The complex topography of the dissected fjord landscape in southwestern Norway bears a close resemblance to the coastal fjord regions at the margins of the modern-day Greenland Ice Sheet. Establishing the timing and pattern of the last deglaciation in southwestern Norway is, therefore, valuable to contextualize predictions of future ice-sheet change under projected climate scenarios. However, the main motivation for this study is to describe and understand the glaciation history of this part of Norway.

During the Last Glacial Maximum (LGM), c. 26.5–19 ka (Clark et al. 2009), the Scandinavian Ice Sheet was connected to the British–Irish Ice Sheet across the North Sea (Fig. 1) and the Norwegian Channel Ice Stream flowed northwards along the west coast of Norway (Fig. 2). The ice stream subsequently broke up at c. 19–18 ka (Lehman et al. 1991; Sejrup et al. 1994, 2016; King et al. 1998; Morén et al. 2018), after which the western margin of the Scandinavian Ice Sheet became more or less parallel with the Norwegian coastline (Mangerud et al. 2017). 10Be exposure dates from Utsira and southern Karmøy (Fig. 2A) suggest a retreat of the ice stream as early as 20 ka (Svendsen et al. 2015), but Briner et al. (2016) proposed that this apparent age is slightly too high owing to the inheritance of muon-produced 10Be. Still, a tip of the southernmost part of western Norway apparently became ice free as early as about 18 ka (Vasskog et al. 2019), followed by ice-margin retreat in Boknafjorden beginning prior to 16 ka (Fig. 2A) (Gump et al. 2017). This early retreat contrasts with deglaciation around the mouth of Hardangerfjorden and further north, which did not become ice free until about 15 ka (Mangerud et al. 2013, 2016, 2017).

Following deglaciation of the outermost coast at c. 15 ka, the ice-sheet margin in the Hardangerfjorden area experienced a net retreat of >100 km (Fig. 2A) during the Bolling and Allerød warm periods (Mangerud et al. 2016). This period was apparently interrupted by an ice-sheet re-advance coinciding with the Older Dryas cold period at c. 14 ka, as inferred from radiocarbon-dated shell-bearing tills (Mangerud et al. 2016). From its restricted Allerød extent in inner Hardangerfjorden, the ice sheet re-advanced about 60 km to the island of Halsnøy in outer Hardangerfjorden during the Younger Dryas (hereafter YD) (Fig. 2A). The re-advance is documented by radiocarbon-dated shell-bearing tills and lake sediment records (Mangerud et al. 2016). The change from falling to rising relative sea level in a well-
dated sea-level curve indicates that the re-advance started as early as 13.6 ka, i.e. in middle Allerød (Lohne et al. 2007). The culmination of the ice-sheet re-advance in the Hardangerfjorden area is dated to the end of the YD at 11.8 ka and the start of its subsequent retreat from this position to 11.6 ka (Lohne et al. 2012). In the early Holocene, the ice sheet then retreated about 120 km to the head of Hardangerfjorden in only c. 500 years, i.e. a retreat rate of ~240 m a⁻¹ (Mangerud et al. 2013; Akesson et al. 2020).

Whilst a wealth of previous research has dealt with the timing and configuration of the Allerød–YD ice-sheet re-advance in Hardangerfjorden, the more detailed geometry of the YD ice-sheet extent between Hardangerfjorden and Matersfjorden (Figs 2A, 3) still remains uncertain. Further, there are only limited chronological constraints on the deglaciation from the outer coast prior to the YD and no regional constraints on the ice thickness during the LGM and subsequent thinning during deglaciation. It has in fact long been debated whether the highest mountain peaks in Norway, including in this area, were ice covered or ice free during the LGM (e.g. Nesje et al. 1987, 1988; Follestad 1990; Nesje & Dahl 1990; Sollid & Sorbel 1994; Brook et al. 1996; Fjeldskaar 2000; Diesen 2003; Goehringer et al. 2008). In this study, we use geomorphological mapping based on high-resolution Light Detection and Ranging (LiDAR) data combined with cosmogenic nuclide ¹⁰Be exposure dating to resolve this problem in our study area.

