The Hageland Hills, Legacies of the Depositional Architecture of the Miocene Diest Sands

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Abstract

The Hageland area is characterized by long, a few tens of metres high hills that form a remarkable array of parallel topographic features. The existing model stating that the Hageland hills are remnants of offshore tidal sandbanks formed during deposition of the Miocene Diest Formation should be abandoned. Also the role of ironstone beds as the primary cause of the hills is questioned. Instead, the hills are thought to be determined by elongated parallel coarser grained sand bodies that were embedded in a more abundant facies of slightly clayey, bioturbated fine sand inside the Miocene Diest Formation. A sedimentary model is presented that interprets the elongate coarse sand bodies as the fill of a series of parallel, linear depressions, each time scoured by locally constricted flow during the continuous lateral fill of a tidal marine embayment. The coarse sand bodies are thought to have been sculpted out as positive relief features much later, during Pleistocene uplift of the area. Their fabric and higher permeability are put forward as important causes of primary preservation during the landscape morphogenesis. The ironstones can only have formed after initial appearance of the long hills. Their presence may have contributed to preserve some hills during further uplift and erosion.

Keywords

Hageland hills • Diest Formation • Tidal scour-and-fill architecture • Tidal sandbank • Differential erosion • Neogene iron sandstone

14.1 Hageland with Its Typical Elongated Hills

Some remarkable long hills appear east of the wide Dijle river plain at Wezemaal, between Leuven and Aarschot (Fig. 14.1). The hills are arranged as parallel ridges. They are a few km long and less than 1 km wide (Fig. 14.1a) and have steep (Fig. 14.1b), often wooded slopes. The top surfaces are flat (Figs. 14.1 and 14.2) and either horizontal or faintly sloping. A gradual convex or sometimes abrupt transition to the flanks surrounds the flat top. The hills are covered by sandy soils and have long been used as extensive pasture land and heath. Many outcrops of brown iron-cemented sandstone can be found in the flanks and road cuts of the hills. The stone has been quarried and used as a building material for centuries (De Clercq et al. 2014). Abandoned quarries dot the flanks of the hills. Some hills where outstanding natural and historical values have been preserved are now protected as nature and heritage reserves, such as Wijngaardberg (Natuurpunt Oost-Brabant 2014) and Kesselberg (Goolaerts 2015).

Hills like those near Wezemaal typify the region named Hageland. The geographical region Hageland is the area between the cities of Leuven, Diest, Aarschot and Tienen (Fig. 14.3). The special geomorphology characterized by

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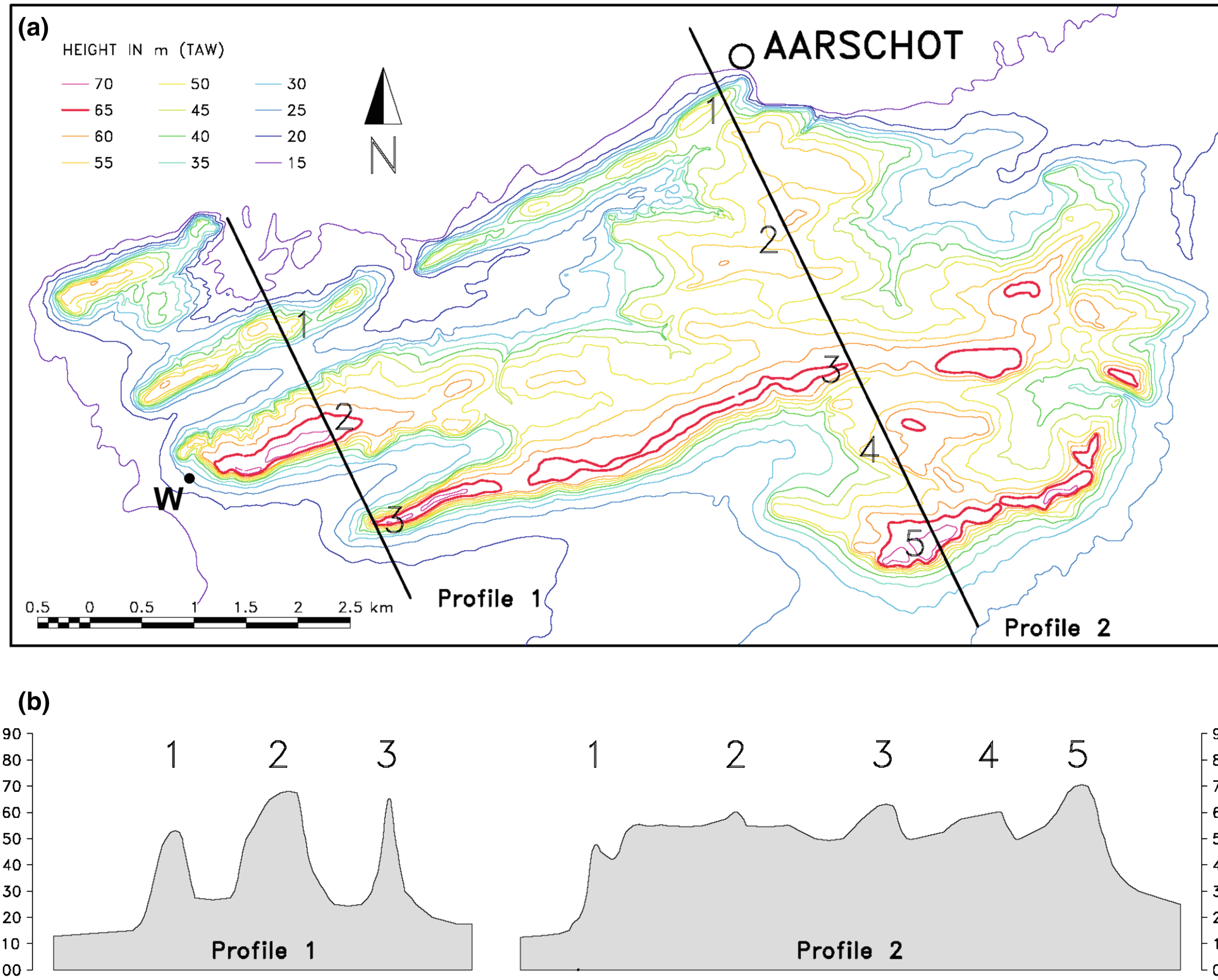

Fig. 14.1 Typical topography of a part of the Hageland area between Wezemaal (*W*) and Aarschot (see Fig. 14.3 for location). **a** Elevation map showing *coloured contour lines* in 5 m steps. Lowest altitude is 15 m near the river Demer in the north and highest is 75 m at the crest of Wijngaardberg (*2*) and Houwaartse Berg (*5*). Source of data: Grid version of the Topographic Map of Belgium at scale 1/10,000, map sheets 24/6, 24/7 and 24/8, NGI 1978–1993). **b** Cross-profiles *1* and *2* in Fig. 14.1. **a** *Note* the strong vertical exaggeration. In *profile 1*, hill *1* is Middelberg, *2* Wijngaardberg and *3* Beninksberg. In *profile 2*, hill *1* is Konijntjesberg, *3* Kratenberg near Nieuwrode, also shown in Fig. 14.4, *4* Benneberg, and *5* Houwaartse Berg

parallel long hills is found in the northern part of the proper Hageland although hills of similar appearance are also found further west (between Leuven and Brussels) and north (southern part of the Kempen (Campine) area). The term “Hageland” is used here in a wider sense to designate the area with elongated hills.

