sect. 12.4.4).

12.3.2.3 The Development of Dune Landscapes

The Campine Plateau is covered by a discontinuous mantle of homogeneous and fine-grained aeolian sands, which occasionally occur in dry valleys or build up a distinct dune relief. In the revised Belgian Quaternary lithostratigraphy, these aeolian deposits belong to the Opgrimbie Member of the Gent Formation, as defined in the revisited type-locality in the dune complex of Opgrimbie (Derese et al. 2009). Here, and at other places, age control suggests that the aeolian sediments were deposited during the Pleniglacial and Late Glacial under conditions of decreasing landscape stability and increasing soil erodibility associated with declining vegetation during cold interstadials (Kasse et al. 2007). Phases of landscape stability are recorded in these aeolian sands as witnessed by the presence of buried (palaeo)soils which developed during the Allerød and Holocene in response to climatic amelioration and vegetation expansion (Hoek 2001). In many occasions, the sands are blown into parabolic or longitudinal dunes (Gullentops 1957) (Fig. 12.1). A map showing the distribution of dune complexes on the Campine Plateau and surrounding areas is given in Fig. 12.9. Most of them are situated on interfluves between small brook valleys along the (north)western edge of the Campine Plateau, and the Meuse-Scheldt water divide. Famous examples of dune complexes can be found in Lommel (Hoeverbergen and Kattenbos) and near Hechtel. Towards the east, an impressive dune complex can be found between the towns of Gruitrode and Opglabbeek (the "Duinengordel", see also Chap. 24).

The distribution of dune complexes on the Campine Plateau is determined by the availability of erodible sand, and the presence of fluvial processes. The latter hamper the formation of dunes through complex fluvial–aeolian interactions, such as reworking of aeolian material, fluvial erosion, and shallow groundwater tables. In the southeastern plateau area, coarse gravel is outcropping and the lack of suitable sand prevented the development of massive dunes south of the Bosbeek valley. The availability of erodible sand increases towards the northwest, where the sand content of the Early-Middle Pleistocene substrate is high, post-depositional cementation as a result of goethite formation (As Soil, see above) is usually absent, and the deposition of aeolian sand from the north, prior to dune formation

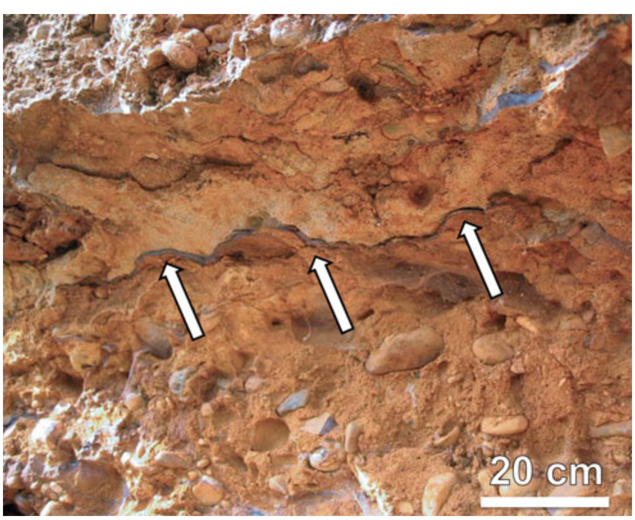

Fig. 12.8 Development of subhorizontal goethite crusts (*arrows*) as part of the As Soil formed in the Zutendaal Gravels (from Dreesen et al. 2005)

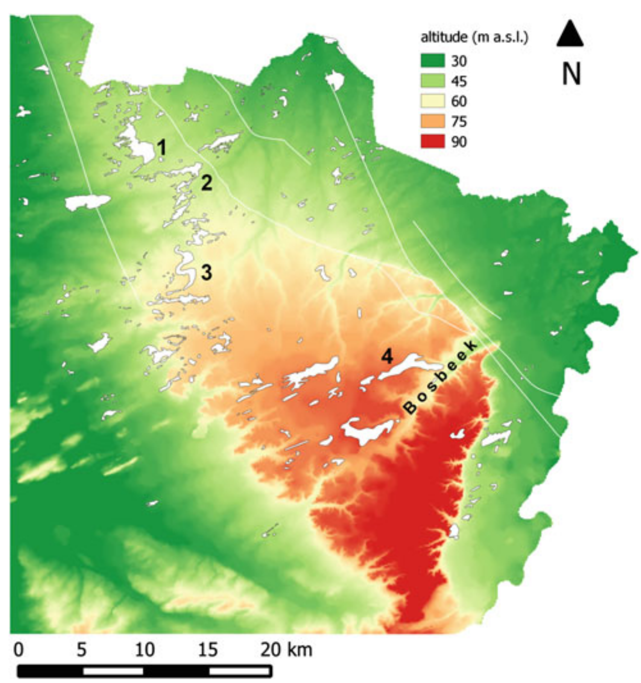

Fig. 12.9 Map of the dune areas (*white spots*) on the Campine Plateau and its surroundings. *1* Kattenbos; *2* Hoeverbergen; *3* Hechtel; *4* Duinengordel. Data from the Quaternary geological map sheets (Beerten 2005a, b, 2006) and the compilation map of Flanders (Bogemans 2005)

during the Late Glacial and Holocene, was more widespread. Finally, dunes are virtually absent in the central-northeastern plateau area; this region is characterized by a relatively dense river network incising deep into the plateau through regressive erosion that is presumably initiated by fault movement along the Feldbiss Fault. Floodplain processes in the valleys, and runoff processes on the interfluves prevented the development of dune systems here.

Forest clearing and intensified land use during the Neolithic gradually destabilized the landscape again, which led to reactivation of aeolian processes and the formation of drift sand dunes (Paulissen 1984). Drift sand dunes can often be interpreted as secondary features associated with the original dune form. Further to the west, off the plateau, the majority of these drift sands is thought to have been deposited during the last millennium (Derese et al. 2010; Beerten et al. 2014). Reforestation and leveling during the late nineteenth and early twentieth century stabilized the landscape again. Today, only few dune areas are still active.

12.3.2.4 The Relation Between Fens and Subsoil Characteristics

The typical fens that are distributed all over the Campine Plateau landscape are rainwater-fed bogs that can be typified according to their position in the landscape (Paulissen 1984) (Fig. 12.10). One type is typical for the southern part of the Campine Plateau, southeast of the Bosbeek valley, where groundwater levels are very low, down to 20 m and sometimes even almost 40 m below the surface. This type of bog developed where aeolian deflation removed any fine sand material that could have been accumulated on top of a hydraulically impermeable clay or silt bed in the top layers of the Zutendaal Gravel. Consequently, a shallow lake could develop in this depression in which peat could accumulate. Another fen type is typical for the northwestern part of the plateau, northwest of the Bosbeek, where groundwater levels are much shallower, less than 5 m deep. Here, the hydraulic barrier developed as an iron-rich B-horizon in response to soil formation processes in blowout hollows. During episodes with elevated groundwater level, bog iron could further develop from precipitation of dissolved iron in seepage water, enhancing the role of the hydraulic barrier.

12.4 Geomorphological Highlights

12.4.1 Escarpments

The shape of the Campine Plateau is determined by three prominent escarpments, with different origins. The southwestern escarpment developed in response to the formation of a pediment. This pediment forms the transition from the Campine Plateau to the river valleys that drain the Scheldt basin (Fig. 12.11). The top of the plateau is studded with residual gravels hampering erosion at this location. The total vertical difference between the plateau and the Demer floodplain that drains the pediment in the west is ca. 50 m. The escarpment itself is ca. 15 m high, while the pediment surface shows a regular dip from ca. 60 to 30 m. The steepest slope is less than 10% over a horizontal distance of 50 m.

