

Jurassic sedimentation in the Cleveland Basin: a review

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39 **SUMMARY:** This review combines two Presidential Addresses (2005; 2006) and
40 aims to provide an up-to-date overview of the stratigraphy and sedimentation of the
41 Jurassic sequence of the Cleveland Basin (Yorkshire), including poorly known data
42 from the western outcrop. These fascinating rocks have been the focus of geological
43 research since the 18th century and have had a profound influence on the
44 development of the geological sciences. Throughout the 20th century, the excellent
45 coastal exposures have acted as a magnet for palaeontologists, stratigraphers,
46 sedimentologists and geochemists, as a natural geological laboratory, and in recent
47 decades, the coastal exposures received increased scientific interest as a result of
48 their analogy with hydrocarbon source and reservoir rocks in the North Sea.
49 Designation of the international Global Stratotype Section and Point (GSSP) for the
50 Sinemurian-Pliensbachian stage boundary in Robin Hood's Bay, the establishment of
51 the Dinosaur Coast, and development of the Rotunda Museum in Scarborough have
52 all given the regional geology additional importance.

53
54 The Lias Group (Hettangian to Toarcian age; 199.6 to 175.6 Ma), exposed in the well
55 known coastal sections, is illustrated by the fully cored Felixkirk Borehole, located at
56 the western margin of the outcrop, and is one of the best examples of shallow marine
57 sedimentation in an epeiric shelf-sea setting. It comprises two large-scale, upward
58 coarsening cycles, namely the Redcar Mudstone to Staithes Sandstone cycle,
59 followed by the Cleveland Ironstone to Blea Wyke Sandstone cycle. Within this broad
60 pattern, smaller scale transgressive-regressive cycles are described from
61 stratigraphically expanded and reduced successions. Detailed ammonite
62 biostratigraphy provides a finely calibrated temporal framework to study the
63 variations in sedimentation, which include storm-generated limestones and
64 sandstones ('tempestites') interbedded with mudstone deposited during fair-weather
65 periods. Hemipelagic mud, occasionally organic-rich, reflects deeper-water anoxic
66 events that may indicate a response to global climate change.

67
68 In cores, the tempestite beds (Hettangian to Sinemurian) are characterized by sharp
69 bases that, at outcrop, are often masked by downward penetrating burrows. Cyclicity
70 on a centimetre scale in the overlying Pliensbachian 'Banded Shales' may be the
71 result of orbitally induced, climatic cycles. Gradational upward coarsening to the
72 Staithes Sandstone Formation marks a transition to sand-rich tempestite deposits,
73 characterised by low angle and swaley cross-lamination, interbedded with sand-
74 starved units (striped siltstones). The sands were probably deposited from sediment-
75 laden, storm-surge and ebb currents in inner- and mid-shelf settings; the sandy
76 substrate was, at some levels, extensively bioturbated by deposit feeding organisms

77 that produced a spectacular range of trace fossil assemblages characteristic of
78 shoreface, inner-, mid-, and outer-shelf settings. Intrabasinal tectonics was a
79 controlling factor during deposition of both the Staithes Sandstone and the overlying
80 Cleveland Ironstone (Late Pliensbachian). The influx of sand is attributed to
81 hinterland uplift and increased sediment flux. More marked intraformational uplift
82 during deposition of the Cleveland Ironstone is manifested in a much attenuated
83 succession in the west of the basin (Felixkirk); southwards, towards the Market
84 Weighton High, the Pecten/Main Seam oversteps unconformably onto progressively
85 older beds to rest on the lower part of the Redcar Mudstone Formation. Ironstone, in
86 the form of berthierine ooids and sideritic mud, was deposited during 5-6 cycles (in
87 coastal exposures) of high sea-level stands that cut off siliciclastic influx from the low-
88 gradient hinterland; regressive, upward-shoaling intervals are marked by
89 interbedded, bioturbated siltstone and fine-grained sandstone.

90

91 The Toarcian succession (Whitby Mudstone and Blea Wyke Sandstone formations)
92 continues the second upward coarsening cycle in response to increased subsidence,
93 rising sea-level, and an influx of siliciclastic sand. Oxygenated, open marine mud was
94 deposited during the initial deepening phase, followed by bituminous mud, attributed
95 to ocean-water stratification and the establishment of anoxic bottom conditions; in the
96 west of the basin an upward shoaling sequence suggests that water depths were not
97 as great. Recent research on the geochemistry and stable isotope signatures across
98 this early Toarcian interval indicates a widespread, global anoxic event, possibly
99 attributed to the release of methane hydrate on the ocean floor. The Alum Shale
100 Member represents increasingly oxygenated bottom conditions and an upward
101 coarsening motif with passage to the Blea Wyke Sandstone Formation, which is
102 preserved only in the Peak Trough, an actively subsiding graben. Basin uplift
103 accompanied by gentle folding in late Toarcian to Aalenian times removed much of
104 the late Toarcian succession so that the Middle Jurassic Dogger Formation
105 (Aalenian), a complex, condensed, shallow water unit rests unconformably on beds
106 as low as the Alum Shale over much of the Basin.

107

108 Deep boreholes and revision mapping by the BGS in the west of the outcrop have
109 allowed a fuller, basin-wide synthesis of the palaeoenvironments and the influence of
110 intra-Jurassic tectonics during Mid- to Late Jurassic times. During Mid-Jurassic times
111 the low-lying, paralic coastal plain, typified by braided and meandering fluvial
112 systems and lacustrine deposits was invaded by marine incursions from the south
113 and east. Each transgressive event was different in its geographical penetration
114 across the coastal plain, resulting in varied lithofacies and palaeoenvironments

115 including ooidal ironstone and lime mud (Eller Beck Formation), peloid and ooid
116 carbonate shoals (Lebberston Member), and tidal sand bars, pelloidal limestones and
117 nearshore marine muds (Scarborough Formation). Trace fossils, including dinosaur
118 footprints, and macro-plant fossils tell us much about the palaeoenvironments on the
119 coastal plain, during this time interval (175.6 Ma – 164.7 Ma) that was characterised
120 by a warm, seasonal climate.

121

122 The basin wide transgression and marked global sea-level rise represented by the
123 Cornbrash Formation, marks deposition in a shallow marine environment during the
124 Callovian, followed by sand (Osgodby Formation) and deeper water muds (Oxford
125 Clay Formation) that spread northwards from the East Midlands over the Market
126 Weighton High during the Oxfordian. Subsequent shallowing of the basin resulted in
127 the establishment of a carbonate/siliciclastic platform typified by ooidal shoals, coral
128 patch reefs and sponge spicule-rich marine sands (Corallian Group). Their complex
129 sedimentation pattern was influenced by local infra-Oxfordian tectonics related to the
130 Howardian-Flamborough Fault Belt. Although the Ampthill Clay and Kimmeridge Clay
131 formations, the latter representing the most important regional hydrocarbon source
132 rock, are not well-exposed, recent boreholes in the Cleveland Basin have allowed a
133 much better understanding of the hemi-pelagic marine environment (both oxic and
134 anoxic) during this phase of sedimentation which marks a global sea-level rise.

135 Although well-studied by world standards, the Jurassic sediments of the Cleveland
136 Basin continue to throw up surprises and advances in our understanding of the Earth
137 as a dynamic system over a period of about 30 million years. These studies have
138 directly and indirectly influenced our understanding of the Earth as a system, and
139 have played an important role in educating non-specialists, undergraduates and
140 professional geologists over many decades.

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147 The Jurassic rocks of the Cleveland Basin (Yorkshire) have been a focus of
148 geological research since the 19th century, with studies by William Smith, his
149 nephew John Phillips (1829), George Young and John Bird (1822), Martin Simpson
150 (1884) and the Survey geologist Charles Fox-Strangways (1992, 1915), to name but
151 a few. Throughout the 20th century the excellent coastal exposures have acted as a
152 magnet for palaeontologists, stratigraphers, sedimentologists and geochemists, as a
153 natural geological laboratory, and their study was given additional impetus in the later
154 part of the 20th and early 21st centuries by their analogy with hydrocarbon reservoir
155 and source rocks in the North Sea Basin.

156

157 The Jurassic of Yorkshire (Fig. 1) was treated to an authoritative review by past
158 YGS President Professor John Hemingway (Hemingway 1974) and has been the
159 subject of a number of excellent regional summaries and guides (Black 1934b;
160 Rawson & Wright 1992, 1996, 2000; Cope 2006). Many of these publications
161 focussed on the superb coastal sections so, in my YGS Presidential addresses
162 (2005, 2006), summarized here, I aimed to present an overview of Jurassic
163 sedimentation in the Cleveland Basin, including relatively recent and unpublished
164 data from the western margin of the basin, an area often overlooked. With such a
165 wealth of scientific publications on these fascinating rocks, I have only been able to
166 focus on key topics, but I hope this paper will provide an up-to-date overview of the
167 geology that will stimulate further research into outstanding problems.

168

169

1. STRUCTURAL SETTING

170

171 The Cleveland Basin in Jurassic times formed part of a system of shallow epeiric
172 seas and small extensional tectonic basins, linked via the Sole Pit Basin (a half-
173 graben structure) to the North Sea Basin (Fig. 2; Zeigler 1982). The Cleveland Basin
174 was relatively small, and was bounded to the north-east by the Mid-North Sea High
175 and to the west by the Pennine High. To the south lay the East Midlands Shelf, the
176 northern part of which comprised the Market Weighton High (MWH) (Kent 1955),
177 which remained as a relatively stable unfolded block, probably underlain by a granite
178 intrusion (Bott *et al.* 1978; Donato 1993) and characterized by reduced rates of
179 sedimentation throughout the Jurassic. The MWH is an asymmetrical structure over
180 which subsidence and sedimentation rates were reduced; it separated rapid
181 subsidence and higher sedimentation rates to the north, in the Cleveland Basin, from
182 more gradual subsidence to the south in the Lincolnshire area of the East Midlands
183 Shelf (Kent 1955, 1974).

184

185 The Mid North Sea High underwent tilting to the southwest in Mid Jurassic
186 times, probably in response to doming associated with the Forties-Piper Volcanic
187 Centre in the central North Sea Basin (Sellwood & Hallam 1974; Zeigler 1982;
188 Underhill & Partington 1993). Mid Jurassic sedimentation in the Cleveland Basin was
189 therefore characterized by marine transgressions that advanced in a north-westerly
190 direction across the Market Weighton High, and by progradation of fluvial and deltaic
191 siliciclastics towards the south-east (Hemingway 1974). The shoreline depositional
192 lithofacies of the Lower Jurassic strata were removed following basin inversion and
193 erosion during the Neogene, but projection of lithofacies and thickness trends
194 suggests that the Early Jurassic shoreline lay around the present-day northern
195 Pennines and Southern Uplands (Fig. 2).

196

197 At the regional scale, the Cleveland Basin was affected by a number of
198 extensional faults and probable strike-slip fault complexes (Hemingway 1974; Kirby &
199 Swallow 1987) that roughly define the present-day outcrop (Fig. 3). Some were
200 active during Early and Mid Jurassic sedimentation (Milsom & Rawson 1989; Powell
201 *et al.* 1992). To the south, the east-west trending Asenby–Coxwold–Gilling Graben,
202 the Helmsley-Filey Fault Belt and the Howardian-Flamborough Fault Belt (defining
203 the Vale of Pickering) were intermittently active north of the Market Weighton High
204 during Mid to Late Jurassic times (Kirby & Swallow 1987). The eastern margin of the
205 basin is cut by north-trending structures, such as the Peak Trough and Peak Fault
206 (Milsom & Rawson 1989), the Cayton Bay Fault and the Whitby Fault. The Peak
207 Trough was active in Early Jurassic times and preserves a thicker sequence of Lower
208 Jurassic rocks compared to the surrounding areas. Furthermore, the entire Cleveland
209 Basin was subjected to gentle folding and erosion in late Toarcian time, so that the
210 upper part of the Lias Group was eroded to increasingly lower stratigraphical levels
211 towards the south and southeast of the basin (Black 1934a; Hemingway 1974). The
212 western margin of the present-day outcrop is marked by the north-trending Borrowby
213 Graben (Powell *et al.* 1992), which, like the Peak Trough, shows evidence of
214 synsedimentary faulting during the Mid Jurassic. Less evident is the uplift of part of
215 the western outcrop near Roulston Scar (Hambleton Hills) in the Oxfordian, resulting
216 in local erosion of the Oxford Clay (Powell *et al.* 1992, fig. 16). A number of the major
217 bounding faults are known to have been active during the Cimmerian orogeny,
218 especially the east-west trending Coxwold-Gilling and Howardian-Flamborough fault
219 belts, which show extension in Oxfordian (Late Jurassic) times (Wright, 2009) and
220 renewed movements in post-Cretaceous times (Kirby & Swallow 1987; Starmer
221 1995). Petrographic and fission-track analysis suggest that the Middle Jurassic
222 sediments were buried to a depth of about 2 to 3km (Hemingway & Riddler 1982;

223 Green 1986; Bray *et al.* 1992), prior to inversion and north-south compression during
224 the latest Cretaceous to Neogene. The latter resulted in formation of the complex
225 east-west trending Cleveland Anticline (Fig. 3) and subsidiary folds (e.g. the Lockton
226 Anticline, Goathland Syncline and Robin Hood's Bay Dome) that control the Jurassic
227 outcrop pattern of the North York Moors (Kent 1980a, 1980b).

228

229

2. PALEOGEOGRAPHY

230

231 The palaeogeography of the Cleveland Basin is less well known during the Early
232 Jurassic (Lias Group; 199.6 to 175.6 Ma) than at later times. With notable
233 exceptions such as the London Platform, the region then formed part of the broad
234 epeiric sea that covered much of England and Wales and much of western Scotland
235 (Cope *et al.* 1992; Scrutton & Powell 2006; Fig. 2). During sea-level low stands (e.g.
236 late Pliensbachian), however, clastic sediments may have been derived from an
237 emergent Pennine-Caledonian High. Lias Group 'background' sediments mostly
238 comprise mudstones, but regional variations in interbedded coarser grained
239 bioclastic carbonate and siliciclastic sediments, ranging from carbonate dominated in
240 southern Britain (e.g. Blue Lias) to siliciclastic, storm-dominated sediments in the
241 Cleveland Basin, suggest a northerly source area for the siliciclastic sediments. This
242 may have been the Pennine High or land areas in southern Scotland (Fig. 2).

243

244 By Mid Jurassic time, uplift of the Mid North Sea High and the northern
245 source areas noted above, coupled with uplift (possibly isostatic buoyancy) over the
246 Market Weighton High, defined a more or less circular Cleveland Basin that was
247 linked intermittently to the East Midlands Shelf to the south and the Sole Pit Basin to
248 the south-east. Marine sedimentation continued in a broad epeiric shelf setting in
249 southern Britain and northwest Europe, but tectonic uplift and North Sea
250 doming/rifting resulted in fluvial progradation from the northwest and northeast (Knox
251 *et al.* 1991) into the Cleveland Basin and deposition in paralic environments ranging
252 from river, lake and delta to estuarine. There is also evidence of 'Millstone Grit'
253 quartz granules in the Aalenian sediments of the Howardian Hills, suggesting a
254 western Pennine source area. Only occasionally did sea-level rise result in marine
255 transgression over the low-lying paralic hinterland. These brief marine
256 transgressions advanced generally northwestwards (Knox 1973; Parsons 1977) over
257 the Market Weighton High, and except for the mid-Bajocian sea-level high
258 (ammonite-bearing Scarborough Formation), they did not extend to the northwest of
259 the present-day outcrop (Fig. 4c).

260

261 Rapid (global) sea-level rise throughout Britain during the early Callovian
262 resulted in partial drowning of the Cleveland Basin and widespread marine
263 sedimentation over northern Britain (Fig 8a). Fully marine ammonite faunas indicate
264 connection with the faunal provinces of southern Britain. The Market Weighton High
265 still influenced sedimentation in the region, however, resulting in thinner and
266 distinctive marine calcareous sands during the Callovian (164.7-161.2 Ma) compared
267 to muds on the East Midlands Shelf. Sand was mostly derived from the northwest
268 (Wright 1977), but tectonics resulted in depositional hiatuses and much reworking of
269 sediment during the Callovian, prior to sea level rise that was characterized by the
270 northwards development of deeper water muds (upper Oxford Clay) across the
271 Market Weighton High in the early Oxfordian. During the late Early Oxfordian to early
272 Late Oxfordian, sedimentation in the Cleveland Basin was again distinct from the
273 area south of the Market Weighton High (Rawson & Wright 2000, fig. 3c).
274 Subsidence rates kept pace with sea-level rise and resulted in deposition of
275 distinctive Corallian Group sediments, a complex suite of marine siliceous sands,
276 spiculites, ooidal shoals, micritic limestone and coral/algal patch reefs (Blake &
277 Huddleston 1877; Wilson 1936, 1949; Wright 1972, 1983). This second phase of
278 basin inversion (relative to the gently subsiding East Midland Shelf) resulted in
279 shallower water sedimentation coeval with deeper water muds of the West Walton
280 Formation and Ampthill Clay on the East Midland Shelf and the Seeley Formation in
281 the Sole Pit Basin (Fig. 7). It was not until mid late Oxfordian times that increased
282 subsidence and rising sea-level allowed the mud lithofacies of the Ampthill Clay to
283 spread northward from the Vale of Pickering area (Cox & Richardson 1982). Finally,
284 a major worldwide sea level rise during the Kimmeridgian (Hallam 1988; Haq *et al.*
285 1988; Herbin *et al.* 1991) resulted in deeper water hemipelagic sedimentation over
286 wide areas of present day Britain, by which time the Cleveland area was no longer an
287 active and distinct tectono-depositional basin.

288

289 Relative sea-level changes expressed within the Jurassic succession of the
290 Cleveland Basin are often not in accord with global patterns (Hallam 1988; Haq 1988;
291 Hallam 2001); there is good correspondence with global sea-level rise in the early
292 Hettangian (Calcareous Shales, Redcar Mudstone Formation), early Pliensbachian
293 (Pyritous/Banded Shales; Redcar Mudstone Formation); early Toarcian (Grey
294 Shales/Mulgrave Shale members; Whitby Mudstone Formation) and late Oxfordian
295 (Weymouth Member, Oxford Clay). However, the early Sinemurian, early-late
296 Bajocian and mid-Callovian global sea-level rises are not well expressed. This is due
297 to the effects of local and regional intra-plate tectonics which resulted in hinterland
298 uplift and local basinal subsidence, increased sediment flux and regressive

299 siliciclastic sedimentation (e.g. Ravenscar Group and Osgodby Formation in the
300 Cleveland Basin).

301

302 **3. BIOSTRATIGRAPHICAL AND CHRONOSTRATIGRAPHICAL FRAMEWORK**

303

304 Ammonites have traditionally provided the biostratigraphical and
305 chronostratigraphical framework for the Jurassic of the Cleveland Basin, although
306 their absence from the paralic and fluvio-deltaic lithofacies of the Middle Jurassic
307 Ravenscar Group has resulted in an interesting debate on the timing of basin fill
308 during the Bajocian and Bathonian (Leeder & Nami 1979; Riding & Wright 1989;
309 Butler *et al.* 2005).

