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A reinterpretation of the Neogene emersion of central Belgium based on the sedimentary environment of the Diest Formation and the origin of the drainage pattern

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ABSTRACT. The 20 to 25 m thick sand deposits crowning the Flemish Hills from Cassel to Flobecq are geologically mapped as Diest Formation (Upper Miocene), but they are now argued to belong to a different formation, for which the new name of Flemish Hills Formation is proposed. They are interpreted as a vertical succession of inner shelf, lower shoreface to upper shoreface deposits of an exposed, wave-dominated coast at the south shore of an open shelf sea. They constitute a normal regressive progradation fed by a mixture of shelf and continent-derived clastics. At least near Flobecq, the top may contain a second normal regressive deposit indicative of a southwards coastal shift. The top sands at Oudenberg and Kesterberg, also traditionally mapped as Diest, are reinterpreted as outliers of the Middle-Eocene Lede Sands, based on a sedimentological revision of the immediately underlying Gentbrugge Formation. The Hageland Diest Sands are thought to be the sedimentary fill of submarine current scour troughs, created at the opposite side of a semi-enclosed tidal embayment, tributary to the Roer Valley Graben, that was rapidly filling from the NW during high base level under a high sediment supply. The depositional environments of the Flemish Hills Sands and the Hageland Diest Sands are incompatible. An analysis of the central Belgium river pattern, including the smallest dry upper thalwegs, suggests that it originated on a Miocene, long-lived, slowly tilting surface draining to the Roer Valley Graben in the NE. The Flemish Hills were no morphological barrier and hence were buried at that time. Additional arguments based on basin architecture and the possible link of the coarse continental clastics to the Pyrenean tectonic phase, lead to the hypothesis that the Flemish Hills Sands may be outliers of the lower cycles of the Upper-Eocene Bassevelde Member of the Zelzate Formation. If confirmed, many elements of the geomorphological genesis of central Belgium need revision.

KEYWORDS: Flemish Hills, shoreface deposits, Kesterberg, tidal embayment, Pyrenean phase, consequent rivers

1. Introduction

It is common knowledge that the last major marine transgression in central Belgium and the neighbouring part of northern France was the Upper-Miocene Diest transgression that covered at least the area north of Calais – Cassel – Ronse – Maastricht (Fig. 1). The river drainage system in central Belgium (Fig. 1) is a legacy of this final transgression, as it is seen as a set of parallel water courses, developed consequently to the shore of the Diest sea that left the area definitively retreating to the NNE. On uplift of the area, the consequent drainage pattern was conserved and perpetuated by vertical erosion. This far-reaching geomorphological paradigm is ultimately dependent on one crucial geological interpretation: the assignment of the sand layer found in the upper part of the Flemish Hills (Fig. 1) to the Upper-Miocene Diest Formation.

1.1. The unification of the top Flemish Hills deposits and the Diest Formation

Dumont created the term "diestien" in 1839 to designate the glauconitic, medium to coarse sands and ferruginous sandstones in the area around Diest, i.e. in the Hageland (Fig. 1) (Tavernier, 1954). In 1849, Dumont extended the term to incorporate part of the "sables gris d'Anvers" found near Antwerpen and in the Kempen (Campine) subsurface. After field work in 1850, he applied the stage name to the sand and ferruginous sandstones that crown the Flemish and North-French hills (Dumont, published posthumously by Mourlon in 1878). Dewalque (1868) and Ortlieb & Chellonneix (1870) followed. After an initially different interpretation (Delvaux, 1881), also Delvaux (1884) came around. The name was since then widely adopted by the late 19th and early 20th century geologists. Also the latest official geological maps, published as GIS maps online (dov.vlaanderen.be, geolwal.be and editions.brgm.fr/cartegeol.jsp), follow this well-established interpretation. But is it true?

The top-Flemish Hills sands contain no fossils whatsoever. From the 19th century publications, it appears that their assignment to the Diest Formation relies on the following main arguments:
- they are the uppermost and thus the youngest marine deposit in the Flemish Hills, just like the Diest Sands in the Hageland;
- they contain glauconite;
- they show massive cemented caps of ironstone;
- their internal sedimentary structure contains cross stratification;
- the line of Flemish Hills seems to be a prolongation of the Hageland hills (Fig. 1). The Kesterberg is a significant relay site on this curved line.

Nevertheless, their stratigraphic position has never been settled definitively. Cogels & van Ertborn (1882) stated that ages ranging from Eocene to Quaternary had been proposed for them and launched a call to find fossils, providing clues to where one could find them; but all searches remained without result. Also the sedimentary context puzzled the scientists. A discussion among attenders of the 1884 field trip to Muziekberg and Pottelberg, described in detail by E. Delvaux (1884), is illustrative. After the field observations, the participants proposed several hypotheses

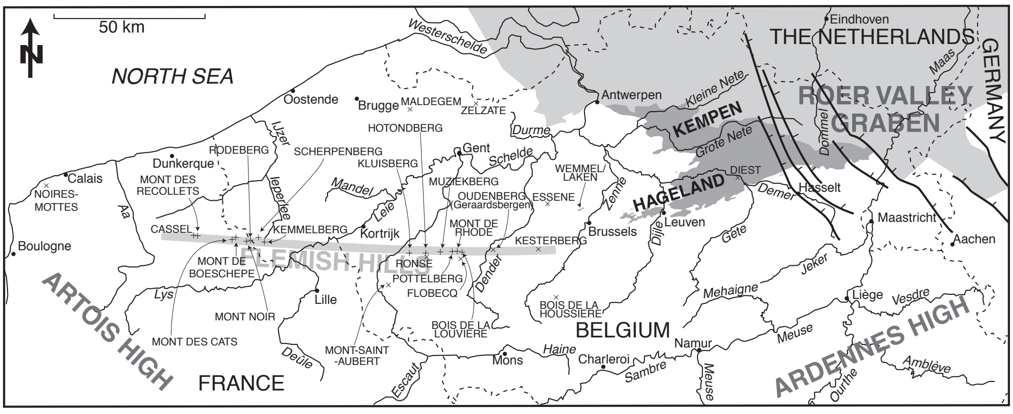

Figure 1. Overview map of north and central Belgium and north France, with indication of the localities named in the text. Outcrop area of the Upper-Miocene Diest Sands in dark grey (Hageland and Kempen). The formation dips to NNE. The younger Neogene cover is shown in pale grey. The line of Flemish Hills is indicated in grey. Hills where Flemish Hills Sands are found are indicated by a "+". Other locations are shown using an "x".

for the depositional environment of the over 20 m thick top deposits: Quaternary fluviatile deposits, a river delta or gravel bars on a beach; the topmost glauconiferous sands were however seen as a sure equivalent of the Hageland Diest Sand, but as a preliminary conclusion: "*Le mot diestien, appliqué aux assises supérieures de la région, n'est que le maintien provisoire d'une dénomination attribuée par Dumont, un point d'interrogation jeté à l'avenir*." (Delvaux, 1884, p. 98)

However, the stratigraphic fate was sealed. Though some doubt occasionally shimmered through, such as the quoted use "diestien" (e.g. in Gulinck, 1960; Laga, 1973; Sommé et al., 1999), and some suggested older interpretations were possible (e.g. Pomerol, 1973), "Diest Formation" remains the denomination in use today. The transgression was considered to have occurred "rapidly" because the base of the Diest Formation has conserved irregularities that were flooded before any marine action could have flattened the landforms (Cornet, 1904; Gullentops, 1957).

1.2. Paleogeographic implications

Far-reaching implications were based on the inclusion of the Flemish Hills top deposits in the Diest Formation. This assumption implies that the Upper-Miocene Diest marine incursion had reached disproportionally deep inland to the West and South, into Northern France and hugging the Weald and Artois High, transgressing over Lower-Miocene, Oligocene and Eocene deposits (de Heinzelin's (1964) fig. 1 gives a good impression of this paleogeography). The consequence is that the Diest deposits on regression must have covered any pre-existing continental landforms, and every present-day geomorphological feature in the area flooded by the Diest sea must postdate this single final marine incursion. The pattern of the main central Belgium rivers was thus sure to be younger than the Diest transgression and the very pattern was seen as a proof of this final transgression. The Upper IJzer, Leie, Middle Schelde, Dender, Zenne, and Upper Dijle, Gete and Demer (Fig. 1), which all contain a nearly parallel trunk running (S)SW to (N)NE, were seen as the present-day incised valleys derived from a set of consequent rivers developed on a NNE-wards retreating Diest sea coast. This idea was already expressed in the last decades of the 19th century, though some opposed it. Lohest (1900) (in Cornet (1904), Demoulin (1993)) interpreted the direction of most of the central Belgian rivers consequent to the Oligocene Chattian sea, not taking into account and thus implicitly rejecting the Diest transgression.

The apogee of Diest-related "consequentalism" is the well-argued treatise by Cornet (1904). He demonstrated that the northeast flank of the Artois High represented the maximum extent of the Diest sea. In central Belgium, the Diest coast was thought to have reached south of Mons and Charleroi, and probably south of Namur and Liège as well. His main arguments are isolated occurrences of ironstones and widespread gravel layers containing weathered flint pebbles and rolled white quartz pebbles, like those found in the Flemish Hills Diest Sand, present at the base of the Quaternary löss and river deposits north of the maximum extent line he proposed. All main rivers flowing NNE to N from this maximum flooding line were interpreted as Diest-consequent water courses. Cornet's (1904) theory was criticized in the sense that flint pebbles and ironstone relics could be no proof of an exclusive reworking of now-eroded Diest deposits; they might equally well have been reworked from other flint and ironstone containing strata. The maximum-extent line of the Diest Formation was pushed north, while remaining south of the row of Flemish Hills; the basic idea of Diest-consequent rivers

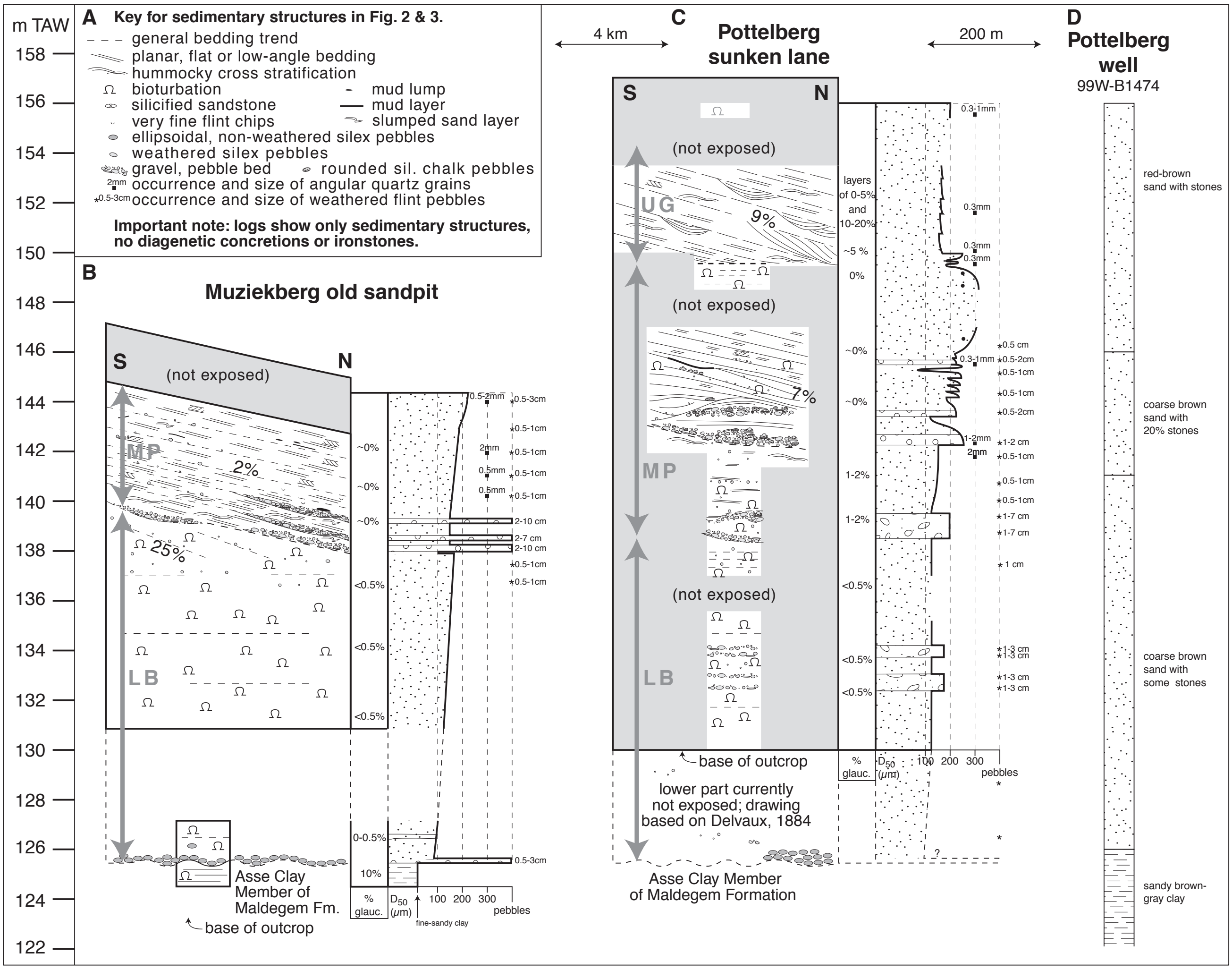

Figure 2. Vertical succession and grain size log of the sedimentary facies of the Flemish Hills Sands. A. Key of symbols for sedimentary structures in Fig. 2 & 3. B. Muziekberg abandoned sandpit. C. Pottelberg sunken lane. Location of outcrops: see text. D. Interpretation of well 99W-B1474. Vertical bed succession indicated in grey (B, C): LB: lower bioturbated beds; MP: middle pebbly beds; UG: upper glauconiferous beds.

north of this maximum-extent line was maintained (de Heinzelin, 1964; Tavernier & De Moor, 1975; Laurant, 1976 and references therein).

Another well-known theory (Gullentops, 1957) that interprets the Hageland Diest Sands as the deposit of an offshore tidal sandbank system, also depends on the stratigraphic position of the Flemish Hills top deposit in the Diest Formation. The Flemish Hills are seen as a coastline of the Diest sea; the Hageland hills are relics of marine tidal sandbanks and their shape has been preserved, though accentuated by differential erosion afterwards, due to a rapid regression. The start of iron cementation is situated early after regression and high positions in the bank crest were thought to be preferential iron cementation areas; this explains the present-day preservation of the sandbanks as high relief landforms (Gullentops, 1957).

The far western extent of the Diest transgression provided Gullentops (1988) a mechanism to account for the incised base of the Hageland Diest Sands and the coarse sand fill. He supposed a seaway connection existed between the English Channel and the North Sea through the Weald-Artois High, established during the Diest transgression, that funnelled tidal currents through the strait. The strong currents eroded the WSW to ENE linear depressions in the Hageland Diest base and supplied sand and glauconite from the English Channel to settle in this area east of the straits.

1.3. Aim of this paper

Considering the scarce and dubitable evidence for the stratigraphic position of the Flemish Hills Diest Sands and the many geomorphologic implications, it is justified to re-examine existing data and bring forward new facts to help settle the open questions.

In this paper, I describe, interpret and compare the sedimentary structures of both the Flemish Hills and the Hageland Diest Sands. The sedimentary structures of a sediment deposit, along with field observations of the vertical and lateral variations in grain size and petrographic components, allow to reconstruct the sedimentation environment and help constraining the paleogeography. Lateral facies variations and concepts of sequence stratigraphy assist in providing realistic transitions inside the depositional basin. As all deposits considered here are at the surface and may locally have been the last marine deposits before definitive emersion, I also re-analyse the pattern of the present-day hydrographic drainage system of central Belgium, searching for genetically primary directions.

As the different Diest Sand outcrops are reputedly sterile, any paleontological approach is precluded. This paper does not consider the characteristics and genesis of palaeosols nor of the ironstones of the Flemish Hills and the Hageland hills. After all, iron cemented stones are found in many Cenozoic formations, often in elevated locations (sites for the Nete and Dijle basins are given by Bos & Gullentops, 1990). Also, no specialist attention was paid to the petrographic composition of the studied sand deposits, as the outcrops studied are affected by strong alteration which would complicate interpretation. However, advanced new studies on these matters are required and will contribute to decide on the final interpretation and stratigraphic position of the different sand deposits that are until today united in the Diest Formation.

As this paper will show that the sands on top of the Flemish Hills, from Cassel to Bois de la Louvière (Flobecq) (Fig. 1), do not belong to the Diest Formation, the new lithostratigraphic name "Flemish Hills Sands" is used throughout this paper for these sands. The typical Diest Sands such as they occur on hill tops near Brussels, Leuven and Diest are called Hageland Diest Sands here. The Kempen subcrop Diest Sands are not considered in this article. Their relation to the Hageland Diest Sands is discussed in Vandenberghe et al. (2014). The focus area of this study is called here "central Belgium"; the name is used informally to indicate the area enclosed between Cassel, Gent, Hasselt, Liège, Namur, Mons and Lille (Fig. 1). Throughout this paper, many stratigraphic names spanning a good part of the Cenozoic are used. Reference is made to the lithostratigraphic tables published at http://dov.vlaanderen.be and http://natstratcommbelgium.drupalgardens.com.

2. Methods

Most new observations in this paper were done during classical fieldwork. Locations were determined in the field based on a hand-held GPS device and elevations were measured based on elevation contours on the 1/10,000 topographic map, with reference to the Belgian TAW datum. Their precision is in the order of a metre. Field estimates were made of the median grain size, content and size of coarse elements, and glauconite pellet content. The focus of the observations are the primary sedimentary structures. They were observed in freshly cleared sections. Many outcrops contain numerous epigenetic ironstones and capricious discolorations due to weathering and differential precipitation of mainly iron oxides. While observing and taking down the sedimentary structures, abstraction was made of these secondary features; and though sometimes they may help observation when primary structures guided cementation, in general, they obscure the primary structures.

2.1. Locations studied for the sedimentary structures of the Flemish Hills Sands

A composite sedimentary section was made of the top deposit in the Muziekberg hill NE of Ronse (hill summit at Belgian Lambert 72 coordinates X 98,720; Y 161,750; altitude 148 m TAW), using data published in Delvaux (1884), new observations in dispersed small outcrops in the woods covering the south side of Muziekberg and larger outcrops in the north and east faces of an abandoned sandpit just north of the summit of Muziekberg (Lambert 72 coordinates X 98,500; Y 161,820).

A second composite section was made using dispersed outcrops in a sunken lane leading from the south to the summit of Pottelberg (summit at X 102,690; Y 161,615; altitude 158 m). This very road was also visited and discussed by a group of notorious late 19th century geologists, among whom J. Ortlieb, A. Rutot, E. Van den broeck, G. Velge, P. Cogels, O. van Ertborn, led by E. Delvaux (1884). They had more fresh outcrops, which showed essentially the same characteristics as were observed now. The section also compares very well with two sections of the nearby, now abandoned sandpits, at Pottelberg and Bois de la Houppe, published by de Heinzelin (1962), and with the abandoned Mont de Rhode sandpit (X 104,250; Y 162,050).

The field observations have been summarized in the sedimentary logs of Fig. 2 and photographs of typical structures are presented in Plate 1.

2.2. Locations studied for the sedimentary structures of the deposits in the top of the Kesterberg

The Kesterberg (summit at X 131,355; Y 162,915; altitude 111 m) Diest outlier was, due to its position between the Flemish Hills and the Hageland hills (Fig. 1), seen as a proof of the originally continuous character of the Diest marine deposits (Delvaux, 1884; Cornet, 1904; Leriche, 1914).