In addition, we use lake sediment cores, dated by accelerator mass spectrometry (AMS) ¹⁴C and the stratigraphic position of the 12.1 ka Vedde Ash (Mangerud et al. 1984; Lohne et al. 2013, 2014), to: (i) resolve whether the highest mountain peaks in the Hardangerfjorden area remained ice free or not during the LGM, and if not, date when they protruded as nunataqs and reconstruct the subsequent thinning history of the ice sheet; (ii) define the extent of the YD ice sheet between Hardangerfjorden and Matersfjorden; and (iii) identify if there were any independent ice caps in the higher mountain massifs during the YD.

Study area

Hardangerfjorden, the second-longest fjord in Norway, is more than 1000 m deep if all of the Holocene sediments are removed (Holte.dahl 1975; Aarsen 1997) and it is bordered by steep mountains ranging from ~500 m a.s.l., around the mouth of the fjord, to ~1800 m a. s. l. near the head of the fjord. Across the outer part of the fjord lies the prominent Halsnøy moraine (Fig. 3), deposited by the Scandinavian Ice Sheet during the YD (e.g. Undás, 1963; Holte.dahl 1967; Follestad 1972; Mangerud 2000; Lohne et al. 2012; Mangerud et al. 2013).

In this study, we focus on the mountains of Melderskinn, Laurdalstind, Solfjell, Engla.fell and Ulvanosa (Fig. 3), as well as the lower mountain Sigheiro (474 m a.s.l.) on the island of Bomlo at the mouth of Hardangerfjorden (Fig. 2A). We also investigate lower-lying areas between Hardangerfjorden and Matersfjorden.

Fig. 1. Location of the study area (Fig. 2A) and the maximum ice-sheet extents (Last Glacial Maximum, LGM) during the last glacial period in northwestern Europe (modified from Hughes et al. 2016).
Material and methods

**LiDAR and field mapping**

The area was analysed for ice-marginal landforms on digital terrain models (DTMs) derived from LiDAR data compiled by the Norwegian Mapping Authority (https://hoydedata.no/LaserInnsyn/). From the DTMs, altitudes can be determined with a precision of <10 cm, relative to the Norwegian Normal zero (NN2000), in practice mean sea level. The remote sensing was complemented by ground truthing in the field.

**Cosmogenic $^{10}$Be exposure dating**

To constrain the deglacial history we used cosmogenic $^{10}$Be exposure dating of glacial erratic boulders ($n = 66$) and bedrock surfaces ($n = 2$). As the ice surface is expected to have thinned upstream during the retreat of the ice margin, we have collected rock samples on summits along Hardangerfjorden in order to date when they first protruded through the ice surface. These data are in turn used to reconstruct ice-sheet profiles along Hardangerfjorden during the deglaciation.

The rock samples were collected from the uppermost surfaces of large erratic boulders and exposed bedrock

Fig. 2. A. Location map (boxes in Figs 1 and 2B) of areas discussed in the text. Ice margins are shown as white lines with ages in cal. ka BP (based on Anundsen 1972; Follestad 1972; Mangerud et al. 2016, 2017; Gump et al. 2017). B. Map of southern Norway and the adjacent North Sea showing the LGM ice margin and corresponding ice flow of the Norwegian Channel Ice Stream. The Younger Dryas (YD) ice margin (thin white line) is also shown (modified from Mangerud et al. 2013).
surfaces using a portable diamond saw, hammer and chisel. We sampled mostly flat surfaces and avoided edges. All of the samples were collected from above the local marine limit. Sample coordinates were determined using a handheld GPS and the elevations were later determined from the LiDAR DTMs. As many of the highest summits are covered with blockfields (Fig. 4), extra caution was taken to only sample large boulders perched on top of the blockfields in positions that were best explained by glacial transport and deposition (Fig. 4B).

