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Tectonic and climatic signals in the Oligocene sediments of the Southern North-Sea Basin

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ABSTRACT. The Oligocene sediments formed between the Pyrenean and Savian tectonic pulses. The earliest Oligocene was characterized by a widespread shallow water transgression. Global cooling coincided with the subsequent retreat of the sea which is also the time of the Grande Coupure faunal turnover. Renewed stepwise transgression resulted in the deposition of the Boom Clay during the Rupelian. High-frequency cyclic changes in water depth of the Boom Clay are driven by waxing and waning of ice masses while lower-frequency cycles can be tectonic signals. By the end of the Rupelian, differential vertical tectonics resulted in considerable erosion west of the Campine subsidence area and in shallower water depth in the eastern part of the southern coastal area. Subsidence of the Lower-Rhine graben resumed at the start of the Chattian. The sea could only briefly transgress over the area outside the graben but in the graben thick Chattian sediments are preserved. Outside the graben, erosion continued to dominate during the Chattian and the Aquitanian. This long period above sea level is due to a combination of the Savian tectonic uplift pulse and a global low sea level.

KEYWORDS: Belgium, Boom Clay, Chattian, cyclicity, Rupelian, Rupel Group, stratigraphy, Tongeren Group.

1. Oligocene chronostratigraphy and paleogeography of the North-Sea Basin

This paper discusses the Oligocene in the southern North-Sea basin, in honor of Ernest Van den Broeck, the eminent 19th century specialist of the Oligocene in Belgium (Van den Broeck, 1883, 1884, 1887, 1893). The Oligocene sediments discussed in detail in this paper (Table 1 and Fig. 1) are comprised between two regional unconformities that are well recognized in Western Europe: the Pyrenean and the Savian unconformities.

The Oligocene is the geological Series and time Epoch between 33.9 and 23.03 Ma. It is subdivided in the older Rupelian (33.9-28.1 Ma) and younger Chattian (28.1-23.03 Ma) stages (www.stratigraphy.org). All Paleogene stages, except the Priabonian, amongst which the Rupelian and the Chattian, are historically defined in the North Sea Basin realm. However the required precision for defining the boundaries between the stages, expressed by the often tedious search for a suitable Global Boundary Stratotype Section and Point (GSSP), cannot be found in the North-Sea Basin area. The reasons for this absence of suitable boundary

Waasland-Boom and West Campine area	**Brabant area**	**East Campine and Tongeren area**
	RUPEL GROUP	
		Voort Formation
		Veldhoven Member
Eigenbilzen Formation		Eigenbilzen Formation
Boom Formation	Boom Formation	Boom Formation
Boeretang Member		Boeretang Member
Putte Member	Putte Member	Putte Member
Terhagen Member	Terhagen Member	Terhagen Member
Belsele-Waas Member		
		Bilzen Formation
		Kerniel Member
		Kleine-Spouwen Member
		Berg Member
	TONGEREN GROUP	
Zelzate Formation	Borgloon Formation	
	Heide Horizon	Alden Biesen Member
	Henis Member	Henis Member
	Boutersem Member	
Ruisbroek Member (s4)	Kerkom Member	
	Hoogbutsel Horizon	
	Sint-Huibrechts-Hern Formation	
Watervliet Member (a4)		
Bassevelde Member (s3)	Kesselberg Member	
(Ba1, Ba2, Ba3 sequences)	Neerrepen Member	Neerrepen Member
	Grimmertingen Member	Grimmertingen Member

Table 1. General lithostratigraphic table of the Tongeren and Rupel Groups in northwest (Waasland, Boom, West Campine), central (Brabant) and northeast (East Campine and Tongeren) Belgium. The location of the Waasland, Boom, Campine, Brabant and Tongeren areas can be found on the location map in Figure 1. For stratigraphic details the reader is referred to the website of the National Commission for Stratigraphy of Belgium http://ncs.naturalsciences.be/paleogene-neogene/paleogene-lithostratigraphy.

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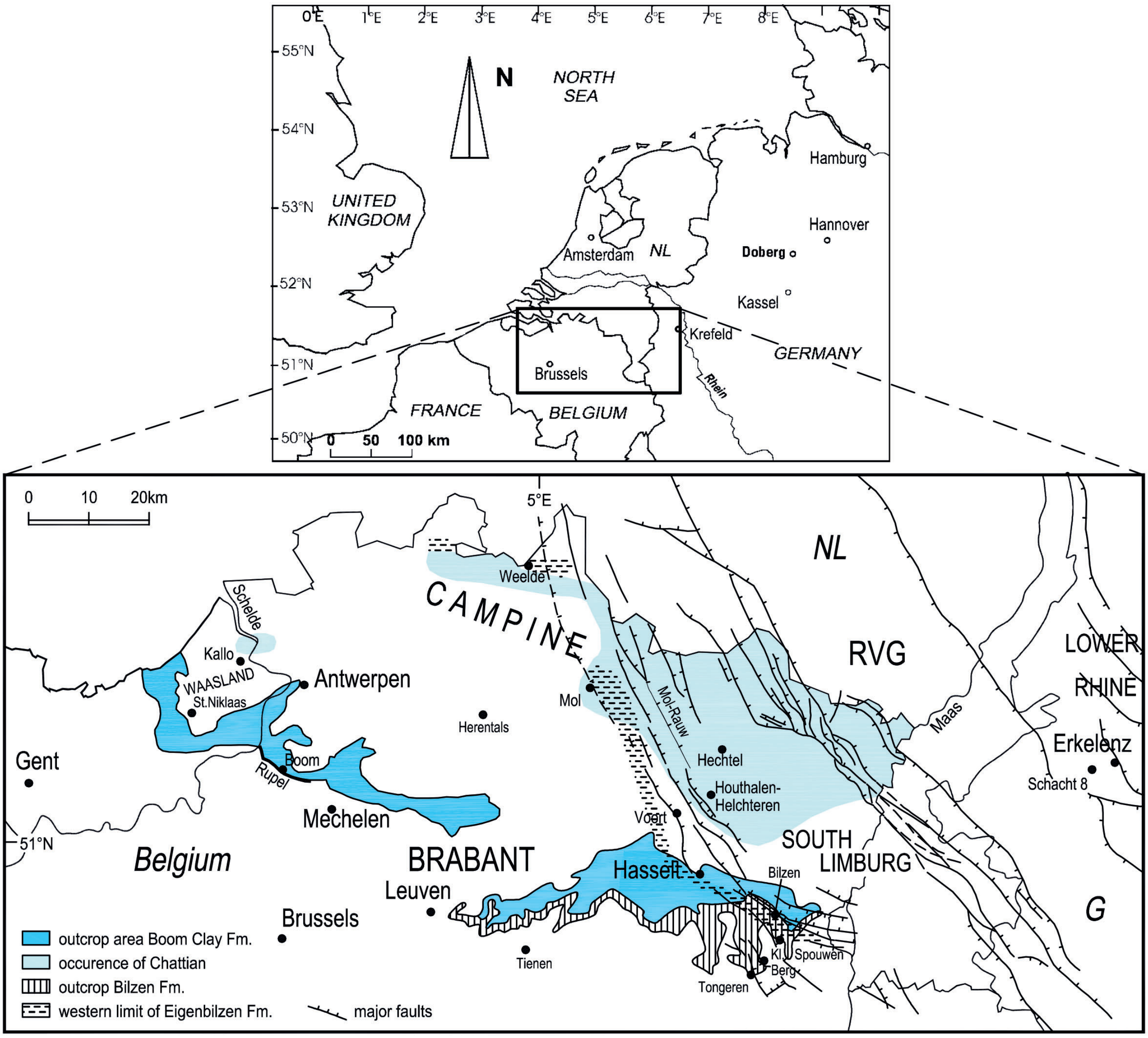

Figure 1. Location map of the North Sea area and detail of the southern North Sea Oligocene outcrop area discussed in the paper. NL stands for The Netherlands and RVG for Roer Valley Graben. The position of the Mol-Rauw fault is indicated.

sections relate to the paleogeography. The map of Europe at the time of the Oligocene (Fig. 2) shows a confined North-Sea Basin with the connecting passages to other seas in all directions small and shallow. This relative isolation has led to a mostly indigenous biological evolution in the Basin making biostratigraphic correlations with sections in other areas in the world difficult. In addition, Paleogene marine sediments in the outcropping areas around the Basin are all shallow water facies, by nature not very well suited for inter-basin biostratigraphic correlations. Also, Paleogene tectonic activity in Europe, driven by the Alpine evolution and the northwards extension of the North-Atlantic Mid Ocean Ridge, has also caused many uplift phases leading to hiatuses and erosional unconformities. Historically, such events were often chosen to delimit stratigraphic units and although they are good correlation levels they are only so regionally, and in addition they miss the stratigraphic resolution required in modern stratigraphy. Therefore, the boundary stratotype section for the base of the Rupelian stage is defined in the Massignano section in Italy (Premoli Silva & Jenkins, 1993), although the Rupelian stage name refers to the area of the Rupel river in Belgium, and in a similar way the Chattian stage refers to an area in Germany (De Man et al., 2010) whilst the recently defined boundary stratotype for its base is also defined in Italy (Coccioni et al., 2016). In a similar way the already defined GSSP's for the Paleocene and Eocene stages are all located outside the North-Sea Basin area (Vandenberghe et al., 2012).

2. The Pyrenean unconformity

In the southern North-Sea Basin, regional stress relaxation during most of the Priabonian has caused broad doming inversion and mild fault reactivation, the Pyrenean tectonic phase (Deckers et al., 2016). The Roer Valley Graben area, already inverted during the Late Cretaceous Sub-Hercynian phase and subsequently partly eroded, was uplifted again above sea level at the beginning of the Priabonian. The tilting and erosion of the strata can be observed as a low angle unconformity in sections from east to west in the Campine of North Belgium (see sections in Vandenberghe et al., 2003; Saeys et al., 2004; Deckers et al., 2016). In the east and southeast the land remained emerged and was subject to erosion till the start of the Rupelian whilst in the west sedimentation resumed already during the Priabonian. In these latter sediments 2 pulses are identified, Bassevelde-1 developed during the nannoplankton NP18 biozone and Bassevelde-2 during the nannoplankton NP19-20 biozone; the latter is onlapping slightly further southeastwards than the Bassevelde-1 pulse (Vandenberghe et al., 2003; Saeys et al., 2004). The Pyrenean reorganization of the provenance areas for the sediments in the North Sea basin must have been considerable as shown by the abrupt change in clay mineral composition (Fig. 3). During the Eocene up to the top of the Bartonian just below the Pyrenean unconformity, smectite is very dominant whilst at the base of the Priabonian just above