A new composite, though not complete, sedimentary section is compared in Fig. 3 with sections published by Rutot (1882c), with the section derived from the present-day DOV geological map, with the description of borehole 101w-B79 (DOV; Geets, 2001) and with De Ceukelaire's (2009) interpretation of boreholes 101w-B49 and B79. My observations were done in a 1989 drinking water tank construction pit just south of the summit of Kesterberg (X 131,260; Y 162,705) and in some small roadside outcrops at the site of the former motocross, just west of the summit. They were combined with observations of the deeper deposits at sites not far from Kesterberg. A unique 2010 outcrop of a drinking water tank construction pit on the Oudenberg hill at Onkerzele near Geraardsbergen, 14 km west of Kesterberg, at X 116,915, Y 163,085, and a few nearby small construction pit outcrops were key in deciding on the stratigraphic position of the layer immediately below the Kesterberg top sand. Finally, construction pits in Kester village 2 km SE of Kesterberg complete the section at its lowest level.

The field observations have been summarized in the sedimentary logs of Fig. 3 and photographs of typical structures are presented in Plate 2.

2.3. Locations studied for the sedimentary structures of the Hageland Diest Sands

Good outcrops in Hageland showing clear sedimentary structures are relatively hard to find. Many outcrops show iron cemented stone layers, plates and blocks with often capricious boundaries, which obscure the primary sedimentary structures. However, fresh excavations often consist of completely loose glauconite sand devoid of ironstones and crusts (see also Bos & Gullentops, 1990).

The main outcrops studied are Kesselberg in Leuven (Kessel-Lo; X 174,860; Y 177,440), Oude Diestsebaan in Tielt-Winge (Sint-Joris-Winge; X 185,130; Y 178,440) and temporary construction pits near Gasthuisberg Hospital in Leuven (X 171,280; Y 174,040). Supplementary outcrops used in this study are roadside outcrops in Wezemaal (Benniksberg), Herent (Mechelsesteenweg), Holsbeek (Sukkelpotweg, Meesberg), Kessel-Lo (Meesbergpad), Nieuwrode (Aarschotsesteenweg), Linden (Zavelstraat), Molenstede (N127), Webbekom (Oude Tiensebaan) and Everberg (E40), and temporary building site outcrops in Veltem (Nikbergenstraat), Laken (Diepestraat and Romeinse Steenweg), Assent (Kwadestraat and Struikstraat) and Kessel-Lo (Leming). Local sedimentary structures and grain size characteristics were described in the field.

The observations allowed the definition of five sedimentary facies. Photographs of typical structures are presented in Plate 3. Attention was further paid to the vertical and lateral facies transitions.

2.4. Methodology to study the pattern of the central Belgium drainage system in order to detect a primitive drainage direction

River patterns, once established, tend to persist (Twidale, 2004), i.e. when an emerging land surface is uplifted and tilted, the rivers will incise, at least over significant parts, their original course and thus provide information of the first emersion of a land surface. A re-evaluation was carried out of the central Belgium drainage system, taking all valleys into account, also the smallest, dry upper parts.

A GIS database was made of all valley thalwegs in central Belgium (Fig. 4) that can be interpreted in the 1/10,000 topographic maps' signature of the elevation contour lines. This method was preferred to deriving thalwegs automatically from a digital elevation model (DEM). Working with DEMs allows to obtain in addition longitudinal slope profiles, but they were not needed in the present analysis that focuses on the geographical pattern. Interactively interpreting thalwegs on contour line maps allows to produce a well-structured dataset that contains coherent, uninterrupted draining channels. Clearly anthropogenic water courses, such as mill feeding diversions and the many dug ditches in wide valley bottoms, have been left out. In the lower reaches where rivers have been canalised, the original, more sinuous water course has been digitised when still identifiable on the maps. The valley thalwegs were digitised as polylines beginning at their upstream origin and proceeding downstream. The digitisation was carried out on a scale of 1/10,000. Some generalisation was applied, especially of meandering water courses, in the sense that vertices defining the polylines were spaced about 50 to 200 m. Thalwegs were categorized as follows:

- "dry": upper part of thalweg lacking a permanent water course. This category thus includes well-recognizable dry landscape thalwegs and upper intermittent brooklets;
- "upper": upper part of thalweg having a permanent water course such as indicated on the topographic map. Upper courses were for the purpose of the present analysis defined as the part of the

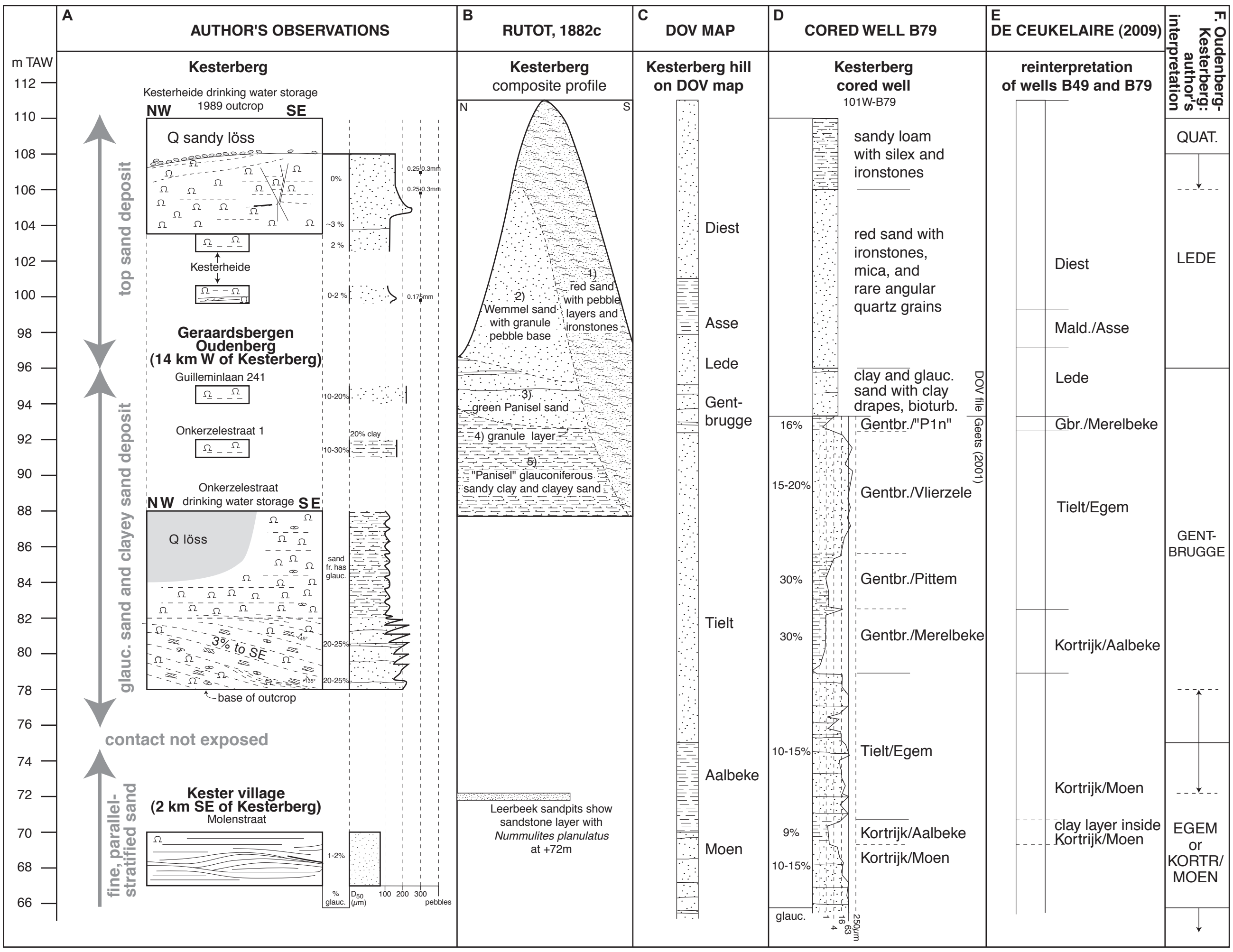

Figure 3. Vertical succession and grain size log of the sedimentary facies of Kesterberg hill. A. Author's observations on Kesterberg and Oudenberg hills. Symbol key in Fig. 2. Location of outcrops: see text. B. Drawing and geological interpretation of Kesterberg by Rutot (1882c). C. Vertical log from the Flemish Region geological map (DOV). D. Cored well 101W-B79: upper part from description in DOV files; lower part with detailed grain-size curve and interpretation from Geets (2001). E. Reinterpretation of wells B49 and B79 by De Ceukelaire (2009). F. Author's stratigraphic interpretation.

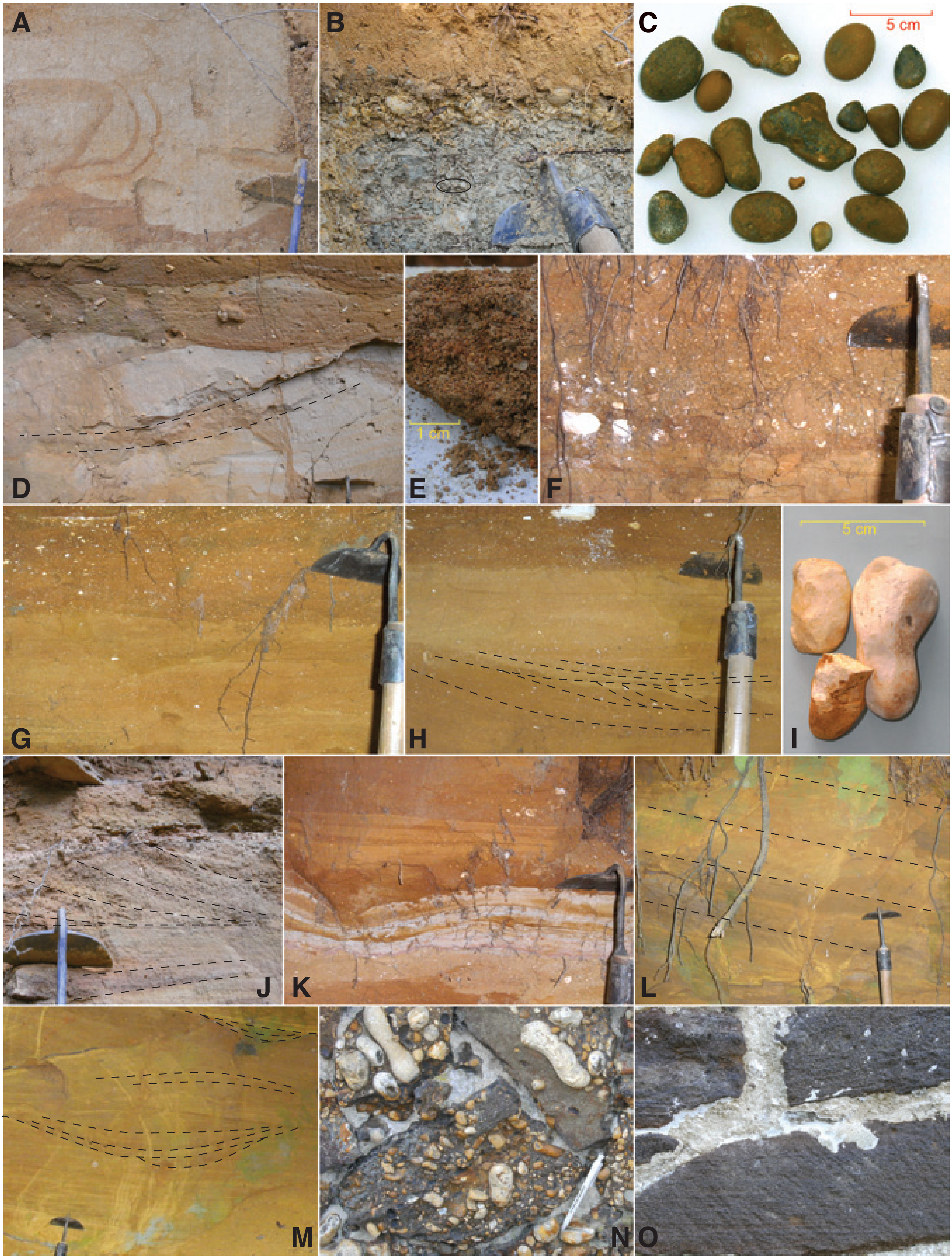

Plate 1. Photos illustrating the variation in Flemish Hills Sands sedimentary structures. In all photos, the blade of the scraper tool is 11 cm wide. **A**. Lower bioturbated (LB) beds, Bois de la Louvière abandoned sandpit, Flobecq. **B**. Contact between the Asse Clay (grey lower unit) and the Flemish Hills Sands (orange upper unit) in Muziekbos, Ronse. The contact surface is covered by flint pebbles. Some pebbles (marked by ellipse) are found a few centimetres below the contact surface. **C**. Flint pebbles from the Flemish Hills Sands base at Muziekbos. **D**. Top of the LB beds in the Muziekberg abandoned sandpit, Ronse. The top contact with the middle pebbly (MP) beds is here accentuated by the appearance of the brown-orange colour. The top part of the LB beds contains more flint pebbles and they are here arranged in cross-beds with foresets dipping left (NW); much of the primary cross-bedding has been homogenized by bioturbation. **E**. Coarse sand containing a high percentage of coarse, angular, 1-2 mm quartz and flint grains, and rounded 5-10 mm flint pebbles. Small outcrop in top of MP beds in building site at La Houppe, Flobecq. **F**. Pebble layer on cross-bedded and bioturbated sand in the MP beds, Pottelberg sunken lane, Flobecq, near base of MP beds. The pebbles float in a sand matrix. Most pebbles are round, weathered flint. The white, broken pebbles are interpreted to be silicified chalk. **G**. Dispersed small flint pebbles in horizontal bedded, low-angle inclined bedded and bioturbated sand, higher up in the MP beds. Pottelberg sunken lane. **H**. MP beds section containing a pebble bed in sand matrix on horizontal bedded, low-angle

inclined bedded and cross-bedded sand with many flint and chalk granules. Pottelberg sunken lane. **I**. Silicified rounded chalk pebbles from the MP beds of the Pottelberg sunken lane. The upper left pebble contains small, grey, round flint pebbles at its upper left and lower left corner in the photo. **J**. Thin cross bed in sand containing many angular coarse sand grains; foreset apparent dip to right (SE). The cross bed's truncated upper surface is draped by a 1 cm thick pinkish clay layer. Muziekberg abandoned sandpit. **K**. Parallel-bedded section near the upper boundary of the MP beds at Pottelberg sunken lane. Prominent package in the centre of thick, pinkish clay layers separated by sand laminae, showing synsedimentary deformation. **L**. Straight inclined bedded upper glauconiferous (UG) beds in Pottelberg sunken lane. Beds dip to right (N). Primary bedding obscured by weathering and lichen cover, but the primary lamination is visible through differences in glauconite content. **M**. Close-up of the UG beds in the Pottelberg sunken lane, with some units of swaley (top and bottom) and hummocky cross-stratification highlighted. North to right. **N**. Building stone of the viewpoint parapet in Cassel, N. France. Conglomerates of flint pebbles in ironstones are similar to those near Ronse. Pen is 14 cm long. **O**. Building stone in the church of Oxelaëre near Cassel, N. France. This facies is iron-cemented MP of the Ronse area. The stone layers are about 10 cm thick.

thalweg down to a distance of 10 km from the upper divide;
- "middle": central part of thalweg, the section between a distance of 10 km and 20 km from the upper divide;
- "lower": the section downstream of the middle part.

It was decided not to categorize streams using branching-dependent criteria, such as the Horton-Strahler number. This provides information on the number of headwaters at each junction. This information is not needed in the geographical analysis undertaken here. Moreover, the Strahler-Horton laws provide no information on the structure or origin of a stream network (Kirchner, 1993; Hodgkinson et al., 2006).

The category defining distances of 10 and 20 km have no morphological meaning, but do in broad terms correspond to structural changes of the drainage pattern in the investigated area; this was noted during the digitization: after some 10 km from the divide, often important direction changes in the central draining stream are found, while a further 10 km downstream, the main draining trunk has often (in some sections of its course) returned to the drainage direction found in the headwaters. This will also appear from the processed results.

In an area of about 12,500 km², all thalwegs were digitised (Fig. 4). This area covers all of central Belgium and includes adjoining areas of northern France. Some outlier plateaus in north Belgium and a neighbouring area south of central Belgium were also covered in the database, but not included in the quantitative analysis of "central Belgium". Inside "central Belgium", the database contains a total length of 29,400 km of thalwegs of which 67% dry, 27% upper, 3% middle and 3% lower courses. Dry and upper thalwegs are predominant and not including them in the direction analyses would result in missing out much information.

The database was submitted to a direction analysis. Each polyline segment, defined by its upper and lower vertex, was attributed a direction class. Direction is expressed in degrees with 0° the thalweg descending due north, 90° due east, etc. Direction classes were defined in 10° steps. All directions from 355° to 360° and from 0° to 5° were attributed the direction class 0; all directions from 5° to 15° were assigned to class 1; and so on, producing 36 classes. All orientations are with reference to the Belgian Lambert 72 grid. The length of each segment was summed per class. The class importance was then expressed by the percentage of length of the segments in the class versus total length. The class with the peak percentage thus constitutes the dominant direction. The result of the analysis is a direction frequency chart either for all river segments together or per thalweg category.

Separately from the quantitative analysis, the GIS database was also examined visually. The human eye is a powerful pattern analyser, and preferential directions can be drawn and displayed

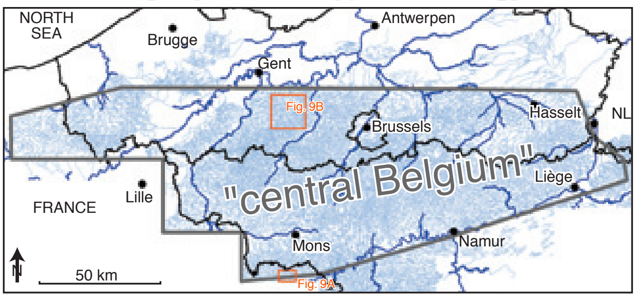

Figure 4. Overview map of the area analysed for drainage pattern. "Central Belgium" = area taken into account in the direction diagrams in Fig. 8.

in a GIS. The visual analysis was intended to detect preferential directions in function of a possible primitive drainage direction. Features such as those labelled 1-4 in Fig. 9 were used as indicators. Special attention was paid to find the dominant drainage direction of the dry and upper thalwegs on the dividing plateau areas. It appeared that the dominant direction was often repeated in sections of the incised, lower reaches of main watercourses. In the visual analysis, attention was also paid to the divide outline shape of the many catchments and to the position of the central water course inside its catchment. The resulting direction found representative of an approximately 20 by 20 km area (and not looking to the adjoining area) was shown by drawing an arrow. The complete thalweg database was stepped through area by area until the map was covered by arrows. Afterwards, the orientation of the arrows was calculated.

3. Sedimentary structures and facies variations in the Flemish Hills Sands

The Flemish Hills Sands can in the Ronse area (Fig. 1) be subdivided in three units that form a vertical succession (Fig. 2): a lowermost bioturbated sand; a middle part with low-angle parallel lamination and numerous pebbles, dispersed and in layers; and a top part with inclined parallel lamination and relatively high glauconite content. The subunits are not separated by sharp boundaries. The deposits are devoid of carbonates and fossils.

3.1. The lower bioturbated beds ("LB beds")

The LB beds are mostly pale yellow-grey, but can locally be orange or pink. They contain few to no ironstones. As its distinctive feature, the LB unit is homogenized by bioturbation (Plate 1A).

This unit is at the base of the Flemish Hills Sands. Its thickness varies from 11 to 14 m. The underlying Asse Clay Member of the Maldegem Formation is in the Ronse area a 4 to 5 m thick unit of fine-sandy clay with relatively coarse glauconite pellets. The top of this layer is eroded and burrowed from above. The erosion surface is reported in the literature to be covered by a gravel of well-rounded, dark grey flint pebbles. Delvaux (1884) mentions various thicknesses of the gravel layer, from a few centimetres up to over 0.5 m. I found a small outcrop on the south side of Muziekberg. Here, the gravel is only one layer of 1 to 5 cm flint pebbles thick (Plate 1B). The pebbles are iron stained and have ellipsoidal or rounded, irregular shapes (Plate 1C). Some of them are partially weathered; their surface displays patches of a weathered (cacholong) rind. Presumably by burrowing activity, occasionally a pebble is found a few centimetres down in the underlying clay (Plate 1B).

On top of the base gravel, a glauconiferous, clayey sand is exposed. The glauconite content of that sand layer is very low and the glauconite pellets are much finer than in the underlying clay.

