Geologie en Mijnbouw **70**: 339–346, 1991.

Eustatic sea level and depth of a Late Cretaceous epicontinental sea: an example from NW Europe

Martin J.M. Bless
Lab. Paléontologie végétale UL, 7 Place Vingt Août, B-4000 Liège, Belgium

Received 4 February 1991; accepted in revised form 24 April 1991

Key words: Campanian-Maastrichtian chalk deposits, Late Cretaceous sea level, NW Europe, paleobathymetry, tectonic overprinting

Abstract

The claim that the Late Cretaceous (and more specifically the Campanian-Maastrichtian) global sea level was considerably higher than that of today is partly based on the assumption that one can recognize areas which were tectonically immobile from the Cretaceous onwards. Careful examination of some of these regions in NW Europe suggests, however, that differential (inverse) warping and tilting during and/or since the Cretaceous has invalidated their use as reliable gauges for Late Cretaceous high stands.

In addition, the original depth of Late Cretaceous epicontinental seas may have been much shallower than generally assumed. In the literature it is frequently suggested that the NW European flint-bearing chalk was deposited at a depth of 100 to 300 m or more. But this study shows that, for example, the Late Campanian and Maastrichtian flint-bearing chalk in the Hautes Fagnes area of NE Belgium accumulated at a depth of less than 45 to 65 m, matching the Modern deposition of coccolith ooze in the Caribbean Belize Lagoon at depths of less than 40 m at places.

Introduction

Various views have been presented on the depth of epicontinental seas and the magnitude of eustatic sea level rise and fall during the Late Cretaceous. Hays & Pitman (1973) estimated that the highest Late Cretaceous sea level (during the Cenomanian-Turonian transgression) stood 521 m above the present level. Even higher estimates were made by Hancock & Kauffman (1979), who argued that the Late Campanian sea level could have been as much as 580 to 645 m and the Early Maastrichtian one 645 to 660 m higher than that of today.

A much lower maximum Late Cretaceous sea level was suggested by Pitman (1978: 350 m instead of the 521 m calculated by Hays & Pitman 1973), by Watts & Steckler (1979: 150 m), by Robaszynski (1981: 250 to 450 m), by Kominz (1984: minimum 45 m, maximum 365 m, most probable height 230 m) and also by Barron et al. (1985: 400 m). Nowadays a Late Cretaceous high stand of some 100 to 300 m above the present sea level is commonly accepted (cf. Ziegler 1982, pp. 75–76; Schlanger 1986, p. 62; Haq et al. 1987).

To accurately estimate Late Cretaceous sea level it is crucial that the presumed depth of the epicontinental sea and the alleged stability of the area be considered. Such an approach was advocated by e.g. Robaszynski (1981, pp. 206–208: 'It would appear that a thickness of between 150 and 250 m for the mid- and late-Cretaceous is usual for formations deposited on apparently stable zones in northwest Europe... If one now assumes a sea water depth of between 100 and 200 m off the coast