We examine the primary origin of this remarkable ensemble of hills, i.e. only the possible original causes of the topography but not its later geomorphological evolution. It is believed that, once the hills appeared as low positive landforms, further uplift of the country and erosion made the hills stand out more prominently. Steep slopes and large altitude differences are found near the present-day main river valleys and plains that are part of the eastern branches of the “Flemish Valley” (see Chap. 18). This is indicative of the ongoing morphogenic processes, though human activities such as the widespread extraction of iron sandstone also produced many steep slope segments, originating from ancient quarry walls.

14.2 Description of the Hageland Relief

The Hageland hills are characterized by their elongated shape and often stand out as single relief features in a flat surrounding landscape. They all contain outcrops of iron

Fig. 14.2 Aerial photograph of Wijngaardberg taken on 8 March 2015. View from the south. *Two contour lines* have been added to enhance the hill shape

sandstone and consistently show the same geological substrate of Upper Miocene highly glauconiferous Diest Sands (Vandenberghe et al. 2014).

Most of the hills (Fig. 14.3) are between 0.2 and 1 km wide, 0.5 and 5 km long, and 20 and 50 m high, though smaller forms rising only a few metres above the surrounding landscape can also be ranked as Hageland hills (e.g. near Westerlo and Tessenderlo, see Fig. 14.3). While the most prominent hills have steep slopes with values of 20–40% and more (Fig. 14.1), many instances are found of much gentler slopes. Also, the inselberg character is no defining feature. Some hills that sharply emerge from the landscape at one end merge with others at their other extremity in a plateau-like area (Figs. 14.1b and 14.4). Such plateaus are found, e.g. between Brussels and Leuven, around Pellenberg, Scherpenheuvel and Linden.

Although a closer look shows that individual hills may have a range of orientations, the hill crest lines are dominantly oriented WSW-ENE (Figs. 14.1 and 14.5). However, a few ridges, mostly low ones, are almost perpendicular to this direction. Mapping of the Hageland hills relied on (1) their sharp topographic emergence from the surrounding topography, and (2) their independence of the hydrographic network, and thus also included the low hills near Westerlo and Tessenderlo. In total, 275 km of crest lines have been identified.

Figure 14.3 shows the intrinsic tie between the Hageland hills and the outcrop area of the Upper Miocene Diest Sands. These sands are a marine deposit of mostly well rounded medium-size quartz grains and up to 50%, and locally more, transported glauconite pellets (Vandenberghe et al. 2014; Adriaens 2015). Clay is also present in drapes and up to a few centimetres thick layers. Both the glauconite pellets and the clay mineral composition attest the marine provenance of the sedimentary matter (Adriaens 2015).

Without doubt, the Diest Formation plays a key role in the story of the Hageland hills. The dominant WSW to ENE topographic direction is also characteristic of the internal structure of the Diest Sands. Many outcrops show thin and thick cross-bed sets (Fig. 14.6). The great majority of the cross-bed foreset laminae dip to ENE (Vandenberghe and Gullentops 2001), indicating that flow currents were chiefly oriented from WSW to ENE during deposition of the sand. Moreover, the basal surface of the Diest Formation shows long, linear depressions oriented in the same WSW-ENE direction, especially in the Hageland (Fig. 14.7).

14.3 Existing Theories on the Origin of the Topography

14.3.1 Role of the Ferricretes

The Hageland hills have always been considered as a classical example of iron sandstone-capped positive topographic features; this was reported since the nineteenth century (Gullentops 1957; Vandenberghe and Gullentops 2001). Obviously, subhorizontal beds of fairly resistant ferricrete are observed in the hills. They would, on uplift of the area, have protected the present-day hills from erosion.

14.3.2 The Tidal Sandbank Model

Gullentops (1957) correctly stated that the previous theory was insufficient to explain the alternation of narrow hills and

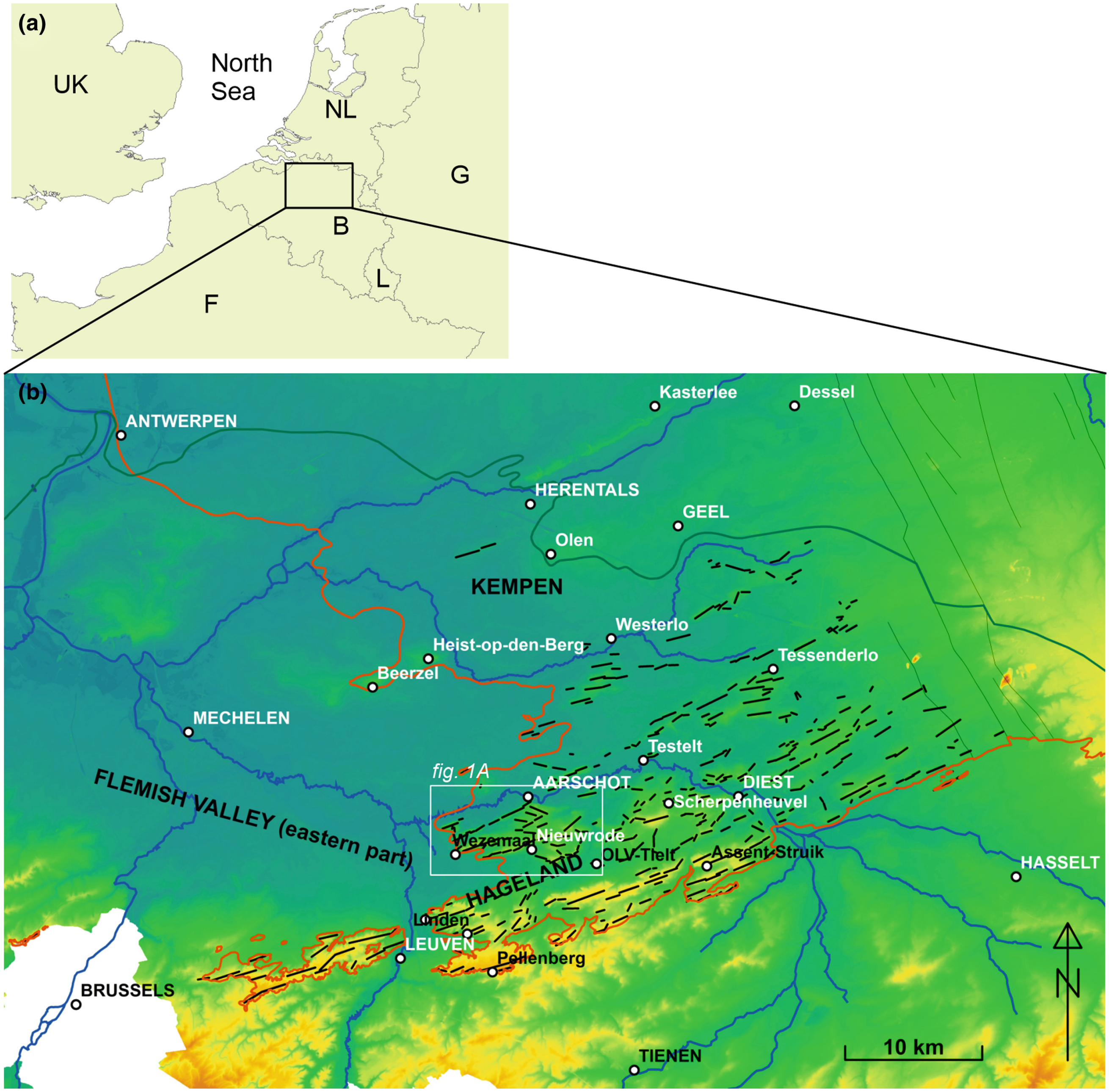

Fig. 14.3 **a** General location of the study area in NW Europe. **b** Elevation map of the Hageland and southern Kempen (Campine) area. Lowest areas in *greenish blue* are at sea level, highest areas in *red* are around 120 m TAW. *Source of data* 5 m digital elevation grid (VMM, Watlab and AGIV (2006) DHM-raster5). *Black lines* indicate crest axes of the Hageland hills. *Red line* is the contour of the Diest Formation outcrop; *dark green line* is the contour of the Kasterlee Formation outcrop. Both formations continue in the subsurface in the north and northeast

