Late Pleistocene sediments, landforms and events in Scotland: a review of the terrestrial stratigraphic record

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ABSTRACT: Lithostratigraphical studies coupled with the development of new dating methods has led to significant progress in understanding the Late Pleistocene terrestrial record in Scotland. Systematic analysis and re-evaluation of key localities have provided new insights into the complexity of the event stratigraphy in some regions and the timing of Late Pleistocene environmental changes, but few additional critical sites have been described in the past 25 years. The terrestrial stratigraphic record remains important for understanding the timing, sequence and patterns of glaciation and deglaciation during the last glacial/interglacial cycle. Former interpretations of ice-free areas in peripheral areas during the Last Glacial Maximum (LGM) are inconsistent with current stratigraphic and dating evidence. Significant challenges remain to determine events and patterns of glaciation during the Early and Middle Devensian, particularly in the context of offshore evidence and ice sheet modelling that indicate significant build-up of ice throughout much of the period. The terrestrial evidence broadly supports recent reconstructions of a highly dynamic and climate-sensitive British–Irish Ice Sheet (BIIS), which apparently reached its greatest thickness in Scotland between 30 and 27 ka, before the global LGM. A thick (relative to topography) integrated ice sheet reaching the shelf edge with a simple ice-divide structure was replaced after the LGM by a much thinner one comprising multiple dispersion centres and a more complex flow structure.

KEY WORDS: chronostratigraphy, ice sheet, last glacial–interglacial cycle, lithostratigraphy.

The Pleistocene covers the greater part of the Quaternary period, from 2.58 Ma to 11.7 ka (Gibbard & Lewin 2016). Its defining characteristic is a series of climatic cycles resulting in repeated glaciations that had a profound but variable effect in shaping the modern landscape of Scotland, adding cumulatively to the effects of longer-term landscape evolution (Hall *et al.* 2018). During the glacial stages, mountain glaciers developed in the uplands and ice sheets extended periodically out across the continental shelf and into the North Sea basin (Stewart *et al.* 2019, and references therein). During the intervening interglacial stages, temperate climatic conditions prevailed, leading to more equable environments. The record of these climatic and environmental changes is documented in the terrestrial sedimentary record, typically by a variety of glacial and periglacial deposits, organic deposits and palaeosols. However, because much of Scotland lay close to ice accumulation centres, successive glaciations have generally removed traces of earlier events. Hence, there is dominance of the last glacial cycle and, especially, its final deglaciation, in the terrestrial stratigraphic record. Exceptionally, in some areas, the former presence of cold-based ice appears to have allowed the local preservation of relatively long sedimentary sequences, as, for example, in NE Scotland (Hall *et al.* 2018). The preservation of earlier events is fragmentary, both in space and time. Nevertheless, the terrestrial evidence represents a significant and accessible record.

The glacial landform and sedimentary archives, including the BRITICE database of landforms and features related to the last British–Irish Ice Sheet (BIIS) (Clark *et al.* 2004, 2018; Evans *et al.* 2005), provide good opportunities to elucidate the history of Quaternary glaciation in Britain, and particularly for understanding the dynamics of the last BIIS, its response to the changing climate and the links between ocean circulation and climate change. The BIIS was largely marine-based (Stewart *et al.* 2019), and because of the extreme maritime location of the British Isles, it is considered to have been highly sensitive to climate changes (Hibbert *et al.* 2010; Thierens *et al.* 2012).

The aim of this paper is to review the Late Pleistocene terrestrial stratigraphy of Scotland over the time period 130–14.7 ka and spanning the last glacial–interglacial cycle (Lisiecki & Raymo 2005). We first summarise recent advances in understanding the lithostratigraphic and chronostratigraphic framework for Scotland over the last 25 years since the publication of the *Quaternary of Scotland* Geological Conservation Review volume (Gordon & Sutherland 1993a), then review the evidence from a number of critical regions. Finally, we consider outstanding questions, new directions and challenges. The focus of the paper on Late Pleistocene terrestrial stratigraphy complements the accompanying review of the terrestrial record of Early and Middle Pleistocene environments, landforms and sediments in Scotland (Hall *et al.* 2018), offshore stratigraphy

Figure 1 Summary of Late Quaternary chronostratigraphic stages (Great Britain) and their correlation with MIS (after Lisiecki & Raymo 2005; Bradwell *et al.* 2008a; Otvos 2015). The approximate timings of the LLGM (*ca*.30–27 ka) (Ballantyne & Small 2018) and global LGM (23–21 ka) (Hughes *et al.* 2016) are shown. Note that some terrestrial stage boundaries are not coincident with MIS boundaries. Abbreviations: DS = Dimlington Stadial; WI = Windermere Interstadial; LLS = Loch Lomond Stadial.

(Stewart *et al.* 2019) and the geomorphological evidence for the last Scottish Ice Sheet and its deglaciation (Ballantyne & Small 2018). Lateglacial landforms and landscape evolution, including evidence relating to the glaciation of Scotland during the Loch Lomond Stadial (Younger Dryas), are reviewed by Ballantyne (2018), whereas evidence for sea level change is reviewed by Smith *et al.* (2018).

Radiocarbon ages are given as '[14]C ka BP' (thousands of years before present). Calibrated ages are quoted as 'cal ^{14}C ka BP'.

1. The Quaternary stratigraphic framework

Onshore Quaternary deposits are classified and correlated principally using lithostratigraphy, chronostratigraphy and geochronometry. The first is based on the observable physical characteristics of the sediments and their relative stratigraphic positions; the second requires establishing the time relationships between different sedimentary units in relative terms and through numerical dating methods (geochronometry). Climatostratigraphy is also used where climatic changes are represented in the deposits. In some cases, deposits may also be classified according to their biostratigraphy, or microfossil content; while in others, geomorphological subdivision may be possible using morphostratigraphy, which incorporates the landform record. Periglacial and organic sediments are rarely preserved within the Pleistocene sequence, but are particularly important stratigraphical markers providing evidence of ice-free episodes. A key element in stratigraphical analysis is the

designation of type sections and type localities establishing named stratigraphic units, which provide a common standard of reference for their identification (Bowen 1999; McMillan & Merritt 2012). Details of all units referred to in the following are fully documented by McMillan *et al.* (2011a).

1.1. Climatostratigraphy and chronostratigraphy
Quaternary stratigraphy was revolutionised during the latter part of the 20th Century by the recovery of deep-sea cores, which provided continuous sedimentary records for the whole of the Quaternary. A series of Marine Isotope Stages (MIS), based on the oxygen isotope composition of the microfaunas contained in the cores (Bowen 1978; Lisiecki & Raymo 2005; Thierens *et al.* 2012; Railsback *et al.* 2015), now provides a reference climatostratigraphic framework for the correlation of Pleistocene terrestrial sediments and events in Scotland (Hall *et al.* 2018, fig. 1). In the conventional British chronostratigraphic scheme, the Late Pleistocene spans the Ipswichian (∼130–116 ka) and Devensian (∼116–11.7 ka) stages (Fig. 1), corresponding to MIS 5e and MIS 5d-2, respectively (Lisiecki & Raymo 2005). The Devensian is subdivided into Early Devensian (∼109–57 ka) (MIS 5d-4), Middle Devensian (∼57–31 ka) (∼MIS 3) and Late Devensian (∼31–11.7 ka) (∼MIS 2). The Late Devensian includes the Dimlington Stadial (∼31–14.7 ka) (Rose 1989), the Lateglacial or Windermere Interstadial (14.7–12.9 ka) and the Loch Lomond Stadial (12.9–11.7 ka). The Lateglacial period, or Last Glacial Transition, is generally acknowledged to have occurred between *ca*.21 ka and 11.7 ka (Lowe & Walker 2015).
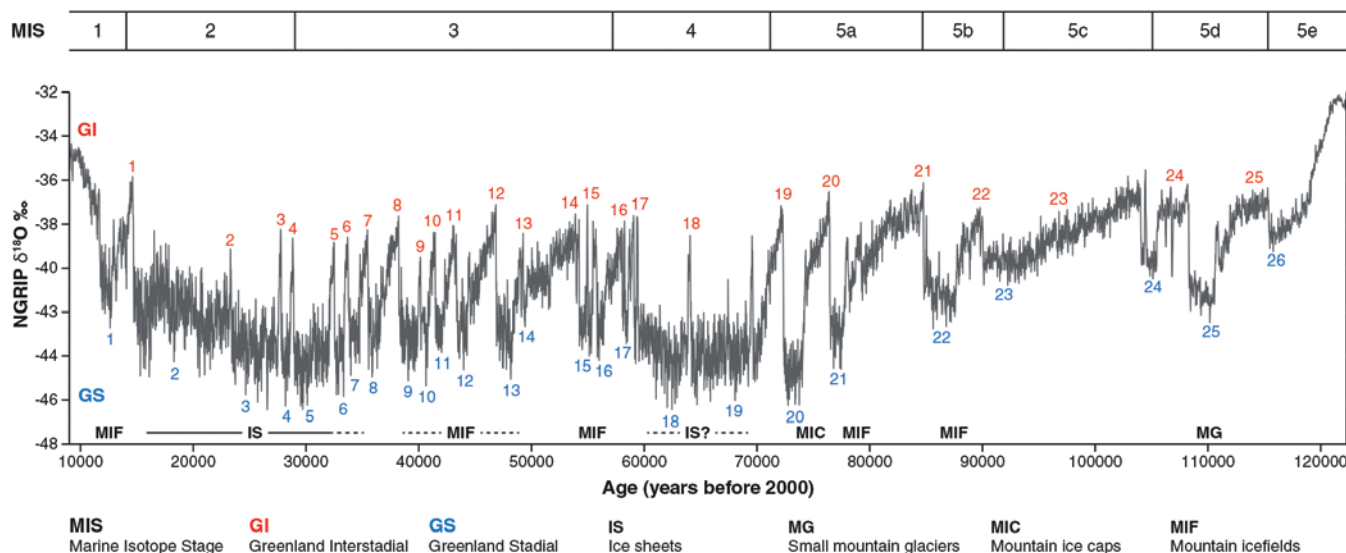
Figure 2 Greenland oxygen isotope record from the North Greenland Ice Core Project (NGRIP) ice core (NGRIP Members 2004) and the main event stratigraphy (Rasmussen *et al.* 2014). The age scale is based on the GIC05 age model as plotted by Abbott *et al.* (2012). (Reprinted from *Quaternary Science Reviews*, **36**, 59–77, P. M. Abbott *et al.* 'A detailed framework of marine Isotope Stages 4 and 5 volcanic events recorded in two Greenland ice-cores', Copyright (2012), with permission from Elsevier.) MIS boundaries are based on Lisiecki & Raymo (2005), Railsback *et al.* (2015) and Otvos (2015). The indicative presence of glaciers in Scotland is adapted from Clapperton (1997). Abbreviations: MIS: Marine Isotope Stage; GI = Greenland Interstadials; GS = Greenland Stadials; MIF = mountain icefields; IS = ice sheets; MIC = mountain ice caps; MG = small mountain glaciers.

There are different interpretations of the timing of the Last Glacial Maximum (LGM) globally, regionally and locally, according to different indicators – the timing of global sea level minimum, the oxygen isotope signal in marine records and terrestrial records (Clark *et al.* 2009; Hughes *et al.* 2013; Lambeck *et al.* 2014; Hughes & Gibbard 2015). Different sectors of the last BIIS apparently reached their maximum extent asynchronously (Ballantyne & Small 2018). The estimate of ∼30–27 ka for ice reaching the Atlantic Shelf (Ballantyne & Small 2018) is earlier than the timing of some estimates of 26–21 ka (Clark *et al.* 2009), but is in line with the time-slice reconstruction of 27–26 ka for this sector from Hughes *et al.* (2016) (see also Section 3.2). In line with Ballantyne & Small (2018), we use the abbreviation 'LGM' to denote the last global glacial maximum of 26.5–19 ka (Clark *et al.* 2009) and 'LLGM' (Local Last Glacial Maximum) to refer to the maximum extent of different sectors of the last BIIS (Fig. 1), some of which apparently reached their outermost limits millennia before others.

In a second major development, analyses of the stable-oxygen isotope signals in cores drilled through the Greenland Ice Sheet have enabled the interpretation of continuous records of climate change, defining a series of stadials and interstadials extending back to the last (Ipswichian) interglacial (Lowe & Walker 2015). Because of the high stratigraphic and temporal resolution, these records form the basis for an event stratigraphy and a framework for the whole of the last glacial–interglacial cycle for the North Atlantic region (Fig. 2) (Rasmussen *et al.* 2014; see Section 2).

The development and application of new dating techniques applied to terrestrial materials has also significantly improved the temporal resolution of onshore Quaternary stratigraphy and the identification of events (Lowe & Walker 2015). In particular, surface exposure dating using terrestrial cosmogenic nuclides (cosmogenic dating) has greatly enhanced the understanding and timing of the pattern of growth and decay of the last ice sheet (e.g., Ballantyne 2010; Ballantyne & Small 2018). Nevertheless, it is clear that only a fragmentary record of the Quaternary is preserved in terrestrial deposits.

1.2. Lithostratigraphy

The traditional approach of establishing contact relationships, sedimentology and lithostratigraphical correlations of named units has helped to decipher the sequence of events and glacial interactions that have occurred in Scotland. The formal 'top-down' lithostratigraphical framework of McMillan *et al.* (2011a) is followed here. In this scheme, the *Great Britain Superficial Deposits Supergroup* includes seven named groups, including (in Scotland) the *British Coastal Deposits Group* embracing raised marine and coastal deposits, the *Britannia Catchments Group*, comprising predominantly fluvial deposits, and a *Residual Deposits Group*, including mainly remanié deposits and other units that have undergone prolonged dissolution (Table 1). Two groups are assigned for glacigenic deposits, namely the *Caledonia Glacigenic Group*, of MIS 5e-2 age, and the *Albion Glacigenic Group*, representing older glaciations. Deposits of these groups in Scotland are included in a series of subgroups that have been divided geographically, some by natural boundaries, such as the Great Glen, but mostly on the basis of overall lithology and provenance of the constituent formations (Fig. 3a). The concept of subdivision is not new; for example, in NE Scotland three 'series' of glacigenic deposits have been recognised traditionally relating to distinct bodies of ice that existed in the region during past glaciations (Merritt *et al.* 2003, 2017a). The glacigenic sequences include some non-glacial, terrestrial sediments, periglacial deposits and soils that provide important biostratigraphical and geochronometric evidence and stratigraphical markers. These units are included in the *Britannia Catchments Group*.

A first step towards a systematic geographical delineation of surficial till units in Britain was undertaken in 2009 (Entwisle & Wildman 2010), utilising the British Geological Survey (BGS) Superficial and Bedrock Geology themes draped on hill-shaded digital elevation models (DEMs; http://www.bgs.ac.uk/data/maps/home.html) (Fig. 3b). Archive BGS mapping, digital air photos and other relevant data were used in that exercise, including the BRITICE database (Clark *et al.* 2004; Evans *et al.* 2005). Similar protocols were followed as for the delineation of glacigenic subgroups, using literature review and prior

Table 1 Summary details of the groups within the Great Britain Superficial Deposits Supergroup represented in Scotland (from McMillan *et al.* 2011a). Abbreviation: MIS = Marine Isotope Stage.

| Group name	Genesis	Lithologies and textures	Age
Britannia Catchments Group	Fluvial, organic and mass movement	Clastic deposits; boulder, gravel, sand, silt and clay grades; periglacial slope deposits/gelifluctate/ 'head'	Holocene to pre-Anglian, MIS 1-pre-12
Caledonia Glacigenic Group	Glacigenic	Clastic deposits; boulder, gravel, sand, silt and clay grades; diamictons (including till)	Devensian (Weichselian), MIS 2–5d
Albion Glacigenic Group	Glacigenic	Clastic deposits; boulder, gravel, sand, silt and clay grades; diamictons (including till)	pre-Devensian (pre-Weichselian), pre-MIS 5e
Residual Deposits Group	Residual and weathering products	Clastic deposits; gravel, sand, silt and clay grades	Tertiary to Early Pleistocene
British Coastal Deposits Group	Marine and coastal	Clastic deposits; gravel, sand, silt and clay grades	Holocene to pre-Anglian, MIS 1-pre-12
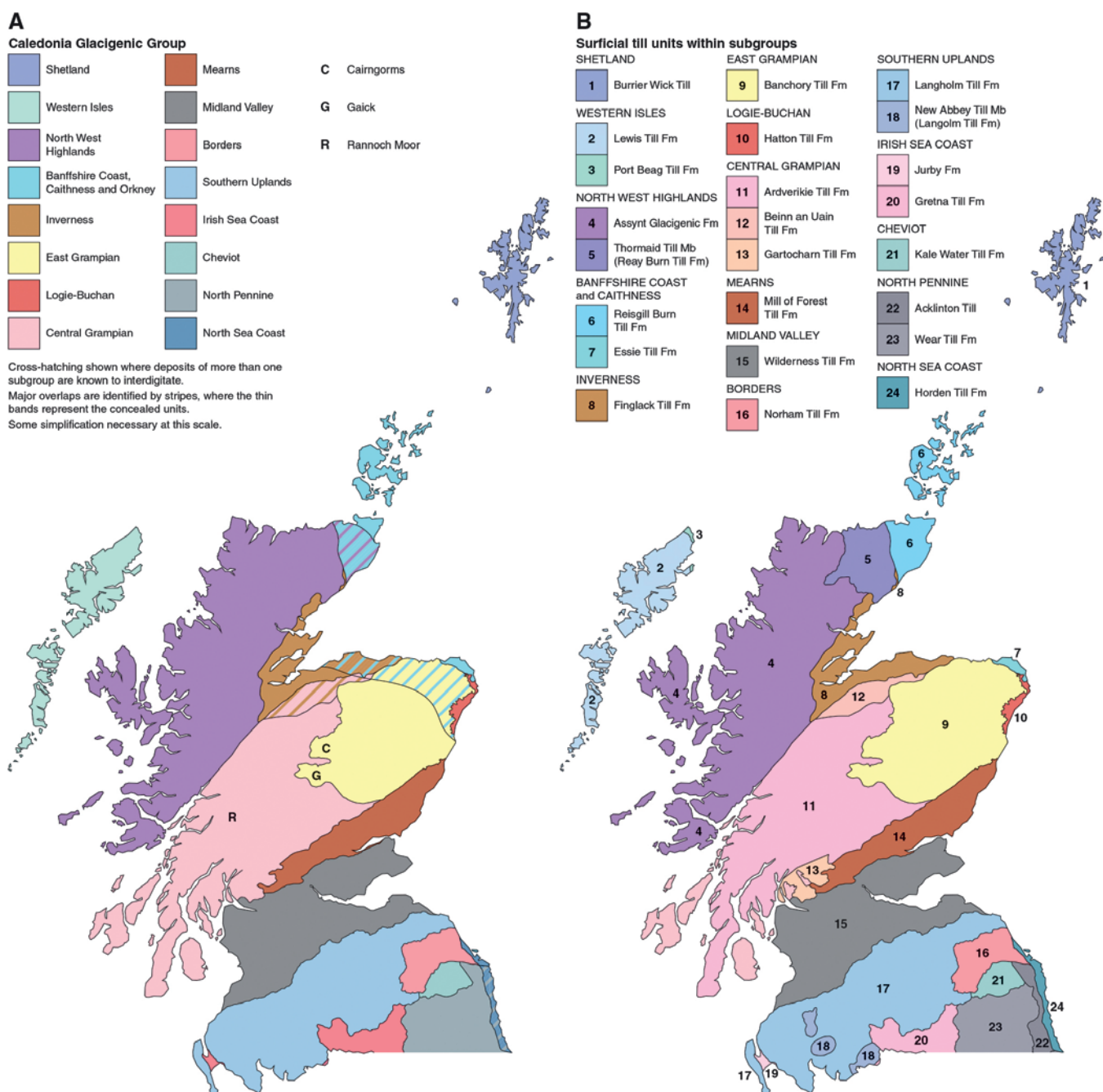
Figure 3 Distribution of glacigenic subgroups (A) and superficial till units (B) (modified from McMillan *et al.* 2011a). Abbreviations: C = Cairngorms; G = Gaick; R = Rannoch Moor.

knowledge of general ice-flow trajectories, glacial history and simplified bedrock geology. As the matrices of tills are known to bear a strong relationship with underlying bedrock (Benn & Evans 2010), many boundaries are positioned a short distance 'down ice' of significant changes in bedrock geology. The use of descriptive qualifiers (such as 'till') and other issues of protocol are discussed by McMillan & Merritt (2012).

2. The Late Pleistocene record

The climate records inferred from offshore sediment cores and Greenland Ice Sheet cores highlight the variability of Late Pleistocene climate (Fig. 2). For example, they reveal 26 clearly defined stadials and interstadials, with numerous sub-events (Rasmussen et al. 2014). However, events in Scotland, and particularly the timing and pattern of glacier and ice sheet development, remain uncertain because of the absence of unambiguous local stratigraphic evidence and dating. Nevertheless, a broad picture may be inferred from the Greenland event stratigraphy (Lowe et al. 2008b; Rasmussen et al. 2014), offshore evidence from the Atlantic Shelf adjacent to Scotland (e.g., Hibbert et al. 2010), the speleothem record from caves in Northern England (Latham & Ford 2013) and comparison with the fluctuations of the Fennoscandian Ice Sheet (FIS) (Hughes et al. 2016; Patton et al. 2017), particularly regarding evidence of its presence in the North Sea basin (Phillips et al. 2017; Stewart et al. 2019).

Clapperton (1997) used the Greenland Ice Core Project (GRIP) Summit ice core $\delta^{18}O$ record to infer time intervals during the Devensian when climate conditions in Scotland could have supported different types of ice mass, including ice sheets, mountain ice caps and corrie glaciers (Fig. 2). Although such extrapolation is over-simplified, it nevertheless suggests that glaciers in one form or another are likely to have existed in Scotland for extended parts of the Devensian.

Core MD04-2822 from the Rockall Trough, beneath the Atlantic Ocean 300 km W of Scotland, provides lithological evidence for the build-up and expansion of the BIIS across the Hebridean Shelf during the Early Devensian (MIS 4), fed by ice sources in Scotland and Northern Ireland. Pulses of ice-rafted debris (IRD) occurred at 70.2 and 61.7 ka (Hibbert et al. 2010). The same core, and core MD95-2006 from the Barra-Donegal Fan, some 200 km W of the Scottish mainland (Hall et al. 2018, fig. 2), demonstrate limited delivery of IRD diagnostic of British Palaeogene Igneous Province sources in Scotland and Northern Ireland to these sites during the early part of MIS 3. However, pulses followed at 41 and 38.5 ka and a steep increase occurred at 27.4 ka (Knutz et al. 2001; Peters et al. 2008; Hibbert et al. 2010). This pattern of IRD delivery is consistent with the periods of cooling in the North Atlantic region revealed in the Greenland ice cores, and indicates BIIS growth during the Early Devensian of sufficient size and extent to deliver IRD to the coring sites, followed by dynamic and fluctuating glacial responses to climate change. It also supports previous indications of Early Devensian ice sheet glaciation (Sutherland 1981; Sissons 1983; Bowen & Sykes 1988; Bowen et al. 2002).

Comparison with SW Norway also lends support to dynamic glacial responses following MIS 5e. For example, extensive ice-free conditions are indicated during the warm phases of MIS 5c and 5a (Helle et al. 1981; Hughes et al. 2016), whereas during the MIS 5b and 5d cold intervals, glaciers extended to the present coast at limits similar to, or just outside those of glaciers in the Younger Dryas (Mangerud 2004). During MIS 4 and early MIS 3, all of southern Norway was ice covered, with the ice sheet extending into Denmark at 65–60 and 55–50 ka (Larsen et al. 2009) and onto the bed of the northern North Sea (Carr et al. 2006). Later in MIS 3, the timing of glacial-interstadial phases is constrained by sediments in coastal caves in western Norway (Larsen & Mangerud 1989; Valen et al. 1995, 1996). Ice advanced beyond the present coastline at ~41 ka, with withdrawal by 38 ka. During Greenland Interstadials GI 8–GI 7, at 38.2–34.5 ka, coastal areas were free of glacier ice during the Ålesund Interstadial, with sufficient vegetation cover for reindeer to graze (Mangerud et al. 2011). Expansion of the last FIS may have occurred later than in Scotland, with ice first reaching limits in northern Denmark after 29 ka. Radiocarbon dates for marine shells indicate that the Witch Ground Basin beneath the central North Sea remained ice free until after 34.7–32.4 ka (Sejrup et al. 2015) or 36.1–33.7 ka (Graham et al. 2010). Ice withdrew from northern Denmark at 27–23 ka before advance to maximum limits at 23–21 ka (Larsen et al. 2009)

Most of the glacigenic deposits in Scotland were laid down during the Late Devensian. The last Scottish Ice Sheet, which dominated the BIIS, appears to have expanded sometime after ~35–32 ka based on radiocarbon dates from sediments underlying glacial deposits in the Central Lowlands (Section 5.4). It reached its maximum extent on the Atlantic Shelf at ~30–27 ka, although this is now thought to have been neither synchronous across different parts of the ice sheet (Ballantyne & Small 2018), nor in tandem with the FIS (Section 3.2). Numerical modelling indicates that the ice sheet was highly dynamic with rapid changes in extent (Boulton & Hagdorn 2006; Hubbard et al. 2009). The terrestrial stratigraphy of Scotland, therefore, needs to be viewed in the context of a highly dynamic and climate-sensitive ice sheet that may have existed throughout much of the Devensian (Ballantyne & Small 2018).

There have been significant advances in understanding of the lithostratigraphic and chronostratigraphic framework for Scotland since the publication of the Quaternary of Scotland (Gordon & Sutherland 1993a). These are reviewed below for a number of critical regions that are grouped broadly into three categories with reference to former ice accumulation areas:

A. Peripheral areas: Shetland, Orkney, Caithness and the Outer Hebrides (Western Isles). These are significant because they include deposits that have been preserved locally beneath cold-based ice and provide key information on the timing of ice sheet advance and retreat.
B. Core areas: the North-West Highlands and Central Highlands. These are significant because they were the main centres of ice accumulation beneath which older deposits have been preserved locally due to relatively little sub-glacial erosion.
C. Lowland depocentres: Inverness and the Great Glen, NE Scotland, Southern Argyll and Arran, the West Central Lowlands and the Solway Lowlands. These are significant for thick sequences of multiple tills and intercalated sediments.

3. Peripheral areas

Scottish ice reached its maximum limits during the Middle and Late Devensian, marked by a series of moraine ridges close to the NW Atlantic shelf edge (Bradwell et al. 2008a; Clark et al. 2012a). At this time the BIIS is believed to have been confluent with the FIS in the North Sea (Graham et al. 2011;

Ballantyne & Small 2018). Earlier claims that ice-free areas existed at the LGM at low elevations in peripheral parts of Scotland (Sutherland 1984), particularly in Lewis (Sutherland & Walker 1984), Caithness and Orkney (Bowen & Sykes 1988) and in Buchan (Synge 1956, 1963), have been shown to be inconsistent with new stratigraphic and dating evidence for LGM ice cover in these areas (Hall & Whittington 1989; Hall 1995; Whittington *et al.* 1998; Ballantyne & Hall 2008; Hall & Riding 2016). Nonetheless, the stratigraphic record in these peripheral areas remains of high importance for understanding the timing, sequence and pattern of the last ice sheet glaciation and associated environmental changes.

3.1. Shetland

Shetland, Orkney and Caithness together form a topographic divide between the NE Atlantic and the North Sea (Hall *et al.* 2018, fig. 2). On Shetland, there has been debate since the 19th Century as to whether the islands supported a local ice cap (Milne-Home 1881; Flinn 1978) or if they were over-whelmed by the FIS during the last glaciation (Peach & Horne 1879; Sutherland & Gordon 1993b). That debate continues (Ross 1996; Golledge *et al.* 2008; Flinn 2009; Carr & Hiemstra 2013). The identification of only one Scandinavian erratic on the islands (Finlay 1926), the absence of Permian to Neogene palynomorphs in tills derived from the bed of the North Sea and evidence of radial ice-flow patterns from erratics, glacial bedforms and striae (Fig. 4) provides little support for former flow of ice across Shetland (Hall 2013). There is no evidence from erratics that Scottish ice reached Shetland (Hall 2013), except Fair Isle, midway between Shetland and Orkney (Hall & Fraser 2014). Consequently, the balance of evidence strongly supports the existence of an independent Shetland ice cap (SIC) during the Late Devensian (Hall 2013).

The oldest Quaternary sediment recognised on Shetland is from Fugla Ness (Fig. 4; Table 2), on the island of North Roe, where coastal erosion has exposed a shallow basin developed in bedrock within glacially scoured 'cnoc-and-lochain' terrain (Chapelhowe 1965; Fig. 5). Resting above 'rock-face breccia' (gelifractate?) on the striated floor of the depression is the *Fugla Ness Lower Till*, with a fabric indicating ice flow to the W or NW (Hall *et al.* 1993b, 2002). The overly-ing, compacted organic deposits of the *Fugla Ness Peat Bed* thicken towards the centre of the basin, comprising a thick lower unit containing some wood fragments and thinner layers above with more wood, small pine stumps and pine cones (Birks 1993a). The presence of thermophilous pollen and macrofossils indicates that the peat formed during an inter-glacial (Birks & Ransom 1969). The pollen record indicates that *Pinus* and possibly *Ilex* grew locally in sheltered areas in a mild, oceanic climate, possibly with wind speeds lower than today (Hall *et al.* 2002). The maritime heath flora has been compared to that of the Gortian Interglacial in Ireland (Birks & Ransom 1969), but it is also consistent with existing environ-mental reconstructions for the Ipswichian/Eemian interglacial in maritime NW Europe (Hall *et al.* 2002). A uranium-series age of $110 + 40/-35$ ka supports the case that the Fugla Ness Peat was formed during the last interglacial (MIS 5e) (Hall *et al.* 2002). The later change in vegetation indicated by the pollen record suggests a drop in annual temperature and an increased frequency of frosts (Birks & Ransom 1969). The high representation of grasses in the pollen reworked into the overlying *Fugla Ness Breccia* may reflect a climatic deteriora-tion to periglacial conditions, which resulted in granular and blocky disintegration of the local bedrock accompanied by wash and solifluction of the debris into the basin (Hall *et al.* 2002). The apparent continuity of deposition has been used to suggest that the breccia formed during MIS 5d, when periglacial

conditions prevailed throughout NW Europe (Aalbersberg & Litt 1998).

Organic sediments and peat are also found beneath till and periglacial deposits at Sel Ayre, on the Sandness peninsula of Mainland Shetland (Birks & Peglar 1979; Birks 1993b; Hall *et al.* 1993a; Figs 4, 6). The *Sel Ayre Lower Breccia* records periglacial conditions. These coarse sediments retain no record of any vegetation cover, but are tentatively assigned to MIS 5d (Hall *et al.* 2002). The overlying *Sel Ayre Organic Sands and Gravels* are attributed to MIS 5c on the basis of pollen stratigraphy and luminescence ages of *ca.*105–98 ka from intercalated sands (Duller *et al.* 1995). The pollen record indi-cates that an earlier grassland environment was followed by ericaceous heath with associated changes in soil type, with a brown earth of the grassland being replaced by the bleached sand of a more podzolic soil (Birks 1993b). This was followed by open, grass-dominated vegetation with a variety of herbs characteristic of skeletal mineral soils. The Sel Ayre sequence appears to record major environmental changes through the MIS 5c interstadial on Shetland (Hall *et al.* 2002). The over-lying *Sel Ayre Upper Breccia* is devoid of organic material, but apparent continuity of deposition suggests that it may date from the succeeding cold interval, MIS 5b, when periglacial conditions returned to the lowlands of NW Europe (Aalbersberg & Litt 1998).

Till fabric data from the *Fugla Ness Upper Till* (Fig. 5) and *Sel Ayre Till* (Fig. 6) are consistent with the patterns of ice flow during the local ice cap phase of glaciation (Hall *et al.* 2002). This phase is usually attributed to the Late Devensian, although firm dating evidence is lacking (Gordon *et al.* 1993; Ross *et al.* 1993; Ross 1996). Despite suggestions on geomor-phological criteria that ice-free areas existed during the LGM on northern Unst and Yell (Flinn 1983; Long & Skinner 1985), and also Ross (1996), offshore data indicate that the SIC extended well beyond the current landmass to the E and N, and close to the edge of the continental shelf W of Shetland (Johnson *et al.* 1993; Stoker *et al.* 1993; Ross 1996). Sets of moraines on the West Shetland Shelf have been interpreted to represent multiple phases of expansion of the SIC during the Middle and Late Devensian (Davison 2004). Stacked till sequences have not been widely recognised on Shetland (Ross 1996), where till is generally thinly developed (Hall 2013). Only single till units are recognised above the MIS 5 organic sediments at Fugla Ness and Sel Ayre (Hall *et al.* 2002), these units having been re-assigned to the *Burrier Wick Till Forma-tion* (McMillan *et al.* 2011a; Fig. 3b). However, marked differences in erratic content and lithology present within the till units on, for example, North Roe (Chapelhowe 1965) and on Unst (Golledge *et al.* 2008; Fig. 4), suggest that a more complex glacial stratigraphy awaits elucidation.