All samples were prepared for $^{10}$Be analysis at the University at Buffalo Cosmogenic Nuclide Laboratory. After the samples had been crushed and sieved and the quartz isolated, the samples were digested, and beryllium was isolated following procedures described in Young et al. (2013). $^{10}$Be/$^{9}$Be ratios were measured at the Center for Mass Spectrometry, Lawrence Livermore National Laboratory, and normalized to standard 07KNSTD3110 with a reported ratio of $2.85 \times 10^{-12}$ (Nishiizumi et al. 2007; Rood et al. 2010). Procedural blank ratios range from $1.1 \times 10^{-15}$ to $3.9 \times 10^{-15}$, equating to background corrections of 1–4% of the sample total.

All $^{10}$Be ages were calculated using version 3 of the online CRONUS Earth exposure age calculator (hess.ess.washington.edu; Balco et al. 2008; Balco 2017) using the regionally constrained production rate for Scandinavia of $4.13 \pm 0.11$ at $g^{-1} a^{-1}$ (Stroeven et al. 2015; close to the locally determined production rate ($4.15 \pm 0.15$ at $g^{-1} a^{-1}$) of Goehring et al. 2012a, b), and time-dependent (Lm) scaling (Lal 1991). We also note that if, instead of the Scandinavian production rate (Stroeven et al. 2015), we used the global ‘primary’ $^{10}$Be production rate by Borchers et al. (2015), the ages would only be ~1% younger. Reported age uncertainties reflect the one sigma AMS analytical uncertainty. Ages refer to the sampling year (AD 2015–2017). We assume an erosion rate of $1.72 \times 10^{-4}$ cm a$^{-1}$ for all samples based on differential quartz vein relief observed on erratic
boulders at the Halsnøy moraine, outer Hardangerfjorden (Goehring et al. 2012a). We made no corrections for shielding by snow cover.

Elevation changes, owing to isostatic rebound, during a rock surface’s exposure history will result in a time-varying rate of $^{10}$Be production at the sampling site. The Marine Limit is 80 m a. s. l. near the Dyrrinda moraine, and thus the samples exposed at 11.6 ka BP are now located 70 m higher relative to sea level. Sample elevations can potentially be corrected for this using regional sea-level curves (e.g. Jones et al. 2019) and to exemplify the impact of using different elevations when calculating $^{10}$Be ages we use sample 96-BM3 from the Dyrrinda moraine, for which we obtain a value of 12.2$\pm$0.3 ka. If instead we used an altitude 80 m lower (the YD relative sea level for the sample site), the age would be 13.1$\pm$0.3 ka. However, the isostatic uplift here was very fast during the first period and thus a 25 m lowering would probably be a more realistic ‘mean’ altitude, giving 12.4$\pm$0.3 ka. Nevertheless, atmospheric variability at the sampling sites, such as varying proximity to the ice margin, may counteract the effects of isostatic rebound on $^{10}$Be production rates (Staiger et al. 2007). Our $^{10}$Be ages are, therefore, calculated applying the current altitude of the samples.

Sediment coring, bathymetric and seismic surveying

Glacier-fed lake sediment records can be used to determine fluctuations of glacier and ice-sheet margins as glaciolacustrine sediments can often be distinguished from other lacustrine sediments (e.g. Mangerud et al. 1979; Lohne et al. 2012). In this study, two palaeolakes, Svartatjønna (314 m a.s.l.) and Kleivahaugtjønn (240 m a.s.l.) (Fig. 3), which are now partly mires, were cored using a 10-cm-diameter Russian peat corer. Lake Langelivatnet (Fig. 3) was cored using a modified Nesje piston corer (Nesje 1992). The cores were brought back to the University of Bergen and stored at ~4 °C before further sediment logging and laboratory analyses. A bathymetric survey of Langelivatnet was performed using a Lowrance Elite-5 HDI echo sounder, and a bathymetry map was produced in the software Reefmaster. The sedimentary infill of Langelivatnet was mapped with a portable EdgeTech 3100P chirp system and the seismic stratigraphy visualized using EdgeTech software.

