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A synopsis of the Ordovician System in its birthplace - Britain and Ireland

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Abstract

Rock successions in Britain and Ireland, and more especially those in North Wales, were instrumental in the founding and naming of the Ordovician System, and the Anglo-Welsh series established both initially and subsequently were used widely as a standard for Ordovician chronostratigraphy. Although now largely superseded in the global scheme of series and stages, they retain their local and regional importance. The Ordovician System in Britain and Ireland documents the history of a segment of the Earth's crust that incorporated opposing peri-Gondwanan and peri-Laurentian/Laurentian margins of the Iapetus Ocean during its closure, and is accordingly complex. The complexity arises from the volcanic and tectonic processes that accompanied oceanic closure coupled with the effects of eustatic sea-level changes, including the far-field effects of the Late Ordovician glaciation. For the past three decades, Ordovician successions in Britain and Ireland have been discussed in terms of terranes. Here we review Ordovician successions in each terrane, incorporating the results of recent research and correlating those successions via biostratigraphical schemes and radiometric dates to the global Ordovician series and stages.

The Ordovician rocks of Britain and Ireland, and especially those of Wales, are historically important as being those on which the system was established. The Anglo-Welsh series – originally Tremadoc (formerly included in the Cambrian System), Arenig, Llanvirn, Llandeilo, Caradoc and Ashgill – provided important points of reference for Ordovician chronostratigraphy, and although now largely superseded in global Ordovician stratigraphy, still do so locally and regionally. Moreover, the work carried out over many years on Ordovician successions throughout Britain and Ireland has illuminated the history of a segment of crust that incorporated the opposite margins of the Iapetus Ocean during its closure. Ordovician stratigraphy in Britain and Ireland is accordingly complex, a consequence of the volcanic and tectonic processes that accompanied oceanic closure on the peri-Gondwanan, peri-Laurentian and Laurentian margins, coupled with the effects of eustatic sea-level changes. The latter include, for example, far-field effects of the end-Ordovician glaciation, which affected peri-Gondwanan successions, and Early–Middle Ordovician eustatic changes that were widespread across Laurentia and are seen in the succession of the Hebridean Terrane.

The history of the Ordovician System from its inception by Charles Lapworth (1879) to the present has been reviewed by Harper *et al*. (this volume). The Anglo-Welsh series, stages and substages used in this account follow the revisions introduced by Fortey *et al*. (1995) and used by Fortey *et al*. (2000). Correlation of the Anglo-Welsh chronostratigraphical divisions with standard global series and stages follows Cooper and Sadler (2012) and Goldman *et al*. (2020).

The Ordovician stratigraphy of Britain and Ireland is reviewed here in terms of terranes - geologically coherent volumes of crust separated from each other by major faults that have hosted regionally significant displacement. First developed in the Cordilleran belt of western North America (Coney *et al.* 1980), the terrane concept was soon applied to Britain and Ireland (Dewey 1982) and a scheme was developed (Bluck *et al.* 1992) that has wide agreement. Following an outline of the terranes recognised in Britain and Ireland, the Ordovician successions in each terrane are summarized and illustrated. In addition to references cited in the text, details of many key Ordovician localities in Great Britain are included in Geological Conservation Review series volumes 17 (Stephenson *et al*. 1999) and 18 (Rushton *et al*. 2000). Place names mentioned below in England, Wales, Scotland, the Isle of Man and Northern Ireland can be located using the British Geological Survey GeoIndex and those in the Republic of Ireland can be located on the Geological Survey of Ireland's online maps: see note on online resources below.

Tectonic terranes

The scheme of terranes developed by Bluck et al. (1992) was the basis of the terrane map (Woodcock 2000) used in the comprehensive review of Ordovician stratigraphy in Britain and Ireland by Fortey *et al.* (2000). The map used here (Fig. 1) shares many features with the map published by Woodcock (2000) but incorporates significant developments, particularly concerning the terranes southeast of the Iapetus Suture.

Most of the terranes host exposed Ordovician rocks, but the relationships of those rocks to the bordering terrane boundaries varies (Fig. 1). Where the latest displacement on a terrane boundary is post-Ordovician, there are contrasts in Ordovician stratigraphy across the boundary. Where the latest displacement is pre- or intra-Ordovician, then all or the upper part of the Ordovician stratigraphy is continuous across the terrane boundary. Examples of such overstepped terrane boundaries are those separating the components of the Avalon Composite Terrane of southern Britain. These components were amalgamated in latest Neoproterozoic or earliest Cambrian time (Pharaoh and Carney 2000). It has been argued, partly on North American evidence, that a continuous late Ediacaran to Ordovician sequence once also overstepped onto the Monian and Leinster-Lakesman terranes (Landing *et al*. 2022). Here, the possibility of large early Palaeozoic displacement on the northwestern edge of the Avalon terrane is retained.

The most noteworthy terrane boundary in Britain and Ireland is the Iapetus Suture (Fig. 1), with peri-Laurentian terranes to the NW and peri-Gondwanan terranes to the SE. This boundary marks the site of final closure of the Iapetus Ocean, which separated the palaeocontinents of Laurentia and Gondwana from at least late Ediacaran until mid-Silurian time (550–425 Ma). The terranes are briefly described from NW to SE.

Laurentian terranes

The *Hebridean Terrane* has a basement of Archean and Palaeoproterozoic metamorphic rocks (Lewisian Complex) overlain unconformably by Meso- and Neoproterozoic sedimentary rocks (Stoer, Sleat and Torridon groups). These are unconformably overlain in turn by a Cambrian–Ordovician marine sequence (Ardvreck and Durness groups). Detrital zircons from the Ardvreck Group (Cawood *et al.* 2007) and faunas from the Durness Group (Prigmore and Rushton 2000) confirm that the Hebridean Terrane was part of the Laurentian continent in the Cambrian–Ordovician.

The *Northern Highlands Terrane* is separated from the Hebridean Terrane by the gently southeast-dipping Moine Thrust Zone. The Northern Highlands Terrane is dominated by Neoproterozoic metasedimentary rocks assigned to the Moine Supergroup. Krabbendam *et al.* (2021) correlated the lowest Morar Group of the Moine with the Sleat and Torridon groups of the Hebridean Terrane, a correlation that implies facies continuity across the Laurentian margin on length scales greater than the 100 kilometre or more displacement on the thrusts of the Moine Thrust Zone. This also questions the role of the Moine Thrust as a major terrane boundary (Krabbendam *et al.* 2021). Regardless, any former Cambrian-Ordovician cover that might have been present in the Northern Highlands Terrane had been eroded by the Devonian.

The *Grampian (Central Highlands) Terrane* and the detached *Connemara Terrane* are dominated by the mostly metasedimentary upper Neoproterozoic to Lower Palaeozoic rocks of the Dalradian Supergroup. The Grampian Terrane is separated from the Northern Highlands Terrane by the steep Great Glen Fault on which strike-slip displacement of at least 700 kilometres has been postulated (Dewey and Strachan 2003). However, recent work on the Neoproterozoic stratigraphy of these blocks suggests that there is greater depositional and structural continuity than has generally been assumed (Krabbendam *et al.* 2021). Tentative correlations of lower Neoproterozoic successions in the Grampian and Northern Highlands terranes imply that either the 700 km estimates of displacement are too large or that facies similarities along the Caledonian–Scandian belt persisted over such large distances (Krabbendam *et al.* 2021). The relationship of the Grampian Terrane to the Laurentian margin, as represented in the Hebridean Terrane, nevertheless remains problematic.

Of relevance to Ordovician stratigraphy is whether and how far the Dalradian Supergroup extends up into the Lower Palaeozoic. A possible Ordovician acritarch, but very poorly preserved, has been reported from the upper part of the Dalradian Southern Highland Group in NE Scotland (Molyneux 1998). This result has not been replicated, however, and the putative Ordovician age of the relevant sample remains unconfirmed and uncertain. Nevertheless, upward continuity of the Dalradian Supergroup into the Highland Border Complex, which includes Ordovician rocks along the SE margin of the Grampian Terrane, was argued by Tanner and Sutherland (2007; see also Stephenson *et al*. 2013 and Tanner *et al*. 2013).

Peri-Laurentian terranes

The *Midland Valley Terrane* is bounded by steeply dipping faults – the Highland Boundary Fault to the NW and the Southern Upland Fault to the SE – that bring down predominantly younger rocks between. The Scottish Midland Valley Terrane is dominated at outcrop by upper Palaeozoic rocks. At its southern margin, however, is the fragmented Early Ordovician Ballantrae ophiolite/arc igneous complex overlain by the Llanvirn to Ashgill (Darriwilian to Hirnantian) sedimentary cover of the Girvan area. Xenoliths in upper Palaeozoic intrusions suggest that other Ordovician volcanic arc rocks underlie the Midland Valley (Badenszki *et al.* 2019). The Tyrone Igneous Complex of Northern Ireland is closely analogous to the Ballantrae Complex (Graham 2009; Cooper *et al*. 2011; Hollis *et al*. 2012, 2013a, 2013b).

In western Ireland, the *South Mayo Subterrane* lies between the Fair Head–Clew Bay Line to the north and the Doon Rock Fault to the south. The subterrane has a thick Ordovician sequence of volcanic and sedimentary rocks spanning at least the late Tremadocian (Williams and Harper 1994) to Caradoc (Sandbian) (Harper *et al.* 2010).

The steep Southern Upland-Skird Rock faults mark the northern boundary of the *Southern Uplands Terrane*, the major terrane that is farthest outboard on the Laurentian margin. Ordovician rocks of the Southern Uplands Terrane are interpreted as being part of an accretionary prism of basinal sediments scraped off the northwestward-dipping subducting slab of Iapetus Ocean lithosphere (Leggett *et al.* 1979). Accretion continued southeastwards until the late Wenlock (Homerian).

Subterranes along the Iapetus Suture

The northward dipping Iapetus Suture is obscured in northern England by upper Palaeozoic and later deposits, and is better defined by seismic data offshore to the NE and SW (Soper *et al.* 1992). The Iapetus Suture in eastern Ireland was first placed (Phillips *et al.* 1976) along the Navan Fault (Fig. 1) but reinterpretation of the faunal evidence (Harper and Parkes 1989; Harper *et al.* 1990; Owen *et al.* 1992) led to recognition of the Slane Fault between the *Grangegeeth Subterrane* to the north and the *Bellewstown Subterrane* to the south as its steep surface trace. This conclusion is supported by contrasts in deformation style across the fault (Vaughan and Johnston 1992). The Grangegeeth Subterrane has a Darriwilian to Katian (Llanvirn–Ashgill) volcano-sedimentary succession with Sandbian (lower Caradoc) Laurentian shelly faunas. The Bellewstown Subterrane, by contrast, has a distinctive Tremadocian to Caradoc (Sandbian–lower Katian) sedimentary-volcaniclastic succession with a Gondwanan shelly fauna near the top.

The course of the Iapetus Suture in western Ireland is imperfectly known. Discussions of the possible western course of the suture have been provided by many authors (McKerrow and Soper 1989; Murphy *et al.* 1991; Todd *et al.* 1991; Church *et al.* 1992; Ford *et al.* 1992; Soper *et al.* 1992). The hypotheses vary between a northern course through the Shannon estuary and a southern course south of the Dingle peninsula. The compromise continuous line (Fig. 1) is that of Todd *et al.* (1991).

Peri-Gondwanan terranes

The terranes south of the Iapetus Suture are all thought to have been derived from the northern margin of the Gondwana continent, although probably from a variety of locations, from opposite Amazonia to West Africa (van Staal *et al.* 2021). An alternative case has been made by Landing *et al*. (2022) that the Ganderian, Avalonian and Megumian terranes (Fig. 1) formed as an isolated ribbon continent along a major strike-slip fault, and that any later compositional links were with Baltica rather than Gondwana. The peri-Gondwanan terrane boundaries closest to the Iapetus Suture have the same NE–SW Caledonian strike as most of the Laurentian terrane boundaries, but farther SE the boundaries have a more variable strike (Fig. 1).

The *Leinster-Lakesman Terrane* borders the length of the Iapetus Suture (Fig. 1). On all terrane maps, a southeastern boundary to the Leinster-Lakesman Terrane with the *Monian Terrane* has been taken to the NW of Anglesey (Fig. 1). Traced southwestwards into Ireland, this boundary has variously been continued (Fig. 1) into the Ballycogly Mylonite Zone (Bluck *et al.* 1992; British Geological Survey 1996), the Courtown–Tramore Line (Gibbons *et al.* 1994) or, most recently, the Wicklow Fault Zone (Pharaoh 2018; Pharaoh *et al.* 2020). The Wicklow Fault Zone is adopted here (Fig. 1) because it is compatible with correlation of the Cambrian Cahore Group in Ireland with the South Stack Group in Anglesey (Tietzsch-Tyler and Phillips 1989; Gibbons et al. 1994; Rushton *et al*. 2011) and with the close match between the Cahore Group SE of the Courtown–Tramore Line and the Bray Group to the NW (Brück and Molyneux 2011). Ordovician rocks crop out extensively and are assumed to underlie the whole terrane. No Precambrian basement has been proved within the terrane, however, and the presence of autochthonous Cambrian rocks is uncertain. Possible Cambrian rocks are present at one locality in the Skiddaw Group of NW England (Cooper *et al*. 2004) and although the Cambrian Bray Group crops out to the NW of the Wicklow Fault Zone in SE Ireland, it is considered to be thrust north-westwards, the displacement being attributed to transpressive movements on the Wicklow Fault Zone (Max *et al*. 1990).

To the northeast of Anglesey, the southern boundary of the Leinster-Lakesman Terrane has been drawn eastwards along the Pendle Lineament (British Geological Survey 1996) and into the Vale of Pickering–Flamborough Head Fault Zone (Woodcock 2011; Holdsworth *et al.* 2012). However, Soper and Dunning (2005) argued that the Ingleton Group (Fig. 1) could belong to the Monian Terrane, with the southern boundary of the Leinster-Lakesman Terrane running instead up the Dent Line. Pharaoh *et al.* (2020) used this line and continued the boundary eastwards along the Stainmore faults and offshore to truncate the southwestern boundary of the Southern North Sea Terrane (Fig. 1). No Lower Palaeozoic rocks are exposed to constrain these more easterly boundaries.

The *Monian Terrane,* alternatively called the Rosslare-Monian-Ingleton Terrane (Pharaoh 2018; Pharaoh *et al.* 2020), is characterised by a late Neoproterozoic basement that is not known in the Leinster-Lakesman Terrane. An overstepping succession, commencing in the Floian, follows a Lower Ordovician unconformity ascribed to the Monian-Penobscottian Orogeny (Schofield *et al.* 2016, 2020). In both Anglesey and SE Ireland, predominantly crystalline rocks to the SE are faulted against a weakly metamorphosed Cambrian clastic sequence to the NW: the Monian Supergroup in Anglesey and the Cahore Group and Cullenstown Formation in SE Ireland. The southeastern boundary to the Monian Terrane is the Menai Strait Fault System (Gibbons 1987), a pre-Floian sinistral strike-slip zone (Gibbons 1987). Followed southwestwards, this terrane boundary is represented by the Llŷn Shear Zone on mainland Wales but must lie offshore of SE Ireland (Fig. 1). To the NE, the Menai

Strait zone is assumed to run offshore parallel to the North Wales coast, then to underlie the Pendle line and the southern edge of the Flamborough Head Fault Zone through northern England (Pharaoh 2018; Pharaoh *et al.* 2020). Pharaoh (2018; Pharaoh *et al*. 2020) showed a northeastern termination to the Monian Terrane along the South Craven Fault (Fig. 1). However, this boundary would isolate to the east the Ingleton Group, which shows the supposedly correlative Monian-Penobscottian deformation event, and it is more likely that the Monian Terrane continues into eastern England (Fig. 1).

Attempts to correlate the peri-Gondwanan terranes of Europe with terranes in the northern Appalachians were initially based mainly on lithological comparisons (Williams 1978; van Staal *et al.* 1998) and faunal evidence (e.g. Harper *et al.* 1996). The correlations have been strengthened in the past decade by acquisition of detrital zircon U-Pb age spectra and Nd- and O-isotopic data from Lower Palaeozoic rocks of Britain and Ireland (Waldron *et al.* 2014; Schofield *et al.* 2016; Fritschle *et al*. 2018). These studies have confirmed the affinity with the Appalachian Gander Terrane of both the Leinster-Lakesman Terrane and the Monian Composite Terrane. The alternative view of Landing *et al*. (2022) is that these two terranes are off-platform successions that are laterally continuous with the marginal and inner platform areas of the Avalon Composite Terrane.

The *Avalon Composite Terrane* lies SE of the Monian Terrane and SW of the offshore Dowsing Fault–South Hewett fault zones that separate it from the poorly known *Southern North Sea Terrane* (Fig. 1). The Avalon Terrane was explicitly named by correlation with the Avalon Peninsula of Newfoundland and contiguous rocks of maritime eastern North America (Bluck *et al.* 1992). Four component Neoproterozoic sub-terranes have been recognised by Pharaoh and Carney (2000): *Cymru, Wrekin, Charnwood* and *Fenland* (Fig. 1). It is possible that all these terranes had amalgamated by the late Neoproterozoic (e.g. Landing *et al*. 2022). However, the evidence on the best exposed of the three intervening boundaries – the Welsh Borderland Fault Zone – suggests stratigraphical linkage across the zone only after the later Katian (mid-Ashgill). Terrane-scale displacements, probably dextral strike-slip, in or before the mid-Katian cannot be ruled out (Woodcock and Gibbons 1988). The Cymru Terrane is the type area for the Ordovician System (Lapworth 1879) and exposes abundant rocks of that age. Outcrop becomes more sporadic in the Wrekin Terrane, and Ordovician rocks are known only from the subsurface in the Charnwood and Fenland terranes.

The correlation of the UK Avalon Terrane with the Appalachian Avalonian terrane has increasingly been questioned in the past decade. Waldron *et al.* (2011) matched lithologically the Cambrian successions in the Harlech Dome of North Wales with those of the Meguma Terrane in Nova Scotia. Detrital zircon U-Pb age spectra from this and later studies (Waldron *et al.* 2014; Pothier *et al.* 2015) support this correlation as do age and Nd- and O-isotopic data (Schofield *et al.* 2016; Waldron *et al.* 2019). The implication of the North Wales results is that the Cymru Terrane may have been part of Megumia (Waldron *et al.* 2011) with large necessary displacements relative to the remaining components of the Avalon Terrane. However, a Megumian zircon age spectrum from the Cambrian of the Wrekin Terrane (Waldron *et al.* 2019) now questions the affinity, and of course the name, of the whole of the Avalon Terrane. Landing *et al*. (2022) are more sceptical about using detrital zircon age spectra for defining terranes. They simply regard the coherent Avalon Terrane as the northern platform to Avalonia, with the Meguma Terrane as the off-platform succession on its southern margin.

The southern edge of the Avalon Terrane is overprinted by the late Carboniferous phase of Variscan deformation. The *Variscan Front* (Fig. 1) does not have terrane-scale displacements, the true southern boundary of the Avalon Terrane being the Bristol Channel Fault Zone. This late Variscan dextral strike-slip fault has been postulated to link with the Bray Fault in northern France and host around 400 km of dextral strike-slip displacement (Holder and Leveridge 1986b; Woodcock *et al.* 2007).

The *Cornubian Terrane* is the northernmost peri-Gondwanan element that was transported northwestwards along the Bristol Channel-Bray Fault in the latest Carboniferous from a position opposite Belgium/SW Germany. The terrane lacks exposed rocks older than Devonian; Ordovician rocks only occur as large redeposited slabs in a mid-late Devonian mélange along the Lizard-Dodman Thrust (Fig. 1; Barnes 1983), formerly taken as the southern boundary of the Cornubian Terrane (Woodcock 2011). However a strong case has been made by Dijkstra and Hatch (2018) for a terrane boundary within the Cornubian terrane, based on contrasts in the isotope geochemistry of mantle-sourced lamprophyre dykes (Fig. 1). The two halves of the Cornubian Terrane are provisionally labelled on Figure 1 as the *North* and *South Cornubian terranes*. Dijkstra and Hatch (2018) regard the South Cornubian Terrane as Armorican, Alexander *et al.* (2019) considered it a terrane distinct from both Normannia and North Armorica but of uncertain allegiance, and Nance *et al.* (2015) correlated the isotope geochemistry of its Cornubian Batholith with the Appalachian Meguma Terrane.

South of the Lizard-Dodman Thrust, the *Normannian Terrane* comprises poorly known orthogneisses unexposed below the English Channel (Holder and Leveridge 1986a), but assumed to be contiguous with components of the North Armorican terrane north of the Paimpol Fault (Fig. 1). The Normannian Terrane and the adjacent *North Armorican Terrane* (Strachan *et al.* 1996) are regarded as part of the peri-Gondwanan Cadomian continent (Linnemann *et al.* 2008).

Cymru Terrane (Welsh Basin)

Ordovician outcrop

The Cymru Terrane equates to the Welsh Basin, which lies between the Menai Strait Fault System in the NW and the Welsh Borderland Fault System to the SE and south. The latter includes the Pontesford Lineament, focussed on the Pontesford-Linley Fault which delineates the eastern edge of the Shelve Ordovician outcrop, and the Tywi Lineament, a complex fold and fault structure with a curvilinear trace that runs north-eastwards from SW Wales to eastern mid Wales; in current interpretations, it is thought to extend north-eastwards into the Severn Valley Faults (e.g., Cherns *et al.* 2006, fig. 4.5). In SW Wales, the E–W trending Fishguard–Cardigan Fault Zone (Davies *et al.* 2003) exerted an influence on sedimentation and volcanicity throughout much of the Ordovician. The Shelve area was probably a sub-basin along the eastern margin of the Welsh Basin (Cave and Hains 2001), like the Builth Ordovician inlier, and is included here.

Ordovician rocks crop out around the Silurian core of the Welsh Basin, from SW Wales north-eastwards along the Tywi Lineament to the Shelve area, and thence north-westwards and westwards across North Wales (Fig. 2). Rocks of all five Anglo-Welsh Ordovician series are represented, and it is convenient to discuss them under those headings. Biostratigraphical zonations based on graptolites (Rushton 1990a), conodonts (Bergström *et al.* 1987; Ferretti *et al.* 2014a;

Ferretti and Bergström 2022) and chitinozoans (Vandenbroucke 2008; Vandenbroucke *et al*. 2008, 2009, Challands *et al.* 2014; Amberg *et al*. in press) have all been applied (Figs 3–5), and trilobites, other shelly fauna and acritarchs have been used in dating and biostratigraphical correlation.

Tremadoc Series (Tremadocian Stage)

Tremadocian rocks in North Wales comprise grey, locally bioturbated mudstones with subordinate sandstones (Dol-cyn-afon Formation, Figs 3, 4) that are conformable on black upper Cambrian mudstones and represent a continuation of late Cambrian mud deposition (Rushton and Howells 1999). A U–Pb zircon age (489 ± 0.6 Ma) has been obtained from sandstones close to the Cambrian–Ordovician boundary east of the Harlech Dome (Landing *et al.* 2000). The *flabelliformis* and *tenellus* graptolite biozones and the *salopiensis* (formerly *pusilla*) and *sedgwickii* trilobite biozones have all been recorded, but latest Tremadocian rocks are generally not present (Pratt *et al.* 1995; Howells and Smith 1997; Rushton and Fortey 2000; Young *et al.* 2002). East of the Harlech Dome, the Dol-cyn-afon Formation is overlain on an irregular surface by the calc-alkaline, arc-related, late Tremadocian Rhobell Volcanic Group (Bevins *et al*. 1984; Kokelaar 1986; Bevins *in* Stephenson *et al*. 1999; Fig. 3), which was folded prior to the Arenig marine transgression (Allen and Jackson 1985).

A more complete succession is present in the Shelve area (Fig. 3), where grey mudstones and siltstones (Shineton Shale Formation) are overlain by siltstones, mudstones and shales with sandstones (Habberley Formation). Trilobites indicate that the Habberley Formation extends into the latest Tremadocian (Fortey and Owens 1992). The succession is interpreted as shallowing upwards (Fortey and Owens 1992; Cave and Hains 2001).

Tremadocian rocks are poorly represented in South Wales but include grey-green mudstone of the *tenellus* Biozone SW of Carmarthen (Owens *et al.* 1982; Prigmore *et al.* 2000) and uppermost Tremadocian or basal Floian sandstones with interbedded shales east of Carmarthen (Molyneux and Dorning 1989; Molyneux *et al.* 2007; Fig. 5). Farther west, andesitic lavas and tuffs (Treffgarne Volcanic Formation; Fig. 4) above upper Cambrian rocks (Fortey and Rushton 2000a) are inferred to be Tremadocian based on petrographical and geochemical similarity to the Rhobell Volcanic Group (Bevins *et al.* 1984).

Arenig Series (Floian–lower Darriwilian stages)

Across Wales, basal Arenig conglomerates, sandstones, siltstones and mudstones record a transgression and subsidence associated with the development of a rapidly deepening marine basin. Lithostratigraphical units include the Allt Lŵyd Formation across much of North Wales, the Wig Bâch and St Tudwal's formations on the Llŷn Peninsula, the Ogof Hên Formation or undivided Triffleton Group in South Wales, and the Stiperstones Quartzite Formation at Shelve (Figs 3–5). The shallow marine *Neseuretus* trilobite assemblage has been widely reported (Beckly 1987). Basal Arenig formations are generally unconformable on Tremadocian or older rocks, although the hiatus below the Stiperstones Quartzite Formation in the Shelve area was probably brief and the disconformity is slight (Cave 2008). Only in the more basinal setting of west Wales, south of the Harlech Dome, has the contact been regarded as transitional and conformable (Pratt *et al.* 1995).

The age of the basal Arenig rocks in North Wales is poorly constrained but the transgression is likely to have started during the early to mid Floian (Fl1–Fl2). Trilobites from the Allt Lŵyd Formation (Henllan Ash Member) in the Arenig area east of the Harlech Dome indicate a Moridunian (early

Floian) age (Fortey and Owens 1987; Fig. 3). Graptolite control is poor but graptolites from the type Arenig area may indicate the *simulans* and perhaps higher biozones, i.e., a latest Moridunian to Whitlandian (Floian–early Dapingian) age (Zalasiewicz 1984, 1986; cf. Zalasiewicz *et al.* 2009). Trilobites and graptolites indicate that all three Arenig stages (Moridunian, Whitlandian and Fennian; Floian–lower Darriwilian) are represented in the Wig Bâch Formation on the Llŷn Peninsula (Beckly 1987; Gibbons and McCarroll 1993; Owens *et al.* 2000; Young *et al.* 2002), which is thought to have accumulated in a narrow basin parallel to the Menai Strait Fault System (Young *et al.* 2002). Marine conditions subsequently extended across the margins of the basin with Fennian sandstones and mudstones (Bryncroes Formation) unconformable on Ediacaran rocks to the NW and coarse- to medium-grained muddy sandstone of Fennian or possibly slightly older age (St Tudwal's Formation) on upper Cambrian rocks to the SE (Beckly 1987; Young *et al.* 2002). Within the Menai Strait Fault System between Caernarfon and Bangor, the base of the Arenig succession was considered by Beckly (1987) to be Whitlandian. The underlying Neoproterozoic–Cambrian succession, however, was folded into tight upright folds, with a pervasive slaty cleavage in places, prior to overstep of the Arenig succession. The timing of this deformation is indicated by resetting of the Rb–Sr system in the Neoproterozoic Twt Hill Granite at *c.* 470 Ma (Dapingian, Fennian) (Schofield *et al.* 2008, 2020).

Debris flows, slump deposits and turbiditic sandstones (Porth Meudwy Formation) of latest Fennian (Beckly 1988) or possibly Llanvirn (Gibbons and McCarroll 1993) age overlie the Wig Bâch Formation on the Llŷn Peninsula, and mudstones (Llanengan Formation; Rushton and Howells 1999; Young *et al.* 2002) of Fennian age overlie the St Tudwal's Formation. Elsewhere, graptolite-bearing siltstones of late Arenig (Fennian) age (Olchfa Shales *sensu* Rushton and Fortey 2000, fig. 11, col. 15; Fig. 3, column 3) are unconformable on the Allt Lŵyd Formation in the Arenig district (Beckly 1987; Zalasiewicz 1992).

Arenig rocks above the basal coarse clastic units in South Wales record deepening along the southern margin of the Welsh Basin. Fortey and Owens (1987) established the Moridunian, Whitlandian and Fennian stages (Floian–lower Darriwilian) and a local trilobite biozonation in mudstones and shales with turbidites in the Whitland and Carmarthen areas (Fig. 5). The Fennian (Dapingian–lower Darriwilian) part of the succession has yielded a large and important fauna of benthic animals, including sponges, brachiopods and carpoids, along with trilobites and graptolites. The trilobites from the Fennian Pontyfenni Formation (anoxic, black and dark grey mudstone) indicate a deep marine setting (Fortey and Owens 1987) and graptolites place the base of the *artus* Biozone (Arenig–Llanvirn and Darriwilian Dw1–Dw2 boundaries) in the overlying oxic grey mudstones (Llanfallteg Formation: Fortey and Owens 1978, 1987; Owens 2000). Cocks and Popov (2019) reported on brachiopods from the succession; acritarchs and chitinozoans have been recorded (Molyneux 1987; Amberg *et al*. in press).

Farther west, anoxic mudstones of inferred Whitlandian–Fennian age (Penmaen Dewi Shales Formation) overlie the basal Arenig rocks (Burt *et al.* 2012; Fig. 4). Overlying beds comprise a mixture of trilobite-bearing anoxic mudstone (Pontyfenni Formation), oxic mudstone (Foel Tyrch Formation) or silty mudstone (Ysguborwen Formation; Burt *et al.* 2012; Fig. 4). The distribution of the oxic units is delineated by faults to the south of the Fishguard–Cardigan Fault Zone, notably the Bronnant, Llanglydwen and Penfordd faults (see Burt *et al*. 2012, fig. 1), suggesting contemporaneous control. One or more of the same faults might have enabled emplacement of the Sealyham Volcanic Formation, which overlies the Penmaen Dewi Shales Formation locally (Fig. 4) with a U-Pb age of

471.7 ± 1.5 Ma (Evans and Boulton in Burt *et al.* 2012). Its calc-alkaline geochemistry suggests an arc-related origin (Bevins *et al.* 1992; Burt *et al*. 2012).

In the Shelve area, bioturbated siltstones with sandstones (Mytton Flags Formation) above the Stiperstones Quartzite Formation have yielded trilobites and graptolites that establish correlation with the regional Moridunian, Whitlandian and Fennian stages (Fig. 3). The shallow-water *Neseuretus* trilobite biofacies occurs throughout (Whittard 1979; Owen *et al.* 2000). The uniform litho- and biofacies through a thickness of 900 m suggests a balance between subsidence and sedimentation in a shallow marine setting.

Llanvirn-Caradoc series (mid Darriwilian, Dw2, to mid Katian, Ka2, stages)

The establishment of major volcanic centres in a back-arc basin, the effects of volcano-tectonic faulting, eustatic changes in sea level, and lateral and vertical changes from deep anoxic facies to shallower oxic facies, have all contributed to the complexity of Llanvirn–Caradoc stratigraphy in Wales.

Mid Ordovician unconformity

There is widespread evidence across North Wales for a mid-Ordovician unconformity that cuts out upper Llanvirn rocks (Zalasiewicz 1992; Howells and Smith 1997; Young *et al.* 2002; British Geological Survey 1997, 1999), although a continuous succession through the Nant Ffrancon Subgroup may be preserved locally (Smith *et al*. 1995; Fig. 3, column 2). Caradoc (Sandbian) rocks in north Snowdonia are unconformable on Arenig strata (Smith *et al.* 1995; Fig. 3), and Llanvirn mudstones of probable *artus* Biozone age (Darriwilian, Dw2) on the Llŷn Peninsula are overlain by rocks of probable late Burrellian (late Sandbian, Sa2) age (Rushton and Howells 1999, fig. 4; Young *et al.* 2002). The upper part of the Aran Volcanic Group (see below) is also considered to overlie the lower part of the group above a non-sequence or unconformity and oversteps westwards to overlie the Allt Lŵyd Formation (Zalasiewicz 1992; British Geological Survey 1995; Pratt *et al.* 1995, table 1).

Volcanic successions

Volcanism was widespread with major volcanic successions comprising the Aran, Builth, Fishguard, Llewelyn, Snowdon, Upper Lodge and Llanbedrog volcanic groups (Figs 3–5). Minor occurrences of volcanic and volcaniclastic rocks occur throughout. Stephenson *et al*. (1999) provide details of volcanic rocks at outcrop across Wales.

The bimodal Aran Volcanic Group (acid and basic tuffs, basaltic and andesitic lavas, volcaniclastic deposits and intercalated mudstones; Fig. 4, column 6) crops out mainly to the south and east of the Harlech Dome, thinning to a feather edge towards the north. Its base marks the onset of back-arc basin volcanism in Wales (Rushton and Howells 1999). Graptolites establish the *artus* Biozone (Darriwilian, Dw2) and the *gracilis* and/or *foliaceus* biozones (Sandbian), but an unconformity or non-sequence in the middle of the group (the Mid Ordovician unconformity) equates to the *murchisoni* and *teretiusculus* biozones (upper Darriwilian, Dw2–Dw3) (Zalasiewicz 1992; Pratt *et al.* 1995). Mudstones with tuffs, volcanic breccias and conglomerates in the Arenig district (Serw Formation; Fig. 3, column 3) are equivalent to units in the Aran Volcanic Group, although the use of 'Serw Formation' as a lithostratigraphical unit is no longer appropriate (Rushton and Howells 1999) as it is evident that the succession includes an unconformity or non-sequence (Zalasiewicz 1992).

The Fishguard Volcanic Group of SW Wales (bimodal with basalt, dacite and rhyolite lavas and acid tuffs; Bevins *et al.* 1992; Phillips *et al.* 2016 and references therein) is placed in the *murchisoni* Biozone (Williams *et al*. 2003; Burt *et al*. 2012; Fig. 4). Most of the Fishguard Volcanic Group rocks were extruded subaqueously, with restricted instances of subaerial volcanism (Bevins *et al.* 1992; Bevins *in* Stephenson et al. 1999; Phillips *et al.* 2016 and references therein). Abrupt thickness changes northwards across the Fishguard-Cardigan Fault Zone are consistent with active faulting, which probably provided conduits for magma extrusion (Davies *et al.* 2003; Burt *et al.* 2012). The abrupt thickening suggests that the Fishguard Volcanic Group accumulated within an active fault-bounded trough or graben (Kokelaar 1988), the relationships of the group and the volume of erupted material being in keeping with the rapid infilling of a collapsed caldera sited along and at least partly defined by the fault zone (Davies *et al*. 2003; Burt *et al*. 2012).

The Builth Volcanic Group (basalt, andesite and dacite lavas, acid and basic tuffs, intercalated mudstone and sandstone) in the Builth Wells Inlier is also placed in the *murchisoni* Biozone (Schofield *et al.* 2004; Fig. 5). There is evidence for subaerial weathering and erosion and lateral facies and thickness variations also suggest the influence of syndepositional faulting, although sea floor topography created by lava domes might have had an effect (Schofield *et al.* 2004). Shoreline erosion of an emergent basaltic and andesitic lava pile in the upper part of the Builth volcanic succession created an irregular unconformity, now overlain by tuffaceous sandstone and boulder conglomerate (Newmead Sandstone Formation). A higher sequence of volcanic breccias and agglomerates (Trelowgoed Volcanic Formation) overlies Llanfawr Mudstones of *gracilis* Biozone age (Fig. 5, column 14).

Later Llanvirn volcanism towards the northern end of the Tywi Lineament is indicated by subaqueous acid ash-flow tuff interbedded with dark grey mudstone, tuffaceous siltstone, sandstone, conglomerate, debrite, basalts and thin ooidal ironstones (Llanwrtyd Volcanic Formation; Schofield *et al.* 2004). The formation is placed in the *teretiusculus* and *gracilis* biozones (Cave and Rushton 1996), thus straddling the Llanvirn–Caradoc (Darriwilian–Sandbian Stage) boundary (Fig. 5, column 13). Its base is not seen.

The Llewelyn and Snowdon volcanic groups represent respectively the first and second of two major Caradoc (Sandbian to early Katian) eruptive cycles in North Wales. Brachiopod-dominated faunas from intercalated sedimentary units suggest a Soudleyan (mid Burrellian, Sandbian, Sa2) age for the Llewelyn Volcanic Group (rhyolite, trachyandesite and basaltic andesite lavas and acid tuffs; Howells *et al.* 1991; Bevins *in* Stephenson *et al*. 1999). The sedimentary units intercalated with the Llewelyn Volcanic Group are variable in thickness and composition, resulting from the interplay of alluvial, fluvial and shallow-marine facies. Shallow-marine sedimentation was initially predominant between the early and latest phases of Llewelyn volcanism, with the later development of subaerial alluvial fan sequences in the north passing southwards into braidplain fan deltas, shallow-marine tidal flats and marine shelf deposits (Howells *et al.* 1991).

The base of the Snowdon Volcanic Group (acidic tuffs and tuffites, basaltic tuffs, volcanic breccias, rhyolites, basalts, marine sedimentary rocks) is at about the Soudleyan-Longvillian boundary (Sandbian, Sa2) (Howells *et al.* 1991; Howells and Smith 1997; Bevins *in* Stephenson *et al*. 1999). Brachiopods and trilobites from higher units indicate late Longvillian and/or Woolstonian ages (latest Burrellian–early Cheneyan, late Sandbian, Sa2–early Katian, Ka1; Howells *et al.* 1991; Howells and

Smith 1997; Rushton and Howells 1999). There is evidence for subaerial as well as submarine emplacement of the Snowdon volcanic rocks (Howells and Smith 1997).