The total height difference between the plateau and the Meuse Valley in the east is also ca. 50 m. However, the morphology of the eastern escarpment is completely different from the western one (Fig. 12.11). Much of the height difference takes place over a horizontal distance of less than 1 km, causing the eastern slope to be much steeper than the western one. The steepest slope sections can easily reach 20% over a distance of 50 m. The remaining 5–10 m height difference is accommodated by river terraces that are separated by small escarpments, usually only several m high, until the actual floodplain is reached.

The northeastern escarpment (Bree Fault Scarp) is tectonic in origin, and can clearly be observed near the town of Bree (Fig. 12.11). Its morphology takes an intermediate position in between the western and eastern escarpments. About two-third of the 30 m vertical height difference between the Campine Plateau and Bocholt Plain occurs within a horizontal distance of 1 km. The remaining 10 m is due to the accumulation of slope deposits at the foot of the escarpment. The steepest part of the slope may reach 10–15% over a horizontal distance of 50–100 m.

12.4.2 Development of the Hydrographical Network on the Campine Plateau

The water divide between the Meuse (east) and Scheldt (west) river basins runs through the town of Waterschei (which literally means ‘Waterdivide’). It is situated closer to the western boundary of the plateau than the eastern one (Fig. 12.12). Drainage towards the north and northeast developed through small brooks and rivers that penetrated deep into the plateau. Brooks that drain in northward direction have a very low gradient and run in very shallow floodplains. Other brooks and rivers make a distinct bend towards the northeast when passing the Bree Fault Scarp along the northeastern edge of the plateau. Although their upper courses trend N–S, they seem to have been responding to tectonic activity along the Feldbiss fault, finding a new course straight into the sinking Bocholt Plain. The gradient and valley depth are larger than those of the northwards flowing rivers, because of the larger and more abrupt height difference between the Bocholt Plain and Campine Plateau along this fault segment. The most striking feature is the

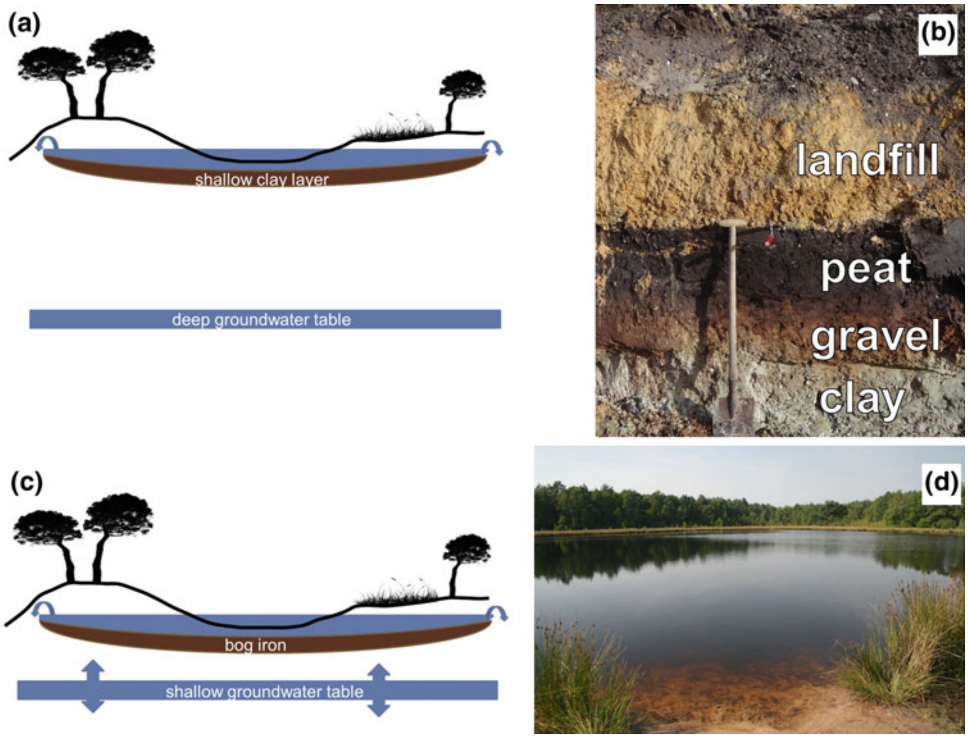

Fig. 12.10 Fen development on the Campine Plateau. **a** Deep groundwater table and a shallow impermeable clay layer. **b** Cross-section through the clay layer and overlying fen deposits of a drained fen (Nieuw Homo, Lanklaar). **c** Shallow groundwater table and impermeable bog iron. **d** Fen developed as a result of the formation of impermeable bog iron (Ruiterskuilen, Opglabbeek). (**a** and **c** from Van Uytven and Dreesen 2015; *photographs* D. Van Uytven)

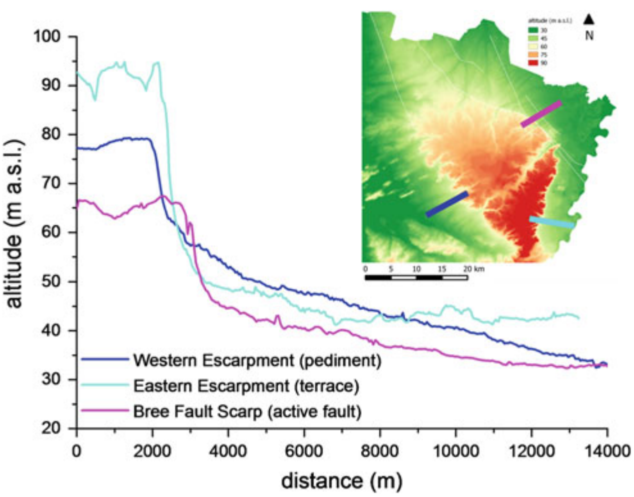

Fig. 12.11 Topographic cross-sections of the escarpments bordering the Campine Plateau on its SW, E and NE sides, respectively. Note the contrasted styles of the escarpments of different origins (SW: cryopedimentation of the Beringen-Diepenbeek Glacis; E: fluvial erosion by the Meuse; NE: Bree fault scarp)

Bosbeek valley, also running southwest-northeast. It has cut through the Zutendaal Gravels and now drains the Neogene substrate (Fig. 12.13).

The Bosbeek valley shows a marked asymmetry in the slopes that border the plateau edges. The evolution of the northwestern slope of the valley (i.e., facing southeast) can be understood in the frame of pediment development under periglacial conditions during the Saalian glaciation (Gullentops et al. 1993). This slope would have functioned as a snow trap for northern winds that would continuously blow in snow from the flat and bare plateau area. Long-lasting snowmelt during summer would have increased active layer processes accelerating slope retreat, the effect of frost cycles breaking up cobbles to transportable dimensions, and the volume of melt water evacuating the debris. These processes finally resulted in the formation of a cryopediment on the west bank of the Bosbeek. The continuous supply of coarse debris from the development of this cryopediment forced the river to flow along the eastern valley side leading to continuous undercutting and erosion of the east bank. This explains why the Bosbeek valley escarpment is about 30 m high along the southeastern valley wall, and only about 10 m along the northwestern valley wall (Fig. 12.13). The northwestern valley edge is drained by perennial tributaries that penetrated into the plateau quite deep, while the southeastern edge is drained by narrow and rather steep dry valleys. This phenomenon is directly related to plateau uplift and fluvial incision by the Meuse, which led to drastic groundwater table lowering along the eastern side of the plateau (Fig. 12.13). As a result, groundwater table is higher along the northwestern valley wall of the Bosbeek compared to the southeastern one. Differences in groundwater table depth also explain why the As Soil developed much more intensely east of the Bosbeek.