wide depressions, oversize with respect to the small creeks that drain them. He proposed instead an explanation inspired by the plan shape of the hill crest pattern and interpreted the hills as remnants of offshore tidal sandbanks, whose submarine relief has been somehow preserved in the present topography (Fig. 14.8). He theorized that at some time, the Diest Sea left the area so suddenly that the sandbank seascape was preserved. In the subsequent subaerial environment, the glauconite present in the sand dissolved and precipitated as iron (hydr)oxides to form resistant sandstone beds, which indeed were able to protect the sandbank relief against erosion during further uplift.

Fig. 14.4 Aerial photo of Nieuwrode, south of Aarschot, taken on 8 March 2015. *Note* the gentle slopes of the long hill, which is connected with steeper hills at its far west end, near Wezemaal. *Two contour lines* have been added to highlight the hill slope

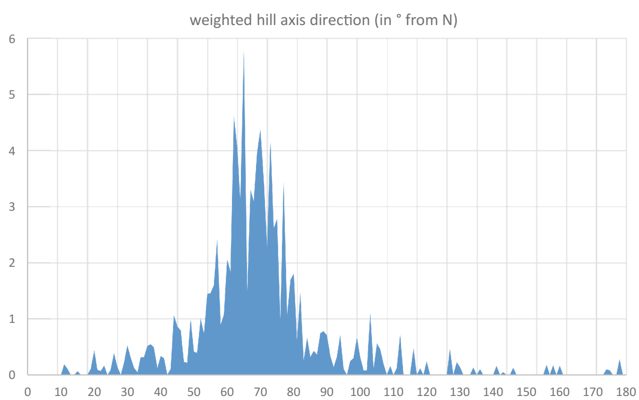

Fig. 14.5 Frequency diagram of structural hill crest axis lines in Hageland, weighted by their length. 90° is east

The tidal sandbank model (Gullentops 1957, 1988; Broothaers 1996; Vandenberghe and Gullentops 2001) invokes the following stages:

- the Upper Miocene Diest Sea transgressed farther SW than the present outcrop area of the Diest Sands, as suggested by similar indurated iron-rich sands on top of the Flemish Hills (as far as Cassel in northern France, more than 100 km west of Hageland), creating a marine connection between the English Channel and the southern North Sea;
- the Diest Formation base surface (Fig. 14.7) was eroded by the tidal currents that imported glauconite-laden sand from the English Channel through this strait and built a topography of shifting tidal sandbanks;
- at the end of the Diest Sands deposition, there was a seascape of linear offshore sandbanks similar to the present Flemish Banks in the southern North Sea (Fig. 14.8);
- the sea level lowered sometime so quickly that this seascape or, at least, the linear banks emerged. The Zanclean flooding event was suggested to have provoked this sudden sea-level drop (see below);
- on initial emersion of the seabed, soil processes under warm conditions during the Pliocene induced iron cementation of the sediments at the top of the emerged sandbanks;
- on further emersion, the ferricretes protected the sandbanks from erosion;
- on further uplift during the Pleistocene, the hills were accentuated by erosion of their surroundings.

Fig. 14.6 Typical Diest Sands cross-bed in a temporary pit west of Leuven. Scraper tool is 40 cm long

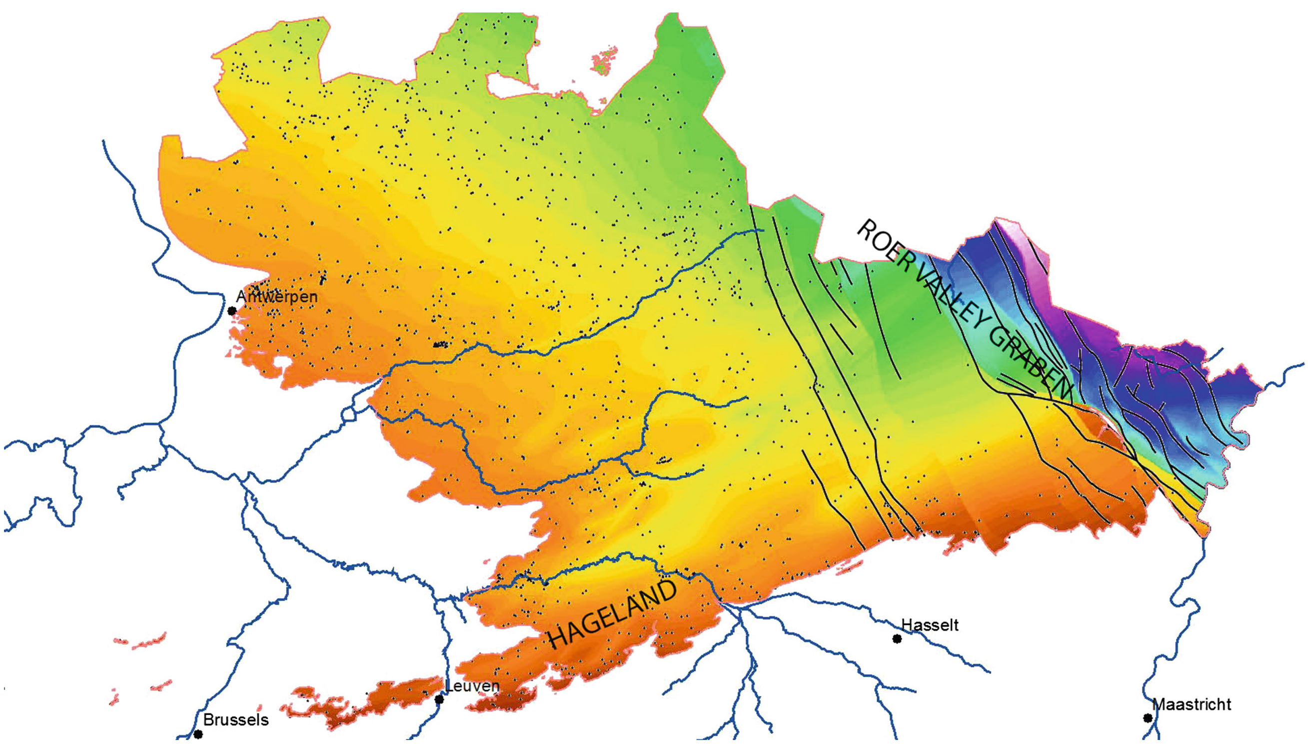

Fig. 14.7 Elevation map of the basal surface of the Diest Formation. *Colours* represent 10 m elevation steps, descending from +100 m in the southwest (*orange brown*) to below −500 m in the northeast (*light purple*). *Dots* represent wells where the base surface has been identified. Source of data: Matthijs et al. (2013), G3Dv2_0203_ Formatie van Diest basis, available on http://www.dov.vlaanderen.be