Accumulation of the Upper Lodge and Llanbedrog volcanic groups on the Llŷn Peninsula was contemporaneous with that of the Snowdon Volcanic Group (Rushton and Howells 1999). Stratigraphical relationships of the Upper Lodge Volcanic Group (basaltic trachyandesite with rhyolitic tuffs) indicate a late Burrellian (Sandbian) age (Rushton and Howells 1999; Bevins *in* Stephenson *et al*. 1999; Young *et al.* 2002). The later Llanbedrog Volcanic Group (basaltic trachyandesite, trachydacite and rhyodacitic lavas, rhyolitic tuffs, volcaniclastic rocks) is likely to be of Woolstonian (early Cheneyan, early Katian, Ka1) age (Bevins *in* Stephenson *et al*. 1999; Young *et al.* 2002).

North Wales sedimentary units

Sedimentary rocks from the base of the Arenig (Floian) Allt Lŵyd Formation to the top of the upper Caradoc (mid Katian) Nod Glas Formation in North Wales are placed in the Ogwen Group (Rushton and Howells 1999). Graptolites of the *artus* and *murchisoni* biozones (Darriwilian) have been recorded from a predominantly mudstone-siltstone unit with sandstones, oolitic ironstones and some volcanic rocks north of the Harlech Dome (Nant Ffrancon Subgroup: Howells and Smith 1997; Rushton and Howells 1999; Fig. 3, columns 1, 2); there is no unequivocal evidence for rocks of late Llanvirn (Llandeilian, Darriwilian Dw3) although they may be preserved locally (Smith *et al*. 1995). A mélange in the Nant Ffrancon Subgroup above the mid-Ordovician unconformity in Snowdonia (Fig. 3, column 2) is attributed to gravity-induced slumping during renewed tectonic activity (Smith 1987; Howells and Smith 1997). Graptolites, shelly faunas, chitinozoans and acritarchs indicate the *gracilis* and/or *foliaceus* graptolite biozones and the upper Aurelucian–lower Burrellian stages (Sandbian, Sa1–Sa2; Howells and Smith 1997) for the upper part of the subgroup. Interbedded with the Nant Ffrancon sedimentary rocks are volcaniclastic debris-flow deposits, acid tuffs and tuffites (Moelwyn Volcanic Formation).

Westwards, on the Llŷn Peninsula, Nant Ffrancon Subgroup rocks of Llanvirn (Darriwilian) age are overlain unconformably by the Snowdon and Upper Lodge volcanic groups (Gibbons and McCarroll 1993; Howells and Smith 1997; Young *et al.* 2002). The absence of Nant Ffrancon Subgroup rocks of Caradoc (Sandbian) age in these areas is possibly due to volcano-tectonic uplift.

Shallow marine siliciclastic rocks (Cwm Eigiau Formation) are interposed between the Llewelyn and Snowdon volcanic groups or overlie the Nant Ffrancon Subgroup conformably where the Llewelyn Volcanic Group is absent (Fig. 3, columns 1, 2). They record an influx of coarse detritus with marked lateral variations in lithology (Howells and Smith 1997). Restricted alluvial fans indicate local emergence while contemporaneous slumping and mass transport probably reflect ongoing volcano-tectonic activity. Shelly faunas establish a mid–late Burrellian (Sandbian, Sa2) age (Howells and Smith 1997; Young *et al.* 2002; Fig. 3). On the Llŷn Peninsula, dark siltstones with intercalated tuffs and sandstones (Dwyfach Formation) are interposed between the Upper Lodge and Llanbedrog Volcanic groups or are laterally equivalent to the latter (Fig. 3, column 1). They were possibly deposited during fault-controlled subsidence following emplacement of the Upper Lodge Group. Shelly faunas indicate a Longvillian–Woolstonian (late Burrellian–early Cheneyan, Sa2–Ka1) age (Young *et al.* 2002).

South of the Harlech Dome, grey mudstones of *gracilis* and/or *foliaceus* biozone age (Ty'r Gawen Mudstone Formation) are interposed within the Aran Volcanic Group (Fig. 4, column 6). The Ty'r Gawen Mudstone Formation, with an oolitic ironstone (Fron Newydd Member) developed widely at its base, overlies the lower part of the Aran Volcanic Group above a non-sequence or unconformity, and oversteps westwards to overlie the Allt Lŵyd Formation (British Geological Survey 1995; Pratt et al. 1995). Upwards, the Ty'r Gawen Mudstone Formation interdigitates with formations of the Aran Volcanic Group (Pratt *et al.* 1995) and is overlain by either the highest Aran Volcanic Group unit or by turbiditic silty mudstones and laminated hemipelagic mudstones (Ceiswyn Formation: British Geological Survey 1995; Rushton and Howells 1999) with graptolites and trilobites that indicate the *foliaceus* graptolite Biozone and middle Burrellian Stage (Sandbian, Sa2; Fig. 4, column 6). An upward increase in the proportion of hemipelagic mudstones probably reflects gradual drowning during a period of eustatic sea-level rise.

The uppermost hemipelagite-rich parts of the Ceiswyn Formation are equivalent to successions deposited in shallower settings farther east. These include mudstones and siltstones in the Bala area (Nant Hir Mudstone, Glyn Gower Siltstones, Allt Ddu Mudstones) with calcareous mudstone (Derfel Limestone), acid tuffs (Cefn Gwyn Tuff, Frondderw Tuff), and shelly and graptolite faunas that indicate a late Aurelucian–mid Burrellian (late Sa1–mid Sa2, *foliaceus* graptolite Biozone) age (Bassett *et al.* 1966; British Geological Survey 1986; Zalasiewicz 1992; Rushton and Howells 1999; Fig. 3, column 3). Overlying calcareous volcaniclastic sandstones, tuffs and tuffites (Gelli-grin Calcareous Ash Formation) contain brachiopods and trilobites that establish a late Burrellian–early Cheneyan (Sa2–Ka1) age (Bassett *et al.* 1966; Rushton and Howells 1999).

Grey mudstones with thin siltstones and fine sandstones, intercalated acid tuffs and middle to upper Burrellian (Sa2) shelly faunas form the main part of the succession farther east in the north and west Berwyn Hills (Brenchley 1978). These have been interpreted as the products of a mainly shallow sub-tidal marine environment with periodic emergence and deepening in the upper part of the succession (Brenchley 1978).

Siltstones and sandstones to the SE, but west of the Severn Valley Faults (Gaer Fawr Formation, Sa2–Ka1), also reflect a shallow shelf environment (Brenchley 1978; Cave 2008), and overlie mudstones, silty mudstones and fine sandstones of Burrellian (Sa2) age (Stone House Shale Formation with trilobites and graptolites of the *foliaceus* Biozone, and the Pwll-y-glo Formation: Cave 2008; Fig. 3, column 4).

East of the Severn Valley, sandstones interbedded with grey mudstone at the base of the exposed succession ('Black Grit', Stone House Shale Formation) are possibly of *gracilis* Biozone age (Cave and Hains 2001); the graptolitic mudstones of the rest of the formation belong to the *foliaceus* Biozone (Cave 2008). Rhyolitic tuffs and breccias (Middletown Quarry Member) signal a localised, explosive volcanic centre. Overlying lithologies include coarse conglomerate with rounded clasts of andesite (Bulthy Formation) and grey, bioturbated, silty mudstone with sandstones and Burrellian (Sa2) trilobites (Hill Farm Formation). Laterally equivalent shaly mudstones with thin sandstones and andesitic conglomerates (Forden Mudstone Formation) also contain Burrellian trilobites (Cave 2008; Fig. 3, column 4).

Shelve

Medium to dark grey mudstone with siltstone laminae (Hope Shale Formation; Fig. 3, column 5) and graptolites and trilobites of *artus* Biozone age constitute the lower Llanvirn (Darriwilian, Dw2) succession in the Shelve area (Cave and Hains 2001). Most of the trilobite genera are blind, reflecting a deep-marine environment that was also dysaerobic with anoxic mud deposition (Cave and Hains 2001). Two laterally impersistent volcanic units (Hyssington and Stapeley volcanic members) are present, comprising acid tuffs, tuffites and volcaniclastic rocks (Hyssington), and andesitic and basaltic tuffs and lavas, rhyolitic tuffs and volcaniclastic sandstones (Stapeley) (Cave and Hains 2001; Cave 2008).

The mudstones are overlain by sandstones with locally burrowed siltstone and mudstone (Weston Flags Formation; Fig. 3). A low-diversity shelly fauna resembles the *Neseuretus* community (Cave 2008) and is thought to indicate deposition during early *murchisoni* Biozone times in a high energy, shallow-marine setting (Cave and Hains 2001). Silty and micaceous mudstones (Betton Shale Formation; Fig. 3) with a fauna of graptolites, brachiopods and trilobites pass up into a succession with impersistent beds of shelly, commonly burrowed calcareous sandstone, siltstone and limestone (Meadowtown Formation; Fig. 3) and then into laminated hemipelagite (Rorrington Shale Formation; Fig. 3). The varied fauna of the Meadowtown Formation includes graptolites (Strachan 1986; Hughes 1989) of the *teretiusculus* Biozone. Graptolites from the Rorrington Shale Formation place its lower part in the *teretiusculus* Biozone and the upper part in the *gracilis* Biozone, so spanning the Llanvirn–Caradoc (Darriwilian–Sandbian) boundary. The Rorrington Shale Formation also contains relatively abundant orthocones, which together with the graptolites reflect pelagic influence, along with trilobites, ostracods and brachiopods (Cave and Hains 2001). Vandenbroucke *et al.* (2009) established a chitinozoan biozonation from the top of the Betton Shale into the Rorrington Shale Formation. The *Eisenackitina rhenana* chitinozoan Biozone (the *rhenana* Subzone of the *Laufeldochitina stentor* Biozone in Nõlvak and Grahn 1993) is only tentatively identified in Anglo-Welsh successions ('?' on Fig. 3) and spans the Llanvirn–Caradoc series boundary in the Rorrington Shale Formation (Vandenbroucke 2008; Vandenbroucke *et al*. 2009).

Overlying calcareous sandstones (Spy Wood Sandstone Formation; Fig. 3) stand out from the underlying and overlying mudstone formations and are the likely accumulations of bioclastic sand debris transported from the basin margin following rapid inundation of an adjacent shelf area (Cave and Hains 2001; Cave 2008). Graptolites, trilobites and ostracods indicate the *gracilis* and *foliaceus* biozones and a late Aurelucian (Sandbian, Sa1) age (Cave and Hains 2001). Brachiopods, trilobites and graptolites from overlying mudstones (Aldress Shale Formation; Fig. 3) indicate an early–mid Burrellian (Sa2) age and the *foliaceus* Biozone (Williams 1974; Strachan 1986; Hughes 1989).

Two volcanic units (the lower Hagley Volcanic Formation and upper Whittery Volcanic Formation) are present in the Caradoc succession of the Shelve area (Fig. 3, column 5) and comprise similar sequences of mainly volcaniclastic rocks with rhyolitic tuffs and probable minor lava flows (rhyolites and andesites) (Cave and Hains 2001; Cave 2008). Between the two volcanic formations are grey-green mudstones with thinly bedded sandstones (Hagley Shale Formation). Similar olive and grey shales with laminated sandstones and tuffites (Whittery Shale Formation) occur at the top of the succession. Trilobites and graptolites indicate a mid Burrellian (Sa2) and *foliaceus* Biozone age (Hughes 1989; Cave and Hains 2001; Cave 2008). The relationship between the Burrellian rocks in the Shelve area and those to the west is obscured beneath Silurian strata.

South Wales anoxic facies

Llanvirn–Caradoc (Darriwilian–Sandbian) sedimentary units across South Wales can be assigned to anoxic, mixed or oxic facies, with the anoxic units mainly in the more basinal settings. Anoxic dark grey graptolitic mudstones with siltstone laminae (Aber Mawr Formation), deposited in an outer shelf or basinal setting, are confined to the northern part of the outcrop in SW Wales, where they overlie the Penmaen Dewi Shales Formation or Sealyham Volcanic Formation (Fig. 4). Thickness variations suggest contemporaneous movements on the Fishguard-Cardigan Fault Zone (Burt *et al.* 2012). Tuffaceous mudstones in the upper part of the formation may signal the onset of volcanism that culminated in the Fishguard Volcanic Group. The Aber Mawr Formation crops out at the type locality of the Anglo-Welsh Llanvirn Series and has accordingly yielded graptolites of the *artus* Biozone (Darriwilian, Dw2) and faunas assigned to the *murchisoni* Biozone (Dw2–Dw3) (Kennedy 1989; Owens 2000; Burt *et al.* 2012).

Younger anoxic facies are placed in the Drefach Group, which comprises mudstones, limestones and tuffs of Llanvirn–Caradoc age (Darriwilian–Sandbian) on the SW margin of the Welsh Basin. Black, pyritic and graptolitic mudstones predominate in undivided Drefach Group sequences of SW Wales that overlie rocks of the Abergwilli Formation (see below), Aber Mawr Formation and unconformably the Fishguard Volcanic Group northwards across the Fishguard–Cardigan Fault Zone (British Geological Survey 2010; Burt *et al.* 2012; Fig. 4). The lithofacies indicates accumulation in an anoxic distal shelf setting and faunas of the *murchisoni*, *teretiusculus*, *gracilis*, *foliaceus* and *clingani* biozones have been recorded (Williams *et al.* 2003; Burt *et al.* 2012).

Farther south and east, the anoxic mudstones that comprise the upper part of the Drefach Group (Mydrim Shales Formation) have yielded faunas of the upper *gracilis* or *foliaceus* (Fortey 2006) and *clingani* graptolite biozones (Zalasiewicz *et al.* 1995). They might extend into the *linearis* Biozone, in the *Normalograptus* proliferation interval of Zalasiewicz *et al.* (1995; Fig. 5, column 10), although Vandenbroucke *et al.* (2008) suggested that this interval correlated instead with the upper *clingani* Biozone. The lower Drefach Group there comprises different facies of turbiditic and hemipelagic mudstone with calcareous sandstones and limestones (see below).

South Wales mixed facies

The Abergwilli Formation is the southern equivalent of the Aber Mawr Formation between the Fishguard area and the Llandeilo–Llandovery district (Figs 4, 5). Oxic and anoxic bottom conditions alternated during its deposition so that it comprises interbedded pale, burrow-mottled oxic mudstones, locally silty, and dark grey anoxic hemipelagic mudstone with abundant *Didymograptus artus*; tuffs and sandstones occur sporadically (Wilby *et al.* 2007; Schofield *et al.* 2009; Burt *et al.* 2012).

Basal units of the Drefach Group above the Abergwilli Formation (Fig. 5, column 10) comprise dark grey, graptolitic, hemipelagic and turbiditic mudstones of *murchisoni* Biozone age (Felin-wen Formation), locally with bioturbated, calcareous sandstone and limestone (Wilby *et al.* 2007). The overlying Asaphus Ash Formation (tuff, tuffaceous sandstone and grey mudstone) can be distinguished across much of SW Wales, and is locally overlain by bioturbated mudstones, calcareous sandstones and thin limestones (Lan Flags: Fortey and Rushton 2000a) with a graptolite and shelly fauna indicative of the *teretiusculus* Biozone (Wilby *et al.* 2007; Burt *et al.* 2012).

The overlying Hendre Shales Formation of the Drefach Group (calcareous turbiditic and hemipelagic mudstone with sporadic thin limestones, calcareous sandstones and ashy siltstones) contains a graptolite-trilobite fauna that spans the *teretiusculus*–*gracilis* Biozone boundary (Owens 2000; Wilby *et al.* 2007; Burt *et al.* 2012; Fig. 4). Flaggy, thin-bedded calcareous sandstones and limestones (Mydrim Limestone Formation), deposited under local and temporary shallow marine conditions, lie between the Hendre and overlying Mydrim Shales in places. On the coast at Abereiddi, conodonts from a similar unit (Castell Limestone Formation) suggest a level low in the *tvaerensis* Biozone, and graptolites and trilobites from shales immediately above indicate the *gracilis* Biozone (Bergström *et al*. 1987; Owens 2000). In contrast to the Drefach Group to the north, the lithofacies and fauna indicate shallower, calcareous sediment accumulation and trilobite colonization of the sea floor (Burt *et al.* 2012).

Mudstones and siltstones of *artus* to *murchisoni* Biozone age in the Builth Wells Inlier (Camnant Mudstone Formation; Fig. 5, column 15) were deposited in a range of environments from shallow marine to deeper offshore settings (Davies *et al.* 1997; Botting and Muir 2008). Faunas from these beds and the overlying Builth Volcanic Group and Llanfawr Mudstone Formation have been extensively documented (Botting and Muir 2008; Muir and Botting 2015 and references therein). Graptolites and trilobites from the Llanfawr Mudstone Formation (dark grey, locally finely laminated, fossiliferous mudstone; Davies *et al.* 1997; Schofield *et al.* 2004) indicate the *murchisoni*, *teretiusculus* and *gracilis* biozones. Ash beds and laminae are present in the *teretiusculus* Biozone, but volcanic influence declines upwards.

The Llanwrtyd Volcanic Formation in the Tywi Lineament is overlain sharply by dark grey, variably sandy, micaceous, poorly bedded and blocky mudstones that are locally slumped or debritic in origin. These are overlain by black, graptolitic and pyritic hemipelagic mudstones (respectively the Pistyllgwyn and Sugar Loaf members of the St Cynllo's Church Formation; Davies *et al*. 1997; Schofield *et al*. 2004; Fig. 5, column 13). Graptolites indicate the *foliaceus* and *clingani* biozones. Unbioturbated graptolitic mudstones at the top of the St Cynllo's Church Formation correlate with the Nod Glas Formation of North Wales (Figs 3, 4; see below). The postulated transgression that they indicate has been attributed to regional subsidence following the cessation of volcanism (Brenchley *et al*. 2006).

South Wales oxic facies

The Abergwilli Formation is laterally equivalent to and overlain by the Ffairfâch Grit Formation at Llandeilo while the lower part of the Drefach Group is equivalent to the upper Ffairfâch Grit and Llandeilo Flags formations (Fig. 5, column 11). The Ffairfâch Grit Formation (formerly the Ffairfâch Group, e.g. Fortey *et al*. 2000) is a sandstone-dominated succession of Abereiddian to Llandeilian (Darriwilian) age with medium- to coarse-grained arkosic and pebbly sandstones, fine calcareous sandstones, siltstones, shales, limestones and bentonites (Schofield *et al.* 2009). Felsic tuff and rhyolite (Coed Duon Volcanic Formation) from a local volcanic centre indicate brief contemporaneous volcanic activity. The Llandeilo Flags Formation (equivalent to the Golden Grove Group of Sutton *et al*. 1999) overlies both the Ffairfâch Grit and Coed Duon Volcanic formations and comprises hummocky cross-stratified sandstone with interbedded limestone and bioturbated sandstone and siltstone (Owens 2000; Schofield *et al.* 2009). The Ffairfâch Grit and Llandeilo Flags formations have been interpreted as either an upwardly deepening sequence, with sublittoral to intertidal facies in the Ffairfâch Grit Formation passing upwards into intertidal to open shelf facies in the Llandeilo Flags Formation (Williams *et al*. 1981; Wilcox and Lockley 1981; Lockley 1983; Owens 2000); or alternatively as an accumulation of resedimented material deposited in an intrashelf, possibly fault-bounded depression (Ffairfâch Grit Formation) overlain by more proximal shelf or shoreface facies (Llandeilo Flags Formation) laid down during marine regression (Schofield *et al*. 2009).

Conodonts show that the succession ranges from the *lindstroemi* Subzone of the *serra* Biozone in the Ffairfâch Grit Formation to the *variabilis* Subzone of the *tvaerensis* Biozone in the upper Llandeilo Flags Formation (Bergström *et al.* 1987; Ferretti and Bergström 2022). The base of the Caradoc Series (base Sandbian Stage) is placed at a level in the lower part of the Llandeilo Flags Formation (Ferretti and Bergström 2022, fig. 2). Argillaceous limestones and flags farther west, near Haverfordwest (Bryn-Banc Limestone, Narberth Group in Fortey and Rushton 2000a), have yielded conodonts of the *variabilis* Subzone and trilobites indicative of the upper Aurelucian Stage (lower Caradoc; Sandbian, Sa1) (Owens 2000; Ferretti and Bergström 2022).

Penyraber, Cwm-yr-Eglwys Mudstone and Dinas Island formations

The succession above the Fishguard Volcanic Group northwards along the South Wales coast differs from the anoxic Drefach Group to the south and represents the fill of a graben that formed along the Fishguard–Cardigan Fault Zone (Davies *et al*. 2003; Fig. 4). Grey mudstones with siltstones, sandstones, rare hemipelagic mudstones and debrites (Penyraber Mudstone Formation) rest on an irregular erosion surface on top of the Fishguard volcanic rocks. Burrow-mottling and diagenetic phosphate nodules indicate oxygenated bottom conditions. Graptolites from a raft in a debrite (Williams *et al.* 2003) suggest the *foliaceus* Biozone. The formation indicates progressive deepening and marine transgression during a period of sustained instability.

Dark grey turbiditic and hemipelagic mudstones (Cwm-yr-Eglwys Mudstone Formation) above the Penyraber Mudstone Formation have yielded graptolites of the *clingani* and *linearis* biozones (Katian Stage) in the south (Williams *et al.* 2003) but are no younger than the *caudatus* Subzone (*clingani* Biozone) in the north where they pass laterally into the Dinas Island Formation (Davies *et al.* 2003; Fig. 4). The latter comprises turbidite sandstones and mudstones with debritic mass-flow deposits and laminated hemipelagic mudstones; graptolites indicate a probable *morrisi* Subzone age (Williams *et al.* 2003). The distribution and relationships of the Cwm-yr-Eglwys Mudstone and Dinas Island formations suggest that major fractures of the Fishguard–Cardigan Fault Belt continued to influence sedimentation (Davies *et al.* 2003; Williams *et al.* 2003). Laminated hemipelagic mudstones in both formations indicate anoxic bottom water conditions.

Post-volcanic sedimentation: Nod Glas Formation and equivalents

Black, pyritous, graptolitic mudstones were deposited across North Wales following the cessation of major volcanic activity and are now all included in the Nod Glas Formation (see Rushton and Howells 1999; Fig. 3). They are considered to have been deposited in anoxic settings, including astride former intrabasinal highs, during a period of high sea level and post-volcanic subsidence (Brenchley *et al*. 2006b). Different facies are present from west to east and the onset of black shale deposition appears to have been later in the east.

The contact with underlying volcanic rocks is gradational south of the Harlech Dome and on the Llŷn Peninsula (Pratt *et al.* 1995; Young *et al.* 2002) but abrupt elsewhere (Owens *et al.* 2000). Graptolites indicate the *clingani* Biozone (Pratt *et al.* 1995; Howells and Smith 1997; Owens *et al.* 2000); *linearis* Biozone faunas are possibly also present (Price 1984; Owens *et al.* 2000; Young *et al.* 2002). Trilobites and brachiopods from calcareous tuffs beneath the Nod Glas Formation at Conwy indicate an early Cheneyan (Katian, Ka1) age (Owens *et al.* 2000).

Eastwards, the lower part of the Nod Glas Formation passes laterally into the Gelli-grin Calcareous Ash Formation (Lockley 1980; Rushton and Howells 1999; Fig. 3, column 3). Farther east, around Welshpool, the Nod Glas Formation includes a thin basal unit of phosphatic mudstone and limestone (Pen-y-garnedd Phosphorite Member). Graptolites indicate the *clingani* Biozone, and trilobites the Onnian Substage (upper Streffordian Stage: Owens *et al.* 2000; Cave 2008; Fig. 3, column 4). In contrast, conodonts from the Nod Glas Formation in eastern Wales have been reported to indicate a younger, *ordovicicus* Biozone age (Savage and Bassett 1985; Owens *et al.* 2000; see also Ferretti *et al.* 2014a). Nevertheless, a hiatus equivalent to at least the upper Cheneyan–lower Streffordian stages (Ka1) separates the Nod Glas Formation from underlying beds. Similar phosphatic facies are present in the north and west Berwyn Hills, where a hiatus of comparable extent is inferred beneath the formation (Brenchley 1978).

Ashgill Series (Pusgillian–Rawtheyan stages; upper Katian Stage)

The black anoxic facies (Nod Glas, Mydrim Shale and upper St Cynllo's Church formations) that became widespread across the Welsh Basin with the cessation of major volcanic activity was replaced across the central part of the Welsh Basin in the Ashgill by bioturbated oxic facies. In basinal settings, the lower part of the Ashgill succession consists predominantly of burrow-mottled silty turbiditic mudstones with hemipelagic mudstones (Nantmel Mudstones, Broad Vein Mudstone and Narrow Vein Mudstone formations: Pratt *et al.* 1995; Davies *et al*. 1997, 2003, 2006; Burt *et al.* 2012; Figs 4, 5). The Nantmel Mudstones Formation has a broad outcrop along the basinal flank of the Tywi Lineament in South Wales, overlying the Cwm-yr-Eglwys Mudstone Formation, Dinas Island Formation and Drefach Group across the Fishguard–Cardigan Fault Zone. In contrast to its earlier control, the fault zone appears to have had no influence on sedimentation. The base of the Nantmel Formation is generally taken to approximate to the Caradoc–Ashgill series boundary (Fortey and Rushton 2000a; but see Vandenbroucke *et al.* 2008 who suggested a Cautleyan age). Where the contact is not faulted, the formation is conformable on underlying units (Davies *et al.* 1997; Schofield *et al*. 2004, 2009).

A hemipelagic mudstone-dominated interval in the *anceps* Biozone is characteristic of upper levels of the Nantmel Formation and its equivalents across the basin ('Red Vein' of Pugh 1923; Pratt *et al.* 1995; Davies *et al*. 2006; Wilby *et al*. 2007; Challands *et al*. 2009). In the centre of the basin, the Nant-y-môch Formation in the Plynlimon Inlier (Fig. 4, column 7), comprising turbiditic mudstone with thinly interbedded sandstones and siltstones, also includes subordinate thinly bedded hemipelagic mudstone of *anceps* Biozone age (Cave and Hains 1986). The changes in lithofacies in the Nantmel Formation and its equivalents have been attributed to changes in sea level, the change from anoxic to oxic facies at the base being possibly the result of lowered sea level, and the hemipelagic mudstone interval in the *anceps* Biozone being attributed to widespread deepening (Davies *et al*. 2006; Page *et al*. 2007; Wilby *et al.* 2007). A further explanation, either as an alternative or complementary to changes in sea level, is that the hemipelagic mudstones in the *anceps* Biozone, which are rich in organic carbon, were deposited during periods of enhanced coastal upwelling and surface organic productivity. This, in turn, has been attributed to strengthening trade winds brought about by Hadley cell migration in response to Late Ordovician changes in ice volume (Challands *et al*. 2009). Contemporaneous alternations between anoxic and oxic facies in the Southern Uplands of Scotland have been attributed to changes in thermohaline circulation, climate and upwelling, the contemporaneity suggesting that such processes may have been ocean-wide and reflected fundamental changes in the Late Ordovician climate-ocean system (Armstrong and Coe 1997; Challands *et al*. 2009). The late Katian Boda event has been attributed to such episodes of pre-Hirnantian climate change, initially global warming (Fortey and Cocks 2005; see also Cocks and Torsvik 2021) and then reinterpreted by Cherns and Wheeley (2007) to signal global cooling.

The extent to which any deepening might be attributed to eustatic sea level rise, intra-Ashgill tectonism or a combination of both is not clear (Schofield *et al.* 2004). Coarse clastic turbidite deposits of limited extent (Taliaris, Bryn Nicol and Doldowlod Conglomerate formations; Schofield *et al*. 2004, 2009; Fig. 5, columns 10, 11) west of the Tywi Lineament were derived from the east and are equivalent to the upper part of the Nantmel Mudstones succession. They are interpreted as the deposits of small fault-controlled depressions that developed during the mid to late Ashgill Shelveian Event (Toghill 1992) of the Welsh Borderland, which is thought to record the collision of Baltica with Eastern Avalonia (Smallwood 1986; Toghill 1992; Schofield *et al.* 2004, 2009; Davies *et al*. 2009). The Tridwr Formation (see below) might also demonstrate the combined effects of late Ashgill tectonism and glacio-eustatic sea level changes, having been subjected to uplift and deformation, subaerial weathering and erosion prior to deposition of the transgressive late Hirnantian Cwm Clŷd Sandstone (Schofield *et al*. 2004, fig. 2; Barclay *et al*. 2005, fig. 3; British Geological Survey 2005; Davies *et al*. 2009).

Across North Wales (Fig. 3), the Nod Glas Formation is overlain by bioturbated siliciclastic and calcareous mudstones and siltstones, locally with limestones and sandstones, that were deposited under oxygenated conditions in mid- to outer-shelf settings. Most of these deposits have yielded Cautleyan or Rawtheyan (Katian, Ka4) shelly faunas (Price 1981; Price and Magor 1984; Cocks and Rong 1988; Owens *et al.* 2000; Young *et al.* 2002). An impersistent limestone with mudstone partings (Rhiwlas Limestone) at the base of the succession in the Bala district (Fig. 3, column 3) has yielded *ordovicicus* Biozone conodonts (Owens *et al.* 2000). There is no definite evidence for the Pusgillian Stage or the *linearis* or *complanatus* graptolite biozones, although some correlations raise the possibility that the base of the succession In the Berwyn Hills is in the upper Pusgillian (Brenchley 1978; Hiller 1981; Rushton and Fortey 2000). A widespread unconformity is recognised at the base of the Ashgill succession across North Wales (Fig. 3) and is attributed to local uplift.

Between Haverfordwest and Llandovery in South Wales, the Mydrim Shales Formation is overlain locally by impersistent limestones with calcareous shale and siltstones (Sholeshook, Robeston Wathen, Birdshill and Crûg limestones; Fig. 5, columns 10, 11). The change from black shale to carbonate deposition is inferred to reflect a decrease in water depth and perhaps cooling (Owens 2000). A gradual passage from the Mydrim Shales Formation into the Sholeshook Limestone near Whitland (Zalasiewicz *et al.* 1995) implies that there is no hiatus between them (but see Brenchley *et al.* 2006b on the possibility of a break). Conodonts indicate the lower part of the *ordovicicus* Biozone for these carbonate developments, with the proviso that that the *superbus* Biozone might be represented in the lower part of the Crûg Limestone (Ferretti *et al.* 2014a). Slightly different ages are indicated by shelly faunas, with trilobites from the Robeston Wathen Limestone reported to indicate a Rawtheyan age whereas faunas from the Birdshill Limestone suggest a late Pusgillian–early Cautleyan age and those from the Crûg Limestone a probable Cautleyan age (Owens 2000). Graptolites from the Sholeshook Limestone suggest the *anceps* Biozone, and shelly faunas indicate a Cautleyan–Rawtheyan age with an apparently diachronous base, possibly extending down into the Pusgillian at Whitland (Owens 2000). Corals from the Robeston Wathen Limestone are common to Ashgill limestones elsewhere (e.g., the Portrane Limestone of Ireland; cf. Baars 2013).

The Birdshill and Crûg limestones have restricted outcrops on the NW flank of the Tywi Lineament near Llandeilo where they are overlain by the Nantmel Formation (Fig. 5, column 11). The Sholeshook and Robeston Wathen limestones on the SE flank of the lineament farther west, near Haverfordwest, are overlain by the Slade and Redhill Mudstone Formation, predominantly mudstones with subordinate sandstones, which contains a shelly fauna of Rawtheyan age (Cocks and Price 1975; Owens 2000; Brenchley *et al.* 2006b; McCobb *et al*. 2019; Fig. 5, column 10). Upward constraints on the age of the Slade and Redhill Mudstone Formation are poor (Brenchley *et al.* 2006b), but black shale (Scotchwell Shale Member) at the base of the overlying Portfield Formation (Fig. 5, column 10) might indicate a change to anoxic conditions that could correspond to the *anceps* Biozone deepening inferred in the Welsh Basin (Brenchley *et al.* 2006b). Higher units of the Portfield Formation are unfossiliferous but possibly Hirnantian (Ingham and Wright 1970; Cocks and Price 1975; Brenchley *et al.* 2006b; see below).

A mid to outer shelf facies of mudstone with intercalated shelly sandstones (Tridwr Formation; Fig. 4) is equivalent to the Slade and Redhill Mudstone Formation along the Tywi Lineament towards the NE and is correlated with the lower Nantmel Formation of basinal successions (Fig. 5, column 14). A Rawtheyan shelly fauna and *anceps* Biozone graptolites have been collected from the upper part of the Tridwr Formation (Cocks *et al.* 1984; Challands et al. 2014). Intercalations of Nantmel Mudstones facies are probably of late Rawtheyan age (Schofield *et al.* 2004). The Nantmel Mudstones Formation passes upwards and laterally along the Tywi Lineament into bioturbated sandy mudstone and muddy sandstone of Rawtheyan age (Cribarth Formation; Fig. 4, column 14), the upper beds of which mark a transition into the Hirnantian succession (Schofield *et al.* 2004, 2009).

Ashgill Series (Hirnantian Stage)

Overlying the Nantmel Mudstone Formation and its equivalents in basinal successions are silty turbiditic mudstones, locally interbedded with turbidite sandstones and debrites (Garnedd-wen Formation, Fig. 4, columns 6, 7; Yr Allt Formation, Figs 4, 5). The change from bioturbated mudstone facies to the silty mudstone facies of the Yr Allt Formation and its correlatives is taken to mark the onset of the Hirnantian glacio-eustatic regression (Waters *et al.* 1992). Slump beds are characteristic of the Yr Allt and Garnedd-wen formations (Davies *et al.* 2009; Pratt *et al.* 1995).

The lower part of the Yr Allt Formation passes south-eastwards across the Tywi Anticline into burrowed silty mudstone with calcareous siltstone and sandstone (Ciliau Formation; Fig. 5, column 14), interpreted as a lower shoreface facies deposited during the early part of the Hirnantian regression (Davies *et al*. 2009). Overlying sandstones (Cwmcringlyn Formation; Fig. 5, column 14) record deposition during the glacial maximum, when much of the area to the east was emergent and subjected to subaerial erosion (Schofield *et al.* 2004, 2009). Shelly fossils belonging to the *Hirnantia* fauna are present in both the Ciliau and Cwmcringlyn formations (Davies *et al.* 1997; Schofield *et al.* 2004).

Silty mudstones (Foel Y Ddinas Mudstone Formation) are also the dominant rock-type in the Bala district of North Wales (Bassett *et al.* 1966; Fig. 3, column 3), with sandstones at the top of the formation. The Hirnant Limestone Member, at the same level as the sandstones, is a pisolitic limestone in mudstone and silty mudstone and is interpreted as a channel fill (Owens *et al.* 2000). Eastwards, in the Berwyn Hills, fine- to coarse-grained sandstones and thin siltstones (Glyn Formation) with a locally developed limestone at the base (Glyn Limestone Member) were deposited in a shallow sublittoral environment (Hiller 1981). A shelly fauna from the limestone suggests a Rawtheyan age. A more restricted assemblage from higher in the formation contains elements of the *Hirnantia* fauna (Hiller 1981), but Rong *et al.* (2020) considered the assemblage to be latest Katian rather than Hirnantian. Turbiditic sandstones and siltstones on the North Wales coast (Conwy Castle Grit Member; Fig. 3, column 2) also contain an allochthonous *Hirnantia* fauna (Owens *et al.* 2000).

Higher Hirnantian deposits across the Tywi Lineament record the postglacial rise in sea level. The Garth House Formation conformably overlies an Hirnantian regressive sequence in more westerly sections (Fig. 5, column 14). Towards the east, the Cwm Clŷd Sandstone, interpreted as a transgressive beach and upper shoreface facies, is interposed beneath the Garth House Formation and lies above an unconformity between the regressive and transgressive parts of the Hirnantian succession; the magnitude of the unconformity increases eastwards (Schofield *et al.* 2004; Barclay *et al*. 2005; British Geological Survey 2005; Davies *et al*. 2009, 2011; Fig. 5, column 14). The Garth House Formation comprises smooth, dark grey mudstone with beds of lenticular sandstone (Schofield *et al*. 2004). An upwards decrease in the abundance and thickness of the sandstones is considered to indicate a transition from upper and lower shoreface facies into offshore mud-dominated facies and deepening during the initial stages of a pulsed post-glacial transgression (Schofield *et al.* 2004; Davies *et al*. 2009, 2016). Poorly preserved chitinozoans from the Garth House Formation suggest the *taugourdeaui* Biozone (Davies *et al*. 2013; Fig. 5, column 14).

Highest Hirnantian deposits across much of Wales comprise mudstones of *persculptus* graptolite Biozone age (Pratt *et al*. 1995; Warren *et al.* 1984; Hiller 1981; Bassett *et al.* 1966; Figs 3–5). In North Wales, grey silty mudstones above the Conwy Castle Grit Member (Fig. 3, column 2) with trilobites that include *Mucronaspis mucronata* and *M.* cf. *olini* are overlain by grey mudstones with graptolitic and bioturbated bands that have yielded a *persculptus* Biozone fauna (Warren *et al*. 1984; cf. Gyffin Shales of Rushton and Fortey 2000, fig. 11). In the west of the Bala district, the silty mudstones of the Foel Y Ddinas Mudstone Formation are overlain by dark blue mudstones and laminated siltstones of the Cym yr Aethnen Mudstones Formation (Fig. 3, column 3), the basal beds of which have yielded *Glyptograptus persculptus*, while farther east mudstones and siltstones of the Fron Frys Formation are also of proved *persculptus* Biozone age (Bassett *et al*. 1966; British Geological Survey 1986).

