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Cover: Sea cliff with raised Holocene spit deposits, western side of Skagen Odde in northernmost Jutland. The cliff exposes beach sand deposits with root structures in the upper part formed above the high-water line, followed by a thin rusty horizon and a layer of swale peat (black), overlain by aeolian sand. See this volume pp. 17–28: Clemmensen, L.B., Glad, A.C., Hansen, K.W.T. & Murray, A.S. 2015: Episodes of aeolian sand movement on a large spit system (Skagen Odde, Denmark) and North Atlantic storminess during the Little Ice Age. Photo: Lars Clemmensen.

Agraulos longicephalus and *Proampyx*? *depressus* (Trilobita) from the Middle Cambrian of Bornholm, Denmark

THOMAS WEIDNER & ARNE THORSHØJ NIELSEN

Weidner, T. & Nielsen, A.T. 2015. *Agraulos longicephalus* and *Proampyx*? *depressus* (Trilobita) from the Middle Cambrian of Bornholm, Denmark. © 2015 by Bulletin of the Geological Society of Denmark, Vol. 63, pp. 1–11. ISSN 2245-7070. (www.2dgf.dk/publikationer/bulletin).

The trilobite genus *Agraulos* Hawle & Corda 1847 has within Scandinavia been recorded only from Bornholm, Denmark, where its representatives occur in the Middle Cambrian *Paradoxides paradoxissimus* Superzone of the Alum Shale Formation. Only cranidia have been found so far, representing *Agraulos longicephalus* (Hicks 1872) and the rare "*Agraulos*" *depressus* Grönwall 1902.

The two species from Bornholm are redescribed and discussed based on museum collections in combination with newly collected material from Borggård, Øleå. *Agraulos longicephalus* occurs commonly in the lower and upper part of the *Acidusus atavus* Zone as well as in the *Ptychagnostus punctuosus* Zone. It closely resembles the coeval *Agraulos ceticephalus* (Barrande 1846) known from Bohemia and eastern Newfoundland. A lectotype for "*Agraulos*" *depressus* is designated and re-illustrated; this taxon is hesitantly assigned to *Proampyx*? It is known only from the *Ptychagnostus punctuosus* Zone and may represent an early, atypical *Proampyx* or maybe a precursor that should be separated in a new genus. Emended diagnoses of *Agraulos* Hawle & Corda 1847 and *Proampyx* Frech 1897 are presented.

Keywords: *Agraulos*, *Proampyx*, trilobites, Middle Cambrian, Bornholm, Denmark.

Thomas Weidner [to.we@paradis.dk], Ravnholtvej 23, Rårup, DK-7130 Juelsminde, Denmark. Arne Thorshøj Nielsen [arnet@snm.ku.dk], Natural History Museum of Denmark, University of Copenhagen, Øster Voldgade 5–7, DK-1350 København K, Denmark.

Corresponding author: Arne Thorshøj Nielsen

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The trilobite family Agraulidae is in Scandinavia represented by the genera *Proampyx* Frech 1897 and *Agraulos* Hawle & Corda 1847. *Proampyx* is common and widespread and includes the five species *P. difformis* (Angelin 1851), *P. aculeatus* (Angelin 1851), *P. acuminatus* (Angelin 1851), *P. anceps* (Westergård 1953), and *P. cornutus* Buchholz 1997, all occurring in the Middle Cambrian *Paradoxides forchhammeri* Superzone of Norway, Denmark and Sweden (Brøgger 1878; Grönwall 1902; Westergård 1953) (Fig. 1). Various species have also been reported from glacial erratic boulders of northern Germany (Rudolph 1994; Buchholz 1997). Representatives of *Agraulos* Hawle & Corda 1847 are, on the other hand, rare in Scandinavia and have been recorded solely from Bornholm, Denmark, and from glacial erratic boulders of northern Germany (Grönwall 1902; Rudolph 1994; Buchholz 1997; Weidner & Nielsen 2014). The latter contain a fauna showing that they likely derive from the Bornholm area.

So far, only cranidia of *Agraulos* have been found. Grönwall (1902) reported three cranidia of *A. ceticephalus* (Barrande 1846) and two cranidia of *A. depressus* Grönwall 1902 (here assigned to *Proampyx*?) from the *Ptychagnostus punctuosus* Zone at Borggård, Øleå (Fig. 2). Rudolph (1994) figured one cranidium of *A.* cf. *ceticephalus* (Barrande 1846) from the *P. punctuosus* Zone. There are two further cranidia assigned to *A. ceticephalus* (Barrande 1846) from the same zone in the collection of Buchholz (1997; personal communication 2010). All mentioned material of *A. ceticephalus* is here identified with *A. longicephalus* (Hicks 1872). This species strongly resembles *A. ceticephalus* which occurs in strata of similar age in the Czech Republic (Šnajdr 1958) and eastern Newfoundland, Avalonian Canada (Fletcher *et al*. 2005), and the only conspicuous difference is the outline of the occipital ring (see below). About 25 additional cranidia of *A. longicephalus* were collected during the years 2004–2007 from the older *Acidusus atavus* Zone at Borggård (for details, see

Weidner & Nielsen 2014) and one additional young cranidium of *P.*? *depressus* was collected from the basal layer of the *P. punctuosus* Zone.

The two species presently known from Bornholm, *Agraulos longicephalus* (Hicks 1872) and *Proampyx*? *depressus* (Grönwall 1902), are treated in this paper.

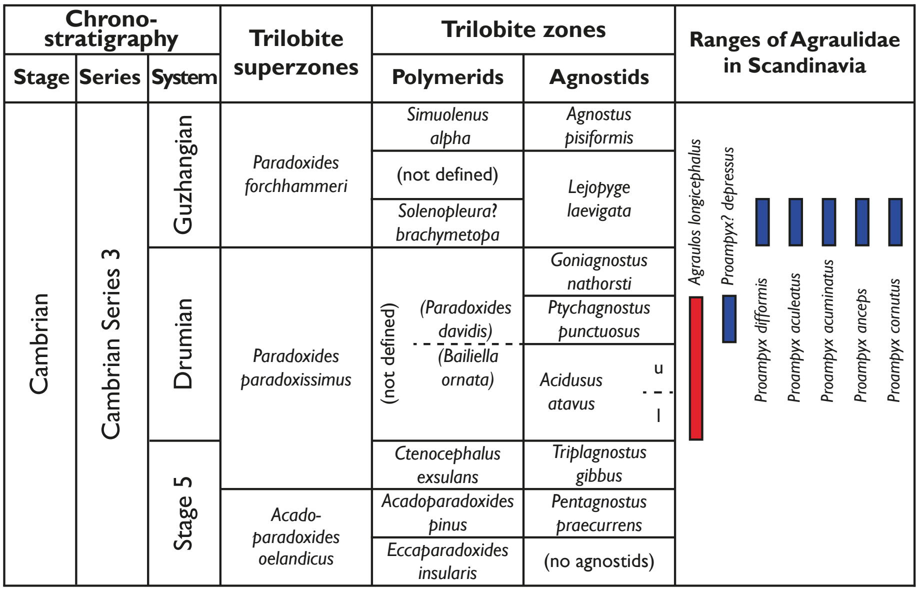


Fig. 1. Biozonation of the Middle Cambrian in Scandinavia. The updated zonation includes data from Axheimer & Ahlberg (2003), Weidner & Nielsen (2014) and Nielsen *et al*. (2014). On the right hand side are shown ranges of Scandinavian taxa. Abbreviations: l: lower part, u: upper part; these units are informal.

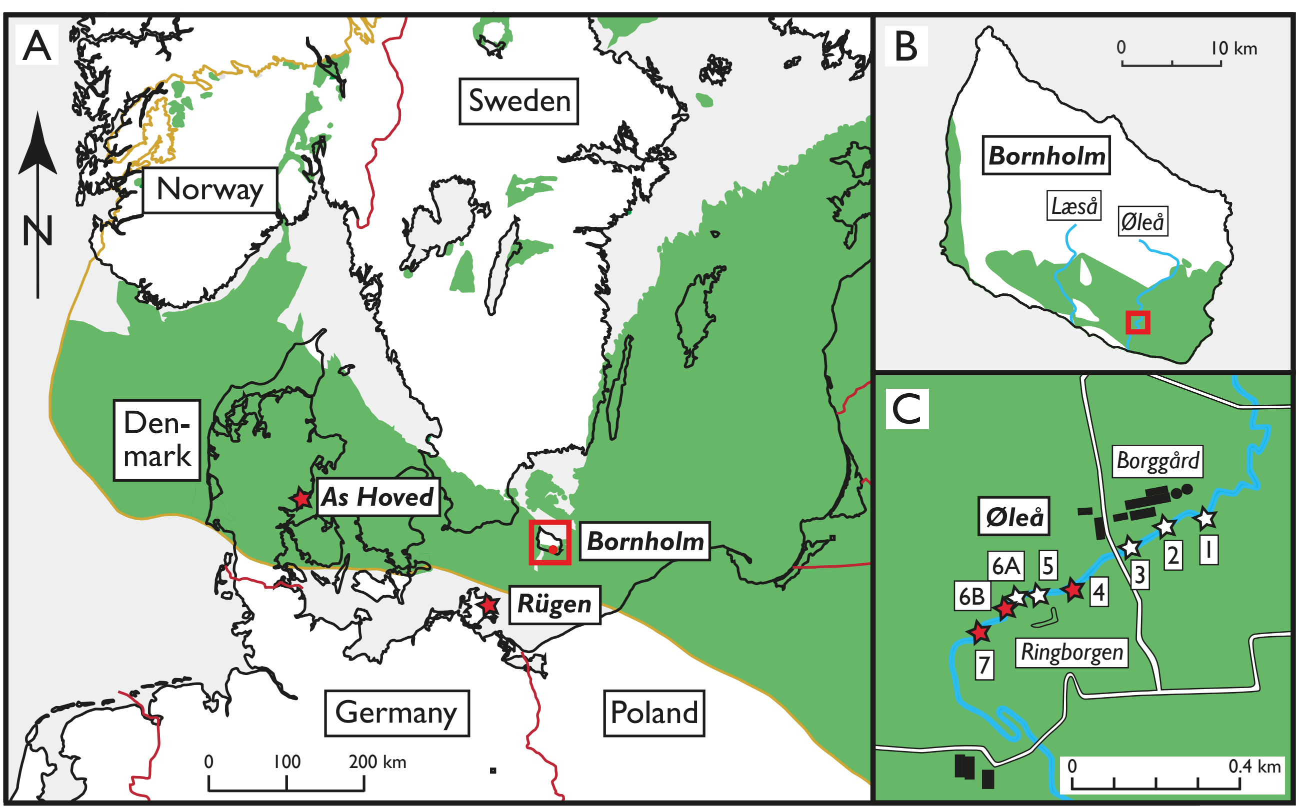


Fig. 2. **A**, map of Scandinavia showing location of Bornholm and the As Hoved locality in Jutland. **B**, map of Bornholm; insert shows location of Borggård. **C**, detailed map of the Øleå rivulet near Borggård showing localities; those marked in red are referred to in the text. (Map modified from Hansen 1945). Areas where Lower Palaeozoic strata are preserved are shown in green; the Caledonian Front is shown in brown.

Taxonomy

The available specimens of *A. longicephalus* and *P.*? *depressus* derive from Borggård, Øleå, Bornholm (for description of exposures, see Weidner & Nielsen 2014). Comparative material of *Proampyx difformis* is from the Andrarum Limestone Bed and the Exporrecta Conglomerate Bed, including material found in glacial erratic boulders at As Hoved, Denmark. Illustrated specimens are deposited at the Natural History Museum of Denmark, University of Copenhagen (MGUH), except for SB-MK 143 from the private collection of A. Buchholz, Stralsund, Germany. A cast of the latter is kept at the Natural History Museum. Measured features are shown in Fig. 3.

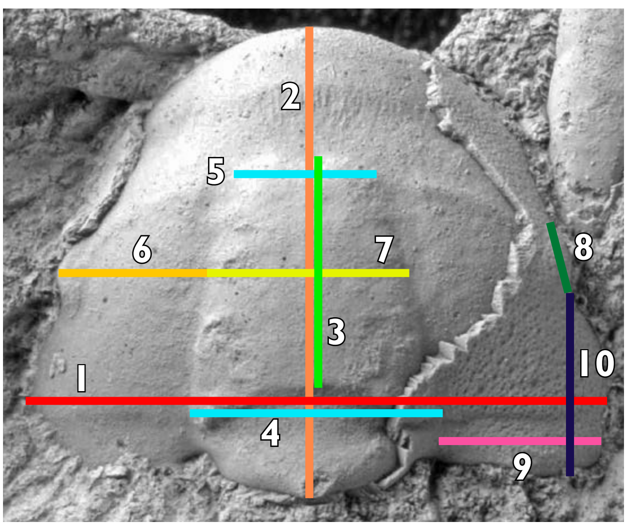


Fig. 3. Measurements referred to in the text (illustrated cranidium, MGUH 30.144, represents *Agraulos longicephalus*). 1: Cranidial width. 2: Cranidial length. 3: Glabellar length. This is the value of the glabella *s. str.* (without occipital ring) on the sagittal line only, due to the forward curvature of the occipital furrow. 4: Posterior glabellar width. 5: Anterior glabellar width. 6: Width of fixigena across palpebral lobe. 7: Width of glabella level width palpebral lobe. 8: Length of palpebral lobe. 9: Width of posterior border. 10: Length of posterior fixigena.

Family Agraulidae Howell 1937

The authorship of this family is often attributed to Raymond (1913). He published several papers that year but mentioned the family Agraulidae only once (Raymond 1913, p. 64), however, without stating that it is new; neither did he provide authorship, diagnosis etc., the name is just a bare heading. We do not consider that this justifies citing Raymond as the author of Agraulidae. Westergård (1953) and Henningsmoen (1959) attributed the family Agraulidae to Howell (1937) where the taxon was described as new and with diagnosis (Howell 1937, p. 1187). We here follow Westergård and Henningsmoen and consider Howell as the author of Agraulidae.

Diagnosis. See Howell (1937) and Henningsmoen (1959).

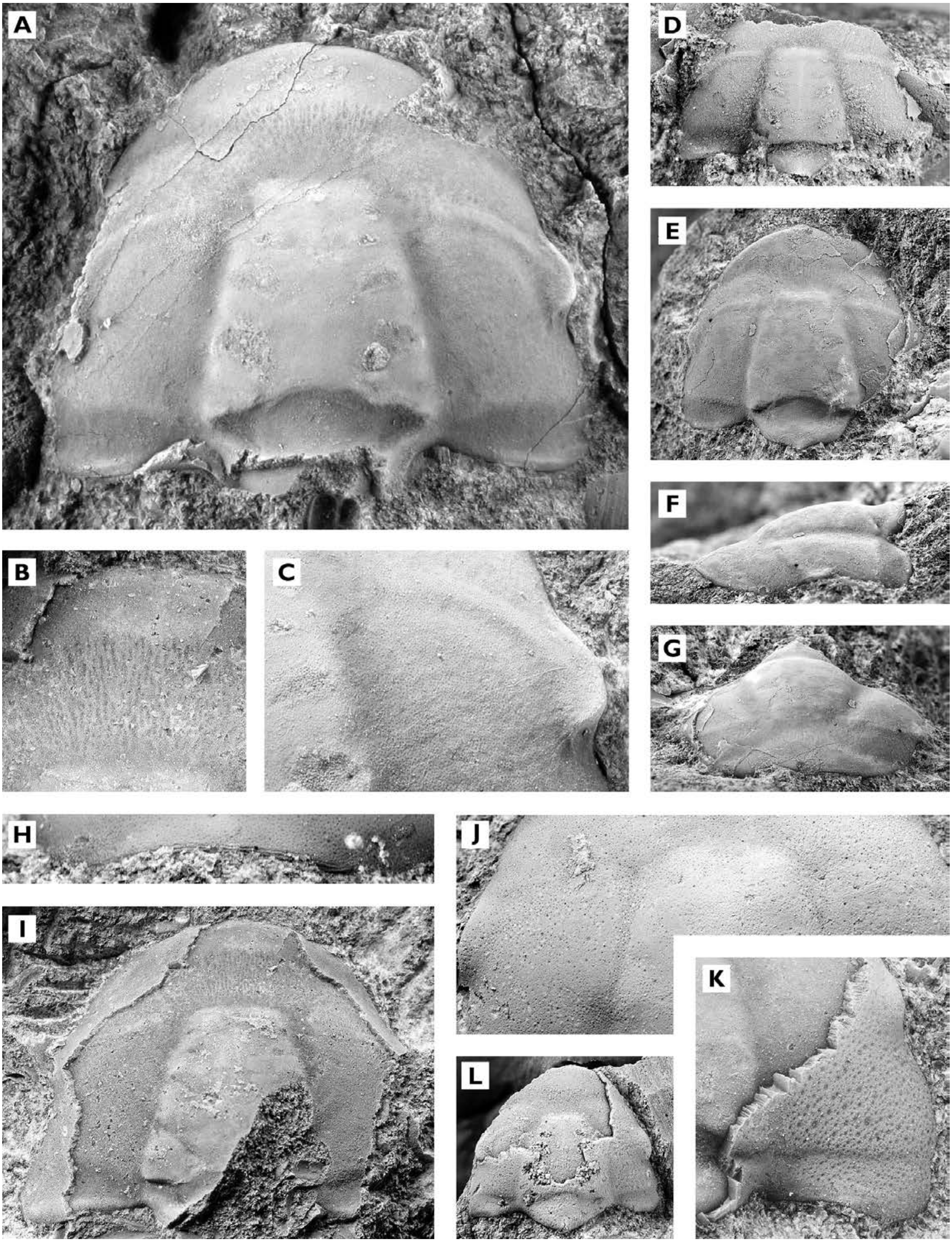
Remarks on diagnosis. Howell's (1937) original diagnosis emphasized that the glabella is almost or quite flush with the fixed cheeks in convexity and that the preglabellar field and border slope evenly towards the anterior margin, in addition to the presence of small eyes and fixed cheeks. He also stated that the posterior fixed cheeks are bluntly triangular, but this is not the case in *A. ceticephalus* or in *A. longicephalus* (Fig. 4). The diagnosis given by Henningsmoen (1959) stresses the forward tapering glabella, the presence of a preglabellar field and the small palpebral lobes situated opposite the anterior half of the glabella.

Geyer & Landing (2001) added as general characters of the cranidium an overall moderate convexity and subtrapezoidal outline. Glabella is also of low to moderate convexity and shows three pairs of short and relatively faint lateral glabellar furrows. However, *A. ceticephalus* and *A. longicephalus* have four pairs of lateral glabellar furrows. Geyer & Landing (2001) further noted that an axial crest tends to be developed and that the cephalic furrows generally are shallow and rather wide, often poorly developed.

Howell (1937) included *Agraulos* Hawle & Corda 1847, *Micragraulos* Howell 1937, *Agrauloides* Howell 1937 and *Conagraulos* Howell 1937 in the Agraulidae. Westergård (1953) assigned only *Agraulos*, but the Scandinavian species he attributed to this genus were later transferred to *Proampyx* Frech 1897 (see Ahlberg & Bergström 1978). Šnajdr (1958) and Henningsmoen (1959) assigned only *Agraulos* and *Skreiaspis* Růžička 1946 to the Agraulidae; Henningsmoen (1959) considered *Agraulos* as a senior synonym of *Proampyx*. Since then, a large number of genera have been assigned to the family; for a summary see Jell & Adrain (2003), who listed more than 35 taxa as a "broad grouping". Geyer & Landing (2001) accepted only *Agraulos* Hawle & Corda 1847, *Skreiaspis* Růžička 1946 and *Proampyx* Frech 1897 in Agraulidae *s.str.*

Bentley & Jago (2004) placed *Agraulos* together with 16 other genera in Agraulidae, whereas *Proampyx* and *Skreiaspis* were transferred to the Gondwanian family Wuaniidae, including 21 genera.

Skreiaspis differs from *Agraulos* and *Proampyx* in having a shorter frontal area, pronounced axial furrows and large eyes (Šnajdr 1958, 1990; Henningsmoen 1959). *Proampyx* differs from both *Agraulos* and *Skreiaspis* by having the frontal area extremely extended or even drawn out into a spine; the palpebral lobes are also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relatively long and strongly curved (Öpik 1961). According to Fletcher *et al*. (2005), details of the type species *Agraulos ceticephalus* exclude *Proampyx* and *Skreiaspis* as close relatives, and the Agraulidae may constitute a monogeneric family.

Genus *Agraulos* Hawle & Corda 1847

Type species. Arion ceticephalus Barrande 1846, designated by Miller (1889); *Eccaparadoxides pusillus* Zone; Skryje, Bohemia, Czech Republic.

Diagnosis (emend. Fletcher & Greene 2013). Test thick, cephalic furrows practically effaced on the outer surface; glabella and fixigenae typically show almost even convexity; glabella forward tapering, often slightly bowed out laterally; four pairs of lateral glabellar furrows, S1 bifurcated; eye ridges start at midpoint of palpebral lobes and meet axial furrows level with front of glabella; internal moulds show indistinct parafrontal band on anterior glabellar lobe; long frontal area; anterior border mesially expanded; no anterior border furrow, border defined by a smooth area on the internal mould devoid of caecal ridges; short palpebral lobes situated anteriorly, centre level with S2; converging anterior branches of facial suture; posterior branch of facial suture meets the posterior border in a gentle curvature; occipital ring backwards extended; spinose librigenae and some pygidial rings; exoskeleton smooth to punctate.

Remarks on diagnosis. Already Lake (1932) observed that the test is thick in *Agraulos* and masks the cephalic furrows and hollows, and specimens with intact test therefore look very different from exfoliated specimens. Lake (1932) also noted that the eye ridge starts at the middle of the palpebral lobe in *A. longicephalus.* This appears to be a salient distinguishing feature of *Agraulos* in comparison with most other trilobites where the eye ridge meets the palpebral lobe at its anterior end. This important feature is often incorrectly shown in drawings of *Agraulos* (e.g. Barrande 1846; Šnajdr 1958; Sdzuy 1961).

Agraulos longicephalus (Hicks 1872)
Fig. 4

1872 *Arionellus longicephalus* Hicks, p. 176, pl. V, figs 20–26.
1902 *Agraulos ceticephalus* BARR., Grönwall, pp. 158–159, pl. 4, fig. 25.
1916 *Agraulos* cf. *holocephalus* Matthew, Nicholas, pp. 461–463, pl. 29, fig. 7.
1932 *Agraulos longicephalus* (Hicks), Lake, pp. 157–159, pl. 19, fig. 10; pl. 20, figs 1–10.
1958 *Agraulos longicephalus* Hicks, Lotze, pp. 731, 737.
1961 *Agraulos longicephalus* (Hicks 1872), Sdzuy, pp. 620–622 (338–340), pl. 23, figs 7–17; text-figs 32, 33.
1988 *Agraulos longicephalus* (Hicks 1872), Martin & Dean, pp. 21–22, pl. 3, figs 9–13.
1988 *Agraulos longicephalus* (Hicks 1872), Morris, p. 13. (Listed).
1994 *Agraulos longicephalus* (Hicks 1872), Young *et al*., p. 343.
1994 *Agraulos* cf. *ceticephalus* (Barrande 1846), Rudolph, p. 217, pl. 24, fig. 5.
1997 *Agraulos ceticephalus* (Barrande 1846), Buchholz, p. 252. (Listed).
2002 *Agraulos longicephalus* Hicks, Young *et al*., p. 17, pl. 4, fig. 5.
2006 *Agraulos longicephalus* (Hicks), Fletcher, pl. 34, fig. 34.
2014 *Agraulos longicephalus* (Hicks 1872), Weidner & Nielsen, pp. 47–48, fig. 41.

Lectotype. Complete specimen SM A3495a, original of Hicks (1872, pl. V, fig. 20), refigured by Lake (1932, pl. 19, fig. 10) and designated as lectotype by Morris (1988); Menevian, St David's, Wales.

Material. Approximately 25 more or less intact cranidia (see also Weidner & Nielsen 2014). Only one of Grönwall's (1902) specimens from Øleå could be identified in the collection of the Natural History Museum, Copenhagen; the whereabouts of the other specimen from Øleå and the single specimen from Læså are unknown.

◀ **Fig. 4**. *Agraulos longicephalus* (Hicks 1872), cranidia. All except L are from Borggård, Øleå, Bornholm. **A,** internal mould previously illustrated as *A. ceticephalus* by Grönwall (1902, pl. 4, fig. 25), × 8. *Ptychagnostus punctuosus* Zone, MGUH 200, see also C. **B,** close-up showing caeca on the internal mould illustrated in full in I, × 8. **C,** close-up of the internal mould illustrated in A showing eye ridge and palpebral lobe; note fine threads at rear end of lobe, × 12. **D,** internal mould of small cranidium showing distinct parafrontal band on anterior glabellar lobe, × 6, *A. atavus* Zone, MGUH 31.236. **E–G,** dorsal, side and frontal views of internal mould, × 4, *A. atavus* Zone, MGUH 30.143. **H,** close-up of frontal margin (anterior view) showing terrace lines on test surface, × 12, *A. atavus* Zone, MGUH 31.237. **I,** largest cranidium recorded, partly exfoliated, × 4, *A. atavus* Zone, see also B, MGUH 31.238. **J,** close-up of test surface showing prosopron, × 8, *A. atavus* Zone, MGUH 30.145. Cranidium previously illustrated by Weidner & Nielsen (2014, fig. 41F). **K,** close-up of test surface showing prosopron, × 8, *A. atavus* Zone, MGUH 30.144. Cranidium previously illustrated in full by Weidner & Nielsen (2014, fig. 41E). **L,** partly exfoliated cranidium showing triangular occipital ring. Ice-rafted boulder from Rügen, Mecklenburg, Germany, × 4, *Ptychagnostus punctuosus* Zone, SB-MK 143.

Occurrence. This species occurs commonly in the Øleå section on Bornholm. The material derives from concretions of bituminous limestone in the lower and upper part of the *Acidusus atavus* Zone (localities 6B and 4 on Fig. 2C, respectively); the specimens described by Grönwall (1902) came from the *Ptychagnostus punctuosus* Zone where they are associated with *P. punctuosus*, *Lejopyge elegans* and *Doryagnostus incertus* amongst others. Three cranidia are known from ice-rafted boulders of limestone found in northern Germany; these concretions contained the same agnostids and likely derive from the Bornholm area (Rudolph 1994; Buchholz 1997). The species is also reported from the *Paradoxides hicksii* Zone of eastern Newfoundland, Avalonian Canada (Martin & Dean 1988; Fletcher 2006) and from the *Tomagnostus fissus* Zone to the lower part of the *P. punctuosus* Zone in Avalonian Great Britain (Thomas *et al.* 1984). Outside Baltica and Avalonia it occurs in Spain in the *Pardailhania* and *Solenopleuropsis* substages (Caesaraugustian Stage) (Lotze 1958; Sdzuy 1961, 1972), which correspond to the *A. atavus* and *P. punctuosus* zones of the *Paradoxides paradoxissimus* Superzone in Scandinavia (Geyer & Shergold 2000).

Description. Most of the described features can only be observed in exfoliated specimens and if not otherwise stated the following description is based on internal moulds. In testaceous material many features are masked by the thick test. Cranidium subtriangular in outline, with rounded anterior margin; length corresponds to about 85 % of max. width (n = 8); largest specimen is 14.2 mm long (Fig. 4I). Glabella occupies *c.* 50 % of the total cranidial length and 40 % of max. cranidial width in adult specimens (measured on exfoliated specimens, see Fig. 3:3–4); it tapers evenly and rather strongly, slightly bowed out laterally, front truncate and the anterior width corresponds to only 50 % of glabellar width at the occipital furrow (Fig. 3:4–5). When the exoskeleton is preserved, the axial furrows are shallow, on internal moulds well-impressed, broad; the preglabellar furrow is shallow, on internal moulds mostly very faint. On exfoliated specimens four faint and broad lateral glabellar furrows extend on each side across one-third of the glabellar width (Fig. 4A; see also Weidner & Nielsen 2014, figs 41A–B, D–E). S1 and S2 are relatively large, elongate and directed rearwards-inwards; S1 tends to bifurcate. S3 and S4 are comparatively small, elongate rounded. Anterior and lateral views show the moderate convexity (tr. and sag.) of the cranidium and a glabella distinctly raised above the level of the downsloping fixigenae and preglabellar field (Fig. 4F–G). A faint glabellar crest is commonly present. When the test is preserved, the occipital furrow appears abaxially just as two curved, short and deep lateral impressions (see Weidner & Nielsen 2014, fig. 41F); on internal moulds the occipital furrow is very broad and curving forwards. Occipital ring triangular in testaceous specimens (Fig. 4L; see also Sdzuy 1961, text-fig. 32 and Weidner & Nielsen 2014, fig. 41A); on internal moulds the occipital ring expands into a distinct node at the posterior margin (Fig. 4E; see also Weidner & Nielsen 2014, fig. 41E). The anterior margin of the cranidium is smoothly curved; the border is long (sag.), mesially swollen and projected forward, sloping downwards and separated from the glabella by a preglabellar field which is slightly longer (sag.) than the border. Border furrow not developed. The preglabellar field carries caeca, the anterior border is smooth (Fig. 4A–B, I). Eyes far from glabella; width of fixigena corresponds to *c.* 60 % of the glabellar width across the centre of the palpebral lobe (Fig. 3:6–7). Palpebral lobes are short (corresponding to *c.* one third of glabellar length, Fig. 3:8), gently falcate, widening slightly posteriorly, with its centre level with S2. The anterior and posterior ends of the palpebral lobe divide into thin, short threads which run parallel with the suture and form narrow, tiny grooves between the threads and the suture (Fig. 4C). The eye ridges start at the palpebral lobes just above their midlengths (Fig. 4C; see also Whittington 1992, pl. 51; Fletcher *et al.* 2005, fig. 11–1; Weidner & Nielsen 2014, fig. 41C). This is an important diagnostic feature not seen in the majority of other trilobites. The eye ridges continue gently curved towards anterior lobe of glabella; they meet the axial furrows nearly level with front of glabella and bifurcate just before reaching the axial furrows (Fig. 4A). The anterior thread continues as an indistinct parafrontal band across the anterior lobe, forming a semi-continuous ridge uniting the eye ridges (Fig. 4D–E). In some specimens a small mesial boss is seen centrally on the parafrontal band (Fig. 4A). A distinct furrow, approximately of the same width as the eye ridge, runs parallel to it on its rear side and amalgamates with the palpebral furrow, which continues to the posterior end of the palpebral lobe where it fades out (Fig. 4C). The facial suture is directed inwards in front of and slightly outwards behind the palpebral lobe and generally runs parallel to the axial furrow, forming almost one single gentle curve, just interrupted by the short palpebral lobe (Fig. 4A; see also Weidner & Nielsen 2014, fig. 41B–F). The posterior branch meets the posterior border in a gentle curvature. The posterior border furrows are adaxially weakly curved impressions on the shell exterior; on internal moulds they are wide (exsag.) and deep and broaden abaxially. Length (exsag.) and width (tr.) of posterior fixigena (Fig. 3:9–10) are equivalent to about 70 % of glabellar length and about 60 % of the width of the occipital ring, respectively. Test thick, with three sizes of punctae scattered over a background of tiny granules. A few terrace lines are seen along the anterior margin (Fig. 4H).

Remarks. The cranidium of *Agraulos longicephalus* strongly resembles that of *A. ceticephalus*, and the two species are likely closely related (see also Fletcher *et al.* 2005). Various width/length ratios and other parameters of *A. ceticephalus* are similar to those discussed above for *A. longicephalus*. The only conspicuous difference is a "short (sag.) occipital ring strongly curved backward into a blunt point" in *A. ceticephalus* (cit. Fletcher *et al.* 2005, fig. 11:4) versus the massive triangular occipital ring in *A. longicephalus* (Fig. 4L) and the latter is considered a possible early variant of *A. ceticephalus*.

The cranidium figured by Grönwall (1902, pl. 4, fig. 25) as *A. ceticephalus* was shown with an intact occipital ring. However, re-investigation of the specimen revealed that the occipital ring is not preserved (see Fig. 4A), except that the base of a broad extension of the occipital ring can be observed on the right side of the cranidium, suggesting that this specimen represents *A. longicephalus*.

In their description of topotype material of *A. ceticephalus* from the *Eccaparadoxides pusillus* Zone of Bohemia, Fletcher *et al.* (2005) showed details not noted previously: the ornament consists of three sizes of punctae scattered over a background of tiny granules with fine terrace ridges along the anterior cranidial margin and the posterior edge of the occipital ring; four pairs of lateral glabellar furrows arching backwards; weakly impressed anterior border furrow; upstanding palpebral lobes. We here note that the eye ridges commence at midlength of the palpebral lobes and thus they do not form a direct continuation of the palpebral lobes as often reconstructed (Barrande 1852, pl. 10, figs 6, 14; Šnajdr 1958, fig. 37; Henningsmoen 1959, fig. 205, 1; Sdzuy 1961, text-fig. 32).

Genus *Proampyx* Frech 1897

Type species. Proetus? *difformis* var. *acuminatus* Angelin 1851, by original designation; from the Andrarum Limestone Bed of the *Lejopyge laevigata* Zone at Andrarum, Scania, Sweden.

Diagnosis (emend. Öpik 1961 and Ahlberg & Bergström 1978). Test thick, cephalic furrows practically effaced on the outer surface; forward tapering glabella elevated above fixigenae and frontal area; glabella parallel-sided, rarely bowed out laterally; three to four pairs of lateral glabellar furrows, S1 bifurcated; median keel on glabella comparatively well marked; eye ridges start at front of palpebral lobes and meet axial furrows slightly behind front of glabella; long to very long flat frontal area, some species with extended cusp; anterior border mesially expanded forwards, rarely backwards; no anterior border furrow, and border is poorly separated from preglabellar field; long palpebral lobes situated posteriorly, centre level with S1; diverging anterior branches of facial suture in front of palpebral lobes; posterior fixigena short (exsag.), distinctly triangular; occipital ring expanded backwards, carrying node or spine on anterior margin; posterior margin evenly rounded; spinose librigenae and thorax; exoskeleton punctate; comparatively large pygidium.

Remarks. Ahlberg & Bergström (1978) assigned several Lower Cambrian species to *Proampyx*; they would now be allocated to various other genera (e.g. Geyer 1990; Geyer & Landing 2004; Geyer *et al*. 2004).

***Proampyx*? *depressus* (Grönwall 1902)**
Fig. 5A–I

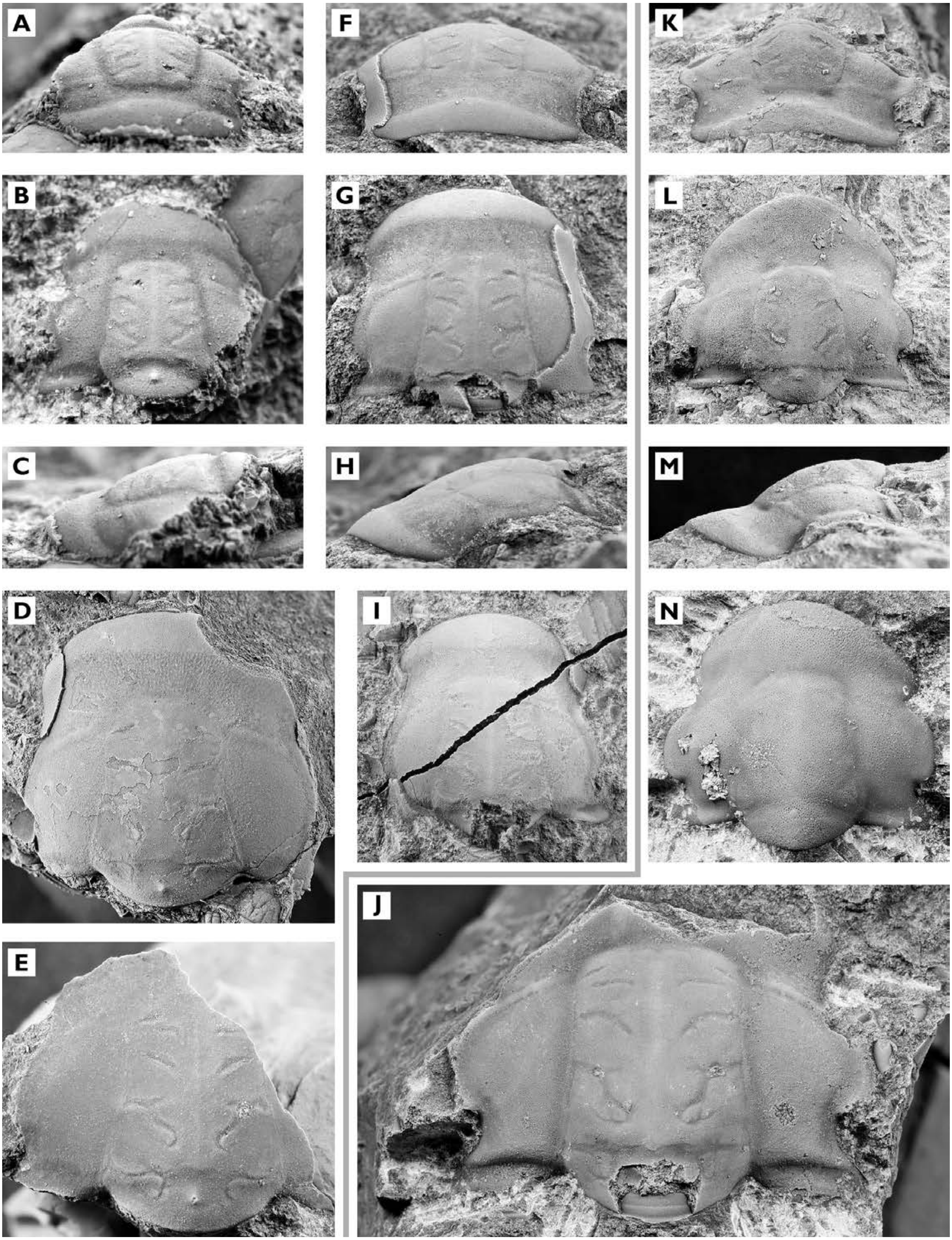
1902 *Agraulos depressus* n. sp. Grönwall, pp. 159–160, pl. 4, fig. 24.

Lectotype. Cranidium MGUH 199, original of Grönwall (1902, pl. 4, fig. 24); here designated as lectotype and re-illustrated in Fig. 5F–H; *Ptychagnostus punctuosus* Zone at Borggård, Øleå, Bornholm.

Material. In addition to the lectotype specimen figured by Grönwall (1902), three more cranidia were located in the collection of the Natural History Museum of Denmark. A young cranidium was further collected from bituminous limestone in the basal part of the *Ptychagnostus punctuosus* Zone at Borggård, Øleå (locality 7, Fig. 2C). It is associated with abundant *Cotalagnostus lens*, rare specimens of *Acidusus atavus* and *P. punctuosus*, as well as *Onymagnostus ciceroides*, *Hypagnostus mammillatus*, *Diplorrhina depressa* and *Diplagnostus planicauda bilobatus.*

Occurrence. The species occurs in the *Ptychagnostus punctuosus* Zone and is only known from Borggård, Øleå, Bornholm.

Description. The cranidia are of low convexity and show one common curvature for glabella and fixigenae (Fig. 5F). The available material is largely exfoliated, allowing the study of details which are masked when the thick test is intact. The largest specimen is 13.3 mm long. The cranidium is almost quadrate in outline, the length corresponds to 90 % of max. width, with a broadly rounded anterior margin. The glabella is not inflated and is defined by shallow and narrow axial furrows; the preglabellar furrow is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very faint to absent, but the gently rounded glabellar front is slightly raised and outlines the anterior delimitation of glabella. The glabella occupies *c*. 60 % of the cranidial length (measured from the occipital furrow) and the maximum width corresponds to almost 50 % of the cranidial width (measured across posterior fixigenae). The glabella tapers to midway between S1 and S2, where glabella is narrowest, then expands faintly forwards; width of the frontal lobe corresponds to *c*. 77 % of the max. glabellar width. Four pairs of lateral glabellar furrows are developed. A fifth pair of furrows is located immediately in front of the occipital furrow and runs in a sinuous course across the entire glabella. The furrows are fairly narrow and well incised and obviously represent a pair of auxiliary furrows not homologous to S1. S1 in front of the auxiliary furrow is biramous, with a fairly long posterior branch directed inwards-backwards, ending in a pair of faintly raised apodemes. Approximately at midlength of the posterior branch, a short anterior branch commences directed inwards-forwards. S2 is composed of simple, faintly curved furrows, starting distant from axial furrows; they are slightly narrower and shorter than the posterior branches of S1 and directed inwards-backwards. S3 is a pair of short, slightly curved, transversely directed furrows located close to glabellar midline. S4 is also short and narrow, but comparatively deep, almost comma-shaped furrows starting close to axial furrows and running slightly anteriorly. The occipital ring shows a small mesial node. In the lectotype specimen a small portion is broken off from the posterior edge of the occipital ring, exposing the ventral doublure. It shows two shallow furrows that may have functioned as stopping devices for the anterior thoracic segment.