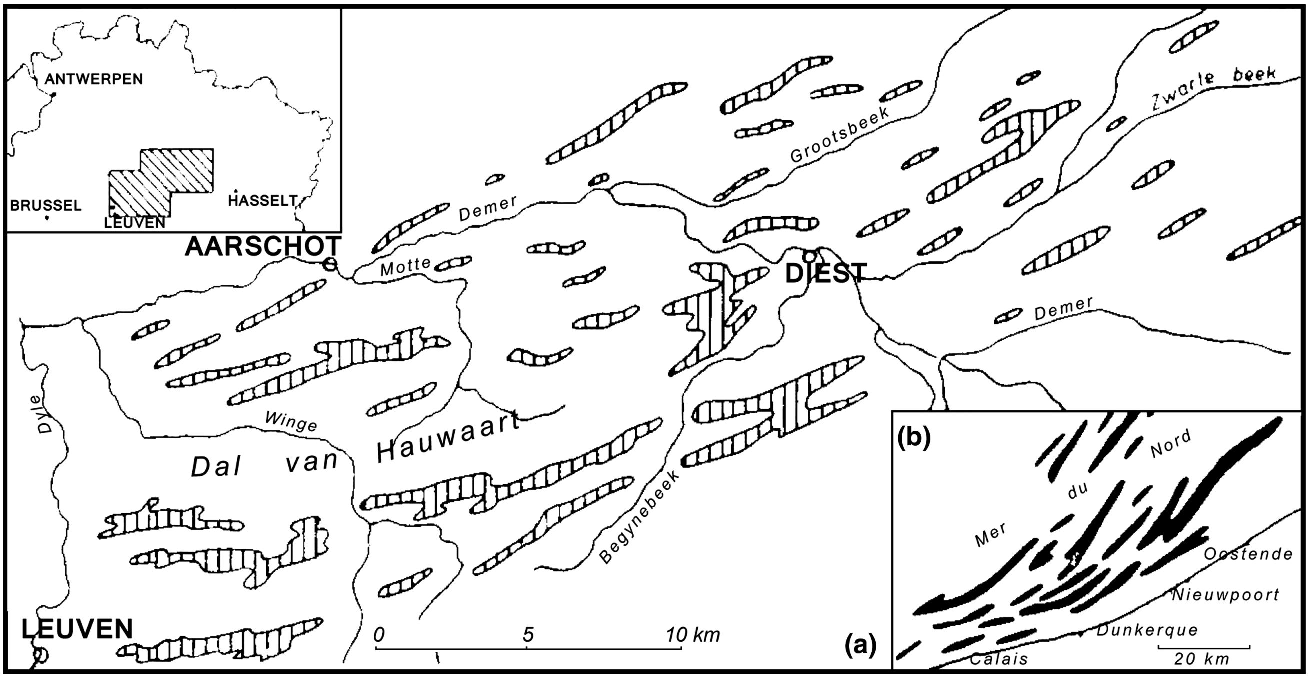

Fig. 14.8 Map of the Hageland hills. *Inset* For analogy, outline of the Flemish Banks on the present-day French–Belgian continental shelf (from Gullentops 1957)

14.4 New Evidence

14.4.1 The Diest Formation Is Absent in the Flemish Hills Area

The sands found on top of the Flemish Hills have been deposited in the shoreface environment of an exposed sea (Houthuys 2014). By contrast, the Diest Sands represent a marine embayment fill (see below). These two environments are incompatible. Instead, Houthuys (2014) suggested an Upper Eocene age for the Flemish Hills Sands, making them ca. 30 My older than the Diest Formation. While the precise age of deposition of the Flemish Hills Sands remains uncertain, the different sedimentary environments and the basin context indeed rule out a coeval Upper Miocene sedimentation. Therefore, a paleogeographic configuration with an E-W coastline parallel to the line of the Flemish Hills as depicted in Broothaers (1996) most likely never existed during the deposition of the Diest Sands.

14.4.2 The Basal Surface of the Diest Formation

The elevation map of the base of the Diest Formation describes an irregular surface dipping $\sim$0.5% to NE and marked by WSW-ENE linear depressions (Fig. 14.7). Vandenberghe et al. (2014) interpreted the depression pattern as the signature of a paleodrainage system of parallel rivers on a relatively flat emerged surface gently sloping to the sea. Indeed, the depressions are perpendicular to the coastline, assumed to coincide with the border faults of the Roer Valley Graben before the Diest marine transgression. While some depressions are well documented (Fig. 14.7), others are more hypothetical, the base surface having been interpolated with the model of a connected channel pattern in mind (Lanckacker, personal communication 2014). However, based on P. Van Calster's data, a detailed map of the Hageland part of the basin revealed a pattern of elongated closed depressions (Houbolt 1982).

14.4.3 The Sedimentary Environment of the Diest Sands

The sedimentary characteristics of the Diest Sands were revised by Houthuys (2014). He identified in the outcrop area a facies suite consisting in "thick wedge-shaped cross-beds" (XT), "tabular descending cross-beds" (XD), "bioturbated cross-beds" (Bx), "bioturbated beds" (B), and "massive breaching beds" (M). Although additional data are needed to support the outcrop evidence, some important deductions about the sedimentary environment can unequivocally be made: (i) the overwhelming majority of foreset directions is to ENE, with a small spread of the data;

(ii) thick cross-beds with long depositional faces can only be the product of very large bedforms formed in relatively deep water (several tens of metres) and very strong flow; (iii) the numerous clay drapes and clay drape packages indicate unsteady currents in relatively confined flow channels where mud can settle during slack waters; (iv) fairly sheltered areas that allowed deposition of slightly clayey fine sand and intense bioturbation were present in the depositional environment; (v) bedding surfaces have a systematic downcurrent dip (i.e. to ENE) and often a transversal dip component pointing to the centre of the hill (Fig. 14.9a); (vi) as the facies are interwoven, they bear witness to a single sedimentary environment; (vii) sedimentation was fast in an environment with slopes steep enough to allow breaching to produce M facies (Van den Berg et al. 2002), and (viii) no sedimentary structures indicative of waves have ever been found.

Reconstruction of the sedimentary architecture is so far partial, restricted to the outcrop-scale. In the longitudinal ENE direction, the beds show downcurrent progradation; in the transverse direction, the bedding indicates a channel fill, while in the vertical direction, a stacked fill is observed (Fig. 14.9a). These elements suggest an environment of eroding channels or linear troughs where filling takes place immediately in the wake of the erosion centre, which migrates downstream (Fig. 14.9b). This mechanism seems to imply that a very high sediment supply drove all other processes involved in the Diest Sands deposition. This is further detailed below.

14.4.4 The Top Surface of the Diest Sands

In the Kempen, the Diest Sand Formation is covered by the Uppermost Miocene marine Kasterlee Formation (Matthijs et al. 2013) (Fig. 14.3), which is also present at Heist-op-den-Berg and Beerzel as an 8–10-m-thick layer at the top of the hills (Fobe 1995; Verhaegen et al. 2014). The contact surface of the Kasterlee and Diest Sands contains rounded and weathered flint pebbles near Heist-op-den-Berg and at Olen (Verhaegen et al. 2014). Near Leuven and in NE Hageland, a residual gravel of well rounded, up to 10-cm-long flint pebbles is present on top of the Diest Sands, marking the base of the Quaternary (Vandenberghe and Gullentops 2001).