Going upwards, the package becomes more purely sandy. The median grain size is between 120 and 150 µm and shows an upward coarsening trend. The sand consists mostly of rounded quartz grains and also contains mica-plates. The observed glauconite content is less than 0.5%.

In the Muziekberg abandoned sandpit, the lower sand unit contains no pebbles, except in the upper 2 m, where they appear to be aligned in strata inclined at 25% dipping north (Plate 1D). In most outcrops however, the LB beds contain dispersed small pebbles often measuring about 1 cm diameter. They are weathered, round flint pebbles. The weathering is expressed by the presence of a white, often iron-stained cacholong rind. The pebble content is low, around 1%, locally a bit more. The arrangement of the pebbles in the LB beds often suggests a subhorizontal original

Plate 2. Photos illustrating the variation in sedimentary structures of the top deposits of Kesterberg (Gooik) and Oudenberg (Geraardsbergen) hills. In all photos, the blade of the scraper tool is 11 cm wide. **A**. Bioturbated fine sand of the Kesterberg top sand deposit, originally horizontal but tilted and faulted possibly by Quaternary hill side movements. Note double clay layer displaced vertically by 20 cm due to the faulting. 1989 water storage construction pit at Kesterberg. **B**. Interbedded interval of fine, bioturbated sand and medium, stratified sand. Note 5 cm thick cross bed in the middle, with some glauconiferous laminae near the top. Top sand deposit at Kesterberg, near base. **C**. Inclined sand and mud layers in the lower cross-bedded unit of the glauconiferous sand and clayey sand deposit exposed in the 2010 water storage construction pit at Onkerzele, Geraardsbergen. The pit face is oriented NNW (far) to SSE (near end). **D**. Originally cross-bedded sand with mud drape couplets, disturbed by burrows of type *Ophiomorpha* and small *Palaeophycus*. Detail of the face depicted in photo C. Foresets dip right. **E**. Thin cross beds and stacks of mud-drape rich bottomsets in the lower cross-bedded unit. Detail of the face depicted in photo C. Note that here, foresets dip left. Arrows indicate a few instances of double mud drapes. **F**. Package of descending thin cross beds in the lower cross-bedded unit. Note disturbances due to bioturbation. Many mud drape couplets occur in the foresets and bottomsets; a few have been indicated by arrows. North face of Onkerzele water storage construction pit, ENE to right. **G**. Typical silicified sandstone

of the lower cross-bedded unit, broken at the excavation of the Onkerzele water storage construction pit. **H**. Middle bioturbated very clayey sand, here in a sandy clay section. Onkerzele water storage construction pit. **I**. Transition of the lower cross-bedded unit to the middle bioturbated very clayey sand in the Onkerzele water storage construction pit. Face oriented WSW (left) to ENE. Note gradual increase in grey clay content above the scraper tool. **J**. Middle very clayey sand, here a few metres higher than photo H. The sediment is less clayey than in photo H. Temporary outcrop in Onkerzelestraat, Geraardsbergen. **K**. Upper bioturbated, glauconiferous sand. Temporary outcrop in Guilleminstraat, Geraardsbergen.

stratification. They are not arranged in coarser layers; they are dispersed in the surrounding, fine sand. The content of pebbles seems to increase upwards, but there is no clear trend. Delvaux (1884) mentions dispersed pebbles in hand-drilled cores near the base of the Pottelberg sunken lane section.

However, going upwards, also concentrated pebble layers occur. The currently exposed Pottelberg sunken lane section shows at least four pebble layers inside the lower bioturbated sand. They are around 10 to 15 cm thick and contain 1 to 3 cm thick, weathered, rounded flint pebbles in a fine sand matrix. My 1988 observations in the nearby, now abandoned Mont de Rhode sandpit showed the pebble layers to be concentrated at the base of shallow channels cut in the surrounding sand.

3.2. The middle pebbly beds ("MP beds")

This unit is 4.5 m thick at Muziekberg and 11 m at Pottelberg.

In the Muziekberg abandoned sandpit, the contact between the LB beds and the MP beds slopes down 2% to the north. It sits at an altitude of 135 m in the north face and 140 m in the south face of the sandpit. In the Pottelberg sunken lane outcrop, the contact is around 138 m. According to sections recorded by de Heinzelin (1962), in the Pottelberg sandpit, the base of the middle pebbly sand is at 142 m and in the Bois de la Houppe sandpit at 130.5 m. The contact is gradual; from the outcrops, it is clear that the LB and MP beds interfinger. This was also observed by Tavernier & de Heinzelin (1962).

The MP unit may stand out in the outcrops by its brown colour and the high content of iron cemented stones. The unit is coarser grained than the unit below; the median grain size is 150 to 250 µm, coarsening upward, with near the top some layers coarser than 300 µm. The sand consists dominantly of well-rounded quartz grains, but typically contains an admixture of angular, coarse quartz and flint grains, measuring 0.5 to 2 mm. The content of angular grains increases upward (Plate 1E). The glauconite content is low, less than 2% and often around 0%.

The main sedimentary structure is inclined parallel stratification. Stratification dips are about 2% in the Muziekbos sandpit and up to 7% in the Pottelberg sunken lane, both to the north.

A striking feature are the pebble beds (Plate 1F-G), mostly near the base of the unit, but also higher up. The pebbles may have long axes of 7-12 cm, sometimes even more, in the lower pebble beds, while higher up, the pebbles are about 1-3 cm long. The thickness of the pebble beds is about 10-30 cm, with the thicker beds occurring near the base of the unit. The pebbles occur in a matrix of sand. The pebble beds fill shallow depressions and may also slope according to the inclined master bedding. The pebble beds contain mostly well-rounded irregular and ellipsoid brown flint pebbles often with a weathered rind. In the finer fraction, partly rounded quartz and quartzite occur. According to Gulinck (1960), around 0.5‰ of the mass of the grain size fraction below 30 mm are kieseloolites (silicified oolitic limestone pebbles, with a Jurassic limestone origin) while around 4.5‰ are "gaizes globulaires" (siliceous cemented fine sandstone with microfossils, with a Late Jurassic Oxfordian sandstone origin). In the uppermost pebble beds of the Pottelberg outcrop, a considerable portion of the pebbles are well rounded to flattened-ellipsoidal, 1 to 5 cm long pebbles that are thought to be silicified chalk pebbles (Plate 1H-I). Though they have the characteristic rounded, somewhat dented, shape of chalk pebbles, they consist of weakly crystalline quartz and some amorphous opal (XRD analysis by Rieko Adriaens, pers. comm.). They were also mentioned by Gulinck (1960), without specification of the vertical trends.

In spite of the many pebbles, most of the unit is sandy with dominant arrangement in inclined parallel stratification. Small pebbles, finer than 1 cm, and fine flint chips occur throughout the unit. Rare clay pebbles occur also, dispersed through the unit. Some clay pebbles may be longer than 10 cm. Subdominant structures include very small ripple lamination in beds thinner than 1 cm. Often, up to 10 cm thick cross beds fill scours in the master bedding (Plate 1H & J). In one outcrop in the abandoned Mont de Rhode sandpit, cross-bedding topsets and foresets dipping east were seen. Furthermore, shallow depressions (channels or scoops) with channel fill cross bedding are rather common. Some depressions show a sand fill containing thin or up to 10 cm thick clay lamination (Plate 1K). Downdip sand slumps are observed, but the structure is obscured by iron cemented capricious forms and it is not clear whether the slumps are a primary feature. Subdominant burrows and local beds homogenized by bioturbation are present. Some of the sloping strata show hummocky lamination.

3.3. The upper glauconiferous beds ("UG beds")

This unit was only found in the top metres of the exposures on Pottelberg and Mont de Rhode, where it is about 3.5-4 m thick. The present description is from the sunken lane exposure. Pale yellow quartz sand laminae, almost devoid of glauconite, alternate with brown-green sand laminae that contain up to 20% of fine, dark green glauconite. The contact with the MP beds is not so well exposed, but the available small outcrops indicate a gradual and probably interfingering contact; there is certainly no sharp, erosive, or gravel-lined lower bounding surface.

The unit consists of fine sand of around 150-175 µm.

The major sedimentary structure is inclined parallel bedding, dipping up to 9% to the north (Plate 1L). The sloping parallel sand beds are arranged in sharply delineated, flat planes, traced out by contrasting glauconite contents. The inclined beds are northwards prograding, downlapping strata with tangential bases.

Intercalated in the general inclined bedding are hummocky cross beds and shallow channel-fill cross beds (Plate 1M).

This unit contains no bioturbation; in the sunken lane exposure, it contains only few iron-cemented stones.

3.4. The top layer of the Flemish Hills

There is a hiatus of 1.5 m in the observations on top of the UG beds in the Pottelberg outcrop. The upper metre is homogenized, probably due to soil processes. It is not clear whether the original structure was like in the UG unit. The sand is a bit coarser than the UG beds, around 200 µm, and again contains stray coarse, angular quartz grains, which may be an indication of yet another sedimentary facies. Capricious ironstone plates and chunks occur up to the summit. There is no top gravel or plateau gravel.

Similar homogenized profiles without pebbles occur everywhere on the summits of the Hotondberg, Muziekberg and Pottelberg. All observations of small outcrops in the crest areas of the Flemish Hills were free of a Quaternary base gravel or any aeolian cover.

3.5. Geographical occurrence and vertical stacking of the facies

Some confirmation of the Pottelberg sunken lane succession can be found in the description of the Pottelberg flush well 99w-B1474, situated at 200 m of the top of the lane (Fig. 2D). Unit LB covered by MP has been observed in small and larger outcrops at Kluisberg, Muziekberg, Pottelberg and Mont de Rhode. Both units are also clear on de Heinzelin's (1962) sections of the Pottelberg and Bois de la Houppe sandpits, and his grain-size trend matches the present field observations. De Heinzelin (1962) had no observations at altitudes above 149 m and so missed out the UG unit. This unit though, along with the lower units, was clearly observed by Delvaux (1884). It is also present in the now abandoned Mont de Rhode sandpit. The upper glauconiferous unit occurs at the top from 149 m to the hill summit at 152.5 m, in the same vertical succession as at Pottelberg. It forms a horizontal layer with clear downlapping progradation of the inclined beds to the north; the lower limit was not too well exposed but appeared to be gradual just like in the Pottelberg sunken lane.

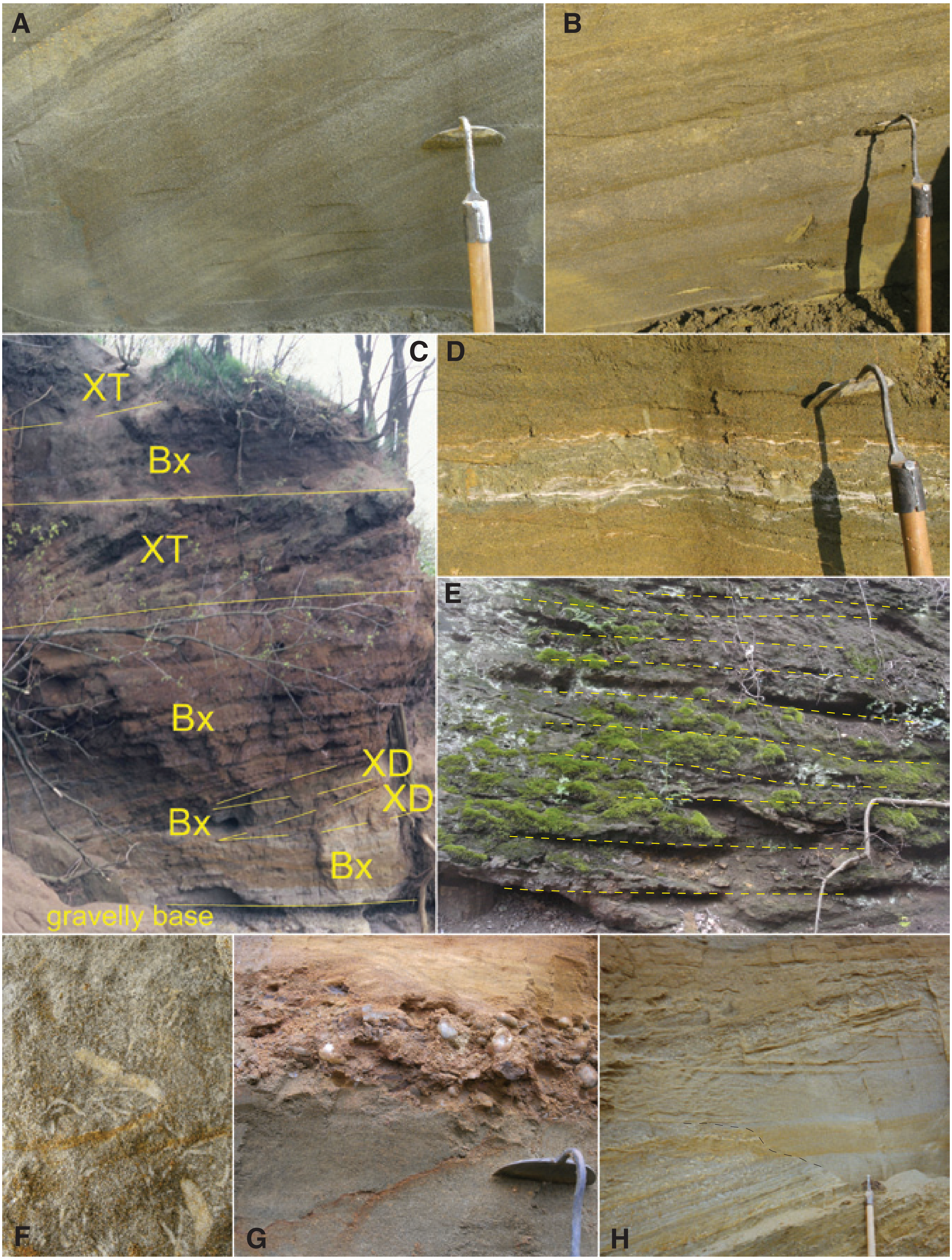

Plate 3. Photos illustrating the variation in sedimentary structures of the Hageland Diest Sands. In all photos, the blade of the scraper tool is 11 cm wide. **A**. Thick cross bed of facies XT in temporary outcrop at Veltem, Nekwinkelstraat. Note foreset slope change packages, here terminated above by whiter sand. In upper left, one such package has been bioturbated. Pit face oriented WSW (right) to ENE. **B**. Strike section (apparent slope to left is due to photo perspective) of the same thick cross bed of photo A. Note foreset packages, often terminated above by whiter sand. Some packages are burrowed from the top. Pit face oriented SSE (right) to NNW. **C**. Abandoned sandpit face, about 4 m high, at W-tip of Kesselberg, Leuven (situation in 1989). Section oriented WSW (right) to ENE. Note vertical stacking of different facies. **D**. Strike section of facies Bx in temporary outcrop at Veltem, Nekwinkelstraat. The light grey mud layers, disrupted by burrows, form a bottomset package of about 25 cm thick, descending, almost completely bioturbated cross-beds (note: the beds are not clear on this picture). Pit face oriented SSE (right) to NNW. **E**. About 2 m high section of iron-cemented facies XD in Meesbergpad sunken lane, Kessel-Lo (Leuven). Section oriented S (left) to N. The descending cross-bed boundaries are indicated by fine dashed lines. **F**. Facies Bm. Photograph of a loose sand lump (top of section probably to left) from the outcrop of photo G. White burrows are approximately 10 mm and 2 mm wide. **G**. Facies Bm in construction pit at Gasthuisberg hospital, Leuven, covered by round flint pebbles of Quaternary base gravel, covered by 5 m of sandy loam and löss. **H**. Thick cross-bed (bottom) truncated by an irregular erosion surface (partly shown by fine dashed line) and facies M. Its quasi structureless sand is in turn covered by thinner cross-beds. 1987 road excavation at Nieuwrode, central Hageland.

According to Rutot (1882a), the facies are similar in all of the Flemish Hills. He describes the summit deposits of Rodeberg near the French border: they contain 17 m of heterolithic, rather coarse sand, with 3 to 4 pebble beds near the base ("MP beds") on 4 m of "regularly stratified", glauconiferous sand ("LB beds") with a thick layer of rolled pebbles at the base covering the Asse Clay, very similar to those at Muziekberg. Ironstones used in historic buildings in the Belgian-French border area testify of the occurrence of both units mentioned by Rutot. The pebbles in the building stones have weathered rinds just like in the Ronse area. The pebble sizes are also comparable (Plate 1N-O). New observations in a few roadside outcrops on Rodeberg confirm the presence of facies LB and MP, including the coarsening upwards trend observed in the Ronse area. It remains to be sorted out whether the UG sand is found there. Kieseloolites and quartzite are mentioned by Briquet (1909) to occur in the easternmost Flemish Hills only, i.e. near Ronse.

The vertical succession described in this paper is thus representative of both the Cassel-Kemmelberg and the Ronse area.

The linear series of remnants formed by the Flemish Hills throughout northern France and western Belgium seems to indicate an original, linear area of occurrence; there are no outliers. One outlier 10 km south of Ronse, Mont-Saint-Aubert (Fig. 1), has traditionally been mapped with "Diest Sand" on its summit. There are presently no good outcrops, but the "Diest Formation" interpretation for the ironstone top deposit is doubtful (Vanneste & Hennebert, 2005); Rutot (1882a) thought they were Wemmel Sands. Weathered flint pebbles are found in a wide environment, tens of kilometres both north and south of the Flemish Hills (Cornet, 1904). While it is difficult to prove the pebbles are derived from the Flemish Hills Sands, nevertheless their presence may be an indication that the deposit occurred in a wider area than the line of Flemish Hills alone.

3.6. Interpretation of the sedimentary environment

The *basal erosive contact* covered by unequally distributed, unweathered or slightly weathered, rounded flint pebbles, marks a transgressive surface. The source of the unweathered flint pebbles is thought to be fresh, probably beach erosion of exposed late Mesozoic chalk. The presence of slightly weathered pebbles may be indicative of some subaerial exposure before transfer to the coast. The rounding shows some transport, such as by wave processes. After deposition, they are concentrated during renewed transgressive wave erosion, where the coast belt is swept landward. This transgressive erosion removes also the top of the possibly already truncated Asse Clay. The burrows below the erosion surface show the transgressive surface was transformed in a firmground. The deposit from which the round pebbles have been concentrated, has completely been eroded. No Eocene sediments containing such pebbles are known in the sedimentary record up to the Asse Member. Therefore, the eroded sediment was a younger deposit, possibly even the transgressive systems tract of the Flemish Hills Sands depositional cycle.

The *clayey sand*, covering the base gravel, indicates a relatively deep coastal shelf environment, well below fair-weather wave base. An inner shelf setting is a likely environment. Bioturbated inner shelf sands are common throughout the Belgian Cenozoic sedimentary record.

The bulk of the *lower bioturbated (LB), slightly glauconiferous sand* with a low clay content has also the typical characteristics of an inner shelf sediment deposited below fair-weather wave base. The scattered presence of weathered flint pebbles makes this deposit stand out. No other marine sand with pebbles dispersed through the deposit is known in the Belgian Cenozoic. However, in itself, it is nothing special (Hart & Plint, 1995; Clifton, 2005).

The flint pebbles must have been eroded from late Mesozoic chalk layers or reworked from such pebbles present in younger strata. Due to the irregular shape of many pebbles, de Heinzelin (1964) thinks they were the product of continental erosion. Indeed, they must have been exposed on a land surface to account for the weathering. Afterwards, they were transited to the coast. As in the LB-beds, especially in the lower part, the pebbles are fine and dispersed in bioturbated sand, it is thought they were eroded from the beach during storm and carried in suspension during the storm events and dispersed over the nearshore inner shelf. Offshore transport by storm waves will move the pebbles only over very short distances. The lower bioturbated sand was thus deposited not far from the coast; depending on the shoreface slope, most likely less than 1 km. The transfer during storm events of small pebbles into a bioturbated environment is an indication of a relatively steep coastal profile. Therefore, the sedimentary environment of the lower bioturbated sand unit is the base of the lower shoreface, which does show many characteristics of inner shelf sands (Clifton, 2005).