310

311 Ammonite zonation is precise and is based on the benchmark work of Arkell
312 (1933, 1945) and later workers. Up to 65 ammonite zones have been recognized in
313 the Cleveland Basin (Figs 5, 6, 7), together with many subzones that allow fine
314 temporal resolution and correlation throughout Britain (Buckman 1909-30; Dean *et al.*
315 1961; Howarth 1955, 1962, 1973; Cope *et al.* 1980a, b; Callomon 1995). The
316 duration of ammonite zones is difficult to determine as rates of extinction and the
317 incoming and acme of new species are likely to have varied through Jurassic time.
318 As a common 'rule of thumb', the duration of an ammonite zone was estimated to be
319 about 1 million year (Ma). However, where recent radiometric ages have been
320 determined based on U-Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ ratios, the duration of Jurassic ammonite
321 zones has been estimated to be between 0.4 and 1.6 Ma (Palfy & Smith 2000),
322 although this duration has been questioned for the Toarcian by McArthur *et al.*
323 (2000), who consider the variation in duration to be much greater.

324

325 Ammonites typical of the Boreal (northern) and Tethyan (southern) realms are
326 present in the Cleveland basin as a result of periodic connection between these two
327 palaeobiogeographical provinces via the Faeroes Rift, the Anglo-Welsh Basin, the
328 Paris Basin and the open Tethys Ocean located to the south (Cope 2006). Howarth
329 (1976) recognized the incoming of Tethyan forms during the Early Jurassic,
330 Sinemurian–Aalenian interval. At other times, Boreal faunas were dominant,
331 especially during the Callovian transgression and the Oxfordian, and are typified by
332 cardioceratid and kosmoceratid ammonites (Cope 2006). Of considerable note is the
333 selection of the Global Stratotype Section and Point (GSSP) for the base of the
334 Pliensbachian Stage at Wine Haven, Robin Hood's Bay (Meister *et al.* 2006), at a
335 level interpreted as coinciding with a major deepening of the sea manifested in the
336 lower part of the Pyritous Shales Member (Redcar Mudstone Formation, Lias Group).

337

338 The ammonite zonal scheme in relation to the chronostratigraphy and
339 lithostratigraphy of the Cleveland Basin and correlative strata in the Sole Pit Basin
340 and East Midlands Shelf is outlined in [Figures 5, 6 and 7](#). The zonal scheme is based
341 on the Boreal ammonite distribution, most commonly used in the UK; in this paper
342 ammonite zones are used as chronozones, so their names are capitalized using the
343 Standard Zone terminology species name (e.g. Planorbis Zone), but for ease of
344 reference to the genus and species (e.g. *Psiloceras planorbis*), the names are written
345 as biozones in Figures 5, 6 and 7. Ogg *et al.* (2008) have revised the Jurassic time
346 scale so that the geochronological age of the base of the Jurassic is 199.6 Ma and
347 the base of the Cretaceous is 145.5 Ma, a duration of 54 million years, considerably
348 shorter than earlier estimates of 205.7 Ma (base) and 142.0 Ma (top) and 63.7 years
349 duration (Gradstein & Ogg 1996). However, as a result of Late Cimmerian (latest
350 Jurassic to pre-Cretaceous) erosion (Rawson & Riley 1982), the youngest beds
351 preserved in the Cleveland Basin belong to the Pectinatus Zone, c.151 Ma (Ogg *et al.*
352 *al.* 2008), or possibly the higher Pallasioides Zone (Herbin *et al.* 1991) ([Fig. 7](#)).

353

354 The standard north-west European sequence of Lower Jurassic ammonite
355 chronozones and sub-chronozone for the Hettangian, Sinemurian, part of the
356 Pliensbachian and Toarcian stages has been recognized in the Lias succession in
357 well-exposed coastal sections (Buckman 1909-30, 1915; Bairstow 1969; Howarth
358 1955, 1962, 1973, 2002). Most of these zones have also been identified in the
359 Felixkirk cores ([Fig. 9](#)) (Ivimey-Cook & Powell 1991, fig. 2; Powell *et al.* 1992). The
360 ammonite zonation for the Middle and Upper Jurassic is based on Cope *et al.*
361 (1980a, b), especially the work of Wright (1980).

362

363 Other fossil groups, particularly microfossils, have aided biostratigraphical
364 zonation and correlation, especially with the North Sea Basin and for the paralic and
365 marginal marine successions where ammonites are not present. Bate (1964, 1965,
366 1967) used ostracods to correlate thin transgressive marine units of the Middle
367 Jurassic Ravenscar Group with the fully marine succession of the East Midlands
368 Shelf. Although less precisely resolved, dinoflagellate cysts were used by Woollam &
369 Riding (1980) to establish up to 16 zones calibrated against the standard north-west
370 European ammonite scheme (76 zones). Dinoflagellates have enabled correlation of
371 the Middle Jurassic succession in the Cleveland Basin with the southern North Sea
372 (Hancock & Fisher 1981) and northern North Sea (Butler *et al.* 2005), and have
373 helped to resolve the age of the Bajocian to Bathonian succession onshore (Riding &
374 Wright 1989).

375

376

4. LITHOSTRATIGRAPHY OF THE CLEVELAND BASIN

377

378 A brief outline of the lithostratigraphy of the Jurassic succession is presented in this
379 section (Figs 5, 6, 7), and the former nomenclature is shown in Tables 1 and 2.
380 Further details of the succession are presented in later thematic sections (Section 5).

381

382 Lower Jurassic sediments comprise the **Lias Group** (Fig 8b) of Hettangian to
383 Toarcian age, with a maximum thickness of 454 m; subdivisions are based on Powell
384 (1984), Knox (1984), Ivimey-Cook & Powell (1991), Howard (1985) and Rawson &
385 Wright (1992). Lias Group sediments (Figs 5, 9) rest conformably on the Upper
386 Triassic (Rhaetian) **Penarth Group** (Benfield & Warrington 1988; Ivimey-Cook &
387 Powell 1991). Black, anoxic, fissile mudstones of the Westbury Formation and the
388 overlying grey-green smectitic claystones of the Cotham Member (Lilstock
389 Formation), both formations of the Penarth Group, were deposited in brackish and
390 restricted lagoons and are dominated by monospecific faunas. The first truly marine
391 interbedded limestones/mudstone beds typical of the Lias Group occur about 10 m
392 below the first marine ammonite fauna represented by *Psiloceras planorbis* (Ivimey-
393 Cook & Powell 1991), which marks the base of the Hettangian Stage in the region
394 (Fig. 9). However, the GSSP for the base of the Jurassic System and the Hettangian
395 Stage is placed at the incoming of *Psiloceras spelae* in the mid-European Tethyan
396 realm, slightly earlier than the '*planorbis* event' in the UK (Page & Bloos 1998; Lucas
397 & Tanner 2007).

398

399

4.1 Lower Jurassic Succession

400

401
402 The Lias Group (Figs 5, 9) in the Cleveland Basin is divided into five formations
403 (Powell 1984), described here in ascending order. The **Redcar Mudstone**
404 **Formation** (c. 283 m thick) forms the greater part of the group, and consists of clay
405 and silt grade siliciclastic sediments interbedded with carbonate-rich shell beds of
406 various types, concretion beds and fine-to medium-grained siliciclastic beds. The
407 coarse-grained beds enable subdivision of the formation into five informal members
408 on the coast (Tate & Blake 1876; Fox-Strangways 1892; Fox-Strangways & Barrow
409 1915; Buckman 1915; Hemingway 1974; Knox *et al.* 1991; van Buchem & McCave
410 1989; Hesselbo & Jenkyns 1998; van Buchem & Knox 1998). In upward sequence,
411 these are the Calcareous Shales (with numerous oyster-rich limestone beds),
412 Siliceous Shales (bioturbated, sand-rich beds), Pyritous Shales (pyritous nodules and

413 beds of concretionary siderite), Banded Shales (regular alternations of siltstone and
414 mudstone beds) and Ironstone Shales (iron-rich, silty laminations). However, in the
415 west of the basin, the distinction between the last three members is less apparent.
416 There, a gradational upward coarsening trend within the Pyritous/Banded/Ironstone
417 interval (lower Pliensbachian) is clearly shown on the gamma-ray logs of the Felixkirk
418 Borehole [SE 4835 8576] (Figs 1, 9; Powell & Ivimey-Cook 1991; Powell *et al.* 1992).
419 The formation is considerably thinner (194 m) in the west of the basin. The marked
420 'saw-tooth' expression of the gamma-ray and sonic geophysical logs in the
421 Calcareous Shales and Siliceous Shales members (Fig. 9) is due to the intercalation
422 of fine-grained 'background' sediments (mudstone) and coarse-grained bioclastic or
423 sand-rich beds that form the characteristic 'benches' in these Hettangian to
424 Sinemurian strata in Robin Hood's Bay (Fig 10a). The origin of these beds is
425 considered in Section 5.

426

427 The Ironstone Shales pass gradationally upward with increasing sand-grade
428 sediment to the **Staithe Sandstone Formation** (c. 30 m) (Hemingway 1974;
429 Howard 1985). The formation forms a coastal cliff and inland scarp feature, and
430 consists of grey, yellow weathering, fine- to medium-grained sandstone and siltstone
431 of late Pliensbachian age. On the coast, at Staithe, the upper part of the formation
432 has a higher ratio of siltstone to sandstone than in the western outcrop. Sandstone
433 beds are often characterized by low-angle, wavy and hummocky cross-bedding
434 (Howard 1985), and the beds are often heavily bioturbated with a rich suite of
435 ichnofossils (Figs 10c,f) (Knox *et al.* 1991).

436

437 The **Cleveland Ironstone Formation (CIF, c. 28 m thick)** and the underlying
438 Staithe Sandstone Formation form a marked mappable feature inland, hence the
439 earlier term 'Middle Lias' for these two formations. The formation, on the coast, is
440 subdivided into a lower Penny Nab Member (Howard, 1984), including five mineable
441 ironstone seams (in upward sequence: the Osmotherly, Avicula, Raisdale and Two
442 Foot seams). The overlying, the Kettleless Member, which includes the Pecten and
443 Main Seam, unconformably oversteps successively younger Lias Group units to the
444 south towards Market Weighton, and to the west (Fig. 16).

445 Ironstone represents only a small part (c. 30%) of the CIF, which consists of
446 grey mudstone and sandy mudstone interbedded with sideritic and berthierine
447 (chamosite)-rich ooidal ironstone (Sorby 1857; Lamplugh 1920; Hemingway 1951;
448 Whitehead *et al.* 1952; Chowns 1968; Howard 1985). Intervening siliciclastic beds
449 show fine parallel lamination, wave ripple lamination and erosional gutter casts
450 (Greensmith *et al.* 1980; Rawson *et al.* 1983; Howard 1985). The formation is

451 thickest at Staithes, but thins to the south and west where the siliciclastic interbeds
452 are reduced in thickness, and as a result of an intraformational unconformity below
453 the Pecten Seam (Fig. 9), only three seams including the Main Seam are present at
454 Felixkirk, with a total thickness of 9m (Powell *et al.* 1992). This is a result of the Main
455 Seam overlapping the Pecten Seam to rest with overstep on successively older
456 strata to the south. In the Howardian Hills, the formation is only 2m thick (Chowns
457 1968)

458

459 The beds informally known as the 'Upper Lias' comprise the **Whitby**
460 **Mudstone Formation** and the **Blea Wyke Sandstone Formation** (Rastall 1905;
461 Powell 1984; Knox 1984). The Whitby Mudstone (c. 105 m thick) consists
462 predominantly of grey to dark grey mudstone and siltstone with abundant shelly
463 fossils at some levels. Uplift and erosion prior to deposition of the Dogger Formation
464 in Aalenian times has resulted in the full succession being preserved only in the
465 syndepositional Peak Trough (Milsom & Rawson 1989), where five members are
466 present (Knox 1984). The **Grey Shale Member** (c. 13.5 m max.) comprises the
467 eponymous silty mudstone with beds of calcareous siderite concretions. A change to
468 more anoxic bottom conditions is recorded in the overlying **Mulgrave Shale Member**
469 **(formerly Jet Rock Member)** (Rawson & Wright 1992) (c. 31 m max.), which
470 consists of fissile, bituminous, dark grey mudstone with abundant ammonites. This
471 unit was long exploited for the mineral jet (dense, water-logged, araucarian wood),
472 mined on the coast and sporadically inland for the manufacture of jewellery
473 (Hemingway 1974, p.174). The **Alum Shale Member** (max. 37 m) is generally less
474 fossiliferous and comprises grey silty mudstone with bands of calcareous and siderite
475 concretions, and bands of phosphatic nodules in the upper part; the shales were
476 formerly worked in large quarries for alum used for 'fulling' wool and in chemical
477 industries (Gad *et al.* 1969; Hemingway 1974). The Middle Jurassic Dogger
478 Formation rests unconformably on this unit over much of the Cleveland Basin (e.g. in
479 the Felixkirk Borehole; Fig. 9). Where the full succession is preserved, the upper part
480 of the Alum Shale Member shows a gradual coarsening upward trend to the overlying
481 **Peak Shale Member** (Knox 1984). The upward coarsening trend continues into the
482 **Fox Cliff Siltstone Member**, comprising muddy siltstone with beds of calcareous
483 and sideritic concretions and with small phosphatic nodules. The coarse-grained end-
484 member of this trend is the **Blea Wyke Sandstone Formation** (18 m max.),
485 comprising grey, mud-rich sandstone (Grey Sandstone Member) passing up to
486 'cleaner' yellow sandstone (Yellow Sandstone Member). When traced southwards
487 towards the Market Weighton High (MWH), pre-Dogger erosion cuts downwards
488 through the Whitby Mudstone so that the Dogger Formation rests on the Mulgrave

489 Shale Member in the Brown Moor Borehole (Gaunt *et al.* 1980; Fig 1), located north
490 of the MWH. At Market Weighton, the highest Lower Jurassic strata below the sub-
491 Cretaceous unconformity comprise sandstone with ironstone nodules attributable to
492 either the Cleveland Ironstone Formation or Staithes Sandstone Formation (Whitham
493 *in* Scrutton & Powell 2006).

494

495 In the Southern North Sea Basin, the Lias Group is between 200 and 300 m
496 thick, but reaches up to 820 m in the Sole Pit Basin. It thins northwards towards the
497 Mid North Sea High (Lott & Knox 1994). Formations offshore are recognized largely
498 from their geophysical wireline log characteristics (Fig. 5), but are broadly equivalent
499 to the onshore equivalents. Hence, in upwards sequence, the Penda and Offa
500 formations are equivalent to the Redcar Mudstone Formation, the ferruginous and
501 sandy Ida Formation is equivalent to the Cleveland Ironstone and Staithes
502 Sandstone formations, and the Cerdic Formation is equivalent to the Whitby
503 Mudstone Formation. As a result of latest-Toarcian folding, the sandy late Toarcian
504 Phillips Member of the southern North Sea Basin, broadly equivalent to the upward
505 coarsening Blea Wyke Formation, was removed by erosion and is identified only in a
506 few wells (e.g. 47/3b-4; 42/29-1), possibly restricted to extensional rifts similar to the
507 better known Peak Trough (Lott & Knox 1994).

508

509

510 **4.2 Middle Jurassic Succession**

511

512 The **Dogger Formation**, up to 13 m thick (Hemingway 1974), is Aalenian in age
513 (Black 1934a; Parsons *in* Cope *et al.* 1980b). It rests unconformably on the Lias
514 Group, generally on the Alum Shale Member but disconformably on the Blea Wyke
515 Sandstone (Knox 1984) within the Peak Trough. In coastal sections, the marine
516 Dogger Formation is generally represented by thin ferruginous sandstone, locally rich
517 in berthierine and calcareous ooids. Intense bioturbation is common, and soft-
518 sediment burrows penetrate downward into the underlying, mudstone (Alum Shale).
519 Near Whitby (East Cliff), for example, the Dogger, c. 1 m thick, consists of highly
520 bioturbated, ferruginous sandstone with rounded black phosphatic pebbles, locally
521 with endolithic borings. Inland, the Dogger Formation is a lithologically
522 heterogeneous unit, comprising conglomerate, sandstone, mudstone, ooidal and
523 bioclastic limestone and ironstone, and including marine and brackish lithofacies.
524 When traced southwards towards the northern margin of the Market Weighton High,
525 the Dogger Formation rests unconformably (overstep) on older units of the Lias
526 Group down to the Redcar Mudstone Formation (Hemingway, 1974).

527

528 The majority of the Middle Jurassic (Aalenian to Bathonian) succession is
529 represented by the **Ravenscar Group** (240 m max.) (Smithson 1934, 1942;
530 Hemingway 1949; Hemingway & Knox 1973; Lott & Humphreys 1994; Cox &
531 Sumbler 2002) comprising mostly paralic, including fluvial and lacustrine, lithofacies,
532 and three distinctive transgressive marine units (Fig. 6): the Eller Beck Formation, the
533 Lebberston Member of the Cloughton Formation, and the Scarborough Formation.
534 The Ravenscar Group thins rapidly southwards to 57 m in the Fordon Borehole [TA
535 058 758], south of the Vale of Pickering (Fig. 8b), and a similar thickness was
536 recorded in the Brown Moor Borehole (see below).

537 The paralic units in the succession (formerly known as 'Estuarine' or 'Deltaic'
538 units; Table 2) comprise the Saltwick, Cloughton and Scalby formations. The
539 **Saltwick Formation** (57 m max.) generally overlies the Dogger Formation, but rests
540 unconformably on the Lias Group (Alum Shales) where the Dogger is absent due to
541 erosion or non-deposition. It consists mostly of medium- to coarse-grained, cross-
542 bedded channel sandstones with fine-grained, planar laminated and ripple cross-
543 laminated sandstone and micaceous mudstone; drifted plant fragments and *in situ*
544 plant rootlets are common in some beds. The **Eller Beck Formation** (c. 8 m max.)
545 represents the first transgressive marine incursion that advanced northwestwards
546 across the basin; it comprises sandstone rich in berthierine ooids, ooidal ironstone
547 and mudstone (Barrow 1877; Knox 1973). The overlying **Cloughton Formation** (85
548 m) is lithologically similar to the Saltwick Formation, but includes a marine
549 limestone/sandstone unit, the **Lebberston Member** (up to 9 m), which, where
550 present in the south of the basin, divides the formation into a lower **Sycarham**
551 **Member** and an upper **Gristhorpe Member**. In southern coastal exposures, where
552 the Lebberston Member comprises sandy ooidal limestone and calcareous
553 sandstone, it is known as the 'Millepore Bed' lithofacies. In the Hambleton and
554 Howardian Hills, it is more calcareous, and is referred to as the 'Whitwell Oolite'
555 lithofacies (Richardson, 1912). When traced southwards to the Market Weighton
556 High (e.g. in the Brown Moor Borehole, Gaunt *et al.* 1980), the attenuated sandy
557 paralic Cloughton Formation succession (56 m thick) between the Eller Beck
558 Formation and the Scarborough Formation becomes more 'marine' in character, and
559 includes 16 m of ooidal limestone (in 3 beds) and sandstone with bivalves and
560 scattered ooids.