The top surfaces of the Hageland hills, generally coincident with that of the Diest Sands, are often fairly flat and the most elevated of them form a relatively flat surface gently sloping to the north and smoothly merging into the Kasterlee Formation base surface. As evidenced by the sedimentary structures in the Diest Sands, this surface truncates the top part of the formation (Houthuys 2014). All this indicates that the top surface of the Diest Sands is a marine abrasion surface formed by wave erosion during the following

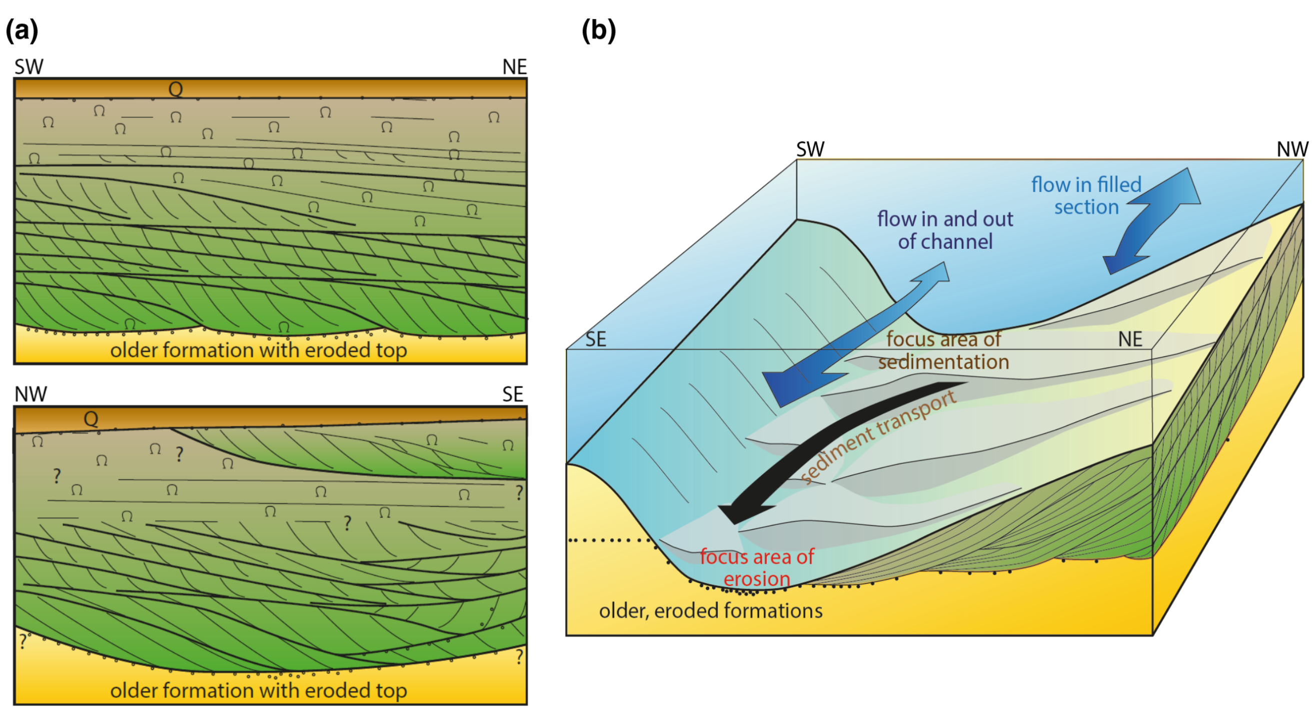

Fig. 14.9 **a** Typical longitudinal (*top*) and transverse (*bottom*) sections through a Hageland hill. Idealized sections of width × height in the order of 50–100 m × 10–20 m. Ω bioturbation. **b** Reconstruction of the sedimentary processes that governed deposition of the Hageland Diest Sands. *Grey-shaded* bedforms are large dunes. They were most probably covered by smaller dunes (*not indicated in sketch*)

transgression, as usual for most Cenozoic marine formations in the southern North Sea basin. It might even be a multi-stage abrasion surface, as the time period between the estimated ages of the Hageland Diest Sands (11–9 Ma, Vandenberghe et al. 2014) and the base of the Kasterlee Formation (7.5 Ma, Louwye and De Schepper 2009) could have accommodated several marine transgression cycles.

14.4.5 On the Iron Sandstones

Iron sandstones are very common in the Diest Sands. The original loose deposit has been cemented by iron minerals of the limonite group through precipitation of iron dissolved in the groundwater. This required oxidizing conditions, and thus emersion. Outcrops of Diest Sands frequently show loosely cemented liesegang rings indicative of slow precipitation in concentric areas according to local chemical gradients (Fig. 14.10a). Though not a prerequisite for iron cementation, clay layers are often found inside and at the bottom of the cemented zones, illustrating the role of perched water tables.

These ferricretes often display capricious shapes, with harder zones jutting out from the outcrop, separated by uncemented or weathered sandy zones (Dusar 2014) (Fig. 14.10a). Though most indurated layers follow the original stratification, some, including concentric, nested spheres ('klapperstenen'), also crosscut the primary layering. However, they generally appear either as thin plates, hard and brittle, often found near the top of the hills, or sheet-like or isolated thick massive or crumbling blocks (Dusar 2014). Only these iron sandstone masses, from ~0.2 m to several metres in thickness, are believed to have had a possible morphogenic role (Fig. 14.10b, c). According to our observations, confirmed by those of others (Van Calster, pers. comm. 2015), such well-developed ferricretes are never found in the deeper subsurface that has always been below the water table. Wells drilled in valleys and throughout the low-lying areas of the Campine never encounter ironstones, apart from local occurrences of siderite (Laga 1972). As observation shows, they also do not occur generally in the centre of plateaus or wide hills but concentrate predominantly along the sides of the hills (Bos and Gullentops 1990; Dreesen et al. 2010; Dusar 2014), often at depths between 10 and 20 m below the hill tops. Outcrops extending over more than a few tens of metres feature multiple (sub)horizontal ferricrete levels (Fig. 14.10b), mostly discontinuous and of varying structure and thickness (Fig. 14.10c).

Iron sandstones are not exclusive to the Diest Sand Formation (e.g. Bos and Gullentops 1990). Famous are the ironcrust-capped Flemish Hills, between Cassel in N France and Flobecq in N Hainaut, and the ferricretes crowning the hills between Geraardsbergen and Kester. While the former indurated sediments are of possible Upper Eocene age, the latter belong to the Middle Eocene Lede Formation (Hout-huys 2014). Ironstone layers locally over 5-m-thick are also found in the Middle Eocene Brussels Sands, often in plateau areas or underneath valley heads, but they had no