Radiocarbon dating and identification of the Vedde Ash

The sediment cores were sampled at selected intervals for AMS $^{14}$C dating, and the samples were sieved through 100- and 63-µm sieves before organic macrofossils were picked by tweezer under a light microscope. Although terrestrial material was preferred, some samples only contained fragmented macrofossils and it was therefore impossible to determine if they were all from terrestrial plants. Bulk sediment samples were dated where no macrofossils were found. The macrofossils were dried at 50°C and placed in sterilized and sealed vials before they were sent to the Radiocarbon Dating Laboratory at Lund University, Sweden, for AMS $^{14}$C dating. All $^{14}$C ages were calibrated using the IntCal20 calibration curve (Reimer, 2020) and are reported relative to AD 1950 (cal. a BP).

In addition, the core chronologies were also supported by the occurrence of the YD-age (12.1 cal. ka BP) Vedde Ash (Mangerud et al. 1984), based on its characteristic shard morphology and colours, the fact that it contains both the rhyolitic and basaltic components (Mangerud et al. 1984), its stratigraphic position and the fact that the Vedde Ash is the only visible ash layer found in western Norway (e.g. Lohne et al. 2012).
Table 1. $^{10}$Be sample data and ages. All samples have a rock density of 2.65 g cm$^{-3}$.

| Sample          | Sample type  | Latitude (°N) | Longitude (°E) | Elevation (m a.s.l.) | Sample thickness (cm) | Topographic shielding factor | $^{10}$Be concentration (at g$^{-1}$) | $^{10}$Be age (ka)$^1$ |
|-----------------|--------------|---------------|----------------|----------------------|-----------------------|-----------------------------|----------------------------------------|-------------------------------|
| Sigjø            | Boulder      | 59.96258      | 5.85370        | 997                  | 1                     | 1                           | 159 285 ± 3013                      | 41.9 ± 0.3                   |
| Skorafjell       | Boulder      | 59.86477      | 5.78675        | 1001                 | 2                     | 1                           | 126 441 ± 2568                      | 11.8 ± 0.2                   |
| Englafljell      | Boulder      | 59.85364      | 5.87827        | 1171                 | 2                     | 1                           | 494 043 ± 6579                      | 41.9 ± 0.6                   |
| Ulnavosa summit  | Boulder      | 59.86435      | 5.94005        | 1111                 | 1.4                   | 1                           | 211 543 ± 9777                      | 18.1 ± 0.4                   |
| Solfjell         | Boulder      | 59.93310      | 5.94193        | 827                  | 2                     | 1                           | 134 089 ± 513                       | 14.6 ± 0.3                   |
| Melderskin       | Boulder      | 60.00284      | 6.06856        | 1332                 | 2                     | 1                           | 1 652 382 ± 18 595                  | 144.8 ± 2.1                  |
| Laurdalstind     | Boulder      | 59.99479      | 6.08192        | 1304                 | 2.4                   | 1                           | 457 769 ± 5964                      | 34.5 ± 0.5                   |
| Ulnavosa vertical transect | Boulder | 59.86741      | 5.93622        | 1091                 | 1.6                   | 1                           | 155 593 ± 2929                      | 13.5 ± 0.3                   |

(continued)
Results

Cosmogenic $^{10}$Be exposure dating results

To date the retreat and thinning of the ice sheet, 66 boulders and two bedrock samples were dated (Table 1). Samples were collected from the mountains of Ulvanosa, Melderskin and Laurdalstind in the east (previously mapped as being above the YD ice-sheet surface; Folkestad 1972) and the island of Bømlo (previously mapped as being beyond the YD limit; Folkestad 1972) on the outer coast to the west (Figs 2A, 3).

