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ORIGINAL PAPER

The stratigraphy of latest Devonian and earliest Carboniferous rocks in Ireland

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Abstract

Thick successions of latest Devonian and early Carboniferous siliciclastic rocks of fluviatile and shallow marine origin are well exposed in numerous coastal sections in the south of Ireland. Correlations based on palynostratigraphy demonstrate that an overall northward transgression, in detail consisting of several fluctuations of relative sea level, resulted in the replacement of fluvial coastal plain environments by shallow marine conditions. The most conspicuous lithological record of transgression occurred at or close to the base of the Carboniferous, when a regionally distributed mudstone deposited in a shallow sea, replaced the generally sandy strata that had prevailed in the latest Devonian. There is no evidence for very large glacioeustatic changes of sea level related to the coeval western Gondwanan glaciation. Of the three horizons for a new GSSP for the base of the Carboniferous currently under review, the base of the *Protognathodus kuehni* Zone/basal *Siphonodella sulcata* Zone and coastal plant extinction is the most easily applied in Ireland and can be identified using palynology and corresponds to an easily recognised, extensively developed lithological change in the south of the country.

Keywords Palynostratigraphy · Glaciation · Sea level · Hangenberg · Famennian · Tournaisian

Introduction

Latest Devonian and earliest Carboniferous rocks are well exposed along the rocky coastline of south and southwest Ireland. They are also seen in inland natural exposures and quarries and have been proved in boreholes further north. Research on the stratigraphy, sedimentology and biostratigraphy of these sections over the past 60 years, summarised by Clayton et al. (1981), Graham (2009a) and Sevastopulo and Wyse Jackson (2009), has established that the successions in the south of Ireland are exceptionally thick in comparison with coeval successions elsewhere in western Europe; that the rocks represent fluviatile and shallow marine environments; and that despite high levels of organic maturation (Clayton 1989), a well-resolved palynological biostratigraphical framework provides correlations that allow detailed analysis of the changing palaeoenvironments. These exceptionally thick successions provide an opportunity to analyse the far-field effects of the Late Devonian glaciation reported from western Gondwana (Caputo 1985; Isaacson et al. 1999, 2008; Caputo et al. 2008) in terrestrial/shallow marine settings and to evaluate the changes in sea level associated with the so-called Hangenberg Event (Kaiser et al. 2016) in facies that contrast with the condensed, deeper marine successions of many parts of the world, particularly the Rheinisches Schiefergebirge, Germany, where it was first recognised.

Regional geology

The rocks of concern here occur within the Munster Basin (Fig. 1), which developed in the south of Ireland during the Middle and Upper Devonian (see Graham 2009a for a detailed account). The northern and eastern margin of the basin has been equated with the 1-km isopachyte, which correlates approximately with an abrupt change of thickness; the locations of the southern and western margins of the basin are not known. The basin fill, more than 6-km thick, is dominated by terrestrial, dominantly fluviatile, red bed facies, which become finer grained both up section and also southward away from the basin

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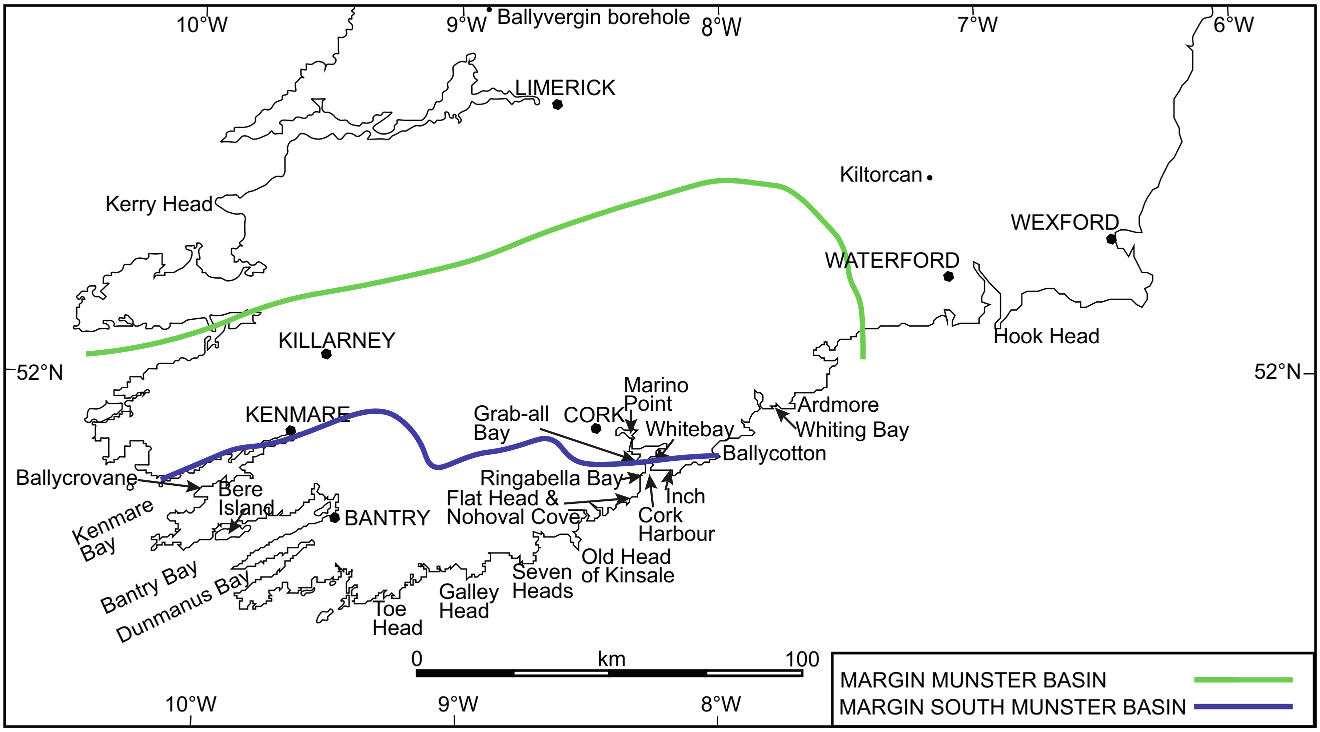


Fig. 1 Map of the south of Ireland showing locations of principal towns, places mentioned in the text and the boundaries of the Munster and South Munster Basins

margin. The oldest rocks identified are of Givetian or Frasnian age, but where the Devonian succession is thickest, its base is not exposed. In the late Famennian, thickness changes across the margin of the former Munster Basin diminished and a new basin, the South Munster Basin (George et al. 1976), developed with its northern margin controlled by a set of linked, en echelon faults along the so-called Cork-Kenmare line (Fig. 1). At approximately the same time, the fluviatile red bed sequences in the South Munster Basin were succeeded by dominantly grey and green sandstones and shales, interpreted as having been deposited on a low gradient, coastal plain. These and the overlying shallow marine sandstones and shales of both late Famennian and early Tournaisian age are described in more detail below. During the remainder of the Mississippian, the South Munster Basin became starved as a marine transgression progressed northward across Ireland and a carbonate ramp, and subsequently a platform developed north of the basin margin. The youngest sedimentary rocks preserved within the South Munster Basin are of Serpukhovian and early Pennsylvanian age (see Sevastopulo 2009 for details).