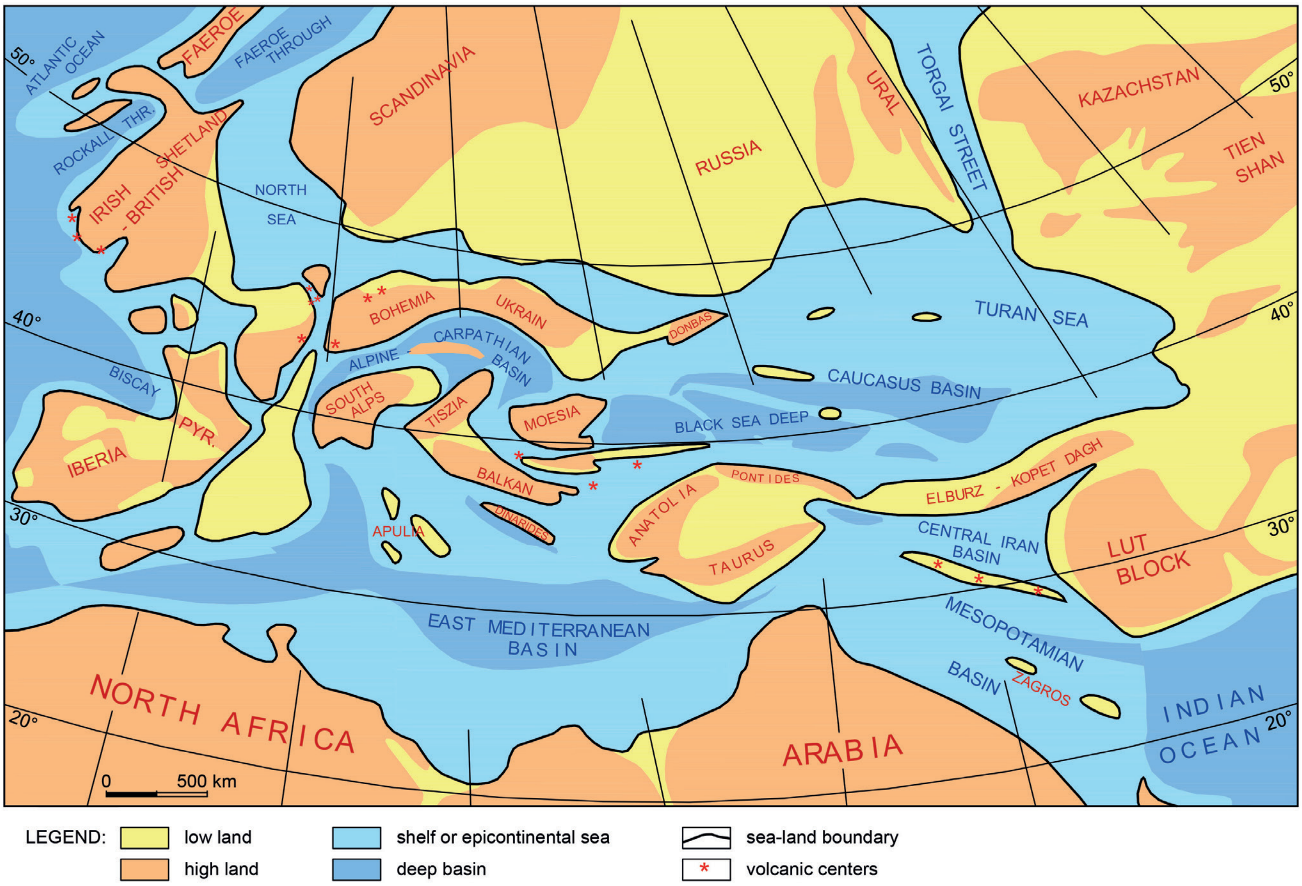

Figure 2. Paleogeographic sketch during the early Oligocene of Europe between North-Africa, Asia, and the Atlantic and Indian Oceans (adapted from and based on Popov et al. (2009), Laenen (1998), Rögl (1998), Vandenberghe et al. (2003a) and Vandenberghe & Mertens (2013)).

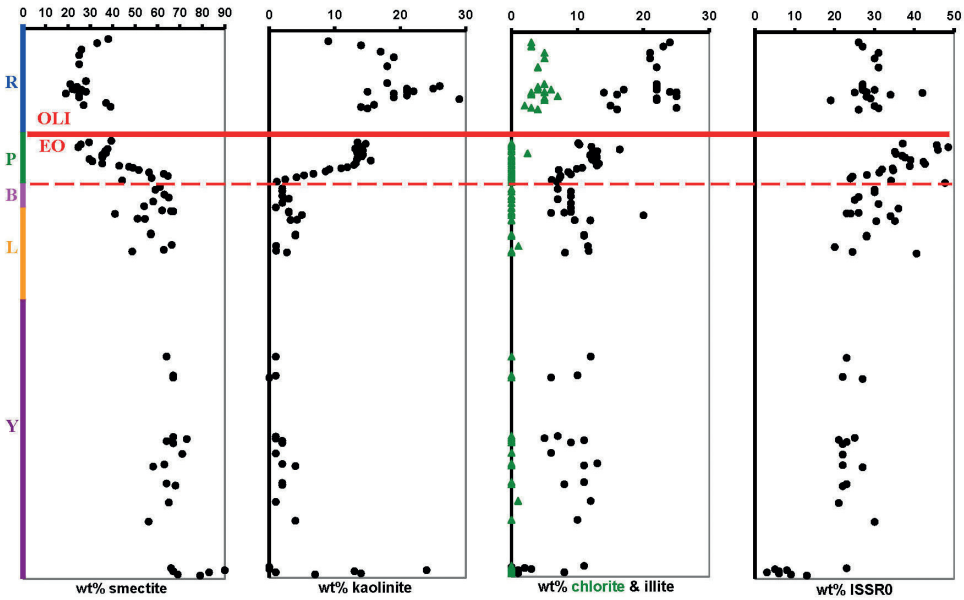

Figure 3. Clay mineral evolution during the Eocene and the Lower Oligocene in northern Belgium. ISSRO stands for Random interstratified illite-smectite clay minerals; Y for Ypresian, L for Lutetian, B for Bartonian, P for Priabonian and R for Rupelian. The full red line is the Eocene/Oligocene boundary and the interrupted red line is the level of the Pyrenean unconformity at the Bartonian/Priabonian transition (from Zeelmaekers, 2011).

the unconformity very rapidly a mixture is installed of illite, smectite, random mixed layers illite-smectite, and two types of kaolinite (Zeelmaekers, 2011). The Eocene smectite type, its iron content and layer charge (Zeelmaekers, 2011) are typically related to volcanism (Środoń et al., 2009) and more precisely derived from the alteration or weathering of basic pyroclastic rocks (Christidis, 2006). This interpretation is consistent with the association in other parts of the North Sea Basin and in the North Atlantic realm of Ypresian smectite with alteration of basic pyroclastics originating from volcanism accompanying the opening of the North Atlantic Ocean between Greenland and Europe at that time (Huggett & Knox, 2006). The intense argilization into smectite in the soils which developed on the volcanic rocks, was helped by the warm and humid Eocene climate. During the Lutetian-Bartonian, the area between the Paris and the Belgian Basins was uplifted leading to slight changes in clay mineralogy but the significant changes in clay mineral assemblage occurred by the early Priabonian after the Pyrenean inversion (Zeelmaekers, 2011). The Pyrenean tectonic episode led to the uplift and consequent erosion of kaolinite-rich Mesozoic paleosols that covered the Artois-Brabant-Ardennes axis in the southern hinterland of the Basin (Yans et al., 2003). Also at the same time or very shortly after, a major uplift of Fennoscandinavia started (Faleide et al., 2002; Huuse, 2002) becoming a major sediment source even for the southern North Sea during the Oligocene (Fig.

11 inset). The difference between a warm Eocene and cool Oligocene climate invoked for the sudden change in clay mineralogy (Chamley, 1989; Saeys et al., 2004) is less probable as the mineral change suddenly occurs at the base of the Priabonian whilst the cooling is more gradual and its main step occurred in the earliest Rupelian (De Man et al., 2004; De Man, 2006a).

3. Earliest Rupelian deposits and stratigraphy at the southern rim of the North Sea Basin

The land area emerged and eroded during the Pyrenean pulse at the end of the Eocene, was submerged again by the start of the Oligocene (Fig. 4). The stratigraphy of this new transgression became well established by recent detailed litho- and biostratigraphic work (De Coninck, 2001; Vandenberghe et al., 2003; De Man, 2006b). It confirmed older suggestions (Batjes, 1958; Drooger, 1969) for a chronostratigraphic correlation between what was originally mapped as Tongrian by Van den Broeck (1883) and Van den Broeck & Rutot (1883) in outcrops in east and central Belgium and the upper part of what was originally described in the subsurface of the Antwerp-Mechelen area as the Kallo Complex between the Asse Clay and the Boom Clay by Gulinck (1969). In modern lithostratigraphic terminology (see ncs.naturalsciences.be) (Fig. 4 and Table 1) this relationship is expressed as follows: the Sint-Huibrechts-Hern Formation of the outcrop area in Brabant and South Limburg (Fig. 1 and geological maps of Flanders 1:50 000), corresponding to the formerly called marine Tongrian and abbreviated in former literature as Tg1, correlates with the upper sequence of the Bassevelde Member namely the Ba3 unit and the overlying Watervliet Clay Member (a4 unit in the former Kallo Complex terminology of Gulinck (1969)), both Members of the Zelzate Formation in the subsurface to the west. Nannoplankton and especially dinoflagellate biozonations allow to date this transgression at the very base of the Rupelian stage of the Oligocene (Stover & Hardenbol, 1993; Vandenberghe et al., 2003; Vandenberghe et al., 2004) (Fig. 4).

This earliest Rupelian transgression was very widespread, certainly extending into the present Condroz in south Belgium (Gulinck, 1963; Soyer, 1978; De Coninck, 1996) and probably also over the present Ardennes as part of the Pierre de Stonne, cemented sediment in the Lorraine area of Luxembourg and Northeast France, is considered to be earliest Oligocene (see Quesnel et al., 2003). The very fine sandy lithology points to shallow marine deposits and only in the west the deeper water Watervliet Clay facies developed (Table 1). A glauconitic Neerrepen sand Member consistently exists in the top the Sint-Huibrechts-Hern Formation in the outcrop area. Current and bioturbation structures characterize this Neerrepen sand Member (Fig. 5) apparently bordered to the north by a coastal barrier quartz sand body exposed in the classical outcrop of the Kesselberg near Leuven (Vandenberghe & Gullentops, 2001; Vandenberghe et al., 2004). Over the whole outcrop area a well expressed podzolic-type soil, first described in detail by Buurman & Jongmans (1975), can be observed in the top of the glauconitic Neerrepen Member (Fig. 6) which is pointing to a regression of the sea with the installation of land conditions. The regression was widespread and also the top of the Watervliet Member clay in the west is eroded (Vandenberghe et al., 1998) (Fig. 6).

Although this installation of land conditions occurring around the nannoplankton biozones NP21 to 22 transition and after the earliest Oligocene marine sequence was very short-lived, it was a remarkably important time in the history of climate and bio-evolution, the significance of which exceeds by large the regional geology of the North Sea Basin area.

At that time the most pronounced rapid oxygen isotope increase in the Cenozoic occurred indicating a sharp decrease in temperature; this global climatic change is related to the rapid growth of a large continental-scale ice cap over Antarctica marking the change-over from the greenhouse climate regime to the present ice-house time in which we still live today (Miller et al., 1991; Zachos et al., 2001). This cooling is considered a consequence of the installation of circum-Antarctic deep-water currents after South America and Tasmania were sufficiently separated from the Antarctic continent to form a deep oceanic realm around the now isolated continent (Ivany et al., 2003; Miller et al., 2005). The opening of the Southern Ocean passages resulted in increased wind-driven mixing of ocean water which is expressed by the observed large inter-basin variations in benthic foraminifera oxygen isotope gradients initiated in the early Oligocene (Cramer et al., 2009).

De Man (2006a) and De Man et al. (2004) have reconstructed in detail the oxygen evolution from the Eocene to the Oligocene in the southern North-Sea area and concluded that although stepwise deterioration of the climate occurred already during the earliest Oligocene, the rapid global climate cooling is situated between the end of the earliest Oligocene transgression-regression cycle described above and the start of the following new inundation of the landscape.

In the Hoogbutsel Bed swamp (Table 1, Fig. 5) that developed upon this renewed inundation of the land, a rich tetrapod fauna is found amongst which the first mammals of Asian origin indicating the faunal turnover known as the Stehlin Grande Coupure (Misonne, 1957; Glibert & de Heinzelin, 1952). This faunal change occurred around the same time in Europe, North America and Asia, suggesting the impact of the major climate change at the Eo-Oligocene transition. In Asia the change is named the Mongolian Remodelling (Meng & McKenna, 1998) where cooling had replaced dense forests by more open forest and grassland leading to a change in mammalian fauna (Bender, 2013, p. 149). Migration of Asian species to Europe was made possible by the creation of a passage over a drying Turgai Street and northern Paratethys branch (Popov et al., 2009, plates 3 & 10; Woodburne & Swisher, 1995) (Fig. 2).