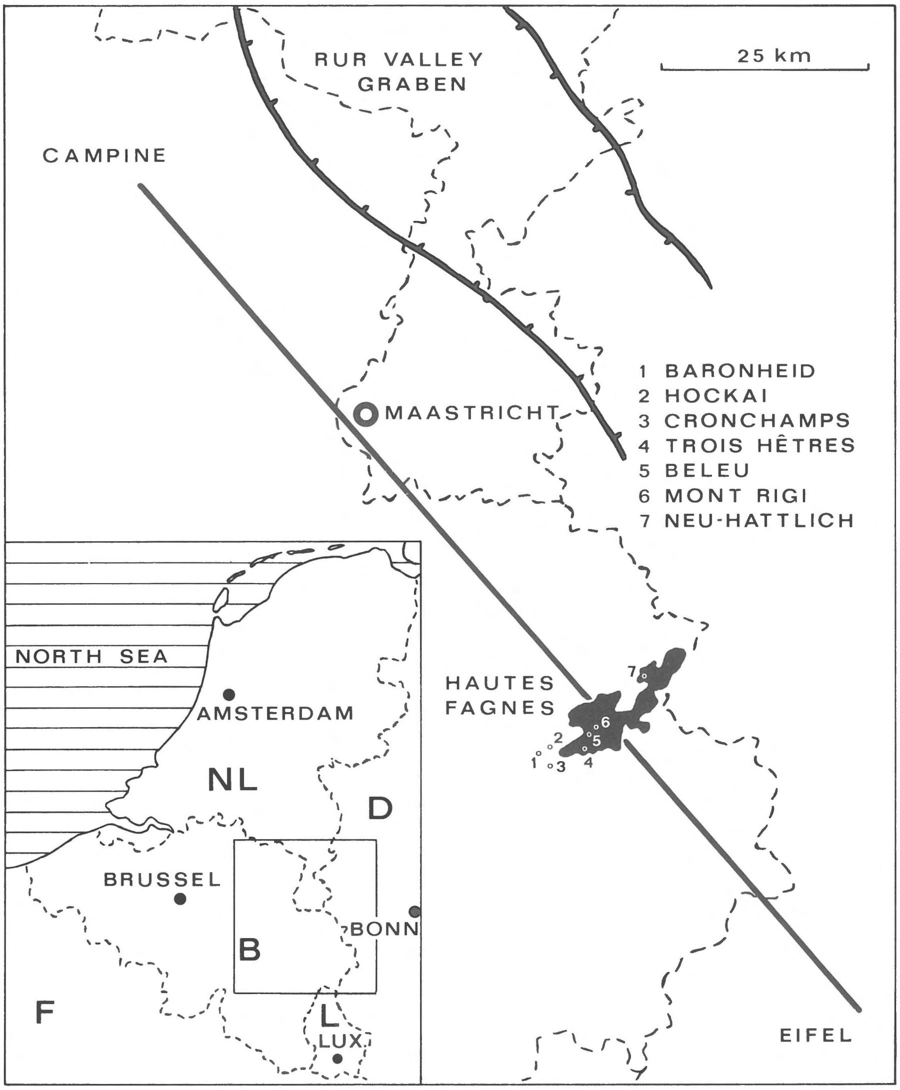

Fig. 1. Location of the Hautes Fagnes area in NE Belgium (largely restricted to the area above 600 m altitude, here shown in black). The German extension of this area is known as 'Hohes Venn'. Residual Late Cretaceous deposits at the localities 1 to 7 have been dated by means of forams, ostracodes, baculitid ammonites and brachiopods (Bless et al. 1991). The heavy straight NW-SE running line shows the position of the cross-section in Fig. 2.

of the marginal areas, the maximum eustatic elevation of sea level during the mid- and late-Cretaceous must be between 250 and 400–450 m.'), and Schlanger (1986, pp. 67–68: 'At the time of the union of the Tethyan and Nigerian arms of the trans-Saharan sea in Early Turonian the water depth in northwest Nigeria, where topographic highs were submerged, was 20–30 m. Considering the stability and elevation of Africa the sea level rise ... could not have been less than ~ 150 meters.').

However, these assumptions are questionable as is illustrated here using the Late Cretaceous in Belgium and the Netherlands as an example. Hancock & Kauffman (1979, p. 183, table 4) suggested an 'original depth of sea' for the 'Middle Maastrichtian' in the 'Ardennes, south-east of Liège, Belgium' of '100 m'. The present height above sea level

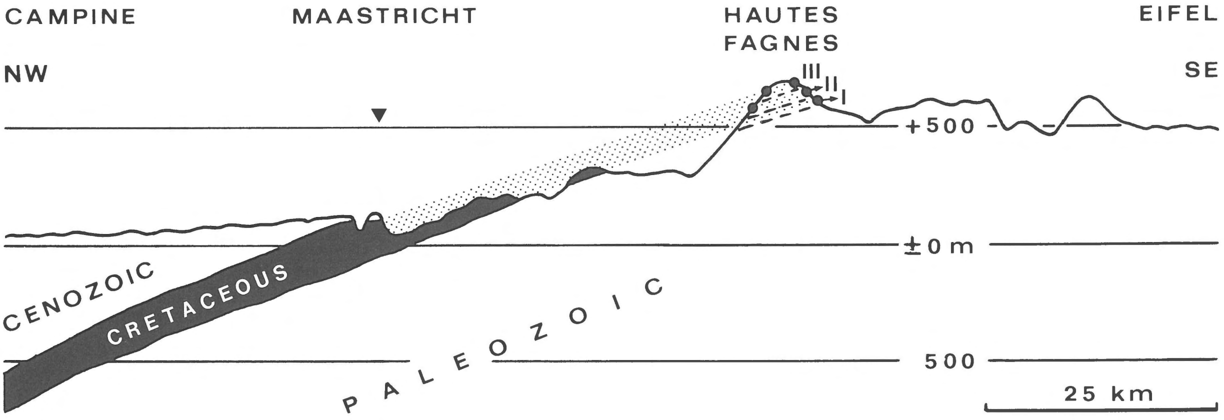

Fig. 2. Relationship between the Late Cretaceous deposits in the Hautes Fagnes and those in the Maastricht region and the Campine, indicating regular post-Cretaceous tilting of this part of the section. Black dots in the Hautes Fagnes relief indicate residual Late Cretaceous deposits. Numbers I to III refer to step by step flooding of the Hautes Fagnes 'monadnock' during Campanian-Maastrichtian. I: sequences starting with Early Campanian basal conglomerate; II: Late Campanian basal conglomerate; III: Late Maastrichtian basal conglomerate.

of that area was indicated to be '545–560 m'. From these data, based on a region 'in which the Cretaceous is nearly horizontal and over a stable massif', they assumed a Middle Maastrichtian sea level to have been '645–660 m' higher than the present level.