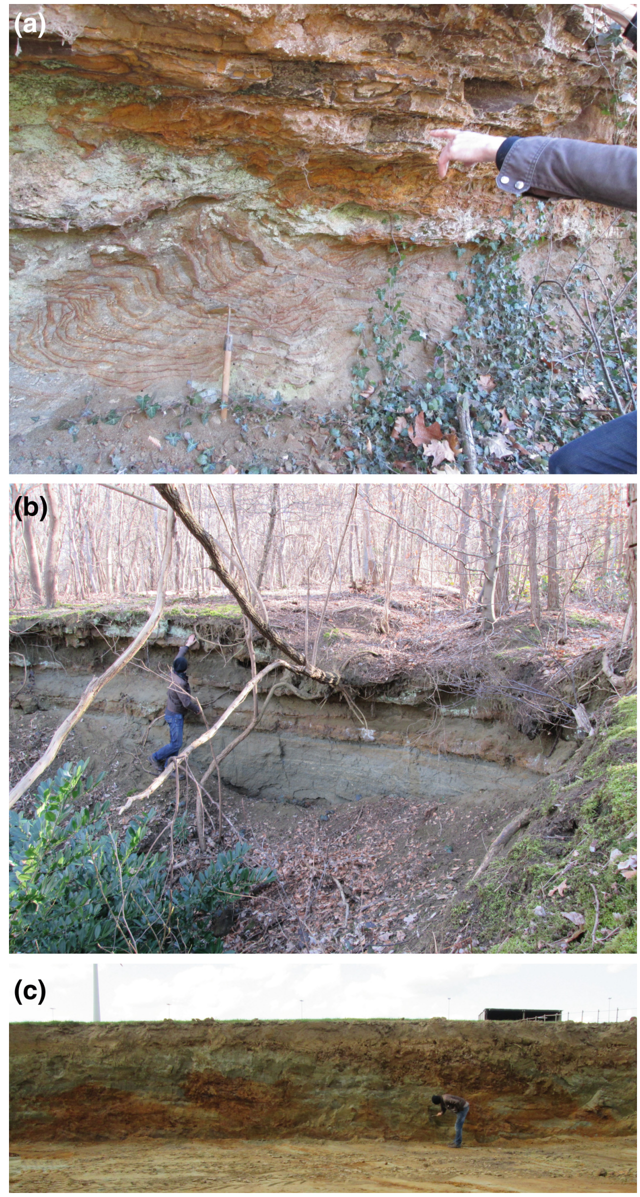

Fig. 14.10 **a** Compact ironstone beds (*top*) and Liesegang structures (*bottom*) in the Diest Sands at Pellenberg. **b** *Horizontal*, a few decimetres thick iron sandstone beds at Kesselberg. The slight elevation of the ground surface may be related to the presence of the *upper sandstone bed*. *Note* the presence of a thin inclined iron crust running across the loose sand layer between the *lower* and *middle* horizontal indurated beds. **c** Irregular orange iron sandstone zones in a temporary outcrop of the Diest Sands at Assent Struik. The *top* half metre is late Pleistocene sandy loess. No influence of the indurated zones is observed in the surface topography. Section is oriented WSW (*left*) to ENE

topographic effect. The more or less horizontal development of all extensive ironstones, regardless of the formation in which they were formed, suggests a link with paleowater tables, whose oscillations provided the alternation of oxidizing and reducing conditions needed to produce such thick ferricretes.

While authors often state that conditions warmer than those of today were necessary for the mobilization of iron oxides (Bos and Gullentops 1990; Vandenberghe and Gullentops 2001; Dusar 2014), some observations in the Hageland suggest a fairly recent origin of the iron sandstones. Differential compaction is nowhere observed near thick massive ironstones, indicating that compaction occurred prior to induration. Moreover, while Bos and Gullentops (1990) mention hard crusts formed in a few decades, water seepage at excavation sites in the Diest Sands is often observed to cause the formation of a thin, crumbly crust of iron-cemented sand in only a few weeks. In brief, though no dating is available of the time of duricrusting, there is evidence of a very long period in which it may have occurred, groundwater ferricretes requiring only a continental environment but being not climate-dependent. However, some evidence points to recent induration, which may be going on today.

14.5 Problems with the Offshore Sandbank Model

In view of the above observations, there are several insuperable problems in the offshore sandbank model of the Diest Sands and Hageland hills, which we list hereafter.

14.5.1 The Internal Structure Is not of a Sandbank

Offshore tidal ridges, or sandbanks, are the largest bedforms in seas with strong tidal currents, i.e. near-surface mean spring peak tidal currents well over 0.5 m/s (Belderson et al. 1982). They may reach a few tens of km in length, a few km in width and a few tens of m in height, dimensions that indeed match the Hageland hills. They often occur as groups of parallel linear forms extending over hundreds of square kilometres (see inset in Fig. 14.8), aligned at a small angle to the direction of the peak tidal flow. Actively maintained sandbanks show large dunes (megaripples) on their flanks and top. The crest of these asymmetric bedforms (with lee side slope $\leq 6°$) is barely lower than low-tide level. Observations of recent and well-documented ancient offshore tidal sandbanks consistently point to the following internal sedimentary architecture and structures (Stride et al. 1982; Houthuys 1990; Trentesaux et al. 1999; Reynaud et al. 1999; Mathys 2009; Reynaud and Dalrymple 2012; Michaud and Dalrymple 2016; Leva Lopez et al. 2016):

- sandbanks have either a flat basal surface or developed over an older core of any origin;
- the internal structure predominantly shows cross-bedded sand produced by the dunes that mantle the ridges; the cross-stratification dip tends to change gradually upwards, in accordance with the dunes becoming more parallel to the bank crest when they approach it;
- the master bedding surfaces are parallel to one or both sandbank flanks, and slope at angles of a few degrees; in the case in which both slope directions have been preserved, the left and right flank master beds truncate each other in the middle;
- the base deposits are more bioturbated, finer grained and more poorly sorted than the top beds;
- wave structures, such as flat wave planation surfaces, may be found near the sandbank crest; also internal gravel lags, coarse graded storm beds, or even hummocky cross-stratification may be present;
- ebb- and flood-oriented cross-beds should be more or less equally developed, especially near the top of the form, where cross-beds are short and separated by a multitude of internal bounding surfaces.

None of these features are found in the Diest Sands.

14.5.2 The Flemish Hills Top Sands Should Be Part of the Diest Formation

The paleogeographic configuration compatible with offshore sandbanks similar to the present Flemish Banks (Fig. 14.8) requires that the Flemish Hills top sands are part of the Diest Formation. If they are of a different age than the Diest Sands, as proposed by Houthuys (2014), the coastline configuration needed by the offshore sandbank model is highly unlikely to have occurred.

14.5.3 The Basal Incision of the Diest Sands Is not Compatible with Sandbank Formation

Present-day sandbanks are not found in sea straits, though they do occur in their vicinity. If incision at the base of the Diest Sands is explained by strait erosion, the location of sandbanks inside such a strait setting is problematic as tidal sandbanks are accretionary forms developing on a flat surface or a positive pre-existing form.

14.5.4 A Timing Problem

Tying a supposed post-Diest Sands fast sea-level fall to the event of the Zanclean flood is inconsistent with available ages. The drop in global sea level entailed by the still debated Zanclean flood, which would have ended the Messinian salinity crisis (base level fall of 1.5–2 km in a landlocked Mediterranean Sea) by breaching the Gibraltar land barrier and refilling the Mediterranean basin, has been dated at 5.33 Ma (Krijgsman et al. 1999). By contrast, the Diest Sands are relatively well constrained in time during the first half of the Tortonian (DN8, from 11 to 9 Ma) (Vandenberghe et al. 2014). A Zanclean sea level fall ending deposition of the Diest Sands is therefore excluded. However, this timing problem is not crucial because eustatic sea-level changes in the order of 10 m, corresponding to third- and fourth-order sequences of Vail et al. (1977), occurred throughout the geological times.

14.5.5 The 9-Ma Submarine Topography Is Unlikely to Preserve

Even the quickest ~10-m sea-level drop following deposition of the Diest Sands would have lasted long enough to allow the currents and waves that had built the sandbanks to fill up intervening spaces and flatten the seascape on lowering of the sea level, in accordance with the sequence stratigraphy scheme in which a relative sea level fall causes the lowering wave base to plane off the soft sea floor sediments and create a 'regressive surface of marine erosion' (Catuneanu 2006). In order to get around this problem, Gullentops (1957) invoked fast, quasi syn-regression ironstone formation. However, there is nothing to document this assumption and, as explained above, the iron sandstones are likely to have formed millions of years after deposition of the Diest Sands and, in any case, only when a subaerial topography provided the required oxidizing conditions.