The rocks discussed below were all deformed during the Variscan Orogeny with the development of folds that strike ESE/NNW, with wavelengths of several kilometres and with axial surfaces that are generally vertical to steeply inclined. Strain indicators show vertical and along strike extension and across strike shortening of probably more than 40%; pressure solution cleavage is ubiquitous, and there are numerous faults, oriented along strike and at high angles to strike (see Graham 2009b for details of structural geology). The metamorphic grade, established using measurements of vitrinite reflectance, clay crystallinity and conodont colour alteration, is everywhere in the anchizone or higher (Clayton 1989; Graham 2009b).

Biostratigraphical framework

Available biostratigraphical information of the latest Devonian and earliest Carboniferous rocks of southern Ireland is based largely on the ranges of miospores, with subordinate information from ammonoids, brachiopods and conodonts. Because the miospore occurrences are largely in facies in which there is no other biostratigraphical information, correlations to the standard conodont/ammonoid biozonal schemes for the latest Devonian/earliest Carboniferous rely on comparisons with successions elsewhere (Fig. 2). These correlations are based on relatively few tie points, and there is a need for additional research to confirm them.

The oldest miospore assemblages under discussion here are dominated by *Retispora lepidophyta* (Higgs et al. 1988); they are assigned to the LL miospore Biozone, first introduced (as a sub-Biozone) by Paproth et al. (1983). The lowest occurrences of this zonal assemblage in Ireland are controlled by sedimentological/taphonomic factors because they occur in grey and green mudrocks above, or sporadically developed within generally barren red bed sequences. The boundary between

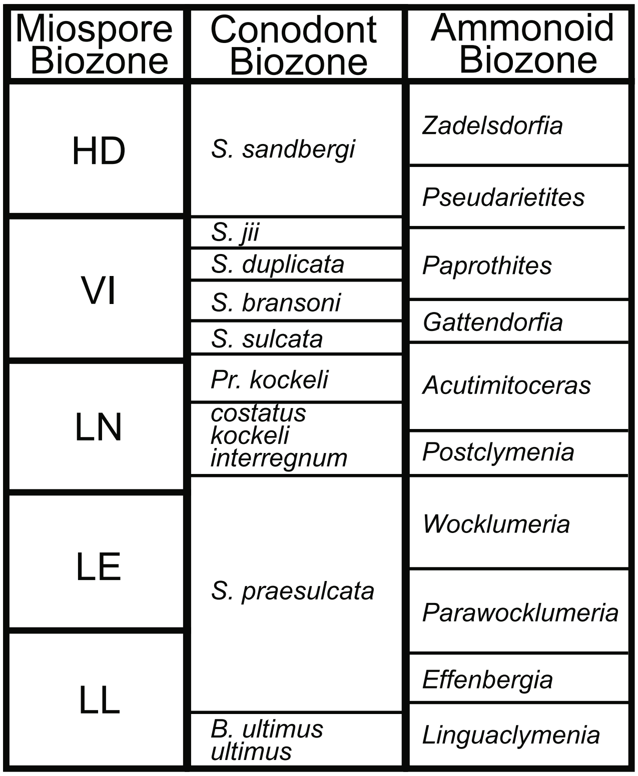


Fig. 2 Correlation of Irish Late Devonian/early Carboniferous miospore Biozones (Higgs et al. 1988) with conodont and ammonoid Biozones. Conodont Biozones after Kaiser et al. (2009) and conodont/ammonoid Biozone correlations after Becker et al. (2016). *B. Bispathodus*, *Pr. Protognathodus*, *S. Siphonodella*

the LL assemblage and the immediately older (VH) assemblage in Belgium equates with a horizon within the Famennian Middle *expansa* conodont Biozone (Higgs et al. 2013), but the VH miospore Biozone has not been identified in Ireland.

The base of the LE miospore Biozone (Paproth et al. 1983) is identified by the first occurrence of *Indotriradites explanatus*. Kaiser et al. (2011; Fig. 1) showed the boundary between the LL/LE Biozones as correlating with a horizon high within the *S. praesulcata* conodont Biozone of the revised scheme of Kaiser et al. (2009). Becker et al. (2016), on the basis of palynological analyses by Rahmani-Antari and Lachkar (2001), reported that in Morocco it falls within the *Parawocklumeria paradoxa* ammonoid Biozone (Fig. 2).

The first occurrence of *Verrucosisporites nitidus* defines the base of the LN miospore Biozone (Clayton et al. 1974). Prestianni et al. (2016) have argued that the stratigraphical distribution of *V. nitidus* is so facies-dependent (being tied to 'proximal' environments) that the LN Zone should be considered an 'ecozone' and that its value in correlation in Belgium is negligible. This is not the case in Ireland where *V. nitidus* occurs consistently in assemblages above its first occurrence and where the thickness of strata assigned to the LN Biozone from place to place varies approximately in concert with the thickness of other miospore Biozones. Higgs et al. (1993) reported that in the youngest assemblages assigned to the LN Biozone, both in the Rheinisches Schiefergebirge and in southern Ireland, *Retispora lepidophyta* makes up a conspicuously smaller proportion of the total spore assemblage than in older horizons. Higgs et al. (1993) recorded the oldest occurrence of the LN Biozone in the Stockum II trench, Rheinisches Schiefergebirge, Germany (Clausen et al. 1994), within black shales (Bed 162) immediately above the youngest level with the conodonts *Bispathodus ziegleri ziegleri* and *Palmatolepis gracilis gracilis,* that is at the base of the *costatus-kockeli* interregnum of the revised conodont biostratigraphical scheme of Kaiser et al. (2009).