The flow of ice from SW Norway across Shetland is a recurrent feature of both classic 19th Century (Croll 1870; Peach & Horne 1879, 1880) and modern ice sheet reconstruc-tions (Bradwell *et al.* 2008a; Graham *et al.* 2010; Sejrup *et al.* 2016). Yet, firm evidence for the former presence of the FIS (or BIIS) on Shetland is absent. In western Shetland, flow sets interpreted from DEMs (Golledge *et al.* 2008) generally conform to patterns of westward ice flow recognised from the dispersal of erratics, striae and rôches moutonnées (Mykura & Phemister 1976; Flinn 1978; Hall 2013). Con-flicting evidence has been reported from eastern Shetland, however, for the interpretation of mega-scale glacial lineations (MSGLs) (Golledge *et al.* 2008), and kinematics derived from glacitectonic structures (Carr & Hiemstra 2013) suggests westward ice flow, whereas erratic dispersal, striae and rôches moutonnées indicate eastward ice flow (Flinn 2009; Hall 2013; Fig. 4a).
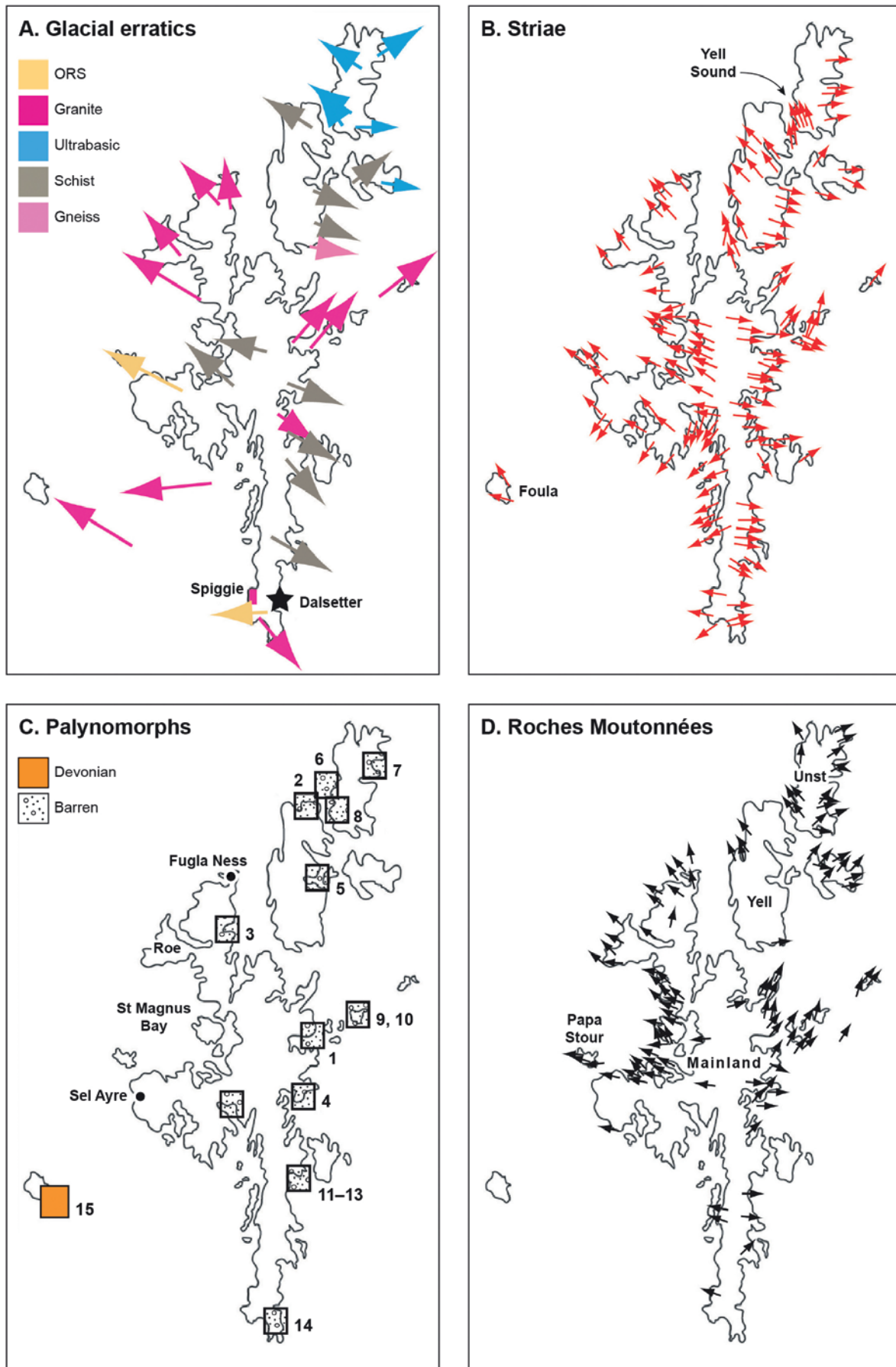
Figure 4 Evidence for the pattern of ice flow on Shetland (modified from Hall 2013). (A) Directions of glacial erratic carry. (B) Measurements of striae, excluding disputed early records (from Peach & Horne 1879). (C) Palynomorph records for till matrix at key sites (J. B. Riding, pers. comm. 2016). Sample sites: 1. Vidlin Voe; 2. Sand Voe; 3. Colla Firth; 4. South Nesting; 5. Salt Wick; 6. Wick of Collaster; 7. Norwick; 8. Westing; 9. The Bugg; 10. Sand Wick; 11–13. Breiwick; 14. Little Bixter; 15. Sand of Hayes; 16. Da Ristie Nessies. (D) Ice-flow directions from rôches moutonnées.

Table 2 Summary of events in the Shetland Islands. Abbreviations: B = glacial 'binge'; C = cold-periglacial; I = Interglacial; IS = Interstadial; P = glacial 'purge'.

| Age	Events, palaeo-environments, important sites and units	
MIS 5e	Formation of the *Fugla Ness Peat* under a warm, humid climate	I
MIS 5d	Climatic deterioration with development of periglacial slope deposits at Fugla Ness and Sel Ayre during MIS 5d	C
MIS 5c	Climatic amelioration and formation of the *Sel Ayre Organic Sand and Gravel*	IS
MIS 5a–b	Climatic deterioration and renewed periglacial activity, with multiple phases of periglacial deposition at Sel Ayre	C/IS
MIS 4	No deposits firmly attributable to these intervals	
MIS 3		
LLGM		
GI-2		
LGM	Development of a large ice cap over Shetland laying down the *Burrier Wick Till*. Ice confluent with the FIS along the western edge of the Norwegian Channel and extending towards the shelf edge	B
*ca.*15 ka	Active retreat of an independent Shetland ice cap	P
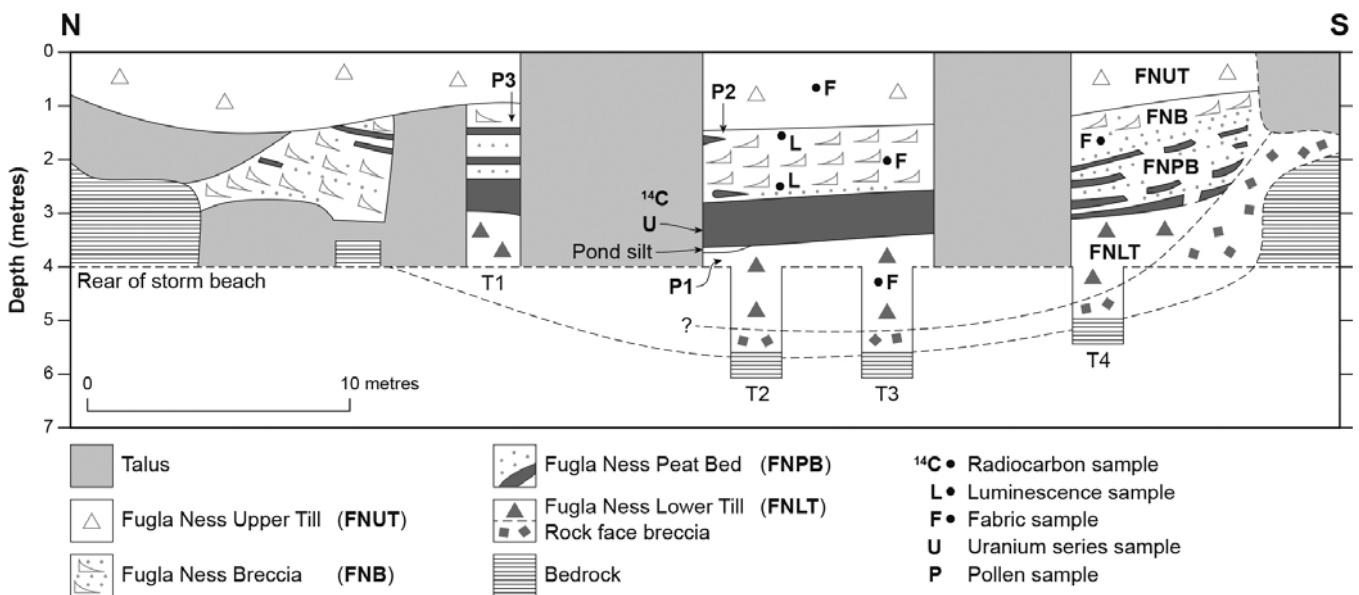
Figure 5 The stratigraphy at Fugla Ness, Shetland (modified from Hall *et al.* 2002).

The last glaciation of Shetland likely involved the development of an extensive ice cap that interacted with the Norwegian Channel Ice Stream (NCIS) between Shetland and Norway, and with ice flowing between Orkney and Shetland. Early build-up, but limited extension of the SIC, are features of recent mathematical models (Hubbard *et al.* 2009). Earlier suggestions that an initial phase of FIS glaciation on Shetland was followed by the build-up of a local ice sheet (Hoppe 1974; Mykura & Phemister 1976) lack support on account of: (i) the scarcity of Scandinavian erratics; (ii) the absence of glacial transport of Permo-Triassic debris across the islands from basins located to the E; and (iii) the apparent absence of glacial deposits from the FIS, especially in zones of lower glacial erosion in the ice shed zone of the SIC and in western areas of the archipelago where MIS 5 deposits are preserved (Hall 2013). Furthermore, the tills on Shetland also appear to lack palynomorphs derived from sedimentary rocks and sediments in the northern North Sea basin (Fig. 4c). MSGLs along the margins of Yell Sound, between Yell and Mainland, and Bluemull Sound, between Unst and Yell (Fig. 4; Golledge *et al.* 2008) are associated with distinctive mud-rich, grey tills (Chapelhowe 1965; Ross 1996) and probably record fast flow within SIC-sourced ice from the heads of both

sounds. Patterns of erratic dispersal and glacial lineaments at the northern and southern extremities of Shetland can be attributed to ice sheds off the present eastern coast (Ross 1996). Striae (Peach & Horne 1879) and glacitectonic structures (Carr & Hiemstra 2013), previously interpreted as indicating widespread onshore movement of ice in eastern Shetland, are both considered by Hall (2013) to have been interpreted incorrectly.

The pattern of deglaciation around Shetland has become much clearer in recent years with the advent of detailed imagery of bedforms on the adjacent sea bed (Bradwell *et al.* 2008a). The SIC retreated back towards the spine of the islands from both the W and E, with moraines on Papa Stour marking a late stage of ice retreat into St Magnus Bay (Fig. 4; Mykura & Phemister 1976). Rogen moraines on the floor of the Pobie Basin, to the E of Unst, indicate ice retreat towards the S (Bradwell *et al.* 2008a), with moraines on Unst (Golledge *et al.* 2008) marking positions of the western flank of this retreating ice lobe.

3.2. Orkney and Caithness
More complex glacial sequences occur in Orkney and Caithness (Hall & Whittington 1989; Gordon 1993a; Hall *et al.* 2011),
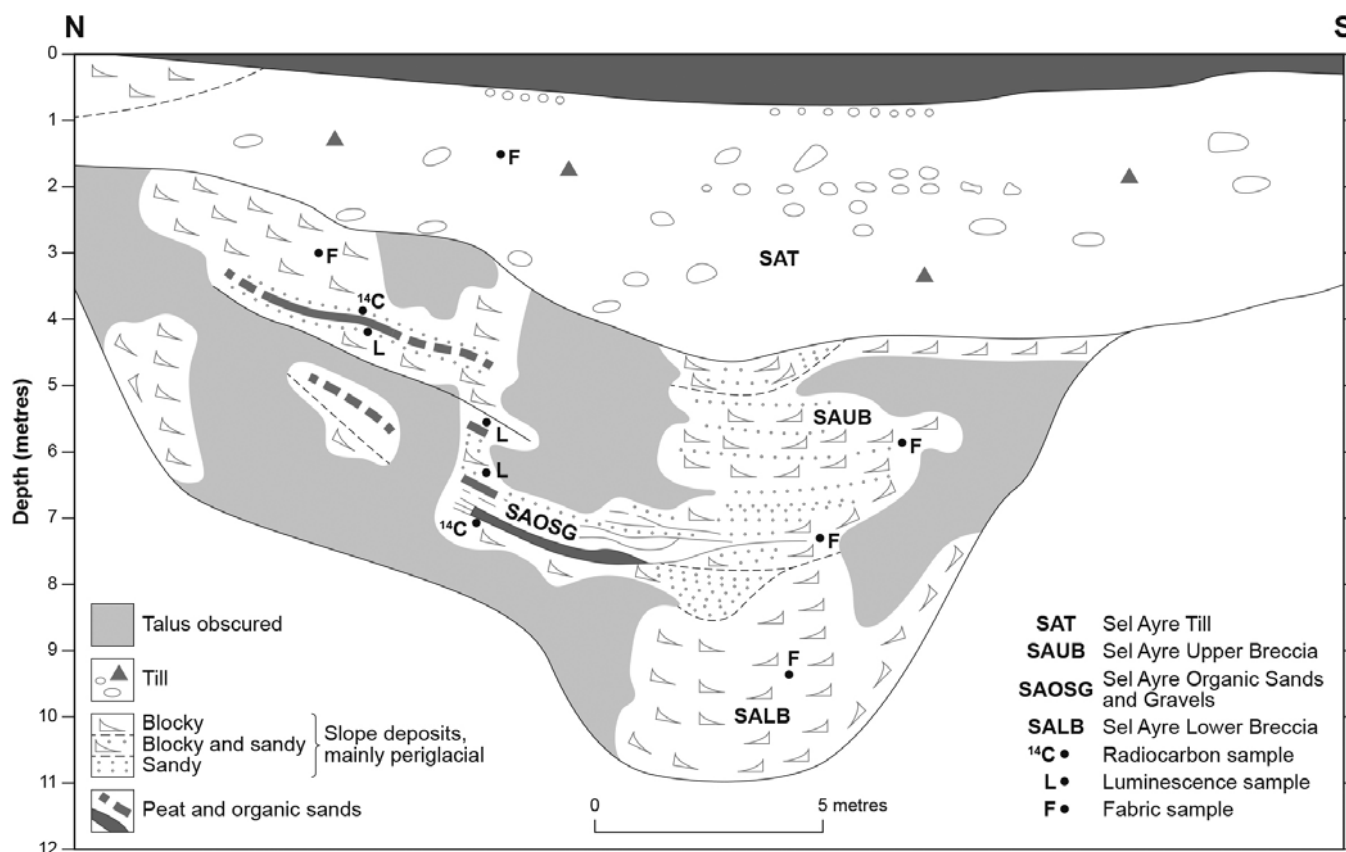
Figure 6 The stratigraphy at Sel Ayre, Shetland (modified from Hall *et al*. 1993a).

from which a three-stage glaciation has been proposed recently for the last glacial cycle, based on the preservation of stacked till sequences (Hall & Riding 2016; Hall *et al*. 2016a; Fig. 7a; Table 3) (Section 3.2.2).

The oldest features recognised in Caithness and Orkney are raised and intertidal rock platforms and associated rock cliffs buried by till (Crampton & Carruthers 1914; Smith *et al*. 2018). On NW Hoy (Fig. 7a), raised beach gravels of the *Muckle Head Gravel* rest on a raised shore platform at 6–12 m OD and are overlain by periglacial slope deposits and a thin, brown, unnamed till (Wilson *et al*. 1935; Sutherland 1993a). From Orkney Mainland northwards a mainly brown diamicton with few far-travelled clasts is found locally (Rae 1976). This *Digger Till* contains Carboniferous and Permian palynomorphs (Fig. 7a), which, taken together with the evidence of striae (Fig. 7b), indicates deposition by ice that flowed to the NW from the platform E of Orkney. The overlying *Scara Taing Till* (Figs 7a, 8) is a red to brown diamicton with locally high contents of marine shell fragments and a far-travelled clast component derived mainly from around the Moray Firth basin, but also including rock types typical of the Forth approaches (Carboniferous limestone and sandstone) and southern Fennoscandia (Fig. 9).

The uppermost *Quendale Till* (Figs 7a, 8) is a locally derived diamicton that records a final period of ice flow N and W across the low ground of the Orkney archipelago and offshore to the adjacent Atlantic shelf (Phillips *et al*. 2008; Hall *et al*. 2016a; Fig. 7b). This event may correlate with an ice sheet readvance to the Reay Ridge, a sea-bed moraine lying 10–20 km W of Orkney Mainland (Bradwell & Stoker 2015). The final retreat of ice is marked by moraine systems on the Atlantic shelf, on W Rousay (Fig. 9) and central Orkney Mainland, as well as by moraines and glaciolacustrine sediments in the valleys of western Hoy (Wilson *et al*. 1935; Hall

1996b; Bradwell & Stoker 2015; Hall *et al*. 2016a). The pattern of retreat indicated by moraine ridges was SE across Orkney, into Scapa Flow (Fig. 9) and SSE towards the Moray Firth. This evidence clearly shows that an independent Orkney ice cap did not exist at this time (Hall *et al*. 2016a). Recalibrated cosmogenic exposure ages indicate that low ground on W Mainland became ice-free by *ca*.16 ka (Phillips *et al*. 2008; Ballantyne & Small 2018).

The orientations of striae (Fig. 7b) together with the origin and distribution of glacial erratics (Figs 9, 10) suggest that ice flowed across Caithness and Orkney from the S and E, out of the Moray Firth (Peach & Horne 1880, 1881). The ice laid down mud-rich, shelly diamictons assigned to the Banffshire Coast, Caithness and Orkney Glacigenic Subgroup (Hall & Riding 2016; Fig. 3a). The occurrence of sparse Scandinavian erratics noted by Peach & Horne (1880, 1893) on the northern isles of Orkney has been confirmed by Hall *et al*. (2016a) and on Fair Isle (Hall & Fraser 2014; Fig. 10). However, the palynomorph content of till matrices suggests that these indicator erratics are likely to have been reworked into Scottish tills from older sediments in the Outer Moray Firth and the Witch Ground basin of the central North Sea (Hall *et al*. 2016a; Fig. 10). Hence, there is no firm evidence that the FIS flowed over northern Orkney during the last glaciation. Instead, episodic development of fast ice flow between Orkney and Shetland is likely, sourced from an ice saddle that developed across the northern North Sea at the LGM (Sejrup *et al*. 2016). The Scandinavian erratics may also derive from one or more earlier incursions of Fennoscandian ice into the western part of the northern North Sea (Flinn 1978).

Recent numerical models (Hubbard *et al*. 2009) and ice sheet reconstructions (Clark *et al*. 2012a) suggest the former existence of a local ice cap on Orkney, a possibility that was rejected in earlier stratigraphic studies (Wilson *et al*. 1935;
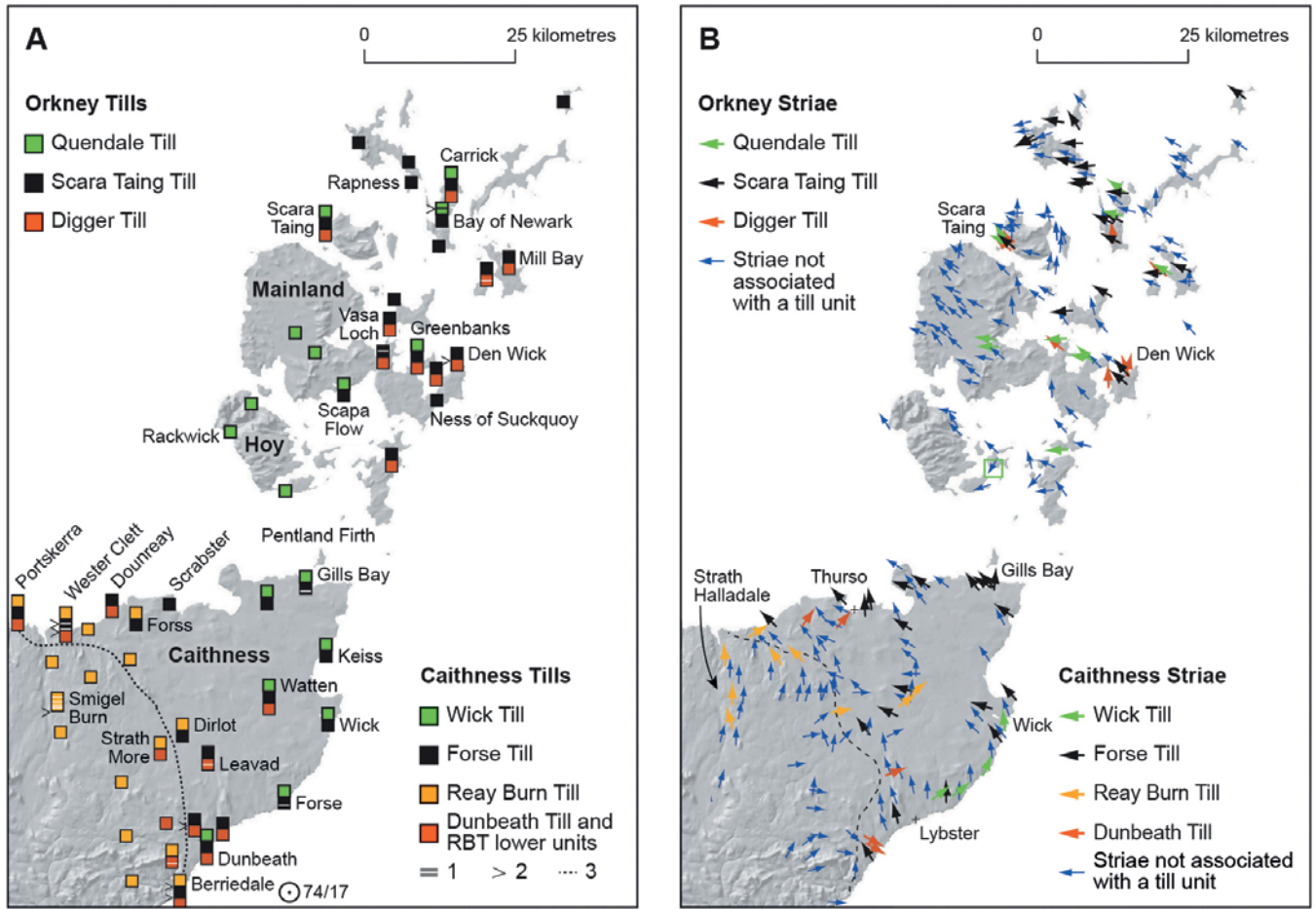
Figure 7 Till stratigraphy (A) and striae (B) in Caithness and Orkney (modified from Hall & Riding 2016 and Hall *et al*. 2016a). 1 = unconformity or disconformity within a till unit; 2 = sand or gravel layer; 3 = western limit of the Forse Till derived from the inner Moray Firth in Caithness and Sutherland.

Table 3 Summary of events in Caithness and on the islands of Orkney. Abbreviations: B = glacial 'binge'; C = cold-periglacial; I = Interglacial; P = glacial 'purge'.

Age	Events, palaeo-environments, important sites and units	
MIS 5e	Raised rock platforms in Caithness (till covered) and on Hoy beneath raised beach deposits (*Muckle Head Gravel*)	I
MIS 5a–d	Periglacial deposits on Hoy resting on Muckle Head Gravel	C
MIS 4	No deposits are known from these intervals	
MIS 3		
LLGM	Ice advances NW across Orkney depositing the *Digger Till*. Advance of NW Highland ice to Reay and onto the plain of Caithness laying down the *Portskerra Till*	B
GI-3	Withdrawal of NW Highland ice from Caithness	P
GI-3 to GI-2	Three advances of Moray Firth ice into Caithness depositing the *Drumhollistan, Eastern Gully* and *Wester Clett tills* at Wester Clett	P/B
LGM	Major expansion of ice from the Moray Firth across Caithness and Orkney reaching the shelf edge, depositing the *Scara Taing Till* in Orkney and the *Forse Till* in Caithness	B
~18–~15 ka	Partial deglaciation of Orkney and Caithness	P
15 ka	Final readvance of ice NW across Orkney and Caithness laying down the *Quendale Till* on Orkney and the *Wick Till* in Caithness	B
15–14 ka	Ice retreat towards the Inner Moray Firth interrupted by local readvances towards Thurso and Wick and by final stillstands at Strath More (Dalemore) and Berriedale	P

Rae 1976). The very limited presence of far-travelled erratics, including marine shells, in Hoy and W Mainland and the absence or weak development of subglacial bedforms on the hills of Hoy is consistent with the build-up of a dome of cold-based ice of sufficient thickness to prevent incursion of Moray Firth ice into this part of Orkney for long periods. However, no till, striae or erratics have been reported that might relate

to radial flow from an Orkney ice cap (Hall *et al*. 2016a). Instead, the ice-flow indicators all point to ice flow across Orkney from an ice shed that lay to the E or SE, laying down the Quendale Till (Figs 7b, 9, 10) (see Section 5.2).

In Caithness, aside from older glacial deposits found at depth (Tait 1912; Hall & Whittington 1989; Hall *et al*. 2018), the oldest till units attributed to the Late Devensian are the
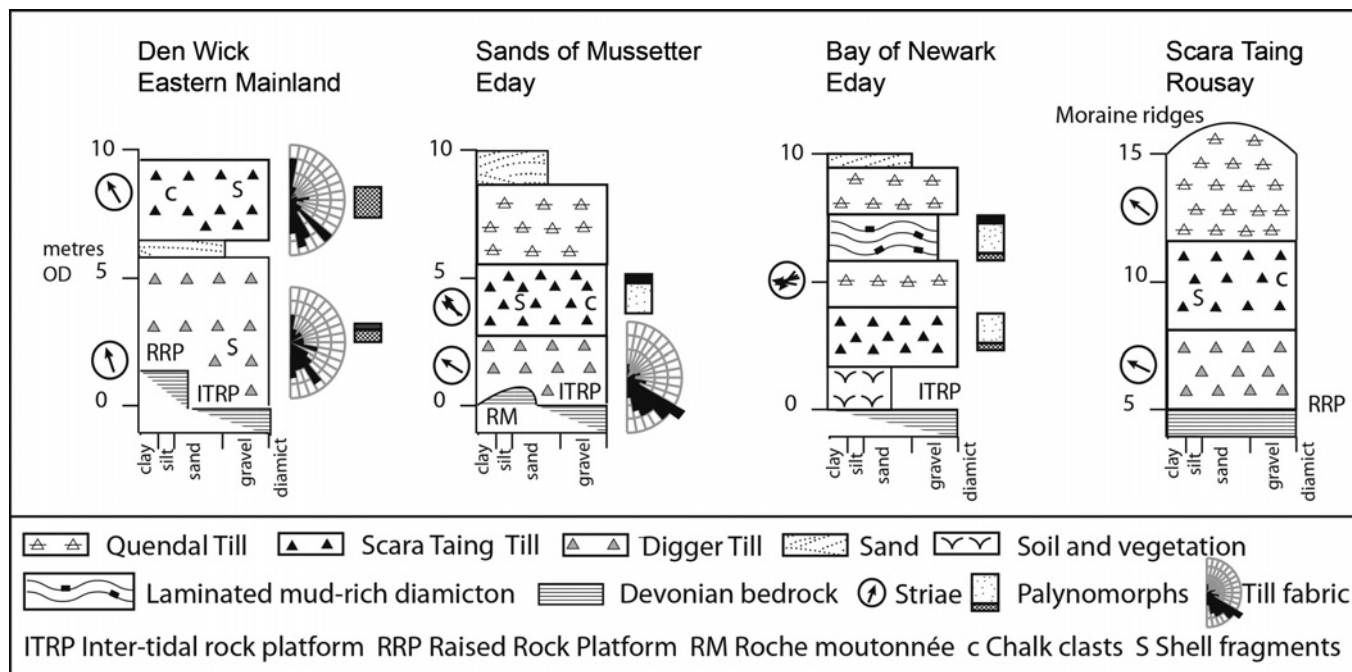
Figure 8 Sections of stacked tills on Orkney (modified from Hall *et al*. 2016a).

Ballantrath Till around Dunbeath (Gordon 1993b) and the *Portskerra Till* around Melvich (Hall *et al*. 2011; Fig. 7a; Table 3). The latter unit includes igneous and metamorphic clasts derived from the Sutherland–Caithness border, along with abundant Devonian sandstone and conglomerate debris, and records an early expansion of the NW Highland ice at least as far E as Watten (Omand 1973). At Wester Clett (also known as Drum Hollistan) (Gordon 1993c; Fig. 7a) the Portskerra Till is overlain by cryoturbated gravel and by the lowermost of three calcareous, mud-rich till units of the *Reisgill Burn Till Formation* (Figs 3b, 7a, 11; Table 4) derived mainly from Devonian sedimentary rocks, but including shells, fossils and palynomorphs derived from the Moray Firth (Hall *et al*. 2011). This *Drumhollistan Till Bed* is overlain by a fan gravel and by the *Eastern Gulley Till* derived from the SE; it is also overlain by the *Wester Clett Till* derived from the S. Late advances of inland ice deposited younger till units and formed moraines on the surface of the Reisgill Burn Till in N and W Caithness. These were probably contemporaneous with a final advance of Moray Firth ice, which deposited the *Wick Till* (Fig. 7a) and moraine ridges on low ground bordering the Pentland Firth. MSGLs in Caithness (Hughes *et al*. 2010, 2014) conform to the main flow paths for the glaciers that laid down the Reay Burn Till (inland ice) and Reisgill Burn Till (Moray Firth ice) (Peach & Horne 1881; Fig. 3b), but the multiple events identified from units in both formations demonstrate that these subglacial landforms are composite products of multiple ice-flow events following similar paths (Hall & Riding 2016; Hall *et al*. 2016a).

3.2.1. Correlation of events on the Atlantic–North Sea divide. The reconstruction offered by Hall *et al*. (2016a) for the last glaciation of Caithness, Orkney and Shetland proposes ice flow both towards the Atlantic shelf edge and into the North Sea basin during three main phases, at 32–28, 24–18.5 and 17.5–16.2 ka (Fig. 12). In recent years, the local LGM in northern Britain has been placed in the first of these intervals (e.g., Bradwell *et al*. 2008a; Merritt *et al*. 2017a; Ballantyne & Small 2018), yet the timing of the event elsewhere in NW Europe was generally coincident with the global LGM at 26.5 ka to 19 ka (Clark *et al*. 2009; Fig. 1). The FIS margin was close to its southernmost limits in Europe during the

periods 24.3–23.4 ka, 22.5–21.3 ka and, again, 20.3–18.7 ka (Toucanne *et al*. 2015). The FIS advanced into Denmark at 23–21 ka (Houmark-Nielsen & Kjaer 2003). Peak IRD flux, coincident with Heinrich Event 2 at 24 ka, is a feature of all records of the BIIS on its western margin (Scourse *et al*. 2009). The LGM on the West Shetland shelf has also been placed at 22–19.5 ka (Davison 2004), comparable to the mid-Norwegian shelf (Dahlgren & Vorren 2003). A late LGM is consistent with cosmogenic isotope exposure ages on deglaciation of high ground (217–530 m OD) in Orkney, Caithness and Buchan, which occurred as late as 18.2 ± 2.6 ka (Ballantyne 2010).