561 The **Scarborough Formation** (Bate 1965; Parsons 1977, 1980; Gowland &
562 Riding 1991; Butler *et al.* 2005), marks a major marine transgression over the whole
563 Basin in the early Bajocian. In the coastal type section at Hundale Point (Fig. 27) it is
564 dominated by mud- and sand-rich sediments with thin argillaceous limestones,

565 subdivided into seven members (Table 2; Gowland & Riding 1991). Ammonites such
566 as *Dorsetensia* and *Teloceras*, and marine palynomorphs in the Ravenscar Shale
567 Member, indicate the Humphriesianum Zone (late early Bajocian). In the coastal
568 outcrop, marine siliciclastic sediments with hummocky cross-bedding (e.g. at
569 Ravenscar cliff) and thin silty limestones that yield bivalves (*Gervillella*,
570 *Pseudomontis*, *Trigonia*, *Astarte* and *Lopha*), belemnites and sparse ammonites,
571 together with a diverse suite of shallow marine trace fossils including *Rhizocorallium*,
572 *Teichichnus* and U-shaped *Diplocraterion* (Hemingway 1974; Miller *et al.* 1984;
573 Gowland & Riding 1991). This contrasts with the different succession in the western
574 outcrops of the Hambleton Hills (Table 2), where a lower unit, the **Brandsby**
575 **Roadstone Member**, comprising peloidal (faecal peloids) planar cross-bedded
576 limestone, is overlain by medium-grained, fossiliferous sandstone, the **Crinoid Grit**
577 **Member** (Powell *et al.* 1992).

578

579 A return to fluvio-deltaic and paralic lithofacies is marked by **Scalby**
580 **Formation** (c. 60 m) (Black, 1928; Leeder & Nami, 1979). At the base, the Moor Grit
581 Member consists of medium- to coarse-grained, locally pebbly, cross-bedded,
582 channel sandstone unconformably overlying the Scarborough Formation. It passes
583 gradationally up to the **Long Nab Member**, which is characterised by micaceous
584 mudstone and finer-grained sandstone locally with abundant plant remains; channel
585 sandbodies are less common and smaller in size compared to those in the Moor Grit,
586 hence its former name the 'Level-bedded Series' (Hemingway 1974).

587

588 The biostratigraphical framework of the Ravenscar Group is poorly
589 constrained. Based on ostracod faunas (Bate 1967), the marine Eller Beck Formation
590 and Leberston Member are thought to be of late Aalenian-early Bajocian and early
591 Bajocian age respectively, coeval wholly or in part with the Discites Zone of the
592 Lincolnshire Limestone. Sparse ammonites collected from the Scarborough
593 Formation suggest the 'mid-Bajocian' Humphriesianum Zone (Romani to Blagdeni
594 subzones) (Parsons 1977). Correlation of the western inland succession with the
595 typical coastal exposures is, however, tentative, and the Scarborough Formation in
596 the Hambleton Hills may be representative of the early Sauzei Zone (Fig. 6; Parsons
597 1980). Fluvial and paralic parts of the succession contain few biostratigraphical
598 indicators, but given the ages indicated for the marine units, the Cloughton Formation
599 probably spans the Discites, Laeviuscula and possibly Sauzei zones (Fig. 6; Cope *et al.*
600 *et al.* 1980b). The basal part of the Scalby Formation (Moor Grit Member) on the coast
601 has yielded a dinoflagellate cyst assemblage of probable late Bajocian age (Riding &

602 Wright 1989); the overlying Long Nab Member ranges from late Bajocian in the lower
603 part to Bathonian in the upper.

604

605 The upper boundary of the Ravenscar Group is defined by the base of the
606 marine Cornbrash Formation or, where absent, by the base of the Osgodby
607 Formation (Powell *et al.* 1992; Gaunt *et al.* 1980). The overlying marine succession
608 represents a condensed sequence of Callovian age (Wright 1977), equivalent to the
609 Upper Cornbrash of the succession on the East Midlands Shelf and southwards to
610 Dorset (Page 1989). The latter author renamed the berthierine-rich limestone unit as
611 the Fleet Member of the Abbotsbury Cornbrash Formation, but this name and earlier
612 terminology are used variably in Rawson and Wright (2000), and the traditional name
613 is preferred here (cf. Douglas & Arkell 1932).

614

615 The base of the Cornbrash Formation marks the base of the Callovian Stage
616 (Macrocephalus Zone) in the Cleveland Basin. The formation consists of oyster-rich
617 sandy limestone with berthierine ooids. This distinctive unit is about 1 m thick on the
618 coast (Cayton Bay) and up to 3.6 m in Newtondale (Wright 1977; Page 1989), but is
619 absent at Brown Moor on the northern flank of the Market Weighton High (Gaunt *et al.*
620 1980). The Cornbrash Formation has not been positively identified in the
621 Hambleton Hills, where the basal sandstone (Redcliff Rock Member) of the Osgodby
622 Formation rests directly on the Scalby Formation (Senior 1975; Powell *et al.* 1992).
623 The formation is much thinner than the equivalent Upper Cornbrash of southern
624 England (Page 1989), where it forms a brashy (stoney) soil best suited to growing
625 corn (hence its name). The overlying **Cayton Clay Formation** (formerly 'Shales of
626 the Cornbrash') consists of dark grey calcareous mudstone and siltstone with
627 phosphatic nodules; ammonites indicate the Herveyi Zone (Wright 1978; Rawson &
628 Wright 2000). Recent boreholes (2009) at Knipe Point, near Osgodby, prove up to 4
629 m of Cayton Clay Formation overlying 1 m of Cornbrash limestone.

630

631 Sandstone and siltstone characterize the overlying **Osgodby Formation**
632 (Wright 1978) of Callovian age (Fig. 6). In typical Yorkshire coast sections, the
633 formation was subdivided into the following members, in ascending order: Kellaways
634 Rock (now the Redcliff Rock Member), Langdale Member and Hackness Rock
635 Member (Buckman 1913; Walker 1972; Wright 1968a, 1968b, 1978). The
636 stratigraphy of the Callovian (and Oxfordian) rocks on the Cleveland Basin has been
637 refined by Wright (1968a, 1977, 1978, 1983), particularly for the eastern part of the
638 Basin. Only the Redcliff Rock and the Hackness Rock are present in the Hambleton
639 Hills, where the Osgodby Formation ranges in thickness from 20 to 23 m (Powell *et*

640 *al.* 1992; Frost 1998), and at Brown Moor only 5.5 m of fine- to medium-grained,
641 poorly lithified, bioturbated sand with a marine fauna (belemnites and *Kosmoceras*)
642 are present. Wright's studies, and BGS re-surveys in the western part of the basin
643 (British Geological Survey 1992, 1994), have demonstrated two unconformities within
644 the Callovian succession. At the base, the **Redcliff Rock Member** (Page 1989),
645 named after Red Cliff, Cayton Bay, ranges from 11.5 to 23 m in thickness and
646 consists of orange, yellow and grey, fine- to medium-grained, thick-bedded
647 sandstone, locally with scattered berthierine ooids. Large bivalves and belemnites,
648 often preserved as decalcified moulds and casts, are conspicuous in some beds,
649 particularly in the upper part of the member. Some beds show cross-bedding and
650 the rock is usually soft and decalcified at outcrop. Vertical burrows and burrow-
651 mottling are common in some beds. Bivalves include the oysters *Gryphaea dilobotes*
652 and *Liostrea* sp., as well as *Chlamys fibrosa*, *Meleagrinnella braamburiensis*, *Trigonia*
653 sp. and *Unicardium* sp.; rhynchonellid brachiopods are also present. The Redcliff
654 Rock has yielded ammonites indicating the Koenigi Zone (Page 1989) and is
655 equivalent in part to the Kellaways Clay Member of southern England. The **Langdale**
656 **Member** (Wright 1968a, 1978) at Red Cliff, Scarborough [TA 07 84] is locally cut out
657 below the unconformable Hackness Rock. In the Hackness Hills, it consists of about
658 15 m of greenish brown, fine- to medium grained sandstone and siltstone, often
659 heavily bioturbated with sparse chamosite ooids and clay laminae. On the coast
660 (Castle Hill; Osgodby Nab) a hard, brown fine-grained sandstone bed is present at
661 the base. The bivalve fauna is similar to that found in the underlying Redcliff Rock
662 Member, and belemnite guards are also common in places. Ammonites include
663 species of *Erymnoceras*, *Kosmoceras* indicating the upper Coronatum Zone. The
664 sandstone members, together with the overlying Hackness Rock Member, are
665 equivalent in part to the Peterborough and Stewartby members of the Oxford Clay
666 south of the Market Weighton High (Cox *et al.* 1993). The **Hackness Rock Member**,
667 where present, is about 3 m thick, and consists of buff-grey siltstone with alternating
668 soft and hard calcite-cemented bands; fossils include bivalves, belemnites and
669 sparse ammonites, the latter indicating the Athleta and Lamberti zones (Wright 1978;
670 Page 1989).

671

672 **4.3 Upper Jurassic Succession**

673

674 The lithostratigraphy of the Upper Jurassic in the Cleveland Basin was established by
675 Fox-Strangways *et al.* (1886) and Fox-Strangways (1892), and was later refined by
676 Wright (1972, 1983, 1996a, 1996b, 2009), who formalized the nomenclature and
677 provided a detailed chronostratigraphical framework based on ammonite zones (Fig.

678 7). The base of the Upper Jurassic is defined at the lower boundary of the Oxfordian
679 Stage (Cope *et al.* 1980*b*), which corresponds to the base of the Oxford Clay
680 Formation in the Cleveland Basin. The youngest Jurassic strata in the Cleveland
681 Basin, the Kimmeridge Clay, belong to the Pectinatus Zone (151 Ma) or possibly the
682 overlying Pallasioides Zone of the Tithonian Stage (Herbin *et al.* 1991; Ogg *et al.*
683 2008). Later Kimmeridgian sediments were removed during the late Jurassic-early
684 Cretaceous Cimmerian earth movements.

685

686 The Upper Jurassic rocks are wholly of marine origin and mark a continuation
687 of the major marine transgression that began during the Callovian Stage. Eustatic
688 sea-level rise in north-west Europe (Hallam 1988; Haq *et al.* 1988) was interrupted
689 locally by a regressive phase during the deposition of the Corallian Group,
690 comprising carbonates and calcareous sandstones, but culminated in restricted,
691 basinal environments with anoxic bottom-conditions during the deposition of the
692 bituminous Kimmeridge Clay.

693 The **Oxford Clay** ranges in thickness from 0 to 44 m and consists of grey-
694 green calcareous mudstone and silty mudstone. South of the Market Weighton High,
695 the formation comprises three members, the Peterborough, Stewartby and
696 Weymouth members in upwards succession, but only the Weymouth Member of
697 early Oxfordian age is present in the Cleveland Basin (Cox *et al.* 1993). The lithology
698 is more silt-rich compared to its occurrence on the East Midlands Shelf, where the
699 formation is about 70 m thick. In the Roulston Scar area of the Hambleton Hills, the
700 absence of the Oxford Clay is the result of uplift and subsequent sub-marine erosion
701 of the Oxford Clay, and in places the underlying Hackness Rock, prior to deposition
702 of the Lower Calcareous Grit.

703

704 An abundant ammonite fauna has been collected from a number of levels in
705 the Oxford Clay of the Hambleton Hills, and indicates the Mariae Zone,
706 Scarburgense Subzone (Cox *in* Powell *et al.* 1992). Other sections near the top of
707 the formation have yielded small casts of *Cardioceras praecordatum* Douvillé,
708 proving the later Praecordatum Subzone.

709

710 The **Corallian Group** ranges in thickness from 70 m to 150 m and
711 predominantly comprises ooidal and micritic limestone and calcareous, spiculitic,
712 fine-grained sandstone. The group is subdivided into three formations (Wright 1972,
713 1983, 1996*a*, 1996*b*) separated by disconformities, and spans the upper Lower,
714 Middle and lower Upper Oxfordian stages (Fig. 7: Wright 1980; Rawson & Wright
715 1992, 2000). Disconformities are also present within the formations over parts of the

716 basin and lateral lithofacies changes have enabled Wright to recognize numerous
717 impersistent members (Fig. 7). The group is equivalent to the mud-rich West Walton
718 Formation and the lower part of the Amphill Clay of the East Midlands Shelf.

719

720 At the base of the group, the **Lower Calcareous Grit (LCG)** crops out along
721 the upper part of the bold, west-facing escarpment of the Hambleton Hills and at
722 classical localities such as Castle Hill (Scarborough) and Filey Brigg (Rawson &
723 Wright 2000). The Lower Calcareous Grit ranges from 22 to 48 m thick in the
724 Hambleton and Howardian Hills, and reaches 50 m on the Yorkshire coast. It
725 consists predominantly of yellow, buff, fine- to medium-grained, calcareous
726 sandstone, with subsidiary beds and concretions of blue-grey, micritic limestone;
727 both lithologies are variably ooidal and peloidal. Siliceous spicules of the sponge
728 *Rhaxella perforata* form much of the clastic component (Sorby 1851; Wilson 1939;
729 Hemingway 1974), and diagenesis of these has produced secondary thin beds of
730 chert, particularly in the lower part of the formation. *Thalassinoides* burrows are very
731 common on bedding planes at some horizons; the backfilled burrows have a higher
732 spicule content and are more resistant to weathering, giving an irregular, nodular
733 appearance to weathered faces. The micritic limestone concretions reach up to 1.5
734 m diameter, and are locally concentrated in the upper part of the formation (the 'Ball
735 Beds' of Arkell 1945).

736 The contact between the LCG and the Oxford Clay is gradational, except on
737 the Roulston Scar 'block' where the Oxford Clay is absent. Near Sutton Bank,
738 between [SE 5156 8121] and [SE 5327 8206], the **Oldstead Oolite Member** (Wright
739 1980) is locally distinguished in the lower part of the LCG. It consists of grey to
740 yellow-grey, bioclastic, ooidal wackestone-grainstone, up to 11 m thick, and cross-
741 bedded in part. The base is an erosive, unconformable junction with the underlying
742 Redcliff Rock Member in the Raven's Gill area [SE 5295 8186] (Fig. 33). To the east,
743 in Shaw's Gill, the Oldstead Oolite overlies Oxford Clay with a sharp base. The
744 proportion of ooids (wackestone texture) decreases gradationally upwards through
745 passage to the spiculitic calcareous sandstone of the 'typical' Lower Calcareous Grit
746 (Powell *et al.* 1992; fig. 16), indicating increasing water depths through time.

747

748 The boundary between the LCG and the overlying Hambleton Oolite Member
749 (Coralline Oolite Formation) is gradational in the Hambleton Hills, the percentage of
750 ooids increasing upwards at the expense of spiculitic sandstone. However, farther
751 east around Givendale, Dalby Forest [SE 854 863] and at the Bridestones [SE 878
752 907], a poorly consolidated yellow sand unit, cross-bedded in part, with calcareous
753 concretions rich in bivalves and brachiopods and termed the Passage Beds Member

754 (Wright 1972) or Yedmandale Member (BGS 2000), is present below the Hambleton
755 Oolite, which has an erosive base. On the coast, at Filey Brigg, the Passage Beds
756 Member consists of bioturbated calcareous sandstone interbedded with grey
757 limestone. The beds are rich in shell debris, including *Nanogyra* and *Gervillella*.
758 Cross bedding indicates a south-east palaeoflow (Wright 1992). In addition to the
759 fauna noted above, the formation has yielded a benthic assemblage that includes
760 bivalves and brachiopods (*Avicula*, *Pecten*, *Trigonia*, *Modiola*, *Ostrea* and
761 *Rhynchonelloidea*; Hemingway 1974), but these are rarely well preserved. The
762 carbonate concretions contain the richest fauna and have yielded many large, well
763 preserved ammonites that indicate the Bukowskii Subzone of the Cordatum Zone
764 (Fig. 7; Wright 1980).

765

766 The **Coralline Oolite Formation** (Wright 1972) comprises the following five
767 members, in upward sequence: Hambleton Oolite, Birdsall Calcareous Grit, Middle
768 Calcareous Grit, Malton Oolite and Coral Rag (Fig. 7). The estimated thickness of
769 the formation ranges from 60 to 70 m. The Coralline Oolite Formation consists of a
770 varied sequence of grey, predominately ooidal and peloidal limestone (ooidal
771 wackestone to ooidal grainstone texture) intercalated with wedges of buff-yellow,
772 sparsely ooidal, calcareous fine-grained sandstone. Subsidiary lithologies include
773 micritic limestone and reefal boundstone rich in corals and algae. Over most of the
774 Cleveland Basin, from Scarborough in the east to Northallerton in the west, the
775 stratigraphical relationship of the members assumes a 'layer-cake' sequence (Wright
776 1972; Hemingway 1974, fig. 53). As the formation is traced from the north-west of
777 the district to the south-east and beyond to the Howardian Hills, however, lateral
778 changes in lithofacies are prevalent, particularly in the lower three members (Fig. 7).
779 South-east of Murton Common [SE 509 885], the Hambleton Oolite is separated into
780 'upper' and 'lower' leaves by the intervening Birdsall Calcareous Grit (Wright 1972).
781 On parts of Byland Moor, south of Cold Cam [SE 542 813], the ooidal limestones
782 cannot be traced and there is a continuous sequence of calcareous spiculitic
783 sandstone from the top of the Lower Calcareous Grit through the Birdsall Calcareous
784 Grit up to the base of the Middle Calcareous Grit (Fox-Strangways *et al.* 1886; Powell
785 *et al.* 1992), the last being marked by a topographical feature. The top of the
786 formation is defined by the base of the Upper Calcareous Grit (Wright 1972) which
787 rests disconformably on the Coral Rag Member.

788

789 The **Hambleton Oolite Member** (up to 34 m thick) caps the escarpment of
790 the Hambleton Hills and forms extensive dip slopes north of Pickering on the North
791 Yorks Moors. It consists of pale grey to white ooidal limestone (packstone to

792 grainstone texture), with a variable proportion of quartz sand, peloids and fragmented
793 shells; chert nodules are common in places. Thin beds of calcareous sandstone with
794 scattered ooids are present in the southern part of the outcrop. Cross-bedding and
795 shallow scours are locally common in the ooidal limestone and the beds are
796 frequently penetrated by circular, vertical burrows, up to 1 cm in diameter. Wright
797 (1972) showed that the oolite member splits into an upper and lower leaf in parts of
798 the Hambleton Hills (Powell, *et al.* 1992) and in the Howardian Hills (S. Price *pers.*
799 *comm.* 2008; Wright 2009). Penecontemporaneous slump structures and injection
800 phenomena (Fig 10b; Hemingway & Twombly 1963; Powell *et al.* 1992) are locally
801 present at Shaw's Gate Quarry [SE 5233 8236] and Old Byland Grange Quarry [SE
802 5454 8567]. The fauna includes the ammonites *Cardioceras*, *Goliathiceras*,
803 *Aspidoceras* and *Perisphinctes*, as well as sporadic bivalves including *Exogyra*,
804 *Lima*, *Astarte*, *Ostrea*, *Modiola* and *Pholadomya*. Echinoids are common in some
805 beds and include *Cidaris*, *Nucleolites* and *Hemicidaris* (Hemingway 1974). Rare
806 specimens of *Rhaxella perforata* and a brittle-star have been collected. Ammonites
807 indicate an age ranging from the Cordatum Subzone to the Vertebrale Subzone,
808 spanning parts of the Cordatum and Densiplicatum zones (Fig. 7; Wright 1972). As
809 noted above, the sharp erosive base of the Hambleton Oolite on the coast suggests
810 a disconformity, and Wright (1972) has demonstrated that where the Passage Beds
811 Member is absent in the west of the outcrop, the Costicardia Subzone is missing.

