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Reconstructing middle to late Holocene sea-level change: A methodological review with particular reference to 'A new Holocene sea-level curve for the southern North Sea' presented by K.-E. Behre

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A number of disciplines are involved in the collection and interpretation of Holocene palaeoenvironmental data from coastal lowlands. For stratigraphic frameworks and the assessment of relative sea-level (RSL) change, many non-specialists rely on existing regional models. It is, however, important that they are aware of major developments in our understanding of the factors controlling coastal change and of the potential sources of error in sea-level reconstructions. These issues are explored through a critical evaluation of a new sea-level curve presented by Behre (2003, 2007) for the southern North Sea. In contrast to most sea-level curves published from this region over the last 20 years, the curve shows strong fluctuations that are interpreted as representing vertical movements of sea level. We present a detailed examination of the data used by Behre. From this analysis it is clear that many of the data points used are unsuitable for high-resolution (centimetre or decimetre) sea-level reconstruction. This paper also gives an overview of possible sources of error with respect to the age and altitude of sea-level index points and of changes in our understanding of the processes that underpin the interpretation of the organic and occupation levels used as index points. The constraints on the spatial scale over which sea-level reconstructions can be applied (changes in palaeotidal range and crustal movements) are also considered. Finally, we discuss whether the large-amplitude centennial-scale sea-level fluctuations proposed by Behre can be reconciled with the known mechanisms of sea-level change and other recent high-resolution studies from this region. We conclude that such fluctuations are highly unlikely to be real features of the sea-level history of the southern North Sea.

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Knowledge of Holocene sedimentary successions and former sea levels is important for geologists, geographers and archaeologists who work in coastal lowlands and need to interpret their findings within a context of palaeoenvironmental change. This has been particularly true since the Convention of Malta, whereby archaeological investigations supported by a geological survey are required before construction works. At a site scale, detailed information is often lacking, and practitioners turn to regional stratigraphic schemes and sea-level curves. For such information these communities rely on publications produced by sea-level specialists, and it is important that they are aware of changes in the conceptual frameworks that underpin this discipline and of potential sources of error in the collection and interpretation of sea-level data.

Sediment variability in the late Holocene deposits on the eastern side of the southern North Sea (henceforth referred to as the southern North Sea) was for most of the 20th century interpreted within the context of transgressions and regressions controlled by vertical changes in sea level. In the last 30 years, the factors controlling coastal changes have been re-evaluated and the errors associated with the construction of sea-level curves have received greater recognition. The latter are based on a series of index points, usually sediment samples, which can be used to reconstruct former sea level when information on their geographic position, indicative meaning (the relationship of the sample material to a tidal level), age and altitude can be obtained. The major factors that determine the error associated with sea-level index points are uncertainties in the age and altitude estimates, considering (i) the spatial scale of the study area, including crustal movements and the coastal setting; (ii) the temporal scale, including changes in tidal range since the time of deposition; (iii) the indicative meaning of the sample; and (iv) compaction (Shennan 2007).

To demonstrate these major factors, as well as other issues that should be considered when reconstructing sea level, we present a critical evaluation of a sea-level curve from the southern North Sea presented by Behre (2003, 2007). The data used by Behre consist of index points selected from geological publications, along with many from archaeological investigations and some from personal communications. The latter are also available in Freund *et al.* (2004). During the middle and late Holocene, for which the temporal resolution is particularly good, the Behre curve consists of a series of sub-millennial oscillations of decimetre to a metre magnitude. Moreover, vertical changes in sea level are

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regarded as being synonymous with transgressions and regressions. Behre (2003, 2007) subdivides the Holocene deposits of the southern North Sea into two units: Calais and Dunkerque. Both are attributed to several transgressions separated by regressions represented by peat beds. It is claimed (2007: p. 82) that this is 'the only system that can be used for supraregional contexts because it is based on dated transgressive and regressive trends, and it is the only terminology that is applicable throughout the southern North Sea'. The Calais–Dunkerque system of stratigraphic subdivision, familiar to many, has been a subject of debate since the early 1970s. Despite the validity of this system being repeatedly questioned it is still in use (e.g. Augustyn 1992, 1995; Meurisse *et al*. 2005; Frouin *et al*. 2007; De Moor *et al*. 2010). The constraints of the Calais–Dunkerque system will not be discussed here: for such critical assessment the reader is referred to Streif & Zimmermann (1973), Barckhausen *et al*. (1977), Streif (1978), Baeteman (1981, 1983, 1991, 2005), van Loon (1981); Berendsen (1984), Westerhoff *et al*. (1987), Denys (1993, 1999), Wheeler & Waller (1995), Vos & van Heeringen (1997), Ervynck *et al*. (1999), Ebbing *et al*. (2003) and Weerts *et al*. (2005).

The area from which data have been compiled by Behre (2003, 2007) extends from the West Frisian part of the Netherlands, across the Weser and Elbe estuaries, to the North Frisian Islands (Fig. 1). This area

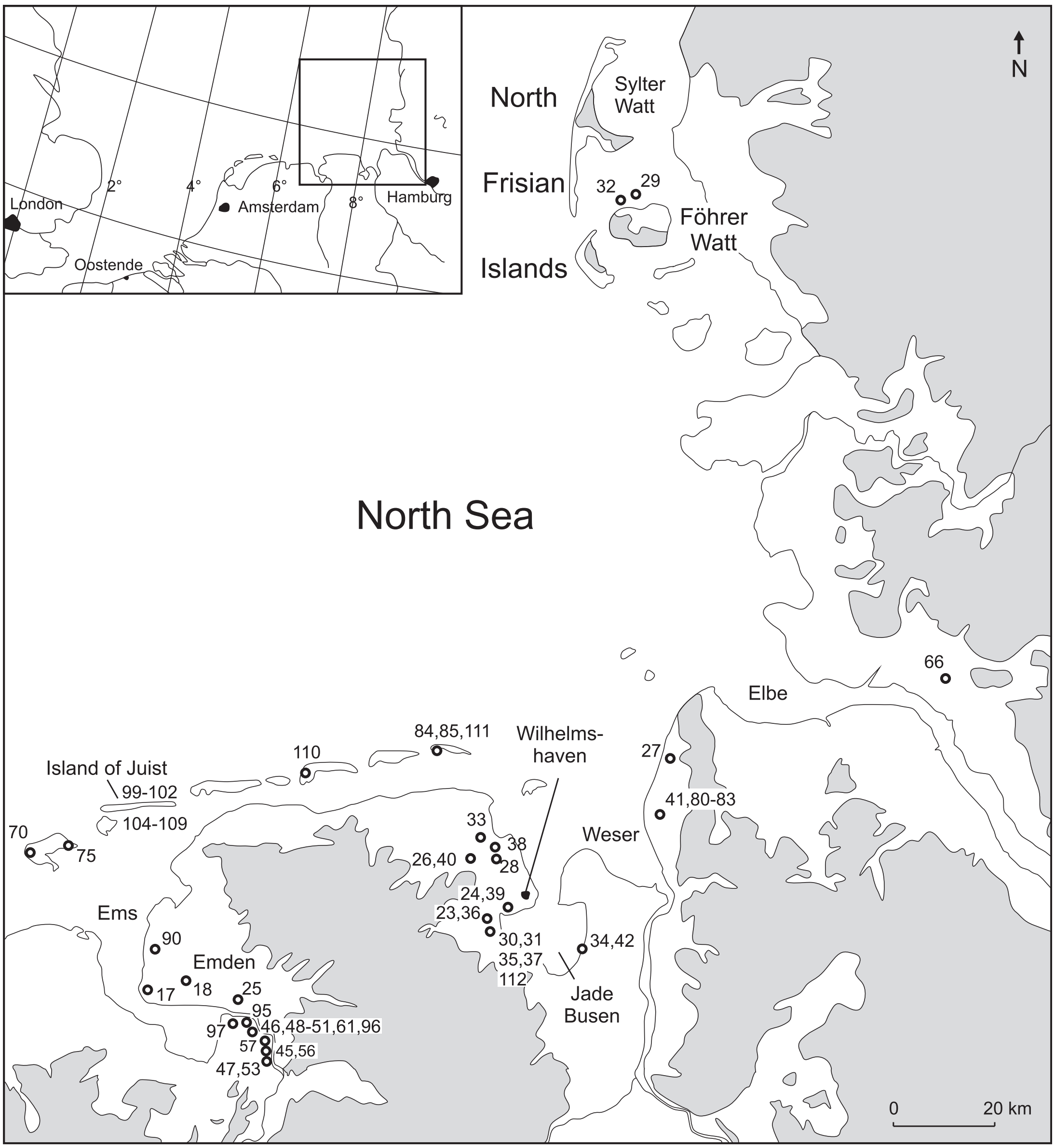

Fig. 1. Map showing the study area of Behre (2007) and the index points discussed in the text (redrawn from Behre 2007). Index point numbers as in Behre (2003, 2007).