Flow of the FIS across southern Shetland, Fair Isle and northern Orkney is also a recurrent feature of recent ice sheet reconstructions (Bradwell *et al*. 2008a; Graham *et al*. 2010; Sejrup *et al*. 2016). This interpretation mainly derives from: (i) a supposed connection between westward ice flow across northern Orkney and NW flow indicated by MGSLs in the Witch Ground basin; and (ii) the presence of Scandinavian erratics on northern Orkney and, probably, Fair Isle. Neither criterion, however, is secure. MGSLs are developed across the Coal Pit Formation and overlain by the Swatchway Formation; both sedimentary units include chalk debris (Graham *et al*. 2009; Davies *et al*. 2011), implying that erosion of the MGSLs was bracketed by phases of ice flow that crossed the chalk outcrop in the outer Moray Firth. The FIS also appears to have remained largely confined to the Norwegian sector of the North Sea basin during the last glaciation, heavily influenced by the Norwegian Trough (Sejrup *et al*. 2009). The extension of an ice stream from the SW flank of the Norwegian Channel to the North Atlantic shelf edge (Bradwell *et al*. 2008a) is glaciologically implausible because the ice surface slope necessary to drive ice flow to the shelf edge gives an ice surface elevation offshore from southern Norway that is incompatible with the proximity of the southern LGM limit in Denmark (Phillips *et al*. 2017, fig. 3). As originally recognised by Graham *et al*. (2007), the MGSLs were probably cut by ice flowing to the SE out of the Moray Firth.

On northern Orkney, the very low concentrations of Scandinavian erratics are accompanied by larger numbers of erratics of Scottish origin in the Scara Taing Till (Hall *et al*. 2016a;
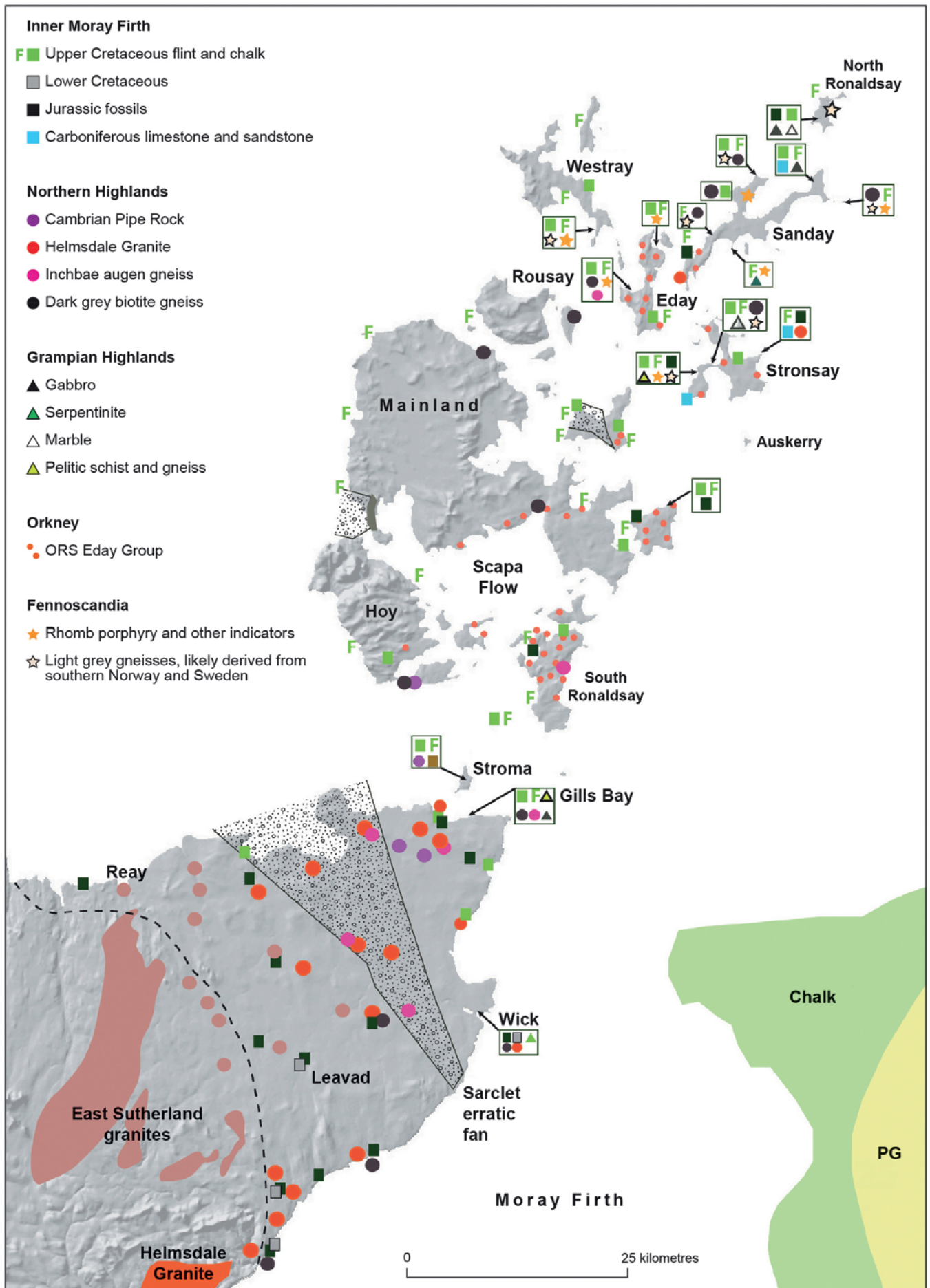
Figure 9   Records of glacial erratics in Caithness and Orkney (modified from Hall & Riding 2016 and Hall *et al.* 2016a).
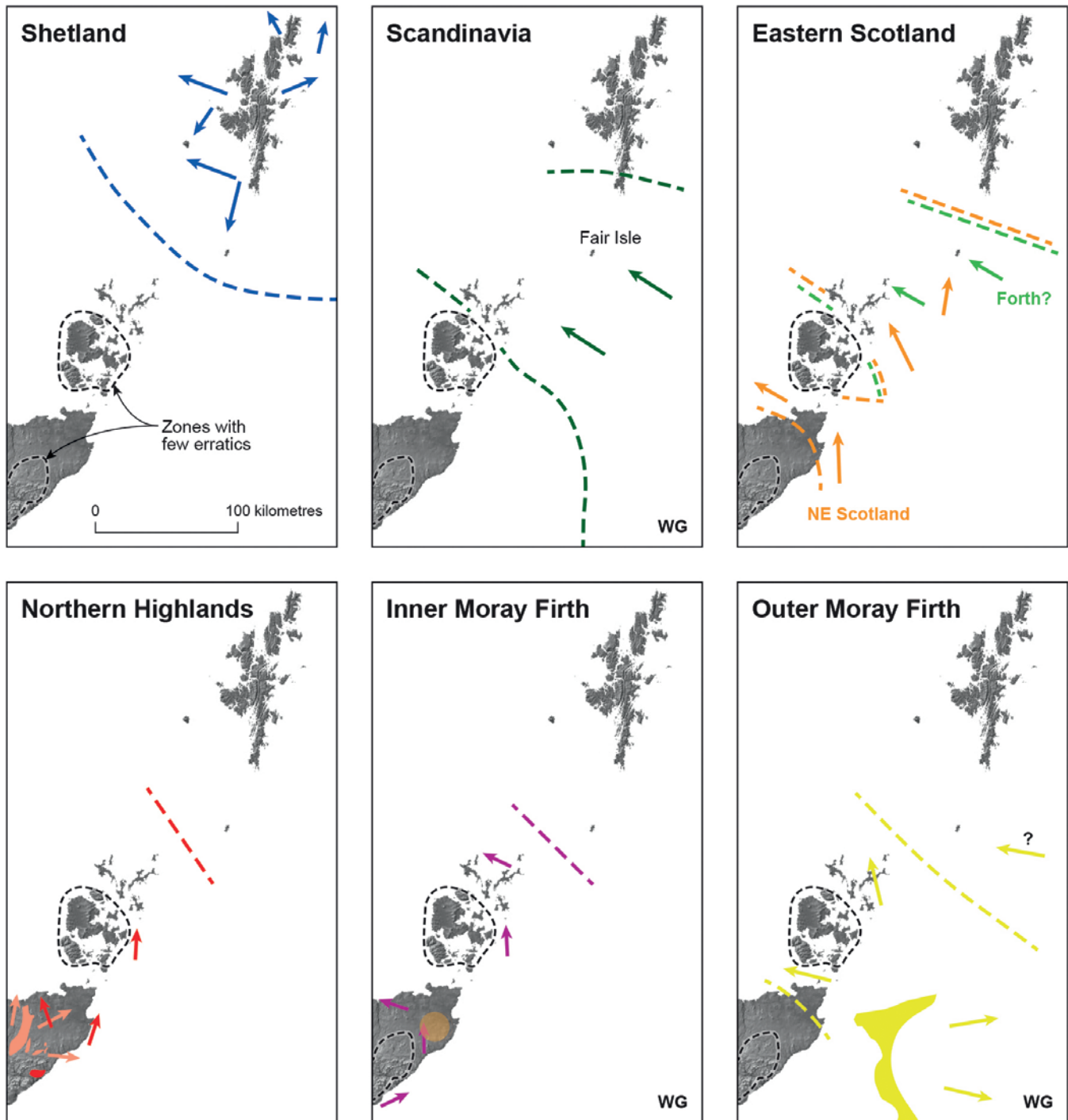
Figure 10 Domains and transport directions for glacial erratics recorded on Caithness, Orkney and Shetland. Shetland rocks (blue arrows) are confined to Shetland, including Fair Isle, and the surrounding shelf. Scandinavian erratics (dark-green arrows) occur on N Orkney, Fair Isle and at one location in southernmost Shetland. Carboniferous limestone, sandstone and volcanic clasts from the Firth of Forth and eastern Scotland (pale-green arrows) are recorded only from Fair Isle and N Orkney. Igneous and metamorphic erratics from NE Scotland (orange arrows) occur along the shores of the Pentland Firth, in N Orkney and on Fair Isle. Erratics from the Northern Highlands (red arrows) are recorded as far north as N Ronaldsay. Jurassic and Lower Cretaceous rocks and fossils from the inner Moray Firth (purple arrows) occur in Caithness (Gordon 1993d) and N Orkney. Erratics of Late Cretaceous chalk and flint (yellow arrows and outcrop) are found N of Wick and along the Pentland Firth, locally in high concentrations, and in N Orkney. Flint is found in tills on Fair Isle, but may be derived originally from sources in the Danish sector of the North Sea. Abbreviation: WG = Witch Ground.

Fig. 8). Palynomorphs indicate that this till is of Scottish origin, deposited by ice passing through the Moray Firth (Peach & Horne 1880). Hence, it is likely that the Scandinavian erratics are reworked and part of complex patterns of erratic distribution on the Atlantic–North Sea divide (Fig. 10).

3.3. Outer Hebrides
The presence of erratics from the mainland in the Outer Hebrides indicates that at least parts of the archipelago were over-run by ice from the E during the last glaciation (Geikie 1873, 1878; Sissons 1980). For example, clasts of Torridonian
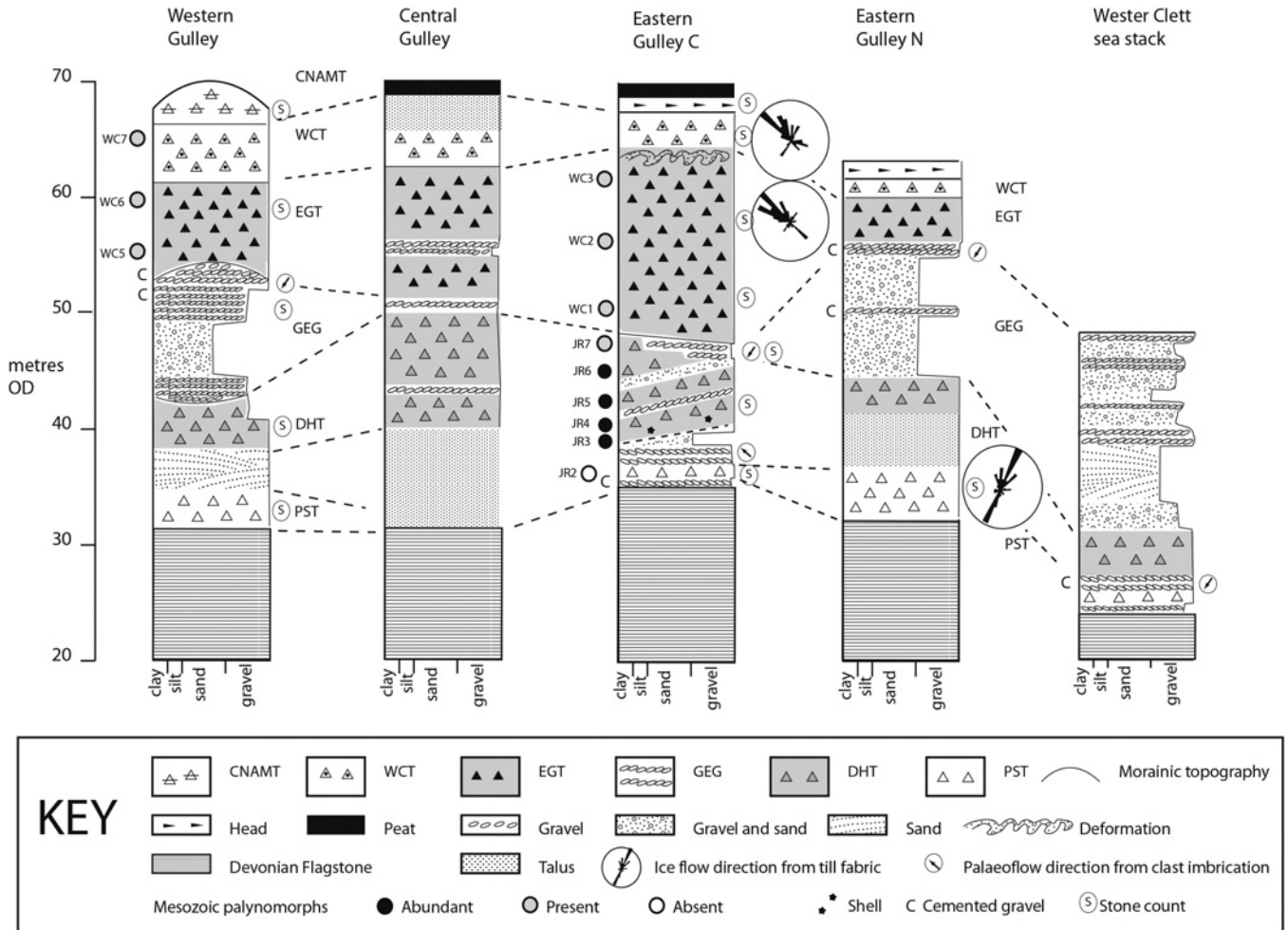
Figure 11 The stratigraphy at Wester Clett, Caithness (from Hall *et al.* 2011). Abbreviations: CNAMT = Cnoc a' Mhàil Till; WCT = Wester Clett Till; EGT = Eastern Gulley Till; GEG = Geodh' Easgaidh Gravel; DHT = Drumhollistan Till; PST = Stack Till.

Table 4 Stratigraphy at Wester Clett, Caithness (after Hall *et al.* 2011).

Group	Sub-group	Formation	Member	Beds (identified at Wester Clett)	Member	Formation	Sub-group
North Britain Glacigenic Group	North West Highland Glacigenic Sub-group	Reay Burn Till	Thormaid Till	*Cnoc a' Mhàil Till (**CNAMT**)*	Forse Till	Reisgill Burn Till	Banffshire Coast, Caithness and Orkney Glacigenic Sub-Group
			Broubster Till	*Wester Clett Till (**WCT**)*			
				*Eastern Gulley Till (**EGT**)*			
				*Geodh' Easgaidh Gravel (**GEG**)*			
				*Drumhollistan Till (**DHT**)*			
		Dunbeath Till	Portskerra Till	*Stack Till (**PST**)*			

sandstone, Cambrian quartzite and also sedimentary rocks derived from the floors of The Minch and the Sea of the Hebrides basins are found on the S and W coasts of South Uist (Jehu & Craig 1925), along the Sound of Harris and on the Flannan Isles (Stewart 1933; Fig. 13). On NW Lewis, mainland erratics are abundant in deposits derived from The Minch (Baden-Powell 1938; von Weymarn 1979; Gordon 1993j). However, the presence of mainland erratics in a till that locally underlies the Ipswichian raised 'Galson Beach' on NW Lewis (Peacock 1984; Fig. 14) indicates that this part of Lewis was over-run during an earlier glaciation (Hall *et al.* 2018).

Local Outer Hebridean erratics, striae and glacial lineations indicate that an independent ice cap developed over the Outer Hebrides during the LGM. The ice cap had divides located close to the W coast of the Uists, across the mountains of Harris and S Lewis and extending N towards the Butt of Lewis (von Weymarn 1979; Peacock 1984; Hall 1996a). Indeed, computer modelling of the last ice sheet suggests that independent ice masses developed on northern Lewis, the mountains of Harris and, later, the Uists, and subsequently coalesced (Hubbard *et al.* 2009). Ice streams developed in the Sounds of Harris and Barra, discharging ice from The Minch and the Sea of the Hebrides onto the North Atlantic shelf (Fig. 13).
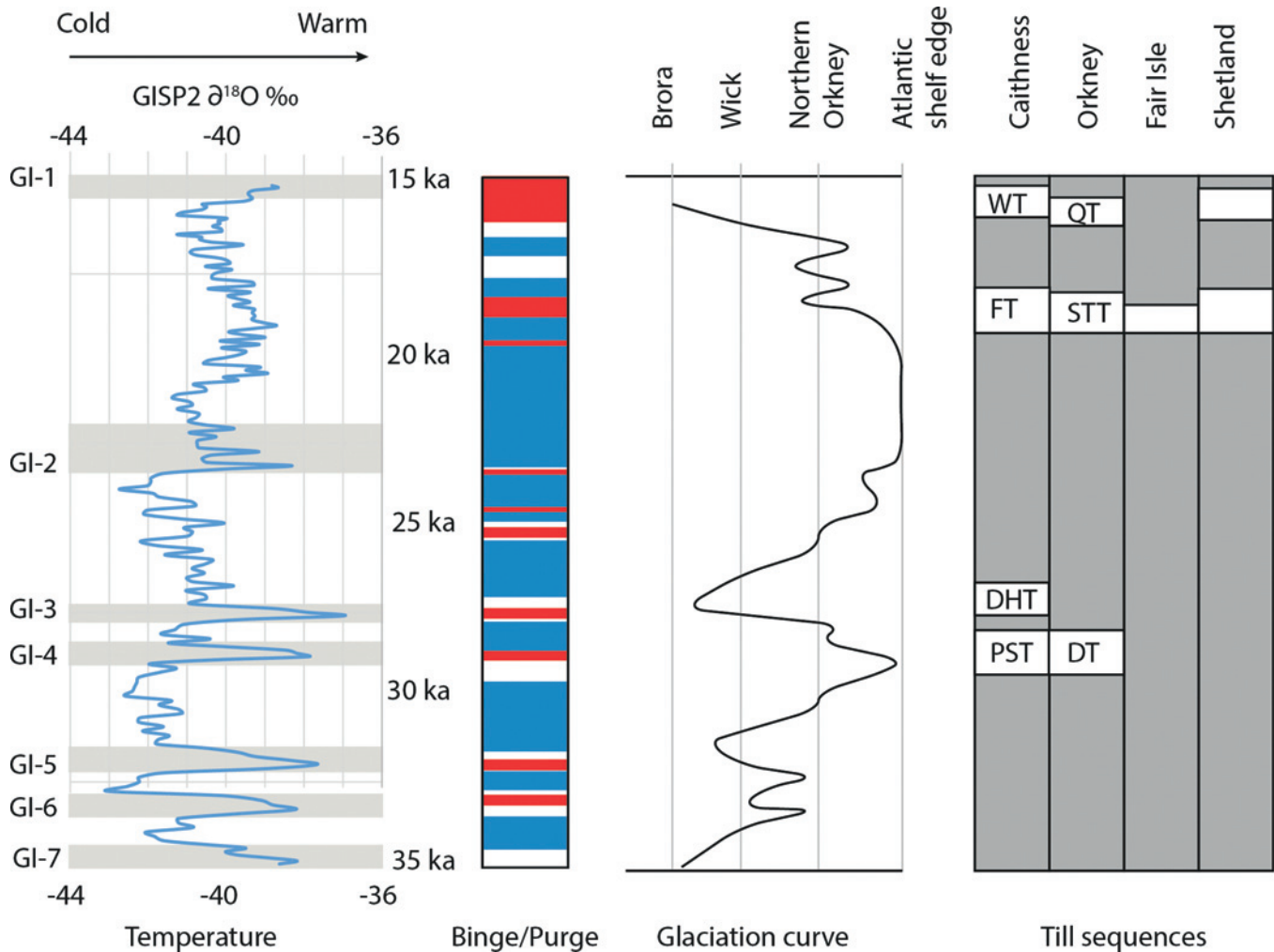
Figure 12 Time frame and glaciation curve for the main events of the last ice sheet on the Atlantic–North Sea Divide. GISP2 ice core data from Stuiver & Grootes (2000); binge-purge cycles and ice extent from Hubbard *et al.* (2009); till stratigraphy from Hall & Riding (2016) and Hall *et al.* (2016a). Abbreviations: DT = Dunbeath Till; DHT = Drumhollistan Till; FT = Forse Till; PST = Portskerra Till; QT = Quendale Till; STT = Scara Taing Till; WT = Wick Till.

The 'Galson Raised Beach' and its underlying raised shore platform at 4–10 m OD provides a valuable stratigraphic marker (Hall 1996a; Table 5). The raised beach rests on organic deposits at Toa Galson (Gordon 1993i) (here renamed based on its location as Sgarbh Sgeir), which, in turn, rests on weathered till and rock (Sutherland & Walker 1984). The beach gravel is cryoturbated N of Borve (Fig. 14), perhaps in more than one periglacial episode, and overlain by a single till unit derived from inland (*Lewis Till Formation* of Fig. 3b) (Hall 1995). The Galson raised beach has escaped glacial erosion beneath cold-based ice cover on NW Lewis (Hall 1995). To the N, on both coasts of the Butt of Lewis, the Galson Beach is overlain by a distinctive, interbedded sequence of glacigenic and water-lain sediments (Baden-Powell 1938; Gordon 1993i). These deposits, assigned to the *Port Beag Till Formation* were deposited by the Minch Palaeo Ice Stream (MPIS) crossing the northern tip of the island (Bradwell & Stoker 2015; Fig. 13). The SW boundary of this formation (Fig. 3b) is marked by a large ridge, either a marginal moraine (Sutherland & Walker 1984) or a wedge of sediment that accreted below a former shear margin within the ice sheet that separated warm-based, fast-flowing ice from sluggish cold-based ice (Bradwell & Stoker 2015). Radiocarbon dates from marine shells in the sequence have yielded ages of 34.4–39.5 cal ^{14}C ka, which, if correct, imply episodic establishment of ice-free conditions in The

Minch during the latter part of the Middle Devensian and a constraining maximum date for the ice advance that deposited these sediments (Sutherland & Walker 1984; Small *et al.* 2017).

Stronger evidence for ice-free conditions prior to the last glaciation has been obtained from the northern side of Tolsta Head (von Weymarn & Edwards 1973; Birnie 1983; Gordon & Sutherland 1993b; Whittington & Hall 2002; Figs 13, 15). The organic sediments preserved at Tolsta Head have been dated to 38–32 cal ^{14}C ka and contain an important record of dominantly open grassland vegetation during the 'Tolsta Interstadial' of MIS 3 (Whittington & Hall 2002; Table 4). Variations in palaeotemperature inferred from the pollen assemblages and sedimentology are provisionally matched by these authors with GI-8 to GI-5 (Fig. 2). The overlying Port Beag Till is a brown, matrix-supported, silty, sandy diamicton, with crude sub-horizontal structure. Clasts are predominantly Lewisian gneiss, but include hard, siliceous brown sandstones probably derived from the Torridonian inlier on the Shiant East Bank Ridge (von Weymarn & Edwards 1973; Whittington & Hall 2002) that lies some 40 km SE of Tolsta Head (Chesher *et al.* 1983). Von Weymarn & Edwards (1973) reported that the till fabric showed a dominant NW clast orientation that corroborates deposition by ice flowing from the SE, but striae inland of Tolsta Head trend E–W (Peacock 1981). The till was probably deposited within a zone of shifting ice flow where
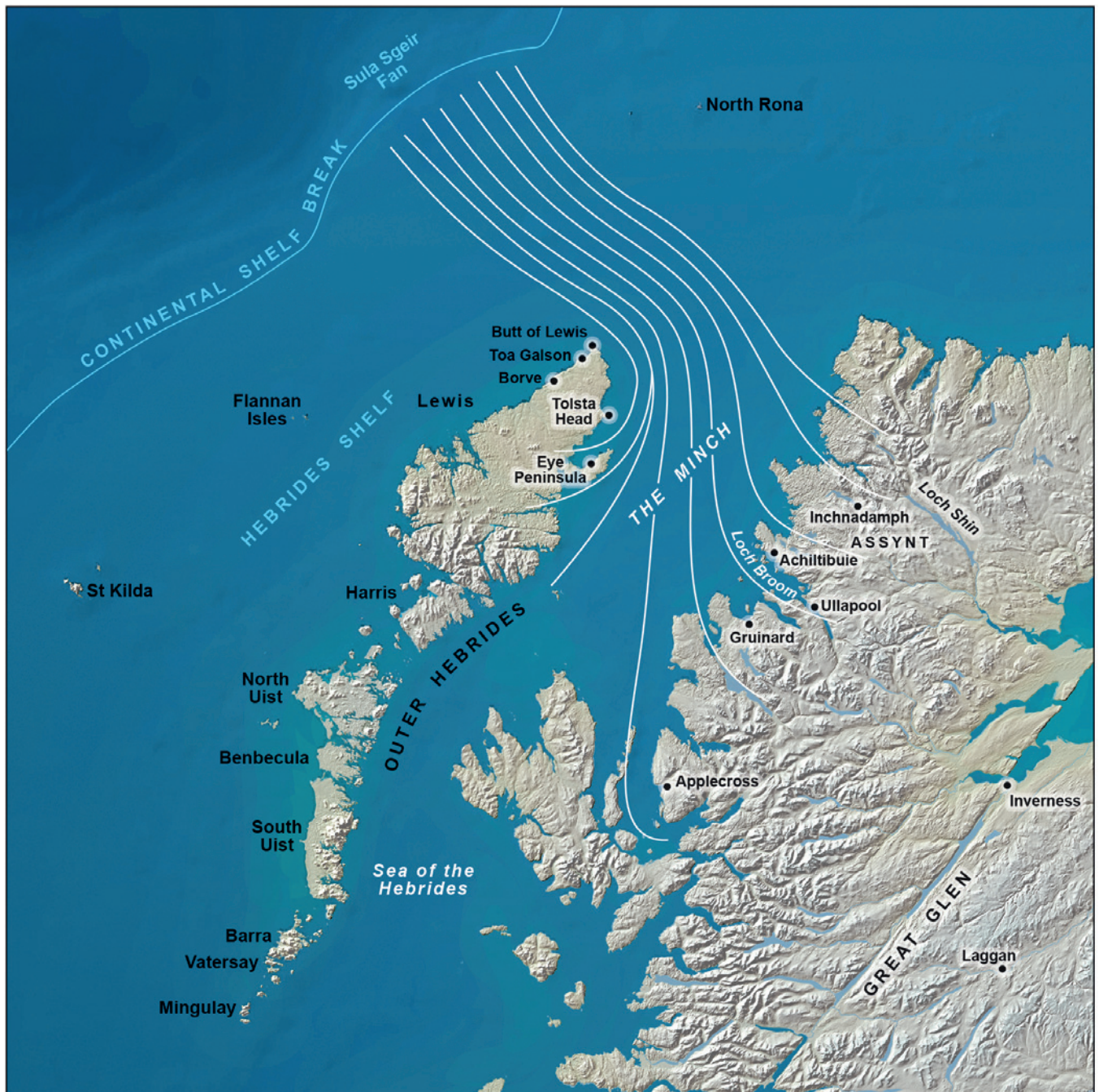
Figure 13 Physiography of the NW Highlands and Outer Hebrides showing the position of the former Minch Ice Stream (modified after Bradwell & Stoker 2015). Hill-shaded surface model built from Intermap Technologies NEXTMap Britain elevation data.

local Lewis ice bordered the western flank of the MPIS, which laid down a thick sequence of glacially overridden and tectonised diamictons, sands and gravels, with fragmented marine shells (Bradwell & Stoker 2015). Broadly equivalent deposits occur on the Eye Peninsula (Sutherland & Walker 1984; Fig. 13).

The outlying islands of St Kilda (Figs 13, 16) occupy a key location for understanding the extent of the last ice sheet on the Hebridean shelf (Sutherland *et al*. 1982). Earlier work on Hirta identified an older, weathered diamict, the *Ruaival 'Drift'*, assigned to a phase of local glaciation before the Late Devensian, and a younger till, the *Village Bay Till*, deposited by cirque glaciers on Hirta during the Late Devensian (Sutherland *et al*. 1984). The *Abhainn Ruaival* organic bed, radiocarbon-dated to 24.7 ka BP (SRR-1809b), occurs beneath the *Ruaival Head*, one of two units of periglacial slope deposit seen to lie between the two tills. The palynology of the organic

bed indicates the establishment of a species-poor grassland with virtually no shrub cover under interstadial conditions (Sutherland *et al*. 1984; Walker 1984). The bed is tentatively assigned to the 'Tolsta Interstadial' (Hiemstra *et al*. 2015), but the pollen record differs from that for the same interval at Tolsta Head, on Lewis, where *Fagus* and *Picea* are absent (Whittington & Hall 2002). Hiemstra *et al*. (2015) suggest returning to the earlier, simplified interpretation of the stratigraphy on Hirta (Wager 1953), challenging whether the Ruaival Drift is a glacial deposit.

Small erratic clasts of red sandstone and heavy minerals derived from Lewisian gneisses occur in the oldest till units recorded on St Kilda (Sutherland *et al*. 1984). The source of the sandstone is probably the Flannan Trough, lying 35–50 km to the E, which is floored by red-brown Permo-Triassic sandstone (Stoker *et al*. 1993). The heavy minerals are likely

Figure 14 The Galson Raised Beach on NW Lewis. (A) South Galson. Undisturbed storm beach gravel of likely MIS 5e age at ~5 m OD, overlain by cryoturbated gravel with erect clasts and layers of gravel-rich diamicton (Lewis Till) deposited by local ice. (B) Eoropie. A small raft of Galson Raised Beach gravel within brown mud-rich diamicton (Port Beag Till) deposited by the Minch Ice Stream.

Table 5 Summary of events in the Outer Hebrides. Abbreviations: B = glacial 'binge'; C = cold-periglacial; I = Interglacial; IS = Interstadial; P = glacial 'purge'.

Age	Events, palaeo-environments, important sites and units	
MIS 5e	Re-occupation and weathering of shore platforms at 4–10 m OD (excluding St Kilda). Creation of the Galson Raised Beach. Marine transgression and storm beach progradation on Lewis, Barra and Vatersay (Selby 1987)	I
MIS 5a–d	Establishment of interstadial grassland at Sgarbh Sgeir	C/IS
MIS 4	Cryoturbation of the Galson Raised Beach	C
MIS 3	Establishment of grassland at Tolsta Head and on St Kilda during the 'Tolsta Interstadial'. Marine and glaciomarine sedimentation in The Minch to provide shell fragments in later glacial deposits	C/IS
LLGM	Growth of a local ice cap between Lewis and Barra, with episodic ice stream development in the Sounds of Harris and Barra. Deposition of the *Lewis Till*. Cirque glaciers on St Kilda deposit the *Village Bay Till*	B
LGM	Incursion of the Minch Ice Stream into NE Lewis to deposit the *Port Beag Till*	P
18.9–17.1 ka	Retreat of ice into the mountains of Harris. Deglaciation of Mingulay and Barra	P

derived from the extensive submarine platform formed by Lewisian rocks E of St Kilda (Stoker *et al*. 1993). The last ice sheet glaciation of St Kilda probably occurred before MIS 5e (Sutherland *et al*. 1984; Hiemstra *et al*. 2015), consistent with the presence of large tors and pinnacles on ridges down to <100 m OD on Hirta (Fig. 16). Any more recent passage of fast-moving, erosive ice should have been capable of removing or modifying these fragile landforms. Recently acquired cosmogenic isotope ages support the conclusion that St Kilda supported only locally sourced glaciers during the LGM (Ballantyne *et al*. 2017). However, this interpretation contradicts the view that the last ice sheet terminated to the W of the archipelago at, or near, the Atlantic shelf edge (Bradwell *et al*. 2008a; Chiverrell & Thomas 2010; Gibbard & Clark 2011; Clark *et al*. 2012a, b).