Although not stated explicitly, these authors possibly referred to the eluvial deposit at Hockai in the western part of the Hautes Fagnes area in NE Belgium (Fig. 1), the base of which is c. 545 m above the present sea level (cf. Bless et al. 1991). The 11 to 12 m thick fossiliferous flint 'conglomerate' at Hockai was described already in the last century by a.o. Dewalque (1886), who recorded the presence of 'Senonian' and 'Maastrichtian' fossils. Rich and diverse, completely silicified microfossil assemblages (forams and ostracodes) allow a distinction to be made between an Early Campanian basal conglomerate (on top of Cambrian basement), an early Late Campanian glauconite-rich layer, and late Late Campanian and early Late Maastrichtian flint-bearing strata, all completely decalcified by dissolution and consequently strongly compacted (Bless & Felder 1989, Bless et al. 1991).

Hancock & Kauffman (1979, pp. 181–182) based their calculations on three 'cornerstones':

- the supposed stability of an area from the Cretaceous onwards;
- the original thickness of the deposit; and
- the supposed depth of the sea at the time of deposition.

The following discussion illustrates the limited value of these arguments.

Apparent stability

The post-Cretaceous instability of NE Belgium was assessed already by Dumont in 1847, and later by a.o. Renier (1902), Macar (1954), Legrand (1968), Robaszynski (1981) and Gullentops (1987). Therefore, the Hautes Fagnes area 'cannot be advanced as an example of a stable massif' (Robaszynski 1981, p. 208), because 'the actual relief of the northern Ardennes is the result of important post-Cretaceous tectonic activity' (Gullentops 1987, p. 365). Figure 2 illustrates the important post-Cretaceous tilting of NE Belgium and the southeasternmost part of the Netherlands (the classic Maastricht area where the 'Maastrichtian' or 'système maestrichtien' was defined by Dumont in 1849).

Furthermore, rapid changes in relative thickness and facies as well as local sedimentary gaps emphasize the impact of differential and inverse move-