The top of the LN miospore Biozone (and by definition the base of the younger *Vallatisporites verrucosus-Retusotriletes incohatus* (VI) miospore Biozone (Clayton et al. 1974)) is identified by the disappearance of *Retispora lepidophyta* and associated taxa. Because miospores in these rocks, particularly those of shallow marine facies, are clearly allochthonous, the abrupt disappearance of this suite of taxa cannot be explained by facies changes in the environments where the spores were deposited and points to a real and probably global extinction event affecting the parent plants, as reviewed by Marshall (2020). The LN/VI Biozone boundary has been identified in sections in the Rheinisches Schiefergebirge, Germany, which have yielded both ammonoids and conodonts (Higgs and Streel 1984; Higgs et al. 1993; Higgs and Streel 1994). In the Stockum II trench, the boundary occurs in a 0.2-m-thick interval between beds 102 and 104, both of which are of shale. Bed 103 is a limestone which is reported to contain *Protognathodus* spp. including *Pr. kockeli* (Clausen et al. 1994), which is therefore assigned to the *kockeli* Biozone of the revised conodont biostratigraphical scheme of Kaiser et al. (2009); this bed also contains ammonoids (*Acutimitoceras* spp., including *A. prorsum*). In the Hasselbachtal stream section, the boundary is identified within Bed 85 of the Hangenberg Shale, 0.14 m below the base of the Hangenberg Limestone, and in the Hasselbachtal Borehole, drilled nearby, at a comparable level within shales between 0.45 and 0.10 m below the base of the Hangenberg Limestone. It is worth noting that although it is commonly stated that the LN/VI biozonal boundary lies just below the base of the *Siphonodella sulcata* conodont Biozone (whose base is shown as coincident with the base of the Hangenberg Limestone), no conodonts have been reported from the upper few centimetres of the Hangenberg Shale above the earliest VI miospore assemblage at Hasselbachtal. The only faunal record of any significance from this interval is that of a specimen of *Acutimitoceras* sp. from 90 to 110 mm below the Hangenberg Limestone (Becker 1996).

The HD miospore Biozone (Higgs et al. 1988) is identified by the lowest occurrence of *Krauselisporites hibernicus*. The VI/HD biozonal boundary has been correlated with a level at the base of or within the *Siphonodella sandbergi* conodont

Biozone (Higgs et al. 1992) and is therefore considerably younger than the Devonian/Carboniferous boundary, the subject of this paper (Fig. 2).

Reference is made below to the BP and PC miospore Biozones (Higgs et al. 1988) which equate with the late part of the range of *Siphonodella* and younger conodont Biozones.

Details of the successions across the Devonian-Carboniferous boundary

The succession spanning the Devonian-Carboniferous boundary is exposed in numerous coastal sections from the west to the east of the South Munster Basin, as well as along the coast east of Cork and at various localities (natural exposures, quarries and cored boreholes) further north (Fig. 1). In outline, the successions (Fig. 3) demonstrate a diachronous change from non-marine to shallow marine environments that becomes younger northward, with fluctuations in the depths of deposition, perhaps mostly of small magnitude, but culminating in an abrupt deepening coincident with the base of the VI Biozone. This resulted in a transgression that carried the zone of accumulation of shallow water sands to north of the margin of the South Munster basin. After the initial phase of deepening, shallow water sands gradually prograded southward into the basin again.

The lithostratigraphic terminology that has been employed in the Munster and South Munster Basins is complex, reflecting the large number of researchers who have worked in the area and difficulties of correlation. In what follows, the evolving palaeoenvironments are described using a simplified lithostratigraphy in which the number of formations to which reference is made is kept to a minimum. Formations that span the Devonian-Carboniferous boundary and that are typical of the South Munster Basin are described first, followed by those to the north of the basin. Figure 3 shows fence diagrams using selected sections to illustrate the stratigraphical and thickness relationships within the basin.

South Munster Basin

Castlehaven Formation

The Castlehaven Formation (Graham 1975a) is the youngest dominantly red bed formation found in western part of the South Munster Basin and has been mapped from Kenmare Bay (Pracht and Sleeman 2002) to the south coast of County Cork, as far east as the Seven Heads (Fig. 1). It is several hundred metres thick and consists of predominantly purple mudrocks with minor sandstone, interpreted as the deposits of flood plains and associated fluvial channels. There are no reports of miospores from the formation, but the overlying Toe Head Sandstone Formation (see below) at the Seven Heads and Galley Head (Figs. 1 and 3) were reported to contain LL Biozone assemblages (Clayton and Higgs 1979).

Toe Head Formation

The Toe Head Formation (Graham 1975b) succeeds the Castlehaven Formation and has been recognised from Ballycrovane in the west (Pracht and Sleeman 2002), across Bantry and Dunmanus bays (Naylor 1975; Naylor et al. 1977) and along the south coast to the Seven Heads (Fig. 1). It has also been identified at Grab-all Bay on the west side of the mouth of Cork Harbour (Higgs and Higgs 2015) (Fig. 1), where it overlies a red bed sequence similar to the Castlehaven Formation. The thickness of the formation is generally several hundred metres. It consists mostly of grey-green, generally fine-grained sandstones, with subordinate mudrocks, which, in contrast to the underlying Castlehaven Formation, are mostly grey-green in colour, with only a small proportion of purple beds. Mud cracks are common. Flow directions measured from sandstone beds are broadly unidirectional at any one locality, generally from the west in the south west and from the north further north (Fig. 4). No marine macrofauna has been found although bioturbation is widespread in the mudrocks. Plant fragments have been recorded from many localities, and well-preserved specimens of *Archaeopteris hibernica* and *A. macilenta* were noted from Toe Head by Connery (1999). There are also records of the large non-marine bivalve *Archanodon* (Clayton et al. 1981). The formation has been interpreted as having been deposited in non-marine environments on a coastal plain. This is consistent with the general lack of scolecodonts in palynological preparations, although Higgs and Higgs (2015) interpreted the occurrence of scolecodonts from mudrocks near the top of the formation at Grab-all Bay, Cork Harbour as indicating the development of estuarine environments.

Everywhere that productive palynological samples have been obtained from the lower part of the Toe Head Formation, they have yielded LL Biozone assemblages (see MacCarthy 2007; Fig. 3), but it is not possible to establish whether its lower boundary is diachronous. The boundary with the overlying Old Head Sandstone Formation is clearly diachronous, recording a marine transgression, and is discussed below.

Old Head Sandstone Formation

The Old Head Sandstone Formation is recognised throughout the South Munster Basin. Its base is identified by the lowest significant occurrence of heterolithic rock types—flaser-bedded sandstones and linsen-bedded mudrocks—marking the establishment of marine conditions. The type section is at the Old Head of Kinsale (Naylor 1966) (Fig. 5). Originally, the formation was divided into a lower Bream Rock Member (550-m thick on the Old Head of Kinsale, where the base is not seen; 261-m thick on the east side of the Seven Heads) and an upper Holeopen Bay Member (290-m thick on the Old Head of Kinsale; approximately 229-m thick on the east side of the Seven Heads) (Kuijpers 1972; Naylor et al. 1974). Kuijpers (1972) interpreted the lower part of the Bream Rock Member as the deposits of intertidal mud flats and the upper part as representing subtidal environments with low energy tidal currents. Although the Bream Rock Member has not been mapped away from the south coast sections, Quin (2008) has described two members, which can be recognised throughout the basin, which together constitute most of the former Holeopen Bay Member. Along the south coast and in the east of the basin, the older Daunt Member is sand rich and has been tentatively interpreted as having been deposited close to a major distributary system under the influence of waves and tides. In the west of the basin, the depositional environments have been interpreted as representing tidally influenced coastal settings and more mud-rich, wave influenced, shallow shelf settings. The boundary between the Daunt and Tower Members was shown by Quin (2008; Fig. 11) as a transgressive surface, and a second transgressive surface was used to divide the Tower Member into two. The member is in general more mud-rich than the Daunt Member and also shows more evidence of storms. The evolving palaeogeography during the deposition of the Bream Rock Member, the Daunt Member and the younger Castle Slate Formation is shown in Fig. 4.

