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Key Points:
- Eighteen new ^{14}C dates constrain the culmination of the last glaciation in Scotland to the late Allerød–early Younger Dryas
- Our ^{14}C chronology contrasts with the current paradigm and suggests the glacial advance peaked ~1,000 years earlier than previously thought
- The Scottish glacial record supports a broader pattern of stadial deglaciation in the North Atlantic basin despite severe winter cooling

Supporting Information:
- Supporting Information S1

Correspondence to:
G. Bromley,
gordon.bromley@nuigalway.ie

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Interstadial Rise and Younger Dryas Demise of Scotland's Last Ice Fields

G. Bromley[1,2], A. Putnam[2], H. Borns Jr[2], T. Lowell[3], T. Sandford[4], and D. Barrell[5]

[1]School of Geography and Archaeology, National University of Ireland, Galway, Ireland, [2]Climate Change Institute and School of Earth and Climate Sciences, University of Maine, Orono, ME, USA, [3]Department of Geology, University of Cincinnati, Cincinnati, OH, USA, [4]Civil and Environmental Engineering, University of Maine, Orono, ME, USA, [5]GNS Science, Dunedin, New Zealand

Abstract Establishing the atmospheric expression of abrupt climate change during the last glacial termination is key to understanding driving mechanisms. In this paper, we present a new ^{14}C chronology of glacier behavior during late-glacial time from the Scottish Highlands, located close to the overturning region of the North Atlantic Ocean. Our results indicate that the last pulse of glaciation culminated between ~12.8 and ~12.6 ka, during the earliest part of the Younger Dryas stadial and as much as a millennium earlier than several recent estimates. Comparison of our results with existing minimum-limiting ^{14}C data also suggests that the subsequent deglaciation of Scotland was rapid and occurred during full stadial conditions in the North Atlantic. We attribute this pattern of ice recession to enhanced summertime melting, despite severely cool winters, and propose that relatively warm summers are a fundamental characteristic of North Atlantic stadials.

Plain Language Summary Geologic data reveal that Earth is capable of abrupt, high-magnitude changes in both temperature and precipitation that can occur well within a human lifespan. Exactly what causes these potentially catastrophic climate-change events, however, and their likelihood in the near future, remains frustratingly unclear due to uncertainty about how they are manifested on land and in the oceans. Our study sheds new light on the terrestrial impact of so-called "stadial" events in the North Atlantic region, a key area in abrupt climate change. We reconstructed the behavior of Scotland's last glaciers, which served as natural thermometers, to explore past changes in summertime temperature. Stadials have long been associated with extreme cooling of the North Atlantic and adjacent Europe and the most recent, the Younger Dryas stadial, is commonly invoked as an example of what might happen due to anthropogenic global warming. In contrast, our new glacial chronology suggests that the Younger Dryas was instead characterized by glacier retreat, which is indicative of climate warming. This finding is important because, rather than being defined by severe year-round cooling, it indicates that abrupt climate change is instead characterized by extreme seasonality in the North Atlantic region, with cold winters yet anomalously warm summers.

1. Introduction

Deciphering patterns and the precise timings of past climate events is fundamental to developing a robust understanding of the mechanisms, effects, and consequences of climate change. Terrestrial and marine paleoclimate proxies indicate that the transition from full-glacial to Holocene conditions was interrupted by the prominent Younger Dryas (YD) stadial or Greenland stadial 1 (Lowe et al., 2008), a climate shift in the North Atlantic region between 12.9 and 11.6 ka. This stadial, and others, was characterized by amplified seasonality (Buizert et al., 2014; Denton et al., 2005; Isarin et al., 1998) attributed to extensive winter sea ice across the northern North Atlantic Ocean (Brauer et al., 2008). Furthermore, accumulation data from Greenland ice cores suggest that high-magnitude transitions in mean climate state may have occurred as rapidly as years to decades (Alley et al., 1993; Steffensen et al., 2008). However, despite the abundant evidence for the occurrence of abrupt climate shifts during late-glacial time, important questions remain concerning the seasonality of those climatic episodes and thus their impact on terrestrial environments and ice masses.

We examine this issue by reconstructing the timing of the late-glacial resurgence of ice in Scotland, termed the Loch Lomond Readvance (LLR) and widely agreed upon as being a consequence of the YD stadial as registered with high chronological precision in the Greenland ice cores (Severinghaus et al., 1998;

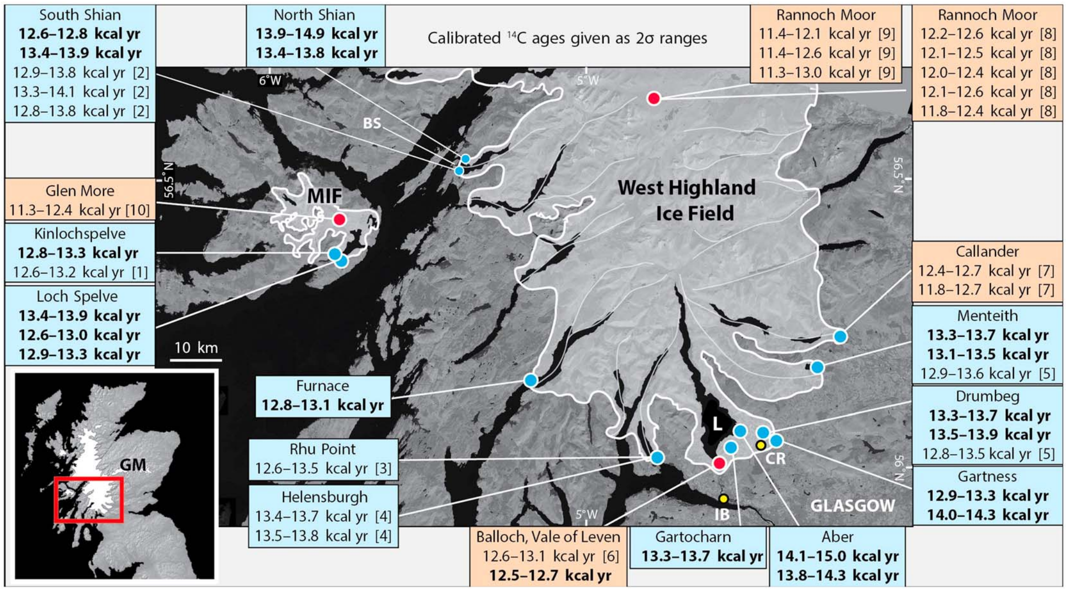

Figure 1. Former limits and radiocarbon chronology (in kcal yr B.P.) of the late-glacial (LLR) ice fields in the SW Scottish highlands. The blue circles represent sites from which maximum-limiting ^{14}C ages are derived, while the red circles indicate the locations of minimum-limiting ages, including the Balloch borehole in the Vale of Leven. Also shown are the locations of the Inverleven borehole (IB; Browne & Graham, 1981) and Croftamie (CR) (yellow dots), the Loch Lomond basin (L), and Balure of Shian (BS; Peacock et al., 1989). "GM" (inset) denotes Grampian Mountains. Dates are shown as calibrated age ranges (2σ) using the Marine13 and IntCal13 curves, as given in Tables 1, S1, and S2. New ages (this study) are shown in bold. Sources of previously published ages, recalculated here, are given in brackets as follows: [1] Gray & Brooks, 1972; [2] Peacock, 1971a; [3] Rose, 1980a; [4] Browne et al., 1983; [5] Sissons, 1967; [6] Browne & Graham, 1981; [7] Lowe, 1978; [8] Bromley et al., 2014; [9] Walker & Lowe, 1979; [10] Walker & Lowe, 1982. For Rannoch Moor, only the five oldest basal ages from [8] are included for brevity.

Steffensen et al., 2008). Because glaciers are influenced primarily by summer temperature (Denton et al., 2005; Oerlemans, 2001; Zemp et al., 2015), the geologic record of glaciation from Scotland affords a valuable opportunity to understand the seasonality of past abrupt climate change events in the North Atlantic and to place events there in a global context. We present a suite of 18 new ^{14}C dates from basal tills formed during the glacial resurgence, as well as one ^{14}C date from above the till, and discuss our data in view of other paleoclimatic evidence, with an aim toward elucidating the mechanics of abrupt climate change and its impact on terrestrial ice masses.