4. The definition of the base of the Oligocene

The Grande Coupure is a particular stratigraphic reference horizon which has been proposed in the past as the Eocene/Oligocene limit (Cavelier, 1979; Pomerol, 1982). It has turned out to coincide with the major oxygen isotope excursion discussed above and which can be considered as a potentially better marker and correlation event for the base of the Oligocene than the presently accepted foraminifer genus *Hantkenina*. The latter criterion characterizes the IUGS-ratified base Oligocene at the Massignano GSSP (Premoli Silva et al., 1988) and is about half a million years older than the isotope change and the Grande Coupure event. The *Hantkenina* extinction level criterion was chosen to fit as good as possible the base of a classical threefold subdivided Oligocene in which the Rupelian and Chattian were the classical middle and upper/late Oligocene stages of Beyrich (1854, 1856). However as the lower/early Oligocene unit stage in this threefold subdivision was not coherently defined, Tongrian in Belgium, Latdorfian in Germany and Sannoisian in France, the ICS at the IGC of Washington 1989 has then decided (Jenkins & Luterbacher, 1992) to extend the Rupelian stage to the base of the Oligocene defined by the *Hantkenina* extinction hereby installing a two-fold Oligocene subdivision as was proposed earlier by Hardenbol & Berggren (1978). However dinoflagellate studies by Brinkhuis & Visscher (1995) revealed that the Massignano boundary stratotype occurs well below the base of the originally defined Rupelian in the Rupel area in Belgium and also within the top of the classical Priabonian stratotype defined by Hardenbol (1968), and therefore these authors suggested a reappraisal of the upper boundary of the Eocene Series. Defining the base of the Oligocene by the major oxygen isotope event would better correspond to the original meaning of the Rupelian stage sensu Beyrich (1854, 1856), however it would bring the short regional stages Latdorfian, Sannoisian and the marine part of the Tongrian, which are lower Oligocene in the Beyrich (1854) threefold division, back to the upper Eocene. Classifying Latdorfian, Sannoisian and the marine part of the Tongrian in the Eocene has also been proposed by Cavelier & Pomerol (1983, 1986),

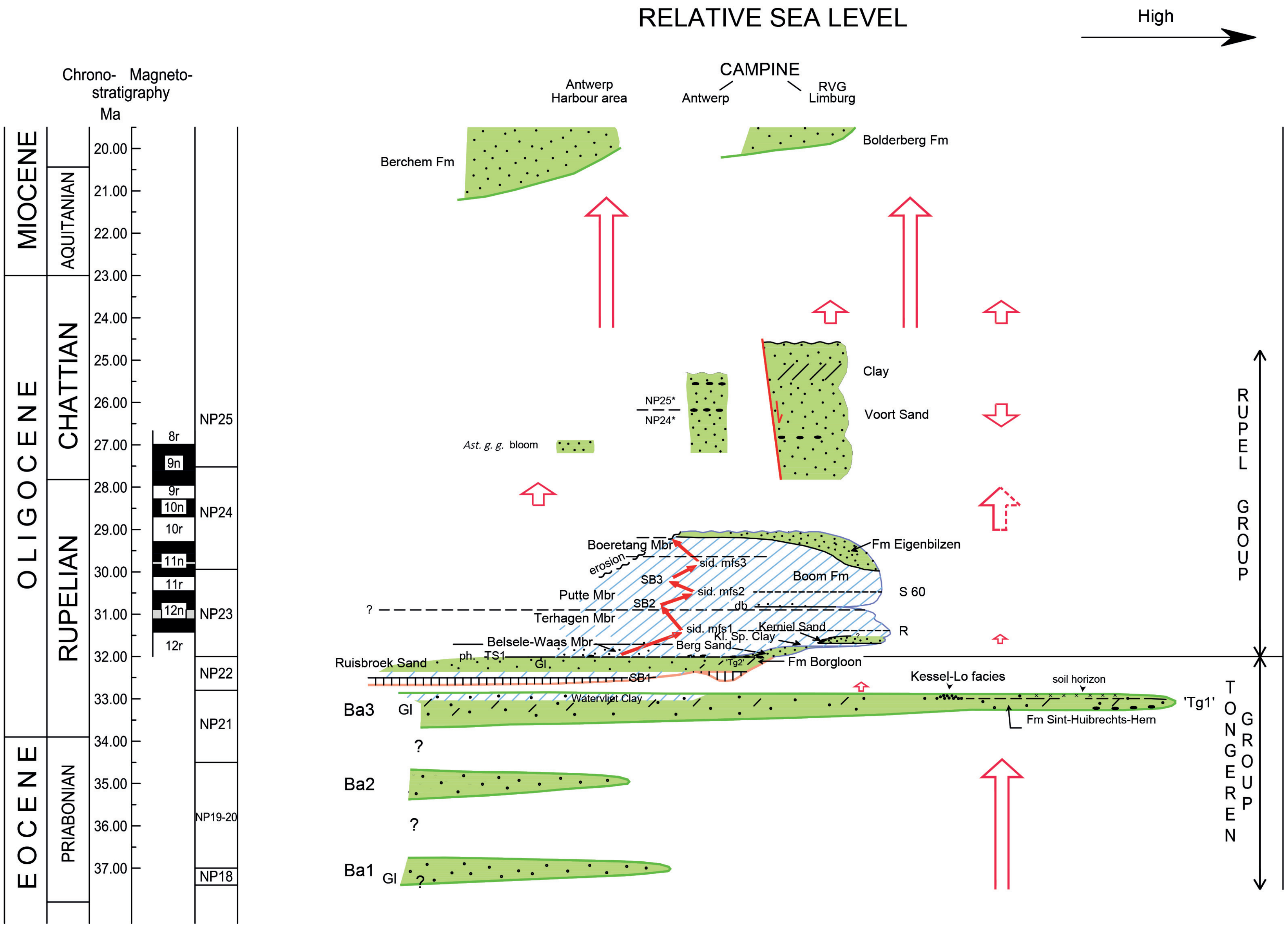

Figure 4. The evolution of the sedimentary and vertical tectonic history in the area from end Eocene to early Neogene. Note that the lithostratigraphic units, grouped in sedimentary sequences, are plotted against the chronostratigraphy and the time scale developed by Coccioni et al. (2016); the latter is differing slightly from the time-scale in Vandenberghe et al. (2012). Ba stands for Bassevelde Sand, Tg for 'Tongrian', ph for phosphate, *Ast. g. g* for the Asterigerina horizon, other abbreviations are explained in the text. The full red arrows in the Boom Clay indicate trends in grain-size evolution and the larger vertical open arrows point to vertical tectonics. Note the discrepancy between the NP24/25 boundary derived by Śliwińska et al. (2014), used as reference to construct the figure (see text), and the same boundary marked by an asterix also considered as proxy in the North Sea for the standard zonation boundary.

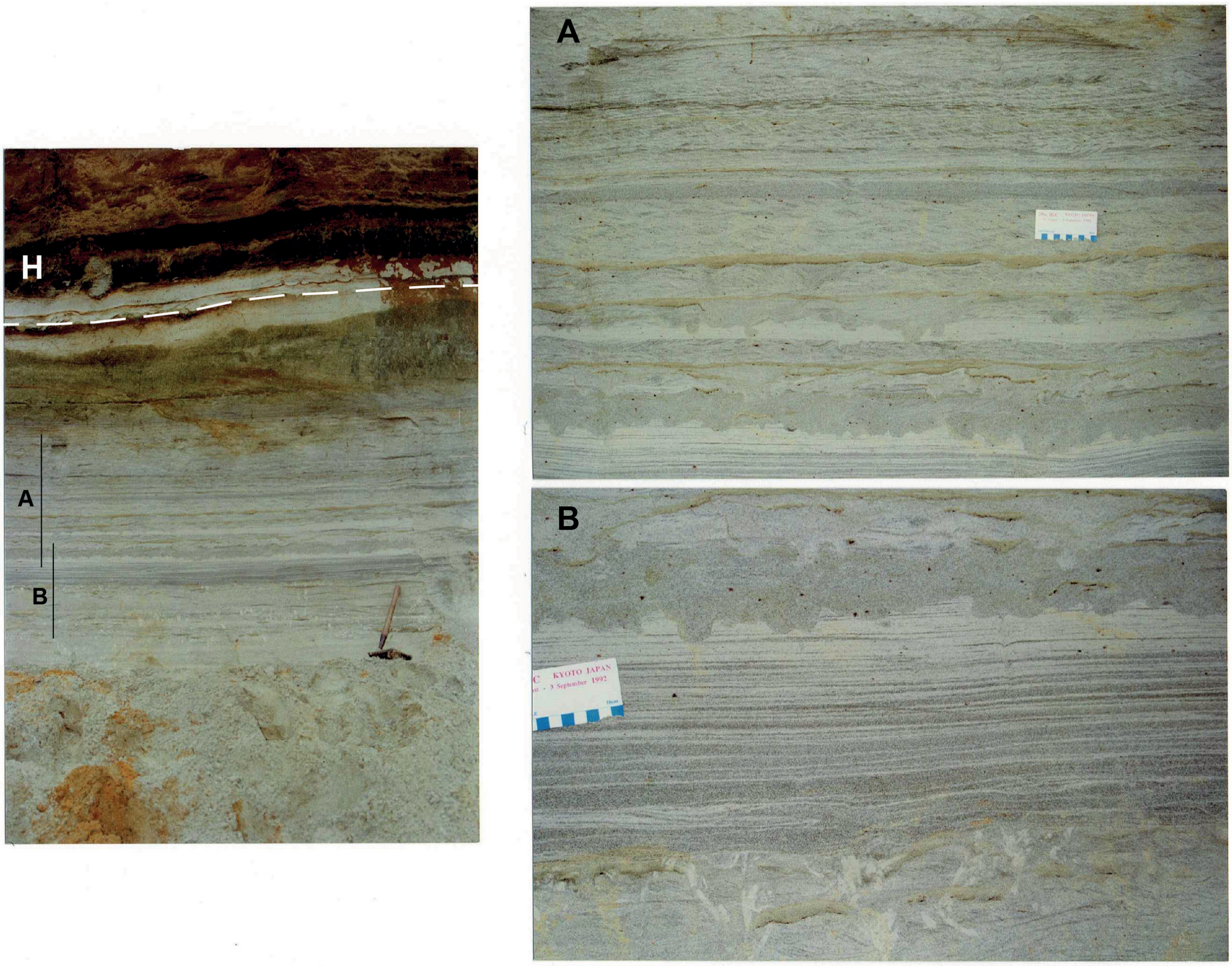

Figure 5. The interrupted white line indicates the limit between the Sint-Huibrechts-Hern and Borgloon Formations (Table 1) in a former sand pit near the Pellenberg academic hospital between Leuven and Tienen (Fig. 1); the green-grey Neerrepen Sand Member of the Sint-Huibrechts-Hern Formation ends at its top in a soil and is overlain by the dark swamp clays of the Hoogbutsel Horizon (H) (Table 1). The A and B intervals in the Neerrepen Sand Member on the photograph are shown in detail and display a succession of decimeter scale master beds consisting of a variety of current, fluidization and bioturbation structures.

Haq & Van Eysinga (1987) and Schuler et al. (1992). However by the choice of the GSSP *Hantkenina* criterion, these short regional stages were intentionally included into the Oligocene, honoring the definition of the Oligocene by Beyrich (1854, 1856) (Vandenberghe et al., 2012). In the detailed North-Sea Basin stratigraphic review (Vinken, 1988) the threefold Oligocene subdivision was maintained (see also Gullentops, 1990). A reintroduction of the threefold Oligocene subdivision by the International Stratigraphic Commission of IUGS could also be a possible scenario (Stover & Hardenbol, 1993; Vandenberghe et al., 2003a).

5. The 'Rupelian' transgression: from coastal plain to marine clay deposits

Over less than 50 km distance, from the Leuven-Tienen area in the south to the Waasland and Rupel area more northwards (Fig. 1), several outcrops have allowed to study and reconstruct the way the raising sea level invaded the land installed after the lowermost short-lived earliest Rupelian sedimentary cycle containing the Sint-Huibrechts-Hern Fm, Bassevelde 3 sand and Watervliet clay Members as discussed above (Table 1) (Vandenberghe et al., 2002, 2003b) (Fig. 6).