The eastern plateau escarpment is drained by deeply incised and very small dry valleys, with a very large gradient. They developed as a response to uplift-triggered incision

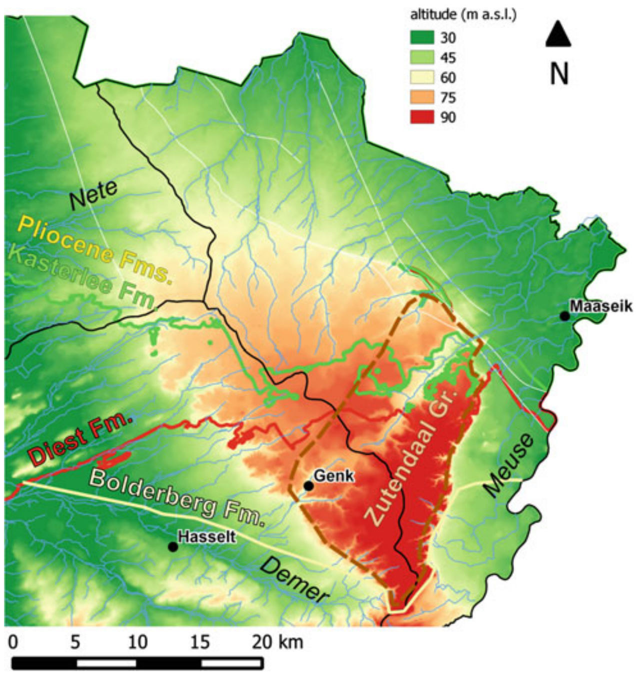

Fig. 12.12 Drainage network of the Campine Plateau, opposing long wide and shallow valleys in the north and west and relatively short narrow deep (and dry) valleys in the east and south. This reflects contrasting conditions of groundwater table (deep in the southeast and shallow in the northwest), and resistance to erosion of the substrate. *Thin white lines* are Quaternary faults and *thin black lines* the limits of the Meuse basin and Nete and Demer catchments (both belonging to the Scheldt Basin) in Belgium. The distribution of Zutendaal Gravels is indicated by a *brown dashed line*, while the southern limits of the Bolderberg, Diest, and Kasterlee and various Pliocene formations are shown as *solid lines of different colors*

of the Meuse river. This led to a significant lowering of the groundwater table, resulting in dry valleys. Their main phase of development is situated in the Saale and earlier (Gullentops et al. 1981). Only in the southern part, where the Boom Clay subcrops below the Zutendaal Gravels, some valleys are drained by perennial groundwater-fed streams (Fig. 12.12). The erosion-resistant role of the Zutendaal Gravels probably hampered the development of these eastern escarpment valleys.

The western escarpment is drained by less deeply incised rivers belonging to the Scheldt basin (Fig. 12.12). They developed as a result of regressive erosion, which finally led to the relief inversion of the Campine Plateau, and penetrated into the plateau edge. This process is also associated with the development of the BDG. In contrast to the valleys draining the eastern escarpment, these valleys are permanently drained by small rivers and brooks. The main reason for this is the relatively high groundwater table compared to the more intensely uplifted eastern part of the plateau. Along this western edge, two trends can be observed in the valley shapes, going from south to north. First, the incision depth and overall gradient decrease in this direction as a result of the differential uplift of the plateau, which is larger in the south than in the north. Second, the width of valleys draining the escarpment increases from south to north, because of the successively less erosion-resistant outcropping sediments, from Zutendaal Gravels in the south to Diest Sands in the middle, and Kasterlee, Mol, and Rhine sands in the north (see also Fig. 12.4). A prominent bend in the western escarpment is visible in the north. The exact cause of this

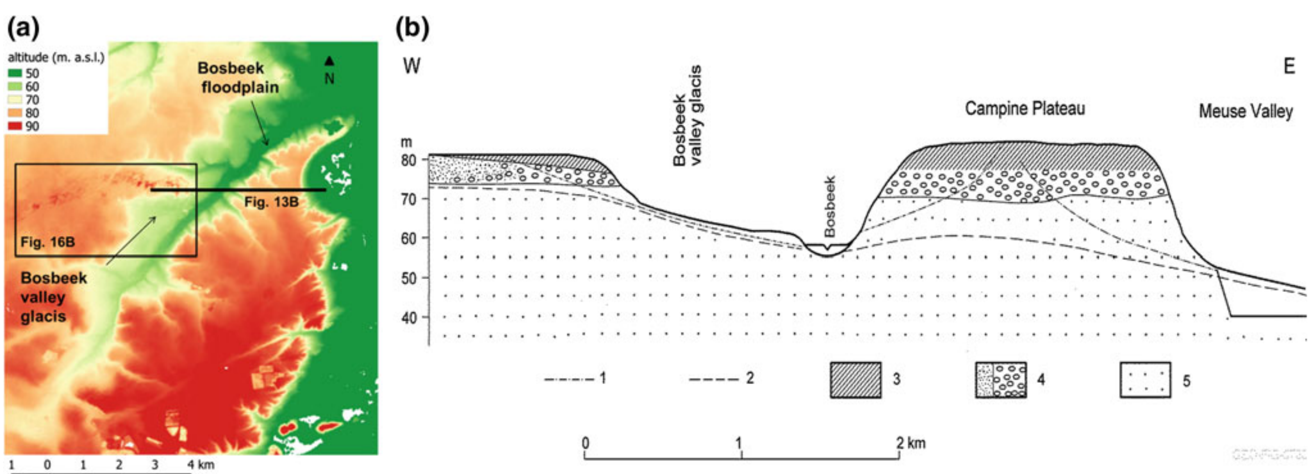

Fig. 12.13 **a** DTM of the Bosbeek valley, showing a steep southeastern valley side and a smooth northwestern side developed as a cryopediment during the Saalian glaciation. **b** Schematic cross-section through the Bosbeek valley and its divide with the Meuse (redrawn after Gullentops et al. 1993). *1* Projected longitudinal profile of tributaries; *2* present-day groundwater table; *3* as soil; *4* sands and gravels of the Meuse; *5* neogene sands

bend is not clear, but it is probably related to the presence of the Rauw fault, the difference in resistance to erosion between the indurated Diest Sands in the south and the Mol Sands in the north, and the role of regressive erosion in the Nete catchment (see above).

12.4.3 Preserved Natural Resources: Meuse Gravels, Rhine Sands, Mol Sands and Bolderberg Sands

12.4.3.1 Meuse Gravels

The Zutendaal Gravels run in a southwest-northeast direction as a 5 km wide strip on the eastern side of the plateau, eroding the underlying Winterslag Sands and Neogene units. The gravel sheet is up to 15 m thick but is thinning towards the north and south. Extent, thickness, and elevated position of the gravel unit make it a unique morphological feature in Western Europe. The Zutendaal Gravels are characterized by a basal channel-lag deposit consisting of cobbles and larger (ice-rafted) boulders (De Brue et al. 2015) (Fig. 12.14a). The gravel deposits form an aggrading unit, deposited during different aggrading cycles, each cycle consisting of a coarser channel-lag deposit at the base, fining up to fine gravel or even a sand layer or a clay lens at the top (Gullentops et al. 1981; Paulissen 1983). Each cycle signifies deposition in a broad shallow channel. Usually, the gravels are matrix-supported by sand or finer material. They are generally interpreted as a cold-climate braided river deposit. Their characteristics are completely different than those of the Holocene Meuse, which displays graded point bar gravels, several meters thick, capped by silty and clayey floodplain deposits, 2–3 m thick. During deposition of the Zutendaal Gravels, sediment supply was guaranteed by frost weathering during cold periods, under periglacial circumstances. Subsequent snowmelt during early summer caused peak discharge in the tributaries and main channel, promoting the transport of very coarse material.