812

813 The **Birdsall Calcareous Grit Member** (Cordatum Subzone) is a yellow-buff,
814 calcareous, fine-grained spiculitic sandstone with scattered ooids and lenses of grey
815 chert, which was deposited coevally with the Hambleton Oolite. It is up to 12 m thick
816 in the Hambleton Hills, but reaches 30 m in the Howardian Hills to the south,
817 suggesting a provenance from that direction. Nodular texture is common and is due
818 to abundant silica-rich *Thalassinoides* burrow-fill; *Chondrites* burrows are locally
819 present in thin-bedded siltstone. The Birdsall Calcareous Grit has yielded the
820 subzonal ammonite *Cardioceras cordatum* (Wright 1972, 2009) as well as bivalves,
821 including *Chlamys fibrosa*.

822

823 The Birdsall Calcareous Grit is well exposed between Cleave Dyke Quarry
824 [SE 507 863] and Boltby Scar [SE 506 857], but wedges out along the main
825 escarpment south of Boltby Scar, and also along the Caddell valley, so that the upper
826 and lower 'leaves' of the Hambleton Oolite are not distinguishable there (Powell *et al.*
827 1992, fig. 18).

828

829 The **Middle Calcareous Grit Member** (Wright 1972) crops out in the south-
830 east of the Hambleton Hills, on Byland Moor [SE 54 81], where it is about 12 m thick.
831 It is similar in lithology to the Birdsall Calcareous Grit, and the rock is often
832 decalcified at outcrop, so that only relict ooids can be seen. The unit probably
833 belongs to the upper part of the Vertebrale Subzone and the lower part of the
834 Maltonese Subzone (Figs 7, 36; Wright 1980).

835

836 The **Malton Oolite Member** (up to 20 m thick), formerly known as the
837 Osmington Oolite, separates the Middle Calcareous Grit from the stratigraphically
838 higher Coral Rag Member (Figs 7, 36; Wright 1972), and comprises variably shelly,
839 ooidal limestone. Quarries in the Malton area show large scale cross-bedding,
840 indicating deposition as laterally migrating ooidal shoals, similar to parts of the
841 present-day Bahama Banks (Twombly 1964). Sparse ammonites indicate the
842 Antecedens Subzone (Wright 1972; Rawson & Wright, 1992).

843

844 The uppermost unit, the **Coral Rag Member** (up to 9m thick), belongs to the
845 Parandieri Subzone (Wright 1972), and comprises coral-algal patch reefs, coral-shell
846 inter-reef debris and micritic limestone; both fore-reef and off-reef bioclastic (ooidal-
847 coral-shell) debris with the echinoid *Hemicidaris* and the oyster *Lopha* are common.
848 Isolated patch reefs had a relief of up to 3.5 m high above the surrounding substrate
849 (Twombly 1964; Hemingway 1974). Similar patch reefs are found in the Ayton area
850 [TA 002 856] (Hemingway 1974), although Rawson & Wright (2000) regarded this
851 locality as uppermost Malton Oolite. There, colonial corals include *Thamnasteria*,
852 *Isastraea*, *Rhabdophyllia* and *Thecosmilia*, in life position and as abraded fragments.
853 In the Hambleton Hills, the member was formerly seen in a small inlier exposed by
854 quarrying of the overlying Upper Calcareous Grit at Snape Hill Quarry [SE 508 787].
855 Fox-Strangways *et al.* (1886, p. 367) recorded up to 1.4 m of Coral Rag ('crystalline
856 limestone') with the colonial coral *Thecosmilia annularis*, the echinoid *Hemicidaris*
857 *florigemma* and the oyster *Lopha gregarea*. Farther east, Twombly (1964)
858 recognized a similar micrite-biomicroite facies rich in corals, echinoids and bivalves,
859 which he interpreted as a back-reef facies that developed to the north of a reefal
860 boundstone facies typically found in the Howardian Hills.

861

862 As the name suggests, the **Upper Calcareous Grit Formation** marks a
863 return to spiculitic sand sedimentation. It is between 12 and 15 m thick and consists
864 of very fine- to fine-grained, calcareous, spiculitic sandstone and siltstone, with
865 abundant beds of clayey, micritic limestone in the middle of the unit. In the Asenby –
866 Coxwold Graben, the clayey carbonate lithofacies is at least 6 m thick and is

867 equivalent to the **North Grimston Cementstone** facies of Wright (1972, 1980), who
868 assigned it to the Nunningtonense subzone. Farther east, in the Kirkdale to Pickering
869 outcrop, the formation was divided into three members by Wright (1972), spanning
870 the Nunningtonense Subzone to early Serratum Zone. In upward sequence, these
871 are the **Newbridge Member**, **Spaunton Sandstone** and **Snape Sandstone**. The
872 Newbridge Member consists of buff, thin-bedded siltstone, marl and fine-grained
873 sandstone. The Spaunton Sandstone is a buff, thin-bedded, bioturbated, calcareous
874 sandstone with abundant sponge spicules and siliceous nodules. The fauna includes
875 belemnites and sparse bivalves. Ammonites collected from the Spaunton Sandstone
876 indicate the Glosense Zone (Wright 1983; Sykes & Callomon 1979; Cox *in* Powell *et*
877 *al.* 1992). The Snape Sandstone Member is about 8 m thick, and consists of buff,
878 flaggy, cross-laminated siltstone and fine-grained sandstone with abundant
879 ammonite fragments and, locally, bioclastic limestone. Ammonites collected from the
880 Snape Sandstone indicate the Serratum Zone of the Upper Oxfordian (Wright 1972,
881 1980; Cox *in* Powell *et al.* 1992). The junction between the Upper Calcareous Grit
882 and the overlying Upper Jurassic clays (Amphill Clay and Kimmeridge Clay
883 formations) is a burrowed, gradational boundary (Cox & Richardson 1982)

884

885 In the Cleveland Basin, the Kimmeridge Clay was formerly thought to overlie
886 the Upper Calcareous Grit directly (Fox-Strangways *et al.* 1886), but more recent
887 studies of outcrops and borehole cores from the western end of the Vale of Pickering
888 (Cope 1974; Richardson *in* Institute of Geological Sciences 1974; Pyrah 1977; Cox
889 and Richardson 1982; Wignall 1993) show that mudstone (c. 48 m thick) with
890 subsidiary beds and nodules of siderite, equivalent to the **Amphill Clay** of southern
891 England and spanning the Upper Oxfordian Serratum, Regulare and Rosenkrantzi
892 zones (Fig. 7), is present between the top of the Corallian Group and the base of the
893 Kimmeridge Clay. The Amphill Clay is probably present at depth in the Asenby-
894 Coxwold Graben, and its correlative, the Woodward Formation, has been proved (22-
895 90 m thick) in the southern North Sea (Cox *et al.* 1987; Lott & Knox 1994).

896 An abrupt change in geophysical log signatures at the top of the Corallian
897 Group in the Hunmanby Borehole (Fig. 34) suggests that the Amphill Clay is faulted
898 out here (Whittaker *et al.* 1985), but an attenuated succession (25 m thick) is present
899 in the Brown Moor Borehole below the Cretaceous unconformity (Fig. 34). The
900 Amphill Clay has yielded infaunal and epifaunal bivalves, gastropods and echinoid
901 spines, suggesting that it was deposited in an oxic shallow marine environment in
902 response to rising sea-level and increased subsidence of the Cleveland Basin, as
903 mud and silt sediments spread to the north of the basin in Serratum Zone time, about
904 157 Ma (Fig. 37). The presence of co-eval mud-dominated sediments of

905 Tenuiserratum Zone age (equivalent to the uppermost Coralline Oolite Formation)
906 towards the south (Brown Moor Borehole; Fig. 34) indicates the gradual diachronous
907 younging of mud sedimentation to the north as the basin gradually subsided.

908

909 The **Kimmeridge Clay** (Baylei to Pectinatus and Pallasioides zones) is the
910 youngest Jurassic formation in the Cleveland Basin (Fig. 7). The Kimmeridge Clay
911 succession comprises, shelly mudstone, often bioturbated, interbedded with
912 bituminous (oil-rich) mudstone; abundant small "*Discinisca latissima*" and ammonite
913 fragments are preserved in some beds. The succession represented by the
914 Cymodoce, Mutabilis, Eudoxus, Autissiodorensis, Elegans, Scitulus and
915 Wheatleyensis zones comprises cycles of mudstone, bituminous mudstone, and
916 coccolith-rich limestone (Fig. 38). Above the Eudoxus Zone the formation becomes,
917 overall, more calcareous and less fissile; hard beds with low gamma-ray and high
918 sonic signatures comprise coccolith-rich limestone laminae. Outcrops are sparse, but
919 dark grey fissile mudstone with yellow-brown weathering, organic-rich laminae crop
920 out in the east of the Asenby-Coxwold Graben, and in the Vale of Pickering where
921 the Amphill and Kimmeridge formations form low ground, largely covered by
922 superficial deposits; they have also been proved in a number of groundwater and
923 hydrocarbon exploration boreholes (Fox-Strangways *et al.* 1886; Falcon & Kent
924 1960; Cox 1982; Cox *et al.* 1987; Herbin *et al.* 1991, 1993, 1995). The Kimmeridge
925 Clay is intermittently exposed in the Vale of Pickering below thin Devensian till near
926 Low Pasture House [SE 5540 7830], Riseborough Bridge [SE 7568 8428] and at
927 Brink Hill [SE 540 786] where it was worked for brick clay. Comparison of the
928 Kimmeridge Clay sequences recorded in boreholes in the Vale of Pickering and the
929 southern North Sea (Cox *et al.* 1987; Herbin *et al.* 1991, 1993) suggests that the
930 beds at Brink Hill are younger than the Eudoxus Zone. Ammonites collected from the
931 uppermost Kimmeridge Clay in boreholes on the north side of the Vale of Pickering
932 (Herbin *et al.* 1991) suggest the Pallasioides Zone (Fig. 7).

933

934 On the coast, the Kimmeridge Clay is overlain unconformably by the Speeton
935 Clay Formation (Lower Cretaceous) at Speeton Sands [TA 140 763], where c. 10 m
936 of dark grey, finely laminated mudstones contain ammonites that indicate the
937 Hudlestoni to lower Pectinatus zones (Rawson & Wright 2000). The unconformity
938 represents a considerable time gap, spanning the late Kimmeridgian (Pectinatus
939 Zone and higher) and the Portlandian, plus the early part of the Cretaceous, so that
940 the Upper Ryazanian 'D' Beds of the Speeton Clay rest on the Kimmeridge Clay.

941

942 The Kimmeridge Clay represents the main source of hydrocarbons in the
943 northern part of the North Sea, where it has been buried to greater depths and at
944 higher pressure than onshore and hence is mature enough to allow migration of
945 lighter hydrocarbons to reservoir rocks such as the Brent sandstones (Herbin *et al.*
946 1993). Although the Kimmeridge Clay in the Cleveland Basin and Sole Pit Basin is
947 bituminous, it is not sufficiently mature to have been a major source of hydrocarbons
948 in these areas.

949

950 The Fordon No. 1 Borehole (Falcon & Kent 1960) proved 385 m of
951 Kimmeridge Clay, but this figure has been questioned by Cox *et al.* (1987) who noted
952 that the Amphill Clay was included in the total thickness in this borehole; they
953 suggest that the total thickness of the formation in the Vale of Pickering is about 305
954 m. However, the unit is much thicker here than the attenuated and eroded
955 succession south of the Market Weighton High where, below the unconformable
956 Cretaceous Carstone Formation, 7.5 m of Kimmeridge Clay belonging to the Baylei
957 and Cymodoce zones, overlie the Amphill Clay (Gaunt *et al.* 1992). The formation
958 thickens southwards, reaching c.115 m in The Wash area (Gallois 1994).

959

960 Lateral lithofacies changes and attenuation of the Amphill/Kimmeridge Clay
961 succession southwards towards the Market Weighton High are illustrated by the
962 Hunmanby [TA 131 759] and Brown Moor [SE 813 620] boreholes (Fig. 34)
963 (Whittaker *et al.* 1985). At Hunmanby, the logs suggest that the Kimmeridge Clay
964 lithofacies did not extend northwards until Cymodoce Zone times, whereas farther
965 south at Brown Moor, near Market Weighton, the Amphill Clay lithofacies extends
966 from Tenuiserratum Zone times (Malton Oolite equivalent) to Serratum Zone times,
967 the Kimmeridge Clay being highly attenuated or not present due to pre-Cretaceous
968 erosion over the high. However, the sharp boundary (Fig. 34) between the Corallian
969 Group and the overlying mudstone of Cymodoce Zone age (c.f. lower Kimmeridge
970 Clay) may indicate a faulted boundary here, with the Amphill Clay cut out.

971

972 Offshore in the Southern North Sea Basin (Fig. 7), the post-Callovian
973 succession (200-300 m thick) is defined as the Humber Group (Lott & Knox 1994).
974 The subtle lithostratigraphical characteristics that allow the sequence to be
975 subdivided onshore are not so apparent in the offshore geophysical logs. Offshore
976 equivalents of the Corallian Group are represented by calcareous sandstone overlain
977 by ooidal limestone (Corallian Formation; 70-100 m thick), but this passes laterally
978 southwards to the mudstone dominated Seeley Formation (equivalent to the West
979 Walton Formation of the East Midlands Shelf). Overlying these is the Woodward

980 Formation (c. 50 m thick), a mudstone unit of late Oxfordian age broadly equivalent
981 to the Amphill Clay (Cox *et al.* 1993), overlain by the Kimmeridge Clay Formation,
982 which is about 250 m thick in the Sole Pit Basin.

983

984

985 **5. JURASSIC SEDIMENTATION IN THE CLEVELAND BASIN: INTERPLAY OF** 986 **TECTONICS, RELATIVE-SEA LEVEL AND CLIMATE**

987

988 The Jurassic succession of the Cleveland Basin, with its broad record of terrestrial,
989 shallow marine and relatively deep-water sedimentation, provides an opportunity to
990 assess the relative importance of the role of global and relative sea-level change,
991 tectonics, climate and sediment flux over about 49 million years of Earth history
992 (Hettangian 199.6 Ma to Late Kimmeridgian 150.8 Ma) (Ogg *et al.* 2008). However,
993 application of the concepts of classical sequence stratigraphy (Van Wagoner *et al.*
994 1988; Haq *et al.* 1988) is difficult because successions preserved in the Cleveland
995 Basin do not record the full range of tectonic and environmental settings. The Early
996 Jurassic marine succession (Lias Group), for example, does not show the shoreline
997 or terrestrial lithofacies, and conversely the non-marine units of the Ravenscar Group
998 cannot be traced laterally to shoreline and deeper water facies. Consequently, for a
999 large part of the succession, only changes in relative sea-level such as upward
1000 shoaling parasequences, cycles and breaks in sedimentation can be deduced (van
1001 Wagoner *et al.* 1988; Knox *et al.* 1991; Coe 1995). Generally, it is not possible to
1002 trace these events laterally to show coastal onlap or offlap, or subaerial
1003 unconformities in coeval settings. We can, however, interpret the fluctuation in
1004 relative sea-level within the Cleveland Basin from the sedimentary record, and
1005 assess the importance of global sea-level fluctuations against the role of intra- and
1006 extra-basinal tectonics. Comparison of the much studied coastal exposures with
1007 lesser known areas in the west of the basin throws more light on the
1008 palaeogeography and tectonics, especially in Mid-Jurassic times.

1009

1010 In this section, the relative importance of these intra- and extra-basinal factors
1011 will be assessed against an outline of significant events, and their sedimentological
1012 characteristics illustrated by the Jurassic succession.

1013

1014 **5.1 Lias Group: tempestites, shoaling cycles, and anoxia**

1015

1016 A shallow epicontinental sea extended throughout northwest Europe in Late Triassic
1017 times, following deposition of the brackish to shallow marine Penarth Group

1018 (Warrington *in* Powell *et al.* 1992; Ivimey-Cook & Powell 1991). The Lias Group was
1019 deposited as two major, upward shoaling cycles: the Redcar Mudstone to Staithes
1020 Sandstone cycle and the Cleveland Ironstone to Blea Wyke Sandstone cycle (Knox
1021 *et al.* 1991). The first cycle has been subdivided into 3 second-order cycles by Van
1022 Buchem & Knox (1998), represented by the Planorbis-Liasicus, Liasicus-Jamesoni
1023 and Jamesoni-Ibex intervals, the base of each cycle marking a significant rise in sea-
1024 level. Hesselbo & Jenkyns (1995, 1998) recognized smaller, third-order cycles of
1025 about 300-500 Ka, which approximate to the estimated duration of many ammonite
1026 zones and subzones, and suggest a link between cyclicity, sea-level change and
1027 extinction or faunal turn over. Each of the third-order cycles was further subdivided
1028 into a number of smaller scale cycles or parasequences that were deposited in
1029 response to tectonic control and sediment flux. These cycles can be recognized in
1030 the gamma-ray and sonic geophysical logs of the Felixkirk Borehole (Fig. 9) (Ivimey-
1031 Cook & Powell 1991).

1032

1033 The Hettangian to early Pliensbachian Redcar Mudstone comprises
1034 stratigraphically expanded and reduced successions (van Buchem & McCave 1989;
1035 Ivimey-Cook & Powell 1991). Background sedimentation was predominantly
1036 hemipelagic mud, but during periods of low sedimentation rates, condensed
1037 sequences rich in iron ooids, glauconite or winnowed shell fragments were
1038 deposited, especially on highs such as the Market Weighton area. Early Hettangian
1039 fair-weather sedimentation (lower Calcareous Shales) is represented by hemipelagic
1040 mud, but periods of deposition within fairweather wave base are indicated by thin,
1041 winnowed laminae comprising thin shelled bivalves (Fig. 11a). The absence of
1042 bioturbation at this level suggests that sedimentation rates were high and/or that the
1043 substrate was not sufficiently oxygenated for colonization by epifauna and infauna.
1044 However, the gamma-ray inflections of the Felixkirk Borehole log (Figs 9, 11a)
1045 indicate a significant change up-sequence. The upper part of the Calcareous Shales
1046 is characterized in this borehole (Fig. 9) and in coastal exposures at Redcar and
1047 Robin Hood's Bay by 0.10 m to 0.40 m thick and, exceptionally, 1 m thick, beds of
1048 coarser grained calcareous siltstone and fine-grained sandstone beds with abundant
1049 bivalve fragments and disarticulated shells (mostly *Gryphaea*) in various orientations
1050 (Fig. 11b,c). The coarse-grained beds have sharp bases, often with erosional scours,
1051 but the junction often appears gradational due to downward penetrating burrows
1052 (especially *Chondrites*) that give a superficial appearance of gradational upward
1053 shoaling (Sellwood 1970). Multiple cycles are also present (Fig. 11d,c) and the tops
1054 of these beds show a sharp return to background mud sedimentation, although
1055 occasional winnowed siltstones are present in the upper few centimetres. These

1056 beds are interpreted as the result of storm-generated waves and bottom currents that
1057 redeposited calcareous sand from the nearshore zone as bioclastic sand layers
1058 (tempestites) resulting from powerful storm-surge and ebb currents (Aigner &
1059 Reineck 1982). The storm-generated beds were deposited rapidly and punctuate the
1060 quieter, slower background sedimentation represented by grey mud and silt, often
1061 rich in nektonic ammonite/belemnite faunas (Buckman 1915; Bairstow 1969; Ivimey-
1062 Cook & Powell 1991; Knox *et al.* 1991; van Buchem & McCave 1989). Deposition
1063 was in the form of laterally continuous bioclastic sheet sand, traceable over many
1064 kilometres, possibly resulting from major hurricanes driving sea water onto the
1065 coastal zone (storm surge), followed by ebbing offshore bottom currents (with
1066 erosional bases) or sediment suspension clouds.