Exposure ages from the highest peaks and the outer coast. – The oldest $^{10}$Be exposure ages were obtained from areas above 1100 m a.s.l. on the mountain summits of Englafljell, Ulvanosa, Melderskin and Laurdalstind (Figs 5,6). The ages displayed a large

![Diagram](https://via.placeholder.com/150)

Fig. 5. $^{10}$Be ages of boulders from the summits of Melderskin (1426 m a.s.l.), Laurdalstind (1307 m a.s.l.) and Ulvanosa (1246 m a.s.l.) plotted against elevation. Note that the two older ages from Ulvanosa are bedrock samples and the two breaks in the time scale at 36–75 and 115–140 ka. Note also that there are samples younger than 22 ka on both Melderskin and Ulvanosa.

Table 1. (continued)

| Sample | Sample type | Latitude (°N) | Longitude (°E) | Elevation (m a.s.l.) | Sample thickness (cm) | Topographic shielding factor | $^{10}$Be concentration (at g$^{-1}$) | $^{10}$Be age (ka)$^1$ |
|--------|-------------|---------------|----------------|---------------------|-----------------------|-----------------------------|---------------------------------|-------------------|
| 85-UN51 | Boulder | 59.83941 | 5.91997 | 613 | 1.9 | 1 | 104 060±2428 | 13.7±0.3 |
| 85-UN52 | Boulder | 59.83941 | 5.91997 | 613 | 1.7 | 1 | 109 000±3119 | 14.3±0.4 |

Scandinavian production rate (Stroeven et al. 2015).
scatter with four ages as old as 155–75 ka and the remaining nine between 35 and 18 ka.

All $^{10}$Be samples from the summit of mount Siggio on the western coast yielded ages between 14.3 and 14.8 ka (average = 14.5 ka), after removal of an outlier of 6.1 ka (Fig. 6).

On the western slope of Solfjell, farther into Hardangerfjorden, lie two conspicuous and parallel moraine ridges called Langhaugene and Sandhaugene, at 560–620 m a.s.l. (Fig. 7). From their appearance, it is difficult to tell whether they are lateral moraines and, if so, whether they were deposited by ice from the north or south. We dated 14 boulders from the moraines and surroundings (Fig. 7). All but one span 12.4–11.2 ka (average = 11.7 ka), indicating that the moraines were emplaced during the YD between ice flowing from both the north and the south.

Above the Langhaugene and Sandhaugene moraines, at 807–827 m a.s.l. on Solfjell, $^{10}$Be ages from six erratic boulders fell between 14.1 and 15.5 ka and one boulder gave a value of 11 ka (Fig. 6).

Vertical transect of exposure ages on Ulvanosa mountain. – To address the regional thinning history of the ice sheet, 19 boulders were sampled in a vertical transect from 1091 to 613 m a.s.l. on the southern slope of Ulvanosa mountain (Fig. 8). The resulting ages range from 17.3 to 11.5 ka, with older ages at lower elevations than on higher ground (Fig. 9). Surprisingly, even the lowest samples yielded pre-YD ages (13.7–14.4 ka) even though they are located well inside the YD ice extent as mapped by Follestad (1972) (Fig. 3), indicating that larger areas remained ice free during the YD than previously thought.

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**Fig. 6.** Location map with $^{10}$Be ages from summits around Hardangerfjorden. YD ice margins (green lines) after Mangerud *et al.* (2016) and Gump *et al.* (2017). Note that, based on our results in this study, we have modified the ice margin between Ulvanosa and Måleråsfjorden (dashed red line) as compared with the previous work by Follestad (1972) (Fig. 3), and we have reconstructed a local ice cap on the Ulvanosa massif (Fig. 14).
Exposure ages from the Dyrrinda moraine and mapping of terraces in Tveitedalen. – Our $^{10}$Be ages from just outside the Dyrrinda moraine (Figs 8,10) suggest pre-YD ages, contrary to earlier interpretations, in which this area was covered by YD ice (Folkestad 1972). We therefore decided to also sample the Dyrrinda moraine for exposure dating.