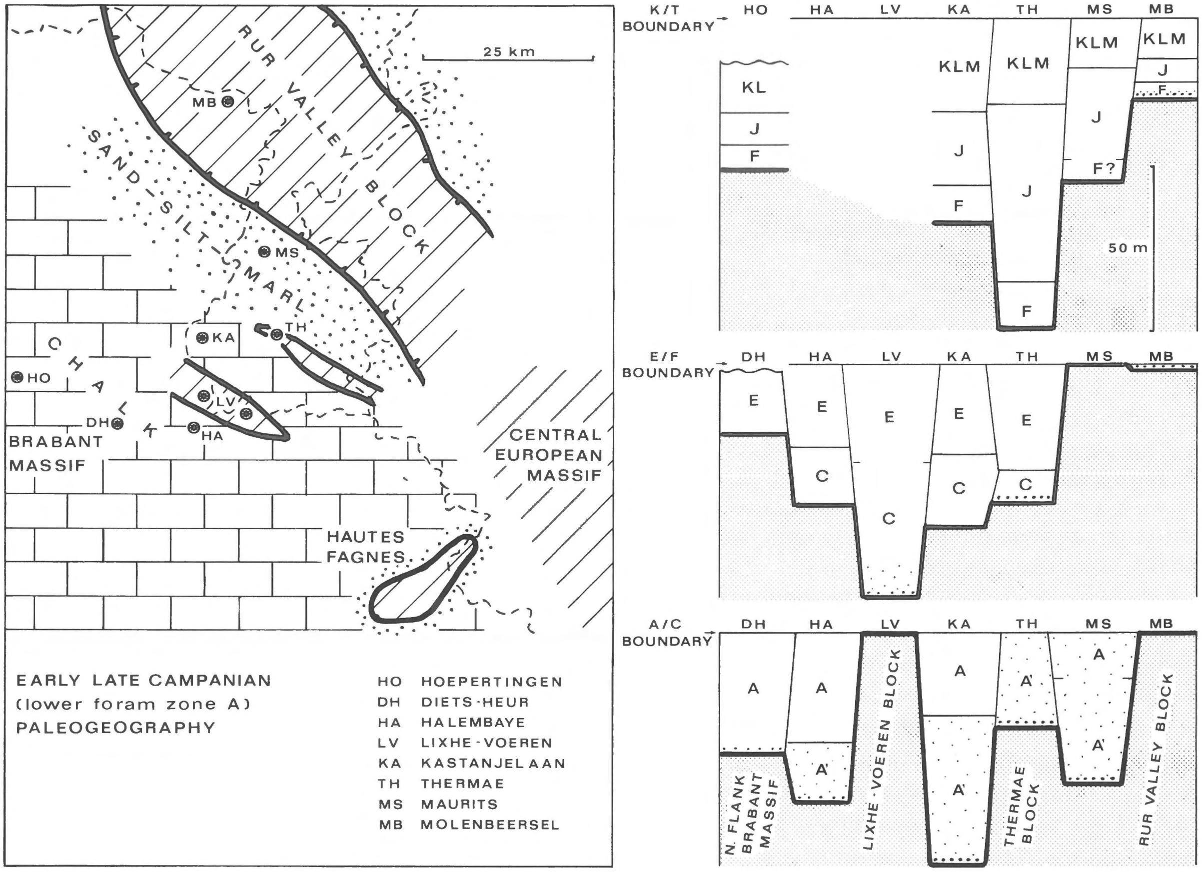

Fig. 3. Variation in thickness of Campanian-Maastrichtian deposits in NE Belgium and the SE Netherlands reflecting differential movement, partly compensating subsidence and/or tilting, of block-faulted basement. A′, A to M: foram zones. A′, A: Early and Late Campanian; C, E: early Late Maastrichtian; F, J: middle Late Maastrichtian, KLM: late Late Maastrichtian. Note reduced total thickness and important gaps on Rur Valley block to the north (inversion structure since Early Campanian) and northern flank of Brabant Massif to the south.

ments of the block-faulted basement of this area during the Late Cretaceous (Fig. 3; Bless & Bouckaert 1988). In addition, it is here emphasized that there are no traces of any sedimentation during the Cenomanian-Turonian transgression in NE Belgium or the SE Netherlands, although this transgression would have taken place during the most important Late Cretaceous global sea level rise (e.g. Wiedmann 1988). Differential post-Cretaceous block movements and tilting of the Belgian Brabant Massif are easily deduced from the maps and cross-sections by Legrand (1968, Figs 5 and 9, planches I-III).

In other words, both synsedimentary and post-Cretaceous regional tectonic movements have overprinted any Late Cretaceous sea level change in Belgium and the Netherlands. Similar statements may be made for the Paris Basin (Robaszynski 1981: 'moderation of Cretaceous transgressions by block tectonics'), Germany (Kockel 1987, Rossa 1987) and the Dutch part of the North Sea area (Van Wijhe 1987, Kooi et al. 1989) where inversion tectonics have controlled the sedimentary history during the Late Cretaceous and Early Tertiary.

The presumed stability of north Antrim (northern Ireland) during and since the Late Cretaceous (Hancock & Kauffman 1979, table 4) raises questions as well (Gullentops 1987). Antrim is one of the regions surrounding the Rockall Trough that are marked by Early Tertiary plateau basalts and dykes (cf. Ziegler 1982), which may have caused considerable post-Cretaceous uplift. It is not unlikely that this area subsided during the Late Cretaceous expansion of the Rockall Trough. Davies (1970) and Mitchell (1981, p. 233) already suggested that 'the remarkable difference in elevation of the various Tertiary deposits in Ireland points to extensive tectonic activity... Thus the sub-basaltic Chalk-surface varies in height from + 500 m at Slieve Gullion to − 800 m in the Langford Lodge Boring, a range of 1300 m', whereas Wilson (1972, p. 57 and also fig. 76) mentioned block-faulting and folding affecting the distribution of Maastrichtian beds between Portrush and Ballycastle in north Antrim ('Maestrichtian beds are preserved in the down-faulted block between the Portnakillew and Tow Valley faults, and also in a relatively down-folded area near Porthrush').

The alleged stability of Devon (SW England) from the Albian onwards (Hancock & Kauffman 1979, table 4) is also open to criticism. This region is located between the Bristol Channel Trough to the north and the Western Approaches Trough to the south, and marked by NW-SE running faults which were active during the Cretaceous and Tertiary (cf. Ziegler 1982). Moreover, the major Late Campanian transgression in NW Europe (cf. Hancock & Kauffman 1989) presumably coincides with a peak in the Late Cretaceous to Early Tertiary inversion tectonics. It is not before the middle Late Maastrichtian that chalk deposits overstepped the inverted Rur Valley Block (cf. Fig. 3). The same appears to be true for several other inversion axes (Van Wijhe 1987).