14.6 A New Depositional Model

Houthuys (2014) proposed a sedimentologically consistent depositional model that accommodates the basal incision and the internal architecture of the Diest Sands, based on an analogy with the Brussels Sands depositional history (Hout-huys 2011).

The depositional environmental similarities between the Eocene Brussels Sands and the Miocene Diest Sands are impressive (Houthuys 2014). In the well-exposed Brussels Sands, the fill mechanisms and architecture could be thoroughly documented, showing that the sedimentary system was characterized by an overall eastward lateral accretion and flow constriction occurring on closure of successive pre-existing local depressions (Houthuys 2011). Such a constriction implied strong reinforcement of the flow currents that caused scouring, immediately followed by cross-bedded fill.

The result is an internally differentiated fill style consisting of a background of bioturbated muddy fine sand, in which a series of parallel elongated bodies exist, corresponding to erosive troughs filled with coarser, cross-bedded sand. The troughs have straight, non-meandering axes in relation with the following driving mechanisms:

- a continuous external sand supply, inferred to have included a coastal drift pathway, feeding into the embayment;
- tide action inside the embayment, causing much water to enter it during rising tide, including over the low-tide flats, and leave it at falling tide through increasingly narrower channels at one side of the filling basin.

As local flow constriction becomes more and more dominant when nearing complete fill of the basin, the elongate troughs filled with coarse cross-bedded sand are preferentially found at the downdrift side of the embayment, e.g. to the east in the Brussels Sands case.

Applied to the Diest Sands, this model accommodates key observations (Fig. 14.11), namely (1) the complex erosion history and the shape of the basal surface, (2) the marine nature of the sand fill and (3) the high-stand style of the fill. In contrast to the Brussels Sands, the Diest Sands appear to have undergone a syn-sedimentary tilt towards the Roer Valley Graben. However, this tilt was probably slow enough not to interfere with the concomitant sedimentary processes.

Waiting for a definitive proof that the Brussels Sands depositional model can be applied to the Diest Sands, we note that unknown key elements remain the exact large-scale progradation direction at the basin scale, the valley- or closed-trough-like character of the longitudinal basal depressions, and the precise style of lateral interweaving of the depression fill bodies within the finer, bioturbated sand. If the analogy holds, then:

- no exceptional tectonic event or unrealistic paleogeographic configuration is needed to explain strong incision underneath Hageland, where the parallel troughs would have been created by internal, fill-related erosion cycles;
- this sedimentary process chain created longitudinal internal bodies built of coarser grained, cross-bedded, trough-fill deposits;
- like every sedimentary cycle, sedimentation ended with a complete basin fill, probably followed by the development of a relatively flat surface of marine erosion truncating the upper part of the Diest Sands upon regression.

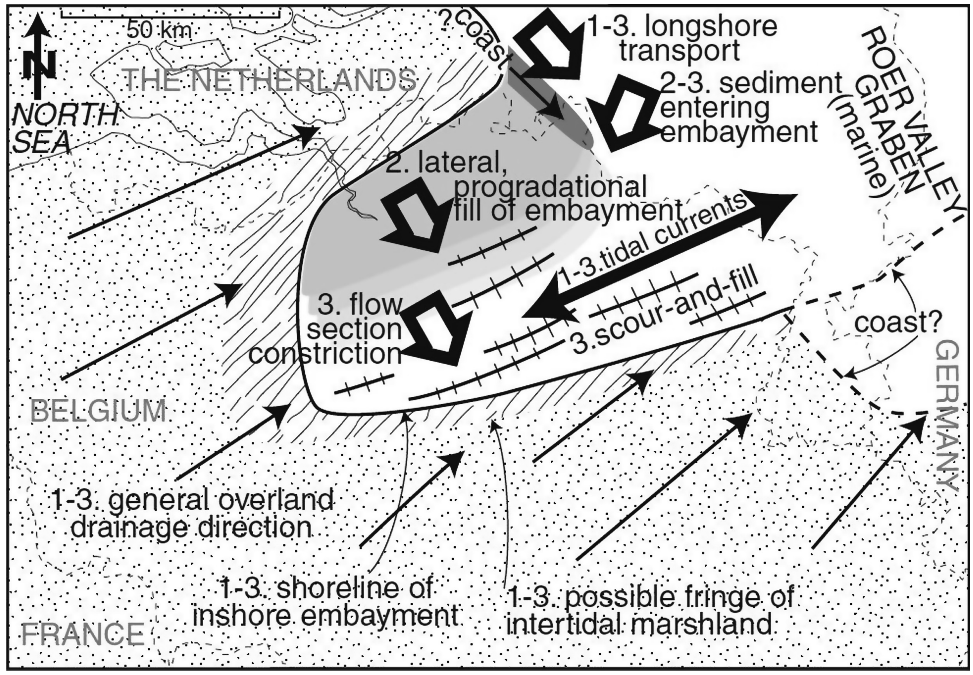

Fig. 14.11 Reconstruction of the paleogeography during deposition of the Diest Sands. Present-day coastline and state boundaries are shown in *thin grey lines* for reference. Stages in the paleogeographical evolution: *1* drowning during relative sea-level rise of a river mouth area and establishment of an inshore marine embayment; *2* first stage of filling, prograding from NW to SE; *3* last filling stages with subtidal flow section constriction and subaqueous formation of scour-and-fill elongate troughs. Stippled area is emerged (from Houthuys 2014)

The differential subsidence during deposition of the Diest Sands also remains to be confirmed. If indeed the NE part of the Diest Embayment experienced stronger subsidence than its SW part during deposition, a decoupling between basal incision and internal incision would be expected in the Diest Formation, with coarse trough fills not necessarily positioned on top of basal incisions.

The model implies a flat, low-lying environment during the Diest Formation deposition, flooded and filled under high sea level conditions. Inferring marine transgressions postdating the deposition of the Diest Sands also implies a significantly younger start of the continental morphogenesis.

14.7 Alternative Explanations of the Development of the Hageland Topography

Beyond purely sedimentary considerations, uncertainties remain about the precise timing and characteristics of the local uplift history, the original extent of the iron sandstones and the conditions and timing of their origin. Regarding the Hageland topography, some new deductions can now be made, which are however necessarily somewhat speculative.

14.7.1 Role of the Iron Sandstones

It seems unlikely that thick ironstones were present in the crucial stage when the topography started to get its grain. Within the flat, low-lying topography that existed prior to significant regional uplift, the Diest Sands lay mostly below the water table, which was unfavourable for iron oxidation and duricrusting anywhere in the area. Ironstone formation required a minimum relief, with an unsaturated zone above an oscillating water table, to take place. Moreover, the repeated iron sandstone levels found at some sites probably reflect a succession of paleo-groundwater tables related either to a stepped uplift history or to perched water tables above clay beds.

Assuming that an iron crust formed very early after regression, one would expect the formation of an extensive sheet-like ferricrete rather than elongated parallel indurated zones that would later turn to hills through differential erosion. In addition, many Hageland plateaus are free of ironstone beds and the isolated hill tops are often underlain by several metres of loose, often clayey sand before compact, thick ironstone layers are encountered in deeper position. Extensive layers of ironstone have nowhere been evidenced underneath hill tops.