These land conditions had lasted only for a short while but are well documented by the existence of an erosion surface and by soil development. North of Antwerp the first deposits above the erosive surface are a thin lagoonal clayey sand with many freshwater algae and containing reworked Lower Cretaceous to Eocene microfossils pointing to erosion of an uplifted Artois area (in Doel 2b borehole, p. 427 in Vandenberghe et al., 2003a). The regionally occurring shallow marine sand of the Ruisbroek Member overlies these lagoonal deposits and also contains many reworked microfossils. The change in microfossils between the earliest Oligocene sediments of the previous cycle and these newly deposited sediments also indicate the important cooling that occurred during the time of the regression and installation of the land surface. As discussed above, this important cooling is global and associated with a considerable sea-level drop of more than 50 m (Miller et al., 2005); therefore the boundary between the earliest Oligocene cycle and the newly deposited sediments of the Ruisbroek Member is suggested to be a global sequence boundary (Vandenberghe et al. 2003a) (Fig. 4). Contemporaneously with this shallow marine Ruisbroek sand Member and laterally towards the continent in the south, occur the coastal plain deposits of the Borgloon Fm, formerly designated in regional literature as the 'continental Tongrian' with symbol Tg2 (Table 1). The Borgloon Fm consists of several lithofacies (Table 1). Above the basal Hoogbutsel swamp clay bed (Fig. 5), corresponding to the Grande Coupure level discussed above, several brackish water facies occur: fossil-rich sand and marl (Boutersem, Alden Biesen Members), a broad sandy estuary (Kerkom Sand) (Fig. 7) and a green clay (Henis Clay). This last Henis clay Member is part of the family of early Oligocene green clays in France, the Isle of Wight and Belgium described by several authors (Jung, 1954; Gabis, 1963; Porrenga, 1968; Thiry & Dupuis, 1998; Thiry et al., 2006). Environmental conditions must have changed quickly in these coastal deposits; this is shown by

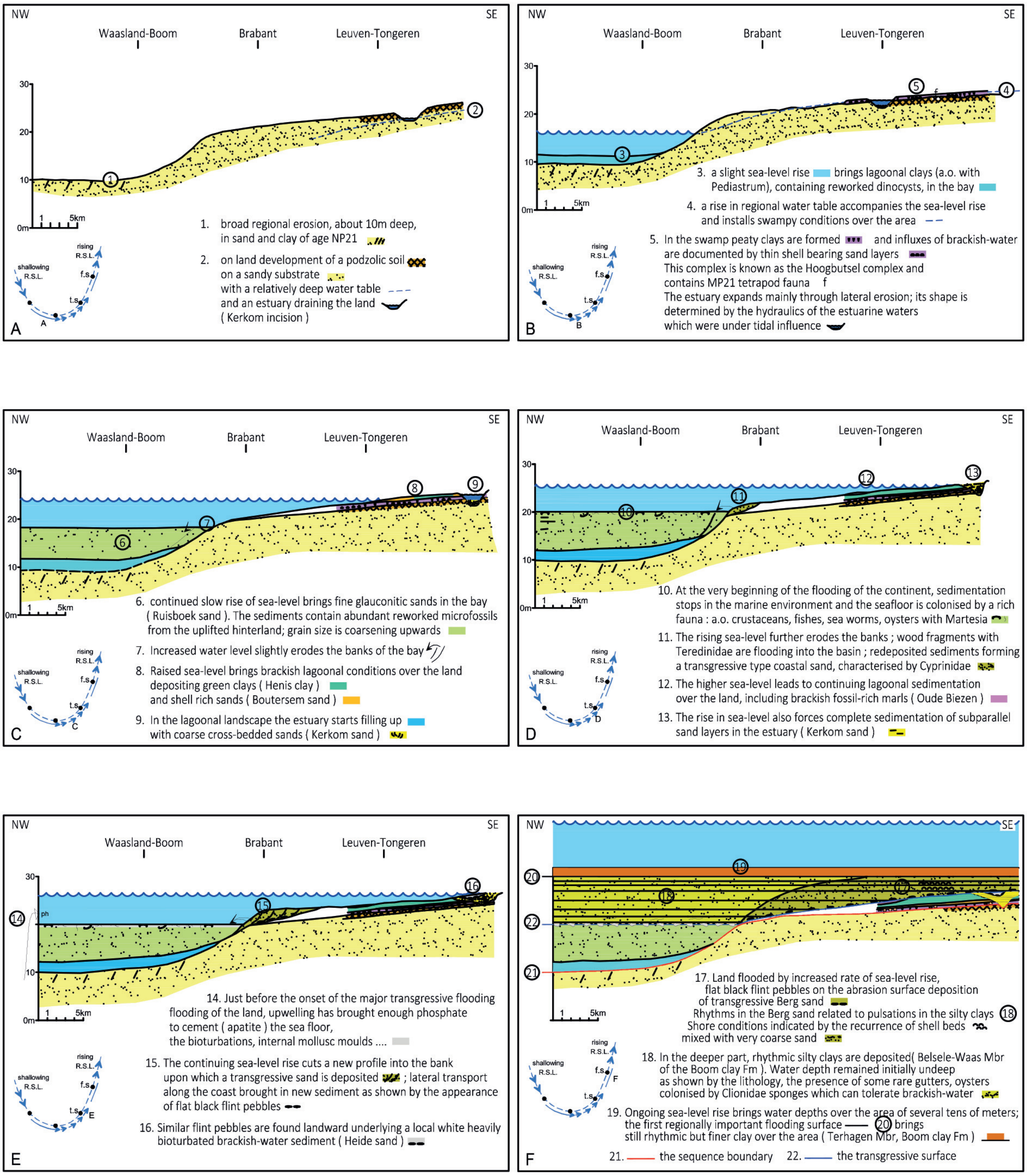

Figure 6. A detailed representation of the successive steps at the start of the main early Rupelian sedimentary sequence, starting in biochron NP22 (see Fig .4), inundating the area from north to south and ultimately bringing Boom Clay over the area (based on Vandenberghe et al. , 2002)

the isotope indications of cooling (De Man et al., 2004) and the influx of cool microfossils at the base of the Ruisbroek Sand (De Coninck, 1999) occurring together with the presence of subtropical and humid environment pollen and spores in the Henis Clay (Roche & Schuler 1976, 1979, 1980), also by the presence of small amounts of the mild-aridity indicator mineral sepiolite in the swamp and lagoonal clays (Huggett et al., 1996; Verbeeck et al., 1998) and by the sedimentological and paleontological interpretation of variable fresh, brackish and marine water conditions (references in Vandenberghe et al., 1998, 2003a). Also in western Europe between the Atlantic and the Tethys realms, terrestrial stable isotope data at the Eocene-Oligocene transition show more variable temperature and humidity conditions explained by changing air trajectories influenced by the rise of the Alpine-Dinaride mountain range (Kocsis et al., 2014). Walliser et al. (2017) have also linked the uplift of the Alps to the pattern change in decade range climate fluctuations in Central Europe from a southern ocean influenced pattern in the Eocene to an Atlantic type variability from the Oligocene onwards.

Within these coastal sediments of the Borgloon Formation a brown organic horizon is present (Fig. 7) described already by Van den Broeck (1883) as altered algae, briefly described as 'sables chocolatés' overlain by a thin soil by Glibert & de Heinzelin (1954), considered as a giant podzol by Gullentops in Gullentops et al. (1988) and challenged by van Riessen &

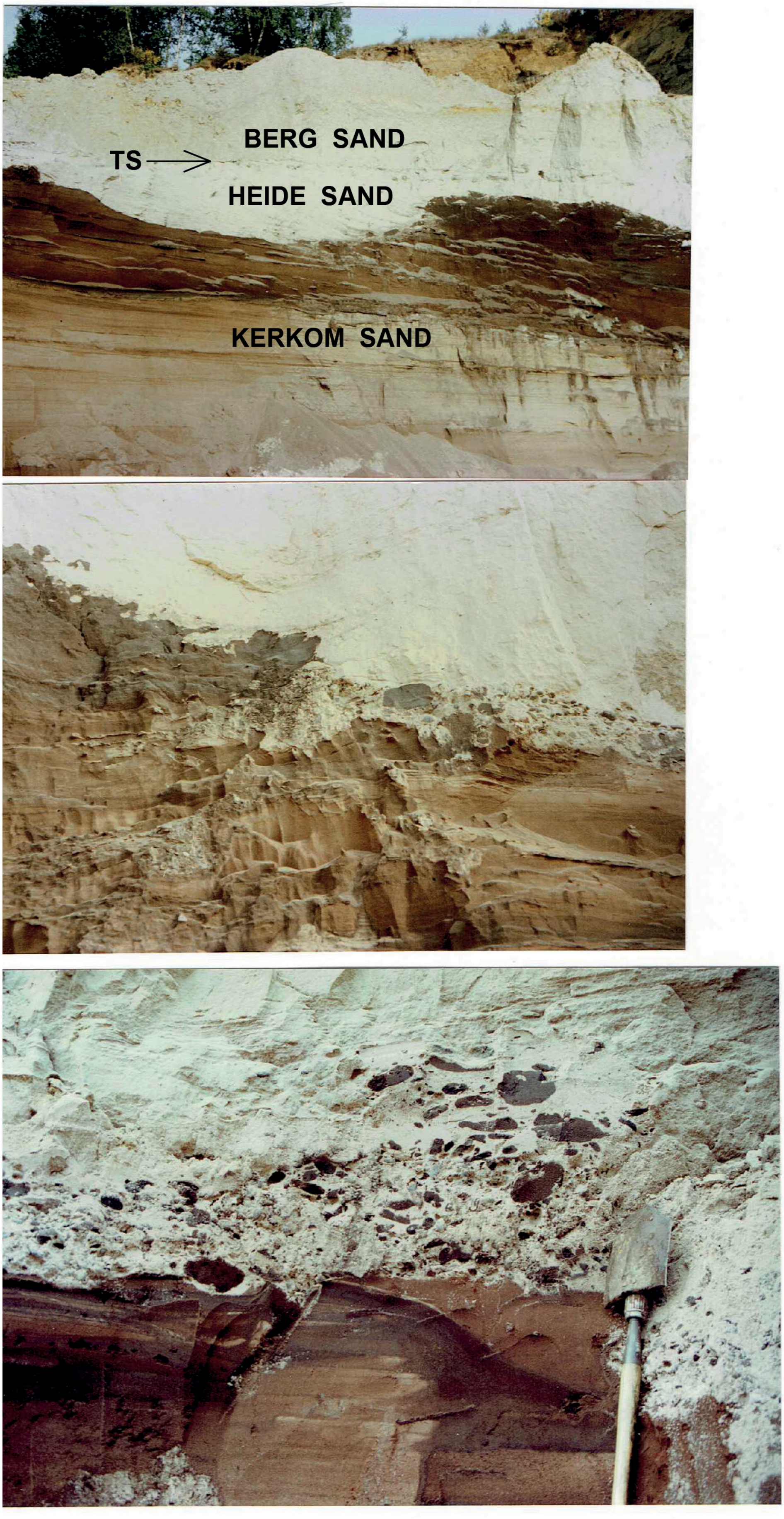

Figure 7. The shallow marine transgressive Berg Sand overlying along a transgressive surface (TS) the continental Heide Sand and Kerkom Sand (Table 1). The Heide Sand fills an erosive topography on top of the 'sables chocolatés' which originally cemented the top of the Kerkom Sand. The middle and lower photographs show details of the contact between the 'chocolate' colored Kerkom Sand and the overlying Heide Sand illustrating an erosion process that can only occur at a well cemented channel floor. Sand Pit Roelants, Pellenberg-Lubbeek.

Vandenberghe (1996) who argued that it is the remnant of an early Oligocene oil seepage that was spilled by tilting as the consequence of the Pyrenean tectonic pulse, hardened at the surface (Fig. 7) and eroded soon afterwards, and at present intensively leached by the infiltrating ground water. Several authors however insist on a soil origin for this brown layer (Buurman et al., 1998; Van Herreweghe et al., 2003; van Riessen & Vandenberghe, 1999).