2. Geologic Setting of the Loch Lomond Readvance

A variety of glacial geomorphologic indictors, such as moraines, eskers, and drumlins, have been used to interpret the extent of the British-Irish ice sheet during the Last Glacial Maximum (LGM), and chronological constraints on the ice sheet history have been developed via a number of dating techniques (Clark et al., 2012). It is generally accepted that the ice sheet over Scotland progressively diminished following the LGM (27–19 ka; Clark et al., 2012), with the British and Irish ice sheets separating into individual entities by ~16,000 years ago (Clark et al., 2012). The subsequent history of post-LGM ice withdrawal in Scotland, and the late-glacial resurgence of ice there during the LLR, has been a matter of debate (Ballantyne, 2012, and references therein).

Despite considerable mapping efforts to delineate glacial-geomorphic features and sedimentologic characteristics, uncertainty remains about the limits of the LLR in parts of the Scottish Highlands, especially in relation to whether there was an extensive and relatively thick ice cap, or more restricted and thinner ice bodies (Gollege, 2010). This uncertainty underscores that the ice limits at the culmination of the LLR are not everywhere marked by prominent, definitive end moraines, or drift limits. The currently favored model is of an interconnected complex of ice fields, termed the West Highland ice field (Ballantyne, 2012). The southern half of this ice field (hereafter WHIF) is considered to have been nourished in the Grampian Mountains during the LLR, with the greatest thickness of ice located over Rannoch Moor (Figure 1). Beyond this central ice mass,

Table 1
Sample Details and Ages for Bracketing Marine Samples

Sample lab ID	Site	Context	Material	$\delta^{13}C$	Uncorrected ^{14}C age (^{14}C yr)	Calibrated 2σ age range (kcal yr B.P.)[a]	Calibrated 2σ age range (kcal yr B.P.)[b]
AA15940	Kinlochspelve	Max.	Astarte elliptica	1.9	11,621 ± 117	13.3–12.8	13.1–12.7
AA15941	Loch Spelve	Max.	Arctica islandica	1.6	12,167 ± 130	13.9–13.4	13.7–13.1
AA15942	Loch Spelve	Max.	*Astarte elliptica*	2.2	11,352 ± 92	13.0–12.6	12.8–12.5
AA15943	Loch Spelve	Max.	Nuculana pernula	0.1	11,668 ± 86	13.3–12.9	13.1–12.7
AA15945	South Shian	Max.	Astartes elliptica	1.9	11,192 ± 78	12.8–12.6	12.7–12.4
AA15946	South Shian	Max.	Arctica islandica	2.5	12,157 ± 120	13.9–13.4	13.6–13.1
AA15947	North Shian	Max.	Arctica islandica	2.7	12,700 ± 116	14.9–13.9	14.3–13.6
AA15948	North Shian	Max.	Unidentified shell fragments	1.3	12,179 ± 85	13.8–13.4	13.6–13.3
OS-2077	Furnace	Max.	Unidentified shell fragments	1.4	11,450 ± 45	13.1–12.8	12.8–12.7
AA15951	Gartocharn	Max.	Astarte elliptica	0.1	12,021 ± 89	13.7–13.3	13.4–13.1
AA15949	Aber	Max.	Unident. shell fragments	0.8	12,816 ± 86	15.0–14.1	14.6–13.8
AA15950	Aber	Max.	Unidentified shell fragments	0.6	12,528 ± 94	14.3–13.8	14.0–13.6
AA15952	Drumbeg	Max.	Unidentified shell fragments	−1.1	12,021 ± 89	13.7–13.3	13.4–13.1
OS-2076	Drumbeg	Max.	Chlamys islandicus	3.19	12,250 ± 50	13.9–13.5	13.6–13.4
AA15953	Gartness	Max.	Unidentified shell fragments	0.8	11,593 ± 79	13.3–12.9	13.0–12.7
OS-133096	Gartness	Max.	Astarte borealis	0.55	12,650 ± 35	14.3–14.0	14.0–13.8
AA15938	Menteith	Max.	Mytilus edulis fragments	0.8	12,058 ± 89	13.7–13.3	13.5–13.1
AA15939	Menteith	Max.	Unidentified shell fragments	3.0	11,843 ± 88	13.5–13.1	13.3–12.9
OS-2078	Balloch	Min.	Portlandia arctica	−0.2	11,050 ± 45	12.7–12.5	12.5–12.1

[a]Age calibrated according to the Marine13 reservoir curve (Reimer et al., 2013). [b]Age calibrated assuming a maximum 600-year reservoir correction (Bondevik et al., 2006).

smaller ice fields accumulated on the islands of Mull and Skye, while individual glaciers occupied high cirques and uplands throughout Scotland (Ballantyne, 1987, 1989; Sissons, 1977), northern England (Hughes et al., 2012; McDougall, 2001; Sissons, 1980; Wilson et al., 2013), Wales (Hughes, 2002; Lowe, 1994), and Ireland (Colhoun & Synge, 1980; Wilson, 2004). The detail of the existing geomorphic record has allowed the nature of deglaciation following the LLR to be examined closely: In places, recession involved the progressive retraction of actively flowing ice for much of this period, with glaciers depositing conspicuous moraines as their termini retreated to higher ground (Benn et al., 1992; Bennett & Boulton, 1993). Elsewhere, in situ stagnation is inferred from chaotic moraine topography.

The LLR has been attributed to cold conditions during the YD, yet direct chronologic constraints on the timing of the LLR and subsequent recession are sparse. Thus, the timings of the LLR—onset, maximum, and recession—are incompletely known (Ballantyne, 2012). Here we contribute 18 new ^{14}C ages from marine shell material incorporated into basal tills of tidewater outlet glaciers of the WHIF and the smaller Mull ice field (MIF), located ~20 km to the west (Figure 1). The particular significance of the shell material is that it provides maximum-limiting ages for the LLR (Table 1). In addition, we provide a new minimum-limiting age from near the terminus of the Lomond piedmont glacier, a southern outlet glacier of the WHIF. We then discuss the implications of this new data set in the context of existing minimum-limiting ^{14}C ages constraining the final deglaciation of Scotland.

Along much of its western and southwestern margins, the WHIF terminated in tidewater outlet glaciers (Ballantyne, 2012), as did the eastern margin of the MIF (Figure 1). Former ice limits in these fjords are marked by conspicuous moraines, deltas, and terraces, many of which contain reworked marine sediments and the shells of marine fauna typical to late-glacial Scotland (Gray & Brooks, 1972). As these organisms inhabited the fjords prior to glaciation, ^{14}C dates from their shells afford maximum-limiting ages for the advance of the WHIF and MIF to their full extents. However, although this approach has been employed since the 1960s (Gray & Brooks, 1972; Peacock, 1971a, Peacock et al., 1989; Rose, 1980b; Sissons, 1967), many of these results include large uncertainties. Moreover, ambiguity in the magnitude of the marine reservoir effect in this part of the North Atlantic has further complicated accurate determinations of shell ages (Sutherland, 1986). To help address these shortcomings, our new chronology is based entirely on multiple AMS measurements and considers the impact of a time-dependent marine reservoir correction in the eastern North Atlantic (Bondevik et al., 2006).

3. Methods and Results

3.1. Sample Collection

Shells and shell fragments were collected from 10 exposures in till and terminal moraines located around the southern and western margins of the former WHIF and the eastern edge of the MIF (Figure 1). Sites were chosen largely on the basis of favorable reports by previous researchers and include Kinlochspelve and Loch Spelve (Isle of Mull), South Shian and North Shian (Loch Creran), Furnace (Loch Fyne), Gartocharn (Loch Lomond), Aber (Loch Lomond), Drumbeg Quarry at Drymen (Loch Lomond), Gartness (Loch Lomond), and Lake of Menteith (upper Forth Valley). Additionally, we dated material recovered from the British Geological Survey borehole at Balloch in the Vale of Leven (Figure 1). With the exception of the borehole data, all samples are derived from basal tills and associated sediments and thus afford maximum-limiting ages for glacier advance and moraine deposition. In contrast, the Balloch age is from shelly material overlying till and, as detailed in section 3.4, represents a minimum age for the LLR. Details of each site are given below.