De Brue et al. (2015) investigated the possibility of a classical fluvial transport mode for the largest boulders—most of which exceed 50 cm and may reach up to 200 cm in intermediate axis and some of them weighting several tons—on the base of empirical hydraulic transport thresholds for gravel bed rivers. Results indicate that hydraulic transport of boulders with intermediate diameters <1 m could have occurred within limited reaches of the palaeoriver, more specifically in the palaeo-Amblève tributary. However, the small slope gradient of the palaeo-Meuse most probably inhibited boulder movement by hydraulic forces only.

Therefore, ice-rafted transport of large boulders is favored here, referring to the existence of recent analogues ("blocs glaciers") in the estuary of the St-Lawrence river of Québec (Dionne 2003) and according to observations and calculations of the transport capacity of ice in the catchment area of the Amblève river by Collard et al. (2012, see also Chap. 6). Furthermore, huge blocks of rock types sensitive to physical weathering, such as phyllites and microgranites (often affected by diaclases), would never have reached the Campine area by normal fluvial transport without breaking into smaller pieces.

The composition of the gravel undoubtedly points to a southern origin, in particular the Ardennes and the Vosges area (northern France). Therefore, the Meuse and its tributaries (including also the Moselle) can be considered as the only transport agent for these gravels. The Zutendaal Gravels display a broad lithological spectrum of weathering-resistant rock types, derived from various Palaeozoic through Meso-Cenozoic geological sources. These include various hard Cambrian quartzites and vein quartz, Lower-Middle Devonian conglomerates and sandstones, Lower-Carboniferous phtanites and silicified oolitic limestones, Cretaceous flints, Tertiary quartz arenites (silcretes) and flint gravels. The striking absence of Palaeozoic and Mesozoic limestone pebbles or boulders is related to chemical weathering before or after transport and deposition of the gravels. Exceptionally, Cambrian-Ordovician phyllites, Late Devonian microgranites (e.g., the "porphyre de Mairupt") and diabases occur within the Zutendaal Gravels as huge ice-rafted boulders. The latter igneous rocks have been derived from dyke swarms in the Rocroi area, SW Ardenne. Finally, pebbles of Palaeozoic granites and Triassic Buntsandstein represent some minor but most characteristic constituents that point to a particular provenance area—the Vosges—and hence transport via the Moselle River (Bosch 1974, 1992; Dreesen et al. 2014). Flint, derived from upstream Cretaceous chalk formations, is also a very typical component.

12.4.3.2 Rhine Sands

The Rhine facies of the Campine Plateau consists of medium to coarse sand, with low clay and silt content, and occasionally containing fine and coarse gravel (Paulissen 1983). Sedimentation took place in wide and shallow channels, probably in a braided river system during glacial stadials or interstadials because a typical fine-grained floodplain facies is absent in these deposits (Fig. 12.14b). Usually, the deposits are arranged in fining upward several-meters-thick sequences. Large boulders and blocks are lacking. Rhine deposits can be found as a deep southwestern bend, covering the western and northwestern part of the Campine Plateau. Along the western edge of the plateau, the sands are eroded and preserved as a residual gravel layer, often containing ventifacts (van Mechelen 1982). Rhine deposits occasionally contain gravel that originates from the Alps, the German middle mountains and the Ardennes-Eifel massif. Gravel, and most notably granule gravel (2–4 mm), is rich in quartz, usually above 60%, while flint fragments are very rare.

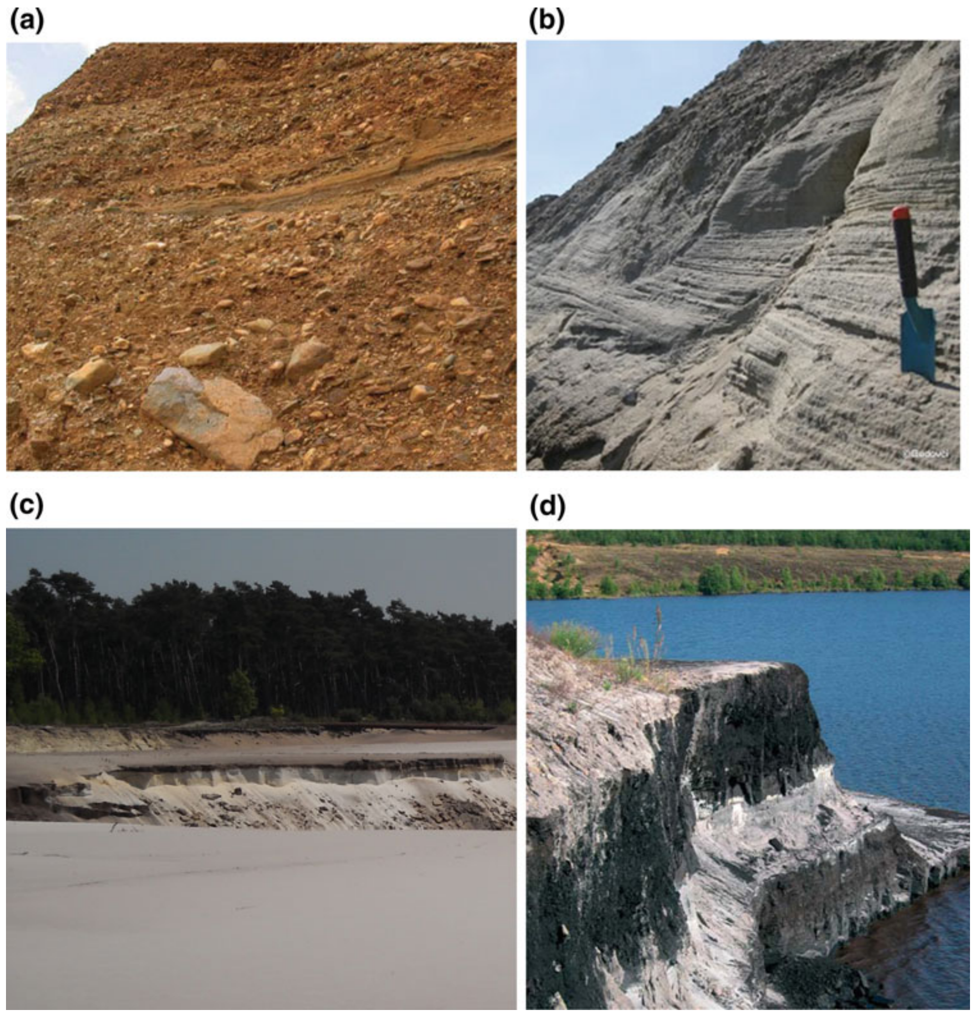

Fig. 12.14 Natural resources in the Campine area. **a** Coarse matrix-supported gravel deposited by the Meuse and weathered into the As Soil; the length of the longest axis of the biggest boulder is ca. 0.6 m ('Kikbeek' quarry, Maasmechelen; reproduced from Dreesen et al. 2005). **b** Coarse sand deposited by the Rhine, no obvious weathering ('Blauwe Kei' quarry, Lommel; *photograph* M. Gedeon). **c** White quartz sand and lignite of the Mol Formation (Blauwe Kei quarry, Lommel; *Photograph* K. Beerten). **d** Outcrop in the Bolderberg Formation showing alternations of bleached quartz sand and lignite (Kikbeek quarry, Opgrimbie; reproduced from Dreesen et al. 2005)

These deposits are probably laterally equivalent to the Winterslag Sands, which occupy more or less the central part of the plateau, running southwest-northeast along a 10 km wide zone (Fig. 12.5).