The initial ingress of the sea was along the south coast between Galley Head and the mouth of Cork Harbour (Fig. 4) during LL biozonal time (Keegan 1977; Sleeman et al. 1978; Clayton and Higgs 1979; Higgs and Higgs 2015). In the west of the basin, marine conditions spread northward, reaching Dunmanus Bay (Naylor et al. 1977) by LN biozonal times and extending to the margin of the basin in Kenmare Bay (MacCarthy 2007; Fig. 3) also during LN biozonal times. In the east of the basin (Clayton et al. 1974; Sleeman et al. 1978), the base of the Old Head Sandstone becomes younger across Ringabella Bay (Fig. 1). On the south side of the bay (South Ringabella), the lowest beds contain an LL biozonal assemblage; on the north side (North Ringabella), where the formation is much thinner (67 m in contrast to more than 580 m to the south), the base of the formation lies just above the LE/LN biozonal boundary, and further north in Cork Harbour, LN biozonal assemblages occur below the base.

Macrofossils, mostly poorly preserved moulds of brachiopods and molluscs in sandstones, have been recorded from several localities, such as Bere Island, where they were reported to be abundant (Pracht and Sleeman 2002), but have not been identified to species level with the exception of those listed by Turner (1939, 1951) and Naylor et al. (1969). Naylor et al. (1969) provided a discussion of faunas from the Bream Rock Member on the Old Head of Kinsale and from the Daunt Member on the South Ringabella section that include spiriferids (*Sphenospira julii*), and rhynchonellids (identified as *Camarotoechia* cf. *letiensis* and *C.* cf. *moresnetensis*). On Flat Head (Fig. 1), the youngest beds of the Daunt Member include thin limestones, from which Turner (1951) recorded *Rhipidomella michelini*, *Spirifer tornacensis* and *Syringothyris cuspidata*, which he referred to the Zaphrentis Zone (Z_2) of Vaughan (1905). The latter equates with a horizon in the late Tournaisian. The fauna from Flat Head requires re-examination, and all of the faunas from the Old Head Sandstone Formation deserve further study.

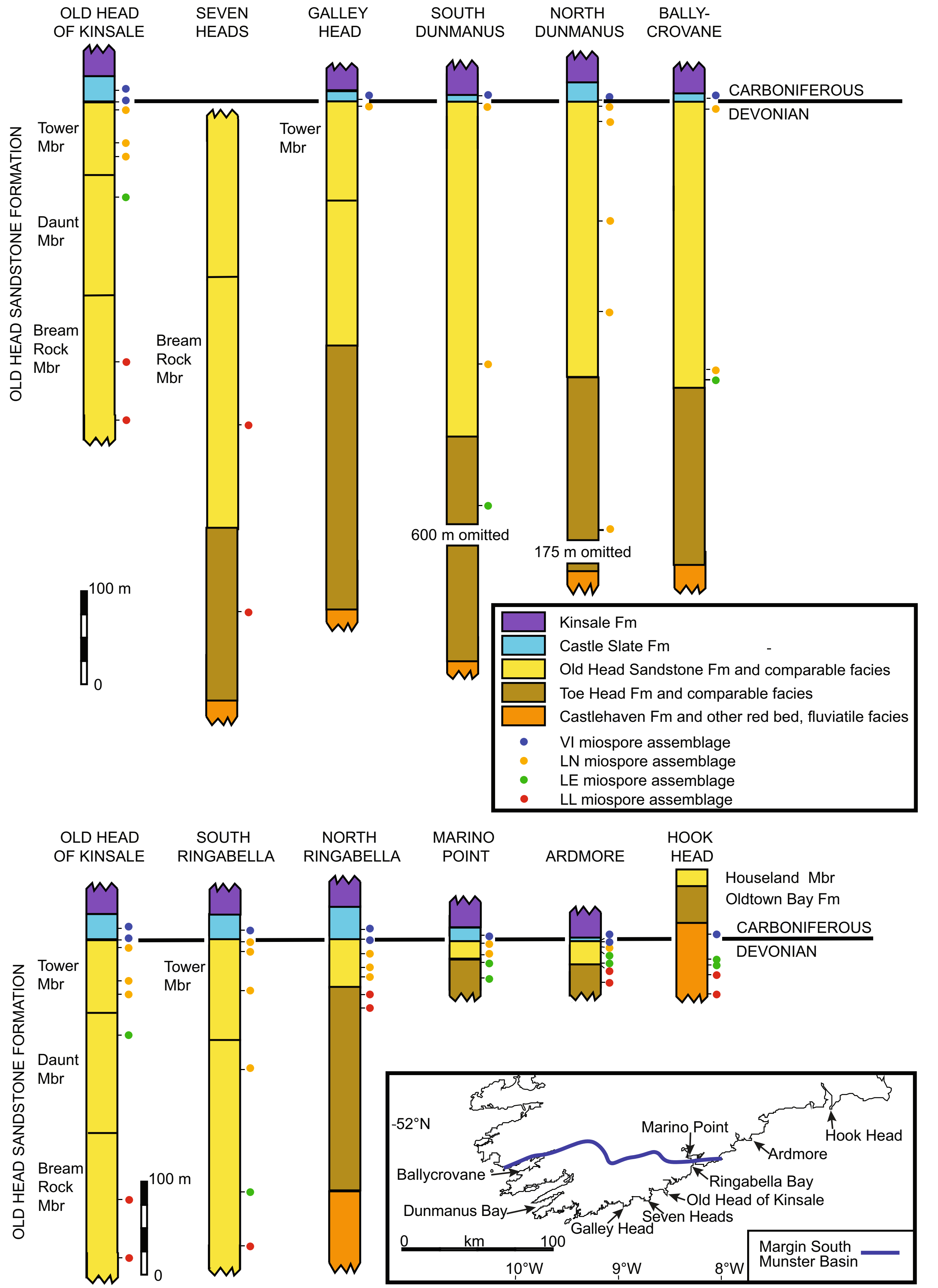
Castle Slate Formation

This formation has been recognised throughout the South Munster Basin (Fig. 3) and oversteps the basin margin in the east (see discussion of Ardmore below). It consists of dark grey, laminated mudrocks, containing common non-calcareous nodules and, at some levels, thin stringers of limestone. The upper part of the formation contains increasing proportions of silt and fine sandstone heralding the sandstones of the overlying Kinsale Formation. The type section is on the Old Head of Kinsale (Figs. 5, 6), where the formation is 62-m thick and where its base was selected as the stratotype of the Courceyan Regional Stage (now redundant) by George et al. (1976). It thins to 10 m at Ballycrovane (Fig. 1) in the north-west of the basin, to 25 m at Marino Point in Cork Harbour and to 4.8 m at Ardmore (see below).