All these coastal plain deposits in the south are covered by the same transgressive shallow marine sand unit with some shell beds in its basal part and known as Berg Sand in Belgium (Fig. 7) and the Netherlands and as Walsumer Sand in Germany. At

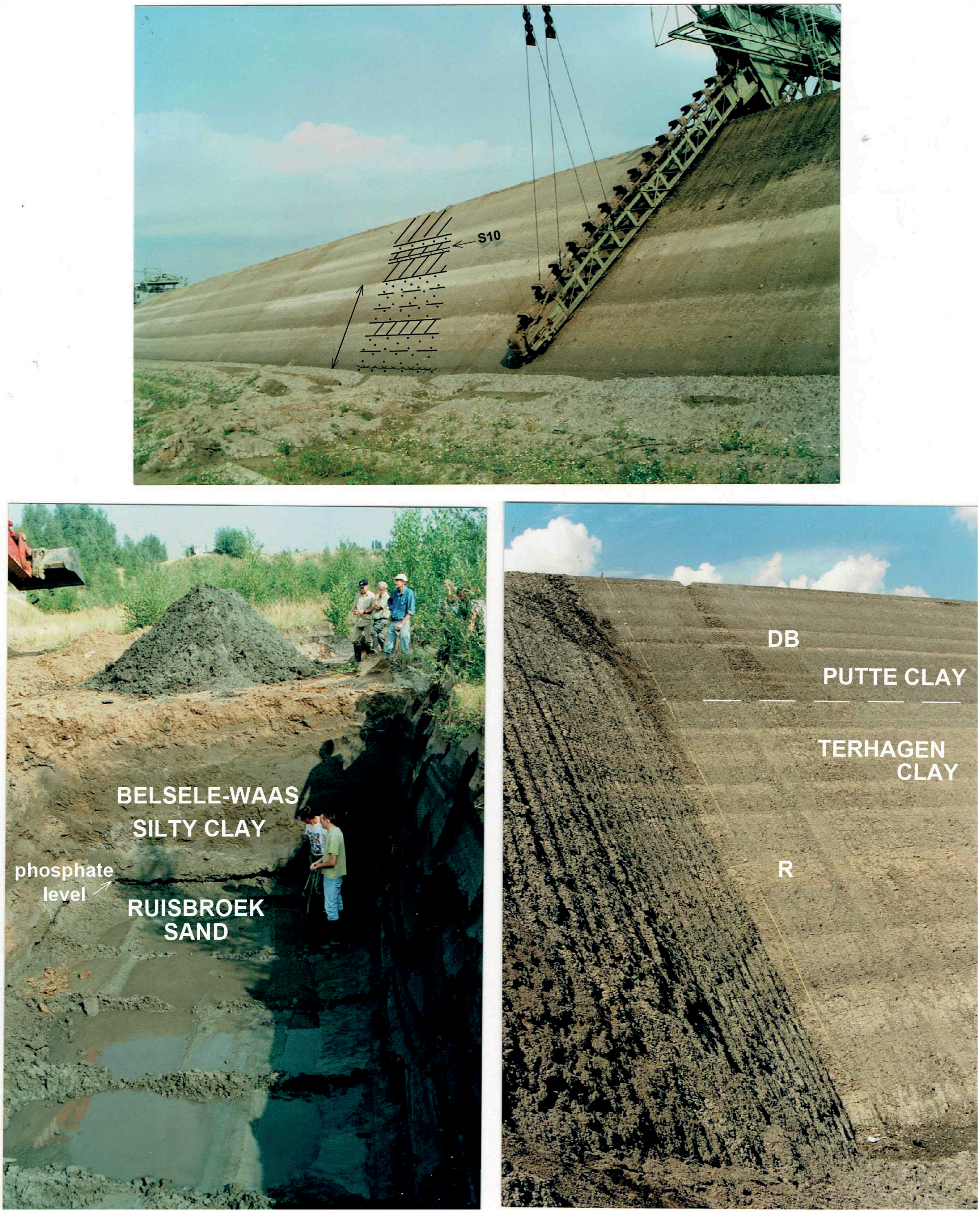

Figure 8. Upper photograph: aside the arrow occurs the Belsele-Waas Member of the Boom Clay Formation (SVK pit in Sint-Niklaas, Fig.1) consisting in the lower half of the section of 2 composite (interrupted lines) thick silt layers (stippled) and overlain by the Terhagen Member of the Boom Clay showing an alternation of clay-rich (hatched) and silt-rich (stippled) layers. S10 indicates the position of the septaria horizon; the clay layer containing S10 is paleontologically shown to correspond to the flooding event bringing Boom Clay over Berg Sand in the south (see text). Note at the very base of the clay face the level with shallow erosive features interpreted as a wave erosion level (see also Fig. 10).
Lower left photograph: an excavation below the base of the SVK pit in Sint-Niklaas (Fig. 1) showing the level of the phosphate bed in the top of the Ruisbroek Sand underlying the Belsele-Waas clay Member (Table 1).
Lower right photograph: a typical image of the Boom Clay showing an alternation of clay-rich and silt-rich layers; the interrupted white line marks the boundary between the pale grey Terhagen Member and the dark grey Putte Member of the Boom clay Formation (DB stands for double band and R for pinkish layer) (Steendorp clay pit).

its base occurs a horizon with typical flat black flint pebbles and some streaks of purple sand ; the latter is eroded from the tops of the 'sables chocolatés' which were sufficiently solid at that time to stand in relief (Fig. 7). In fact this chocolate-brown horizon, occurring at most a couple of meters below the base of the Berg Sand, was shortly after its formation already hard to become eroded by the arrival of new a coastal sand (Heide Sand) transporting also the first black flint pebbles (Fig. 7).

The arrival of flint pebbles indicates a new phase of intense erosion of chalk cliffs existing in the west; the base gravel of the Sint-Huibrechts-Hern Formation, the earliest Oligocene sequence overlying the Pyrenenan unconformity surface, still consists of Paleozoic siliciclastic rocks derived from the hinterland in the south and southeast. The base of the Berg Sand Member is a good example of a transgressive surface (Figs 4, 6, 7) (Vandenberghe et al., 1998). It occurs at the turn of the nannoplankton NP22 to 23 biozones (Fig. 4).

At exactly the same biozone turnover more northwards in the direction of the basin, the sedimentation of the shallow marine Ruisbroek Sand is halted and a fauna develops on the sea bottom which is then impregnated by apatite (Fig. 6); the excess phosphorus is probably brought by the upwelling water accompanying the transgression (Vandenberghe et al., 2002). Above the transgressive surface, marked by a concentration of the slightly reshuffled phosphate-impregnated mollusc molds and bioturbation traces with some sparse black flint pebbles, occurs a very silty and heavily bioturbated clay, the Belsele-Waas Member of the Boom Clay (Table 1; Fig. 8), which is time equivalent with the shallow marine Berg Sand (Fig. 7) in the south.

Land emerged after the short earliest Oligocene sedimentary sequence and upon resumption of the flooding of the land facies were deposited during the initial slow sea-level rise; the transgressive surface above these lowstand systems tract (LST) deposits, corresponding to the base of the Berg Sand, marks the important increase in rate-of-rise of the sea level (Vandenberghe et al., 1998, 2004) (Fig. 6). The Berg Sand above the transgressive surface (Fig. 7) is fining upwards reflecting the increasing water depth until a sudden and even more pronounced deepening of the basin occurs (flooding surface in sequence stratigraphic terminology) starting the Boom Clay deposition over the area. This same flooding surface event in the southern coastal area can also be observed in the grain-size evolution within the Boom clay deeper in the basin in the north (Fig. 4), a correlation which is confirmed by paleontological correlation data (dinoflagellates, molluscs, fish otoliths, see Vandenberghe et al., 2001 p. 75); this flooding can also be followed in the subsurface as a sharp resistivity log signature drop at the top of the silty Belsele-Waas Member (Figs 8, 12) (Vandenberghe et al. 2001). In northeast Belgium and the adjacent Lower-Rhine area in Germany, the base of the clay sedimentation above the Berg (Walsumer) Sand is the Kleine-Spouwen or *Nucula comta* clay (in Germany named Ratinger Ton or Tonmergel); it is separated from the main clay mass of the Boom Clay Formation above it by a sand wedge, the Kerniel Sand, which thickens towards the Lower Rhine area. To the west, the Kleine-Spouwen Clay is time equivalent of the clay interval occurring just on top of the silty Belsele-Waas Clay and containing the septaria horizon S10 (Fig. 8) (references in Vandenberghe et al., 1998). Van den Broeck (1883) had already identified and mapped the succession of the Berg, Nucula comta Clay, Kerniel Sand and Boom Clay in his geological studies of the Tongeren area (Table 1).

These successions of lithofacies at the beginning of the Rupelian transgression in eastern Brabant and in the type area more northwards in the Waasland-Antwerp-Boom area can easily be understood as a backstepping of facies responding to the southwards shift of the coastline (Vandenberghe et al., 2003b) (Fig. 6).

6. High-frequency cyclic sedimentation in the Boom Clay Formation

6.1. Lithology and high-frequency cycles

The sections of Boom Clay in the pits excavated by a digging ladder display a remarkable layered pattern of alternating bands expressed by different shades of gray and several tens of cm thick (Vandenberghe, 1978; Vandenberghe et al., 2014) (Fig. 8). The most regularly alternating darker and paler grey layers represent changes in the concentration of silt-sized particles, i.e. a varying ratio of clay-fraction/silt-plus-very-fine-sand content; in addition, while the very-fine-sand coarsest particle size itself remains constant in all layers, the distribution of the fine sand and sortable silt population systematically and gradually changes its skewness (Vandenberghe, 1978). Such evolution of the grain-size characteristics shows that the same sediment continuously arrived in the basin but was periodically sorted and winnowed with a gradually evolving intensity.

Black clay layers of several cm thick are another striking feature of the Boom Clay. The Putte Member of the Formation appears in outcrops as a black clay because the systematic presence of such black layers just above each silty clay layer at the base of the overlying clay-rich layer (Fig.8). The black color is due to a few weight percent of land-derived Oligocene phytoclasts (Vandenberghe, 1976). In some instances large flattened tree trunks have been observed (Fig. 9).

Pale grey horizons of some tens of cm thick are carbonate-rich horizons that occur independently of the grain-size and black organic matter layering. The originally marly clay layers have now evolved into septaria horizons, so typical for the

Figure 9. A black heavily pyrite impregnated and completely flattened tree trunk swept in the basin observed in the middle of the Boom Clay section (Terhagen clay pit).

Boom Clay as to become named Septarienton in Germany.

6.2. Sea-level model for the genesis of the high-frequency cycles

The layering observed in the Boom Clay has a regionally very widespread and regular occurrence. West of the Mol-Rauw fault all layers observed in outcrop and additionally in the subsurface, can be correlated one by one; only in a direction from the depositional center in the north to the coastal area in the south marked thickness differences in the individual layers are observed, and towards the coast clay-enriched layers can even become absent and silt-enriched layers stacked on top of each other as is the case in the Belsele-Waas Member (Fig. 8) (Van Echelpoel, 1991; Vandenberghe et al., 2001, 2014). East of the Mol Rauw fault (Fig. 1) the correlation is still possible for the lower part of the Boom Clay, even as far as the Achterhoek in The Netherlands and the Lower Rhine area in Germany (Vandenberghe et al., 2001). Therefore the mechanism driving the cyclicity must have had a wide areal impact and consequently wave turbulence is a prime candidate mechanism for the observed grain-size sorting in the layers. This mechanism is supported by the presence of thinner levels with flat wave turbulence structures in the most silty layers near the base of the Boom Clay that can be correlated between clay pits (Fig. 10 and photographs in Vandenberghe, 1978). The systematic relationship between the grain-size sorting and the land-derived plant particles in the black layers (Fig.11) strongly suggests that varying water depths are controlling the wave turbulence reaching the sea bottom or not. Indeed the alternative, periodically changing the average wavelength to increase the depth of the wave turbulence, would not explain the observed coupling with the organic matter layering. Periodically changing the water depth will also influence the land area along the basin rim resulting precisely in the observed geometrical relationship between land-plant detritus, the black layers, and the grain-size sorted silt and clay layers (Vandenberghe et al., 2014). Indeed, at the lowest relative sea-level position, turbulence can maximally sort the arriving sediment and the middle part of a silt layer is formed; slow sea-level rise will gradually lose its sorting capacity and upon the switch to a rapid rise the silt layer changes into a clay layer with minimal sorting but now also the land around the basin becomes rapidly inundated and the plant cover is destroyed delivering the phytoclasts to the basin thereby appearing as a black layer at the base of the clay-rich layer. When the sea level approaches its highest level with minimal sorting in the middle of the clay-rich layer, the transgression over the land stops and no more plant cover is destroyed marking also the top of the black layer. A new silt layer starts when the rate of sea-level lowering increases and the wave turbulence can again reach the bottom of the sea to efficiently sort the sediment (Fig. 11). The slightly increased kaolinite content in the clay-enriched layers compared to the adjacent silt-enriched layers (Vandenberghe & Laenen, 1999; Zeelmaekers, 2011) is explained by the more intense erosion of the top soils in the northern provenance areas during the high sea-level phases (Fig. 11 inset map). The very constant mineralogy of the Boom Clay (Zeelmaekers et al., 2015), reflecting the varying silt/clay fraction ratio, is supporting the continuous arrival of the same sediment in the basin. Note that the position of the carbonate layers bears no relationship to the position of the other lithological components; the precise reason for the development of carbonate horizons in the clay during the sedimentation is not yet understood.