12.4.3.3 Mol Sands

In the northwest, the Rhine sands overlay a thick unit of white to brown relatively pure quartz sands with two distinct lignite layers. These sands and lignite layers are known as the Mol Formation (Fig. 12.14c). They were deposited during the Pliocene in a large estuary fed by the rivers Rhine and Meuse, as attested by the presence of "kieseloliths" (pebbles of silicified oolitic limestone) near the base of the formation. During periods of graben subsidence, thick peat layers developed in a swampy environment, and were transformed into lignites during subsequent burial under quartz sands. The sands in the Mol Formation are fine to medium-grained, while clay particles and glauconite are virtually absent. Post-depositional bleaching of the sand as a result of dissolved organic acids in the infiltrating rain water gave the sands their pureness and white appearance, such that is very suitable in the glass, ceramics, chemical and metallurgical industries (Gullentops and Wouters 1996).

12.4.3.4 Bolderberg Sands

On the southeastern part of Campine Plateau, the Zutendaal Gravel blanket overlays about 40 m of fine-grained white to slightly yellowish quartz-rich sands (Fig. 12.14d). These sands belong to the Lower Miocene (Burdigalian) Bolderberg Formation and correspond to shallow-marine deposits in a North Sea bay bordering the Roer Valley. The source material probably originated from the weathered and therefore quartz-rich material at the margins of the southern North Sea Basin in Belgium and adjacent areas. Their uniform fine grain-size distribution (average 200 µm), good sorting, as well as mineralogical composition (slightly glauconite-bearing) and the presence of herringbone cross-bedding, point to longshore transport, coastal sorting by tidal currents in an estuarine depositional environment (Gullentops 1973). The Bolderberg Sands display a threefold subdivision due to a continental phase interrupting the shallow-marine sand sedimentation. During a short withdrawal of the sea, extensive coastal marshes and back-swamps developed (analogous to the actual Everglades in Florida or the Okefenokee swamps in Southern Georgia, USA) leading to the formation of the so-called Kikbeek Lignite (averaging 3 m in thickness). This lignite is mainly composed of driftwood although swamp trees grew in place, as evidenced by numerous roots that penetrate the underlying sands. This seam is of the same age as the lignite seams of the Roer Valley in the Netherlands (Heksenberg Member of the marine Breda Formation) that correlate well with the well-known thick lignite deposits exploited for fuel in the Garzweiler, Frimmersdorf and Hambach quarries of the Lower Rhine Basin in Germany (Van der Meulen et al. 2009). Early-diagenetic leaching by humic acids originating from the peat affected the underlying sands: so, non-quartz components were dissolved resulting in the creation of the famous Miocene glass-sands or "silversands" of Maasmechelen. These worldwide reputed glass-sands are still extensively quarried on the Campine Plateau because of their pureness (iron content is less than 0.01% Fe_2O_3, Al_2O_3 content less than 0.025%). Moreover, this extraction activity was also responsible for the creation of deep extraction pits, resulting in the creation of important geological outcrops that have been recently reclaimed into highly valuable nature reserves and important geoheritage sites (Dreesen et al. 2005).

12.4.4 Polygonal Soils and Cryoturbations

In the uppermost part of the Zutendaal Gravel, right below the Late Pleistocene-Holocene aeolian sands, spectacular examples of a particular type of cryoturbation occur (Fig. 12.15). In section, these features look like vertical tongues or flames ("involutions") up to 2–3 m in length, whereby the pebble axes are vertically oriented. In plane section (at the surface), these tongues form a characteristic network of large polygons with diameters of less than 1 m up to several meters. The core of these gravel polygons is essentially made of sandy material, and their center is slightly raised (Fig. 12.15b). According to Paulissen (1970), this patterned ground type deformation (called "sorted polygons") is characteristic of periglacial areas undergoing permafrost conditions. These peculiar forms most probably originated within the active layer of the permafrost. Frost heave is generally seen as the main mechanism behind the formation of sorted patterned ground (Pissart 1987; Van Vliet-Lanoë 2014; Yamagishi and Matsuoka 2015). Dependent on grain size (fine sands and silts are most frost-susceptible), freezing rate, and available moisture (ensured where a perched water table is present above the frozen ground during summer), lenses of segregation ice develop gradually within the active layer with seasonal freeze-thaw cycles, causing soil heave and erecting elongated gravel (Pissart 1969). While differential frost heave, mainly related to lateral variations in grain size or soil moisture availability, may be responsible for the polygonal pattern, sometimes also guided by a net of desiccation cracks, the contrast between gravel involutions and cells of finer material is attributed to frost sorting, vertical and lateral, by freezing and thawing. These spectacular polygons bear witness to the harsh (cold and dry) climatic conditions that prevailed in Campine area during the last ice ages, particularly during the Late Weichselian (ca. 20 ka). They are currently characteristic of arctic regions or tundra areas like those located around the Arctic Circle (Canada, Spitsbergen, Greenland).

12.4.5 Late Glacial and Holocene Dunes

Among the most striking landforms topping the Campine Plateau are parabolic and longitudinal dunes. They can easily be identified on DTM images (Figs. 12.1 and 12.9) and historical maps. The dune complex of Hoeverbergen (Fig. 12.1c) was originally studied by Gullentops (1957). It is a parabolic dune, ca. 1 km wide (between tips), ca. 2 km long (along the dune axis) and up to 10 m high. Hand augerings allowed for reconstructing the original surface by mapping the palaeosoil that stabilized the dune, in response to climatic amelioration and vegetation development. The dune itself probably developed during the Late Glacial (Younger Dryas), since it is underlain by a bleached horizon, which is interpreted as the Usselo soil of Allerød/Early Younger Dryas age (Derese et al. 2012). The dominant wind direction during dune development was from the southwest as is evidenced by the steep dip angle (26°) of individual sand beds at the northeastern leeward side of the dune (Fig. 12.16a). The soil that is preserved inside the blown-out surface is described as a hydromorphic soil, such that deflation is supposed to have continued until the water table was reached. The northeastern tip of the dune is blown out at a later stage during landscape development, the blowout sand being deposited towards the east as small isolated dunes 1–2 m high. A similar dune complex is present near Kattenbos and Maatheide (Lommel). The core of the dune at Maatheide is shown to contain the Late Glacial Usselo soil (Derese et al. 2012). Holocene reactivation of the Kattenbos dune as a result of agricultural activity (deforestation, herding, repeated use of tracks for herding, cutting of pieces of heath with roots and humus, called plaggen, as bedding for cattle) caused newly eroded sand to drift onto the older