It is therefore most probable that there is no area in NW Europe that can be considered to have been tectonically stable from the Cretaceous onwards (cf. Kooi et al. 1989). Consequently, the present altitude of a Late Cretaceous deposit in NW Europe should not be used as a gauge to assess the eustatic sea level during its sedimentation.

Original water depth

It is commonly accepted that Late Cretaceous chalk accumulated on a relatively deep shelf. Robaszynski (1981, p. 208) assumed 'a sea water depth of between 100 and 200 m off the coast of the marginal areas', whereas Wilson (1981, p. 210) suggested that 'the depth in which the ooze was formed was probably 300 m or more.' However, the assessment of the original water depth is one of the most difficult aspects of paleoenvironmental reconstructions. This is perhaps best illustrated by the wide range of estimates proposed for the depositional environment of the Late Cretaceous Chalk in the Harlingen gas field in the northern Netherlands (Van den Bosch 1983). These vary from 'not much more than 70 m' (based on 'the high percentage of intact coccoliths') through 'a depth range of 100–600 m' ('based on a large number of criteria') to 'approximately 250 m' (based on 'seismic-stratigraphic analysis').

Hancock & Kauffman (1979, table 4) suggested an original depth of the 'Middle Maastrichtian' sea at Hockai of 100 m. The bulk of the (now completely decalcified and compacted, eluvial) deposit consisted of flint-bearing chalk, its lithology and fossil content are similar to those of the coeval Late Campanian 'craie blanche' (Zeven Wegen Chalk and possibly also Beutenaken Chalk) and of the early Late Maastrichtian 'craie grise' (Vijlen Chalk) and 'craie tigrée' (Lixhe Chalk) in the classic area around Maastricht (Bless & Felder 1989).

Liebau (1980, p. 209) placed these deposits of the 'Gulpenkreide' (Gulpen Formation) in his 'sublittoral below the pterygocline' (elittoral/lower sublittoral). Comparison with Liebau's Figs 3 and 4 suggests a depth between 30–60 m (storm wave base or pterygocline) and 150–200 m (lower limit of shelf environment, which, according to Liebau, is marked by 50% plankton ratio in foraminifer thanatocoenoses). Liebau was unable to find indications of a lower limit of the circalittoral zone (marked by the rhodocline or lower limit of red algae, which he accorded a waterdepth between 80 and 200 m). Some years later, the same author showed that the 60 m storm wave base is characteristic of oceanic conditions, whereas a 30 m (or even shallower) storm wave base marks an inland sea (Liebau 1984, Fig. 4). Similarly, the rhodocline