3.1.1. Loch Spelve

On Mull, an east-flowing outlet of the MIF advanced through Loch Spelve and deposited a series of terminal moraines within ~50 m of the current shoreline (Figure 1; Ballantyne, 2002). Previously, Gray and Brooks (1972) dated shell fragments collected from an exposure in a terminal moraine near Kinlochspelve Farm (56.36871°, −5.79811°; sample I-5308; Table S1 in the supporting information), on the northwest side of the loch. We revisited the site, which comprises ~2.5 m of clay-rich till overlain by sands and gravels, and collected shell and barnacle fragments from the till to provide further maximum-limiting age constraint for the moraine (sample AA15940; Table 1). We also visited a 2-m-high till exposure on the south side of Loch Spelve (56.36498°, −5.78563°) where stream erosion has incised the LLR moraine to reveal a stony, shell-rich till overlain by 1 m of beach sediments (Gray, 1977). We collected three samples (AA15941, 15942, and 15943; Table 1) of shell fragments from within the till unit.

3.1.2. South and North Shian

We collected shell samples from two sites at the mouth of Loch Creran (Figure 1), which was occupied during the LLR by a WHIF outlet glacier (McCann, 1966; Peacock, 1971b; Peacock et al., 1989; Synge, 1966). At South Shian (Figure 1; 56.52516°, −5.40181°), part of a terminal moraine complex has been incised by coastal erosion to reveal a clayey till with abundant stones, disarticulated shells, and shell fragments. The till is overlain in turn by sands and gravels. This stratigraphy was reported by Peacock (1971a) and Peacock et al. (1989), who also dated shell fragments from the till to provide a maximum age for the advance (samples IGS C14/16–18; Table S1). In our study, we analyzed shells (samples AA15945-6; Table 1) from the basal meter of the same unit. Additionally, we collected shell fragments from clayey till exposed in a meter-high section on the shoreline at North Shian (Figure 1; 56.53452°, −5.38952°).

3.1.3. Furnace

Along its southern margin, the WHIF was drained via a number of large outlet glaciers, one of which occupied Loch Fyne (Sutherland, 1981; Figure 1). Near the former terminus of the Fyne glacier, LLR till is exposed in a 1.2-m-high shoreline section immediately south of Furnace (56.14978°, −5.19265°). First described by Sutherland (Sutherland, 1981), the till comprises a gray, clay/silt-rich diamicton containing pebbles, cobbles, and shell fragments. We collected sample OS-2077 (Table 1) from this unit.

3.1.4. Loch Lomond Basin—Gartocharn, Aber, Drumbeg, and Gartness

South of Loch Lomond, we collected samples from four sites inside the terminal limits of the Lomond outlet glacier (Figure 1). At Gartocharn, which gives its name to the local till of LLR age, a compact red clay-rich basal till is exposed in an ~3-m-high stream cut on the village outskirts (56.03916°, −4.53051°). Sample AA15951 (Table 1) comprised a complete *Astarte elliptica* bivalve from the basal meter of this till unit. Farther north at Aber (56.05641°, −4.25795°), basal till outcrops in section along the shore of Loch Lomond, where it comprises a highly compacted, dark-brown diamicton overlain by sandier till (Rose, 1981). The lower unit contains abundant striated clasts and broken shells derived from the underlying marine deposits (e.g., Rose, 1981). Samples AA15949–50 (Table 1) were collected from this layer near the base of a 1.2-m-high exposure. East of Drymen, the Drumbeg Quarry (Figure 1; 56.05729°, −4.43722°) forms a large pit in a terminal moraine of the outlet glacier. First described by Sissons (1967), deformed foreset sands and gravels overlain by basal till are exposed in the ~12-m-high quarry wall and indicate glaciofluvial deposition in standing water, prior to overriding by the Lomond glacier (Evans & Wilson, 2006). Samples AA15952 and OS-2076 are from the shell-rich sand and gravel unit and provide maximum ages for the overlying till (Table 1).

Two kilometers southeast of Drumbeg Quarry, where Endrick Water has incised the outer terminal moraine of the Lomond outlet glacier, several meters of Late Quaternary deposits are exposed in section at Gartness (Figure 1; 56.04713°, −4.41240°) and have been described in detail by Rose (1980a, 1981). We collected shells (samples AA15953 and OS-133096; Table 1) from the red basal till, which can be seen overlying the lacustrine sediments of glacial Lake Blane (Rose, 1980a) and the underlying Clyde Beds (Figure S1 in the supporting information), a widespread unit of marine clays representing a period of higher sea level following the LGM (Peacock, 1981).

3.1.5. Lake of Menteith

East of the Loch Lomond basin, a southeast-flowing outlet of the WHIF deposited a conspicuous terminal moraine complex in the western Forth River valley (Figure 1). At Lake of Menteith, the basal till and overlying sediments are exposed in an ~10-m-high lake-shore exposure (56.17312°, −4.27452°; Simpson, 1933; Sissons, 1967; Table S1). The till, which is considerably faulted and distorted due to glacial tectonism, comprises a clay-rich, shell-bearing diamicton overlain by sands and gravels. We collected two samples of shell fragments and complete valves (AA15938–9; Table 1) from the basal 2 m of this till unit.

3.1.6. Balloch Borehole, Vale of Leven

In the Vale of Leven, the ~50-m deep Balloch borehole was extracted from a site 2 km inside the terminal moraine of the Lomond outlet glacier (Figure 1). The borehole stratigraphy includes a basal till overlain by ~40 m of water-lain clays and silts capped by 10 m of sand, gravel, and terrestrial sediments (Browne & Graham, 1981). We dated shells of *Portlandia arctica* that are present in the lower 1.5 m of clay but absent from the overlying ~35 m of water-lain sediments. Previously, Browne and Graham (1981) reported a radiometric age from barnacles on cobbles atop the basal till (sample SRR1530; Table S2).

3.2. Sample Preparation, Measurement, and Correction for Marine-Reservoir Effects

Samples comprised pristine shell material (Figure 2), frequently exhibiting intact periostracum and with no indication of postdepositional modification, such as secondary calcite precipitation. Shells first were cleaned with deionized water and dried in a low-temperature oven at the University of Maine, before being submitted to either the NOSAMS Facility (Woods Hole) or the University of Arizona AMS Facility for ^{14}C and $\delta^{13}C$ measurement.

To account for the marine reservoir effect, we calibrated our ^{14}C ages using the Marine13 global reservoir curve (Reimer et al., 2013), acknowledging that the precise magnitude of the marine correction in Scotland has not been determined independently. While this calibration method refines the assumption of a uniform 400-year late-glacial marine reservoir effect (Harkness, 1983), we recognize the potential for uncertainty due to local variations in reservoir effect, such as might arise due to influx of meltwater from the advancing ice margins. Under that scenario, the magnitude of the local reservoir effect would decrease, thereby increasing our ^{14}C age determinations. To test the impact of a time-variable reservoir effect, we also calculate our ages assuming a uniform correction of 600 years, which is the maximum reservoir effect reported from western Norway (60–62°N) during the late glacial (see Bondevik et al., 2006). We note that this first-order approach provides a conservative estimate, since the maximum correction applies to the interval 12.6–11.6 ka (Bondevik et al., 2006), which is younger than most of our ages. Nonetheless, we consider that this approach provides a conservative lower limit for our age calculations. Radiocarbon ages, along with reservoir-corrected, calibrated (calendar years before 1950) age ranges (2σ), are given in Table 1 and shown in Figures 1 and S2 (plotted via OxCal v4.2.4 (Bronk Ramsey, 2013)). Values for $\delta^{13}C$ are given in Table 1.