stretches about 120 km north–south and about 140 km west–east, and covers a multiplicity of coastal settings including open coasts, barrier islands and tidal basins of various sizes. Considerable variation can therefore be expected in the site-specific processes (for example in tidal range, wave attack, longshore current, and variations in sediment budget) that influence the sedimentary record and the interpretation of sea-level index points. The tidal range varies from less than 1.50 m in the east to 2.50 m in the north, with ranges of more than 3.50 m at the Jadebusen and the mouth of the river Weser. The altitude of the index points given by the original authors was transformed by Behre to Mean High Water (MHW) and corrected to the standard tide gauge of Wilhelmshaven, Normal Null (NN).

In this paper we first analyse in detail the data presented by Behre (2003, 2007), to determine whether the interpretation of sea-level fluctuations of the magnitude shown can be sustained. For some of the data points, the original publications have been consulted to gain a better understanding of their stratigraphic context. In the second part of the paper, the concepts underpinning the sea-level curve are critically evaluated. The final part of the paper addresses whether the magnitude of the sea-level fluctuations for the middle and late Holocene proposed by Behre are plausible, namely whether they are consistent with our current understanding of the underlying processes and with other reconstructions of sea-level change from the region.

The data points

Behre (2003, 2007) attributes the Holocene deposits of the southern North Sea to four Calais (C) and six Dunkerque (D) transgressions (called Calais and Dunkerque series), which are separated by regressions (R) as reproduced here in Fig. 2. The data points used to compile the sea-level curve, referred to below, are shown in Behre's (2007) fig. 1 and listed in his tables 2 and 3.

The Calais series

As recognized by Behre, a clear distinction between the deposits of C I and C II is not possible. The evidence for separating C I and C II is the occurrence of wood near Emden, which, it is stated, may indicate a slowing down or a stagnation of sea-level rise *c.* 5350 cal. a BC (at −10 m NN). The context for this index point (18), given in his table 3, is wood from a river bank. According to the original publications (Barckhausen & Streif 1978; Barckhausen 1984), the sample was taken from channel-related mud deposits that occur locally in depressions in the pre-Holocene surface of the palaeovalley of the Ems (Streif 1989). These deposits formed subaquatically, as a result of freshwater flooding. It is therefore likely that their formation was dependent on local hydrological conditions rather than on changes in sea level. After this apparent stagnation, based on the first extensive demonstrable [*sic*] transgression (C II phase), the sea-level curve is shown as steepening again, and 'the basal peats around Wilhemshaven were also covered by marine deposits' (Behre 2007: p. 85). However, a lateral extension of marine deposits (an extensive transgression) does not necessary indicate a significant rise in relative sea level (RSL). The degree of lateral extension mainly depends on the relief of the pre-transgression surface, here characterized by valley systems separated by divides (cf. Caspers *et al.* 1995).

The data available to separate the C II and C III transgressions are also scant. Behre refers to intercalated peats dated during geological mapping near Emden (Barckhausen & Streif 1978) and in the lower Weser (Streif 1993). However, the age ranges available for these deposits are not included in the data set, but rather Behre selects other points (23, 24 and 25) to draw the curve (Figs 3, 4). These are from the top of peat beds and wood in mud and therefore are not indicative of the beginning of peat growth. Moreover, point 23 is from a basal peat.

Between *c.* 3900 and 3000 cal. a BC, the sea-level curve (Behre 2007: p. 84) is shown as steepening again (until −2.00 m NN), coinciding with the C III transgression (Figs 3, 4). Dates to document the start of C III are not discussed. The two points (24 and 25) listed in Behre's (2007) table 3 that are dated to around 3960 cal. a BC, however, are at −7.10 and −5.94 m NN, while the curve is drawn at −5 m NN. The altitude of point 24 has not been corrected from the original NN to local MHW, although no explanation is provided. All index points (26–30) to document the maximum of the C III transgression are situated far below the curve, namely between −7.00 and −4.40 m NN, with their ages ranging between 3518 and 3355 cal. a BC. Behre uses points 32–34 (at −3.93, −3.60 and −3.20 m NN) to argue for compaction, as he assumes that at points 26–30 MHW reached the same altitude, with the curve indicating that a correction of 2 m has been applied. However, the ages of points 32–34 range between *c.* 2660 and 2770 cal. a BC, and they therefore actually coincide with his overlying Regression I (cf. Fig. 2). Point 32, which is from the top of basal peat in Förher Watt (Fig. 1), is not a suitable index point, as Hoffmann (1980) describes the basal peat here as being clearly eroded. On the island of Förher Watt and on the neighbouring island of Sylter Watt, three other dated basal peats at an altitude of between −5.30 and −5.13 m NN are available (Averdieck 1980; Hoffmann 1980), but are not used by Behre. At Föhrer Watt, the basal peat is younger (*c.* 4400 cal. a BC) than that at Sylter Watt (*c.* 5700 cal. a BC), although from a similar altitude. This demonstrates the importance of local factors in controlling peat growth, even over short distances. The index points for the C III transgression (points 24–30) are spread over ~100 km and are located in different coastal settings, which, moreover, have different tidal ranges and orientations to the open sea

	Calibrated age AD (+) BC (–)	MHW m NN	Duration (years)	Amplitude (m)	Rate of Change (mm a^{-1})
Dunkerque IV		+1.70	300	+1.30	4.3
	+1700				
Regression 7		+0.40	250	–1.00	4.0
	+1450				
Dunkerque IIIb		+1.40	350	+1.40	4.0
	+1100				
Regression 6		±0.00	250	–0.80	3.2
	+ 850				
Dunkerque IIIa		+0.80	150	+0.30	2.0
	+ 700				
Regression 5		+0.50	350	–0.35	1.0
	+ 350				
Dunkerque II		+0.85	300	+1.50	5.0
	+ 50				
Regression 4		–0.65	200	–1.25	6.2
	– 150				
Dunkerque Ib		+0.60	250	+2.20	8.8
	– 400				
Regression 3		–1.60	400	–0.20	0.5
	– 800				
Dunkerque Ia		–1.40	200	+0.20	1.0
	–1000				
Regression 2		–1.60	500	–1.60	3.2
	–1500				
Calais IV		±0.00	900	+2.50	2.8
	–2400				
Regression 1		–2.50	600	–0.50	0.8
	–3000				
Calais III		–2.00	900		
	–3900				
	–4150				
Calais II		–5.00	1200		
	–5350				
Calais I		–10.00			
	(–6650)				

Fig. 2. Ages and MHW levels of the Calais and Dunkerque transgressions and separating regressions as shown in table 1 of Behre (2007). The duration, amplitude and rate of sea-level change have been added for the post-3000 cal. a BC transgressions and regressions.

(Emden, the Jadebusen, Wesermarsch and North Frisian Islands; Fig. 1). The key point to draw the curve higher (at −2.00 m NN at *c*. 3000 cal. a BC, cf. Fig. 4) than index points 24–30 is point 35, which is from the base of the Upper Peat (= Regression 2). However, according to Behre's table 1 (Fig. 2), Regression 2 lies on top of the deposits of the C IV transgression.