4. Core areas

Considerable research has been undertaken in the NW and Central Highlands since the review by Sutherland (1993b). Both regions were centres of ice accumulation beneath which older deposits have been preserved locally due to relatively little subglacial erosion.

4.1. NW Highlands
The ice that flowed from an ice divide situated over the NW Highlands deposited tills, outwash and glaciolacustrine deposits assigned to the *North-west Highlands Glacigenic Subgroup* (Auton 2003; Bradwell 2003; McMillan *et al*. 2011a; Fig. 3a).

Three formations have been recognised to the W of the main N–S watershed following recent onshore and offshore mapping (Bradwell 2003, 2010a; Stoker *et al*. 2009; Bradwell & Stoker 2010), namely the *Loch Broom Till*, *Assynt Glacigenic* and *Ullapool Gravel* formations. To the E of the watershed, in eastern Sutherland and Caithness, the *Reay Burn Till* and *Dunbeath Till* formations also form part of the subgroup (Auton 2003) (Section 3.2; Fig. 3b). The Loch Broom Till has been mapped alongside, and beneath, Loch Broom and Gruinard Bay, and is interpreted as a subglacial traction till that was deposited by the MPIS (Fig. 13). The till has a streamlined surface morphology with lineations trending NW.

The Assynt Glacigenic Formation comprises clast-rich to clast-supported diamictons interbedded with sands and gravels. It is distributed discontinuously to the W of the main watershed, extending offshore, and includes well-developed, nested moraine ridges (Stoker *et al*. 2009; Bradwell & Stoker 2010). Several
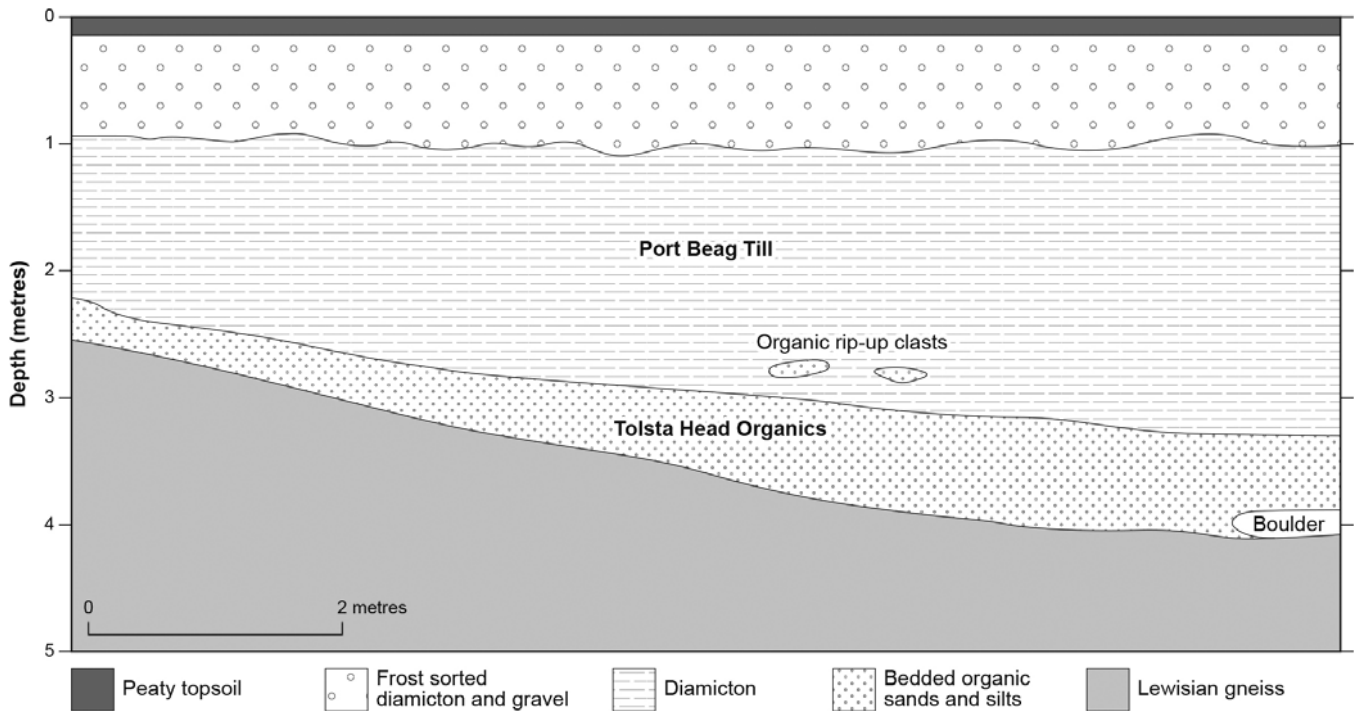
Figure 15 Generalised stratigraphy at Tolsta Head on NW Lewis (modified from Gordon & Sutherland 1993b).

Figure 16 Hirta, St Kilda, looking SW to Dùn, showing tor landforms mantled by thick periglacial slope deposits.

geographically distinct members have been recognised associated with late-stage glacier oscillations, namely the *Allt na h-Airbhe Member* (predominantly offshore), and the *Allt an t-Srathain*, *Rhiroy* and *Glen Douchary* members. The last of these occurs within the ice limits correlated with the Loch Lomond Stadial in the region (Ballantyne 2018). The Ullapool Gravel Formation includes glaciofluvial fan, deltaic and outwash deposits that formed during high relative sea levels (RSLs) around Ullapool, along margins of Loch Broom and adjoining sea lochs, and locally along the coast of the NW Highlands (Stoker *et al.* 2009; Bradwell 2010b). It is assigned to the Windermere Interstadial.

The tills of the NW Highlands typically contain clasts derived from local crystalline metamorphic and igneous rocks, Cambrian quartzites and Torridonian sandstones, but observations on the dispersal of sparse indicator erratics are revealing. For example, Cambrian Pipe Rock is found from NW Lewis to Hoy and into Moray, indicating that the ice shed once straddled the outcrop of this distinctive rock type in Assynt (Sutherland 1984; Fig. 13). The location of the ice shed is also suggested by the general distribution of erratics of 'Carn Chuinneag-Inchbae augen gneiss' (Fig. 17), which are found across Easter Ross and from Elgin eastwards towards Buchan (Sissons 1967a; Peacock 1970). However, these distinctive erratics also occur along Loch Broom, implying that the ice shed once lay E of the outcrop of the gneiss in Easter Ross, requiring a very different ice sheet configuration. Interestingly, in the ice sheet model of Hubbard *et al.* (2009) the ice shed is located over the coastal hills of Sutherland and Easter Ross for a time during the last glaciation, at 29.4 ka.

The significance of the region as an ice dispersal area has been enhanced by the recognition of the MPIS, which was fed by a number of tributaries sourced in the mountains of Kintail, Wester Ross, Assynt and Sutherland (Stoker & Bradwell 2005; Bradwell *et al.* 2007; Fig. 13). Geomorphological mapping based on NEXTMap imagery has elucidated the role of glaciation in shaping the landscape (Lukas & Bradwell 2010). For example, multiple cycles of glacial erosion during the Pleistocene have removed pre-existing regolith and decomposed bedrock along the western fringe to form extensive ice-scoured 'cnoc-and-lochain' topography (Sissons 1967a; Krabbendam & Bradwell 2014). Most notably, smooth, sub-parallel linear furrows and flutes in bedrock have been identified as 'megagrooves' (*cf.* Goldthwaite 1979; Bradwell 2005; Bradwell *et al.* 2007, 2008b; Krabbendam *et al.* 2016). Another swathe of megagrooves, whalebacks and rock drumlins has been identified to the E of the former regional ice divide in the valley of the River Endrick (Figs 17, 18), N of the Great Glen (BGS 2012). These features are likely to be the legacy of a former topographically unconstrained ice stream within the last ice sheet that flowed NE towards the Moray Firth basin (Merritt *et al.* 2013; Hughes *et al.* 2014).

The caves of the Allt nan Uamh and Traligill valleys in Assynt, including the so-called 'Bone Caves' of Inchnadamph (Fig. 13), provide critical evidence for Quaternary landscape evolution and chronology in the region (Hebdon *et al.* 1997; Table 6). Further investigations since the review by Lawson (1993) have provided new information on the cave sediments,
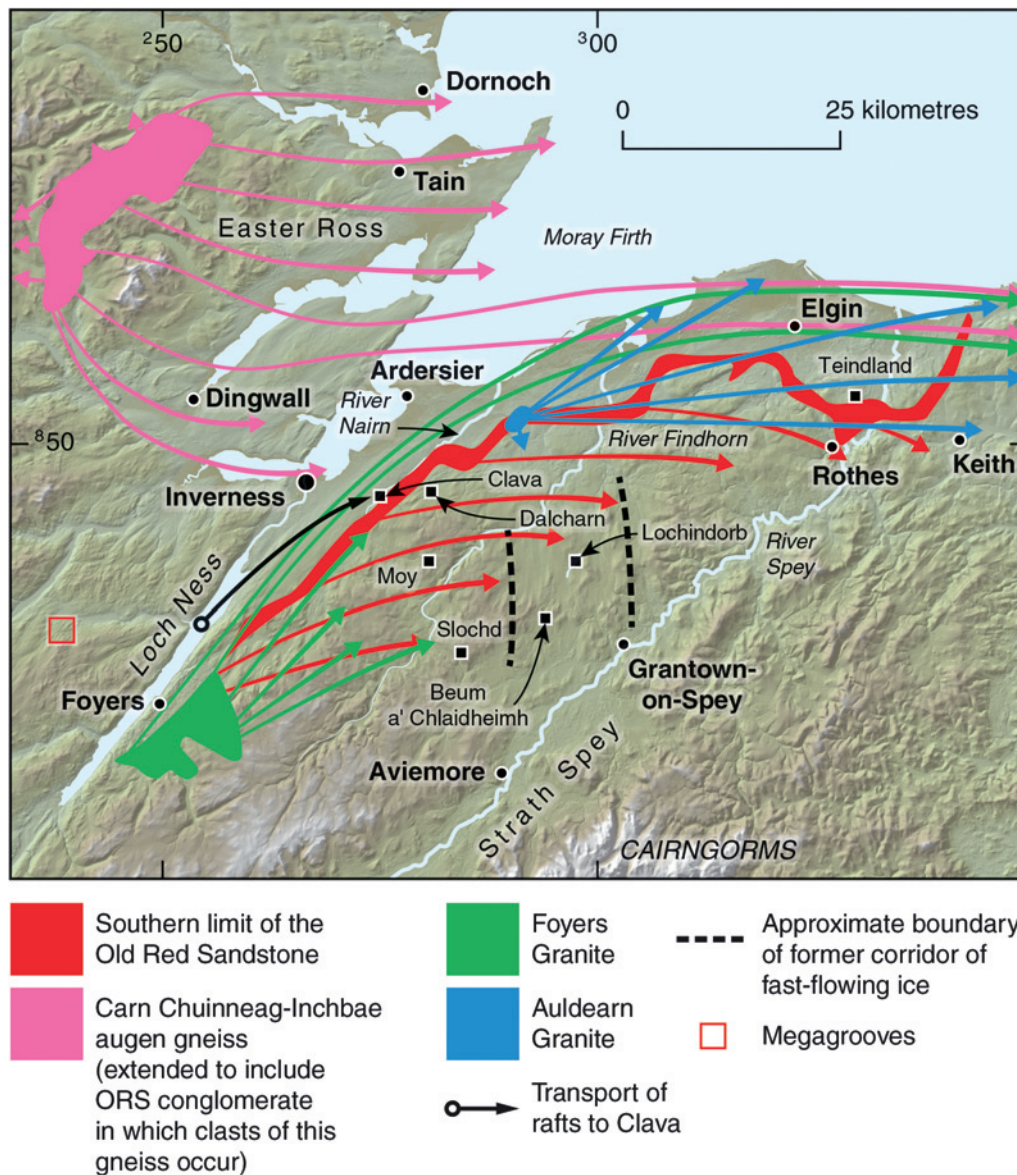
Figure 17 Transport paths of some indicator erratics across the inner Moray Firth (after Mackie 1905; Sissons 1967a; Fletcher *et al*. 1996; Merritt *et al*. 2017a). Hill-shaded surface model built from Intermap Technologies NEXTMap Britain elevation data.

faunal remains and chronology (Lawson 1995a, b, 2010; Lawson & Atkinson 1995; Kitchener & Bonsall 1997; Lawson *et al.* 2014). Uranium disequilibrium series dates on speleothems indicate ice-free conditions during the last (MIS 5e) interglacial, between 95 and 56 ka, at c. 38 ka and between 30 and 26 ka (Lawson 1995a; Lawson & Atkinson 1995). Radiocarbon dates on faunal remains (reindeer and brown bear) also imply that at least parts of the area were ice-free between ∼50 and ∼47 cal ^{14}C ka BP and between ∼36 and ∼26 cal ^{14}C ka BP (Lawson 2010; Lawson *et al.* 2014). Contrary to other evidence (Section 2), these dates appear to preclude extensive glaciation in the NW Highlands during the Early Devensian (particularly during MIS 5b and 5a where there is a cluster of dates) and conflict with interpretations of early build-up of the last ice sheet and its maximum extent on the Atlantic Shelf between ∼30 and 27 ka (Bradwell *et al.* 2008a; Ballantyne & Small 2018).

During deglaciation, the 'Wester Ross Readvance' produced a series of end and lateral moraines between Applecross and Achiltibuie (Ballantyne *et al.* 2009a; Ballantyne & Small 2018; Fig. 13). Recalibration of earlier terrestrial cosmogenic nuclide age estimates places this event at 15.3 ± 0.7 ka at the beginning

of the Last Glacial–Interglacial Transition (Ballantyne & Small 2018). On the E side of the ice divide, an oscillation of the ice sheet margin is indicated by a currently undated moraine system alongside Loch Shin (Lukas 2010; Fig. 13).

4.2. Central Highlands

The Pleistocene deposits in upper Strathspey (Fig. 17), the middle reaches of the Findhorn and to the W of the Pass of Drumochter have been assigned to the *Central Highland Glacigenic Subgroup* (Fig. 3a). The eastern boundary of this subgroup divides those areas affected by ice flowing from the main N–S ice divide of the last ice sheet, situated over the western Highlands, from mountains affected by locally sourced ice, notably the Cairngorms and Gaick (British Geological Survey 2008a, b, 2013; Fig. 3a). The oldest known glacigenic unit in the sequence is the *Pattack Till*, which most probably antedates the last interglacial (MIS 5e) judging by its weathered condition (Hall *et al.* 2018; Table 7). It is locally overlain by the *Ceardaich Sand and Gravel*, which was deposited as glaciofluvial outwash during an early stage of the last glaciation when ice advanced eastwards up reverse-draining valleys

Figure 18 Megagrooves on the western flank of Suidhe Ghuirmain in the Endrick valley, N of the Great Glen, looking SE. Inset: rock drumlin associated with the megagrooves. Ice flowed towards the Moray Firth, to the NE (left). See Figure 17 for location.

Table 6 Summary of events in the NW Highlands. Abbreviations: B = glacial 'binge'; C = cold-periglacial; I = Interglacial; IS = Interstadial; P = glacial 'purge'.

Age	Events, palaeoenvironments, important sites and units	
MIS 5e	Speleothem growth in the Allt nan Uamh caves indicate fully ice-free conditions	I
MIS 5a–d	Intermittent growth of speleothems in the Allt nan Uamh caves indicate mainly ice-free conditions, especially during MIS 5a–b	C/IS
MIS 4	Intermittent growth of speleothems in the Allt nan Uamh caves indicate mainly ice-free conditions	C/IS
MIS 3	Intermittent speleothem growth in the Traligill caves and radiocarbon dates on reindeer antlers and brown bear bones from the Allt nan Uamh caves indicate periods of ice-free conditions	C/IS
LLGM?	Speleothem growth in the Allt nan Uamh caves and radiocarbon dates on reindeer antlers and brown bear bones from the Allt nan Uamh caves suggest ice-free conditions 36–26 ka	C/IS
LGM	Westward expansion of the last ice sheet with ice streams flowing offshore (e.g., MPIS). Deposition of the *Loch Broom Till*	B
ca.21–15.3 ka	Partial deglaciation accompanied by deposition of moraines and outwash of the *Assynt Glacigenic Formation*	P
15.3 ka	Wester Ross Readvance	B
15.3–14.7 ka	Renewed deglaciation interrupted by local readvances and deposition of moraines, outwash and deltaic deposits	P

(Merritt 1999; British Geological Survey 2002). Deltaic glacio-fluvial sequences locally grade downwards into laminated clay and silt with sparse lonestone pebbles (*Linn of Pattack Silt*). Similar sequences occur to the S of the Gaick plateau, where sections displaying till-capped fine-grained sediments correlated with the Linn of Pattack Silt Formation indicate that ice first blocked the lower reaches of southward draining valleys, causing ponding, before it overrode the deposits from the W and impinged against the plateau (Merritt 2004a). No evidence for an early expansion of ice from the Gaick has been reported, but granite-rich till derived from the Cairngorms is locally overlain by psammite (metasandstone)-rich till derived from the W, recording an early expansion of Cairngorm ice (Barrow *et al*. 1913; British Geological Survey 2013).

The laminated deposits of the Linn of Pattack Silt Formation typically have been deformed into glacitectonite (*sensu* Benn & Evans 2010) at the base of the overlying *Ardverikie Till* (Merritt 1999). The latter includes most surficial tills in the region and is typically a stony diamicton with boulders of grey porphyritic granodiorite from the Rannoch area in addition to locally sourced psammitic rock types (Fig. 3b; McMillan *et al*. 2011a). The Ardverikie Till extends N across the Spey–Findhorn catchment divide where it commonly overlies very sandy diamicton of the *Inverness Glacigenic Subgroup* (Horne & Hinxman 1914; Fig. 3b). The Ardverikie Till is correlated with the *Beinn an Uain Till* at the Dalcharn and Moy sites E of Inverness (Figs 3b, 17; Section 5.1).

Table 7 Summary of events in the Central Highlands. Abbreviations: B = glacial 'binge'; P = glacial 'purge'.

Age	Events, palaeo-environments, important sites and units	
MIS 5e	No reported evidence	
MIS 5a–d	No reported evidence	
MIS 4	Possible deposition of the *Pattack Till* during an Early Devensian glaciation	B
MIS 3	No reported evidence	
LLGM	Ice advanced from W ponding lakes into which the *Ceardaich Sand and Gravel* and *Linn of Pattack Silt* were deposited as fan-deltas	B
	Most of district excepting the Cairngorms and eastern Gaick engulfed by ice from W depositing the *Ardverikie Till*	B
LGM	Ice streamed N from upper Strathspey across the Spey–Findhorn divide through a narrow corridor centred on Lochindorb (Fig. 17)	P
18–15 ka	A huge outlet glacier becomes established in upper Strathspey when ice could no longer flow across the Spey–Findhorn divide	P
15.1 ka	Local readvance of Strathspey glacier into Rothiemurchus followed by active retreat towards Loch Etteridge	B

Northward flow of ice across the Spey–Findhorn divide is indicated by streamlined subglacial bedforms (Hughes *et al.* 2014), the orientation of glacial striae, the clast-composition of the surficial till units (Fletcher *et al.* 1996) and the dispersal of erratics of Cairngorm Granite towards Lochindorb (Sutherland 1984). The northward flow was concentrated within a narrow corridor of strongly glacially modified terrain centred on Lochindorb, whereas the landscape to either side, and particularly to the E, appears to have experienced relatively little cumulative glacial modification (Merritt *et al.* 2013, 2017b; Hughes *et al.* 2014; Fig. 17). Evidence for strong northward flow towards the corridor includes clusters of large-scale ('mega') crag-and-tail features in the lower Dulnain valley, W of Grantown-on-Spey (Young 1977b; British Geological Survey 2013) and mega-scale rôches moutonnées (Creagan a' Choin and Ordan Shios), 25 km SW of Aviemore (British Geological Survey 2008b; Merritt *et al.* 2013; Merritt 2017a).

The absence of far-travelled erratics within the Cairngorms and eastern Gaick indicates that local ice caps had developed there before the district became overwhelmed by ice sourced from the W (Bremner 1929; Sugden 1970; Gordon 1993h). The Cairngorms were entirely buried beneath ice, based on evidence of lee-side joint-block removal, glacially transported tor blocks and stripping of blockfields at elevations of up to 1200 m (Hall & Glasser 2003; Phillips *et al.* 2006). Ice flowed against the Cairngorm massif, where it was deflected by local cold-based ice (Sugden 1968; Hall & Sugden 1987; Sutherland 1993e), but probably penetrated eastwards into the catchment of the River Dee (Barrow *et al.* 1913; Merritt *et al.* 2004; Merritt 2004b). Cairngorm ice apparently re-occupied the uppermost catchment of the Dee during deglaciation following the retreat of ice sourced from the W (Merritt 2004b, fig. 24).

A topographically constrained corridor of relatively fast-flowing ice (ice stream) centred on upper Strathspey probably operated late in the LGM at the time when ice was flowing northwards across the Spey–Findhorn divide, via the narrow 'Beum a' Chlaidheimh Breach' and Lochindorb (Merritt *et al.* 2017a, b; Fig. 17). Considering the elevation of the breach at 360 m, ice is likely to have stopped flowing across the divide soon after deglaciation of the Cairngorm tops, which commenced at *ca.* 21.3 ka (Phillips *et al.* 2006; Table 7). A recalibrated mean cosmogenic age of 18.2 ka (Ballantyne *et al.* 2009b; Ballantyne & Small 2018) for rockslide debris suggests early deglaciation in Strath Nethy, in the NW Cairngorms. After ice became too thin to flow across the divide, a huge outlet glacier became established in upper Strathspey. Recent cosmogenic dating of boulders on an ice-marginal moraine in Rothiemurchus, a large depression lying to the W of the Cairngorms, suggests that the Strathspey glacier

was still active and readvancing at *ca.* 15.1 ka (Hall *et al.* 2016b). Widespread ponding occurred during and following the decoupling of local Cairngorm glaciers from the Strathspey glacier sourced in the W (Brazier *et al.* 1998). This was associated with the aggradation of glaciofluvial fans and deltas on the southern flanks of the Spey–Findhorn divide at elevations of 300–350 m OD (Merritt *et al.* 2013). Meltwaters carrying clasts of Old Red Sandstone (ORS) lithologies flowed southwards through a deep channel on the divide, at Slochd Mòr. They were then deflected eastwards along the margin of the outlet glacier in Strathspey before flowing back northwards across the divide through the Beum a' Chlaidheimh Breach (British Geological Survey 2013; Merritt 2017a; Fig. 17).

There is abundant evidence, such as glacial drainage channels on the valley flanks, for the retreat of the Strathspey outlet glacier (Young 1974, 1975a, b, 1977a, b, 1978; British Geological Survey 2004, 2013; Merritt *et al.* 2013). Some ice-marginal channels and benches are intimately associated with lateral moraine ridges, and evidence from Raitts Burn, 16 km SW of Aviemore (Phillips & Auton 2000, 2013; Phillips *et al.* 2007), indicates that the glacier remained active during its retreat towards Loch Etteridge, 8 km E of Laggan (Fig. 13), an important late-glacial site that had become free of ice by 15.6 cal ^{14}C ka BP (Lowe *et al.* 2008a; Ballantyne *et al.* 2009b; Ballantyne & Small 2018). Meltwaters were firstly ponded behind temporary barriers of stagnant ice and glacial sediment in the Middle Findhorn during deglaciation, before being constrained to follow the present axis of the valley towards the coast (Bremner 1939; Auton 1990, 1998, 2017a, b). In contrast, much of the meltwater from decaying ice in the upper reaches of the Findhorn Valley drained NW into the Nairn valley, depositing broad spreads of kettled outwash sand and gravel around Moy (Fletcher *et al.* 1996; British Geological Survey 1997; Merritt *et al.* 2017b; Fig. 17)

5. Lowland depocentres

5.1. Inverness and Moray
This region, lying between the Inverness area and the mouth of the River Spey, E of Elgin, includes two of the few sites in Scotland at which organic material has been assigned to the Ipswichian Interglacial (Sutherland & Gordon 1993c; Hall *et al.* 2018, fig 13; Table 8). At Dalcharn, 15 km E of Inverness, the *Dalcharn Palaeosol Formation* (Figs 17, 19) contains compressed and disseminated organic material that includes pollen of full interglacial affinity reflecting the middle and later stages of an interglacial cycle (Walker *et al.* 1992; Auton 1993; Merritt & Auton 2017). The organics appear to be relatively unweathered, but they rest on the *Craig an*

Table 8 Summary of events around Inverness and Moray. Estimated ages after Merritt *et al.* (2017a). Abbreviations: B = glacial 'binge'; C = cold-periglacial; I = Interglacial; IS = Interstadial; P = glacial 'purge'; R = glacial readjustment.

Age	Events, palaeo-environments, important sites & units	
MIS 5e	Formation of palaeosols at Dalcharn and Teindland	I
MIS 5a–d	Accumulation of the *Odhar Peat* at Moy, probably during MIS 5c.	C/IS
MIS 4	Possible deposition of the *Woodside Diamicton* at Teindland and concealed sandy tills around Dalcharn and Moy during an Early Devensian glaciation.	B
MIS 3	Possible accumulation of the Odhar Peat early in MIS 3. Deposition of the *Clava Shelly Clay* in the Loch Ness basin.	C/IS
LLGM	ESE flow across the Great Glen and inner Moray Firth lays down the sandy *Athais Till*. Dislocation and transport of rafts to Clava.	B
∼24 ka?	Ice flow to E deposits the *Cantray Till* at Dalcharn (Fig. 19).	R
∼23 ka?	Northward flow across the Spey-Findhorn divide lays down the *Ruallan Till* at Dalcharn (Fig. 19).	P
LGM	Long period of glaciation laying down *Beinn an Uain Till* and *Finglack Till* in part.	B
∼16–15 ka	Partial deglaciation of the Middle Findhorn and establishment of the Moray Firth Ice Stream (MFIS) that lays down the *Finglack Till* around the inner Moray Firth following the Elgin Oscillation.	P
	Minor oscillations of the MFIC during retreat. Deposition of the glaciomarine *Ardersier Silts* at ∼30 m OD.	P
∼15–14.7 ka	Major readvance of the MFIS at Ardersier lays down the *Baddock Till* on top of the Ardersier Silts.	R/B
	Stillstand at Alturlie and deposition of the *Bothyhill Gravels*.	R

Daimh Gravel, which is substantially decomposed, indicating that it has been subjected to prolonged weathering under warm, humid conditions. The pollen record indicates that closed pine forest with birch, alder and holly was succeeded by pine-heathland, which was replaced initially by birch and later by heath and open grassland (Walker *et al.* 1992). As the Dalcharn site lies near to the present northern limit of holly in Britain, the relative abundance of *Ilex* pollen in the profile almost certainly reflects a climate somewhat warmer than that of today. The pollen assemblage is similar in some respects to that described from Fugla Ness, on Shetland (Section 3.1), which was originally equated with the Gortian of Ireland (Birks & Ransom 1969), and, hence, with the Hoxnian of southern Britain. However, samples of sandy matrix collected from a unit of white, cryoturbated gravel taken from beneath the uppermost organic part of the palaeosol at Dalcharn have yielded maximal luminescence ages of 68 and 50 ka (Duller *et al.* 1995). The ages are not at variance with an infinite radiocarbon date of >41.3 ^{14}C ka BP obtained from a bulk sample of the organic unit above, but suggest that the thermophilous arboreal pollen in the palaeosol must be either allochthonous (Whittington 1990) or reworked.

At another important site at Teindland (Fig. 17), near Elgin, where a preserved section reveals a truncated podzolic soil, the *Teindland Palaeosol* contains pollen of thermophillous species and grasses and has also been assigned to the latter part of the Ipswichian Interglacial (Sutherland 1993c; Hall *et al.* 1995a; Hall *et al.* 2018, fig. 21). Luminescence ages of around 79 and 67 ka obtained by Duller *et al.* (1995) from the *Badentinian Sand Bed* (Table 9) above the main organic horizons support the age suggested. Although the genesis of the overlying *Woodside Diamicton* is unclear, the presence of an isoclinal fold and shear zones within the sand bed strongly suggests post-depositional glacitectonism (Merritt *et al.* 2003; Connell & Phillips unpublished). It follows that the uppermost part of the sand may have been eroded and, thus, the Thermoluminescence (TL) dates could significantly predate the initial glaciation of lower Strathspey during the last glacial cycle. However, the simplest interpretation would suggest that glaciation occurred soon after its deposition (Table 8).

A compressed bed of fibrous peat was discovered beneath till at the Allt Odhar site on the Moy estate, some 16 km SE of Inverness, at 371 m OD (Walker *et al.* 1992; Merritt 2017b; Fig. 20). This *Moy Peat Formation* contains pollen, plant-macrofossils and coleopteran remains that reveal climatic

deterioration towards the end of a cool interstadial as reflected in the replacement of birch woodland and willow scrub with grassland and heath, and then by open communities of grass and sedges (Walker *et al.* 1992). An Early Devensian MIS 5d or 5b age is not at variance with three infinite radiocarbon age determinations (>51.1 ^{14}C ka BP) (Walker *et al.* 1992), nor with an experimental U-series disequilibrium measurement of 106 ka for the peat (Heijnis & van der Plicht 1992). However, samples collected from sand-rich lenses within the peat unit have subsequently yielded luminescence ages of between 37 and 58 ka (Duller *et al.* 1995). These ages are consistent with a single finite radiocarbon age of 47.7 ^{14}C ka BP obtained from the alkali insoluble component of samples of the peat (Walker *et al.* 1992). The time span of 20 ka between the TL ages for the top and bottom of the peat is difficult to accept, but not totally at variance with the entomological palaeo-environmental evidence. On reflection, a Middle Devensian age is possible, especially as pollen grains of *Bruckenthalia spiculifolia* (Balkan heath) are not recorded (M. J. C. Walker, pers. comm. 2017). This species has been reported from Early Devensian interstadial deposits in NE Scotland (Whittington 1994; Section 5.2), although at much lower altitudes.