3.3. Maximum-Limiting ^{14}C Ages

Calibrated ages for shells from the glacial tills give a distribution with a 2σ interval of 15.0–12.6 kcal yr B.P. using the Marine13 correction and 14.6–12.4 kcal yr B.P. with the 600-year correction (Table 1). Such broad age ranges are typical of maximum-limiting data sets (Lowell et al., 1999; Strelin et al., 2011), since advancing glaciers incorporate both recently living and long-dead material into their basal tills. Upon recalibration with our two marine-reservoir corrections, nine previously published ^{14}C dates (Browne et al., 1983; Gray & Brooks, 1972; Peacock, 1971a; Peacock et al., 1989; Rose, 1980b; Sissons, 1967) from shells in tills deposited by the WHIF and MIF give 2σ intervals of 14.1–12.6 kcal yr B.P. (Marine13 correction) and 13.8–12.2 kcal yr B.P. (600-year correction; Table S1). While the older end of this range is as much ~900 years younger than that of our new data set, the younger end is in close agreement with our data.

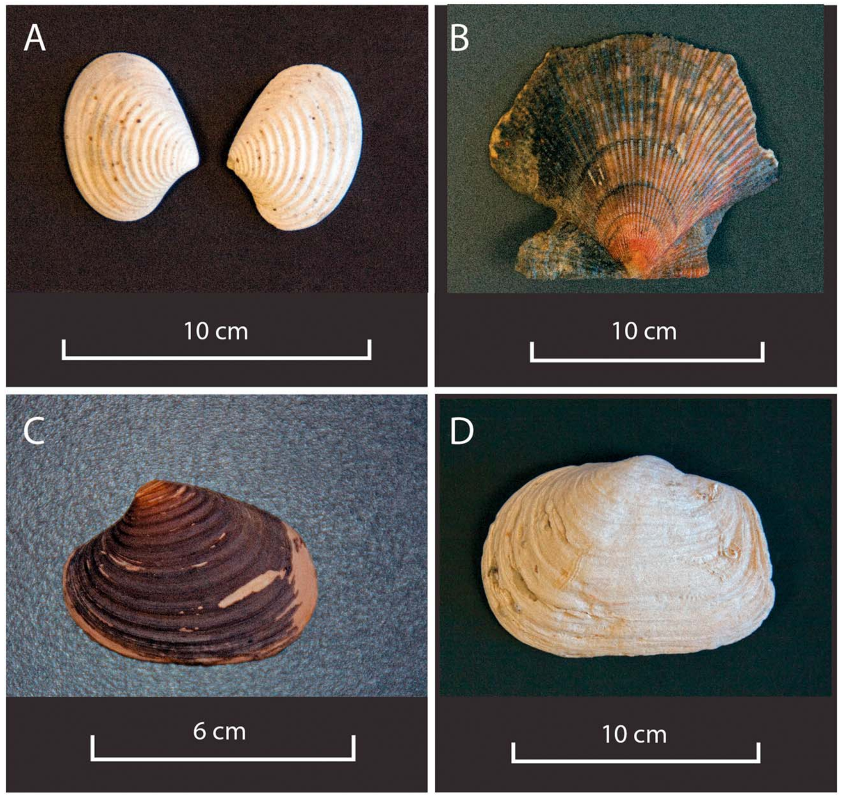

Figure 2. Examples of well-preserved shells collected from Loch Lomond Readvance basal tills. (a) *Astarte elliptica*. (b) *Chlamys islandica*. (c) *Astarte elliptica* valve, with intact periostracum. (d) *Portlandia arctica*.

Because of this high internal consistency and their comparable stratigraphic contexts, we combined these two maximum-limiting data sets (ours and previously published) to provide a more comprehensive bracket for the timing of the culmination of the LLR. This collection of 27 ^{14}C ages was calibrated first with the Marine13 correction and second with the 600-year correction to produce two specific age groups. We then generated summed probability curves for each group to identify their respective 95% confidence intervals, which constitute the young tails of the distribution curves. According to the Marine13-corrected probability curve (Figure S3a), the youngest probable age in the maximum-limiting data set is 12.7 kcal yr B.P., whereas the 600-year correction gives a slightly younger age of 12.8 kcal yr B.P. at 95% confidence (Figure S3b). We reemphasize that these ages represent the timing of mollusk growth and thus that the ice fields reached their maximum extent thereafter. Moreover, the relatively broad spatial distribution of our samples renders these first-order estimates only. Nonetheless, the close agreement between the two maximum-limiting distributions indicates that our interpretations are independent of correction scheme. Henceforth, we discuss maximum-limiting ages as calculated using the Marine13 curve.

3.4. The Balloch Borehole: ^{14}C Data and Stratigraphic Implications

In the Balloch borehole, sample OS-2078 comprised fragments of *P. arctica* collected from glaciomarine sediments at a depth of 48.9 m. This sample gives a calibrated 2σ age range of 12.7–12.5 kcal yr B.P. (Marine13; or 12.5–12.1 kcal yr B.P. with 600-year correction; Table 1) and is in stratigraphic order with a conventional radiocarbon age of 13.1–12.6 kcal yr B.P. (Marine13; 13.0–12.4 kcal yr B.P. with 600-year correction) from barnacles found deeper in the core, on gravel at 50.5 m depth (sample SRR1530; Browne & Graham, 1981; Table S2). In their original interpretation of the Balloch stratigraphy, Browne and Graham (1981) correlated the basal till with the British-Irish ice sheet, not based on characteristics of the till itself but on their assumption that the overlying clays represent the post-LGM Clyde Beds. If correct, the radiocarbon-dated barnacles from the till surface would serve as a maximum-limiting age for both the water-lain deposits and the LLR, when the Lomond outlet glacier of the WHIF overran the site. However, several lines of evidence lead us to conclude that this model is incorrect and that Balloch has not been glacially overridden since deposition of the basal till, in which case the ^{14}C data do not provide maximum-limiting age control for the LLR.

First, there is no till overlying the water-lain clays, which is surprising considering the voluminous late-glacial deposits elsewhere in the southern Loch Lomond basin, including the conspicuous terminal moraine complex (up to 7 m tall and 40 m wide). Browne and Graham (1981) suggested that the till was removed by wave action following the LLR maximum, yet even this model would be expected to leave a measurable stony lag deposit (similar to the Gartness site described by Rose, 1981) and noticeable unconformity. In addition to the absence of till, the water-lain sediments themselves exhibit neither a fissile texture nor deformation structures, both of which would be expected in overridden sediments (Phillips et al., 2007).

Alternatively, Browne and Graham (1981) suggested that the glacier terminus was floating and thus did not deposit a till. As floating ice margins are largely restricted to polar settings, where ice is below the pressure-melting point and has sufficient tensile strength to resist calving (Powell, 1984), this scenario is unlikely. Moreover, where ice shelves do exist, melt-out of englacial and supraglacial material produces undermelt diamicton (Gravenor et al., 1984), which would be preserved in the sedimentary record. Finally, a floating ice front at Balloch is incompatible with deposition of the prominent LLR moraine system immediately down-valley.

Second, the geotechnical data presented as evidence for glacial overriding of the borehole site (Browne & Graham, 1981) are inconsistent with overconsolidation. For one, undrained shear strength values from the basal ~20 m (~100 kN/m^2) show that these are normally consolidated, buoyant sediments (Terzaghi et al., 1986); hence, their close alignment with normally consolidated glaciomarine clays at nearby Inverleven (Browne & Graham, 1981; Figure S4). Had Balloch been overridden, the basal clays would exhibit consistently higher shear strength values typical of overconsolidation, as in profiles of Boston Blue Clay (Ladd & Foott, 1974; Figure S4), since any consolidation must be transmitted the full depth of the profile (Terzaghi et al., 1986). Specifically, for the dimensions of the former Lomond outlet glacier (several kilometers wide in this area), whatever pressure was exerted on the surficial sediments was also imposed on the sediments at 50 m depth. We calculate that for a glacier thickness of 50 m (a conservative minimum at Balloch) and an ice unit weight of 9 kN/m3, the entire sediment column from 0 to 60 m depth would be stressed an additional 450 kPa. The sediment for the full depth would gain about 110 kPa in shear strength from the ice overburden (the strengths would be offset 110 kPa to the right of the normally consolidated dashed line in Figure S4). As this is not the case in Figure S4, we conclude that there is no evidence for overriding at Balloch.