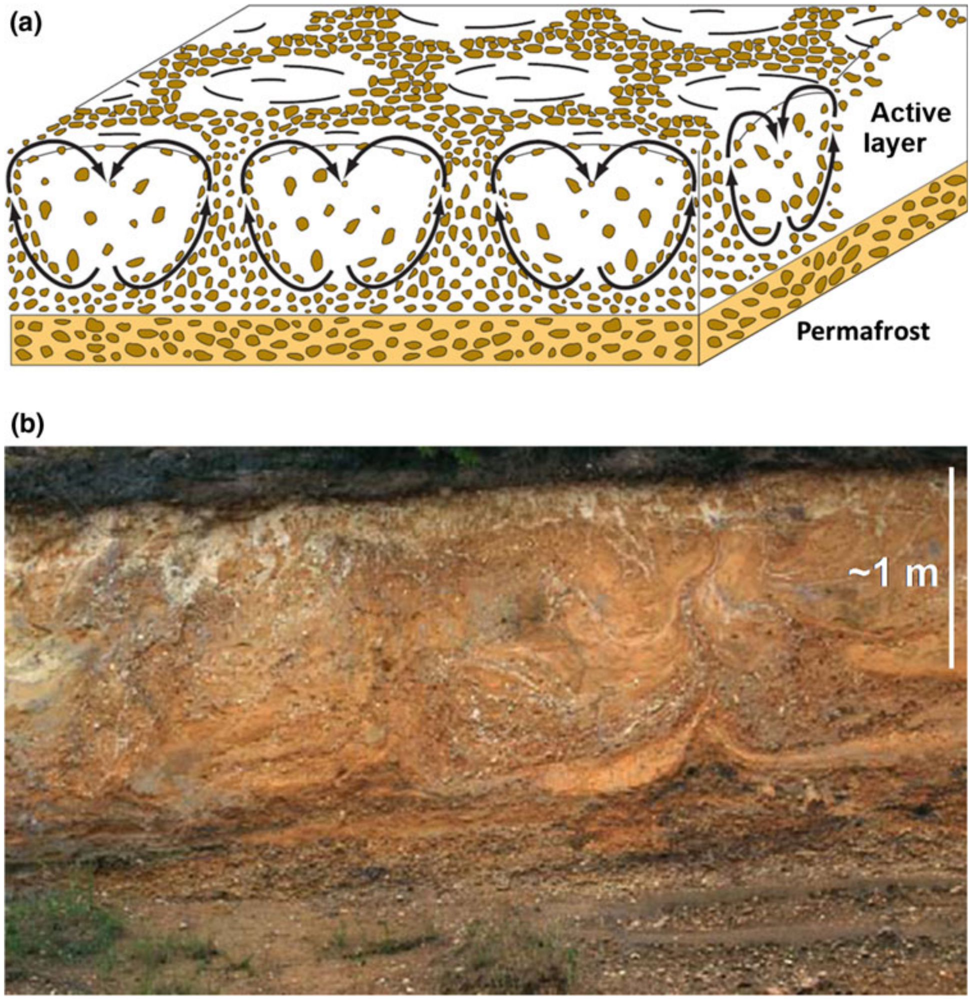

Fig. 12.15 **a** Development of large cryoturbation structures in the active layer above a permafrost table in gravel deposits, according to Paulissen (1970) (modified from Dreesen et al. 2005). **b** Polygonal soil pattern in the top layers of the Zutendaal Gravels, with large-scale involutions and flames containing vertically aligned gravel ('Salamander' quarry, Maasmechelen; reproduced from Dreesen et al. 2005)

dune surface posterior to the development of a podzol soil (Fig. 12.16b).

A different type of dune complex is the one on the west bank of the Bosbeek, called the "Duinengordel" (Fig. 12.16 c). It consists of an array of small parabolic dunes, ca. 200 m wide, 200–400 m long (along the dune axis) and up to 15 m high. The complex itself is about 8 km long, and more or less oriented east-west. Typically, the northeast-facing leeward slope of the individual dunes is much steeper than the southwest-facing slope, indicating that they developed under influence of southwesterlies. The dunes are considered to be relatively young. On the one hand, the absence of Late Glacial palaeosoils in their core suggests that they did not develop from a Late Glacial dune, as is the case in Lommel. On the other hand, traces of soil formation on the dune top are absent, suggesting that they are still active, especially in those areas where vegetation is lacking. A particular feature in this respect is the dune top of the Oudsberg, the highest top of the dune complex (Fig. 12.16d). Aeolian deflation facilitated by over-recreation of the dune led to a system of excavated tree roots. The development of the dune complex is probably related to deflation from ploughed fields south of it.

About 200 circular depressions with an average diameter of 140 m and rims averaging 2.5 m in height have been discovered by using panchromatic and LIDAR-imaging (Laser Imaging Detection And Ranging) in the Bocholt Plain, NE of the Campine Plateau. Field surveys demonstrated that these landforms (with low conservation potential) actually represent blowouts or lunettes (Dusar et al. 2008). Their exact age is still not clear but most probably they were formed as a result of deforestation during the early Neolithicum.

12.4.6 Groundwater Silcretes

About 1.5 m below the base of the Kikbeek lignite seams, ash-white quartz arenitic sandstones occur (the so-called

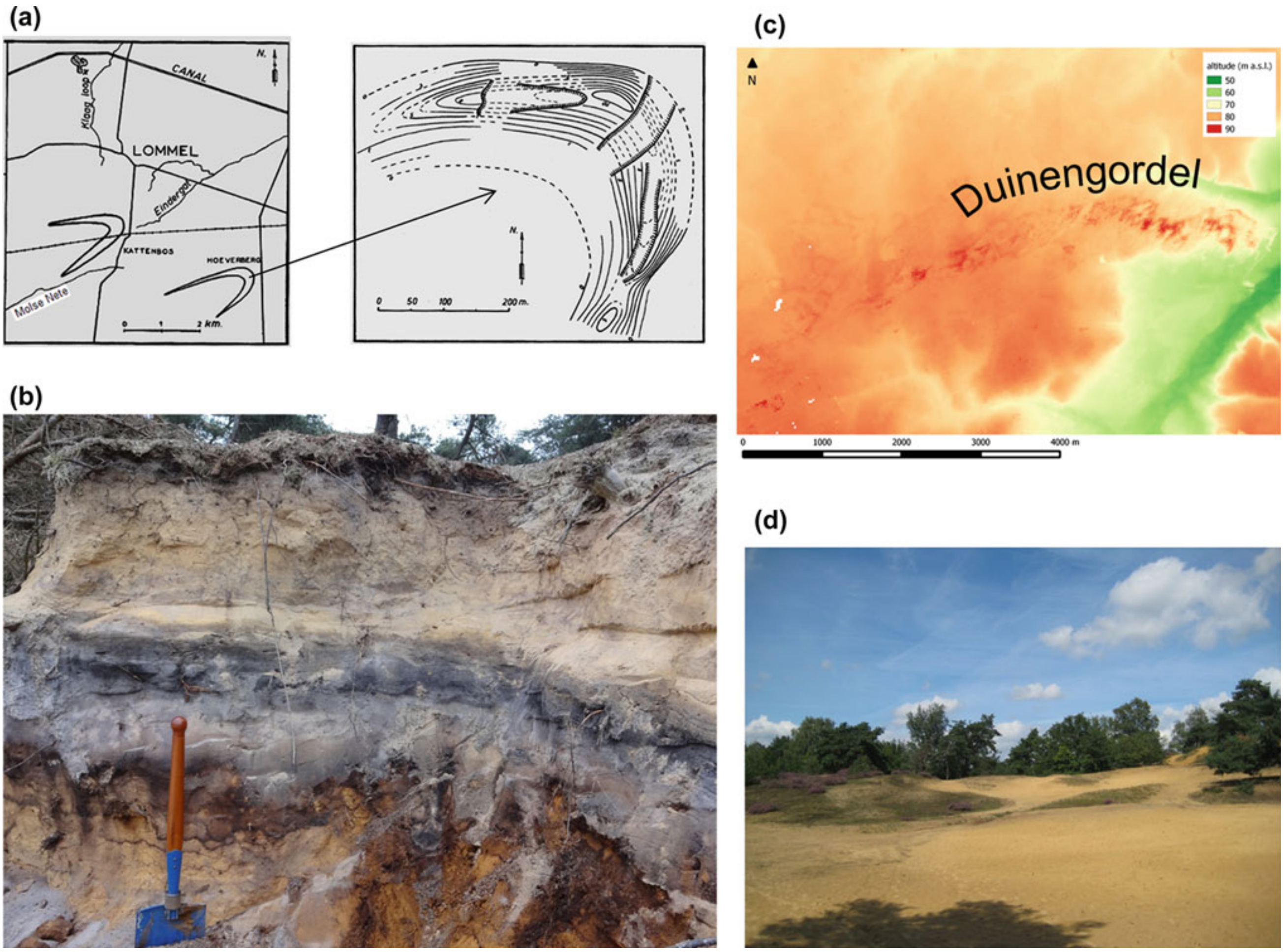

Fig. 12.16 **a**. Detail topographic maps of the Hoeverbergen dune (from Gullentops 1957). **b**. Typical profile showing drift sand migrating over a podzol soil developed on top of the Kattenbos dune near Lommel (*photograph* D. Van Uytven). **c** Detail DTM of the 'Duinengordel' dunes. **d** The Oudsberg, the highest summit of the dune system (*photograph* D. Van Uytven). See also Fig. 12.1c for the location of the Hoeverbergen and Kattenbos dunes