The C III transgression is followed by the 'first general fall of the MHW' (Behre 2007: p. 87), called Regression I (between *c*. 3000 and 2400 cal. a BC) and represented by an intercalated (the middle) peat at −2.5 m NN (Fig. 2). However, it is not explained how this peat bed could develop (at an altitude of −2.50 m NN) on top of the deposits of the C III transgression, which are said to lie 0.50 m higher. Behre describes this middle peat as a common and synchronous phenomenon along the German coast, which extends beyond the present-day coastline and therefore cannot be regarded as being caused solely by local processes. It is accepted that peat development over this period is not a local phenomenon; however, other interpretations are possible (see below). Peat development has to be seen in the context of factors such as the general trend of the RSL rise and the amount of sediment and accommodation space (the height difference between a sediment surface and high tide) available. Other authors (e.g. Denys & Baeteman 1995; Long *et al*. 2000; Beets *et al*. 2003) indicate that this period coincides with a reduction in the rate of RSL rise.

Regression 1 is followed by the C IV transgression, which starts at *c*. 2400 cal. a BC and is said to represent

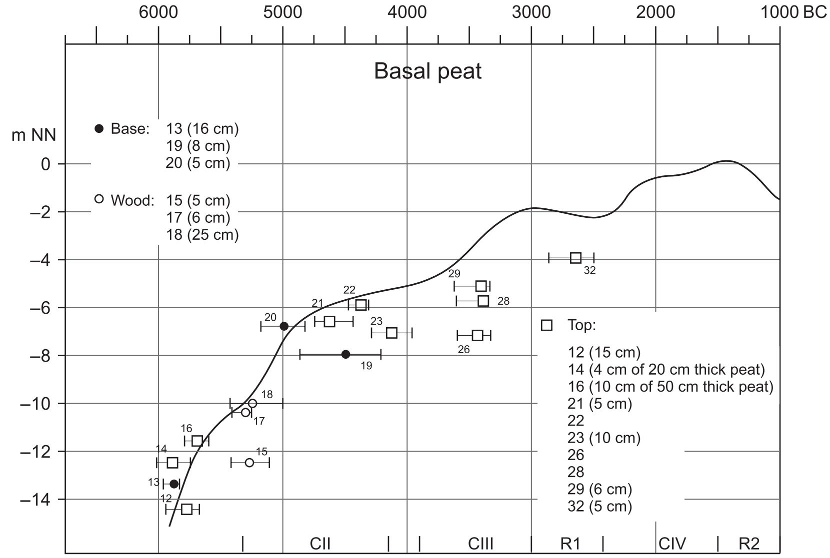

Fig. 3. Basal peat index points: showing the 1σ range of the calibrated dates and the sampling position within the peat and the sample thickness (where specified in the original publications). The sea-level curve and the Calais transgressions (CII, CIII, CIV) and regressions (R1, R2) are redrawn from Behre (2007).

another considerable rise in sea level. Of five dates available from the base of the 'Upper Peat' overlying C IV, one is selected (point 36) to draw the curve at −1.0 m NN at *c*. 2150 cal. a BC (Fig. 4). However, the maximum altitude of the C IV transgression is given as 0.0 NN, and the start of Regression 2 (Upper Peat) placed at *c*. 1500 cal. a BC (Behre 2007: table 1; Fig. 2), seemingly on the basis of point 40. The latter date is a bulk sample of 8 cm from the whole of the Upper Peat collected during geological mapping, with an inexact altitude.

Regression 2 ('Upper Peat') is described by Behre (2007) as the best known phenomenon in the Holocene sediment succession along the southern North Sea, covering very large areas (*sic*, Behre 2007: p. 90). Such a widespread change from marine to freshwater conditions, it is stated, can only be explained by a fall in sea level. The Upper Peat started to grow between 1550 and 1300 cal. a BC (p. 90), and at the height given (−1.60 m NN, Behre 2007: table 1; Fig. 2) would represent a considerable fall in MHW. However, the five dates from the base of the Upper Peat (points 35, 36, 37, 39, 41) vary from *c*. 2500 to 1300 cal. a BC and they have an altitudinal range of between −2.32 and −0.58 m NN (see Fig. 4). The altitude of Regression 2 (−1.60 m NN, Fig. 2) was established on the basis of the first settlements in the German Clay District at Rodenkirchen (Weser area), although it is acknowledged they start only at the beginning of the subsequent period of sea-level rise. The lowering of the groundwater level by as much as postulated (1.60 m) could be expected initially to be followed by a phase of soil development rather than peat growth. Behre (2007) notes that in many places changes occurred in the composition of the peat-forming vegetation from reed to sedge, and ending in

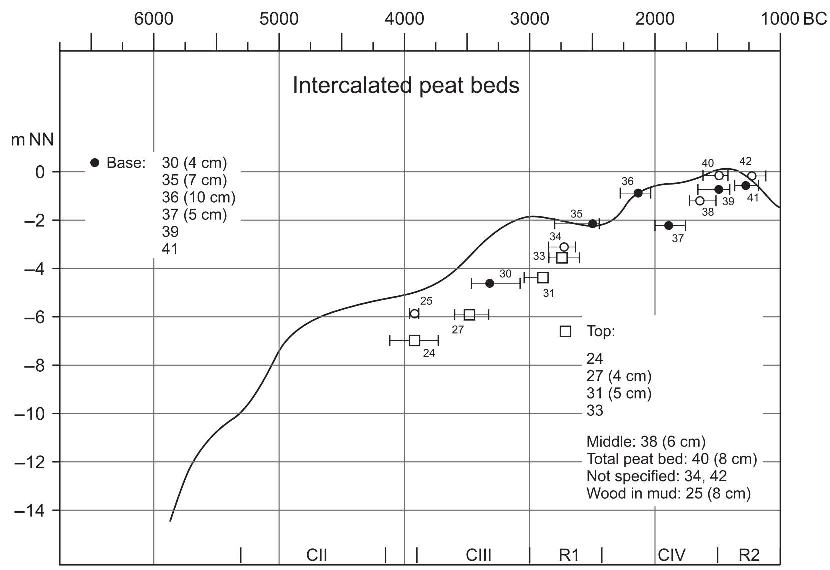

Fig. 4. Intercalated peat index points: showing the 1σ range of the calibrated dates and the sampling position within the peat and the sample thickness (where specified in the original publications). The sea-level curve and the Calais transgressions (CII, CIII, CIV) and regressions (R1, R2) are redrawn from Behre (2007).

fen peat. Such sequences are not necessarily indicative of a falling watertable but may simply result from sediment accumulation (see below).

The Dunkerque series

Regression 2 is followed by the Dunkerque transgressions. The identification of six transgressions and five regressions over a period of about 2700 years rests largely on archaeological evidence, with a few age determinations from intertidal shells and peat beds within coastal barrier deposits. The archaeological data used by Behre are occupation layers from which the height of the earliest houses is used, data largely compiled by Brandt (1980) (Fig. 5). Brandt (1980) indicated that the original altitude of these data points must have been higher, because of compaction and probably crustal movements, and showed that settlements of the same period occur at different elevations. To correct to local MHW level, Behre lowers the original height of most of these points by 1 m and consequently draws his curve below the data points (Fig. 5). However, in the case of occupation layers there is no need to correct to local MHW because human settlements on salt marshes are likely to be at least at MHW level.