Where sediments do exhibit overconsolidation (e.g., due to desiccation and overriding), shear strength values tend to be uniformly higher (Bjerrum, 1973; Figure S4). Yet in the Balloch profile, values above ~37 m depth are highly disparate (~60–700 kN/m^2; Figure S4), at face value, implying that normally consolidated sediments are interbedded with highly overconsolidated sediments. Since that scenario is implausible without artesian pressure, the incongruent values in the Balloch borehole most likely represent erroneous shear strength measurements. Indeed, it is well documented that vane shear tests can be affected by the occurrence of shells, stones, sand layers, and varves in the sediment profile (Holtz et al., 2011; Wilson, 1964), any of which might explain the alternating high and low shear strength values in the Balloch clays.

We prefer a more straightforward interpretation of the Balloch borehole stratigraphy, in which the basal till corresponds to the Lomond outlet of the WHIF. Subsequently, during deglaciation, the overlying clays and silts were deposited initially in a marine environment (as indicated by *P. arctica*) and then under lacustrine conditions as the basin became isolated from exchange with marine waters. This scenario agrees with evidence for marine conditions in the Loch Lomond basin immediately following the LLR (Browne & McMillan, 1989). If true, the Balloch samples constitute minimum-limiting constraint for the LLR in the Loch Lomond basin and show the Lomond glacier attained its full extent and began retreating as early as ~13.0 kcal yr B.P.

3.5. Minimum-Limiting ^{14}C Ages

To test how representative this model is of the broader retreat following the LLR, we applied the same approach to published terrestrial minimum-limiting ^{14}C data as for the maximum-limiting ages in order to identify the closest probable age for the onset of deglaciation. First, we calibrated terrestrial minimum-limiting ages for the WHIF and MIF using IntCal13 (Table S2), noting (i) that this compilation includes data from sites close to and far inland from the LLR limits and (ii) that the latter are more numerous than the former. Both factors would make our results generally younger than the initial age of retreat. We generated a

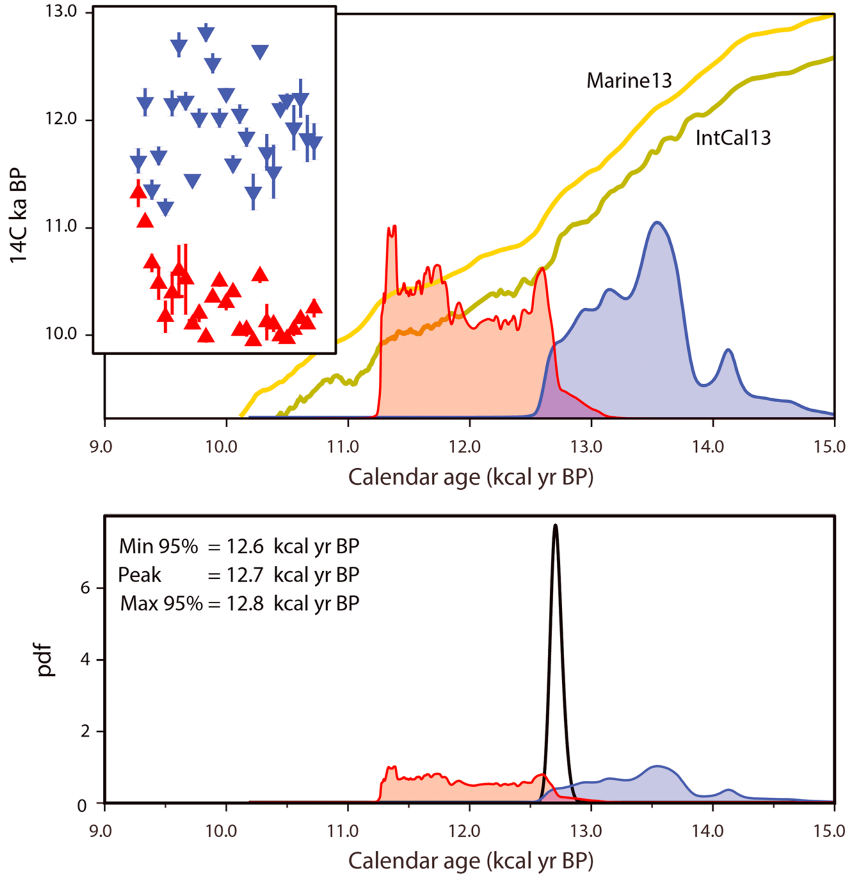

Figure 3. Probability distribution functions (PDFs) for maximum- and minimum-limiting age populations. (top) Maximum- (blue) and minimum-limiting (red) ages for the Loch Lomond Readvance (LLR) plotted as individual ^{14}C dates (inset) and as summed probability distributions of calibrated ages, showing both the IntCal13 (green) and Marine13 (yellow) calibration curves. (bottom) PDF of the period between maximum- and minimum-limiting populations, showing the most probable period for the culmination of the LLR as being 12.8–12.6 kcal yr B.P.

summed probability curve for the population (Figure S3c), the 95% confidence interval of which gives an oldest probable age of 12.7 kcal yr B.P. Although we include the two ^{14}C ages from the Balloch borehole in this assessment, as shown in Figure S3d, the 95% confidence interval for this minimum-limiting data set is minimally affected by the inclusion or exclusion of the Balloch ages, owing to the large number of terrestrial samples (*n* = 27) relative to marine ages (*n* = 2).

3.6. Probability Distribution Functions for Maximum- and Minimum-Limiting Ages

To identify the statistically most probable timeframe for the culmination of the LLR using the available ^{14}C data, we calculated the probability distribution function of the interval between maximum- and minimum-limiting populations (Figure 3). This calculation yielded minimum, best, and maximum estimates, whereby the minimum and maximum are at the 95% confidence interval of the probability distribution function. Further details on this treatment can be found in Kelly et al. (2015). Our results show that the two data sets form discrete populations with minimal overlap and give a probable age range for the peak of the LLR of 12.8–12.6 kcal yr B.P. (Figure 3). This range shifts to 12.7–12.5 kcal yr B.P. if one applies a 600-year reservoir correction (Figure S5).

4. Discussion

4.1. A Maximum-Limiting Age for the LLR

Our radiocarbon data set provides new maximum constraint for the LLR along the southwestern margins of the WHIF and on Mull, as well as additional constraint for the subsequent onset of deglaciation along the southern margin of the WHIF. Climatically, the spread of ages indicates that mollusk growth—and thus ice-free conditions—persisted at these sites for much of the Bølling-Allerød/Greenland Interstadial 1 (GI-1, 14.7–13.1 ka: Lowe et al., 2008), a scenario that is supported by previously published maximum-limiting ages from the same region (Table S1). Due to local variations in topography, glacier hypsometry, and climate, specific outlet glaciers and individual ice fields may not have reached their respective late-glacial maxima in lockstep (Golledge et al., 2008). Such considerations are particularly relevant in view of the scale difference between the MIF (~145 km^2) and WHIF (~9,500 km^2). However, by combining these maximum-limiting data sets (Figure 3), we propose that it creates a working estimate for the maximum age of the LLR. Additionally, the minor difference between our statistical estimates suggests that the timing of late-glacial events in western Scotland probably was relatively uniform on account of the short distances involved (Figure 1) and the dominant maritime climate. This scenario is supported by the consistent response of circum-North Atlantic glaciers—both maritime and continental, terrestrial, or tidewater—to modern warming (Dyurgerov & Meier, 2000; Zemp et al., 2015).

A conspicuous feature of the large maximum-limiting data set is the absence of ages younger than ~12.7 kcal yr B.P., suggesting that the margins of the WHIF and MIF had ceased advancing over extant marine sediments by that time and thus were at or close to their outer moraines. An alternative explanation—that the dearth of younger ages reflects a reduction in marine deposition due to sea level lowering (Peacock, 1971a)—is improbable considering the large number and variable depths of the (paleo) fjord sites investigated. Following the LLR, the minimum-limiting ages presented here suggest that deglaciation was underway by 12.6 kcal yr B.P. (Figure 3). The distributions of both maximum- and minimum-limiting data used in this assessment are also shown in Figure 4.