Across the centre and southern parts of the basin, highest Hirnantian deposits are assigned to the Cwmere Formation (Figs. 4, 5). A unit of burrow-mottled hemipelagic and turbiditic mudstones (Mottled Mudstone Member) at the base of the formation sharply overlies the Yr Allt Formation. The Mottled Mudstone is considered to indicate predominantly oxic conditions during the marine transgression that followed the late Ordovician glaciation (Cave and Hains 1986; Davies *et al*. 1997, 2006; Schofield *et al.* 2004; Wilby et al. 2007; Fig. 4). The overlying anoxic facies of the Cwmere Formation, consisting predominantly of thinly interbedded turbidite mudstones and laminated hemipelagites, spans the Ordovician–Silurian boundary, persisting into the lower part of the Aeronian Stage (middle Llandovery), and is considered to have developed in response to subsequent rapid deepening (Davies *et al*. 1997). Locally, to the NW of the Tywi Lineament, the Cwmere Formation above the Mottled Mudstone passes laterally into turbidite conglomerates and sandstones of the Caban Conglomerate Formation (Fig. 5, column 11). The coarser turbidites have been interpreted as proximal resedimented facies at the base of a nested sequence of channel deposits (Davies and Waters 1995). They mark the site of a stable submarine channel that persisted into the Aeronian (Davies *et al*. 1997). Three separate facies have been identified in the lower part of the Caban Conglomerate Formation (Davies *et al*. 1997). The Cerig Gwynion Grits facies comprises medium- to thick-bedded turbidite sandstones and subordinate thin-bedded turbidite sandstones and mudstones, with sparse intercalated laminated hemipelagites, and probably lies entirely within the *persculptus* Biozone. The lowest of five stratigraphically discrete sequences of Caban Conglomerates facies, consisting of medium- to thick-bedded turbidite conglomerate and sandstone couplets with rare very thin turbidite sandstone/mudstone couplets, ranges from the upper *persculptus* Biozone into the lower *acuminatus* Biozone (basal Silurian). The Dyffryn Flags facies of thin-bedded turbidite sandstone/mudstone couplets and laminated hemipelagites typically has gradational contacts with the Cwmere Formation; it occupies a transitional position between the Cwmere Formation and the other facies of the Caban Conglomerate Formation from the *persculptus* Biozone to the *acuminatus* Biozone. Deposition of the lower parts of the Caban Conglomerate Formation was initiated and continued through the late Hirnantian post-glacial rise in sea level (Davies *et al*. 2016, fig. 6). The transport of coarse turbidite sediment to the basin at this time might reflect tectonic uplift of source areas, although recalibrations suggested by Davies *et al*. (2013, 2016) imply that eustasy might have been more influential than previously believed. Supply routes from the east have not been identified, however, although it is possible that they may have been offset by strike-slip displacement along the Tywi Lineament (Davies *et al*. 1997).

Eastwards across the Tywi Lineament, facies of *persculptus* Biozone age pass laterally from the Cwmere Formation through a mixed oxic-anoxic succession of extensively slumped turbidite and hemipelagic mudstones (Tycwtta Mudstone Formation; Davies *et al.* 1997), into unfossiliferous but locally bioturbated silty mudstones, considered to be distal shelf/upper slope facies (Chwefri Formation) (Davies *et al.* 2009; Schofield *et al.* 2004, 2009) and then into fossiliferous sandy mudstone (Bronydd Formation) and muddy sandstone (Crychan Formation, Goleugoed Formation) (Schofield *et al.* 2004, 2009; Davies *et al*. 2013; Fig. 5, columns 12–14). The Chwefri, Bronydd, Goleugoed and Crychan formations record a major progradational event (Schofield *et al.* 2004, 2009; Davies *et al*. 2013) that was initiated during the interval of post-glacial deepening in late *persculptus* Biozone time and persisted into the early Silurian (Rhuddanian; Davies *et al*. 2016). It must therefore record high levels of sediment supply and perhaps tectonic rejuvenation of clastic source areas (Davies *et al.* 1997, 2016).

Davies *et al*. (2009) identified the unconformity beneath the Cwm Clŷd Sandstone and Garth House Formation as a major sequence boundary that they were able to trace across the basin. In the Plynlimon area of mid Wales, a laterally persistent turbidite sandstone unit (Pencerrigtewion Member at the top of the Drosgol Formation; Fig. 4, column 7) was correlated with the Cwmcringlyn Formation (Fig. 5, column 14) and interpreted as having been deposited during the glacio-eustatic lowstand. The slumped mudstones of the overlying Brynglas Formation (Fig. 4, column 7) are noticeably finer grained than the sandy and silty mudstones of the Drosgol Formation, a feature attributed to an abrupt reduction in the grade of sediment coming into the basin as the transgression progressed. Davies et al. (2009) noted the same change in the Yr Allt Formation; units of smooth, fine-grained mudstone in the upper part of the formation differ from the siltier lithologies in its lower part, and more closely resemble those of the Garth House Formation. Similarly, a unit of smoother mudstones in the upper part of the Garnedd-Wen Formation (Fig. 4, columns 6, 7; Upper Garnedd-wen Beds of Pugh 1928, 1929) can be distinguished from an underlying, coarser division. The contact between the two units in the Garnedd-Wen Formation is taken to mark the glacial maximum regression (James 1985; Davies *et al*. 2009) and the upper unit is regarded as having been deposited during the initial post-glacial transgression. The Brynglas Formation and the upper parts of the Yr Allt and Garnedd-Wen formations are now correlated with the Garth House Formation to the east, replacing previous correlations of the Garth House Formation with the Cwmere Formation in basinal successions (Figs 4, 5). The Mottled Mudstone Member at the base of the Cwmere Formation (e.g. Fig. 5, columns 11, 12) is now correlated eastwards with a bioturbated interval at the base of the Tycwtta Formation and with the burrow-mottled Ystradwalter Member at the base of the Chwefri Formation (Davies *et al*. 2009; Fig. 5, column 14). Morphometric analysis of normalograptid graptolites from the *persculptus* and basal Silurian *acuminatus* biozones of central Wales (Blackett *et al*. 2009) was instrumental in correlating the Mottled Mudstone and Ystradwalter members. Blackett *et al*. (2009) identified four successive morphospecies that could be used for refined biostratigraphy and correlation in the late Ordovician and early Silurian of central Wales, distinguishing two successive lower *persculptus* Biozone intervals, an upper *persculptus* to lower *acuminatus* Biozone interval, and a middle *acuminatus* Biozone interval.

The sequence of sedimentological and faunal events recognised across the Tywi Lineament and traced into the basin might be discernible elsewhere near the basin margin (Davies *et al*. 2009). Towards the eastern margin of the basin, near Welshpool, an impersistent sandstone (Graig-wen Sandstone Formation; Fig. 3, column 4) has been interpreted as part of a complex channel fill, overlain by bioturbated shales with graptolites of probable Hirnantian age (Laundry Mudstone Formation; Fig. 3, column 4; Brenchley et al. 2006a; British Geological Survey 2008), or as a transgressive sand deposited during the post glacial rise in sea level and probably of *persculptus* Biozone age (Cave 2008). The Laundry Mudstone Formation has blue-black mudstones lacking benthic shelly fauna at its base (Brenchley *et al*. 2006a), inviting comparison between these basal beds and the Garth House Formation. In SW Wales, the Portfield Formation is overlain by the Haverford Mudstone Formation (Cocks and Price 1975; Fig. 5, column 10), its paler colour and bioturbation being suggestive of regression (Brenchley *et al*. 2006b). Davies *et al*. (2009) suggested that unfossiliferous mudstones at the top of the Portfield Formation (Cuckoo Shale Member) might also be equated with the Garth House Formation, the underlying Cethings Sandstone Member in the middle Portfield Formation possibly correlating with the Cwm Clyd Sandstone and/or Cwmcringlyn formations. Davies *et al*. (2009) further suggested that bioturbated olive-green mudstone with a *Hirnantia* fauna at the base of the Haverford Mudstone Formation (St Martin's Cemetery Bed) might then be equivalent to the Mottled Mudstone and Ystradwalter members, in which case the *Hirnantia* fauna would be from the post-glacial maximum succession (Davies *et al*. 2009) and possibly from the lower *persculptus* Biozone (cf. Rong and Harper 1999; Rong *et al*. 2002; Rong *et al*. 2020). Above the

St Martin's Cemetery Bed, the lower part of the Haverford Mudstone contains a shelly fauna with *Mucronaspis mucronata* and *Kozlowskites gracilis*. The upper part of the formation has a rich Rhuddanian shelly fauna, however, and the succession appears to be unbroken from the Rawtheyan into the Silurian with no documented evidence for emergence at the glacial maximum (Brenchley *et al.* 2006b).

Wrekin, Charnwood and Fenland terranes

Ordovician rocks are poorly exposed in the Wrekin, Charnwood and Fenland terranes across SE Wales and southern and eastern England but are widely distributed at depth. The most significant exposures are along the western edge of the Wrekin Terrane where the type area of the Caradoc Series is situated east of the Church Stretton Fault, the easternmost component of the Welsh Borderland Fault System. The Wrekin and Charnwood terranes formed a stable area (Midland Platform) to the east of the Welsh Basin throughout the Ordovician, except during the Tremadocian when narrow rift-basins developed in which thick sequences accumulated. Ordovician rocks of post-Tremadocian age are mainly absent on the platform and were either removed or the area remained emergent throughout most of the early Palaeozoic. To the east, the Fenland Terrane formed a basinal area (Anglian Basin) that includes Ordovician rocks of post-Tremadocian age, including arc volcanics and associated intrusions.

Tremadocian Stage

Tremadocian rocks are widespread across the Wrekin and Charnian terranes and, where their base is seen, they are conformable on upper Cambrian beds (Fig. 6). They are of fairly uniform lithology, comprising mainly grey or greenish grey mudstones and shales with siltstones and sandstones. They generally contain a shelly fauna and dendroid graptolites that form the basis of the Tremadocian biozones used for these successions, including the *Rhabdinopora flabelliformis* and *Clonograptus tenellus* graptolite biozones and the *Shumardia salopiensis* trilobite Biozone.

Tremadocian rocks (Shineton Shale Formation) are overlain unconformably by the type Caradoc Series east of the Welsh Borderland Fault System, and similar Tremadocian rocks are found in the Shelve area to the west (see Cymru Terrane). Other outcrops (with different formation names) are at the southern end of the Malvern Hills and southwards along the Malvern Line in the Tortworth Inlier (Fig. 6). The latter is part of the infill of a graben, variously termed the 'Tremadoc Worcester Graben' (Smith and Rushton 1993) or the 'Worcester Proto-graben' (Pharaoh 2018), that accumulated a significant thickness of Tremadocian rocks (>2km: Pharaoh 2018). Similarly, the Shineton Shale Formation in the Caradoc Series type area, where it is >1km thick, accumulated in a half-graben, the 'Cressage-Cardington sub-basin'. The grabens are interpreted as transtensional basins that formed in response to the opening of the Rheic Ocean (Smith and Rushton 1993). Elsewhere, Tremadocian rocks of similar lithology have been proved beneath Silurian rocks on the Wrekin Terrane (e.g. in the Fownhope-1 Borehole, Pharaoh 2018) or beneath rocks of Carboniferous or Mesozoic age in boreholes across the Charnian Terrane (Molyneux 1991; Rushton 2000; Pharaoh 2018; Fig. 6).

Apart from the widespread Tremadocian mudstones, a minor succession of tuffs, tuffaceous sandstones, siltstones and mudstones (Barnt Green Volcanic Formation) in a small, fault-bounded inlier in the Lickey Hills, SW of Birmingham, is tentatively placed in the Tremadocian (Old *et al.* 1991; Rushton 2000).

Arenig Series (Floian–lower Darriwilian stages)

There are no known rocks of Arenig age at either outcrop or subcrop across these terranes, with the possible exception of a quartzite (Lickey Quartzite) in the Lickey Hills (Old *et al.* 1991; Rushton 2000).

Llanvirn Series (middle and upper Darriwilian stages)

Llanvirn rocks (Abereiddian Stage) have been proved in boreholes along the eastern and northern flanks of the Midland Platform and comprise grey mudstones and siltstones of probable *artus* Biozone (Dw2) age in the Eyam, Great Paxton and Huntingdon boreholes (Molyneux 1991; Pharaoh 2018; Fig. 6).

Several boreholes in the northern Fenland Terrane, NE of the Charnwood Terrane, have proved rocks of a calc-alkaline volcanic suite in a NW–SE trending belt about 150 km long (Pharaoh 2018). A crystal-lithic tuff in one borehole has yielded a U-Pb zircon age of 449 ± 13 Ma, and a Rb–Sr whole rock isochron age of 466 ± 11 Ma has been obtained from a succession of andesite, dacite and rhyolite lavas in a second borehole; the latter is interpreted as being close to the age of eruption (Pharaoh 2018). A phase of plutonic magmatism at around 460 Ma (late Darriwilian) is considered to be penecontemporaneous and comagmatic with the volcanic activity (Pharaoh 2018; Fig. 6).

Caradoc Series (Sandbian–lower Katian stages)

The *gracilis* transgression, possibly enhanced by movement on faults of the Welsh Borderland Fault System (WBFS), brought marine conditions eastwards from the Welsh Basin onto the western edge of the Midlands Microcraton towards the end of the early Sandbian (Sa1). Sediment was deposited in three small sub-basins (Cressage-Cardington, Onny, Ludlow) east of the Church Stretton Fault, the easternmost fault of the WBFS (Smith and Rushton 1993; Brenchley *et al.* 2006b). The northerly Cressage-Cardington sub-basin and the Onny Sub-basin immediately to its south constitute the type area of the Caradoc Series. The Caradoc rocks in the sub-basins are unconformable on Precambrian or Tremadocian rocks and are unconformably overlain by Silurian rocks (Figs 6, 7). They comprise shallow marine deposits, mainly sandstones, siltstones, mudstones and shales with some carbonate deposits and with deeper marine mudstone in the upper part of the succession (Brenchley *et al.* 2006b). The successions in the two sub-basins are broadly similar but differ in detail. Hummocky cross-stratification is rare in sandstones of the Cressage-Cardington Sub-basin but prevalent in equivalent deposits of the Onny Sub-basin (Brenchley *et al.* 2006b), and the succession in the Cressage-Cardington Sub-basin is less complete, with hiatuses developed within the succession (Bowdler-Hicks *et al.* 2002, text-fig. 1; Fig. 7). A shelly fauna of brachiopods and trilobites is present throughout and forms the basis of subdivision of the type Caradoc succession into stages and substages (Fortey and Rushton 2000b; Owen *et al*. 2000). Also present are conodonts (Savage and Bassett 1985), acritarchs (Turner 1982, 1984) and chitinozoans (Jenkins 1967; Vandenbroucke *et al.* 2009).

Sedimentary rocks of Caradoc age are virtually absent elsewhere across the Wrekin, Charnwood and Fenland terranes, an exception being a sequence of steeply dipping grey micaceous siltstones and sandstones beneath Jurassic rocks in the Bobbing Borehole of SE England (Fig. 6). These have yielded a shelly and graptolite fauna of Late Ordovician age, and acritarchs and chitinozoans of Caradoc age (Molyneux 1991; Pharaoh 2018).

Monian Terrane

Anglesey

The island of Anglesey hosts a complex of metamorphosed Neoproterozoic–Early Ordovician(?) igneous and metasedimentary rocks, overlain by cover sequences of Ordovician, Silurian, Devonian and Carboniferous age. It is separated from the Welsh Basin/Cymru Terrane by the Menai Strait Fault System; the NW boundary is mapped offshore along the SW–NE trending Môn–Deemster Fold–Thrust Belt (Pharaoh *et al.* 2020).

Tectonostratigraphical framework and Neoproterozoic–Lower Ordovician(?) succession
The Neoproterozoic–Early Ordovician(?) rocks at the base of the succession on Anglesey have been referred historically to the 'Monian Supergroup'. These rocks were affected by Early Ordovician deformation at *c.* 474 Ma (mid Floian, Fl1 Stage Slice; Fig. 8), based on a K–Ar age of 474 ± 9 Ma from phengite in the New Harbour Formation, at the top of the succession in NW Anglesey (Asanuma *et al.* 2017; Schofield *et al.* 2020). The deformation episode is referred to as the Monian Orogeny on Anglesey and approximates temporally to the Penobscotian Orogeny of the northern Appalachians (see Schofield *et al.* 2020 for discussion). The timing of deformation means that the pelites, semipelites and subordinate psammites at the top of the Monian Supergroup could be Early Ordovician rather than older. This is not yet confirmed, however, in the absence of direct evidence for their age of deposition (Schofield *et al.* 2020).

Lower–middle Arenig Series (Floian–Dapingian stages): Cemaes Group
The oldest rocks considered to be of certain Ordovician age on Anglesey are now placed in the Cemaes Group and are unconformable on Monian Supergroup rocks in northern Anglesey. They constitute the Porth Trefadog Volcanic Formation (formerly the Church Bay Tuffs) overlain by the Porth Swtan Formation (Schofield *et al.* 2020; Fig. 8).

The Porth Trefadog Volcanic Formation comprises felsic ash flow tuff and tuffaceous sandstone that pass gradationally into the Porth Swtan Formation. The latter constitutes a mega-conglomerate mass-transport facies. There is no direct evidence for the depositional age of either formation, but they post-date the 474 Ma deformation event and are overlain unconformably by rocks of late Arenig–Llanvirn (Dapingian–Darriwilian) age. An Arenig, probably Floian–Dapingian, age is therefore inferred. The Cemaes Group formations are considered to record an Early Ordovician episode of extension, volcanism and subsidence with active faulting (Schofield *et al.* 2020).

Upper Arenig–Llanvirn series (upper Dapingian? – Darriwilian stages)
The Cemaes Group is overlain unconformably in northern Anglesey by the Porth Wen Group (Bates 1972; Schofield *et al.* 2020). Elsewhere across Anglesey, the Cemaes Group or the Monian Supergroup are overlain unconformably by rocks that are now placed in the Llyn Alaw Group (Schofield *et al.* 2020; Fig. 8). The Llyn Alaw Group includes rocks of late Arenig–Llanvirn age on Anglesey, but as the names of the units included in that group have not yet been published, the succession is discussed below using the formation names of Bates (1972).

The Porth Wen Group comprises a lower unit of purple pebble-cobble conglomerates (Porth Cynfor Conglomerate Formation) overlain by conglomerates and sandstones with mudstone and siltstone partings (Torllwyn Formation). The Porth Cynfor Conglomerate Formation infills incisions in the underlying rocks and is overlapped by the Torllwyn Formation (Brenchley *et al.* 2006b). Interbedded mudstones and thinly bedded sandstones in the Torllwyn Formation have yielded brachiopods of probable late Arenig (Fennian) age (Bates 1968, 1972; Beckly 1987). The brachiopods are apparently not transported and indicate a shallow-marine environment (Brenchley *et al.* 2006b). The Porth Wen Group is variously interpreted as debris-flow deposits with sandstone turbidite beds derived from a nearshore location (Brenchley *et al.* 2006b), or as an alluvial fan-fluvial-shoreface succession that developed on the northern margin of the basin filled by the Llyn Alaw Group (Schofield *et al.* 2020).

The succession in the main area of Ordovician outcrop on Anglesey (the 'Principal Area' of Bates 1972) and outliers in SE Anglesey consists mainly of dark grey mudstone and mudstone-dominated mass transport deposits, overlain by green oxic-facies mudstone (Schofield *et al.* 2020; their Llyn Alaw Group). At the base of the succession in the Principal Area are either fossiliferous, predominantly well-sorted coarse sandstones with a brachiopod-trilobite fauna (Carmel Formation; Fig. 8), or a thinner succession of unfossiliferous pebbly sandstone, conglomerate and flaggy sandstone (Foel Formation; Bates 1972). The distribution of the two facies is fault-controlled (Bates 1972; Brenchley *et al.* 2006b). Trilobites from the Carmel Formation indicate the nearshore *Neseuretus* Community (Owens *et al.* 2000), and the brachiopods from this and other formations in the Llyn Alaw Group are representative of the Celtic Province (Williams 1973; Neuman and Bates 1978).

The overlying Treiorwerth Formation (Fig. 8) comprises a thick succession (up to 650 m) of predominantly pebbly sandstone and conglomerate in mass debris-flow deposits, banked against an east-dipping fault scarp and probably derived mainly from the Monian Supergroup in the west (Bates 1972; Beckly 1987; Owens *et al.* 2000). A rich shelly fauna, mainly brachiopods, occurs in lenticular beds and has evidently been transported (Owens *et al.* 2000). Graptolite fragments identified as *Tetragraptus headi*? have been recorded (Bates 1972; Owens *et al.* 2000). Cocks and Popov (2021) included the brachiopod fauna from the Treiorwerth Formation in their assessment of brachiopod benthic communities of the Mediterranean Province, referring to it as latest Dapingian and early Darriwilian. They assign the fauna to the Celtic provincial group, concurring with earlier studies (Harper *et al*. 1996; Harper *et al*. 2013).

The Treiorwerth Formation is overlain or replaced laterally by gritty shales, sandstones, pebbly sandstones and a slide conglomerate (Nantannog Formation; Fig. 8). Faunas that include graptolites, brachiopods and trilobites indicate an age ranging from the late Arenig *cucullus* Biozone to the mid Llanvirn *murchisoni* Biozone (Owens *et al.* 2000). Uplift on the basin margins resulted in the deposition of cobble-conglomerate debris flows (Bod Deiniol Formation; Fig. 8) that are interbedded locally with the Nantannog Formation and are probably of Abereiddian age (Bates 1972; Rushton and Fortey 2000).

The Nantannog Formation passes upwards and laterally into blue-black, micaceous and pyritous shales that are locally interbedded with thin sandstones and siltstones. Graptolites indicate the *artus* and *murchisoni* biozones and trilobites have also been recorded (Bates 1972). Alternating mudstones and sandstones resembling turbidites (Dulas Formation) are thrust over the shales but are of uncertain age; elsewhere they overlie the Foel Formation. Overlying oxic turbiditic mudstones (the Parys green shales of Greenly 1919) have yielded an *artus* Biozone fauna (Greenly 1919); Schofield *et al.* (2020) cited an unpublished report of a sparse Dapingian–Darriwilian graptolite fauna.

In the Berw Fault Zone to the SE of Bates' (1972) Principal Area, a basal sandstone unit (Berw-uchaf Grits Formation) is overlain by turbiditic greywacke sandstones with interbedded shales (Dryll Formation), with blue shales (Glanmorfa Shales) above the turbidites (Fig. 8). A shelly fauna and the graptolite *Didymograptus hirundo* indicate that the Berw-uchaf Grits Formation lies in the *cucullus* Biozone (Bates 1972, fig. 2). The succession extends upwards into the *murchisoni* Biozone (Bates 1972, fig. 2). A late Arenig isograptid graptolite facies with *Isograptus gibberulus* has been recorded from dark mudstones within the Menai Strait Fault System in eastern Anglesey and is taken to indicate a deep-water environment (Greenly 1919; Bates 1972; Beckly 1987). Rushton and Fortey (2000) recorded Abereiddian graptolites in the Bryn Celyn Ironstone (Fig. 8) from the same area.

The Carmel and Foel formations are interpreted as the deposits of an initial gentle late Arenig transgression over a surface of some relief (Bates 1972), probably due to a combination of eustatic sea-level rise and subsidence (Beckly 1987). More rapid fault-controlled subsidence followed, resulting in foundering of the Arenig-Llanvirn basin on Anglesey. Faunas from the Berw-uchaf Grits are comparable to those from the Treiorwerth and Nantannog formations and may indicate that the initial transgression was slightly later in the Berw Fault Zone than in the areas to the north (Bates 1972; Rushton and Fortey 2000). In contrast, Beckly's (1987) palaeogeographical model and facies distribution maps envisage a marine transgression spreading across Anglesey from the SE during the late Arenig (Fennian) that did not reach north Anglesey until mid–late Fennian times. The presence of an isograptid fauna within the Menai Strait Fault System indicates a deep marine setting during the late Arenig with a probable oceanic connection.

The upper Llanvirn *teretiusculus* Biozone has only been reported from eastern Anglesey (Llangoed; Fig. 8) and at the NW and NE extremities of the Principal Area (Bonw Ironstone locality and Mynydd Eilian respectively). Bates (1972) questioned whether these might be occurrences of the *gracilis* Biozone without the zone fossil, but Rushton and Fortey (2000, fig. 13) accepted the Llangoed occurrence, which lies within the Menai Strait Fault System. Ironstones in North Wales, however, are associated with the acme of the Middle Ordovician regression and other occurrences on Anglesey are from the *gracilis* Biozone. In the Mynydd Eilian area, alternating shales and coarse pebbly grits (Porth Corwgl Formation) are overlain by shales with *Glyptograptus teretiusculus* (Tyllau Duon Shales) and conglomerates with interbedded shales (Fresh Water Bay Conglomerates); grey shales and siltstones with small scale slumping, pebbly grits and boulder beds (Porth-y-gwichiaid Formation) separated across a fault have yielded *Glyptograptus teretiusculus euglyphus* and *Climacograptus scharenbergi* (Bates 1972).

Lower Caradoc Series (Sandbian–lower Katian, Ka1, stages)

Lower Caradoc beds overlie the upper Arenig–Llanvirn and older beds on Anglesey above a disconformity or unconformity (Bates 1972; Brenchley *et al.* 2006b; Schofield *et al.* 2020). Cherty and sandy shales of *gracilis* Biozone age (Gynfor Shales, Bates 1972; Fig. 8) in northern Anglesey rest with disconformity on the Torllwyn Formation, locally on an erosion surface and with a conglomerate at the base. They are overlain by flaggy black ironstones (Penterfyn Ironstone) followed by dark mudstones, fine ferriferous sandstones with shale partings and bedded oolitic ironstones (Porth Pridd Formation; Fig. 8).

At the NW extremity of the Principal Area of the Ordovician outcrop, around Carmel Head, the base of the Caradoc succession comprises shallow-marine sandstones, grits and sandy breccia (Crewyn Formation) and quartz conglomerates (Gader Conglomerates), and massive poorly sorted breccias interbedded with shales (Garn Formation) (Fig. 8) that were deposited in deeper water from mass flows (Bates 1972; Brenchley *et al.* 2006b). All are placed in the Aurelucian Stage (Sandbian, Sa1 to basal Sa2) of the Caradoc Series and are considered to be lateral equivalents (Rushton and Fortey 2000). Shelly faunas have been obtained from the Crewyn Formation and from blocks of Ordovician limestone in the Garn Formation (Bates 1972). Other clasts in the Garn Formation breccia were derived locally from the Monian Supergroup (Bates 1972). Interbedded shales in the Garn Formation have yielded *gracilis* Biozone graptolites (Bates 1972). From their fauna, the Ordovician limestone blocks appear to be little older than the unit that contains them, but similar limestones are not known *in situ* on Anglesey (Brenchley *et al.* 2006b). The contrast between shallow-marine deposits above the pre-Caradoc (pre-Sandbian) unconformity in some places but debris-flow deposits elsewhere suggests that the *gracilis* transgression coincided with fault movements, with mass-flow deposits focussed on fault scarps (Brenchley *et al.* 2006b). The basal Caradoc beds are overlain by grey, principally unfossiliferous shales that are probably of Caradoc age, locally with siltstone and fine sandstone bands (Fig. 8). The upward passage from the breccias of the Garn Formation into shale is gradual, with intercalations of breccia in the shale.

Farther south in the Principal Area, Caradoc rocks overlie Llanvirn beds with a probable disconformity or unconformity (Fig. 8). In upwards succession, they consist of oolitic ironstone and ferriferous grit (Fferam Ironstones), dark blue graptolitic shales (Fferam Shales) and pebbly grits alternating with shales (Llanbabo Church Grits). The last have yielded a Costonian (upper Aurelucian; Sandbian, upper Sa1 to lower Sa2) brachiopod fauna and graptolites of the *gracilis* Biozone. The ironstones at Llanbabo and elsewhere possibly mark the *gracilis* transgression. A *gracilis* Biozone graptolite fauna accompanied by a brachiopod-trilobite fauna has also been recorded from soft, grey micaceous shales (Tandinas Shales; Fig. 8) in the Ordovician succession of SE Anglesey (Bates 1972). The Llanbabo Church Grits are overlain by shales of the *foliaceus* and *clingani* graptolite biozones (Bates 1972), but no later Ordovician (Katian, Ka2–Hirnantian) rocks are known from Anglesey (Fig. 8).

SE Ireland

Rosslare: Tagoat Group, Floian–lower Darriwilian (Dw1) stages

The small Rosslare Ordovician outcrops (see Harper and Parkes 2000), situated on the south-east tip of Ireland, have a Precambrian basement with Gondwanan affinities (Murphy 1990). The Tagoat Group unconformably overlies the Precambrian Rosslare Complex and the mylonites of the Cambrian Ballycogly Group (Fig. 8). The Tagoat Group comprises three formations, the highest of which (Ballybro Formation) consists of greywacke sandstones and siltstones that contain a diverse brachiopod fauna almost identical to broadly coeval assemblages from Anglesey (Neuman and Bates 1978). The brachiopod faunas (Bates *in* Brenchley *et al*. 1967), which include species of *Ffynnonia*, *Hesperonomiella*, *Paralenorthis*, *Palaeoneumania*, *Productorthis*, *Rhynchorthis, Rugostrophia* and *Tritoechia*, were considered to constitute the core of the Celtic Province (Williams 1973), and further research places these faunas in island and continental-margin settings at medium to high latitudes (Neuman and Harper 1992; Harper *et al.* 1996, 2013). Mid to late Arenig graptolite faunas occur in the preceding Milltown Formation (Brenchley *et al*. 1967; Harper and Parkes 2000). The Tuskar Group, which occurs offshore, may include lateral equivalents of the Tagoat Group, although it is metamorphosed by the Carnsmore Granite (Tietzch-Tyler and Sleeman 1994).

Ribband Group SE of the Wicklow Fault Zone: Cambrian–Dapingian Stage(?)

The Ribband Group in SE Ireland historically comprised numerous local units (Brück *et al*. 1979) but its lithostratigraphy has since been rationalised (see McConnell *et al*. 1999 and references therein; Harper and Parkes 2000). Four tracts were recognised by McConnell *et al*. (1999). Tracts 1 and 2 are both NW of the Wicklow Fault Zone and are therefore included in the Leinster-Lakesman Terrane (see below). Tracts 3 and 4 are SE of the Wicklow Fault Zone, are separated from each other by the outcrop of the Duncannon Group, and the Ribband Group of tract 4 is separated from the Ordovician rocks of Rosslare by the outcrops of the Cambrian Cahore Group, the Cullenstown Formation and the Ballycogly Group mylonites. The base of the Ribband Group in tract 4 is placed in the Wuliuan Stage of the Cambrian Miaolingian Series (Brück and Molyneux 2011).

The Ribband Group of tracts 3 and 4 comprises dominantly distal turbidite successions (Brück *et al*. 1979). The oldest rocks exposed in tract 3, between the Wicklow Fault Zone and the Duncannon Group outcrop, comprise the Ballylane Formation (Fig. 8), inferred to be of Tremadoc or early Arenig (Tremadocian or early Floian) age. The Ballylane Formation consists of laminated green to grey slates and shales interbedded with green or pale-grey siltstones and occasional andesitic flows and tuffs. These rocks are overlain by red-purple, green, grey and buff, laminated slate and siltstone with green greywackes (Oaklands Formation: McConnell *et al*. 1999; descriptions from the Geological Survey of Ireland 2022). The Oaklands Formation has yielded lower Floian graptolites (Rushton 1996).

The Ordovician succession in tract 4 commences with dark grey to black slates, commonly striped with pale siltstone laminae (Ballyhoge Formation), overlain by green-grey and dark grey slaty mudstones with thin siltstone bands and laminae (Seamount Formation). The overlying Riverchapel Formation comprises reddish-purple, buff and green mudstones with siltstone laminae and is capped by feldspathic sandstones (Geological Survey of Ireland 2022). Bioturbation is extensive in the upper two formations (Crimes and Crossley 1968). Graptolites from the Riverchapel Formation indicate a probable early-mid? Arenig (Floian–Dapingian?) age (Brenchley *et al.* 1967; Crimes and Crossley 1968; Skevington *in* Brenchley and Treagus 1970). The underlying formations are inferred to be Tremadoc–early Arenig (Tremadocian–early Floian) (Harper and Parkes 2000); dendroid graptolites are present in the Seamount Formation (Crimes and Crossley 1968). Unfossiliferous dark grey shales with subordinate grey/brown siltstone and sandstone of the Tramore Shale Formation (Fig. 8), around Tramore, Co. Waterford, are inferred to be Cambrian–Early Ordovician on the basis of lithological comparisons with Ribband Group units elsewhere (Carlisle 1979).

Duncannon Group: upper Darriwilian–Katian stages

The Duncannon Group (Gardiner 1974) comprises subaqueous volcanic and interbedded sedimentary rocks in the Wicklow-Waterford belt of magmatic rocks in SE Ireland (Stillman and Williams 1979; McConnell 2000). They overlie the Ribband Group unconformably. The Wicklow-Waterford volcanic rocks are dominated by rhyolite, although basalt, andesite and dacite are present

(Stillman and Williams 1979) and are interpreted to have been generated in a suprasubduction extensional tectonic setting (McConnell *et al*. 1991; McConnell 2000; Fritschle *et al*. 2018). At Courtown on the east coast of Ireland, the succession commences with the Courtown Limestone Formation, comprising fossiliferous limestones, calcareous sandstones and siltstones of Caradoc, Aurelucian (early Sandbian) age with a discontinuous basal conglomerate (Harper and Parkes 2000; Geological Survey of Ireland 2022; Fig. 8). On the south coast near Tramore, the Llanvirn–Caradoc (Darriwilian–Katian) succession grades from a shallower-water shelf facies in the east to a basinal facies in the west (Carlisle 1979; Liljeroth *et al*. 2017). In the east, the succession commences with carbonate facies comprising calcareous sandstones, siltstones, and shales with impure limestone bands containing a rich shelly fauna (Tramore Limestone Formation: Harper and Parkes 2000; Geological Survey of Ireland 2022) that extends down into the Llanvirn (upper Darriwilian) *murchisoni* or *teretiusculus* graptolite biozones (Harper and Parkes 2000; Fig. 8). Over 20 brachiopod species have been described from the Tramore Limestone Formation including the endemic *Hibernobonites* (Liljeroth *et al*. 2017); the faunas confirm an oceanic setting. The Ordovician succession terminates with the Caradoc (lower Katian) Raheen Formation, containing rich brachiopod (Harper *et al.* 2017) and trilobite (Owen *et al*. 1986) faunas generally indicative of a deeper-water environment on the outer-shelf. At both Courtown and Tramore, the hiatus between the Ribband and Duncannon groups embraces at least a significant part of the Darriwilian and possibly also the Dapingian and upper Floian stages (Arenig–Llanvirn series).

The basal carbonate rocks at Courtown are overlain by dark grey to black slaty mudstones of the Ballinatray Formation, considered to be of Caradoc (Sandbian) age and probably the *gracilis* and *foliaceus* graptolite biozones, followed by the Campile Formation (Harper and Parkes 2000; Geological Survey of Ireland 2022; Fig. 8). The Campile Formation has been mapped throughout the Duncannon Group outcrop, extending up into the Ashgill Series (Owen and Parkes 1996), and consists of rhyolite and rhyolitic tuffs or agglomerates in grey and brown slaty mudstones with occasional andesitic tuffs or agglomerates (Geological Survey of Ireland 2022). Near Wicklow, dark grey slate with minor pale sandstone and some tuffaceous levels (Kilmacrea Formation) are overlain by rhyolites, rhyolitic tuffs and breccias and interbedded dark grey slate of the Avoca and Ballymoyle formations. The Avoca Formation has yielded graptolites of the *foliaceus* and *clingani* biozones and a Longvillian (upper Burrellian) shelly fauna (Harper and Parkes 2000), indicating a Caradoc (late Sandbian–early Katian) age (Fig. 8). A rhyolite of the Avoca Formation has been dated to 463.6 Ma (U-Pb, Darriwilian: Fritschle *et al*. 2018; Fig. 8). U-Pb zircon ages of 454 ± 1 Ma (Gallagher *et al*. 1994), 455.4 ± 2.8 Ma and 456.9 ± 2.4 Ma (Fritschle *et al*. 2018) from the Croghan Kinshelagh Granite, considered to be contemporaneous with and related magmatically to the Duncannon Group volcanics (Gallagher *et al*. 1994; Fritschle *et al*. 2018), are Sandbian–early Katian (Fig. 8). The Croghan Kinshelagh Granite post-dates a regional S1 foliation affecting the country rock (Gallagher *et al*. 1994; Fritschle *et al*. 2018), which comprises the Ballylane Formation of the Ribband Group (Geological Survey of Ireland 2022; Fig. 8). The deformation event (D1) associated with the foliation is thought to have occurred at *c.* 460 Ma (late Darriwilian; Fritschle *et al*. 2018).