The D I transgression is subdivided into D Ia and D Ib. During D Ia, sea level rose by 20 cm over a period of 200 years (Fig. 2). It is considered as a weak transgression because the Upper Peat continued to grow over that period in areas away from the coast. D Ia and D Ib are considered as separate transgressions, as the oldest house of a settlement erected on top of the earliest D Ia deposits at *c.* 900 cal. a BC was followed by a second occupation phase around *c.* 800 cal. a BC built on a thin clay layer from an intervening flooding episode. The settlement concerned is Rodenkirchen, which was also used to document the maximum altitude (−1.60 m NN) of Regression 2. This classical archaeological site was discussed by Petzelberger (2000) with respect to sediment compaction. A 2-m-thick peat bed occurs under the site, within a 10-m-thick Holocene sediment succession. The archaeological excavation also revealed a ditch constructed in the Middle Ages. Compaction of the peat underlying the Bronze Age settlement is likely to have occurred as a result of the thicknesses of compressible strata and drainage associated with the construction of the ditch. As a result, both the interpretation of the thin clay as being caused by a rise in sea level and the use of this site to determine the altitude of sea level during Regression 2 are thrown into doubt. To support a fall in sea level between D Ia and D Ib (Regression 3), Behre simply refers to settlements established on former salt marshes from around *c.* 600 cal. a BC in the Netherlands.

During the subsequent D Ib transgression, between *c.* 400 and 150 cal. a BC, sea level is said to have risen from −1.60 to +0.60 m NN (Figs 2, 5). The coastline was characterized by elevated levees, also called ridges by Behre, although their context is not discussed. The altitude of these features has been used to estimate the MHW level at +0.60 m NN. The 2003 publication provides more detail (p. 35), with the maximum altitude of D Ib drawn at +0.60 m NN but the highest index points for D Ib shown at an elevation slightly below 0 m NN (Fig. 6). Both samples (70 and 75) are from intertidal shells and originate from the same island, although Behre applies different altitudinal corrections (15 and 30 cm, respectively). In addition, if the ages given in the original publication (which include a necessary correction for the reservoir effect) are followed then these index points actually coincide with Regression 4 and the D II transgression that follow (Fig. 6). Point 70, which is derived from *Hydrobia ulvae*, should

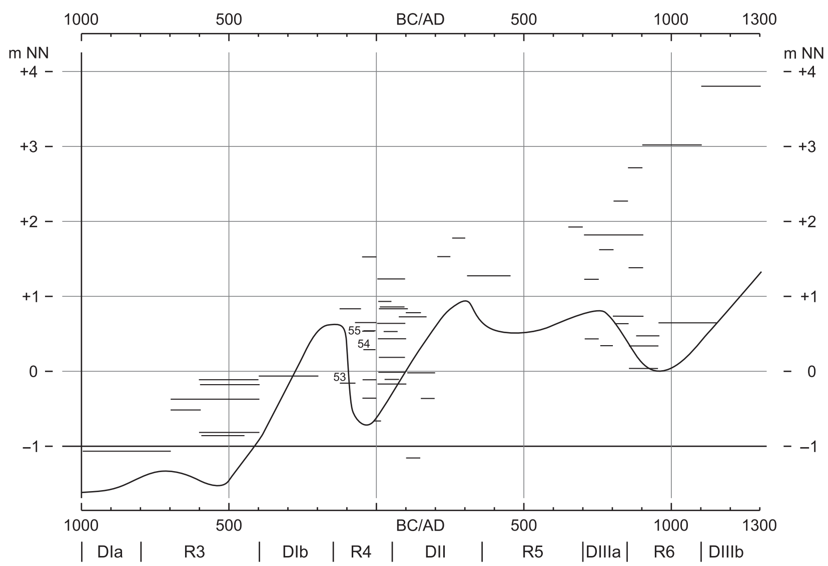

Fig. 5. Age range and altitude of the earliest occupation layers from the marshes between the Ems and the Elbe mouth redrawn from Brandt (1980). The earliest houses used as index points by Behre (2003, 2007) are numbered. The data show the wide altitudinal range at which broadly contemporary occupation layers now occur. The sea-level curve and the Dunkerque transgressions (DIa, DIb, DII, DIIIa, DIIIb) and regressions (R3, R4, R5, R6) are redrawn from Behre (2003).

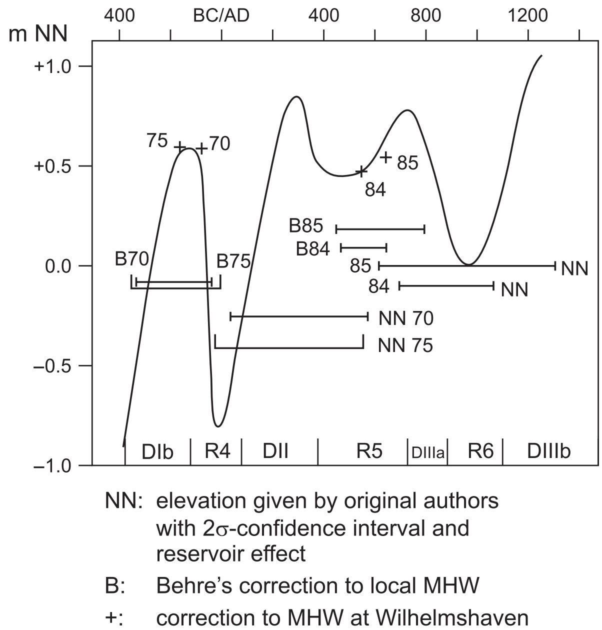

Fig. 6. Age and altitude of the tidal shells used as index points by Behre (2007). Altitudinal corrections by Behre to local MHW and to MHW Wilhelmshaven are shown, along with the altitude given by the original authors. The age errors shown on the latter are based on the 2σ confidence interval and the marine reservoir effect, compared with the age errors with the 1σ confidence interval used by Behre (2007). The index points used to define D1b and R5 clearly lack chronological precision and appear to post-date the sea-level fluctuations they are used to define. The sea-level curve and the Dunkerque transgressions (DIb, DII, DIIIa, DIIIb) and regressions (R4, R5, R6) are redrawn from Behre (2003).

not have been used as a sea-level index point because, as stated by Freund *et al.* (2004), such gastropods can float for a long time in the water and be transported over great distances.

Following the D Ib transgression, Behre indicates a fall in sea level by 1.25 m (Regression 4, *c.* 150 cal. a BC to AD 50, Behre 2007: table 1; Figs 2, 6). The arguments for a lowering of sea level, which apparently was rapid and sudden, are a change from fen to raised bog, the development of a widespread soil, and settlements on non-elevated marshland surfaces. Three occupation layers (53–55), which have been unnecessarily reduced in altitude by Behre, are used to set the minimum of the regression at −0.65 m NN, with the curve drawn beneath almost all of the index points (Fig. 5).

Of the many index points shown in the younger part of the sea-level curve (Behre 2007: fig. 3), few are discussed in the context of the D II transgression. The start of D II is given as *c.* AD 50 on the basis of the presence of a pier on a gully, at −0.15 m NN (−0.25 m NN after correction to local MHW, point 66), dated to *c.* AD 35. The maximum altitude of the D II transgression is drawn on the basis of one index point (81, cf. Fig. 7A), a dwelling mount. A major argument for a sea-level rise is the building of such mounds in response to increasing storm flood levels. An estimated difference of *c.* 1 m between MHW and 'general storm flood level' (Behre 2007: p. 94) was used to provide a maximum altitude of the D II transgression of +0.87 m NN. Regression 5, which follows, starts at AD 350 and is drawn at +0.45 m NN on the basis of two occupation layers (points 82, 83) along the Weser and ages derived from two *Scrobicularia plana* shells (points 84, 85) on the island of Wangerooge (Fig. 1). As illustrated in Figs 5 and 6, both the archaeological data and the shell dates have a wide age range. Both the vertical estimates of sea-level change and precise periods of transgression/regression inferred are therefore questionable.