From the LiDAR-DTMs we identified a lobate continuation of the Dyrrinda moraine south of Bergsdalen (Fig. 10). Two boulder samples were collected

Fig. 7. LiDAR DTM with $^{10}$Be ages from Sandhaugene and Langhaugene moraines (location in Fig. 6).

Fig. 8. LiDAR DTM with $^{10}$Be ages from lower altitudes on the Ulvanosa massif (location in Fig. 6). Note the Dyrrinda moraine (green) in the southern end of Bergsdalen.
along this ridge and another four from the Dyrinda moraine proper. The Dyrinda samples yielded ages ranging from 14.4 to 10.5 ka. In addition to the oldest (14.4 ka) and the youngest (10.5 ka) ages, the other four ages are within 12.2–11.3 ka (Fig. 10), spanning the early Preboreal and late YD but with an average in the YD. The Dyrinda moraine is the first moraine proximal to the lakes with undisturbed Allerød–YD sequences (see below), and thus a YD age is far more probable than a Preboreal age.

About 2 km south of the Dyrinda moraine, across the mouth of Tveitdalen and only 400 m from the shore of Matersfjorden, lies a segmented moraine ridge (Fig. 10) that we consider to be a lateral moraine corresponding to the YD end moraine at the mouth of Matersfjorden (Fig. 6). On the distal side, i.e. north, of the moraine there are several glaciofluvial terraces located well above the marine limit (78 m a.s.l.), the largest being at 108–110 m a.s.l. and close to the moraine (Fig. 10). The high elevations suggest that these terraces were formed in ice-dammed lakes along the Matersfjorden glacier and that they were fed by a glacial river from the Dyrinda moraine (Fig. 10). If correct, this supports the hypothesis that the Dyrinda moraine was formed late in the YD.

Sediment cores and radiocarbon dating

To further investigate the extent of the YD ice sheet, we mapped the area between Ulvanosa and Matersfjorden and collected sediment cores from the three lakes Svartatjørna (314 m a.s.l.), Kleivahaugtjørn (240 m a.s.l.) and Langelivatnet (279 m a.s.l.) (Figs 3, 10). All three sediment cores were radiocarbon dated (Table 2).

Svartatjørna and Kleivahaugtjørn. – At Svartatjørna (Fig. 10) we cored from the floating fen surface surrounding the present lake. The corer stopped in coarse sediments, most likely till or glaciofluvial gravel, at 550 cm depth. The retrieved sediment succession (450–550 cm) displays six distinct units (Fig. 11). At the base is grey clayey silt with pebbles followed by 9 cm of brown gyttja. Above the gyttja is a 22-cm-thick, massive grey clayey silt including a 7-cm-thick Vedde Ash layer (12.1 cal. ka BP). Above the clayey silt is brown gyttja towards the top of the core, which represents typical Holocene lake sediments in western Norway. Four samples of plant macrofossils were extracted from the lower gyttja layer for AMS 

14C dating, resulted in ages between c. 12.9 and 13.5 cal. ka BP (Fig. 11). These ages suggest that the early Allerød is a minimum age for deglaciation, and that the area remained ice free throughout the YD.

Lake Langelivatnet and the extent of the YD ice sheet in Matersfjorden. – As the stratigraphies in Svartatjørna and Kleivahaugtjørn (Fig. 11) as well as the 10Be exposure ages show that the areas distal to the Dyrinda moraine remained ice free throughout the YD, we wanted to map the YD extent of the Scandinavian Ice Sheet in this area.

At the southern end of Lake Langelivatnet, we identified a moraine ridge (Fig. 10) with a configuration suggesting that it was formed by an ice lobe filling the lake basin. The only possible source for such an ice lobe is the major outlet glacier filling the Matersfjorden (Fig. 6). On the valley side SW of Langelivatnet, the DTMs reveal an ~800 m long and narrow terrace at ~308 m a.s.l., i.e. ~30 m above the lake level (Figs 10, 12A). Smaller terraces were found at the corresponding elevation on the eastern valley side. As the valley in which Langelivatnet
resides currently drains northwards, we suggest that the terraces are shorelines formed in a lake dammed by the glacier tongue that formed the moraine south of Langelivatnet.