1067

1068 Similar 'tempestite' beds characterize the overlying Siliceous Shales, the
1069 base of which corresponds to the mid Turneri Zone in the Felixkirk Borehole but is
1070 slightly higher at Robin Hood's Bay (Hesselbo & Jenkins 1995). As the name
1071 suggests, the coarse-grained beds comprise fine-grained sand but few shells.
1072 Sedimentary structures such as scoured basal contacts and low-angle cross-
1073 lamination indicate high-energy bottom currents (Fig. 12). However, the primary
1074 sedimentary structures are often obscured by extensive bioturbation, including
1075 *Diplocraterion*, *Rhizocorallium*, *Teichichnus* and *Chondrites* burrows that 'piped'
1076 sediment downward, below the scoured erosional base, again giving a superficial
1077 appearance of an upward shoaling succession (Fig. 12). There is no indication that
1078 the individual coarse tempestite beds represent upward coarsening (shallowing)
1079 sequences (Sellwood 1970). The change from the shell-rich to sand-rich storm beds
1080 suggests provenance of the coarse-grained fraction from different shoreface facies or
1081 was perhaps a result of greater sand flux into the nearshore zone during the late
1082 Sinemurian. Although early Lias Group sedimentation is characterized by alternating
1083 hard/soft beds in both southern England and the Cleveland Basin, the Redcar
1084 Mudstone differs markedly from the climate- or diagenetically controlled 'rhythmic
1085 couplet' sedimentation (Blue Lias) of the Dorset and South Wales provinces (Hallam
1086 1964; House 1985; Weedon 1986; Weedon & Jenkyns 1990). Sedimentation in the
1087 Cleveland Basin was influenced by tectonic uplift of the Pennine High hinterland (Fig.
1088 2), increased siliciclastic sediment flux and storm events that redistributed coarser
1089 material (bioclasts and sand) offshore from the nearshore sublittoral zone. The
1090 presence of iron-rich ooids and glauconite (Fig. 11c) in some of the sandy layers
1091 suggests shoaling conditions during condensed phases, possibly a result of a relative
1092 sea-level fall (Knox *et al.* 1991; Hesselbo & Jenkyns 1995). The change in storm

1093 sedimentation sequences from the lower part of the Calcareous Shales to the
1094 Siliceous Shales as seen in the Felixkirk cores is summarized in Fig. 12.

1095

1096 5.1.1 Early Pliensbachian sea-level rise and climatically induced sedimentation

1097

1098 The Sinemurian-Pliensbachian boundary (base Jamesoni Zone) marks a major sea
1099 level rise (Sellwood 1972; Knox *et al.* 1991; Hesselbo & Jenkyns 1995; van Buchem
1100 & Knox 1998) and is reflected clearly in the 'smooth' gamma and sonic log interval
1101 above 164.38 m depth in the Felixkirk borehole (Fig. 9). The same deepening event
1102 is clearly visible in the GSSP section at Wine Haven, Robin Hood's Bay (Knox *et al.*
1103 1991; Meister *et al.* 2003). The global sea-level rise is manifested in deposition of
1104 dark grey to black hemipelagic mud, deposited in quiet bottom conditions that
1105 resulted in local anoxia on the sea-floor, below storm wave base. Coarse-grained
1106 storm layers are absent, but the sediment interface was not wholly anoxic or
1107 inimicable to life as semi-infaunal bivalves such as *Pinna* sp. are found partially
1108 buried, *in situ*, together with pyritized burrows and concretions and a nektonic fauna
1109 of pyritized belemnites (locally current aligned) and ammonites. The last include the
1110 zonal fossil *Echioceras raricostatum* in the lower part. Partial burial of large *Pinna*
1111 shells in life position indicates rapid sedimentation rates in a distal offshore setting.

1112

1113 The section in Robin Hood's Bay and the Felixkirk Borehole logs show a
1114 gradual increase in the sand:mud ratio above the Taylori Subzone, in the form of
1115 discrete pale/dark banding, and the lower part of the Ironstone Shales is locally
1116 known as the 'Banded Shales' (van Buchem & McCave 1989). Sand is present as
1117 thin (2-3cm) graded and delicately laminated beds, but unlike the Siliceous Shales,
1118 there is no evidence for erosive traction currents. The Banded Shales represent
1119 distal hemipelagic sedimentation with silt and fine-grained sand being introduced into
1120 the deeper parts of the basin. Isolated lenses of sand preserved as large 'gutter
1121 casts' with low angle cross-lamination suggest periodic fluxes of sand that were
1122 dispersed offshore from the littoral zone. Again in contrast to the Siliceous Shales,
1123 these sand lenses are not bioturbated, suggesting rapid sedimentation rates and little
1124 time for colonization by benthic and infaunal organisms. Lighter bands probably
1125 represent the fallout of plankton and silt-grade sediment during more intense
1126 weathering of the hinterland. Regular banding in these beds has been attributed to
1127 alternating climatic cycles (van Buchem *et al.* 1992, 1994); analysis of the cycles
1128 suggest a regular periodicity, possibly linked to orbitally induced Milankovitch climatic
1129 cycles of perhaps 26 Ka periodicity (precession cycles). In contrast to the cycles
1130 induced by storm events in the lower part of the succession (Calcareous Shales and

1131 Siliceous Shales), the Banded Shales cycles are therefore probably the result of
1132 astronomically induced climatic variations.

1133

1134 A glauconitic-oidal bed at the base of the Ibex Zone (Figs 9, 14) in Robin
1135 Hood's Bay marks a depositional hiatus at the top of the Banded Shales, and the
1136 overlying Ironstone Shales show a strong upward-coarsening gamma-ray signature
1137 as pulses of sand were deposited offshore as distal pro-delta deposits, a precursor to
1138 progradation of the littoral sands of the overlying Staithes Formation (near base of
1139 the Davoei Zone; Fig. 9).

1140

1141 *5.1.2 Sandy tempestites and upward shoaling cycles*

1142

1143 The base of the Staithes Sandstone at 94.85 m in the Felixkirk Borehole and Bed 1
1144 at Staithes (Howarth 1955; Powell 1984; Howard 1985) marks an influx of shallow
1145 marine sand that extended over much of England during late Pliensbachian times
1146 (Davoei Zone). Individual beds have shallow, scoured erosive bases and parallel and
1147 low-angle cross-lamination, and were deposited as extensive sublittoral and
1148 shoreface sands. Sand influx was probably due to shallowing of all the English sub-
1149 basins (e.g. Wessex Basin, East Midlands Shelf, Cleveland Basin) as a result of
1150 tectonic uplift of the source areas. The effects of this relative sea-level fall are more
1151 pronounced over the Market Weighton High and the East Midlands Shelf where the
1152 Staithes Sandstone equivalents (and the lower part of the Cleveland Ironstone) were
1153 removed by erosion prior to deposition of the Marlstone Rock Formation (broadly
1154 equivalent to the Cleveland Ironstone) in Spinatum Zone times (Fig. 16; Howard
1155 1985).

1156

1157 Sand-rich tempestites typified by the Staithes Sandstone beds are sheet-like
1158 in general form, with planar, low-angle and hummocky cross stratification; some beds
1159 have wave-rippled tops (Figs. 10 c, f.). Erosional gutter casts at the base of
1160 tempestite beds (Fig. 10d) indicate an east to west palaeoslope, and internal low-
1161 angle cross-lamination measurements suggest that dominant storm-surge currents
1162 flowed predominantly to the east. Sheet sands were deposited by storm-surge-ebb
1163 currents that re-distributed sand from the nearshore zone to offshore locations
1164 (Howard 1985). Consequently, beds often have sharp erosive bases, occasionally
1165 with shelly lags (Fig. 10e). Low angle cross-lamination and hummocky cross-
1166 stratification represent reworking by waning, wave-generated, oscillatory currents
1167 (Nottvedt & Kriesa 1987). These sedimentary structures are locally destroyed by
1168 intense bioturbation where the residence time of the sand was longer and the

1169 substrate colonized by infaunal organisms (Plate xx). However, where sedimentation
1170 rates were high, the internal structures are well preserved. The location offshore
1171 below fair-weather wave base also aided preservation of the internal structures (Plate
1172 xx). Ichnofossil (trace fossil) assemblages provide a further insight into
1173 palaeoenvironmental conditions during 'Staithes time' (Seilacher 1967; Howard 1985;
1174 Knox *et al.* 1991). Assemblages from shoreface environments are dominated by
1175 deposit feeders such as *Chondrites* and *Planolites*, which exploited nutrient-rich
1176 laminae as opposed to clean sand (Fig. 10f); assemblages from innermost shelf
1177 environments, below fair weather wave-base, are characterized by dwelling burrows
1178 of infaunal deposit feeders such as *Skolithos*, *Diplocraterion* and those with
1179 specialist sediment-mining behaviours such as *Rhizocorallium*, *Asterosoma* and
1180 *Thalassinoides*. Distinctive sediment-mining burrows, such as *Siphonites* and
1181 *Teichichnus*, are present in the mid-shelf zone; the outer shelf assemblage is again
1182 typified by traces of sediment mining activities of deposit feeding organisms.

1183

1184 Upward coarsening (shoaling) cycles in the Staithes Formation are well
1185 displayed between Bed 17 (Howarth 1955) and Bed 23 near Penny Nab, Staithes
1186 (Howard 1985). These trends are also apparent in the attenuated sequence in the
1187 Felixkirk cores (Fig. 9, above 82 m depth; Fig. 10d) and are interpreted as
1188 progradation of shoreface and inner-shelf sands during rising sea-level as
1189 accommodation space increased on the shelf, a pattern seen also in the overlying
1190 Cleveland Ironstone cycles.

1191

1192 5.1.3 Transgressive cycles and iron-rich lagoons

1193

1194 The Cleveland Ironstone Formation comprises a succession of ooidal ironstone
1195 (formerly mined seams) separated by mudstone and fine-grained sandstones. These
1196 fine-grained sediments were deposited in laterally extensive lagoons and shallow
1197 littoral seas during a period of much reduced input of terrigenous, siliciclastic
1198 sediment, probably as a result of rising sea-level (second-order rise) during the
1199 Margaritatus Zone, which blanketed the geomorphologically low-lying hinterland. The
1200 late Pliensbachian appears to have been a time of reduced tectonic uplift of the
1201 surrounding landmass, including the Pennine High and the Mid-North Sea High,
1202 which were the main sources of clastic sediment from Hettangian to early
1203 Pliensbachian time. However, Chowns (1966) showed that the ironstone-rich
1204 succession was deposited over a long time period (i.e. it is a condensed sequence)
1205 and included a major phase of folding and erosion that produced an intraformational
1206 unconformity below the Pecten Seam (base *Spinatum* Zone) (Chowns 1966;

1207 Hemingway 1974; Howard 1985). The ironstone seams and the intervening
1208 siliciclastic sediments thin towards the south (Market Weighton High) and the west
1209 (towards the basin margin) (Figs 15, 16, 18). In addition, the five seams below the
1210 base Spinatum Zone unconformity are cut out successively southwards as result of
1211 uplift that removed up to 25 m of beds (Figs. 18,19). Flexuring of the shallow marine
1212 basin may also have contributed to extensive shallow water conditions over the shelf
1213 during this period, and may have been related to gentle isostatic uplift of the Market
1214 Weighton High where the 'sub-Spinatum unconformity' is most pronounced (Howard
1215 1985) (Fig. 16). In the west of the basin, only the Osmotherly and Avicula seams are
1216 present below the Main Seam unconformity, represented by thin, condensed units
1217 exhibiting burrowing and boring, consistent with a long period of accumulation with
1218 little siliciclastic sediment input (Fig. 19).

1219

1220 The coastal succession at Penny Nab (Howarth 1955; Howard 1985; Knox *et*
1221 *al* 1991) is generally characterized by upward coarsening (shoaling) siliciclastic
1222 cycles, namely mudstone, siltstone and fine-grained sandstone (with tempestite
1223 laminae) capped by transgressive, berthierine-rich ooidal ironstone. The latter have
1224 sharp erosional bases, commonly with reworked bored siderite and phosphatized
1225 nodules and occasional shelly lags, overlain by intensely bioturbated ooidal
1226 ironstone. Berthierine (formerly chamosite) ooids are set in a matrix of
1227 microcrystalline siderite; ooids commonly exhibit crushing and unravelling of
1228 concentric layers (spastolithization) (Marley 1857; Hallimond 1925; Chowns 1966;
1229 Hemingway 1978, fig. 45). Iron, in colloidal form, was probably derived from
1230 dissolution of lateritic soils as a result of marine flooding of the low-lying hinterland
1231 during punctuated periods of sea-level rise. The Cleveland Basin lay at about 30
1232 degrees north during late Pliensbachian times, within the equatorial zone, and a hot
1233 climatic regime conducive to lateritic weathering is likely (cf. warm seasonal climate
1234 during the early Mid Jurassic; Morgans *et al.* 1999). Iron colloids may have been
1235 introduced into the shallow marine lagoons by rivers draining from the hinterland
1236 areas or through marine flooding of deeply laterized coastal areas; preservation of
1237 colloidal iron would have required reducing anoxic conditions, possibly a result of
1238 high levels of organic matter in the bottom sediment (Curtis & Spears 1968).

1239

1240 The lower part of the succession (Penny Nab Member, 19 m) on the coast,
1241 includes four upward coarsening (shoaling) parasequences, generally capped by
1242 transgressive ironstones (Osmotherly, Avicula, Raisdale and Two Foot seams) and
1243 an incomplete cycle between the Two Foot Seam and the unconformable base of the
1244 Pecten Seam (base Spinatum Zone) (Fig. 15). A typical cycle pattern is illustrated by

1245 the Avicula to Raisdale interval (Howarth 1955; Chowns 1966; Howard 1985; Knox *et*
1246 *al.*, 1991). The Avicula Seam (Fig. 17a,b) was deposited during a marine
1247 transgression and comprises berthierine-rich, ooidal ironstone with a basal lag
1248 deposit that includes worn bivalve fragments and reworked and bored siderite
1249 mudstone nodules with pyritized rims. The overlying upward coarsening (shoaling)
1250 phase comprises laminated mudstone with thin, graded, low-energy tempestites
1251 ('striped beds') locally, and with east-west orientated gutter-casts (Greensmith *et al.*
1252 1980) produced by strong, storm-generated, erosive, helicoidal bottom currents that
1253 flowed down a gentle eastward dipping palaeoslope. Similar gutter-casts are seen in
1254 the 'striped siltstones' immediately below the Raisdale Seam (Fig.17c), which, in turn,
1255 marks the next punctuated sea-level rise and marine transgression. Microfacies
1256 analysis of the Cleveland Ironstone (Macquaker & Taylor 1996) confirmed systematic
1257 upward-coarsening and upward-fining unit at various scales; the small-scale upward-
1258 coarsening trends (0.1 – 1.0 m thick) are attributed to shoaling parasequences whilst
1259 the large –scale (1.0 - 3.0 m thick) upward-fining and upward-coarsening packages
1260 are interpreted as retrogradational and progradational units, respectively. Ironstone
1261 beds are interpreted as marking the interval (sequence boundary) between
1262 progradational and retrogradational units; phosphate-rich horizons probably
1263 represent maximum flooding surfaces across the shallow lagoons.

1264

1265 Up sequence, the Kettlecess Member (10 m thick), which includes the Main
1266 Seam, oversteps successively younger Lias Group units to the south towards Market
1267 Weighton, and to the west (Fig. 16). Just north of the Market Weighton High, at
1268 Whitwell-on-the-Hill, it rests unconformably on beds as low as the Redcar Mudstone,
1269 but can be traced southwards over this structure to the East Midlands Shelf where it
1270 is equivalent to a highly condensed iron-rich sandstone unit, the Marlstone Rock
1271 Formation. In turn, the Marlstone Rock Formation can be traced southwards to the
1272 Wessex Basin where it is represented by the Dyrham Siltstone (Cox *et al.* 1999). In
1273 the Cleveland Basin, the condensed succession mostly comprises the Pecten and
1274 Main seams with thin siltstones, but south of Osmotherly even the Pecten Seam is
1275 cut out below the unconformity (Frost 1998), so that the Main Seam rests
1276 unconformably on siliciclastic beds above the Avicula Seam (Figs 16, 18,19). Overall,
1277 the Kettlecess Member/Marlstone Rock succession marks a basin-wide shallowing
1278 phase. The base 'base Spinatum unconformity', south and westerly thinning, and
1279 condensed ooidal ironstone seams suggest that shallowing is attributable to regional
1280 tectonic uplift rather than global sea-level fall. Condensed sequences of ironstone
1281 and siliciclastic sediments indicate that the hinterland areas were not uplifted nor
1282 supplying large volumes of sediment to the basin. If, as noted above, the iron was

1283 derived from the weathering of thick lateritic soils, it is likely that the marginal areas
1284 were geomorphologically low lying, so that minor sea-level rises that characterize the
1285 transgressive ironstone parasequences (small-cycles) extended over a wide
1286 geographical area. The presence of intensely bioturbated horizons, ooidal ironstone,
1287 sub-rounded phosphatic pebbles, locally bored surfaces and a prolific benthic fauna
1288 confirm shallow-water conditions. Minor isostatic tectonic fluctuations (e.g. uplift of
1289 the Market Weighton High) would therefore have had a profound effect over a wide
1290 area.