Everywhere in the basin, the lower part of the formation has been assigned to the VI Biozone. In some early studies (e.g. Sleeman et al. 1978; Gardiner and MacCarthy 1983), it was reported that in places the base of the formation contains LN biozonal assemblages, but these records have now been revised (Clayton et al. 1982, 1986). Cephalopods, including both ammonoids and nautiloids, have been identified at several localities. Matthews (1983) identified *Imitoceras* cf. *prorsum* and *Imitoceras* sp. from Nohoval Cove (Fig. 1). The limestone stringers in the south coast sections contain crinoids, ostracodes and conodonts. MacCarthy (1974; Fig. 4) and MacCarthy et al. (1978) recorded crinoids and orthoconic nautiloids at Whitebay, Inch and Ballycotton (Fig. 1). The conodont faunas at the Old Head of Kinsale (Matthews and Naylor 1973) contain *Bispathodus aculeatus*, *Bispathodus stabilis*, *Patrognathus variabilis* and *Polygnathus communis*. The presence of *Patrognathus* and the absence of protognathodids and siphonodellids suggest a shallow water depositional environment. Further sampling for conodont analysis is required to substantiate this conclusion.

The contact between the Old Head Sandstone and the Castle Slate Member

The contact between the Old Head Sandstone Formation and the Castle Slate Formation is always sharp (Fig. 6) and records


OLD HEAD OF KINSALE
SEVEN HEADS
GALLEY HEAD
SOUTH DUNMANUS
NORTH DUNMANUS
BALLY-CROVANE
CARBONIFEROUS
DEVONIAN
OLD HEAD SANDSTONE FORMATION
Tower Mbr
Daunt Mbr
Bream Rock Mbr
600 m omitted
175 m omitted
100 m
0
Kinsale Fm
Castle Slate Fm
Old Head Sandstone Fm and comparable facies
Toe Head Fm and comparable facies
Castlehaven Fm and other red bed, fluviatile facies
VI miospore assemblage
LN miospore assemblage
LE miospore assemblage
LL miospore assemblage
OLD HEAD OF KINSALE
SOUTH RINGABELLA
NORTH RINGABELLA
MARINO POINT
ARDMORE
HOOK HEAD
Houseland Mbr
Oldtown Bay Fm
CARBONIFEROUS
DEVONIAN
OLD HEAD SANDSTONE FORMATION
Tower Mbr
Daunt Mbr
Bream Rock Mbr
100 m
0
-52°N
Marino Point
Hook Head
Ardmore
Ballycrovane
Ringabella Bay
Old Head of Kinsale
Dunmanus Bay
Seven Heads
Galley Head
0 km 100
10°W
9°W
8°W
Margin South Munster Basin

◀ **Fig. 3** Stratigraphical sections and positions of miospore assemblages of Late Devonian and early Carboniferous rocks of the south of Ireland from the Old Head of Kinsale west and north to Ballycrovane and east and north to Hook Head. Sources of data are cited in the text

a regionally important deepening event. The lower part of the Castle Slate Formation has not been studied in the detail it deserves, most authors reporting that it consists of dark grey mudrock with small siliceous nodules; regional differences, apart from thicknesses, cannot be gleaned from the literature. However, the highest beds of the underlying Old Head Sandstone Formation have received considerable attention. The details of the highest beds vary regionally but generally contain the coarsest sandstones of the formation, medium-grained sandstone with crinoid ossicles from Galley Head to the Old Head and granules and pebbles further east (Fig. 4). The succession has been interpreted to represent high energy barrier beach and associated lagoonal deposits (Graham 1975a) or a transgressing storm-dominated microtidal barrier (Gardiner and MacCarthy 1983). Quin (2008) has suggested differing origins for the sandstones at the top of the Old Head Sandstone Formation in different areas as shown in Fig. 4. In