The D IIIa transgression (between AD 700 and 850) is based on a further rise in the height of dwelling mounds. Behre (2007) indicates that there are few reliable indicators for the maximum altitude of this transgression, but it is said that the minimum was above +0.80 m NN. However, the seven index points presented in his data set (table 3) range in altitude between −0.50 and +0.60 m NN (after correction to local MHW; see Fig. 7A). Thereafter, with the construction of sea defences, archaeological sources are acknowledged by Behre (2007) as having limited value in sea-level reconstruction. This lack is compensated for by dates from fossil vegetation horizons from the island of Juist dated between AD 1120 and 1452 (points 99–101 and 104–111, Fig. 7A). In principle, barrier islands are suitable locations for sea-level studies because of the compaction-free subsoil. According to diatom and pollen analyses (Freund *et al.* 2004), these vegetation horizons developed in the upper part of low salt marsh and on high salt marsh. Some of them are likely to have originated at even higher levels because they developed on storm-surge deposits. Behre should have selected which of the sub-fossil horizons he used because, as noted by Freund *et al.* (2004), some of the samples used as index points have probably been contaminated. At a number of locations, several horizons were found superimposed one upon another, and penetration with younger rootlets seems therefore possible. Fig. 7B shows altitude and age ranges of the vegetation horizons from the data presented by Freund *et al.* (2004). There are differences in the altitude of broadly contemporaneous deposits of more than 1 m, and the index points bare no obvious relationship to Behre's curve. To determine their relationship to sea level more accurately, each sample dated should have been analysed individually to establish its indicative range. Freund *et al.* (2004) undertook such an exercise and present the minimum and maximum levels of mean high tide for the last 2000 years based on palaeoecological analyses. They demonstrate a slow rise of mean high tide level from 0 to +0.40 m NN over the period between 340 cal. a BC and AD 1015, with a short phase of lowering at about AD 1350. Freund *et al.* (2004) suggest that the short period of lowering may be the result of the

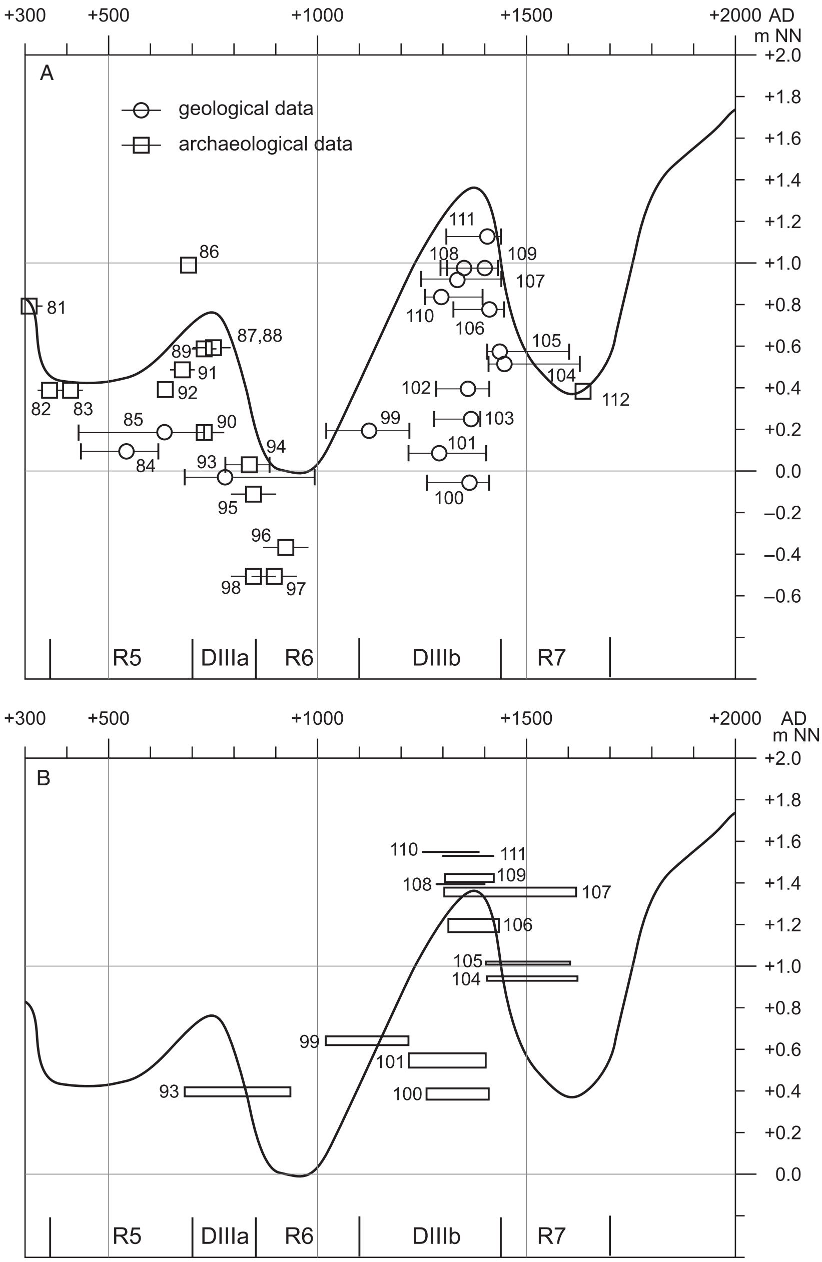

Fig 7. Index points used by Behre (2007) to reconstruct sea level between AD 300 and 1650. A. All geological and archaeological points from Behre (2003, 2007). B. The vegetation horizons from the island of Juist presented by Freund *et al.* (2004), with the age ranges and altitudes from the original publication. Comparison of the figures shows the effect of Behre's unnecessary correction to local MHW and then to MHW Wilhelmshaven, which is the basis for the sea-level curve. The data also illustrate differences in the elevation of broadly contemporaneous deposits. The sea-level curve and the Dunkerque transgressions (DIIIa, DIIIb) and regressions (R5, R6, R7) are redrawn from Behre (2003).

creation of new tidal bays, such as Jade Bay and the Dollart, following the breaching of dikes during the Middle Ages.

The ages of the vegetation horizons of the Frisian Islands are used to set the end of the D IIIb transgression at *c*. AD 1450. According to Behre (2007), there followed a rapid 1-m fall in MHW (to +0.40 m NN, Regression 7) that lasted until AD 1700. This is based on index point 112 alone (Fig. 7A), which is from the land surface beneath a dike. The dike, at Sande, is said to rest on Pleistocene deposits; however, a contour map showing the base of the Holocene deposits in this area (Streif 1981) indicates that Pleistocene material does not crop out and that the thickness of the Holocene deposits is between 3 and 6 m. With dike building implying artificial drainage, the surface is likely to have undergone considerable subsidence. Regression 7 therefore has no reliable basis.

This analysis of the data used by Behre (2003, 2007) to construct his sea-level curve leads us to conclude that the fluctuations are largely a consequence of the application of the Calais–Dunkerque system and its underpinning rationale, which, by the production of a curve with fluctuations, Behre seeks to perpetuate in the face of its abandonment by most North Sea coastal scientists.

The concepts underpinning the construction of sea-level curves

From the detailed analysis above it is clear that many of the concepts and methods underpinning the sea-level curve of Behre (2003, 2007) require critical re-evaluation. We discuss first the problems associated with the interpretation of index points (age and altitudinal errors and their indicative meaning), before considering issues that relate to the geographical extent from which Behre collated his data (changes in palaeotidal range and crustal movements).

Age and altitude errors

Many of the radiocarbon dates used by Behre were obtained in the 1970s and 1980s for the purpose of geological mapping and reconstructing coastal evolution. All such age estimates were derived using bulk samples and the conventional radiocarbon method. These dates lack precision. More than a third of the ages have a standard deviation larger than 100 years. The sampling interval ranges between 4 and 16 cm (as indicated in Figs 3 and 4). For basal peats, bulk-sample sources of age error include contamination with older carbon from underlying palaeosols (Törnqvist *et al.* 1998). Obtaining accurate dates for the end of peat growth is notoriously difficult. Waller *et al.* (2006) show that some erosion and reworking is always likely during the initial stages of inundation, and the tops of peat beds may also be subject to root penetration (Streif 1971; van de Plassche 1980; Törnqvist *et al.* 1992).