The upward increase in dispersed pebble content is explained by the beach environment prograding seawards in a regressive setting.

The transition to the *middle pebbly (MP) beds* is marked by inclined beds with pebble layers especially near the base. They look like gravel lag layers, but as they consist of coarse pebbles, of which some measure over 12 cm, they cannot have been derived by relative concentration of the underlying lower, bioturbated sand, but show a fresh supply of continent-derived thick pebbles.

Very severe storm waves can transport such thick pebbles, but also rip channels can transport pebbles and even boulders offshore (Hart & Plint, 1995). In sandy settings, coarser material accumulates as lags in the base of the rip channels. Elongate and nearly shore-parallel troughs associated with longshore bars have been recognized as broadly lenticular conglomerates with concave upward bases cut in sandstones (Hart & Plint, 1995). Gravel is however not restricted to the troughs, but can also be found on the bar crests.

The admixture with quartz, quartzite, kieseloolite and silicified chalk pebbles is a clear indication of continental denudation going on during the supply and deposition of this marine, near-coastal sand. The appearance is in agreement with a regressive setting coupled to nearby uplift of chalk and Jurassic strata. The uplift may have triggered continental erosion and released a large supply of sediment to the coastal environment, causing normal regression. It cannot be told whether the silicification of the chalk pebbles occurred before or after the transport.

The major sedimentary structure of the MP beds is inclined parallel stratification. The dip is systematically to the north. The sharply delineated, straight laminae and the lack of bioturbation is consistent with high-energy wave stratification such as on upper shoreface bars.

The intercalated hummocky and swaley cross-stratification is another indication of this environment. Hummocky cross-stratification (Harms et al., 1982) is formed under a combination of unidirectional and oscillatory flow that is generated by relatively large storm waves in the ocean. The shallow depressions (swales) between the storm-wave generated hummocks are filled by conformal strata of fine sand, but also mud can be deposited. Pure hummocky and swaley cross-stratification typify the lower shoreface (Saito, 2005), but as they are intercalated in inclined strata, an interpretation of upper shoreface is preferred here. This interpretation is also consistent with the lower content or lack of glauconite pellets, assuming they are carried in from the offshore environment, the presence of an admixture of angular quartz grains, and the presence of mud pebbles, derived from ripped-up mud layers. In this environment, longshore currents occasioned by oblique incidence of waves, create low dunes with foresets dipping in a coast-parallel direction (Clifton, 2005). They produced the many up to 10 cm thick cross beds that are present in the MP beds.

The observed slopes (2 to 7% to the north) of the inclined master bedding indicate a steep upper shoreface. The possible occurrence of slumped sand layers is compatible with steeper slopes.

As a general rule, coastal profiles subject to strong swell (long waves) have steeper slopes (Komar, 1998). As long waves can only form at large fetch conditions, the remarkable steep cross-shore coastal slopes found in some parts of the MP beds strongly suggest an exposed coast at the shore of a wide, open sea. Fair-weather processes would tend to flatten the shoreface and wipe out the high-energy features. The fact that they are preserved may indicate a strongly depositional, prograding coast environment. The progradation is also supported by the northward dipping master bedding.

The vertical succession of inner shelf, lower shoreface and upper shoreface sediments indicates a progradational beach in a wave- or storm-dominated sandy coast. The succession would in the case of complete preservation be followed by foreshore and backshore sediments (Saito, 2005). The swash or wedge-shaped cross-stratification typical of foreshore sands was not encountered in the outcrops available in this study.

Instead, the *upper glauconiferous (UG) beds* are finer-grained than the MP beds. The straight inclined parallel laminae, deposited on a slope dipping up to 9% to the north, showing many intercalated shallow troughs filled with swaley cross-stratification, and sometimes hints of hummocky cross-stratification, continue to indicate an upper shoreface environment of a high-energy open coast. In contrast to the MP beds, no pebbles and only a minor, though upward increasing amount of angular quartz grains are present, and the content of glauconite pellets is much higher. It is therefore thought that this unit might represent a new sedimentation cycle. But no clear separation surface with the middle pebbly sand was found.

The northward progradation seen in the upper glauconiferous sand is again typical for a regressive environment, but this time in a setting with less continental relief: either the continental hinterland was too flattened to produce weathered flint pebbles or the continental hinterland was too far.

The stacking of both deposits without clear bounding surface is interpreted as the result of a base-level rise that occurred during the deposition of the Flemish Hills Sands, after the lower and middle unit had been formed. Both deposits represent what is called in sequence stratigraphy a normal regressive deposit (Catuneanu, 2006). Clearly, after the thicker, lower regressive unit, a base-level rise caused the shoreline to shift southward, after which, in a similar environment to the lower deposit, again normal regression took place, laying down the UG beds.

Apart from the sedimentary structures, the presence of at least the LB and MP beds over a nearly 100 km long, almost rectilinear outcrop area is another indication of an exposed coast at the south shore of a large sea.

Summarizing, the Flemish Hills Sands are a deposit of certainly one, but most probably two regressive successions of a progradational, open marine, high-energy wave-dominated coast. During the deposition of the lowermost thickest series, formed by the LB and MP beds, sediment of clearly continental origin was admixed with shallow marine sand. The continental relief contained exposed chalk. The second series, represented by the UG beds, indicates a landward shift of the coastline to the south in response to a relative base-level rise, followed by a normal regressive deposit in conditions similar to the lower series, but this time without continental detritus.

4. Sedimentary structures in the Kesterberg summit deposits

As they occupy an intermediate position between the Flemish Hills "Diest" Sands and the Hageland Diest Sands (Delvaux, 1884; Cornet, 1904; Leriche, 1914), the correct interpretation of the Kesterberg hill top sands is important in the question of the last marine transgressions in central Belgium. They are mapped as Diest Sands on the late 19th century and recent maps.

The results of the new field observations are summarized in Fig. 3A.

4.1. Top sand deposit

4.1.1. Description

The exposure of the 1989 drinking water storage construction pit, in the south flank near the summit of Kesterberg, showed at the top about 2 m of Quaternary löss and sandy loam with a base gravel of flint pebbles. This gravel is unevenly developed but is on average about 10 cm thick. The pebbles have a weathered rim and some are up to 10 cm long. Below the Quaternary cover, about 4 m of fine, white-to-orange sand were exposed over a lateral extent of some 20 m in two perpendicular faces. The deposit is devoid of carbonates and fossils. The dominant sedimentary structure is complete bioturbation (Plate 2A), though original horizontal bedding can be inferred from colour differences. The sand is fine (100-200 µm), though locally levels can be found containing slightly coarser sand (about 250 µm). Lower in the exposure, the sand is slightly finer and contains some clay. Angular, 250-300 µm grains are present in a minor fraction throughout the exposure. The deposit is free of glauconite in the upper part, while near the bottom part some levels contain a very low glauconite content. Also near the base, in a small roadside outcrop, inclined stratification was found in which occurred one 5 cm thick cross-bed with laminae containing glauconite alternating with laminae devoid of glauconite, with foresets dipping ESE (Plate 2B). This level was also slightly coarser (about 175µm). One horizontal 1 cm thick mud layer in the middle of the deposit consisted of two thin mud layers separated by a thin sand layer (Plate 2A). Overall, in the outcrops no ironstones were found, though loose chunks of platy or massive ironstones are easily picked up in the fields nearby. Also noteworthy is the absence of pebbles in the sand deposit. The exposure shows locally high-angle, crossing faults with vertical displacements of up to 20-30 cm; the blocks separated by the faults are slightly tilted (Plate 2A).

The base of this sand unit is probably very fine (120-150 µm), poorly sorted sand. The lower contact on the underlying clay contains scarce, 1-2 cm thick, unweathered, blackish, rounded flint pebbles. These observations, however, were made in a small outcrop at altitude 96 m for which it was difficult to ascertain the deposits are *in situ*; they are therefore not included in the sedimentary log of Fig. 3A. The underlying grey, bioturbated clay contains some very fine sand.

4.1.2. Interpretation of the sedimentary environment

This is an inner shelf, near-coast marine deposit. The admixture of up to 300 µm angular quartz grains indicates a nearby continental source whose sediment is mixed with fine, marine sand. Sedimentation is quiet and below fair-weather wavebase. The two levels of coarser grains are interpreted as horizons winnowed by storm. The depositional environment was relatively shallow and not too far from the shore. The fault activity was most probably due to recent hill-side instability.

4.2. Underlying glauconiferous sand and clayey sand deposit

The construction in 2010 of a new underground drinking water storage in Onkerzele near Geraardsbergen provided very good outcrops of the second highest geological deposit in the Oudenberg, a hill of similar internal structure and equal elevation as the Kesterberg (Fig. 1 & 3A).

4.2.1. Lower cross-bedded unit

The lower contact of this deposit was not exposed. At the base of the outcrop, a 2 m thick sand unit is arranged in master beds dipping up to 3% SE (Plate 2C). These beds have tangential bases where the bedding slopes grade to horizontal. The beds consist mainly of bioturbated 100 to 150 µm sand with many mud drapes and mud lined burrows; frequently, beds of slightly coarser 225 µm sand occur (Plate 2D-E), arranged in thin cross-beds, 5 to 15 cm thick, with plenty single and double mud drapes (Plate 2E-F) The mud drapes unite at the base of the foreset laminae to form clayey sand bottomsets that slope according to the master bedding. Most of the foreset laminae dip to NE. Dips to the opposite, subordinate direction SW are relatively rare but not exceptional.

The sand consists mostly of rounded quartz grains. The glauconite content varies around 20 to 25%. The dark green glauconite is relatively coarse, in size similar to the quartz grains.

This unit contains "field stone" lumps ("veldsteen"), i.e. silicified sandstones, usually in the sandy cross-beds (Plate 2G).

4.2.2. Middle bioturbated very clayey sand

The upper 4 m in the outcrop were occupied by a bioturbated, very clayey sand (Plate 2H). The clay content is an estimated 20 to 40%. The clay was originally deposited as horizontal, thick mud layers, separated by sand. Due to bioturbation, the original horizontal stratification is barely visible (Plate 2H). The sand is the same as in the lower cross-bedded unit, and contains a similar, relatively high content of dark and fairly coarse glauconite pellets. There are little or no silicified stones.

The lower contact is gradual and interfingering (Plate 2I). There is no coarse sand, gravel or sharp transition. The clayey sand layers interfinger with the lower cross-bedded mildly sloping sand beds.

In some nearby outcrops, the same layer was found. Their altimetric position allowed to establish possibly 10 m thickness for this unit. A weak coarsening upwards trend could also be recognized (Plate 2J).

4.2.3. Upper bioturbated sand

The contact between the previous and the present unit was not found in outcrop. The upper bioturbated sand is a 225 µm quartz sand with an estimated 10 to 20% of coarse dark green glauconite pellets. The sand is homogenized by bioturbation (Plate 2K). The available outcrop was too small to establish a vertical grain size trend. Quartz grains are well rounded.

4.2.4. Geometry of the sedimentary facies

The upper part of the vertical succession described for Geraardsbergen fits remarkably well Rutot's (1882a & 1882c) description of the Kesterberg (Fig. 3B). He mentioned a 20 cm thick sand layer containing coarse grains or granules between his layer "3", the upper "green Panisel Sand" (= upper bioturbated sand), and his layer "5", the "Panisel glauconiferous sandy clay and clayey sand" (= middle bioturbated very clayey sand). This layer with coarse grains was taken by G. Velge (1882) for a gravel separating two formations, but actually is part of the Panisel Sand, according to Rutot (1882c). In the present outcrop descriptions, the coarse grains were not found, but the facies mentioned by Rutot for the Kesterberg were also found near Geraardsbergen. It can be concluded that the description shown in Fig. 3A is accurate and represents the regional stratigraphy of the second geological formation found in Kesterberg and Oudenberg.

4.2.5. Interpretation of the sedimentary environment

This deposit consists of a vertical stacking of sand containing mud laminae, a bioturbated very clayey sand and an upper unit of bioturbated sand. Constant in all levels is the presence of a fair percentage of relatively coarse, dark green glauconite. The *lower cross-bedded unit* is a lateral accretion from NW to SE of a tidal shoal or channel bank, in a subtidal area sheltered from waves or below wavebase. The rounded quartz grains and transported glauconite indicate sand supply from a coastal path or shallow shelf. The *middle bioturbated very clayey sand* may be a quiet-water deposit in abandoned tidal channels. The originally centimetres-thick pure clay layers, separated by fine sand, may have been laid down as flocculated mud, or, as currents were present in the tidal environment, as advected fluid mud. Though here superposed on the cross-bedded sand, it is, in view of the lateral variations in the area and the gradual and interfingering contact in the water storage outcrop, probable that the quiet-water, mud-rich deposits are intercalated in the lateral sand accretion. The *upper bioturbated, relatively coarse sand* consisting of rounded quartz grains may represent a sandy channel fill or shoal in the same environment as the lower units. The overall coarsening upwards trend for the middle to upper unit may indicate a regressive deposit.

4.3. Underlying fine, parallel stratified sand

Underneath the glauconiferous sands and clayey sands, very fine sand with low-angle parallel stratified beds or hummocky cross bedding is found in outcrops in Kester village (Fig. 3A). This sand deposit contains a minor fraction of fine glauconite pellets and has a powdery feel because it contains no clay. Occasional cm-thick clay layers and clay lumps do occur. The deposit is unbioturbated. In the area, the deposit is decalcified and contains no fossils.

This deposit is interpreted as the lower shoreface of an open coast.

4.4. Comparison with previous observations and stratigraphic position of the top deposit

The uncertainty about the correct stratigraphic position of the Kesterberg top deposit is far from new. It existed clearly in an 1882 dispute between M. Velge and A. Rutot (Rutot, 1882a & 1882c). More recent reconnaissance work has up till now not settled the question. A simple comparison of different wells drilled on the summit of Kesterberg (DOV wells n° B/2-0039, B/2-0040, 101W-B49, B79 and B108) shows completely different interpretations of layers occurring at the same level. In spite of this, the geological maps assign the top deposit to the Diest Formation.

A west-east profile through Kesterberg (Fig. 5B & 11B), linking the Pottelberg section to the plateaus south of Brussels, shows that the Diest or Flemish Hills Formation on the summit of Kesterberg would imply a significantly lower base than at Pottelberg. Yet, the underlying layers have conformal, nearly horizontal boundaries. Also north-south profiles show a regular, conformal pattern of the underlying strata (Fig. 5C & 6).

The lowest unit described here belongs, based on its stratigraphic position on top of the thick clay of the Kortrijk Formation and sedimentary structures, to the very-fine sand layer, formerly labelled "Yprésien d", which constitutes, in a wide area south of Brussels, an areally continuous, well recognizable stratigraphic level. The sedimentary structures are in agreement with the shoreface facies in outcrops at Egem published by Steurbaut (2006). The grain size correlates with the eastern facies of the Tielt Formation, according to Geets (2001). Biostratigraphically, this layer, named "Mons-en-Pévèle

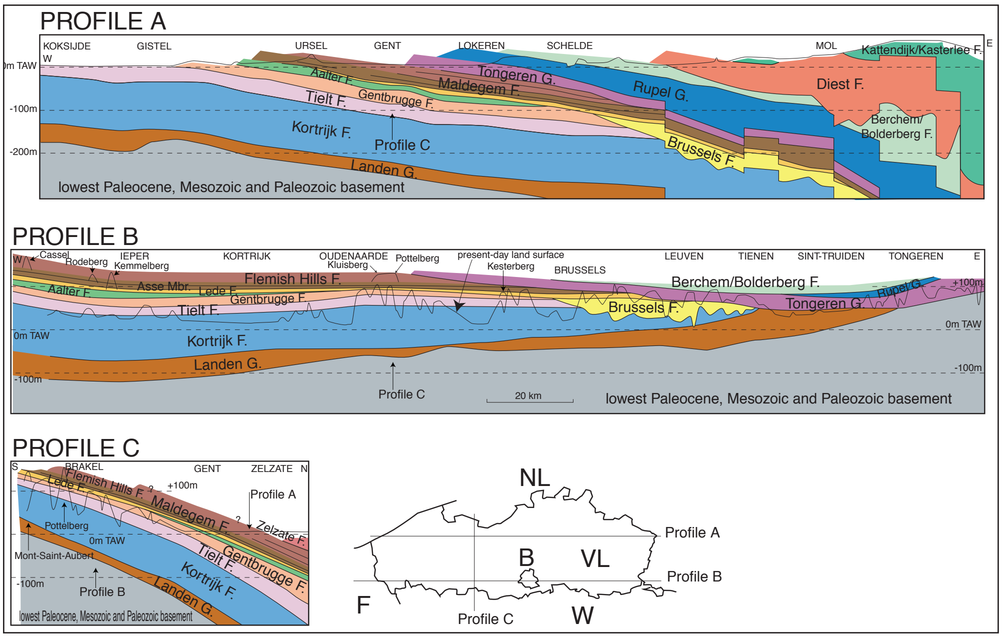

Figure 5. Schematic geologic cross profiles through northern and central Belgium. Profile A, B and C are based respectively on profile 6, 7 and 2 published on dov.vlaanderen.be (Mattijs et al., 2003). Vertical exaggeration 100x. Geologic layers are shown on the geological formation (F.) or group (G.) level. Only in the Maldegem Formation, an indication of the several members is shown using fine lines and two colour shades of brown. The layers have been extrapolated above the present-day ground surface. Data described in the text on the different Flemish Hills have been used to supplement the profiles. In profile B, the Cassel hill section has been shifted down by 5m, Rodeberg by 14m, Kemmelberg by 17m, Pottelberg by 24m and Kesterberg by 21m, according to the distance of each hill south of the profile line and assuming a uniform tectonic dip north of 3.5 m per km. In profile C, the Mont-Saint-Aubert and Pottelberg sections have been added without vertical shift.

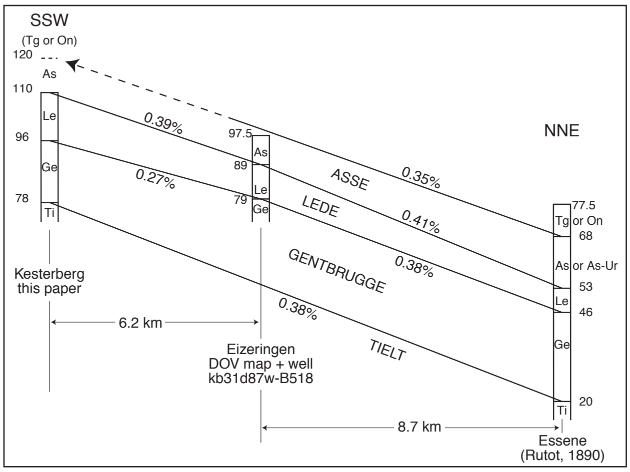

Figure 6. Strata boundary tectonic dips in a section between Essene and Kesterberg (location, see Fig. 1). Formation or member names from bottom to top: Ti = Tielt Formation; Ge = Gentbrugge Formation; Le = Lede Formation; As = Asse Member; Ur = Ursel Member; On = Onderdale Member; Tg = Tongeren Group.

Sand" by Steurbaut & Nolf (1986), was shown by these authors, based on calcareous nannoplankton microfossils, to correlate with the Roubaix (also named Moen) Member of the Kortrijk Formation. Lithologically, the unit should be interpreted as the Egem Member of the Tielt Formation. While the question of the local lithostratigraphy has not been settled, an assignment to either Tielt Formation/Egem Member or Kortrijk Formation/ Moen Member is certain.

The contact between the lowest deposit considered here, and the glauconiferous sand and clayey sand unit was not seen in outcrop. The cored well 101W-B79 description (Geets, 2001) contains no details on the contact and does not mention a gravel layer.