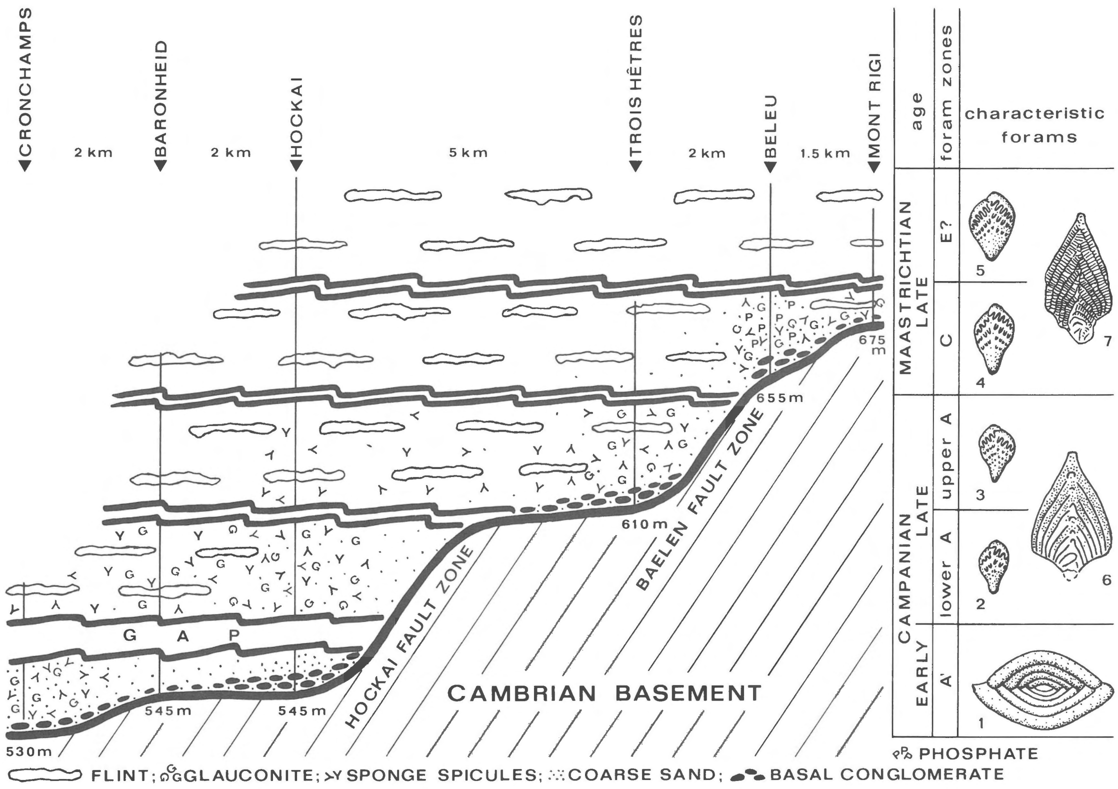

Fig. 4. Step by step flooding of the southwestern flank of the Hautes Fagnes during Campanian-Maastrichtian. Note the rapid lateral and vertical change in lithofacies. Early and Late Campanian transgressions came to a halt at nowadays active Hockai and Baelen fault zones. This indicates that these were active already during the Late Cretaceous. Presumably, their throw has increased by Late Tertiary and Quaternary movements (modified after Bless et al. 1991). Characteristic forams in column: 1. *Spirophthalmidium cretaceum* (Reuss); 2–3. *Bolivinoides* gr. *decoratus* (Jones) with 3 and 4 pustules on the last chamber, respectively; 4–5. *Bolivinoides* gr. *australis* Edgell with 5 and 6 pustules on the last chamber, respectively; 6. *Neoflabellina leptodisca* (Wedekind); 7. *Neoflabellina reticulata* (Reuss).

occurs at shallower depths in inland seas than in the open ocean.

There is little doubt that the conditions of the Late Cretaceous sea in the SE Netherlands and NE Belgium more closely compared to those of an inland sea than those of an ocean, because this region was surrounded by the Central European Massif to the east, the inverted blocks of the former and later Western Netherlands and Rur Valley grabens (Van Wijhe 1987) to the north, and the Brabant Massif (marked by a very incomplete Late Cretaceous succession; cf. Fig. 3) to the west and southwest. Low numbers of planktonic species and specimens in foraminifer assemblages (usually well below 1%; personal observations) of the Campanian-Maastrichtian in this area corroborate the virtual absence of oceanic influence.

These observations suggest that for the area under discussion the maximum depth of the storm wave base was at 30 m or even less, and that there is really no paleontological evidence for the depth of the lower limit of Liebau's (1980) 'lower sublittoral'. One could just as well assume it to have been at 50 m than at 250 m or more.

Paleogeomorphological data suggest that the Campanian-Maastrichtian flint-bearing chalk in the Belgian Hautes Fagnes area must have been deposited in a rather shallow sea. The closely spaced sections in that area record the stepwise inundation of the Hautes Fagnes during the Campanian and Maastrichtian (Fig. 4). The pre-Cretaceous basement rocks of the Hautes Fagnes formed a morphological high or monadnock, rising some 100–150 m above the pre-Late Cretaceous (pre-Santonian) peneplain, comprising the Maastricht area and the Campine to the northwest (cf. Fig. 2). During the Late Cretaceous transgression this high became an island of gradually decreasing extent until it was completely flooded at the start of the early Late Maastrichtian (Bless et al. 1991).

The basal conglomerates (with poorly rounded pebbles, up to 20 cm in diameter) in the various sections are assumed to have formed in the intertidal zone at 0 m water depth. These conglomerates can be correlated with flint-bearing chalk (now completely decalcified) in more offshore sections at distances of a few kilometres only (Fig. 4). The depth of the water in which this chalk accumulated cannot have exceeded the present-day difference in altitude between the successive residual deposits, for the Late Cretaceous topography of the Hautes Fagnes (at least along the line of the sections from the SW to the NE) appears to have been very similar to the contemporary one (Bless et al. 1991).

This line of reasoning leads to a maximum water depth of 45 m for the early Late Maastrichtian and late Late Campanian flint-bearing chalk at Trois Hêtres (610 m altitude) in comparison with the early Late Maastrichtian basal conglomerate at Beleu (655 m altitude) some 2 km to the north, or to a maximum water depth of 65 m for the early Late and late Late Campanian flint-bearing chalk at Baronheid (545 m altitude) and Hockai (545 m altitude) in comparison with the late Late Campanian basal conglomerate at Trois Hêtres (610 m altitude). These figures are to be reduced by taking into account the original thickness of the present residual deposits (Bless et al. 1991), and the possibly (late Cenozoic) enhanced throw of the Hockai and Baelen faults (Fig. 4).