With respect to altitudinal uncertainty, most importantly, Behre fails to acknowledge the pervasive influence of compaction. Compaction has long been recognized as having a major impact on the altitude of sea-level index points (e.g. Bennema *et al.* 1954; Skempton 1970; Paul & Barras 1998; Allen 1999, 2000a), with highly compressible peat and fine-grained minerogenic deposits more susceptible to lowering than sands (van Asselen *et al.* 2009). Peat can compact by up to one-eighth of its original thickness as a result of overburden loading – a continual process that has an ongoing effect on the structure and strength of the peat fabric (Kay & Barghoorn 1964; van de Plassche 1980). Behre (2007: p. 83) acknowledges the potential influence of compaction on the altitude of index points from intercalated peat beds, while stating that data collected from such beds have been used if they developed on sand, or if it can be assumed that compaction came to an end before peat formation started. However, Behre does not consider that tidal-flat deposits and peat compact as soon as fresh sediment accumulates on top, or that compaction by simple de-watering continues, although at an exponentially decreasing rate, for thousands of years. The initial compaction of water-saturated fine-grained sediments with low permeability takes considerable time, after which period compaction continues as a result of secondary consolidation (Baeteman 1994).

Behre also fails to consider the overall stratigraphic context of index points. The thickness of the sediment column and morphology of the pre-Holocene subsurface are influential, as deeper Holocene deposits will continue to compact, lowering the overlying succession whether it is organic or clastic. Considerable local altitudinal variation in the top and base of spatially extensive intercalated peat beds as a result of factors such as thickness of the overlying and underlying Holocene deposits, and subjacent lithological variability, have been documented in cross-sections. For example, from the area where many of Behre's data points originate, Streif (1972), Barckhausen & Streif (1978), Preuss (1979) and Freund *et al.* (2004) report variations in the elevation of a peat bed of between 2 and 5 m. Variations of this magnitude have been documented elsewhere by Waller (1994), de Groot & de Gans (1996), Bertrand & Baeteman (2005), Long *et al.* (2006) and Törnqvist *et al.* (2008). Estimates of compaction by Behre (2007) seem far too low (e.g. up to 0.2 cm: p. 91). Significantly greater lowering of the altitude of the intercalated peat beds from which Behre's index points derive can be expected. For example, in the area west of the Jade Busen, the Holocene sediment succession is between 10 and 17 m thick (Streif 1981), while around Emden, the geological map indicates thicknesses of between 4 and 6 m (Barckhausen & Streif 1978). The order of magnitude of the altitudinal error is likely to be as much as 3 to 5 m.

Currently, compaction is a source of error that cannot be quantified, despite attempts to use models of decompaction calibrated against field data (Allen 2000b; Edwards 2006). Much of the information required, such as the initial state of the sediment (including composition, moisture content, pore-water space) and the water and sediment loading histories, simply cannot be obtained.

Further altitudinal error is likely, as with some index points the altitudinal reference given by the original authors is depth below the modern surface, with correction to any datum inevitably therefore an approximation. Twenty-one of the dates used by Behre, taken from the literature, were collected as a result of systematic geological mapping. As noted above, they were not intended to serve as sea-level index points, with the altitude of the boreholes inferred from topographic maps giving errors of at least 10 to 25 cm (H. Streif, pers. comm).

Interpretation of peat beds

Intercalated peat beds are generally considered by Behre (2007) as representing a regression both in the sense of a retreat of the sea and a vertical decline in sea level. It is stated that intercalated peat beds that are synchronous across wide areas signify with certainty a fall in sea level. However, neither the broader context of a decelerating rate of RSL rise in the middle and late Holocene (e.g. Baeteman 1999; Beets & van der Spek 2000; Long *et al.* 2000; Beets *et al.*, 2003) nor the interplay between the rate of RSL rise, sediment supply and accommodation space is considered. When sediment supply surpasses the creation of accommodation space by RSL rise, tidal basins will rapidly accrete sediment to highwater level. As a consequence, the frequency of tidal inundation declines and drainage networks are liable to become redundant and silt up. Eventually salt marsh encroaches upon mudflat, followed, if sufficient time is available, by peat accumulation. Thus, given sufficient sediment supply, there exists an asymptotic relationship between sediment-accretion rates and time that will produce stratigraphies traditionally interpreted as representing a vertical decline in sea level, even though RSL is rising (Jennings *et al.* 1995).

Interferences on water-level movements made from vegetation successions are similarly open to a number of possible interpretations. For example, changes from reed, to sedge and then wood peat (Behre 2007: p. 90) certainly show that a rise occurred in the sediment surface level in relation to the groundwater level. However, they are not necessarily indicative of falling water levels. Such a sequence would be expected given stable water levels, through the gradual accumulation of organic matter, and may also occur when water levels are rising if the quantity of organic material deposited is able to keep pace. Fen woodland communities produce large quantities of detritus and it can be argued (e.g. Waller *et al.* 1999) that they need rising water levels to be maintained over any length of time (for more than a few hundred years).

The development of acidiphilous vegetation and ombrotrophic bog in coastal lowlands certainly requires isolation from base-rich water, although it is by no means obvious that this inevitably occurs as a result of vertical isolation produced by a prolonged lowering in RSL (as indicated by Behre 2007: p. 90). In the English Fenland and on Romney Marsh (Waller 1994; Waller *et al.* 1999), bog initiation is associated with periods of laterally extensive peat formation, suggesting that spatial isolation from base-rich water is the main requirement, although factors such as the preceding vegetation, the mobility of the peat surface (Giller & Wheeler 1988) and climatic wetness (e.g. van Geel *et al.* 1996) are also likely to be influential. Once established, the growth of a raised bog is dependent upon rainfall, and with the bog surface decoupled from the ground water table any sea-level inferences are negated.

Finally, Behre (2007) is inconsistent, both in terms of using intercalated peat beds for stratigraphic division and in terms of interpretation. Intercalated peat beds in the Weser area have been dated by Preuss (1979) to *c.* 7900 cal. a BC, *c.* 6050 cal. a BC and *c.* 5050 cal. a BC. Logically, extra transgressions and regressions should be added to the Calais series. Another early intercalated peat (dated *c.* 4000 cal. a BC) is used to separate C II and C III, but is not thought to represent a fall in sea level as 'it is unlikely that a fall in sea level occurred during this time' (Behre 2007: p. 85).

The use of archaeological material

For the late Holocene, the sea-level curve of Behre (2003, 2007) is heavily reliant on dates from archaeological contexts. While changes in palaeogeography (which may or may not be directly related to vertical changes in RSL) unquestionably influenced the settlement and the use of coastal areas in the southern North Sea over this period, they were clearly not the only influence (see for example Behre 2004). This lends an inevitable degree of uncertainty to interpretations, and archaeological contexts, unlike index points based on sedimentary and fossil evidence, have no inherent reference water table or indicative meaning. The use of archaeological remains as sea-level markers has been considered by Antonioli *et al.* (2007) and Auriemma & Solinas (2009). The occupation layers of settlements are considered simply to have a 'functional height' of above high tide. The suggestion of Behre (2004) that dwelling mounds were adapted to maximum storm surge level (p. 83) would have required foreknowledge not always matched by modern occupants of the coastal zone. In addition, as noted previously, the present position of specific levels will not coincide with their levels at deposition, owing to consolidation of the underlying sediments and of the dwelling mound itself (see Baardman, in Vos 1999). Antonioli *et al.* (2007) also discuss the use of structures such as piers and quays and define their functional height as 0.6–1.00 m above m.s.l., based on comparison with modern harbour structures in a microtidal environment (the Mediterranean). Behre (2007) gives no such range, and even the vertical range of Antonioli *et al.* (2007) negates most of the sea-level change inferred by Behre (2007: p. 92) from the use of a pier gangplank as an index point.