Bolderian Sandstones) in the bleached Bolderberg sand, forming thick plates or thick lenticular beds (Gulinck 1961) that reach volumes of several m^3 (Fig. 12.17). These silcretes often display particular organic or anthropomorphic forms, leading to their naming "*grès mamelonné*" in French. Silicification of the sand, often incomplete, commonly occurs through quartz overgrowth on the quartz grains. The silcrete blocks contain small circular pits at their upper surface and vertical root traces, pointing to the presence of a peat (lignite) layer above. Although different stratigraphic levels display silicification in the subsurface of the province of Limburg, silcretes most frequently formed during continental episodes of the Thanetian (Upper Paleocene) in Belgium and Northern France (Dreesen and Dusar 2012). Silcrete formation is related to post-depositional groundwater circulation and the release of silica accompanying acid leaching of the host material or adjacent strata (Ullyott and Nash 2006). Demoulin (1990) distinguished several types of silicified Tertiary sandstones in Southern Limburg (NL) and at the northern border of the Ardennes (Hautes Fagnes and Pays de Herve). The Miocene quartz arenites of the High Campine area belong to Demoulin's Nievelstein type, typically associated with lignite deposits. The Miocene silcretes show strong analogies with the sarsen stones of southern England and the "Braunkohlequartzite" of the German Lower Rhine area. Moreover, they have been locally used as a vernacular building stone in the province of Limburg (Dreesen et al. 2001). Finally, when some of these huge, massive stones (megaliths) became unearthed through erosion of the enveloping sand, they have been locally used as a grinding stone for flint tools during prehistoric times, such as the sandstone blocks of the Holsteen, in the Zonhoven area (Huyghe 1990).

Fig. 12.17 Blocks of groundwater silcrete indurating Bolderberg sands at the Holsteen, near Zonhoven (Dreesen and Dusar 2012)

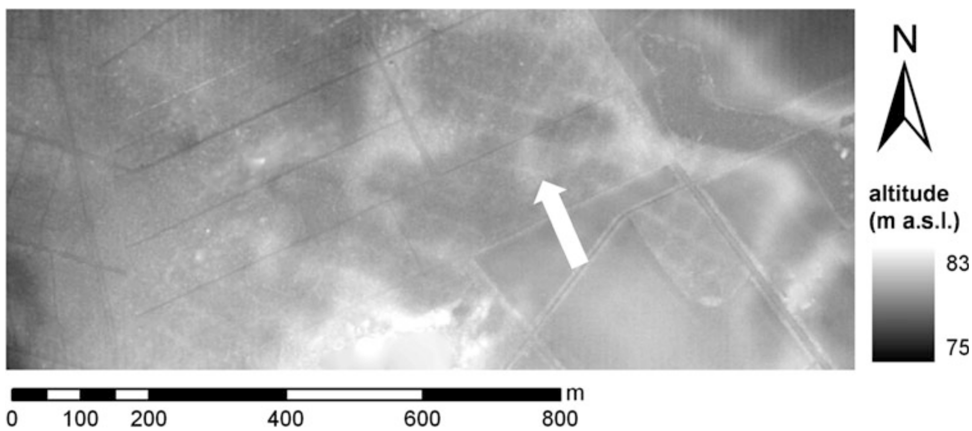

Fig. 12.18 DTM of the Muisven-Ophoven Celtic Field complex (Meeuwen-Gruitrode, see location in Fig. 12.1; modified after Creemers et al. 2011). The squared pattern of the Celtic Field is indicated by an *arrow*

12.4.7 Celtic Fields

Detailed LIDAR images often show a striking pattern of slightly elevated banks, arranged in a multiple crossroad pattern. They are called Celtic Fields and characterized as 'a prehistoric parceling system of which the parcels are completely surrounded by low banks consisting of stones, sand or a mixture of the two, and laid out for agricultural purposes' (Creemers et al. 2011). The banks may only be 20–30 cm high, and the parcels which they surround typically measure 40 × 40 m. A representative example is the Celtic Field of the Muisvenner-Ophovenerheide in Meeuwen-Gruitrode (Fig. 12.18). The fields were used to cultivate, among others, emmer wheat (*Triticum dicoccum*) and gold-of-pleasure (*Camelina sativa*). The term Celtic Fields is somewhat misleading: the Celts were initially thought to have developed this agricultural system, but later investigations showed that the fields date back from as early as the late Bronze Age. The fields were abandoned during the Roman period and developed into heathland until the region became reforested by pine during the late nineteenth and early twentieth century. Their location, as well as that of Mesolithic settlements, seems to be related to the presence of nearby fens and wetlands (see above).

Fig. 12.19 Overgrown coal mining tip ("terril") of Eisden. Groundwater crops out in abandoned gravel pits (*Photograph* M. Bex)

12.4.8 Impacts of Coal Mining on the Geomorphology of the Campine Plateau

Coal mining tips (spoil heaps or waste dumps, named "*terrils*" or "*steenbergen*") represent enormous beacons in the Campine landscape and they are regarded as true sociocultural heritage (Fig. 12.19). They represent silent witnesses of the luxurious tropical rain forests that once ruled the area during the Late Carboniferous, generating numerous coal layers in the deep subsurface of the Campine basin. Their discovery in 1901 meant the beginning of a coal rush. From the first coal production in 1917 until the closure of the last colliery in 1992, over 441 million tons of coal have been produced from seven different collieries, with an equivalent amount of coal waste. The latter mainly consists of mudstone (over 70%) with minor amounts of siltstone, sandstone, coal, and carbonaceous shale. After final closure of the collieries, the coal mining tips and coal sludge basins (liquid waste generated by washing of the crushed coal) were partly reclaimed from 1993 to 2000, because of environmental regulations. Reclaiming measures concentrated on remodeling the dumps, aiming at stabilizing the slopes and avoiding acid coal mine waste drainage. In order to further reduce erosion (e.g., through gully formation and debris sliding) greening of the dumps was carried out by hydroseeding. Due to the latter remodeling, the original outline, shape and location of the dumps have been considerably changed (Fig. 12.20). The coal mining tips originally reached between 70 and 110 m in height. Although they represent enormous reserves of secondary raw materials for the ceramics industry (Dreesen et al. 2008), the ones that have not been remodeled or reclaimed have become nature reserves, hosting thermophilic and halophilic fauna and flora.

Moreover, this deep coal mining activity caused important and extensive ground subsidence. Maps with estimated iso-subsidence lines have been calculated in collaboration with the Campine Collieries between 1981 and 1985 (Vansteelandt 1996) showing subsidence values exceeding 6–8 m (e.g., in the central Zwartberg–Waterschei area). This subsidence is related to the extraction of several superimposed coal layers and to strong groundwater (mine water) pumping (Fig. 12.21). Moreover, with the help of radar interferometry (PS-inSAR) using satellite data from 1992 through 2005, contrasting ground movements have been reported over the abandoned Campine coal mining area (Devleeschouwer et al., 2007). The collieries in the eastern part of the coal basin closed much earlier (1966–1987) than those in the western part (1992). As a result and in contrast with residual mining subsidence in the western part (−4 to −14 mm/year between 1992 and 2000), a distinct uplift has been recorded in the eastern part (+3 to +23 mm/year). This uplift apparently results from poro-elastic rebound of the eastern area when flooding of the underground mine working subsequent to the end of mine water pumping induced a pore-pressure increase. Meanwhile, also the western part of the Campine coal field has reversed to uplift mode (Vervoort 2016; Bejarano-Urrego et al. 2016).