Since the fossil assemblages (forams, ostracodes, echinoderms, molluscs, bryozoans) in the coeval flint-bearing chalk (soft, friable calcilutites and calcisiltites) in the Maastricht area are very similar to those of the Hautes Fagnes (Bless et al. 1991), one may assume these to reflect roughly comparable water depths, not exceeding 45 to 65 m. These observations obviously contradict Hancock & Kauffman's (1979) assumption of an original depth of 100 m in this area. However, this does not mean that these figures also apply to chalk deposits elsewhere in NW Europe. Robaszynski's (1981) suggestion of 'a sea water depth of between 100 and 200 m off the coast of the marginal areas' may still be valid. However, such assumptions need to be checked carefully in individual cases and should not lead to generalisations for large areas.

The suggestion that the succession from glauconitic sand/silt/clay to (flint-bearing) chalk (observed in many Late Cretaceous sequences, including the Maastricht area and Hockai) 'certainly implies a sharply deepening sea' (cf. Schlanger 1986, p. 64) is open to criticism if one considers the Holocene deposition of coccolith ooze in the Caribbean Belize Lagoon at depths of less than 40 m at places (Scholle & Kling 1972, Simien 1987). Moreover, as already emphasized by Simien (1987), the succession of siliciclastics, glauconite and coccolith ooze in the Southern Belize Lagoon forms a perfect Modern counterpart of the Late Cretaceous sequences in North America, NW Europe and the European part of the USSR.

These observations illustrate that coccolith ooze and flint-bearing chalk (calcilutites and calcisiltites) may accumulate and have, in fact, done so in very shallow environments, at least on a local or regional scale. In this context, it seems worthwhile to refer to a recent paleobathymetrical analysis of the Late Cretaceous Gulf Coast Plain Chalk (USA), for which a depositional depth of between 35 and 85 m has been proposed (Puckett 1990).

Perhaps the average depth of Late Cretaceous shelf seas has been overestimated at times. And the same may be true for calculations of the Late Cretaceous global sea level.

References

Barron, E.J., M.A. Arthur & E.G. Kauffman 1985 Cretaceous rhythmic bedding sequences: a plausible link between orbital variations and climate – Earth Planet. Sci. Lett. 72: 327–340

Bless, M.J.M. & J. Bouckaert 1988 Suggestions for a deep seismic investigation north of the Variscan mobile belt in the SE Netherlands – Ann. Soc. Géol. Belg. 111: 229–241

Bless, M.J.M. & P.J. Felder 1989 Note on the Late Cretaceous of Hockai (Hautes Fagnes, NE Belgium) – Ann. Soc. Géol. Belg. 112: 47–56

Bless, M.J.M., A. Demoulin, P.J. Felder, J.W.M. Jagt & J.P.H. Reynders 1991 The Hautes Fagnes area (NE Belgium) as a monadnock during the Late Cretaceous – Ann. Soc. Géol. Belg. 113: 75–101

Davies, G.L. 1970 The enigma of the Irish Tertiary – In: Stephens, N. & R.E. Glasscock (eds): Irish Geographical Studies – Queens Univ. Belfast, Dept. Geogr.: 1–16

Dewalque, G. 1886 Session extraordinaire à Spa tenue les 30 et 31 août et 1er septembre 1885 – Ann. Soc. Géol. Belg. 13: 29–56

Dumont, A. 1847 Mémoire sur les terrains ardennais et rhenan de l'Ardenne, du Rhin, du Brabant et du Condroz, Première Partie (terrain ardennais) – Acad. R. Belg. Mém. Cour. 20: 163 pp

Dumont, A. 1849 Rapport surt la carte géologique du Royaume – Acad. R. Belg. Bull. 16: 351–373

Gullentops, F. 1987 The Maastrichtian sea level rise – Ann. Soc. Géol. Belg. 109: 363–365

Hancock, J.M. & E.G. Kauffman 1979 The great transgressions of the Late Cretaceous – Geol. Soc. Lond. J. 136: 175–186

Hancock, J.M. & E.G. Kauffman 1989 Use of eustatic changes of sea level to fix Campanian-Maestrichtian boundary in Western Interior of USA – 28th Intern. Geol. Congr. Washington 1989, Abstr. 2: 23

Haq, B., J. Hardenbol & P.R. Vail 1987 Chronology of fluctuating sea level since the Triassic (250 million years to present) – Science 235: 1156–1167