The anterior margin of the cranidium is smoothly curved. The anterior border slopes gently downwards; it is long (sag., exsag.) and slightly projected backwards medially; border furrow lacking, the preglabellar field is of same length (sag.) as the border. In the lectotype specimen the posterior border furrow is broad and shallow on the internal mould (left hand side), but shallow to nearly effaced on the exterior (right hand side). The width of fixigena across palpebral lobe corresponds to *c*. 66 % of the glabellar width at the same level (Fig. 3:6–7) and the eyes are thus far from glabella. The palpebral lobe is long, corresponding to *c*. 50 % of glabellar length, narrow, gently curved, posteriorly slightly wider than at the anterior end, stretching from mid-part of L1 to almost the level of S3. Eye ridges distinct, forming a continuous arc with palpebral lobes, adaxial end located slightly behind anterolateral corners of frontal glabellar lobe. The anterior branch of the facial suture consists of a faintly diverging posterior section and an anterior section turning towards sagittal line where the anterior branches meet smoothly. The posterior branch runs backwards-outwards from the posterior end of the eye at an angle of *c*. 45° from sagittal line, and the posterior fixigena is distinctly triangular in outline. Length (exsag.) of posterior fixigena corresponds to about one third of glabellar length. Width (tr.) of posterior border corresponds to about 50 % of occipital ring. The few patches of intact test show a prosopon of very fine, densely spaced pits.

A small specimen, only 4.5 mm long, is interpreted as a juvenile specimen of *P.*? *depressus*, despite some morphological differences (Fig. 5A–C). We especially emphasize that the pattern of lateral glabellar furrows is identical. Besides, the anterior cranidial margin is also smoothly rounded, and the glabella and fixigenae form one smooth convexity. The juvenile cranidium differs from the adult specimen by having a parallel-sided glabella, which is a little inflated and bounded by deeper axial furrows, the fixed cheeks are relatively narrower, the eyes are shorter and more strongly falcate, and the preglabellar field is distinctly shorter.

Remarks. Agraulos depressus Grönwall 1902 is here allocated to *Proampyx*? because

1) eye ridges spring from the anterior end of palpebral lobes
2) palpebral lobes are long, centre situated level with S1
3) diverging anterior branch of facial suture
4) occipital ring carries a node, situated anteriorly

◀ **Fig. 5**. **A–I:** Cranidia of *Proampyx*? *depressus* (Grönwall 1902), all from the *Ptychagnostus punctuosus* Zone at Borggård, Øleå, Bornholm. **A–C,** anterior, dorsal and side views of exfoliated juvenile specimen; note that the side view (C) is shown reversed in order to facilitate comparison with the other specimens, × 8, MGUH 31.239. **D,** largest specimen recorded, largely exfoliated, × 4. MGUH 31.240. **E,** internal mould showing distinct lateral glabellar furrows; note sinuous auxillary furrows just in front of the occipital furrow, × 5, MGUH 31.241. **F–H**, anterior, dorsal and side view of partly exfoliated lectotype cranidium, previously illustrated by Grönwall (1902, pl. 4, fig. 24), × 4, MGUH 199. **I,** internal mould showing distinct lateral glabellar furrows, × 5, MGUH 31.242. **J–N**: Cranidia of *Proampyx difformis* (Angelin 1851), all from the Andrarum Limestone Bed, *Lejopyge laevigata* Zone. **J,** internal mould showing distinct lateral glabellar furrows; note small auxillary furrows just in front of the occipital furrow, × 4, Borggård, Øleå, Bornholm, MGUH 31.243. **K–M,** anterior, dorsal and side views of exfoliated cranidium, × 4, found in glacial erratic boulder, As Hoved, Jutland, Denmark, MGUH 31.244. **N,** cranidium showing finely pitted test, × 5, Borggård, Øleå, Bornholm, MGUH 31.245.

5) posterior fixigena triangular
6) four pairs of lateral glabellar furrows arranged in virtually the same pattern are shared with *P. difformis* (see e.g. Fig. 5J, N; Álvaro *et al*. 2013, fig. 5c–e). Another *Proampyx* species showing constantly four pairs of lateral glabellar furrows is *Proampyx aculeatus* (see Westergård 1953, pl. 1, figs 9, 10a).

However, *Proampyx*? *depressus* differs from the Scandinavian species of *Proampyx* in the following features:
1) a non-inflated glabella (adult condition)
2) very faint to effaced preglabellar furrow
3) comparatively short, downsloping frontal area
4) the palpebral lobes are only gently curved, narrow
For these reasons the generic assignment is uncertain.

All specimens at hand of *P.*? *depressus* show a laterally waisted glabella; this feature is only occasionally observed in *Proampyx* species e.g. *P. difformis* (Fig. 5J, N) and *P. anceps* (Westergård 1953, pl. 2, fig. 2a). Most *Proampyx* species show a laterally bowed out glabellar shape. *P.*? *depressus* also shows a fifth pair of auxillary furrows behind S1 on glabella. So far, this feature has only been observed in *P. difformis* (Fig. 5J), but potentially it offers strong support for an assignment of *depressus* to *Proampyx*.

Öpik (1961) stressed the similarity between the Australian species *Proampyx agra* and the Scandinavian *P. difformis* and *P. acuminatus*. The Australian species occurs in the *Proampyx agra* Zone of the Selwyn Range, Queensland, Australia (Öpik 1961, pp. 146–148, pl. 12, figs 1–6), which corresponds to the *Lejopyge laevigata* Zone of Scandinavia. Some of the features found in *P.*? *depressus* are also reminiscent of *P. agra*. The two species thus share an almost quadrate shape of the cranidium, not so strongly curved palpebral lobes, shallow palpebral furrows and deeply incised auxiliary furrows on the posterior glabella (Fig. 5G; Öpik 1961, pl. 12, fig. 5a). However, *P. agra* differs from the Scandinavian species of *Proampyx* by "being much more effaced, having wider fixed cheeks, a more evenly rounded anterior margin of the cranidium, a less tapered glabella and a much less variable anterior border" (cit. Bentley & Jago 2004 p. 184). The latter authors transferred *agra* to their genus *Arminoepikus* of the Gondwanan family Wuaniidae.

Typical species of *Proampyx* occur in Scandinavia in the *Lejopyge laevigata* Zone (Andrarum Limestone Bed), whereas *Proampyx*? *depressus* occurs in the older *Ptychagnostus punctuosus* Zone. *P.*? *depressus* may represent an early, atypical form of *Proampyx* or maybe even a precursor that should be separated in a genus of its own.

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Book Review: Finally, all Steno's scientific papers translated from Latin into English

JENS MORTEN HANSEN

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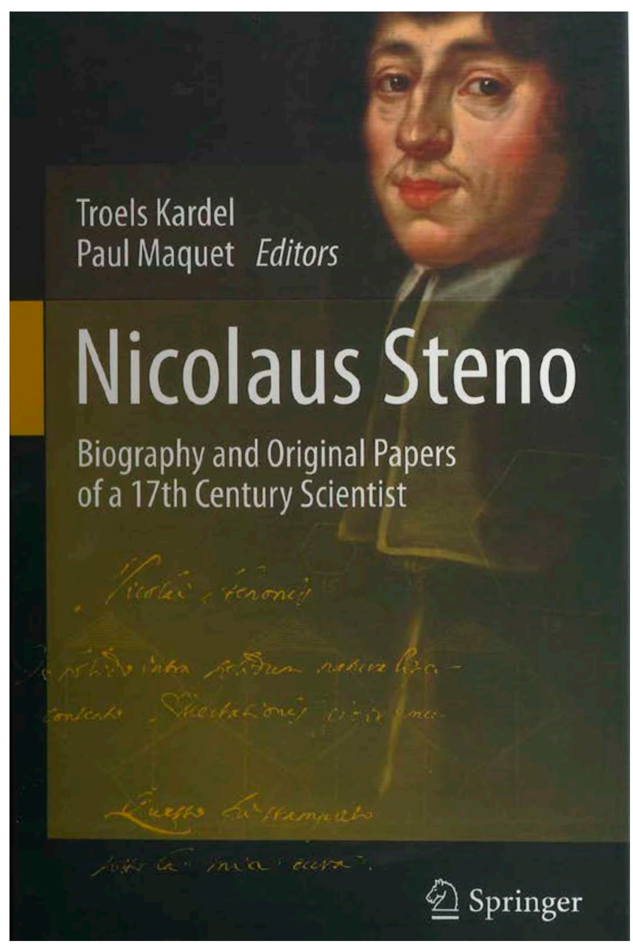
Jens Morten Hansen [jmh@geus.dk], Geological Survey of Denmark and Greenland, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark.

Kardel. T. & Maquet, P. (eds) 2013: Nicolaus Steno. Biography and Original Papers of a 17th Century Scientist. Springer-Verlag, Berlin, Heidelberg, 739 pp.

The Danish-Italian scientist Nicolaus Steno (Niels Stensen, 1638–86) is considered to be the founder of geology, including palaeontology and mineralogy, as a discipline of modern science. He is also considered to be the founder of modern scientific conceptions of the human glands, muscles, heart and brain. Steno also made important novel contributions in fields of anatomy involving, for example, comparative anatomy (humans/animals; animals/fossils), embryology and the diverse roles of body fluids.

The general outset of Steno's science and his philosophy of science constitutes an important step from the medieval and renaissance way of thinking into the appearance of modern science in the 17th century and the Enlightenment of the 18th century (e.g. Kardel 1994; Hansen 2009). The appearance in the 18th century of the contradistinction between the traditional creationistic understanding and Lyell's (1797–1875) and Darwin's (1809–1882) paradigmatically new interpretation of the evolution of the Earth and life on Earth can, to some extent, be traced back to Steno and his methods.

However, Steno's ideas and great influence on geology and the general principles of modern science remain relatively unrecognized. Latin was the common scientific language in Steno's time, but it soon became replaced by modern, national languages. The declining role of Latin in protestantic Europe, and the fact that most of Steno's philosophical statements were not translated from Latin into English until recently, have probably contributed to 19th and 20th century oblivion of Steno in English-speaking countries, whereas Steno's achievements remained known among geologists, anatomists and historians of science in Italy, France, Germany and Russia. In addition, Steno's strong religious faith and his conversion from his native Danish protestantism to roman catholi-



cism when he had become established as a scientist at the Medici's (Ferdinand II and Cosimo III) court in Florence, has puzzled modern scientists. Steno was certainly aware of his strong faith and, perhaps as a consequence, never confused science with religion. On the contrary, in the harsh post-reformation environment of the just-ended counter-reformation wars of Europe, he humbly claimed the freedom of science from religious belief. Moreover, his appointment by the pope as *titular bishop* of the former city of Titiopolis in the East Roman Empire, but stationed in Schwerin with secret duties in protestant and monocratic Denmark (at that time including northern Germany and Norway), has not made it easier to comprehend Steno's extremely stressed situation involving loyalty to science, the roman catholic church and the absolute monarchy of the protestantic Denmark.

Moreover, it may have contributed to the lack of recognition of Steno's achievements that several of his ideas were simply not understood. This relates in particular to his modern understanding of muscles about which professor of anatomy and biographer of Steno, Harald Moe, as late as in 1988 wrote that Steno's work on muscles is "among his weakest". In his 1994-edition of the biography from 1988, after Kardel's studies and comparisons of Steno's descriptions of muscles with computer animations of the human motion apparatus, Moe completely changed his opinion, considering Steno's muscle theory to be amongst his most brilliant achievements.

Kardel and Maquet's new monograph on Steno's science

The present book on Steno, edited and translated by Troels Kardel and Paul Maquet, is the most important monograph for studies of Steno's scientific discoveries and life yet written. Besides translations into modern English of all of Steno's known scientific publications, the book includes the comprehensive biography by the German (Danish citizen from 1938), anti-Hitlerian Steno-scholar and catholic priest, Gustav Scherz, shorter biographies, the editors' comments, many illustrations and footnotes, a comprehensive bibliography and indexes to all the people and places mentioned. Kardel and Maquet's translation (1994) from Latin into English of Steno's general view on scientific recovery and levels of knowledge ('Prooemium') is also included, making it available to a broader public. Steno's first scientific thesis ('De Thermis'), which was rediscovered as late as 1960 in Philadelphia (USA), is also included.

With this wonderful book a complete edition of Steno's 34 known scientific works is available in English for the first time. A previous almost complete edition by Vilhelm Maar (1910) was in Latin (apart from one paper originally written in French). Many of Steno's scientific papers have been translated one by one into English, German, French, Danish, Italian, Russian and Japanese over the years, but the absence of collected translations of both geological and anatomical papers has made it difficult to fully comprehend his more encompassing, general understanding of fundamental scientific principles. Thus, only Steno's second geological paper (1669, 'De Solido'), which was translated into English already in 1671 by Henry Oldenburg, secretary of The Royal Society of London, can be considered to be generally known by historians of science. Accordingly to *Encyclopaedia Britannica,* 'De Solido' is among the 100 most important works of modern science.

Some other sources and recent papers on Steno's science

Another important and comprehensive publication on Steno's scientific thoughts, a complete translation of and commentary on Steno's large student manuscript 'Chaos', was published by August Ziggelaar in 1997. Because this manuscript is already available in English and would require many more pages it is not included in the new monograph. The name of the manuscript is derived from Steno's exclamation written on the manuscript: "In nomine Jesu, Chaos", expressing his opinion on the state of science at the time. 'Chaos' (now 520 printed pages) constitutes a basis for understanding what inspired him at an early stage of his academic life.

Likewise, Steno's correspondence with his Dutch friend, the great philosopher Baruch Spinoza (1632–1677), as well as his correspondence with his German friend and admirer, G.W. Leibniz (1646–1716), are not included, although they give important clues to Steno's understanding of the role of science in contrast to theology. The two editors have wisely chosen a form of monograph that exclusively includes what Steno himself decided to publish.

Introductions to the study of Steno's scientific papers

Direct reading of Steno's scientific papers without introduction to his time and the state of renaissance science would not make it easy to understand his genius, achievements and role in posterity. Such introductions may be found in a relatively new, popular description in English of Steno's life, science and sainthood by Alan Cutler (2003), in my own book (2000) on perspectives of Stenonian geology, and in my review (2009) of Steno's philosophy of science. Other modern interpretations of Steno's scientific work can be found in e.g. Stephen Jay Gould's introduction (1981) to Steno's first geological paper (comparison of modern shark teeth which he showed were identical with fossil shark teeth, so-called *Glossopetrae*).

Such introductions to Steno may make it easier to understand that Steno's way of study and reasoning differs fundamentally from that of his contemporaries, e.g. Athanasius Kircher (1602–80), by seeking causes and not explaining nature by causes already given in the Bible or other authoritative theological sources. In Steno's thinking the role of science is not to explain effects by means of *a priori* given causes. That is the role of religion, and in our time, I may venture to say, the role of forecasting and mathematical modelling. In Steno's thinking the role of science is to study the things visible in nature and the human body and thereafter, e.g. by assuming that the forces active today were also active in the past, to 'back-strip' and reconstruct, and thereby understand, the causes of the observed effects.

Especially for students of geology, I consider that it will be an eye-opener to read Steno's own introduction in 'De Solido' in order to learn that most of the basic principles applied in modern geology originate from Steno and his breathtaking study of northern Italy, and that these principles have been practiced ever since.

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Episodes of aeolian sand movement on a large spit system (Skagen Odde, Denmark) and North Atlantic storminess during the Little Ice Age

LARS B. CLEMMENSEN, ASLAUG C. GLAD, KRISTIAN W. T. HANSEN & ANDREW S. MURRAY

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Late Holocene coastal dune successions in north-western Europe contain evidence of episodic aeolian sand movement in the recent past. If previous periods of increased sand movement can be dated sufficiently precisely and placed in a correct cultural and geomorphological context, they may add to our understanding of storminess variation and climate change in the North Atlantic during the later part of the Holocene. In this study, coastal cliff sections of Holocene dune sand were investigated in the north-western part of the Skagen Odde spit system in northern Denmark. Four units of aeolian sand were recognized. Optically stimulated luminescence (OSL) dating indicates that aeolian sand movement took place in four phases: around AD 1460, between AD 1730 and 1780, around AD 1870, and since about AD 1935. The first phase of sand movement occurred during cooling in the first part of the Little Ice Age. A change in the atmospheric circulation, so that both the North Atlantic Oscillation (NAO) and the Atlantic Multidecadal Oscillation (AMO) were negative, apparently led to an increased number of intense cyclones causing inland sand movement and dune building. The second and third phase of aeolian sand movement during the Little Ice Age also took place in periods of increased storminess, but during these events it appears that negative NAO values were coupled with positive AMO values. The final phase of sand movement is intimately linked to the modern formation of frontal dunes which takes place during moderate storminess. These findings are important as they indicate three major periods of aeolian sand movement and storminess during the Little Ice Age.

Keywords: Aeolian sand movement, storminess, climate change, Little Ice Age, Skagen Odde.

Lars B. Clemmensen [larsc@ign.ku.dk], Aslaug C. Glad [hrm158@alumni.ku.dk], Kristian W.T. Hansen [hansenkw@gmail.com], Department of Geosciences and Natural Resource Management, University of Copenhagen, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark. Andrew S. Murray [anmu@dtu.dk], Nordic Laboratory for Luminescence Dating, Department of Geoscience, Aarhus University, DTU Risø Campus, DK-4000 Roskilde, Denmark.

Coastal dunefields in north-western Europe have experienced episodic activation throughout the later part of the Holocene (e.g. Clarke & Rendell 2009, 2011; Clemmensen *et al*. 2009). If dated sufficiently precisely and placed correctly in a cultural framework, periods of aeolian sand movement within the dunefields may act as proxy records of past wind climate and North Atlantic storminess variation (e.g. Clemmensen *et al*. 2001a; Clemmensen *et al*. 2007; Madsen *et al*. 2007; Clarke & Rendell 2009; Clemmensen *et al*. 2009).

Numerous studies (e.g. Pye & Neal 1993; Knight *et al.* 1998; Wilson *et al.*2001, 2004; Clarke *et al.* 2002; Clarke & Rendell 2009; Clemmensen *et al*. 2009; Reimann *et al.* 2011) report extensive aeolian sand movement and transgressive dune formation along north-western European shores during the Little Ice Age (LIA) between approximately AD 1350 and AD 1900 (Hass 1996). This phase of large-scale aeolian sand movement has been related to increased North Atlantic storminess (e.g. Clarke & Rendell 2009; Clemmensen *et al.* 2009). Aeolian sand invasion was particular severe along the Danish west coast of Jutland where transgressive dunes formed by persistent westerly storm winds migrated up to 10 km inland during this period (Clemmensen *et al.* 2007; Clemmensen *et al.* 2009). Also the Skagen Odde spit system in northernmost Denmark (Fig. 1) saw extensive inland sand movement in this period, and the northernmost part of the spit landscape became almost completely covered by aeolian dunes and sand plains (Clemmensen & Murray 2006). The majority of these transgressive (parabolic) dunes are now stabilized by vegetation (Anthonsen *et al.* 1996; Clemmensen & Murray 2006; Clemmensen *et al.* 2014).

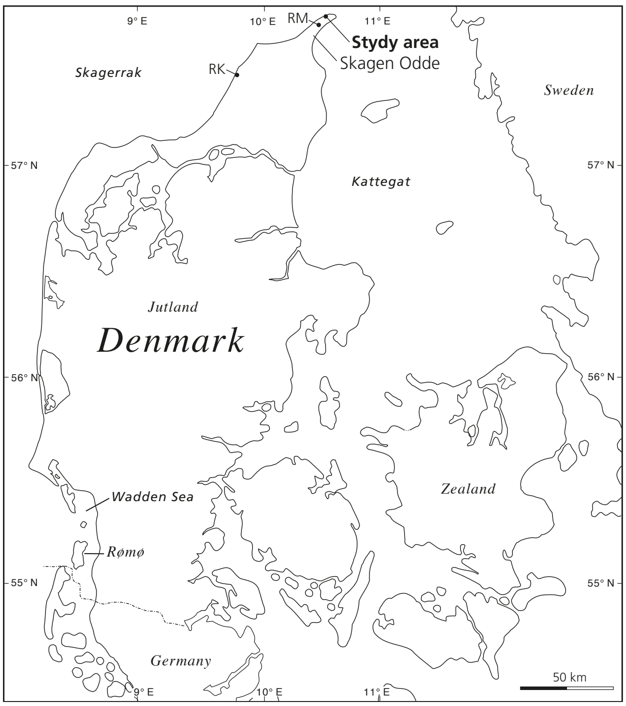

Fig. 1. Location map. The study area is situtated at the north-western side of the Skagen Odde spit system (see Fig. 2). Locations of Råbjerg Mile (RM) and Rubjerg Knude (RK) are given.

On Skagen Odde and in nearby areas along the west coast of Jutland, the termination of sand movement and transgressive dune formation during the LIA is relatively well dated to the end of the 19th century (Clemmensen & Murray 2006). However, the initiation and early history of this event is poorly dated. This is primarily thought to reflect that the first sand deposited during this event was frequently reworked during later periods of sand movement, so that material available for dating of the initiation of the event therefore is missing at most places (Clemmensen & Murray 2006). Aeolian sediments deposited during the LIA are exposed in coastal cliffs south of Højen on the north-western side of the Skagen Odde spit (Fig. 2). Due to coastal retreat the aeolian sediments represent wind-blown sand originally deposited up to 1300 m inland.

The purpose of this study is to investigate these exposures, to document the sedimentary characteristics of the aeolian sand, and to use optically stimulated luminescence (OSL) dating to obtain ages of the initiation and later phases of the LIA sand invasion on Skagen Odde. The OSL ages presented in this study are discussed in relation to human impact on the dune environment (e.g. Brüel 1918; Jessen 1936; Hansen 1964; Hauerbach 1992), historical records of sand invasion (Brüel 1918; Clemmensen & Murray 2006), observational and instrumental records of wind climate (Anthonsen *et al.* 1996; Clemmensen *et al.* 2014), and theoretical considerations on North Atlantic storminess during the LIA (e.g. Raible *et al.* 2007, 2008; Trouet *et al.* 2009; Trouet *et al.* 2012;Van Vliet-Lanoë *et al.* 2014). Data extracted from the sedimentary characteristics of the coastal dune succession at Skagen Odde, together with new age control of the aeolian activity phases, allow us to discuss the conditions that led to increased storminess and widespread aeolian sand movement in large parts of north-western Europe during the LIA.

Study area

Skagen Odde (northern Jutland, Denmark) is one of the largest spit systems in Europe. The spit, which now has a length of more than 30 km, developed over the past 7000 years and spit growth towards the north-east is linked to a relatively continuous supply of marine sand and gravel transported north-eastward by longshore currents (Petersen 1991; Hauerbach 1992; Clemmensen *et al.* 2001b; Nielsen & Johannessen 2008). Skagen Odde lies between the waters of Skagerrak (high-energy wave climate) and Kattegat (less high-energy wave climate) (Fig. 1). The spit experiences uplift due to isostatic rebound with present values around 1.5 mm/year (Clemmensen *et al.* 2001b). The western part of the spit from Højen and southwards (Fig. 2) is retreating and nice exposures of the spit deposits are frequently seen, especially after storm and wave erosion of the cliffs, or after strong winds.

The spit deposits are composed of a lowermost marine succession overlain by swale peat (martørv) and an uppermost succession of aeolian deposits (Clemmensen *et al.* 2001b; Clemmensen & Murray 2010). For this study, four closely situated sites immediately north of Pælebakke Klit 300–1000 m south of Højen on the north-western side of the spit were selected (Fig. 2). Frontal dunes reaching heights up to 20 m above sea level have developed; the frontal dunes are separated by large wind gaps leading inland to dome-shaped dunes or incipient parabolic dunes. Wind gaps and related dome-shaped dunes reach up to about 300 m inland. At the study area the exposed part of the marine spit succession is relatively thin and composed of upper shoreface and beach conglomerates typically

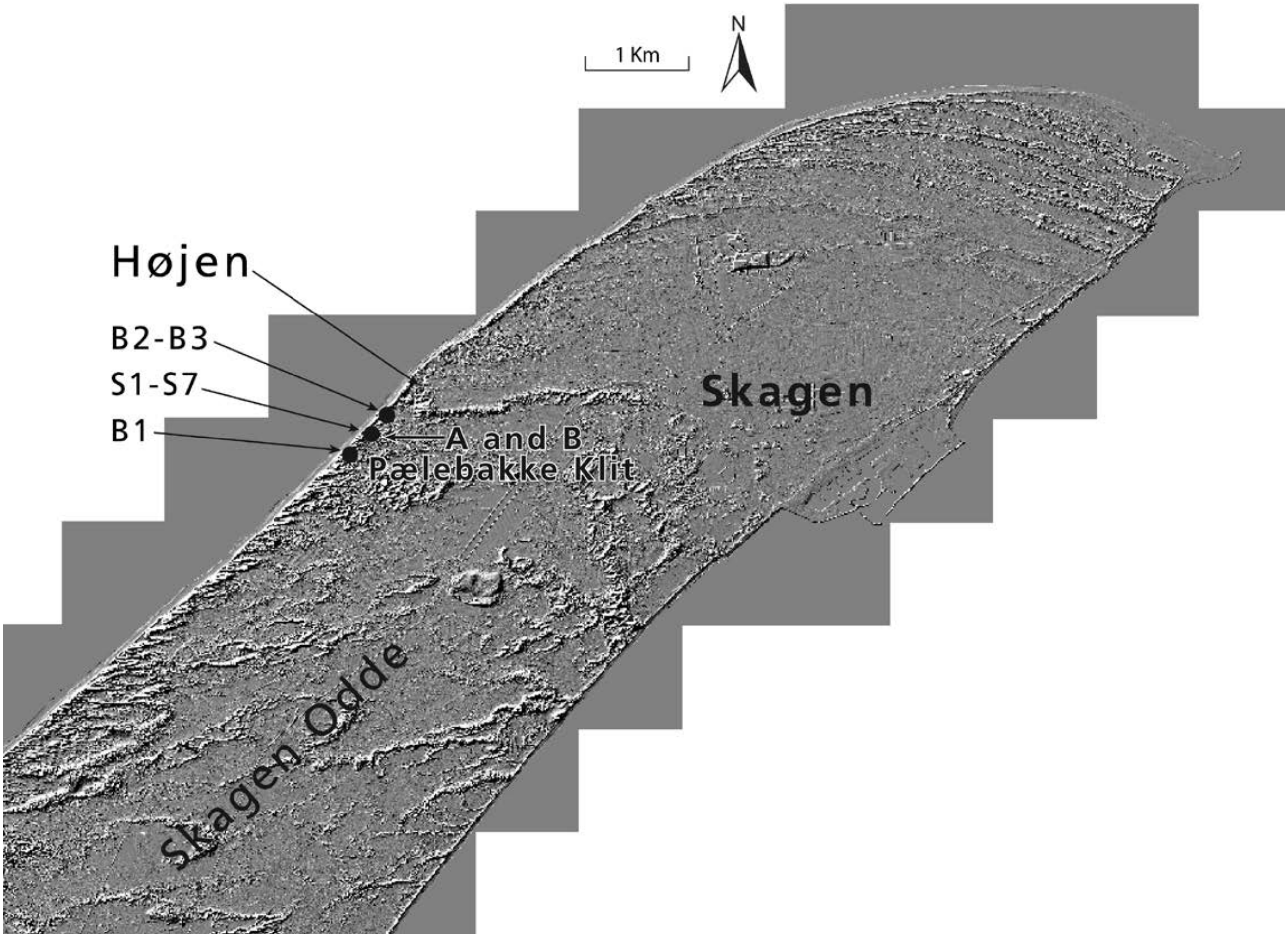

Fig. 2. LIDAR map of the northernmost part of the Skagen Odde spit system. Study area with sample sites at Pælebakke Klit (sites A and B, samples S1–S7), as well as sample sites for beach deposits (samples B1–B3), are indicated (Table 1). Note the slightly curved, W–E trending dune ridges and swales along the northwards facing and prograding spit coast, as well as active blowouts and incipient parabolic dunes along the westward facing and retreating coast south of Højen (see also Clemmensen *et al.* 2014). Coordinates (UTM zone 32; WGS 84): site A: E0589902, N6398834; site B: E0589889; N6398819.

overlain by a thin package of stratified beach sand. The beach deposits are at many places overlain by martørv; at the study area the martørv is only up to about 0.2 m thick, but elsewhere the martørv reaches thicknesses up to 1 m and locally 1.3 m (Clemmensen *et al.* 2001b). The aeolian sand that covers the martørv shows lateral variation in thickness and stratigraphy, but four units divided by thin soil horizons are recognized (Figs 3, 4). As the soil horizons locally are weakly developed or missing due to wind erosion at the beginning of the next aeolian phase, it is difficult to measure a complete succession at one site. The aeolian stratigraphy given here is therefore based on sedimentological observations at two closely situated sites (Figs 3, 4).

Skagen Odde is located in a cool, temperate climate region with an annual precipitation of 635 mm, a potential evaporation of 555 m/year, and an average temperature of 7.9° C (Scharling 2000). The Skagen Odde spit lies in a high-energy wind belt, and modern (AD 1991–2010) values for drift potential (DP) are as high as 274 vector units (VU, m/s) or 2014 VU (knots) (Clemmensen *et al.* 2014). Resultant drift potential (RDP) is 142 VU (knots) or 1044 VU (m/s) and directed towards north-north-east. Storminess is relatively high and wind events of Beaufort 8 and higher occur with typical annual frequencies between 5 and 10 percent (Clemmensen *et al.* 2014).

Methods and sampling

This study is based on map information of shoreline history and dune development, field studies of aeolian sedimentology and geomorphology, and optically stimulated luminescence (OSL) dating of aeolian sand.

The first detailed map of the study area is the one in Resen's Atlas Daniscus from AD 1677 (see map on page 40 and 41 in Lønstrup & Nielsen 1995); then follows the map of The Royal Danish Academy of Sci-

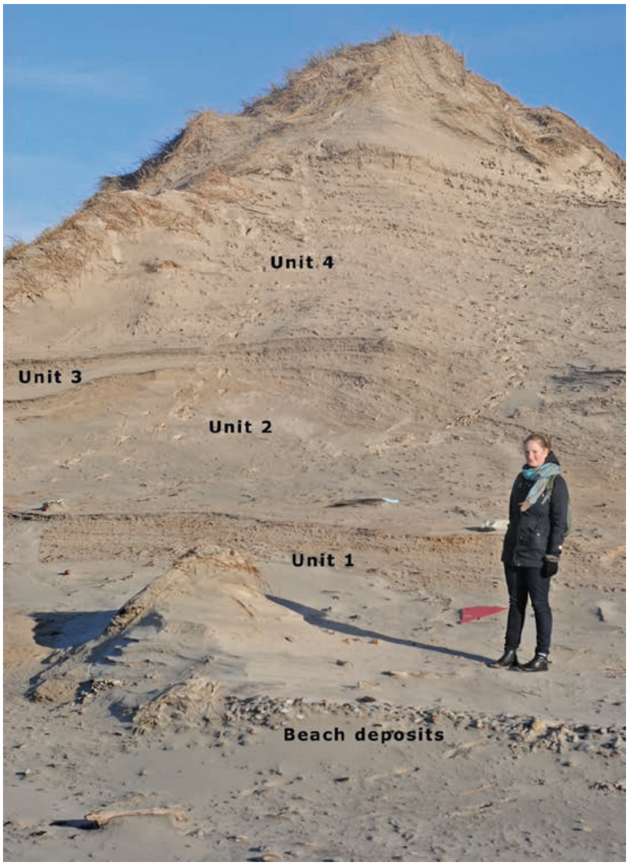

Fig. 3. Photo of coastal exposures at Pælebakke Klit (site B). Four aeolian sand units are recognized, divided by thin soil horizons. Note beach conglomerate at the base of the section. The beach deposits are overlain by swale peat, martørv (not exposed on the photo). The base of the swale peat has an age around AD 500 (Clemmensen *et al*. 2001b).

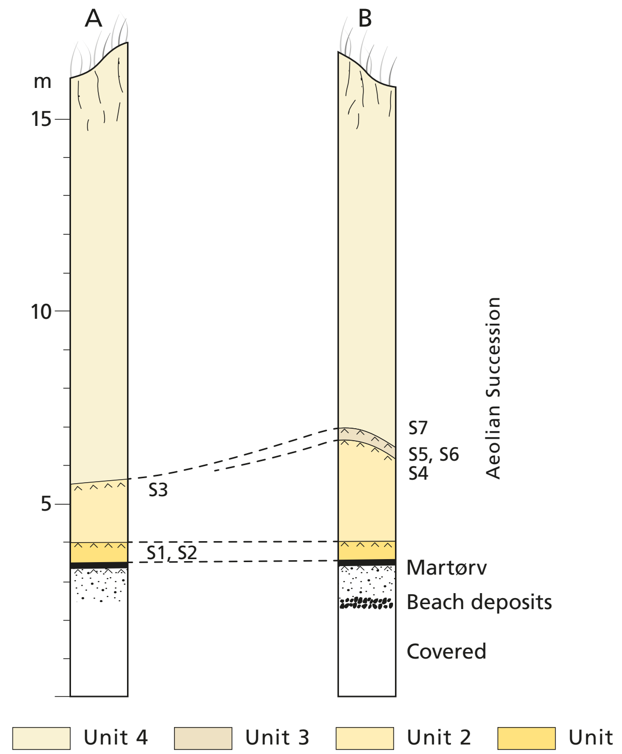

Fig. 4 Sedimentary successions at Pælebakke Klit (sites A and B). The distance between the two sites is around 20 m. Aeolian sand units 1–4 and OSL sample levels (S1–S7) are indicated. Soils are indicated by thin black lines and inverted v's. For details of OSL samples, see Table 1.

ences and Letters from AD 1793, and topographical maps from AD 1887, AD 1968 and AD 1977 (source: Geodatastyrelsen; http://eng.gst.dk). The most recent development of the area is documented by orthophotographs.

Due to changing degrees of exposure (from almost completely covered by slope deposits to windswept and clean) it is rarely possible to restudy the same exposure. The study area was initially visited in May 2012 (site A; Fig. 2). martørv at the base of the exposure was overlain by about 15 m of aeolian sand, and aeolian units 1, 2 and 4 were identified (Fig. 4). A closely situated section was studied in February 2014 and August 2014 (site B). Beach deposits and martørv at the base were here overlain by about 15 m of aeolian sand, and aeolian units 1, 2, 3, and 4 were identified (Figs 3, 4).

Samples for OSL dating were collected in the two adjacent coastal sections (sites A and B). In section A, two samples (124803=S1 and 124802=S2) were taken in unit 1, and one sample (124801=S3) was taken in unit 2 (Fig. 4; Table 1). In the closely situated section B, an additional sample (144701=S4) was taken in unit 2, two samples (144702=S5 and 144703=S6) were taken in unit 3, and one sample (144704=S7) was taken in unit 4 (Fig. 4; Table 1).

In the underlying marine spit sediment one OSL sample (024811=B1) was taken immediately south-west of sites A and B, and two OSL samples (024809=B2 and 094801=B3) were collected immediately north-east of sites A and B (Fig. 2). These OSL samples were collected in sandy beach deposits at the top of the marine succession approximately 0.5 m below the martørv.

OSL samples were dated at The Nordic Laboratory for Luminescence Dating, DTU Risø Campus. For sampling procedures and dating methodology, see Clemmensen & Murray (2010). Results are shown in Table 1 where ages are given as years before dating in the laboratory; in the text the ages are also given as calendar years.

Table 1. OSL ages of sand samples from Pælebakke Klit, Skagen Odde

Risø no.	Field no.	Unit	Depth, cm	Age (Kyr) ± 1σ	Dose (Gy)	n	Dose rate (Gy/Kyr)	w.c. %
124803	S1	AU 1	530	0.57 ± 0.04	0.56 ± 0.02	27	0.98 ± 0.05	16
124802	S2	AU 1	530	0.53 ± 0.04	0.52 ± 0.02	26	0.98 ± 0.05	11
124801	S3	AU 2	420	0.28 ± 0.02	0.30 ± 0.002	26	1.09 ±0.05	10
144701	S4	AU 2	560	0.237 ± 0.015	0.255 ± 0.006	24	1.08 ± 0.06	12
144702	S5	AU 3	510	0.162 ± 0.012	0.200 ± 0.010	23	1.23 ± 0.06	14
144703	S6	AU 3	510	0.130 ± 0.009	0.152 ± 0.008	23	1.17 ± 0.06	14
144704	S7	AU 4	490	0.080 ± 0.007	0.084 ± 0.005	23	1.05 ± 0.05	20
024811	B1	Be. dep.	500	1.87 ± 0.16	1.97 ± 0.14	27	1.05 ± 0.05	5
024809	B2	Be. dep	500	1.83 ± 0.15	1.88 ± 0.10	30	1.03 ± 0.06	5
094801	B3	Be. dep	500	1.35 ± 0.09	1.80 ± 0.06	29	1.33 ± 0.07	4

AU: aeolian unit. Be. dep: underlying beach deposits. Depth is given as 50% of the present depth beneath the top of the cliff face as it is assumed that aeolian sand accumulated continuously over time. Year of dating is given by the first two figures in the Risø no., i.e. sample 124803 was dated in 2012. Gy: Dose measured in Gray units, the metric (SI) unit of absorbed radiation dose of ionizing radiation. n: number of aliquots measured to give the Dose. w.c.: water content.

Spit evolution

Spit growth

During the last 5000 years the spit has been growing to the north-east at a rate of 2–10 m/year, and since AD 1695 more than 2 km has been added (Clemmensen *et al.* 2001b; Hauerbach 1992). Spit growth was accompanied by the formation of slightly curved, W–E trending and coast-parallel dune ridges separated by low-lying swales (Nielsen & Johannessen 2008; Clemmensen *et al.* 2014; Fig. 2). Peat developed relatively quickly in these swales and was with time transformed into relatively dense martørv. Radiocarbon dating of selected swale peats along a 15 km stretch of the spit forms the basis for an age model of spit evolution (Clemmensen *et al.* 2001b). According to this model, swale peat (martørv) started to develop at the study site at about AD 500. It is not documented when swale peat formation ended, but it is likely that peat formation in the swale continued until the swale was covered by aeolian sand during the beginning of the LIA.

Age results

In this work, supplementary age control on the spit deposits that underlie the aeolian sediments is provided by the three OSL samples B1–B3 from the topmost part of the beach deposits immediately below the swale peat (Fig. 2). These samples (094801, 024809, 024811; B3, B2, B1) have ages of, respectively, 1350±90 years (AD 659±90), 1830±150 years (AD 172±150), and 1870±160 years (AD 132±170) (Table 1).

Formation of aeolian sand units

Between AD 500 and the initiation of large-scale aeolian sand movement, spit growth continued, and the studied site at Pælebakke Klit was gradually shifted from a position at the relatively protected northward growing part of the spit to a position on the westward facing and more wind-exposed part of the spit. This shift in position also placed the study site in an area of coastal retreat.

Map information shows that the studied part of the coastal section has retreated landwards about 2.5 m/year since AD 1793 and about 1.3 m/year in most recent years. This indicates that the study area was situated almost 600 m inland when the map by The Royal Danish Academy of Sciences and Letters was measured in AD 1793 and therefore probably up to 1250 m inland 500 years ago. The map in Resen's Atlas Daniscus from AD 1677 shows a line of relatively tall frontal dunes along the west coast of the spit, grading inland into smaller and less well defined dunes; aeolian sand covered most of the spit already in AD 1677. The map from AD 1793 also indicates that most of the northern part of Skagen Odde was covered by aeolian sand and sand dunes. The map is, however, not detailed enough to show the geomorphological form of these inland dunes. On the topographical map from AD 1887 the inland dunes are named and their morphology is mapped in great detail with contour intervals of 5 feet (1.5 m). Most larger dunes are parabolic in shape and their morphology indicates that they have moved from the west toward the east; these dunes are only partly stabilized by vegetation. Comparisons of the map from AD 1887 with the maps from the 20th century indicate that vegetation cover has increased with time (Anthonsen *et al.* 1996) and most inland dunes are now to a large degree covered by vegetation; most dunes have not moved detectably since the end of the 19th century. Map and orthophoto studies also show that frontal dunes, wind gaps and incipient parabolic dunes along the north-western shore of the spit are shifted inland in pace with coastal retreat, but the incipient parabolic dunes do not develop into true transgressive dunes (Clemmensen *et al.* 2014).