Ingleton

Lower Ordovician (?): Ingleton Group

Grey-green turbiditic sandstones, siltstones, mudstones and conglomerate (Ingleton Group) crop out in two inliers along the Craven Faults at the southern margin of the Carboniferous Askrigg Block in northern England and have been proved at depth beneath Carboniferous rocks on the block itself (Wilson and Cornwell 1982; Arthurton *et al.* 1988; Soper and Dunning 2005). The Ingleton Group was intruded by early mafic dykes, folded into isoclinal upright folds, was subjected to low-grade metamorphism and was eroded prior to deposition of the Upper Ordovician Dent Group (Soper and Dunning 2005), but its age of deposition is problematic. An Ordovician assignment is based on acritarchs from a single borehole sample that indicated an age no older than Early Ordovician (Wilson and Cornwell 1982); the record of *Veryhachium trispinosum* suggests a late Tremadocian age at the oldest (Servais *et al.* 2018). Rb–Sr isochron ages obtained from the Ingleton Group are 494 ± 18 Ma (O'Nions *et al.* 1973, recalculated from 505 Ma: see Soper and Dunning 2005 and Dodson and Robinson 2006), interpreted as dating isotopic homogenization during either diagenesis or low-grade metamorphism, and 465 ± 10 Ma (Dodson and Robinson 2006), thought to be close to the age of metamorphism although a diagenetic cause could not be discounted.

The Ingleton Group is lithologically distinct from and has a different provenance to the Lower Ordovician Skiddaw Group of NW England (Moore 1992; Soper and Dunning 2005), which is the closest Lower Ordovician succession at outcrop. On petrographical and geochemical grounds, the Ingleton Group is interpreted as being the product of an actively eroding continental arc, in contrast to the quartzose recycled orogen/craton interior source of the Skiddaw Group rocks (Soper and Dunning 2005). The latter authors pointed out that the lithofacies and provenance characteristics of the Ingleton Group matched those of Ediacaran inter-arc basin deposits in the English Midlands and Welsh Borderland rather than those of the Skiddaw Group.

A further difference between the Ingleton and Skiddaw groups is the style of early, pre-Late Ordovician deformation (Soper and Dunning 2005). Unlike the predominantly soft-sediment, gravity-induced deformation involving slumping and olistostrome-formation seen in the Skiddaw Group, the folding in the Ingleton Group is more likely to be tectonic, although pre-cleavage; the cleavage affecting these rocks is superimposed obliquely across the isoclines and is the regional Acadian (end-Early Devonian) cleavage (Soper and Dunning 2005). The major folds had an upright attitude prior to erosion at the sub-Dent Group unconformity and folding was associated with anchizonal metamorphism; it probably also post-dated emplacement of early mafic dykes (Soper and Dunning 2005). The Ingleton Group deformation could be coincident with the mid-Floian 474 Ma episode that affected the Monian Supergroup on Anglesey, but that probably indicates a late Tremadocian–earliest Floian depositional age closely followed by deformation if the acritarch evidence is accepted (Fig. 8). However, recognition of a late Middle Ordovician deformation (D1) event at *c.* 460 Ma affecting Ribband Group rocks now assigned to the Monian Terrane in SE Ireland, coupled with dating of the syn-D1 Dhoon Granite on the Isle of Man in the Leinster-Lakesman Terrane to 457.2 ± 1.2 Ma (late Darriwilian–early Sandbian; Fritschle *et al*. 2018; Fig. 9 and see below), suggests that the possibility of a Middle Ordovician deformation event affecting the Ingleton Group cannot be ruled out.

Ashgill Series (upper Katian–Hirnantian stages)

The Ingleton Group is overlain unconformably by rocks of the Dent Group (Fig. 8), the unconformity at the base of the latter thus overstepping the boundary between the Monian and Leinster-Lakesman terranes. Calcareous siltstones and argillaceous limestones above the unconformity (Norber Formation) have a diachronous base ranging in age from Cautleyan to Rawtheyan, with a siliceous conglomerate in places and a notable neptunian dyke at one locality. A non-sequence probably separates Cautleyan and Rawtheyan strata where Cautleyan beds are present and is probably the result of widespread regression (Arthurton *et al.* 1988).

Above the Norber Formation is a varied succession of siltstones and mudstones with interbedded tuff, sandstone and conglomerate (Sowerthwaite Formation). A sparse fauna is locally present (Arthurton *et al.* 1988) and is indicative of the Rawtheyan and Hirnantian stages of the Ashgill Series. At the base of the formation (Fig. 8), the laterally equivalent Jop Ridding Sandstone (volcaniclastic rocks) and the Dam House Bridge Tuffs (tuffs, probably rhyolitic, interbedded with siltstones and mudstones) indicate localized volcanic activity. A pebble conglomerate (Wharfe Conglomerate; Fig. 8) towards the top of the formation marks a widespread non-sequence and erosional event. It is overlain by laminated sandy siltstones with a Hirnantian fauna that are correlated with the Ashgill Formation of the Lake District and Cautley inliers. Calcareous mudstones with abundant shell debris overlying the Sowerthwaite Formation are likely equivalents of the Spengill Mudstone Member of *persculptus* Biozone age at the base of the Skelgill Mudstone Formation in the Lake District and Cautley areas (Arthurton *et al.* 1988; Kneller *et al.* 1994).

Leinster-Lakesman Terrane

The major Ordovician lithostratigraphical units of the Leinster-Lakesman terrane are the Skiddaw Group of NW England, the Manx Group of the Isle of Man, and those formations of the Ribband Group NW of the Wicklow Fault Zone in SE Ireland, all of Tremadoc–late Arenig or Llanvirn (Tremadocian–Darriwilian) age; the Eycott Volcanic Group and Borrowdale Volcanic Group of Caradoc (Sandbian–early Katian) age in NW England; and the Dent Group of late Caradoc–Ashgill (Katian–Hirnantian) age, also in NW England (Fig. 9). Other Ordovician rocks, not assigned to groups, occur in inliers close to the NW margin of the terrane in Ireland, from the Dingle Peninsula in SW Ireland through Kildare to the area around Portrane and Balbriggan on the east coast of Ireland, extending to Lambay Island. The rocks in these inliers are important for establishing the character of Irish Ordovician successions along the NW margin of the terrane.

Tremadoc–Llanvirn series (Tremadocian–mid Darriwilian stages: Skiddaw, Manx and Ribband groups)

The Skiddaw Group crops out in several inliers in northern England (Cooper *et al*. 1995, 2004) and is divided into two discrete successions by the WSW–ENE trending Causey Pike Fault (CPF; Figs 2, 9), which traverses the Skiddaw Group outcrop in the Lake District and extends to Cross Fell in the Pennines. Contrasting styles of sedimentation, opposing slump directions and differences in provenance maturity either side of the CPF suggest that the fault partitioned the Skiddaw Group basin.

Biostratigraphical dating and correlation is based on graptolites and acritarchs (Cooper *et al*. 1995, 2004). North of the CPF, graptolites define a succession of zones from the *murrayi* Biozone to the *artus* Biozone (Fig. 9). An acritarch zonation developed north of the CPF is correlated with the graptolite zonal scheme. Chitinozoans have also been recorded from the Watch Hill Formation (Amberg *et al*. 2017) and form a local *symmetrica-destombesi* assemblage in the upper Tremadocian. Sparse trilobites, mainly bathypelagic cyclopygids and atheloptic forms, are also present (Fortey *et al*. 1989; Cooper *et al*. 1995, 2004). The trilobites have Gondwanan affinities and are consistent with a deep-marine basinal setting. South of the CPF, acritarchs, graptolites and trilobites have been instrumental in establishing the ages of olistoliths in the Buttermere Formation olistostrome.

The succession north of the CPF, about 5 km thick, comprises turbidite mudstone with thin turbidite sandstones, hemipelagite mudstone and sporadic pebbly mudstone (Cooper *et al.* 1995, 2004). Bioturbation is common at some levels. More substantial sandstone packages form the Watch Hill and Loweswater formations (Fig. 9). Sedimentary breccias associated with slump deposits are present in the upper part of the succession, at about the same level as the Buttermere Formation olistostrome south of the CPF (Fig. 9). The lowest mapped unit (Bitter Beck Formation) is placed in the *murrayi* graptolite Biozone (late Tremadocian) with a faulted base. There is no evidence for older Tremadocian rocks in the succession north of the CPF, although there is some evidence to suggest that deposition might have been ongoing since at least the late Cambrian (Millward and Molyneux 1992; Cooper *et al*. 2004).

The Buttermere Formation, at least 1.5 km thick, is at the base of the succession south of the CPF (Fig. 9). It contains olistoliths of dark grey mudstone with pale grey siltstone and sandstone laminae, homogenous dark grey mudstone, and sandstone, all set in an argillaceous matrix. Olistoliths and matrix are intensely deformed by minor folds and shears. Olistoliths vary in size from granules to blocks up to a kilometre or more across, and range in age from Tremadoc to Arenig (Tremadocian to late Dapingian) (*gibberulus* Biozone). Olistostrome emplacement is inferred to have taken place at about the *gibberulus*–*cucullus* biozone boundary (Cooper *et al*. 1995, 2004). The overlying Tarn Moor Formation (Cooper *et al*. 1995, 2004), at least 1 km thick, comprises siltstone and mudstone of late Arenig–Llanvirn (Darriwilian) age (*cucullus*–*murchisoni* graptolite biozones; Fig. 9), sparsely bioturbated in places. Volcaniclastic turbidite sandstone and bentonites in the middle of the formation provide the first indications of volcanic activity in the Skiddaw Group, which seems either to have been deposited at some distance from major volcanic centres or to have only received material from relatively small-scale and localised eruptions (Cooper *et al*. 2004).

The Skiddaw Group sandstone is considered to have been derived mainly from an old inactive volcanic arc founded on continental crust, but whereas there is increased volcaniclastic input in younger strata, Nd isotope data show a general decrease upwards of juvenile component (Cooper *et al*. 1995, 2004; Stone and Evans 2002). This suggests that the upwards increase in volcaniclastic input is countered by increasing maturity of provenance as a whole (Cooper *et al*. 2004).

The Manx Group is similar in age and facies to the Skiddaw Group, comprising laminated turbidite mudstone and siltstone with hemipelagite deposits and some typically fine-grained sandstone units and pebbly mudstones. The succession is repeated across the island in a series of SW–NE trending fold and thrust belts (Chadwick *et al*. 2001); representative sections are shown in Figure 9 NEW. Petrographical and geochemical data are consistent in showing that the Manx Group was derived mainly from mature sedimentary successions and cratonic quartzose basement, but with a minor volcanic component (Burnett and Quirk 2001).

The oldest parts of the succession in the NW and along the east coast of the island (Glen Dhoo and Lonan formations respectively) are similar in age and facies (Woodcock *et al*. 1999; Chadwick *et al*. 2001; Millward and Stone 2012). They comprise pale to dark grey mudstones with variable proportions of siltstone laminae and sandstone, deposited from turbidity currents. The Lonan Formation is overlain locally by quartz arenite turbidity current deposits (Creg Agneash and Mull Hill formations). The Creg Agneash Formation passes upwards into dark grey mudstone with laminae and thin beds of pale grey siltstone (Maughold Formation) deposited from hemipelagic fallout and low-concentration turbidity currents (Woodcock *et al*. 1999; Chadwick *et al*. 2001; Millward and Stone 2012). This facies also contains pebbly mudstone and is bioturbated. Similar facies occur in the centre and NW of the island (Injebreck, Creggan Moar and Lady Port formations). However, both the Creggan Moar and Lady Port formations also include thin beds of manganiferous ironstone, and the Lady Port Formation contains major olistostrome and pebbly mudstone deposits, interpreted as the product of deposition in a mudstone-dominated deep marine basin subject to slumping and debris flows (Woodcock *et al*. 1999; Chadwick *et al*. 2001; Millward and Stone 2012).

Biostratigraphical dating of the Manx Group relies on acritarchs, which enable correlation with the zones in the Skiddaw Group (Molyneux 1999; Chadwick *et al*. 2001). The Glen Dhoo and Lonan formations are dated as late Tremadoc to early Arenig (late Tremadocian to early Floian), the Creggan Moar Formation as Arenig (late Floian to early Darriwilian), and the Lady Port Formation as late Arenig (early Darriwilian). The Lady Port Formation is structurally isolated with faulted contacts, is internally complex and is resedimented, so the early Darriwilian age might be considered a maximum (Chadwick *et al*. 2001). Assemblages from other formations are not diagnostic of age, although they are similar to those from Floian parts of the Skiddaw Group. Poorly preserved graptolites from one locality in the Lonan Formation indicate an Arenig age only, and material from a locality now placed in the Injebreck Formation could be Tremadoc or Arenig (Rushton 1993a).

Unlike the Skiddaw Group, the Manx Group contains evidence for Arenig (Floian and possibly Dapingian) volcanic activity. An arc-related origin has been suggested (Fritschle *et al*. 2018) for andesitic tuff and volcanic breccia in the Glen Dhoo Formation near Peel on the west coast of the island, dated palynologically as early Arenig (Floian) (Molyneux 1999), and a volcaniclastic deposit identified in the Maughold Formation yielded a mid Arenig (Floian) age of 472.7 ± 2.8 Ma (Fritschle *et al*. 2018). A further outcrop of andesitic tuff approximately 4.5km SW of the Peel locality (at Ballaquane Farm; see Chadwick *et al*. 2001 for the position of both occurrences) is probably within either the Glen Dhoo Formation or the Creggan Moar Formation (McConnell *et al*. 1999; Chadwick *et al*. 2001).

Tracts 1 and 2 of the Ribband Group (McConnell *et al*. 1999 and see Monian Terrane above) are both NW of the Wicklow Fault Zone and are respectively NW and SE of the Leinster Granite. In tract 1, thinly bedded turbidite greywacke siltstone, slate and quartzite (Aghfarrell Formation) are overlain by dark slate and schist (Butter Mountain Formation). The latter formation contains a distinctive package of coticule (spessartine-bearing quartzite) and tourmalinite that has been attributed to an exhalative event in the Caledonian-Appalachian Orogen (Kennan and Kennedy 1983). The coticule package can be traced into the dark slates and schists of the Maulin Formation in tract 2 (McConnell *et al*. 1999). Kennan and Morris (1999) postulated that the manganiferous ironstone beds of the Creggan Moar and Lady Port formations on the Isle of Man were a low-grade coticule protolith produced by the same event (cf. also the Dapingian–Darriwilian coticule package in New Brunswick and New England: van Staal *et al*. 2021). From this, McConnell *et al*. (1999) proposed a correlation of the Butter Mountain and Maulin formations with the Lady Port, Creggan Moar, Injebreck and Maughold formations, and the more pelitic base of the Butter Mountain Formation with the Barrule and Glen Rushen formations (Fig. 9). They further proposed correlation of the Aghfarrell Formation and the Ballybeg Greywackes (beneath the Maulin Formation) with the Lonan Formation. Arc-related volcanic rocks in the Aghfarrell (Dowery Hill basalts), Butter Mountain (Donard andesites) and Maulin formations (Kilcarry andesites) were considered to be possible equivalents of the arc-related rocks in the Manx Group. There are no biostratigraphical data from these rocks. No equivalent manganiferous ironstones, coticule or Lower Ordovician arc-related volcanic rocks are known from the Skiddaw Group. Tourmaline veins are present in the Kirk Stile Formation of the Skiddaw Group in the Crummock Water aureole but very little bedded tourmalinite has been recorded (Cooper *et al*. 2004).

Rocks similar to those of the Ribband Group occur on the Dingle Peninsula of SW Ireland, and in inliers at Kildare and Herbertstown. The Annascaul Formation (*sensu* Todd *et al*. 2000) on the Dingle Peninsula is dominated by green, grey and purple mudstone and slate with subordinate turbiditic and slumped sandstone and conglomerate and includes a sedimentary mélange (Bealacoon Member: Todd *et al*. 2000; Geological Survey of Ireland 2022). Acritarchs indicate a Tremadoc–mid Arenig (Tremadocian–Floian) age, including an Arenig (Floian) age for the mélange (Todd *et al*. 2000; Molyneux *et al*. 2007). At Kildare, dark green-grey shales of the Conlanstown Formation have yielded a graptolite fauna of Llanvirn age (Darriwilian, Dw2, *artus* Biozone: Parkes and Palmer 1994), and unfossiliferous banded red, green and grey mudstone and siltstone (Fournocks and Snowtown formations) are placed in the lower Ordovician on structural grounds (Murphy 1987; Harper and Parkes 2000).

Caradoc Series (Sandbian–lower Katian stages: Eycott and Borrowdale Volcanic groups)

The Skiddaw Group is overlain unconformably by volcanic successions comprising the Eycott and Borrowdale volcanic groups (Fig. 9). Volcanic rocks of equivalent age also occur in the inliers along the NW margin of the terrane in Ireland. There are no Ordovician sedimentary or volcanic rocks overlying the Manx Group or the Ribband Group of tracts 1 and 2, but the age of the Dhoon Granite on the Isle of Man (457.2 ± 1.2 Ma: Fritschle *et al*. 2018; Fig. 9) suggests that it is related to the magmatic event responsible for the Eycott and Borrowdale volcanic groups.

The Eycott and Borrowdale volcanic groups comprise thick, subaerial, subduction-related volcanic successions that aggraded in opposing half-graben, respectively north and south of the Skiddaw Group outcrop (Millward 2002, 2004). Uplift attributed to mantle hydration, thermal expansion and melt generation resulted in a major regional unconformity between the Skiddaw and the volcanic groups, with an estimated removal of at least 2km and perhaps as much as 5km of Skiddaw Group strata (Branney and Soper 1988; Hughes *et al*. 1993). Given their origin as the products of subaerial volcanism, the preservation of both groups is attributed to subsidence resulting from both extensional tectonics and the local movement of magma to the surface keeping pace with the emplacement of new material (Millward 2002).

The Borrowdale Volcanic Group (BVG), more than 6km thick, comprises a succession of basaltic to rhyolitic lavas, sills and pyroclastic rocks of calc-alkaline continental margin type. Following initial phreatomagmatic eruptions, the lower BVG is dominated by andesite lavas produced by low-profile volcanoes. The upper BVG consists of the products of a more paroxysmal phase with the emplacement of widespread sheets of densely welded silicic ignimbrite in at least five depositional centres, with the development of two major silicic calderas (Millward 2002, 2004).

Eruption of the BVG is constrained stratigraphically between the youngest Skiddaw Group strata south of the CPF, which are assigned to the Llanvirn *murchisoni* graptolite Biozone (Darriwilian) and the oldest rocks of the overlying Dent Group, there correlated with the Cautleyan Stage of the Ashgill Series (Katian Stage Slice Ka4; McNamara 1979; Millward and Stone 2012). It is thought, however, to have taken place over a period of less than 5 Ma within that interval (Piper *et al*. 1997; Millward 2002). U–Pb zircon ages from tuffs in the BVG and from associated intrusions cluster between *c.* 453 and 450 Ma (Hughes *et al*. 1996; Millward and Evans 2003) in the Caradoc (latest Sandbian–early Katian) (Fig. 9). Intrusion of granitic bodies is judged to have taken place late in the volcanic cycle (Millward 2004) and consequently a Caradoc (Sandbian–early Katian) age is inferred for the volcanic succession.

Between major silicic eruptions, tephra deposits of the upper BVG were reworked and deposited in basins in the volcanic field, interpreted as fluvio-lacustrine basins developed through regional extensional faulting and caldera collapse (Millward 2004). Arthropod trace fossils were described from one such deposit by Johnson *et al*. (1994) and their origins in a lacustrine versus a littoral environment have been the subject of debate (Shillito and Davies 2019a, b; Briggs *et al*. 2019). Marine acritarchs recorded from a dark grey mudstone in the BVG (Holehouse Gill Formation) are all now thought to be reworked from the Skiddaw Group, however, together with the mudstone in which they occur (Millward 2002).

The Eycott Volcanic Group (EVG) is a succession of basaltic, andesitic and dacitic lavas and sills up to 3.2 km thick with subordinate pyroclastic rocks. It resembles the lower part of the BVG (Millward *et al*. 2000; Millward 2002). The maximum age of the Eycott Volcanic Group is poorly constrained biostratigraphically to no older than Llanvirn (mid-Darriwilian), given the *artus* Biozone age of the youngest underlying Skiddaw Group rocks (Millward and Molyneux 1992), but a U-Pb zircon date of 452.4 ± 4.1 Ma (Millward and Evans 2003) from the co-genetic Carrock Fell Complex (layered gabbro-microgranite: Millward 2002) provides a minimum age. The minimum age is further constrained by the mid-Caradoc (Longvillian, late Burrellian, Sandbian Sa2) age of the Drygill Shale Formation of the Dent Group (Rushton and Ingham 2000), which is in faulted contact with but post-dates the EVG (Millward 2002). Given the comparison of the EVG with the lower BVG, a Caradoc (Sandbian) age is inferred (Fig. 9).

Andesites, basalts, tuffs and interbedded sedimentary rocks of Caradoc (Sandbian–Katian) age occur in inliers from Lambay Island to Kildare in eastern Ireland and are assigned to the Lambay Belt, interpreted to represent arc magmatism (McConnell 2000; McConnell *et al*. 2015; Fritschle *et al*. 2018). Mudstone and siltstone with interbedded andesite sheets at Herbertstown (Clashford House Formation) contain a shelly fauna of probable Soudleyan (mid Burrellian, Sandbian, Sa2) age; at Portrane, the succession of andesites overlain by volcaniclastic rocks (Portrane Volcanic Formation) is Burrellian to Pusgillian (Sa2 to Katian, Ka3) in age; a similar succession at Balbriggan (Belcamp Formation) is also of Sandbian, Sa2 to Katian, Ka1, age; at Kildare, andesites of the Allen Andesite Formation are bracketed by Sandbian, Sa2, shelly faunas in sedimentary successions below (Grange Cottage Formation) and above (Grange Hill Formation) (Harper and Parkes 2000).

Upper Caradoc–Ashgill series (upper Katian–Hirnantian stages: Dent Group)

Rocks of Katian and Hirnantian age are represented mainly by the Dent Group of northern England, and by formations in inliers close to the NW boundary of the terrane in eastern Ireland (Portrane, Balbriggan, Kildare: Harper and Parkes 2000). The Dent Group (Kneller *et al*. 1994) comprises shallow marine clastic lithologies and limestone with some pyroclastic rocks and rhyolitic lava in the Lake District of NW England and in the Cross Fell and Cautley-Dent inliers. These rocks were deposited when cooling of the granites emplaced during the later Borrowdale Volcanic episode (Millward 2004) led to thermal subsidence, enabling transgression and the establishment of shallow marine shelf conditions. The entire Dent Group and its constituent formations were reviewed most recently by Millward and Stone (2012), who provide details of the succession in the various outcrops, and the Dent Group in the Cautley-Dent inliers was included in a revision of the parent Windermere Supergroup by Rickards and Woodcock (2005).

The Dent Group is interpreted in terms of four depositional cycles (Kneller *et al*. 1994; Rickards and Woodcock 2005; Millward and Stone 2012). The cycles are most evident across the southern Lake District, where deposition took place in a shallower setting than elsewhere and where each cycle is separated by a period of emergence and erosion. There, the first cycle, consisting of grey fossiliferous calcareous siltstone with nodular limestone, thin units of pebbly coarse-grained sandstone and a basal conglomerate (Stile End Formation), is unconformable on and fills irregularities in the surface of the Borrowdale Volcanic Group. Sedimentation commenced in the Cautleyan (Katian, Ka4; Shelly Zone 2 of Ingham 1966; Millward and Stone 2012; Fig. 9). The second cycle, with the Kirkley Bank Limestone Formation at its base (Fig. 9), oversteps the rocks of cycle 1 to rest directly on rocks of the Borrowdale Volcanic Group and is also dated as Cautleyan (zones 2 and 3 of Ingham 1966; McNamara 1979; Millward and Stone 2012). A nodular and bioturbated micritic limestone (Broughton Moor Limestone Formation; Fig. 9) with a thin unit of chamositic ooids at its base is also placed in the second cycle, the ooidal unit being considered to represent the maximum marine transgression and the limestone a condensed deposit. A trilobite fauna recorded locally indicates Rawtheyan zones 5 or 6 of Ingham (1966; McNamara 1979; Millward and Stone 2012). The third cycle comprises rhyolitic tuff resedimented by gravity flows (Appletreeworth Volcanic Formation) of limited and stratigraphically restricted extent across the southern Lake District, also assigned to Ingham's (1966) Zone 6 (McNamara 1979; Kneller *et al*. 1994).

More continuous deposition during the first three cycles is evident in successions on the northern and eastern fringes of the Lake District volcanic field, including those in the Cross Fell and Cautley-Dent inliers. Sedimentation started earlier in the northern Lake District and at Cross Fell in the Longvillian Substage (late Burrellian, late Sandbian–earliest Katian), with grey to black calcareous mudstone and siltstone containing an abundant shelly fauna in the northern Lake District (Drygill Mudstone Formation – see above) and partly calcareous siltstone and mudstone with silty limestone in the Cross Fell inlier (Dufton Mudstone Formation). A basal sandier facies ('Corona' Beds) of the Dufton Mudstone Formation includes debris from the Borrowdale Volcanic Group. An unconformity separates the Dufton Mudstone Formation from younger beds, the strata immediately below the unconformity ranging from Pusgillian to early Cautleyan in age (Owen and Rushton 2000). Grey nodular limestone with pale grey mudstone above the unconformity (Swindale Limestone Formation) is possibly equivalent to the Broughton Moor Limestone Formation and is either mid-Rawtheyan or at about the Cautleyan–Rawtheyan boundary. In the Cautley-Dent inliers, massive and bioturbated calcareous mudstone with calcareous nodules, impersistent limestone beds and some sandstone beds (Cautley Mudstone Formation) extend from at least the upper Streffordian (Onnian, mid Katian) to the upper Rawtheyan (Ka4). Rhyolitic volcaniclastic rocks (Cautley Volcanic Member) occur in the upper part.

An upper Rawtheyan unconformity (upper Ka4) separates the Cautley Mudstone Formation from the Ash Gill Mudstone Formation, a unit of grey mudstone that is distinguished by its lower carbonate content and is present above an unconformity throughout the Dent Group outcrop. The Ash Gill Mudstone Formation, so named after the locality of Ash Gill by Millward and Stone (2012) to distinguish the lithostratigraphical unit from the chronostratigraphical Ashgill Series, represents the fourth of the depositional cycles of the Dent Group (Kneller *et al*. 1994). It is late Katian and Hirnantian in age and contains a *Hirnantia* brachiopod fauna (Owen and Rushton 2000). An upwards reduction in carbonate content and bioturbation in the Ash Gill Formation has been interpreted as a response to increased clastic input and cooling sea water prior to the Hirnantian glaciation (Rickards and Woodcock 2005). Sandstones and conglomerates occur at a high level in the Ash Gill Formation across its outcrop and are presumed to represent lowstand deposits (Rickards and Woodcock 2005).

Biostratigraphical zonation and correlation of the Dent Group was based initially on its shelly, mainly trilobite faunas (Ingham 1966; McNamara 1979), but more recent attention has focussed on graptolites (Rickards 2002), conodonts (Bergström and Ferretti 2015), chitinozoans (Vandenbroucke *et al*. 2005) and ostracods (Williams *et al*. 2001). Rickards (2002) reviewed and reported on previous and new graptolite collections, but the faunas need further revision.

A succession of chitinozoan biozones in the Dufton and Cautley mudstone formations of the Cross Fell and Cautley-Dent inliers (Vandenbroucke *et al*. 2005) are correlated with zones in Baltoscandia, Avalonia and Gondwana. The lower three zones of *Fungochitina spinifera*, *Tanuchitina bergstroemi*? and *Conochitina rugata* were originally established in Baltoscandia (Nõlvak and Grahn 1993). Correlations since then have generally placed the base of the Ashgill Series in the *bergstroemi* Biozone, but Vandenbroucke *et al*. (2005) showed that it correlated with a level in the *spinifera* Biozone (Fig. 9).

Conodonts from the Keisley Limestone, a carbonate mud mound in the Cross Fell Inlier, were reviewed and reported on by Bergström and Ferretti (2015). These authors also drew lithological and faunal comparisons with the Kildare Limestone Formation of eastern Ireland and the Boda Limestone of Sweden and carried out a preliminary investigation of carbon isotope stratigraphy. They concluded that the conodont fauna indicated the *Amorphognathus ordovicicus* Biozone and was of late Katian (Ka4) age, and that the $\delta^{13}C_{carb}$ values obtained indicated a level below the Hirnantian Isotopic Carbon Excursion (HICE) for the sampled interval. A variant of the *Hirnantia* fauna that might have inhabited deep-water environments (Rong *et al*. 2020) occurs in thin limestone bands overlying the massive Keisley Limestone mud mound. The brachiopods and associated trilobites from this level were included by Temple (1968, 1969) in the Silurian lower Llandovery Series but are now placed firmly in the Hirnantian Stage (Ingham and Wright 1972; Wright 1988; Owen and Rushton 2000). The carbonate beds are overlain by siltstones containing graptolites of the *persculptus* Biozone (Owen and Rushton 2000 and references therein; Bergström and Ferretti 2015).

In the southern Lake District and Cautley-Dent area, the Ash Gill Formation is overlain by the Spengill Mudstone Member at the base of the Skelgill Mudstone Formation (Stockdale Group). The member comprises nodular limestone with pale grey, pyritous mudstone (Kneller *et al*. 1994 and references therein; Rickards and Woodcock 2005; Millward and Stone 2012). It represents a condensed carbonate sediment with a low-diversity, deep-water shelly benthos and a *persculptus* Biozone graptolite fauna (Kneller *et al*. 1994; Rickards and Woodcock 2005; Millward and Stone 2012). The Spengill Member passes up into black graptolitic mudstone of the Skelgill Formation of early Silurian age, widely interpreted as being deposited in anaerobic bottom waters during the transgression that followed the Hirnantian glaciation (e.g. Rickards and Woodcock 2005). This unit contains a relict deep-water *Hirnantia* brachiopod fauna (Harper and Williams 2002).

Rocks of equivalent age to the Dent Group are present in inliers at Portrane, Balbriggan and Kildare in eastern Ireland (Harper and Parkes 2000). Shelly faunas and conodonts indicate a Cautleyan age, equivalent to the basal Ka4 stage slice of the Katian Stage, for the Portrane Limestone Formation, deposited in a shallow water setting, possibly on the slope of a volcanic island (Ferretti *et al*. 2014b and references therein). The unit contains a diverse silicified shelly fauna dominated by brachiopods, mainly non-articulates and orthides (Wright *et al*. 2022). Correlation and comparison of the Kildare Limestone Formation and its putative Rawtheyan (late Katian) age have been noted by several authors (e.g. Parkes and Palmer 1994; Bergström and Ferretti 2015). The Kildare Limestone is overlain by a mudstone that has yielded a *Hirnantia* fauna (Wright 1968; Parkes and Palmer 1994). Black shales in a fault block at Balbriggan have yielded graptolites of the *anceps* Biozone (Katian, Ka4; Rickards *et al*. 1973).

Irish subterranes along the Iapetus Suture: Bellewstown and Grangegeeth

Bellewstown

The Ordovician succession at Bellewstown comprises the Prioryland, Hilltown and Carnes formations in upward succession, the last with the Bellewstown Member at its base (Harper and Parkes 2000; Fig. 10). This succession contrasts with that of the adjacent Grangegeeth Subterrane and has no obvious counterpart in Britain.

The Prioryland Formation consists of laminated grey, purple and buff-coloured mudrocks and siltstones with slump breccias and coticule material (Harper and Rast 1964; Murphy 1987; Murphy *et al.* 1991; Kennan and Murphy 1993; McConnell *et al*. 2015). Acritarchs of possible Tremadocian age have been reported from the lower part of the formation (McKee 1976). The nature of the contact between the Prioryland Formation and the overlying Hilltown Formation is uncertain.

The Hilltown Formation is mainly volcanogenic and contains an early Llanvirn (Darriwilian) brachiopod fauna related to the Celtic Province (Harper *et al.* 1990; Harper *et al.* 2013); *D. artus* Biozone graptolites have been reported from associated siltstones (Harper and Rast 1964). Zircons separated from sandstone within the volcanogenic breccia, however, have yielded a Floian age of *c.* 474 Ma, which McConnell *et al*. (2015) interpreted as the age of volcanism at Bellewstown. Graptolites assigned to the *Nicholsonograptus fasciculatus* Biozone of Maletz (1997) have been obtained from cleaved dark grey mudrocks more than 24 m above the volcanogenic rocks. The fauna is considered to be no older than the upper part of the British *D. artus* Biozone (McConnell *et al*. 2015). Middle Ordovician volcanism in the Bellewstown Subterrane is inferred to have ceased after emplacement of the volcaniclastic deposits at *c.* 474 Ma (McConnell *et al*. 2015).

The Hilltown Formation is overlain by a unit of calcareous sandstone and bioclastic limestone (Bellewstown Member of the Carnes Formation). The contact with the Hilltown Formation was described by Harper and Rast (1964) as conformable but the precise age of the Bellewstown Member remains uncertain. It may represent a condensed unit within the upper Llanvirn, an interpretation supported by conodont data (Bergström and Orchard 1985). The Bellewstown Member is overlain in the Carnes Formation by greywackes, tuffaceous shales and dark mudstones (Harper and Rast 1964; Murphy 1987; McConnell *et al*. 2015). The volcanic lithologies signal renewed volcanism in the Bellewstown Subterrane during the Sandbian. A *Nemagraptus gracilis* Biozone graptolite fauna is present within shales of the formation and a middle Caradoc shelly fauna with an Anglo-Welsh character occurs higher in the succession (Brenchley *et al*. 1977). The shelly fauna contains the brachiopod *Aegiromena* (Harper and Rast 1964), which is only known from Gondwana and related terranes at this time (Colmenar *et al*. 2022), a palaeogeographical position confirmed by McConnell *et al*. (2015).

The upper part of the succession at Bellewstown is comparable in some respects to successions farther south in Ireland. Upper Llanvirn carbonate deposits at Bellewstown and Tramore, and the slightly younger lower Sandbian carbonates at Courtown, pass upwards into Upper Ordovician volcanic successions. There, however, the carbonate deposits are unconformable on rocks of the Ribband Group (see above), with much of the Llanvirn and therefore equivalents of the upper part of the Hilltown Formation missing. In contrast, the fine-grained Conlanstown Formation at Kildare is correlated with the *artus* Biozone.

Grangegeeth

The Grangegeeth succession includes, in ascending order, the Slane, Grangegeeth and Melifont Abbey groups (Fig. 10), ranging in age from earliest Llanvirn to mid Ashgill (Darriwilian to Katian). An equivalent succession has not been identified in northern England or Scotland. The Slane Group is mainly volcanogenic, with shales within the Hill of Slane Formation. Graptolites of the Llanvirn (Darriwilian) *Didymograptus artus* Biozone are present in the Hill of Slane Formation while the overlying White Island Bridge Formation correlates with both the *D. artus* and *D. murchisoni* biozones (Harper and Parkes 2000). The graptolite faunas have a peri-Gondwanan affinity (Owen *et al.* 1992). This provincial signal has been challenged, however, by zircon data from the White Island Bridge Formation, which are characteristic of Laurentian crust or peri-Laurentian metasedimentary successions (McConnell *et al.* 2010).

The middle Caradoc (Sandbian) Grangegeeth Group has abundant and diverse shelly faunas of Scoto-Appalachian Laurentian-marginal affinity within the tuffaceous sandstone and shale of the Knockerk Formation, including brachiopods, bivalves, bryozoans, cephalopods, corals, and echinoderm debris (Harper 1952; Romano 1980; Owen *et al.* 1992; Romano and Owen 1993); the Fieldstown Formation (olive to grey mudstone and tuff) contains a depauperate graptolite fauna that is undiagnostic of age (Brenchley *et al*. 1977). Acritarchs considered to be of Arenig (Floian–early Dapingian) age and reworked into the Knockerk Formation include typical peri-Gondwanan genera (Smith 1981; Molyneux *et al*. 2013).

The Broomfield Formation (black shale with chert) at the base of the Melifont Abbey Group contains graptolites of the *D. clingani* and *P. linearis* biozones, whereas the overlying Oriel Brook Formation (grey pyritic mudstone) is a deep-water facies that contains cyclopygid trilobites (Romano 1980;

Owen and Romano 2011) and a *Foliomena* brachiopod fauna (Harper and Mitchell 1982; Rong *et al.* 1999). The faunal data suggest an origin in higher latitudes for this subterrane, moving to lower latitudes during the Ordovician and subsequently supporting deep-water facies not unlike the later Ordovician environments of the Girvan district, SW Scotland (Harper 2001).

Southern Uplands and Longford-Down

Ordovician rocks of the Southern Uplands of Scotland and the Longford-Down massif along strike in Ireland form part of an accretionary thrust complex that developed above a northward-directed subduction zone. The complex is commonly referred to as an accretionary prism, following the introduction of the concept by McKerrow *et al*. (1977) and Leggett *et al*. (1979). The structure of the accretionary complex means that Ordovician rocks crop out in the northernmost NE–SW striking fault-bounded tracts that represent successive accreted packages (Northern Belt of the Southern Uplands) and at the base of successions in subsequently accreted tracts (Central Belt).

The successions in each tract comprise pelagic mudstone and chert overlain by turbidite units. As subduction proceeded and progressively younger rocks approached the subduction zone, the minimum age of pelagic mudstone in each successive tract became younger (Fig. 11), as did the age of the turbidite beds deposited upon it. The earliest accreted tracts, which are in the north of the terrane, therefore comprise only the older units. Younger tracts were thrust beneath and structurally rotated to the vertical or beyond to build up the accretionary thrust complex, giving rise to the Southern Uplands outcrop pattern of elongated, NE–SW-striking tracts delineated by major strike-parallel faults (originally accretionary thrusts). The successive tracts therefore become younger to the south whereas the effect of structural rotation is such that, within each tract, the sense of younging is to the north. Despite challenges, an accretionary origin for the Southern Uplands Terrane is now the consensus (Stone and Merriman 2004; Stone 2014b). The principal alternative interpretation regards the Southern Uplands' Ordovician succession as having been deposited in an extensional, continental forearc environment (Armstrong *et al*. 1996; Owen *et al*. 1999).