The bathymetric and seismic data revealed an undulating, high-amplitude reflection (acoustic basement), which we interpret as bedrock and/or a basal till, draped by acoustically layered sediments up to ~3 m in thickness (Fig. 12B). There were no signs of any moraine ridges on the lake floor, but point reflectors in the seismic record appearing just above the basement are interpreted as glacial boulders. The sediment core displays a simple stratigraphy with laminated clay and silt with intercalated sand layers in the lower 45 cm, followed by a sharp boundary to the overlying brown gyttja (Fig. 12D). The laminated silt and clay are interpreted as glaciolacustrine sediments deposited during ice-margin retreat from the lake basin. We assume that the deposition of the Holocene gyttja started soon after the ice margin retreated north of the valley and the inflow of glacial meltwater and sediments into Lake Langelivatnet stopped. Three radiocarbon dates from the lower part of the gyttja resulted in early Holocene ages, with the bottom two being c. 11.2 cal. ka BP (Fig. 12D), providing a minimum deglaciation age for Langelivatnet, and therefore also for the moraine ridge in its southern end.

Discussion

Emerging nunataqs and the onset of deglaciation

The oldest $^{10}$Be exposure ages in this study, ranging from 155–18 ka, were obtained from the high mountain summits of Englafljell, Ulvanosa, Melderskind and
Table 2. Radiocarbon and Vedde Ash sample data and ages from Svartatjørna, Kleivahaugstjørn and Langelivatnet. Median ages are rounded to the nearest decade.

| Depth (cm) | Weight (mg) | Weight C (mg) | Material | Laboratory ID | ¹⁴C ages (a BP) | Calibrated ages (cal. a BP) |
|-----------|-------------|---------------|----------|---------------|-----------------|-----------------------------|
|           |             |               |          |               | From | To | From | To | Median |
| Svartatjørna |             |               |          |               | From | To | From | To | Median |
| 524–525   | 2.4         | 1             | Betula nana and Salix leaves | LuS 14482 | 10 970±60 | 12 767 | 12 961 | 12 760 | 13 065 | 12 890 |
| 527–528   | 2.6         | 1             | Small stick | LuS 14483 | 11 750±60 | 13 505 | 13 741 | 13 493 | 13 758 | 13 600 |
| 531–532   | 2.7         | 1.2           | Mixed plant remains | LuS 14484 | 11 410±60 | 13 183 | 13 333 | 13 168 | 13 419 | 13 280 |
| 532–533   | 1.7         | 0.4           | Mixed plant remains | LuS 14485 | 11 605±70 | 13 363 | 13 577 | 13 319 | 13 594 | 13 460 |
| 511–518   |             |               | The Vedde Ash |               |       |     |       |     |       |
| Kleivahaugstjørn |         |               | Mixed plant remains | LuS 15345 | 8620±45 | 9534 | 9659 | 9528 | 9692 | 9590 |
| 729–730   | 13.9        | 1.3           | Bulk sediment | LuS 15346 | 855±50 | 9807 | 10 148 | 9730 | 10 172 | 9960 |
| 741–742   | 1.7         |               | Bulk sediment | LuS 15347 | 9745±45 | 11 164 | 11 234 | 10 893 | 11 248 | 11 190 |
| 751–752   | 1.1         |               | Bulk sediment | LuS 15348 | 480±19 | 12 192 | 12 622 | 12 060 | 12 680 | 12 460 |
| 758–759   | 6.5         | 1.4           | Mixed plant remains | LuS 15349 | 10 480±70 | 12 064±48 | 12 121±57 | (Lohne et al. 2013, 2014) | (Rasmussen et al. 2006) |
| 782–787   |             |               | The Vedde Ash |               |       |     |       |     |       |
| Langelivatnet |             |               | Small stick | LuS 15349 | 8330±50 | 9292 | 9427 | 9142 | 9471 | 9350 |
| 49–60     | 9.2         | 1.5           | Bulk sediment | LuS 15350 | 9825±50 | 11 201 | 11 259 | 11 175 | 11 390 | 11 240 |
| 63–64     | 1.0         |               | Bulk sediment | LuS 15351 | 9770±45 | 11 185 | 11 235 | 11 111 | 11 257 | 11 210 |