Sedimentological field studies indicate that the aeolian sand that lies on top of the martørv is composed of four units (Figs 3, 4). Unit 1 lies directly on top of the martørv and is up to 1 m thick. It forms a sheet-like to low-relief undulatory sand unit and is covered by a 2–10 cm thick peaty soil horizon underlain by a dense network of roots. The sand has a pale yellowish grey colour (Fig. 3), is medium-grained with a mean grain size of 0.27 mm and is well sorted (moment sorting=0.36). The main part of the sediment is characterized by horizontal to low-angle stratification, but locally steeper dipping sets are developed. The stratification is outlined by subtle variations in grain size with the coarsest grains concentrated in thin layers. The sediment of unit 1 is interpreted as a sand accumulation formed by inland sand drift and dune migration. Similar low-angle stratified aeolian sheet sand and associated cross-stratified dune deposits have been described from Holocene dunefield deposits in Thy, north-western Denmark, and are also related to inland sand movement (Clemmensen *et al.* 2001a; Pedersen & Clemmensen 2005).

The second aeolian sand unit is up to 3.0 m thick and is covered by an immature soil. It forms a sheet-like to low-relief undulatory unit (Figs 3, 4). The sand has a pale yellowish grey colour, is medium-grained with a mean grain size of 0.29 mm and is well sorted (moment sorting=0.36). The main part of unit 2 is characterized by low-angle stratification, but in its upper part convex-up cross-bedding with steeper dip angles is developed. Also this unit is interpreted to have formed by inland sand drift and dune migration.

Aeolian unit 3 is only locally developed. It lies between the two incipient soil horizons (Figs 3–4) and has a thickness of about 0.5 m; it has a pale yellowish grey colour, is medium-grained with a mean grain size of 0.29 mm and is well sorted (moment sorting=0.36). Also this unit is characterized by low-angle to convex-up stratification, and the stratification seems to drape the underlying incipient soil. Sedimentary structures have typically been obliterated in its upper part due to soil formation and the presence of small roots.

The uppermost aeolian sand unit is up to 10 m thick and its upper part forms the modern frontal dunes (Figs 3, 4). The sand is fine- to medium-grained; a sample from the base of the unit has a mean grain size of 0.28 mm and is well sorted (moment sorting= 0.36). The sand is paler than the underlying aeolian sand; it has horizontal and low-angle stratification but eastward dipping cross-strata are also present, particularly in the uppermost part of the unit. The sediment of unit 4 formed by accumulation of sand on the frontal dunes and on closely associated dome-shaped dunes. Cross-strata structures indicate inland transport of sand.

Age results

OSL dating results are shown in Table 1. Two closely spaced samples in aeolian sand unit 1 (S1 and S2, site A) gave ages of 570±40 years (AD 1442±40) and 530±40 years (AD 1482±40), respectively. Two samples in aeolian unit 2 (S3, site A, and S4, site B) gave ages of 280±20 years (AD 1732±20) and 237±15 years (AD 1777±15), respectively. Two closely spaced samples in

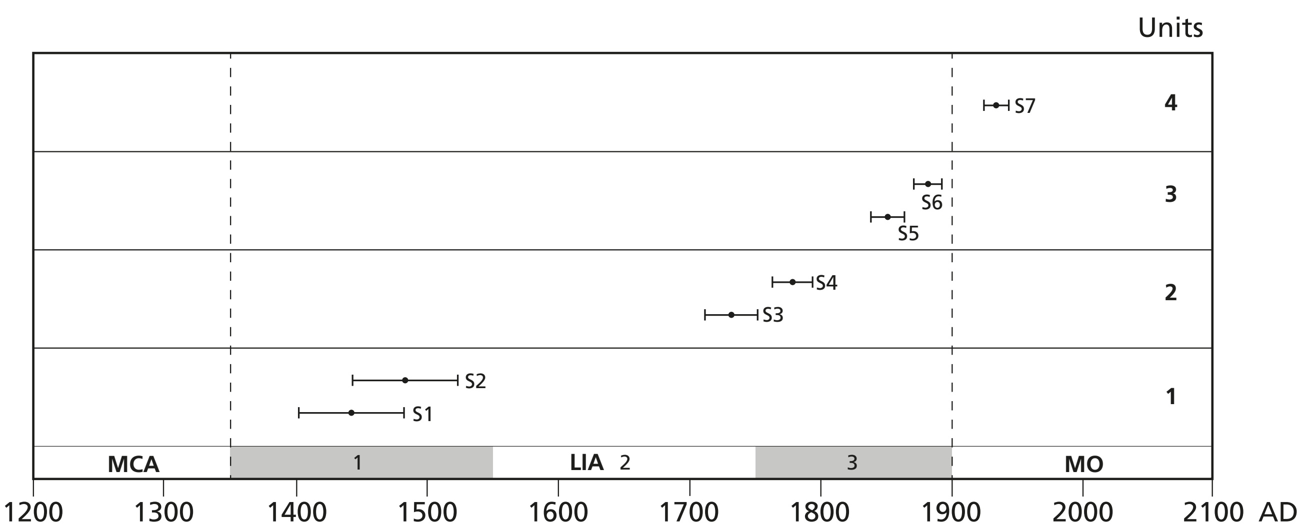

Fig. 5. OSL ages of the aeolian sand units, Pælebakke Klit. Calculated ages are given as small dots with error bars. Note that the first three units were deposited during the the Little Ice Age (LIA) while the fourth unit is modern. Units 1–3 are well separated in time, suggesting episodic aeolian sand movement during the LIA. MCA=Medieval Climatic Anomaly; MO=Modern Climatic Optimum. Climatic subdivision of LIA into stormy phases 1 and 3 (grey), and a relatively calm phase 2 (white) is indicated (Hass 1996).

unit 3 (S5 and S6, site B) gaves ages of 162±12 years (AD 1852±12) and 130±9 years (AD 1884±9), respectively, and one sample (S7, site B) from the basal part of the frontal dune unit (unit 4) gave an age of 80±7 years (AD 1934±7) (Fig. 4; Table 1).

Discussion

Timing of events

The stratigraphy of the study site and the OSL dating results suggest that inland sand movement and transgressive dune formation took place in three phases prior to the most recent formation of frontal dunes (Fig. 5). Each episode of sand mobilization was separated by a phase of stabilization or near-stabilization of the aeolian landscape as suggested by the thin soil horizons of which the lowermost one is best developed. The three phases of aeolian sand movement all fall within the climatic period defined as the Little Ice Age (LIA). The Little Ice Age was the most recent cold period in much of the North Atlantic region; it followed the warm period during the Medieval Climate Anomaly (MCA; Fig. 5) (Trouet *et al.* 2012). The duration of the LIA has been defined differently by different authors. According to Hass (1996), the LIA covers the time span between AD 1350 and AD 1900; Trouet *et al.* (2012) places the LIA between AD 1400 and AD 1800, while Clarke & Rendell (2009, 2011) restrict the LIA to the time period between AD 1570 and AD 1890 (1900). Data from raised peat bogs in south-western Sweden suggest that the LIA took place between AD 1350 an AD 1850 (Björck & Clemmensen 2004). We here follow the definition of Hass (1996) and define the LIA to cover the time span between AD 1350 and AD 1900.

Multi-proxy data on temperature on the northern hemisphere indicate that there were climatic fluctuations during the LIA, and a reconstruction of climate change during the last 2000 years by Christiansen & Ljungquist (2012) shows that the MCA peaked around AD 1000; thereafter temperature decreased stepwise with cold intervals around AD 1300, AD 1450, and AD 1650. After AD 1650 temperatures increased again but there was a final cold interval around AD 1850. According to Hass (1996), there were three climatic periods during the LIA: a first phase between AD 1350 and AD 1550 with increased storminess, a second phase between AD 1550 and AD 1750 with decreased storminess, and a third phase between AD 1750 and AD 1900 again with increased storminess (Fig. 5).

Large-scale sand movement and transgressive dune formation at the northernmost part of Skagen Odde is here interpreted to have started around AD 1460 (the mean age of samples S1 and S2; Fig. 5). The preservation of low-relief dune forms in the lowermost unit indicates that at least part of the sand movement inland was related to dune migration. Modern dunes in the northern part of Skagen Odde have mean grain sizes between 0.20 and 0.23 mm (Clemmensen *et al.*2014) and are thus more fine-grained than the sand of the lower sand unit. This is tentatively taken to indicate that inland transport of sand during formation of the lowermost sand unit took place during strong winds.

As the study site was probably situated up to 1250 m inland 500 years ago (see above), it can be deduced that sand movement at the coast most likely was initiated earlier than around AD 1460. Knowing that the modern parabolic dune Råbjerg Mile migrates around 10 m inland per year (Anthonsen *et al.* 1996), it is inferred that inland sand movement at the coast near Højen may have started 125 years earlier or around AD 1335. If the sand moved inland as sand drift (sand sheets), however, sand could have shifted inland much faster. According to Pye & Tsoar (2009), sand grains in the 0.28–0.48 mm size range have a mean forward velocity of 0.6–1.8 cm/s during sand drift. Using 1 cm/s as a preferred value, inland movement of sand could be as much as 860 m per day, indicating that sand drift initiated at the coast could reach the study site in about two days. Such a rapid inland transport of sand is of course unlikely as existing vegetation and topography would have slowed down the sand drift, but it is possible that sand could have been transported inland from the coast to the study site within a few years.

Historical reports first mention sand drift on Skagen Odde around AD 1571 (Brüel 1918; Hansen 1964). Farther south along the west coast of Jutland there are reports of sand drift around AD 1553 (Brüel 1918). The new OSL dates indicate that the first phase of sand movement started somewhat earlier than mentioned in historical records.

The second phase of inland sand movement and transgressive dune formation is documented by OSL ages from both sites and is interpreted to have taken place between AD 1730 and AD 1780 (Fig. 5). The study area was then situated some 700 m inland, suggesting that sand movement at the coast could have started 70 years earlier or around AD 1660. This is in agreement with historical sources that mention severe sand drift in the middle and later part of the 17th century (Brüel 1918; Hansen 1964). If sand movement was associated with sand drift and sand sheet formation, inland movement could be much faster as discussed above. Sand unit 2 at Skagen Odde is coarser grained than modern dune sand in the area, suggesting that inland transport was linked to strong winds. Sand movement on Skagen Odde during this second phase was severe and sand reached the Sanct Laurentii Church near the town of Skagen around AD 1775; the church was abandoned in AD 1793. As the distance between the study site and the church is about 2800 m in a downwind direction, it is likely that sand from nearby sources (stabilized or partly stabilized dunes) was reactivated during this event. This second phase of sand movement may have affected large coastal parts of northern Jutland, and the basal part of aeolian sand in the cliff-top dune at Rubjerg Knude (a coastal cliff 55 km south of the study area; Fig. 1) has an OSL age of 274±14 years (AD 1726+14) (Saye *et al*. 2006).

The third episode of sand movement is dated to around AD 1868 (mean age of samples S5 and S6; Fig. 5). As the study area 150 years ago was situated only some 375 m inland, this event only records limited inland transport of sand from frontal dunes and closely situated dome-shaped dunes. Also this episode may have affected a large part of Skagen Odde, and a sample from the basal part of an uppermost aeolian dune unit 10.9 km south of Pælebakke Klit has an age of 189±11 years or AD 1876±11 (Clemmensen & Murray 2010).

The fourth and final episode of sand movement is related to the formation of frontal dunes along the present coast. OSL dating at the base of this unit at site B indicates that these frontal dunes started to develop around AD 1934 (Fig. 5). However, as these frontal dunes are a type of roll-over dunes constantly reforming inland in pace with coastal retreat, this date only yields an age of the oldest frontal dune material presently preserved in the exposure.

Thus it is concluded that there were three phases of aeolian sand movement on Skagen Odde during the LIA; the first one was initiated around AD 1460 (or possible as early as AD 1335), the second phase was initiated around AD 1730 (or possible already around AD 1660), while the third phase took place around AD 1870. Sand drift was episodic as indicated by the development of thin soils. The soil separating unit 1 and 2 is best developed, suggesting that after the first sand drift event the dune landcape became stabilized or near-stabilized. Judged from the new OSL dates this period of near-stabilization took place after AD 1550 but before AD 1700.

Causes and climate connections

Having established that sand invasion on Skagen Odde during the LIA took place in three main phases, it remains to be discussed what caused the sand mobilization and how these phases of sand movement may be linked to climate and/or anthropogenic influence. Skagen Odde lies in a high-energy wind climate, but observational and instrumental data indicate that the wind climate has varied a lot the last 150 years (Clemmensen *et al.* 2014). The storminess level was extremely high around AD 1870 and has been decreasing since then. Also the drift potential (DP) was outstandingly high around AD 1870, reaching annual values of 9600 VU (knots), while the DP levels were around 2000–2500 VU over the last decades and are still decreasing (Clemmensen *et al.* 2014). Yizhaq *et al.* (2009) found that dunes are typically active when DP values are higher than 2000–3000 VU (knots), irrespective of yearly precipitation. In agreement with these data, dunes on Skagen Odde have been stabilized during the last 150 years in response to the

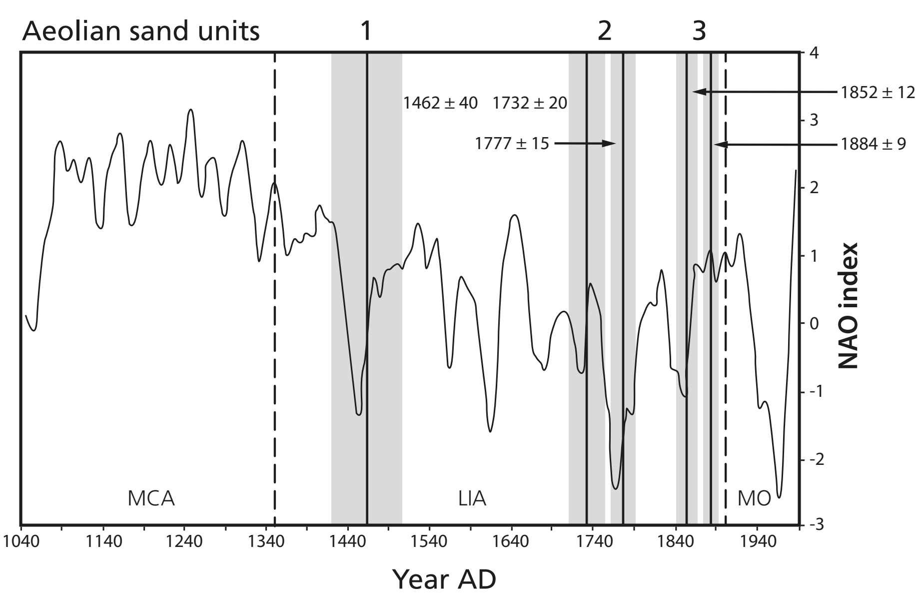

Fig. 6. Values of the North Atlantic Oscillation (NAO) index versus the ages of the first three phases of aeolian sand movement (units 1–3), Pælebakke Klit. The age of the first event is given as the mean of samples S1 and S2 as they overlap in age. The grey bands are the uncertainty intervals. MCA=Medieval Climatic Anomaly; MO=Modern Climatic Optimum. The three phases of aeolian sand movement occurred after AD 1460 during the LIA when overall NAO index values were low. In particular the first phase of sand movement around AD 1460 is clearly linked to a marked drop in NAO index. NAO index values from Trouet *et al*. (2012).

decrease in storminess and dune management work (Clemmensen *et al.* 2014). In the same period Råbjerg Mile has changed form from a crescentic dune to a parabolic dune (Anthonsen *et al.* 1996).

The third phase of aeolian sand movement (unit 3), dated to have taken place around AD 1870, occurred during an interval when wind energy was exceptionally high at Skagen Odde (Clemmensen *et al.* 2014). Sand activation took place in spite of ongoing coastal dune management. If this third event is taken to be representative for the two earlier events, it may be inferred that also these events were linked to an exceptional high-energy wind climate. Judged by the amount of sand that was shifted inland during the firsts two events, storminess then may have been even higher. In addition, the removal of parts of the vegetation on the frontal dunes by local inhabitants during the early and middle part of the LIA made these dunes more vulnerable to remobilization. In support of a cultural imprint, especially on the first sand drift event, is the fact that the Danish King Christian III in AD 1539 promulgated a law forbidding the removal of vegetation in the dunes, implying that dune vegetation may have been partly removed at the beginning of the LIA (Kjærgaard 1994).

According to Trouet *et al.* (2009, 2012), the transition from the MCA to the LIA was characterized by a major change in atmospheric circulation. A reconstruction of the North Atlantic Oscillation (NAO) index from proxy data shows that NAO values were positive during the MCA and that there was a marked drop to negative values around AD 1450, at the beginning of the LIA (Fig. 6). NAO values fluctuated between negative and moderately positive during the LIA. Intervals with particularly negative values are seen around AD 1620 and AD 1760. During intervals with persistent positive NAO values, enhanced winter storminess is assumed to influence north-western Europe (Trouet *et al.* 2009, 2012). In contrast, blocking anticyclones at mid-latitudes during negative NAO values should result in decreased winter storminess. However, according to Trouet *et al.* (2012), periods of negative NAO values as seen frequently during the LIA are also characterized by an increased intensity of the cyclones.

Van Vliet-Lanoë *et al.* (2014) relate storminess during the late Holocene to variations in NAO as well as in the Atlantic Multidecadal Oscillation (AMO). The AMO describes variations in the sea surface temperatures of the North Atlantic Ocean. Cool (negative AMO) and warm (positive AMO) phases alternative on decadal scale.

According to Van Vliet-Lanoë *et al.* (2014), increased storminess in the North Atlantic region is linked to two different climatic scenarios: negative NAO mode combined with positive AMO values, and negative NAO mode combined with negative AMO mode. The first climatic scenario characterized the period between AD 1830 and AD 1890 (Van Vliet-Lanoë *et al.* 2014) and could probably explain the enhanced storminess around AD 1870 documented by wind data from Skagen Odde (Clemmensen *et al.* 2014) and by the new OSL ages. A similar climatic setting may also explain the enhanced storminess at Skagen Odde between AD 1730 and AD 1780 deduced from the new OSL dates, as much of the 18th century was characterized by a negative NAO mode (especially around AD 1750) but also had a number of prominent positive AMO events (Trouet *et al.* 2012; Van Vliet-Lanoë *et al.* 2014). The second climatic scenario characterizes the period between AD 1400 and AD 1480 when both the NAO and AMO modes were negative (Trouet *et al.*2012; Van Vliet-Lanoë *et al.* 2014). Enhanced storminess and the initiation of the first aeolian sand movement at Skagen Odde can probably be linked to this second climatic setting. Studies of wind data at Skagen Odde since AD 1860, however, indicate that there are no statistical links between storminess and NAO and AMO variation on an annual basis (Clemmensen *et al.* 2014). It may therefore be that storminess variation is controlled by decadal or longer term variation in NAO and AMO values, as suggested in this study.

Regional significance

In order to test the regional importance of the new age determination of the initial phase of the LIA event of aeolian sand movement, care should be taken when looking at OSL ages of dune sand because, due to reworking as in most Danish examples, many north-

Table 2. Proxy data of North Atlantic storminess during the beginning of the Little Ice Age

Locality	Data	Age	Reference
Skagen, Denmark	Aeolian sand, coastal dunefield	Around AD 1460	This paper
Wadden Sea, Denmark	Aeolian sand, salt marsh deposits	Around AD 1460	Szkornik *et al*. 2008
Outer Hebrides, Scotland	Aeolian sand, coastal peat mosses	After AD 1400	Dawson *et al*. 2004
Halland, Sweden	Aeolian sand, raised peat bogs	Around AD 1475	Björck & Clemmensen 2004
Iceland	Loess profiles	Around AD 1500	Jackson *et al*. 2005

Ages date onset of aeolian sand movement presumably linked to increased storminess. See text for details.

west European coastal dune deposits would rarely be expected to give the age of the first sand movement. Published dates therefore in most cases would postdate the initiation of sand movement. As an example, transgressive dunes on the barrier island Rømø in the southern part of the Danish Wadden Sea (Fig. 1) have ages of 270 years (AD 1740) and 370 years (AD 1640), and these ages most likely postdate the initiation of aeolian sand movement because the old beach surfaces on top of which the dunes were deposited have ages between 630 years (AD 1380) and 690 years (AD 1320), cf. Madsen *et al*. (2007).

More reliable regional evidence of the initial LIA event would therefore arise from studies of sedimentary archives with relatively continuous and complete records of late Holocene aeolian activity (Table 2). One such record is provided by raised peat bogs in Halland, south-western Sweden. These bogs contain a continuous record of aeolian sand influx (ASI) through time and also cover the transition from the MCA to the LIA. This time interval was characterized by increased aeolian sand influx to the bogs with a first pronounced ASI peak around AD 1475 (Björck & Clemmensen 2004). There was a second pronounced ASI peak around AD 1720 (De Jong *et al.* 2006; Raible *et al.* 2008). These peaks of increased sand influx to the raised bogs, and in particular the abundance of medium-sized sand grains, were used as proxies for storm frequency and intensity. Thus, there is a good match between the timing of onset of sand drift on Skagen Odde around AD 1460 and increased aeolian sand influx to the peat bogs in south-western Sweden around AD 1475.

A second reliable record in this connection comprises aeolian soil profiles in Iceland. Grain size studies of these soil profiles have indicated a number of intervals with increased grain size related to cold and stormy intervals during the Holocene; the last of these intervals was initiated around AD 1500 (Jackson *et al.* 2005). Supplementary evidence on this early event of sand movement is given by aeolian sand layers embedded in fine-grained coastal deposits. An aeolian sand sheet buried in salt marsh deposits in the northern part of the Danish Wadden Sea (Fig. 1) was deposited between AD 1460 and AD 1540 (Szkornik *et al.* 2008). Finally, Dawson *et al.* (2004) examined landward tapering layers of wind-blown sand in coastal peat mosses in the Outer Hebrides, Scotland. These aeolian sand layers are considered to constitute well dated records of increased storminess and many were first produced after AD 1400 at the onset of the LIA (Dawson *et al.* 2004).

Thus, five records from aeolian sediments across large regions in the North Atlantic consistently indicate an increase in storminess between AD 1400 and 1500 (Table 2) around the onset of overall negative NAO values at the beginning of the LIA. The aeolian records are supplemented by data from high-energy estuaries in northern France indicating that an important storm phase in the North Atlantic was initiated around AD 1500 (Sorrel et *al.* 2012).

The new data from Skagen Odde indicate renewed sand movement between AD 1730 and 1780, and around AD 1870. OSL dating of the basal part of aeolian sand sheets suggests that these latter episodes of sand movement also affected areas on the Danish west coast of Jutland south of the study area. However, the chronology of these episodes in a wider regional perspective is poorly known as existing OSL ages of dune sand typically postdate the onset of sand movement.

Conclusions

Sedimentological studies of coastal sections along the north-western part of the Skagen Odde spit system, Denmark, reveal the presence of four aeolian sand units overlying late Holocene beach deposits.

Optically stimulated luminescence (OSL) dating of samples from the coastal exposures indicate four phases of aeolian sand movement: around AD 1460, between AD 1730 and 1780, around AD 1870, and since about AD 1935.

Sand movement in the first three phases occurred during the Little Ice Age and was related to extensive inland dune migration during this period. Periods of stabilization or near-stabilization of the dune landscape separated the three phases of aeolian sand movement. The final fourth phase represents the recent formation of frontal dunes and related coastal retreat.

The first phase of sand movement around AD 1460 was initiated during the beginning of the Little Ice Age. Negative North Atlantic Oscillation (NAO) and Atlantic Multidecadal Oscillation (AMO) at the same time apparently caused a change in the atmospheric circulation leading to more severe cyclones forcing sand to move inland and build dunes.

Both the second and third phase of aeolian sand movement occurred during the Little Ice Age during increased storminess; however they are distinguished from the first phase as negative NAO values probably were linked to positive AMO values during these periods.

Aeolian sand movement during the first phase (AD 1460) at Skagen Odde is in accordance with regional studies of aeolian activity in the North Atlantic, indicating extensive changes in the atmospheric circulation around that time.

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Proterozoic basement and Palaeozoic sediments in the Ringkøbing–Fyn High characterized by zircon U–Pb ages and heavy minerals from Danish onshore wells

METTE OLIVARIUS, HENRIK FRIIS, THOMAS F. KOKFELT & J. RICHARD WILSON

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New data from the Proterozoic basement and scattered Palaeozoic sediments in the Ringkøbing–Fyn High including zircon U–Pb geochronometry, heavy mineral compositions and whole rock geochemistry is presented here to provide a frame of reference for detrital provenance studies. The Ringkøbing–Fyn High is a WNW–ESE trending structural high including subcropping basement rocks, and the results indicate that it is a southerly extension of the Fennoscandian Shield. The zircon data show matching age distribution patterns in crystalline basement rocks obtained from two drill sites, the Glamsbjerg-1 and Grindsted-1 wells. They both record a characteristic Telemarkian accretionary event at 1.51 and 1.48 Ga and a Sveconorwegian metamorphic overprinting at 1.08 Ga. Furthermore, the dominant age intervals in the Glamsbjerg High (1.55–1.48 Ga) and the Grindsted High (1.51–1.44 Ga) suggest that rocks of the Gothian orogeny (that ended at 1.52 Ga) are only present in the eastern part of the Ringkøbing–Fyn High. Thus, the buried basement in central Denmark may be youngest towards the west, which is consistent with the general westward age progression trend in the Sveconorwegian Orogen. The basement breccia in the Arnum-1 well on the southern flank of the Ringkøbing–Fyn High has zircon ages (c. 1.54–1.53 Ga) that resemble those of gneiss in the Glamsbjerg High. The conglomeratic sandstone in the Ringe-1 well on the Glamsbjerg High has a dual age distribution as the matrix has late Palaeoproterozoic to early Mesoproterozoic ages, whereas the granitic clasts have a distinct middle Neoproterozoic age (*c.* 0.76 Ga) that may indicate an Avalonian source. The quartzite in the Slagelse-1 well on the northern flank of the Ringkøbing–Fyn High has a broad age span with late Palaeoproterozoic to late Mesoproterozoic zircon ages.

Supplementary material: Detailed documentation of U/Pb analytical procedures, results and analysed zircon spots are available at http://2dgf.dk/publikationer/bulletin/189bull63.html.

Keywords: Ringkøbing–Fyn High, detrital zircon provenance, Proterozoic basement, Palaeozoic sediments, zircon U–Pb age dating, heavy mineral analysis, Denmark.

Mette Olivarius [mol@geus.dk], Thomas F. Kokfelt [tfk@geus.dk], Geological Survey of Denmark and Greenland (GEUS), Øster Voldgade 10, DK-1350 Copenhagen K, Denmark. Henrik Friis [henrik.friis@geo.au.dk], J. Richard Wilson [jrw@geo.au.dk], also Mette Olivarius, Department of Geoscience, Aarhus University, Høegh-Guldbergs Gade 2, DK-8000 Aarhus C, Denmark.

The Ringkøbing–Fyn High (Fig. 1) consists of elevated basement blocks with a local cover of pre-Zechstein deposits. The overlying sedimentary succession has a thickness of 1–2 km and ranges in age from Zechstein to Quaternary, the main part being Mesozoic. Earlier studies have suggested that the Ringkøbing–Fyn High represents a possible provenance area for some of the sediments deposited on and around the high during the Palaeozoic and Mesozoic eras (Larsen 1966; Poulsen 1969; Dahlgren & Corfu 2001). However, it has not previously been possible to determine whether sediments in the Norwegian–Danish Basin and the North German Basin were in fact sourced by the Ringkøbing–Fyn High, the Fennoscandian Shield or somewhere else because the provenance signatures of the Ringkøbing–Fyn High basement, and the locally overlying Palaeozoic sediments, were unknown. The purpose of this study is to provide data from the

Ringkøbing–Fyn High that can be compared with the thick sedimentary packages deposited north and south of the high and in the grabens that divide its basement blocks.

Zircon age constraints on the timing of accretion and metamorphism of the buried Danish crystalline basement are presented here for the first time. The data reveal that the basement in the Ringkøbing–Fyn High was formed during the same Proterozoic orogenic events that affected south-western Sweden and southern Norway during the Gothian and Telemarkian accretions and the Sveconorwegian metamorphism. Concurring ages are found in a basement breccia deposited on the southern flank of the Ringkøbing–Fyn High. In contrast to the zircon age distributions, the heavy mineral assemblages demonstrate considerable compositional variation within individual gneiss complexes in the high.

Geological setting

The Ringkøbing–Fyn High is a WNW–ESE-oriented structural high that separates the Norwegian–Danish Basin to the north from the North German Basin to the south (Fig. 1). The Ringkøbing–Fyn High is transected by grabens which separate the Grindsted, Glamsbjerg and Møn Highs in the Danish onshore area. It was proposed by Lassen & Thybo (2012) that the thick crust of the Ringkøbing–Fyn High was formed during break-up of Rodinia in the late Ediacaran. Crustal thinning was relatively minor in this area as it was positioned farthest from the Iapetus and Tornquist rifts. The Ringkøbing–Fyn High was identified as a basement high on the basis of the Grindsted-1 and Glamsbjerg-1 wells where cores reaching Proterozoic basement are available (Noe-Nygaard 1963; Sorgenfrei & Buch 1964). The Palaeozoic sedimentary cover on

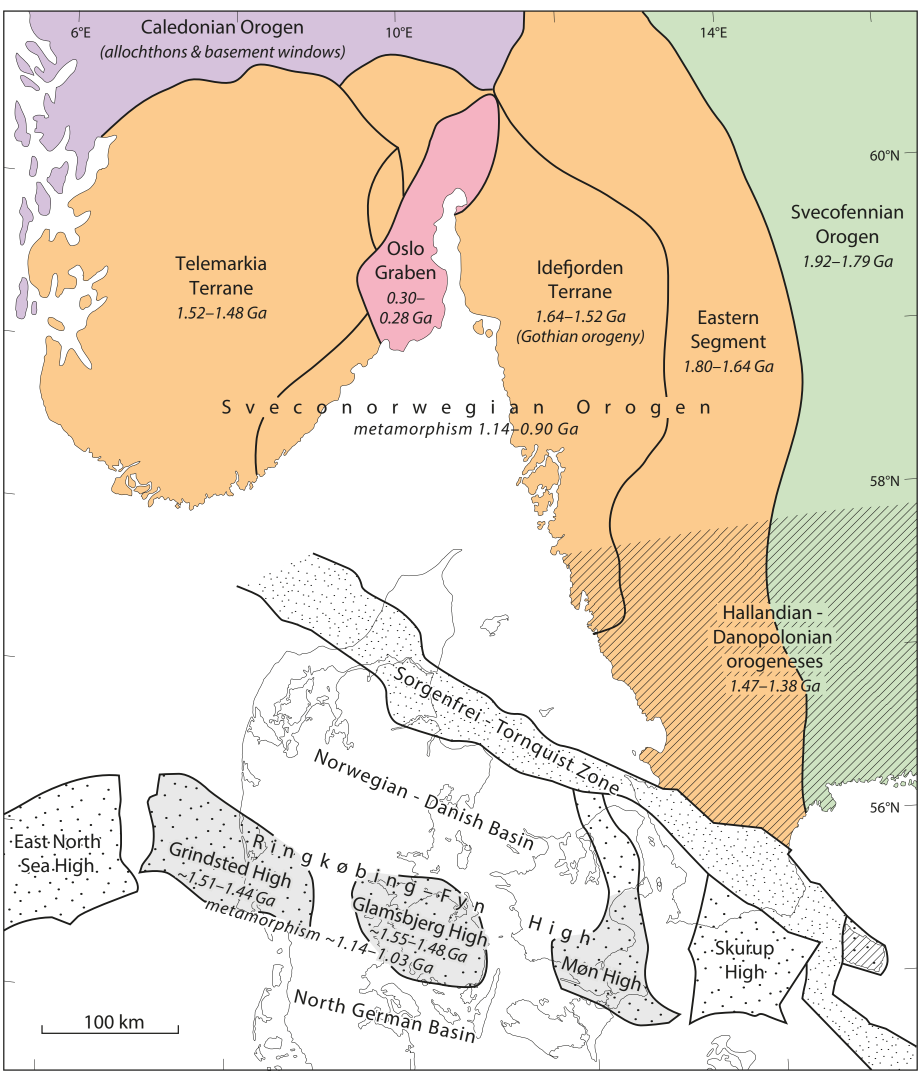

Fig. 1. Position of the Ringkøbing–Fyn High and other structural elements in the Danish area (Vejbæk & Britze 1994; Nielsen 2003) in relation to the major basement provinces in southern Norway and south-western Sweden (Bingen *et al*. 2008a; Bingen & Solli 2009). The basement ages of the Ringkøbing–Fyn High are from this paper (Fig. 9). The ages are in billion years (Ga).

the Ringkøbing–Fyn High was eroded during late Carboniferous to early Permian uplift (Cartwright 1990; Lassen & Thybo 2012), but is locally preserved as documented by the Arnum-1, Ringe-1 and Slagelse-1 wells (Fig. 2) (Nielsen & Japsen 1991). Further uplift of the Ringkøbing–Fyn High took place during Jurassic, Cretaceous and Cenozoic times (Nielsen 2003; Rasmussen 2009).

Lassen & Thybo (2012) suggested that the basement in the Ringkøbing–Fyn High has the same origin as the south-western part of the Fennoscandian Shield that was formed during several orogenic events.

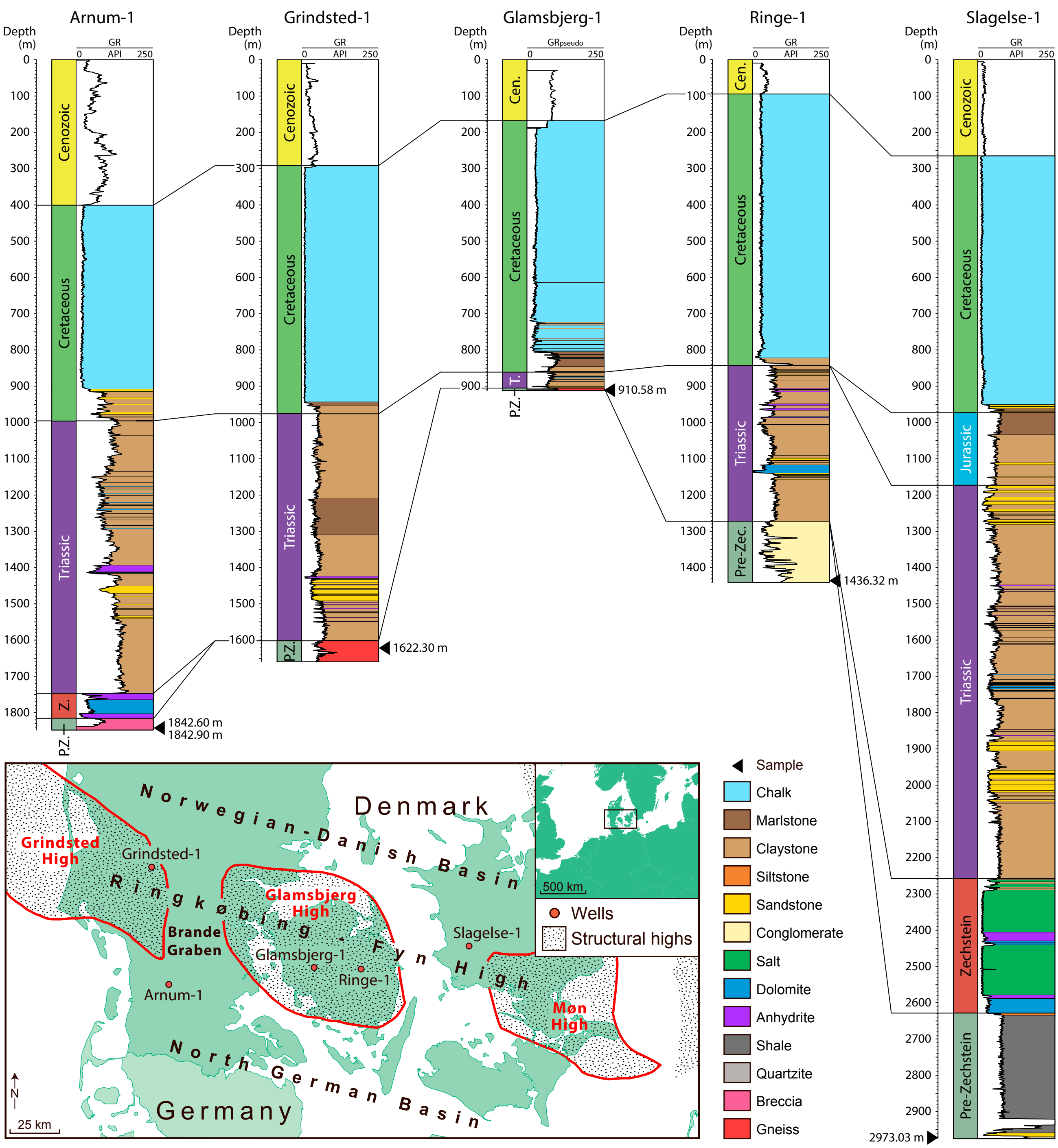

Fig. 2. Location map, lithological columns and gamma-ray logs from the five sampled wells in the Ringkøbing–Fyn High in central Denmark. The log in the Glamsbjerg-1 well is modified from the spontaneous potential log. The outline of the Ringkøbing–Fyn High is from Vejbæk & Britze (1994) and Nielsen (2003).

The Gothian accretion of south-western Sweden (the Idefjorden Terrane) and the Telemarkian accretion of southern Norway (the Telemarkia Terrane) took place at 1.64–1.52 Ga and 1.52–1.48 Ga, respectively (Fig. 1) (Bingen *et al.* 2008a). The Hallandian and Danopolonian orogenic events affected southern Sweden and Bornholm at 1.47–1.38 Ga (Möller *et al.* 2007; Bogdanova *et al.* 2008) where the basement age on Bornholm is constrained to 1.46 Ga (Waight *et al.* 2012). Pre-Sveconorwegian activity was widespread in southern

Fig. 3. Photographs of core pieces from the sampled wells. The two pieces from the Arnum-1 well give an impression of the large variability in the basement breccia.

Norway at 1.34–1.14 Ga (Brewer *et al*. 2004; Bingen *et al*. 2008a). Sveconorwegian metamorphism and associated magmatism at 1.14–0.90 Ga are recorded in the crystalline rocks of both south-western Sweden and southern Norway (Bingen *et al.* 2008b) which were primarily metamorphosed in amphibolite facies.