The Crawford and Moffat Shale groups are at the base of the succession. The Crawford Group (mudstones, cherts and basaltic lavas) comprises the oldest rocks in the Southern Uplands Terrane and represents the oceanic substrate on which pelagic mudstones of the Moffat Shale Group were deposited (Fig. 11). The lavas are associated locally with blue-grey radiolarian chert and brown mudstone (Raven Gill Formation) for which conodonts indicate a probable Arenig (Floian–Dapingian) age (Armstrong *et al.* 1990). A second unit of radiolarian chert (Kirkton Formation) has yielded conodonts of the *anserinus* Biozone (Armstrong *et al.* 1996), which spans the Llanvirn–early Caradoc (Darriwilian–early Sandbian) interval. A mixture of within-plate lava types and tholeiitic basalts of possible mid-ocean affinity has been reported in association with both formations by Phillips *et al.* (1995) and Barnes *et al.* (1995).

The Moffat Shale Group comprises four formations of graptolitic shales, mudstones and siltstones, with cherts, bentonites and greywackes, and represents distal turbidite as well as hemipelagite deposition. The Glenkin Shale and Lower Hartfell Shale formations are present in the Northern and Central belts, but the Upper Hartfell Shale and Birkhill Shale formations are only known from the

Central Belt. Only the Glenkiln Shale Formation is present in northernmost tracts of the Northern Belt, the occurrence of the pelagic mudstone units illustrating the point made above regarding the youngest mudstones becoming successively younger southwards (Fig. 11).

The Glenkiln Shale Formation overlies the Kirkton Formation and comprises black cherty shales of the Caradoc (Sandbian) *gracilis* and lower part of the *bicornis* graptolite biozones, separated and succeeded by grey shales. The Lower Hartfell Shale Formation consists of dark grey and black shales with cherty bands and ranges from the upper part of the *bicornis* graptolite Biozone to the *linearis* Biozone (Caradoc to Ashgill). The Upper Hartfell Shale Formation is composed largely of unfossiliferous grey shales with dark graptolitic bands of the Ashgill (Katian) *complanatus*, *anceps* and (Hirnantian) *extraordinarius* biozones. The overlying Birkhill Shale Formation extends into the Llandovery Series and includes the Global Boundary Stratotype Section and Point (GSSP) for the base of the Silurian System at Dob's Linn, Moffatdale (Williams 1988; Rong *et al*. 2008; Fig. 11). The lowest dark grey muddy siltstones and black shales of the formation are in the *persculptus* Biozone. The predominance of grey rather than black mudstones in the Upper Hartfell Shale Formation has been ascribed to the cooling effect of the Hirnantian glaciation (e.g. Finlay *et al*. 2010; see also Armstrong and Coe 1997). A trace element geochemical study at Dob's Linn, however, has favoured the warming effect of large-scale volcanicity with resulting anoxia as the primary cause of the lithological variation (Bond and Grasby 2020).

Numerous thin layers of volcanic ash (metabentonites) are interbedded within the mudstone succession and demonstrate contemporaneous volcanicity (Merriman and Roberts 1990). Radiometric dates reported from zircon in the metabentonite (Tucker *et al*. 1990) include a U–Pb age of 445.7 ± 2.4 Ma from the *anceps* biozone, revised to 448.88 ± 1.17 Ma by Melchin *et al*. (2012, p. 546).

Ordovician turbidite units in the Northern Belt are included in three groups (Tappins, Barrhill and Scaur) of the Leadhills Supergroup (Floyd 2001; Stone 2014b), which overlie the Moffat Shale Group. The Tappins Group crops out in the most northerly structural tracts (Fig. 11 NEW: Downan Point Lava Formation and Marchburn Formation). Limited biostratigraphical evidence (graptolite and conodont: Floyd 2001) indicates a Llanvirn–Caradoc (Darriwilian–Sandbian) age but underlying Moffat Shale Group strata are not present. The succession comprises red and green mudstone, chert, turbidite sandstone and conglomerate. Most of the sandstones were derived from the north and NW and contain a high proportion of igneous detritus, much of it likely to have been derived from ophiolitic rocks similar to those forming the Ballantrae Complex. The conglomerates form stacked, lenticular bodies interpreted by Kelling *et al*. (1987) as proximal or inner-fan features deposited in laterally migrating channels with a minimum width of 2–3 km.

At the northern margin of the Southern Uplands Terrane in SW Scotland, the Downan Point Lava Formation comprises tholeiitic and pillowed basalt lava with a geochemistry indicative of within-plate, oceanic island eruption (Thirlwall and Bluck 1984). It was thought to be part of the Early to Middle Ordovician Ballantrae Complex, but the apparent interbedding of similar pillow lavas with the Tappins Group raised the possibility that the Downan Point lavas were younger. The Downan Point Formation is now included in the Tappins Group in one of the oldest, northernmost tracts of the Southern Uplands accretionary complex (Floyd 2001).

Southwards, thick turbidite sandstone successions with ages that range through the Caradoc (Sandbian and lower Katian) and into the Ashgill comprise the Barrhill and Scaur groups (Fig. 11). Variations in the compositions of the sandstones in each tract indicate a range of provenance. These compositional differences are not only on a tract scale, but also appear as the interbedding of differently sourced sandstone units. Within the Barrhill Group, the quartzo-feldspathic wacke succession of the Kirkcolm Formation (Fig. 11) has interbeds of volcaniclastic wacke (Galdenoch Formation) and polymict wacke and conglomerate (Blackcraig Formation). Within the Scaur Group, the Portpatrick Formation comprises volcaniclastic wackes but is interbedded with quartzo-feldspathic wacke assigned to the Glenwhargen Formation (Fig. 11). The Shinnel Formation (Fig. 11) is also mostly quartzo-feldspathic, whereas the Glenlee Formation is more mixed, with quartzose and andesitic components. Turbidites of the Glenlee Formation, in the southernmost Ordovician tract of the SW Southern Uplands, have yielded Ashgill (Katian and possibly Hirnantian) faunas (Fig. 11; Floyd and Rushton 1993).

The largest coherent volcanic assemblage within the Southern Uplands Terrane is the Bail Hill Volcanic Group (Fig. 11) which crops out in the central part of the Leadhills Supergroup. It is a mildly alkaline, oceanic seamount assemblage, a product of within-plate volcanism (Hepworth *et al.* 1982; Phillips *et al.* 1999). It comprises a heterogeneous succession of submarine lavas and volcaniclastic rocks up to 2 km thick, cut by a vent breccia and several minor intrusions. The volcanic rocks range in composition from alkali basalt to trachyandesite and are of Caradoc (Sandbian) age. The oldest part of the volcanic succession rests on Moffat Shale Group mudstone of early Caradoc *gracilis* Biozone age whilst younger parts laterally interdigitate with and overlie turbidite sandstone beds of the Kirkcolm Formation (Barrhill Group), also of *gracilis* Biozone age. Elsewhere, there are sporadic occurrences of mafic lava subjacent to the Marchburn Formation (Tappins Group in the NE Southern Uplands) and either underlying or interbedded with the basal sandstone of the Kirkcolm Formation; for these lavas a Caradoc (Sandbian) age is also likely.

Graptolites provide the primary biostratigraphical control throughout the Southern Uplands (e.g. S.H. Williams 1982a, 1982b, 1983, 1987, 1988, 1994; Rushton 2001; Zalasiewicz *et al*. 2009), but a derived shelly fauna from the Kirkcolm Formation (Kilbucho horizon, Fig. 11) enables correlation with the adjacent Midland Valley Terrane. The shelly fauna (trilobites, gastropods, corals and brachiopods) is contained within mass flow deposits interbedded within the sandstone succession (Clarkson *et al.* 1992). The fauna is of Caradoc (Sandbian or earliest Katian) age (as is the Kirkcolm Formation) and is equivalent to *in situ* shelly faunas known from the southern margin of the Midland Valley Terrane at Girvan (Scotland) and Pomeroy (Northern Ireland). Despite its relative proximity to Girvan, the Kirkcolm Formation fauna compares most closely to that of Pomeroy (Candela and Harper 2010), which has relevance to any debate concerning relative strike-slip movement between the two terranes along the Southern Upland Fault. Whelan (1988) provided a preliminary assessment of acritarch and chitinozoan distributions across the Ordovician–Silurian boundary in the GSSP for the base of the Silurian System at Dob's Linn. Verniers and Vandenbroucke (2006) investigated the chitinozoans from the Silurian GSSP further. They recorded *Armoricochitina reticulifera* in a monospecific assemblage from the upper part of the *Dicranograptus clingani* Biozone, and therefore from the same stratigraphical interval as in SW Wales, and *Ancyrochitina ellisbayensis* across the base of the Silurian, coincident in part with its uppermost Ordovician range in the Ellis Bay Formation on Anticosti Island, Québec (Soufiane and Achab 2000).

The stratigraphy currently applied to the Longford–Down area along strike in Northern Ireland (Anderson 2004) follows that of the Southern Uplands Terrane (Floyd 2001), with the Crawford Group, Moffat Shale Group and Leadhills Supergroup all recognized (Fig. 11). Crawford Group rocks in the Longford-Down area are confined to outcrops at Acton in the Central Belt and along the southern shore of Belfast Lough in the Northern Belt (Anderson 2004). At Acton, a *c.* 5 m succession of purplish red and green mudstone, black chert and pale grey radiolarian chert resembles the Kirkton Formation of the Southern Uplands, with which it is correlated (Fig. 11). A sparse conodont fauna from the mudstone indicates the boundary between the *Pygodus serra* and *P. anserinus* biozones (Bergstrom *in* Anderson 2004) and therefore a late Llanvirn (mid-Llandeilian, late Darriwilian) age. These are the oldest Ordovician rocks in the region below the Moffat Shale Group. On the shore of Belfast Lough, a succession of pillow lavas and conglomerate is succeeded by *gracilis* Biozone mudstones (Glenkiln Shale Formation) of the Moffat Shale Group. All four Moffat Shale Group formations have been recognised and show a pattern of occurrence that replicates that seen in Scotland, i.e., with increasingly younger beds seen beneath turbidite units in a southerly direction (Anderson 2004; Fig. 11).

Occurrences of Moffat Shale Group rocks in County Monaghan previously mapped by the Geological Survey of Ireland (GSI) have been given formation status and dated by graptolite biostratigraphy in successive studies by O'Connor (1975), Morris *et al.* (1986), Rushton (1990b, 1991, 1993b) and Geraghty (1996). Re-mapping by the GSI of Moffat shale horizons in County Monaghan has since been completed using their strong conductivity signal in the airborne geophysical survey of the Tellus Border project (Beamish *et al.* 2010; Cooper *et al.* 2016; Cooper 2021; Young and Donald 2013). Combined with borehole drilling and fieldwork, this has resulted in a revised distribution of Moffat Shale Group rocks and their associated tract-bounding faults (McConnell *et al.* 2016).

The revised GSI stratigraphy comprises alternating Gala Group (Silurian) greywacke formations and intervening Moffat Shale Group packages. The GSI sequence of formations from south to north is: Taghart Mountain Formation, Laragh Formation (MSG), Shercock Formation, Tullyraghan Shale Formation (MSG), Drumagelvin Greywacke Formation, Kehernaghkilly Formation (MSG), Oghill Formation, Corderrybane Shale Formation (MSG), and Lough Avaghon Formation. Entries on the GSI online map viewer (https://www.gsi.ie/en-ie/data-and-maps/Pages/Bedrock.aspx#) assign the Laragh Formation to the *anceps* Biozone, the Kehernaghkilly Formation to the Ashgill (*linearis* Biozone?), and the Corderrybane Shale Formation to the undivided Caradoc Series (O'Connor 1975; Morris *et al*. 1986; Rushton 1990b, 1991, 1993b). Biostratigraphical work has been carried out on graptolite-bearing mudstones, mostly from the Moffat Shale Group (Rushton 2011b, 2014a; Williams and Zalasiewicz 2014).

Midland Valley

The main Ordovician successions in the Midland Valley Terrane are the ophiolitic Ballantrae Complex on the coast of SW Scotland, immediately north of the Southern Uplands Terrane, the Girvan succession of shallow to deep marine sedimentary rocks that overlie the Ballantrae Complex, and their equivalents around Tyrone and Pomeroy in Northern Ireland.

Ballantrae Complex

The Ballantrae Complex consists mainly of serpentinized ultramafic rocks, plutonic intrusions and mafic pillow lavas that originated as Cambro-Ordovician oceanic lithosphere and mantle. The components of this ophiolitic assemblage were juxtaposed tectonically at the margin of the Iapetus Ocean during the Early to Middle Ordovician, then obducted onto the Midland Valley continental block, at that time the leading edge of Laurentia (see Stone 2014a for a review and bibliography). Radiometric dates from a dynamothermal metamorphic aureole adjacent to one of the major serpentinite bodies show that tectonic assembly was in progress by about 477 Ma (around the Tremadocian–Floian boundary): 478 ± 8 Ma (K-Ar, hornblende in amphibolite, Bluck *et al*. 1980) and 477.6 ± 1.9 Ma (Sm-Nd, garnet in amphibolite/granulite, Stewart *et al*. 2017). Two plutonic bodies intruded into serpentinite have given late Cambrian to Tremadocian ages (Bluck *et al*. 1980): 483 ± 4 Ma (U-Pb, zircon in leucotonalite) and 487 ± 8 Ma (K-Ar, gabbro). Radiometric dates obtained from lavas are imprecise, spanning much of the Cambrian and the Early to Middle Ordovician: 501 ± 12 and 476 ± 14 Ma (Sm-Nd, basalt, Thirlwall and Bluck 1984).

The volcanic successions are all included in the Balcreuchan Group (Stone and Smellie 1988; Fig. 12). The dated lavas have the geochemical characteristics of oceanic island arc rocks. Other lavas differ geochemically and probably originated in a within-plate volcanic setting. Whereas the island arc lavas and breccias associated with them are devoid of fine-grained sedimentary interbeds, the within-plate volcanic successions include conglomerates, mudstones and cherts. Locally, these contain graptolite faunas, most of which establish various Arenig (Floian) ages (Stone and Rushton 1983, 2018; Rushton *et al*. 1986; Fig. 12).

One structurally isolated unit of lava and clastic rock contains an Arenig (Dapingian) graptolite fauna (Stone and Rushton 2003; Maletz 2004). A conglomerate interbedded with the graptolitic strata contains pebbles of altered serpentinite with algal coatings. This has been taken as evidence for exposure and erosion of altered ultramafic (mantle) rock during obduction, hence dating that process to the Dapingian. There is conflicting evidence, however, from detrital zircon populations in Balcreuchan Group tuffs and sandstones with distribution peaks as young as 464 Ma (Fujisaki *et al.* 2015), that volcanicity continued into the Llanvirn (Darriwilian), so precluding Dapingian obduction (Stone and Rushton 2018). Nevertheless, the entire Ballantrae Complex had been tectonically emplaced and eroded by the mid-Darriwilian, based on conodonts from a limestone near the base of the unconformably overlying sedimentary succession (Bergström 1990). The age of the basal strata above the unconformity is probably about 463 Ma (Fig. 12), confirming that after a protracted, polyphase history, the final stages of tectonic assembly and obduction of the Ballantrae Complex proceeded rapidly.

Girvan

The Girvan district exposes some of the most complete and informative sections through the Ordovician part of the Midland Valley Terrane. The key importance of the Girvan succession, as recognised by Lapworth (1882), lies in its intercalated graptolite, shelly and latterly its conodont faunas that enable correlation between slope/basinal facies and platform facies. Ingham (2000) provided detailed charts and a glossary of all the stratigraphical units used for Girvan until then and that stratigraphy is adopted here (Fig. 12).

The rocks of the Ballantrae Complex were eroded prior to deposition of a shallow to moderately deep marine succession of conglomerates, sandstones, siltstones and mudstones with minor reefal limestones. The Ordovician units comprise the Barr, Ardmillan and Albany groups, with the Albany Group partially coeval with the Barr Group (Fig. 12). The Ordovician part of the succession ranges in age from Llanvirn to Ashgill (Darriwilian to Hirnantian) and is nearly 4 km thick.

The Girvan succession has been placed in a proximal fore-arc setting modified by listric faulting (Ingham 2000). Deposition was accompanied by movement on broadly east–west-trending faults developed in sequence northwards, with downthrow to the south. Two consequences follow from this: (i) the base of the succession is diachronous, with the oldest beds becoming younger northwards as successive faults were overstepped in the same direction; (ii) shallow marine deposits in the north are laterally equivalent to deeper marine deposits in the south.

Three key sectors, (i) Penwhapple Burn-Stinchar Valley, (ii) Girvan foreshore and adjacent areas and (iii) the Craighead Inlier, demonstrate, from south to north, sequentially overlapping, younger parts of the Ordovician (Fig. 12). The Penwhapple Burn-Stinchar Valley sector ranges in age from early Llanvirn to early Ashgill (Darriwilian–Katian) and exposes the lowest parts of the Girvan succession, notably, in ascending order, the Kirkland Conglomerate Formation, the Confinis Formation, the Stinchar Limestone Formation and the Superstes Mudstone Formation; the Confinis Formation and Stinchar Limestone have rich conodont and shelly faunas, whereas the Superstes Mudstone is graptolitic. Conodonts place the base of the *Pygodus anserinus* Biozone in the Stinchar Limestone Formation (Bergström 1990) and graptolites from the Superstes Mudstone Formation indicate the *gracilis* Biozone (Rushton *et al.* 1996).

These units are capped by the Benan Conglomerate Formation, concluding the rocks assigned to the Barr Group. Farther to the south, the Albany Group containing cyclopygid trilobites and graptolites in its lower part, and graptolites and shelly faunas in its upper part, is an offshore equivalent of the Barr Group (Rushton *et al*. 1996). Within the Girvan foreshore and adjacent areas, a thinner development of the Benan Conglomerates forms the base of the succession, overlain by the Balclatchie Formation, with rich shelly faunas, and its lateral equivalents, together with the Ardwell Farm Formation, comprise the lower parts of the Ardmillan Group (Ardwell Subgroup, lower to middle Caradoc). This succession is capped by the Whitehouse Subgroup (Fig. 12) with mixed graptolitic and shelly faunas, and with cyclopygids and the *Foliomena* fauna and its upper part indicating deep-water facies. The highest part of the Ordovician succession is exposed in the Craighead Inlier and includes the Drummuck Subgroup (Ashgill) of the Ardmillan Group, its summit marked by the Hirnantian High Mains Formation (Fig. 12).

Rapidly deepening sequences in the Barr Group and the Ardwell, Whitehouse and Drummuck subgroups of the Ardmillan Group are generally capped by thick conglomerates or turbidites throughout the succession, cycles first identified by Lapworth (1882). The palaeontological data thus reflect a wide range of water depths, from deep-water graptolite associations to more shallow-water assemblages dominated by shelly faunas such as brachiopods and trilobites, and remain a fertile area for research. Brachiopods are the dominant element of the benthos, described in detail from the Barr and Lower Ardmillan groups by Williams (1962) and the Upper Ardmillan Group by Harper (1984, 1989, 2006), but also noted in shorter papers (e.g. Harper 1979, 1981, 1982; Harper and Owen 1986); their diversity through the succession is reviewed by Harper (2001) and Harper and

Stewart (2008); Candela and Harper (2014) synoptically revised the brachiopods from the Barr and Lower Ardmillan groups. The trilobite faunas from the Barr and lower Ardmillan groups were described in a series of papers by Tripp (e.g. 1955, 1962, 1965, 1967, 1976, 1979, 1980) and those from the Doularg Formation (Albany Group) by Ingham and Tripp (1991). Trilobites from the upper Ardmillan Group have received sporadic modern attention despite their abundance and diversity in the Lady Burn Starfish Beds (Harper 1982; Thomas *et al*. 1984; Owen 1986). Ostracod faunas have been described recently in some detail; their distribution, ecology and taxonomy have been discussed from key parts of the Ordovician succession (Craighead Limestone Formation: Williams and Floyd 2000; Mohibullah *et al*. 2011, 2013). The distribution and palaeoecology of bivalve and gastropod faunas from the Middle and Upper Ordovician rocks have been charted (Stewart 2012), whereas Ebbestad (2008) has redescribed and reassigned the hitherto Gondwanan tergomyan *Carcassonnella*, based on new data from the upper Whitehouse and Drummuck subgroups. The hyoliths from the Upper Ordovician part of the succession have been redescribed by Malinky (2003); a number of species are no longer recognizable, and one is assigned to the Gastropoda. Echinoderms from the Lady Burn Starfish Beds have been described including camerate crinoids from the Lady Burn Starfish Beds (Donovan and Gilmour 2003). Further crinoid material has been located in the intestine of *Helminthochiton* from the Lady Burn Starfish Beds, suggesting that those animals were an important part of the diet of the chiton (Donovan *et al.* 2010, 2011). Crinoids from the Craighead Inlier have been discussed by Donovan and Harper (1993) and Donovan and Clark (2020). The studies of Rushton (2001, 2003), Williams (1987) and Williams *et al.* (2004) have documented and revised parts of the graptolite fauna. Vandenbroucke *et al.* (2002) have refined parts of the Upper Ordovician biostratigraphy using chitinozoans, allowing calibration with the graptolite zonation and better global correlation of these parts of the succession. The Sandbian Lagerstätte, within the Laggan Member of the Balclatchie Formation, exposed in Dalfask Quarry has been documented in detail by Stewart (2005; Stewart and Owen 2008) following its initial discovery (Harper and Owen 1996).

For much of the succession, the faunas have affinities with coeval faunas in the Appalachians, comprising, together with faunas in the west of Ireland, the Scoto-Appalachian province (Williams 1962; Ross and Ingham 1970). During the late Caradoc and Ashgill (Sandbian–Katian), the faunas were more cosmopolitan; the period terminated with the near-globally distributed *Hirnantia* brachiopod fauna in the High Mains Formation (Harper 1981; Rong *et al.* 2020).

Northern Ireland

Tyrone Igneous Complex

The Tyrone Igneous Complex of Northern Ireland (Fig. 13) is one of the largest remnants of an Ordovician Laurentian-affinity arc system preserved in the British and Irish Caledonides. It is broadly divisible into a tectonically dissected ophiolite (=Tyrone Plutonic Group), a volcanic arc/backarc (=Tyrone Volcanic Group), and a late intrusive suite of continental arc granitoids (Cooper *et al.* 2011). Both the Tyrone ophiolite and its associated arc structurally overlie a sequence of Dalradian-affinity paragneisses termed the Tyrone Central Inlier, which was deformed and metamorphosed up to sillimanite grade during the *c.* 475–465 Ma Grampian orogeny (Chew *et al.* 2008).

Interpreted as the uppermost portions of a suprasubduction affinity ophiolite, the Tyrone Plutonic Group is dominated by layered, isotropic and pegmatitic gabbros, with lesser sheeted dolerite dykes, and rare pillow lavas (Hutton *et al.* 1985; Cooper and Mitchell 2004; Draut *et al.* 2009). Its ultramafic section is almost entirely absent. U-Pb zircon geochronology (c. 484-479 Ma) has constrained its formation to the Tremadocian (Cooper *et al.* 2011; Hollis *et al.* 2013a), which suggests broad correlations to *c.* 481-478 Ma Annieopsquotch Ophiolite Belt of Newfoundland (Lissenberg *et al.* 2005), and the Ballantrae ophiolite of western Scotland (Stone and Rushton 2018; this chapter). The ophiolite most likely formed immediately outboard of the Tyrone Central Inlier (a probable outboard microcontinental block; Chew *et al.* 2008) above a NW dipping subduction system (Hollis *et al.* 2013a). Although obduction onto the Tyrone Central Inlier occurred prior to the intrusion of tonalite intrusions at *c.* 470 Ma (Cooper *et al.* 2011), unpublished U-Pb zircon geochronology suggests obduction may have occurred at *c.* 477 Ma.

The Tyrone Volcanic Group is characterized by basaltic to rhyolitic lavas, tuffs, banded chert, ferruginous jasperoid, and rare mudstones, cut by numerous high-level intrusive rocks (Cooper and Mitchell 2004; Draut *et al.* 2009). Geochemical signatures are consistent with formation in an evolving Laurentian-affinity arc and back-arc that underwent several episodes of intra-arc rifting prior to its accretion to the Laurentian margin (Cooper *et al.* 2011; Hollis *et al.* 2012, 2013b). Rift packages are considered highly prospective for volcanogenic massive sulphide (VMS) style mineralization (Hollis *et al.* 2014, 2015, 2016). In contrast to the Lough Nafooey arc system of western Ireland (Draut *et al.* 2004), the accreted Tyrone Volcanic Group is considerably younger and very short-lived. U-Pb zircon and graptolite constraints range from upper Floian (*c.* 473 Ma) to lowermost Darriwilian (*c.* 469 Ma) in age (Cooper *et al.* 2008; Hollis *et al.* 2012, 2013b; Fig. 13). The Tyrone Volcanic Group most likely correlates with parts of the VMS-rich Buchans-Robert's Arm arc system of Newfoundland of the Annieopsquotch Accretionary Tract (as argued in Hollis *et al.* 2012, 2014), and locally the Charlestown Group of County Mayo (Herrington *et al.* 2018).

The late suite of calc-alkaline continental arc intrusions, which stitch both the Tyrone Plutonic Group and Tyrone Volcanic Group in their present position above the Tyrone Central Inlier, include large intrusions of diorite, tonalite, granodiorite, biotite granite and quartz-porphyry (Cooper and Mitchell 2004). These have yielded U-Pb zircon ages ranging from *c.* 470 to *c.* 464 Ma (i.e. Dapingian to Darriwilian; Cooper *et al.* 2011; Fig. 13). Equivalent late continental arc intrusions occur within the Slishwood Division (Flowerdew *et al.* 2005) and Connemara (Draut and Clift 2002; Friedrich *et al.* 1999a, b; see below).

Late Ordovician sedimentary rocks

The Ordovician succession at Pomeroy contains diverse shelly and graptolite faunas and is comparable with more complete and better exposed coeval rocks along strike in the Girvan district. The Bardahessiagh Formation (Mitchell 1977), over 150 m thick, unconformably overlies the Tyrone Volcanic Group and in ascending order consists of: Member (I), including locally fossiliferous, fine grey micaceous mudstones with some siltstones; Member (II), mainly unfossiliferous grey micaceous mudstones with laminated siltstones and decalcified nodules in its upper part; and Member (III), with sandstones and siltstones, the middle part of which is highly fossiliferous (Candela 2003; Scrutton *et al.* 1997). Abundant and diverse brachiopod faunas have been described (Mitchell 1977; Candela 2003); the faunas correlate the formation with the middle Caradoc (Sandbian) and occur in four ecological associations distributed along an onshore-offshore gradient (Candela 2001). The brachiopod faunas from the Bardahessiagh Formation and the enigmatic coral *Kilbuchophyllia*

(Scrutton *et al.* 1997) link this part of Northern Ireland with the coeval Kilbucho and Wandell Water faunas in the Northern Belt of the Southern Uplands (Candela and Harper 2010).

The overlying Killey Bridge Formation probably ranges in age from Cautleyan to Rawtheyan and contains a diverse shelly fauna that includes brachiopods (Mitchell 1977), bivalves (Tunnicliff 1982), cephalopods (Evans 1993), trilobites (Thomas *et al*. 1984, pp. 42, 44) and a stem-group chordate (Cripps 1988). The Tirnaskea Formation includes calcareous siltstones; a late Rawtheyan deep-water fauna in its middle part is succeeded by a few elements of the *Hirnantia* brachiopod (Harper *et al.* 1994; Rong *et al.* 2020) and *Mucronaspis* trilobite faunas (Temple 1952).

Peri-Laurentian arcs of NW Ireland

Lough Nafooey arc system and South Mayo Trough

The upper Cambrian and Ordovician rocks exposed in western Ireland from Clew Bay to Galway Bay are well documented in the literature. They have been interpreted as preserving a remarkably complete record of arc-continent collision within this sector of the *c.* 475–465 Ma Grampian orogeny (Dewey 2005). Such collisions resulted in the widespread deformation and metamorphism of the Dalradian Supergroup across Ireland and Scotland. The Mweelrea-Partry Syncline of South Mayo includes a thick succession of Ordovician sedimentary rocks and lesser volcanic rocks, exposed along its northern and southern limbs (Fig. 14). Together, this succession of rocks has been interpreted as representing remnants of the Lough Nafooey arc (i.e. the Lough Nafooey and Tourmakeady groups) that collided with the Laurentian margin, its proposed fore-arc basin (the South Mayo Trough), and evidence for post-collisional continental arc magmatism following a reversal in subduction polarity (see Dewey 2005; Ryan and Dewey 2011; Graham 2019).

Evidence for juvenile magmatism of the Lough Nafooey arc is exposed on both limbs of the Mweelrea-Partry Syncline. This includes 'boninitic' to island arc tholeiitic affinity lavas of the Bohaun Volcanic Formation on the northern limb, and island arc tholeiitic basalts of the Bencorragh Formation (=lowest Lough Nafooey Group; Fig. 14) on the southern limb (Clift and Ryan 1994; Ryan and Dewey 2011; Graham 2019). Both have geochemical characteristics indicative of an origin far from Laurentia. As the Lough Nafooey arc system approached the Laurentian margin, continental material entered the subduction channel and arc magmas evolved to more silicic, calc-alkaline and contaminated compositions (Draut and Clift 2001). This 'soft collision' is also marked by a dramatic increase in LREE enrichment in whole rock geochemical analyses, and significantly lower εNd_t values with stratigraphical height (Draut *et al.* 2004). The basaltic to andesitic Finny Formation, which overlies the Bencorragh Formation, has been constrained to the late Cambrian through U-Pb zircon geochronology (ages of *c.* 489.9 ± 3.1 and 487.8 ± 2.3 Ma from plagiogranite boulders: Chew *et al.* 2007). The overlying Knock Kilbride Formation has been constrained by late Tremadocian graptolites (La2; Williams and Harper 1994) and shows increased levels of crustal contamination (Draut *et al.* 2004). The Derry Bay Member (Graham 2019) marks the onset of marine volcaniclastic sandstone deposition. An upper limit for the age of the Lough Nafooey Group is provided by: 1) a Floian U-Pb age (474.1 ± 2.9 Ma) from zircons within the unconformably overlying Currarevagh Limestone Formation (Graham 2019); and 2) a *D. protobifidus* (i.e. mid-Arenig; Chewtonian) assemblage from the Knock Kilbride Argillites (Dewey *et al.* 1970; Ryan and Dewey 2011). A full description of the Lough Nafooey Group stratigraphy is provided by Graham (2019).

Hard collision between the Lough Nafooey arc and the Laurentian margin most likely occurred during the late Floian, prior to the deposition of the strongly LREE-enriched Tourmakeady Volcanic Group (Draut *et al.* 2004). The Tourmakeady Volcanic Group (Fig. 14) is divisible into the Mount Partry, Tourmakeady and Srah formations. The Mount Partry Formation includes a '*D. nitidus*' graptolite assemblage (mid-Arenig; Chewtonian) within a succession of predominantly volcanogenic rocks, cherts and mudstones. The succeeding Tourmakeady Formation includes carbonate rocks (limestones and limestone breccias), built on a series of extrusive rhyolite domes and their volcanic derived products (i.e. volcanic breccias and clastic sediments). Brachiopods (Williams and Curry 1985), trilobites (Adrain and Fortey 1997), bryozoans (Taylor and Curry 1985) and 'pelmatozoans' (Donovan 1986) are present in the limestone sequence, and Bergström and Orchard (1985) recorded conodonts. The faunas are of Laurentian aspect, possibly associated with marginal volcanic archipelagos. The Srah Formation is dominated by conglomerates and sandstones, interbedded with cherts and mudstones with an *Undulograptus austrodentatus* Biozone fauna indicating the uppermost Arenig (lower Darriwilian) (Dewey *et al.* 1970).

Broadly coeval with the deposition of the Tourmakeady Volcanic Group, a thick sedimentary succession is preserved on the northern limb of the Mweelrea-Partry Syncline. This includes the Letterbrock, Derrymore and Sheeffry formations (see Ryan and Dewey 2011; Fig. 14). Numerous tuff bands of calc-alkaline affinity occur in the sequence, particularly within the last formation. Towards its base the Sheeffry Formation has also yielded an Arenig (Dapingian) graptolite assemblage (references in Ryan and Dewey 2011). Overlying sedimentary deposits of the largely turbiditic Derrylea Formation (Darriwilian) are contemporaneous with the Rosroe Formation, discussed below.

On the southern limb of the Mweelrea-Partry Syncline, the Rosroe Formation (Fig. 14) consists of conglomerates and sandstones and extends westwards from Lough Nafooey to the southern shores of Killary Harbour (Archer 1977; Stouge *et al.* 2016). Clasts within the Rosroe conglomerate are well-rounded and include granite, rhyolite and psammite. Metamorphic clasts increase in proportion upwards, indicating a metamorphic source region being rapidly exhumed. The Rosroe Formation also contains blocks of allochthonous limestone. Two separate conodont assemblages in different blocks (Fig. 14) were identified by Stouge *et al.* (2016), the lower of which (Floian–Dapingian) was from the recently defined Currarevagh Limestone Formation (Graham 2019). The upper assemblage from the Rosroe Formation, from Killary Harbour, is of Darriwilian age (Stouge *et al.* 2016) and suggests continued carbonate deposition throughout the Arenig. Limestone blocks were derived from short-lived carbonate successions that accumulated on offshore 'peri-Laurentian' islands. During collapse of the carbonate system in the late Mid Ordovician, limestone blocks travelled down a steep slope into deep-water in debris flows, mixing with other components in the coarse, polymict clasts of the Rosroe Formation (Stouge *et al.* 2016).

The Derryveeny Formation crops out west of Lough Mask, on the southern limb of the Mweelrea-Partry Syncline. It is dominated by boulder conglomerates and sandstones (Graham *et al*. 1991). Conglomerates contain clasts of migmatite, schist, gneiss, granite, acid porphyry, spilite and vein quartz, and trace-element analysis of granite clasts highlight a similarity to the Rosroe Formation (Clift *et al*. 2002). Schist clasts have yielded muscovite - whole-rock Rb-Sr ages of 471 ± 8 Ma and 462 ± 7 Ma, constraining the maximum age of deposition (Graham et al. 1991).

Overlying sedimentary sequences of the Mweelrea-Partry Syncline include the Llanvirn–Caradoc (Darriwilian–Sandbian) Glenummera and Mweelrea formations (described in detail by Pudsey 1984a, b). The Glenummera Formation is composed of green-grey slates, mudrocks and cherty argillites with thin sandstones. It has yielded a Darriwilian 3 age on the northern limb of the syncline (Fig. 14). The overlying sandstone-dominated Mweelrea Formation was deposited by humid alluvial fans, fan deltas and braided rivers. Three marine shale units/horizons have been recognized from the northern limb, and include fossil material (Fig. 14). The lowest is correlated with the lower Llanvirn (Harper *et al.* 1988; Darriwilian 3-4), the middle part with the middle Llanvirn (Williams 1972) and the upper with the lower Caradoc (Harper *et al.* 2010; Sandbian). All three assemblages have Laurentian provincial affinities, the first two with the Toquima-Table Head province and the highest with Scoto-Appalachian faunas, occurring also at Girvan, SW Scotland. A number of rhyolitic ignimbrites in the Mweelrea Formation are associated with late post-subduction reversal continental arc magmatism. The lowest ignimbrite has yielded a U-Pb zircon age of 464.4 ± 3.9 Ma (Dewey and Mange 1999). Biostratigraphical constraints from the Lough Shee Mudrocks suggest a correlation between the Maumtrasna Formation exposed to the east, and the Mweelrea Formation (Harper *et al.* 1988; Ryan *et al.* 2010).

The timing for the exhumation of different components of the orogen into the South Mayo Trough is well documented by sediment geochemistry and heavy mineral studies (Wrafter and Graham 1989; Dewey and Mange 1999; Ryan and Dewey 2011). The Letterbrock Formation, the lowest sedimentary succession of the northern limb of the Mweelrea-Partry Syncline, comprises slates, greywackes and conglomerates containing ophiolitic and mature continental detritus. This was most likely derived from the Deer Park Complex (a dismembered ophiolite) and Killadangan Formation (an accretionary mélange of Laurentian provenance). The timing of exhumation for the ophiolitic Deer Park Complex has been constrained to *c.* 482 Ma (Tremadocian; Chew *et al.* 2010), which is consistent with the likely Arenig (Floian) age of the Letterbrock Formation. Whereas the overlying turbiditic Sheeffry and lower Derrylea formations (of Dapingian and Darriwilian age respectively) are dominated by ophiolitic and evolved arc detritus, the upper Derrylea Formation and equivalent upper Rosroe Formation of the southern limb record the sudden appearance in detritus of garnet and staurolite (Dewey and Mange 1999; Ryan and Dewey 2011). This is indicative of the abrupt unroofing of Grampian Barrovian amphibolite facies metasedimentary rocks to the north (Dewey and Mange 1999). Graptolites from just below this marker horizon comprise a fauna ascribed to the Darriwilian 3 stage of Australasia (Ryan and Dewey 2011).