1291

1292 *5.1.4 Subsidence, anoxia and a second major shoaling cycle*

1293

1294 The Toarcian 'Upper Lias' succession marks a second major shallowing cycle (cf.
1295 Redcar Mudstone to top Staithes Sandstone cycle) comprising the Whitby Mudstone
1296 to Blea Wyke Sandstone formations (Dean 1954; Hallam 1967a; Powell 1984; Knox
1297 1984). Following a relatively quiescent tectonic phase during the late Pliensbachian,
1298 the early Toarcian (Tenuicostatum Zone) was a period of major basin subsidence
1299 throughout England. More rapid subsidence of the Cleveland Basin, resulting in
1300 relatively deeper open water conditions during deposition of the Grey Shales Member
1301 (Tenuicostatum Subzone), was accompanied by increased sediment flux. **Fluctuating**
1302 oxic and slightly anoxic bottom conditions are present through the Grey Shales but
1303 increasingly higher levels of organic carbon are present from the Semicelatum
1304 Subzone (**Figs 9, 20**) through the Exaratum Subzone (Falciferum Zone), spanning
1305 the uppermost Grey Shales and the 'Jet Rock' (Mulgrave Shale Member). The high
1306 organic carbon content has been long recognized as indicating ocean-wide anoxia at
1307 the sediment-water interface (Hallam 1967a, 1967b; Morris 1979; Jenkyns 1988;
1308 Jenkyns & Clayton 1997). Minor shoaling cycles with striped siltstone laminae (**Figs**
1309 **17d, 20**) suggest that water depths were in the region of tens of metres, similar to
1310 depths for the Siliceous Shales. However, anoxia on the ocean floor may have been
1311 due to a density-stratified ocean (O'Brien 1990) with little overturn of the water
1312 column, resulting in stagnant bottom conditions. Prolific ammonite faunas testify to
1313 oxygenated water above, conducive to nektonic faunas. The cause of the Toarcian
1314 Oceanic Anoxic Event (OAE) anoxia during Exaratum Subzone time in NW Europe
1315 has been the subject of much debate. Hesselbo *et al.* (2000) demonstrated that
1316 isotopically light carbon isotopes were present ocean-wide and they attributed this to
1317 a discharge of methane from the overturn of gas hydrate (a solid ice-like substance
1318 rich in biogenic methane found today on deep continental marine margins, e.g.
1319 Antarctica). Further studies using osmium, carbon and strontium isotopes (Jenkyns
1320 *et al.* 2002; Cohen *et al.* 2004; Kemp *et al.* 2005) have demonstrated a sharp

1321 negative excursion in isotope ratios, such as the short-lived carbon ($\delta^{13}\text{C}_{\text{org}}$) isotope
1322 excursion within the Exaratum Subzone that is broadly coincident with a high osmium
1323 ($^{187}\text{Os}/^{188}\text{Os}$) isotope ratio excursion (Fig. 20). The marked change in organic carbon
1324 isotope ratios occurs in three marked steps at the base of the Falciferum Zone
1325 (Exaratum Subzone) and these steps are inferred to coincide with a marked
1326 reduction of benthic species in 'shifts' of minus 67% and minus 50%. High $\delta^{13}\text{C}_{\text{org}}$
1327 values result from an increase in the burial of biogenic carbon whereas low $\delta^{13}\text{C}_{\text{org}}$
1328 values result from oxidation of carbon in the sediment or introduction of reduced
1329 carbon. Kemp *et al.* (2005) attribute the anoxic phase to the release of carbon
1330 dioxide into the atmosphere as a result of overturn of methane hydrates normally
1331 trapped as a solid phase on the ocean floor. Both methane and, its oxidation product
1332 carbon dioxide, are significant 'greenhouse gases' and their release into the
1333 atmosphere may have caused global warming, in turn causing higher ocean
1334 temperatures, sea-floor anoxia and extinction of benthic, planktonic, and nektonic
1335 species worldwide. The high osmium isotope ratio over this interval was interpreted
1336 by Kemp *et al.* (2005) to have resulted from warm-climate, lateritic weathering of the
1337 hinterland (c.f. the origin of iron colloids in the Cleveland Ironstones). A subsequent
1338 fall in the osmium isotope ratio was interpreted to be due to subsequent take up of
1339 carbon dioxide in the atmosphere through the incorporation of the gas in the
1340 formation (weathering) of lateritic soils. However, alternative explanations have been
1341 proposed, first that the rapid negative $\delta^{13}\text{C}_{\text{org}}$ value excursions are attributable to
1342 recycling of isotopically light carbon from the lower water column in a local euxinic
1343 (oxygen-starved) epeiric sea (McArthur *et al.*, 2000; Wignall *et al.* 2006), and
1344 secondly, based on fossil leaf stomatal frequency, the fluctuations in carbon dioxide
1345 and the injection of isotopically light carbon into the atmosphere is a result of the
1346 release of thermogenic methane into the atmosphere generated through the intrusion
1347 of Gondwana coals by dolerites in the Karoo-Ferrar region (McElwain *et al.* 2005)

1348

1349 Renewed influx of fine-grained terrigenous siliciclastics kept pace with
1350 gradual subsidence of the basin during deposition of the overlying Alum Shales
1351 Member under more oxygenated conditions (Pye & Krinsley 1986). Periods of lower
1352 sedimentation rate are indicated by bands of phosphatic nodules in the upper part of
1353 the unit. A paucity of benthic fauna compared to typical Lias mudstones may have
1354 been due to climatic conditions. A shallowing, upward-coarsening trend and a
1355 concomitant increase in burrowing benthic fauna (including brachiopods and
1356 bivalves) (Knox 1984) is seen in the uppermost Toarcian sediments, only locally
1357 preserved (e.g. Peak Trough) below the Dogger unconformity. Within the Peak
1358 Trough and its onshore extension at Blea Wyke (Fig. 23), the shoaling trend (Figs 21,

1359 22) is indicated by the Peak Mudstone and Fox Cliff Siltstone and the uppermost unit
1360 of the Lias Group, the Blea Wyke Sandstone. The latter comprises fine-grained
1361 sandstone, and like much of the late Pliensbachian Staithes Sandstone, is heavily
1362 bioturbated, the reworking destroying most of the primary sedimentary structures.
1363 However, low-angle cross bedding (southwards palaeocurrent) is present, as well as
1364 scoured erosive surfaces with bivalve fragments (*Trigonia*; *Nerinea*), indicating
1365 deposition in the littoral zone. As in Staithes Sandstone times, the shallowing trend is
1366 attributed to regional tectonic uplift, probably doming of the Mid-North Sea High
1367 (Sellwood & Hallam 1974; Underhill & Partington 1993) and the Pennine highlands,
1368 rather than global sea-level fall.

1369

1370

1371 **5.2 Early-Mid Jurassic intra-basinal tectonics and heterolithic condensed** 1372 **marine sedimentation**

1373

1374 Regional tectonics (doming and related uplift) that resulted in shoaling marine
1375 environments during deposition of the late Toarcian Lias Group succession continued
1376 during early Aalenian times. The Cleveland Basin was gently folded during a
1377 compressional event in late Moorei Subzone to early Opalinum Subzone times (Black
1378 1934a; Hemingway 1974), which resulted in erosion of unconsolidated and semi-
1379 consolidated mud and sand. Up to 60 m of mud, silt and sand (uppermost Lias
1380 Group) was eroded in the central coastal areas (e.g. Whitby) and in the Hambleton
1381 Hills, where variable thicknesses of the Alum Shales are preserved below the Dogger
1382 Formation or, where that formation is absent, below the erosive base of the Saltwick
1383 Formation (Powell *et al.* 1992; Frost 1998). Southwards, towards the Market
1384 Weighton High, the Lias formations become thinner and the group is eroded down to
1385 lower stratigraphical levels (Redcar Mudstone Formation).

1386

1387 The Dogger Formation is a heterolithic unit deposited as a condensed
1388 succession over a considerable period of time (Opalinum Zone to Murchisonae Zone)
1389 and probably over the whole basin, although in some areas, e.g. Botton Head and
1390 along parts of the western escarpment, it was removed by erosion prior to deposition
1391 of the fluvio-deltaic Saltwick Formation. Despite intense bioturbation and reworking of
1392 this predominantly ferruginous, berthierine-rich sandstone unit, two subdivisions have
1393 been defined within the formation, namely a lower Opalinum Zone succession in the
1394 east of the basin, and a Murchisonae Zone succession that oversteps it, in the west
1395 (Macmillan 1932; Black 1934a; Rastall & Hemingway 1939, 1940, 1941, 1943, 1949;
1396 Hemingway 1974). However, some of the Dogger sands represent homogenized

1397 Opalinum Zone sediments reworked as a palimpsest unit in late Murchisonae time.
1398 The relationship between the Dogger Formation and the underlying and overlying
1399 beds is critical in unravelling its depositional history. Near Whitby (East Cliff), where
1400 the Dogger unconformably overlies the Alum Shales (Fig. 17e), the presence of
1401 downward penetrating soft-sediment burrows (*Thalassinoides*), backfilled with
1402 Dogger sand and phosphatic pebbles that were 'piped' up to 25 cm into the
1403 underlying Alum Shale muds, indicates that the latter were still relatively
1404 unconsolidated in this area during deposition of the Dogger. This suggests that the
1405 post-Toarcian compressional folding and subsequent erosion of semi- or un-
1406 consolidated sediment took place over a short time interval, or perhaps that the
1407 compressional folding was ongoing from late Toarcian times as a result of Mid North
1408 Sea volcanic doming. The Peak Trough (Milsom & Rawson 1989), a narrow syn-
1409 depositional graben that preserves the uppermost Lias at Blea Wyke Point (Fig. 23),
1410 may have resulted from transtension rather than compression during this tectonic
1411 event. Furthermore, the top of the Dogger sandstone at Blea Wyke and Whitby East
1412 Cliff is penetrated by carbonaceous-lined plant rootlets from the overlying paralic
1413 Saltwick Formation, demonstrating that the Dogger sands were also poorly
1414 consolidated when the Saltwick fluvio-deltaic sediments were deposited.

1415 In the west of the basin, where the Dogger is thought to be mostly of early
1416 Murchisonae Subzone age (Hemingway, 1974), the formation varies laterally over a
1417 few tens of metres from ferruginous berthierene-rich ooidal ironstone to ferruginous
1418 sandstone and cross-bedded ooidal limestone at Cleave Quarry [SE 497 828]
1419 (Powell *et al.* 1992; Fig. 29). Ooidal limestone is also present in the Mowthorpe area
1420 [SE 67 68], where there is a thick development of bi-modal, trough cross-bedded
1421 bioclastic limestone that includes the bryozoan *Collapora* [formerly *Haploecia*]
1422 *straminaea*, which is more usually associated with the Bajocian Lebberton Member
1423 (J Ford *pers comm.* 2007). These atypical carbonates, preferentially preserved in
1424 small shallow sub-basins, indicate the presence, locally, of a Dogger carbonate
1425 platform in Murchisonae times. Carbonate shoal deposits are preferentially preserved
1426 and may once have been more widespread ooidal banks that formed on shoaling
1427 highs and were subsequently swept offshore to accumulate in local downwarps. In
1428 the Howardian Hills, well-rounded calcareous mudstone pebbles and phosphatic
1429 pebbles (Spy Hill lithofacies) are found together with fossil wood 'raffle' in some of
1430 the more typical ferruginous sandstones, which here include quartz granules that
1431 were probably derived from the Carboniferous Millstone Grit, suggesting a westerly
1432 provenance from the Pennine highlands. In the west of the basin, there is no
1433 evidence, such as downward penetrating burrows, that the underlying Alum Shale
1434 muds were unconsolidated during deposition of the Dogger, probably due to longer

1435 residence time and early diagenesis of the folded Lias Group sediments in the west.
1436 Lithification of these pre-Dogger sediments is further indicated by the presence, in
1437 the Cold Moor area [NZ 558 002], of rounded, bored concretions containing Toarcian
1438 ammonites (including *Dactyloceras*, *Hildoceras* and *Harpoceras*) and other derived
1439 Toarcian ammonite clasts (Black 1934a), which show that the Whitby Mudstone
1440 sediments (down to the stratigraphical level of the Grey Shales Member) had, in
1441 contrast to the soft sediment burrows at Whitby, gone through early diagenesis and
1442 were lithified prior to the pre-Dogger tectonic folding, uplift and erosion.

1443

1444 Locally, in the Rosedale area [SE 729 946], the Whitby Mudstone was deeply
1445 eroded into a series of shallow 'boat-shaped' depressions, about 500 m long by 30 m
1446 wide (Marley 1870). The depressions were filled with a distinctive Dogger lithofacies,
1447 a 'magnetite' ironstone (now shown to be a form of ferric, chromstedtite spinel;
1448 Hemingway 1974) overlain by ferruginous sandstone (Fox-Strangways *et al.* 1885;
1449 Rastall & Hemingway 1949). The iron ore probably represents a condensed, remanié
1450 deposit of early Opalinum Zone times that was preserved locally in shallow
1451 depressions on the pre-Dogger sea-floor.

1452 The regional depositional setting of the enigmatic highly variable Dogger unit
1453 is outlined below:

1454

1455 (a) Compressional and transpressional folding of poorly consolidated muds, at least
1456 in the upper part of the sediment profile; transpression and local rifting resulted in the
1457 preservation of latest Toarcian Blea Wyke sands locally in the Peak Trough, but
1458 elsewhere up to 60 m of relatively unconsolidated sediment was removed in late
1459 Toarcian to earliest Aalenian times.

1460

1461 (b) In the west of the basin, uplift resulted in the development of local submarine
1462 highs within fairweather wave-base; here, during Murchisonae times, warm water
1463 ooidal shoals developed with an abundant shelly fauna that included bivalves,
1464 gastropods, solitary corals and bryozoa. These carbonate platform deposits were
1465 probably more extensive than the present outcrop; mobile carbonate sediments were
1466 preserved in local sub-basins on the lithified undulating pre-Dogger sea floor. Erosion
1467 to lower stratigraphical levels, including erosion of the the Grey Shale Member,
1468 reworked already lithified carbonate concretions and phosphatic nodules, which often
1469 contain Toarcian ammonites, as a basal conglomeratic lag deposit in the Cold Moor
1470 area [NZ 558 002].

1471

1472 (c) Uplift of the hinterland, balanced by basin subsidence, resulted in the southward
1473 and eastward progradation of the Middle Jurassic rivers and delta fronts. During the
1474 late Opalinum to Murchisonae zones, siliciclastic sediments were deposited in a pro-
1475 delta, shallow marine environment characterized by reworking of earlier carbonate
1476 platform shoals; iron colloids, in the presence of plant remains derived from the
1477 adjacent floodplains, were precipitated in shallow lagoons. Reworking in a shallow,
1478 restricted, partly anoxic tidal regime produced berthierine ooids interbedded with
1479 ferruginous sandstones. An abundant benthic infauna resulted in intense bioturbation
1480 and mixing of the sandy sediments. In places, the underlying Toarcian sediments
1481 were still poorly consolidated, enabling crustaceans to develop back-filled burrows
1482 into the underlying muds.

1483

1484 (d) Locally, the nearshore shallow lagoons became stagnant, resulting in deposition
1485 of laminated black mudstones. On adjacent highs, coeval sediments include trough
1486 cross-bedded sandstones with phosphatic pebble lags.

1487

1488 (e) Southward and eastward advance of the rivers and deltas and high sediment flux
1489 from the doming hinterland resulted in infilling of the Dogger shelf, lower relative sea-
1490 level and colonization of the floodplains by plants, resulting locally in rootlet
1491 penetration from plant-rich mires through to the Dogger sands. Basin subsidence in
1492 post-Dogger, late Aalenian times was paced by uplift of the surrounding areas and a
1493 high siliciclastic sediment flux, so that the coastal plain remained above sea-level
1494 until the late Aalenian marine transgression (Eller Beck Formation) that advanced
1495 from the south (Fig. 4a).

1496

1497 **5.3. Mid-Jurassic paralic sedimentation in a rapidly subsiding basin:** 1498 **transgressions versus regressions**

1499

1500 Differential subsidence in the Cleveland Basin, as compared to the East Midlands
1501 Shelf, was more pronounced from Aalenian to latest Bathonian times (175.6 – 161.2
1502 Ma), but was at its greatest during the early Bajocian to mid Bathonian, when up to
1503 300 m of fluvial, paralic and thin marine sediments were deposited (Ravenscar
1504 Group). During this interval, however, sedimentation rates kept pace with subsidence
1505 to fill the available accommodation space. This balance of sedimentation versus
1506 subsidence maintained the fluvial to paralic coastal plain in the region at or close to
1507 sea-level, so that relatively small increases in sea-level resulted in marine
1508 transgressions across large parts of the basin or, in the case of the late Aalenian
1509 Eller Beck Formation transgression and especially the Bajocian Scarborough

1510 Formation transgression, throughout the whole basin. At other times, such as during
1511 the early Bajocian Leberston transgression, the northern part of the basin remained
1512 above sea-level (Fig. 4b).

1513

1514 The Saltwick Formation was deposited in Murchisonae to Concavum zone
1515 times when paralic (i.e. fluvial, deltaic and brackish lagoonal) conditions were rapidly
1516 established throughout the basin as result of rapid uplift of the hinterland and high
1517 sediment flux from the north and west (Fig. 28). From the present-day coastal
1518 exposures to the western escarpment, muds and fluvial sands were deposited rapidly
1519 in river channel and interdistributary environments on the coastal plain, where the
1520 interfluves were colonized by plants, locally forming peaty mires thick enough to form
1521 thin coals.

1522

1523 The Ravenscar Borehole (Fig. 26) illustrates a typical Saltwick succession;
1524 initial progradation of muds and fine-grained sands across the low-gradient coastal
1525 plain, followed by progradational fluvial and deltaic sands in distributary channels.
1526 These sand-dominated, stacked distributary channels pass upwards to an
1527 interbedded mud and sand succession that illustrates a change to more isolated
1528 channel sands, some meandering and laterally accreted, intercalated with marshy
1529 overbank muds, frequently colonized by a diverse plant assemblage (Murchison
1530 1832; Harris 1953; Morgans *et al.* 1999; van-Konijnenburg & Morgans 1999). Overall
1531 upward-fining of sediments in the borehole suggests rapid subsidence and high initial
1532 sedimentation rates, followed by a lowering of the geomorphological gradient,
1533 reduction in accommodation space and a resultant change to low-sinuosity
1534 meandering streams. If this trend is true for the basin in general, the lowering of the
1535 gradient may reflect rising sea-level in late Aalenian times that peaked during the
1536 Eller Beck marine transgression (Knox 1973; Powell & Rathbone 1983).

1537

1538 The transgressive Eller Beck Formation illustrates the subtle balance between
1539 paralic and fully marine environments across the basin (Fig. 4a). The sea advanced
1540 from the south across the Market Weighton High, depositing micritic lime mud in
1541 shallow lagoons; the resulting unit was formerly known as the Hydraulic Limestone. A
1542 mixed lime mud and ooidal, ferruginous sand lithofacies was deposited to the north,
1543 in the Hambleton Hills (Powell *et al.* 1992), marking a boundary with the ooidal
1544 ironstone and ferruginous sand lithofacies to the north (Fig. 4a). Detailed
1545 sedimentological studies of the Eller Beck Formation (Knox 1970, 1973; Hemingway
1546 1974) show that the iron was derived in colloidal form from a lateritic hinterland
1547 subject to warm, humid weathering, and was deposited in shallow lagoons under

1548 anaerobic, reducing conditions (cf. Cleveland Ironstone, Dogger ironstone). Ooids
1549 indicate that the lagoons were influenced by tidal currents that resulted in oscillation
1550 and aggradation of individual berthierine (formerly chamosite) ooids. Dispersed ooids
1551 were also incorporated into the overlying sand member that is characteristic of the
1552 upper part of the Eller Beck Formation. Palaeocurrent analysis (Knox 1973) suggests
1553 influx of sand from the north that was reworked in a shallow shoreface and tidal
1554 regime in the north of the basin (e.g. Goathland [NZ 833 022]). Later, these marine
1555 ferruginous sand lithofacies extended to the south of the Hambleton Hills where they
1556 overlie the earlier micritic limestone lithofacies (Powell & Rathbone 1983).

1557

1558 The gamma-ray log of the Ravenscar Borehole (Fig. 26) shows the lower part
1559 of the overlying Sycarham Member (Cloughton Formation) to be dominated by minor
1560 channel sands and overbank coastal plain muds. However, the upper part of the
1561 Sycarham succession is represented by a sharp-based sand unit marking a major
1562 progradation of fluvio-deltaic sand. These sands represent stacked, erosively based
1563 channels similar to those seen in the lower part of the Saltwick Formation at Whitby
1564 West Cliff (Knox *et al.* 1991), and were deposited by a major river, at least in the
1565 Ravenscar area (Alexander 1986). The major river channels may have been aligned,
1566 in this area, close to the axis of a major channel system that persisted through
1567 Aalenian to early Bajocian times (Alexander 1986), although Hemingway (1974)
1568 proposed that the marked difference in the channel-sand dominance east and west
1569 of the Whitby Fault (trace of the River Esk) was due to later, probably Cenozoic
1570 transcurrent fault movement.