The remaining basement terranes in southern Norway (Fig. 1) include the allochthons and basement windows of the Caledonian Orogen, which are of diverse origin and age (Corfu *et al.* 2014), and the Carbon-

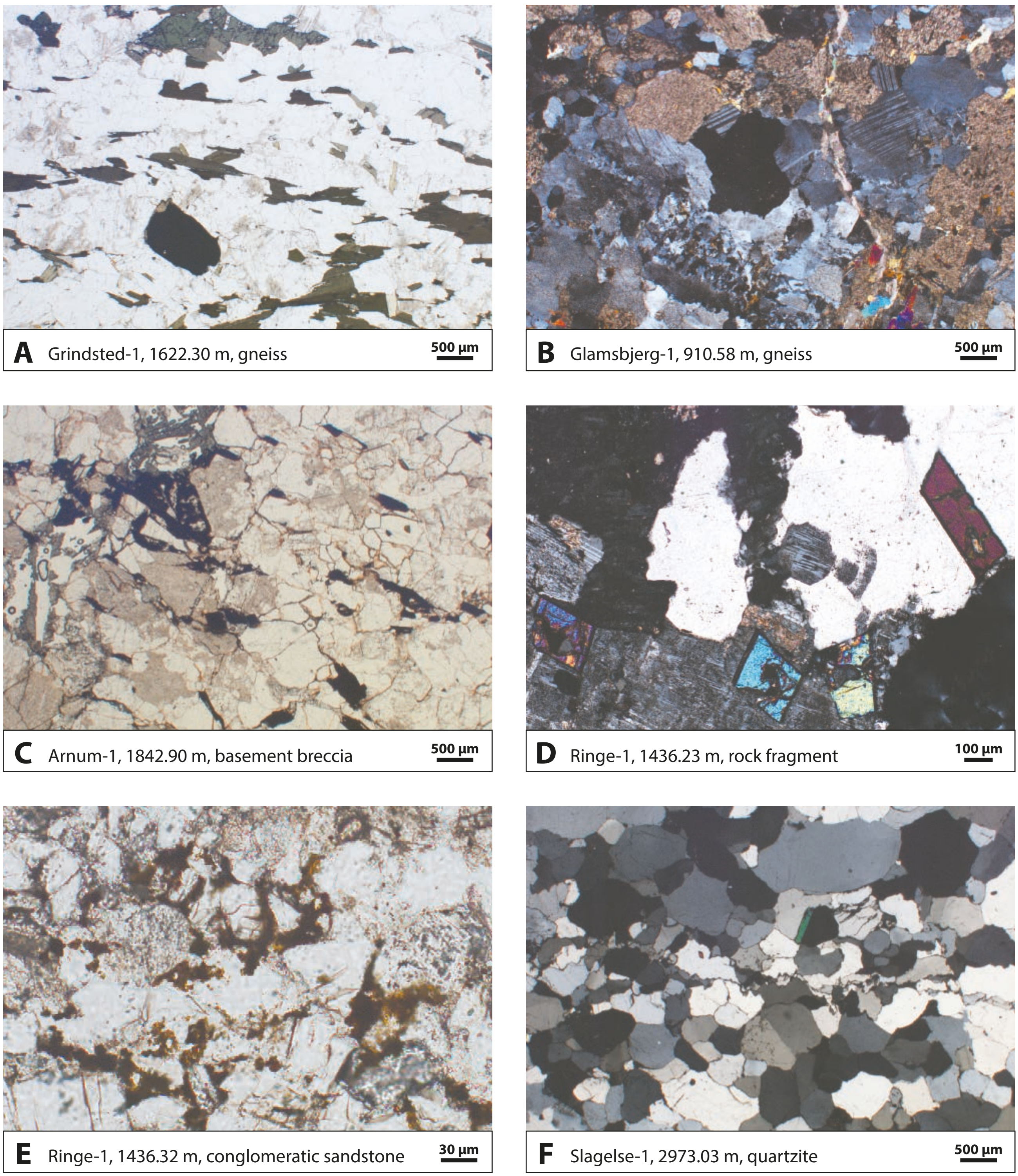

Fig. 4. Thin section photographs with plane polarised light in A, C and E and crossed polars in B, D and F. Note zircon crystals in D.

iferous–Permian Oslo Graben formed at 0.30–0.26 Ga (Bingen & Solli 2009). In southern Sweden, the Eastern Segment with zircon ages of 1.80–1.64 Ga (Bingen & Solli 2009) forms the eastern part of the Sveconorwegian Orogen (Fig. 1). South-eastern Sweden is part of the Svecofennian Orogen with zircon ages of 1.92–1.79 Ga (Lahtinen *et al.* 2008).

Samples and methods

Zircon age dating, heavy mineral studies and geochemical analyses were performed on six samples from the Ringkøbing–Fyn High, comprising two from the basement, three from redeposited basement and one from a quartzite (Fig. 2). Proterozoic gneisses (Noe-Nygaard 1963) were sampled in the Grindsted-1 and Glamsbjerg-1 wells which are the only wells that have reached the basement in the Ringkøbing–Fyn High. Crystalline basement relics were sampled in breccia from the Arnum-1 well and in conglomeratic sandstone from the Ringe-1 well in order to estimate the regional variation of the basement lithologies. A quartzite was sampled from metasediments in the Slagelse-1 well. The Arnum-1 and Slagelse-1 wells are located in the North German Basin and Norwegian–Danish Basin, respectively, on the flanks of the Ringkøbing–Fyn High. About 10 cm long core pieces were cut lengthwise into two samples, one small and one large. The small sample was used for geochemical analysis and the large sample was crushed and split into two samples used for respectively zircon age dating and heavy mineral analysis.

Zircon U–Pb geochronometry

Samples were crushed and zircon grains were handpicked after sorting on a Wilfley shaking table. The U–Pb data were obtained by laser-ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) at GEUS (Frei & Gerdes 2009). The specific analytical procedure is described in Supplementary data file 1 and the full analytical details are reported in Supplementary data file 2. In-house GEUS software was used for the data reduction. The full table of the U–Pb data are reported in Supplementary data file 3, and the individual spots in the zircon grains are documented in Supplementary data file 4.

A total of 552 zircon ages are provided for the six samples. The vast majority of data points are highly concordant; only 32 analyses are more than 10% discordant (Supplementary data file 3). The following criteria were used to identify and remove data of poor quality from the final dataset: (1) analyses with large analytical errors ($^{238}U/^{206}Pb$ >10%) were removed from the dataset; (2) analyses with evidence of containing a significant component of common Pb, reflected in significant ^{204}Pb count rates or exceedingly low $^{206}Pb/^{204}Pb$ ratios (<<1000), were removed. In the filtered dataset some samples remain with fairly low $^{206}Pb/^{204}Pb$ ratios (<1000); however these were assessed not to be affected by common Pb as the corrected and uncorrected $^{207}Pb/^{206}Pb$ ages did not vary significantly (outside error). Finally, for calculating the $^{207}Pb/^{206}Pb$ ages only highly concordant analyses with a discordance of <5% were used.

Heavy mineral analyses

Heavy mineral concentrates were produced from the grain-size interval 45–710 μm by heavy liquid separation of grains with a density >2.8 g/cm^3. The element composition of 1200 heavy minerals was measured by computer-controlled scanning electron microscopy (CCSEM) at GEUS (Keulen *et al.* 2008, 2012). The grains were classified as various minerals using in-house GEUS software.

Bulk geochemical analyses

Whole rock geochemical analyses were carried out on 10 g crushed samples by Acme Labs, Canada. Major oxides and minor elements were determined by inductively coupled plasma emission spectrometry (ICP-ES) and trace elements by inductively coupled plasma mass spectrometry (ICP-MS). Rare earth elements (REE) were included in the measurements and comprise the light REE (LREE, La–Sm) and the heavy REE (HREE, Eu–Lu). Loss on ignition (LOI) was measured after heating to 1000°C.

Results

The rocks are characterized with respect to their macroscopic and microscopic appearance (Figs 3, 4), their heavy mineral assemblage (Figs 5, 6, 7) and their zircon U–Pb age distribution patterns (Figs 8, 9). All zircon U–Pb data are reported in Supplementary data file 3 and are presented in concordia diagrams (Fig. 8). Only zircon grains with a discordance of <5% are included in the U–Pb age distribution diagrams (Fig. 9). The zircon grains are generally rounded to well-rounded (Supplementary data file 4) and there does not seem to be a correlation between the age of the grains and their grain size and shape. The only exception is found in the Ringe-1 well, as described below. Biotite, muscovite and the mica-like mineral chlorite are collectively referred to as mica minerals in the following. They are reported separately from the non-mica heavy mineral assemblage (Fig. 5).

Grindsted-1: crystalline basement

The crystalline rocks in the Grindsted-1 well were sampled at 1622.30 m depth (Fig. 2). Triassic sediments overlie the crystalline rocks at 1602 m depth. The rock was described by Noe-Nygaard (1963) and Larsen (1971) as medium-grained biotite gneiss dominated by quartz, plagioclase and biotite. The gneiss does not vary much in the two cored sections of *c*. 1 m length each, but the amount of banding varies. Horizontal light pinkish grey and dark grey layers are visible in the sampled section (Fig. 3A), where the light layers are rich in quartz and plagioclase whereas the dark layers contain abundant biotite and hornblende. Preferred orientation of the mafic minerals is evident in thin section (Fig. 4A) (Aghabawa 1993). Pegmatitic veins cut the gneiss in several places.

Mica grains, mainly biotite, make up 84 wt% of the total heavy mineral content (Fig. 5). The non-mica heavy mineral assemblage is dominated by 89 wt% amphibole which is primarily hornblende (Fig. 6), as Ca-amphibole lies between Ca-rich and Ca-poor pyroxene compositions in a CaO–MgO–FeO diagram according to Wilson & Larsen (1985). K-feldspar occurs as clusters of very small inclusions in plagioclase. The rock has a felsic composition with 67 wt% SiO_2 (Table 1). The rock contains the whole range of Mesoproterozoic zircon ages, of which the highest concentration is in the interval 1.51–1.44 Ga (Fig. 8A, Fig. 9).

Glamsbjerg-1: crystalline basement

The crystalline rocks in the Glamsbjerg-1 well were sampled at 910.58 m depth (Fig. 2) and are immediately overlain by a Triassic succession at 907 m depth. Noe-Nygaard (1963) described the rock as a medium-grained granodioritic hornblende gneiss, and it was suggested that the intervals dominated by quartz and feldspar should be named granodioritic gneiss, and those with much hornblende should be called amphibolitic gneiss. The gneiss varies in composition within the *c*. 2 m-long cored section (Noe-Nygaard 1963; Larsen 1971). Oblique banding of light pinkish grey and dark greenish grey layers is visible in the sampled section (Fig. 3B). In thin section anhydrite cement is observed to fill the thin fractures (Fig. 4B) which are also visible in hand specimen (Fig. 3B). The sample used in this study has a high content of calcite that apparently has replaced some K-feldspar.

Amphibole is present as hornblende (Fig. 6) and constitutes 94 wt% of the non-mica heavy mineral assemblage (Fig. 5). Mica grains make up 11 wt% of the total heavy mineral content and are mainly biotite. The rock has a low SiO_2 content of 49 wt% and a high CaO content of 12 wt% (Table 1). Two age populations are present in the zircon age distribution (Fig. 9). The oldest has age peaks in the interval 1.55–1.48 Ga and is dominated by a large peak at 1.51 Ga, whereas the

Table 1. Bulk geochemistry of Precambrian basement and Palaeozoic sediments in the Ringkøbing–Fyn High

Well	**Grindsted-1**	**Glamsbjerg-1**	**Arnum-1**	**Arnum-1**	**Ringe-1**	**Slagelse-1**
Depth, m	1622.30	910.58	1842.60	1842.90	1436.32	2973.03
Major elements, wt%						
SiO_2	67.3	48.6	46.6	48.5	65.5	97.3
Al_2O_3	14.6	15.5	14.4	16.2	15.0	0.5
Fe_2O_3	5.3	7.4	10.5	10.5	4.5	1.4
MgO	2.0	4.7	8.5	8.3	0.1	0.1
CaO	2.9	12.1	4.9	2.7	1.4	0.1
Na_2O	4.2	3.1	1.6	2.3	3.3	0.0
K_2O	2.3	2.1	3.3	3.4	5.9	0.1
TiO_2	0.5	1.0	0.7	0.8	0.6	0.0
P_2O_5	0.1	0.1	0.1	0.2	0.1	0.0
MnO	0.1	0.3	0.1	0.1	0.1	0.0
LOI	0.6	4.9	9.2	6.8	3.2	0.4
Total	99.9	99.8	99.8	99.8	99.6	100.0
Trace elements, ppm						
Zr	245	68.0	67.1	87.5	1348	75.5
La	22.5	14.0	9.9	14.3	122	3.4
Lu	0.3	0.3	0.3	0.3	0.9	0.1

Total iron is given as Fe_2O_3. LOI: Loss on ignition.

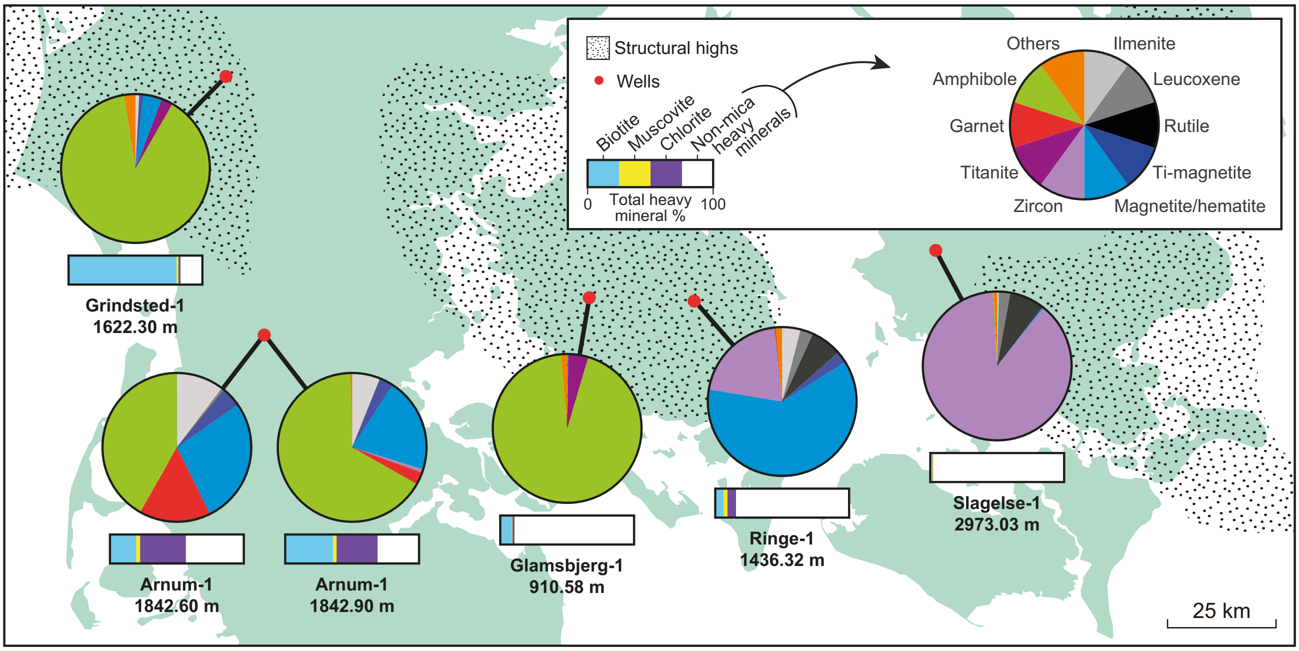

Fig. 5. Heavy mineral assemblages of the Precambrian basement and Paleozoic sediments in the Ringkøbing–Fyn High. The total heavy mineral content in the bar charts includes the mica minerals (biotite, muscovite and chlorite) whereas the non-mica heavy mineral assemblage is reported separately in the pie charts. The CCSEM method cannot distinguish magnetite and hematite. The "others" group includes epidote, apatite, corundum, monazite and chromite in decreasing abundance.

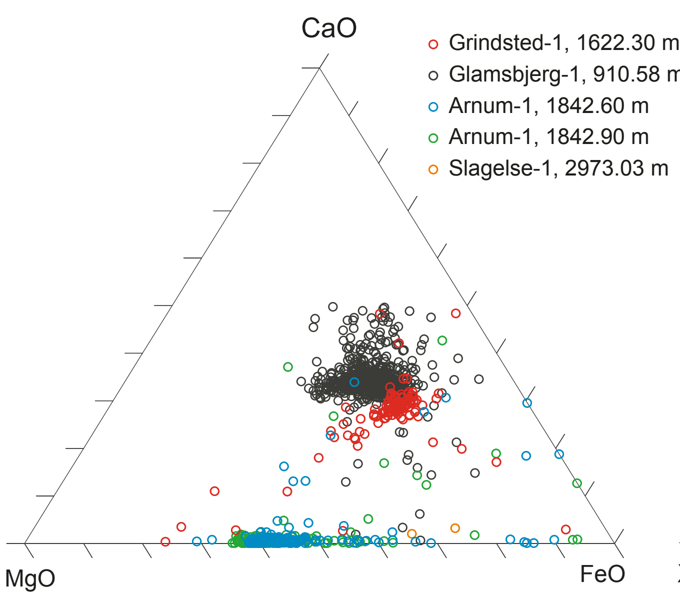

Fig. 6. CaO–MgO–FeO (total iron) composition of the amphibole grains. Ca-amphibole lies between the compositions of clinopyroxene and orthopyroxene in this diagram, and Ca-poor amphibole has the same composition as Ca-poor pyroxene (Wilson & Larsen 1985). The Ca-amphibole in the Grindsted-1 and Glamsbjerg-1 samples is hornblende whereas the Ca-poor amphibole in the Arnum-1 samples is presumably gedrite as it contains Al.

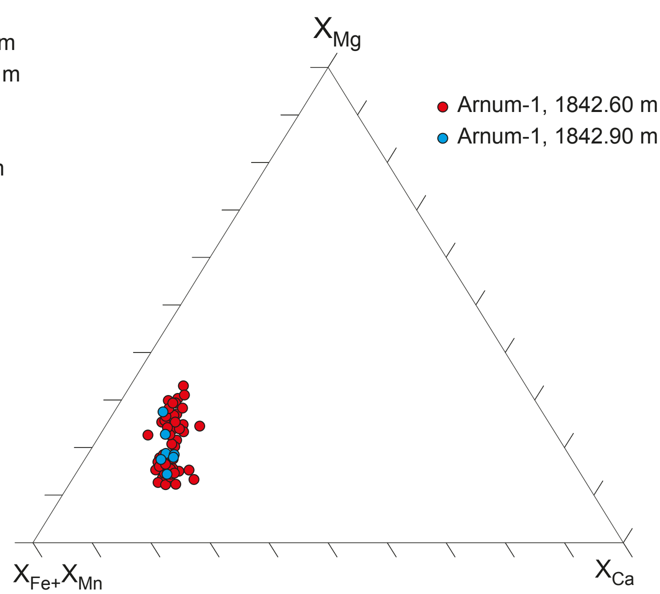

Fig. 7. Garnet composition of the basement breccia in the Arnum-1 well. The end-members are calculated as element wt% of Mg, Fe+Mn and Ca. The garnet grains also contain Al and are therefore almandine-rich.

younger age peaks in the interval 1.14–1.08 Ga are less prominent. The two age populations are also visible in the concordia diagram (Fig. 8B).

Arnum-1: basement breccia

The brecciated rocks in the Arnum-1 well were sampled at 1842.60 and 1842.90 m depth (Fig. 2). Zechstein deposits overlie the breccia at 1814 m depth. The rock was described by Noe-Nygaard (1963) as a fanglomerate and the rock fragments were identified as biotite-hornblende gneiss, hornblende-biotite gneiss, garnet-hornblende-biotite gneiss and fine-grained quartz-microcline rock. The rock is a very poorly sorted breccia with angular basement blocks up to 10 cm in diameter. The composition of these rock fragments varies through the 30 m-long cored section and they become less abundant upwards. The breccia appears as dark reddish grey to dark greenish grey gneiss in some intervals (Fig. 3C) and as angular basement fragments embedded in a dark red conglomeratic sandstone matrix in other intervals (Fig. 3D). The sandy matrix in the deepest of the studied samples consists of rock fragments, quartz and feldspar, as well as completely chloritized and locally illitized clasts which presumably represent strongly altered biotite and hornblende (Fig. 4C). The thin section contains two centimetre-large fragments; one is hornblende-biotite gneiss and the other is granitic gneiss. In the hornblende-biotite gneiss the hornblende is pervasively altered to chlorite and illite in a similar way to the individual clasts in the matrix.

The non-mica heavy mineral assemblage consists of a range of minerals with a dominance of 42–66 wt% amphibole, 21–27 wt% magnetite/hematite, 3–16 wt% garnet and 6–10 wt% ilmenite (Fig. 5). The amphibole grains are Ca-poor (Fig. 6) and according to their chemistry they are most likely gedrite as they contain Al. Mica constitutes 57–69 wt% of the total heavy mineral content and both biotite and chlorite are abundant. The chlorite is atypical with much Fe, Mg, K and Si, so most of it is presumably biotite that has been partially altered into chlorite and hematite, but some of the chlorite could also be altered hornblende. The garnets have a quite uniform composition with 56–67 wt% FeO, 1–10 wt% MnO, 12–33 wt% MgO and 8–21 wt% CaO (Fig. 7), and they are almandine-rich since they also contain Al. The rock has a low SiO_2 content of 47–48 wt% and a very high LOI of 7–9 wt% (Table 1). The zircon population is dominated by 1.54–1.53 Ga grains with few grains of Archean, Palaeoproterozoic and Sveconorwegian age (Fig. 9).

Ringe-1: conglomeratic sandstone

The conglomeratic sandstone in the Ringe-1 well was sampled at 1436.32 m depth (Fig. 2). The conglomerate is overlain by Triassic deposits at 1272 m depth. The conglomeratic succession was proposed by Sorgenfrei & Buch (1964) to be of early Permian or Eocambrian age, and Nielsen & Japsen (1991) suggested that it could be of Permian age. Cores from the pre-Triassic rocks constitute a total of 25 m collected from 11 intervals. Greenish grey reduction spots are present in the upper part of the pre-Triassic succession. The rock fragments in the conglomerate consist mainly of quartz and feldspar. They are numerous in the lower part of the pre-Triassic succession and become smaller and fewer up through the succession, and clasts consisting of only quartz or feldspar are primarily found towards the top. Feldspar is the dominant light mineral in the lower part of the succession but the content decreases upwards whereas the quartz content increases and the grains become more rounded. Light pinkish grey subangular clasts are abundant in the sampled rock (Fig. 3E). They consist of metamorphosed alkali feldspar granite and contain angular heavy minerals (Fig. 4D). The clasts are dispersed in a dark brownish red matrix which contains Fe-oxides/hydroxides (Fig. 4E). A thin vein between a large granite clast and its surrounding matrix is filled by hematite, quartz, calcite, kaolinite, barite and locally monazite. The clasts dominate in both the gravel and the coarse sand fraction.

The non-mica heavy mineral assemblage is dominated by 61 wt% magnetite/hematite and 21 wt% zircon (Fig. 5). Mica grains make up 19 wt% of the total heavy mineral content, and SiO_2 constitutes 65 wt% of the rock (Table 1). The clasts in the sampled rock have a high content of Zr and LREE. The zircon population is mainly dominated by ages in the interval 1.66–1.51 Ga and a younger group culminating at 0.76 Ga (Fig. 8E, Fig. 9). The youngest zircon grains are larger and clearly more angular than the late Palaeoproterozoic to early Mesoproterozoic grains (Fig. 10), and also contain numerous mineral inclusions.

Slagelse-1: quartzite

The quartzite in the Slagelse-1 well was sampled at 2973.03 m depth (Fig. 2). Poulsen (1969, 1974) proposed that the quartzite could be of early Cambrian age as fossils of early Cambrian to early Silurian age were found in the thick overlying shale succession. Alternating layers of quartzitic sandstone and shale present at 2960–2972 m depth separate the quartzite from the shale succession (Fig. 2). The quartzite is very uniform in the *c.* 1 m-thick cored section but is interrupted by a shale layer. The quartzite is light grey with oblique

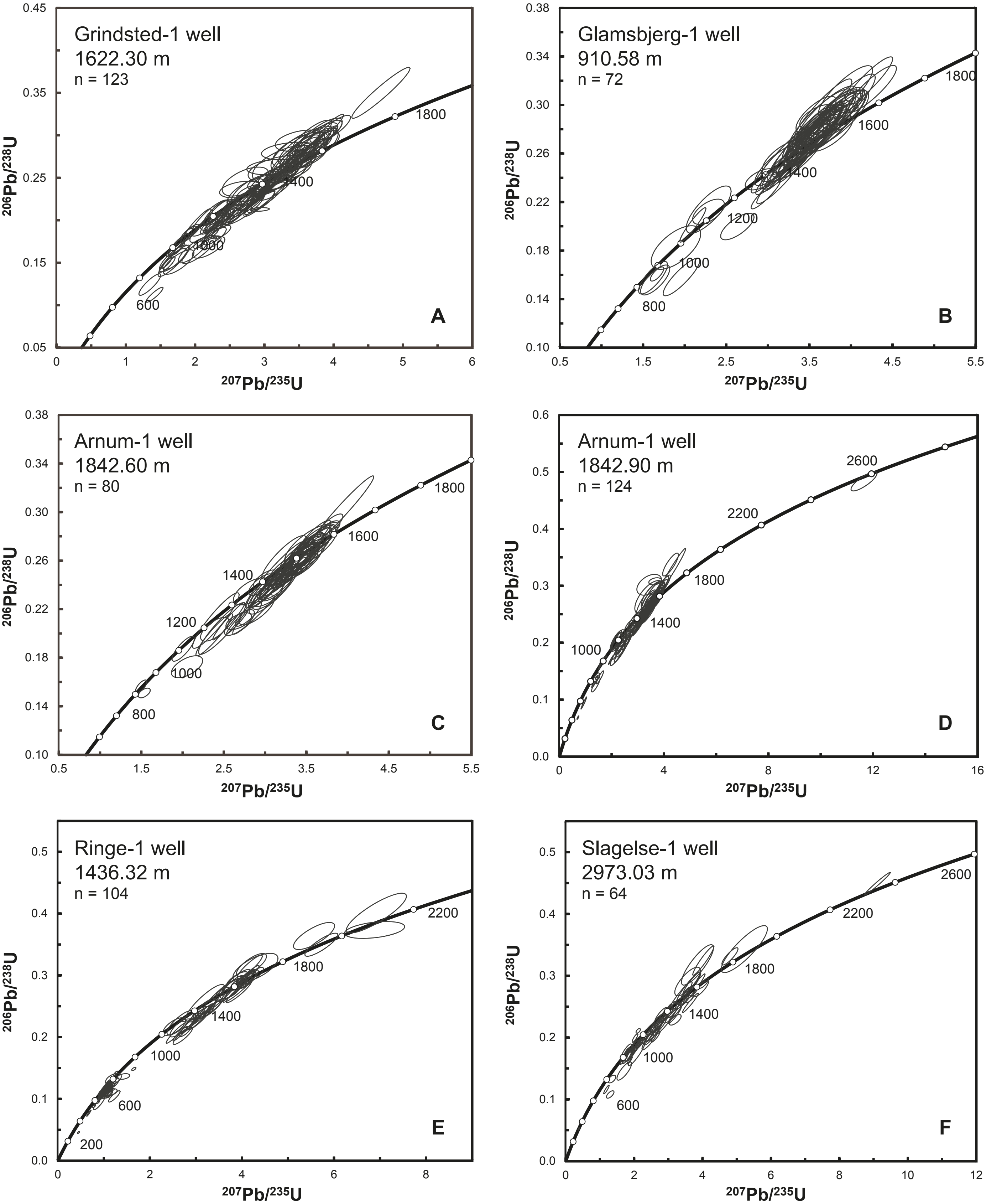

Fig. 8. Concordia diagrams displaying all the U–Pb data. For more details see Supplementary data file 1.

grey banding (Fig. 3F), and inherited stylolites from the sandstone precursor are seen in thin section (Fig. 4F).

Zircon constitutes 88 wt% of the heavy mineral assemblage in the sampled rock (Fig. 5) together with 8 wt% rutile and 3 wt% leucoxene plus a few other grains. Mica grains constitute 2 wt% of the total heavy mineral content, and the rock contains 97 wt% SiO_2 (Table 1). The rock has a quite homogeneous dispersal of ages within the interval 1.72–1.00 Ga (Fig. 8F, Fig. 9).

Discussion

Proterozoic basement

The zircon age distribution patterns from gneisses in the Ringkøbing–Fyn High (Fig. 9) are consistent with formation during the orogenic events that affected south-western Sweden and southern Norway in the early Mesoproterozoic. Thus, the Ringkøbing–Fyn High can be considered a southerly extension of the

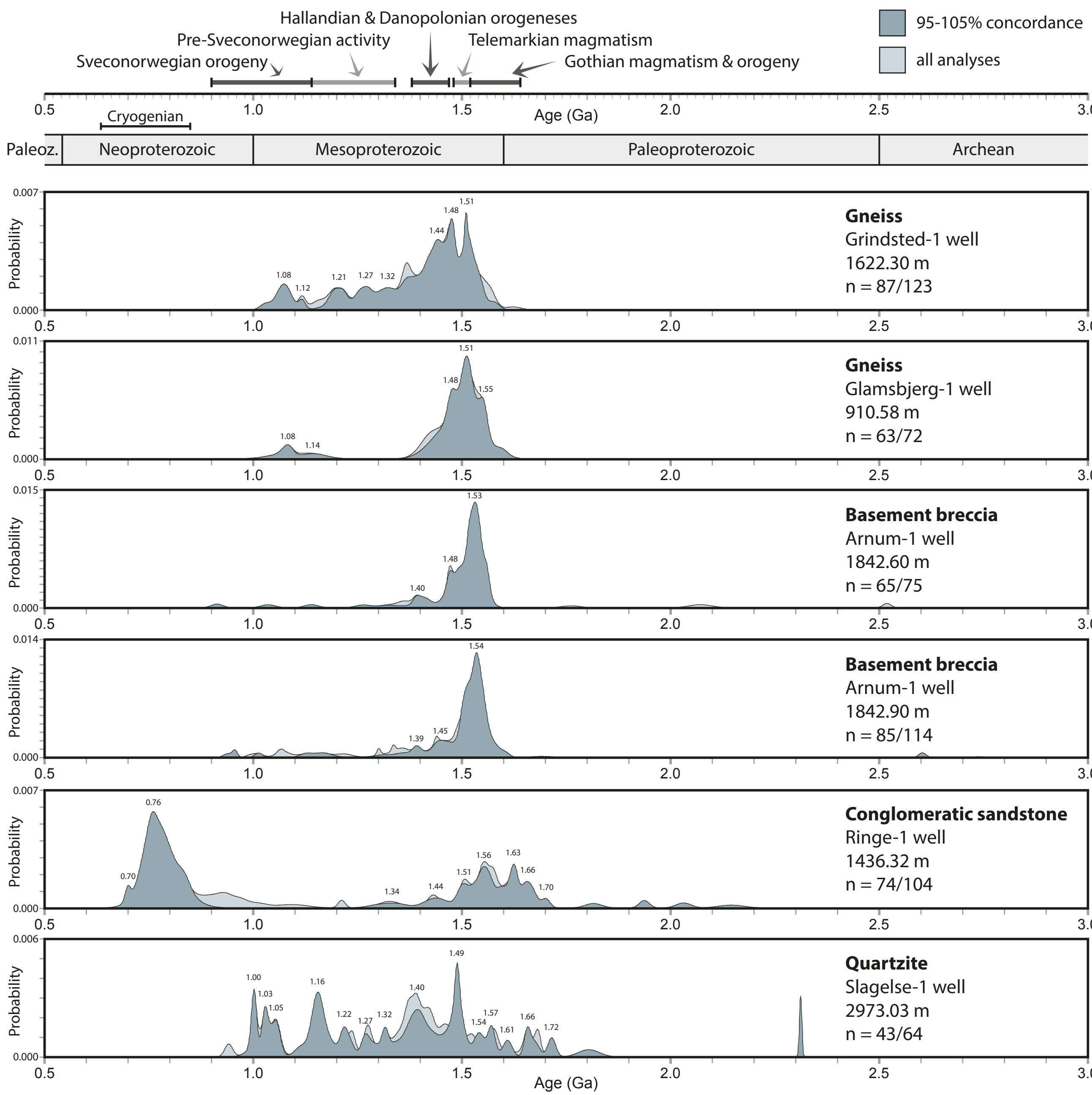

Fig. 9. Zircon U–Pb age distributions of the Precambrian basement and Paleozoic sediments in the Ringkøbing–Fyn High. The age peaks are indicated in billion years. n is the number of concordant grains relative to the total number of analyzed grains.

Fennoscandian Shield. The dominant age intervals in the Glamsbjerg and Grindsted Highs indicate that rocks of the Gothian orogeny may only be present in the Glamsbjerg High whereas the Telemarkian accretion and Sveconorwegian metamorphism affected both areas. This is consistent with the general westward age progression trend in the Sveconorwegian Orogen, so that the buried basement in central Denmark may be youngest towards the west.

Pre-Sveconorwegian magmatic activity (Brewer *et al.* 2004; Bingen *et al.* 2008a) must be included in the interpretation to account for the entire age range of the gneiss in the Grindsted-1 well (Fig. 9). The ages of the gneiss in the Glamsbjerg-1 well form two well-defined zircon populations; the oldest denotes the age of formation and the youngest represents the metamorphic overprint, as it was evident in measurements of both core and rim. The roundness of the zircon grains in the basement was probably caused by metamorphic modification (Corfu *et al.* 2003) or perhaps by earlier sedimentary processes.

K/Ar age dating of biotite from the Grindsted-1 gneiss and hornblende from the Glamsbjerg-1 gneiss were performed by Larsen (1971) and gave ages of 0.87 ± 0.2 Ga and 0.82 ± 0.2 Ga, respectively. The Sveconorwegian zircon ages in these gneisses (Fig. 9) and the late Sveconorwegian zircon ages in south-western Fennoscandia (Söderlund *et al.* 2002) are older than this, probably because zircon can retain Pb at higher temperatures.

The heavy mineral assemblages in the Grindsted-1 and Glamsbjerg-1 gneisses (Fig. 5) correspond very well to those described by Noe-Nygaard (1963). The Fe-rich compositions of the amphibole grains (Fig. 6)

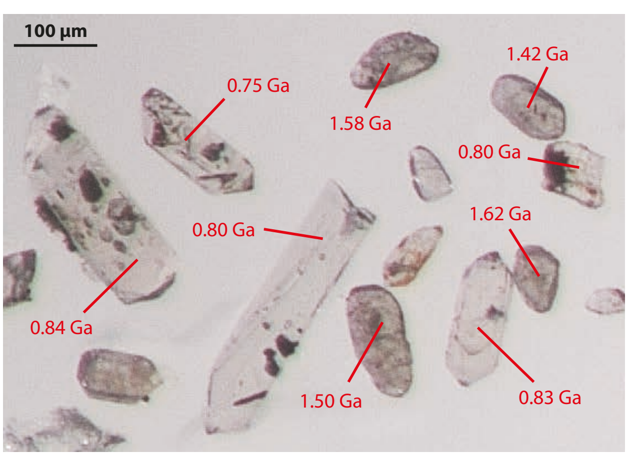

Fig. 10. Photograph of zircon grains from the conglomeratic sandstone sampled in the Ringe-1 well at 1436.32 m depth. The grains of Cryogenian age (age peak culminating at 0.76 Ga) contain mineral inclusions and are generally angular whereas the older grains (major age peaks in the interval 1.66–1.51 Ga) are rounded. The uncertainties are in the order of 0.01–0.02 Ga.

are typical for granitic rocks which were the probable precursors for the gneisses. The high SiO_2 content in the Grindsted-1 gneiss probably reflects the original felsic composition (Table 1), whereas the SiO_2 content in the Glamsbjerg-1 gneiss is diluted by a high CaO content reflecting anhydrite and calcite precipitations.

Redeposited basement

The two zircon age distributions from the basement breccia in the Arnum-1 well are very similar to each other (Fig. 9), despite the large variation in embedded rock fragments and the 30 cm-spacing between the samples. The rock fragments were presumably derived from the same source (1.54–1.53 Ga), even though the relative proportions of heavy minerals vary (Fig. 5). Noe-Nygaard (1963) described this wide range of metamorphic rocks but noted that they grade into each other and therefore must originate from the same gneiss complex. The Fe-rich garnet composition in both samples (Fig. 7) is also in accordance with Noe-Nygaard (1963) who mentioned almandine as the only garnet in the rock fragments. Almandine is typical for clay-rich rocks metamorphosed into garnetiferous gneiss. The large content of magnetite/hematite in the samples (Fig. 5) may represent hematite precipitations in the matrix and hematite formed from biotite degradation. The high LOI (Table 1) may reflect the presence of chlorite and illite in the matrix and the partial replacement of biotite and hornblende in the gneissic clasts. The rock became fragile as a result of the replacement process so the replacement probably took place after deposition of the breccia.

The Arnum-1 well does not penetrate the basement. However, the rock fragments in the Arnum-1 breccia were very likely derived from the Ringkøbing–Fyn High basement, as the large size of the rock fragments indicates that they cannot have been transported very far (Fig. 3D). These clasts can therefore be assumed to represent the local basement of the Ringkøbing–Fyn High, and the age peak at 1.54–1.53 Ga probably pinpoints the age when this basement was formed. The zircon ages of the breccia are quite similar to those of the Glamsbjerg-1 gneiss (Fig. 9) which could have been outcropping over a large area, at least on the Glamsbjerg High. A biotite age of 0.69 ± 0.02 Ga from a basement fragment in the Arnum-1 breccia was rejected because of potassium loss (Larsen 1971), but a new K/Ar age of a less weathered sample from the deepest level of the Arnum-1 well was reported by Larsen (1972) as 0.86 ± 0.02 Ga which is in accordance with the K/Ar ages from the gneisses in the Grindsted-1 and Glamsberg-1 wells (Larsen 1971).

Noe-Nygaard (1963) suggested that the breccia was deposited by an alluvial fan that developed along the margin of the Ringkøbing–Fyn High. The measured heavy mineral assemblage of the Glamsbjerg-1 gneiss is not identical to the assemblage in the Arnum-1 breccia (Figs 3, 5), but this may be explained by the alteration of some of the dark minerals into chlorite and hematite in the breccia. Similar alteration has taken place in the uppermost weathered part of the gneiss (Noe-Nygaard 1963). In addition, the breccia contains gedrite, a metamorphic mineral that mostly occurs in Ca-poor ultramafic amphibolite facies rocks.

The zircon grains with younger ages than the major peak in the Arnum-1 breccia may represent metamorphic overprints, but no age cluster is apparent (Fig. 9). Alternatively, they could represent detrital zircons from the sandstone matrix and may have been transported a considerable distance from a provenance outside the Ringkøbing–Fyn High. The Sveconorwegian metamorphic influence in the area probably varied considerably, as was generally the case in south-western Fennoscandia (Bingen *et al.* 2008b), which, combined with the presumed minor isotopic resetting during metamorphism, may explain the rarity of zircons with this age in the breccia. The Lower Triassic Röt Formation was deposited directly on top of the Glamsbjerg-1 gneiss (Fig. 2) and erosion must have occurred before this time. The time of deposition of the basement breccia in the Arnum area is unknown, but it must have taken place before deposition of the overlying upper Permian Zechstein sediments (Fig. 2).

The youngest zircon age population in the conglomeratic sandstone in the Ringe-1 well probably occurs in the metamorphosed alkali feldspar granite clasts, as the youngest zircons are more euhedral than the rest and contain mineral inclusions (Fig. 10) like the zircons observed in the clasts (Fig. 4D). These zircon grains of Cryogenian age with a prominent age peak at 0.76 Ga (Fig. 9) are very characteristic with their large size and angular shape. Similar zircon ages are absent in Sweden and Norway (Olivarius *et al.* 2014) as Cryogenian magmatism in the region was very limited (Bingen & Solli 2009). The size and shape of the zircon grains and of their host clasts (Fig. 3E) indicate that they have a local source. The granite clasts are likely derived from Avalonia since Neoproterozoic ages are more common in Avalonia where a phase of arc activity began at 0.76 Ga (Murphy *et al.* 2000). The Ringkøbing–Fyn High is located at the southern margin of Baltica with Avalonia positioned immediately south of the area (Pharaoh 1999).

The oldest zircon age population in the Ringe-1 conglomeratic sandstone is probably present in the sandy matrix, as the oldest zircons are smaller and less angular than the rest (Fig. 10), like the zircons observed in the matrix. The major age peaks at 1.66–1.51 Ga partly overlap with the formation ages of the basement rocks recorded in the Ringkøbing–Fyn High (Fig. 9). However, the matrix may have been sourced from the Gothian basement in south-western Sweden where the entire age span is present, and the grain size also points to more than just a local transport distance. The few older zircon grains probably also represent a distant provenance and indicate that the sediment was deposited at a time when sediment transport from Fennoscandia across the Ringkøbing–Fyn High area was possible. This could have been during the Proterozoic or at the beginning of the late Carboniferous to early Permian uplift and erosion of the Ringkøbing–Fyn High. The high contents of mechanically stable zircon and rutile (Fig. 5) confirm that some components in the sediment had probably been transported a considerable distance. The hematite-quartz-calcite-kaolinite-barite-monazite vein in the basal part of the succession may indicate hydrothermal activity after deposition of the conglomerate but does not point to any specific age. It could have been related to late Carboniferous to early Permian igneous activity recorded from the area (Thybo 1997). A Permian depositional age is also possible, as suggested by earlier studies by Sorgenfrei & Buch (1964) and Nielsen & Japsen (1991).

In conclusion, the rock fragments in the Arnum-1 breccia and the Ringe-1 conglomeratic sandstone represent provenance areas that may have been strictly local and related to large faults developed during uplift of the Ringkøbing–Fyn High. The sediments in which the rock fragments reside might have been eroded and thereby influenced the composition of younger sediments deposited around the high. As the Arnum-1 breccia and the Ringe-1 conglomerate consist mainly of local basement reworked due to weathering or faulting they must have transmitted a strong age signal from local rocks to younger sediments. This signal is accompanied by the matrix ages which are of mixed and possible Fennoscandian origin.