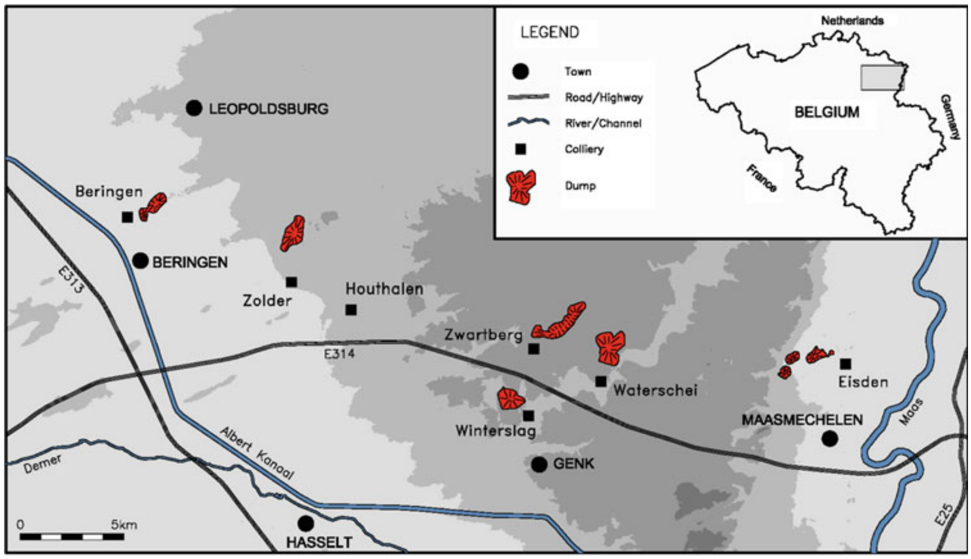

Fig. 12.20 Location map of the remaining spoil heaps (terrils) in the Campine area (modified after Dreesen et al. 2008)

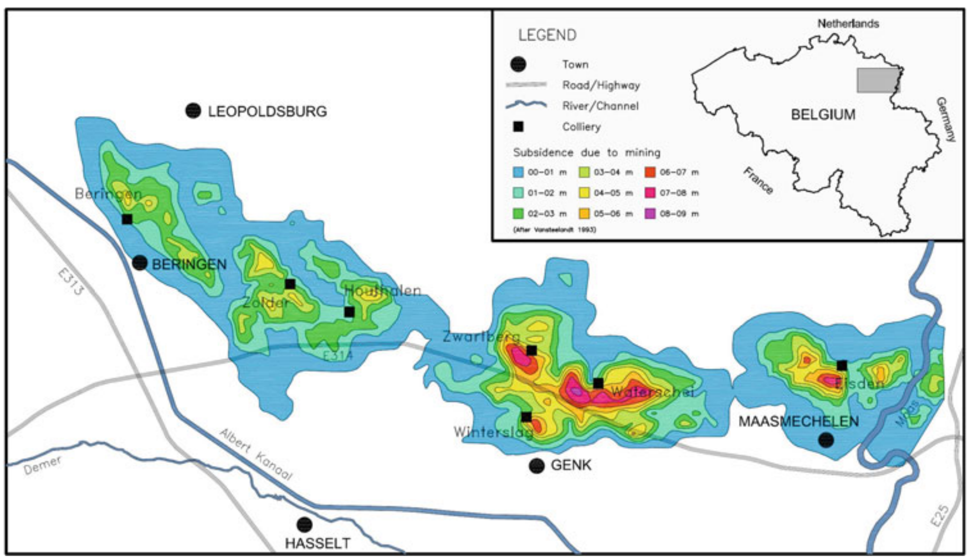

Fig. 12.21 Lines of iso-subsidence above the former coal mining areas in the Campine Plateau (*drawing* J. Matthijs, modified after Vansteelandt 1996)

12.5 Conclusions

In 2011, the Hoge Kempen (High Campine) Rural-Industrial Transitional Landscape (HKRIL) has been nominated for the Tentative list of UNESCO's World Heritage Sites. It bears unique geological and geomorphological witnesses of rather extreme global changes that once affected this area, grading from warm and humid tropical swamps to cold and dry permafrost conditions. Moreover, the effects of dramatic climatological changes that affect our living planet can clearly be demonstrated in natural or artificial outcrops that occur within its geographical boundaries. Abandoned sand and gravel pits in particular represent great geological windows for educational purposes: they offer unique opportunities to display the layer-cake structure and to unravel contrasting climatological conditions and depositional settings in which the exposed sediments and rocks have been formed. Several geotouristic initiatives have been launched by volunteer organizations (e.g., the Geological working group of LIKONA—the Limburgse Koepel voor Natuurstudie) stimulating outreach and triggering public awareness. The latter working group has been involved in the reclamation of abandoned glass sand and gravel pits in the Hoge Kempen National Park, where rather unique geological features are particularly well exposed, including spectacular Pleistocene polygonal soils and gravelly braided river deposits (with ice-rafted boulders up to 2 m in diameter) witnessing the harsh periglacial conditions during the youngest ice age. A particular dune complex represents a key reference section for the Late Glacial sandy aeolian deposits in the southernmost part of the

NW European coversand belt, because of excellent preservation of fossilized soils, rarely seen elsewhere.

Huge ice-rafted boulders embedded in braided river deposits display a broad lithologic spectrum representative of the geology of the upstream area of the Meuse river basin. Below this Pleistocene gravel blanket, pure white quartz sands occur, the origin of which is related to the presence of lignite formed during the Miocene in coastal marshes, analogous to the recent Everglades or Okefenokee swamps in the USA. The lignite as well as the rocks and fossils in the coal mining tips in the direct surroundings provide strong evidence for past global climate change. A selection of ice-rafted boulders, representative of different geological stages, were removed from the gravel deposits and set into a geological "rock garden" offering a new destination for geography school classes and their teachers. A colorful brochure allows the general public to identify pebbles that occur everywhere throughout the National Park and that represent its hallmark (Dreesen 2007). Recently a new geological biking route has been set up, explaining the complex relationship between subsurface, landscape, flora, and cultural history of the High Campine area (Van Uytven and Dreesen 2015).

The Campine Plateau is a unique geomorphological object and one of the best-kept examples in Europe of an enormous alluvial fan-like gravel and sand deposit formed during the Middle Pleistocene. Its present shape is the result of interfering erosional and tectonic processes. The southern tip, the western and northeastern boundaries are all (in)directly fault-related, this particular tectonic configuration controlling the fluvial development of the area around the Early-Middle Pleistocene transition. Once tectonic uplift had raised the Eastern Campine Block, that largely coincides with the Campine Plateau, the fluvial deposits were directly responsible for the preservation of the plateau as a morphological entity in topographic inversion. Aeolian processes actively reshaped the plateau surface during the Late Pleistocene and Holocene. Soil erosion and depletion of nutrients due to intensive grazing had to be countered by massive pine planation on top of the plateau. These could easily thrive on the relatively dry sandy and stony soils. Locally, clay or loam lenses on top of the gravel deposits entail completely different hydrological conditions such that rainwater-fed fens could develop. Coal extraction has led to specific anthropogenic geomorphic features, such as big spoil heaps and subsidence phenomena above the former Campine collieries.

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