6.3. The climatic beat of the high-frequency sea-level cycles

But which was the mechanism driving the cyclic sea-level variations? Tectonic periodicity does not seem to be involved in the formation of these high-frequency cycles because of the very regular layering over a wide area and the absence of local erosional features or changes in provenance of the sediment.

A Milankovitch-driven climatic periodicity on the other hand has been proven by spectral analysis of the cyclicity (Van Echelpoel 1991; Van Echelpoel & Weedon, 1990; Abels et al., 2007). The study by Abels et al. (2007) benefited from the most recent detailed biostratigraphic control and concluded that the major beat was the 40 ka orbital obliquity. This is in line with ODP (Ocean Drilling Program) studies (Wade & Pälike, 2004) demonstrating the obliquity beat during the Oligocene and estimating the eustatic sea-level fluctuations in the order of several tens of meter, undoubtedly related to the waxing and waning of the large Antarctic ice-caps that formed shortly after the start of the Rupelian as shown by the large oxygen isotope jump. Stable oxygen isotope data (De Man 2006a; Vandenberghe et al., 2014) show that the Eocene water temperatures of 20 to 25 °C start to drop during the Priabonian to just below 20 °C, coinciding with the last appearance of the Nummulites spp in the North-Sea Basin, and to further drop during the Eocene-Oligocene transition and reach around 10 °C during the upper Rupelian, compared to the 5 °C of present-day North Sea bottom waters. The temperature seasonality dropped during the Rupelian, in the Boom Clay and the North Sea in general, to very low values of just a few degrees. This is in contrast with Rupelian data in the Gulf Coast area (Ivany et al., 2000), a difference

Figure 10. Flat gully type structure in the silty parts of the Belsele-Waas Member interpreted as the effect of erosion and sorting of the sediment by wave turbulence acting on the sea bottom; individual levels with such turbulence features extend regionally and can be correlated between clay pits.

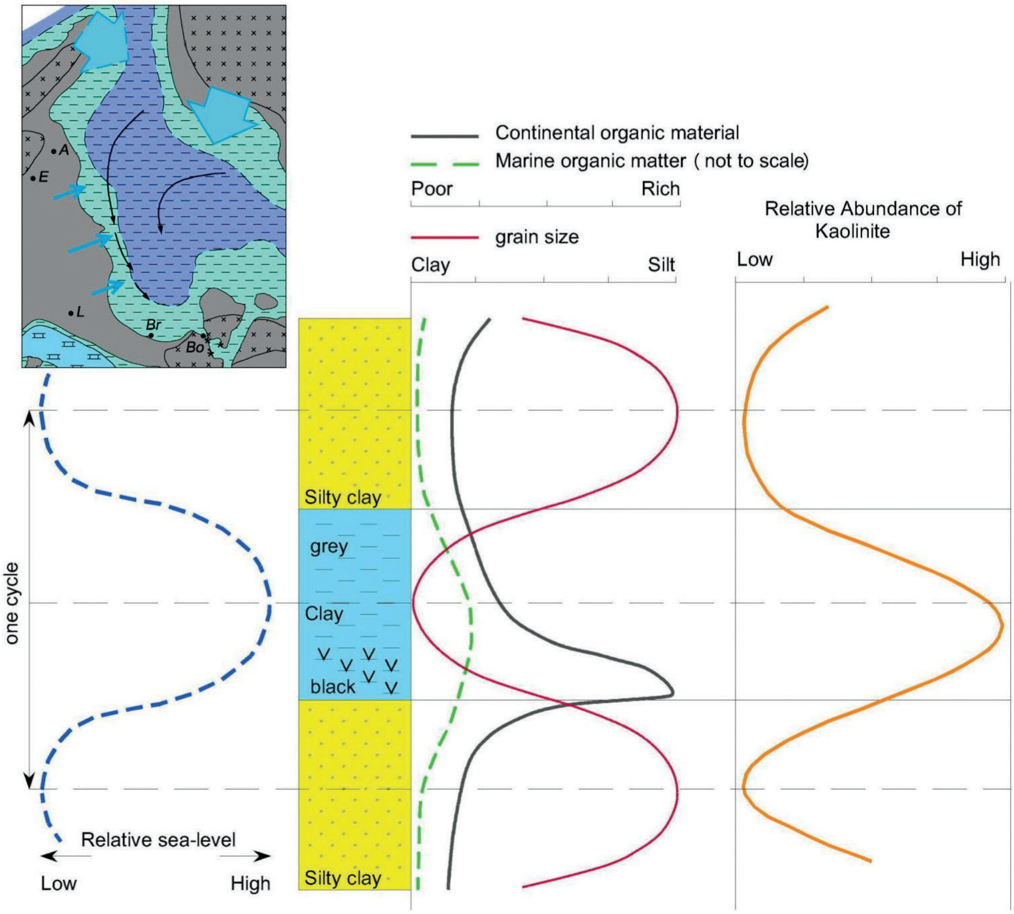

Figure 11. High-frequency cycle or parasequence model for the alternating silt-rich and clay-rich layering in the Boom Clay. All variations shown are relative and not to scale.
Inset map showing the provenance areas (blue arrows) to the North Sea Basin during the Rupelian (A Aberdeen, E Edingburgh, L London, Br Brussels, Bo Bonn). Crossed areas are pre-Mesozoic cratons, the greenish border of the North Sea represents deposition of shallow marine siliciclastics, the blue horizontally hatched area represents deeper-marine clay deposition and the blue area south of the uplifted Artois-Brabant axis are carbonate containing shallow marine deposits.

tentatively explained by De Man (2006a) as the influence of the newly installed Gulf Stream current gyres. Also Cramer et al. (2009) have suggested changes in wind-driven ocean circulation patterns since the Eocene-Oligocene transition to explain the large difference in benthic foraminifera oxygen isotopic compositions in the ocean basins since that time.

On top of the obliquity cycles, the spectral analysis of the Boom Clay section shows also short and long eccentricity signals; a remarkable result of this analysis is the presence of asymmetric evolutions at the low eccentricity pace in the resistivity log data which are the proxy for grain-size reflecting the climatic evolution. Such asymmetric climatic cycle pattern resembles very closely the climatic evolution during Pleistocene glacial-interglacial cycles (fig. 6 in Abels et al., 2007).

7. Lower-frequency cycles in the Boom Clay Formation

7.1. Low frequency grain-size cycles

The long eccentricity 400 ka beat is becoming the stable metronome for the Cenozoic time scale (Vandenberghe et al., 2012). Abels et al. (2007) identified in the Boom Clay this long eccentricity 400 ka signal but could not tune it exactly to the Laskar et al. (2004) reference solution. Laenen (1997) (see also Vandenberghe et al., 1997) has recognized intermediate cycles of about 800 to 1000 ka duration in the stable C and O isotope contents and which are supported by small variations in clay mineral composition and chemical weathering indexes. These cycles have been interpreted as intermediate sea-level variations but their precise significance has remained unclear.

The presence of 2 lower-frequency cycles as an envelope of the high frequency grain-size cycles in the outcrop area was recognized in the early sedimentological study by Vandenberghe (1974, 1978). Vandenberghe & Van Echelpoel (1987), Stover & Hardenbol (1993), Vandenberghe et al. (1998) have associated these low frequency grain-size cycles with global so-called 3th order sea-level sequences. The 2 cycles identified in the outcrop area have been used in the Haq et al. (1987, 1988) global sequence chart for the Rupelian. Unfortunately the biostratigraphic relationship between the outcrop area and subsurface Rupel Group to the north was poorly understood at that time and became only clear with geophysical well log studies and further biostratigraphic work (Vandenberghe et al., 2001, Van Simaeys, 2004; De Man, 2006a). It appeared that the sections in the outcrop area only represent about half of the total global Rupelian time span and in the complete subsurface Rupelian Boom Clay section, now 4 low-frequency cycles are recognized (Vandenberghe et al., 2001) (Fig. 12). The lower-frequency cycles are expressed as grain-size variations, and consequently by resistivity cycles on the logs, and the change-over points in grain-size evolution trend are also expressed by some peculiar layers observed in field sections such as the pink to brownish colored layer (R layer), two successive very silty horizons (DB or double layer) and two sideritic septaria levels (SID in Fig. 12). The carbonate in most septaria is ferroan calcite and the presence of considerable siderite in the two levels, interpreted as main flooding surfaces (mfs's) (Fig.12), is attributed to particular slow sedimentation rates at these levels (Laenen & De Craen, 2004). The relationship of these 4 sea-level sequences in the Boom Clay with the three Rupelian sequences on the European Sedimentary Basin Sequence Chronostratigraphic Framework chart in Hardenbol et al. (1998, chart 1) remains unclear at present.

7.2. The lower frequency cycles: climatic or tectonic signals?

The argument to equate the lower frequency cycles with sea-level fluctuations has only been based on their expression by grain-size variations comparable to the grain-size variations in the high-frequency cycles which are of a well understood glacio-eustatic origin. This was further corroborated by the estimated time duration of the low-frequency cycles which is in the order of the duration of global 3th order sea-level fluctuations. They were also taken up in the Haq et al. (1987, 1988) chart as model for the Rupelian.

Also it was believed that tectonic influence was expressed in another way, namely by the thickness variation in the silt-clay couples; these couples take the same time to form as