Metasediment

The zircon age distribution from the quartzite in the Slagelse-1 well (Fig. 9) constitutes a mixture of all the ages present in the basement areas of south-western Sweden and southern Norway (Bingen *et al.* 2008b). The heavy mineral assemblage and geochemical signature are extremely mature (Fig. 5, Table 1) and the sedimentary precursor is thus likely to have been reworked many times before its final deposition. Poulsen (1969) suggested that the quartzite was formed from weathered gneiss in the Ringkøbing–Fyn High, but the much narrower age distributions found in the gneisses (Fig. 9) indicate that this was probably not the case. The Fennoscandian Shield is thus presumed to be the initial provenance area of the quartzite.

The Slagelse-1 quartzite is supposed to be of early Cambrian age and correlates with lower Cambrian sandstones in Sweden and on Bornholm (Poulsen 1969). The broad spectrum of zircon ages and the relative abundance of older zircon grains may support that the provenance of the metasediment was Fennoscandia and not a local source. This could point to a relationship to a basin-wide sandstone complex, such as the lower Cambrian Hardeberga Formation. Erosion of the Slagelse-1 quartzite may have occurred, thereby adding a homogenized Fennoscandian signature to some of the younger sediments.

Conclusions

The Proterozoic basement in the Ringkøbing–Fyn High and that in south-western Fennoscandia were formed during the same orogenic events and exhibit matching metamorphic overprints as recorded by zircon U–Pb ages. The age spans in the Ringkøbing–Fyn High are mainly restricted to the 1.55–1.44 Ga interval, culminating at 1.51 Ga. Consequently, zircon U–Pb ages may be used to appoint or exclude the basement high as a provenance area for sediments in the Norwegian–Danish Basin and the North German Basin. Basement that has been redeposited from the Ringkøbing–Fyn High due to weathering or faulting, and Palaeozoic sediments that are locally present on the Ringkøbing–Fyn High, show various mixtures of local and Fennoscandian provenance signatures. The conglomeratic sandstone from the Ringe-1 well contains granitic rock fragments with Cryogenian zircon ages representing an igneous event not previously recorded in the area.

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The first evidence of trace fossils and pseudo-fossils in the continental interlava volcaniclastic sediments on the Faroe Islands

RICHARD POKORNÝ, LUKÁŠ KRMÍČEK & UNI E. ÁRTING

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Interlava volcaniclastic sediments, mostly sandstones, from the Palaeogene Faroe Islands Basalt Group (Malinstindur and Sneis Formations) contain rare ichnofauna and well-preserved pseudo-fossils in the form of linear structures. Five specimens of two ichnogenera have been identified, which include *Helminthoidichnites* isp. and ?*Palaeophycus* isp. The linear structures are interpreted to be desiccation cracks. This association indicates an environment with low to moderate hydrodynamic energy, which confirms a mosaic landscape of floodplains with rivers and shallow lakes.

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Richard Pokorný [richard.pokorny@ujep.cz], Jan Evangelista Purkyně University, Faculty of Environment, Králova Výšina 3132/7, 40096 Ústí nad Labem, Czech Republic. Lukáš Krmíček [l.krmicek@gmail.com], Brno University of Technology, Faculty of Civil Engineering, AdMaS Centre, Veveří 95, CZ-602 00 Brno, Czech Republic (also Institute of Geology, v.v.i., Czech Academy of Sciences, Rozvojová 269, CZ-165 02 Prague 6, Czech Republic). Uni E. Árting [Uni.Arting@jardfeingi.fo], Jarðfeingi, Brekkutún 1, Postsmoga 3059, FO 110 Tórshavn, Faroe Islands (also University of the Faroe Islands, J.C. Svabos gøta 14, FO 110 Tórshavn, Faroe Islands).

The Faroe Islands are a 1400 km^2 remnant archipelago of the extensive Palaeogene subaerial volcanic succession that covers large parts of the eastern margin of the North Atlantic. Beds of volcaniclastic rocks, generally sandstones and conglomerates, are often intercalated with the lava flows in the succession, and Palaeogene fossil material on the Faroe Islands occurs almost exclusively in these volcaniclastic beds. The fossil material is not rich, and trace fossils have not been recorded before. This study describes and interprets some structures, believed to be trace fossils and pseudo-fossils, that were recently discovered at two localities, Eiði and Hundsarabotnur, during a survey of intra-volcanic sedimentary rocks at 32 localities across the Faroe Islands (Fig. 1).

Geological setting

The term Faroe Islands Basalt Group (FIBG) covers the basaltic lava succession onshore as well as the offshore continuation on the Faroe Platform, the Faroe–Shetland Channel, and the banks south of the Faroe Platform (Jolley & Bell 2002). The onshore parts of the FIBG are presumed to rest on a continental crust consisting of Precambrian metamorphic rocks (Casten 1973; Bott *et al*. 1974; Gariépy *et al*. 1983; Holm *et al*. 2001). Passey & Jolley (2009) described the FIBG as comprising both the onshore parts and the offshore continuation of the lava succession to the east and southeast into the Faroe–Shetland Basin, covering an estimated area of *c*. 120 000 km^2. Their work was based on the study of both borehole material and subaerial exposures. Passey & Jolley (2009) also re-defined and renamed the formations originally described by Rasmussen & Noe-Nygaard (1969). Seven lithostratigraphic formations are now formalized with a total thickness of 6.6 km. The formations, in stratigraphic order, are as follows: Lopra Formation, Beinisvørð Formation, Prestfjall Formation, Hvannhagi Formation, Malinstindur Formation, Sneis Formation and Enni Formation (Figs 1, 2). Noe-Nygaard & Rasmussen (1968) distinguished three major petrological types of subaerial lava flows, and subordinate interlava units within the Faroe Islands. The lava flow types are aphyric basalts that lack phenocrysts and typically have a fine-grained groundmass; plagioclase-phyric basalts that also typically have a fine-grained groundmass in which the phenocrysts sometimes form stellate clusters; and olivine-phyric basalts with small, typically stout, olivine phenocrysts. Radiometric data from Waagstein *et al.* (2002) and Storey *et al.* (2007) provide an age range of 60.5 Ma to 55.2 Ma. These data are however in disagreement with the biostratigraphic ages from the FIBG (Jolley 2009). For further discussion, see Passey & Jolley (2009).

Beds of volcaniclastic rocks, generally sandstones and conglomerates, are often intercalated with the lava flows of the FIBG. Mudstone and coal also occur here, but in smaller amounts (Ellis *et al*. 2002; Passey 2004; Passey & Jolley 2009). Frequent intervals of fine-grained clayey sandstone are recorded in the lower and upper part of the Lopra Formation in the Lopra-1 borehole on Suðuroy, where up to several tens of metres in thickness of such rocks were recorded (Ellis *et al*. 2002; Boldreel 2006). Thin beds up to 2 m thick of laminated sandstone or conglomerate are present mostly in the lower and middle part of the Beinisvørð Formation (Hald & Waagstein 1984). In the upper 200 m of this formation, there are abundant layers of mudstone that are sporadically accompanied by coal seams and sandstone (Rasmussen & Noe-Nygaard 1970; Passey 2004; Passey & Bell 2007; Passey 2008).

The Prestfjall Formation is a 3–15 m thick sedimentary succession comprising coal-bearing layers along with clay, shale, and occasionally sandstone and conglomerate. Palaeo-depressions created after the deposition of the Prestfjall Formation were consequently filled by sediments from the Hvannhagi Formation. This mostly volcaniclastic facies also comprises poorly sorted mudstone, conglomerate and sandstone, with rare coal clasts (Rasmussen & Noe-Nygaard 1970; Passey 2004; Passey & Jolley 2009).

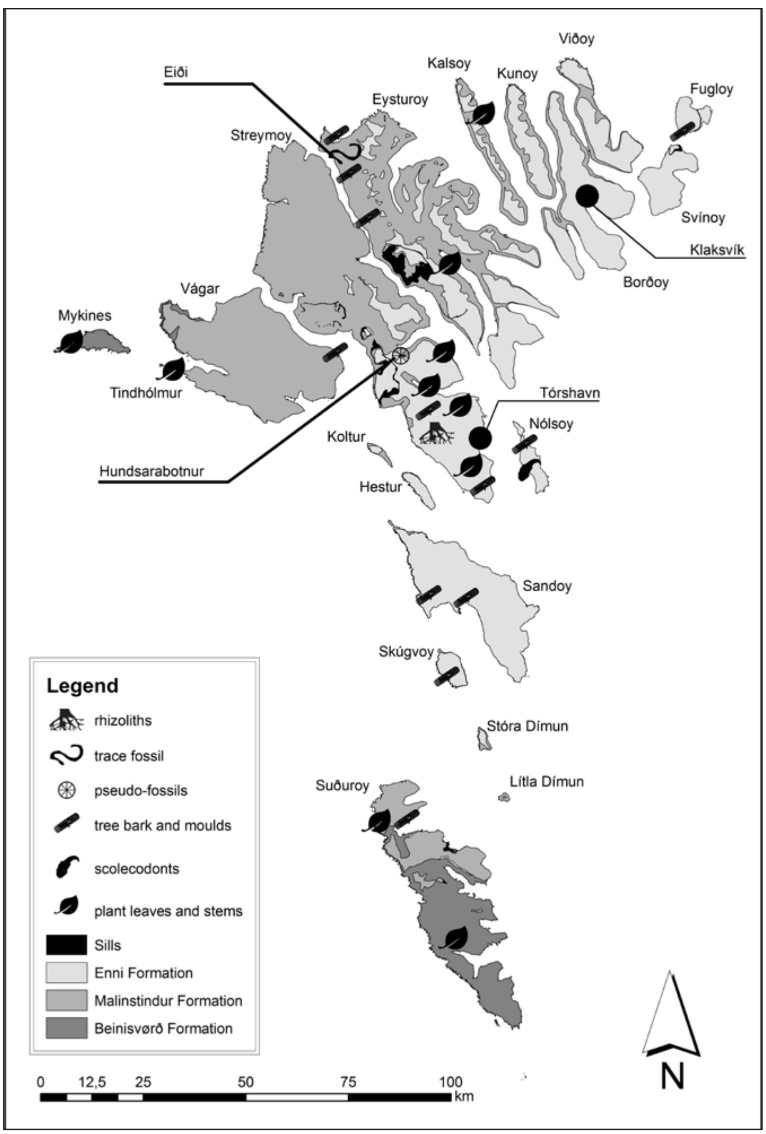

Fig. 1. Simplified geology and evidence of Palaeogene macrofossil localities on the Faroe Islands. Modified and supplemented after Hartz 1903, Noe-Nygaard 1940, Rasmussen & Koch 1963, Rasmussen & Noe-Nygaard 1990, Ellis *et al*. 2002, Passey 2004, Passey 2008 and Passey & Jolley 2009.

The Malinstindur Formation overlies the upper erosional base of the locally occurring Prestfjall and Hvannhagi Formations. Volcaniclastic sandstones and mudstones, generally several metres thick, are present throughout the lava succession, but they are most abundant in the upper part of the Malinstindur Formation (Passey 2004; Passey & Jolley 2009). Passey & Jolley (2009) defined a significant, up to 7 m thick horizon named the Kvívík Beds. The Kvívík Beds are located approximately 780 m above the base of the Malinstindur Formation. They consist of a red basal sandstone and, in the upper part, coarse-grained conglomerate and grey-coloured mudstone (Passey & Jolley 2009; Passey & Varming (2010).

The Sneis Formation comprises a laterally extensive, up to 30 m thick sandstone and conglomerate sequence. A distinct sediment horizon, the Sund Bed, can be found at the base of the Sneis Formation and marks a major hiatus in the volcanic activity. It is up to 1 m thick and composed of coarse-grained volcaniclastic sandstone and clasts of zeolitised wood. This layer is overlain by greenish-grey conglomerate (Passey 2009; Passey & Jolley 2009).

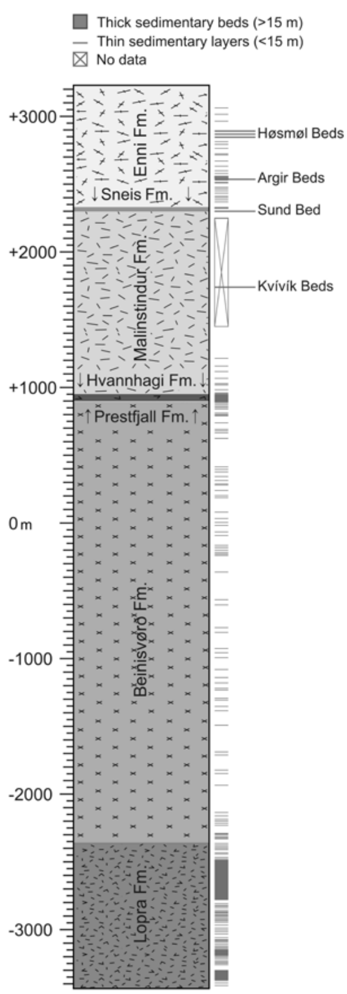

Fig. 2. Schematic stratigraphical log of the Faroe Islands with volcaniclastic sediment layers indicated to the right. The vertical scale is the stratigraphic height in metres; the part of the succession below the 0 m level is only known from drilling. After Rasmussen & Noe-Nygaard 1969, Hald & Waagstein 1984, Waagstein & Hald 1984, Waagstein 2006, Passey & Jolley 2009 and personal communication from S. Passey.

The youngest lithostratigraphic unit of the Faroe Islands is the Enni Formation. It often contains sediment horizons similar to the Malinstindur Formation. The best-developed horizon, the Argir Beds, is situated approximately 250 m above the base of the Enni Formation. The Argir Beds are 1–13 m thick and consist of thinly laminated fine-grained claystone and mudstone in the lower portion, which passes upwards into coarse-grained conglomerate (Passey 2009; Passey & Jolley 2009; Passey & Varming 2010).

A distinct sediment horizon of volcaniclastic sandstone, approximately 15 m thick, occurs on Nólsoy. Passey (2008) referred to it by the working term the Høsmøl Beds. Its stratigraphical position is 200–250 m above the base of Argir Beds, and the Høsmøl Beds are potentially the youngest significant sedimentary unit of the Faroe Islands (Fig. 2).

Review of the palaeontological record and palaeoecology

The preservation of fossil organisms on the Faroe Islands is almost wholly associated with the volcaniclastic rocks. These can form as part of the volcanism (e.g., tuffs), but they mostly represent quiescent periods during which there was sufficient time for sedimentation to take place between the lava flows (Passey 2004; Passey & Jolley 2009). The interval between single volcanic eruptions is commonly characterised by reddened tops of subaerial lava flows, where subaerial weathering of volcanic soil occurred to form the soil profile and sedimentary layer in the humid subtropical climate (Parra *et al*. 1987; Passey 2004).

The variability of the Paleocene to Eocene palaeoecosystems of the FIBG was studied in detail by Laufeld (1965), Lund (1981, 1983, 1989), Ellis *et al*. (2002), Jolley (2009) and Passey & Jolley (2009). They identified a number of fungal spores, water algae, acritarchs, dinoflagellates, bryophyte spores and flowering plant pollen grains, which predicate sedimentation in a wide range of palaeoenvironments from estuarine and shallow marine, littoral to rivers and lakes to wetlands and swamps with coal development.

Macrofossils are represented by coalified and zeolitised plant fragments that usually cannot be specified taxonomically because of missing or badly preserved morphological features (Passey 2004, 2008; Passey & Jolley 2009).

The amount of identified fossils is relatively small (Fig. 1). Fragments of *Metasequoia occidentalis* Chaney 1951 found in the shale, claystone and coal beds of the Beinisvørð Formation on Mykines were described by Hartz (1903), Rasmussen (1925), Noe-Nygaard (1940) and Rasmussen & Koch (1963). Passey & Jolley (2009) described the poorly preserved broadleaved angiosperm *Corylites hebridicus* Seward & Holttum 1924 and the pteridophyte *Equisetites* sp. at the same locality. These same horsetails were also mentioned by Passey (2008) and Passey & Jolley (2009) from the volcaniclastic sandstones of the Malinstindur Formation at several localities on Streymoy, Borðoy and Kalsoy. A well-preserved sample of *Metasequoia occidentalis* foliage was also found in the volcaniclastic sandstone of the Enni Formation on the southern part of Streymoy (Rasmussen & Noe-Nygaard 1990).

Tree moulds, often with preserved shape of the inner imprint of the bark, represent a specific type of the palaeoflora preserved on the Faroe Islands. These are usually cylindrical cavities within a lava flow with diameters of approximately several centimetres to decimetres and lengths of several tens of decimetres. They occur, for example, in lava flows of the Malinstindur Formation on Vágar and Eysturoy, as well as in the Enni Formation on Nólsoy, Fugloy and Streymoy (e.g. Passey 2004; Passey & Bell 2007; Passey 2008). Ellis *et al*. (2002) supposed that the cavities from the Sundshálsur locality on Streymoy, which are situated closely above the layers of volcaniclastic sandstone, could represent moulds of thin trunks or branches.

Ellis *et al.* (2002) described a range of ferrous rhizomorphs, namely rhizoliths and rhizoconcretions, in fine-grained volcaniclastic sediments on the southern part of Streymoy. These are residues of plant root systems that colonised the terrestrial and wetland biotopes during the periods without volcanic activity. Beds of sandstone and siltstone with rhizoliths typically have a mottled appearance.

Animal fossils are extremely rare on the Faroe Islands. Probably the only remnants of fossil animal life in the Palaeogene volcaniclastic sediments are undetermined scolecodonts from the central part of the Enni Formation on Nólsoy (Ellis *et al*. 2002).

Fossils from the period after the end of the Palaeogene volcanic activity on the Faroe Islands are very rare. The unusual find of an 8–10 Ma old rostrum from the beaked whale *Choneziphius planirostris* Cuvier 1824, found at the bottom of the sea shelf north-west of Mykines, has been only roughly described (Post & Jensen 2013).

From the Quaternary period, a well-known and published vegetation succession with a rich palaeobotanic record in lacustrine sediments and peats dates from the late Quaternary after the retreat of glaciers (e.g. Persson 1968; Jóhansen 1975, 1985, 1989; Edwards & Craigie 1998; Andresen *et al*. 2006; Jessen *et al*. 2008; Hannon *et al*. 2010; Olsen *et al*. 2010).

Rasmussen (1972) described frustules of diatoms and wood fragments, probably *Picea* sp. and/or *Larix* sp., in sandy clays of glaciomarine origin from the southern part of Borðoy. Greve (2001), Wastegård *et al*. (2005) and Abbott *et al*. (2014), following the presence of palynological material and tephrochronological analysis, dated them to the end of the Eemian Interglacial, *c*. 120 ka BP. Jóhansen (1989) mentions fragments of *Pinus* sp. from the same locality.

Holocene sand and organic conglomerate from a shallow lake on Suðuroy are interpreted as deposited by a tsunami caused by a large submarine slide, the Storegga slide, dated to 7320–6400 ^{14}C yr BP. These sediments contain small marine shell fragments, foraminifera and marine diatoms (Grauert *et al*. 2001).

A complex of Holocene organic sediment was found in borehole material from the fjord bottom collected in the southern part of Eysturoy. It contained palynomorphs, residues of insects, crustaceans and bryozoans of freshwater origin (Bennike *et al*. 1998), and was covered by clayey silt with marine shells and shell fragments, foraminifera, dinoflagellate cysts, diatoms and acritarch assemblages (Juul 1992; Roncaglia 2004; Witak *et al*. 2005). Holocene insect fauna studies from several localities of the Faroe Islands were produced by Buckland & Dinnin (1998), Buckland *et al*. (1998a, b), Edwards *et al*. (1998) and Gathorne-Hardy *et al*. (2007).

Trace fossils

Probably no trace fossils have been recorded on the Faroe Islands, except for the previously mentioned rhizomorphs from the Enni Formation on Streymoy (Ellis *et al.* 2002). Problematic are thin tubes in ironstones and claystones and the circular micro-cavities in siderite spherules from the Prestfjall Formation on Suðuroy described by Passey (2004, 2014). He considered them as the evidence of bioturbation, i.e. burrows and borings. This assertion, however, requires detailed ichnological analysis and the relationship to trace fossils remains unclear.

In the broad area around the archipelago, fossil expressions of the life processes of organisms were found only in young submarine sediments. Fu & Werner (1994) published a short list of biogenic structures (e.g. *Chondrites*-like, *Planolites*, *Scolicia*, *Teichichnus*-like etc.), probably of Holocene age, coming from the sands and silts on the Iceland–Faroe Ridge slopes approximately 150–300 km west and north of the Faroe Islands. The presence of postglacial bioturbation structures in contourites on the Faroe–Shetland Channel slopes, approximately 250 km southeast of the Faroe Islands, is mentioned by Masson *et al*. (2010). Signs of bioturbation from Holocene marine sediments on the fjord bottom close to Eysturoy of a similar origin are mentioned by Bennike *et al*. (1998) without any details.

Predation traces in foraminifera tests (*Dipatulichnus rotundus* Nielsen & Nielsen 2001) from Holocene sediments on the sea floor, approximately 200 km southwest of the Faroe Islands, were mentioned by Nielsen & Nielsen (2001).

Locality descriptions

Eiði

This exposure is situated in the north-western part of Eysturoy, on the eastern edge of the Eiði village, in a roadside cutting near road No. 23 (62°17′56.8″ N, 07°05′09.4″ W).

A 0.75–1.00 m thick succession of volcaniclastic sediments is situated between two amygdaloidal basalt lava flows of the Malinstindur Formation. The sediments form part of the Kvívík Beds (Fig. 2), and their lithology and mineralogy was described in detail by Passey (2004) who defined two different lithological units that are separated from the lava flows by sharp and planar surfaces at the top as well as at the base.

Brownish-yellow to reddish-yellow, tuffaceous, laminated, moderately sorted sandstone, approximately 50 cm thick, overlies the underlying basalt. The sandstone alternates with coarser-grained layers of darker colour in the lower part of the profile. The darker layers are characterised by highly palagonitised basalt glass and sub-rounded basalt fragments. Lighter, fine-grained layers contain palagonitised glass as well as angular plagioclase crystals, but basaltic clasts are missing. Sandstone fills abundant vertical fissures in the underlying lava flow (Passey 2004).

The lower sediments pass upwards into a reddish-brown, laminated, poorly to moderate sorted, variably grained sandstone up to 50 cm thick, containing a large amount of angular to sub-rounded clasts of palagonitised basalt glass in different stages of alteration (Fig. 3). Some of them contain plagioclase phenocrysts (Passey 2004). Sporadic trace fossils were found in this horizon.

Ellis *et al*. (2002) mention the occurrence of a palynoflora with a dominance of *Inaperturopollenites hiatus* (*Metasequoia*) with *Retitricolpites* species (possibly a riparian plant), *Monocolpopollenites tranquilus* (*Ginkgo*) and bryophyte spores (*Stereisporites* spp.) around Eiði. We found here recently also macroscopic fragments of plant fossils.

Hundsarabotnur

This exposure is in the active quarry of Hundsarabotnur, situated in the western part of Streymoy, on the eastern slope of the Skælingur mountain (767 m a.s.l.), near road No. 50.

Rocks exposed in the Hundsarabotnur quarry belong to the upper part of the Malinstindur Formation, the volcaniclastic sedimentary Sneis Formation, and the lower part of the Enni Formation. The lowest unit in the Sneis Formation is the Sund Bed, the base of which is visible in the quarry floor. The Sund Bed is overlain by conglomerates of the Sneis Formation, with a thickness of approximately 25 m. On the quarry

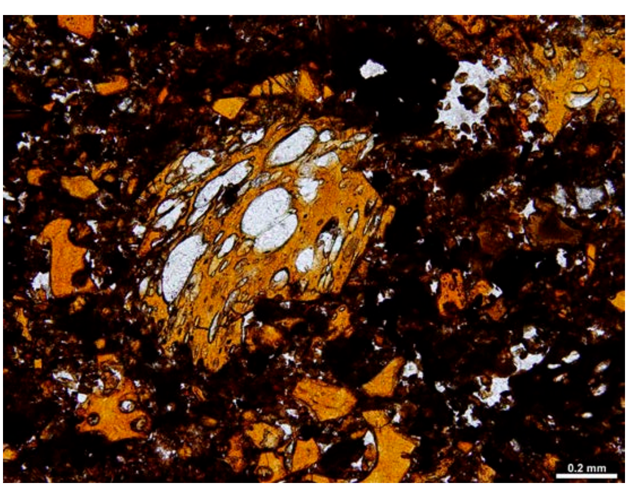

Fig. 3. Photomicrograph of fine-grained laminated sandstone of the Kvívík Beds (Malinstindur Formation) from the Eiði exposure, containing highly vesicular palagonitised basalt glass fragments. Width of picture 1.75 mm.

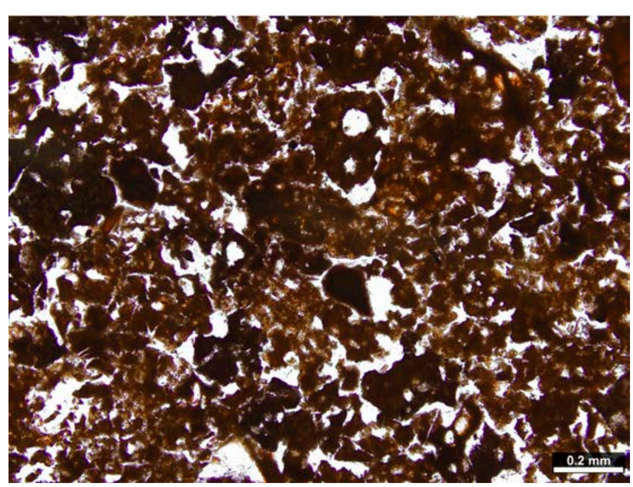

Fig. 4. Photomicrograph of fine-grained laminated sandstone of the Sund Bed from the Hundsarabotnur quarry (the base of the Sneis Formation), showing abundant palagonitised basalt glass. Width of picture 1.95 mm.

wall, the relatively simple stratigraphic arrangement is affected by the intrusion of two segments from the large Streymoy Sill. These apophyses juxtapose the Malinstindur Formation and the Enni Formation (Passey 2008; Árting & Petersen 2012).

Exposure of the volcaniclastic sediments, belonging stratigraphically close to the basal Sund Bed (62°05′36.85″ N, 06°58′44.56″ W), is seen in the local road cutting 750 m north-west of the technical buildings. The sediment horizon is 1 m thick and comprises two distinguishable units. Compact reddish-brown, medium-grained sandstone is situated in the lower part. This layer passes gradually upwards into lighter, ochre-reddish, finer-grained laminated sandstone

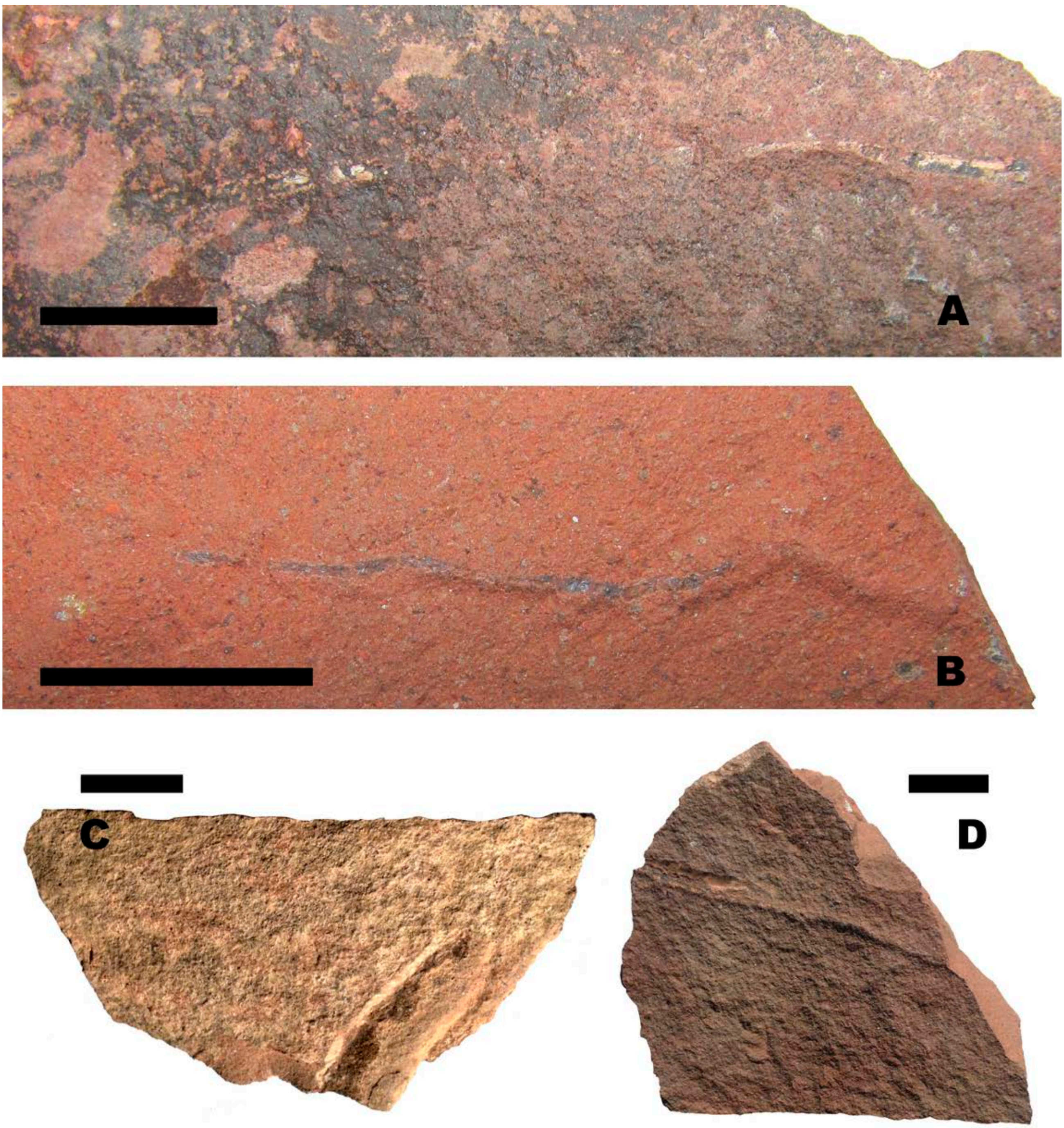

Fig. 5. Trace fossils from the Eiði exposure (Malinstindur Formation). **A**, *Helminthoidichnites* isp.; specimen Eiði2a/1. **B**, *Helminthoidichnites* isp.; specimen Eiði1b. **C**, ?*Palaeophycus* isp.; specimen Eiði4a. **D**, *Helminthoidichnites* isp.; specimen Eiði3a. The scale bar is 2 mm long in all pictures.

predominantly consisting of sub-angular fragments of highly palagonitised basaltic glass and sericitised feldspar accompanied by clay minerals and abundant opaque minerals in the matrix (Fig. 4). The upper layer contains carbonated and zeolitised fragments of plant fossils, and sporadic geometrically straight lines are found on the bedding surfaces. The contacts to the compact lava flows below and above the sediment horizon are not exposed.

Systematic ichnology

***Helminthoidichnites* isp.**
Figs. 5A, 5B, 5D, 6A

Diagnosis: Horizontal, small, thin, unbranched, simple, straight or curved, irregularly meandering or winding trails or burrows with occasional loops that commonly overlap among specimens, but lack self-overcrossing (Buatois *et al*. 1998; Schlirf *et al*. 2001).

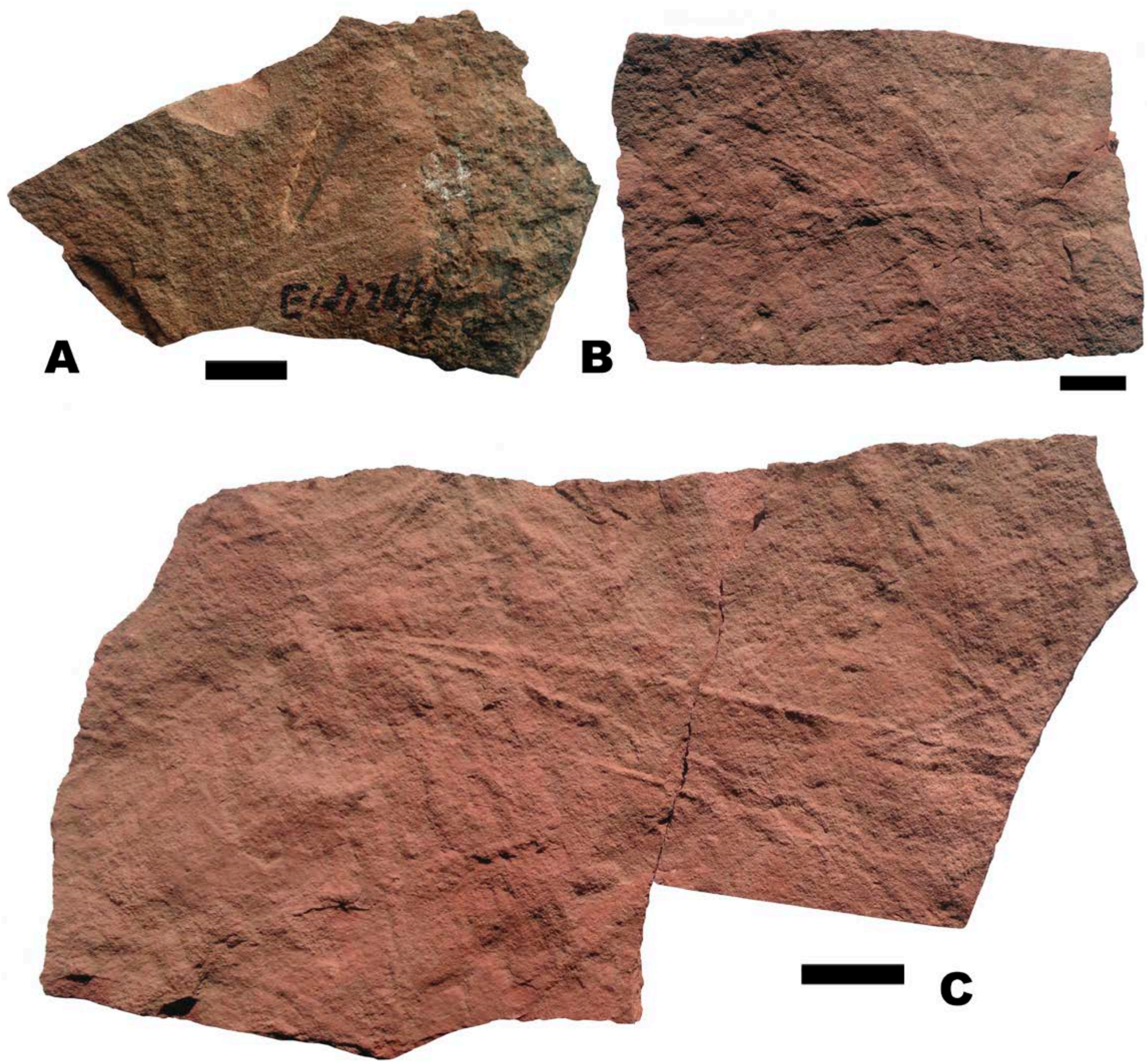

Fig. 6. **A**, trace fossil from the Eiði exposure (Malinstindur Formation), *Helminthoidichnites* isp.; specimen Eiði2b/1. **B**–**C**, linear structures from the Hundsarabotnur quarry (Sneis Formation). **B**, specimen Hund2a. **C**, specimen Hund1a/1+2. The scale bar is 2 mm long in all pictures.

Material: Four specimens – Eiði1a and Eiði1b (upper and lower bedding surface); Eiði2a/1+2 (upper bedding surface), Eiði2b/1 and Eiði2b/2 (fragments of lower bedding surface); Eiði3a and Eiði3b (upper and lower bedding surface).

Description: Thin, slightly winding to almost straight, horizontal, non-branching and non-meandering simple ridges or furrows, without self-overcrossings. The structures are preserved in a weakly convex to almost flat relief on the upper bedding surface and in a concave to flat relief on the lower bedding surface. The trace width ranges from 0.7 to 1.8 mm and is constant over the whole length. Maximal length reaches 150 mm.

Remarks: The traces are poorly preserved, without any detailed or diagnostic features. The structures resemble the segments of *Treptichnus pollardi* Buatois & Mángano 1993 identified by these authors in horizontal view (Buatois & Mángano 1993a), but the pits or nodes at the adjacent segment contacts are missing here. Moreover, the traces only have a two-dimensional character. The gently curved shape of the traces, without a tendency to meander, is different from that in *Helminthopsis* or *Cochlichnus*. The absence of self-overcrossing distinguishes them from *Gordia* (Buatois *et al*. 1997). A detailed overview of morphological differences can be found in the literature, e.g. Jensen (1997), Schlirf *et al*. (2001), Uchman *et al*. (2005) and Baucon & Neto de Carvalho (2008).

Helminthoidichnites probably represents a grazing trace (pascichnia), where the tracemakers are assumed to be unspecified Nematomorpha, Arthropoda, or insect larvae (Buatois *et al*. 1997, 1998). In addition to occurring in marine sediments, *Helminthoidichnites* is mentioned in many freshwater palaeoecosystems, including floodplains (Buatois *et al*. 1997; Schlirf *et al*. 2001; Buatois & Mángano 2002; Uchman *et al.* 2004; Melchor *et al*. 2006), estuaries, lacustrine deltas, shallow lacustrine facies (Buatois *et al*. 1998, 2000; Melchor 2004; Voigt 2005; Melchor *et al*. 2006), glaciolacustrine lakes (Buatois & Mángano 1993b; Gaigalas & Uchman 2004; Uchman *et al*. 2008, 2009), and turbidites of deep lakes (Buatois *et al*. 1996).

?*Palaeophycus* isp.
Fig. 5C

Diagnosis: Branched or, more typically, unbranched, straight to slightly curved to slightly undulate or flexuous, smooth or ornamented, typically lined, elliptical to circular in cross-section, predominantly horizontal structures interpreted as originally open burrows; infill typically structureless, massive, of same lithology or similar to host rock. Where present, the bifurcation is not systematic, nor does it result in swelling at the site of branching (Pemberton & Frey 1982; Fillion & Pickerill 1990; Kim *et al*. 2001).

Material: One specimen – Eiði4a (upper bedding surface).

Description: Short fragment of the horizontal, slightly curved, weakly convex, non-branching trace, 40 mm long and 15 mm wide. The trace is flat-oval in cross section, with a height of 3.5 mm. The surface is smooth, and a wall is faintly visible as a dark coloured thin coating. The burrow-fill is massive and similar to the host rock.

Remarks: A fragment of only one specimen prevents a more accurate determination. The presence of wall coating distinguishes this trace from *Planolites*. Due to the absence of surface sculpture, we suggest a relationship to *Palaeophycus*, probably *P. tubularis* Hall 1847.

Palaeophycus is interpreted as a dwelling or deposit feeding structure (domichnia or fodinichnia), produced infaunally and thereafter passively filled in. The tracemaker may be different types of vermiform animals, probably polychaetes or annellids (Pemberton & Frey 1982; Keighley & Pickerill 1995). *Palaeophycus* is a representative of eurybathic traces in a wide range of marine and freshwater palaeoenvironments including floodplains (Buatois *et al*. 1997; Buatois & Mángano 2002; Kim *et al*. 2002; Krapovickas 2012), rivers (Wang *et al*. 2014) and lakes (Buatois *et al*. 2000; Melchor 2004; Melchor *et al*. 2006). Ekdale *et al*. (2007) and Krapovickas *et al*. (2010) described *Palaeophycus* from aeolian sediments.

Linear structures

Fig. 6B, 6C

Material: Two specimens - Hund1a/1+2 (upper bedding surface); Hund2a and Hund2b (upper and lower bedding surface).

Description: Fragments of sandstone containing numerous simple, straight and thin convex lines, subcircular to Λ-shaped in cross-section. The lines reach up to 100 mm in length and 1.5 mm in width. The lines intersect each other at many places. The infill of the structures is identical to the host rock.