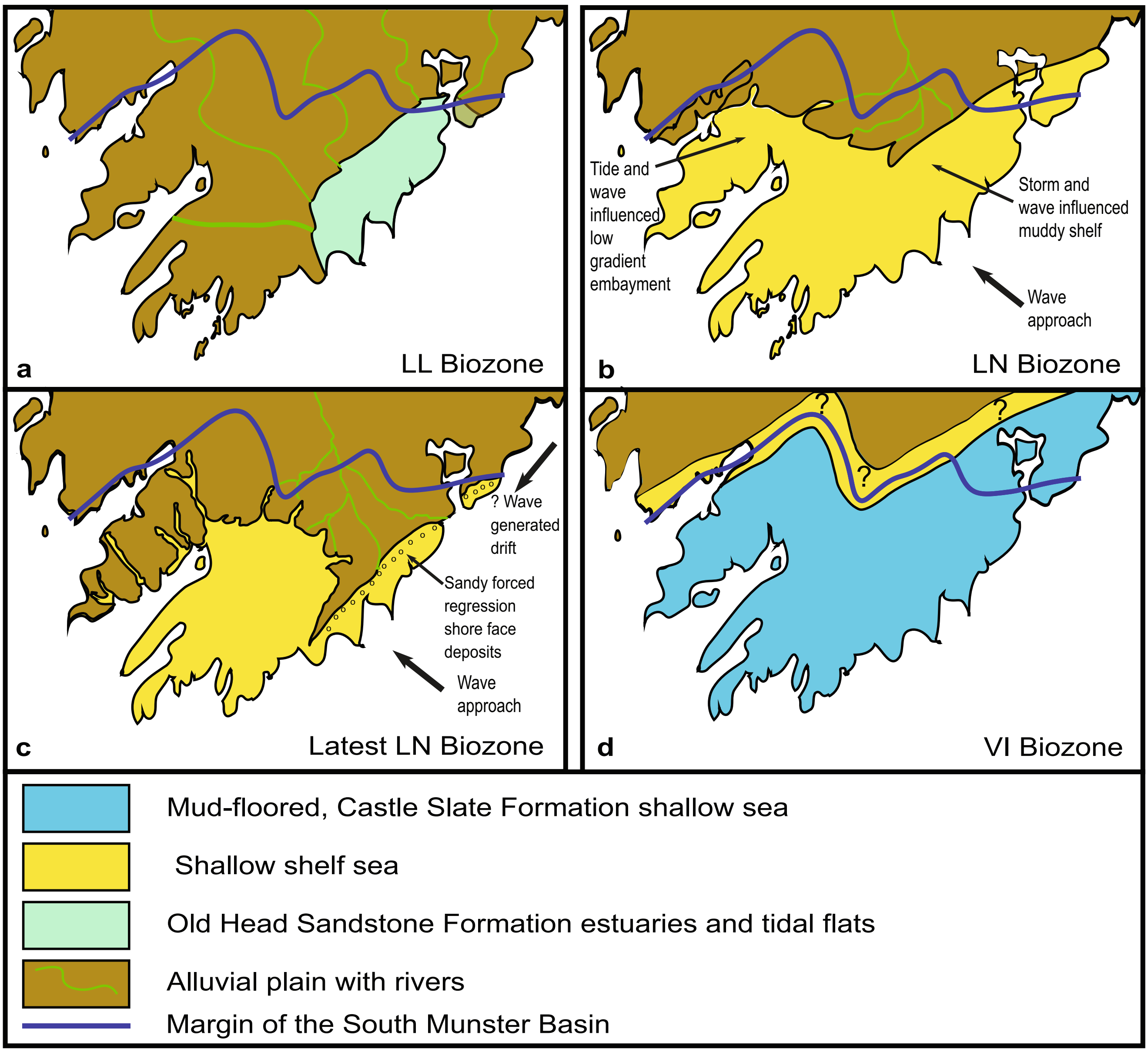


Fig. 4 Palinspastic maps of the evolving palaeogeography of the South Munster Basin during the latest Devonian and early Carboniferous. **a** Initial marine transgression in LL biozonal time, the Bream Rock Member of the Old Head Sandstone Formation. **b** High sea level during LN biozonal time, the Tower Member of the Old Head Sandstone Formation. **c** Low sea level during latest LN biozonal time, top of the Tower Member of the Old Head Sandstone Formation. **d** The Castle Slate Formation transgression in VI miospore biozonal time. (**a**) incorporates information from Higgs and Higgs 2015; (**b**) and (**c**) are adapted from Quinn (2008; Figs. 12C and 12D, respectively)

several localities, such as Flat Head (Fig.1), the uppermost sandstone body is overlain by a thin unit of mudrock, containing siltstone linsen and thin limestones, in some cases capped by a thin sandstone bed (Quin 2008). These facies have been interpreted as related to high frequency oscillations of sea level, prior to the Castle Slate transgression.

Volcanic tuffs in the Late Devonian and early Carboniferous

Claystone bands, shown to have originally been volcanic tuffs, occur in the Toe Head Formation, the Old Head Sandstone Formation and the Castle Slate Formation (Naylor 1966; Graham and Reilly 1976; Pracht and Batchelor 1998). Some samples were reported by Pracht and Batchelor to contain euhedral zircons. It would be worthwhile to attempt to obtain radiometric ages from these samples because most of them are well constrained stratigraphically with associated palynological samples.

Region north of the South Munster Basin

In general, the earliest marine strata north of the South Munster Bain are of VI Biozone or younger age (Clayton and Higgs 1979). This can be illustrated by brief descriptions of the successions at a few selected localities (Fig. 1).

Kenmare

The Kenmare section is located immediately north of the margin of the basin. According to MacCarthy (2007; Fig. 3), fluvial facies persisted there through VI and HD biozonal times. Shallow marine environments first occurred during the younger BP Biozone, well into the Tournaisian.

Kerry Head

The marine transgression at Kerry Head has been described by Diemer et al. (1987). The succession consists of more than 700 m of sandstones and mudrocks deposited by southerly flowing rivers, overlain by tidally influenced, shallow marine deposits. The uppermost fluviatile Kilmore Formation, 208-m thick, consists of grey and green sandstone interbedded with mudrocks with abundant plants. It has been reported to be of VI biozonal or younger age. Miospore assemblages from close to the base of the overlying shallow marine deposits are of PC biozonal age, well into the Tournaisian.

Ballyvergin borehole

Clayton et al. (1980) recorded a HD biozonal assemblage from the base of marginal marine sandstones, which overlie predominantly red sandstones and mudrocks of fluviatile origin.

Kiltorcan

The Kiltorcan Formation (Colthurst 1978) is characterised by thick, non-red sandstones and grey, green and yellow mudrocks, interpreted as the deposits of rivers flowing south across a low gradient, coastal plain. At Kiltorcan, it contains a rich flora and non-marine fauna, including *Archaeopteris hibernica*, *Cyclostigma kiltorkense*, the bivalve *Archanodon jukesi*, the fish *Groenlandaspis* and eurypterids. Jarvis (1990) demonstrated that the Kiltorcan biota is bracketed by two horizon containing miospores of the LE and VI Biozones, respectively. Clayton et al. (1977) recorded a BP biozonal assemblage from higher in the formation, so the earliest marine deposits in the region are clearly within the Tournaisian.

Hook Head

The Hook peninsula (Figs. 1 and 3) provides a complete section of fluviatile Old Red Sandstone (approximately 350-m thick) overlain by shallow marine facies, from which Higgs (1975) reported several miospore assemblages. Assemblages of LL biozonal age were recovered from two horizons within pebbly sandstones and red mudrocks of fluviatile origin, 90 m and 72 m below the earliest marine strata. An assemblage of LE biozonal age was identified from 6 m above the younger LL assemblage and is separated from the oldest VI assemblage by 20 m of red mudrock. Above this horizon, the succession (the Oldtown Bay Formation; Fig. 3) probably represents an alluvial coastal plain analogous to the Toe Head Formation of the South Munster Basin. The oldest shallow marine deposits (Houseland Sandstone Member; Fig. 3) contain a miospore assemblage of HD biozonal age.

Ardmore

The sections at Ardmore and neighbouring Whiting Bay (Figs. 1 and 3) are of particular interest because facies typical of the South Munster Basin, notably a development of the Castle Slate Formation, are succeeded by the carbonate-dominated succession typical of the region north of the South Munster Basin. The Ballyquin Member of the Gyleen Formation (MacCarthy et al. 1978; Sleeman and Pracht 1994), consisting of several hundred metres of thick red and grey sandstones alternating with thick red mudrocks, is succeeded by the Ardmore Member, over 100 m of alternating grey and pale red sandstones and grey and yellow mudrocks. Both members were interpreted as being of fluviatile origin. MacCarthy et al. (1978) described the Dysert Member (30 m of grey sandstones and mudrocks), which overlies the Ardmore Member and recorded flaser bedded mudrocks and linsen sandstones, suggesting a tidally influenced environment. Clayton et al. (1982) recorded LL biozonal assemblages from the Ardmore Member and LE biozonal assemblages from all