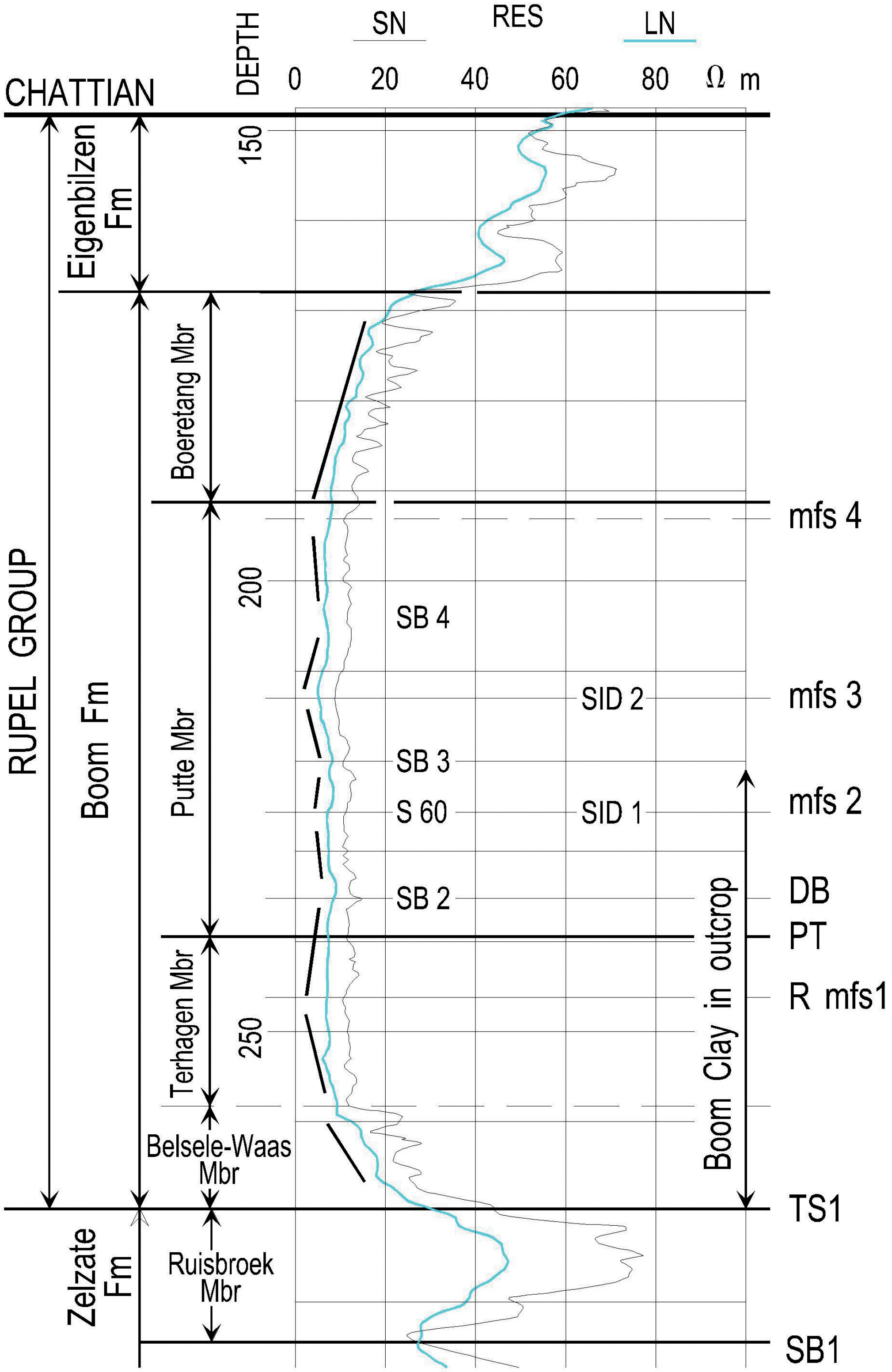

Figure 12. Trends in the well log resistivity curves (short SN and long LN normal) illustrating the presence of 4 lower frequency cycles (heavy black line tracks) and the position of their boundaries (SB1,2,3,4) and main flooding surfaces (mfs 1,2,3,4); note that in the outcrop area only two such cycles are present. The presence of one long trend in resistivity (blue line) is interpreted as part of a major transgression-regression cycle spanning the Rupelian (see also Vandenberghe et al., 2012). The log is taken from the Herentals area (Fig. 1). The subdivisions at the left side are the lithostratigraphic units in the Rupel Group (details at http://ncs.naturalsciences.be/paleogene-neogene/paleogene-lithostratigraphy).

they are controlled by a Milankovitch orbital beat. The fairly constant silt/clay thickness ratio in the high-frequency cycles over the whole section (Van Echelpoel, 1991; Vandenberghe & Mertens, 2013) points to a similar turbulence duration during all cycles and therefore total cycle thickness variations, and also clay bed thickness variations, can only be due to variations in sediment supply. This high-frequency cycle thickness curve (Fig. 13) can be described as one long term variation with high sediment supply at the base, followed by lower sediment supply, lowest at the DB layer, and back to very high sediment supply towards the top of the Boom Clay; note that this trend in sediment supply is similar to the long term resistivity curve for the total Rupel Group section (Fig. 12). The amount of sediment supply is considered to be directly related to the amount of tectonic uplift in the provenance area around the Rupelian North-Sea Basin (Fig. 11 inset); however in the sequence stratigraphic interpretation (Stover & Hardenbol, 1993; Vandenberghe et al., 1998, 2004) the DB layer level (Fig. 13) is also interpreted as the expression of a glacio-eustatic sea-level low, occurring at a sequence boundary, and in addition the long eccentricity orbital signal also shows a maximum at the DB level (Abels et al., 2007). Apparently the origin of the very silty DB level can be interpreted in terms of a climatic as well as a tectonic evolution. Therefore the question arises how distinguish a climatic from a tectonic signal.

Thicknesses of all individual layers in the Boom Clay have been compared with each other using geophysical borehole logs of 17 boreholes calibrated to cores in 6 of these boreholes (Mertens, 2005; Vandenberghe & Mertens, 2013). To improve the reliability of the results, the thicknesses of each layer were compared to and subtracted from the thickness of the corresponding layer in a reference well and the cumulated

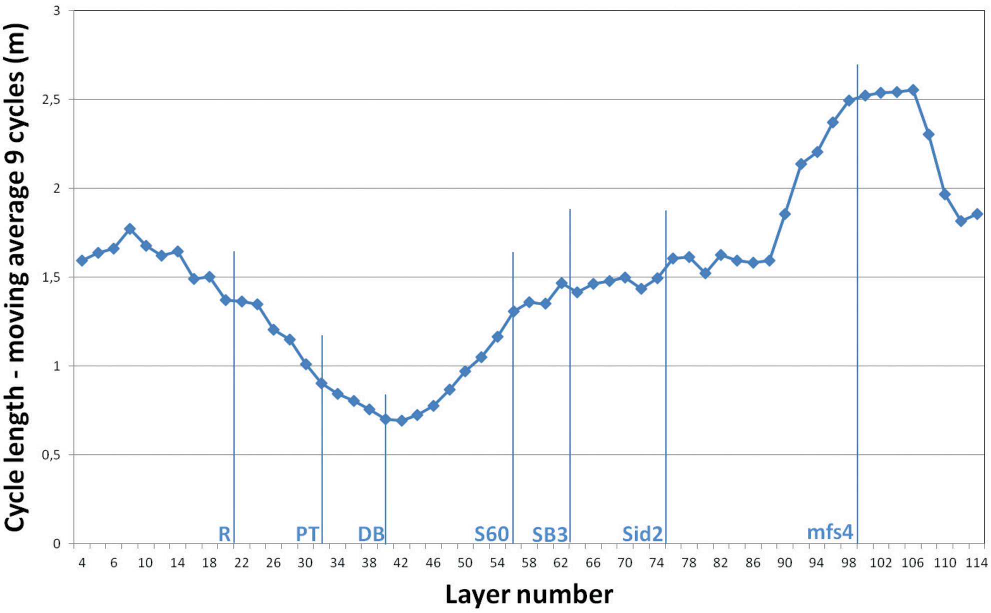

Figure 13. The evolution of the length of the high-frequency cycles, represented by the sediment thickness, in the Boom Clay Fm from base (left) to top (right), representing the sediment delivery rate to the basin (from Vandenberghe & Mertens, 2013); the position of the stratigraphic marker levels (R, PT,….) in the usual vertical lithostratigraphic column of the Boom Clay can be found in Figure 12.

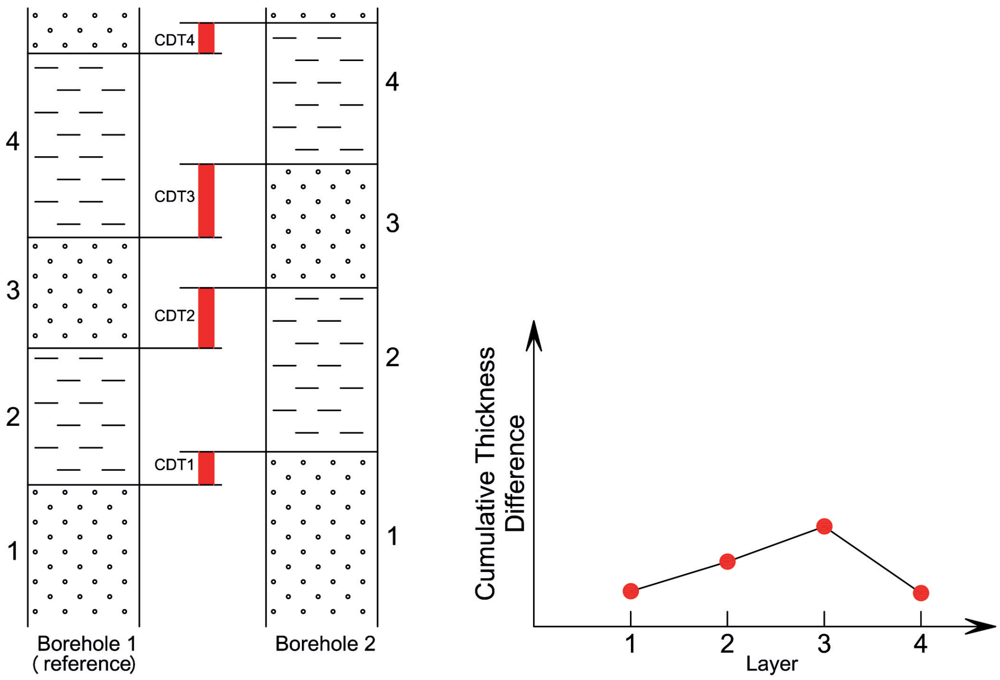

Figure 14. Graphic presentation of the procedure for comparing thicknesses of the same individual layers in the Boom Clay; cumulative thickness differences are measured between the same individual layers in a borehole section and in a reference borehole; the cumulative thickness values are plotted in a stratigraphic order. This procedure eliminates the eustatic component and changes in gradient or reversal of the gradient point to changes in the sediment delivery rate to the basin (from Mertens, 2005).

thickness differences of individual layers above the base of the Boom Clay were compared (Fig. 14). In particular the subtraction of the thicknesses eliminates any eustatic component and only reflects sediment supply and therefore the local tectonic intensity. Tectonic intensity changes are reflected by changes in the gradient of the cumulative curve for each borehole and the location of tectonic compartments in the basin is shown by mapping cumulative differences in all boreholes at a particular level in the clay. Surprisingly the cumulative curves show gradient changes at the levels which were interpreted as glacio-eustatic 3[th] order sequences, namely the SB's like the DB and the mfs's (R, siderite horizon S60, base Boeretang Member) of the 4 sequences identified (see levels in Figs 12, 13) (Vandenberghe & Mertens, 2013; Vandenberghe et al., 2014). However as they cannot be eustatic, these low-frequency cycles must be relative sea-level variations induced by local tectonics. Mapping the evolving uplift-subsidence development delineates different areal compartments in the basin; the history of the uplifting southwestern part is well known in detail by previous studies and will be briefly discussed hereafter.

8. Tectonic activity during the Rupelian and Chattian

8.1. End Rupelian tilting

Independent geological, geophysical and geotechnical evaluations of the Boom Clay pre-consolidation state at a site along the Scheldt river northwest of Antwerp consistently estimated a maximal loading of the clay equivalent to 80 to 90 m; a geological evaluation concluded that this thickness could have been eroded from the Boom Clay itself during the hiatus stretching from end Rupelian to early Burdigalian (Schittekat et al. 1983). In fact the geometric reconstruction of the Boom Clay geometry by correcting for post-Rupelian subsidence of North Belgium together with the micropaleontologically proven presence of an early Chattian relict on top of the eroded Boom Clay in the Antwerp harbor area, confirmed the erosion of about 80 m Boom clay at the end Rupelian and early Chattian (Vandenberghe & Laga, 1986; Vandenberghe et al., 2003b). In boreholes in the northern Campine, Rupelian sections are more complete and a sudden influx of reworked silicified Upper Cretaceous microfossils identified in these boreholes close towards the end of the Rupelian (Van Simaeys et al., 2004) is considered to coincide with the relative tilting of the Campine and the erosion to the south and southwest (De Man et al., 2010).