Charlestown Group

The Charlestown Group is exposed across approximately 45 km^2 of County Mayo and records the progressive evolution of a Laurentian-affinity Ordovician arc system. Early volcanism of the Horan Formation (Fig. 14) is dominated by tholeiitic to calc-alkaline basalt, which is interbedded with lesser crystal tuff and sedimentary rocks (O'Connor 1987; Long *et al.* 2005). The latter provide graptolite constraints indicative of a Mid Ordovician, latest Dapingian age (Yapeenian Ya2 of Australasia; Rushton 2014b) equivalent to *c.* 470 Ma using the GTS 2020 timescale (Goldman *et al*. 2020). U-Pb zircon constraints from volcanic and volcaniclastic rocks of the Horan Formation have yielded comparable ages of 472–471 Ma, with a slightly younger age for a diorite intrusion (c. 469 Ma; Herrington *et al.* 2018). The overlying and undated Carracastle and Tawnyinah formations record a progressive evolution from calc-alkaline andesitic tuffs, flows and volcanic breccias, to more silicic tuffs and volcanic breccias higher in the stratigraphy (O'Connor 1987; Long *et al.* 2005). Geochemical evidence for crustal contamination and zircon inheritance indicate the Charlestown arc system was founded upon continental crust - either the composite margin or a microcontinental block (Herrington *et al.* 2018). The Charlestown Group is best correlated with either the *c.* 474–469 Ma Tyrone Volcanic Group, or the syn-collisional arc volcanics of the Lough Nafooey arc system (i.e., Tourmakeady Volcanic Group), or both (Herrington *et al*. 2018).

Intrusive rocks of Connemara and the Delaney Dome Formation

The Connemara terrane of western Ireland is dominated by variably metamorphosed and deformed metasedimentary rocks of the Dalradian Supergroup intruded by an extensive belt of Ordovician and Early Devonian (c. 420–400 Ma) intrusive rocks (Fig. 14; Chew and Stillman 2009). Four major deformation events have been recognized during the Grampian orogeny, which are well constrained by the emplacement of syn- to post deformation, intrusive rocks (c. 475–463 Ma; see Friedrich *et al.* 1999a, b). The intrusive complex comprises tholeiitic mafic intrusions (mainly gabbro) and younger calc-alkaline intrusions of quartz diorite to granite (Friedrich *et al.* 1999b). It is interpreted to represent the roots of a continental arc (Yardley and Senior 1982).

The location of the Dalradian rocks of the Connemara terrane, south of the accreted arc-ophiolite complexes, was for many years interpreted to be the result of post-Grampian sinistral strike-slip displacement (Hutton and Dewey 1986; Hutton 1987). Dewey and Ryan (2015) presented an elegant alternative model, whereby the Connemara terrane is not displaced with respect to the rest of the orogen but is broadly *in situ* and was overridden by the Lough Nafooey arc, its fore-arc basin (South Mayo Trough), frontal ophiolite complex (Deer Park) and accretionary complex (Killadangan Formation) (also see Ryan and Dewey 2019).A rare package of Ordovician volcanic rocks also occurs in the Connemara region. The Delaney Dome Formation (Fig. 14) crops out within a structural window, separated from the overlying Connemara metagabbro and orthogneiss suite by the late Mannin Thrust. Now composed of monotonous, mylonitic quartzofeldspathic rocks, their protolith was interpreted by Leake and Singh (1986) as felsic and igneous, most likely rhyolites or ignimbrites. Additional amphibolitic rocks, some mylonitic, occur within the window (the Lough Nacorrussaum metabasites of Morris *et al.* 1995). These are geochemically distinct from overlying metabasic rocks immediately above the Mannin Thrust (the Ballyconneely amphibolites) and most likely represent metamorphosed Ordovician extrusive or intrusive units (Leake and Singh 1986). Whole rock geochemistry, Nd isotope constraints and U-Pb zircon geochronology from a meta-rhyolite (475.9 ± 2.2 Ma) suggest the Delaney Dome Formation is a probable along-strike equivalent to the syn-collisional Tourmakeady Volcanic Group (Draut and Clift 2002).

Ordovician magmatism in the Slishwood Division

The Slishwood Division is a thick sequence of mainly Proterozoic metasedimentary rocks exposed across three inliers of NW Ireland: NE Ox Mountains, Rosses Point and Lough Derg. It is composed of psammitic paragneisses, minor pelite, and abundant metabasite intrusives, with occurrences of ultrabasic rocks (Daly *et al.* 2022). Detrital zircons have a youngest population of *c.* 926 Ma, with the minor basic intrusions constrained to *c.* 595 Ma (Flowerdew and Daly 2005; Daly *et al.* 2022). The Slishwood Division is relevant here due to its complex early metamorphic history and the presence of Ordovician tonalite and granite intrusions (Fig. 14) which help constrain the timing of Grampian deformation and metamorphism. Tonalite and granite intrusions have yielded ages from 474 ± 5 to

467 ± 6 Ma (Flowerdew *et al.* 2005). Early eclogite-facies and high pressure-granulite facies metamorphic assemblages occurred before or during the early Grampian orogeny (Flowerdew *et al.* 2005; Daly *et al.* 2022). These high-grade rocks were extensively retrogressed under amphibolite-facies conditions during the Grampian orogeny. Juxtaposition of the Slishwood Division with the Dalradian Supergroup through SE-directed shearing occurred between *c.* 476 and 463 Ma (Flowerdew *et al.* 2000).

South Connemara Group

The South Connemara Group is exposed along the Southern Upland Fault in western Ireland. It is composed of a tectonic melange of MORB-affinity pillow lavas, abyssal cherts, sandstones and conglomerates with amphibolite intrusive rocks (Williams *et al.* 1988). The early stratigraphy of Ffrench and Williams (1984), comprising five formations with a total thickness of >3km, was reinterpreted by Ryan and Dewey (2004) as three formations with several structural repetitions. The Gorumna Formation is interpreted as the oldest formation and is composed of mafic volcanic rocks, mostly sheared basalts, though now occurring as six imbricate slices (Ryan and Dewey 2004). The Golam Formation is dominated by grey chert, with maroon, green and grey argillite, and cherty argillite interbeds. The Lettermullen Formation exhibits a gradational contact with the underlying Golam Formation, and is dominated by turbiditic sandstones and conglomerates, with four facies described by Ryan and Dewey (2004). One of the numerous mélange zones of the South Connemara Group contains an Arenig–Llanvirn microfossil assemblage (Williams *et al*. 1988). The South Connemara Group most likely forms an early component of the Southern Uplands and Longford-Down accretionary complex, forming in a late Ordovician trench associated with late north-dipping subduction along the Laurentian-margin (Ryan and Dewey 2004; Ryan and Dewey 2011).

Highland Border Complex

A discontinuous belt of much-faulted igneous and sedimentary rocks extending across Scotland from Stonehaven to the Isle of Arran and associated with the Highland Boundary Fault has been referred to the Highland Border Complex. Ordovician rocks within this belt are variably exposed and locally fossiliferous. A detailed stratigraphy was assembled by Curry *et al*. (1984) that formed the basis for an allochthonous model for the complex as part of exotic Midland Valley Terrane (Bluck 2002). Conodont and trilobite faunas from the Dounans Limestone Formation considered by those authors to be near the base of the succession, above the Aberfoyle Ophiolite, yielded an uncontroversial Arenig age. Less secure were reports of Upper Ordovician chitinozoans from the Margie Limestone and Sandstone together with the Achray Sandstone. The chitinozoan identifications, however, are not sustainable (Tanner and Sutherland 2007).

In a much-simplified re-interpretation of the stratigraphy (Tanner and Sutherland 2007), the complex is now divided into a lower part, which consists of the Trossachs Group, with the Leny Limestone and Slate Member (Cambrian Series 2; Fletcher and Rushton 2007) and the Arenig Margie Limestone Member, and an upper Garron Point Group with the Arenig Dounans Limestone overlying an obducted ophiolite. In this model, the lower part of the complex (Trossachs Group) comprises autochthonous Dalradian rocks of the Grampian Terrane whereas the Garron Point Group is an obducted ophiolite complex and cover succession, part of the Midland Valley Terrane (Tanner and

Sutherland 2007). The start of ophiolite obduction has been dated to *c.* 490 Ma (late Cambrian; Chew *et al*. 2010). The revised stratigraphy thus provides continuity along strike with that in the South Mayo Trough of NW Ireland, where the ophiolitic Deer Park Complex, equivalent to the Garron Point Group, is overlain by a Lower–Middle Ordovician (Arenig–Llanvirn) succession (Harper *et al*. 1989). Younger ages have been claimed for higher parts of the Highland Border Complex, but these remain controversial.

Ethington (2008) documented a conodont fauna from the Margie Limestone Member that includes *Paracordylodus gracilis*, which has a range from high in the *Paroistodus proteus* biozone to the Floian middle *Oepikodus evae* biozone (Goldman *et al.* 2014). This equates to a Floian age (c. 477–472 Ma; all absolute ages based on Goldman *et al.* 2020). The Dounans Limestone Formation has a conodont fauna (Ethington and Austin 1991) that includes *Jumudontus gananda* and *Periodon flabellum*. The presence of *J. gananda* constrains the age of the Dounans Limestone Formation between the uppermost *Oepikodus communis* Biozone and the top of the *Tripodus laevis* Biozone (473–470.2 Ma).

Hebridean Terrane

The Highlands of Scotland constituted the SE margin of Laurentia during the Ordovician, and the Hebridean Terrane, west of the Moine Thrust, was an integral part of the Great American Carbonate Bank, a non-uniformitarian, continent-wide area of carbonate deposition (Derby *et al.* 2012a, b; Raine and Smith 2012). This palaeogeographical location means that the Scottish margin was contiguous with western Newfoundland to the palaeo-west and NE Greenland to the palaeo-north (Swett and Smit 1972; Swett 1981; Smith and Rasmussen 2008; Ryan and Dewey 2019).

By far the most complete and most studied Ordovician succession in the Scottish Highlands is the carbonate-dominated Durness Group that crops out from Skye in the south to Durness and Loch Eriboll in the north (Fig. 15). It forms the youngest part of the foreland to the Moine Thrust Zone and is also incorporated into the lowermost thrust sheets. The Durness Group overlies the clastic Ardvreck Group (Cambrian Series 2) and ranges from the base of Cambrian Series 3 (Faggetter *et al.* 2018) to the Middle Ordovician (mid-Dapingian). Although the outcrop belt is more or less continuous, it rarely extends into the younger, Ordovician units, which are only present on Skye (where they are heavily faulted), at Stronchrubie in Assynt (where the succession only just extends into the Lower Ordovician Sailmhor Formation) and in the type area around Durness. The latter is the most complete and most structurally intact Ordovician north of the Highland Boundary Fault and is representative of the Laurentian margin in the British Isles, so is the focus of the remainder of this account (Fig. 16).

Lithostratigraphy

The Durness Group was first described in detail by Peach and Horne (1884) and Peach *et al.* (1907), who erected the stratigraphical divisions that continue to be used. Although the formation names have been stable and units consistently defined and recognised, the published thicknesses for the group have been variable, ranging from 460 m (Peach *et al.* 1907) to almost 1600 m (Phemister 1948). Recent bed-by-bed logging, corrected for faulting, has confirmed a thickness of at least 930 m, of which the Ordovician component has a minimum thickness of 736 m (the lowest two units, the

Ghrudaidh and Eilean Dubh formations are entirely Cambrian in age). The group has a strong diagenetic/metamorphic overprint with extensive recrystallisation, but the recent work has enabled a detailed appraisal of the depositional environments, biostratigraphy and sequence stratigraphy (Raine *et al.* 2011; Raine and Smith 2012, 2017).

Sailmhor Formation. The unit comprises 113 m of mostly dark, mottled dolostone forming metre-scale parasequences; white cherts are particularly abundant in the lower half of the formation. The formation is well exposed at its type section along the shores of Balnakeil Bay (Fig. 17) but is heavily faulted; the upper half of the formation is best exposed on the shore at Smoo Cave, to the east of Durness (Fig. 18). The base of the formation is taken at a sharp colour and lithological change from light grey, peritidal, finely crystalline dolostones of the Eileann Dubh Formation to dark grey, mottled dolostones exhibiting locally common cherts (Fig. 17). Oolite beds are common in the basal half of the formation. The mottling results from burrow networks and thrombolites and becomes more prominent up section, where the parasequences are commonly capped by ripple- and parallel-laminated, pale grey dolostones displaying stromatolites.

Sangomore Formation. The 55 m-thick formation is exposed in its entirety along the type section in Balnakeil Bay and is distinguished by light-grey and buff, finely laminated dolostones, with some mid-grey, thrombolitic limestones, stromatolites and bioclastic, peloidal and ooidal wackestones or packstones occurring locally. The lower boundary is gradational, with a series of chert breccias and dolomite sands spanning the boundary, which is placed at the top of a 60 cm-thick dolomite sand that forms a distinctive notch in the cliff. Above the formation boundary the dolostones become lighter in colour and the cherts are dominantly orange in colour.

Balnakeil Formation. The Balnakeil Formation has a minimum thickness of 86 m, and the lower boundary is marked by a distinctive oncoid and pebble bed at Balnakeil Bay. Lithologically, the formation is dominated by mid- to dark grey, stromatolitic and thrombolitic dolostones and limestones, with ribbon carbonates and bioclastic wackestones and packstones.

Crosaphuill Formation. A composite measured section provides a minimum thickness of 350 m. The lower part of the formation comprises some 135 m of cliff-forming, strongly burrow-mottled, purplish-grey, dolomitic limestones. Fossils are commonly found within brownish black cherts in the basal 30 m, and include rostroconchs, cephalopods, gastropods, brachiopods and sponges. Most of the fossils are poorly preserved, with some replaced by dolomite, but the majority by chert (beekite). The upper part of the formation is distinct from the underlying succession and dolostone beds are more abundant. Several light grey, structureless, dolomitic-limestone intervals up to 3 m thick are present at various levels and burrow-mottled, dolomitic limestone beds persist. Parasequences become increasingly apparent in this upper section, with lighter, parallel-laminated dolostones commonly capping the cycles.

Durine Formation. The highest unit of the Durness Group is not exposed in any one complete section but a series of inland sections provide a minimum thickness of 132 m. The basal boundary is gradational and placed where light-grey, fine-grained dolomites become abundant. The basal part of the Durine Formation contains some beds of burrow-mottled carbonate, a higher proportion of dolostone, a change in the colour of the cherts from black to orange-pink and an increasing proportion of parallel and ripple-laminated, light grey dolostones. The latter predominate further up section.

Biostratigraphy

Macrofossils are generally rare in the Durness Group due to a combination of lithofacies control, dolomitisation and low-grade metamorphism in the footwall of the Moine Thrust Zone that resulted in re-crystallisation. Where macrofossils do occur, they provide points of biostratigraphical correlation, and this is the case for both trilobites (Fortey 1992) and cephalopods (Evans 2011). In contrast, conodonts have been recovered throughout the Ordovician part of the group, although both yields and preservation are generally poor.

Conodonts were first recorded from the Durness Group in reconnaissance studies by Higgins (1967, 1971, 1985), but systematic collecting and investigation has taken place more recently (Huselbee 1998; Raine 2010). The Cambrian–Ordovician boundary interval was sampled for conodonts by Huselbee (1998). In common with many other shallow water sections, the key zonal taxon *Iapetognathus fluctivagus* was not recovered, but the last appearance of the Cambrian taxon *Eoconodontus notchpeakensis* and the first appearances of *Cordylodus lindstromi*, *Acanthodus* sp. and *Semiacontiodus nogamii* all occur within a few metres of the Eilean Dubh–Sailmhor formation boundary (Fig. 17). Similar relationships are seen at the global stratotype for the Cambrian–Ordovician boundary at Green Point, Newfoundland (Cooper *et al.* 2001; Goldman *et al.* 2020) and in the graphically correlated composite reference section for the Furongian–Darriwilian of Laurentia (Sweet and Tolbert 1997). It may be concluded that, within approximately a metre, the Cambrian–Ordovician boundary coincides with the Eilean Dubh–Sailmhor formation boundary.

In the Laurentian conodont biozonation, the base of the Floian lies in the upper part of the *Acodus deltatus/Oneotodus costatus* Biozone (477 Ma), with the base of the *Oepikodus communis* Biozone at around 475.5 Ma, coincident with the Flo 1 sea-level highstand (Goldman *et al.* 2020). *O. communis* has a FAD 8.5 m above the base of the Croisaphuill Formation, indicating that the base of the Floian lies close to the poorly exposed top of the underlying Balnakeil Formation.

The Floian–Dapingian (Lower–Middle Ordovician) boundary is defined by the first occurrence of the conodont *Baltoniodus triangularis*, which in Laurentia approximates to the base of the *Microzarkodina flabellum/Tripodus laevis* conodont biozone – in the boundary stratotype the FAD of *M. flabellum* is 20 cm above that of *B. triangularis* (Goldman *et al.* 2020). In the Durness Group, *M. flabellum*? occurs only in the topmost sample collected, and *T. laevis* is also scarce, but has a FAD 10 cm below the Croisaphuill–Durine formation boundary. In the composite reference section of Sweet and Tolbert (1997), *Pteracontiodus cryptodens* has a FAD at 965 composite standard units (csu), above the range-base of *T. laevis*. *P. cryptodens* is present 53 m below the top of the Croissaphuill Formation, and 162 m above the base of the recessive upper member of the formation. No conodonts were recovered in the 130 m of dolostone underlying the first occurrence of *P. cryptodens*, and typical Floian faunas were present below that interval. The Floian–Dapingian boundary therefore lies in the interval 167–297 m above the base of the Croisaphuill Formation, and 32–162 m above the base of the dolostone-dominated recessive interval of the upper member. This lithology is typical of the basal Dapingian (Whiterockian) sea-level lowstand in Laurentia, and it is likely that the base of the Dapingian lies toward the base of the barren interval.

The Durine Formation records the youngest deposition in the Scottish sector of the Laurentian margin prior to Grampian and Scandian orogenic collisions, and its youngest level therefore provides an important constraint on tectonic interpretations. The uppermost horizons have yielded moderately diverse faunas that include the zonal taxon *Histiodella altifrons* and, more particularly, indicate a position within the lower *H. altifrons* Biozone, corresponding to a late Dapingian age of 470 Ma.

The age of the youngest preserved Durine Formation is remarkably consistent with the dates for igneous intrusions associated with the Grampian orogeny, where in the Scottish sector S-type granites derived at peak metamorphism from melted lower crustal sedimentary rocks also cluster around 470 Ma, and Sm–Nd garnet ages constrain peak metamorphism to 473–465 Ma (see Chew and Strachan 2014, p. 59, for review).

Sequence stratigraphy and global correlation

During the Ordovician, concurrent sedimentation across almost the whole of Laurentia constituting the Great American Carbonate Bank means that changes in relative sea-level can be recognised with high precision on a continental scale. Utilising this, Sloss (1963) identified and named a series of megasequences identifiable across Laurentia, of which the oldest is the Sauk Megasequence that ranges in age from Cambrian Series 2 to Dapingian (Derby *et al.* 2012a). The Sauk sequence hierarchy is readily recognizable in the Hebridean Terrane (Raine and Smith 2012, 2017; Faggetter *et al.* 2018), with Sauk I corresponding to the clastic Ardvreck Group underlying the Durness Group (Cambrian Series 2). The Sauk I–II sequence boundary is marked by the switch from clastic-dominated deposition to carbonates at the Ardvreck–Durness group boundary and the sequence boundary, 2 m above the lithostratigraphical change, also coincides with the Cambrian Series 2–Series 3 boundary, the Hawke Bay event, and the later part of the ROECE carbon isotope excursion (Faggetter *et al.* 2018).

All four constituent sequences of the Furongian–Dapingian Sauk III supersequence are identifiable within the Durness Group (Raine and Smith 2012). The Sauk II–III sequence boundary lies 46 m below the top of the Eilean Dubh Formation (Raine and Smith 2012, 2017) and sequence IIIA is also entirely contained within that unit. The base of the Sauk IIIB sequence (131 m thick) is 18 m below the Eilean Dubh–Sailmhor formation boundary, and therefore approximately the same distance below the Cambrian–Ordovician boundary, and is marked by a cream-weathering, clast-supported conglomerate with angular to subangular clasts up to 7 cm in size. An abrupt deepening is evident at the base of the Sailmhor Formation, with parasequences comprising dark subtidal mottled dolostones with thrombolites, sometimes with a thin, peritidal cap. Parasequence thickness and the proportion of subtidal sediment increase upwards (reflecting increasing accommodation space) within this transgressive systems tract (TST) to the unexposed maximum flooding zone (MFZ), and the upper part of the formation then represents the high-stand systems tract (HST). Parasequences progressively thin and become more peritidally dominated in the upper 40 m of the Sailmhor Formation (Fig. 18) and the Sauk IIIB–IIIC boundary is marked by a 45 cm coarse-grained dolostone with an erosive base that also defines the base of the Sangomore Formation.

The Sangomore Formation corresponds to the Sauk IIIC sequence (55 m thick) and the upper sequence boundary (IIIC–IIID) is marked by an erosive surface with a distinctive pebble bed containing subangular dolostone clasts of up to 3 cm diameter, with some local chert. In the upper part of the 20 cm bed, the pebbles are sparser but have oncoidal cortices and are matrix-supported in an ooidal grainstone. Angular carbonate breccias infill erosion surfaces at this level, chert breccias are present, and parasequence tops are frequently karstified. The sequence boundary lies within the *Macerodus dianae* conodont biozone.

Sauk IIID is the thickest sequence in the Ordovician part of the Durness Group, extending from the base of the Balnakeil Formation through the Croisaphuill Formation to the top of the exposed Durine Formation (568 m). Within the Balnakeil Formation, parasequences comprise peloidal and bioclastic wackestones with thrombolites that shallow upwards into microbial laminites and stromatolites, representing the TST of Sauk IIID. Parasequences thicken upwards towards the MFZ and pale peritidal parasequence tops disappear. The lower part of the overlying Croisaphuill Formation is the most fossiliferous part of the group and comprises burrow-mottled dolomitic limestones in which well-developed parasequence boundaries are absent – the MFZ of Sauk IIID is located within the lower part of this succession, within the lowest part of the *Oepikodus communis* conodont biozone. In the upper part of the Croisaphuill Formation well-defined parasequences re-appear, albeit poorly exposed, and typically have peritidal laminite tops with an increasing proportion of dolostone upwards. The Durine Formation is similar, but more dolostone dominated. Fine-grained, pale grey dolostones predominate and cherts preserve evaporite pseudomorphs. There is evidence that the environment became increasingly restricted as both macro- and ichnofaunas are absent, although conodonts (a nektonic component of the fauna) can be abundant and diverse at some levels.

The HST of Sauk IIID is exceptionally thick, suggesting that subsidence slowed and little accommodation space was available for an extended period of time, from the *Reutterodus andinus* Biozone to the *Histiodella altifrons* Biozone, an interval of around 2 million years. Raine and Smith (2012) speculated that this may have been due to the onset of Grampian orogenesis restricting subsidence and the development of accommodation space. A similar effect is seen in western Newfoundland (Knight *et al.* 1991) and back-stripping of subsidence for these two successions lends support to the hypothesis (Smith and Rasmussen 2008). The preserved top of the Durine Formation (470 Ma) is also close to coeval with the Margie Limestone Member at the top of the Dalradian Supergroup (477–472 Ma) and the Dounans Limestone Formation above the Highland Border ophiolite (473–470.2 Ma), lending support for a tectonic end to sedimentation in the Durness Group of the NW Highlands rather than the current top of the group representing an arbitrary erosion surface.

Note on online resources

Further details of successions summarised in this account and localities mentioned are freely available on geological survey and other websites. The British Geological Survey's (BGS) onshore Geoindex (https://www.bgs.ac.uk/map-viewers/geoindex-onshore/) provides access to map data for England, Scotland, Wales and Northern Ireland at 1:50,000 and 1:625,000 scales, plus an array of other datasets. The Geological Survey of Ireland's Map Viewer (https://www.gsi.ie/en-ie/data-and-maps/Pages/default.aspx) performs a similar function for the Republic of Ireland. BGS memoirs, sheet explanations and other publications are available in full text on the BGS TextViewer (https://webapps.bgs.ac.uk/Memoirs/); those cited in this paper are linked directly from the reference list below. The BGS Maps Portal (https://www.bgs.ac.uk/information-hub/bgs-maps-portal/) provides scans of published and printed BGS maps (useful for cross sections and generalised vertical sections; those maps cited are again linked from the reference list), and the BGS Lexicon (https://www.bgs.ac.uk/technologies/the-bgs-lexicon-of-named-rock-units/) provides summary information on lithostratigraphical units. Chapters in the two Geological Conservation Review series volumes cited in this paper (Stephenson *et al*. 1999; Rushton *et al*. 2000) are also freely available and can be accessed by following the links.

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Figure Captions

Figure 1. Palaeozoic terrane map of Britain and Ireland; see text for sources.

Figure 2. Ordovician outcrop in Britain and Ireland. Numbers 1–41 indicate the locations of stratigraphical columns illustrated in Figures 3–12: 1. Llŷn Peninsula; 2. Snowdonia; 3. Arenig - Bala; 4. Welshpool; 5. Shelve; 6. Cadair Idris - Rhobell Fawr; 7. Plynlimon; 8. Cardigan-Fishguard; 9. Whitland-Carmarthen; 10. Llandeilo-Llandovery; 11. W Limb of Tywi Anticline; 12. Llanwrtyd; 13. Garth - Crychan Forest; 14. Builth Inlier; 15a. Caradoc (south), Onny Sub-basin; 15b. Caradoc (north), Cressage-Cardington Sub-basin; 16. Malvern Hills; 17. Tortworth Inlier (Worcester Proto-graben); 18. Nuneaton; 19. Craven Inliers, N England; 20. Anglesey; 21. Tagoat; 22. Tramore; 23. Courtown; 24. SE Wicklow/Central Wexford; 25. Lake District (north); 26. Cross Fell (south); 27. Cautley-Dent inliers; 28. Lake District (south); 29. Isle of Man; 30. West Wicklow; 31. NE Wicklow/NW Wexford; 32. Bellewstown; 33. Grangegeeth; 34. Northern Belt, Southern Uplands; 35. Moffatdale, Central Belt, Southern Uplands; 36. Northern Belt, Longford-Down; 37. S of Orlock Bridge Fault, Central Belt, Longford-Down; 38. Craighead Inlier; 39. Girvan foreshore; 40. Penwhapple Burn-Stinchar Valley; 41. Water of Gregg-Doularg. Number 42 indicates the location of the Tyrone Igneous Complex and overlying Ordovician successions at Pomeroy (see Fig. 13); numbers 43 and 44 show the locations of peri-Laurentian arc successions, respectively at Charlestown and in the South Mayo Trough (see Fig. 14). BH: Berwyn Hills; CSF: Church Stretton Faults; CPF: Causey Pike Fault; HBC: Highland Border Complex: HD: Harlech Dome; PLF: Pontesford-Linley Fault.

Figure 3. Representative stratigraphical columns of the Cymru Terrane from west to east across North Wales, from the Llŷn Peninsula to the Shelve area in the Welsh Borderland. Correlation of global stages, Anglo-Welsh series and stages, Anglo-Welsh graptolite zones and conodont zones is taken with slight modification from TSCreator V. 8.0, based on Goldman *et al.* (2020); the *Baltograptus jacksoni* graptolite Biozone replaces the *Corymbograptus varicosus* Biozone in the Anglo-Welsh scheme (Rushton 2011a). Chitinozoan biozones based on Vandenbroucke (2008), Vandenbroucke *et al.* (2008, 2009) and Challands *et al.* (2009); *lat-onn*: *Acanthochitina latebrosa*–*Ancyrochitina onniensis* Biozone; *dal-deunff*?: *Lagenochitina deunffi* to *Lagenochitina dalbyensis* biozonal interval(?). See Figure 2 for locations. Abbreviations – ADM: Allt Ddu Mudstones Formation; AF: Aran Fawddwy Formation; Br: Bryncroes Formation; Cei: Cwm Eigiau Formation; Dwf: Dwyfach Formation; Dyn: Dol-cyn-afon Formation; GyG: Glyn Gower Siltstones Member (Ceiswyn Formation); Lgn: Llanengan Formation; NFr: Nant Ffrancon Subgroup; NHM: Nant Hir Mudstone; PMu: Porth Meudwy Formation; PT: Pitts Head Tuff Formation; ST: St Tudwal's Formation; ULG: Upper Lodge Volcanic Group. Unit names in italics (Serw Formation, Olchfa Shales, Filltirgerig Beds) are regarded as either superfluous stratigraphical terms (Rushton and Howells 1999: Serw Formation) or are of uncertain status (Fearnsides 1905; see Zalasiewicz 1984 for correlation).

Figure 4. Representative stratigraphical columns from west Wales. See Figure 3 for an explanation of the left-hand columns and Figure 2 for locations. MMb: Mottled Mudstone Member.

Figure 5. Representative stratigraphical columns from the Whitland-Carmarthen district of South Wales north-eastwards along the Tywi Lineament to the Builth Wells Inlier. See Figure 3 for an explanation of the left-hand columns and Figure 2 for locations. Abbreviations - BrF: Bronydd Formation; BRN: Bryn Nicol Formation; CaT: Cwm-amliw Tuff Formation; Cbn: Caban Conglomerate

Formation; CcF: Crychan Formation; Ccy: Cwm Clŵd Sandstone; CF: Ciliau Formation; CgF: Cwmcringlyn Formation; ChF: Chwefri Formation; Cri: Cribarth Formation; db: disturbed beds in the Yr Allt Formation; Dd: Doldowlod Conglomerate Formation; LrM: Llanfawr Mudstone Formation; MMb: Mottled Mudstone Member; Nwd: Newmead Sandstone Formation; PiM: Pistyllgwyn Member; SLM: Sugar Loaf Member; Tal: Taliaris Formation; TrV: Trelowgoed Volcanic Formation; YA: Yr Allt Formation.

Figure 6. Representative stratigraphical columns, borehole data and radiometric dates from the Wrekin, Charnwood and Fenland terranes. Sections in the Wrekin and Charnwood terranes are based on selected columns in Fortey *et al*. (2000, figs 14, 16). The locations of sections 15b, 16, 17 and 18 are shown in Figure 2. The summary of borehole provings in the Fenland Terrane is from Molyneux (1991) and radiometric dates are from Pharaoh (2018). See Figure 3 for an explanation of the left-hand columns.

Figure 7. Stratigraphical sections in the Onny and Cressage-Cardington sub-basins of the type area of the Caradoc Series on the western edge of the Wrekin Terrane (Fig. 2). Modified from Fortey *et al.* (2000, fig. 14) and Bowdler-Hicks *et al.* (2002, text-fig. 1).

Figure 8. Representative stratigraphical columns from the Monian Terrane, based on Fortey *et al*. (2000, figs 13, 33), Schofield *et al.* (2020, fig. 1b) and sources cited in the text. See Figure 3 for an explanation of the left-hand columns and Figure 2 for locations. Abbreviations - BM: Bunmahon Formation; DB: Dunabrattin Shale Formation; DBL: Dunabrattin Limestone Formation; DHBT: Dam House Bridge Tuff Member; JRS: Jop Ridding Sandstone Member; WFC: Wharfe Conglomerate Member.

Figure 9. Representative stratigraphical columns from the Leinster-Lakesman Terrane in NW England, the Isle of Man and SE Ireland, based on sources cited in the text. See Figure 2 for locations and Figure 3 for an explanation of the left-hand columns. Chitinozoan biozones in northern England are from Vandenbroucke *et al.* (2005). Ahl: Ash Gill Mudstone Formation; AVF: Appletreeworth Volcanic Formation; BBF: Bitter Beck Formation; BMr: Broughton Moor Limestone Formation; BRUL: Barrule Formation; BVG: Borrowdale Volcanic Group; CGA: Creg Agneash Formation; CRMR: Creggan Moar Formation; Cmu: Cautley Mudstone Formation; CtV: Cautley Volcanic Member (Cautley Mudstone Formation); Cyo: Cystoid Limestone Member (Ash Gill Mudstone Formation); DnSh: Dufton Mudstone Formation; DrSh: Drygill Mudstone Formation; EVG: Eycott Volcanic Group; GLRN: Glen Rushen Formation; Hbe: Hope Beck Formation; IBK: Injebreck Formation; KkB: Kirkley Bank Limestone Formation; KSt: Kirk Stile Formation; LPT: Lady Port Formation; LWF: Loweswater Formation; MGD: Maughold Formation; MHL: Mull Hill Formation; Sen: Stile End Formation; SPN: Spengill Mudstone Member (Skelgill Mudstone Formation); SwL: Swindale Limestone Formation; WHg: Watch Hill Formation; Yrl: Yarlside Volcanic Formation. Volcanic and volcaniclastic rocks in the Manx and Ribband groups: *p*, tuff and breccia near Peel on the west coast of the Isle of Man; *d,* Donard andesites, and *dh*, Dowery Hill basalts in tract 1 of the Ribband Group; *k*, Kilcarry andesites in tract 2. Mn: manganese-rich rocks on the Isle of Man; C.: coticule in the Ribband Group.

Figure 10. Stratigraphical successions in the Bellewstown and Grangegeeth terranes adjacent to and respectively south and north of the Iapetus Suture in eastern Ireland. After Fortey *et al*. (2000, fig. 31). See Figures 2 and 3 for locations and an explanation of the left-hand columns respectively.

Figure 11. Correlation of representative stratigraphical columns from the Northern (34, 36) and Central belts (35, 37) of the Southern Uplands of Scotland (34, 35) and the Longford-Down area (36, 37). Modified from Fortey *et al*. (2000, fig. 22) for the Southern Uplands and Anderson (2004) for Longford-Down. See Figures 2 and 3 for locations and an explanation of the left-hand columns respectively; Scottish graptolite zones are from Williams *et al.* (2004) and Zalasiewicz et al. (2009). Abbreviations - BAIL: Bail Hill Volcanic Group; BS: Birkhill Shale Formation; CRFD: Crawford Group; DPF: Downan Point Lava Formation; GNE: Glenlee Formation; GS: Glenkiln Shale Formation; KKF: Kirkcolm Formation; LHG: Leadhills Supergroup; LHS: Lower Hartfell Shale Formation; MCHB: Marchburn Formation; SHIN: Shinnel Formation; UHS: Upper Hartfell Shale Formation.

Figure 12. Successions in the Ballantrae and Girvan areas of the Midland Valley of Scotland (Fig. 2). After Fortey *et al*. (2000, figs 23, 24). See Figures 2 and 3 for locations and an explanation of the left-hand columns respectively; Scottish graptolite zones are from Williams *et al.* (2004) and Zalasiewicz *et al*. (2009). Formations in the Whitehouse Subgroup are coloured to identify them more easily and aid correlation between sections.

Figure 13. Geological evolution for the formation of the Tyrone Igneous Complex, and related deformation, metamorphism and magmatism in the Tyrone Central Inlier. U-Pb zircon geochronology is from Chew *et al.* (2008), Cooper *et al.* (2008, 2011), and Hollis *et al.* (2012, 2013a, b, in prep.). Biostratigraphical constraints are from Cooper *et al.* (2008). Argon-argon geochronology from the Tyrone Central Inlier is from Chew *et al.* (2008). Stratigraphical correlations are modified from Hollis *et al.* (2013b, 2014). TPG: Tyrone Plutonic Group; TVG: Tyrone Volcanic Group.

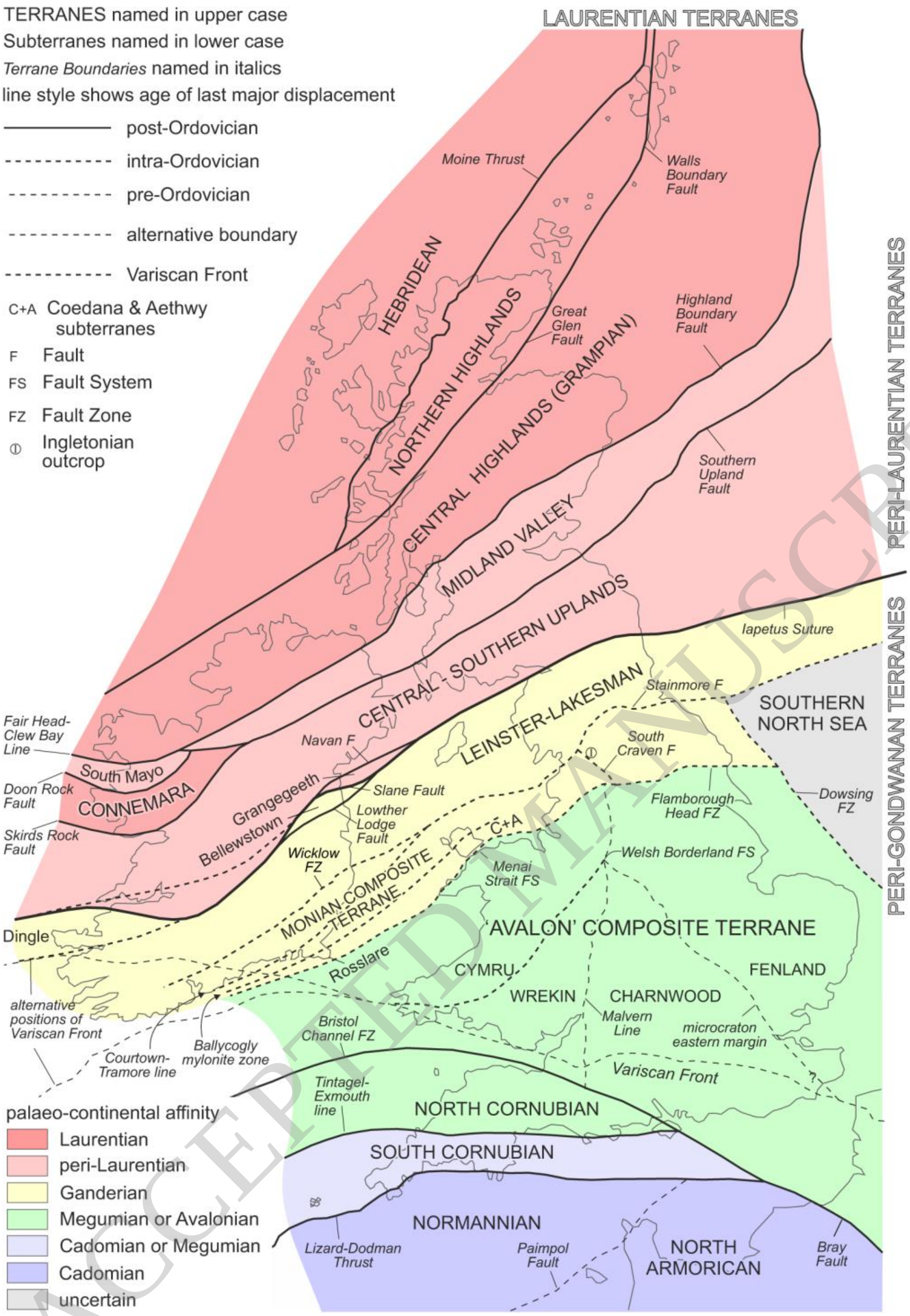
Figure 14. Geological evolution for Ordovician arc systems preserved in western Ireland, modified after Ryan and Dewey (2011) and Herrington *et al.* (2018). Biostratigraphical age constraints are from the review of Ryan and Dewey (2011), Stouge *et al.* (2016) and Herrington *et al.* (2018). U-Pb zircon geochronology is from: Dewey and Mange (1999; = Mweelrea Fm), Friedrich *et al.* (1999a, b; = Connemara magmatism), Draut and Clift (2002; = Delaney Dome Formation); Flowerdew *et al.* (2005; = Slishwood Division), Chew *et al.* (2007; = plagiogranite boulders), Ryan *et al.* (2010; = Rosroe Fm); Herrington *et al.* (2018; = Charlestown Group), and Graham (2019; Currarevagh Limestone Formation). Argon-argon geochronology from the Deer Park Complex is from Chew *et al.* (2010). Charlestown is locality 43 and the South Mayo Trough is locality 44 on Figure 2.