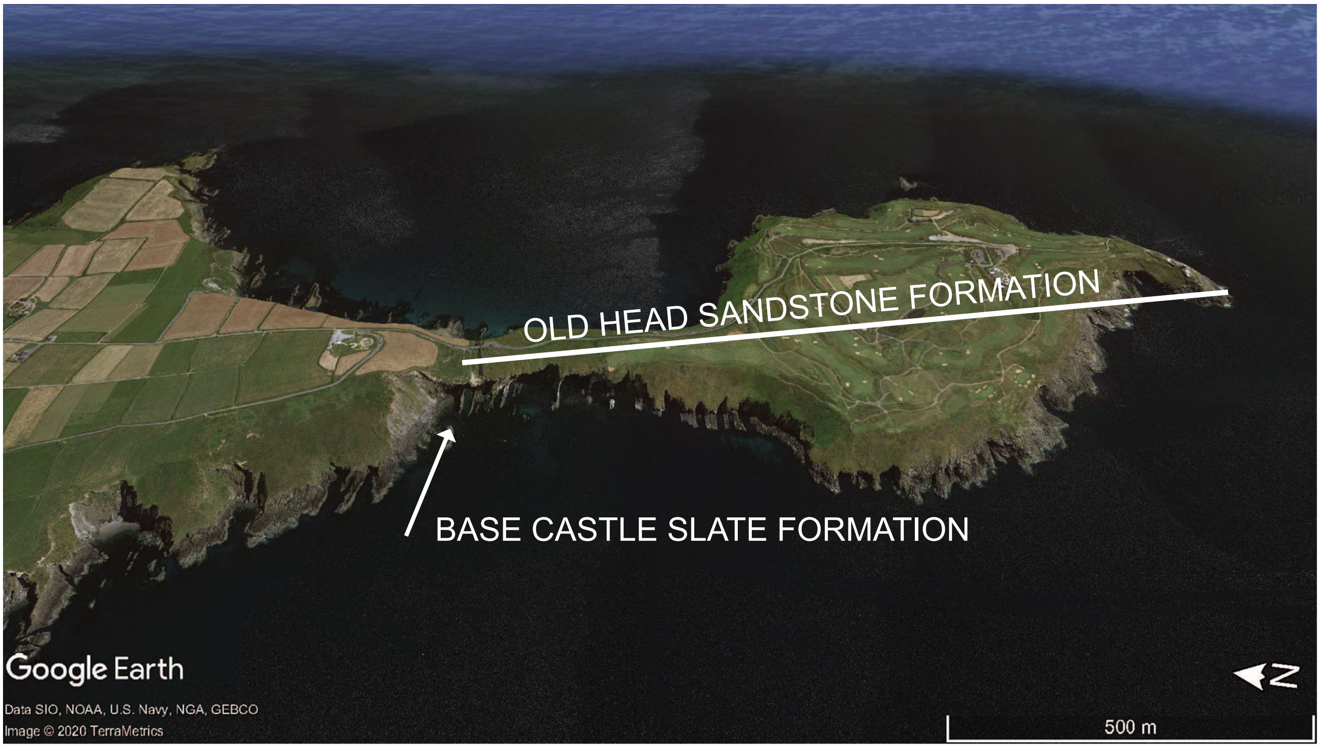


Fig. 5 The southern part of the west coast of the Old Head of Kinsale, showing the Old Head Sandstone (here over 840-m thick) and Castle Slate Formations (Image by Google Earth 2020 TerraMetrics)

but the highest 4.8 m of the Dysert Member. This upper part of the member yielded LN biozonal assemblages and the uppermost 0.4 m a VI biozonal assemblage. The overlying Castle Slate Formation, here 4.8-m thick, yielded VI biozonal assemblages. A calcareous nodule 1.75 m above the base yielded a fauna of conodonts very similar to that from the Old Head of Kinsale with *Patrognathus variabilis*. Other elements of the fauna included fish teeth and scales, silicified ostracodes and a juvenile ammonoid.

Implications of the selection of a new GSSP for identification of the base of the Carboniferous in Ireland

It has been agreed that a new GSSP for the base of the Carboniferous should be sought to replace the current GSSP at La Serre, Montagne Noire, France. A Working group has identified three potential levels at which the base could be chosen. They are in ascending stratigraphical order: the main extinction event at the base of the Hangenberg Black Shale in Germany; the end of the mass extinction and the top of the major regression just below the base of the *Protognathodus kockeli* Zone; and the base of the *Protognathodus kuehni* Zone/basal *Siphonodella sulcata* Zone and coastal plant extinction (i.e. the extinction of *Retispora lepidophyta* and associated taxa).

The lowest of these proposed levels is at the base of the *costatus-kockeli* interregnum of the revised conodont biostratigraphical scheme of Kaiser et al. (2009), equivalent to the base of the LN miospore Biozone, as discussed above. It might be possible to identify this boundary accurately in sections in the South Munster Basin based on miospores. The boundary would not track lithostratigraphical boundaries.

The second of these proposed levels cannot be identified using miospores because it is equivalent to an interval within the LN miospore Biozone. It therefore cannot be recognised in the South Munster Basin.

The youngest of the proposed levels conflates the conodont and miospore records, although it has yet to be proved that the base of either the *Protognathodus kuehni* Zone or *Siphonodella sulcata* Zone is coincident with the base of the VI miospore Biozone. However, if the base of the VI miospore Biozone is accepted as a proxy for the base of the Carboniferous, identification of this boundary in southern Ireland will be straightforward: the boundary will coincide with the base of a widespread lithological unit, the Castle Slate Formation.

Glaciation in Gondwana and the record of changes of sea level in the South Munster Basin

There is abundant evidence of Late Devonian glaciation (Isbell et al. 2003) in basins in western Gondwana (Peru, Bolivia, northern Brazil and Central Africa; Caputo 1985; Wicander et al. 2011) and in more equatorial areas in the Appalachians (Brezinski et al. 2009, 2010; Ettensohn et al. 2020). Lakin et al. (2016) have reviewed the palynostratigraphy of the glacial sequences in west Gondwana and have shown that glaciation certainly occurred during LN Biozone times and may have occurred through the total range of *Retispora lepidophyta* that is through the LL–LN Biozones. The Appalachian examples are also of LN biozonal age (Coleman and Clayton 1987). The duration of the glaciation and therefore the estimations of rates of glacioeustatic sea level changes are poorly constrained because of the paucity of accurate and precise radiometric ages. Becker et al. (2012; Fig. 22.11) showed the combined duration of the LL, LE and LN Biozones as 1.5 Ma. Denayer et al. (in press, this issue) used calculations based on assumed orbital forcing parameters to estimate the duration of the same time interval as 0.84–0.89 Ma. Davydov et al. (2012; Fig. 23.5) showed the equivalent of the early Carboniferous VI Biozone as having a duration of less than 0.5 Ma, and this may be an overestimate, based on the relative thicknesses of miospore