This subsidence of northeast Belgium compared to the west and southwest and the subsequent bending of the Boom Clay mass is invoked to explain the pattern of consistently oriented vertical joints, observed in the uplifted and now outcropping Boom Clay and also to explain the presence of the meter-scale Kruibeke fault zone, unique in the outcrop area and related to a flexure in the deep Caledonian Massif (Mertens et al., 2003; Dehandschutter et al., 2004, 2005; fig. 6 in Dusar & Lagrou, 2007). The consistent joint orientation in the outcrop area points to the influence of a regional stress state, with the maximal principal stress in the horizontal plain and oriented NW-SE, the same orientation as for the normal fault observed in the clay pit at Kruibeke near Antwerp with a NE down-throw. The eroded Boom Clay top surface above the Kruibeke fault zone is not displaced. The joint development also influenced septaria while these were still slightly soft (Mertens et al., 2003). As eroded and perforated fragments of hard septaria occur at the base of the Neogene (Burdigalian) cover overlying the Boom Clay, the jointing is pre-Neogene and most logically related to the end-Rupelian uplift and strong erosion discussed above. Dehandschutter et al. (2004, 2005) have linked the common small striated shear surfaces, the joints and the Kruibeke fault to compactional volume reduction in the context of the minor external tectonic tension caused by the Roer Valley Graben (RVG) development (Fig. 1). NE-SW minimal horizontal stress in the area exists already since the Oligocene (Bergerat & Vandycke, 1994). Note that the upward bulging of the Boom Clay, only observed in the Scheldt river (Wartel, 1980; Henriet et al., 1986) and along its bank (Laga, 1966) is a more recent unloading phenomenon related to the geomorphological evolution of the river valley.

8.2. Uplift and the Roer Valley Graben development during the Chattian

The Rupelian to Chattian transition coincides with the start of renewed tectonic activity in the North-Sea area (de Jager, 2007). On seismic sections in the Belgian eastern Campine the Rupelian-Chattian boundary is expressed as a small angle unconformity. In the western Campine, uplift continued during the whole Late Oligocene until the Miocene and sedimentation only resumed during the Burdigalian. This long tectonic inversion phase is the Savian inversion phase and it is a remote effect of the Alpine tectonic activity. In the Condroz area, north of the Ardennes in south Belgium, the earliest Rupelian sand at the base of large karsts in the Paleozoic limestones, already described by Van den Broeck et al. (1910), is overlain by Miocene clastic fill; logically therefore the important uplift of the Condroz with the lowering of the water table necessary for karst formation can be linked to the Rupelian-Chattian tectonic rearrangement (Vandenberghe et al., 1998).

The subsiding Campine area can be considered as a southwestern shoulder at the rim of the Roer Valley Graben (RVG) (Fig. 1) which became strongly reactivated at the start of the Chattian. The Oligocene thickness distribution map in the North-Sea Basin (Vinken, 1988) shows that the subsiding RVG area in Germany, Belgium and the Netherlands was a particular sediment sink during the Chattian. Sections across the area from Belgium, over the Netherlands into Germany (Hager et al., 1998; Van Simaeys, 2004) (Fig. 15) show up to 400 m Chattian deposits in the RVG whilst the sea only occasionally spilled over the limits of the RVG towards the west depositing much thinner glauconitic deposits over the Campine and even Antwerp area (Figs 1, 15).

The development of the RVG during the Chattian was already preceded by changing sedimentation conditions in the area towards the end of the Rupelian, expressed by the development of the upper Rupelian Eigenbilzen Formation which is a fine sandier facies than the Boom Clay (Fig. 1). This Eigenbilzen Formation is time-equivalent of the upper part of the Boom Clay in the west and is most developed almost exactly overlying the later subsiding RVG area, east of its boundary faults like the Mol-Rauw fault (Fig. 1); its sandier lithofacies indicates a shallower depositional environment compared to the Boom Clay. Therefore the RVG block in that area must have been slightly uplifted during the late Rupelian before its considerable subsidence from the start of the Chattian onwards. The total thickness of non-eroded Rupelian sections, the sum of the Boom Clay and Eigenbilzen Sand, in and outside the RVG is very similar (Fig. 15) (see also fig. 4 in Hager et al., 1998 and fig. 15 in Vandenberghe et al., 2001).

The Chattian sediments deposited mainly in the RVG, and less on its shoulder in the Campine and in the Chattian type area of Doberg between Kassel and Hannover (Fig. 1) in Germany, are shallow marine glauconitic and often shell-bearing sands. The several more clay-rich intervals associated sometimes with lignite along the southern edge of the Lower Rhine Graben are shallow deltaic water deposits (Hager et al., 1998); it is not well established if the clay intervals deeper to

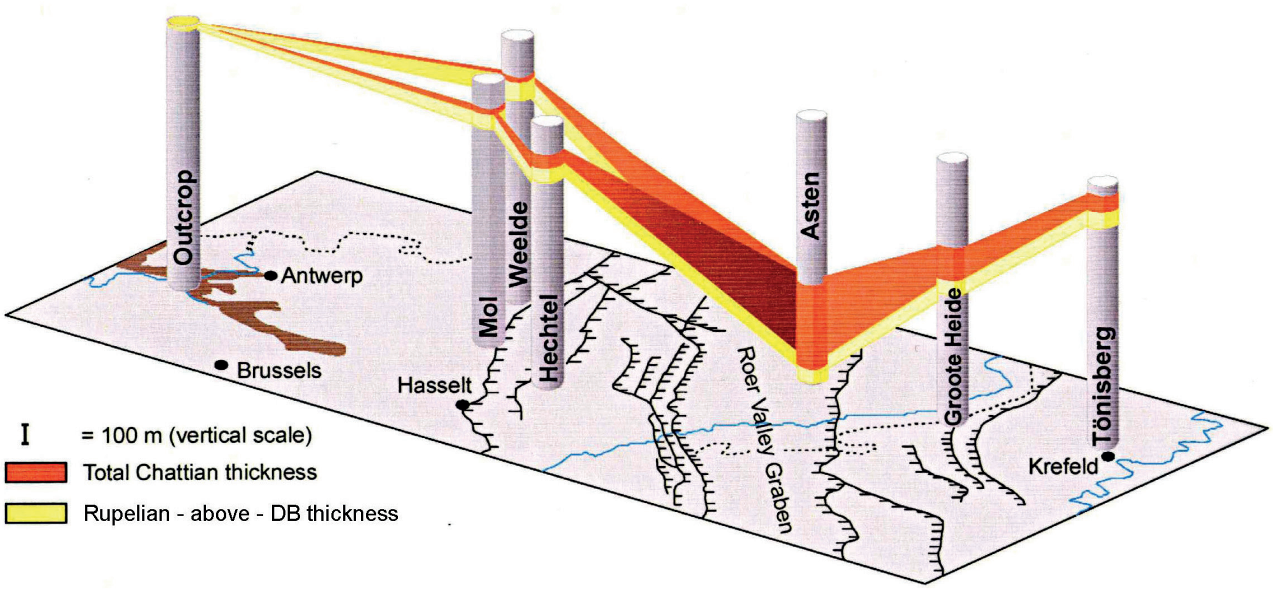

Figure 15. The thickness distribution of the top Rupelian (above the Double Layer) and Chattian sediments over the Roer valley Graben area (from Van Simaeys, 2004); note the almost constant Rupelian thickness in contrast to the increased thickness of the Chattian over the graben.

the northwest in the RVG are also shallow water clays or if they are on the contrary deposited further away from the main river input in waters deeper than the main glauconitic body sand as is suggested by the geometry of the deposits in the profiles by Schäfer et al. (2004, 2005). Chattian stratigraphic successions and their correlations in the contiguous area of the RVG, the West Netherlands Basin and Belgium need further precision.

9. Rupelian-Chattian boundary and Chattian paleoclimate

Although the historical stratotype areas for the Rupelian and the Chattian are located in the North-Sea-Basin realm, a major hiatus occurs between the top Rupelian in Belgium and the base of the Chattian in Doberg, Germany (Fig. 4). This hiatus is the result of the combined effect of Savian tectonism and of a glacio-eustatic sea-level fall (Vandenberghe et al., 2012). Because of the presence of this hiatus the Chattian GSSP cannot be defined in the historical type area. A suitable GSSP has been found in the pelagic sequence in the Umbria-Marche Basin of the NE Apennines in Italy. The recently defined Rupelian-Chattian GSSP defined in the Monte Cagnero section of this area (Coccioni et al., 2016) honours the chronostratigraphy of the historical stratotypes in the sense that the boundary age falls within the time of the hiatus between the Rupelian and Chattian historical stratotypes as studied in De Man et al. (2010). Following the correlation between the Rupelian-Chattian composite section in the North-Sea area and the Monte Cagnero GSSP section by Śliwińska et al. (2014), the duration of the hiatus in the historical area is at maximum 1.7 Ma, situated for a large part in the top of the Rupelian (fig. 12 in Coccioni et al., 2016), and encompassing the glacio-eustatic sea-level low labelled Oi2a (Fig. 4). In the southern North-Sea Basin sedimentation is resumed during the earliest Chattian by transgressive shallow water sediments characterised by the bloom of tropical to subtropical benthic foraminifer taxa, 'the *Asterigerina* horizon', contrasting with the generally colder fauna in the underlying Rupelian sediments (De Man & Van Simaeys, 2004). However another short global cool phase labelled Oi2b and expressed by the southwards migration of the cold dinoflagellate species *Svalbardella*, dated at 27.1 Ma, (Van Simaeys et al., 2004; Śliwińska et al., 2010, 2014; Coccioni et al., 2016) interfered with the warm start of the Chattian '*Asterigerina* horizon' (Śliwińska et al., 2014). In the eastern North Sea a forced regression unit (see Catuneanu, 2006) is identified that formed during this last Oi2b cooling event (Clausen et al., 2012).

Globally during the Chattian, benthic foraminifera oxygen isotope data show a gradual return to warmer temperatures which in the latest part of the Chattian become similar to the end-Eocene temperatures (Pagani et al., 2005, 2009). At the start of the Miocene a global cooling event occurs and global temperatures decrease again (Bender, 2013). The start of the Miocene is not documented in the Belgian sections as it corresponds with a hiatus extending to the Burdigalian and related to the still ongoing Savian uplift pulse.

The Oligocene is the Epoch during which the carbon dioxide levels in the atmosphere show an increased drop to modern pre-industrial levels. This drop has been related in particular to the enhanced weathering of silicates as a consequence of the crustal thickening and uplift of the Tibetan-Himalayan collision (Pagani et al., 2009). That the atmospheric carbon dioxide concentrations did not further lower during the Neogene notwithstanding the increased tectonic uplift and denudation rates is explained by changes in terrestrial vegetation controlling silicate weathering: the decrease in atmospheric carbon dioxide during the Oligocene led to loss of forest cover and expansion of grasslands during the latest Oligocene and early Miocene, weakening the weathering feedback effect and preventing a further drop in carbon dioxide (Pagani et al. , 2009).

10. Conclusions

The Oligocene sediments in the southern North Sea basin are comprised between the Pyrenean and Savian unconformities. This area is the historical reference area for the Rupelian and Chattian stages of the Oligocene. The almost enclosed nature of the basin is the reason for the dominance of endemic fossils and the frequent occurrence of hiati, making the area unsuitable for the modern definition of stage boundaries. The importance of the Pyrenean unconformity is underlined by a sudden major change in clay mineral composition at the base of the late Eocene Priabonian. The earliest Rupelian sedimentary cycle is a short-lived but regionally extensive transgression consisting of shallow water deposits. Upon the regression of the sea from the area a soil is developed. This moment in time corresponds to a major global cooling event and also to a marked change in mammal fauna in West-Europe and could eventually better define the base of the Oligocene than the present slightly older bio-event criterion. The next major Rupelian transgression stepwise retrogrades southwards over the land. The Boom Clay is the dominant lithology of the Rupelian in the southern North Sea basin. Its major sedimentological feature is its regularly layered architecture reflecting the glacio-eustatic obliquity beat of the climate. Lower-frequency cycles in the lithology, traditionally also considered reflecting fluctuating climatic sea-level sequences, may however rather reflect the regional tectonic evolution. By the end of the Rupelian tectonic activity increased and induced uplift and considerable erosion in the west while in the area overlying the later RVG in the east uplift resulted in the development of a shallower fine-sandy facies. The RVG resumed strong subsidence at the beginning of the Chattian accommodating several hundred meters of shallow water deposits. Joints and small unconformity surfaces in the Boom Clay have been related to the differentiated subsidence and uplift history in the area. The generally warmer climate of the Chattian compared to the Rupelian, was also punctuated by glacial cooling events. Outside the RVG, continuous sedimentation resumed only in the Burdigalian overlying the Savian unconformity.

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