A possible exception to our interpretation is a shell chronology from Balure of Shian (Figure 1) presented as maximum-limiting age constraint for the Creran outlet glacier (Peacock et al., 1989). There, ages on shells from raised marine sediments range from 12.3 ± 0.3 to 11.4 ± 0.1 kcal yr B.P. (as computed by the approach here). Because Peacock et al. (1989) interpreted these marine sediments as having been overridden by glacier ice, they concluded that the Creran glacier attained its maximum extent as late as ~11.6 kcal yr B.P. However, considering the dearth of evidence for glacial overriding (e.g., no till or lag deposits atop the marine sediments) and the lack of an unequivocal terminal moraine complex distal to the site (McCann, 1966; Synge, 1966), an alternative possibility is that the Creran outlet glacier did not reach Balure of Shian during late-glacial time. This scenario is supported by minimum-limiting ^{14}C data from farther inland (Figure 1), indicating that area was largely ice-free by 12.5 kcal yr B.P. (Bromley et al., 2014; Lowe & Walker, 1976; Walker & Lowe, 1977, 1979).

4.2. Bracketing the Culmination of the Loch Lomond Readvance

Our statistical treatment of both maximum- and minimum-limiting ^{14}C populations suggests that the most probable window of time for the culmination of the LLR was 12.8–12.6 kcal yr B.P. (Figure 3), a scenario that is also evident from the individual age populations (Figure 4). Although these are bracketing estimates only, they are consistent with several earlier assessments, suggesting that post-LLR deglaciation in Scotland was well underway during the first half of the YD chron (Bowen, 1999; Bromley et al., 2014; Dawson et al., 1987; Golledge et al., 2007; Lowe, 1978; Lowe & Walker, 1976, 1980, 1981, 1984; Tipping, 1985; Walker & Lowe, 1977, 1979, 1981, 982). We note, however, that this model does not align with the currently prevailing hypothesis in which the LLR culminated late in the YD, or earliest Holocene, in response to the termination of stadial conditions (MacLeod et al., 2011; Rose et al., 1988). Specifically, ^{14}C dates of plant fragments from beneath lacustrine sediments at Croftamie, in the southern Loch Lomond basin (Figure 1), have been interpreted as maximum ages for an overlying thin diamict (termed "Gartocharn till") correlated with the LLR, placing the late-glacial maximum as late as ~11.6 ka, at the YD-Holocene transition (Figure 4; MacLeod et al., 2011).

A first impression of these apparently incongruent scenarios is that either the minimum-limiting data set discussed above, including deglacial ages from Rannoch Moor, Callander, and Glen More, is unreliable or that the Croftamie ^{14}C ages are somehow incorrect. An alternative interpretation, which we explore below (section 4.3), is that the ^{14}C dates from the Croftamie section instead provide a minimum-limiting age for the LLR and are thus morphostratigraphically compatible with the broader ^{14}C chronology presented here.

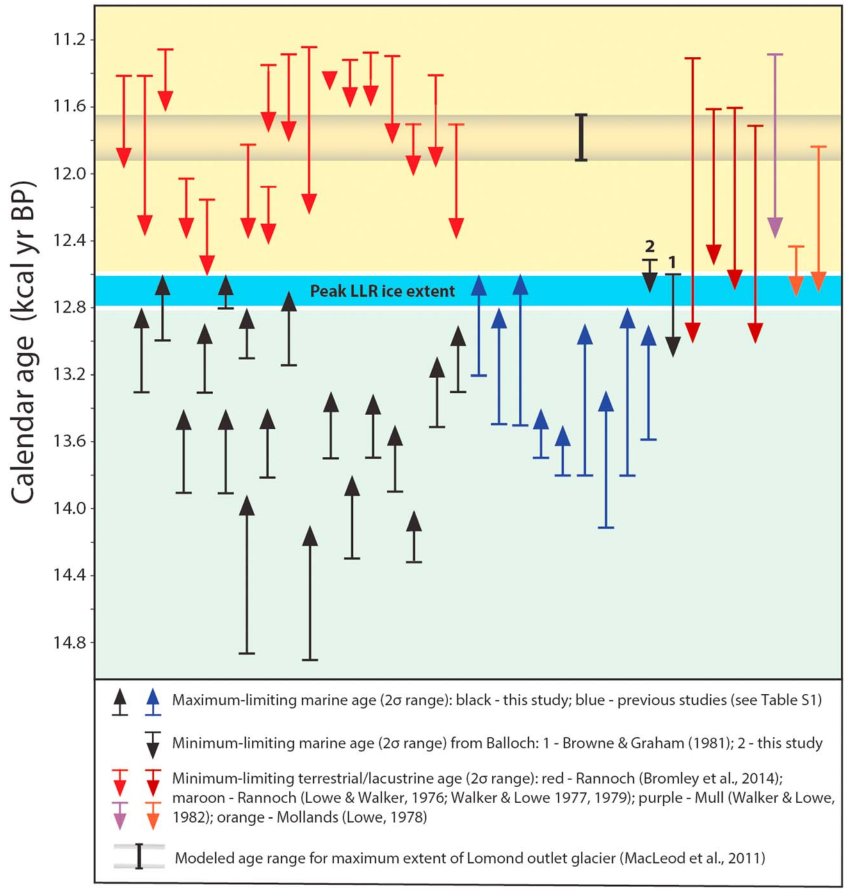

Figure 4. Relative distribution of maximum- and minimum-limiting ^{14}C ages (in calibrated years) for the Loch Lomond Readvance (LLR). All marine dates, both published and new, have been corrected via the Marine13 calibration curve and are shown with 2σ uncertainty. Horizontal blue band represents the 95% confidence interval of the pdf, which these data suggest is the most likely period for the peak of the LLR. If a maximum 600-yr reservoir correction is used instead, this window becomes slightly (~100) years narrower. For comparison, the modeled age range presented by MacLeod et al. (2011) for the maximum extent of the Loch Lomond glacier at Croftamie is shown by the light gray horizontal band.

4.3. An Alternative Hypothesis for the Croftamie Site

The detailed reports of the Croftamie stratigraphy describe a red-colored basal till, attributed to the ice sheet, overlain by a second (WHIF) till, with several centimeters of lacustrine sediments separating the two (MacLeod et al., 2011; Rose, 1981; Rose et al., 1988). By this model, the silts were deposited in a proglacial lake (Lake Blane) that inundated the area as the advancing Lomond outlet glacier blocked local drainage during the LLR, prior to overriding the site. A second deposit of lacustrine silts caps the section (Rose et al., 1988). Finally, a suite of tightly clustered ^{14}C ages from the lower lacustrine unit is interpreted as maximum-limiting constraint for the LLR (MacLeod et al., 2011; Rose et al., 1988). Thus, Croftamie has been described as the type locality for the LLR (Coope & Rose, 2008).

Important questions remain, however, concerning the relationship between the sediments themselves and the events that produced them. Central among these are (1) age control for the basal till, (2) potential stratigraphic discrepancies between Croftamie and nearby sites, and (3) the dearth of sedimentologic structures unequivocally indicative of glacial overriding. With regard to the basal till, an LGM origin for this unit is inferred on the basis of structure, fabric, and composition (Rose et al., 1988). However, because it has not yet been dated directly, there remains the possibility that the till does not correspond to the British-Irish ice sheet but to a later event. Direct age constraint from the till is vital to help resolve this issue.

Second, while we acknowledge that the absence of evidence is not evidence for absence, the lack of marine deposits (i.e., the Clyde Beds) at Croftamie, when they are widely preserved throughout the southern Loch

Lomond basin, nevertheless is potentially significant. At Gartness, located only 2.7 km from Croftamie (Figure 1) and within the same depositional context (i.e., post-LGM marine transgression followed by Lake Blane submergence and LLR glaciation; Rose, 1981), the basal ice sheet till is overlain by several meters of Clyde Beds sediments. This unit is capped in turn by red-colored basal till of the Lomond outlet glacier (Rose, 1980a, 1981), which we note is compositionally identical to the so-called Wilderness till at Croftamie. The Clyde Beds also outcrop beneath late-glacial till at Aber. In contrast, the basal till and overlying diamicton at Croftamie, which sits at a similar elevation to Gartness, are separated by a few centimeters of glaciolacustrine sediments (Rose et al., 1988). One explanation for this disparity is that the Clyde Beds were somehow scoured from Croftamie—but not elsewhere—by wave action during the rise of Lake Blane (Rose et al., 1988).