The glauconiferous sand and clayey sand unit at Geraardsbergen is characterized by coarse glauconite. The unit was previously known as part of the Lower Panisel Formation (Gulinck & Hacquaert, 1954), Anderlecht Clayey Sands (Kaasschieter, 1961) or Panisel Sand Member (Steurbaut & Nolf, 1986). Lithologically, it is now commonly mapped as part of the Gentbrugge Formation. In general, the succession of the different units in the Gentbrugge and also the underlying Tielt and overlying Brussels Formations is complicated by lateral facies variations and, more particularly, by several pulses of erosion into underlying sediment (Vandenberghe et al., 2004). In the area between Zenne and Dender, detailed mapping has often difficulties to bring home the different clay strata in boreholes (De Ceukelaire, 2009). The typical bottom to top succession inside the Gentbrugge Formation of Merelbeke Clay, Pittem Sandy Clay and Vlierzele Sand Members, defined near Gent (Marechal & Laga, 1988; Steurbaut, 2006), is not valid in the area between Zenne and Dender (Kaasschieter, 1961; Geets, 1969). The deposits in that area are in general more clayey (Kaasschieter, 1961) than the Gentbrugge Formation in its type area with which they are lithologically correlated (Marechal & Laga, 1988; Steurbaut, 2006), in spite of the fact that, biostratigraphically, they should be correlated with the Tielt Formation (Steurbaut & Nolf, 1986). Also, the vertical succession of clayey and sandy units seems to show more variation in this area. Rutot (1890) observed that in the area NW of Brussels, between Jette and Essene, the "paniselien" has a continuous, 1 m thick grey clay layer at the top.

The new observations of sedimentary structures and environments found for the lower, cross-bedded unit at Geraardsbergen Onkerzelestraat, are typical for the Vlierzele Member and its lateral facies (Houthuys, 1990). Interestingly, the outcrop proves the glauconite-rich sandy and clayey superposed units grade into each other and are thus lateral facies. At the same elevation, only 14 km to the east, the vertical succession inside the Gentbrugge Formation (Geets, 2001) is different (Fig. 3A & D). Though no continuous outcrops linking both sites exist, it is in line with the fact that the facies arrangement inside the Gentbrugge Formation is variable in the area, with clayey layers occurring either at the bottom, in the middle, or near the top of the formation; but the different units are definitely lateral facies inside a single depositional very shallow, subtidal and probably inshore environment. They thus belong to one formation, with a local thickness of at least 18 m; this latter observation is in agreement with Rutot (1882c).

De Ceukelaire (2009) considered the thin clay package in the Kesterberg well 101W-B79 at the altitude 94 to 95 m, just below the Lede Sands in Geets's (2001) interpretation, as the Merelbeke Member (compare Fig. 3D with 3E), with the other members of the Gentbrugge Formation absent at this location. De Ceukelaire (2009) revised the local lithostratigraphic map, based on lithological descriptions and resistivity logs and inspired by the time correlations shown by Steurbaut & Nolf (1986). The result of this interpretation is raising up the lithological levels. In order to keep the Diest Sand appellation for the Kesterberg top sand, which De Ceukelaire (2009) does, an even more important hiatus or squeezing of layers with respect to the Pottelberg section must be assumed than in Geets's (2001) interpretation.

The lithological and sedimentological observations in Geraardsbergen presented in this paper, however, demonstrate that the glauconiferous sand and clayey sand deposit occurring between the levels of 78 and 96 m TAW in the outlier hills from Oudenberg to Kesterberg, belong to one depositional level, with gradual and laterally dipping contacts between sand and sandy clay layers. This is in agreement with Rutot's (1882c) interpretation of Kesterberg. His profile, which only shows the top of the "paniselien" labelled "3" to "5", has been reproduced in Fig. 3B.

For the present study, the implication is that the clay layer around altitude 95 m, interpreted in the DOV file of Kesterberg well 101W-B79 as Gentbrugge Formation (Fig. 3D) and mapped accordingly on the geological map sheet Brussels (Buffel & Matthijs, 2009) (Fig. 3C), is almost certainly only the top part of the Gentbrugge Formation.

Going upwards, Rutot (1882c) assigns the fine sand labelled "2" to the Wemmel Sands (Fig. 3B). To explain the presence of "red coarse sand with pebble layers and ironstone", labelled "1", in the south flank of the hill, he hypothesizes a ravinement. He interprets this sand as the local equivalent of the sands that crown in considerable thicknesses the Cassel and Ronse hills. It is clear that here is the germ of the Diest Sand interpretation for the upper sand deposit. But Rutot (1882c) had no exposures of the contact between the "red sand" and the Wemmel or Panisel Sand. His sketch of the "red sand" ravinement in the layers "2", "3" and "4" (Fig. 3B) is an unconfirmed interpretation of a series of small outcrops and wells. What the sketch does show, is that ironstone concretions may be found at different levels and probably in different sandy deposits in the summit of Kesterberg Hill. Rutot (1882c) also observed that no glauconiferous Asse Clay or "sable chamois" (a sand layer found in Laken, now considered a local representation of the Bolderberg Formation) was ever found on the Kesterberg.

The geological map (Fig. 3C) and De Ceukelaire (2009) (Fig. 3E) show another clay layer, around altitude 98 m, interpreted as Asse Clay, of which Rutot (1882c) and cored well B79 find no trace. I did not find it either in outcrop. I therefore agree with Rutot (1882c) that no Asse Clay is present at Kesterberg. My interpretation is shown in Fig. 3F. It implies that, on grounds of general basin architecture (see N-S profile in Fig. 5C & 6 and W-E profile in Fig. 5B), the top deposit at Kesterberg must be the Middle-Eocene Lede Sands.

The field observations are perfectly compatible with this Lede Sand interpretation: overall completely bioturbated fine sand, a thickness in agreement with regional values, no pronounced vertical grain-size gradient, low glauconite content, coarser levels with lamination near the base (mentioned by Rutot, 1882c; general occurrence, see Vandenberghe et al., 1998), often multiple erosional intercalations with coarser quartz grains that in undecalcified outcrops also contain shell concentrations and coincide with the Balegem stone layers (Fobe, 1986). Other outcrops not too far (e.g. near Lion's Mound at Waterloo, 2013 underground museum construction site) show similar sedimentary characteristics for the Lede Sands.

The flint pebbles at the base of the Quaternary sandy löss can only be interpreted as erosion lags of younger layers, once present on top of the Kesterberg Lede Sands. Also this observation reinforces the Lede Sands interpretation. Several Middle-Eocene to Oligocene layers are known to cover the Lede Sands in the region.

In conclusion, the top 14 m of Kesterberg are reinterpreted to be Lede Sands. The same interpretation is applied laterally to the top sands in the hill row, going west to Oudenberg. Similarly, the summit of the Mont-Saint-Aubert will probably turn out to be Lede Sands. At the time of writing, no good outcrops are available to verify this.

5. Sedimentary structures and facies variations of the Hageland Diest Sands

The Upper-Miocene Diest Sands in their type area between Diest and Brussels are very glauconiferous fine to medium sands (Tavernier, 1954; Tavernier & de Heinzelin, 1962; Matthijs, 1999; Vandenberghe & Gullentops, 2001). Very frequently, thin and thick layers or irregular volumes inside the Diest Sands are limonite cemented. The glauconite content is around half of the volume (Vandenberghe et al., 2014). The grain size varies from 125 to over 300 µm. Fine sand between 125 and 250 µm and consisting of well-rounded quartz grains dominates, but a subpopulation of coarse, 0.5 to 2 mm, often angular quartz grains may be present. Apart from the base gravel, the sand contains no pebbles, except probably very locally near the base, such as at Kesselberg, where two rounded-flint pebble layers can be found inserted in the formation a few metres above the local base. Vandenberghe et al. (2014) suggest the Diest Formation such as it is mapped now in the Kempen area, may consist of the deposits of two consecutive sedimentation cycles. The present observations are restricted to the Hageland part of the Diest Sands.

5.1. Occurrence and bounding surfaces

The Diest Formation (Fig. 1) occurs in NE Belgium. The outcrop area is about 45 by 70 km. The often 50 to 100 m thick deposit has a tectonic dip of 0.3% (in the Hageland) to 0.6% (in the Kempen) to NNE, in line with the deposits underneath, the Berchem/Bolderberg Formation (DOV profiles). It is covered by uppermost Miocene Kasterlee Sands in the north and their lateral equivalent, the Lower-Pliocene Kattendijk Sands, in the NW (chronostratigraphy according to Louwye & De Schepper, 2010). The formation sits deeper in the subsurface in the Roer Valley Graben (Vandenberghe et al., 2014). The Roer Valley graben is a complex fault-bounded subsidence area, part of the Lower Rhine Embayment, strongly reactivated since the Late Oligocene (Geluk et al., 1994). The graben in NE Belgium and the adjoining areas of The Netherlands and Germany was during the Miocene most of the time a marine bight of the southern North Sea, with coastal and coastal swamp areas at its SE tip (Verbeek et al., 2002; Westerhoff et al., 2008). The Diest Sands or time-equivalent deposits occur in great thickness in the graben, showing that the area was actively subsiding during the Diest Sands deposition (Demyttenaere, 1988).

The base of the Diest Formation is erosional. While several parallel valley-like incisions occur at the base in the Kempen, the Diest Sands in their type area, Hageland, fill a conspicuous, 15 to 20 km wide and at least 80 km long basin connected in the NE to the Roer Valley Graben (Vandenberghe et al., 2014, fig. 2). The erosional base there cuts through tens of metres, locally over 100 m, of older Cenozoic sedimentary layers (Vandenberghe et al., 2014). The basal surface is not one trough but shows several parallel, elongate depressions, hundreds of metres or a few kilometres wide and probably over 10 km long. The long axis of the depressions is parallel to the long axis of the Hageland basin. Some of the elongate depressions seem to be connected like channels in a braided river or channels separating shoals in an estuary (Vandenberghe et al., 2014, fig. 2). Some depressions are closed like troughs (Houbolt, 1982, fig. 7, based on data compiled in the 1960s by Van Calster). The irregular base gravel is thicker in the bottom part of base troughs. It is composed of coarse elements, often flint pebbles, derived from locally eroded, older sediments (Gullentops, 1988; Matthijs, 1999; Vandenberghe & Gullentops, 2001).

As the Hageland Diest Sands now occupy an elevated position in isolated hilltops and a plateau, from about 50 m TAW in the NW to about 100 m in the SE, little can be told with certainty about their original extent. It is possible that only the lower part has been preserved, as the top of the deposit is truncated. It is in some outcrops covered by a well-developed gravel of rounded, weathered, several centimetres thick flint pebbles underneath a Quaternary löss or sandy loam (Plate 3G). They may be reworked form disappeared upper Diest deposits, but as the top of the Diest deposits makes out a fairly regular surface, this surface may have been shaped by marine erosion and the pebbles may thus have been concentrated from younger marine or continental deposits. The current outliers near Brussels and Leuven were certainly connected to the main body of Diest Sands.

The relationship with the Deurne and Dessel Sand Members and the Diest Sand Member occurring in the Kempen is discussed by Vandenberghe et al. (2014). They may partly constitute a lateral equivalent and partly contain a younger depositional cycle. The question is not finally settled yet. The lower cycle to which the Hageland Diest Sands belong, have a Tortonian age (Vandenberghe et al., 2014).

As to the western extent of the Hageland Diest Sands, there is no doubt that they are found at Laken and Wemmel as the locally youngest deposit on a hill top. Other small outliers more to the west on the geological map are uncertain. The relation to the Flemish Hills Sands is the object of the present paper.

5.2. Sedimentary facies in outcrops

All outcrops show one or more of the following sedimentary facies:
- very thick cross beds (XT)
- descending cross beds (XD)
- bioturbated descending beds (Bx)
- bioturbated fine sand (Bf) or medium sand (Bm)
- massive to vaguely laminated, unbioturbated sand (M)

It is remarked that the inventory of facies is not complete. More to the north and east, fine-grained facies occur especially below, but locally also intercalated and on top of the typical Diest Sands (e.g. Matthijs, 1999). Also, in the area studied, more facies may turn up when new outcrops become available.

5.2.1. Very thick cross-beds (XT)

This facies consists of sets of cross-strata dipping at an angle of about 25 to 30° ENE to NE. Out of 55 measurements of foreset direction carried out over the entire Hageland by Van Calster in the 1960s (map published by Houbolt, 1982), 54 give an average direction of 41° (NE) with an even spread across the range 15° to 60°; one measurement near Linden was 302°. The measurements were probably carried out across facies XT and XD (see below).

The XT sets have probably a trough-shaped base and are bounded at their top by an erosive surface. Due to the scale of the sets, the complete geometry of a single set is mostly not exposed in one outcrop. The largest outcrops available indicate that the flow-transverse geometry may be wedge-like rather than symmetrically trough-shaped, with the thickest part either in the left or right part of the wedge near its lateral margin; the top truncation surface dips, so that the thickness of the sets decreases laterally. The maximum observed thickness is up to about 6 m and the maximum width tens of metres. In flow-parallel sections, the length is tens to probably a few hundreds of metres, and the set boundaries are either horizontal or show a slight downcurrent dip.

The foreset lamination is mostly normal-graded (Plate 3A-B), but groups of foreset laminae displaying avalanching lenses may also be found. The downward dip terminations of the long cross-strata are tangential to concave. Bottomsets are well developed and, especially at their base, homogenized by bioturbation.

Mud drapes of a few millimetres thickness are mostly found near the transition of foresets to bottomsets, at irregular distances. Sometimes, several mud drapes separated by a very thin sand layer, unite downdip to form a mud layer of a few centimetres thick.

The thick cross-beds show sometimes convex-upward internal unidirectional reactivation surfaces. They are the result of smaller-scale dunes migrating on the top of the large bedform, which produced the very thick cross beds; on reaching the large bedform's depositional brink, the superposed bedform travels down the depositional lee face and spills all sand present in the superposed bedform to merge with the large bedform's cross-bed. V-shaped animal escape traces may be associated with the unidirectional reactivation surfaces. They are indicators of rapid sedimentation. Only once, a sedimentation pause plane was found, covered by lamination of small ripples climbing up the plane; this ripple lamination had a fine mud drape cover. Bioturbation is rare and occurs in association with surfaces bounding packages of minor foreset-dip change (Plate 3A-B). The trace fossils belong to the *Skolithos* ichnofacies (Pemberton et al., 1992) and affect only a few centimetres underneath the dip change surfaces. These surfaces thus constituted, presumably for a short time, sedimentation standstill surfaces from where burrowing could take place.

Over 10 m thick stacks of very thick cross-beds without intercalation of beds displaying one of the other sedimentary facies may be found, but good outcrops to observe this are rare. Nevertheless, one or two superposed sets of very thick cross-beds do occur in many outcrops.

The median grain size is about 200 to 250 µm, but always a significant subpopulation of 125 to 150 µm grains is found. The bioturbated bottomsets consist mostly of such very fine sand. Sometimes, especially in thick cross-beds near the formation's base, a subpopulation of coarse 0.3 to 1 (sometimes 2) mm angular quartz grains is present.

Interpretation: the very thick cross-beds are the deposit of either very large subaqueous dunes or transverse bars (in the case of symmetrically trough-shaped or tabular very thick cross-beds), or large lateral bars (in the case of wedge-shaped very thick cross-beds) that migrated in channels ENE to NE-wards. Transverse bars and lateral bars are bedforms usually associated with river channels. They are downcurrent-accreting, large, up to a few metres high, sediment bodies. Transverse bars or cross-channel bars occupy the centre of the river channel; lateral or side bars are attached to the sides of the channel, usually alternately at the left and right side of slightly sinuous channels, in which case they are called alternate bars (Bridge & Demicco, 2008). Smaller-scale bedforms such as dunes, ripples, and bed-load sheets are commonly superimposed on alternate bars. Their accretional downstream side is steeper than their upstream flank. The accumulative downstream side may either show a relatively steep avalanche face, or may have a milder slope, over which smaller bedforms migrate (Bridge & Demicco, 2008). The very thick cross-beds of the Diest Sands facies XT must be associated with a steep slip face.

The configuration of the Diest Sand Formation (see below), along with the glauconite content and ichnofossil traces preclude a fluviatile channel environment. The use of "transverse bar" and "lateral bar" in the present context is informal. These bedforms, that are commonly known from fluviatile channels, merely fit best the bedforms inferred to have produced the very thick cross-beds of this Diest Sand facies.

The normal-graded cross strata in long, straight to tangential laminae show that a flow separation vortex existed at the lee side of the large bedform that produced the cross-beds. A lee vortex of the size represented by the very thick cross-beds can only be formed by a very strong current in relatively deep water. Applying the simple rule provided by Allen (1966), water depth for the thickest cross beds of 6 m would have been between 30 and 60 m. The two-step rule cited by Dalrymple and Choi (2007) would result in estimates of water depth between 40 and an (unrealistic) 400 m. The incised channel depths allow to derive a water depth of at least 50 m, probably more. Also the scarce bioturbation proves sedimentation was fast; but nevertheless, the frequent mud drapes and burrowed sedimentation standstill surfaces indicate the current was unsteady. Most likely, variations in sediment supply at the depositional brink of the large dunes or bars caused these temporary standstill surfaces. These and the unidirectional reactivation surfaces indicate smaller dunes were present on the stoss side of the larger bedforms. The overwhelming majority of foresets have the same dip direction, so the current was unidirectional. No clear tidal signature is found in the cross-beds. Mud drapes are typical but not exclusive for tidal environments. Yet, a strongly asymmetric tidal current with a dominance of the NE flowing stage, seems likely: tides were present in the nearby Lower Rhine Embayment, such as shown by the Upper-Miocene shallow near-coastal marine deposits there (Schäfer et al., 2005); tidal currents are unsteady; the rare small-ripple lamination on sedimentation pause planes may be the product of subordinate tidal-current flow, especially as the structure was mud draped. Van Calster's foreset dip measurements (in Houbolt, 1982) report one outcrop showing foresets dipping WNW, a rare proof of the opposed tidal current.

Metres-high unidirectional dunes can be found in modern tidal channels where one tidal current dominates, but also narrow sea straits can have large dunes. In the case of the Golden Gate Strait at San Francisco, very large dunes, up to 10 m high, with very long slip faces, are associated with powerful tidal currents of up to 2.5 m/s (Barnard et al., 2012). However, these very big dunes are found just outside the strait and the erosion maximum, and are likely to form downcurrent climbing sets. In the Diest Sands case, the depositional setting and the sets' wedge-shaped geometry, with downcurrent dipping set boundaries, rather indicate the large bedforms producing the very thick cross-beds were something resembling transverse and lateral bars, but in deep submarine channels. No good analogues of recent environments with bedforms producing this type of very thick cross-beds are available. Better outcrops will probably one time allow to reconstruct the somewhat mysterious bedform that deposited this Diest Sands facies.

5.2.2. Descending cross beds (XD)

This facies consists of cross beds with foresets dipping ENE to NE varying in thickness from 0.2 to about 1 m, arranged in vertical stacks of downcurrent dipping sets (Plate 3E). The sets are tabular in the outcrops. Their tops are bounded by straight, erosive surfaces. At the base, there is always a thin (order of 5 cm) bottomset. The cross-beds descend in the downcurrent direction with a slope of a few degrees, with a maximum observed slope of about 6°.

The foreset lamination is like in facies XT. The downward dip terminations of the long cross-strata are tangential. Also, bioturbation characteristics are similar to those of XT. Millimetre-thick to amalgamated, centimetre-thick mud drapes occur.

The cross-beds show no internal truncating unidirectional reactivation surfaces, other than the parallel bed bounding surfaces and minor foreset dip-change surfaces.

Stacks of descending cross-beds may be metres thick.

Interpretation: the descending cross-beds are the deposit of a succession of large hydraulic dunes descending a downcurrent sloping surface and migrating ENE to NE. A suite of large dunes migrating downslope produced the vertical stacks of descending cross-beds. The paleocurrent conditions that moved the large dunes were similar to these of the very thick cross-beds. The smaller dune heights may also have occurred in less deep water.