Changes in palaeotidal range and MHW levels

Crucial steps in sea-level research are (i) to establish the altitude of a sea-level index point relative to a former water level (often MHW), and (ii) to relate this local water level to coastal sea level (MHW or Mean Sea Level (MSL)) (van de Plassche 1982). The second step necessitates the estimation of the difference between the former local water level and sea level at that time. In the case of a sea-level index point obtained from close to or on the open sea (e.g. on a barrier island), the difference between local MHW level and MHW at sea will be minimal, reducing or eliminating step (ii). However, if the site is located inland (e.g. upstream in an estuary), the difference in water levels may be substantial. Factors influencing this difference include flood-basin and river gradient effects (increasing local water levels in an upstream direction along a (tidal) river) (van de Plassche 1982, 1984, 1995; Van Dijk *et al.* 1991). These factors have to be taken into account explicitly when evaluating and interpreting data points, requiring a thorough understanding of the palaeogeography of the study area (the size and shape of estuaries, tidal basins, distribution of supra-, inter- and subtidal areas, etc.).

Behre (2007) interprets his local water-level index points in relation to local MHW. Because MHW differs along the coast and from the coastal zone inland, he then corrects the altitude of his data points to MHW at Wilhelmshaven, using the tide gauge nearest to the location of each index point. To arrive at a common MHW-level curve for Wilhelmshaven, the present-day difference (period 1985–2002) between MHW at this local tidal station and Wilhelmshaven is added to or subtracted from the MHW altitude indicated by the index point (Behre 2003). This procedure implicitly assumes that the difference between MHW at all index point locations (or local tidal stations) and Wilhelmshaven has remained unchanged over time. Behre does not present any evidence for this assumption. However it is highly likely that long-term palaeotidal changes have affected local MHW in parts of Behre's study area, thereby invalidating the assumption of a constant palaeotidal range and MWH levels. In their analysis of the sea-level data set from northwest Germany, based on comparison with geophysical glacial rebound models, Vink *et al.* (2007: pp. 3269–3270) show that differences between observed and modelled water levels in the Ems estuary can be explained by changes in palaeotidal range.

That the changing relationship between MHW and MSL was not taken into account by Behre (2003, 2007) has serious consequences for his sea-level curve. Behre's data points from the inner Ems estuary (Fig. 1) consistently occupy low time-depth positions (see fig. 13 in Behre 2003 and fig. 3 and table 3 in Behre 2007). Index points 45 to 51, which exclusively determine the low position of Behre's Regression 3, are all from the inner Ems estuary. Likewise, index points 53, 56, 57 and 61 from the inner Ems estuary are at or near the 'MHW minimum' of Regression 4. Three out of the four lowermost index points of Behre's Regression 6 are also from the inner Ems estuary (numbers 95, 96 and 97, cf. Fig. 7A). In contrast, the very pronounced MHW maximum of the D Ib transgression ($\sim$2 m higher than the preceding Regression 3, cf. Figs 5, 6) is based on index points 70 and 75 from the barrier island of Borkum, well out at sea in front of the Ems estuary mouth.

It is clear that the correlation between time-depth and location within the estuary is a consequence of the procedure of Behre (2003, 2007) to 'correct' index points to a common MHW level at Wilhelmshaven while assuming unchanging local MHW levels relative to MHW at sea. If the local MHW level was lower in the past, the upward correction applied to the index point, based on the present-day MHW difference with Wilhelmshaven, will be too small, and the index point will plot at an altitude that is too low. The resulting altitude error depends on the inland MHW-level reduction relative to MHW at sea, which in the extreme case may be up to about half the coastal tidal range. For the Ems area, this may amount to a maximum error of $\sim$1.5 m, assuming a coastal tidal range of $\sim$3 m. The actual value may be less, mainly depending on the palaeomorphology of the estuary and the resulting palaeotidal conditions. Detailed palaeogeographical reconstructions and assessments of the within-estuary evolution of water levels are required before the magnitude of this flood-basin effect can be estimated (e.g. van de Plassche 1982, 1984, 1995; Kiden 1989, 1991, 1995; Vos & van Heeringen 1997).

The above analysis supports the hypothesis of Vink *et al.* (2007) that, for a large part of the middle and late Holocene, strong tidal dampening occurred in the inner Ems estuary, reducing the tidal range and lowering MHW levels relative to MHW at sea. It is noteworthy that a comparably strong flood-basin effect has been observed for the same periods in the Schelde river in northern Belgium and the southwestern Netherlands (Kiden 1991, 1995, 2006), a river that compares well to the Ems in terms of size, fluvial discharge and coastal tidal range.

Consideration of changes in palaeotidal range and MHW levels alone calls into question the validity of the sea-level curve presented by Behre (2003, 2007), regarding not only the amplitude of the reconstructed sea-level fluctuations but even their very existence.

The influence of glacio-isostatic crustal movements

The construction of a new sea-level curve for the southern North Sea by Behre (2003, 2007) is based on the assumption that the German North Sea coastal area is 'the most stable one with respect to tectonic and isostatic movements' (Behre 2007: p. 82). Glacio-isostatic movements cause subsidence and uplift of the Earth's crust owing to loading and unloading by ice. Only after presenting his sea-level reconstruction does Behre provide a limited evaluation of the influence of such crustal processes. His estimate of the long-term tectonic subsidence of the German North Sea since the Eemian of between 0.64 and 0.54 cm per century is comparable to values calculated by others (e.g. Vink *et al.* 2007). However, for isostatic movements, Behre states that 'the available measurements and estimates suggest a subsidence of <0.1 cm/century at present and probably also for the Holocene as a whole. This includes tectonic as well as isostatic movements.' As will be argued below, with this assumption Behre seriously underestimates the magnitude of isostatic movements. Moreover, isostatic influences on the altitude of the sea-level index points should have been taken into account by Behre during data evaluation and interpretation, prior to the construction of the curve itself.

A considerable body of knowledge exists on the influence of isostatic movements on global and regional sea-level records. Since the seminal work of Walcott (1972), Farrell & Clark (1976) and Clark *et al.* (1978), it has become clear that glacio-isostatic crustal movements as a response to the growth and decay of the large ice sheets have influenced regional sea-level histories worldwide, and that no region on the globe has been free from these effects. These initial global studies were followed in northwestern Europe and the North Sea region by increasingly sophisticated integrated sea-level and geophysical modelling investigations (Fjeldskaar 1994; Lambeck 1995; Lambeck *et al.* 1998; Shennan *et al.* 2000a, 2000b). Research along the east coast of the southern North Sea has shown that the differences between the sea-level records of Belgium, the Netherlands and Germany, as well between these areas and adjacent parts of the southern North Sea, can largely be explained by variations in glacio-isostatic subsidence (Kiden *et al.* 2002; Vink *et al.* 2007). The effect of isostatic subsidence on (late) Holocene sea-level change has also been demonstrated for Ho Bugt at the northern boundary of the Danish Wadden Sea area (Gehrels *et al.* 2006b).

The comprehensive study of Vink *et al.* (2007) is particularly relevant. It uses the data set of Behre (2003, 2007) for northwestern Germany as part of a larger sea-level database, in combination with geophysical glacial rebound models. The data assessment by Vink *et al.* (2007) expands on and confirms the results of Kiden *et al.* (2002), and reveals a complex pattern of differential crustal movement between Belgium, the Netherlands, northwestern Germany and the southern North Sea that cannot be attributed solely to tectonic activity. The pattern of movement contains a clear non-linear, isostatic subsidence component, which is small on the Belgian coastal plain but increases significantly towards the northeast in the direction of the Fennoscandian land mass. Vink *et al.* (2007) conclude that northwest Germany was subjected to a total isostatic lowering relative to Belgium of ~7.5 m between 6000 and 2800 cal. a BC, after which differential isostatic subsidence can no longer be unambiguously identified from the sea-level reconstructions. Furthermore, their modelled RSL data suggest that the zone of maximum isostatic subsidence runs in a relatively narrow, WNW–ESE-trending band from Lower Saxony in Germany to the Dogger Bank area in the southern North Sea. Vink *et al.* (2007) thus convincingly show that the German North Sea coast is probably the area with the largest Holocene isostatic subsidence in the southern North Sea, contrary to the claim of Behre (2003, 2007).

The above-mentioned results invalidate the initial assumption of Behre (2003, 2007) that the German North Sea coastal area is stable with respect to Holocene isostatic crustal movements. This has serious consequences for (i) the validity and (ii) the accuracy of the sea-level curve of Behre (2007), which operate at two different spatial scales.