Remarks: Geometrically straight lines usually indicate an abiotic origin. Wetzel *et al.* (in press) describe the trails of modern earthworms with a similar arrangement pattern, but these traces differ from the specimens in the Hundsarabotnur quarry by the not entirely linear trails, and the presence of morphological features such as the imprints of setae, pseudospreiten, blind deviations part of traces and tunnel openings. By studying the thin sections, the lines were found to have the same mineral composition as the surrounding rock. This fact excludes an explanation as pseudomorphs after needle-like minerals formed during diagenesis or due to the low-grade contact metamorphism of sediments by the overlying lava flow. The straight shape of the lines partly resemble the frost cracks arising due to seasonal freezing of sediment, the formation of elongated ice crystals and subsequent infilling of the fissures by sediment after the melting of the ice. Very similar structures were published by Udden (1918) and Fortier *et al.* (2008). Ice crystal marks, however, can be excluded due to the warm character of the Palaeogene climate (Vandenberghe *et al.* 2012). Based on the palaeoenvironment characteristics, the probable process leading to the creation of the structures in the Hundsarabotnur quarry could be described as the gradual drying and shrinking of the sediments in a subaerial environment (e.g. Geoff Tanner 2003; Sadhukan *et al.* 2007). Although the main distinguishing sign of such desiccation structures – a polygonal arrangement – was not observed here, it can be attributed to the fragmentary preservation of the examined specimens. The linear structures are thus pseudo-fossils.

Summary and conclusions

The fine- to medium-grained volcaniclastic sandstones of the Palaeogene Faroe Islands Basalt Group consist mainly of reworked palagonitised basaltic glass and are traditionally associated with fluvial channel environments and their surroundings (Rasmussen & Noe-Nygaard 1970). Based on the lack of brecciation of the overlying lava flows Passey (2004) concluded that the land surface was without deep lakes and was only occasionally affected by seasonal flooding events at the time of lava emplacement. The important evidence of exposure to air in the warm climate is the iron-oxide-rich, red-coloured sedimentary formations and flow tops. This is supported by the occurrence of generally angular to sub-rounded basaltic clasts, indicating short-time/short-distance transport (Passey 2004), as well as the fossil evidence of mostly terrestrial or river- to lake-littoral palynomorphs (e.g. Ellis *et al.* 2002; Passey & Jolley 2009).

The subaerial, occasionally inundated, non-marine moist or wet, soft to loose ground environment corresponds to the *Scoyenia* ichnofacies. This trace-fossil assemblage however, is typically dominated by meniscate burrows, arthropod trackways and simple vertical burrows (cf. Frey *et al*. 1984; Buatois & Mángano 1995).

Helminthoidichnites isp. and ?*Palaeophycus* isp. that were found at Eiði (Malinstindur Formation) indicate a low to moderate hydrodynamic energy environment, which represents suitable conditions for benthic tracemakers (Buatois *et al*. 1997). These ichnogenera are, however, usually present as the diagnostic elements of the *Mermia* ichnofacies, which differs from the previous by occurring in permanently subaqueous zones, represented by the shallow to deep zones of lacustrine systems. Both of the mentioned ichnogenera are occasionally found in the *Scoyenia* ichnofacies (Buatois & Mángano 1995; Buatois *et al*. 1997; Melchor *et al*. 2012).

Based on the above assertions, the new discoveries of trace fossils and possible desiccation cracks confirm that the environmental character of the Palaeogene interlava sedimentary periods of the Faroe Islands was a mosaic landscape of floodplains, interwoven with rivers and wide shore-lined shallow lakes, with the presence of an impoverished *Scoyenia* ichnofacies.

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Glacial erratic boulders from Jutland, Denmark, feature an uppermost lower Cambrian fauna of the Lingulid Sandstone Member of Västergötland, Sweden

THOMAS WEIDNER, GERD GEYER, JAN OVE R. EBBESTAD & VOLKER VON SECKENDORFF

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Reinvestigation of glacial erratic boulders from Jutland, Denmark, and from northern Germany, has revealed a moderately diverse fauna with the trilobites *Holmiella*? sp., *Epichalnipsus anartanus*, *Epichalnipsus* sp. A, *Epichalnipsus* sp. B, and *Berabichia erratica*, three species of lingulid brachiopods, one hyolith species, and trace fossils comparable to *Halopoa imbricata*. Comparison with faunas from the Cambrian of Scandinavia strongly suggested a biostratigraphic position equivalent to the uppermost part of the (revised) *Holmia kjerulfi*–'*Ornamentaspis*' *linnarssoni* to lowermost *Comluella*?–*Ellipsocephalus lunatus* zones sensu Nielsen & Schovsbo (2011), or the lower to middle part of the traditional '*Ornamentaspis*' *linnarssoni* Zone, but probably a particular horizon and biofacies not yet discovered in Scandinavia. Considerations of glacial transport regimes and the distribution of comparable rock units, as well as a petrographical analysis of the material from the studied erratic boulders and rocks from outcrops in Sweden, indicate that the boulders were derived from the Lingulid Sandstone Member of the File Haidar Formation and the source area is situated in the vicinity of the present-day outcrops in the Halleberg–Hunneberg area, Västergötland, Sweden.

Keywords: Trilobites, Lingulid Sandstone Member, lower Cambrian, Cambrian Series 2, Västergötland, Sweden, Denmark, biostratigraphy, glacial boulders.

Thomas Weidner [to.we@paradis.dk], Ravnholtvej 23, Rårup, DK-7130 Juelsminde, Denmark. Gerd Geyer [gerd.geyer@uni-wuerzburg.de], Institut für Geographie und Geologie, Lehrstuhl für Geodynamik und Geomaterialforschung, Bayerische Julius-Maximilians-Universität Würzburg, Am Hubland, D-97074 Würzburg, Germany. Jan Ove R. Ebbestad [jan-ove.ebbestad@em.uu.se], Museum of Evolution, Uppsala University, SE-75236 Uppsala, Sweden. Volker von Seckendorff [volker.von_seckendorff@uni-wuerzburg.de], Institut für Geographie und Geologie, Lehrstuhl für Geodynamik und Geomaterialforschung, Bayerische Julius-Maximilians-Universität Würzburg, Am Hubland, D-97074 Würzburg, Germany.

Corresponding author: Gerd Geyer

Trilobite-bearing lower Cambrian strata (provisional Cambrian Series 2) in Sweden have been studied mainly at scattered localities in the south (Scania) and in numerous outcrops along the Caledonian mountain range (Jämtland, Ångermanland and Lapland), in both the autochthonous and allochthonous sequences (Fig. 1A). Rocks of early Cambrian age predominantly consist of siltstone and sandstone. Trilobites recorded from these strata of Baltica are fairly restricted by comparison to those from other Cambrian continents and belong to the olenelloid family Holmiidae (*Holmia*, *Holmiella*, *Kjerulfia* and *Schmidtiellus*), to the ptychoparioid family Ellipsocephalidae (with species assigned, or provisionally attributed to the genera *Comluella*, *Ellipsocephalus*, *Ornamentaspis* and *Strenuaeva*), and the eodiscid genera *Runcinodiscus*, *Calodiscus*, *Chelediscus* and *Neocobboldia*. The early Cambrian faunas and biostratigraphy of the Swedish regions are portrayed by Bergström (1973), Ahlberg & Bergström (1978, 1983, 1991), Ahlberg (1979, 1980, 1984a, 1984b, 1984c, 1985), Bergström & Ahlberg (1981), Ahlberg *et al*. (1986), Nielsen & Schovsbo (2006, 2011) and Høyberget *et al.* (2015). Studies during the last 15 years have increased our knowledge about these faunas, particularly of those from localities in the northern Caledonides (Moczydłowska *et al*. 2001; Axheimer *et al*. 2007; Cederström *et al*. 2009, 2011, 2012). Due to both outcrop conditions and depositional environments, polymerid trilobites are mostly fragmentary and often

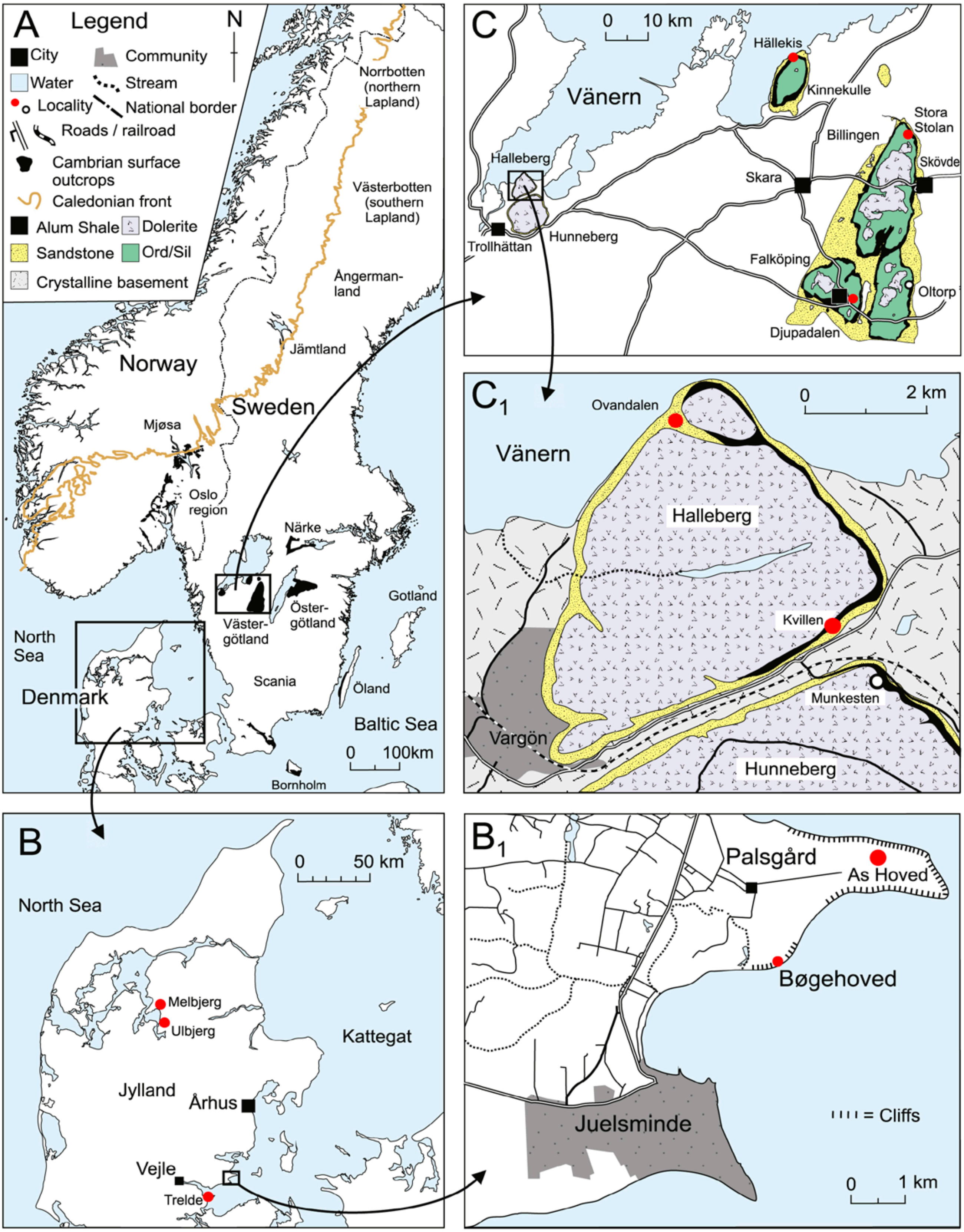


Fig. 1. A, map of Scandinavia showing outcrops of the Cambrian with inserts of the areas in Denmark and Sweden treated in the text; **B** and **B**$_1$, maps of Jutland, Denmark, with localities (red) yielding boulders of the fossil-bearing Lingulid Sandstone; **C**, map of Västergötland showing Lower Palaeozoic outcrops at Halleberg, Hunneberg, Kinnekulle, and Falbygden to the east with localities (red) yielding the fossil-bearing Lingulid Sandstone; **C**$_1$, detailed map of Halleberg with the localities Ovandalen and Kvillen (base map from Andersson *et al*. 1985).

do not permit a precise determination; they belong to genera or species endemic to Baltica and are of limited use for intercontinental or even global correlation, whereas correlation into other Cambrian continents is facilitated by the species of *Calodiscus* and *Chelediscus*. The lower Cambrian biostratigraphy of Scandinavia and Baltica in general has thus been in a state of preliminary subdivision and was recently revised by Nielsen & Schovsbo (2011) as shown in Fig. 2.

Of particular interest is the upper lower Cambrian part of the succession, which provides distinct problems for correlation into other regions as emphasized in the discussions on the Cambrian Series 2–Series 3 and Stage 4–Stage 5 boundary (e.g., Geyer & Palmer 1995; Fletcher 2003; Geyer 2005; Sundberg *et al*. in press). The traditional subdivision and subsequent zonal schemes that recognize a *Holmia kjerulfi* group Zone, a *Proampyx* (or *'Ornamentaspis'*) *linnarssoni* Zone, and a *'Protolenus'* Zone (Fig. 2) are equally unable to characterize the trilobite faunas, lack a recognizable evolutionary development of the trilobites, and are not even readily correlatable with the neighbouring regions of Baltica in northern Poland or the Holy Cross Mountains in southern Poland.

Geyer *et al*. (2004) described trilobites of the ellipsocephaloid genera *Epichalnipsus* and *Berabichia* from glacial erratic boulders collected at As Hoved and other localities in Denmark and northern Germany. These boulders of trilobite-bearing quartzitic sandstones were interpreted as being derived from the Lingulid Sandstone Member of Västergötland and to represent an unidentified stratigraphical interval. Based on close morphological affinities with ellipsocephaloid trilobites from the Moroccan Atlas ranges, Geyer *et al.* (2004) proposed a stratigraphic correlation with the *Sectigena* Zone of Morocco, West Gondwana. However, differences in lithology suggested that this type of Lingulid Sandstone was unknown in Sweden.

Lingulid Sandstone: distribution, lithology, and biostratigraphy

A key unit for the upper lower Cambrian is the Lingulid Sandstone which is generally treated as a member of the File Haidar Formation (Thorslund &

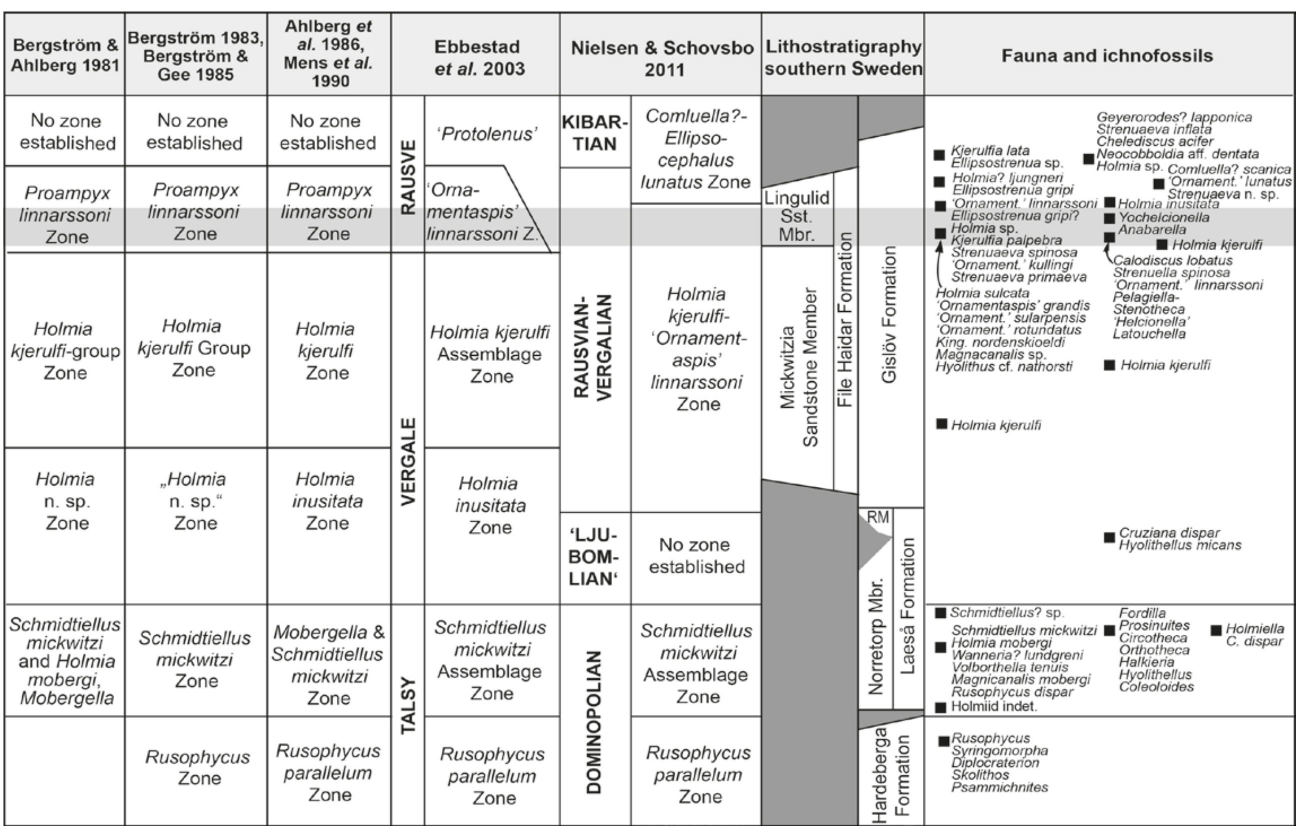


Fig. 2. Biostratigraphic chart of the Scandinavian classical lower Cambrian with the historic development of the lithostratigraphical units for Västergötland and Scania (left and right columns under Lithostratigraphy southern Sweden) and the tentative position of important fossil horizons mentioned in numerous publications (cited in the text) from southern and central Sweden, the Swedish Allochthon and Lapland, the Mjøsa District in Norway and the Digermul Peninsula, Finnmark, Norway. Stratigraphic interpretations that differ from the original concepts follow data in Nielsen & Schovsbo (2011). RM, Rispebjerg Member.

Westergård 1938; Thorslund 1960; Bergström & Gee 1985; Nielsen & Schovsbo 2006). The Lingulid Sandstone was first adequately recognized by Holm (in Holm & Munthe 1901) who separated Linnarsson's Fucoid Sandstone (see Wallin 1868) into a lower and an upper unit termed Mickwitzia and Lingulid Sandstone, respectively, based on differences in lithological characters and bedding thicknesses. It is generally regarded as a depositional tongue of the File Haidar Formation extending from the stratotype on Gotland (Thorslund & Westergård 1938; Bergström & Gee 1985) into Öland, Västergötland, Östergötland and Närke. The two units of the File Haidar Formation form the uppermost part of the traditional lower Cambrian succession, but their vertical extent shows an opposite trend. The Mickwitzia Sandstone Member increases in thickness eastward, whereas the Lingulid Sandstone Member has its maximum thickness in the west. In addition, the boundary between the two is locally (Lugnås, Billingen) marked by a conglomeratic or coarse-grained layer (Linnarsson 1871; Hadding 1927; Westergård 1931; Bergström & Gee 1985; Jensen 1997; Nielsen & Schovsbo 2006) which appears to indicate a hiatus of unknown extent between the units. The Lingulid Sandstone Member appears to testify a change in depositional regime and has been interpreted as a result of deposition in deeper water than the Mickwitzia Sandstone Member, and the sparse amount of fine-grained material has been understood as indication of increased wave action (Hadding 1927; Westergård 1931). Nevertheless, Martinsson (1974) advocated the differences to be a result of deposition at a later stage in the basin. Nielsen & Schovsbo (2011) interpreted both units as two sequences of a continuous depositional history.

Nielsen & Schovsbo (2006) selected the stratotype of the Lingulid Sandstone Member in the Bårstad-2 drill core in Östergötland, which has been described in detail by Eklund (1990).

The Lingulid Sandstone Member is widely distributed in Västergötland in narrow outcrops at the flanks of the renowned table mountains, where it is composed of light grey, moderately to poorly cemented fine-grained quartz sandstone. Its maximum thickness in Västergötland is 28.6 m in the DBH 15/73 core of southern Billingen (Nielsen & Schovsbo 2006).

Only three reports of trilobite findings exist for the Lingulid Sandstone Member of Västergötland (Ahlberg *et al.* 1986). The upper part of the member in the quarry at Stora Stolan, northern Billingen, yielded a large fragment of a cheek with a genal spine that was tentatively assigned to *Holmia grandis* Kiær 1917, a species otherwise known only from the lower part of the *Holmia kjerulfi* Zone of the Mjøsa area, Norway (Kiær 1917; Ebbestad *et al.* 2003). The other report is a fragmentary pygidium identified as *Holmiella* sp. from the lower part of the Lingulid Sandstone Member in a quarry at Hällekis at Lake Vänern, Kinnekulle (Ahlberg *et al*. 1986). However, this specimen had originally been attributed to the Mickwitzia Sandstone (Bergström 1981; Ahlberg 1984c). This incomplete pygidium and numerous cranidial fragments from this locality were assigned to *Holmiella* sp. Both olenelloid taxa were assumed to indicate the *Holmia kjerulfi* Zone and are of little biostratigraphic significance, although Jensen (1997) emphasized the lithological equivalence of the host rocks with the Mickwitzia Sandstone. Furthermore, Jensen (1997) cited the discovery of a relatively complete trilobite by collector Holger Buentke in 1992, which has remained unpublished since and also is uncertain with respect to its precise stratigraphic position. Additional specimens of olenelloids were discovered at the eastern tip of Halleberg by Jan Johansson (personal communication to TW), and ellipsocephalid trilobites were collected by Frank Rudolph in erratic boulders in northern Germany (unpublished, examined by TW). Eponymous linguliform brachiopods are also known from several localities but have rarely found their way into collections. In summary, the fossil content of the Lingulid Sandstone appears to be far less sparse than commonly assumed, and the poor record is primarily an artefact of collecting and sampling.

This macrofossil record permits an unequivocal assignment to one of the established zones. Nielsen & Schovsbo's (2011) sequence stratigraphical approach placed the Lingulid Sandstone Member into the LC2–4 sequence of the newly established *Comluella*?–*Ellipsocephalus lunatus* Zone. This issue is discussed below (under 'Stratigraphy') in some detail.

Acritarchs from the File Haidar Formation were studied in some detail from drill cores obtained on Gotland, the Gotska Sandön and the southern Bothnian Sea (Hagenfeldt 1989a, 1989b; Hagenfeldt & Bjerkéus 1991), and from Östergötland (Eklund 1990). The recovered assemblages include the *Heliosphaeridium dissimilare–Skiagia ciliosa*, *Volkovia dentifera–Liepaina plana* and part of the *Eliasum–Cristallinium* assemblage zones, but are of little significance for a precise determination of the boundary between the members (see discussion in Landing *et al.* (2013) and Sundberg *et al*. in press).

The lithology of the Grötlingbo-1 core from Gotland was used by Hagenfeldt & Bjerkéus (1991) to introduce a subdivision of the File Haidar Formation into four units termed the Viklau Sandstone, När Shale, När Sandstone and Grötlingbo Siltstone, of which the Viklau Sandstone and the Grötlingbo Siltstone were ranked as formal members (Hagenfeldt 1994). The När Shale and När Sandstone, in contrast, have been regarded as subunits of a När Member but were subsequently classified as members in the Öland–

Gotland region by Nielsen & Schovsbo (2006), who subdivided the formation into the Viklau, När Shale, När Sandstone and Grötlingbo Members. Of these, the När Sandstone Member is regarded as more or less equivalent to the Lingulid Sandstone Member.

Localities

As Hoved locality

As Hoved is a *c.* 6 km long cliff-framed headland near Palsgård manor, situated 8 km north of Juelsminde (around 55°45′N, 10°04′E; Fig. $1B_1$) at the east coast of Denmark, facing the Kattegat sea. This locality has an exceptional accumulation of glacial erratic boulders of Cambrian age. Along the shore thousands of boulders of lower Cambrian sandstones occur together with hundreds of boulders of middle Cambrian bituminous limestone (Swedish: orsten), limestone, conglomerate and alum shale, and thousands of boulders of Furongian bituminous limestone and shale (Fig. 3). Glacial erratic blocks of the middle Cambrian *Paradoxides paradoxissimus* Superzone act as index blocks which allow tracking their origin to Sweden, because the rocks of this superzone show particular characteristics for each province in southern and central Sweden (Scania, Öland, Västergötland, Östergötland and Närke). The foremost rock types at As Hoved that unmistakably point to Västergötland (Fig. 1C) as the source area of all the boulders are:

a. a phosphoritic sandstone conglomerate (Swedish: basalkonglomerat) with constituents of the Lingulid Sandstone. This unit marks the boundary between the lower Cambrian and the middle Cambrian sequence in Västergötland;
b. a greenish limestone of the *Triplagnostus gibbus* Zone with *Jincella munsteri* as found at Oltorp, Billingen (Westergård 1953, p. 12);
c. a greenish glauconitic limestone of the *Acidusus atavus* Zone occurring *in situ* at Oltorp and as loose boulders from other localities in southern and northern Billingen and on Mount Kinnekulle;
d. a highly fossiliferous, metamorphosed ('baked') bituminous limestone of the *Acidusus atavus*, *Ptychagnostus punctuosus* and *Goniagnostus nathorsti* zones found *in situ* at Munkesten, Hunneberg;
e. the existence of a thin conglomeratic layer spanning the *Ptychagnostus punctuosus* and *Goniagnostus nathorsti* zones, discovered for the first time at As Hoved; eventually proven in Falbygden and on Mount Kinnekulle (Weidner *et al.* 2004);
f. countless blocks of dolerite that now cap the Lower Palaeozoic strata in Västergötland.
g. Houmark-Nielsen (1987, 1994, locality 92) demonstrated that the four Pleistocene glacial tills exposed at As Hoved cliff were derived from the north (Norway) and the north-east (Västergötland). The glacial erratic boulders transported from the north are dominated by rhomb porphyry and sedimentary rocks from the Danish subsurface, mostly of Palaeogene and Cretaceous age.
h. No index blocks from Scania (*Exsulans* Limestone) or Öland ("*Oelandicus* shales", "*Paradoxissimus* Sandstone") have been found at As Hoved.

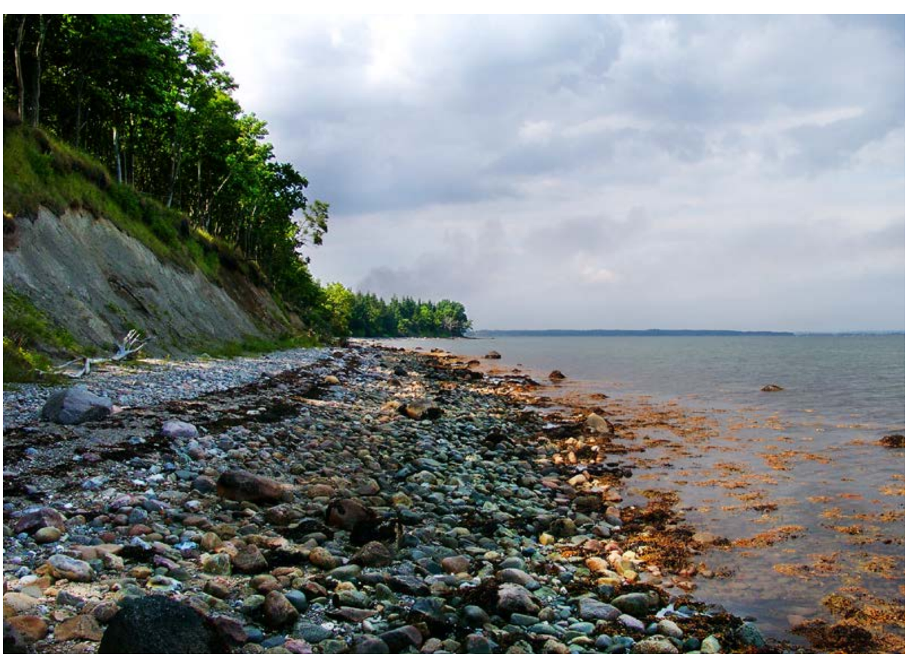


Fig. 3. The cliff section at As Hoved near Palsgård, Jutland, with glacial erratic boulders along the beach.

Additional erratic boulders with Lingulid Sandstone lithology and equivalent fossils of *Epichalnipsus* faunal aspect have already been described from Ulbjerg and Melbjerg, Jutland, Denmark, and Sandesneben (Herzogtum Lauenburg county, Schleswig-Holstein), northern Germany, by Geyer *et al.* (2004). Other material has subsequently been collected from Damsdorf and Stocksee (both Segeberg county, Schleswig-Holstein), and Köhlen (Cuxhaven county, Niedersachsen), Germany.

Halleberg localities

Fossiliferous quarzitic sandstones have a wide geographical distribution in glacial deposits. In Denmark they are found, in addition to As Hoved, at other localities with Pleistocene moraines containing rocks characteristic of Västergötland (Bøgehoved, Trelde, Melbjerg, and Ulbjerg cliffs, Fig. 1). These sandstones are also scattered over the whole of northern Germany. However, they are unknown in Sweden, which prompted one of the authors (TW) to search for their source primarily in the Falbygden and Kinnekulle areas of Västergötland (Fig. 1), from where most of the middle Cambrian material from As Hoved has originated. Ten years of investigation in natural and man-made outcrops, as well as loose boulders in Västergötland, subsequently also in Östergötland and Närke, was unsuccessful. Only the regular Lingulid Sandstone was met in the field, with all boulders devoid of trilobite remains. In 2011, the search was extended to the Halleberg–Hunneberg outliers, two table mountains in the far west of Västergötland. Andersson *et al.* (1985, fig. 11) has shown that at certain parts of Halleberg and Hunneberg (Fig. $1C_1$), a dolerite sill rests directly upon the Lingulid Sandstone or on a thin stratum of middle Cambrian alum shale and that the "diabase intrusion had baked the alum shale" Andersson *et al.* (1985, p. 19). Four exposures, a few metres long, and known since a long time in the literature (Hansen 1933), can be studied in the Ovandalen valley in the north-western corner of Halleberg where the public trail starts to descend from the dolerite platform down to Lake Vänern. Between 50 and 150 m north of the trail, three outcrops expose the dolerite sill where it is underlain (in descending order) by metamorphosed alum shale (*c.* 40 cm), metamorphosed bituminous limestone of the *Triplagnostus gibbus* Zone (*Paradoxides paradoxissimus* Superzone; *c.* 25 cm) and again by metamorphosed alum shale (*c.* 35 cm). The lower boundary of this alum shale bed is covered by debris of dolerite, soil and vegetation (Fig. 4). The foot of the approximately 10 m high slope is strewn with smaller blocks of quarzitic sandstone of a lithology quite similar to the erratic boulders known from As Hoved, but trilobite remains were not discovered in the material. Another section *c.* 100 m to the south of the trail indicates that the alum shale pinches out, and the boundary between the dolerite sill and the

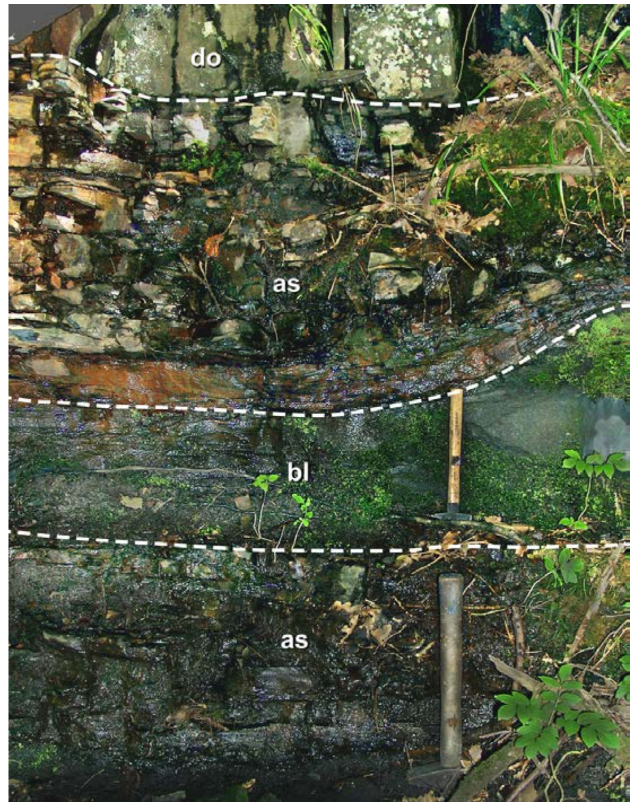


Fig. 4. Section 1 at Ovandalen on Halleberg, with the contact between a dolerite sill (do) overlying a strongly metamorphosed sequence comprising alum shale (as), a bituminous limestone bed of the *Triplagnostus gibbus* Zone (bl), and alum shale (as) the lower part of which is covered by soil. Length of hammer shaft 33 cm.

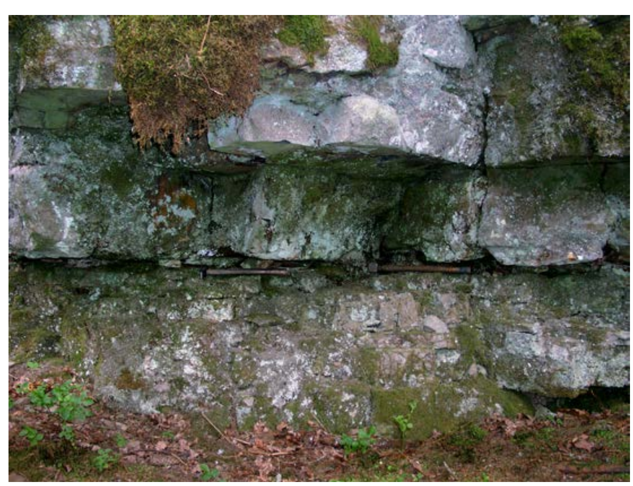


Fig. 5. Section 2 at Ovandalen on Halleberg, showing the boundary between the Lingulid Sandstone Member and the overlying dolerite sill (marked by the two hammers). Length of the right hammer shaft 33 cm.

underlying Lingulid Sandstone is visible (Fig. 5). This section lies 150 m from the nearest outcrop north of the trail. The same quartz-arenitic sandstones are also found as loose pebbles along the prehistoric shore-line at Kvillen close to the south-eastern corner of Halleberg. Trilobite remains from this locality were reported by Jan Johansson (personal communication 2002) but the material has not been published.

Karlsson (2001) demonstrated that the 90 m thick Permo–Carboniferous dolerite cap on Halleberg and Hunneberg, either resting directly on or only separated from the underlying sandstone by a thin cover of alum shales, has influenced the diagenetic development by hydrothermal overprint. This resulted in the regionally restricted occurrence of the quartzitic variety of the Lingulid Sandstone. In the east, at Billingen, in contrast, the dolerite cap attains only 45 m, and the dolerite is further separated by a thick Palaeozoic cover from the sandstone beneath. Therefore, the Lingulid Sandstone remained largely unaffected in the eastern areas of Västergötland.

Thus, we conclude that the quartzitic sandstones found in great quantity at As Hoved were derived from the Halleberg–Hunneberg area or another area where the dolerite was in close contact with the underlying sandstone, but in which the whole succession has been subject to denudation. Ovandalen is the only available outcrop on Halleberg–Hunneberg that reveals the boundary between the Cambrian rocks and the dolerite sill. Only the uppermost metre of the sandstone is exposed (the lower parts being concealed by large dolerite blocks), and therefore the exact position of this level within the Lingulid Sandstone Member of Västergötland remains uncertain.

Lithology and petrology of samples from the As Hoved cliff and the Lingulid Sandstone of Västergötland

The sandstone of the glacial erratic boulders from As Hoved and bedrock samples from Västergötland were studied to allow a reliable comparison of lithological characters. The As Hoved samples and two samples from the Kvillen section, Halleberg, and the Djupadalen locality were examined petrographically in thin sections. See Appendix 1 for systematic comparisons.

The studied sandstone of the As Hoved cliff displays a variety of lithological characteristics. The sandstones are either platy (most frequently separating into slabs with thicknesses of 1–2 cm) or massive blocks with bed thicknesses of up to 30 cm. The general colour is light grey, occasionally with a yellowish, greenish or reddish stain. The sand grains are generally fine-grained, with nearly even sizes, and relatively pure, indicating a high maturity.

Thin sections (Fig. 6A–D) indicate a pore volume of *c.* 1–5 vol.%. Quartz grains constitute *c.* 94 vol.% of the rock and range from 0.03 to 0.3 mm in diameter, with the prevalence of a slight longitudinal shape. They show high compaction and cementation as suggested by inconspicuous concave–convex and saturated grain contacts, and syntaxial overgrowths. Ferritic/limonitic coats are rarely present. Feldspar accounts for *c.* 1 vol.%, mica for *c.* 2 vol.%, glauconite for *c.* 2 vol.%, and zircon and tourmaline together for *c.* 0.5 vol.%. Alkali feldspar as well as plagioclase grains are fairly fresh and present as slightly rounded cleavage fragments (up to 0.3 mm in size, but generally around 0.1 mm and less). Mica consists nearly exclusively of white mica; only a few flakes of dark mica are present. Glauconite grains are generally around 0.1 mm in size and slightly rounded, mostly compacted and partly show a pale brown crust. The argillaceous-chloritic and illitic porous cement is restricted to minute interporous spaces. A secondary diffuse film of rust is occasionally visible between the quartz grains.

Sedimentary structures are usually inconspicuous. Most samples show a fine, nearly parallel bedding or a very low angle cross-stratification, which becomes obvious after splitting.

A sample of typical thick-bedded sandstone of the Lingulid Sandstone Member from the Kvillen section, Halleberg (sample 09/05), shows remarkable similarity to the As Hoved sample. Thin sections (Fig. 6E, F) indicate the same amount of pore space of *c.* 1–5 vol.%, which is in accordance with the results of Karlsson (2001), who determined a porosity of 3–6 vol.% for samples collected at Hunneberg (with a resulting permeability of 0 mD). The quartz grains constitute *c.* 90 vol.% of the rock and are from 0.04 to 0.15 mm in size, with the prevalence of slightly longitudinal grains. Ferritic coats are rarely observed. Feldspar accounts for *c.* 1 vol.%, mica for *c.* 3 vol.%, glauconite for *c.* 5 vol.%, and zircon and tourmaline together for *c.* 0.5 vol.%. Alkali feldspar and plagioclase grains are fairly fresh and present as cleavage fragments. Glauconite grains are generally about 0.1 mm in size and slightly rounded, mostly compacted and partly show a pale brown crust. Mica consists predominantly of white mica, with a minor amount of biotite present. The minute interporous spaces are filled with an argillaceous-chloritic and illitic cement.

Sample 3 of the typical sandstone of the Lingulid Sandstone Member from the Djupadalen locality shows a similar composition as the sandstones from

As Hoved and Halleberg, but differs in a number of minor aspects (Fig. 6G, H). Thin sections of the quartz arenite indicate a distinctly higher pore space of *c.* 10–15 vol.%, which is clearly filled with argillaceous-ferritic cement (illite + goethite) and distinct syntaxial quartz overgrowths. The quartz grains constitute *c.* 97 vol.% of the rock and range from 0.05 to 0.3 mm in size, with slightly rounded grains of low sphericity, and distinctly platy grains with frequent ferritic coats. Feldspar and mica account together for *c.* 1 vol.%, glauconite for *c.* 1 vol.%, and zircon and tourmaline together for *c.* 0.5 vol.%. Alkali feldspar as well as plagioclase grains are fairly fresh, but alkali feldspar grains commonly show a fringe altered to illite or white mica. Glauconite grains are generally around 0.1 mm in size, rounded, and mostly compacted, but are devoid of external crusts. Mica is present as white mica, but also as biotite (partly altered to chlorite).

It should be emphasized that the samples from the Lingulid Sandstone Member of Västergötland studied by Karlsson (2001) have a feldspar content between 2.3 and 6 vol.%, with some samples having a distinct content of rock fragments up to 5.6 vol.%. However, the samples described by Karlsson (2001) are all devoid of any notable amounts of Fe-hydroxide.