5.2.3. Bioturbated descending beds (Bx)

This facies consists of metres-thick stacks of 10-20 cm thick parallel layers of fine, bioturbated sand with a clayey base (Plate 3C-D). The bioturbated beds have a finer-grained basal part and a fine-to-medium sand upper part. The beds are laterally persistent over tens of metres. Together, they build up sets of several metres thickness, with internally parallel, bioturbated descending beds. The beds are arranged on a slope that dips up to 10°, though usually the dip is much less. The slope surface has a downcurrent and a lateral component, normal to the current direction. Some outcrops seem to indicate the beds may be arranged in flow-transverse troughs. Due to the large size of the descending bed sets, the true shape and dip is difficult to observe in outcrops, but it is probably often also somewhat oblique with respect to the paleocurrent direction. Both packages attached to a right channel flank, dipping somewhat to N or NNW, and to a left flank, dipping somewhat to SE, have been observed. Towards their base in a flow-parallel section, the beds are concave, much like foresets in typical cross-bedding. Indeed, from a distance, a package of bioturbated descending beds may resemble a very thick trough-shaped cross-bed; however, the slope angle of the beds is much milder than the typical cross-bed foreset slope. Several sets of bioturbated descending beds can be stacked vertically, with an erosive unconformity surface topping each set (Plate 3C).

The bioturbation is of the *Cruziana* ichnofacies (Pemberton et al., 1992). Sometimes, it can be seen that the sandy part of the bioturbated beds was originally cross-bedded. The cross-bedding, though strongly disrupted by the bioturbation, consisted of foreset laminae dipping ENE to NE.

Interpretation: the bioturbated descending beds are the result of lateral and downcurrent accretion on a sloping surface. The accretion occurred by the deposits of small dunes, migrating ENE to NE over a surface that sloped transversely and downcurrent to the dunes' migration direction. As the main current that moved the dunes was ENE to NE, packages with a bed dip component to north alternating with packages with a bed dip component to SE can be considered to be the deposit of large lateral bars attached to opposing channel flanks with the bars having milder accretional downcurrent slopes than those that produced facies XT. The muddy sand layer at the base of an individual descending bed inside a lateral bar deposit is the amalgamation of mud drapes inside the dune's foresets. The migration was slow enough to allow nearly complete homogenisation of the sandy deposit by endobenthos. The bioturbated lateral bars filled relatively low-current spaces in the same channels where very large and large dunes and lateral bars, associated with facies XT and XD, are the bedforms of the highest currents. This is clear from the fact that the three facies occur intermingled (Plate 3C).

5.2.4. Bioturbated fine sand (Bf) and medium sand (Bm)

This facies consists of completely bioturbated fine sand or medium (Plate 3F-G) sand. In the latter case, a significant fine fraction is present. There is also some mud, probably from disrupted mud drapes. The homogenisation due to bioturbation is complete. Facies Bf was only found in Laken, over a vertical exposure of 0.7 m near the local top of the Diest Formation, while facies Bm was only found at Gasthuisberg, near the top of the formation but at lower levels as well, over several metres of vertical exposure.

Interpretation: this is a mixed deposit of suspended very fine sand and mud and probably only occasionally active, low dunes. Little can be said about the extent of the beds, as large outcrops displaying this facies have not been found yet.

5.2.5. Massive to vaguely laminated, unbioturbated sand (M)

This facies was so far only found at two sites. At Nieuwrode, it fills a ca. 0.5 m deep, irregular erosional cut in the top of a thick cross-bed (Plate 3H). The sand fill, with sand and glauconite grains comparable to the sand in which the erosion is cut, is massive or vaguely laminated, with vague lenticular shapes in cross profile, and unbioturbated. At Kesselberg, two conformal beds, each ca. 0.5 m thick, of very glauconiferous, mixed and unbioturbated sand may also belong to this facies M. The outcrop shows a structureless mass of 2 to 3 cm wide, round-flattened, white sand-rimmed ghost structures, in otherwise very glauconiferous sand. The round shapes may be shell-ghosts or may represent water-escape structures of ripped-up, originally flat quartz sand laminae, such as can by produced by water forced out of quickly buried sediment. Each bed terminates towards the top in a ca. 0.1 m thick layer of slightly finer sand covered with a mud drape.

Interpretation: the shape of the erosive base and the characteristics of the fill at Nieuwrode make this facies a good candidate for a breaching deposit, near the breach or in the breach evacuation channel that may be present downslope of a breach if there is sufficient underwater relief (van den Berg et al., 2002; Houthuys, 2011). The interpretation needs confirmation from the 3D structure of the facies; however, the outcrop at Nieuwrode is no longer available. The beds at Kesselberg may be the depositional lower, conformal deposits near the channel base of a breaching event, containing either ghosts indicating a concentrated shell bed, or water escape structures.

5.3. Sedimentary architecture

The available outcrops allow only limited observations of how the different facies are related in space. XT, XD and Bx make up the bulk of the Hageland Diest Sand exposures. They are intimately mixed. In a single outcrop, a vertical superposition and alternation of XT, XD and Bx can be observed (Plate 3C). Also in the horizontal direction, the facies succeed each other.

Facies M was observed as the fill of an erosional cut in XT at Nieuwrode and as a conformal intercalation between two sets of Bx at Kesselberg.

Vertically, XT may predominate at the base of the Diest Formation, but it occurs at all levels. No systematic vertical succession of the different facies has been observed. Outcrops of Bf and Bm were found near the top but Bm also at lower levels. They were in the vicinity of XD and XT; unfortunately, the contacts were not well exposed. The fine, clayey sands at the base and sometimes the middle or top of the Diest Sands more to the east (Matthijs, 1999) may be equivalent facies.

In the longitudinal direction, parallel to the paleocurrents flowing ENE and NE, bed boundaries are subhorizontal, or slightly to outspokenly downcurrent descending. A progradational downcurrent fill is evident. In this progradational fill, facies XT, XD, Bx and probably also Bm alternate in a downlap fashion.

In the transverse direction, outcrops are too scarce, insufficiently large, and often too obliterated to allow clear observations. XT beds appear to be trough-shaped but even more often wedge-like. Both wedges tapering out to N and to SE have been observed. Also XD beds appear to be arranged in sets with surfaces dipping either to NNW or to SE. Bx beds have a similar transverse dip component, either to NNW or to SE.

The arrangement is not well understood, as alternating left and right dipping master beds seem to suggest alternate bars or point bar lateral accretion deposits of sinuous flow channels, while the range of foreset dips in all cross-bedded facies occupies a narrow sector of flow directions, suggesting rather straight flow channels.

Taking everything together, most of the Hageland Diest Sands appear to be stacked channel fills, with probably very large dune or large transverse bar sets near the base grading upwards to lateral or alternate bars, attached in an alternating fashion to the left and right flank of relatively straight, tens of metres deep, submarine channels.

The ultimate answer to the sedimentary build-up of the Hageland Diest Sands necessarily awaits new outcrops and observations.

5.4. Inferences about the sedimentary environment and the paleogeography

The Upper-Miocene Hageland Diest Sands testify at the same time of marine conditions and of confined, strong, unsteady unidirectional currents directed to the Roer Valley Graben, part of the then marine bight of the Lower Rhine Embayment.

The water depth during deposition of the Diest Sands was considerable, at least 50 m, probably more. Such a water depth is compatible with the size of the channel incisions. However, relatively thick cross-beds occur also at the top of the formation. They indicate, for the top of the preserved deposits, a water depth of at least 15 to 30 m, which implies the upper surface of the Diest Sands is a truncation.

The facies building up the Hageland Diest Sands are clearly not the product of offshore tidal current ridges, such as suggested by Gullentops (1957, 1988). Tidal current ridges are found on relatively flat erosional surfaces, consist of intertwined packages of ebb and flood oriented cross-beds, contain numerous tidal and unidirectional reactivation surfaces, show probably a lateral internal architecture of mildly sloping master bedding planes and a vertical transition of heterolithic, bioturbated facies at the base to well-sorted sand with primary current structures upward, and are expected to show wave sedimentary structures, especially near their top (Houthuys, 1990; Houthuys et al., 1994; Trentesaux et al., 1994). Present-day tidal current ridges in the North Sea often possess a core of older sediments, remnants of lowstand environments (Berné et al., 1994; Mathys, 2009). None of these features is encountered in the Hageland Diest Sands.

As the axial directions of the basal incisions match the current direction shown by the cross-beds, the strong currents responsible for the channel erosion were also the transport agent that deposited the Diest Sands (Vandenberghe & Gullentops, 2001). This is supported by the long foreset laminae that imply a large-scale lee vortex and thus very strong currents, while the architecture of downcurrent sloping internal boundary surfaces of the thick, descending and bioturbated cross-beds imply a forward moving focus of erosive, strong currents followed immediately by a wake of sedimentation. The forward-moving focus of strong currents scoured linear, parallel, elongate troughs, filled with the Diest Sands.

But which mechanism drove the vertical incision? Taking the deposit's dimensions into account, several candidate environments can be thought of:

- local tectonic subsidence creating a fault bounded narrow graben is not supported by the regional tectonic framework;

- a sea strait connecting two marine basins may be characterised by strong unidirectional currents. Gullentops proposed such a setting (1988; Vandenberghe & Gullentops, 2001) where a connection was conjectured between the English Channel and the southern North Sea approximately at the present-day Dover Strait location. The origin of the sediment filling up the Hageland Diest incised valley was thought to be the English Channel. The sediment load would have been deposited, after transiting the straits, at the eastern exit area, where expanding flow lines caused a decrease in transport power. This explanation is problematic. First, a connection with the English Channel is hypothetical: there is no deposit left to prove such a marine link. The Lenham Beds in south-east England are no longer considered to have a proven correlation with the Upper-Miocene Diest Sands (Wood et al., 2000), while the Noires-Mottes (Fig. 1) iron cemented top sands most probably correlate with the Flemish Hills Sands, that are in this paper interpreted as a completely different depositional environment, unrelated to the Diest Formation (see below). It is hard to envisage a sea strait would not have left a strong incision with at least at the base coarse lag sediments. Second, the Hageland Diest Sands are clearly high-current deposits related to erosional channels. An expanding flow area would not be the environment where the current maximum and erosional channels are found. Third, the depositional area in Hageland is relatively far east from the claimed sea strait. Fourth, the waning flow should be accompanied with a general fining-upward and fining-basinward trend. This is not observed. Also, if this deposit was at the mouth of a narrow strait with strong eroding currents, the produced sets would have downcurrent climbing boundaries, while the opposite is the case for the Diest Sands;

- an incised valley system is supported by the configuration of the formation base showing the arrangement of a consequent river system draining to the Roer Valley Graben (Vandenberghe et al., 2014). Such an accommodation space filled by fluvial sediments is clearly no valid option. Incised-valley fills show very diverse facies, different from the observed facies of the Diest Sands, with a high or exclusively continental origin of the fill (Zaitlin et al., 1994; Dalrymple & Choi, 2007);

- an incised valley system, drowned during rising base level and transformed into a tidal delta. A tidal delta is the sand bar or shoaling area left at the mouth of a river by the movement of bottom mud and sand by diurnal tidal currents. However, no important river that may have had its estuary here is known and no deposits of distributaries that should have fed the delta, are found. Also the sedimentary architecture that arises from the outcrops is not compatible with the facies associations and internal buildup of facies in a tidal delta environment (Dalrymple & Choi, 2007; Legler et al., 2013);

- an incised valley system, drowned during rising base levels and transformed into a high-energy estuary. Estuaries (Dalrymple & Choi, 2007; Martinius & van den Berg, 2011) attract marine matter during rising base level. Essentially, estuary fills consist of mutually evasive ebb and flood channels separated by shoals. Though many sedimentary characteristics of the Hageland Diest Sands may be compatible with high-current tidal channel deposits, diagnostic other facies such as tidal shoals, tidal point bars, fine-grained abandoned channel fills and intertidal and supratidal plain and marshland, along with the many different-order bounding and erosion surfaces, are lacking (Chaumillon et al., 2010). There is even no unequivocally tidal signature in the deposits; there is no overall fining-upward trend typical of estuaries; and the very narrow range of ENE and NE flow directions documented by the foreset dips is in contradiction with the often high sinuosity of estuarine channels.

In the several tens of metres' thick stacks of Diest Sand cross-bedded deposits, all bed boundaries are subhorizontal to descending in the downcurrent direction. This can only be the result of an erosive focus area proceeding downstream and followed in its upstream "wake" by strong deposition. The process required to produce this kind of deposits is thought to be similar to the mechanism described for the Eocene Brussels Sands (Houthuys, 2011). Here, in a relatively sheltered tidal embayment, a constant supply of sand fed at one side into the embayment mouth by a longshore current was inferred to have gradually filled the subtidal part of the embayment while the tidal prism, i.e. the intertidal volume entering the embayment at high tide, remained constant. This caused outgoing tidal currents at low tide to be progressively constricted into narrowing channels. The section constriction (note: I used the term "section restriction" in my 2011 publication, but "constriction" appears to be the proper term) resulted in strongly enhanced ebb currents and vertical channel erosion. Continuing sediment input during flood caused the erosion focus in the subtidal parts of the ebb-evacuation channels to shift seawards. The channels affected by this process of section constriction and filled by thick stacks of large cross-beds were situated at the downdrift side of the embayment, and the process occurred just before its sedimentary closure.

In the Brussels Sands case, the lateral facies relationships are clear and they provided the clue to deciphering the depositional process. The lateral relationships of the different depositional facies in the Diest Sands remain to be sorted out. But the Brussels Sands model is likely to be valid for the Diest Sands. The broad paleogeographical context with a semi-enclosed embayment attached to a more open marine environment is similar; erosional channels at the base of the embayment, resembling shallow valleys at one side of the basin and much deeper, partly enclosed, elongate troughs at the other side, are present in both formations; the deposit bodies' dimensions are similar; corresponding facies are found; many characteristics of the thick cross-bedded facies XT, including the downcurrent sloping set boundaries, match those of the Brussels Sands' facies X; in both environments, there is a complete absence of wave-induced sedimentary structures.

If correct, a number of deductions can be made for the paleogeographical context of the Diest Sands. During a preliminary lowstand, the Kempen and Hageland area emerge and shallow valleys incise perpendicular to a coastline at about the border of the Roer Valley Graben (Fig. 1; Vandenberghe et al., 2014, fig. 2). During base-level rise, the sea penetrates in the valleys creating an inshore marine embayment (Fig. 7, stage 1). A possibly thin layer of transgressive sediments, containing transported and possibly also newly generated glauconite, may have been deposited during transgression (part of the Deurne and Dessel Members?). The embayment is during highstand filled by sediments fed by a longshore transport path that is trapped into it. Bioturbated, finer-grained sand is deposited at the embayment's bottom (part of the Deurne and Dessel Members?) and coarser sand in low-angle inclined beds, prograding from NW to SE (part

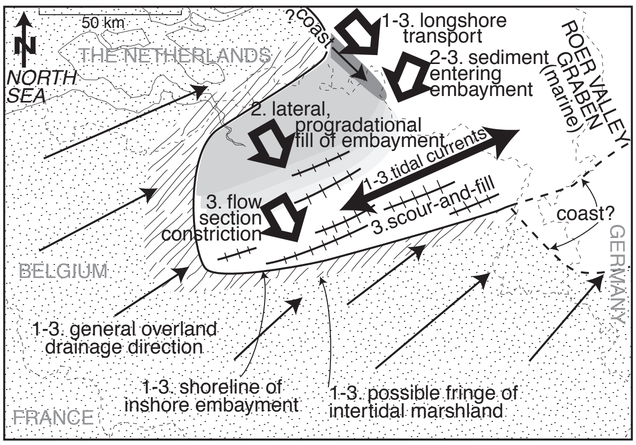

Figure 7. Hypothetical paleogeographical configuration during deposition of the Hageland Diest Sands. Note similarity with configuration proposed for the Eocene Brussels Sands (Houthuys, 2011, fig. 17). Present-day coastline and state boundaries are shown in fine lines for reference. Stages in the paleogeographical evolution: 1. drowning of a river mouth area and establishment of an inshore marine embayment; 2. first stage of filling prograding from NW to SE; 3. last stages with subtidal flow section constriction and formation of scour-and-fill elongate troughs. Stippled area is emerged.

of the Kempen Diest Sand?) (Fig. 7, stage 2). As the volume of water entering the embayment in the intertidal zone remains constant, the drainage of this tidal volume during low tide is gradually being confined to a narrowing area at the SE side of the embayment. In this stage, as the subtidal flow sections continue to decrease, falling-tide ebb currents are strongly reinforced (Fig. 7, stage 3). This causes intraformational channel erosion of the pre-existing, shallow, drowned valleys, which creates 50 to 100 m deep elongate scour troughs. In this last stage of the embayment fill-up, any previous transgressive deposits are now also removed. The exogenous sediment pressure does not relent, and thus the troughs are filled as soon as they are scoured. It is essentially the pressure of continuing external sand supply into the embayment that drives the scour focus area in the ebb-evacuating channels to shift seawards. Once a drowned paleovalley is thus deepened and filled, the next paleovalley SE of it is the scene of a repeated, seaward moving scour-and-fill process. It is possible that several such scour-and-fill stages occurred simultaneously in parallel channels, but always with the SE-most channel subject to higher currents due to increasing constriction of flow sections. The sand pressure into the embayment and the ensuing scour-and-fill mechanism is thus at the origin of the Hageland Diest medium-to-coarse sands, arranged in elongate bodies, with the SE-most bodies thickest and containing more coarse admixture. This process suite also accommodates the marine signature of the sediments. Finally, all of the available space is filled and the sea leaves the area. Lagoonal, marshland or overwash deposits may have followed on top, but the formation is truncated and no such sediments have been preserved. The sequence may have been followed by a next marine cycle, whose basal scour truncated the Diest Sands.

It can be speculated that long after the deposition of the Diest Sands, at the start of the Pliocene-Pleistocene tectonic uplift of the land surface, the coarser channel-fill sand bodies have guided groundwater flow and attracted iron precipitation. On further uplift, both the higher permeability and the ferruginous stones have made the channel fill sands relatively more resistant to continental erosion. If true, the internal sedimentary differentiation of the Diest Sands is thus at the origin of the present-day Hageland hills.

To conclude, it is proposed that the sedimentary model of the Brussels Sands (Houthuys, 2011) can be applied to the Hageland Diest Sands. All available observations are compatible with the model, but a definitive proof will only be provided if the large-scale 3D facies architecture inside the Diest Sands can be reconstructed from observations. The consequences for the Flemish Hill Sands will be detailed in section 7.

6. Direction analysis of the central Belgium river pattern

6.1. Results of the analysis

In an initial analysis step, direction calculations were done over partial areas. Somewhat unexpected, this direction analysis yielded similar results all over central Belgium. That is why, ultimately, all river segments for "central Belgium" (Fig. 4) were taken together in a mass analysis, and only category distinctions were kept in the result plots of Fig. 8. Over the complete study area (Fig. 4) and over all thalweg types, a clear peak frequency of thalwegs running NE (45 to 50°) is observed (Fig. 8A). Also the opposite direction of thalwegs running SW (230°) constitutes a local peak. A weak local peak is the direction north (0°).

When only considering the dry thalwegs (Fig. 8B) and the permanent upper courses (Fig. 8D), the same peak directions over the study area arise. The middle courses (Fig. 8E) have a wide spread of direction frequencies over the whole northern semicircle, but the peak remains NE, with a maximum frequency at 30°. The lower course directions (Fig. 8F) are again more grouped, with two local peaks at 30 and 70°.

When looking at the individual geographical pattern of the dry and upper permanent valleys, many appear to have an upper section running NE followed by a bending, either to left or right, that makes them join another dry or permanent thalweg. Fig. 9A gives an example for an area near Feignies (south of Mons) in north France, but the pattern is found all over the study area. Another striking feature is the peculiar shape of many sub-catchments (Fig. 9A-B): with respect to the central watercourse, the catchments have an asymmetric shape with a relatively larger part of the area situated to the SW, while short tributaries at the NE side, if present at all, flow often SW. Also in the few plateau relics of the northern lowland (Fig. 4), many sub-catchments have an upper part with a striking dominance of drainage to NE.

Fig. 10 presents the geographical variation in the preferential drainage direction supposed to represent primitive drainage. The direction arrows were derived based on a visual appreciation such as described in section 2.4. Areas known to have a young morphogenesis have been left out of the appreciation. The picture shows an impressive consistency in arrows pointing NE or ENE. Geographically, there is a general and gradual peak direction change from a more northerly 40° peak in the area south of the Haine-Sambre-Meuse axis, a peak around 50° in the central area, and approaching 60° in the northern area (Fig. 10). In the Kempen, especially around the SW border of the Kempen Plateau, the opposite direction emerges.