The Ipswichian biogenic deposits at Dalcharn (Fig. 19) and the Early (or Middle) Devensian interstadial peat unit at Allt Odhar (Fig. 20) are overlain by the sandstone-rich *Athais Till Formation* of the *Inverness Glacigenic Subgroup*, which has a fabric indicating ice flow towards the SE (Walker *et al.* 1992; Auton 1993; Merritt 1993; Merritt 2017b; Merritt *et al.* 2017a). The distribution of other concealed sandy diamictons of this subgroup (Fig. 3a) suggests that during a relatively early phase of the last glaciation, if not before, ice flowed across the Great Glen towards the ESE, against the north-facing slopes of the Inner Moray Firth (Sutherland & Gordon 1993c; Fletcher *et al.* 1996). The ice carried sandstone, siltstone and conglomerate from ORS outcrops around Loch Ness eastwards towards Grantown-on-Spey onto ground up to at least 500 m OD on the Nairn–Findhorn watershed and into the middle to lower reaches of the Nairn, Findhorn and Spey valleys (Hinxman & Anderson 1915; Horne 1923; Sutherland & Gordon 1993d; Merritt *et al.* 2017b; Fig. 17). The sandy Athais Till is generally overlain by psammite-rich till (*Beinn an Uain Till Formation*) deposited by ice flowing NNE (Fletcher *et al.* 1996). Following the systematic logging of numerous natural sections E of Inverness, Merritt & Auton (1993) found that the contact between the sandstone-rich and
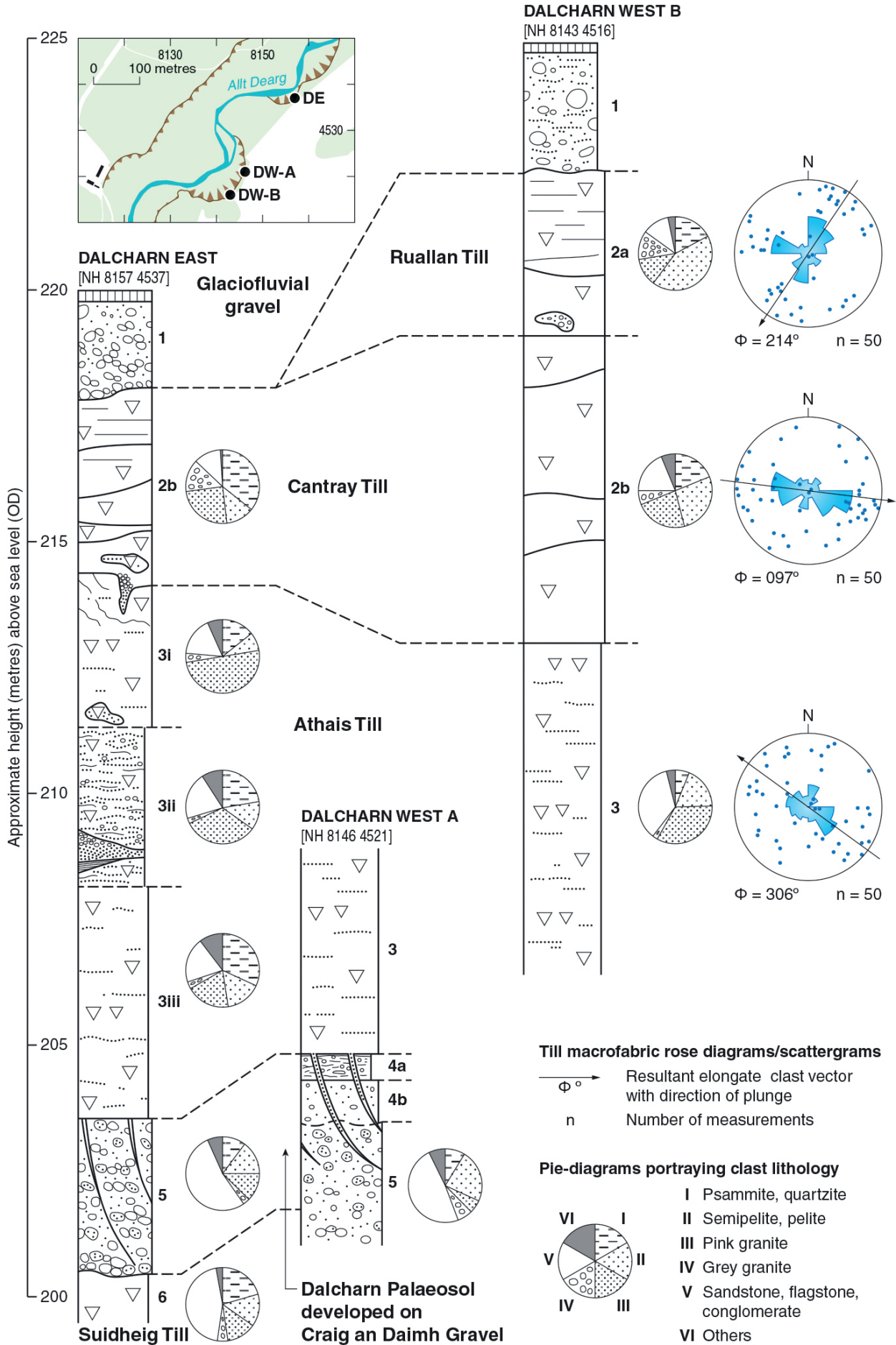
Figure 19 Summary logs of sections recorded at Dalcharn (modified after Walker *et al*. 1992).

Table 9 The lithostratigraphy in the vicinity of Teindland Quarry, near Elgin.

| Unit	Formal name (Merritt *et al.* 2003)	Original name (Hall *et al.* 1995a)	Depositional environment	MIS
9	Waterworks Till Formation		Glacial	2
8	Tofthead Till Formation		Glacial	2
7	Altonside Till Formation		Glacial	?2
6	Woodside Diamicton Formation	Teindland Till	Glacial	?4
5	Badentinian Sand Bed	Teindland Upper Sand	Slopewash	5a/4
4	Teindland Palaeosol Bed	Teindland Buried Soil	Interglacial soil	5e
3	Orbliston Sand Bed	Teindland Lower Sand	Glaciofluvial	6
2	Deanshillock Gravel Formation	Teindland Gravel	Glaciofluvial	6
1	Red Burn Till Formation		Glacial	6
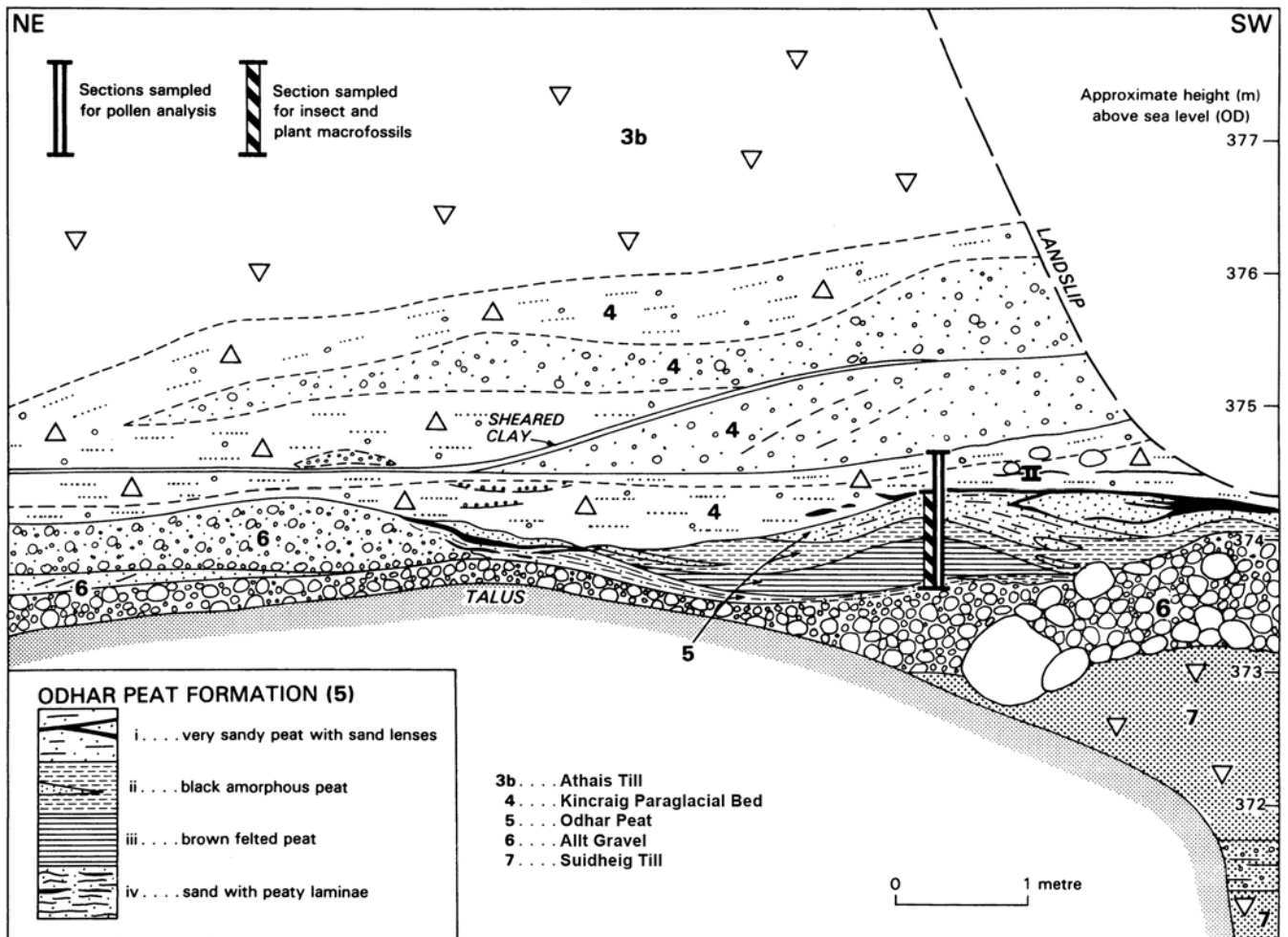
Figure 20 The Allt Odhar section showing details of the Odhar Peat Formation and the position of the sampling points for pollen, insect and plant macrofossil remains (modified from Walker *et al.* 1992). The terminology follows McMillan *et al.* (2011a).

overlying psammite-rich till was generally represented by a sharp change in colour and commonly by a planar discontinuity suggestive of a hiatus in aggradation. However, no major unconformities, distinct weathering profiles, palaeosols or organic deposits were found at the boundary, indicating deglaciation, although the presence of downward-tapering, wedge-shaped pull-apart structures, truncated gravel-filled channels, seams of glacitectonite and rip-up clasts of diamicton suggest that ice-bed boundary conditions changed during the switch in flow (Fletcher *et al.* 1996). At Dalcharn, till fabrics reveal that ice flow first swung around to the E before heading NNE (Walker *et al.* 1992; Fig. 19). This inferred direction of flow suggests that the uppermost till unit at Dalcharn was deposited when ice flowed directly towards the Moray Firth basin across

the Spey–Findhorn divide from centres positioned over the Cairngorms and western Highlands (Fig. 17; Merritt *et al.* 2013; Hughes *et al.* 2014). Both the sandstone- and psammite-rich till units could be Middle to Late Devensian in age if the youngest luminescence age at Moy (37 ka) is accepted (Duller *et al.* 1995; Merritt *et al.* 2017a; Table 8).

At Clava, in the Nairn valley (Fig. 17), there is an important, well-documented site long known for discoveries of 'arctic' shelly clay and shelly diamicton (Merritt 1992; Gordon 1993e; Merritt & Phillips 2017). At elevations of 150 m OD, the *Clava Shelly Clay* contains well-preserved, high-boreal to low-arctic, shallow-water marine shells that have yielded amino acid racemisation (AAR) ratios no younger than the Middle Devensian (Merritt 1992). Furthermore, an associated

folded raft of pebbly diamicton (*Clava Shelly Till*) containing fragments of *Portlandia arctica*, an indicator of shallow-marine, fully arctic conditions, yielded a radiocarbon age of >47.1 cal ^{14}C ka BP. These sediment units are now accepted to be glacial rafts transported eastwards from the Loch Ness basin (Fig. 17), which was probably a marine fjord during the Middle Devensian (Merritt 1992; Phillips & Merritt 2008; Merritt *et al.* 2014). The rafts are overlain by the *Finglack Till Formation*, which was laid down by NE-directed ice that flowed towards the Inner Moray Firth, forming elongated subglacial landforms of this orientation (Hughes *et al.* 2014, flow-set 6; Ballantyne & Small 2018, fig. 4). The sandy Finglack Till crops out around Inverness, Nairn, Forres and across the Black Isle (Fletcher *et al.* 1996; Fig. 3b).

Evidence of generally falling RSLs around Inverness during deglaciation is preserved in flights of raised late-glacial shorelines lying up to about 30 m OD around the inner Moray Firth (Firth 1989; Firth & Haggart 1989; Smith *et al.* 2018). These shorelines truncate widespread deposits of glaciofluvial outwash, silt and clay to the E of Inverness assigned to the *Alturlie Gravels* and *Ardersier Silts formations*, respectively (Fletcher *et al.* 1996; British Geological Survey 1997). These raised deposits, which are interpreted to be glaciomarine in origin on account of the cyclic sedimentation they commonly display, were laid down in front of an active outlet glacier that flowed into the Inner Moray Firth (Merritt *et al.* 1995). The Ardersier Silts were deformed during a substantial readvance that formed a push moraine on the Ardersier peninsula and a capping of diamicton (*Baddock Till Member*) (Merritt *et al.* 1995; Fletcher *et al.* 1996; Merritt & Firth 2017; Merritt *et al.* 2017b). A tidewater fan-delta (*Bothyhill Gravel Member*) formed during a subsequent stillstand at Alturlie Point (Merritt *et al.* 1995, 2017c; Fletcher *et al.* 1996). These glacial oscillations, and an earlier one that formed a possible push moraine at Grange Hill, between Forres and Kinloss (Auton 2017c), are currently undated. They probably result partly from the competing strengths of ice flowing from the Great Glen and through the Beauly Firth basin, and partly from the delicate interplay between contemporary rates of global sea-level rise and local glacio-isostatic uplift (Merritt *et al.* 1995; Smith *et al.* 2018).

The results of recent geophysical surveys and lake-bed sampling in Loch Ness suggest that the frontal margin of the Ness Glacier terminated in water throughout its retreat (Turner *et al.* 2012). The glacier initially had a lightly grounded or possibly floating ice shelf, forming a series of low, cross-valley ridges on the lake bed. It had become fully grounded by Foyers (Fig. 17), where a major moraine ridge was formed during a stillstand. The subsequent retreat of the glacier was dynamic and punctuated by minor stillstands and oscillations that formed smaller moraines beneath the loch.

5.2. NE Scotland

A comprehensive, critical review of the complex lithostratigraphical and geomorphological record of the last glaciation of NE Scotland (River Spey around Buchan to Montrose) has been given by Merritt *et al.* (2017a) and so is not repeated here in detail. Descriptions of the sites in the region that are mentioned below and that have provided important stratigraphical information are given by Sutherland & Gordon (1993a) and Merritt *et al.* (2003, http://earthwise.bgs.ac.uk/index.php/Cainozoic_geology_and_landscape_evolution_of_north-east_Scotland._Memoir_of_the_British_Geological_Survey,_sheets_66E,_67,_76E,_77,_86E,_87W,_87E,_95,_96W,_96E_and_97_(Scotland)

The informal lithostratigraphic 'series' of glacigenic deposits in NE Scotland established by Sutherland (1984) and Hall &

Connell (1991) have been replaced by more formal subgroups (McMillan *et al.* 2011a; Fig. 3a). The 'blue-grey series' of Synge (1956), renamed as the *Banffshire Coast, Caithness and Orkney Glacigenic Subgroup*, relates to a suite of generally thickly developed diamictons, sands and gravels laid down by ice that flowed onshore from the Moray Firth (Fig. 21). The 'red series' of Jamieson (1906) and Synge (1956) formed of generally thickly developed reddish-brown diamictons, sands and gravels has been subdivided. This is in order to separate the typically more vivid reddish brown diamictons, sands, silts and gravels that were laid down by ice that encroached onshore from the North Sea basin (*Logie-Buchan Glacigenic Subgroup*), from those that were derived entirely from ice flowing NE along Strathmore (*Mearns Glacigenic Subgroup*). The 'inland series' of Hall (1984) comprising generally thin, very sandy and commonly weathered diamictons, sands and gravels has become the *East Grampian Glacigenic Subgroup*.

NE Scotland, and particularly Buchan (Fig. 21), includes several sites that are important for understanding pre-Devensian events (Hall *et al.* 2018). At the Kirkhill and Leys sites (Fig. 21), the truncated *Fernieslack (Kirkhill Upper) Palaeosol Bed* (Hall *et al.* 2018, fig. 14) is assigned to the Ipswichian following TL dates on sediments underlying it (Duller *et al.* 1995; Tables 10, 11). Early Devensian stadial/interstadials recording fluctuating cool and colder environmental conditions may be represented at the Camp Fauld and Crossbrae sites in Buchan (Fig. 21; Whittington *et al.* 1993, 1998; Merritt *et al.* 2003). Elsewhere, the *Burn of Benholm Peat Bed* at the Burn of Benholm, S of Inverbervie (Fig. 21), is also now correlated tentatively with MIS 5c on the basis of the presence of pollen of the Balkan heath, *Bruckenthalia spiculifolia*, which occurs at Camp Fauld and Crossbrae (Whittington 1994; Auton *et al.* 2000). At Benholm, the erratics within the shelly, clayey diamicton (*Benholm Clay Formation*) that underlies the peat bed indicate that it was laid down by ice flowing onshore from the North Sea basin, probably during MIS 6 (Merritt *et al.* 2003; Hall *et al.* 2018).

The last ice sheet witnessed abrupt changes in ice-flow direction and geometry (Hughes *et al.* 2014), but the peripheral location of this area in respect to accumulation centres in the W and NW probably resulted in large sectors remaining cold-based for long periods. This is suggested by palimpsest periglacial features (*cf.* Kleman 1994) preserved in central Buchan, where the stratigraphical record is longest (Connell & Hall 1987; Hall *et al.* 2018).

It has been suggested that much of Scotland was glaciated during the Early Devensian (MIS 4) (Sutherland 1981) and Hibbert *et al.* (2010) have recorded contemporaneous IRD with a Scottish provenance in the offshore Barra-Donegal Fan (Hall *et al.* 2018, fig. 1) implying a marine-terminating ice sheet during this period. However, with the possible exception of the *Woodside Diamicton* at Teindland (Hall *et al.* 1995a; Table 8), no tills in the region can be confidently assigned to this glacial event, including the *Hythie Till* (Kirkhill Upper Till) (Hall & Jarvis 1993a), and the *Pitlurg Till*, at Bellscamphie, near Ellon (Hall & Jarvis 1993b, 1995) (Fig. 21). Several sites probably include a record of this period in the form of periglacial phenomena; for example, the *Corsend Gelifluctate Bed* at Kirkhill (Table 11; Hall *et al.* 2018, figs 16, 18). At Oldmill, in Buchan, and Nigg Bay, immediately S of Aberdeen, clastic sediments have yielded luminescence (OSL) age estimates of 74 ± 6 ka and 63 ± 7 ka, respectively (Merritt *et al.* 2003). This suggests, if correct, that the dated glaciodeltaic deposit at Oldmill (truncated and capped by a periglacial land surface displaying ice-wedge casts) and upwards-fining outwash sediments at Nigg Bay may represent terrestrial evidence of an MIS 4 ice sheet (Gemmell *et al.* 2007). At both sites these sediments are overlain unconformably by complex suites of tills,
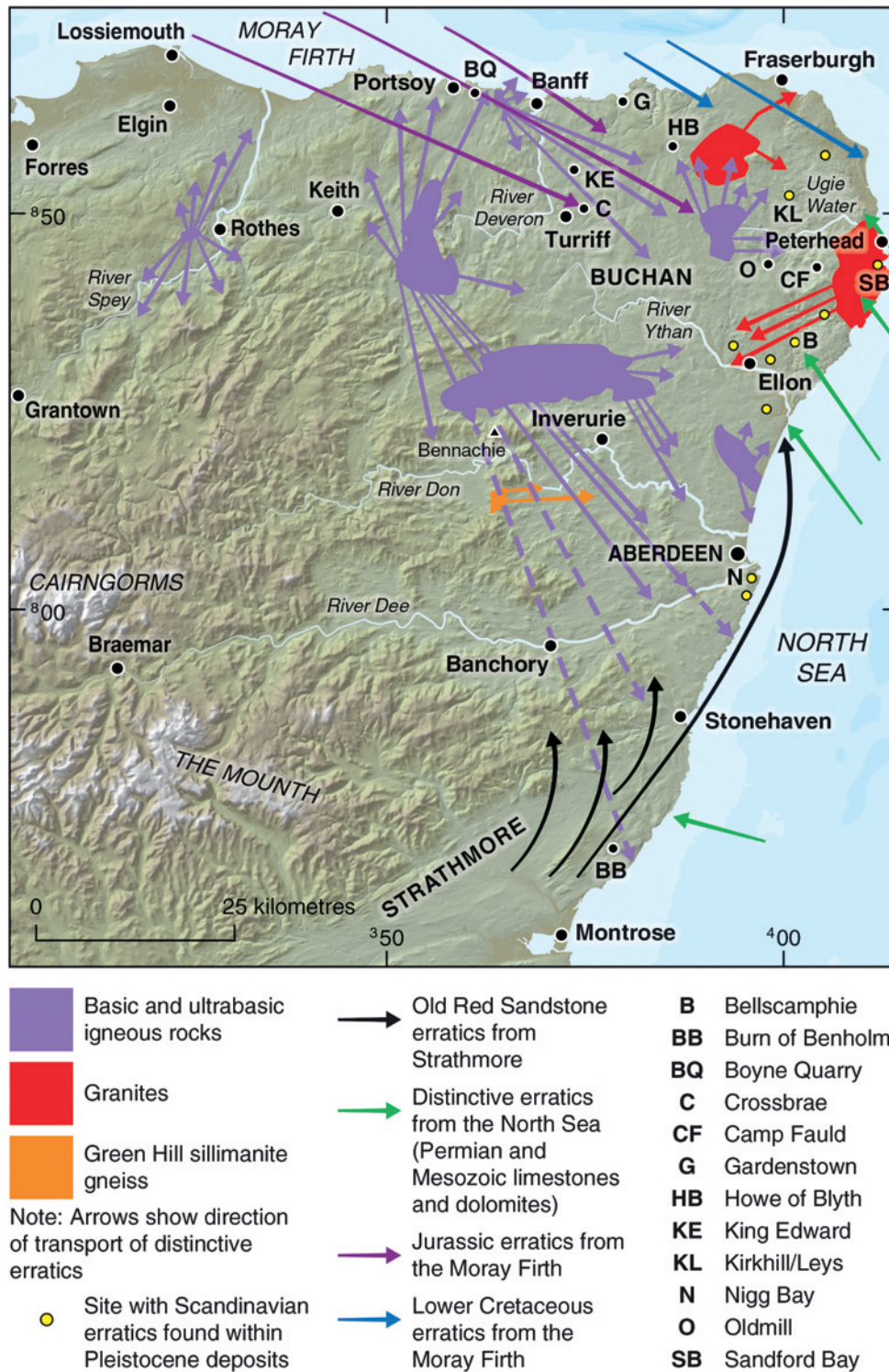
Figure 21 Transport paths of some indicator erratics in NE Scotland (after Read 1923; Bremner 1928; Synge 1956; Sissons 1967a; Merritt *et al*. 2003). Hill-shaded surface model built from Intermap Technologies NEXTMap Britain elevation data.

sands and gravels believed to date to later Devensian events (Merritt *et al.* 2017a).

There are now no known representative organic deposits of Middle Devensian interstadials in the region. The *Crossbrae Peat* at Crossbrae, near Turriff (Fig. 21) was originally thought to date from between 22 and 26.5 ^{14}C ka BP, but it is now correlated with MIS 5a or 5c (Whittington *et al.* 1998). The apparent absence of Middle Devensian deposits may be explained if the region was ice-covered during this period of time. Indeed, this scenario is suggested by evidence obtained

from the gravel pit at the Howe of Byth, near New Pitsligo (Fig. 22), at which the *Howe of Byth Gravel* has yielded luminescence ages of between 45 and 38 ka (Duller *et al.* 1995; Hall *et al.* 1995b). As this gravel unit is interpreted to have been laid down by glacial meltwater (Hall *et al.* 1995b; Hall & Connell 2000), ice had possibly reached the northern coast of Buchan at this time.

Twelve stages in the evolution of the last BIIS in NE Scotland during the Middle to Late Devensian have been recognised by Merritt *et al.* (2017a; Table 10; Fig. 23) and

Table 10 Summary of events in NE Scotland. Stages and estimated ages after Merritt *et al.* (2017a). Abbreviations: B = glacial 'binge'; C = cold-periglacial; I = Interglacial; IS = Interstadial; P = glacial 'purge'; R = glacial readjustment.

Stage	Age	Events, palaeo-environments, important sites and units	
	MIS 5e	Formation of palaeosols at Kirkhill and Leys	I
	MIS 5a–d	Cold & cool stages represented at the Kirkhill, Camp Fauld, Crossbrae and Burn of Benholm sites	C/IS
	MIS 4	Periglacial phenomena at Kirkhill and Leys. ESE flow from shelf edge parallel ice sheet with outwash at Oldmill and Nigg Bay?	C/B
1	MIS 3	No known organic deposits, but early advance from NW deposited the *Howe of Byth Gravel*?	C/IS
2	~32–24 ka LLGM	SE flow of ice from across the Moray Firth towards Aberdeen depositing *Whitehills Glacigenic Formation* at Boyne Quarry, Gardenstown, Oldmill, Ardglassie and King Edward	B
3	~24 ka	Ice flow swings E or ENE across Buchan depositing *Byth Till* at Howe of Byth, *Hythie Till* at Oldmill and Kirkhill	R
4	~23 ka	Possible northward drawdown towards the Moray Firth when an ice divide stretches from the Cairngorms towards Buchan	P
5	22–19 ka LGM	Thickening ice in Moray Firth and Strathmore coalesces in E Buchan where a small enclave is occupied by 80 m Glacial Lake Ugie. Deposition of *Banchory, Essie (part), Hatton & Mill of Forest tills*	B
6.1	22–19 ka LGM	East Grampian ice retreats slowly in periglacial climate W across Buchan leaving reticulate channel system. Enclave in E Buchan enlarges, occupied by 50 m Lake Ugie	C
6.2		Strathmore ice readvances onshore into Logie-Buchan and Glacial Lake Ugie. End moraines formed at Cross-Stone	R
6.3		Stagnation and decay of the Logie-Buchan ice lobe with deposition of *Kippet Hills S and G*	R/P
7.1	~19 ka	Strathmore and East Grampian ice unzips down the eastern coast. Widespread outwash. RSL at ~30 OD. *Tullos Clay*	P
7.2		Minor readvance N of Montrose? RSL at ~22 m OD. Deposition of glaciomarine *Lunan Bay Clay*	R
8	18–17.5 ka	Fladen 1 Readvance SE into the Witch Ground, but contemporaneous events in NE Scotland are unclear	B
9	17.5–16.2 ka	Fladen 2 Readvance SE towards Bosies Bank. Subsequent minor readvance of Moray Firth ice around Fraserburgh?	B
10	~16 ka	Moray Firth ice advances onshore at St Fergus. RSL at ~12 m OD. Deposition of glaciomarine *St Fergus Silts*	R
11	~16–15 ka	Moray Firth ice pinned along Banffshire coast damming ice-marginal lakes. Retreat of East Grampian ice interrupted by limited readvance to N depositing the 'Northern Drift' and *Blairdaff Moraine Fm*	B/C
12	~14–15	Elgin Oscillation. Ponding in the lower reaches of Strathspey	R

Table 11 The lithostratigraphy at Kirkhill and Leys quarries relating to the last glacial–interglacial cycle. Note: Due to the position of the sites in eastern Buchan, and the accommodation space available in channels/basins, this complex stratigraphy does not exceed 2.5–3.0 m in total thickness.

Unit	Formal name (McMillan *et al.* 2011a)	Old name (Hall and Jarvis 1993a)	MIS
16	Manse Gelifluctate Bed	Kirkhill Gelifluctate 5	2
15	Kirkhill Church Sand Formation	Kirkhill Upper Sands	2
14	East Leys Till Formation	East Leys Till	2
13	Hythie Till Formation	Kirkhill Upper Till	3/2
12	Corse Diamicton Formation	Corse Diamicton	3/2
11	Corsend Gelifluctate Bed	Kirkhill Gelifluctate 4	4?
10	Fernieslack Palaeosol Bed	Kirkhill Upper Buried Soil	5e

are summarised in the following sections. Stages 1–5 are established mainly from a parsimonious interpretation of the lithostratigraphical record at a few key sites, whereas stages 6–12 are linked to retreat stages based on the geomorphological record, both onshore and offshore, identified by Clark *et al.* (2012a, b) and Hughes *et al.* (2014). The evidence for each stage is presented fully as 'Supporting Information' in Merritt *et al.* (2017a); however, problems remain regarding the published geochronometry for the region, particularly the integration of ages produced by different dating methods and of adjustments and recalibrations that have been made to them.

At least three 'binge-purge' cycles (*cf.* Alley & MacAyeal 1994) occurred during the last glaciation of Scotland (Hubbard *et al.* 2009; Fig. 12) and appear to be represented in the stratigraphy of NE Scotland (Merritt *et al.* 2017a; Table 10). The 'binge' phases resulted in very limited, patchy subglacial

erosion and till deposition, but were conducive to the initiation, transport and preservation of glacial rafts, which are unusually common along the Banffshire coast and in Buchan (Peacock & Merritt 1997, 2000; Merritt & Connell 2000; Merritt *et al.* 2003). The relatively short 'purge' phases dominate the subglacial geomorphological legacy, in terms of the number of distinct flowsets that have been recognised (Hughes *et al.* 2014), and appear to be more fully represented in the stratigraphical record through deposition of till.

5.2.1. Stage 1. An extensive, early, ESE-directed flow of ice from the Great Glen and the western Highlands has been deduced to have occurred to the W of the Spey across ground rising to at least 500 m OD (Fletcher *et al.* 1996; Fig. 23a; Section 5.1). Several sites in NE Scotland preserve stratigraphical evidence of a possibly contemporaneous early ice advance from the NW across the southern shore of the Moray Firth,
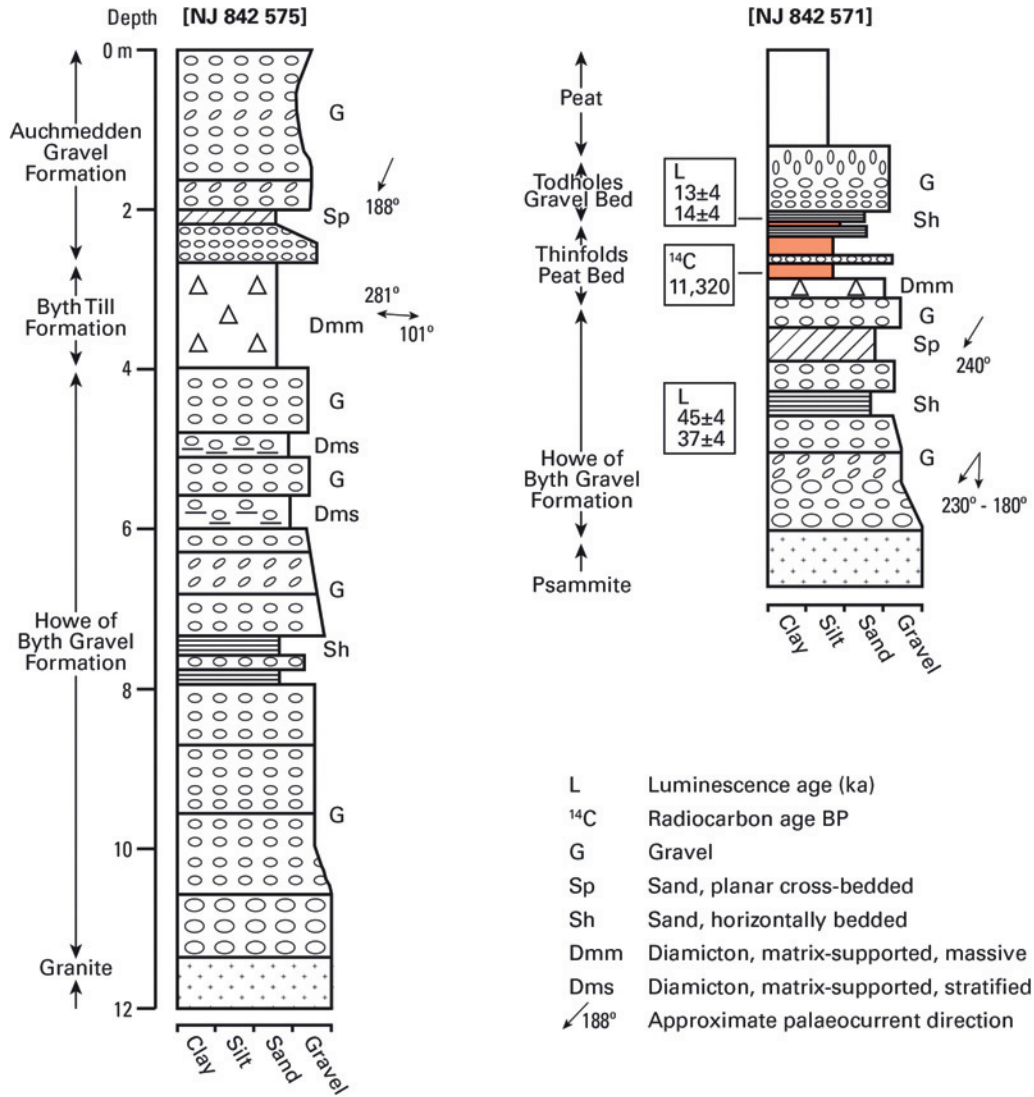
Figure 22 Summary logs of sections recorded at the Howe of Byth Quarry, SW of Fraserburgh (modified after Hall & Connell 2000).

such as at Teindland, Boyne Quarry and Gardenstown (Peacock & Merritt 1997, 2000; Merritt *et al*. 2003; Fig. 21).

5.2.2. Stage 2. The SE flow of ice from the Moray Firth laid down dark grey, shelly diamictons with a rich Jurassic to early Cretaceous palynoflora (*Whitehills Glacigenic Formation*) (Peacock 2000a), for example, at Boyne Quarry (Gordon 1993f; Peacock & Merritt 2000; Figs 24, 25). The ice penetrated inland across central and eastern Buchan towards Aberdeen on at least one occasion (Synge 1963; Fig. 21), laying down large masses of crushed, dark grey Mesozoic mudstone at Gardenstown (Castle Hill) (Sutherland 1993d; Peacock & Merritt 1997), Oldmill (Merritt & Connell 2000a), Ardglassie (Merritt & Connell 2000b) and at Kirkhill and Leys (*Corse Diamicton Member*) (Table 10). Ice was apparently sufficiently thick to create SE-orientated striae on the NE flank of the iconic mountain of Bennachie, at *ca*.335 m OD (Merritt *et al*. 2017a; Fig. 21). AAR ratios (Miller *et al*. 1987) on derived shell fragments in glacigenic sediments, mostly from the King Edward site (Hall & Peacock 2000; Fig. 21), indicate that the BIIS last advanced across Buchan in the latter part of the Middle Devensian (Peacock & Merritt 2000). A substantial shelf-edge parallel ice sheet is speculated to have become established during this stage (Merritt *et al*. 2017a; Fig. 23b).