Fossils

Fossils occur predominantly in platy sandstone and occur as external moulds. They are often poorly preserved as fairly eroded imprints due to perennial wave transport on the present-day shoreline. The

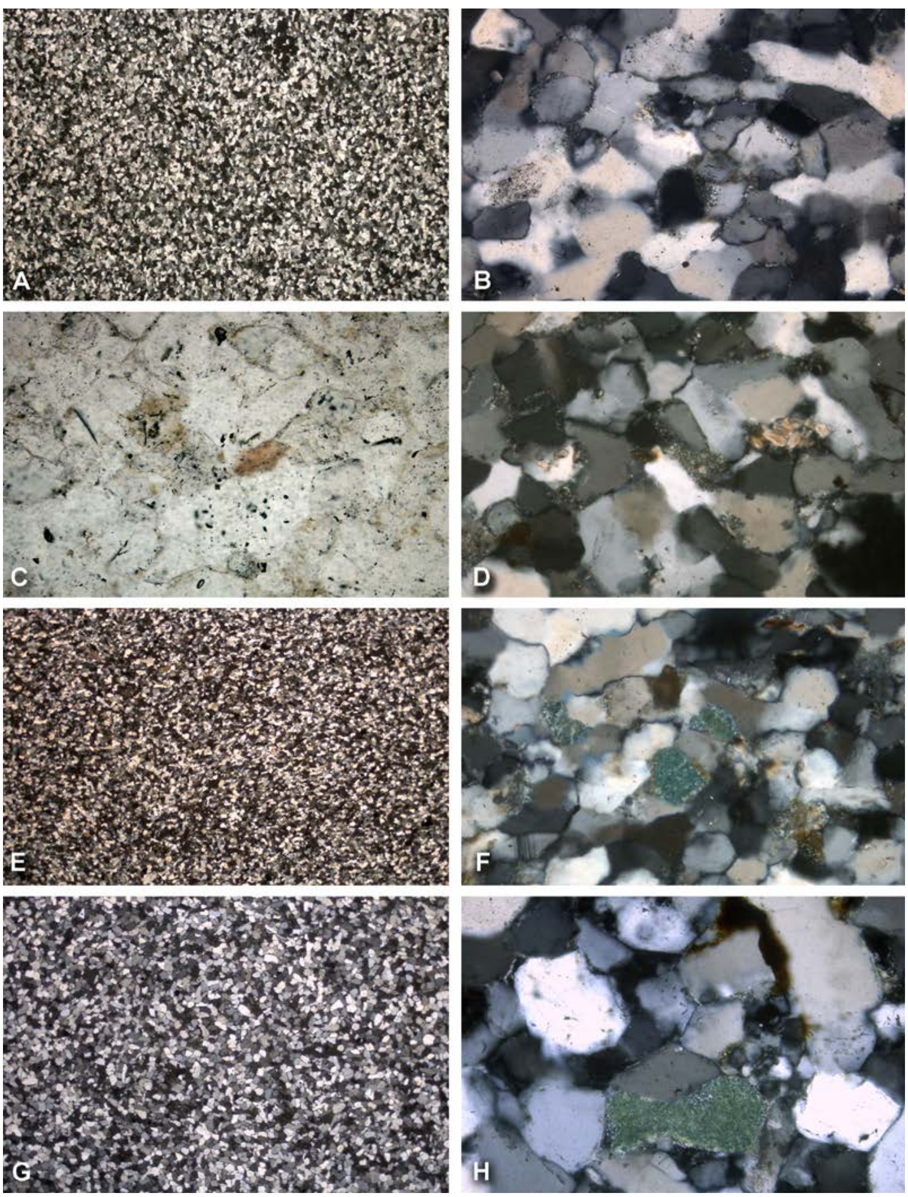


Fig. 6. Thin sections. **A**–**D**, sandstone from erratic boulder at the As Hoved Cliff near Palsgård, Jutland, Denmark (sample PMU 28676). **A**, overview under crossed polarizers, width of image 7.2 mm; **B**, detail with chloritized contact zone, view under crossed polarizers, width of image 0.6 mm; **C**, detail of typical composition with relatively large quartz grains, width of image 0.6 mm; **D**, detail with illitic pore cement, view under crossed polarizers, width of image 0.6 mm.
E–**F**, sandstone representing the typical thick-bedded sandstone of the Lingulid Sandstone Member in the Kvillen section, Halleberg, Sweden (sample PMU 28677). **E**, overview under crossed polarizers, width of image 7.2 mm; **F**, detail with glauconite grains near centre, view under crossed polarizers, width of image 0.6 mm.
G–**H**, quartz arenite from the Lingulid Sandstone Member, Djupadalen, Falbygden, Sweden (sample PMU 28678). **G**, overview under crossed polarizers, width of image 7.2 mm; **H**, detail with large glauconite grain below centre, view under crossed polarizers, width of image 0.6 mm.

best preserved fossil remains come from the larger, massive blocks, in which scattered trilobite remains occur, but unfortunately are often destroyed during splitting of the rock. Internal and external moulds, typical for preservation in sandstone, rarely display the characteristics necessary to determine the species (Table 1). Only cranidia of *Epichalnipsus* and *Berabichia* are preserved fairly completely. Other trilobite remains are fragmentary. Holmiid trilobites were found in the As Hoved boulders only as a genal fragment (Fig. 7A). Due to the prevalence of *Epichalnipsus* in the assemblage, it is here termed the *Epichalnipsus* fauna.

Lingulid brachiopods are eponymous for the Lingulid Sandstone of Sweden and occur as well fairly frequently in the collected erratic boulders (Fig. 7B, C, E–I), but are mostly too poorly preserved to be determinable even to the genus. The only well-known genus and species reported from the Lingulid Sandstone of Sweden is *Glyptias favosa* (Linnarsson 1869), which is also represented in the As Hoved samples (Fig. 7I). A rare and relatively surprising constituent of the fauna is an orthothecid hyolith conch (Fig. 7D) found in unequivocally the same type of erratic boulders at Damsdorf in Germany.

Some of the boulders include large-scale trace fossils with a rather peculiar preservation, probably caused in part as a result of specific weathering processes (Fig. 8). Some specimens appear to record different layers of the original trace as shown in Fig. 8B and 8D. Sören Jensen (personal communication to GG) suggested similarities with *Halopoa*, particularly *H. imbricata* Torell 1870, recorded from lower and middle Cambrian sandstones in Scandinavia, particularly the Mickwitzia Sandstone Member; a genus which has been synonymized with *Palaeophycus* in Jensen (1997).

The fauna described derives from 35 blocks of a collection of 50 units. Of these, 43 units are from As Hoved and seven from other localities in Denmark and northern Germany. All figured and listed specimens in this paper are deposited in the Museum of Evolution, Uppsala University, Sweden (PMU).

Table 1. List of the fossil material collected from the As Hoved locality

Trilobita	Material
Holmiella? sp. A	1 gena
Epichalnipsus anartanus	>100 cranidia, 3 librigenae, 3 thoracic segments
Epichalnipsus sp. A	3 cranidia
Epichalnipsus sp. B	2 cranidia
Berabichia erratica	8 cranidia, 3 tentatively assigned librigenae
Brachiopoda	
Lingulid gen. and sp. A	8 valves
Lingulid gen. and sp. B	2 valves
cf. *Glyptias favosa*	1 ventral valve
Hyolitha	
Orthothecida gen. and sp. incert. A	1 conch

Systematic palaeontology

Superfamily Olenelloidea Walcott 1890

Family Holmiidae Hupé 1953

Genus *Holmiella* Fritz 1972

Type species. *Holmiella preancora* Fritz 1972; by original designation.

***Holmiella*? sp. A**
Fig. 7A

Material. Single gena, PMU 28679, from As Hoved, Palsgård area, Denmark.

Discussion. The single fragmentary gena has a moderately and evenly curved lateral border that proceeds into a long, moderately tapering genal spine with elliptical cross section. The genal field ("ocular platform") is relatively narrow (tr.), weakly convex. The lateral border and border furrow are obsolescent and the posterior librigenal margin has a distinct, large indentation so that the margin defines a broad curvature towards the genal angle.

The specimen is fairly abraded on the cobble so that the recognizable characters do not permit a confident determination, but it matches the morphology known from *Holmiella*. The specimen is tentatively assigned to *Holmiella*.

Superfamily Ellipsocephaloidea Matthew 1887

Family Ellipsocephalidae Matthew 1887

The family Ellipsocephalidae is a notoriously difficult taxon. It includes numerous genera, which are often not only difficult to demarcate from other genera with a similar morphology, but may also pose difficulties in the recognition of species if not based on well-preserved material. Geyer (1990) attempted a reappraisal of the group with strong focus on material from the Moroccan Atlas ranges, but emphasized that a large number of taxa described from Baltica are in need of careful revision.

The difficulty in the study of ellipsocephalid trilobites arises from the coincidence of three particular problems:

Firstly, there is an evolutionary tendency towards a progressive effacement of the cephalic relief that commonly masks important morphological characters. This trend during evolutionary development contributes to the need of well-preserved specimens to permit a confident determination. This trend is also seen during ontogeny of species from the later phase of this development such as *Kingaspioides sanctacrucensis* (Czarnocki 1927) from the Holy Cross Mountains, southern Poland (see Żylińska & Masiak 2007), or *Kingaspidoides frankenwaldensis* (Wurm 1925) from the Franconian Forest, Germany, where the cephalic relief becomes largely effaced in large mature individuals.

Secondly, morphologic trends during the phylogenetic history are decipherable in different lineages so that a strong development of morphological convergence can be recorded. Careful biostratigraphic

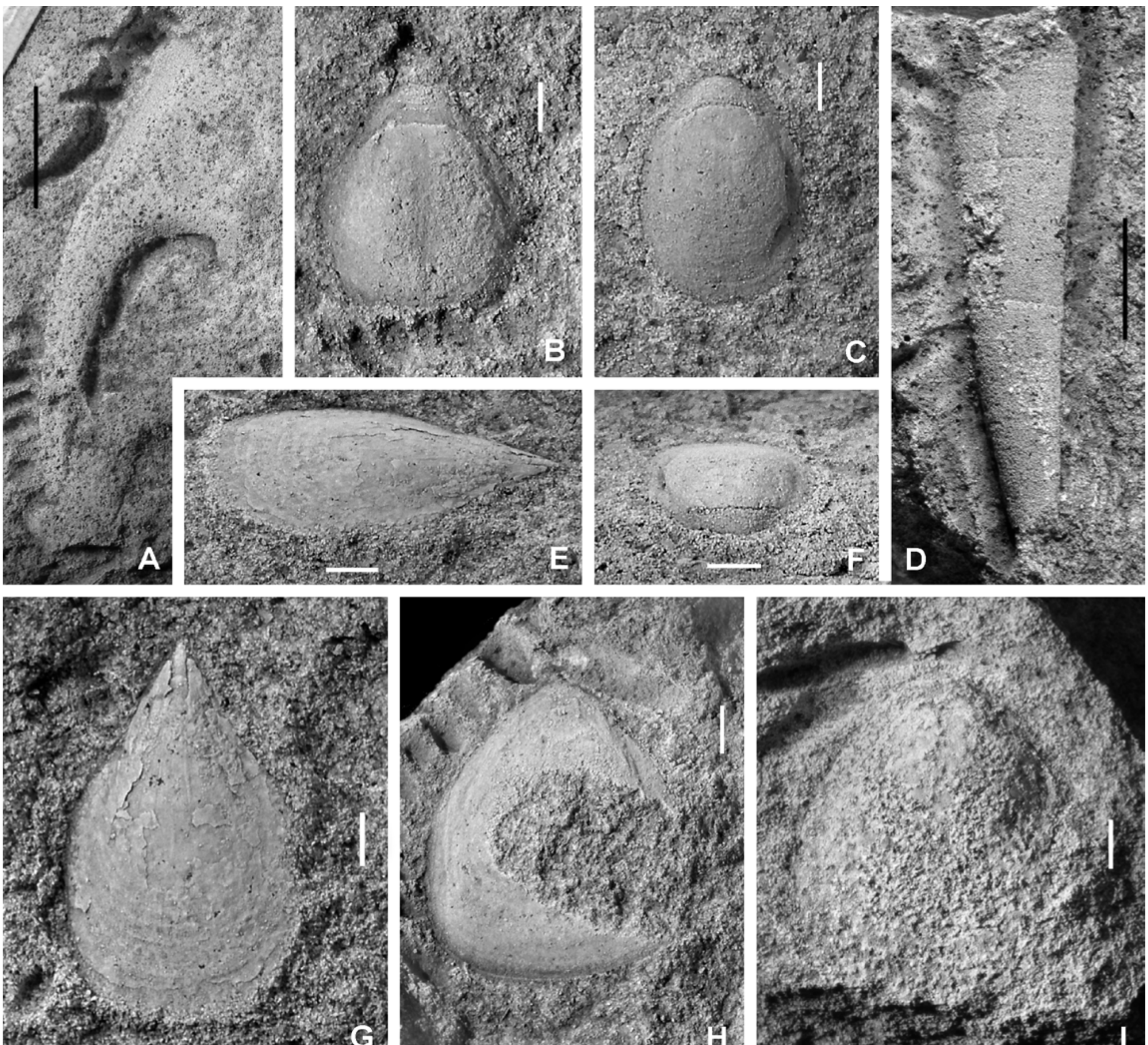


Fig. 7. *Holmiella*? sp. and examples of non-trilobitic body fossils from the erratic boulders at As Hoved, Jutland, Denmark (except for D). **A**, *Holmiella*? sp. A, PMU 28679, incomplete gena **B**, **C**, **E**–**G**, Lingulid gen. and sp. A; **B**, PMU 28680/1, ventral valve; **C**, **F**, 28681, ventral valve, ventral and anterior views; **E**, **G**, PMU 28682/1, dorsal valve, lateral and dorsal views; **D**, Orthothecida gen. and sp. incert., PMU 28683, incomplete conch; from erratic boulder probably from the Lingulid Sandstone collected near Damsdorf, Germany; **H**, Lingulid gen. and sp. B, PMU 28680/2, dorsal valve; **I**, cf. *Glyptias favosa* (Linnarsson 1869), PMU 28684, ventral valve. White scale bars equal 1 mm, black scale bars equal 5 mm.

analysis indicates that these trends partly occur at different phases. This is particularly adverse because biostratigraphically reliable index fossils occur sparsely in the stratigraphic interval in which the ellipsocephalids possessed their acme. This, in turn, either helped to obscure biostratigraphic significance or led to erroneous correlations.

Thirdly, the major stratigraphic interval in which the ellipsocephalids occurred in Earth history coincides with the lower–middle Cambrian boundary interval, which is affected on most Cambrian continents and particularly in their major area in the Acadobaltic faunal realm by sea-level fluctuations. These were responsible not only for an unusually large amount of siliciclastic sediments that were deposited during this interval and which are often unsuitable for favourable preservation of trilobites, but also created gaps that further restrict the amount of available material.

The majority of ellipsocephalid trilobites are known only from cranidia, although it is obvious that pygidia provide important additional information necessary to reveal systematic relations. An example of lumping of species into 'morphotaxa' is the genus *Ellipsocephalus* Zenker 1833, which is used since decades to furnish species with a similar effaced cranidium, regardless of a distinctly differing pygidial morphology as illustrated by *Ellipsocephalus hoffi* Schlotheim 1823, *Germaropyge germari* (Barrande 1852), and *Ellipsostrenua gripi* Kautsky 1945.

It is beyond the scope of this article to discuss the taxonomy of the elllipsocephalid species from Scandinavia. However, it is necessary to emphasize that most generic assignments of the ellipsocephalids from Sweden and Norway are in strong need of a revision. None of the species described from the upper to uppermost lower Cambrian strata of Scandinavia can be assigned correctly to *Ellipsocephalus*, *Proampyx*, and *Ornamentaspis*, although the latter has been suggested by Geyer (1990) to be used as a nickname for the index species *Strenuella linnarssoni* Kiær 1917, to avoid the completely inappropriate assignment to the late middle Cambrian genus *Proampyx* Frech 1897. Recently, Høyberget *et al.* (2015) placed the species in *Ellipsocephalus* although its pygidia differ distinctly from those of typical species of *Ellipsocephalus* such as *E. hoffi* or *E. lejostracus*. Another species dealt with as *Ellipsocephalus lunatus* and used as an index fossil of the *Comluella*?–*Ellipsocephalus lunatus* Zone is cer-

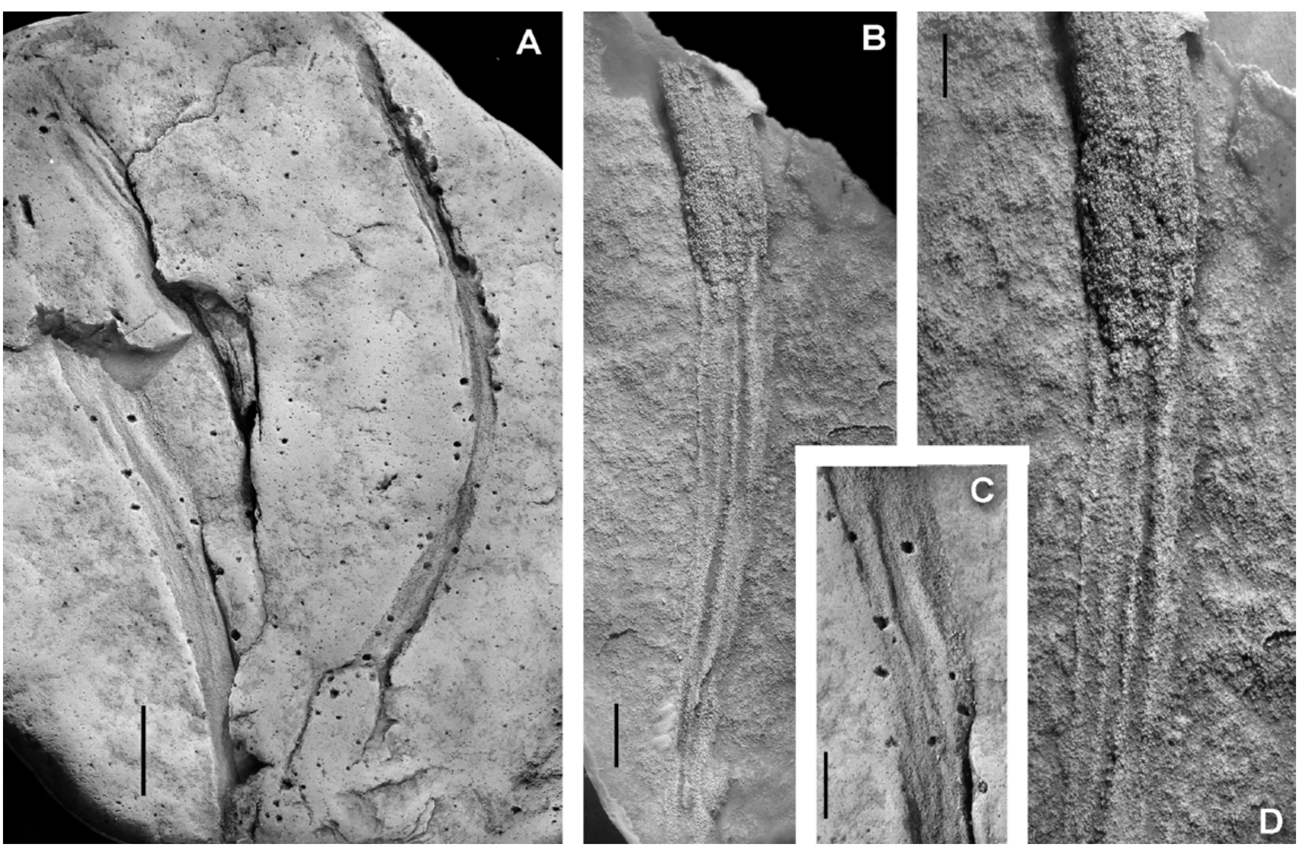
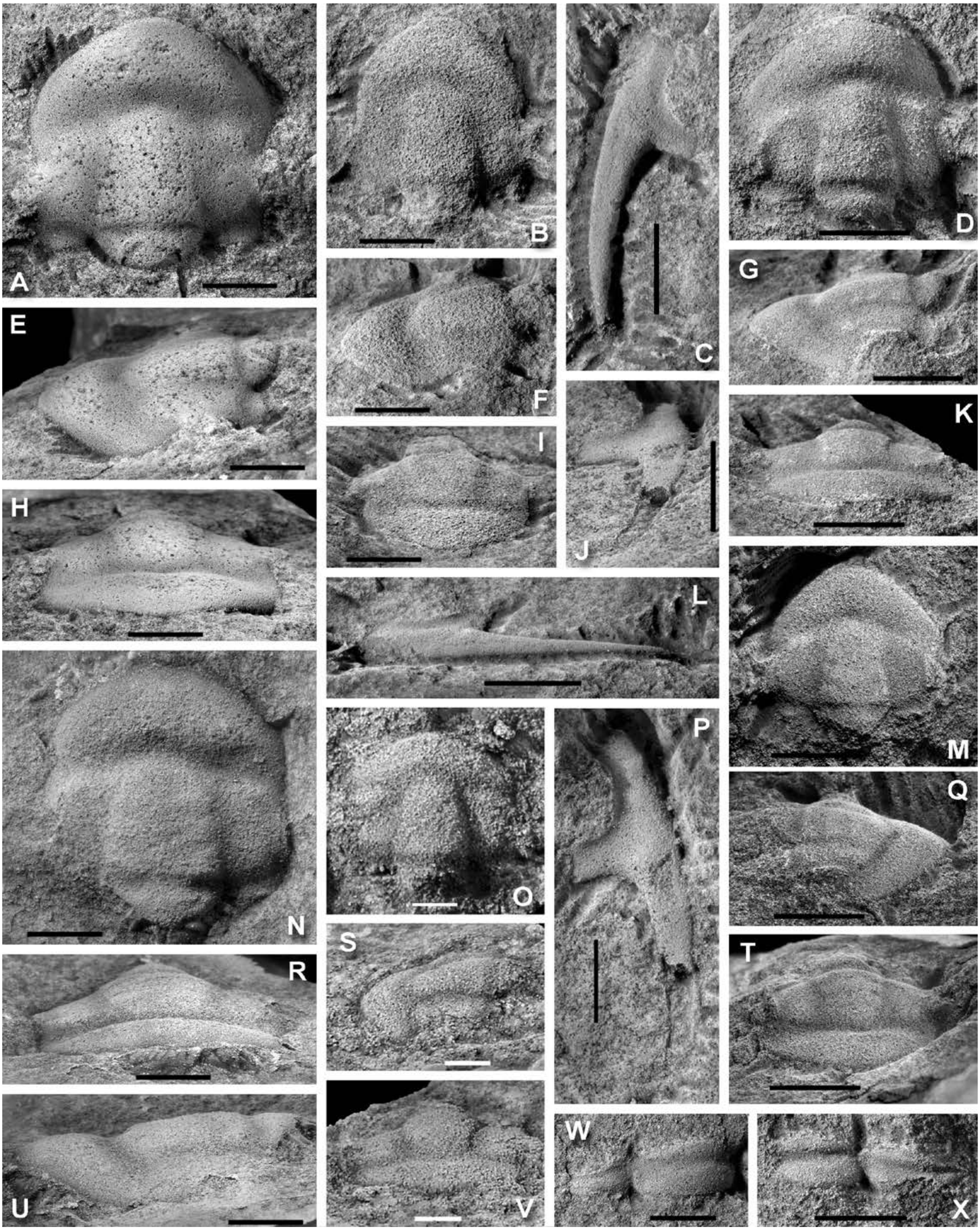


Fig. 8. Trace fossil (cf. *Halopoa imbricata* Torell 1870) from the erratic boulders at As Hoved, Jutland, Denmark. **A**, PMU 28685, several specimens preserved as negative hyporelief; length of figured view *c*. 7 cm; **C**, detail; **B**, **D**, latex cast of PMU 28685, illustrating coating with grains on the wall. All scale bars equal 5 mm.

tainly not a true species of the genus. It is probably more closely related to species of *Kingaspidoides* and (pending future studies) it is termed *"Ellipsocephalus" lunatus* herein, awaiting a careful revision.

The subdivision of the family into the subfamilies Antatlasiinae, Ellipsocephalinae, Protoleninae, and



Strenuellinae appears to be appropriate to date. The subfamily Kingaspidinae is a junior synonym of the Ellipsocephalinae.

Subfamily Antatlasiinae Hupé 1953

Discussion. The Antatlasiinae are the oldest ellipsocephaline trilobites and thus ancestral to all other ellipsocephaloideans. They are unequivocally derived from bigotinid trilobites. Species of *Berabichia* have a relatively general body plan, although the genus *Antatlasia* predates *Berabichia*. A phylogenetic lineage leads towards late early Cambrian genera such as *Sectigena* and *Issafeniella*. The Siberian genera *Chorbusulina* Lazarenko 1962 and *Charaulaspis* Lazarenko 1962 are closely related to *Berabichia* Geyer 1990, although the general similarity in the morphology of the genae is obviously an independent, parallel development in geographically disjunct genera because they can be differentiated by several other characters as discussed in Geyer *et al.* (2004).

Despite of caveats mentioned in Geyer *et al.* (2004), *Strenuaeva* Richter & Richter 1940 is also a genus of the Antatlasiinae. However, species of this genus only include the type species *S. primaeva* (Brøgger 1879) from Norway, *S. inflata* Ahlberg & Bergström 1978 from the Lake Torneträsk area, Sweden (see Cederström *et al.* 2012), *S. baltica* (Wiman 1905) from erratic material of the Gävle Bay, east-central Sweden, *S. nefanda* Geyer 1990 from the western Anti-Atlas of Morocco, and an unnamed species listed as *S. nefanda* from south-eastern Newfoundland (Fletcher 2003, 2006). Other species, previously assigned to *Strenuaeva*, belong to *Issafeniella* Geyer 1990, including *I. orlowinensis* (Samsonowicz 1959) and *I. trifida* (Orłowski 1985), both from the Ociesęki and Kamieniec formations of the Holy Cross Mountains, Poland (Żylińska & Masiak 2007; Żylińska & Szczepanik 2009), and the Spanish species *I. sampelayoi* (Richter & Richter 1940) and *I. melendezi* (Gil Cid 1972).

Genus *Epichalnipsus* Geyer, Popp, Weidner & Förster 2004

Type species. *Epichalnipsus anartanus* Geyer, Popp, Weidner & Förster 2004; by original designation.

Additional species included: *Berabichia inopinata* Geyer 1990, from the lower Cambrian *Sectigena* Zone of Morocco.

Emended diagnosis. Antatlasiine genus with shallow furrows and a generalized overall convexity in transverse section; frontal area convex in sagittal section, distinctly separated from glabella; glabella somewhat raised above shallow dorsal furrows, genae sloping ventrally from dorsal furrows; glabella tapering forward, frontal lobe subacute or with low curvature in dorsal view.

Discussion. *Epichalnipsus* has been introduced by Geyer *et al.* (2004) for species of the *Berabichia* clade that are characterized primarily by a subquadrate cephalon with a well-rounded to faintly subarcuate anterior margin, with a subequal overall convexity in transverse section, and a glabella that is slightly to moderately raised above the genae. The fixigenae themselves are weakly convex in transverse section, without a transverse or diagonal depression, and gently slope abaxially from the axial furrows. The axial furrows are moderately wide and weakly defined from the fixigenae.

As noted by Geyer (1990), the type species of *Berabichia* Geyer 1990, *B. vertumnia*, *B. stenometopa* Geyer 1990, and four of the forms described in open nomenclature are characterized by a cephalon in which the different cranidial regions (such as the glabella, fixigenae, palpebral lobes, and frontal area) have individual convexities so that these different parts are generally well defined by distinct furrows. In turn, *Berabichia inopinata* Geyer 1990 differs in having a cephalon with a rather homogenous overall convexity, at least in a transverse section across the glabella, the fixigenae and the palpebral lobes. This species, provisionally assigned to *Berabichia* in Geyer (1990), has the typical convexities shown by *Epichalnipsus anartanus* and was thus transferred to *Epichalnipsus*.

A specimen of the type species from the erratic boulders was first described by one of the authors (Popp 1999) under the name *Proampyx*? cf. *rotundatus*

◀ **Fig. 9. A**–**V**, **W?**, **X?**, *Epichalnipsus anartanus* Geyer, Popp, Weidner & Förster 2004. **A**, **E**, **H**, PMU 28686, cranidium, occipital ring partly exfoliated, composite mould, dorsal, lateral and anterior views; **B**, **F**, **I**, PMU 28687, cranidium, composite mould, dorsal, lateral and anterior views; **C**, **L**, PMU 28688, librigena, internal mould, dorsal and left lateral views; **D**, **G**, **K**, PMU 28689, incomplete cranidium, internal mould, dorsal, lateral and anterior views; **J**, **P**, PMU 28690, librigena, internal mould, oblique posterior and dorsal views; **M**, **Q**, **T**, PMU 28691, cranidium of small individual, internal mould, dorsal, lateral and anterior views; **N**, **R**, **U**, PMU 28692a, incomplete cranidium, latex cast of external mould, dorsal, anterior and lateral views; **O**, **S**, **V**, PMU 28693/1, incomplete immature cranidium, internal mould, dorsal, lateral and anterior views, scale bars 1 mm; **W**, PMU 28694, fragment of thoracic segment, dorsal view; **X**, PMU 28693/2, partial thoracic segment, dorsal view. All specimens from the As Hoved locality, Jutland, Denmark. Scale bars equal 5 mm except where otherwise noted.

and *Proampyx* cf. *rotundatus*, respectively. The species described under the name *Proampyx rotundatus* (Kiær 1917) bears indeed a great resemblance in respect to proportions, convexity of the preglabellar field and the pattern of lateral glabellar furrows with species of *Epichalnipsus*. However, this species has a tapering glabella and a generally strong convexity in the cranidium. Although the exact convexities of the shell exterior are unknown, it appears to share most characters with *Berabichia* (see below).

Epichalnipsus anartanus **Geyer, Popp, Weidner & Förster 2004**
Fig. 9A–V, W?, X?

v 1999 *Proampyx*? cf. *rotundatus* (Kiær 1917), Popp, p. 3.
v 1999 *Proampyx* cf. *rotundatus* (Kiær 1917), Popp, pp. 4–7, figs. 1–2.
v 2004 *Epichalnipsus anartanus* n. gen., n. sp., Geyer *et al*., pp. 131, 133–134, figs. 3, 4.1, 4.3–4.15.

Material. More than one hundred cranidia or cranidial fragments, mostly incomplete internal moulds, 13 cranidia studied in detail (in repository: PMU 28686, PMU 28687, PMU 28689, PMU 28691, PMU 28692a, b (part and counterpart), PMU 28693/1, PMU 28695/1, PMU 28696), three librigena (in repository: PMU 28688, PMU 28690), three fragments of thoracic segments attributed to the species (in repository: PMU 28694, PMU

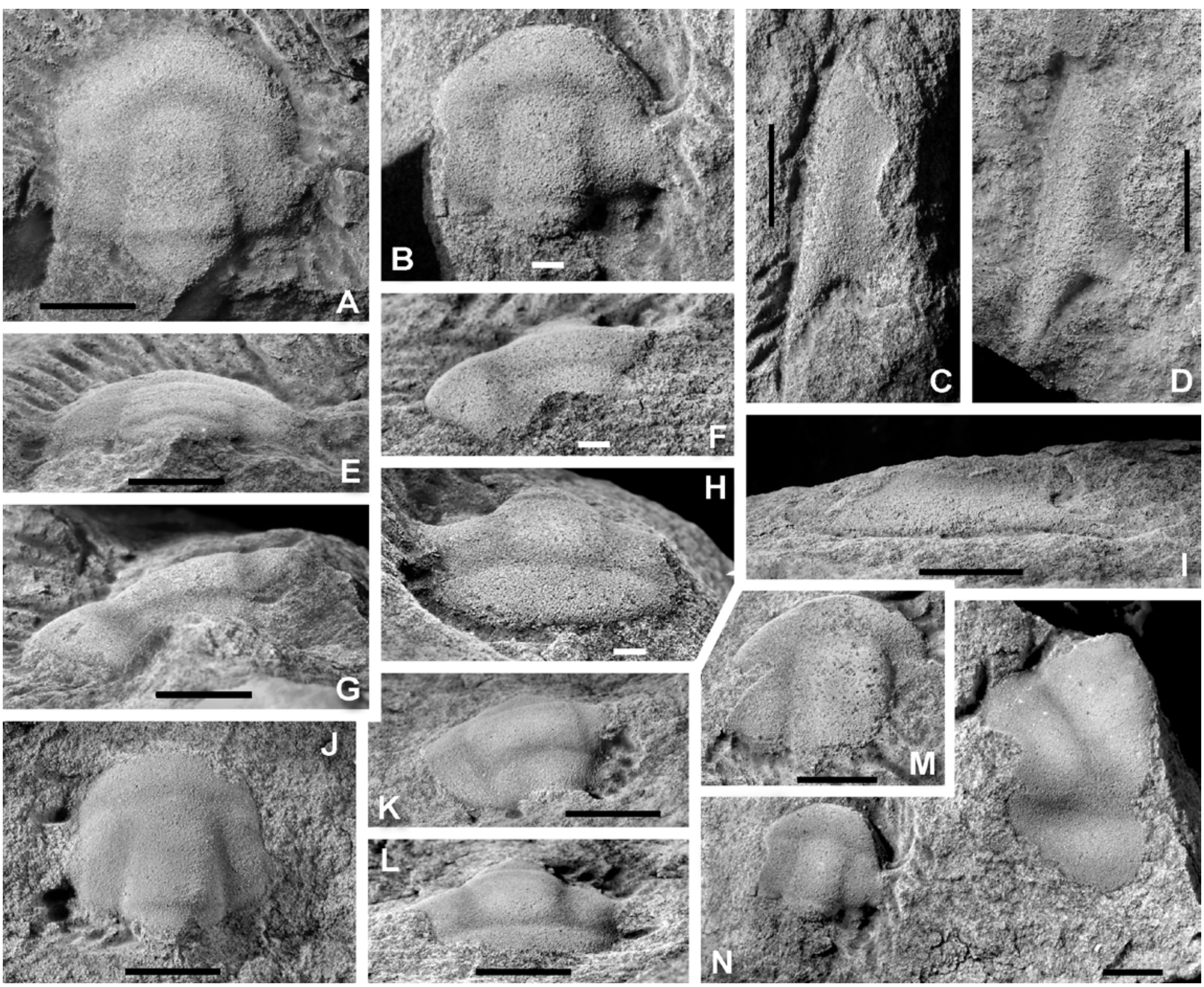


Fig. 10. **A**, **B**, **E**, **F**, **G**, **H**, **N**, *Epichalnipsus* sp. A. **A**, **E**, **G**, PMU 28697, incomplete cranidium, internal mould, dorsal, posterior and lateral views; **B**, **F**, **H**, PMU 28698/1, incomplete cranidium, composite mould, dorsal, lateral and anterior views, scale bar 1 mm; **N**, slab with PMU 28698/1 and PMU 28698/2 illustrating differences in size.
C, **D**, **I**, *Berabichia erratica* Geyer, Popp, Weidner & Förster 2004?; **C**, **I**, PMU 28701, librigena, dorsal and lateral views; **D**, PMU 28702, incomplete librigena, dorsal views.
J–**L**, **M?**, *Epichalnipsus* sp. B; **J**–**L**, PMU 28699, cranidium, dorsal, lateral and anterior views; **M**, PMU 28700, partial cranidium, dorsal views. All specimens from the As Hoved locality, Jutland, Denmark. Scale bars equal 5 mm except where otherwise noted.

28693/2, PMU 28695/2). All specimens from As Hoved, Palsgård area, Denmark, except for PMU 28695/1 and PMU 28695/2 from Damsdorf, Germany.

Emended diagnosis. Species of *Epichalnipsus* with frontal area distinctly inflated that forms a conspicuous platform in lateral view; glabella including occipital ring of *c.* 65–70 % cephalic length; occipital ring with small, subterminal spine; palpebral lobes of about 25 % cephalic length.

Description. Cephalon with strongly curved to well-rounded to faintly subarcuate anterior margin, ranging from 4.0 to 27 mm in the studied specimens. Glabella including occipital ring in adult individuals of 63 to 72 % cephalic length, maximum width across L1 *c.* 45 % maximum cranidial width across centre of palpebral lobes (slightly less in juveniles); tapers forward to anterolateral corners of frontal lobe to *c.* 80 to 85 % width across L1; sides weakly curved. Glabella with three pairs of shallow lateral glabellar furrows, which are fairly well visible on internal moulds, but faint to obsolescent on the shell exterior; S1 moderately long, curved backward; S2 and S3 comparatively shorter, S2 directed slightly backward, S3 almost normal to axis; front terminates in a subacute tip that is usually well visible on internal moulds, but less well developed on the shell exterior. Occipital furrow relatively broad (sag.) and normal to axis in the middle part. Occipital ring sagittally about 14–19 % cephalic length, extreme lateral parts narrow, with moderately long spine in subterminal position.

Fixigenae close to dorsal furrows of about 40 % cephalic length or slightly less, genal width across centre of palpebral lobes *c.* 43 to 49 % maximum glabellar width (across L1); with slightly extended posterior limb. Eye ridge weakly defined on exterior, weakly to moderately well developed on internal moulds, faintly connected with anterolateral corners of glabella; tends to bifid (or perhaps even trifid; Geyer *et al.* 2004, fig. 4.13, also suggested in Fig. 9D, 9M) at dorsal furrows, the anterior thread-like branch of which form an almost obsolete, rarely visible parafrontal line. Palpebral lobes moderately long (exsagittally 22 to 27 % maximum cephalic length in adult individuals), posterior tip located opposite anterior portion of L1 or at about S1, anterior tip opposite L3, faintly oblique to axis.

Preglabellar field long, distinctly convex to slightly inflated, fused medially with anterior border to form an anterior unit of up to more than 30 % cephalic length in adult individuals, distinctly shorter in juveniles. Anterior border narrow, obsolescent, occasionally faintly indicated close to the sutures. Posterior border with straight adaxial part, weakly sigmoidally curved distally and bent forward to palpebral lobes. Posterior border furrow well defined, narrow adaxially, broadening abaxially.

Anterior branches of facial suture diverge from anterior ends of the palpebral lobes to border furrow, with strong inward curvature anteriorly. Posterior branches diverge markedly from posterior ends of palpebral lobes.

Librigena with moderately and evenly curved lateral border that defines a fairly long, moderately tapering genal spine with elliptical cross-section. Librigenal field ('ocular platform') relatively narrow (tr.), weakly convex, grades into distinctly upturned ocular socle. Lateral librigenal border and border furrow obsolescent. Posterior librigenal margin slightly forward directed to narrow curved genal angle.

Rostral plate and hypostome unknown.

Thorax known only from three fragmentary segments assigned to *E. anartanus* (Fig. 9W, X). Axial ring moderately convex in transverse section, without median node or spine, but with faint lateral swelling adjacent to the axial furrow. Transverse furrow shallow to moderately well impressed. Articulating half-ring lenticular in dorsal view, with distinctly curved anterior margin, moderately elevated. Pleurae directed slightly backward, short in comparison to the transverse width of the axial rings. Pleural furrow fairly shallow adaxially, better impressed toward pleural tips, divides pleura into narrow, lenticular anterior strip and a broader transversely triangular posterior section of about double exsag. width near axial furrow; commences at axial furrow close to the anterior margin, directed obliquely to the axis almost straight to the acute pleural tip.

Pygidium unknown.

Discussion. Epichalnipsus anartanus is distinguished from *E. inopinatus* (Geyer 1990) from the *Sectigena* Zone of the Atlas Ranges in Morocco by a strongly inflated frontal area which forms a platform in front of the glabella, whereas the frontal area in the Moroccan species slopes from the glabellar front. Furthermore, *E. anartanus* has slightly shorter palpebral lobes and an occipital ring with a small, subterminal spine rather than a minute node as in *E. inopinatus*.

***Epichalnipsus* sp. A**
Fig. 10A, B, E, F, G, H, N

Material. Three cranidia (PMU 28697, PMU 28698/1, PMU 28698/2). Specimens from As Hoved, Palsgård area, Denmark.

Description. Cephalon with strongly bowed anterior margin. Glabella including occipital ring of slightly less than 70 % cephalic length, maximum width across L1 slightly more than 40 % maximum cranidial width across centre of palpebral lobes; tapering forward or apparently with subparallel sides due to slightly extended anterolateral corners of frontal lobe; sides faintly curved; three pairs of shallow, faint to obsolescent lateral glabellar furrows; S1 moderately long, curved backward; S2 and S3 short and indistinct; frontal lobe with weak curvature anteriorly. Occipital furrow relatively broad (sag.) and normal to axis in the middle part, slightly more distinct laterally. Occipital ring sagittally of 16–18 % cephalic length, reduced to particularly narrow extreme lateral parts, with distinctly curved posterior margin.

Fixigenae close to dorsal furrows of *c*. 40 % cephalic length or slightly less, genal width across centre of palpebral lobes *c*. 60 % maximum glabellar width (across L1). Eye ridge weakly defined. Palpebral lobes moderately long (exsagittally 26 to 33 % maximum cephalic length in the studied specimens), faintly oblique to axis.