Figure 15. Outcrop of the Cambrian–Ordovician rocks in NW Scotland. Ordovician rocks crop out only in Strath Suardal and the Ord window on Skye; Stronchrubie, near Loch Assynt; and in the type area of the Durness Group, around Durness. Location of figure indicated on Figure 2.

Figure 16. Stratigraphical column of Ordovician strata in the Durness area correlated with the ICS standard, Midcontinent conodont biozonation, and the Sauk sequences of the Laurentian interior. Biostratigraphical correlations are to the composite reference sections of Sweet and Tolbert (1997) and Goldman *et al.* (2014). Correlation with Sauk sequences is based on Raine and Smith (2012).

Figure 17. Conodont ranges at the Eilean Dubh–Sailmhor formation boundary in Balnakeil Bay, 1.5 km NW of Durness village. The position of the Cambrian–Ordovician boundary approximates to the formation boundary. Red dots indicate conodont samples, and the red lines mark the position of the formation boundary.

Figure 18. Parasequences, indicated by blue triangles, with subtidal bases and pale peritidal tops, thin upwards in the upper Sailmhor Formation towards the Sauk IIIB–IIIC boundary; Smoo Cave inlet, 1.7 km ESE of Durness.



TERRANES named in upper case
Subterranes named in lower case
Terrane Boundaries named in italics
line style shows age of last major displacement
post-Ordovician
intra-Ordovician
pre-Ordovician
alternative boundary
Variscan Front
C+A Coedana & Aethwy subterranes
F Fault
FS Fault System
FZ Fault Zone
Ingletonian outcrop
LAURENTIAN TERRANES
PERI-LAURENTIAN TERRANES
PERI-GONDWANAN TERRANES
Moine Thrust
Walls Boundary Fault
HEBRIDEAN
NORTHERN HIGHLANDS
Great Glen Fault
CENTRAL HIGHLANDS (GRAMPIAN)
Highland Boundary Fault
Southern Upland Fault
MIDLAND VALLEY
CENTRAL - SOUTHERN UPLANDS
Iapetus Suture
LEINSTER-LAKESMAN
Stainmore F
SOUTHERN NORTH SEA
South Craven F
Fair Head-Clew Bay Line
Navan F
South Mayo
Doon Rock Fault
CONNEMARA
Grangegeeth
Slane Fault
Lowther Lodge Fault
Bellewstown
Skirds Rock Fault
Wicklow FZ
C+A
Flamborough Head FZ
Dowsing FZ
Welsh Borderland FS
Menai Strait FS
MONIAN COMPOSITE TERRANE
'AVALON' COMPOSITE TERRANE
Dingle
Rosslare
CYMRU
FENLAND
WREKIN
CHARNWOOD
alternative positions of Variscan Front
Bristol Channel FZ
Malvern Line
microcraton eastern margin
Ballycogly mylonite zone
Courtown-Tramore line
Variscan Front
Tintagel-Exmouth line
NORTH CORNUBIAN
SOUTH CORNUBIAN
NORMANNIAN
Lizard-Dodman Thrust
Paimpol Fault
NORTH ARMORICAN
Bray Fault
palaeo-continental affinity
Laurentian
peri-Laurentian
Ganderian
Megumian or Avalonian
Cadomian or Megumian
Cadomian
uncertain

Figure 1

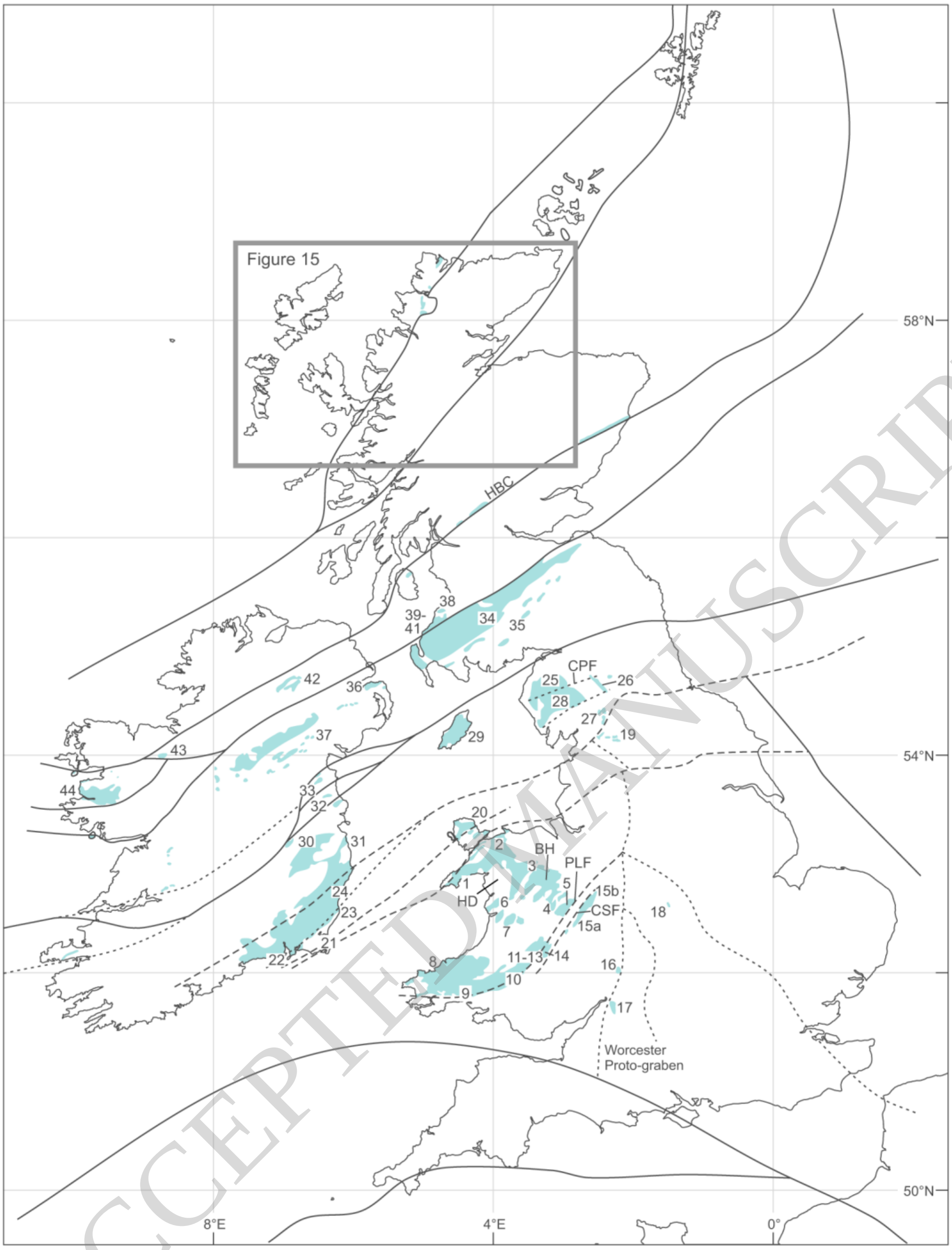


Figure 15
HBC
58°N
54°N
50°N
8°E
4°E
0°
CPF
BH
PLF
HD
CSF
Worcester
Proto-graben

Figure 2

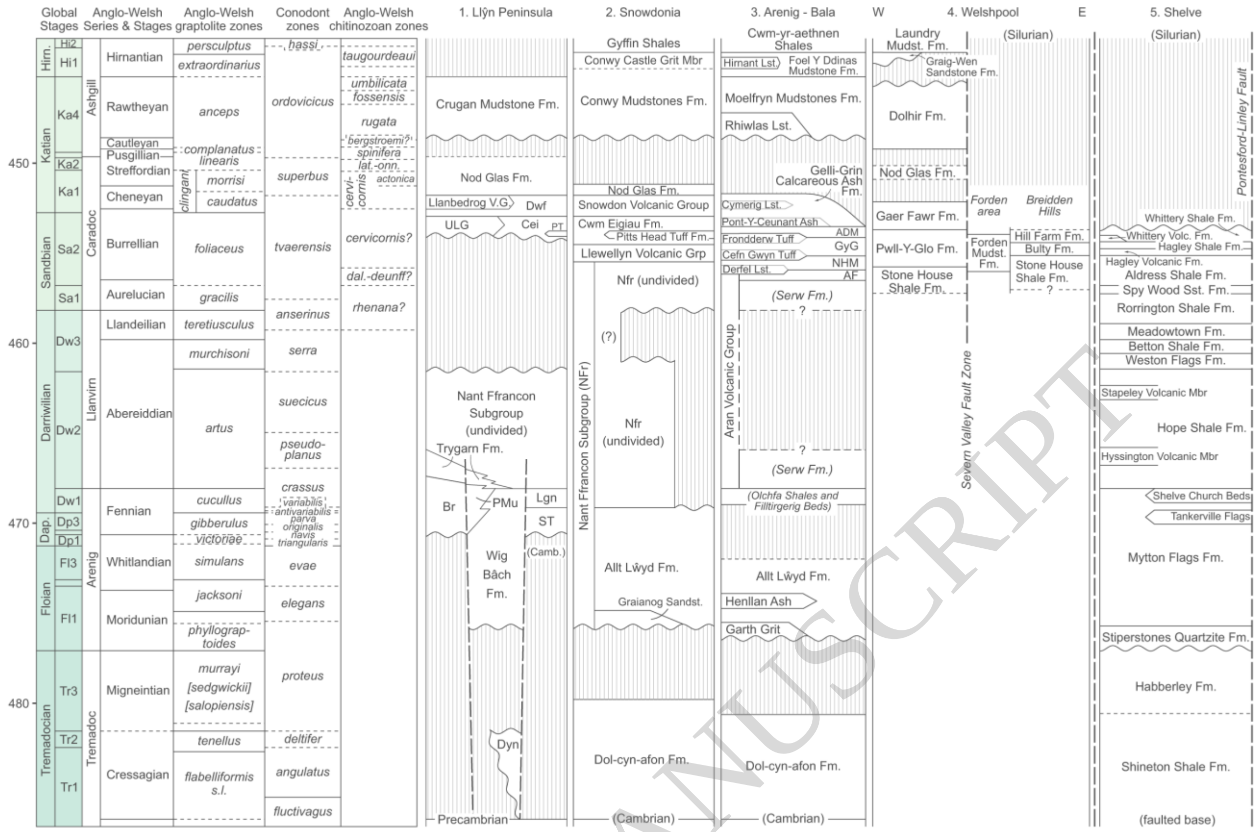

Global Stages
Anglo-Welsh Series & Stages
Anglo-Welsh graptolite zones
Conodont zones
Anglo-Welsh chitinozoan zones
1. Llŷn Peninsula
2. Snowdonia
3. Arenig - Bala
W
4. Welshpool
E
5. Shelve
Hirnantian
Rawtheyan
Cautleyan
Pusgillian
Streffordian
Cheneyan
Burrellian
Aurelucian
Llandeilian
Abereiddian
Fennian
Whitlandian
Moridunian
Migneintian
Cressagian
persculptus
extraordinarius
anceps
complanatus
linearis
morrisi
caudatus
foliaceus
gracilis
teretiusculus
murchisoni
artus
cucullus
gibberulus
victoriae
simulans
jacksoni
phyllograptoides
murrayi
[sedgwickii]
[salopiensis]
tenellus
flabelliformis s.l.
ordovicicus
superbus
tvaerensis
anserinus
serra
suecicus
pseudoplanus
crassus
evae
elegans
proteus
deltifer
angulatus
fluctivagus
Crugan Mudstone Fm.
Nod Glas Fm.
Nant Ffrancon Subgroup (undivided)
Trygarn Fm.
Wig Bâch Fm.
Precambrian
Conwy Mudstones Fm.
Snowdon Volcanic Group
Llewellyn Volcanic Grp
Allt Lŵyd Fm.
Dol-cyn-afon Fm.
(Cambrian)
Moelfryn Mudstones Fm.
Rhiwlas Lst.
Aran Volcanic Group
Allt Lŵyd Fm.
Henllan Ash
Garth Grit
Dol-cyn-afon Fm.
Dolhir Fm.
Nod Glas Fm.
Gaer Fawr Fm.
Pwll-Y-Glo Fm.
Stone House Shale Fm.
Severn Valley Fault Zone
Whittery Shale Fm.
Aldress Shale Fm.
Rorrington Shale Fm.
Meadowtown Fm.
Hope Shale Fm.
Mytton Flags Fm.
Stiperstones Quartzite Fm.
Habberley Fm.
Shineton Shale Fm.
(faulted base)
Pontesford-Linley Fault

Figure 3

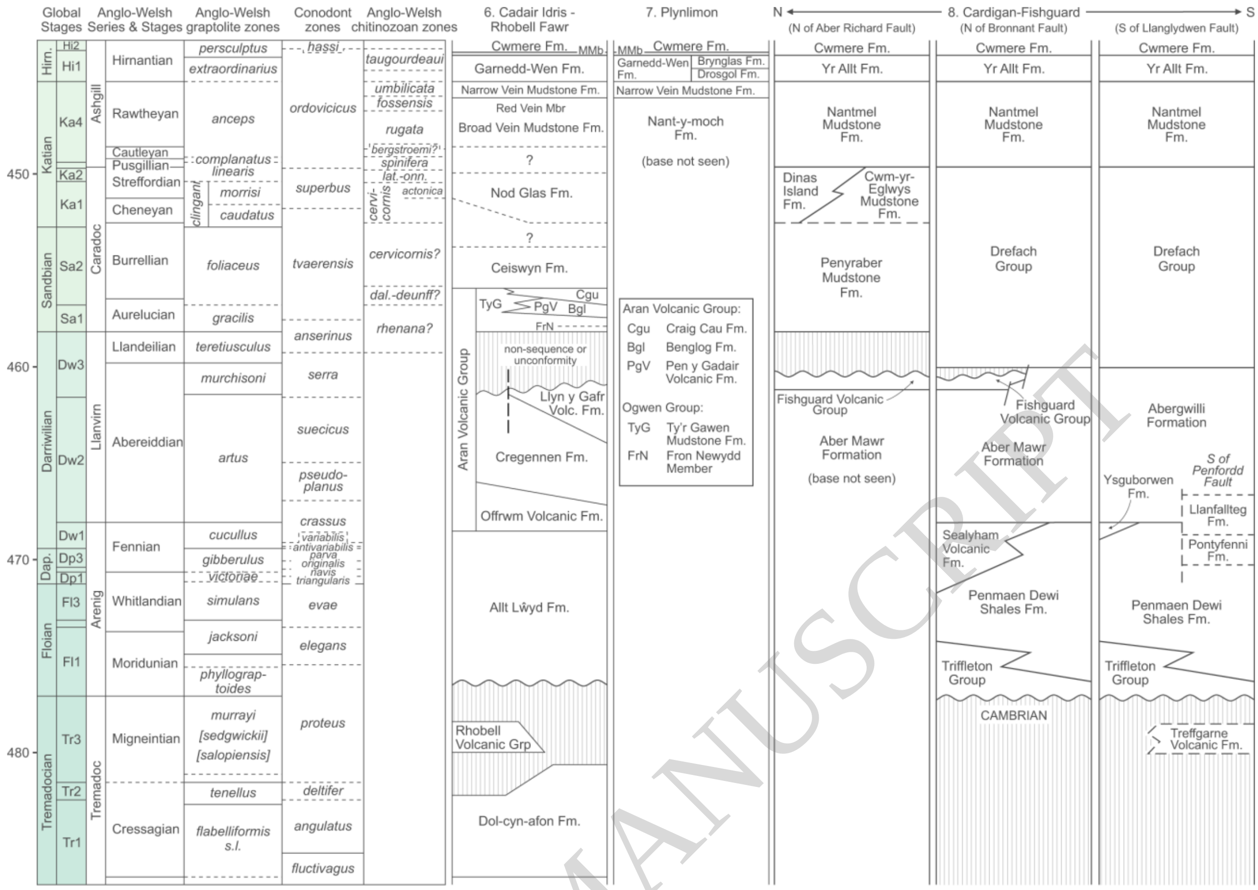


Figure 4

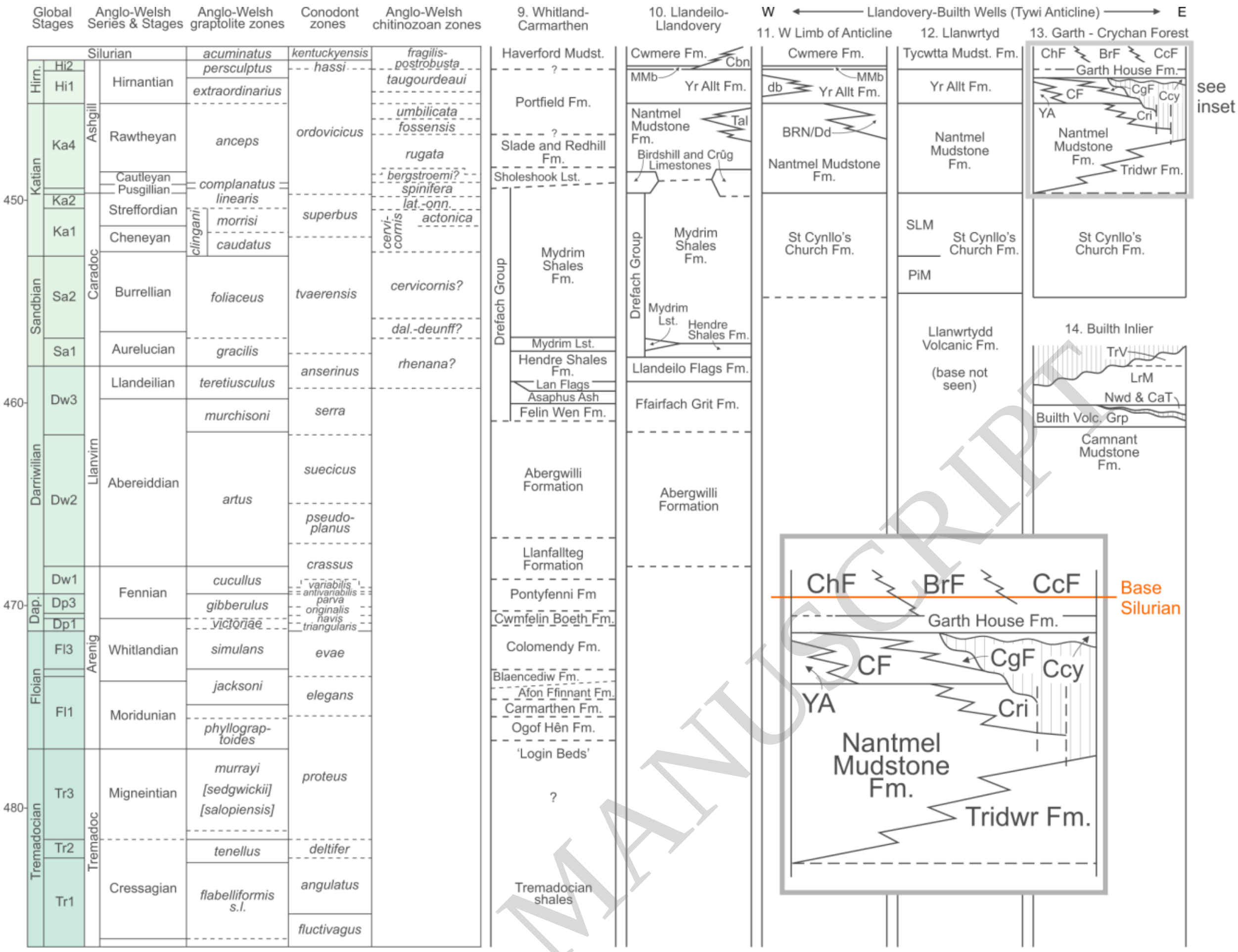


Figure 5

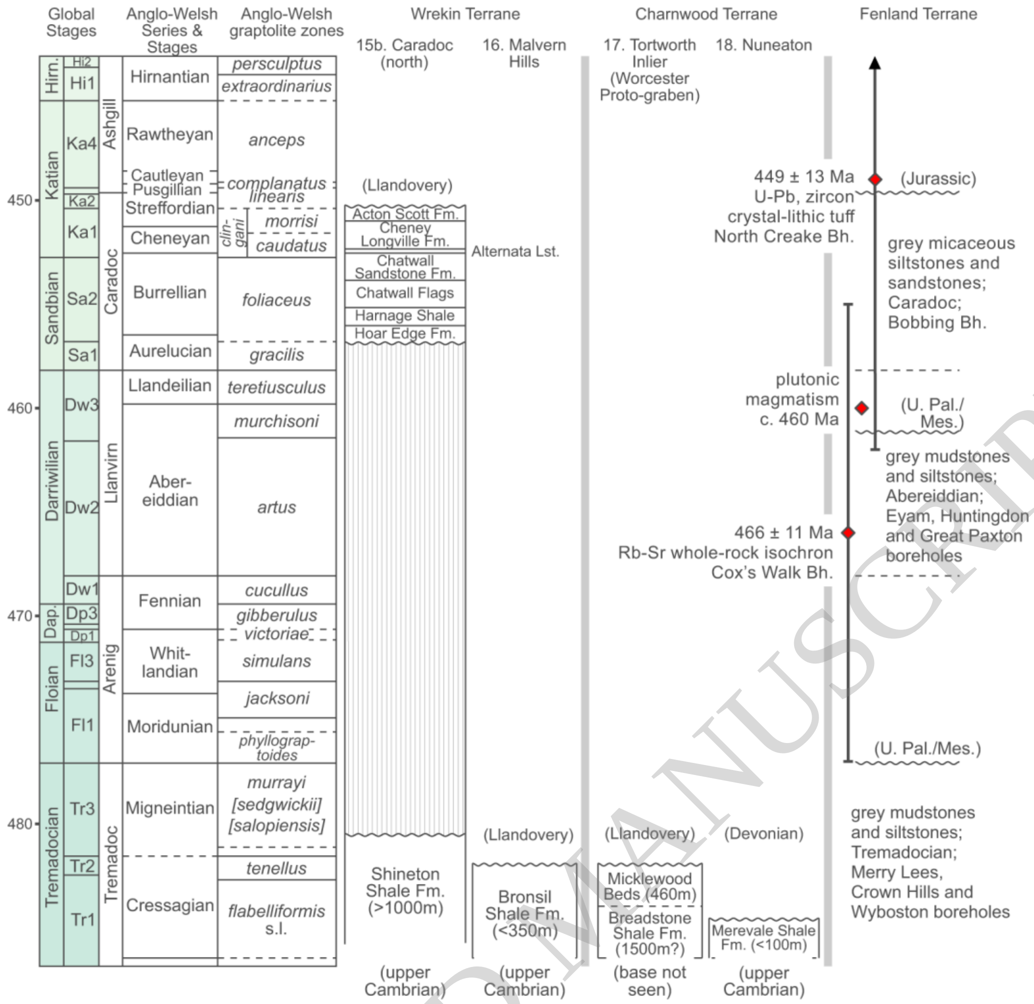
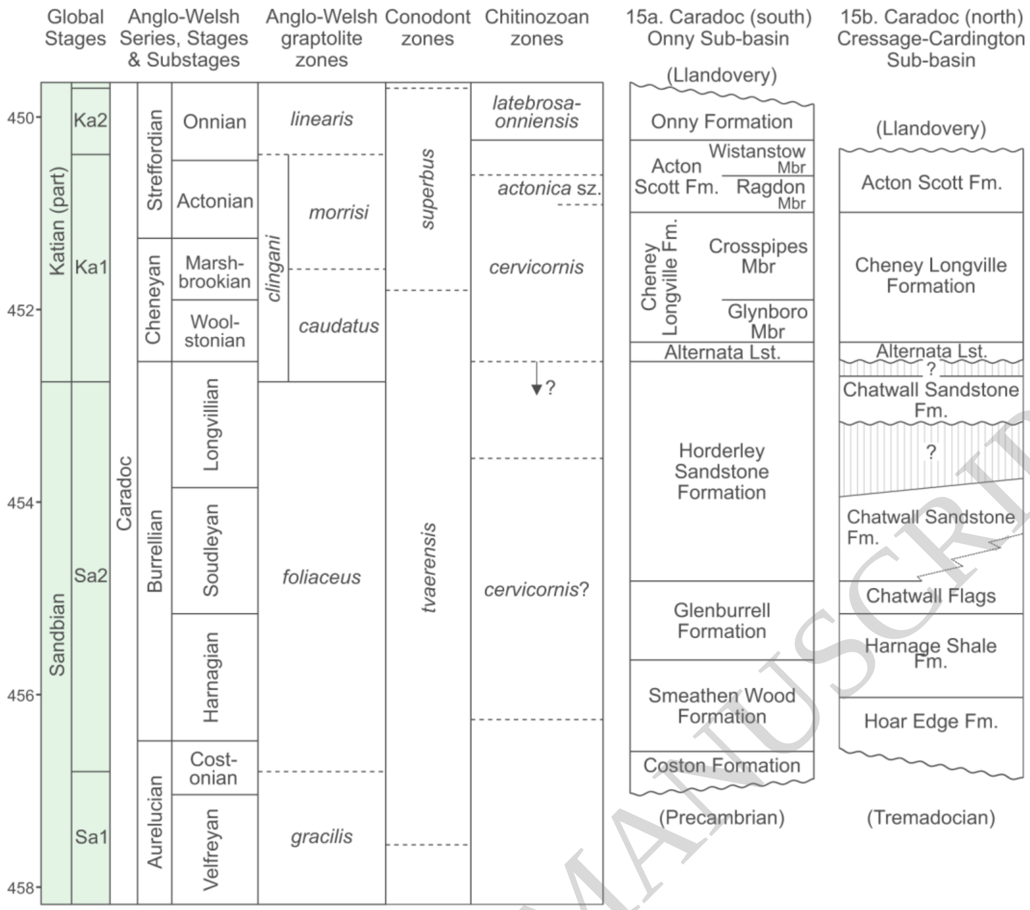

Global Stages
Anglo-Welsh Series & Stages
Anglo-Welsh graptolite zones
Wrekin Terrane
15b. Caradoc (north)
16. Malvern Hills
Charnwood Terrane
17. Tortworth Inlier (Worcester Proto-graben)
18. Nuneaton
Fenland Terrane
Hirn.
Hi2
Hi1
Katian
Ka4
Ka2
Ka1
Sandbian
Sa2
Sa1
Darriwilian
Dw3
Dw2
Dw1
Dap.
Dp3
Dp1
Floian
Fl3
Fl1
Tremadocian
Tr3
Tr2
Tr1
450
460
470
480
Ashgill
Caradoc
Llanvirn
Arenig
Tremadoc
Hirnantian
Rawtheyan
Cautleyan
Pusgillian
Streffordian
Cheneyan
Burrellian
Aurelucian
Llandeilian
Aber-eiddian
Fennian
Whit-landian
Moridunian
Migneintian
Cressagian
persculptus
extraordinarius
anceps
complanatus
linearis
clin-gani
morrisi
caudatus
foliaceus
gracilis
teretiusculus
murchisoni
artus
cucullus
gibberulus
victoriae
simulans
jacksoni
phyllograp-toides
murrayi
[sedgwickii]
[salopiensis]
tenellus
flabelliformis s.l.
(Llandovery)
Acton Scott Fm.
Cheney Longville Fm.
Chatwall Sandstone Fm.
Chatwall Flags
Harnage Shale
Hoar Edge Fm.
Alternata Lst.
Shineton Shale Fm. (>1000m)
(upper Cambrian)
(Llandovery)
Bronsil Shale Fm. (<350m)
(upper Cambrian)
(Llandovery)
Micklewood Beds (460m)
Breadstone Shale Fm. (1500m?)
(base not seen)
(Devonian)
Merevale Shale Fm. (<100m)
(upper Cambrian)
449 ± 13 Ma
U-Pb, zircon
crystal-lithic tuff
North Creake Bh.
(Jurassic)
grey micaceous siltstones and sandstones; Caradoc; Bobbing Bh.
plutonic magmatism c. 460 Ma
(U. Pal./ Mes.)
grey mudstones and siltstones; Abereiddian; Eyam, Huntingdon and Great Paxton boreholes
466 ± 11 Ma
Rb-Sr whole-rock isochron
Cox's Walk Bh.
(U. Pal./Mes.)
grey mudstones and siltstones; Tremadocian; Merry Lees, Crown Hills and Wyboston boreholes

Figure 6



Global Stages
Anglo-Welsh Series, Stages & Substages
Anglo-Welsh graptolite zones
Conodont zones
Chitinozoan zones
15a. Caradoc (south) Onny Sub-basin
15b. Caradoc (north) Cressage-Cardington Sub-basin
450
452
454
456
458
Katian (part)
Sandbian
Ka2
Ka1
Sa2
Sa1
Caradoc
Streffordian
Cheneyan
Burrellian
Aurelucian
Onnian
Actonian
Marsh-brookian
Wool-stonian
Longvillian
Soudleyan
Harnagian
Cost-onian
Velfreyan
linearis
clingani
morrisi
caudatus
foliaceus
gracilis
superbus
tvaerensis
latebrosa-onniensis
actonica sz.
cervicornis
?
cervicornis?
(Llandovery)
Onny Formation
Acton Scott Fm.
Wistanstow Mbr
Ragdon Mbr
Cheney Longville Fm.
Crosspipes Mbr
Glynboro Mbr
Alternata Lst.
Horderley Sandstone Formation
Glenburrell Formation
Smeathen Wood Formation
Coston Formation
(Precambrian)
(Llandovery)
Acton Scott Fm.
Cheney Longville Formation
Alternata Lst.
?
Chatwall Sandstone Fm.
?
Chatwall Sandstone Fm.
Chatwall Flags
Harnage Shale Fm.
Hoar Edge Fm.
(Tremadocian)

Figure 7

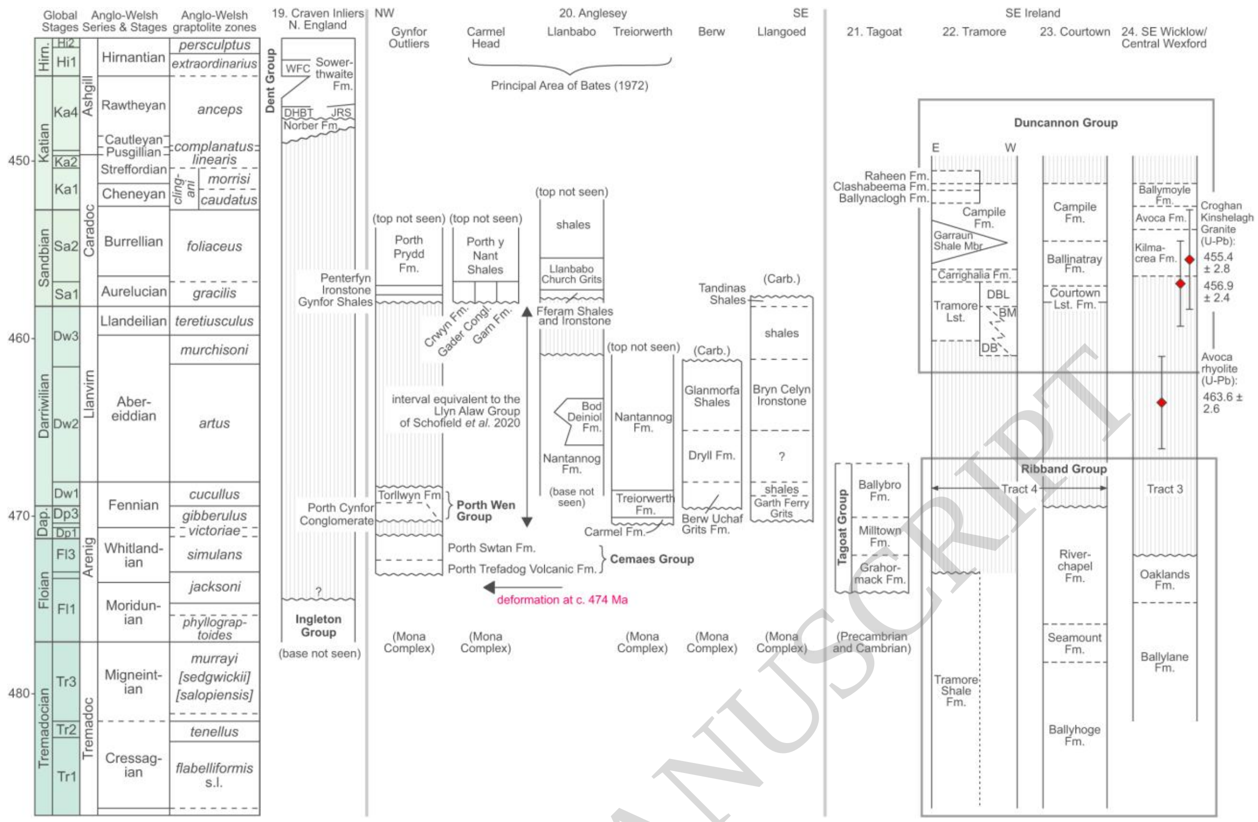
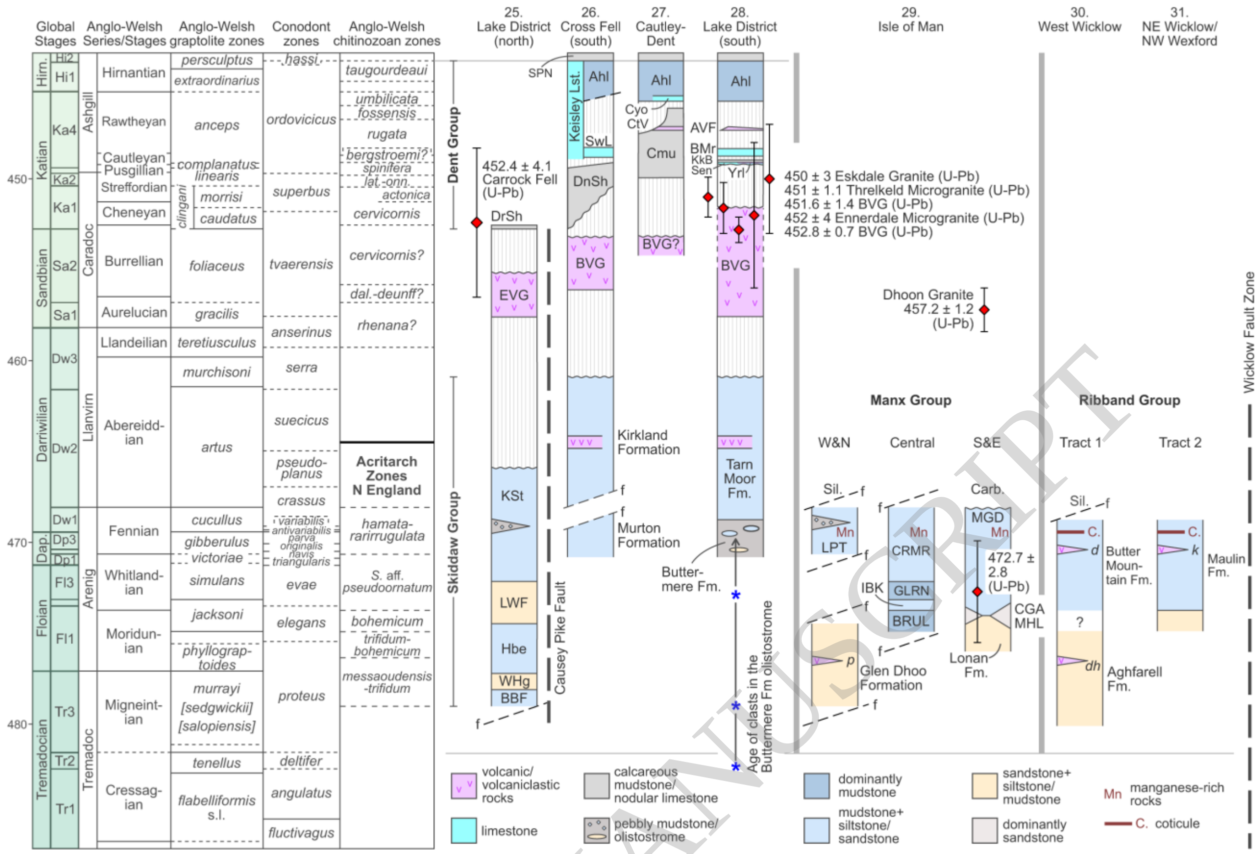
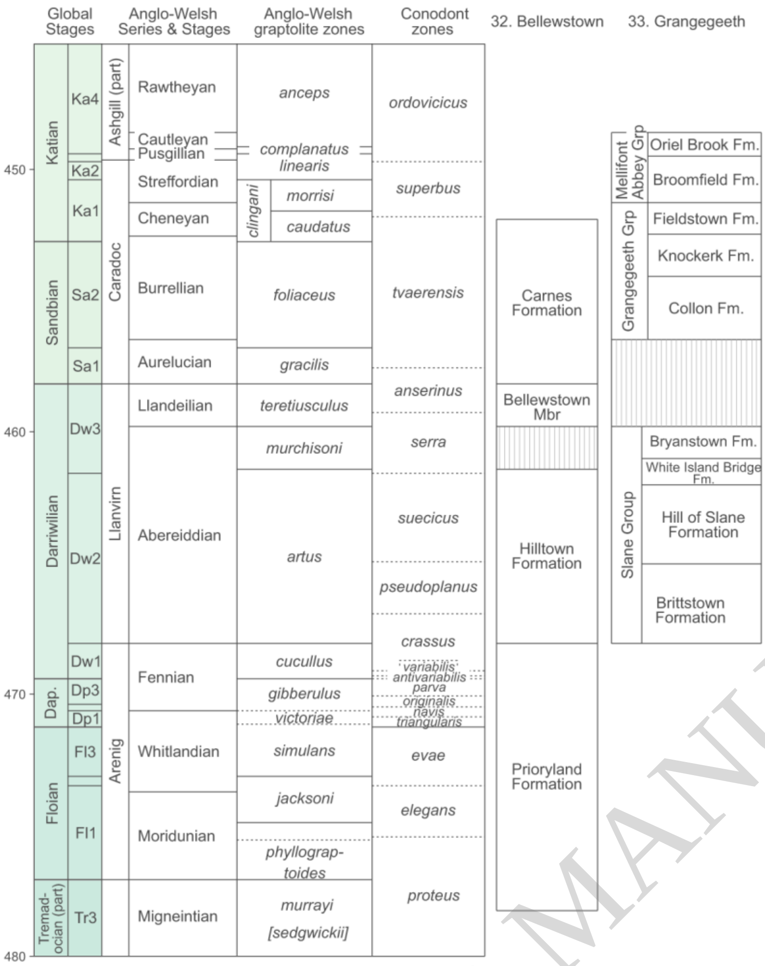