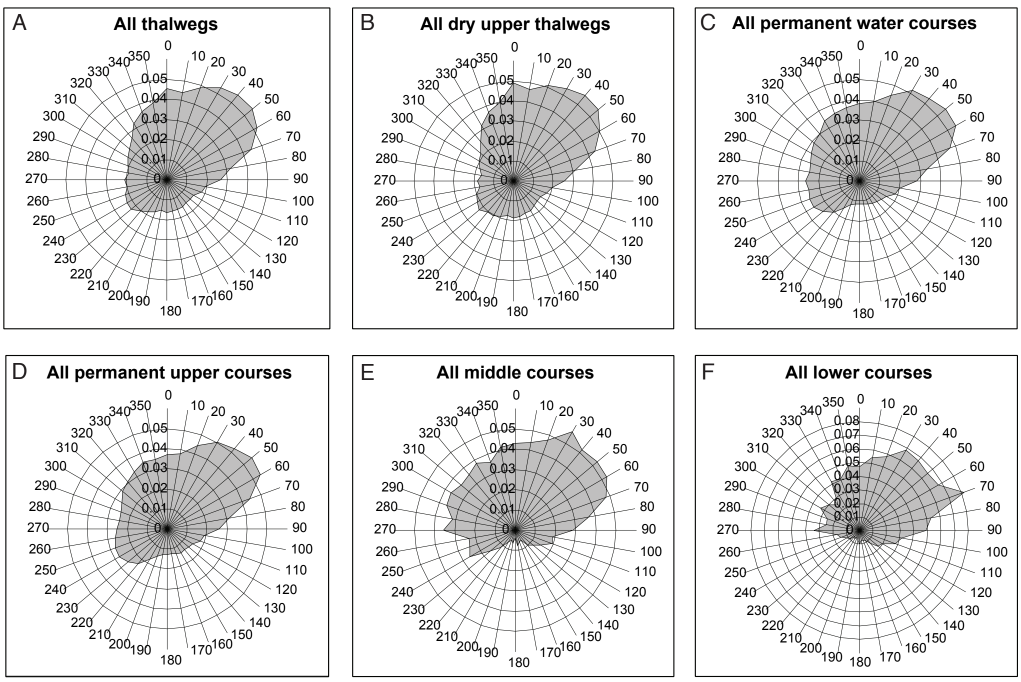

Figure 8. Drainage direction charts for the area labeled "central Belgium" in Fig. 4. Directions in degrees with 0° to north. The circles represent the length frequency values of the 10° direction classes.

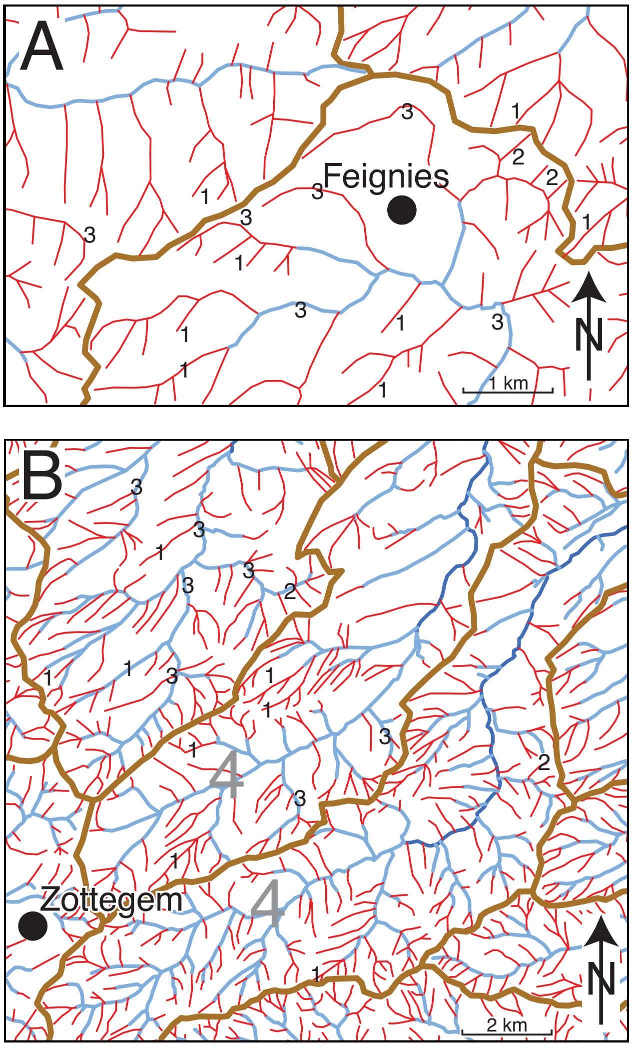

Figure 9. Thalwegs and watersheds in two arbitrarily selected small areas (location, see Fig. 4). Permanent water courses in blue, dry upper thalwegs in red, watershed divides in brown. A. Around Feignies (north France, south of Mons). Note longer thalwegs flowing NE, often bended as if diverted; note hardly any thalweg flowing SW. B. NE of Zottegem, between Gent and Geraardsbergen. Note asymmetrical development of catchments with dominance of thalwegs draining to NE and short right bank tributaries draining to the central water course. A and B. Features used as indicators of the primitive drainage direction (only a few instances have been marked): 1. consistent NE thalweg directions in sometimes long, parallel valleys; 2. short SW thalwegs; 3. bending of NE directions as a later reorganization of the drainage pattern; 4. dominance of NE draining in catchment shapes.

6.2. *Interpretation of the direction analysis*

As the youngest geological layers in north and central Belgium are shallow marine, conformal strata (albeit slightly tilted with a low dip to NNE), it can be inferred that emergence due to tectonic uplift produced a smooth, regular, low-lying land surface. On an initial slope such as a raised and tilted seafloor, runoff forms overland linear flows in a certain density (Twidale, 2004). With a lack of structural interference, parallel patterns develop with rivers flowing downslope. It is put forward here as a hypothesis that such a primitive, uniform drainage pattern was indeed present and developed in a consequent way on the freshly emerging land surface. Together with the youngest geological layers, the emergence surface is now at elevations above present-day sea level of around 50 m in the north to 200 m in the south of central Belgium. The surface has a several-millions-of-years long uplift and erosional history. The present-day plateaus and hill tops can be linked to a fairly flat surface and are interpreted to be close to the original emersion surface. Most of the uplifted area was transformed into a hilly landscape due to overland erosion and valley incision. The hypothetical original uniform drainage pattern must have been modified by the genesis of new,

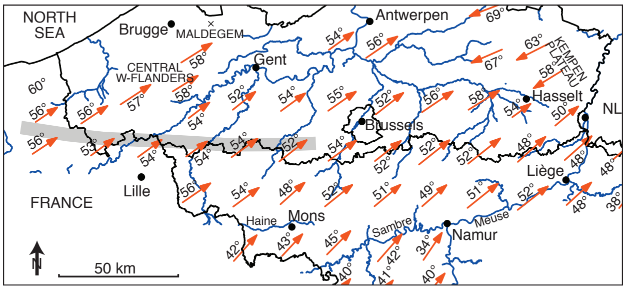

Figure 10. Geographical variation in the derived primitive drainage direction. Arrows based on visual appreciation of elements such as those labeled 1-4 in Fig. 9. Line marked by the Flemish Hills is indicated in grey.

young headwaters, retrograde erosion, river captures, diversions and gradually shifting, incising valleys. Yet, it is thought that an imprint of the original pattern may still be found. The analysis presented here is intended to demonstrate such an imprint.

This is why the uppermost parts of the valleys, situated on the present-day plateaus, have been included in the thalweg database. A drainage area of around 0.5 km² is often found the minimum for headwater rivers to respond by fluvial, as opposed to colluvial, processes to geomorphological triggers (Montgomery & Foufoula-Georgiou, 1993; Wobus et al., 2006; Crosby & Whipple, 2006). It can thus be theorized that, if a primitive pattern was present on an originally nearly flat, now uplifted surface, thalwegs and headwaters on the plateaus near the watershed divides will least have been affected by younger processes in response to changing geomorphologic constraints. If this pattern had a parallel, consequent drainage direction, this may be expressed in the actual direction of especially the upper, now dry valleys. Of course, new, younger headwaters are present in the thalweg databank too, and as they will have developed towards local low areas, they are thought to assume a nearly random distribution of directions; but as they develop along the local shortest path, they will not randomise too much the direction spectrum.

The drainage direction on the plateaus may have been obscured or newly created during the aeolian deposition of löss. In some areas, especially on the flat plateaus, over 5 m, even 10 m thick Late-Pleistocene aeolian löss deposits occur. Theoretically, they may have been deposited, or rearranged after deposition, in ridges with a preferred direction. Published evidence of such structures is scarce. Matthijs (2004) mentions up to 20 m high north-south ridges between Liège and the Jeker river (Fig. 1) consist entirely of löss on a flat top-chalk surface and could be löss depositional ridges. A löss preferential depositional direction has once been invoked as a hypothesis to explain SW – NE löss ridges on Mid-Pleistocene Meuse terraces, west of Maastricht (Boenigk et al., 1995). Both interpretations remain to be proven; taken together, they even invoke intersecting directions. Generally, however, outcrops that show entire sections of the löss and underlying Cenozoic layers in upper thalwegs confirm in most areas on the löss plateaus that dry upper thalwegs have rather been smoothed by blankets of original löss or of colluvium than vice versa (author's observations; see also Goossens, 1997). The very deposition of löss may have been guided by the pre-existing topography, where in some cases flat plateaus developed thick löss and transitions to valley flanks thinner löss (Dirk Goossens, pers. comm.). This implies that, if a thalweg is recognized in surface map contour lines, the pre-löss topography was in many cases even more explicit than the present-day relief, while in the rare cases where löss may have developed thicker deposits on pre-existing flat topography, at least the direction of the valleys and interfluves was conserved. It is therefore thought that löss sediments did not superpose a direction related only to their deposition.

The result of the quantitative direction analysis, i.e. the outspoken dominance of the NE direction over nearly the complete area considered in this study, including the upper, dry thalwegs, thus seems to confirm the hypothesis of an original uniform drainage direction. This is further supported by the visual appreciation of preferential drainage directions derived for the interfluve areas and supplemented by anomalous shapes of catchments (Fig. 10). Typical features that have been used to identify such dominant direction thought to be inherited from a primitive direction, have been marked 1-4 in Fig. 9; such features are found profusely all over "central Belgium". Somewhat unexpected, both types of analysis yield indications of a uniform primitive drainage direction all over central Belgium. It was also observed that many low water courses contain sections showing the same primitive direction.

In the lowest parts of Belgium, the primary NE-direction is also present, specifically on the small plateau remnants, such as in Central West-Vlaanderen and south of Maldegem, but also in the mean direction of many main river segments. Also in the part of the Kempen area at the SW border of the Kempen Plateau, the SW-NE direction occurs, but all rivers drain to SW.

The findings are interpreted as a confirmation of the hypothesis that the present-day river pattern derives from an original, parallel drainage pattern. It would be hard to find an alternative explanation for the pervasive NE direction found in the visual appreciation of the many catchments all over central Belgium, and the many neighbouring, often closely spaced, dry and upper segments flowing NE often in a parallel arrangement. The pattern is therefore thought to show the imprint of a primary drainage pattern developed at the time of the first emersion. Given the areal extent of the pattern, the drainage pattern is the signature of a systematic, long-lasting retreat of the sea to NE due to a slow, persisting, tectonic uplift that most raised the SW, i.e. the area of Boulonnais-Artois. This uplift shifted the coastline to the NE. Throughout the Miocene, a marine bight existed in the Lower Rhine and Roer Valley Grabens. The shoreline location at the Belgian side is inferred to have been determined, especially at low sea level, by the Roer Valley Graben's boundary faults (Demyttenaere, 1988; Vandenberghe et al., 1998, 2014). The parallel river pattern is thus consequent to that shoreline.

Attention is drawn to the fact that the line of Flemish Hills marks no break in the primitive uniform drainage direction. It persists in the area both north and south of the line of outlier hills (Fig. 10). This can only mean the hills were no topographic high at the time of origin of the pattern.

Summarizing, the central Belgium river pattern is interpreted to have originated as a parallel consequent drainage to NE on an extensive low-relief and low-slope regression surface. The direction suggests a Miocene surface, as the coastline was then situated in NE Belgium and had a NW-SE alignment. The mere size of the uniform pattern suggests the uplift and regression to NE was long-lived. The pattern was afterwards modified by Pliocene-Pleistocene uplift in the south. This provoked a cascade of river captures and even inversions, dissecting the original drainage pattern and creating composed main river courses going north and NNE.

7. Discussion on the stratigraphic position of the Flemish Hills Sands and the Neogene emersion of central Belgium

7.1. The Flemish Hills Sands versus the Hageland Diest Sands

This paper shows the depositional environments of both deposits are widely different. The Flemish Hills Sands are a shoreface deposit of a coastline at the south shore of an open sea, in a setting where angular clastics and weathered flint pebbles were being delivered to the coastline. The shoreline had an east-west trend and prograded from south to north; the open sea was to the north. The Hageland Diest Sands are the highstand fill a fast narrowing, inshore tidal embayment developed during rising base level on the site of a number of river valleys that drained into the Roer Valley Graben marine bight. The narrowing process provoked the intraformational scour-and-fill of deep, elongate channels. The channel deposits near the SE border of the embayment contain some admixture of angular clastic grains, but the depositional setting demonstrates no continental uplift. Both deposits are environmentally unrelated and incompatible.

The Kesterberg to Oudenberg hills sit midway between both deposits and were often seen as the geographical link, as their top also contained "Diest Sands". This paper however shows the Kesterberg to Oudenberg hill top sand must be assigned to the Middle-Eocene Lede Formation. The link argument is invalid.

7.2. Candidate correlates for the Flemish Hills Sands

While the Hageland Diest Sands are fairly well constrained stratigraphically (Vandenberghe et al., 2014), the correct stratigraphic position of the Flemish Hills Sands remains to be sorted out. Any Cenozoic layer above the Asse Clay Member of the Maldegem Formation can be considered a candidate correlate for the Flemish Hills Sands.

Was there an emersion after the Asse Clay deposition? Tavernier & De Moor (1975, p. 204) mention a deep red paleosol incised by the "Diest" deposits on Hotondberg near Ronse. This must then be situated in the truncated top of the Asse Clay. I found no confirmation of any weathering in the interface outcrop on Muziekberg. Therefore, though the eroded top of the Asse Clay covered by the base pebble layer of the Flemish Hills Sands indicates a sedimentary record hiatus in the Flemish Hills, the hiatus may have been short. The Late Eocene younger members of the Maldegem Formation are therefore the first depositional units to consider. In the Maldegem stratotype area and the subsurface of north Belgium, the alternation of several sand and clay layers gave rise to the identification of some depositional sequences (Vandenberghe et al., 1998), probably separated by depositional hiatuses (Vandenberghe et al., 2004). A correlative surface linking one of these northern Belgium hiatuses and the Flemish Hills Sands base is theoretically possible. The top of the Maldegem Formation's uppermost Onderdijke Clay Member is burrowed and contains peat. It is recognized as the local signature of the Pyrenean tectonic phase, situated at the turn of the Bartonian to the Priabonian (Vandenberghe et al., 2004). During this stage, the eastern part of Belgium was uplifted and slightly tilted, resulting in important erosion (Vandenberghe et al., 2004).

In the overlying Bassevelde Sand Member of the Zelzate Formation, two lower depositional sequences Ba1 and Ba2 of local extent have been recognized and placed in the Lower and Middle Priabonian, respectively (Vandenberghe et al., 2004; Saeys et. al., 2004). These sequences contain higher amounts of continent-derived clay, possibly indicating deposits related to the Pyrenean uplift (Saeys et. al., 2004).

The upper Ba3 sequence in northern Belgium and the extensively transgressive Grimmertingen sand in east Belgium mark the base of the Oligocene. All of the overlying Oligocene strata are a bit less likely candidates, mainly for basin architectural reasons (see next section).

Another correlation candidate is the Middle-Miocene Bolderberg Formation, e.g. suggested by Leriche (1921, in Tavernier & de Heinzelin, 1962). It crops out in the East of Belgium from near Tongeren (Matthijs, 1999) in a narrow strip over Leuven to Laken (Buffel & Matthijs, 2009). It is of the same Burdigalian to Langhian age as the Berchem Formation that crops out around Antwerpen and occurs in the Kempen subsurface. Just like the Flemish Hills Sands, the formation covers a regionally significant erosion surface and stratigraphic hiatus (Vandenberghe et al., 1998). There is also some similarity in the base gravel. Near Leuven, the ovoid 2 to 7 cm long, bluish flint pebbles, which are thought to be derived from Limburg chalk cliffs (Vandenberghe & Gullentops, 2001), are similar in size and shape to the Flemish Hills Sands base gravel near Ronse. Both base gravels cover a firmground. The Bolderberg Sands are fine at their base (around 80 to 100 µm) and contain a high content (about 50%) of glauconite. The deposit is completely bioturbated and may show thin mud drapes. It is clearly a marine shallow shelf deposit of depths below storm-wave base or of a near-coast, somewhat sheltered environment. In Limburg, where upper levels of the formation have been preserved, several flint pebble layers are found in the Genk Member. These pebble layers mark internal erosion surfaces; the pebbles are not dispersed in the deposit (Matthijs, 1999).

Based on the kieseloolites present in the Flemish Hills Sands, Gulinck (1960) suggested a correlation with the Kempen Kasterlee Formation. This would bring the hill top deposits in the uppermost Miocene.

7.3. Basin architecture

Many Belgian Cenozoic deposits are similar in lithology and depositional structures. Crucial additional information on the stratigraphic position can be derived from the basin architecture.

From Cassel to Flobecq, the Flemish Hills Sands cover the same Asse Clay Member of the Maldegem Formation. The glauconiferous clay with the 30 cm thick stratigraphic marker layer “bande noire” at the base, is present at Cassel, Mont des Récollets, Mont des Cats, Mont de Boeschepe, Mont Noir, Rodeberg, Scherpenberg and Kemmelberg, according to Ortlieb & Chellonneix (1870). The same clay member underlies the Flemish Hills Sands near Ronse (Jacobs et al., 1999). In spite of their erosional base, the Flemish Hills Sands are conformal with the Asse Clay over a west-east distance of almost 100 km.

Profiles A and B in Fig. 5 show the general west-east conformal character of all underlying Eocene strata east of the Schelde river and of Brussels. Fig. 11A is a generalized geological map of north and central Belgium at present sea level. Documented incisions in the base of some formations (Vlierzele, Brussels, Hageland Diest) have been left out as the map’s purpose is to show the broad configuration of the strata at the scale of the basin. Fig. 11B is a generalized west-east geological profile on the Lambert ‘72 Y=162,000 gridline, based on well data available on DOV. Most data are situated on the profile line; the hills from Cassel to Kemmelberg are a few kilometres north of the line and therefore, the contact elevations have been shifted vertically according to a generalized tectonic dip of 0.35% to the north.

The same conformal lay-out for the Paleocene to Eocene strata is clear on the simplified horizontal and vertical section presented in Fig. 11A & B. These strata constitute a conformal package documenting synsedimentary subsidence in and around the West of Belgium.

These figures also clearly show the significant low-angle unconformity at the Eocene-Oligocene boundary that is connected to the Pyrenean tectonic phase (Vandenberghe et al., 2004). In the Dutch sector of the North Sea basin during that phase, tectonic inversion of fault bounded blocks was important and geographically vast (de Lugt et al., 2003). In the Roer Valley Graben, nearly the complete Eocene marine sequence was eroded; locally, even reversed faulting occurred (Verbeek et al., 2002). In the Belgian part of the basin, this phase is marked by a low-angle, erosional unconformity at the base of the Tongeren Group (Grimmertingen Member of the Sint-Huibrechts-Hern Formation) continued to the NW with the base of the third sedimentation cycle inside the Bassevelde Member of the Zelzate Formation (Vandenberghe et al., 2004). The truncation surface cuts, going east starting near Brussels-Antwerpen, over a distance of over 100 km, through progressively older deposits, from the Upper-Eocene Maldegem Formation to the top-Mesozoic. Sediments correlated with the Sint-Huibrechts-Hern Formation covered the now culminating point of the Ardennes (Demoulin, 1995; Vandenberghe et al., 2004). Fig. 11A & B demonstrate that all strata covering the unconformity make out a conformal series. The westward prolongation of the different strata boundaries suggests they would have been at a higher level than the present-day Flemish Hills. If the Bolderberg Formation, which has a following, less erosive, low-angle unconformity at its base, is extended westward, the boundary surface would intersect the Oligocene base unconformity, but even so, the base would be at a level too high to allow a correlation with the Flemish Hills Sands.