Evidence of Kasterlee Sands inside the top part of the hills at Heist-op-den-Berg and Beerzel further discredits the purported relation between iron sandstone and hill formation. Rounded limonite lumps and iron crust fragments have been found at the base of the Latest-Miocene Kasterlee Sands at Beerzel (Verhaegen et al. 2014). They would indicate supply from already emerged and duricrusted Diest Sands around a marine Kasterlee embayment. However, the hills in the Heist-op-den-Berg and Beerzel area have a similar topography to the Hageland Hills, yet contain non-indurated marine sediments (the Kasterlee Sands) significantly younger than the Diest Sands. It seems thus far more probable that ferricretes played no structuring role during the initial stages of definitive emersion of the area.

Once present, the ironstones may have helped to protect locally and temporarily the initial Hageland hills (Fig. 14.10 b). They nevertheless may not be as resistant as they look and their protective efficiency should be put in perspective. For instance, many sunken roads in Hageland show iron sandstone walls suggesting that, if the ferricretes were pre-existing extensive features, they succumbed to a few centuries of passage and runoff. Also, many iron sandstone blocks used in old buildings show significant wear (Fig. 14.12). A critical review of the observations seems to point to a fairly fast ironstone formation (in the order of kyears) at various times since the area is emerged. Bearing in mind the formation of iron crusts is a matter of weeks at excavation sites, one may even wonder whether ironstones in the flanks of sunken roads formed after the latter were incised.

But, regardless of the formation time of the massive iron sandstones, a primitive topography had to exist for such duricrusting to take place, so that the ferricretes cannot have preceded, and have been responsible for, the formation of the hills. At this point, only a few hypotheses independent of the presence of ironstones are thus left to explain these remarkable parallel hills.

14.7.2 Hypothesis 1: Interfluves of a Primary Parallel Overland Drainage System Have Been Preserved

It could be theorized that an overland drainage system with a preferential flow direction to ENE (corresponding to the peak direction in Fig. 14.5) was present after the last marine transgression. After initial uplift and incision, the interfluves would have developed into the typical long hills. In this hypothesis, the internal structure of the Diest Sands is indifferent. Later creation of cross-hill river trunks such as the Dijle, Winge, Motte, Begijnebeek, Demer (Fig. 14.8), river captures and local inversions would have rearranged this original system.

However, the spacing and rhythmicity of the hills seem out of proportion to the intervening small watercourses. Vandenberghe and Desmedt (1979) thought the structures related to the Diest Sands in the eastern part of the Scheldt basin rather imposed river valley directions than vice versa. Also, some typical Hageland hills have a different alignment (Fig. 14.3), barely compatible with the proposed parallel drainage system. Finally, this hypothesis is also weakened by the observation that, on the whole, this parallel arrangement is present only in the rather restricted area where the Diest Sands crop out.

Fig. 14.12 Zoom on the stone front of the Testelt water mill, built in 1608. *Note* wear of the building blocks of local ironstone extracted from the Diest Formation

14.7.3 Hypothesis 2: Denudation Processes Were Guided by Internal Structures

This hypothesis states that the internal structure of the Diest Sands, organized along a preferential direction, has been carved out in the landscape. In other words, the elongated parallel coarser grained, cross-bedded trough fills characterizing the internal architecture of the formation would have been sculpted out as positive relief features during initial uplift and denudation. The regional arrangement of the Hageland hills, their dimensions, and the correspondence between the hill axis direction and the principal internal directions (basal depression long axis, dominant foreset direction) strongly support this hypothesis.

Again, the analogy with the Brussels Sands, where the interpretation is more firmly documented, is striking. There, after removal of a younger cover, differential erosion within the Brussels Sands also allowed internal structures to impose the local topographic grain (Houthuys 2011). Therefore, Hypothesis 2 is a plausible candidate to explain the topographic structure of the Hageland. As grain size is the main differentiating element inside the Diest Sands, we suggest that the more grain-supported fabric of the coarser grained bodies might have resulted in slightly positive relief features after compaction, which in turn would have better resisted to erosion than the surrounding finer clayey sand because their higher permeability favoured infiltration of meteoric water, thus reducing runoff erosion.

14.7.4 Hypothesis 3: Imprint of Directional Structures Inherited from a Younger, Now Eroded Formation

This hypothesis proposes that younger, now eroded Cenozoic deposits were present on top of the Diest Sands in the Hageland. After emersion, they would have imposed topographic directions following mechanisms similar to those exposed in Hypothesis 2 and these directions would have been inherited and preserved within the Diest Sands on further uplift and denudation. Hypothesis 3 has the virtue of accommodating the observation that similarly oriented hills partly cut in younger deposits exist north of the Hageland, notably the hills at Heist-op-den-Berg and Beerzel, covered by Kasterlee Sands, and the long hill between Herentals and Kasterlee, with Poederlee and Kasterlee Sands.

The Kasterlee Formation is made of three stacked members of fairly wide areal extent (Fobe 1995; Verhaegen et al. 2014) but they do not seem to show any spatial internal differentiation, which makes the parallelism of these hills with those of the Hageland very intriguing and depreciates the last hypothesis.

14.7.5 What Happened Once the Direction Appeared in the Topography?

While decisive arguments are currently lacking to choose between the hypotheses presented above, the second hypothesis, i.e. sculpting of internal sedimentary contrasts of the Diest Sands, is suggested to be the most plausible representation of how the Hageland morphology was initiated. Once low embryonic hills had appeared in the emerging landscape, several of the processes discarded above as primary morphogenic causes may have joined forces to perpetuate the differential erosion. The more permeable hills evacuated their share of meteoric water by percolation towards the depressions. Iron sandstone formation at seepage locations may have created an armour that contributed to preserve some hill slopes and make them steeper in a process that may well go on today.

14.7.6 The Story at a Glance

In brief, the Hageland hills were most plausibly created and evolved as follows:

- during deposition, the Diest Sands acquired an internal sedimentary architecture consisting of parallel elongated bodies whose ENE trend was imposed by the lateral fill style that constricted tidal currents; these bodies are cross-bedded, relatively coarse-grained sand fills of longitudinal erosion troughs, embedded in finer grained, clayey and bioturbated sand;
- these coarse sand bodies were sculpted out after first emersion and initial uplift, probably not before the Late Pliocene (depending on the regional uplift history, a currently still unsolved issue); the proposed, but not yet demonstrated, mechanism for differential erosion relies on differential compaction and contrast in permeability between coarse sand and clayey sand;
- ironstone formation started only after an initial topography had formed;
- ironstones may have contributed to the preservation, possibly also the enhancement, of the initial hills during further uplift and associated valley incision.

14.8 Conclusion

The Hageland area has a particular topography: the land surface displays long, parallel hills. The hills are a few tens of metres high and consist internally of an Upper Miocene marine sand deposit, the Diest Sands. Horizontal ironstone beds and variously shaped iron sandstones often crop out in the flanks of the hills.

Building a coherent theory to account for the array of hills is an issue that has boggled the minds of researchers for decades. The present state of knowledge allows discarding the established theory, making two things clear, namely that (i) the morphology of the Hageland hills is not the legacy of an offshore tidal sandbank morphology as the paleogeography and internal structure of the Diest Sands are not compatible with offshore sandbanks, and (ii) massive iron sandstones cannot have been responsible for the creation of the hill array because their formation required an initial topography.

The alternative theory presented here is that the sedimentary architecture of the Diest Sand Formation includes parallel elongated sand bodies and that, a long time after deposition, uplift, probably in the Pleistocene, induced grain-size-controlled differential compaction and erosion that materialized these sand bodies in the topography as positive residual landforms. Some essential elements of this alternative theory are still hypothetical. This paper hopefully will encourage research and new data collection aimed at further unravelling the origin of the intriguing Hageland hills.

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