Hays, J.D. & W.C. Pitman 1973 Lithospheric plate motion, sea level changes and climatic and ecological consequences – Nature 246: 18–22

Kockel, F. 1987 Upper Cretaceous biostratigraphy, the key to the understanding of inversion tectonics in NW-Germany – Ann. Soc. Géol. Belg. 109: 357–361

Kominz, M.A. 1984 Oceanic ridge volumes and sea level change – an error analysis – Am. Assoc. Pet. Geol. Mem. 36: 109–127

Kooi, H., S. Cloetingh & G. Remmelts 1989 Intraplate stresses and the stratigraphic evolution of the North Sea central Graben – Geol. Mijnbouw 68: 49–72

Legrand, R. 1968 Le Massif du Brabant – Mém. Explic. Cartes Géol. Min. Belg. 9: 148 pp

Liebau, A. 1980 Paläobathymetrie und Ökofaktoren: Flachmeer-Zonierungen – N. Jb. Geol. Paläont. 160: 173–216

Liebau, A. 1984 Grundlagen der Ökobathymetrie – In: Luterbach, H.P. (ed.): Paläontologische Kursbücher, 2, Paläobathymetrie – Paläont. Ges. Selbstverlag (München): 149–184

Macar, P. 1954 L'évolution géomorphologique de l'Ardenne – Soc. R. Belge Géogr. Bull. 78: 1–23

Mitchell, G.F. 1981 Other Tertiary events – In: Holland, C.H. (ed.): A Geology of Ireland – Scott. Acad. Press (Edinburgh): 231–234

Pitman, W.C. 1978 Relationship between eustasy and stratigraphic sequences of passive margins – Geol. Soc. Am. Bull. 89: 1389–1403

Puckett, T.M. 1990 Paleobathymetry of the Demopolis Chalk (Upper Cretaceous) based on ostracodes – SE Geol. Soc. Am. Ann. Mtg., Abstract: 1 p

Renier, A. 1902 Le poudingue de Malmédy, Essay géologique – Ann. Soc. Géol. Belg. 29: 145–223

Robaszynski, F. 1981 Moderation of Cretaceous transgressions by block tectonics – An example from the north and northwest of the Paris Basin – Cretaceous Res. 2: 197–213

Rossa, H.G. 1987 Upper Cretaceous and Tertiary inversion tectonics in the western part of the Rhenish-Westphalian coal district (FRG) and in the Campine area (N Belgium) – Ann. Soc. Géol. Belg. 109: 367–410

Schlanger, S.O. 1986 High frequency sea-level fluctuations in Cretaceous time: an emerging geophysical problem – In: Hsü, K.S. (ed.): Mesozoic and Cenozoic Oceans – Geodynamic Ser. 15, Am. Geophys. Un./Geol. Soc. Am.: 61–74

Scholle, P.A. & S.A. Kling 1972 Southern British Honduras: lagoonal coccolith ooze – J. Sediment. Petrol. 42: 195–204

Simien, T.R. 1987 From coastal terrigenous deposits to pelagic rich oozes: the Holocene transgressive complex of southern Belize Lagoon (C.A.), a tool for the past? – Comparative Sed. Lab. Univ. Miami, Rosenstiel School Mar. Atmosph. Sci., Abstract 13: 32–35

Van den Bosch, W.J. 1983 The Harlingen Field, the only gas field in the Upper Cretaceous Chalk of the Netherlands – Geol. Mijnbouw 62: 145–156

Van Wijhe, D.H. 1987 Structural evolution of inverted basins in the Dutch offshore – Tectonophysics 137: 171–219

Watts, A.B. & M.S. Steckler 1979 Subsidence and eustasy at the continental margins of eastern North America – In: Talwani, M., W. Hay & W.B.F. Ryan (eds): Deep drilling results in the Atlantic Ocean: Continental margins and paleoenvironments – Maurice Ewing Series 3, Am. Geophys. Un.: 218–234

Wiedmann, J. 1988 Plate tectonics, sea level changes, climate – and the relationship to ammonite evolution, provincialism, and mode of life – In: Wiedmann, J. & J. Kullmann (eds): Cephalopods – Present and Past – Schweizerbart (Stuttgart): 737–765

Wilson, H.E. 1972 (Regional Geology of Northern Ireland – Her Majesty's Stationary Office (Belfast): 115 pp

Wilson, H.E. 1981 Permian and Mesozoic – In: Holland, C.H. (ed.): A Geology of Ireland – Scott. Acad. Press (Edinburgh): 201–212

Ziegler, P.A. 1982 Geological Atlas of western and central Europe – Elsevier (Amsterdam): 130 pp