5.2.3. Stage 3. Till provenance and the direct superimposition of tills suggest that ice flow swung around to the E or ENE along the coast of the Moray Firth, laying down, for example, the *Tofthead Till* at Teindland (Table 9; Hall *et al*. 1995a, 2018), the *Old Hythie Till Formation* at Boyne Quarry (Peacock & Merritt 2000; Fig. 24), the *Byth Till Formation* at the Howe of Byth (Hall *et al*. 1995b; Fig. 22) and the *Hythie Till Formation* at Oldmill and Kirkhill (Merritt *et al*. 2003; Table 10; Fig. 23c). A major glacial reorganisation appears to have occurred associated with thinning of the ice sheet, but apparently not resulting in widespread glacial retreat across the region.

5.2.4. Stage 4. Sparse, poorly developed, streamlined subglacial landforms in western Buchan (Fig. 26, flow-set 33) indicate possible NNW drawdown of ice towards the Moray Firth basin from an ice divide stretching from the Cairngorms towards Buchan (Hughes *et al*. 2014), but no contemporary till unit has been reported (Merritt *et al*. 2017a). This scenario is supported by crossing striae recorded at two localities SSE of Banff (Fig. 21), where eastward-directed striae are cut by N or NNE-directed ones (Read 1923; Merritt *et al*. 2003), but there is little evidence in eastern Buchan for northward flow, possibly because that area was protected beneath cold-based ice.

5.2.5. Stage 5. Ice flowing through the Moray Firth basin and Strathmore appears to have subsequently thickened, coalescing at the 'Angle of Buchan', between Fraserburgh and Peterhead, as reconstructed by Clapperton & Sugden
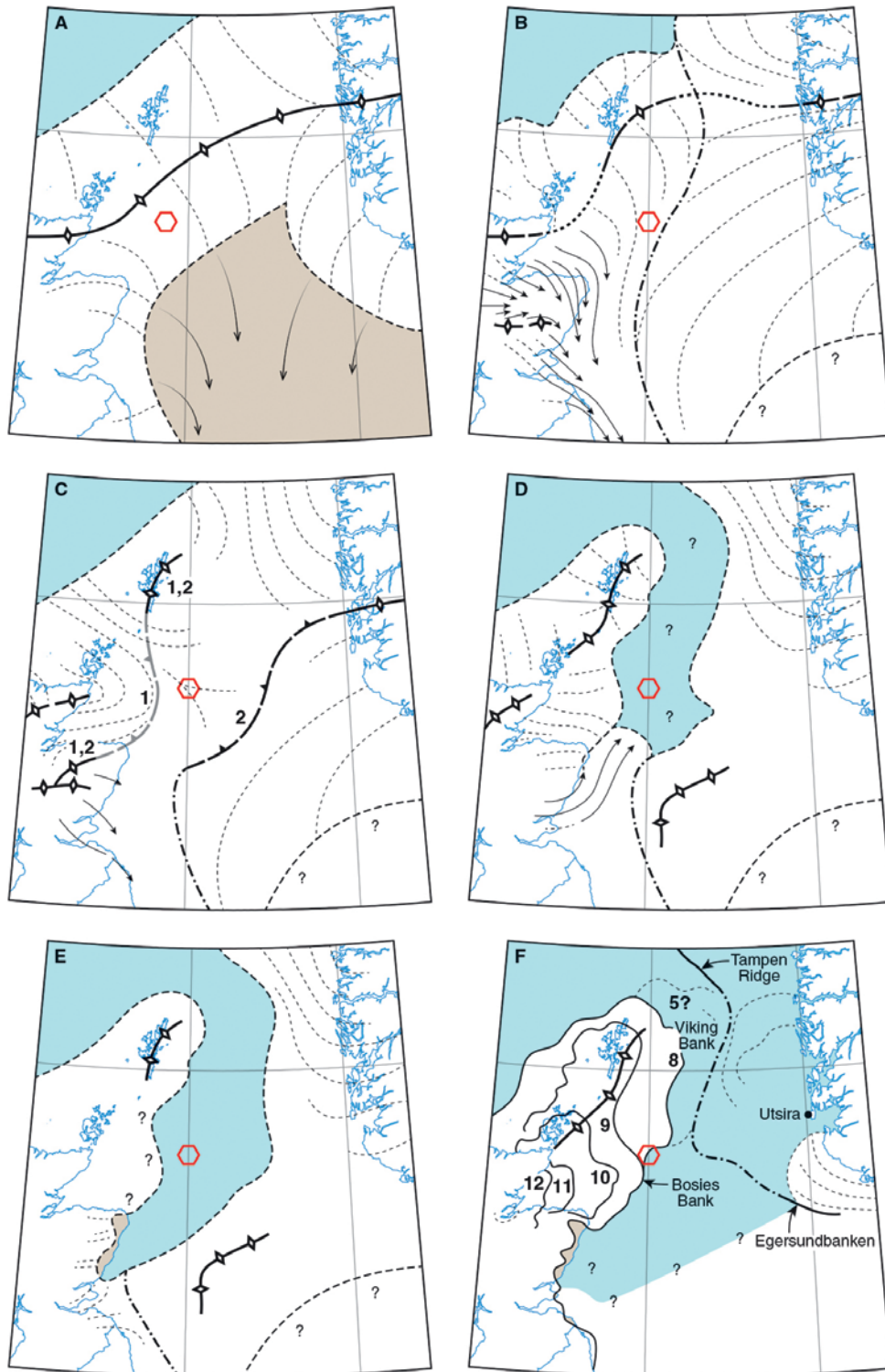
Figure 23 Conceptual glacial reconstructions depicting stages in the evolution of the NE quadrant of the last BIIS (from Merritt *et al*. 2017a with minor modification). (Reprinted from *Journal of Quaternary Science*, **32**, 276–94, Merritt, J. W., Connell, E. R. & Hall, A. M., Middle to Late Devensian glaciation of north-east Scotland: implications for the north-eastern quadrant of the last British–Irish ice sheet. Copyright (2017), with permission from Wiley.) (A) Reconstruction similar to that of Boulton *et al*. (1985, fig. 10) (stage 1). (B) LLGM (stage 2). (C) Considerable thinning and glacial reorganisation during stages 3 and 4. (D) Re-expansion following possible marine ingression into the Witch Ground basin during stage 5, followed by probable coalescence of BIIS and the NCIS at the Tampen Ridge (see reconstruction F for limits). (E) Marine ingression to Lunan Bay during stage 7. (F) Stages 8–12 and other limits after Clark *et al*. (Clark *et al*. 2012a, figs 12, 13). Red hexagons show the centre of the Witch Ground basin. Ice divides shown by thick black lines with open diamonds; single filled ticks denote ice divide receding by ice stream headward scavenging.

(1977) (Fig. 23d). As the inland boundary of the *Hatton Till Formation* (Fig. 3b), laid down by Strathmore ice, can be traced from the Ythan valley NE across ground above 80 m OD (grey dashed line in Fig. 26a), there was probably ice occupying the central North Sea basin, preventing ice from flowing directly offshore from Strathmore. The absence of evidence suggests that Central Buchan was mostly protected from glacial erosion beneath cold-based ice at this time. A small ice-free enclave appears to have opened up soon after in eastern Buchan, at the confluence of Strathmore, East
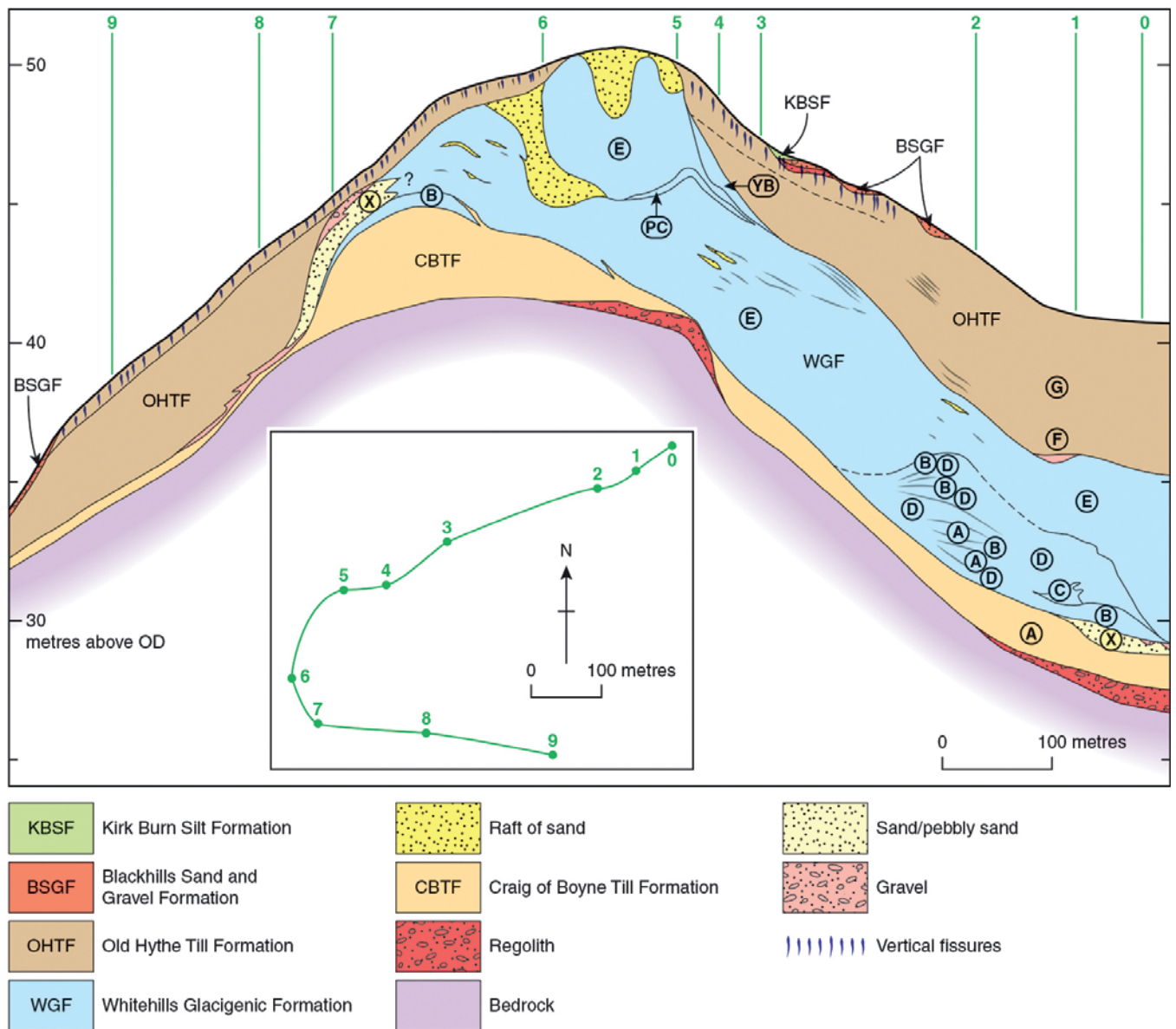
Figure 24 Generalised stratigraphy recorded at Boyne Quarry in 2014, E of Portsoy. Encircled letters refer to units described by Peacock & Merritt (2000).

Grampian and Moray Firth-sourced ice, where deltaic deposits formed in *Glacial Lake Ugie* at its 80 m OD level (Merritt *et al.* 2003, 2017a; Fig. 26a).

5.2.6. Stage 6. East Grampian ice slowly retreated W across Buchan (Fig. 26a), leaving a reticulate network of glacial drainage channels of a type associated with cold-based ice margins (*cf.* Benn & Evans 2010, fig. 8.38) and laying down the *Banchory Till Formation* (Fig. 3b). A local glacial reorganisation followed that involved a further advance of Strathmore ice onshore between Aberdeen and Peterhead, up the valley of the River Ythan to the vicinity of Ellon, where laminated silts and clays occur that formed in a proglacial, ice-dammed lake (Merritt *et al.* 2003) and formed arcuate end-moraines at the Cross-Stone site nearby (Gemmell *et al.* 2007; Fig. 26a). The ice also advanced into the lower Ugie Valley during this event, named as the 'Logie-Buchan Readvance' by Merritt *et al.* (2017a). It dammed Glacial Lake Ugie at its 50 m OD level (Fig. 26b), in which thick sequences of laminated muds were deposited (*Ugie Clay Formation*) (Peacock & Connell 2000). Glaciofluvial deposits

associated with both the former lakes experienced intense periglacial conditions during subsequent stages, testified by widespread preservation of tundra polygons and other cryogenic phenomena in the Ugie Valley (Gemmell & Ralston 1984), but attempts at OSL dating the glaciofluvial sediments proved inconclusive (Gemmell *et al.* 2007).

The ice lobe occupying Logie-Buchan subsequently decayed to form fresh, hummocky topography typical of ice stagnation (Jamieson 1882). The associated deposits of the Logie-Buchan Glacigenic Subgroup are generally undivided as they comprise a complex interbedded sequence of clayey diamict, sand, silt, clay and gravel, typically vivid reddish brown in colour. Sands and gravels forming eskers and kame-terraces are assigned to the *Kippet Hills Sand and Gravel Formation*, which contains a distinctive suite of glacial erratics, in particular Permian limestone, and shells derived from the floor of the North Sea basin immediately to the E (Gordon 1993g; Hall & Jarvis 1995; Merritt *et al.* 2003; Fig. 26). Although East Grampian ice had retreated from at least the lower reaches of the Ythan and Ugie valleys during the Logie-Buchan Readvance
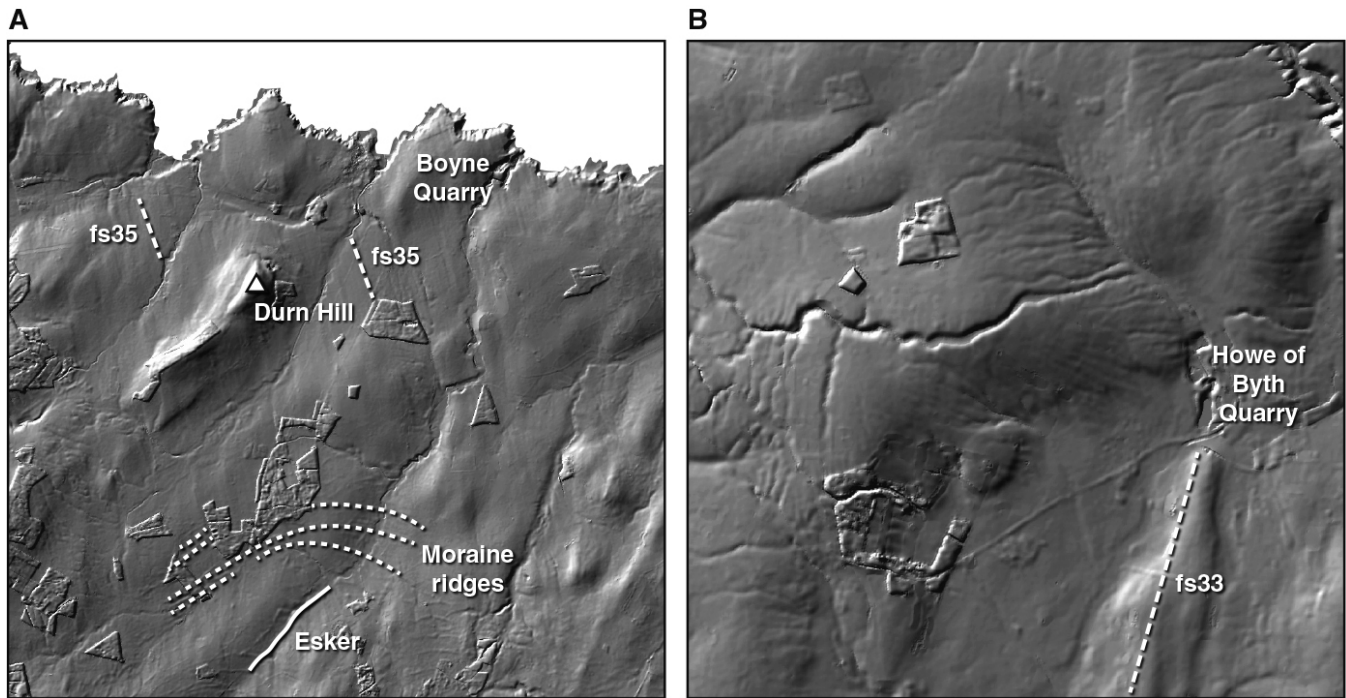
A

B

Figure 25 Subglacially streamlined features and ice marginal glacial drainage channels around Boyne Quarry (A) and the Howe of Byth (B) sites (modified after Merritt *et al*. 2017a). Flowsets (fs) after Hughes *et al*. 2014. Hill-shaded surface model built from Intermap Technologies NEXTMap Britain elevation data.

(Fig. 26b), the two ice masses abutted in the vicinity of Aberdeen where there is both stratigraphical and geomorphological evidence at Nigg Bay (Fig. 21) for limited eastward readvance ('Aberdeen Readvance') (Merritt *et al*. 2003), but no confining dates.

5.2.7. Stage 7. Strathmore and East Grampian ice continued to 'unzip' down the E coast, associated with widespread glaciofluvial and deltaic deposition around Aberdeen (*Lochton Sand and Gravel Formation* and *Glen Dye Silts Formation*) (Simpson 1948, 1955; Merritt *et al*. 2003; Fig. 26b). A marine incursion led to the deposition of red clays of the *Tullos Clay Member* (*Errol Formation*) when RSL stood at about 30 m OD (Peacock 1999). The 'unzipping' continued down the coast to Montrose, where red marine clay of the *Lunan Bay Clay Member* was laid down at Lunan Bay (Fig. 23e). McCabe *et al*. (2007a) conclude that the clay accumulated in open waters following a minor readvance at 20.1–20.9 cal ^{14}C ka BP, when RSL stood at about 22 m OD. However, the basis of these authors' investigation has been questioned by Peacock *et al*. (2007) and Sejrup *et al*. (2009), who comment on how the dates are incompatible with the youngest ages of *ca*.17.5 ka for deglaciation of ground above 200 m on Mormond Hill (Fig. 26b) and Pitfichie, SSW of Inverurie (Fig. 21), provided by cosmogenic techniques (Phillips *et al*. 2008). Furthermore, in a review of RSL changes, glacial isostatic modelling and ice sheet reconstructions from the British Isles since the LGM, Shennan *et al*. (2006) conclude that the area only became ice-free after *ca*.17 ka (Smith *et al*. 2018).

5.2.8. Stage 8. This stage is correlated with the *Fladen 1 Readvance* (17.5 cal ^{14}C ka BP), during which ice advanced SE into the Witch Ground basin of the North Sea (Sejrup *et al*. 2015), but the contemporaneous position of ice in NE Scotland is unclear (Clark *et al*. 2012a; Sejrup *et al*. 2015). The Scottish and Scandinavian ice sheets had separated by this time as the Viking Bank (E of Shetland) had become ice-free by *ca*.18.3 ka (Peacock 1995; Hughes *et al*. 2016; Stewart *et al*. 2019; Fig. 23f).

5.2.9. Stage 9. The mapped extent and provenance of the *Essie Till Formation* (Fig. 3b) combined with evidence from striae around Peterhead and Fraserburgh suggests that a limited readvance of Moray Firth ice occurred prior to its subsequent retreat across the NE tip of Buchan (Peacock 2000b). By this stage, a huge ice lobe would have occupied most of the Moray Firth basin (Clark *et al*. 2012a). It was fed from the NW through to the SE, delimited by sea-bed features that can be traced from Peterhead to the N tip of Shetland via the Bosies Bank, in the outer Moray Firth (Clark *et al*. 2012a; Fig. 23f). The Bosies Bank has been interpreted as a push moraine (Graham *et al*. 2010, 2011) now judged to have formed during the *Fladen 2 Readvance* at 16.2 cal ^{14}C ka BP (Sejrup *et al*. 2015).

5.2.10. Stage 10. Evidence from the raised glaciomarine deposits of the *St Fergus Silt Formation* suggests that the southern margin of the Moray Firth ice lobe terminated in the sea in the vicinity of St Fergus, at a RSL of *ca*.12 m OD, before a minor onshore readvance created the *St Fergus Push Moraine* (Hall & Jarvis 1989; Merritt *et al*. 2003; Fig. 26c). This feature includes faulted, tilted and deformed masses of bedded marine silt and sand formed by ice push from the E or NE. The absence of contemporary late-glacial marine deposits south of Peterhead (Shennan *et al*. 2006; Smith *et al*. 2018) suggests that the sea had retreated from the area owing to rapid glacio-isostatic rebound (Merritt *et al*. 2003). Radiocarbon dates on marine bivalves in the deformed silts suggest that the readvance occurred sometime after 17.5 cal ^{14}C ka BP (Hall & Jarvis 1989; Peacock 1999). The push moraine has been linked tentatively with sea-bed features off the northern tip of Orkney (Clark *et al*. 2012a; Fig. 23f) and it possibly correlates with a minor readvance that deposited the Wick Till in Caithness (Hall & Riding 2016; Section 3.2).

5.2.11. Stage 11. During subsequent retreat, the southern margin of the Moray Firth ice lobe was pinned along the Banffshire coast where the lacustrine *Kirk Burn Silt Formation* accumulated in proglacial lakes, and swathes of glaciofluvial
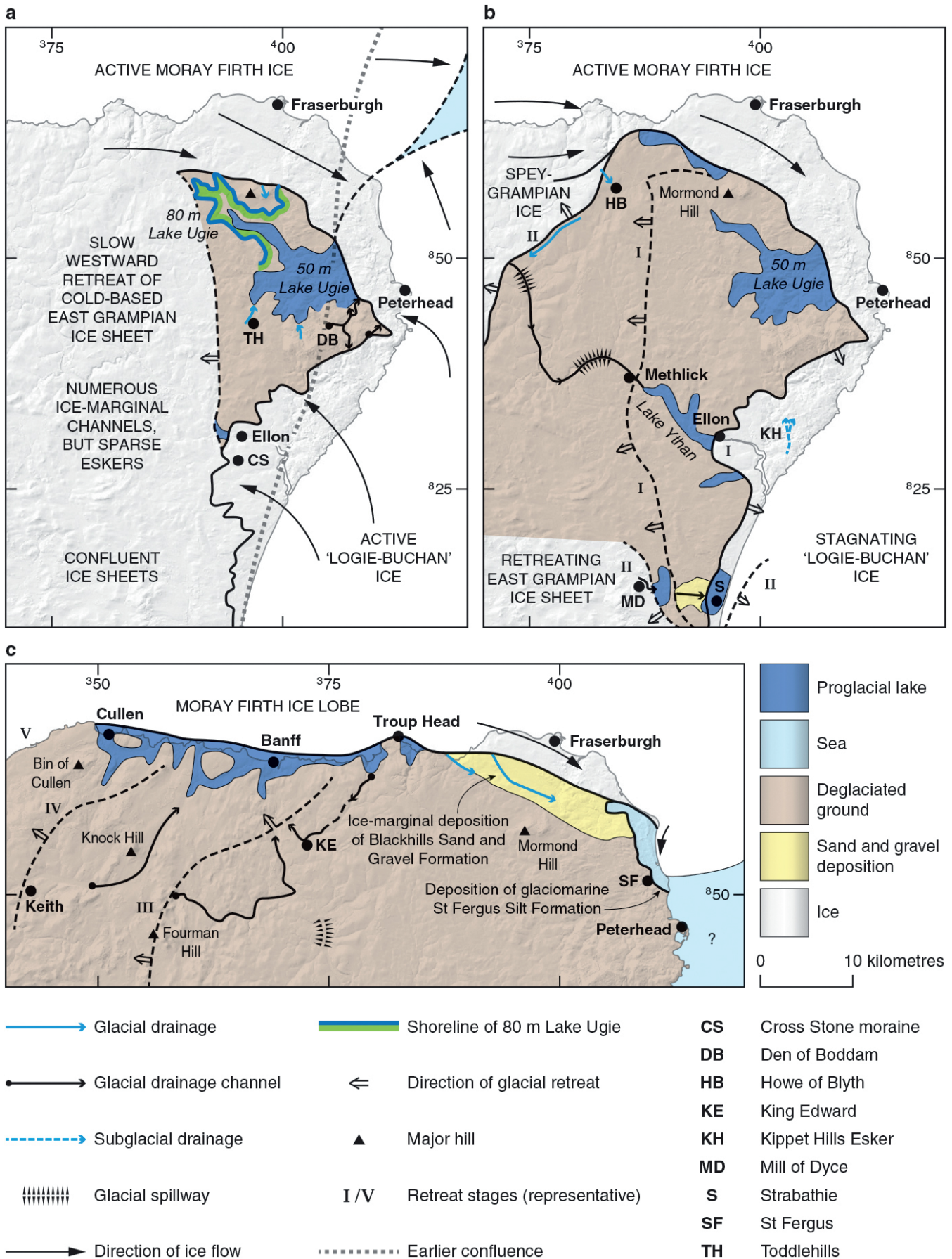
Figure 26 Reconstruction of ice-free enclaves and proglacial lakes in Buchan (modified after Merritt *et al.* 2017a). (a) Confluence between ice lobes during stage 5, creation of 80 m Glacial Lake Ugie (GLU) followed by the 'Logie-Buchan Readvance' and creation of 50 m GLU during stage 6. (b) Decay of the Logie-Buchan ice lobe, further retreat of East Grampian ice followed by readvance of Moray Firth ice during stages 7–9. (c) St. Fergus Readvance and subsequent ponding during glacial retreat along the Banffshire coast during stages 10–11.

outwash (*Blackhills Sand and Gravel Formation*) were laid down in association with ice-marginal drainage channels (Merritt *et al*. 2003; Fig. 26c). Evidence of minor readvances of the Moray Firth ice lobe against the coast have been reported, notably to the W of Banff (Merritt *et al*. 2003). A limit based on sea-bed ridges that swing around from the vicinity of Banff to the NE tip of Caithness has been identified by Clark *et al*. (2012a, b) (Fig. 23f). Merritt *et al*. (2017a) tentatively link this stage with deposition of the so-called 'Northern Drift', which resulted from a northward advance of ice towards the Moray Firth basin (Read 1923; Bremner 1934, 1943). It laid down little till – for example, at Boyne Quarry (Fig. 24) – but left eskers, retreat moraines and numerous blocks of locally derived Huntly/Knock Gabbro (Read 1923; Peacock & Merritt 2000; Fig. 25a).

A significant change in the character of the landscape occurs inland at retreat limits formed by moraine ridges S of Bennachie (*Blairdaff Moraine Formation*) and around Durn Hill (Merritt *et al*. 2003, 2017a; Fig. 25a). For example, periglacial phenomena, specifically ice-wedge pseudomorphs and sheets of gelifractate and gelifluctate are widespread to the NE, across central Buchan (Synge 1956; Romans *et al*. 1966; Connell & Hall 1987), where a reticulate network of ice-marginal drainage channels formed at the retreating ice sheet margin (Clapperton & Sugden 1977). The channels are typically incised into bedrock with little or no associated till (Merritt *et al*. 2003), they are rarely associated with eskers or other ice-contact glaciofluvial deposits (Merritt *et al*. 2003) and they have a style that is typical of those formed at cold-based ice margins (*cf*. Benn & Evans 2010, fig. 8.38). In contrast, relatively fresh-looking moraine ridges and glaciofluvial deposits relatively devoid of periglacial phenomena are more common to the SW, at least on lower ground (Sutherland & Gordon 1993d).

It is suggested that the fresher topography formed after a significant regional re-expansion of ice associated with the 'Northern Drift', following prolonged very cold conditions, but the event is currently undated. At the Howe of Byth site (Hall *et al*. 1995b; Fig. 22) the *Auchmedden Gravel Formation* forms an upper fan that was fed from water flowing eastwards via interconnected glacial drainage channels typical of those associated with westward retreating ice across central Buchan (Merritt & Peacock 2000; Fig. 25b). Gemmell *et al*. (2007) obtained an OSL luminescence age estimate of *ca*.61 ka from the gravel unit, but it is older than the luminescence ages of *ca*.45 and 38 ka obtained from the Howe of Byth Gravel below (Duller *et al*. 1995).

5.2.12. Stage 12. Peacock *et al*. (1968) present geomorphological evidence suggesting that ice re-occupied the lower reaches of Strathspey during a significant, but poorly constrained, readvance that they named as the *Elgin Oscillation* (Fig. 23f). The *Waterworks Till* was probably laid down during this event at Teindland (Table 9; Hall *et al*. 1995a). Ice at the coast ponded lakes (*Glacial Lake Rothes*) in an area previously vacated by East Grampian ice (Bremner 1934), such as at Tofthead, upstream of Fochabers (Connell 2000). The direction of large-scale foreset beds in deltaic deposits preserved in Glen Rothes, S of Elgin, and Glen Rinnes, SW of Dufftown, indicate that meltwater entered the lakes from the N and S, respectively (Merritt *et al*. 2003). Sets of N–S orientated linear transverse ridges around Elgin (Peacock *et al*. 1968) subsequently formed at the westward retreating margin of Moray Firth ice (Merritt *et al*. 2017d). As similar features have not been recorded to the E of the Spey, their presence within the glacial limits of the Elgin Oscillation suggests that glaciological conditions may have changed following that event (still not

firmly dated), associated with the initiation of the Moray Firth Ice Stream (Hughes *et al*. 2014).

5.3. Southern Argyll and Arran

Southern Argyll, including Kintyre, Islay and Arran (Fig. 27), contains evidence of glaciation that provides an important link between the terrestrial geomorphological and lithostratigraphical records of SW Scotland and NE Ireland (Finlayson *et al*. 2014; Table 12). It also links the sedimentary records of the fjord-like coastal terrain of the western Scottish Highlands with the Malin Shelf beneath the Atlantic Ocean to the W (Fig. 27). This is another region that is notable for the presence of shell-bearing clays underlying till, particularly at sites in and around Tangy Glen, on the western coast of Kintyre (Fig. 27a), at elevations of between 40 and 60 m OD (Sutherland 1993f). Deposits containing both cold and temperate marine molluscs, ostracods and foraminiferids also occur on the southern tip of Arran, at up to 55 m OD (Sutherland 1981). At Afton Lodge (Merritt *et al*. 2014), in Ayrshire, the shelly deposits have either been interpreted to be *in situ* (Sutherland 1981) or attributed to glacial rafting (Synge & Stephens 1966) (Section 5.4.2). In either case, erratic dispersal patterns and glacial striae suggest that the localities on Kintyre were subsequently overridden by ice flowing W towards the Malin Shelf (Finlayson *et al*. 2014). However, ice-marginal features on the W coast of Kintyre, together with striae on Arran, suggest that the final flow was towards the S through the Kilbrannan Sound and the Outer Firth of Clyde (Fig. 27g), consistent with the occurrence of red tills derived from Permo-Triassic rocks along the W coast of Kintyre (Synge & Stephens 1966). Two suites of recessional moraines have been recognised in the mountains of Arran (Gemmell 1973), an inner suite attributable to the Loch Lomond (Younger Dryas) Stadial, and more subdued outer limits formed earlier during deglaciation (Ballantyne 2007). The relationship between the outer moraines and raised deltaic deposits in Glen Dougarie, western Arran, indicate that they formed after RSL had fallen from 32 m OD (Finlayson *et al*. 2014).

A raised rock platform attributed to an interglacial period pre-dating the last glacial cycle passes beneath till in the vicinity of Tangy Glen, one of several such occurrences along the W coast of Scotland (Gray 1993; Smith *et al*. 2018). Finlayson *et al*. (2014, fig. 5b) note that at Drumore Burn, S of Glenacardoch Point (Fig. 27g), 15 m of till overlay up to 8 m of cobble-gravel with weakly developed herring-bone cross stratification. Tentatively interpreted as beach gravel, the unnamed unit overlies a rock platform at *ca*.18 m OD, only a few metres higher than the one described by Gray (1993) at Glenacardoch Point.