Preglabellar field long, distinctly convex, fused medially with anterior border to form an anterior unit of up to 30 % cephalic length. Anterior border obsolescent. Posterior border with straight adaxial part, weakly sigmoidally curved distally and bent forward to palpebral lobes. Posterior border furrow well defined, narrow adaxially, widens abaxially.

Anterior branches of facial suture diverge from anterior ends of the palpebral lobes to border furrow, with strong inward curvature anteriorly. Posterior branches diverge markedly from posterior ends of palpebral lobes.

Librigena, rostral plate, hypostome, thorax and pygidium unknown.

Discussion. The As Hoved material includes cranidia that represent a species different from *Epichanipsus anartanus*, here termed *Epichalnipsus* sp. A. Only three more or less completely preserved and fairly well preserved cranidia have been discovered. The large specimen (Fig. 10A, E, G) has a length of 13.5 mm and testifies an adult individual. The other cranidium (Fig. 10B, F, H) is only 7.5 mm long and is regarded as a late immature stage. Both unequivocally represent the same species although they differ in a few aspects, which all can be attributed to ontogenetic development. These characters include a longer (sag.), more strongly extending occipital ring, a more clearly tapering glabella, and slightly longer palpebral lobes in the large specimen.

Epichalnipsus sp. A is well differentiated from *E. anartanus* in having a strongly curved rather than subarcuate anterior cephalic margin, a less inflated frontal area, broader fixigenae, and the absence of a subterminal occipital spine. Nevertheless, the form shares the typical cephalic morphology of *Epichalnipsus* with a clearly convex, fairly homogenous frontal area separated from the glabella by only a shallow transverse furrow and without a dramatic ventral drop in front of the eye ridges, and exsagittally as well as transversely moderately convex fixigenae, which slope ventrally from the axial furrows. *Epichalnipsus* sp. A is distinguished from *Epichalnipsus* sp. B (described below) by its longer frontal area, broader fixigenae, and a longer (sag.) occipital ring.

***Epichalnipsus* sp. B**
Fig. 10J–L, M?

Material. A single well-preserved cranidium, PMU 28699, and a slightly distorted partial cranidium tentatively assigned to the same form, PMU 28700; both from As Hoved, Palsgård area, Denmark.

Description and discussion. A third species of *Epichalnipsus* from the As Hoved locality is represented by a single well-preserved cranidium and a slightly distorted partial cranidium tentatively assigned to the same taxon. The well preserved cranidium is characterized by a broad (width across L1 slightly less than half cranidial width across palpebral lobes), slightly tapering glabella with a faintly subacute front; a relatively short (sag.) occipital ring of *c*. 15 % cephalic length; relatively narrow and thus clearly rectangular fixigenae; a relatively short (sag.; 24 % cephalic length in the well-preserved specimen) and only moderately convex frontal area; and a subequally curved anterior margin of the cephalon. These characters clearly distinguish it from *E. anartanus* and the form is described as *E.* sp. B. The partial cranidium (Fig. 10M) only shows the anterior two-thirds of the cranidium with the same proportions but apparently a less clearly tapering glabella and a broader front which, however, may be a result of the slight deformation.

Genus *Berabichia* Geyer 1990

Type species. Berabichia vertumnia Geyer 1990; by original designation.

Discussion. The genus *Berabichia* is characterized by a cephalon with the glabella moderately raised over a wavy platform formed by the genae. The fixigenae have their most elevated areas close to the palpebral furrows, but a basal line drawn from the axial furrow to the palpebral furrow dips gently abaxially so that a transverse section is subhorizontal.

The genus has been introduced to encompass three species from the Moroccan Anti-Atlas and High Atlas ranges. Another six forms were attributed by Geyer (1990) to the genus *Berabichia* in open nomenclature. Nevertheless, all come from the upper lower Cambrian

Antatlasia guttapluviae and *Sectigena* zones, upper Banian Stage, suggesting that the Moroccan Atlas ranges depict a major area for the diversification of the *Berabichia* clade.

In addition to the Moroccan species, Geyer (1990) pointed out that *Chorbusulina wilkesi* Palmer & Gatehouse 1972 and *Chorbusulina subdita* Palmer & Gatehouse 1972, both from the uppermost lower Cambrian of the Argentina Range, Antarctica, are also species of *Berabichia*. *Chorbusulina* cf. *subdita*, described by Palmer & Rowell (1995) from the Shackleton Limestone of Antarctica, probably represents another species of the genus. An additional species of *Berabichia* has been described from Sweden under the name *Strenuella primaeva* var. *rotundata* by Kiær (1917) and later dealt with as *Proampyx rotundatus* (e.g., Ahlberg & Bergström 1978). This species, first described from the Gislöv Formation at Forsemölla, Scania, agrees well with the Moroccan species but is known only from imperfectly preserved material (see discussion below under *B. erratica*).

Additional species subsequently assigned to *Berabichia* include *B. milleri* Westrop in Westrop & Landing 2000, *B. eslaensis* Álvaro 2007, *B.*? *kiaeri* (Czarnocki 1927) (by Geyer 1990) and *B. oratrix* (Orłowski 1985) (by Żylińska *et al.* 2013).

Berabichia milleri has been reported from the so-called *Kingaspidoides* cf. *obliquoculatus* Zone of the Long Island Member of the Hanford Brook Formation, New Brunswick, which directly overlies Matthew's classical *Protolenus* (*P.*) *elegans* fauna and Zone (Westrop & Landing 2000). Despite the overall similarity of *B. milleri* in the relative sizes of the different areas of the cranidium, fundamental differences exist between it and the unequivocal species of *Berabichia*. These differences include: (i) the overall convexity, which shows independently inflated genae in *B. milleri* that do not slope ventrally from the axial furrows; (ii) a relatively strongly convex and slender glabella; (iii) subequally developed, simple and slightly backward directed S1 to S3; (iv) transverse, well developed eye ridges that describe a distinct angle to the palpebral lobes; (v) long palpebral lobes that reach backward and almost reach to the posterior border furrow; and (vi) a distinct ventral slope just anterior to the eye ridges. These characters clearly indicate that *B. milleri* is a species of the *Protolenus* clade rather than the *Berabichia* clade. It does not readily fit into an existing genus, but shows affinities to *Latoucheia* (*Latoucheia*) Hupé 1953 and particularly *Cambrunicornia* Geyer 1990. Both *Latoucheia* (*Latoucheia*) and the variably developed species of *Cambrunicornia*, the latter known from the Anti-Atlas of Morocco (Geyer 1990), the Tröbitz Formation of Saxony, Germany (Geyer *et al.* 2014), and species tentatively assigned to the genus from the *Redlichops* faunule of the Dead Sea area, Jordan (Elicki & Geyer 2013) and from the Láncara Formation of the Cantabrian Mountains, northern Spain (Álvaro 2007), differ in several aspects, particularly in the absence of a distinctly inflated anterior area, but appear to be closely related. Both genera, however, occur in strata equivalent to the *Protolenus* Zone of New Brunswick or even slightly younger strata so that their stratigraphic occurrence would match with the occurrence of *B. milleri* in the Hanford Brook fauna.

Berabichia eslaensis Álvaro 2007 has been described from a fauna that includes a species assigned to *Kingaspis* from the Láncara Formation of the Cantabrian Mountains, northern Spain. Although the species strongly resembles some of the species of *Berabichia* from Morocco, the overall convexity of the frontal area and the manner in which the palpebral lobes are connected with the fixigenae suggest a close relationship with the *Kingaspis* clade rather than *Berabichia*.

Strenuaeva? *kiaeri* Samsonowicz 1959 from the *Protolenus–Issafeniella* Zone of the Holy Cross Mountains of Poland has been transferred to *Ellipsocephalus* by Orłowski (1985) and then tentatively assigned to *Berabichia* by Geyer (1990). Żylińska *et al.* (2013) emphasized that the only known specimen of this species is strongly effaced, shows shallow lateral and axial furrows, and it may in fact represent a distorted specimen of *Issafeniella orlowinensis* (Samsonowicz 1959) that occurs in the same stratigraphic interval of the Holy Cross Mountains (Żylińska & Masiak 2007).

Comluella oratrix Orłowski 1985 and *Comluella igrzycznae* Orłowski 1985, both described from the *Holmia–Schmidtiellus* Assemblage Zone of the Holy Cross Mountains, have been subject to a recent morphometric analysis by Żylińska *et al.* (2013), who showed that the two species constitute a monospecific assemblage representing a single species dealt with as *Berabichia oratrix* (Orłowski 1985). This species indeed fits very well into the concept of the genus although the relatively strong deformation of the material makes a reliable reconstruction of the species' original morphology difficult.

***Berabichia erratica* Geyer, Popp, Weidner & Förster 2004**

Fig. 10C?, 10D?, 10I?; Fig. 11A–M

v 2004 *Berabichia erratica* n. sp., Geyer *et al*., pp. 134–135, figs. 5.1, 5.2, 5.4.

Material. Eight cranidia (reposited: PMU 28682/2, PMU 28703 (part and counterpart), PMU 28704–PMU 28707). All specimens from As Hoved, Palsgård area, Denmark. Tentatively assigned to *B. erratica*: three librigenae (reposited: PMU 28701, PMU 28702).

Emended diagnosis. Species of *Berabichia* with frontal area moderately inflated, anterior margin tends to be subarcuate; glabella gradually tapering forward; fixigenae with highest elevation abaxially; palpebral furrow obsolete, merely a shallow and wide depression; palpebral lobes of about one-fourth cephalic length.

Description. Cephalon with faintly subarcuate anterior margin. Dorsal furrows moderately wide, poorly defined from fixigenae, grading into weakly to moderately convex fixigenal area.

Glabella including occipital ring of 72 to 76 % cephalic length, tapers gradually forward, frontal lobe 66 to 72 % maximum width across L1; front in dorsal view with low triangular tip or shallow curvature. Three pairs of lateral glabellar furrows very faint to obsolescent on the shell exterior; S1 moderately long, distinctly curved backward, relatively broad; S2 and S3 shorter, S2 backward directed. Occipital furrow broad (sag.) medially, shallower laterally. Occipital ring sag. variably of 15 to 22 % cephalic length, fades to narrow lateral portions, posterior margin thus

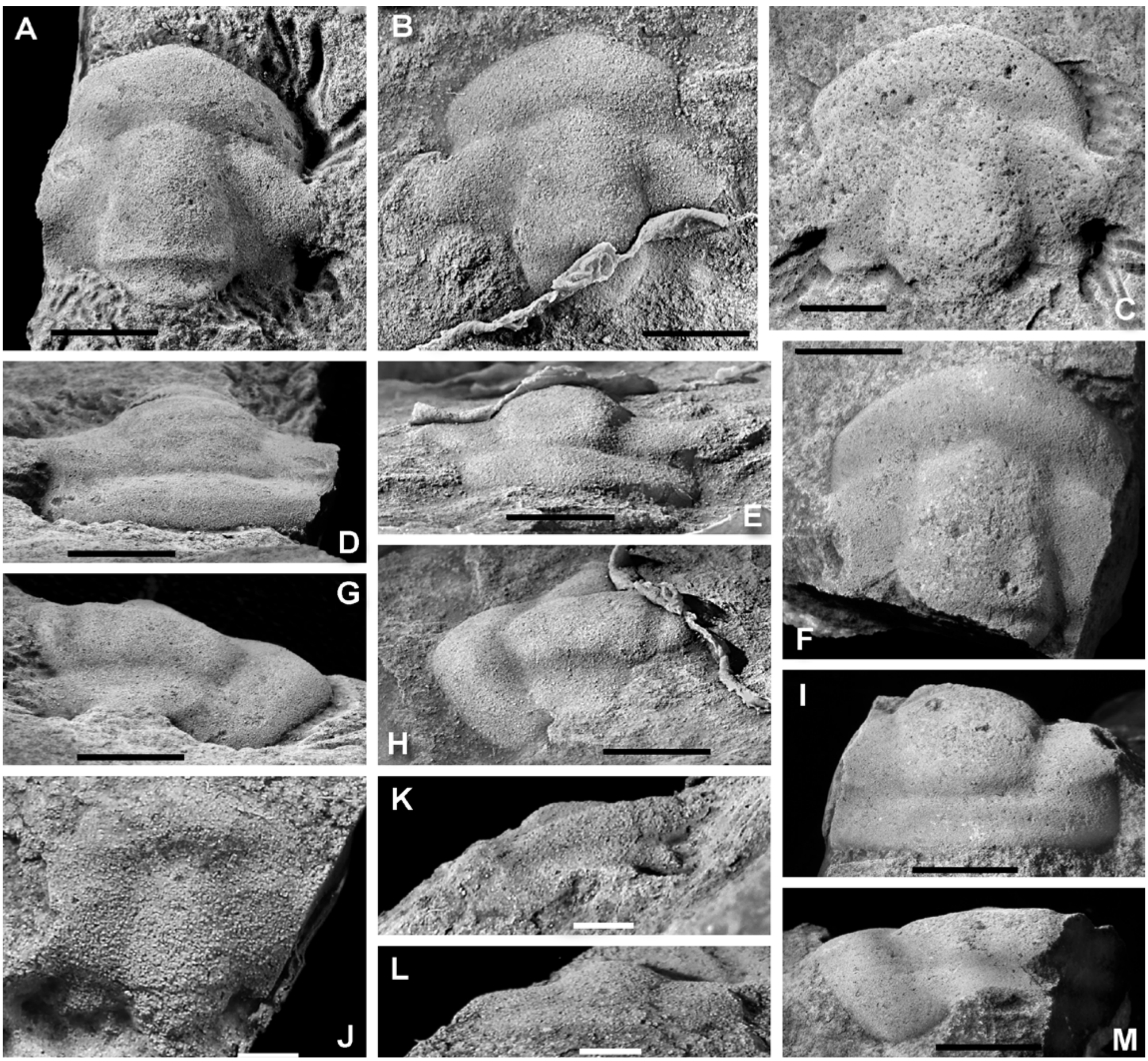


Fig. 11. *Berabichia erratica* Geyer, Popp, Weidner & Förster 2004. **A**, **D**, **G**, PMU 28703a, incomplete cranidium, internal mould, dorsal, anterior and lateral views; **B**, **E**, **H**, PMU 28682/2, incomplete cranidium, latex casts of external mould, dorsal, anterior and lateral views; **C**, PMU 28704, cranidium, internal mould, weathered and partly abraded surface, dorsal view; **F**, **I**, **M**, PMU 28705, partial cranidium, internal mould, dorsal, anterior and lateral views; **J**–**L**, PMU 28706, immature cranidium, dorsal, lateral and anterior views; scale bar 1 mm. All specimens from the As Hoved locality, Jutland, Denmark. Scale bars equal 5 mm except where otherwise noted.

strongly curved in dorsal view, probably with small node (not clearly preserved in any of the studied specimens).

Fixigenae close to dorsal furrows of less than *c.* 40 % cephalic length, width across palpebral lobes 40 to 45 % maximum glabellar width across L1, with slightly extended posterior limb; most elevated point close to palpebral furrows. Eye ridge weakly defined on the shell exterior, faintly connected with indistinct anterolateral corners of glabella. Palpebral lobes moderately long, posterior tips opposite L1, anterior tips opposite S2 or slightly anterior, exsagittally 26 to 31 % of cephalic length, slightly oblique to axis. Palpebral furrows shallow to obsolescent, developed as indistinct depressions or merely a change of convexity.

Preglabellar field moderately long, more or less fused with anterior border to form a slightly inflated frontal area of slightly more than 22 to 28 % cephalic length. Anterior border relatively narrow, defined by shallow and indistinct border furrow laterally. Posterior border weakly sigmoidally curved and with a clear forward bent posterior to palpebral lobes. Posterior border furrow a narrow depression adaxially, widens abaxially, with forward curvature distally to reach posterior tips of palpebral lobes.

Rostral plate, hypostome, thorax and pygidium unknown.

Three librigenae from the As Hoved material are tentatively assigned to the species. Librigena moderately broad (tr.), with a moderately long genal spine of slightly less than half length of the anterior part. Genal spine with a broad base, tapers rapidly backward. Lateral margin with low curvature, progressing with a faint flaw into the genal spine, which has a nearly straight abaxial margin. Posterior margin very shallow, with a shallow but notable indentation towards the genal spine. Librigenal lateral and posterior border and border furrow obsolescent. Librigenal field ('ocular platform') moderately broad (tr.), weakly convex, grades into distinctly upturned ocular socle.

Discussion. The description of *Berabichia erratica* was based on only two well preserved cranidia from a gravel pit at Wilsche near Gifhorn, northern Germany. Additional material discovered at the As Hoved locality slightly complements the information on the species as described above. Particularly noteworthy is that the figured specimens from Wilsche are slightly smaller than the average specimens from As Hoved and have a slightly more slender glabella which is attributable to allometric growth.

As discussed by Geyer *et al.* (2004), *Berabichia erratica* falls within the morphological range of *Berabichia*, although the species appears to differ from all other species of the genus in showing a tendency to develop a subarcuate anterior margin of the cephalon due to a slightly narrower curvature medially. This feature is known, however, from a few specimens of *Berabichia vertumnia* Geyer 1990 as well, the type species of the genus. The palpebral lobes of *B. erratica* are defined from the fixigenae by only a shallow and wide depression, which is known also from *B. vertumnia*. Further differences between *B. erratica* and *B. vertumnia* are discussed in Geyer *et al.* (2004). Particularly noteworthy is that the abaxial part of the fixigenae is developed in *B. erratica* as a rather flat area so that it indicates some similarity to the striking depression that crosses the fixigenae in *Sectigena* and can be used for generic distinction.

The most similar species formally described to date are *Berabichia oratrix* (Orłowski 1985) (see discussion under genus and Żylińska *et al.* 2013) and *B. subdita* (Palmer & Gatehouse 1972). *Berabichia oratrix* differs in having broader fixigenae and slightly longer palpebral lobes. However, specimens of *B. oratrix* may develop a faint subarcuate anterior cephalic margin as well (see Żylińska *et al.* 2013, fig. 6B, 6D, 6E).

Berabichia rotundata (Kiær 1917) is also fairly similar to *B. erratica*. It differs, among other characters, by a sagittally shorter frontal area and a longer occipital ring (see Ahlberg & Bergström 1978, pl. 3, figs. 1, 2).

Stratigraphy

The biostratigraphy of the Scandinavian lower Cambrian (Cambrian Series 2) (Fig. 2) is surprisingly poorly constrained. The major reason is the fairly humble fossil record from scattered localities due to the prevalence of shallow marine, siliciclastic deposits; a dominance of endemic trilobite species and a high amount of endemic genera; and a preliminary stage of generic concepts for the ellipsocephalid trilobites and a preliminary stage of the taxonomic treatment of brachiopods.

A more or less well agreed biostratigraphic framework has been established for the entire Baltica continent since the monographic correlation chart (Mens *et al.* 1990). However, almost none of the recognized trilobite zones could be correlated into any other Cambrian continent with satisfactory precision. The recognition of distinct acritarch zones apparently helped to correlate the trilobite zones into distant regions, but correlations entirely based on acritarch assemblages led to surprising results, which partly contradicted correlations based on trilobites as indicated in Landing *et al.* (2013) and Sundberg *et al.* (submitted).

A profound reappraisal of the biostratigraphy of Baltica backed by sequence stratigraphic criteria led to a partial revision of the established biostratigraphic scheme (Nielsen & Schovsbo 2011). Particularly noteworthy is the abandoning of the Holmia inusitata Zone and the partial amalgamation of the *Holmia kjerulfi* group Zone with the '*Ornamentaspis*' *linnarssoni* Zone. This in fact leads to the assumption that a tremendously shorter time interval is represented by the *Holmia* and '*O.*' *linnarssoni* ranges than previously assumed. Nielsen & Schovbo's (2011) analyses prove that the profoundly incomplete depositional record in many parts of Scandinavia hampers the construction of a sufficiently detailed biostratigraphic framework. They also show that the acritarch zones may be controlled in part by facies. Correlations based on acritarch assemblages are a helpful tool, but they must be used with caution. Nevertheless, the revised zonal scheme with a *Holmia kjerulfi*–'*Ornamentaspis*' *linnarssoni* and a *Comluella*?–*Ellipsocephalus lunatus* Zone sensu Nielsen & Schovsbo (2011) does not allow a direct recognition of the biostratigraphic position for most of the relevant recorded Scandinavian faunas and awaits further revisions of the faunal assemblages and a revision of the ellipsocephaloid trilobites.

The *Epichalnipsus* Fauna from the glacial erratic boulders described herein illustrates the problem of the suggested zones. As shown above, there is little doubt that the boulders from As Hoved, Jutland, are derived from the Lingulid Sandstone Member of the Västergötland region so that the *Epichalnipsus* Fauna portrays a fauna typical for Nielsen & Schovsbo's (2011) LC2-4 sequence, which the authors place in the suggested *Comluella*?–*Ellipsocephalus lunatus* Zone. However, the eponymous index fossils are scarce and almost nowhere distinctive for the fauna in which they are found. In addition, indirect correlation suggests the trilobites of the *Epichalnipsus* Fauna to be slightly older than the assemblages with '*Ellipsocephalus*' *lunatus* due to reasons detailed in the following paragraphs. However, the occurrence of '*E.*' *lunatus* in the assemblages does not appear to precisely determine both the lower boundary of the zone, nor can it be proved that the base of Nielsen & Schovsbo's (2011) LC2-4 sequence coincides with the base of their *Comluella*?–*Ellipsocephalus lunatus* Zone. Therefore it is suggested that the *Epichalnipsus* Fauna is a representative of the boundary interval of the *Holmia kjerulfi*–'*Ornamentaspis*' *linnarssoni* to *Comluella*?–*Ellipsocephalus lunatus* zones although the fauna is not known with certainty from outcrops in Scandinavia.

Key areas for correlation from Scandinavia and Baltica into other Cambrian continents and for biostratigraphic resolution are the Holy Cross Mountains, southern Poland, and subsurface occurrences in central and northern Poland although the stratigraphic sequences from these regions are as well incompletely known. Recent investigations and taxonomic revisions of the lower and middle Cambrian Holy Cross Mountains faunas (e.g., Żylińska 2013a, 2013b; Żylińska & Masiak 2007; Żylińska & Szczepanik 2009; Żylińska *et al.* 2013) indicate different faunal assemblages and a possibility to subdivide the ellipsocephaloid-bearing pre-*Paradoxides* strata. Nevertheless, the significance of the crucial *Holmia*–*Schmidtiellus* and *Protolenus*–*Issafeniella* faunas from the Holy Cross Mountains also suffer from poor outcrops and unresolved taxonomic problems of the generic concepts.

Nielsen & Schovbo's (2011) sequence stratigraphical analyses help to identify the sea-level fluctuations that indicate an extremely varying termination of the lower Cambrian sequences which is partly a result of the late early Cambrian regressive–transgressive events (commonly termed the 'Hawke Bay regression'; Palmer & James 1979) and partly a result of Neogene and Pleistocene erosion that removed parts of the Cambrian strata. Thus, the young to youngest early Cambrian faunas from various areas of Scandinavia are not necessarily of similar age, and the most diverse of these faunas offer a glimpse into characteristic biota. A few of these faunas merit a brief characterisation.

Tømten Member fauna

The traditional fauna of the 'Holmia shale' (now Tømten Member of the Ringstrand Formation; Nielsen & Schovsbo 2006) of the Mjøsa region, Norway (Kiær 1917) includes the trilobites *Holmia kjerulfi*, *Kjerulfia lata*, cf. *Runcinodiscus index*, *Strenuaeva primaeva*, and *Ellipsostrenua* cf. *gripi* (Kiær 1917; Skjeseth 1963; Ahlberg & Bergström 1978; Bergström 1981; Høyberget *et al*. 2015). The overlying Evjevik Member includes two major limestone beds in the lower and upper part, respectively. Both appear to bear a fairly rich fauna of the '*Ornamentaspis*' *linnarssoni* assemblage (earlier termed the *Strenuella* limestone) with trilobites such as *Calodiscus lobatus* (Münster 1900; Nikolaisen 1986; Cederström *et al*. 2012; Høyberget *et al*. 2015). Most determinable material has been collected from weathered loose blocks so that apparently no biostratigraphical distinction between them had been achieved, but Høyberget *et al*. (2011a, 2011b, 2015) showed that a similar fauna with '*O.*' *linnarssoni*, *E. gripi* and helcionelloid molluscs occurs in both limestone levels. In addition, the upper limestone bed yielded '*Ornamentaspis*' *sularpensis*, *Strenuaeva spinosa* and *Calodiscus lobatus*, but it remains doubtful whether these indicate a biostratigraphical change in the trilobite spectrum. More significant seems to be the fact that *Holmia kjerulfi* and *Kjerulfia lata* occur in the lower limestone bed and *H. kjerulfi* is found in the shales between the limestone beds but is not found any higher in the succession of the Ringstrand Formation (Nikolaisen 1986; Høyberget *et al.* 2011a, 2011b, 2015). However, it should be emphasized that Høyberget *et al.* (2011b, 2015) report the helcionelloid *Helcionella antiqua* (Kiær 1917) from the lower limestone bed, which is also known from strata with *'O.' sularpensis* of the middle part of the Gislöv Formation of Scania (see below). The only trilobite identified from the overlying Skyberg Member in Norway is *'Ellipsocephalus'* cf. *lunatus* indicative of the *Comluella*?–*Ellipsocephalus lunatus* Zone (Høyberget *et al.* 2011b, 2015).

Gislöv Formation faunas

A detailed study of the Gislöv Formation of eastern Scania (Bergström & Ahlberg 1981) appears to indicate (despite the strong condensation) a relatively continuous sequence to be present, and the ranges of trilobites portray a slight change in faunal composition from the middle to the uppermost part of the formation. A lower trilobite-bearing assemblage is recognizable in a shaly interval of the middle part of the formation. It bears, according to Bergström & Ahlberg (1981), *Holmia*? sp., *'Ornamentaspis' grandis*, *'O.' sularpensis*, *Berabichia*? *rotundata*, *Kingaspidoides*? cf. *nordenskioeldi* and an obolellid brachiopod assigned to *Magnicanalis* sp. (generic assignment partly revised herein). Nathorst's (1877) report of fossils from the formation at Forsemölla lists *Holmia sulcata*, *Kingaspidoides*? *nordenskioeldi* and *Lingulella*? *nathorsti* as additional species from probably this interval of the Gislöv Formation. Additional faunal elements from this interval were reported by Troedsson (1917) from the Hardeberga quarry. These include *Calodicus* cf. *lobatus*, which was firmly assigned to *Calodiscus lobatus* in Cederström *et al.* (2009). The latter authors confirmed a range of *C. lobatus* from the upper part of the *H. kjerulfia* assemblage into the lowest part of the *'O.' linnarssoni* range zone.

The presence of holmiids and the synchronous absence of *'Ornamentaspis' linnarssoni* prompted the interpretation of these strata and the interval to the *Holmia kjerulfi* group Zone. As discussed by Nielsen & Schovsbo (2011), the occurrence of *Holmia* and holmiid trilobites in general is, however, not restricted to the eponymous *Holmia* 'zones'. The presence of typical late to latest early Cambrian ellipsocephalid trilobites, in contrast, indicates a distinct faunal turnover so that the fauna represented in the middle part of the Gislöv Formation indeed signifies a new, recognizable zone.

A higher, but not the highest, part of the Gislöv Formation (generally developed as light coloured limestone) is characterized by a low diverse fauna with *'Ornamentaspis'* cf. *linnarssoni*, an obolellid assigned to *Magnicanalis* sp. and additional brachiopods. This part has been regarded as the lower part of the *'Ornamentaspis' linnarssoni* Zone, but its faunas are insufficient to distinguish it biostratigraphically from the underlying strata.

The uppermost part of the formation, generally developed as a light or dark limestone with phosphoritic nodules, bears a fauna with *'Comluella' scanica*, *'Ellipsocephalus' lunatus*, *'Strenuaeva'* sp., *Amphigeisina danica*, *Magnicanalis* sp., and *Hyolithellus* cf. *micans* (Bergström & Ahlberg 1981; Cederström *et al.* 2009, 2012).

Remarkably, the well-known *'Ornamentaspis' linnarssoni* fauna from the dark grey shales at Forsemölla includes additional trilobites such as a new species of *Strenuaeva* and a species of *Protolenus s.l.* that will be described elsewhere. They appear to indicate a distinctly younger age of the fauna than previously assumed, probably related to the earliest *Protolenus s.l.* bearing strata in Avalonia and West Gondwana (Geyer, unpublished data). A similar development with *Holmia* spp., '*Strenuella*' and *Berabichia* occurs in the *Holmia*–*Schmidtiellus* Zone of the lower parts of the Ociesęki and Kamieniec Formations and is replaced by species of *Kingaspidoides* and *Issafeniella* in the shallower marine facies of the Ociesęki Formation and *Protolenus* and *Issafeniella* in the deeper marine facies of the Kamieniec Formation. These are followed by a typical middle Cambrian fauna with species of *Paradoxides s.l.* and ellipsocephalines (Żylińska 2013b).

Torneträsk Formation faunas

A fauna from the Torneträsk Formation in the Luobákti section near Lake Torneträsk, northern Sweden, assigned to the *Ornamentaspis*? *linnarssoni* Assemblage Zone includes *Geyerorodes*? *lapponicus*, *Strenuaeva inflata*, *Neocobboldia* aff. *dentata* and *Chelediscus acifer* (Axheimer *et al.* 2007). An earlier report of an assemblage from the Torneträsk Formation by Ahlberg (1980) reported *Strenuaeva inflata* together with *S. triangularis*, *Ellipsostrenua* cf. *gripi*, and *Geyerorodes*? *lapponicus*. *Geyerorodes*, first introduced as *Orodes* Geyer 1990 and replaced because of homonymy (Özdikmen 2009), is a genus indicative of late sub-*Paradoxides* strata in Avalonia (Fletcher 2006) or latest sub-*Paradoxides* to earliest *Paradoxides*-bearing strata in Moroccan West Gondwana (Geyer 1990; Geyer *et al.* 1995). The specimen illustrated by Axheimer *et al.* (2007, fig. 4r) has been re-examined by P. Cederström and appears to be a deformed cranidium of *G.*? *lapponicus* (P. Ahlberg, personal communication 2015). Although a confident generic assignment is difficult, it is tentatively assigned to *Cambrunicornia*, a genus known from the lower–middle Cambrian boundary interval in West Gondwana, where it ranges from the *Hupeolenus* Zone to the *Morocconus notabilis* Zone of the Moroccan biostratigraphical zonation in the lowermost Agdzian Stage (see discussions in Geyer *et al.* 2014).

Grammajukku Formation faunas

The traditional *Ellipsostrenua gripi* fauna of the upper part of the Grammajukku Formation from a ravine at Mount Aistjakk (now commonly spelled Assjatj) in the Laisvall region (Kautsky 1945) includes the trilobites *Holmia*? *ljungneri* and *Ellipsostrenua gripi*, and a bradoriid, the helcionelloid mollusc *Helcionella antiqua*, and three species of linguliform brachiopods. Two additional faunas slightly below Kautsky's original horizon have been reported by Ahlberg (1984a), Cederström in Nielsen & Schovsbo (2011, p. 222) and Cederström *et al*. (2012) from the Mt. Assjatj and the nearby Delliknäs sections. According to Cederberg *et al*. (2012) these include also *Holmia*? *ljungneri* and *Ellipsostrenua gripi* as in the upper faunal horizon, and a fairly diverse assemblage comprising *'Ornamentaspis' linnarssoni*, *Ellipsostrenua gripi*, *Holmia* cf. *inusitata*, *Kjerulfia palpebra*, *Strenuaeva spinosa*, and *S*.? *kullingi*.

The fossil assemblages portrayed above accumulate to a consistent succession of trilobites without detectable biostratigraphic gaps. On the other hand, the similarities with faunas of the Holy Cross Mountains, Poland, and the Atlas ranges, Morocco, appear to indicate that the *Epichalnipsus* fauna cannot be younger than the younger part of the traditional *'Ornamentaspis' linnarssoni* Zone, or Nielsen & Schovsbo's *Comluella*?–*'Ellipsocephalus' lunatus* Zone. Nielsen & Schovbo's (2011) assignment of the member to the LC2-4 sequence appears to be accurate. Accordingly, the endemic character of the *Epichalnipsus* fauna must be regarded as result of biofacies differentiation due to depositional settings of the well washed and reworked, shallow sand accumulations.

Intercontinental and global correlation

A precise, direct correlation of the *Epichalnipsus* fauna with the uppermost part of the *Holmia kjerulfi*–*'Ornamentaspis' linnarssoni* Zone (lower part of the former *'Ornamentaspis' linnarssoni* Zone) or lowermost *Comluella*?–*Ellipsocephalus lunatus* Zone (upper part of the former *'Ornamentaspis' linnarssoni* Zone) into regions outside Scandinavia is almost impossible. As demonstrated by Sundberg *et al.* (submitted), who provide a detailed correlation chart for the Cambrian Series 2–Series 3 boundary interval, a correlation of the late early Cambrian trilobites known from Scandinavia is slightly facilitated by a comparison with faunas from the Holy Cross Mountains, Poland, which provide links to faunas from West Gondwana (particularly Morocco) and western Avalonia (southeastern Newfoundland). The *Epichalnipsus* fauna correlates with part of the *Protolenus*–*Issafeniella* faunal assemblage (Żylińska & Szczepanik 2009) and probably with the lowermost Agdzian Stage of the Moroccan Atlas ranges (*Hupeolenus* Zone; see Geyer & Landing 2004).

An equivalent of the *Epichalnipsus* level in Avalonian Newfoundland is probably Fletcher's (2006) *Orodes howleyi* Zone in the upper Branchian Stage. All other correlation must be derived from indirect correlations of these stratigraphic intervals, e.g. with the *Acimetopus bilobatus* Zone of the upper Dyeran in the Taconic Allochthon of eastern Laurentia and coeval strata in the upper, but not uppermost *Olenellus* 'Zone' of western Laurentia; the mid-Toyonian of the Siberian Platform; and the mid-Duyunian *Arthricocephalus chauveaui*–*Changaspis elongata* Zone of the South China/Yangtze Platform.

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Appendix 1. Petrological characters of the three samples from the studied erratic boulders from the As Hoved cliff, Jutland, and the Lingulid Sandstone Member of Kvillen, Halleberg and Djupadalen, Falbygden

	AS HOVED	KVILLEN	DJUPADALEN
Grain size	Fine	Fine	Coarser than the As Hoved and Kvillen samples
Compaction	High	High	Low
Grain contact	Poorly visible due to general absence of Fe-oxide coats.	Poorly visible due to general absence of Fe-oxide coats.	Irregularly stepped, straight/linear ≥ concave-convex, << point contacts.
Grain bond	Predominantly direct grain contacts and mostly syntaxial quartz overgrowth.	Predominantly direct grain contacts and usually syntaxial quartz overgrowth.	Predominantly direct grain contacts and syntaxial quartz overgrowth.
Pores	1–5 vol.%, syntaxial quartz overgrowths, grain contacts << interporous spaces.	1–5 vol.%, syntaxial quartz overgrowths, grain contacts << interporous spaces.	10–15 vol.%, pore cement, prevailing interporous spaces 0.05–0.1 mm, scattered large, branched interporous spaces (L 0.6 mm, B 0.1–0.15 mm).
Contact cement	Argillaceous-chloritic, rarely ferritic (goethite) on grain margins	Argillaceous-chloritic, small amounts of goethite pigment	Locally argillaceous-ferritic (illite + goethite) at grain contacts, scattered glauconite
Pore cement	Argillaceous-chloritic, argillaceous-illitic, minute interporous spaces (0.07×0.04 mm to 0.25×0.05 mm), small amounts of goethite pigment. Goethite granulose (0.02–0.09 mm), brownish-grey semitransparent.	Argillaceous-chloritic, argillaceous-illitic, minute interporous spaces (0.1×0.03–0.06 mm), small amounts of goethite pigment. Goethite granulose (0.02–0.09 mm), brownish-grey semitransparent.	Locally argillaceous-ferritic (illite + goethite) in minute and scattered large interporous spaces. Syntaxial quartz overgrowths. ferritic (goethite) in large branched interstitial pore (0.35×0.2 mm).
Quartz	*c*. 94 vol.%, 0.03–0.3 mm, maximum 0.3 mm. Monocrystalline, minor subgrain phenomena. Rare Fe oxide coats. Small grains (0.03–0.1 mm) predominantly isometric, subrounded, irregular shapes, moderate to high sphericity. Large grains platy, poorly rounded (L 0.2–0.3 mm, W 0.05–0.1 mm), angular or splintered grains (L 0.13 mm, B 0.01–0.03 mm), subtriangular in section.	*c*. 90 vol.%, mostly 0.04–0.1 mm, maximum 0.15 mm. Monocrystalline, minor subgrain phenomena. Rare Fe oxide coats. Small grains (0.04–0.07 mm) predominantly isometric, subrounded, irregular shapes, moderate to high sphericity, some angular or splintered grains. Large grains (>0.15 mm) platy, poorly rounded (L 0.15–0.25 mm, W 0.04 – 0.05 mm), no angular or splintered grains observed.	*c*. 97 vol.%, 0.05–0.3 mm. Monocrystalline, minor subgrain phenomena. Frequent Fe oxide coats. Small grains (0.05–0.15 mm) predominantly subrounded/with rounded corners, irregular shapes, partly very angular, low sphericity. Large grains (>0.15 mm) either with low sphericity and poorly rounded corners or platy. Some grains with Boehm lamellas possible relics of the parent rocks, scarce inclusions of rutile (sagenite).
Alkali feldspar	0.05–0.1 mm. Fresh cleavage fragments with rounded edges or subrounded, microcline (twinning), rare perthitic unmixing.	0.06 mm. Fresh cleavage fragments with rounded edges or subrounded, microcline (twinning), rare perthitic unmixing.	0.12–0.17 mm. Fresh cleavage fragments with rounded edges or subrounded, microcline (twinning), fringe altered to illite or white mica.
Plagioclase	0.07–0.09 mm, max L 0.18 mm, B 0.03 mm. Fresh cleavage fragments with rounded edges, platy (L 0.18 mm, W 0.03 mm).	0.05–0.07 mm. Fresh cleavage fragments with rounded edges, platy (L 0.08–0.18 mm, W 0.05 mm); altered cleavage fragments (L 0.12 mm, W 0.05 mm; sericite, illite, goethite pigment).	0.15 mm. Fresh cleavage fragments with rounded edges.
Glauconite	0.04–0.12 mm. Subangular, deformed during compaction, partly with pale brown crust.	0.05–0.15 mm, maximum 0.25 mm (angular). Subangular, with rounded edges and angular, deformed during compaction, partly with pale brown crust.	0.05–0.15 mm. Subangular, deformed during compaction, no pale brown crust observed.
Tourmaline	0.04–0.07 mm. Cleavage fragments, generally with rounded edges, rarely polygonal.	0.07–0.08 mm. Cleavage fragments, generally with rounded edges.	0.04–0.1 mm. With rounded edges.
Zircon	0.03–0.04 mm, with rounded edges, partly with clearly recognizable oscillatory zonal composition.	0.04 mm, with rounded edges.	0.07 mm, with rounded edges, partly with clearly recognizable oscillatory zonal composition.
Apatite	L 0.1–0.12 mm W 0.03 mm. Long prisms, transparent.	L 0.05–0.12 mm, W 0.02–0.04 mm. Prismatic, transparent.	Not observed.

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Goethite, detritic	0.05–0.09 mm. Rounded.	0.04 mm. Rounded.	0.06–0.08 mm. Subrounded.
Dark mica (biotite)	L 0.07–0.15 mm. Scarcely present, pleochroism yellow brown to pale brown, altered.	L 0.22 mm. Pleochroism dark yellow brown to pale brown, altered to chlorite and goethite.	L 0.07 mm. Pleochroism yellow brown to pale brown, partly altered to chlorite.
White mica (muscovite)	L 0.06–0.17 mm, maximum 0.3 mm.	L 0.15–0.25 mm, maximum 0.6 mm (bent by compaction).	L 0.04–0.12 mm, maximum 0.3 mm.
Chlorite	L 0.15–0.22 mm, W 0.01–0.04 mm. Pleochroism yellowish green, pale green to colourless, lamellar intergrowth with white mica, goethite (altered dark mica).	L 0.22 mm, W 0.08 mm. Pleochroism yellowish green, pale green to colourless, lamellar intergrowth with white mica (altered dark mica).	Pleochroism yellowish green to pale green.
Chert grains	0.07 mm. Elongated, subrounded.	0.04 mm. Rounded.	0.09 mm. Subrounded.

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