An alternative hypothesis is that the basal till at Croftamie was not deposited by the British-Irish ice sheet but by the Lomond outlet glacier during the LLR and that the overlying water-lain sediments (Rose et al., 1988) were deposited in a lacustrine setting after glacier retreat. This scenario would account for the absence at Croftamie both of the Clyde Beds and of deformation structures (e.g., folds and faults), fissile texture, or evidence of overconsolidation in the water-lain sediments (Phillips et al., 2007) indicative of glacial overriding, as well as the lack of a stony lag deposit atop the basal till suggestive of wave erosion (e.g., Rose, 1980b, 1981). Thus, rather than a lodgement till, we suggest that the thin upper diamicton is instead postglacial colluvium emplaced by slumping of the adjacent drumlin at Croftamie. In this case, the suite of ^{14}C dates from the organic layer would provide minimum, rather than maximum, age constraint for the LLR and would align with the broader radiocarbon chronology for the LLR (Figure 4).

4.4. Estimating the Timing of Deglaciation

By comparing our maximum- and minimum-limiting ^{14}C ages for the WHIF, we can broadly approximate the duration of deglaciation following the LLR. For example, a period of ~300 years separates the youngest probable age (95% confidence) for tills associated with the ice margins from the oldest replicable minimum ages from Rannoch Moor (~12.5 kcal yr B.P.; Bromley et al., 2014), which we take as representing substantial deglaciation of the WHIF. Acknowledging that Rannoch Moor probably was not the final place to become ice free (shrinking glaciers may have persisted throughout the YD in high cirques) and that this site did not supply all the outlet glaciers (e.g., the Lomond glacier) represented by the current ^{14}C data set, we consider that this rate of deglaciation is plausible in light of the rapid recession reported from other midlatitude ice masses (Hall et al., 2013; Moreno et al., 2015; Putnam et al., 2013; Ravazzi et al., 2014; Schlüchter, 1988).

4.5. Broader Climatic Implications

The new ^{14}C ages presented here, in conjunction with recalculated existing data, suggest that the LLR culminated early in the YD stadial and that subsequent deglaciation of the Scottish Highlands was rapid. While we acknowledge that this interpretation contrasts with the classic view of the YD as a period of renewed glacial advance in the British Isles (Phillips et al., 2016), it is not incompatible with findings reported from other sites in the circum-North Atlantic (Figure 5). Indeed, in Norway, which experiences a similar Atlantic-maritime climate to Scotland, Andersen et al. (1995) described numerous moraine systems that, within the resolution of the radiocarbon data, document advances culminating during the Allerød/earliest YD and net retreat during the ensuing stadial. Examples include Oslofjord (59°N), where the late-YD Ås and Ski moraines lie as much as ~40 km inboard of the early-YD Eidert moraine, as well as moraine chronologies from Nordfjord (62°N), Tautra (63°N), Nordland (66°N), and Troms (69°N; Andersen et al., 1995, and references therein). Supporting that model, a reconstruction of relative sea level in western Norway (Lohn et al., 2007) showed that the late-glacial buildup of the Scandinavian ice sheet commenced during the early Allerød (~13.6 ka). Closer to home, Walker et al. (2003) allowed for a similar scenario of interglacial ice buildup in Wales, based on paleoecologic and geochemical proxies, while a recent ^{10}Be moraine data set from northern England suggests that cirque glaciers were more extensive during the Allerød than during the YD (Hughes et al., 2012).

Farther north, a recent ^{10}Be deglacial record from East Greenland (Levy et al., 2016) joins a growing body of glacial-geologic data (Dyke et al., 2014; Hall et al., 2008, 2010; Jennings et al., 2006, 2014; O'Cofaigh et al., 2013; Rinterknecht et al., 2014) indicating recession of both the Greenland Ice Sheet and local ice masses during the YD, while glaciers on Svalbard (Mangerud & Landvik, 2007) and Baffin Island (Briner et al., 2009) were smaller during the YD than during the Little Ice Age. Even the Laurentide ice sheet experienced conspicuous

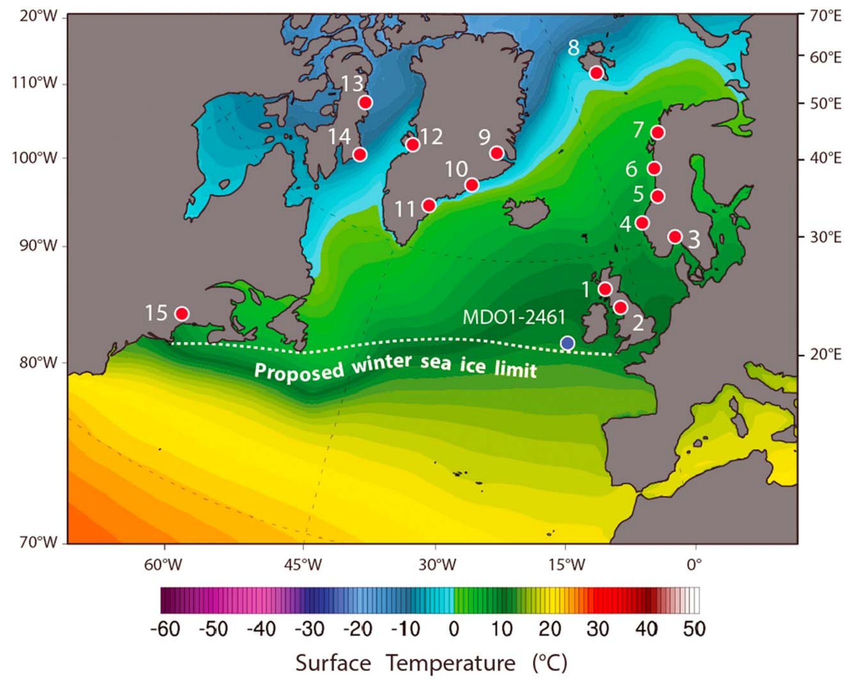

Figure 5. Supporting North Atlantic glacial and marine chronologies. The numbered red circles represent sites discussed in the text that either underwent Allerød-early Younger Dryas (YD) advance and subsequent stadial retreat or where glaciers were less extensive during the YD than during the Little Ice Age: 1 – Scotland (this study); 2 – English Lake District (Hughes et al., 2012); Oslofjord, 4 – Nordfjord, 5 – Tautra, 6 – Nordland, and 7 – Troms (Andersen et al., 1995, and references therein); 8 – Spitzbergen (Mangerud & Landvik, 2007); 9 – Scoresby Sund (Hall et al., 2008; Levy et al., 2016); 10 – Kangerdlugssuaq (Dyke et al., 2014; Jennings et al., 2006); 11 – Bernstorffs Fjord (Dyke et al., 2014); 12 – Disko Bay region (Jennings et al., 2014; O'Cofaigh et al., 2013; Rinterknecht et al., 2014); 13 and 14 – Baffin Island (Briner et al., 2009); 15 – New England (Bromley et al., 2015). The blue circle indicates site of core MD01-2461 (Peck et al., 2008). Also shown is the proposed YD winter sea ice limit in the North Atlantic Ocean (Isarin et al., 1998), based on model simulations. Mean annual temperature (1979 to present) at 2 m above surface (University of Maine Climate Reanalyzer, www.cci-reanalyzer.org).

advances of its southern margins during the Allerød/GI-1 (Bromley et al., 2015; Lowell et al., 1999; Thompson et al., 2017), after which retreat commenced.

An early-YD age for the late-glacial maximum in Scotland places important limits on the timing of the LLR itself. While model simulations suggest the WHIF and surrounding ice masses could have accumulated rapidly at the onset of YD cooling (Golledge et al., 2008), that model was forced with a scaled version of the GRIP δ^{18}O-inferred temperature record and so is not an independent indicator of glacier behavior, particularly in light of a recent assessment (Buizert et al., 2014) of the Greenland isotope record as reflecting wintertime temperature. Therefore, unless it can be shown explicitly that the Scottish ice fields accumulated, peaked, and deglaciated all within the first few centuries of the YD, we suggest a simpler model in which the LLR largely predates the YD stadial.