Figure 8



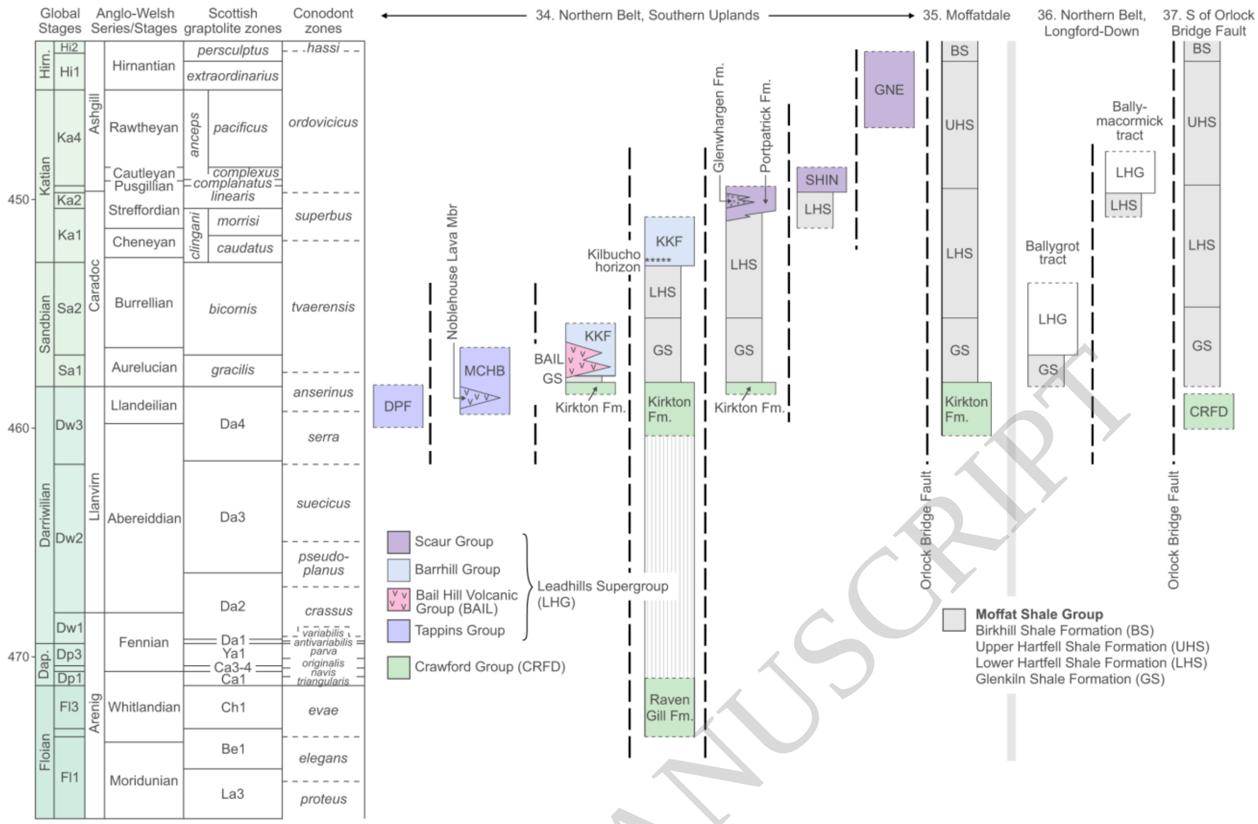
Global Stages
Anglo-Welsh Series/Stages
Anglo-Welsh graptolite zones
Conodont zones
Anglo-Welsh chitinozoan zones
25. Lake District (north)
26. Cross Fell (south)
27. Cautley-Dent
28. Lake District (south)
29. Isle of Man
30. West Wicklow
31. NE Wicklow/ NW Wexford
Dent Group
Skiddaw Group
Causey Pike Fault
452.4 ± 4.1 Carrock Fell (U-Pb)
450 ± 3 Eskdale Granite (U-Pb)
451 ± 1.1 Threlkeld Microgranite (U-Pb)
451.6 ± 1.4 BVG (U-Pb)
452 ± 4 Ennerdale Microgranite (U-Pb)
452.8 ± 0.7 BVG (U-Pb)
Dhoon Granite 457.2 ± 1.2 (U-Pb)
472.7 ± 2.8 (U-Pb)
Manx Group
Ribband Group
Wicklow Fault Zone
Kirkland Formation
Murton Formation
Tarn Moor Fm.
Buttermere Fm.
Age of clasts in the Buttermere Fm olistostrome
Glen Dhoo Formation
Lonan Fm.
Butter Mountain Fm.
Maulin Fm.
Aghfarell Fm.
Acritarch Zones N England
volcanic/ volcaniclastic rocks
limestone
calcareous mudstone/ nodular limestone
pebbly mudstone/ olistostrome
dominantly mudstone
mudstone+ siltstone/ sandstone
sandstone+ siltstone/ mudstone
dominantly sandstone
Mn manganese-rich rocks
C. coticule

Figure 9



Global Stages
Anglo-Welsh Series & Stages
Anglo-Welsh graptolite zones
Conodont zones
32. Bellewstown
33. Grangegeeth
450
460
470
480
Katian
Sandbian
Darriwilian
Dap.
Floian
Tremad-ocian (part)
Ka4
Ka2
Ka1
Sa2
Sa1
Dw3
Dw2
Dw1
Dp3
Dp1
Fl3
Fl1
Tr3
Ashgill (part)
Caradoc
Llanvirn
Arenig
Rawtheyan
Cautleyan
Pusgillian
Streffordian
Cheneyan
Burrellian
Aurelucian
Llandeilian
Abereiddian
Fennian
Whitlandian
Moridunian
Migneintian
anceps
complanatus
linearis
clingani
morrisi
caudatus
foliaceus
gracilis
teretiusculus
murchisoni
artus
cucullus
gibberulus
victoriae
simulans
jacksoni
phyllograp-toides
murrayi
[sedgwickii]
ordovicicus
superbus
tvaerensis
anserinus
serra
suecicus
pseudoplanus
crassus
variabilis
antivariabilis
parva
originalis
navis
triangularis
evae
elegans
proteus
Carnes Formation
Bellewstown Mbr
Hilltown Formation
Prioryland Formation
Mellifont Abbey Grp
Oriel Brook Fm.
Broomfield Fm.
Grangegeeth Grp
Fieldstown Fm.
Knockerk Fm.
Collon Fm.
Slane Group
Bryanstown Fm.
White Island Bridge Fm.
Hill of Slane Formation
Brittstown Formation

Figure 10



Global Stages
Anglo-Welsh Series/Stages
Scottish graptolite zones
Conodont zones
34. Northern Belt, Southern Uplands
35. Moffatdale
36. Northern Belt, Longford-Down
37. S of Orlock Bridge Fault
Hirnantian
Rawtheyan
Cautleyan
Pusgillian
Streffordian
Cheneyan
Burrellian
Aurelucian
Llandeilian
Abereiddian
Fennian
Whitlandian
Moridunian
persculptus
extraordinarius
pacificus
complexus
complanatus
linearis
morrisi
caudatus
bicornis
gracilis
hassi
ordovicicus
superbus
tvaerensis
anserinus
serra
suecicus
pseudo-planus
crassus
evae
elegans
proteus
Noblehouse Lava Mbr
Kilbucho horizon
Glenwhargen Fm.
Portpatrick Fm.
Kirkton Fm.
Raven Gill Fm.
Orlock Bridge Fault
Ballygrot tract
Bally-macormick tract
Scaur Group
Barrhill Group
Bail Hill Volcanic Group (BAIL)
Tappins Group
Leadhills Supergroup (LHG)
Crawford Group (CRFD)
Moffat Shale Group
Birkhill Shale Formation (BS)
Upper Hartfell Shale Formation (UHS)
Lower Hartfell Shale Formation (LHS)
Glenkiln Shale Formation (GS)

Figure 11

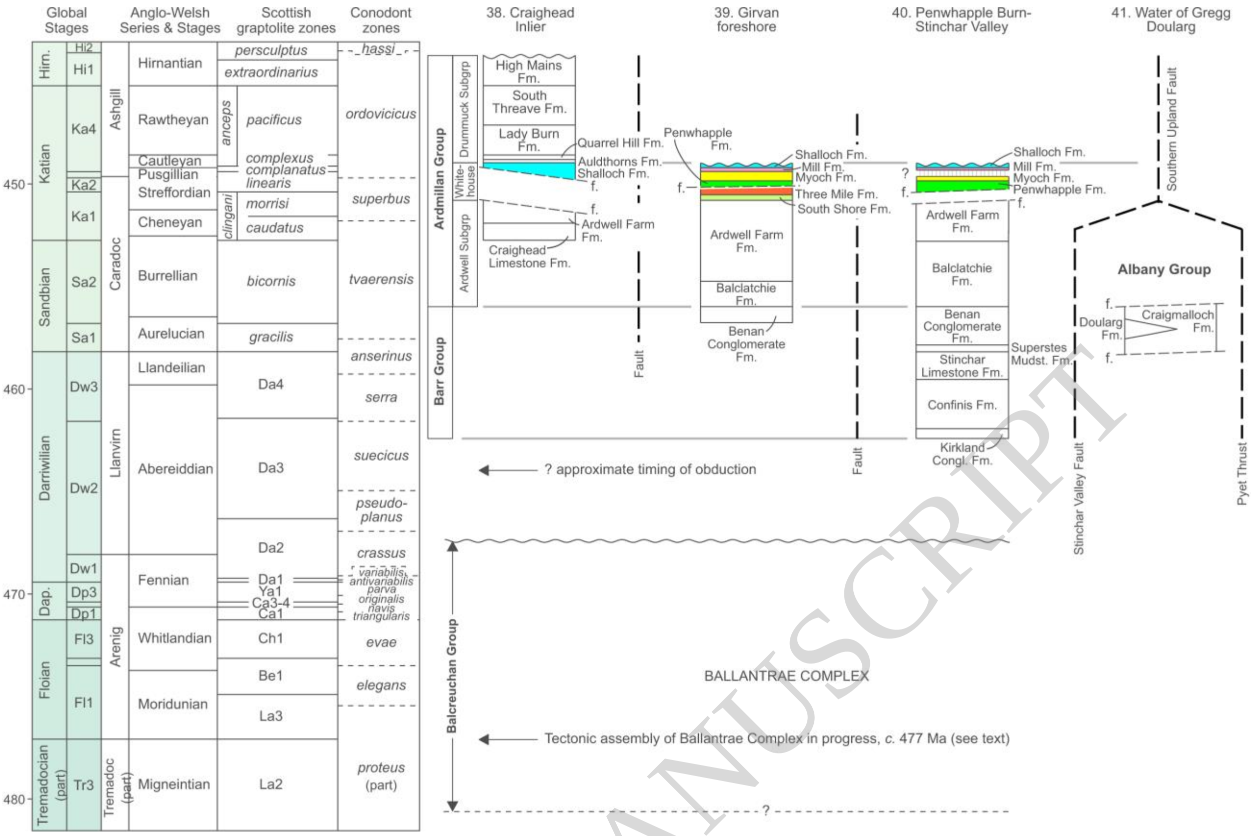

Global Stages
Anglo-Welsh Series & Stages
Scottish graptolite zones
Conodont zones
38. Craighead Inlier
39. Girvan foreshore
40. Penwhapple Burn-Stinchar Valley
41. Water of Gregg Doularg
Ardmillan Group
Barr Group
Balcreuchan Group
Albany Group
BALLANTRAE COMPLEX
? approximate timing of obduction
Tectonic assembly of Ballantrae Complex in progress, c. 477 Ma (see text)
Southern Upland Fault
Stinchar Valley Fault
Pyet Thrust
Fault

Figure 12

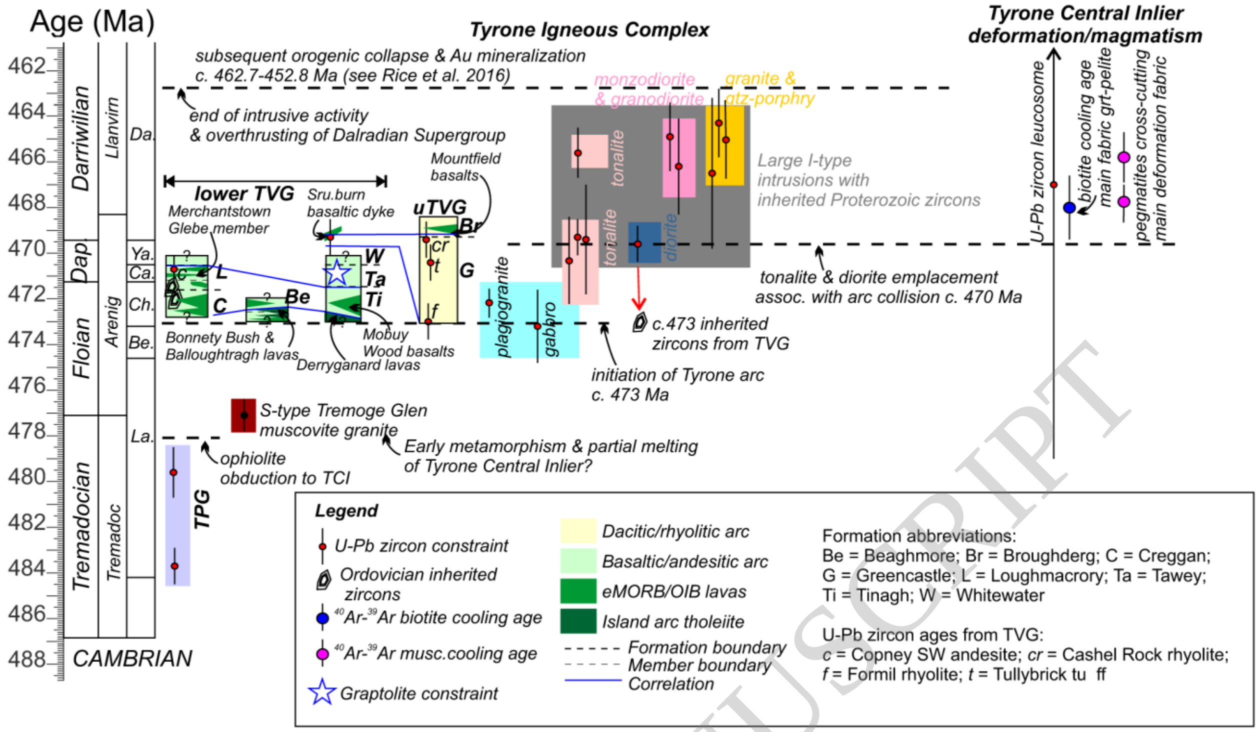
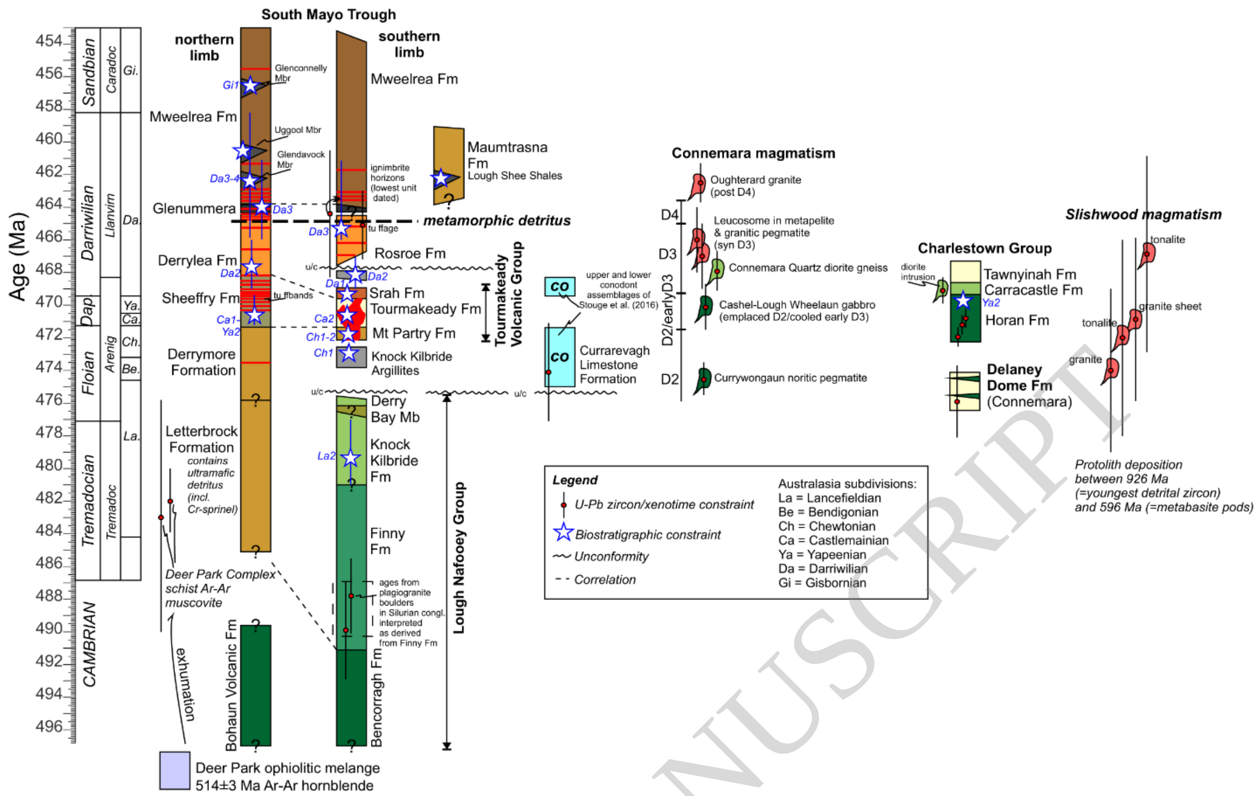
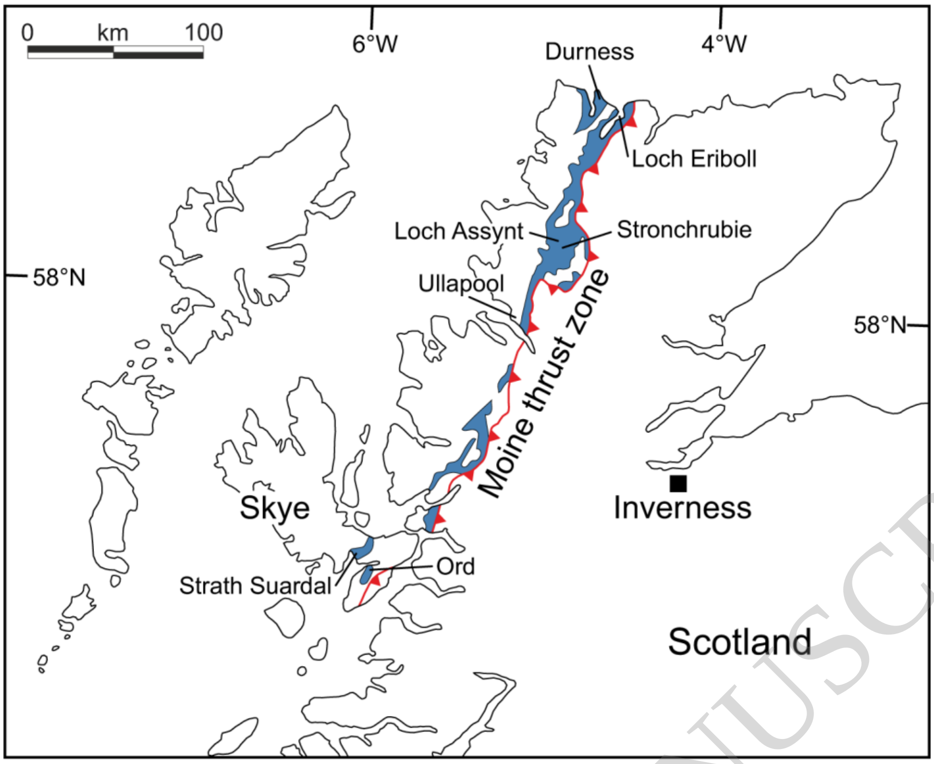

Age (Ma)
Tyrone Igneous Complex
Tyrone Central Inlier deformation/magmatism
subsequent orogenic collapse & Au mineralization c. 462.7-452.8 Ma (see Rice et al. 2016)
end of intrusive activity & overthrusting of Dalradian Supergroup
lower TVG
uTVG
Merchantstown Glebe member
Sru.burn basaltic dyke
Mountfield basalts
Bonnety Bush & Balloughtragh lavas
Derryganard lavas
Mobuy Wood basalts
monzodiorite & granodiorite
granite & qtz-porphry
tonalite
diorite
plagiogranite
gabbro
Large I-type intrusions with inherited Proterozoic zircons
tonalite & diorite emplacement assoc. with arc collision c. 470 Ma
c.473 inherited zircons from TVG
initiation of Tyrone arc c. 473 Ma
S-type Tremoge Glen muscovite granite
Early metamorphism & partial melting of Tyrone Central Inlier?
ophiolite obduction to TCI
TPG
U-Pb zircon leucosome
biotite cooling age main fabric grt-pelite
pegmatites cross-cutting main deformation fabric
Darriwilian
Dap
Floian
Tremadocian
CAMBRIAN
Llanvirn
Arenig
Tremadoc
Legend
U-Pb zircon constraint
Ordovician inherited zircons
40Ar-39Ar biotite cooling age
40Ar-39Ar musc.cooling age
Graptolite constraint
Dacitic/rhyolitic arc
Basaltic/andesitic arc
eMORB/OIB lavas
Island arc tholeiite
Formation boundary
Member boundary
Correlation
Formation abbreviations:
Be = Beaghmore; Br = Broughderg; C = Creggan;
G = Greencastle; L = Loughmacrory; Ta = Tawey;
Ti = Tinagh; W = Whitewater
U-Pb zircon ages from TVG:
c = Copney SW andesite; cr = Cashel Rock rhyolite;
f = Formil rhyolite; t = Tullybrick tuff

Figure 13



South Mayo Trough
northern limb
southern limb
Age (Ma)
Sandbian
Darriwilian
Dap
Floian
Tremadocian
CAMBRIAN
Caradoc
Llanvirn
Arenig
Tremadoc
Mweelrea Fm
Glenconnelly Mbr
Uggool Mbr
Glendavock Mbr
Glenummera
Derrylea Fm
Sheeffry Fm
Derrymore Formation
Letterbrock Formation
Deer Park Complex schist Ar-Ar muscovite
exhumation
Bohaun Volcanic Fm
Deer Park ophiolitic melange 514±3 Ma Ar-Ar hornblende
Maumtrasna Fm
Lough Shee Shales
metamorphic detritus
Rosroe Fm
Srah Fm
Tourmakeady Fm
Mt Partry Fm
Knock Kilbride Argillites
Tourmakeady Volcanic Group
Derry Bay Mb
Knock Kilbride Fm
Finny Fm
Bencorragh Fm
Lough Nafooey Group
Currarevagh Limestone Formation
Connemara magmatism
Oughterard granite (post D4)
Leucosome in metapelite & granitic pegmatite (syn D3)
Connemara Quartz diorite gneiss
Cashel-Lough Wheelaun gabbro (emplaced D2/cooled early D3)
Currywongaun noritic pegmatite
Charlestown Group
Tawnyinah Fm
Carracastle Fm
Horan Fm
Delaney Dome Fm (Connemara)
Slishwood magmatism
Protolith deposition between 926 Ma (=youngest detrital zircon) and 596 Ma (=metabasite pods)
Legend
U-Pb zircon/xenotime constraint
Biostratigraphic constraint
Unconformity
Correlation
Australasia subdivisions:
La = Lancefieldian
Be = Bendigonian
Ch = Chewtonian
Ca = Castlemainian
Ya = Yapeenian
Da = Darriwilian
Gi = Gisbornian

Figure 14



0 km 100
6°W
4°W
58°N
58°N
Durness
Loch Eriboll
Loch Assynt
Stronchrubie
Ullapool
Moine thrust zone
Skye
Inverness
Strath Suardal
Ord
Scotland

Figure 15

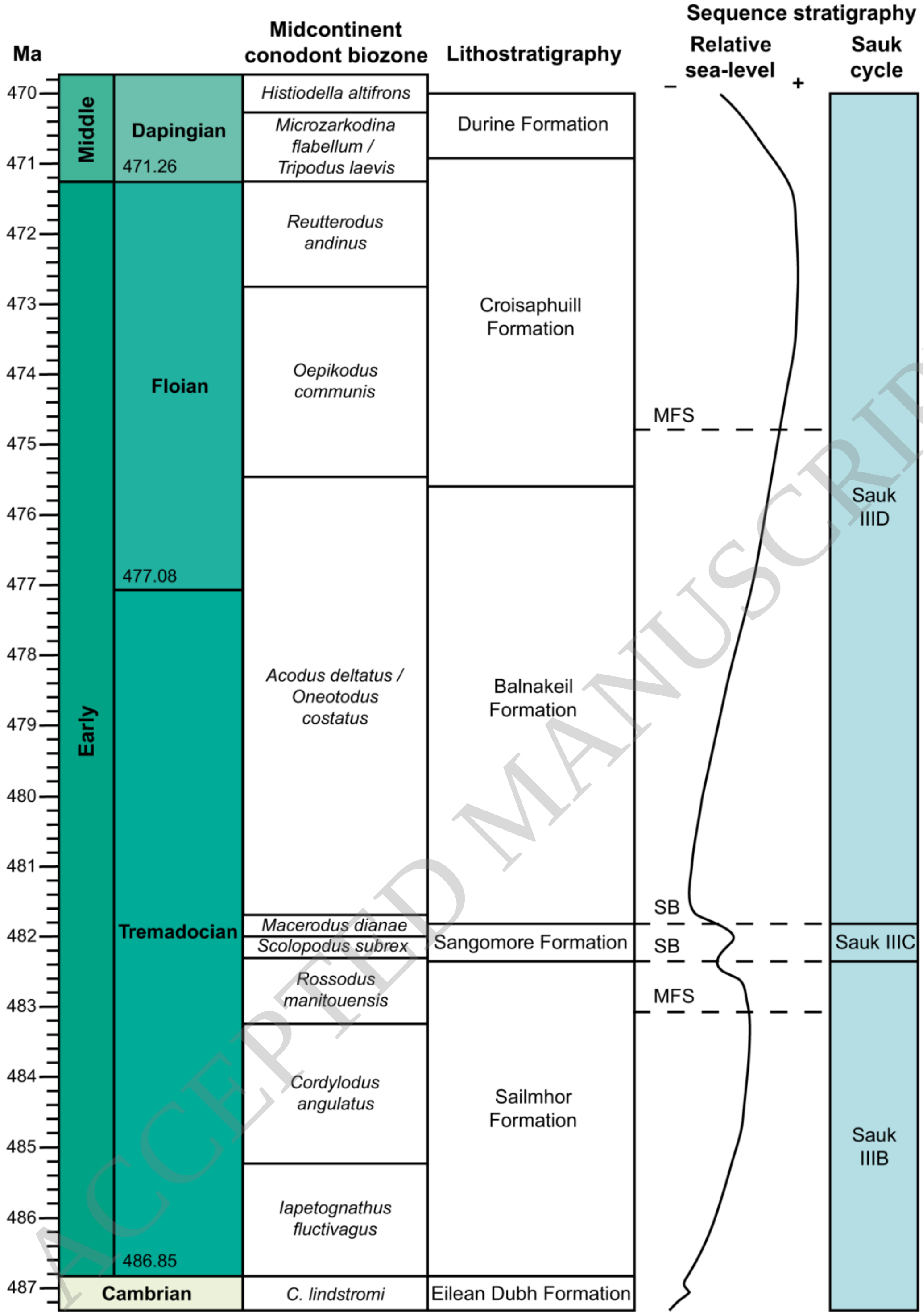

Ma
Midcontinent conodont biozone
Lithostratigraphy
Sequence stratigraphy
Relative sea-level
–
+
Sauk cycle
470
471
472
473
474
475
476
477
478
479
480
481
482
483
484
485
486
487
Middle
Early
Dapingian
471.26
Floian
477.08
Tremadocian
486.85
Cambrian
Histiodella altifrons
Microzarkodina flabellum / Tripodus laevis
Reutterodus andinus
Oepikodus communis
Acodus deltatus / Oneotodus costatus
Macerodus dianae
Scolopodus subrex
Rossodus manitouensis
Cordylodus angulatus
Iapetognathus fluctivagus
C. lindstromi
Durine Formation
Croisaphuill Formation
Balnakeil Formation
Sangomore Formation
Sailmhor Formation
Eilean Dubh Formation
MFS
SB
SB
MFS
Sauk IIID
Sauk IIIC
Sauk IIIB

Figure 16

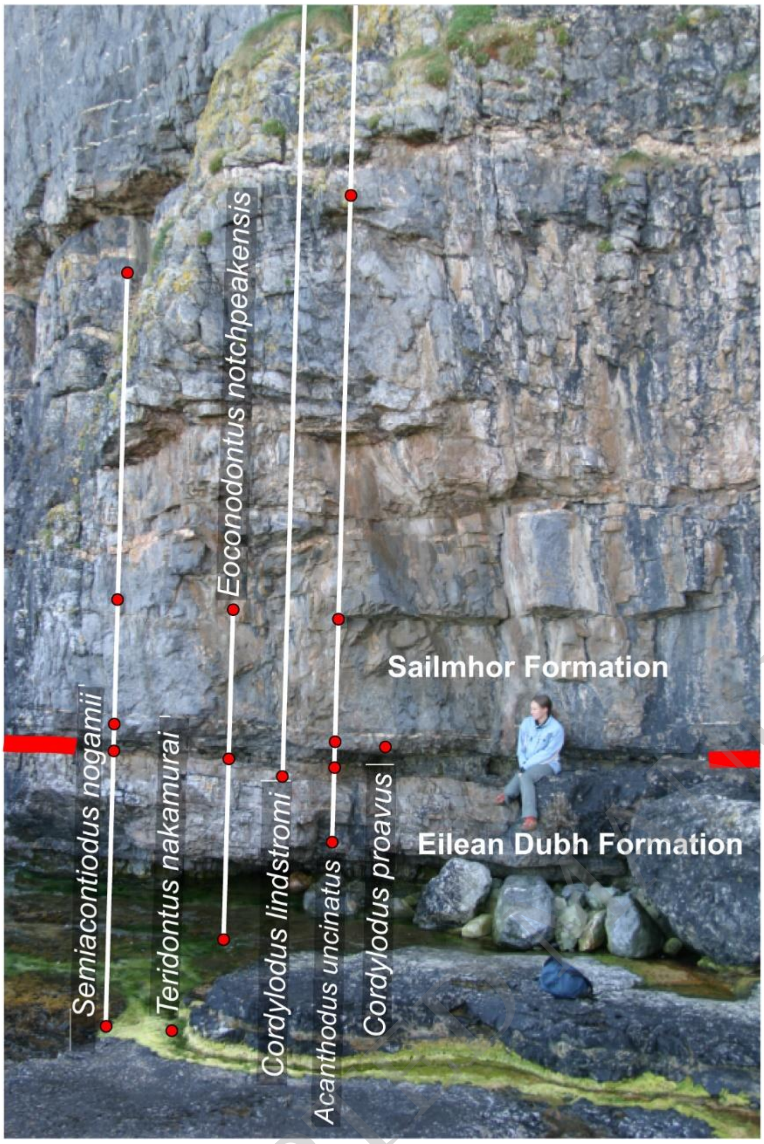

Eoconodontus notchpeakensis
Sailmhor Formation
Semiacontiodus nogamii
Teridontus nakamurai
Cordylodus lindstromi
Acanthodus uncinatus
Cordylodus proavus
Eilean Dubh Formation

Figure 17

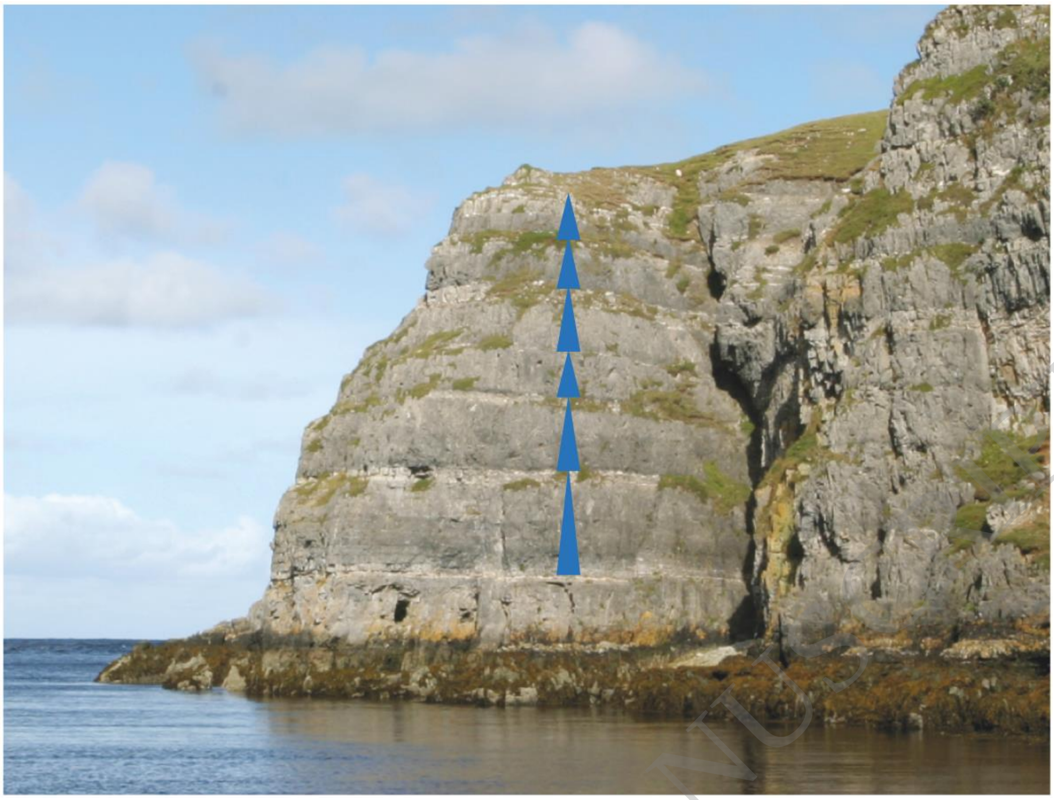


Figure 18