In western Islay, up to 60 m of pink clays, silts and diamicton rest on raised shore platforms (Benn & Dawson 1987). Luminescence dates of 55–40 ka for these supposed glaciomarine sediments suggest an Early Devensian age (Dawson *et al*. 1997), but the dates may be in error due to inadequate exposure to light in turbid water (Peacock 2008). Furthermore, the terraced surface of the sequence at up to 80–90 m above OD is far above the Late Devensian marine limit in western Islay at 23 m OD (Dawson 1982). Considering that global sea level was possibly as low as -75 m and never significantly higher than -50 m in the Early Devensian (Shackleton 1987), considerable glacio-isostatic depression would be required immediately following the retreat of a thick Early Devensian ice sheet for the sequence to have been deposited *in situ* (Sutherland 1981). Whilst a similar explanation has been proposed to account for the occurrence of other 'high level' cold-water
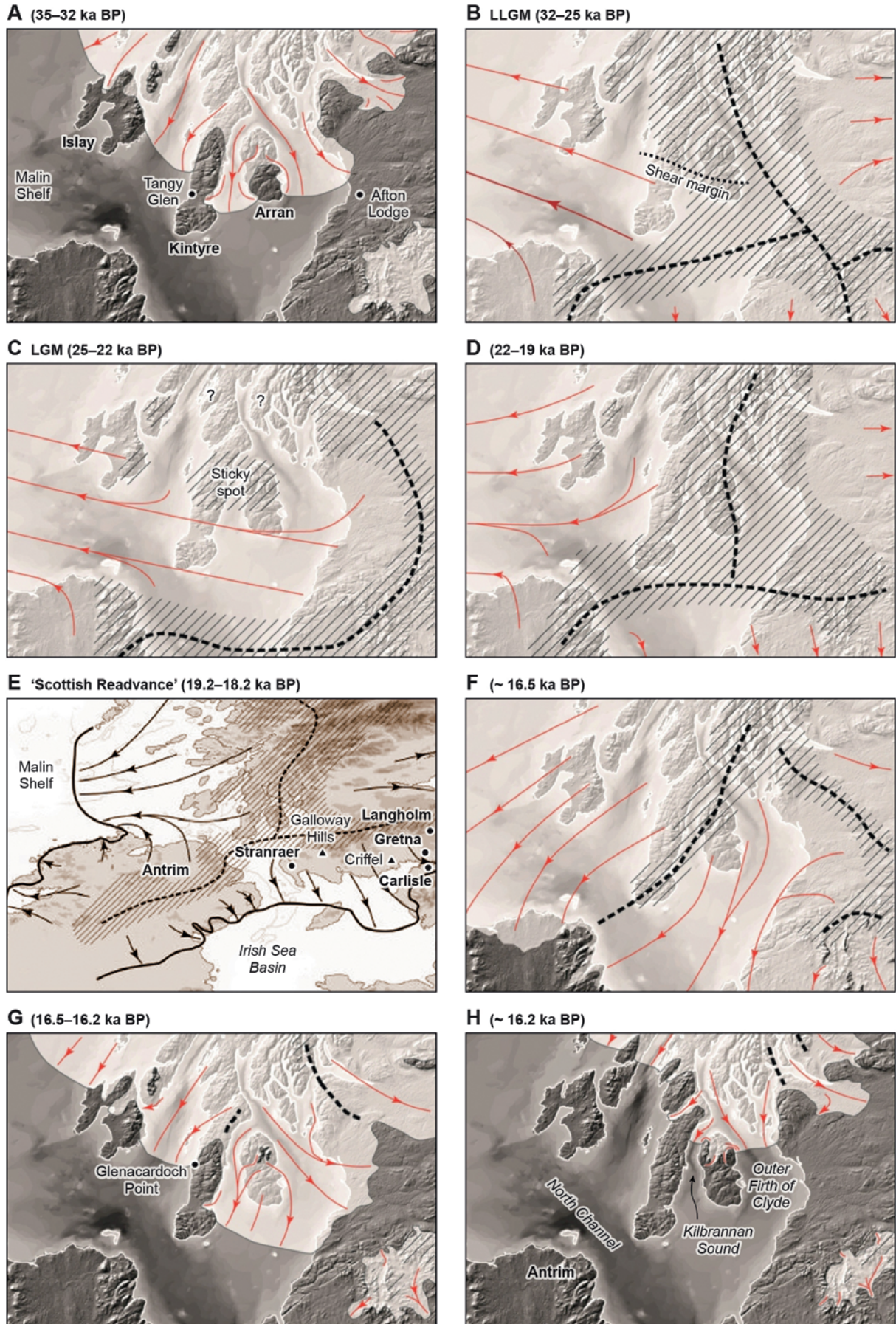
Figure 27 Eight stages (A–H) of the growth and decay of the last BIIS over western Scotland, the North Channel and NE Ireland (modified after Finlayson *et al.* 2014). Diagonal shading denotes probable cold-based ice. Dashed lines denote ice divides. Age estimates modified in respect of the recently published age of the Scottish Readvance determined by Chiverrell *et al.* (2018). Hill-shaded surface model built from Intermap Technologies NEXTMap Britain elevation data.

Table 12 Summary of events in Argyll. Stages and ages modified after Finlayson *et al*. (2014) and Chiverrell *et al*. (2018). Abbreviations: B = glacial 'binge'; I = Interglacial; P = glacial 'purge'.

Stage	Age	Events, palaeo-environments and important units	
	MIS 5e	Raised rock platforms and overlying beach gravels probably formed in western Kintyre (now till covered)	I
	MIS 5a–d	No evidence reported	
	MIS 4	Early Devensian glaciation likely, but no indisputable evidence. Raised glaciomarine deposits claimed to occur in W Islay	B
	MIS 3	Cold-water marine sediment laid down off Kintyre later to be rafted	
1	35–32 ka?	Ice advance against reverse slopes along western Kintyre with emplacement of glacial rafts of shelly marine sediment at Tangy Glen	B
2	32–25 ka? LLGM	Ice sheets over western Highlands and Southern Uplands merge with N–S divide across the Outer Firth of Clyde from which ice flows WNW towards Malin Shelf. Scottish ice advances into Belfast Lough in Northern Ireland before an ice divide forms across over the North Channel linking Scottish and Irish ice sheets	B
3	25–22 ka?	Enhanced drawdown towards shelf edge causes ice divide over the Outer Firth of Clyde to migrate some 60 km to E. Cold-based ice protects mountains of Arran. The ice divide sits across the Clyde basin during the LGM	P/B
4	22–19 ka? LGM	Main N–S ice divide migrates W towards Arran following substantial thickening. E–W divide links Southern Uplands with Irish ice sheet over Antrim. Cold-based ice protects mountains of Arran	B
SR	19.2–18.2 ka	Scottish Readvance, as established across the northern Irish Sea basin and Solway	B
5	~16.5 ka	Eastern half of E–W divide breaches and ice flows SSW through Outer Firth of Clyde. Eventual collapse of western half of the divide allows Scottish ice to advance SW into Northern Ireland	P
6	~16.5–16.2 ka	Ice becomes topographically constrained. Minor readvances of locally sourced ice on Arran at *ca*.16.2 ka BP followed its separation from outlet glaciers occupying the Kilbrannan Sound and Firth of Clyde	P
7	~16.2 ka	Deglaciation of the Outer Firth of Clyde	P

marine/glaciomarine sediments around Scotland (Sutherland 1981), all other occurrences are now considered more likely to have been emplaced by glacial rafting during the last glaciation (Merritt 1992; Peacock 2008; Finlayson *et al*. 2014; Merritt *et al*. 2014).

Although no formal lithostratigraphy has been established, the glacial history of the region has become much clearer following recent analysis of NEXTMap imagery combined with cosmogenic dating and fieldwork. For example, Finlayson *et al*. (2014) have elucidated seven stages (summarised below) combining their findings with ice sheet reconstructions made for W central Scotland (Finlayson *et al*. 2010), SW Scotland (Salt & Evans 2004), northern England (Livingstone *et al*. 2012), NE Ireland (Greenwood & Clark 2009; McCabe & Williams 2012) and the Malin Shelf (Dunlop *et al*. 2010) (Fig. 27; Table 12).

Evidence from erratics reveals that ice advance was initially directed through N–S aligned, geologically controlled basins that have been carved through successive glacial cycles. Ice entered these basins in *stage 1* from a restricted, marine-proximal mountain ice sheet similar to those that probably existed during many cold stages of the middle and late Pleistocene during 'average' conditions (Porter 1989; Fig. 27a). The ice front advanced against reverse slopes along the western coast of Kintyre, leading to the local preservation in hollows of glacial rafts of shelly sediment. The Western Highlands ice sheet subsequently joined with a smaller ice cap centred over the Southern Uplands during *stage 2*, which witnessed a major glacier expansion. An E–W ice divide developed over northern Kintyre, Arran and the Firth of Clyde, protecting the underlying landscape from glacial erosion. The ice sheet became sufficiently thick for ice to flow directly westwards to the Malin Shelf, and was no longer topographically constrained, creating WNW-directed bedforms (streamlined hills, crag-and-tail forms and drumlins) across southern Kintyre, maintaining the same alignment at all elevations. Following a possible early advance of Scottish ice into Belfast Lough, in Northern Ireland (Green-

wood & Clark 2009), an E–W ice ridge developed over the North Channel, bridging the British and Irish ice centres (Fig. 27b).

The N–S ice divide centred over the Firth of Clyde migrated some 60 km to the E during *stage 3*, when enhanced drawdown towards the western marine margins of the BIIS created WNW-orientated streamlined subglacial bedforms across till along the western seaboard of Kintyre (Fig. 27c). Finlayson *et al*. (2014) infer that a shear margin within the ice sheet across southern Arran and central Kintyre separated warm-based ice, which flowed towards the Malin Shelf, from cold-based ice within a 'sticky spot' that protected the delicate mountaintop tors of northern Arran (*cf*. Benn & Evans 2010, p. 472) (Fig. 27b). Ice flowing over southern Arran and Kintyre merged with powerful NW flowing ice to overwhelm Islay (Cousins 2012). Finlayson *et al*. (2014) conclude that the regional N–S ice divide then migrated westwards again, following a substantial, renewed build-up of ice during *stage 4*. It became pinned over Arran and the eastern seaboard of Kintyre, causing minimal subglacial erosion there (Fig. 27d). An ice divide over the North Channel linked Antrim and the Southern Uplands at this time, which is possibly contemporaneous with the Gosforth/Blackhall Wood Oscillation (Scottish Readvance) into the northern Irish Sea basin (Merritt & Auton 2000; Livingstone *et al*. 2012; Chiverrell *et al*. 2018) and the Clogher Head Readvance of Irish ice in NE Ireland (McCabe *et al*. 2007b; Clark *et al*. 2012b; Merritt *et al*. 2018) (Fig. 27e; Section 5.5).

During *stage 5*, ice flowing SW towards the North Channel subsequently formed streamlined, topographically unconstrained subglacial bedforms with this general alignment across thick till mantling western Kintyre (Finlayson *et al*. 2014; Fig. 27f). Similar SSW-orientated bedforms in southern Arran indicate probable contemporaneous funnelling of ice out of the Outer Firth of Clyde, which must have followed breaching of the eastern half of the North Channel ice divide that previously linked Antrim and the Southern Uplands (Fig. 27e). The absence of streamlined landforms on the high ground of

Clyde/Glasgow		Marine Isotope Stage	Greenland	Ayrshire
Gartocharn Till Fm	Linwood Clay Mbr	2(LLS)	GS-1	Clyde Clay Fm
Clyde Clay Fm		2(LI)	GI-1e to GI-1	
Ross Sand Mbr / Broomhouse Sand and Gravel Fm	Paisley Clay Mbr			
Bellshill Clay Mbr				Auchenwinsey Till Mbr
Wilderness Till Fm		2	GS-5 to GS-2a	Eglington Shelly Till Mbr
Broomhill Clay Fm				Sourlie Organic Silt Fm
Cadder Sand Fm (contains bones from woolly rhinoceros)		3		Armsheugh Sand and Gravel Fm
Ballieston Till Fm		4? or earlier		Lawthorne Diamicton Mbr
				Littlestone Till Fm

Figure 28 Simplified glacial lithostratigraphy for the Clyde and Ayrshire basins (after McMillan *et al.* 2011a). Greenland Ice Core GICC05 Chronology events after Lowe *et al.* (2008b). Abbreviations: Fm = formation; Mbr = member; LLS = Loch Lomond Stadial; LI = Lateglacial Interstadial; GS = Greenland Stadial; GI = Greenland Interstadial.
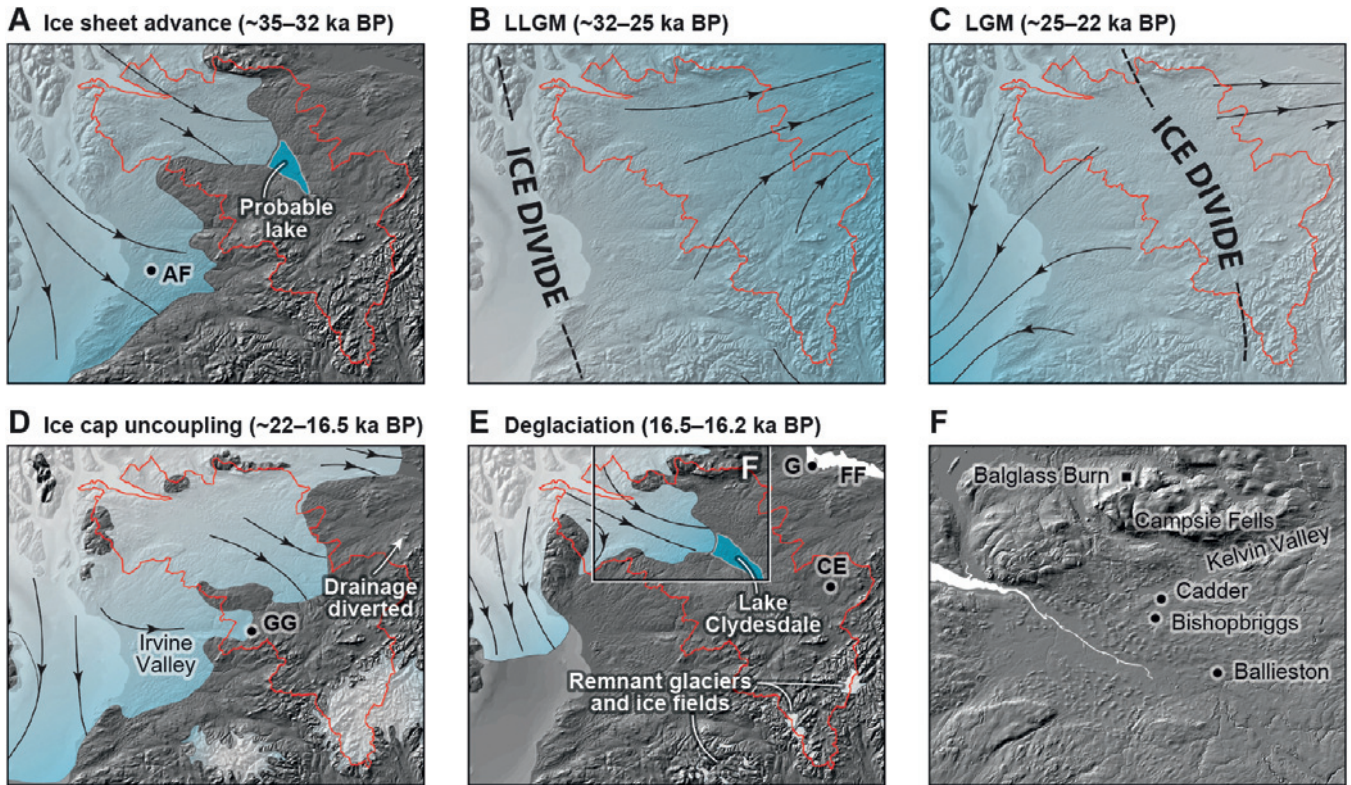

Figure 29 Reconstructed stages (A–F) of glaciation in the West Central Lowlands (modified from Finlayson 2012). Clyde basin outlined in red. Hill-shaded surface model built from Intermap Technologies NEXTMap Britain elevation data. Abbreviations: AF = Afton Lodge; CE = Carstairs Eskers; G = Grangemouth; GG = Glengavel; FF = Firth of Forth.

Table 13 Summary of events in western central Scotland and Ayrshire. Stages and estimated ages modified after Finlayson *et al.* (2010) and Finlayson (2012). Abbreviations: B = glacial 'binge'; C = cold-periglacial; I = Interglacial; IS = Interstadial; P = glacial 'purge'; R = glacial readjustment.

Stage	Age	Events, palaeo-environments, important sites and units	
	MIS 5e	Raised rock platforms form around the Firth of Clyde (till covered)	I
	MIS5a–d	No reported evidence	
	MIS 4	Possible ice flow from NW depositing *Ballieston* and *Littlestone tills* during an Early Devensian glaciation	B
	MIS 3	Deposition of the *Afton Lodge Clay* in the Firth of Clyde basin. Biogenic sediment deposited at Sourlie and Balglass. Bones deposited in the *Cadder Sand* around Bishopbriggs	C/IS
A	~35–32	Ice advances from NW against reverse slopes forming ribbed moraine. Laminated *Broomhill Clay* laid down in proglacial lakes prior to being overridden. Rafting of *Afton Lodge Clay* and deposition of *Eglinton Shelly Till* in Ayrshire. Growth of Southern Uplands ice centre	B
B	32–25 ka LLGM	N–S ice divide develops over the Firth of Clyde from which wide corridor of fast-flowing flows E towards the Firth of Forth depositing the *Wilderness Till* across central Scotland	B
C	25–22 ka LGM	Enhanced drawdown to W causes eastward migration and lowering of the N–S divide linking W Highlands and Southern Uplands ice centres. The ice divide sits across the Clyde basin during the LGM	P/B
D	22–19 ka	The N–S ice divide migrates back towards W and thickens over W Highlands and the Firth of Clyde	B
	19–16.5 ka	Deglaciation commences with widespread cutting of ice-marginal glacial drainage channels and ponding to W of the regional E–W watershed	P
E	16.5–16.2 ka	Ice streams E and SE into the Clyde basin during deglaciation. Ice-front oscillations at Blantyreferme causes extensive ponding in the Clyde valley upstream	P/R

southern Kintyre and northern Arran, together with the local preservation of the mountaintop tors, suggest that either cold-based ice protected these areas from subglacial erosion or they formed nunataks. This likely resulted from the migration of the North Channel ice divide across these areas, linking Kintyre and Antrim.

Following rapid wasting of the Irish Ice Sheet, ice flowing from the more substantial ice centre positioned over the western Scottish Highlands temporarily advanced into Northern Ireland during the 'East Antrim Coastal Readvance' (McCabe & Williams 2012; Finlayson *et al.* 2014; Fig. 27f). Suites of converging subglacial landforms identified by Finlayson *et al.* (2014) across NW Kintyre, parts of southern Kintyre and southern Arran indicate that ice became increasingly topographically constrained during deglaciation (*stage 6*). The distribution and orientation of ice marginal meltwater channels mapped by Finlayson *et al.* (2014) in the region indicate that the major pathways of glacier retreat were through corridors of low-lying ground and glacially over-deepened basins (Fig. 27g).

The pattern of retreat of local glaciers on Arran deduced by Gemmell (1973) has been generally accepted, involving limited advances following separation from the main retreating outlet glaciers occupying the Kilbrannan Sound and Firth of Clyde (*stage 7*). The sediments and geomorphology in Dougarie Glen, western Arran, together with the results of recent cosmogenic dating there (Ballantyne 2007; Finlayson *et al.* 2014), suggest that a limited advance of local ice occurred at *ca.*16.2 ka BP following a fall in RSL from 32 m OD (Fig. 27h). This timing supports RSL simulations for the area, where a falling RSL is modelled between *ca.*16.5 and 15 ka BP (Shennan *et al.* 2006; Smith *et al.* 2018).

5.4. West Central Lowlands

The lithostratigraphy of the region (McMillan *et al.* 2011a), which includes the Glasgow conurbation and Ayrshire, is based on formations originally proposed by Rose (1981, 1989), Browne & McMillan (1989) and Sutherland (1999) (Fig. 28). The history of glaciation has been disputed (Price 1975; Menzies 1976, 1981; Sutherland & Gordon 1993e; Rose & Smith 2008), but recent geomorphological, lithostratigraphical and three-dimensional (3D) mapping investigations have substantially resolved the glacial history of this area (Finlayson 2012; Finlayson *et al.* 2010, 2014; Kearsey *et al.* 2018).

5.4.1. Glasgow area (Clyde Basin). At *Balglass Burn*, N of Glasgow (Fig. 29f), a weathered till containing partially decomposed clasts with weathering rinds underlies organic sands that have been radiocarbon dated to the Middle Devensian (36.0–34.9 cal ^{14}C ka) (Brown *et al.* 2007). The sands are capped by relatively unweathered till formed during the last glaciation. Thus, Balglass is one of very few reliably described and sampled localities in Scotland where ice-free conditions can be proven immediately before the last glaciation, at *ca.*35 cal ^{14}C ka BP (Table 13); others notably include Sourlie (see the following section) and Tolsta Head, on the Isle of Lewis (Whittington & Hall 2002; Hall *et al.* 2018) (Section 3.3).

The considerable degree of weathering observed in the lower till at Balglass suggests that it was formed in a pre-Ipswichian (MIS 5e) glaciation (Brown *et al.* 2007). The unit probably correlates with the *Ballieston Till Formation* in the Clyde basin, which was laid down by a substantial advance of ice from the NW following a significant period of exposure (Price 1975; Browne & McMillan 1989; Sutherland & Gordon 1993e; Fig. 29a). The Ballieston Till is shown from site investigations in parts of northern Glasgow to be overlain by gravelly sands of the *Cadder Sand Formation* (Menzies 1976, 1981; Finlayson 2012; Kearsey *et al.* 2018), from which bones and teeth of woolly rhinoceros have been obtained at two localities in the vicinity of *Bishopbriggs* (Fig. 29f) (Hungryside and Wilderness Pit) (Rolfe 1966). Following ultrafiltration pre-treatment, a revised radiocarbon age of 31.1 ^{14}C ka BP (*ca.*35 cal ^{14}C ka BP) has been obtained from bones collected from Wilderness Pit (Jacobi *et al.* 2009). Excavations at the type site of Wilderness Pit (Rose 1989) showed that the Cadder Sand is locally overlain by water-lain diamictons of the *Broomhill Clay Formation*. Borehole records reveal that the latter comprises up to 15 m of very compact, laminated, glaciolacustrine mud with dropstones (Browne & McMillan 1989; Finlayson 2012). These glaciolacustrine muds, if varved, indicate that ponding occurred for at least 1000 years during the earliest stages of the last glaciation (Fig. 29a) and their elevation suggests that contemporary sea level was at least 25 m below present sea level (Browne & McMillan 1989). The Broomhill Clay, Cadder Sand and Ballieston Till are mainly located within concealed bedrock depressions, particularly a 108-m-deep, 1500-m-wide and 26-km-long elongate depression, probably a tunnel valley (*cf.* Kehew *et al.* 2012), that trends WSW–ENE beneath

the floodplain of the River Kelvin (Kearsey *et al.* 2018) (Fig. 29f).

The Broomhill Clay and/or Cadder Sand are generally overlain by the widespread *Wilderness Till Formation* (Fig. 3b) assigned to the last glaciation (Rose *et al.* 1988; Rose 1989; Sutherland & Gordon 1993e; Forsyth *et al.* 1996; Hall *et al.* 1998). At Wilderness Pit the Sand is sheared, folded and includes seams of diamicton, possibly glacitectonite, and the overlying Wilderness Till includes tectonic slices and rafts of sand and laminated clay from underlying units (Hall unpublished). In eastern Glasgow, the Wilderness Till is overlain by the *Broomhouse Sand and Gravel Formation*, which forms ice-contact topography including eskers, mounds, flat-topped kames and kettleholes (Browne & McMillan 1989; Rose & Smith 2008; Finlayson 2012; Kearsey *et al.* 2018). The formation may be subdivided locally into the *Ross Sand Member* (deltaic sand and gravel) and the *Bellshill Clay Member* (glaciolacustrine laminated clay) (Browne & McMillan 1989). These units were deposited in ice-dammed 'Glacial Lake Clydesdale' (Bell 1871; Sissons 1967a, b) after the ice sheet margin had retreated NW (Sutherland & Gordon 1993e) (Fig. 29e).

Late-glacial raised glaciomarine deposits assigned to two members of the *Clyde Clay Formation* (Fig. 28) were laid down around the Clyde estuary downstream of Glasgow following deglaciation at *ca.*15 ka BP (Peacock 1971, 2003; Rose 2003; McMillan *et al.* 2011a; Fig. 30). The *Paisley Clay Member*, which typically comprises thinly laminated mud with dropstones and includes the cold-water foraminifera *Elphidium clavatum*, occurs up to altitudes of *ca.*40 m OD (Browne & McMillan 1989). The *Linwood Clay Member* is confined to western areas of the estuary, where it commonly overlies the Paisley Clay; it consists of more thickly bedded mud, with a richer, warmer-water faunal assemblage (Peacock 2003).

5.4.2. Ayrshire. The sequence in Ayrshire (Fig. 28) is largely based on sections in a former opencast coal pit at *Sourlie*, near Irvine (Jardine *et al.* 1988; Sutherland & Gordon 1993e; Sutherland 1999; Bos *et al.* 2004; Finlayson *et al.* 2010), and on natural river bank sections at *Afton Lodge*, near Ayr (Merritt *et al.* 2014) (Fig. 30). Both sites lie within a suite of ribbed moraine (Fig. 30). The importance of the Sourlie site lay in the discovery of thin lenses of organic material occurring between two units of till (Jardine *et al.* 1988). These lenses, now assigned to the *Sourlie Organic Silt Formation* (McMillan *et al.* 2011a), occupy shallow depressions within the surface of the underlying glaciofluvial *Armsheugh Sand and Gravel Formation*. They have yielded a very rich flora and fauna deposited within a shallow pond in a treeless, low-shrub to sedge-moss tundra environment, together with bones of woolly rhinoceros and antler fragments of reindeer. Radiocarbon dates on the antler, plant debris and bulk organic matter range between 33.5 and 29.0 cal ^{14}C ka BP, confirming a Middle Devensian age (Bos *et al.* 2004). The basal unit at Sourlie (*Littlestone Till Formation*) comprises dark-grey stony diamicton locally enclosing deformed sheets of sand. The till is overlain by up to 3.5 m of clay-rich gravel of uncertain origin named as the *Lawthorn Diamicton Member*. The two units were assumed to have been deposited during an Early Devensian glaciation by Sutherland (1999), but they could be older.

At Sourlie, the *Wilderness Till* has been divided into two members (Fig. 28). The *Eglinton Shelly Till* is a diamicton containing fragments of marine shells together with far-travelled sedimentary, igneous and metamorphic rocks from the Scottish Highlands. It represents the 'shelly tills' of the region (Sutherland & Gordon 1993e), and some *Arctica islandica* shell fragments have yielded Late Devensian amino

acid ratios (Jardine *et al.* 1988). The overlying *Auchenwinsey Till* is a grey diamict containing local Middle Coal Measures and basic igneous rock lithologies (Jardine *et al.* 1988).

The Eglinton Shelly Till at Afton Lodge is associated with several large bodies of shelly clay (see Gordon 1993k and references therein), which have been assigned to the *Afton Lodge Clay Formation* by Merritt *et al.* (2014). They comprise weakly laminated to massive, stiff, calcareous mud containing sparse shells, foraminiferids, ostracods, fish debris and echinoid spines. The generally well-dispersed pebbles include Carboniferous sandstone, coal and basalt together with more far-travelled crystalline 'Highland' rock types. The shelly clay contains abundant marine macro- and micro-fauna, as well as palynomorphs consistent with its deposition within a shallow marine to estuarine environment. Faunal changes within the main body of marine clay record at least one millennial-scale cycle of Arctic-boreal, to boreal, and back to Arctic-boreal climatic conditions (Merritt *et al.* 2014). A radiocarbon date of >41 ^{14}C ka BP obtained from the foraminifera indicates that these marine clays are older than the surrounding till (Merritt *et al.* 2014).

The location of Afton Lodge suggests that the rafts were transported at least 10 km from their likely source within the Firth of Clyde basin, prior to being accreted at between 70 and 82 m OD (Merritt *et al.* 2014). The rafts of shelly clay, together with closely associated bodies of 'shelly till', which were possibly formed mainly by the attenuation and crushing of other rafts during their transport within the subglacial deforming bed (*cf.* Aber 1985), were emplaced during an early phase of the LGM glaciation of Scotland by ice flowing from the western Grampian Highlands of Scotland. This ice flowed through the topographically confined Firth of Clyde basin where it probably detached the blocks of unfrozen marine sediment prior to splaying-out onshore against reversed slopes, leading to subsequent raft emplacement and creation of closely associated ribbed moraine (Merritt *et al.* 2014). The transport of the rafts was probably facilitated by the development of water-lubricated décollement surfaces along their margins (*cf.* Phillips & Merritt 2008).

5.4.3. Regional interpretation. The published literature (Sissons 1967a; Price 1983; Rose & Letzer 1977; Sutherland 1984; Sutherland & Gordon 1993e; Rose & Smith 2008; Finlayson 2012; Finlayson *et al.* 2010, 2014), together with the numerical ice sheet model of Hubbard *et al.* (2009), indicate that ice initially advanced into the area from the NW (Figs 27a, 29a). The ice front generally advanced against reverse slopes leading to the widespread formation of ribbed moraine perpendicular to flow (Finlayson *et al.* 2010; Hughes *et al.* 2010; Fig. 30). That landforms from such an early stage of glaciation could survive is supported by the preservation of Middle Devensian deposits, such as at Sourlie and Bishopbriggs. Merritt *et al.* (2014) conclude that the rafts of Afton Lodge Clay were detached and entrained into the deforming layer beneath the ice at this stage as it advanced from the Firth of Clyde basin SE towards *Nith Bridge*, near New Cumnock (Sutherland 1993g; Fig. 30), and S to Stranraer, the most southerly known occurrence of shelly diamicton in Scotland (Phillips *et al.* 2009; Peacock & Everest 2016). The advance of ice into the Clyde basin resulted in ponding (Price 1975) and the ribbed moraine may have been initiated locally as ice-marginal sediment ridges generated by folding and thrusting (McMillan & Browne 1983; Finlayson 2012).

Sustained stadial conditions at the LLGM (30–25 cal ka BP) resulted in the development of an ice divide over the Firth of Clyde basin, beneath which there was minimal subglacial erosion and landscape modification. Once ice had become sufficiently thick to overtop the main Clyde–Forth drainage
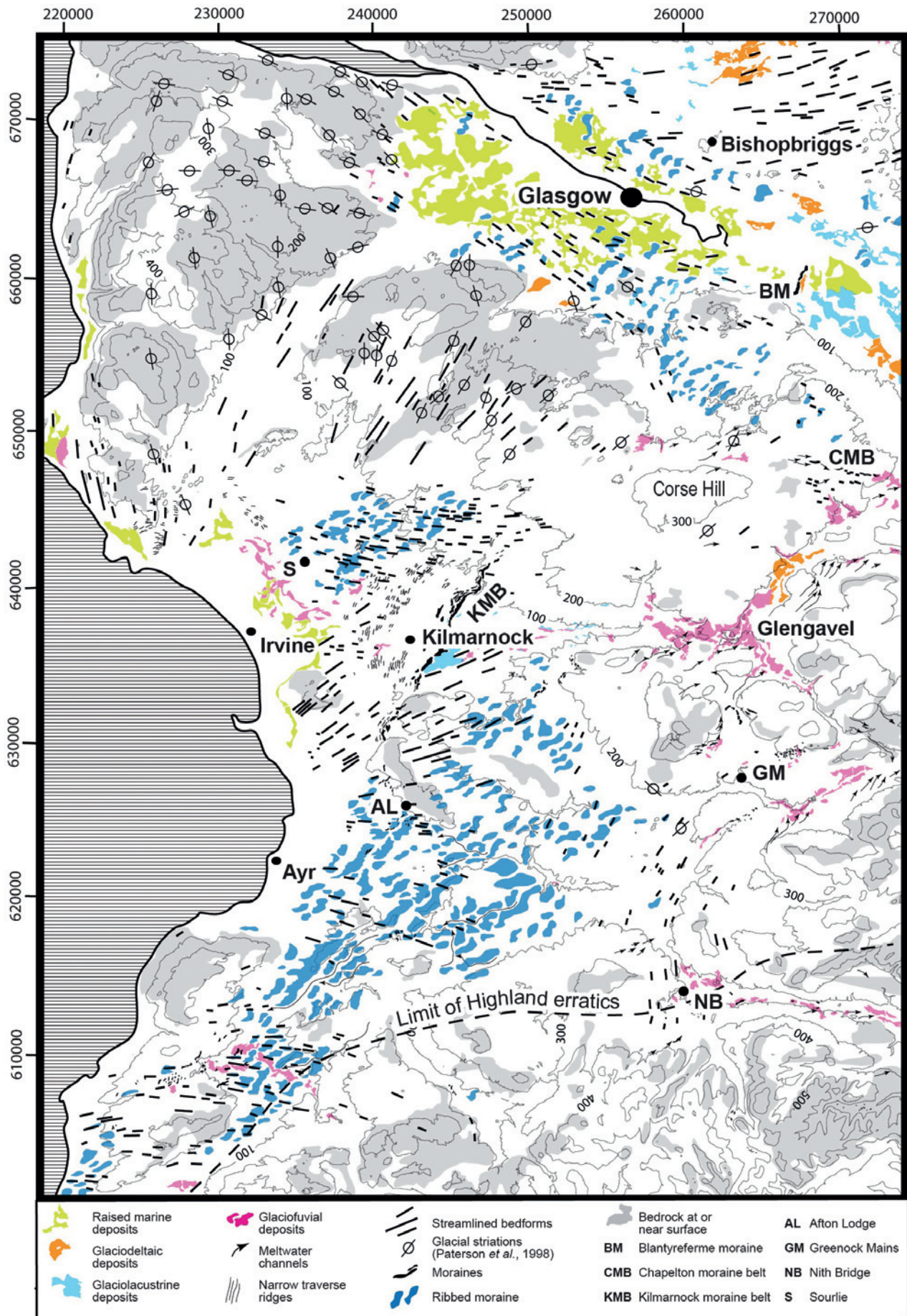
Figure 30 Glacial geomorphology and geology of the Clyde and Ayrshire basins. (Reprinted from *Quaternary Science Reviews*, **29**, 969–88, Finlayson, A., Merritt, J. W. Browne, M., Merritt, J. E., McMillan, A. & Whitbread, K. Ice sheet advance, dynamics and decay configurations: evidence from west central Scotland. Copyright (2010), with permission from Elsevier.) Erratic limits are from Eyles *et al.* (1949) and glacial striations from Paterson *et al.* (1998).

divide, a wide corridor of fast-flowing ice flowed E towards the Firth of Forth forming drumlins (Rose 1987; Rose & Smith 2008; Finlayson 2012), striae (Paterson *et al*. 1998) and a train of essexite erratics that were dispersed from their source near Lennoxtown at the foot of the Campsie Fells N of Glasgow (Peach 1909; Shakesby 1978) (Figs 27b, 29b). A significant dispersal centre had developed over the Southern Uplands by this stage, contributing to deflection of ice in the Clyde basin towards the NE. This flow formed streamlined bedforms and deposited till of Southern Uplands provenance on shelly dia-mictons (Eglington Shelly Till) at Nith Bridge (Holden & Jar-dine 1980; Sutherland 1993e; Fig. 30).