1571

1572 Relative sea-level rise in Laeviuscula Zone times (Butler *et al.* 2005) resulted
1573 in a second transgression and deposition of the Lebberton Member (Cloughton
1574 Formation), with lithofacies similar to those in the Eller Beck Formation, i.e. sands to
1575 the north and carbonates to the south. However, the marine influence in the north of
1576 the basin is much less pronounced, being restricted to thin ferruginous, locally shelly,
1577 sands and sideritic sands and thin limestone, as seen in coastal exposures
1578 (Cloughton Wyke and Yons Nab). This marine transgression advanced from the
1579 south. In the Howardian Hills, near Whitwell-on-the Hill, it is represented by ooidal
1580 and peloidal limestone (grainstone) up to 9 m thick, with an abundant shelly benthic
1581 fauna that includes the bryozoan *Collapora* [*Haploecia*] *straminea*; hence the local
1582 names, Whitwell Oolite and Millepore Bed, 'millepore' being an early name given to
1583 this bryozoan. Shallow tidal conditions are indicated by bi-directional cross-bedding,
1584 and in places colonial corals (*Thamnasteria* and *Thecosmilia*) are present with
1585 bivalves and gastropods. However, the Lebberton transgression was short lived as

1586 a result of progradation of fluvial and paralic sands and muds from the north. Based
1587 on their ostracod faunas, the carbonates are thought to be of Discites Zone age,
1588 equivalent to the Lincolnshire Limestone carbonate platform of the East Midlands
1589 Shelf (Bate 1964, 1967). On the coast, the transgressive and regressive lithofacies
1590 can be clearly seen at Yons Nab [TA 084 844], where the carbonate-rich 'Millepore
1591 Bed' is overlain by mudstones and sandstones with thin limestones termed the Yons
1592 Nab Beds, with a marine fauna that includes crinoid ossicles, bivalves and ostracods
1593 (Bate 1959, 1967; Whyte & Romano, 2006*b*). However, the marine beds disappear
1594 rapidly northwestwards along the coast, and this trend is paralleled inland where the
1595 Lebberston Member thins northward from 9 m thick in the Howardian Hills to less
1596 than 1 m in the Hambleton Hills, and cannot be traced north of a line from Whitby to
1597 Osmotherly (Hemingway 1974; Powell *et al.* 1992; Frost 1998). South of the Market
1598 Weighton High, the transgression is represented by the Cave Oolite [SE 91 31]
1599 (Neale 1958).

1600

1601 Lower delta plain environments with thin crevasse-splay sandstones and
1602 muddy flood plain deposits, characterized by a 'saw-tooth' gamma ray profile, typify
1603 the Gristhorpe Member (Cloughton Formation) in the Ravenscar Borehole (Fig. 26).
1604 Diverse and abundant plant remains have been described from the coastal plain and
1605 delta top sediments (van Konijnenburg-Cittern & Morgans 1999). Interfluves were
1606 colonized by a very diverse fauna (up to 260 plant species identified), including
1607 horsetails (*Equisetum*), ferns, conifers, cycads and tree ferns, with 15 species of
1608 *Ginkgo* alone (Harris 1953). Plants rapidly colonized the substrate and stabilized the
1609 interfluves, resulting in high sinuosity and meandering channels. Classic localities at
1610 Hayburn Wyke, Cloughton Wyke and Whitby are famous for the preservation of
1611 drifted plants deposited in shallow, freshwater pools, and for rooted, *in situ*
1612 specimens of *Equisites* (Black 1929; Harris 1953; van Konijnenburg-Cittern &
1613 Morgans 1999). Charcoal, in the form of fusain among the drifted plant remains, has
1614 been interpreted as evidence of forest fires that periodically destroyed the forest
1615 canopy, and growth ring studies suggest that the climate was humid, sub-tropical and
1616 seasonal (Morgans *et al.* 1999; Hesselbo *et al.* 2003). Thin coals were developed on
1617 the flood plain in shallow mires; downward penetrating rootlet traces and siliceous
1618 seat-earths resemble those more familiar in the Carboniferous Coal Measures
1619 (Fig. 17f). These 'Moor Coals' were worked locally for use in smelting ironstone and in
1620 lime kilns (Wandlass & Slater 1938; Owen 1970*b*). Coals, with underlying rootlet
1621 beds and seatearths, occur in the Cloughton Formation and Scalby Formation and
1622 have been exploited inland from bell-pits and shafts such as Boars Gill [SE 5188
1623 8084] where black mudstone and coal in spoil from an old bell-pit indicates coal of

1624 workable thickness. Also, in the west of the Asenby-Coxwold Graben, opencast coal
1625 exploration boreholes around Burtree House [SE 482 768] have confirmed the
1626 presence of the two coal seams worked in the late 18th century at the former Birdforth
1627 Colliery (Fox-Strangways *et al.* 1886; Owen 1970a; Hemingway & Owen 1975).

1628 Connection with the open Tethys Ocean was only achieved during the
1629 Scarborough Formation transgression (Bate 1965; Parsons 1977, 1980; Gowland &
1630 Riding 1991; Butler *et al.* 2005), a major sea-level rise in the early Bajocian. The
1631 coastal type section at Hundale Point (Fig. 30) is dominated by mud- and sand-rich
1632 sediments with thin argillaceous limestones, subdivided into seven members
1633 (Gowland & Riding 1991). Ammonites such as *Dorsetensia* and *Teloceras*, and
1634 marine palynomorphs in the Ravenscar Shale Member, indicate the Humphriesianum
1635 Zone (mid Bajocian). In the coastal outcrop, marine siliciclastic sediments with
1636 hummocky cross-bedding (e.g. at Ravenscar cliff) and thin silty limestones that yield
1637 bivalves (*Gervillella*, *Pseudomontis*, *Trigonia*, *Astarte* and *Lopha*), belemnites and
1638 sparse ammonites, together with a diverse suite of shallow marine trace fossils
1639 including *Rhizocorallium*, *Teichichnus* and U-shaped *Diplocraterion* (Hemingway
1640 1974; Miller *et al.* 1984; Gowland & Riding 1991), suggest depositional conditions of
1641 a littoral sandy embayment passing offshore to calcareous mud (Fig. 4c). A different
1642 palaeogeographical setting is suggested for the western outcrops of the Hambleton
1643 Hills (Fig. 31), where peloidal (faecal peloids) planar cross-bedded limestone
1644 (Brandsby Roadstone) is overlain by medium-grained, fossiliferous sandstone
1645 (Crinoid Grit) (Powell *et al.* 1992). The limestone exhibits bi-directional planar cross-
1646 bedding, suggesting deposition in a tidal lagoon. The overlying sandstone also
1647 shows tidal influences such as bi-directional trough cross-bedding and current ripples
1648 with surface *Gyrochorte* burrows (Powell 1992). The Scarborough Formation is
1649 absent over the Market Weighton High, and the thicker ammonite-bearing marine
1650 muds and sands (up to 20 m) on the coast suggest that the transgression came from
1651 the east (open marine). The Hambleton Hills were the site of shallow, tidally
1652 influenced carbonate lagoons that gave way to shallow tidal marine sands derived
1653 from the north, a probable precursor to the influx of the fluvio-deltaic system (Moor
1654 Grit Member and Scalby Formation) that unconformably overlies the Scarborough
1655 Formation over the whole of the basin. South of Scarborough, the formation thins to 6
1656 m at White Nab, 3 m at Yons Nab and Gristhorpe Bay [TA 085 842], but is only 1.5 m
1657 thick in the Fordon Borehole [TA 058 758], suggesting a shoreline to the south (Fig.
1658 4c).

1659
1660 It seems probable that there is a significant time gap at the base of the overlying
1661 Moor Grit Member (Scalby Formation). During this interval there was a lowering of

1662 relative sea-level, emergence of the basin, and erosion and uplift of the hinterland
1663 that resulted in high sediment influx from the north, as indicated by the marked
1664 lithofacies changes at the Scarborough Formation-Moor Grit boundary across the
1665 basin. The basal sand body of the fluvio-deltaic Scalby Formation is well exposed
1666 and best known from coastal outcrops (Cloughton Wyke and Black Rocks,
1667 Scarborough). On the coast, the Moor Grit is characterized by a low gamma-ray
1668 signature (Fig. 26) and typically comprises stacked sets of large-scale trough cross-
1669 bedded, medium- to coarse-grained sandstone deposited by a large braided or low-
1670 sinuosity river (Nami 1976; Nami & Leeder 1978; Leeder & Nami 1979; Livera &
1671 Leeder 1981; Kantorowicz 1985). Large-scale bedforms include laterally accreted
1672 channel sandstones with log impressions and pebble lags indicating high current
1673 velocities (Black 1928). In South Bay, Scarborough, and south of Hundale Point,
1674 sequentially stacked river channels pass up to mud-rich, level bedded units with only
1675 occasional laterally accreted channels (Nami & Leeder 1978). Interfluvial beds reveal
1676 tetrapod and sauropod dinosaur footprints (Hargreaves 1913; Sargeant 1970; Whyte
1677 & Romano 1993; Romano & Whyte 2003; Whyte & Romano, 2006a) and
1678 'dinoturbation' structures that the latter authors attributed to large saurian reptiles
1679 producing thixotropically contorted laminae below their footfalls (Fig. 27a). The
1680 coarse-grained stacked channels in the Moor Grit are mostly confined to the coastal
1681 exposures orientated northwest-southeast, and palaeocurrent measurements
1682 suggest a major river channel axis in this orientation, with sediment derived from the
1683 northwest and north. Elsewhere, the Moor Grit unit is definitely not a 'grit'; for
1684 instance in the Hambleton Hills, on the east of the outcrop, it comprises a texturally
1685 mature, fine- to medium-grained white sandstone, lacking major channels or laterally
1686 accreted bedforms (Powell *et al.* 1992). This tabular sand lithofacies was probably
1687 deposited by sheetflows on the interfluves between the major fluvial channels seen in
1688 the coastal exposures, which, though better known, are atypical. The texturally
1689 mature quartzitic sandstone of the inland outcrop suggests a distant provenance and
1690 deposition by a major river system. Large-scale, stacked fluvial channels seen in the
1691 northwest trending coastal sections at Hundale Point and south of Scarborough may
1692 have been controlled by subsidence associated with penecontemporaneous
1693 extensional subsidence within the Peak Trough.

1694

1695 Waning river velocity and reduced sediment flux is clearly seen across the
1696 whole basin in the upper part of the Scalby Formation. The upper 'level-bedded' Long
1697 Nab Member is mudstone-dominated and is characterized by deposition in
1698 meandering low-velocity rivers, muddy interfluves and fresh- to brackish-water lakes.
1699 North of Scalby Mills, the curved meander traces of point bar deposits (Nami &

1700 Leeder 1978) can be clearly seen from the cliff top (Fig. 27b); cross-sections in the
1701 cliff reveal low angle, laterally accreted point bar sands and mud-filled abandoned
1702 channels, together with laterally persistent thin sandstones with ripple marks
1703 representing crevasse-splay and overbank deposits that accumulated in shallow
1704 fresh water and brackish lakes; sphaerosiderite is a common early diagenetic
1705 structure in mud influenced by fluctuating, slightly acidic groundwaters (Kantorowicz
1706 1990).

1707

1708 Studies of well-exposed sections of the Long Nab Member on the Yorkshire
1709 coast (Black 1929; Leeder & Nami 1979) have concluded that it was deposited in a
1710 spectrum of sub-environments, including meandering channels, marshes and flood
1711 basins on an alluvial plain. However, subsequent reports of *Ophiomorpha* burrows,
1712 together with bioturbation and mud-grade sediments in channel-fill deposits (Livera &
1713 Leeder 1981), supported by the discovery of marine microplankton (Hancock &
1714 Fisher 1981), have suggested at least some degree of marine influence. On the
1715 basis of detailed sedimentological and palynofacies investigations of the coastal
1716 succession, Fisher & Hancock (1985) reinterpreted the upper part of the Scalby
1717 Formation as a saline-influenced delta-plain, interrupted by small distributary
1718 channels, some of which may have been tidal. However, it is not known whether this
1719 marine-influenced lithofacies extended over the whole of the basin, and based on the
1720 westward marine incursions in the fully marine units such as the Scarborough
1721 Formation, it might have been restricted to the current-day coastal sections.

1722

1723 A major sea-level rise in Callovian times (Herveyi Zone) resulted in marine
1724 incursion from the south and east across the Market Weighton High, flooding the low-
1725 gradient coastal plain. This major sequence boundary marks the demise of fluvial
1726 and deltaic sedimentation due to a world-wide sea-level rise (Haq *et al.* 1988) and
1727 increased subsidence of the Pennine landmass. The lower Cornbrash Formation of
1728 southern England is absent (Page 1989), indicating that marine flooding of the
1729 Cleveland Basin occurred slightly later during Herveyi Zone time. Bioturbation in the
1730 topmost Scalby mudstones and *Rhizocorallium* burrows that penetrate downwards
1731 from the berthierine ooidal limestone (Cornbrash Formation) (Wright 1977; Rawson &
1732 Wright 2000) indicate that the sea transgressed rapidly across a still poorly lithified,
1733 low-gradient coastal plain. The condensed succession represented by the Cornbrash
1734 limestone with its abundant oysters (*Lopha marshii*) at Cayton Bay [TA 0765 8405]
1735 indicates low levels of sediment flux to the littoral coastal plain during the
1736 transgression.

1737

1738

1739 **5.4 Mid-Late Jurassic carbonates, corals and ooid shoals**

1740

1741 The high Middle and Upper Jurassic succession is characterized mostly by
1742 calcareous sandstone and limestone that form the spectacular cliffs of Castle Hill,
1743 Scarborough, the high ground of the North York Moors and, in the west, the
1744 escarpment of the Hambleton Hills. The major sea-level rise in Callovian times
1745 marked a return to marine sedimentation that continued for the remainder of the
1746 Jurassic, about 20 million years, culminating in relatively deeper water sedimentation
1747 of the Kimmeridge Clay in a broad epeiric sea. The Callovian, Oxfordian and
1748 Kimmeridgian succession is best known from the early pioneering studies of Fox-
1749 Strangways (1892 and references therein), and from Arkell (1933, 1945) and the
1750 extensive publications of Wright (1968a, 1972, 1977, 1978, 1983, 1992, 1996a,
1751 1996b, 2009). Wright's detailed studies, and re-surveys of part of the Hambleton Hills
1752 by the BGS (Powell 1982, 1992) have demonstrated penecontemporaneous tectonic
1753 uplift and erosional unconformities in a seemingly conformable sequence (Coe 1995)
1754 (Figs 7, 33). Representative geophysical logs for the Hunmanby and Brown Moor
1755 boreholes, illustrating the lateral variation southwards towards the Market Weighton
1756 High, are shown in (Fig. 34).

1757

1758 As noted above, the Cornbrash Formation was deposited during the marine
1759 transgression at the base of the Callovian Stage (Herveyi? Zone), but its absence
1760 from the outcrop in the Hambleton Hills (Senior 1975; Powell *et al.* 1992) suggests
1761 that the initial marine transgression did not reach that far west. The overlying Cayton
1762 Clay Formation, albeit thin over most of the basin, probably marks the maximum
1763 flooding across the fluvio-deltaic plain (Wright 1977; Senior 1975; Coe 1995). The
1764 only exposure in the Hambleton Hills was in Northwoods Slack [SE 4982 8920],
1765 noted by Fox-Strangways *et al.* (1886, p. 43). Senior (1975) logged sections through
1766 this part of the sequence and noted that Fox-Strangway's locality was no longer
1767 exposed, but he recorded several small exposures of grey bioturbated, fossiliferous
1768 mudstone with bivalve fragments (c.1 m) at the base of the Osgodby Formation
1769 sandstone.

1770

1771 The overlying sandy marine Osgodby Formation, and the intra-formational tectonics
1772 that resulted in its component members being separated by local unconformities,
1773 have been described in detail by Wright (1968, 1978, 1992). The unconformities at
1774 the base of the Langdale Member and at the base of the Hackness Rock are well
1775 developed on the coast between Castle Hill, Scarborough [TA 05 89] and Cunstone

1776 Nab [TA 10 83] (Wright 1968a, fig.3) and in the Hambleton Hills (Powell *et al.*1992).
1777 The lower two members, the Redcliff Rock and overlying Langdale Member
1778 represent an upward coarsening succession deposited in a shallow-water, littoral
1779 environment, probably the subtidal shoreface zone passing offshore to finer-grained
1780 siliciclastics in the south-east of the basin.

1781

1782 Shallow marine siliciclastic sedimentation continued through the Callovian,
1783 but was interrupted by tectonic events that caused gentle basinal uplift and flexure,
1784 non-deposition and erosion, a precursor to early Oxfordian tectonics. Consequently,
1785 some of the Callovian ammonite zones are missing. In the Hambleton Hills, the
1786 Langdale Member of the Osgodby Formation (Wright 1978) is absent, so the upper
1787 member of the formation, the Hackness Rock (Athleta-Lamberti zones) rests
1788 unconformably on the Red Cliff Rock Member (Koenigi Zone) with strata representing
1789 up to three ammonite zones missing. Callovian to early Oxfordian uplift and erosion
1790 is manifested in the area between Whitestone Cliff and Raven's Gill (Fig. 33), where
1791 the Oxford Clay and underlying Hackness Rock are cut out by a low-angle,
1792 overstepping unconformity at the base of the Lower Calcareous Grit (see below).

1793

1794 A global rise in sea-level in Late Jurassic times (Cope *et al.* 1992; Haq *et al.*
1795 1988) is marked in the Cleveland Basin by the change in facies at the base of the
1796 Oxford Clay (Weymouth Member, Mariae Zone), which represents deeper water
1797 sedimentation and a continuation of the major marine transgression that began in the
1798 Callovian Stage. The later arrival of Oxford Clay lithofacies in the Cleveland Basin,
1799 compared to the East Midland Shelf, was due to lower subsidence rates north of the
1800 Market Weighton High. Furthermore, the global eustatic sea-level rise in north-west
1801 Europe (Hallam 1975, 1988; Haq *et al.* 1988) was interrupted in the Cleveland Basin
1802 by a regressive low sea-level stand during deposition of the Corallian Group.

1803

1804 Early Oxfordian sedimentation (Oxford Clay and Lower Calcareous Grit)
1805 represents an upward coarsening (shallowing) succession consisting of grey-green
1806 mudstone and silty mudstone passing gradationally up to calcareous siltstone and
1807 sandy limestone. The benthic fauna of the Oxford Clay includes sparse, small
1808 bivalves (*Meleagrinella* sp., *Oxytoma* sp., *Gryphaea* sp., *Nuculoma* sp. and
1809 *Rollierella* sp.) and the gastropod *Dicroloma* sp. The fine-grained lithology of that
1810 formation, the absence of current structures, the paucity of its benthic fauna, and the
1811 presence, locally, of a nektonic fauna that includes belemnites (*Hibolites* sp.) and
1812 ammonites (*Quenstedtoceras mariae* (d'Orbigny), *Cardioceras scarburgense* (Young
1813 and Bird) and *Peltoceras* (*Parawedekendia*) sp.), all suggest an offshore, moderately

1814 deep-water environment of deposition. However, some beds, especially in the lower
1815 part, are heavily bioturbated, with abundant *Chondrites* burrows and *Planolites*
1816 burrows, indicating an oxygenated sea-floor.