First, as significant differences in isostatic movements exist between the German North Sea coast and other parts of the southern North Sea area, the sea-level curve of Behre (2003, 2007) can be valid only for the study area in northwestern Germany and not for other regions around the southern North Sea with different isostatic subsidence histories.

Second, differences in isostatic movement within Behre's study area along the German North Sea coast itself affect the relative altitude of his sea-level index points, depending on their location and age. Behre does not correct for these differential movements between his index points. Instead, he amalgamates all index points into one single curve, which at best will not be representative of the sea-level history at any single location, and at worst suggests long-term trends and short-term fluctuations that do not correspond to physical reality.

The latter effect was demonstrated by Lambeck (1997) in his study of sea-level change along the French Atlantic coast, for which Ters (1986) had constructed a strongly fluctuating sea-level curve. Lambeck (1997) showed that the fluctuations were artifacts of the amalgamation of sea-level observations from regions along the coast with different isostatic subsidence histories:

> These predictions illustrate well how oscillations of the order of several meters could be introduced into sea-level curves if data points from the entire French Atlantic coast, non-uniformly distributed in space and time, were combined into a single curve without first applying the corrective terms associated with the unloading of the former ice sheets. The resulting fluctuations would not reflect changes in ocean volume but represent artifacts arising from the neglect of the response of the Earth to the changing surface load. [. . .] if those artifacts in the eustatic sea-level curve are to be avoided it will be necessary to either restrict the analysis of the field data to small regions or to correct the observed age-height relations for the isostatic factors
>
> (Lambeck 1997: p. 11)

He concludes that 'sea-level change along the French Atlantic margin can be expected to exhibit considerable spatial variability because of the effects of isostasy. This variability in Holocene time can attain ± 5 m and observations from the region should not be combined into a single curve if an accurate measure of sea-level change is sought' (Lambeck 1997: p. 18).

Likewise, Bungenstock & Weerts (2010) have taken up the suggestion of Vink *et al.* (2007), breaking down the data set of Behre (2003, 2007) into five regions for which separate sea-level curves are constructed. However, it should be noted that the authors adopted the age and altitude errors in the Behre data set. The resulting regional curves all differ from the original curve of Behre (2007) and from each other. Most of the sea-level fluctuations in the original curve of Behre cannot be supported and are explained as data artifacts or local effects.

Middle and late Holocene sea-level fluctuations

The curve of Behre (2003, 2007) is characterized by sea-level fluctuations of large amplitude. They are most evident over the last 3000 to 4000 years, when they have durations of between 150 and 500 years and amplitudes of up to 2 m. Owing to the short durations and the large magnitudes, the rate of sea-level rise or fall during these fluctuations reaches values of up to 8 mm a^{-1}, with values of 3 to 4 mm a^{-1} being common (calculated from Behre's table 1, cf. Fig. 2). These rates are very high compared with the average rate of sea-level rise over the last 3000 years of 1.1 mm a^{-1}, as calculated by Behre (2007) himself.

The cause of the rapid and large fluctuations as reconstructed by Behre (2003, 2007) is left largely unexplained. Behre (2007) simply suggests a connection to global climate change, the Little Ice Age, for his last sea-level fall (Regression 7) and the subsequent rise to present-day sea level (the D IV transgression). The large amplitude and rate of the sea-level fluctuations as reconstructed by Behre, however, are not consistent with known mechanisms of sea-level change.

The most important processes likely to be responsible for changes in MSL on the time scale of centuries are ocean density changes owing to thermal expansion or contraction (the steric effect) and changes in land ice mass (Church *et al.* 2001). Ocean thermal expansion is a relatively slow process. The large heat capacity of the ocean means that there will be a considerable delay before the full effect of surface warming is felt throughout the depth of the ocean. As a result, sea level will continue to rise even when the rate of global temperature rise remains the same or decreases (Church *et al.* 2001; Van den Hurk *et al.* 2006). Moreover, it takes a relatively large temperature increase to produce a significant amount of sea-level rise on a century time scale.

Together with the slow response and long lag time, this makes steric sea-level change an unlikely mechanism to account for the high-amplitude sea-level fluctuations proposed by Behre, some of which consist of a very rapid sea-level rise followed by an equally rapid sea-level fall within a time span of 400 to 500 years.

The contribution to sea-level rise by changes in land ice mass can be subdivided into three main components: small ice caps and mountain glaciers, and the large ice sheets of Greenland and Antarctica. Mountain glaciers and ice caps are comparatively sensitive to climate change. Changes in their mass lag behind climate change by a few years to several centuries, and they are capable of producing an important contribution to the rate of sea-level change (Church *et al.* 2001; Lemke *et al.* 2007). However, the total amount of water contained in glaciers and ice caps is low: at present equivalent to between 0.15 and 0.37 m of global sea-level rise (Lemke *et al.* 2007). On the other hand, the large Greenland and Antarctic ice sheets contain 7.3 and 56.6 m of sea-level rise equivalent respectively (Lemke *et al.* 2007), but have much longer response times to climate change, of the order of 100s to 1000s of years (Church *et al.* 2001).

As the sea-level curve of Behre represents MHW rather than MSL, one could argue that the maxima of the curve are periods with increased storminess, which might raise (extreme) MHW levels but not necessarily affect the position of MSL. However, this is not the case. Behre has constructed a MHW instead of a MSL curve for the sake of convenience: 'For coastal areas, in particular along flat shores, the use of MSL is not practical because its line runs in front of the coast; therefore, the mean high water level (MHW) was chosen for construction of the sea-level curve' (Behre 2007: p. 83). Consequently, Behre's MHW curve must be considered equivalent to a MSL curve, only vertically translated: 'Therefore, to obtain a MSL curve, 1.70 m must be subtracted from the values presented here' (Behre 2007: p. 99).

The large-amplitude centennial-scale sea-level fluctuations proposed by Behre are also in disagreement with recent high-resolution sea-level studies, many of which use a multi-proxy approach to reconstruct the course of middle and/or late Holocene sea-level rise and investigate the existence of century-scale sea-level fluctuations and climate/sea-level connections (van de Plassche 1991, 2000; Varekamp *et al.* 1992; van de Plassche *et al.* 1998; Gehrels 1999; Gehrels *et al.* 2005, 2006a; Szkornik *et al.* 2008; González & Törnqvist 2009). The general conclusion that can be drawn from these studies is that centennial-scale sea-level fluctuations have a maximum amplitude of the order of decimetres, which is at the resolving limit of existing methodology, and are certainly not greater than ~0.55 m (González & Törnqvist 2009; Gehrels 2010).

Conclusion

Behre (2003, 2007) presents a sea-level curve for the southern North Sea that shows strong fluctuations that are interpreted as vertical movements of sea level. However, neither the methodology adopted nor the underpinning concepts can withstand detailed scrutiny. The importance of sediment compaction in altering the elevation of sea-level index points is greatly underestimated, and the peat layers and archaeological material, from which the sea-level index points used were derived, are open to alternative interpretations. In addition, Behre's 'correction' of his MHW index points to a common MHW level at Wilhelmshaven introduces potentially large altitudinal errors, as it is assumed that the relative palaeo-MHW levels at the time of formation of the index points were the same as they are for the present-day. The assumption of Behre that the German North Sea coastal area was stable with respect to Holocene isostatic movements can also be challenged, and this region appears to have undergone considerably more isostatic subsidence than areas to the south and north. As a result, Behre's sea-level curve cannot be considered representative for the southern North Sea as a whole. Moreover, the amalgamation of sea-level index points with different isostatic subsidence histories into one single sea-level curve has produced data artifacts that may (partly) be the cause of the high-amplitude middle and late Holocene sea-level fluctuations identified by Behre (2003, 2007). Finally, these high-amplitude sea-level fluctuations require forcing mechanisms or climate changes that are currently unknown. They are therefore highly unlikely to be real features of the sea-level history of the southern North Sea.

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