A broad belt of streamlined bedforms in rock, striae and drumlins carved into the earlier-formed ribbed moraine indicate convergent flow into the outer Firth of Clyde basin (Rose & Letzer 1977; Finlayson 2012; Finlayson *et al*. 2014; Fig. 27c). This phase clearly followed a major change in the ice sheet configuration that probably resulted from enhanced marine drawdown to the W, initiating an eastward migration and sub-stantial lowering of the ice divide (Figs 27c, 29c). This glacial reorganisation probably eventually reduced the power of eastward flow towards the Firth of Forth and resulted in the relatively early deglaciation of eastern Scotland (Sissons 1967a, b; Sutherland 1984). An intermediate phase probably occurred when ice covering the Southern Uplands was scavenged and drawn down to the W around the S of Arran, as indicated by the westward transportation of microgranite erratics derived from the island of Ailsa Craig, off the SW Scottish coast (Sissons 1967a; Fig. 27c).

Following substantial thinning and local deglaciation, a subsequent change in ice sheet configuration is indicated in the E of the region by a suite of landforms including local E-trending subglacial bedforms, eastward-descending ice-marginal meltwater channels and suites of ice-contact glacio-fluvial landforms (including the *Chapelton Moraine Belt*) (Finlayson *et al*. 2010; Clark *et al*. 2012a; Finlayson 2012; Fig. 30). Collectively, they demonstrate ice-divide migration back towards the W, coupled with probable thickening of ice located over the Firth of Clyde basin (Finlayson 2012; Finlayson *et al*. 2014; Fig. 27d). The Highland and Southern Uplands ice caps subsequently separated across southern parts of the Clyde Basin in the vicinity of the *Carstairs Eskers* (Finlayson 2012; Fig. 29d), supporting the conclusion of Thomas & Montague (1997) and Huddart & Bennett (1997) that this well-documented esker system (Gordon 1993l), and others nearby (Bennett *et al*. 2007), developed in an inter-lobate sediment sink during uncoupling of northern- and southern-sourced ice caps. The configuration is roughly that proposed for the supposed 'Perth Readvance' in central Scotland (Sissons 1963, 1964, 1967a, b), when the upper catchment of the River Clyde drained eastwards into that of the River Tweed (Shaw & Merritt 1981). A minor readvance of Southern Uplands ice is apparent in the Ayr valley near Greenock Mains (Fig. 30) where the orientation of a nested assemblage of melt-water channels and moraine ridges indicates a late NE advance (Holden & Jardine 1980). Ice-marginal ponding first began on the watershed between the catchments of the NE-flowing Avon Water and the W-flowing River Irvine, in Glengavel (Figs 29d, 30), where thick sequences of laminated glaciolacustrine silt coarsen upwards into deltaic sand and gravel occur up to an elevation of at least 205 m OD (Nickless *et al*. 1978).

A phase of more active and widespread streamlining towards the E and SE accompanied the final phase of ice sheet thinning in the region (Fig. 29e). One or more significant ice-front oscilla-tions occurred late during deglaciation. The streamlined bed-

forms assigned to this phase terminate abruptly in the vicinity of the Kilmarnock and Blantyreferme moraine complexes (Fig. 30), to the E of which suites of glaciolacustrine and deltaic sediments were laid down in the Clyde and Irvine valleys (Fig. 29e). Water previously ponded only within upper parts of the Avon Valley eventually merged with a much larger lake, 'Glacial Lake Clydesdale' (Bell 1871), which occupied the Clyde valley and its tributaries upstream of Glasgow (Fig. 29e). Paterson *et al*. (1998) deduced that the level of Lake Clydesdale dropped in stages as the NW retreat of the ice margin in the lower Clyde Valley made available spillways to the E at progressively lower elevations of 200, 165, 102 and 85 m OD, respectively. The lake finally drained E via a col in the upper Kelvin valley, NE of Glasgow, at about 45 m OD (Forsyth *et al*. 1996; Hall *et al*. 1998).

Lake Clydesdale probably existed during the creation of the 'Main Perth Shoreline' in the Forth estuary (Sissons & Smith 1965; Sutherland 1984), and possibly into the beginning of the Lateglacial Interstadial (GI-1) (Peacock 1999, 2003). The timing, contemporary sea level and manner in which the Late-glacial sea eventually invaded Lake Clydesdale is disputed (see Peacock 2003 for review), but it is generally accepted that the transgression had occurred by 13.1–12.5 ^{14}C ka BP (based on uncorrected, reported ages from marine shells) (Peacock 1971, 2003; Browne *et al*. 1977; Peacock *et al*. 1977; Rose 2003). Eventual disintegration of ice blocking the Clyde estuary resulted in the level of Lake Clydesdale falling from 45 m to no less than 40 m, the contemporaneous sea level (Peacock 2003). Later stages in the deglaciation to the W of Glasgow have been deduced by Finlayson *et al*. (2014) (Section 5.3).

5.5. Solway Lowlands

A review of the complex lithostratigraphical and geomorpho-logical record of the last glaciation within the Solway Low-lands and surrounding areas has been given by Livingstone *et al*. (2012), who identify seven distinct sets of glacially streamlined features relating to six stages of glaciation (Fig. 31). The oldest known unit on the Scottish side of the border is the *Chapelknowe Till Formation* (McMillan *et al*. 2011b; Table 14), which was possibly laid down during an 'Early Advance' of Scottish ice across the Solway Firth, up the Vale of Eden and through the Stainmore Gap into eastern England (Trotter 1929; Hollingworth 1931; Trotter & Holling-worth 1932b; Evans *et al*. 2005; Stone *et al*. 2010; Mitchell 2013; Ballantyne & Small 2018; Fig. 31a). The till contains a suite of Southern Uplands erratics, including Criffel and Dalbeattie granite and greywacke. The Chapelknowe Till is overlain locally by the 'Middle Sands' of a 'tripartite' sequence that occurs around Gretna, pinching out just E and S of Carlisle (Trotter & Hollingworth 1932a). This middle unit is now assigned to the *Plumpe Farm Sand and Gravel Formation*, which is capped by a thin 'upper' red diamicton, the *Plumpe Bridge Till Member of the Gretna Till Formation* (McMillan *et al*. 2011b). The widespread red, sandstone-rich Gretna Till was laid down by ice flowing from the W as indicated by its composition and the predominant orientation of drumlins across the lowlands N of the inner Solway Firth (Livingstone *et al*. 2008; McMillan *et al*. 2011b; Fig. 3b). Relatively little ice flowed into the lowlands from the Langholm Hills, to the N, as revealed by the distribution of the greywacke-rich *Langholm Till Formation* (British Geological Survey 2005, 2006; Fig. 3b).

The occurrence of red, granite-bearing till (unnamed) pass-ing upwards into rubbly, greywacke-rich till in the Langholm Hills indicates that ice overwhelmed the area from the W
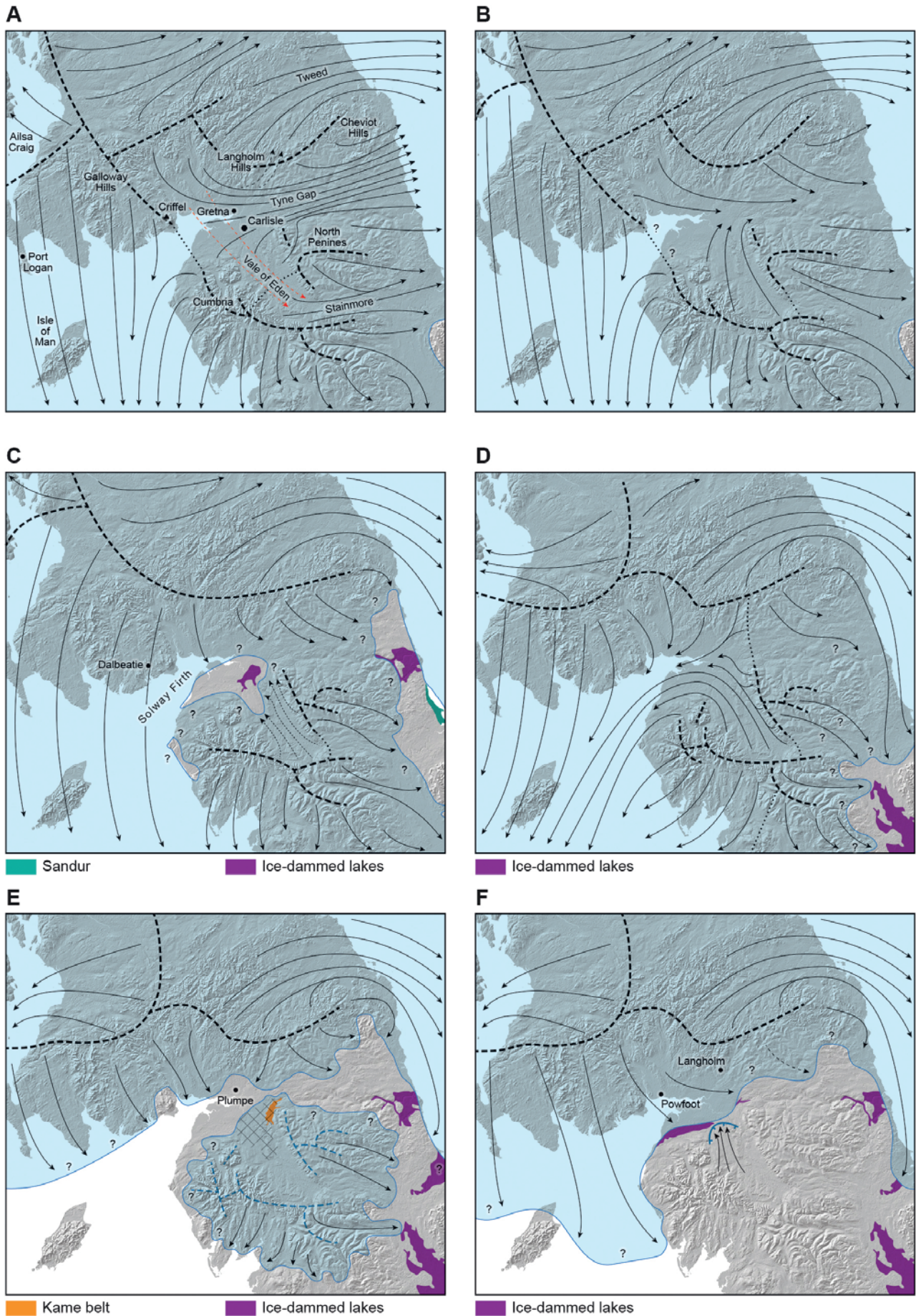
Figure 31 Reconstructed stages (A–F) of glaciation around the Solway Firth (from Livingstone *et al.* 2012, trimmed and with minor modification). (Reprinted from *Earth-Science Reviews*, **111**, 25–55, Livingstone, S. J., Evans, D. J. A., O'Cofaigh, C., Davies, B. J., Merritt, J. W., Huddart, D., Mitchell, W. A., Roberts, D. H. & Yorke, L., Glaciodynamics of the central sector of the last British-Irish Ice Sheet in Northern England. Copyright (2012), with permission of Elsevier.) Dashed-dotted lines refer to ice divides (with the thick dotted lines indicating possible ice-saddles) and the arrows indicate ice-flow vectors (dotted arrows indicate alternative ice-flow scenarios). Hill-shaded surface model built from Intermap Technologies NEXTMap Britain elevation data.

Table 14 Summary of events in the Solway Lowlands. Stages and estimated ages after Livingstone *et al.* (2012) and Chiverrell *et al.* (2018). Abbreviations: B = glacial 'binge'; I = Interglacial; P = glacial 'purge'.

Stage	Age	Events, palaeo-environments, important sites and units	
	MIS 5e	Raised rock platforms formed around Port Logan at *ca.*10 m OD (now till covered) (Sutherland 1993h)	I
	MIS 5a–d	No reported evidence	
	MIS 4	Possible 'Early Advance' of Scottish ice up the Vale of Eden during an Early Devensian glaciation	B
	MIS 3	No reported evidence	
1	28–25 ka LLGM	Likely time of 'Early Advance' of Scottish ice up the Vale of Eden laying down the *Chapelknowe Till* near Gretna	B
2	∼25–22 ka LLGM	An ice divide links the Cumbrian Mountains and Galloway Hills across the Solway Firth, forcing ice E through the Tyne Gap and over Stainmore, and probably NE into the Tweed catchment, laying down the *Gretna Till*. Cold-based ice over the Langholm and Cheviot Hills	B
3	∼22–20 ka LGM	Flow over Stainmore stops, but ice continues to flow E through the Tyne Gap	P
4	∼20 ka	The ice divide over the Solway Firth breaches followed by partial deglaciation of the Solway Lowlands where varved clays are deposited near Carlisle (*Blackhall Wood Glaciolacustrine Fm*). Cessation of flow through the Tyne Gap	P
5	∼20–19.2 ka	Ice readvances into the Solway Lowlands and streams towards the Irish Sea basin partially remoulding the Gretna Till (*Blackhall Wood Readvance*)	B
6	∼19.2–18.2 ka	Another partial deglaciation and glacial reorganisation in the Solway Lowlands. Deposition of the *Plumpe Sand and Gravel* near Gretna	P
7	18.2–∼17 ka	Ice readvances SE across the inner Solway laying down the *Plumpe Bridge Till Mb* of the Gretna Till (*Scottish Readvance*). Ponding of ice marginal lakes including 'Glacial Lake Carlisle'	B
8	∼17–16 ka	Deglaciation punctuated by minor readvances of Southern Upland ice near Langholm and Powfoot (Fig. 31f). Widespread ponding and glaciofluvial deposition. Galloway Hills deglaciated by 15.2 ka	P

during an early phase of the last glaciation (Lumsden *et al.* 1967; McMillan *et al.* 2011b). NEXTMap imagery of the Langholm Hills reveals that this region of the Southern Uplands has been relatively little modified by glacial erosion because it was a minor ice-dispersal centre and probably over-lain by mainly sluggish, cold-based ice (Fig. 31a). A minor, southward readvance of ice from the Langholm Hills occurred following the separation of ice masses during deglaciation (Lumsden *et al.* 1967), laying down rubbly diamicton of the *Mouldy Hills Moraine Formation* (British Geological Survey 2006). Although the Gretna and Chapelknowe till formations have not been mapped to the E of the Langholm Hills, sand-stone and granite-bearing red tills including a suite of glacial erratics sourced in the Galloway Hills have been recorded there, high on the catchment divide (Day 1970). Together with new evidence from NEXTMap and satellite imagery revealing elongate, glacially streamlined landforms arcing NNE towards the divide (Livingstone *et al.* 2008; McMillan *et al.* 2011b; Fig. 31a), this suggests that transfluence of ice may have occurred into the catchment of the Tweed at the LGM.

The Plumpe Bridge Till has been connected with the late-stage 'Scottish Readvance' into the Solway Lowlands (Dixon *et al.* 1926; Trotter & Hollingworth 1932a; Huddart 1991; Merritt & Phillips 2010). Indeed, the thin and laterally discontinuous nature of the till probably results from a short-lived advance across water-saturated sediment adjacent to proglacial lakes dammed up against higher ground to the E of the ice (Phillips *et al.* 2007) (Fig. 31f). The 'Middle Sands' were considered to have formed proglacially in, and around, 'Glacial Lake Carlisle' following partial deglaciation of the Solway Lowlands (Dixon *et al.* 1926; Trotter & Hollingworth 1932a). This interpretation has been confirmed by the identification in the Caldew Valley, SW of Carlisle, of a sequence of laminated mud deposited in a pro-glacial lake (Livingstone *et al.* 2010a, b; Fig. 31b). These varved sediments (*Blackhall Wood Glaciolacustrine Formation*), which are also overlain by a reddish-brown 'upper till' that includes mixed Scottish, Lake District and local clasts, are now assigned to the *Blackhall*

Wood/Gosforth Oscillation (Livingstone *et al.* 2010b, c, 2012; Ballantyne & Small 2018).

Evidence for a subsequent, minor readvance of Scottish ice includes a striated boulder pavement that is intermittently exposed at low tide W of Powfoot (Brookfield *et al.* 2010; Merritt 2010). The pavement consists of a single layer of interlocking, ice-bevelled boulders composed predominantly of wacke sandstone with some red Triassic sandstone and sparse granodiorite derived from Criffell, some 15 km to the W. The boulders have smoothed and striated upper surfaces and both the long axes of the boulders and the striations show a consistent WNW–ESE alignment. The simplest explanation is that the pavement formed subglacially beneath wet-based ice that flowed ESE across previously deposited till and glacio-fluvial deposits. The presence of granodiorite boulders from Criffell in diamicton locally overlying the pavement, but not below it, is compatible with a late-stage readvance of ice that splayed out into the Solway Firth from valleys to the N.

The age, regional correlations and integrity of the Scottish Readvance have been controversial (Merritt & Auton 2000; Evans *et al.* 2005; Livingstone *et al.* 2012), but many issues have been resolved recently following new geochronological, offshore geophysical and stratigraphical evidence collected from around the northern Irish Sea Basin (Chiverrell *et al.* 2018). The latter authors constrain the event to 19.2–18.2 ka and demonstrate it to be a significant regional readvance across the Isle of Man and Cumbrian lowlands. It probably correlates with the Gosforth Oscillation that affected the W Cumbrian coast (Trotter *et al.* 1937; Merritt & Auton 2000; Livingstone *et al.* 2012) and the Clogher Head Readvance at the coast of County Down, in Northern Ireland (McCabe *et al.* 2007a, b; Clark *et al.* 2012b; Merritt *et al.* 2018).

5.6. East Central Lowlands, Tayside, Fife and the Borders
In the apparent absence of new, comprehensive field studies in eastern Scotland, McMillan *et al.* (2011a) assigned existing stratigraphic units to their lithostratigraphic framework (Fig. 3b). Given the historical importance of this region

(Sutherland 1993i, j), and particularly the Lothians and Borders (Gordon 1993m), it merits detailed reinvestigation, for example in the context of recent understanding of ice-flow-pattern evolution and palaeo-ice streams (Everest *et al.* 2005; Golledge & Stoker 2006; Hughes *et al.* 2014; Livingstone *et al.* 2017). For example, the results of mapping in the valley of the Kale Water, W of the Cheviot Hills, revealed that this upland area lay beneath a cold-based, frozen patch to the E of a thermal boundary related to the Tweed ice stream (Mitchell 2008; Fig. 31). In the past 15 years, superficial deposits have been mapped by BGS in the Borders (around Duns and Moffat), central Galloway, Montrose and Crieff, but no descriptive and interpretive accounts have been written.

A digital dataset of over 100,000 boreholes that penetrate the full thickness of Quaternary deposits in the Central Lowlands has been utilised in a recent study of 'buried' valleys and their infills (Kearsey *et al.* 2018). The study identified 18 buried palaeo-valleys in the region ranging from 4 to 36 km in length and from 24 to 162 m in depth, some cross-cutting one another. The deepest features are orientated WSW–ENE beneath the Kelvin valley, NE of Glasgow (Browne & McMillan 1989), the Carron valley, farther to the ENE between Falkirk and Grangemouth (Browne *et al.* 1984; Cameron *et al.* 1998; Browne 2011), and the Devon valley at the foot of the Ochil Hills, to the E of Stirling (Parthasarathy & Blyth 1959). These palaeo-valleys all have undulating longitudinal profiles and are interpreted as tunnel valleys (*cf.* Kehew *et al.* 2012) formed during the LGM (Kearsey *et al.* 2018). The shallower features are generally aligned with ice flow during ice sheet retreat.

6. Advances and setbacks over the past 25 years in understanding the terrestrial stratigraphic record of the last glacial cycle in Scotland

6.1. Advances
Progress in the last 25 years has been facilitated by a number of developments.

The application of a systematic approach to lithostratigraphy (Sutherland 1999; McMillan *et al.* 2011a) has enabled a more rigorous framework for the recognition and correlation of Pleistocene deposits.

Advances in dating techniques have improved understanding and timing of events and the development of a rigorous chronostratigraphy. Refinements in radiocarbon dating through the use of accelerator mass spectrometry have enabled the application to smaller samples and reduced the error limits on older material pre-dating the last glaciation (e.g., Bos *et al.* 2004; Brown *et al.* 2007; Jacobi *et al.* 2009). Recognition of natural variations in the atmospheric concentration of radiocarbon through time and the development of calibration methods to a common calendar timescale have greatly enhanced the comparability and interpretation of dates (Reimer *et al.* 2013). Although there have been limited applications in Scotland, improvements in luminescence dating and amino acid geochronology have helped clarify Ipswichian and Early Devensian chronostratigraphy, particularly in NE Scotland and around Inverness (e.g., Merritt 1992; Duller *et al.* 1995; Hall *et al.* 2002). Surface exposure dating using terrestrial cosmogenic nuclides has greatly advanced understanding of the timing, sequence and pattern of the last ice sheet glaciation (Ballantyne & Small 2018 and references therein).

The use of satellite imagery and digital surface models derived from remote sensing, and particularly NEXTMap Britain topographic dataset, has progressed geomorphological mapping and morphostratigraphy (Chandler *et al.* 2018).

It has enabled the recognition of multiple sequential ice-flow patterns during the life of the last BIIS indicated by superimposed glacial lineations (e.g., Salt & Evans 2004; Golledge & Stoker 2006; Hughes *et al.* 2014) and has considerably improved knowledge of glaciation during the Younger Dryas (Ballantyne 2018; Bickerdike *et al.* 2018).

3D modelling based on borehole data has helped understanding of the architecture of layered sedimentary sequences (e.g., Finlayson 2012; Kearsey *et al.* 2018). When combined, the use of lithostratigraphy, geomorphological mapping, 3D modelling and geochronometry has proved a powerful combination in elucidating the dynamics and pattern of decay of the last ice sheet (Finlayson *et al.* 2010, 2014).

The significant progress in mapping and dating offshore landforms and sediments reviewed by Stewart *et al.* (2019) has provided a more extended lithostratigraphic and chronostratigraphic framework in which to set the incomplete terrestrial record. Ice sheet modelling (e.g., Boulton & Hagdorn 2006; Hubbard *et al.* 2009) has also highlighted the highly dynamic nature of the last ice sheet and provided further context for interpreting the stratigraphic record.

6.2. Constraints
Progress has been hampered by a decline in stratigraphic studies, arising from a number of factors.

There has been a shift in research away from regional studies of glacial deposits and landforms based around detailed field survey, mapping and logging of sections to more systematic, desk-based, remote sensing studies and analysis of digital topographic imagery. While the latter have produced important new results, particularly at regional and national scales, the two approaches must be complementary.

Few new important natural sections have been discovered and then described. This has been compounded by the end of open-cast coal mining, the opening of fewer small gravel pits and increasingly stringent restrictions being applied to access resulting from more rigorous health and safety legislation. Meanwhile, many former sections have been lost to development, infilling, land restoration or natural degradation. There has also been an absence of resources (human and financial) for routine logging of exposures arising from construction and infrastructure projects (e.g., roads, pipelines, power lines, run-of-river hydro schemes and forestry and wind farm access routes) and for re-excavation of former sections that are now degraded. New excavations typically involve 'cut and cover', so that exposures are largely temporary.

6.3. Next steps
Significant advances have been made in elucidating the pattern and timing of deglaciation, but these results have yet to be integrated into numerical ice sheet models in order to test how closely computer simulations match litho- and morph-stratigraphic records. Successful tests would give confidence that model design replicated ice sheet behaviour and so offer the promise of using improved models as frameworks in which to piece together the much more fragmentary records of events during and before the LGM.

Despite the huge step forward that the Natural Environment Research Council (NERC)-funded BRITICE-CHRONO consortium has made in dating events during deglaciation, further dating is needed, including the application of luminescence, radiocarbon and cosmogenic methods. However, it is suggested that the focus of dating should now switch towards the LGM and older events. That could usefully include attempting to extend the Scottish varve chronology back beyond the Younger Dryas (McLeod *et al.* 2011). This, for example, could begin

with detailed examination, targeted drilling and sampling of known exposures of laminated glaciolacustrine clay concealed beneath till, such as in the Solway Lowlands (Livingstone et al. 2010a, b), the Irvine valley upstream of Kilmarnock and around Glasgow (Finlayson et al. 2010), in the catchment of the Ugie Water, in eastern Buchan (Merritt et al. 2003) and near Grantown-on-Spey (Merritt et al. 2017e). More cosmogenic ages on bedrock are required; for example, on glacially roughened and streamlined bedrock forms in order to help establish rates of glacial erosion, check for cosmogenic inheritance and attempt to date glacial lineations formed before the LGM. Attempts could also be made to date shore platforms covered by till (Smith et al. 2018). However, any new work should involve detailed lithostratigraphy as morphostratigraphy alone is of limited value for sediment sequences. Long-term programmes of section logging during major construction projects are required together with sampling boreholes.

The coverage of detailed litho- and morph-stratigraphic investigations remains patchy. Large parts of Sutherland, Easter Ross, the Outer Hebrides, the Eastern Grampians, Angus, Lothian, the Borders and the Tweed catchment lack recent published work. We observe that the focus of BRITICE-CHRONO and recent dating has been on glacial environments and specifically on ice limits. Relatively little attention has been given to Late Pleistocene non-glacial environments during the last 25 years. For example, there are readily accessible exposures of glacially transported rafts of marine and glaciomarine sediment, raised beaches on Lewis, Barra and Vatersay, Hoy and Kintyre, and buried periglacial soils and structures in Buchan and elsewhere. Old records of organic remains, including peat and bones found below, or within the Wilderness Till in the central lowlands (Sutherland 1984), represent targets for new studies.

7. Conclusions

Our knowledge of the sequence of events that have occurred during the last glacial–interglacial cycle in Scotland has vastly improved over the past 25 years, with ten or more separate events now identified in some regions. Research has benefitted from the development and application of a systematic approach to lithostratigraphy, digital surface models, glaciological models and new dating methods, combined with access to offshore seabed mapping, geophysical datasets and sediment core records. This has enabled a more rigorous spatial and temporal framework to be developed for the differentiation and correlation of terrestrial Late Pleistocene deposits.

Few new critical sites pre-dating the Late Devensian have become available, but further research on several existing sites from Shetland and NE Scotland has clarified aspects of Ipswichian (MIS 5e) and Early Devensian (MIS 5d-4) events and environments. New offshore evidence has become available that indicates ice sheet glaciation of Scotland during both the Early Devensian (MIS 4) and the Middle Devensian (MIS 3) is consistent with the climate records from Greenland ice cores and the Norwegian stratigraphic record. However, the terrestrial stratigraphic record in Scotland for MIS 4 and MIS 3 is poorly resolved in space and time.

The greater part of the terrestrial evidence apparently relates to the Late Devensian (MIS 2) and broadly accords with recent computer reconstructions of the last BIIS (Boulton & Hagdorn 2006; Hubbard et al. 2009). However, many truncated units of till occurring locally towards the base of glacigenic sequences are undated and could be older, like many ice-moulded landforms. The last ice sheet clearly was dynamic and probably reached its greatest thickness between 30 and 27 ka, during the local LGM (Ballantyne & Small 2018; Fig. 1). The ice sheet had two modes of flow-pattern geometry: an early, thick (relative to topography) integrated one with a relatively simple ice-divide structure, and a later, much thinner mode of ice sheet comprising multiple dispersion centres and a more complex flow structure (Hughes et al. 2014). The changing flow patterns are reflected in the lithostratigraphy. The ice sheet was episodically warm-based and erosive, but the pattern of warm-based flow was spatially variable. One former hypothesis firmly rebutted is that there were ice-free enclaves at the LGM in Scotland; there were none. The overall picture emerging is that a dynamic glacial system, highly sensitive to climate change drivers, existed in Scotland throughout much of the Devensian, but the terrestrial record so far as known remains incomplete. Nevertheless, terrestrial stratigraphy in its various forms (litho-, chrono- and morpho-stratigraphy) will continue to be a benchmark for constraining the pattern and timing of events during the Late Pleistocene in Scotland.

The enigmatic 'high-level' shelly clay deposits located around the Scottish coast (Sutherland & Gordon 1993a) are now all interpreted as glacially transported rafts of marine sediment rather than remnant in situ deposits formed during periods of high RSL before the Last glaciation (Sutherland 1981). Localities include Afton Lodge, in Ayrshire (Merritt et al. 2014), other localities around the coasts of the Firth of Clyde and Kintyre (Finlayson et al. 2010, 2014), the Rhins of Galloway (Peacock & Everest 2016), Clava, near Inverness (Merritt 1992), sites along the southern coast of the Moray Firth (Merritt et al. 2003) and the North Sea coast of the Grampian Highlands (Auton et al. 2000). Much of the controversy has centred on the local field evidence of ice flow, which Sutherland (1981) concluded to be generally offshore rather than onshore as required in the rafting model. However, the subsequent research has shown that an early phase of onshore flow did indeed take place at all known sites.

Several areas offer opportunities and challenges to address current gaps in understanding. First, improvements in dating methods and their novel applications are crucial, particularly to older deposits pre-dating the LGM. They include luminescence applications, revival of AAR dating, cosmogenic radionuclide (CRN) dating and development of new techniques. Second, digital surface models provide opportunities for refinements of morphostratigraphy and developing closer links with lithostratigraphy. Third, more detailed till provenance studies may help to clarify directions of ice movement, distances of ice transport and sediment recycling, ice movement indicators and rafting, recognising that the simple law of superposition does not always apply. Fourth, further applications of 3D modelling based on borehole data in combination with lithostratigraphy and geomorphological mapping should provide new opportunities to elaborate on the dynamics and pattern of decay of the last ice sheet. Fifth, detailed field mapping is required of the geomorphology and superficial deposits in key areas that have been overlooked in recent decades (e.g., the Lothians and Fife), supported by pit excavations and coring and links with morphostratigraphy and flow patterns from digital surface models. Sixth, studies of intermontane areas (e.g., Strathspey) and loch and sea loch basins may help to elucidate the history of mountain icefields and glaciers during periods of more restricted glaciation during the Middle Devensian. Seventh, digital modellers are now challenged to produce ice sheet reconstructions that accommodate the complex event stratigraphy that has been elucidated, particularly from recent offshore surveys.

8. Acknowledgements

9. References

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