1817

1818 In the Hambleton Hills, the marked local unconformity below the Lower
1819 Calcareous Grit reflects the continuation of tectonic activity that resulted in
1820 depositional hiatuses during the Callovian (Fig. 7). This is most pronounced in the
1821 south-east of the Hambleton Hills, where uplift and tilting of the Roulston Scar 'block'
1822 resulted in the Lower Calcareous Grit resting unconformably (overstep) on the Red
1823 Cliff Rock (Osgodby Formation) (Figs 7, 33). The absence of the Oxford Clay
1824 between Sutton Bank and Raven's Gill along the main escarpment, and also at Hood
1825 Hill [SE 504 813], demonstrates uplift and subsequent sub-marine erosion of the
1826 Oxford Clay, and in places the Hackness Rock, prior to deposition of the Oldstead
1827 Oolite, the ooidal limestone developed locally at the base of the Lower Calcareous
1828 Grit (Wright 1983; Powell *et al.* 1992). This area appears to have been an uplifted
1829 block, tilted gently towards the north, since the Oxford Clay thins gradually
1830 southwards towards Roulston Scar [SE 5110 8153], but is present between Raven's
1831 Gill and Shaw's Gill, where a penecontemporaneous post Oxford Clay-pre Lower
1832 Calcareous Grit fault is invoked (Fig. 33). Uplift and erosion must have been short-
1833 lived because the Oxford Clay and the overlying Lower Calcareous Grit have yielded
1834 ammonites of the Mariae Zone and Cordatum Zone (Bukowskii Subzone),
1835 respectively, in the Shaw's Gill area [SE 507 834] (Powell 1982; Powell *et al.* 1992,
1836 fig. 18). Clay pellets in the unconsolidated sand at the top of the Redcliff Rock
1837 between Whitestone Cliff and Raven's Gill may have been derived from the Oxford
1838 Clay during the erosive phase that removed both the Oxford Clay and the Hackness
1839 Rock and reworked the top of the Redcliff Rock (Powell 1982). The unconformity is
1840 equivalent to five ammonite zones and represents rapid erosion of marine mud and
1841 sand representing a considerable time span (Cope *et al.* 1980*b*, fig. 8).

1842

1843 A broad, shallow carbonate platform was established across the Cleveland
1844 Basin area during mid Oxfordian times, contrasting with more rapid subsidence and
1845 deeper water sedimentation across the East Midlands Shelf. The spicule-rich
1846 calcareous sandstones and micritic, bioclastic, reefal and ooidal limestones that
1847 comprise the Corallian Group (70 to 150 m) were deposited in a warm, shallow sea
1848 during a relative sea-level low stand.

1849

1850 Where the Oxford Clay is present, away from the Roulston Scar 'block', the
1851 gradational upward-coarsening across the boundary between the Oxford Clay and

1852 the Lower Calcareous Grit,, together with the shelly benthic fauna, abundant sponge
1853 spicules and *Thalassinoides* burrows in the latter suggest sedimentation under
1854 shallower water conditions than prevailed during deposition of the underlying Oxford
1855 Clay. The paucity of small-scale sedimentary structures in the Lower Calcareous Grit
1856 is due to intense bioturbation of the substrate soon after deposition; *Thalassinoides*
1857 burrows with a higher spicule content are well preserved on bedding planes, but
1858 intense bioturbation within individual beds has produced a homogeneous fabric.
1859 Despite the general absence of primary current structures, the lithological and faunal
1860 characteristics indicate deposition in shallow to moderate depths (c. 10 to 30 m) in
1861 the offshore zone.

1862

1863 Development of the Oldstead Oolite Member (Wright 1980; Powell *et al.*
1864 1992) at the base of the formation near Roulston Scar [SE 5156 8121 to 5327 8206]
1865 was probably due to the development of a shallow water ooidal shoal lithofacies on a
1866 sub-marine high that developed in response to the local tectonic uplift this area (see
1867 above), the Oldstead Oolite being deposited in turbulent conditions on the south-
1868 eastern flanks of the Roulston Scar 'block', which formed the palaeohigh. Cross-
1869 bedding, ooidal grainstone texture and basal erosional scours indicate shoaling
1870 conditions, within wave-base. A decrease in the proportion of ooids (wackestone
1871 texture) at the top of the member suggests gradually increasing water depths through
1872 time during the deposition of the Lower Calcareous Grit.

1873

1874 Shallowing of the sea, with a change in benthos (sparse *Rhaxella* spicules)
1875 and the development of dynamic, tidally influenced ooidal shoals, is reflected in the
1876 overlying Coralline Oolite Formation. The lowermost member, the Hambleton Oolite,
1877 has a gradational base and in places passes laterally into spiculitic calcareous
1878 sandstone of the Birdsall Calcareous Grit (Wright 1972), indicating the lateral
1879 discontinuity of migrating ooid shoals and passage offshore to siliciclastic lithofacies.
1880 This relationship reflects the original depositional environment of migrating ooidal
1881 shoals passing into slightly deeper water environments typified by the spiculitic sand
1882 'background' sedimentation.

1883

1884 Detailed studies of the Corallian succession in the Howardian Hills by Wright
1885 (2009) have shown that the broadly east-west fault system in the Howardian Hills and
1886 Vale of Pickering was locally tectonically active during the mid Oxfordian (c.f.
1887 Roulston Scar in the Hambleton Hills). Penecontemporaneous extensional
1888 movement on the Coxwold-Gilling faults resulted, locally, in marked changes in

1889 thickness and lithofacies with areas of uplift and erosion along the Corallian ridge
1890 notably at Gilling East and between Malton and North Grimston (Wright 2009, fig. 13)

1891

1892 The distribution of lithologies, fauna and facies suggest that the Coralline
1893 Oolite Formation was deposited in a warm shallow sea that covered an extensive
1894 carbonate platform, across which ooid shoals prograded offshore (south-eastwards)
1895 from the nearshore zone situated to the north of the district. Micritic carbonates
1896 developed in sheltered lagoons that were protected, in part, by coral-algal patch reefs
1897 during deposition of the Coral Rag Member (Reeves *et al.* 1978). Intercalation of
1898 ooidal carbonates and calcareous sandstones in the lower part of the formation, and
1899 lateral passage to increasingly siliciclastic-dominated lithofacies to the south-west,
1900 suggest a south-easterly transition from nearshore to offshore zones.

1901

1902 Ooidal limestone (packstone to grainstone texture) lithofacies (e.g. Hambleton
1903 Oolite) with a variable proportion of quartz sand and fragmented shells are typified by
1904 cross-bedding and shallow scours. Multidirectional cross-bedding azimuths suggest
1905 that the oolite shoals were deposited in an oscillating tidal current regime. Soft-
1906 sediment deformation structures, such as the slump structures and injection
1907 phenomena locally present at Shaw's Gate Quarry [SE 5233 8236] and Old Byland
1908 Grange Quarry [SE 5454 8567] in the Hambleton Hills, might have resulted from the
1909 displacement of pore-waters held in the semi-lithified sediments during seismic
1910 activity associated with local tectonic uplift in the Roulston Scar area in early
1911 Oxfordian times. However, some of the convoluted sandy beds, particularly the basal
1912 bed at Shaw's Gate Quarry, have features such as convoluted clasts, erosive bases
1913 and planar, truncated, upper bedding surfaces, that indicate deposition as submarine
1914 debris flows.

1915

1916 Temporal relationships between the ooidal (Hambleton Oolite Member) and
1917 variably spiculitic calcareous sandstone (e.g. Birdsall Calcareous Grit Member)
1918 lithofacies in the Hambleton Hills are difficult to resolve because of the paucity of
1919 ammonite faunas. The overall lithofacies distribution of these two members, as
1920 deduced from their outcrop pattern, indicates a depositional environment ranging
1921 from shallow-water ooid shoals in the north of the Cleveland Basin, interdigitating
1922 with, and passing offshore to marine siliciclastics towards the south-east. Similar
1923 lithofacies have been described from the modern-day Great Bahama Bank, Andros
1924 Island and the Arabian Gulf (Purdy 1963; Bathurst 1975, p.135; Black 1980; Hine *et*
1925 *al.* 1981). The lithological characteristics of the ooid lithofacies, taken together with
1926 the low-dipping, multidirectional cross-bedding and shallow scours, suggest periodic

1927 migration of oolitic shoals on a shallow-water carbonate platform, influenced by
1928 waves and oscillating tidal currents. Sparse vertical burrows indicate temporary
1929 stability of the substrate that allowed colonization by infauna.

1930

1931 The pattern of fluctuating sea-level is repeated with deposition of the Middle
1932 Calcareous Grit Member, Malton Oolite Member and Coral Rag Member which
1933 together form the second upwards shallowing cycle in the Corallian Group. Shell
1934 beds composed of *Myophorella huddlestoni* in the Middle Calcareous Grit (Vertebrata
1935 Subzone; Wright 1980), together with *Rhizocorallium* burrows and cross-bedding,
1936 suggest a high-energy, shallow-marine environment of deposition (Hemingway
1937 1974). The shoaling cycle is capped by the Malton Oolite and Coral Rag members.
1938 Large-scale foresets and a paucity of benthic faunas in the Malton Oolite indicate
1939 large mobile laterally migrating ooidal shoals, similar to parts of the present-day
1940 Bahama Banks (Twombly 1964), formed during strong flood and ebb storm surges
1941 and preserved as mega-dune foresets. The Coral Rag comprises locally developed
1942 coral-algal patch reefs, coral-shell inter-reef debris, and micritic limestone deposited
1943 in back-reef lagoons.

1944

1945 As sea-level rose towards the end of Corallian Group deposition, the Upper
1946 Calcareous Grit Formation was deposited in slightly deeper water across the Market
1947 Weighton High and northwards into the Cleveland Basin. Very fine- to fine-grained,
1948 highly calcareous, spiculitic sand and silt, with abundant beds of clayey lime-mud in
1949 the middle of the unit, was deposited in moderate depths on the shelf in nearshore to
1950 offshore environments. Increased rates of subsidence and global sea-level rise
1951 around Serratum Zone time resulted in 'drowning' of the shallow siliciclastic and
1952 carbonate platform, with the deposition of mud (Amphill Clay and Kimmeridge Clay)
1953 in deeper water environments.

1954

1955

1956 **5.5 Late Jurassic global sea-level rise and sea-floor anoxia**

1957

1958 Global sea-level rise during the late Oxfordian is manifested in the Amphill and
1959 Kimmeridge Clay formations, which are the youngest Jurassic units in the Cleveland
1960 Basin. The marked change in lithofacies from shallow-water carbonates and
1961 nearshore siliciclastics, represented by the uppermost Corallian Group, to oxic
1962 shallow marine mudstones with thin carbonate beds and nodules (Amphill Clay),
1963 passing upward to organic-rich mudstones (Kimmeridge Clay), reflects subsidence of
1964 the Cleveland Basin and deeper-water conditions during late Oxfordian and

1965 Kimmeridgian times. Infaunal and epifaunal bivalves, gastropods and echinoid spines
1966 suggest that the Ampthill Clay was deposited in an oxic shallow marine environment.
1967 Oxic and anoxic (organic-rich) bottom conditions then alternated as the platform
1968 subsided during deposition of the Kimmeridge Clay; pelagic and hemipelagic muds
1969 were deposited from suspension across the former platform and into the current-day
1970 North Sea, in response to a major worldwide sea-level rise (Haq *et al.* 1988). Muds
1971 were deposited in rhythms indicating fluctuating relatively oxic and anoxic bottom
1972 conditions. The geophysical logs of the Hunmanby Borehole (Cox & Gallois 1981;
1973 Whittaker *et al.* 1985) illustrate the 'saw-tooth', small-scale sedimentary rhythms (Fig.
1974 38). These couplets comprise brown-black, bituminous fissile mudstone (anoxic;
1975 kerogen-rich) and overlying medium-grey and pale grey calcareous mudstone (oxic).
1976 Anoxic and oxic sea-bed conditions are indicated by fossil associations with
1977 increased levels of bioturbation and infaunal bivalves during more oxic periods
1978 (Wignall 1990). Free-swimming fauna such as ammonites, fish and marine reptiles
1979 were preserved after death during periods of sea-floor anoxia, resulting in their
1980 exceptional preservation. Restricted circulation and high organic productivity in the
1981 Tethys Ocean during the late Kimmeridgian (post-Eudoxus Zone) led to deposition
1982 and preservation under anoxic bottom conditions of organic-walled phytoplankton,
1983 which following burial in the northern North Sea Basin and the formation of kerogen,
1984 generated hydrocarbons (Gallois 1976; Herbin *et al.* 1993). Earlier depositional
1985 models attributed the high organic content to high-levels of phytoplankton productivity
1986 (algal blooms) (Gallois 1976). However, Weedon *et al.* (2004) calculated the organic
1987 productivity over time and concluded that the 'Kimmeridge sea' was no more
1988 productive than modern-day continental shelves, and consequently that the high
1989 organic content (3.8% Total Organic Carbon) was due to a dilution or absence of
1990 terrigenous siliciclastics sediment entering a semi-restricted basin. If this was the
1991 case, it suggests a geomorphologically subdued hinterland with very little erosion of
1992 terrigenous material, perhaps akin to the southern margins of the present-day
1993 Arabian Gulf, but with more anoxic bottom conditions (cf. the Black Sea).

1994

1995 The non-turbulent and cyclical anoxic bottom conditions resulted in fine
1996 preservation of ammonites such as *Amoeboceras* sp. (Ampthill Clay), *Pictonia* sp.,
1997 *Aulacostephanus fallax* and *Rasenina evolata*, and these have enabled detailed
1998 biostratigraphical zonation and correlation with the North Sea and Europe (Herbin *et al.*
1999 *et al.* 1995). The Kimmeridge Clay cycles have been attributed to astronomical
2000 (Milankovitch) cycles (Weedon *et al.* 1999; Weedon *et al.* 2004) with the larger (2-4
2001 m wavelengths) representing orbital obliquity (c. 41,000 year periodicity) and the
2002 smaller wavelength cycles (1-2 m wavelengths) to precession cycles (c. 26,000 year

2003 periodicity). Studies based on sequence stratigraphy (Wignall 1991; Taylor *et al.*
2004 2001; Williams *et al.* 2001) have recognized between 9 and 11 depositional cycles in
2005 the North Sea and adjacent areas, with silt-dominated units in the centre of the basin
2006 representing lowstand conditions during which siliciclastic sediment by-passed the
2007 shallow shelf. The thin and siltier succession through the Elegans to Wheatleyensis
2008 zones may have represented significantly shallower basin conditions than during the
2009 earlier Mutabilis to Eudoxus zones (Fig. 38). During the late Kimmeridgian (Bolonian
2010 of Cope 1993), this upward shallowing trend is reflected in the Late Cimmerian
2011 tectonic uplift (Rawson & Riley 1982) that resulted in a basin wide sea-level fall.

2012

2013 Water depth and depositional conditions for the Kimmeridge Clay have been
2014 the topic of much debate (Cope 2006). Early workers (Irwin 1979; Tyson *et al.* 1979;
2015 Myers & Wignall 1987) postulated relatively deep water and fluctuating oxic-anoxic
2016 conditions on the sea-floor as the oxic-anoxic boundary migrated vertically above and
2017 below the sediment-water interface. Wignall (1989) also proposed that storm events
2018 ripped up intraclasts and fine-grained clastic sediments, which were re-deposited in
2019 deeper parts of the basin, thereby oxygenating the bottom sediments. This implies
2020 that water depths, at least at the basin margins, were within storm-wave base (c. 30
2021 m depth). It is generally accepted that the 'Kimmeridge Basin' was 'restricted'
2022 relative to full ocean circulation to some degree, and some authors (Hallam 1975;
2023 Aigner 1980) proposed deposition in stagnant conditions of shallow water depths of
2024 about 10 m rather than the c. 100 m or so postulated by Tyson *et al.* (1979) and
2025 others.

2026

2027 The unconformity between the Kimmeridge and Speeton clays was a result of
2028 latest Jurassic sea-level fall (Haq *et al.* 1988) coupled with the Late Cimmerian earth
2029 movements (Rawson & Riley 1982) that led to a tensional (rifting) tectonic phase
2030 during latest Jurassic to earliest Cretaceous times, probably in response to sea-floor
2031 spreading in the Atlantic Ocean (Rawson 2006).

2032

2033 6. CONCLUSIONS AND FURTHER STUDY

2034

2035 The Jurassic Cleveland Basin, by virtue of its diverse lithofacies, tectonic evolution,
2036 excellent coastal exposures and analogues to North Sea hydrocarbon plays, is one
2037 of the most studied sedimentary basins in NW Europe. Pioneering research in the
2038 geological sciences, supported through the Yorkshire Philosophical Society (later the
2039 Yorkshire Geological Society), were stimulated by the early researches by William
2040 Smith and his nephew John Phillips. Later academic and practical studies on the

2041 coast (e.g. Young & Bird 1822; Phillips 1829, 1858), combined with the search for
2042 energy and industrial minerals (coal, ironstone, alum, jet, building stone and
2043 aggregates) in the 18th and 19th centuries, paved the way for the Geological Survey
2044 Primary Survey at '6 inches-to-the-mile scale' by Fox-Strangways and Barrow in the
2045 late 19th century, which laid the foundations to our understanding of the basin in its
2046 broadest context.

2047

2048 Our knowledge has been refined over the last 100 years or so through the
2049 advent of high resolution biostratigraphy, geochemistry, sedimentology and
2050 sequence stratigraphy, heavy mineral studies, and more recently cyclostratigraphy
2051 and stable isotope geochemistry, techniques that may lead to an astronomical
2052 timescale for the deeper water marine successions. Academic research has
2053 focussed on the well-exposed coastal sections, but new data is still emerging from
2054 detailed studies of the inland exposures and deep hydrocarbons boreholes. The
2055 latter, combined with detailed biostratigraphy and stable isotope geochemical
2056 studies, are likely to provide a better understanding of the influence of the Earth's
2057 orbit and Milankovitch cyclicity on sedimentation and diagenesis. These advances will
2058 no doubt lead to a better understanding of the palaeoceanography, sedimentation
2059 and evolution of the Cleveland Basin. However, there is still a place for detailed
2060 geological mapping, three-dimensional modelling and multidisciplinary
2061 sedimentological/petrological/biostratigraphical studies of parts of the succession that
2062 we still do not fully understand, such as the origin of the Jurassic ironstones, the
2063 nature of sea-bed anoxia, palaeoceanography, the influence of the Howardian-
2064 Flamborough Fault Belt on Jurassic sedimentation north of the Market Weighton
2065 High, intra-Jurassic tectonics, palaeoclimates and atmospheric gasses, and the
2066 evolution of flora and fauna. These and no doubt many new, avenues of research will
2067 ensure that the Cleveland Basin remains a focus of geological research and training,
2068 thereby attracting leading international scientists to reveal more about this fascinating
2069 period of Earth's history – here in Yorkshire!

2070

2071

2072 c.22 k words

2073

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2087 Figure & Table Captions (see separate list)

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2089 References (see separate list)

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