If truly representative of cryospheric behavior in the North Atlantic, this emerging picture begs the question: What caused glaciers to advance during apparently mild interstadial conditions and decay during stadial conditions? Regarding the latter, one possibility is that the extreme seasonality characteristic of North Atlantic stadials, due to weakened Atlantic meridional overturning circulation (McManus et al., 2004) and expanded winter sea ice (Li et al., 2005; Figure 5), generated severe cold during YD winters but not during summer (Buizert et al., 2014; Denton et al., 2005). This process could reduce mean annual air (Severinghaus et al., 1998; Stuiver & Grootes, 2000; Figure 6) and sea-surface temperatures (Bond et al., 1993) throughout the circum-North Atlantic and may even perturb terrestrial ecosystems (Atkinson et al., 1987; Brooks & Birks, 2001; Isarin & Bohncke, 1997; Peteet et al., 1993) through permafrost expansion and the later onset of spring but would not impact adjacent glaciers since mass balance is controlled primarily by summer melting (Oerlemans, 2001; Zemp et al., 2015). Instead, we posit that glaciers in Scotland, and potentially elsewhere around the North Atlantic basin, retreated during the YD stadial due to such processes as

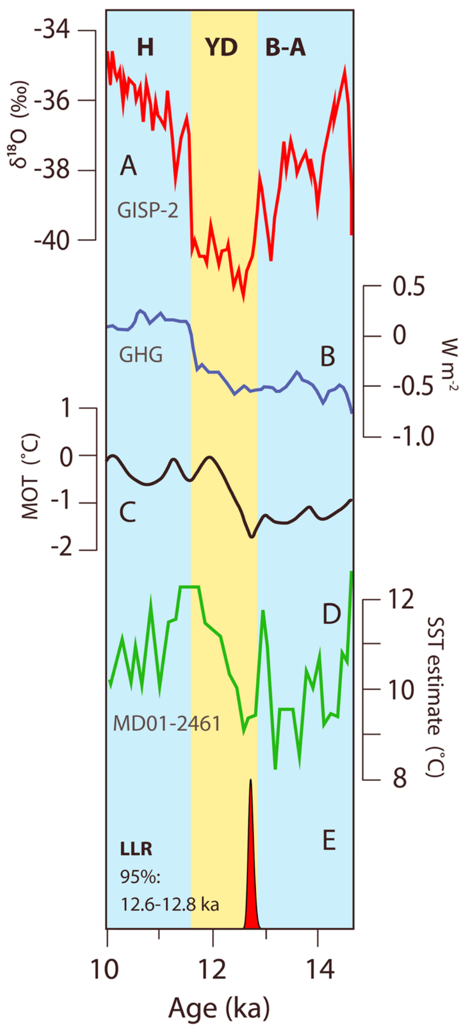

Figure 6. Scotland ^{14}C chronology for the Loch Lomond Readvance (LLR) in the context of North Atlantic and global late-glacial temperature reconstructions. (a) Greenland (GISP2) δ^{18}O record (Stuiver & Grootes, 2000). (b) Greenhouse gas forcing (Marcott et al., 2014). (c) Mean ocean temperature relative to modern values (Bereiter et al., 2018), reconstructed from the temperature-dependent solubility of noble gases. (d) Estimated sea-surface temperature from core MD01-2461 (Peck et al., 2008). (e) Probability distribution function for the culmination of the LLR (from Figure 3) showing that the glacial event culminated early in the Younger Dryas (YD) and thus that the period of ice growth likely preceded the onset of stadial conditions. Viewed in a broader context, this pattern of glacial behavior supports the hypothesis for extreme North Atlantic seasonality during stadials and suggests that YD deglaciation in Scotland was part of a global pattern of pronounced climate warming.

summertime surface heating of the stratified ocean (Bromley et al., 2014; Haug et al., 2005; Hays & Morley, 2004; Peck et al., 2008), cross-equatorial atmospheric compensation for reduced ocean-atmosphere heat transfer (Covey & Barron, 1988; Enderton & Marshall, 2009; Manabe, 1969; McGee et al., 2014), rising greenhouse gas concentrations (Broecker, 2013; Marcott et al., 2014), and/or increasing boreal summertime insolation, or simply in response to the significant upturn in global ocean temperature after ~12.8 ka (Bereiter et al., 2018; Figure 6). Thus, Scottish glaciers reveal the summertime temperature signal that is masked in records of North Atlantic mean annual air temperature (e.g., ice cores) by severely cold winters.

An alternative hypothesis, involving glacial starvation due to expanded winter sea ice, has been invoked in the past to explain the active retreat of British glaciers during the YD (Benn, 1997; Hughes et al., 2012). However, while a decline in winter precipitation might have exacerbated the glacial impact of summertime warming, starvation alone cannot account for the rapidity of YD retreat evident in the Scottish ^{14}C record. Nor can precipitation explain the concurrent retreat of ice masses located upstream of the North Atlantic (i.e., Laurentide ice sheet) and in regions already characterized by winter sea ice, such as Greenland and Baffin Island. A valuable test of the precipitation hypothesis would involve high-resolution reconstructions of glacial behavior in regions located closer to the stadial winter storm track, such as the circum-Mediterranean (e.g., Hughes et al., 2018).

With regard to what caused the LLR itself, the variability in seasonality we invoke to explain stadial deglaciation might also account for ice buildup during the Allerød/GI-1 interstadial. Whereas the YD in Scotland was characterized by high seasonality, we speculate that the opposite was true during the Allerød and thus that summers were sufficiently cool for glacier growth during this period of declining greenhouse gas forcing. Given the relative dearth of information on the glacial response to Allerød climate forcing, we suggest that this key period deserves more focused attention. Whatever the mechanism(s) responsible for driving interstadial (stadial) growth (decay) of glaciers in the British Isles, we note that several North Atlantic sea-surface temperature records, including one immediately adjacent the British Isles (core MD01-2461; Peck et al., 2008; Figures 5 and 6), show a distinct pattern of interstadial cooling and stadial warming (Peck et al., 2008; Ruhlemann et al., 1999), potentially helping reconcile glacier behavior with ocean surface conditions. Moreover, while this pattern is contrary to that supported by transient model simulations (Liu et al., 2009), we stress that those experiments were forced primarily with stadial meltwater from boreal ice sheets. If accurate, that meltwater represents a clear indication that ice masses surrounding the North Atlantic were indeed melting during stadial summers (Toucanne et al., 2009, 2015).

Finally, we point out that our new chronology brings events in Scotland and the circum-North Atlantic closer in line with glacier records from the tropics (Bromley et al., 2011; Jackson et al., 2014; Jomelli et al., 2014; Rodbell & Seltzer, 2000; Stansell et al., 2015) and southern midlatitudes (Kaplan et al., 2010, 2013; Menounos et al., 2013; Putnam et al., 2010; Strelin et al., 2011), raising the prospect of a widespread, rather than regional, signature of abrupt climate change during the last termination (e.g., Barker & Knorr, 2007; Figure 6).

5. Conclusions

We present 18 new radiocarbon dates, adjusted for marine reservoir effects using two separate corrections, which provide bracketing age constraint for the LLR. Additionally, we applied these reservoir corrections to the existing marine ^{14}C data set from deposits associated with the LLR. Our results suggest, to the first order, that fjords draining the western and southern margins of the WHIF, and also the neighboring MIF, remained ice free for much of the Bølling-Allerød, after which both ice fields attained their greatest extent during the late Allerød or earliest YD stadial. This scenario aligns with several previous minimum-limiting radiocarbon ages from Scotland and also with emerging chronologies from elsewhere in the circum-North Atlantic but is incompatible with models invoking culmination of the LLR toward the end of the YD. These incongruent scenarios underscore the pressing need for comprehensive, high-resolution, and reproducible assessments of glacial chronologies for the Scottish late-glacial landscape, which will in turn clarify patterns and causes of deglacial abrupt climate change in the circum-North Atlantic region. Comparison of our maximum-limiting ^{14}C data set to the minimum-limiting chronology from the former center of the WHIF suggests (a) that post-LLR deglaciation could have occurred in as few as 500 years and (b) that deglaciation occurred during full stadial conditions in the North Atlantic.

Acknowledgments

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