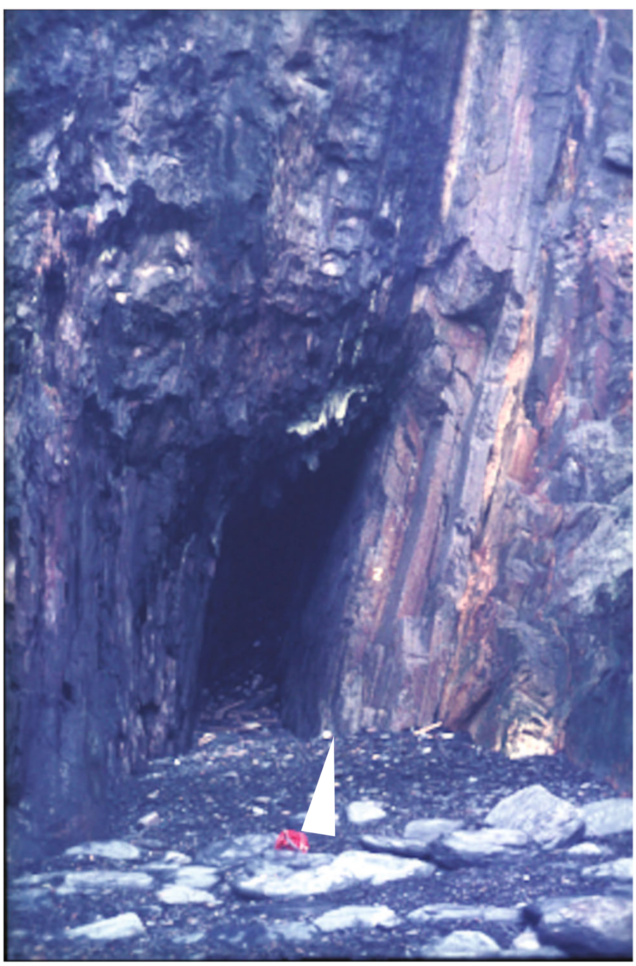


Fig. 6 The contact (arrowed) between the Old Head Sandstone and Castle Slate Formations on the west coast of the Old Head of Kinsale (width of red rucksack in foreground and length of hammer handle both 0.3 m)

Biozones in southern Ireland. Myrow et al. (2014) have reported high resolution ID-TIMS U-Pb ages from zircons from two ash horizons that bracket the ‘Hangenberg’ Black Shale at Kowala Quarry, Poland. The difference in age of the two samples is 310 Ka, and on the assumption that the sedimentation rate was constant, they calculated that the crisis had a maximum duration of 150 Ka and could have been less than 50 Ka. It should be noted, however, that Filipiak and Racki (2010; Fig. 2) showed the base of the LN Biozone at Kowala to be substantially below the dated horizon of black shale.

Many authors have suggested that Late Devonian glacioeustatic changes of sea level were directly or indirectly the cause of the Hangenberg Biocrisis/Event. Kaiser et al. (2016) have provided the most detailed interpretations of sea level changes linked to the crisis. They suggested that the first phase, in the LE Biozone, involved a minor drop in sea level. This was followed in the early part of the LN Biozone by widespread deposition and burial of organic carbon-rich black shale under warm conditions in the equatorial zone. The burial of carbon leads to a massive drawdown of CO_2 which triggered the glaciation in western Gondwana during the LN Biozone and that resulted in a rapid fall in sea level, which Kaiser et al. (2011) estimated to have been on the order of 100 m. In the Rheinisches Schiefergebirge, the sea level fall resulted in the deposition of the Hangenberg Sandstone in what had been a long lived pelagic sedimentary environment. The initial post-glacial warming during the upper part of the LN Biozone (corresponding to the *kockeli* conodont Biozone) resulted in widespread transgression. Further poorly understood minor fluctuations of sea level took place in the youngest part of the Devonian and sea level and then rose through the early part of the Tournaisian.

There are varying opinions on the magnitude of glacioeustatic changes in sea level during the latest Devonian. As Montañez and Poulsen (2013) have pointed out, late Palaeozoic glacial records have been commonly interpreted by analogy with Pleistocene glacial history, particularly in relation to the extent of ice cover. Rygel et al. (2008) reviewed published estimates of glacioeustatic changes of sea level during the Carboniferous and Permian, showing the largest changes (100–160 m) within the time interval late Viséan to lower Permian, approximately corresponding to the times of greatest extent of ice cover in Gondwana. Isbell et al. (2003) concluded that the Late Devonian western Gondwanan record suggests glaciomarine deposition near the tidewater terminus of alpine glaciers and a similar setting also applies to the Appalachian glacial deposits. It seems unlikely that fluctuations of ice centres of small size that fed valley glaciers which debouched into marine environments would have resulted in eustatic changes of sea level, comparable in magnitude of those of the late Viséan to Lower Permian.

During the glaciation of western Gondwana, the Old Head Sandstone Formation and possibly also the Toe Head Formation were being deposited in southern Ireland. The exceptionally thick sequences of rocks deposited in coastal plain and tidally influenced, wave- and storm-dominated, shallow marine environments in the South Munster Basin reflect a near equilibrium of relative sea level over a period of at least 1 Ma, maintained by rapid subsidence (several hundred metres per million years) almost balanced by rapid sediment supply and eustatic sea level change. Although it is not possible to parse the relative contributions of rate of sediment supply, tectonic subsidence, sediment and water loading and eustatic sea level change to the changing depth of the depositional environments in the South Munster Basin, it seems likely that the sedimentary system would have been very sensitive to large and rapid glacioeustatic sea level changes. The transgressive surfaces in the Old Head Sandstone Formation recognised by Quin (2008) record relative sea level fluctuations, but the magnitude of these is unlikely to have exceeded a few tens of metres. A eustatic sea level drop of 100 m as suggested by (Kaiser et al. 2011) seems unlikely.

The most conspicuous change of facies around the Devonian/Carboniferous boundary in the South Munster Basin is the end- or post-glacial transgression signalled by the deposition of the Castle Slate Formation. It is possible to make very approximate estimates of the amount of sea level rise that contributed to the transgression. At Ardmore (Fig. 3), the Castle Slate is 4.8-m thick, which assuming an original porosity of 80% would represent a decompacted thickness of 24 m. The overlying Cuskinny Member of the Kinsale Formation represents a shallow coastal environment with storm-generated gravel barriers and beaches, at first approximation indicating a similar depth to the Dysert Member, which underlies the Castle Slate. Since there is no reason to suggest that the rate of subsidence at Ardmore changed substantially during latest Devonian/Carboniferous time, it is likely that the Castle Slate transgression reflected a maximum sea level rise of substantially less than 24 m. Similar calculations further south, where the Castle Slate Formation is much thicker and subsidence rates much higher, lead to similar results. In conclusion, evidence from the South Munster Basin does not support estimates of glacioeustatic changes of sea level of large magnitude during latest Devonian and earliest Carboniferous time.

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Compliance with ethical standards

Conflict of interest The authors declare that they have no conflict of interest.

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