The westward prolongation of the unconformities, indicated in Fig. 5B and 11B, is a hypothetical line above the present-day land surface, in which all confirmed nearest occurrences (such as the Bolderberg Formation at Laken) have been used and projected. In north-south sections, such as Fig. 5C, it is clear that the North of Belgium had a different subsidence rate throughout the Cenozoic than central Belgium. This causes all strata to be inflected in north-south profiles (in Fig. 5C: near Gent). A similar inflection could theoretically have affected the youngest, now eroded strata, so that Miocene or Pliocene strata might have reached the Flemish Hills area. But this should then also imply an inflection (above the present-day land surface) in the west-east sections of Fig. 5B and 11B. These profiles are situated on top of the ancient Lower-Paleozoic Brabant Massif. A hypothetical inflection of the youngest, now eroded strata, can only be supported if also the underlying strata are inflected. This is not the case. As a conclusion, it is unlikely that Miocene or Pliocene strata would have once stretched to the Flemish Hills.

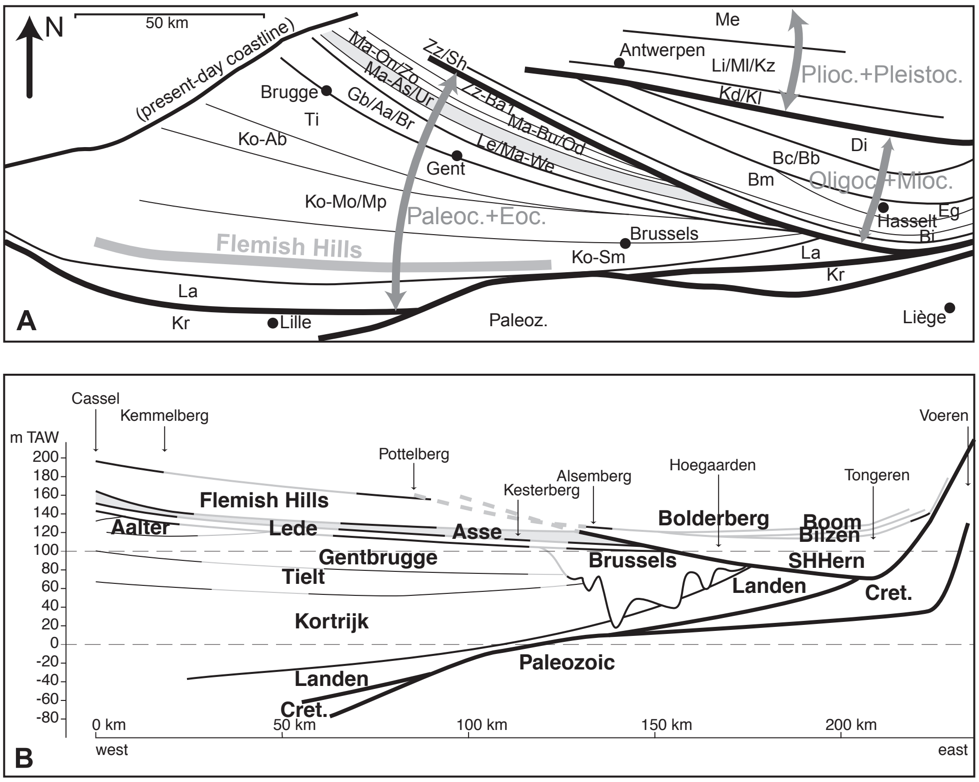

Figure 11. A. Generalized outcrop map of the Cenozoic layers in central Belgium in the 0 m elevation plane. Documented incisions and outliers have been left out. The line of the Flemish Hills is shown in grey. Thick arrows indicate the conformal strata packages. Kr = Cretaceous; La = Landen Group; Ko = Kortrijk Formation with members SM = Saint-Maur, Mo = Moen, MP = Mons-en-Pévèle, Ab = Aalbeke; Ti = Tielt Formation; Gb = Gentbrugge Formation; Aa = Aalter Formation; Br = Brussels Formation; Le = Lede Formation; Ma = Maldegem Formation with members We = Wemmel, As = Asse, Ur = Ursel, On = Onderdale, Zo = Zomergem, Bu = Buisputten, Od = Onderdijke; Zz = Zelzate Formation with Ba1 = first sedimentary cycle of Bassevelde Member; Sh = Sint-Huibrechts-Hern Formation; Bi = Bilzen Formation; Bm = Boom Formation; Eg = Eigenbilzen Formation; Bc = Berchem Formation; Bb = Bolderberg Formation; Di = Diest Formation; Kd = Kattendijk Formation; Kl = Kasterlee Formation; Li = Lillo Formation; Ml = Mol Formation; Kz = Kiezeloöliet Formation; Me = Merksplas Formation. B. Generalized west-east section through central Belgium, at the latitude of the Flemish Hills. Full lines are documented; grey lines are extrapolations based on outcrops at some kilometers more north (elevations corrected for a general 0.35% tectonic slope to north). Thick lines delineate conformal strata packages. Group or formation names in bold (SHHern = Sint-Huibrechts-Hern; Cret. = Cretaceous).

7.4. The original regression surface revealed by the river pattern

All over central Belgium, a primitive, uniform, parallel drainage pattern to NE has been detected (Fig. 10). It is thought that this pattern developed on an originally almost flat regressive surface with a very low slope to NE. Given the relatively large area over which this pattern persists, over a generalized surface in which successively younger strata crop out when going in the same NE direction, towards the Roer Valley Graben, the development is thought to have taken a very long time period, in which the surface was subject to slow, continuous tectonic tilting. The tectonic movement raised the SW of central Belgium and the adjoining part of northern France and SE England, with culmination in the Weald-Artois High. Such an uplift and an associated NE regression may have taken place since the Late Oligocene, and may have persisted throughout the Miocene. The Oligocene sediments are restricted in Belgium to the northeast, though they may also have covered the southwest; the position of the shoreline is not known (Vandenberghe et al., 1998). The start of uplift may then date from the Savian phase, an erosive hiatus well documented in the Netherlands subsurface (Duin et al., 2006), situated at the transition Oligocene to Miocene. The uplift of the Artois-Weald High may have been slow and long-lived. In SE England, the uplift and inversion of the Weald is thought to be at its maximum probably around the transition Oligocene to Miocene (King, 2006).

It is remarked that the general primitive drainage direction is found back in the fossil valleys preserved at the base of the Diest Formation (Vandenberghe et al., 2014, fig. 2). These valleys were eroded after the deposition of the Middle-Miocene Berchem/ Bolderberg Formation, in the Late Serravallian (Vandenberghe et al., 2014). The incision pattern is compatible with the derived original drainage pattern over the whole of the area indicated in Fig. 10. This association provides a date ante quem for the emersion surface of central Belgium, situated landward of the Diest Formation: Late Serravallian. Earlier, Middle-Miocene deposits are preserved as far SW as Wemmel and Laken (de Heinzelin & Missonne, 1958). The preserved deposits indicate an inner shelf; therefore the coast must have been situated further SW. The precise extent of the Miocene transgression cannot be traced.

Taking everything together, the low-relief NE sloping land surface on which the drainage system of central Belgium originated, is thought to have existed during a long time, probably throughout the Miocene. After regression of the Middle-Miocene, only the lower reaches of the river pattern are thought to have incised near their mouth in the Roer Valley Graben sea bight and later to have been covered during the Tortonian by Diest deposits (Vandenberghe et al., 2014). If this cycle is purely a base level cycle, which is supported by the sedimentary model presented for the Diest Sands in this paper, it is only natural that younger river patterns covering the Diest Sands continue to drain to NE. A very long-lived, slow rise of SW Belgium, northern France and SE England is thus a simple driving force that explains both the primitive river pattern and the dominant current direction inside the Diest marine deposits.

The pattern of parallel NE draining rivers probably persisted into the Pliocene and even the Early Pleistocene (Westerhoff et al., 2008, fig. 9). In the central Kempen area SW of the Kempen Plateau, the direction is found but rivers are interpreted to have suffered inversion. This may have occurred during the Pleistocene uplift (van Balen et al., 2002; Demoulin & Hallot, 2009) of the Ardennes and neighbouring areas, followed by incision and capture by the gradually eastward extending Flemish Valley (of which the Lower Dijle, Demer and Nete are remnants). This interpretation is in contradiction to the development suggested by Vandenberghe & Desmedt (1979), that NNE flowing rivers would during gradual incision have adapted to SW-NE structures present in the underground.

7.5. Bringing all arguments together

The Flemish Hills Sands and the Hageland Diest Sands are the deposits of completely different and incompatible sedimentary environments. Based on this observation, both deposits cannot belong to one formation.

In spite of some similarities, the Flemish Hills Sands are thought not to correlate with the Bolderberg Formation. The sedimentation environment is different; the base gravel consists of differently shaped and weathered flint pebbles and other different elements and is not fossiliferous like in the closest confirmed outcrop in Wemmel (de Heinzelin & Misonne, 1958) but also in Limburg (Matthijs, 1999); the intraformational pebble layers are different (Matthijs, 1999); the glauconite content is different.

The main argument against a correlation with the Bolderberg Formation is basin geometry. The Bolderberg Formation and the other Oligocene to Miocene strata form a conformal series in NE Belgium. Extending one of these strata over 100 km westward would call for an exceptional tectonic movement out of line of the long-term trends. Also, an inflection indicative of differential tectonic movements, is not documented in the west-east profiles.

This argument is a fortiori valid for the overlying Diest Sands.

A short-lived tectonic movement bringing a Miocene marine environment over west Belgium is also in contradiction with the primary drainage pattern. This indicates a persistent uniform drainage to NE during all of the Miocene. A slow, long-lived rise of Artois and its NE flank in the widest sense is the logical explanation for this drainage pattern. Local tectonic subsidence of the Kempen and the Roer Valley Graben, coupled with eustatic sea-level changes, brought successive Miocene marine incursions over the Kempen (Vandenberghe et al., 1998, 2004, 2014), without, after sedimentary infill and regression, essentially changing the master drainage direction.

It is noted that the line of Flemish Hills constitutes no break in the overall NE drainage pattern. This indicates the Flemish Hills were covered by younger layers or, at least, did not mark a positive relief, at the time of origin of the drainage pattern. This observation also invalidates the hypothesis that the Flemish Hills would mark a Diest Sands shoreline (Gullentops, 1957). The line of the hills intersects the uniform primitive drainage direction at an acute angle, where river patterns commonly develop at right angles to a regressive shore.

This paper proposes a new correlation of the Flemish Hills Sands to one or both of the lower Bassevelde Member cycles, labelled “Ba1” and “Ba2” in Vandenberghe et al. (2004). The arguments are: (1) The Bassevelde Member of the Zelzate Formation is part of the conformal series of Upper-Eocene southern North Sea Basin strata. The Flemish Hills Sands belong to this conformal series (Fig. 11A & B). (2) The Ba1 base covers a hiatus (Gulinck, 1969, in Vandenberghe et al., 1998); also at the Flemish Hills Sands base is an erosive hiatus. (3) East Belgium was tilted during the Pyrenean phase at the transition of Bartonian to Priabonian (Vandenberghe et al., 2004). The Priabonian lower Bassevelde cycles were thus deposited during uplift of east Belgium. The Flemish Hills Sands contain a considerable content of land-derived elements, especially in the MP beds. Uplift of east Belgium and NE France can have delivered chalk pebbles, weathered flint pebbles, angular quartz grains, kieseloolite pebbles, etc. to the shore. (4) The Ba1 cycle has a transgressive, glauconitic base. So do the Flemish Hills Sands in the LB unit. (5) The clay minerals of the Ba1 cycle are characterized by a rapid upwards decrease of smectite and increase of kaolinite (Saeys et al., 2004, where the cycle is labelled “S3.1”). It was suggested that kaolinite was delivered from older saprolites present in the uplifted Ardennes. The clay minerals from a sample of the Flemish Hills Sands also show a composition different from marine deposits and a relatively high content of kaolinite (Rieko Adriaens, pers. comm.).

The Upper-Eocene Flemish Hills Sands may well have been covered by younger strata, similar to the succession seen in north Belgium. It is thought, however, that renewed uplift of Artois and west Belgium started in the Oligocene and was responsible for emergence of the area, certainly since the Miocene. This is the time of origin of the systematic, parallel overland drainage pattern to NE. The uplift must have been slow for the drainage pattern to have persisted till the end of the Miocene, such as indicated by the consistent NE drainage directions both at the base and on the top of the Kempen and Hageland Diest deposits. It is speculated that red-coloured deep soils, of which remnants are found on the Flemish Hills, but also on other high points such as Bois de la Houssière, may have been formed during that long time, during which a last phase of a subtropical climate was established in our region, especially during the Langhian-Serravallian (e.g. Utescher et al., 2002).

It is further speculated that groundwater circulation was favoured by the relatively porous Flemish Hills Sands where two normal regressive deposits are stacked: a thicker one formed by the LB and MP beds, and a thinner upper one formed by the UG beds. The fastest stage of base-level rise during the deposition of the Flemish Hills Sands may have halted the progradation of the lower normal regressive series right at the point where the sedimentation front, prograding from the south, had reached the west-east line marked by the present-day Flemish Hills. This may even explain the present hill line: due to these changing conditions, the deposits would have developed thickest right on this line. The axis of thickest sand development may have attracted groundwater loaded with dissolved iron from neighbouring areas. The better porosity may have favoured aerobic conditions and the precipitation of iron may have produced, especially at that line, the massive and capricious ironstones that are found today in the Flemish Hills Sands. Miocene and more recent uplift and differential erosion made this line to stand out as a positive relief and become the Flemish Hills.

The eastward extension from Oudenberg to Kesterberg may originally have been covered by a similar deposit of Flemish Hills Sands, with similar ironstones. They were completely eroded, but may still explain the linear occurrence of Lede Sands topped outlier hills here.

7.6. Paleogeographical implications

If a Late-Eocene age for the Flemish Hills Sands is confirmed, many aspects of the Cenozoic paleogeographical evolution of central and north Belgium and northern France need revision. It is too early to elaborate on details, but with the necessary reserve, some major implications for the Cenozoic paleogeographical evolution of the area are given here.

The Flemish Hills Sands might represent the depositional sequence associated directly with Late-Eocene uplift and tilting of east Belgium and NE France during the Pyrenean tectonic phase.

The uniformity of primitive drainage directions across central Belgium is thought to document a Miocene, long-lived, slow rise of Weald-Artois and tilt of the land surface in central and north Belgium to NE. The Flemish Hills Sands mark no break or barrier in the drainage pattern and thus were covered or, at least, did not mark a positive relief feature during the Miocene.

The sedimentary environment and new proposed Upper-Eocene stratigraphic position of the Flemish Hills Sands, and the Miocene low-relief emersion surface, exclude a last, major "Diest transgression" over the entire north of Belgium and France has ever taken place. The Hageland Diest embayment reached to the vicinity of Brussels but extended no farther west or SW. This is in line with the general Miocene trend of stepped marine retreat to NE. The location of the Kempen and Hageland Diest confined basin was probably predisposed by the valleys belonging to the Miocene overland drainage system.

The result is also in contradiction to the widely accepted view that the general direction of all main rivers in central Belgium to NNE marks a final regression to NNE (Cornet, 1904; Leriche, 1914; de Heinzelin, 1964). The well-established view of rivers consequent to a NNE marine retreat, present in many school textbooks and general paleogeographic overviews, such as Gibbard (1988), need at least local revision. When looking closely at the main river courses, it can be noted that they are a concatenation of segments flowing NE linked by segments flowing north. Such an arrangement cannot result from a general, gradual NNE-ward retreat of a coastline.

The present-day central Belgium NNE-flowing main river sections are now thought to be the result of a reorganisation of the primitive drainage pattern, through a chain of river captures, in response to a Pliocene and Pleistocene uplift of the South, later the SE of Belgium. It is thought that the lowest part of the NE directed primitive drainage system in the Kempen (now situated at the west rim of the Kempen Plateau) was inverted in response to strong Pleistocene uplift of east Belgium.

Also part of the Pleistocene history of river incisions and river terraces (e.g. Tavernier & De Moor, 1975) needs revision.

8. Proposed new formation name: Flemish Hills Formation

The new formation name of "Flemish Hills Formation" is proposed to designate the medium to locally coarse-grained, ferruginous sands, up to now mapped as "Diest Sands" on the summits of the Flemish Hills from Cassel to Flobecq. The Noires-Mottes ferruginous top sands (Van Vliet-Lanoë et al., 2010) may turn out to belong to this formation too, but this remains to be verified. The Flemish Hills Sands contain one, possibly two sedimentary cycles of an open, west-east oriented, marine shoreline prograding to the north. As a stratotype, the permanently accessible sunken lane NW of Flobecq village leading to the Pottelberg hill summit (Lambert 72 X 102,685; Y 161,490) is proposed. This can be supplemented by the deserted sandpit on Muziekberg NE of Ronse (X 98,515; Y 161,830). Based on basin geometry arguments, an age correlation with the lowest cycle or cycles of the Upper-Eocene Bassevelde Member of the Zelzate Formation is suggested. It is realized that the age correlation is speculative and remains to be sorted out.

9. Conclusion

The new lithostratigraphic formation name of Flemish Hills Sands is proposed to designate the medium to locally coarse-grained, ferruginous sands crowning the Flemish Hills from Cassel in France to Bois de la Louvière in Flobecq (Fig. 1). The sedimentary structures and lateral and vertical gradients present in the Flemish Hills Sands prove they were the shoreface deposits of an open, west-east, marine shoreline prograding to the north. The nearly 25 m thick succession at Pottelberg is interpreted as a stacking of a lower normal regression formed by the LB and MP beds, followed by a second normal regression, the UG beds. The Oudenberg to Kesterberg "Diest Sands" are reinterpreted as outliers of the Middle-Eocene Lede Sands. The Upper-Miocene Hageland Diest Sands have a setting, basin dimensions, and detailed sedimentary structures and facies transitions suggesting an analogy with the last stages of sedimentation of the Eocene Brussels Sands (Houthuys, 2011). The proposed analogy needs further support, yet to be established, from the large-scale internal fill structure of the Diest Sands. If confirmed, the reason for the strong-current fill of narrow, elongate channels would be continuous sand supply by a coastal pathway under highstand conditions into a semi-enclosed embayment over the Kempen and the Hageland, tributary to the Roerdal Valley Graben and the Lower Rhine Marine Embayment. The fill would have proceeded under highstand conditions from NW to SE. The last stages of the fill would be associated with ebb flow section constriction, strong ebb currents and the intraformational channel erosion and simultaneous deposition of the Hageland Diest Sands.

In any case, the depositional environments of the Flemish Hills Sands and the Diest Sands belong to completely different paleogeographical configurations. The broad buildup of conformal series of Cenozoic deposits in central Belgium and an analysis of the primitive drainage pattern of the land surface lead to a consistent picture of the emersion of central Belgium. The emersion is the result of a gradual, slow and long-lived uplift of SE England, northern France and SW Belgium during most of the Miocene, with the development of a uniform parallel drainage system directed to the Roerdal Valley Graben. The line of Flemish Hills constitutes no break in this primitive pattern. The implication is that the Flemish Hills Sands must be much older than the Miocene uplift and the Upper-Miocene Diest Sands. It is argued that they might be outliers of the lowest cycle or cycles of the Upper-Eocene Bassevelde Member of the Zelzate Formation. Their high content of continental clastics would be a product of continental uplift of east Belgium and NE France during the Late-Eocene Pyrenean tectonic phase. Their present-day elevated position is the result of differential erosion after the emersion of central Belgium.

The implications for our understanding of the origin and development of central Belgium's geomorphology are far-reaching.

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