¹This age is subtracted by 50 years from their B2K age in order to be comparable with calibrated ¹⁴C ages.

Laurdalstind (Figs 5, 6). These ages may be used as arguments for the summits remaining ice free through LGM (e.g. Diesen 2003). However, we consider that the spread of the older ages, combined with the five much younger ages between 22 and 18 ka (Fig. 5), strongly suggests that the older ages are due to inheritance. Nuclide inheritance occurs when erratics (or bedrock) have been exposed during previous ice-free periods, or if the parent bedrock has not been sufficiently eroded (>2 m) to remove previously produced ¹⁰Be, resulting in apparent ages older than the time of deglaciation (e.g. Briner et al. 2005). Several studies have demonstrated that inheritance increases with altitude in Norway (Egholm et al. 2017; Andersen et al. 2018; Jansen et al. 2019). We therefore conclude that the oldest ages are due to inheritance. We also conclude that the youngest ages show that all summits were ice covered during the LGM, and protruded as nunataqs c. 22–18 ka. However, owing to the inheritance, and indeed other sources of error, we do not consider the face values of these ages as very reliable.

The age range of 21–18 ka overlaps with age estimates (20–18.5 ka) of the shut down and subsequent break-up of the Norwegian Channel Ice Stream (Lehman et al. 1991; Sejrup et al. 1994, 2016; King et al. 1998; Svendsen et al. 2015; Morén et al. 2018). We acknowledge the low precision in the age determinations for both events, but we postulate that the retreat of the ice-stream margin would result in a lowering of the ice-sheet surface over Hardangerfjorden and thus the first emergence of the higher mountain summits as nunataqs (Fig. 13).

Deglaciation from the outer coast
At the westernmost locality, the summit of mount Sigjø, the ¹⁰Be ages grouped around 14.5 ka (after removal of an outlier of 6.1 ka), and thus overlap with previous ¹⁰Be ages from lower elevations on southern Bømlo (Mangerud et al. 2013) and from Ryvarden south of Bømlo (Gump et al. 2017) (Fig. 13). The similar ages obtained near sea level and on summits along the outer coast suggest such a fast retreat/thinning that the time differences are within the uncertainties of the method, as also shown slightly farther north (Mangerud et al. 2017). A main point is that following deglaciation of the Norwegian Channel, at c. 20–18.5 ka, there was little or no ice-sheet retreat until c. 14.5 ka, when the mouth of Hardangerfjorden became ice free, again a parallel situation to the adjacent area to the north (Mangerud et al. 2017). We suggest that the halt in ice recession following the collapse of the Norwegian Ice Stream is due to the many islands along the coast and the shallow sill at the mouth of Hardangerfjorden and other fjords (Fig. 13) acting as pinning points that stabilized the ice.


margin. \(^{10}\)Be ages from Boknafjorden slightly farther south (Fig. 2A) suggest that its deglaciation initiated as early as c. 16 ka (Fig. 2; Svendsen et al. 2015; Gump et al. 2017), i.e. some 1500 years before the mouth of Hardangerfjorden. The earlier deglaciation of Boknafjorden is potentially due to the absence of pinning points in its deep and wide trough (>200 m depth and >8 km width).

Following deglaciation from the outer coast at c. 14.5 ka, the ice margin retreated fast during the mild Bølling period and was located inside the Halsnøy moraine after only a few hundred years (Mangerud et al. 2016). However, the retreat in Bølling was interrupted by an Older Dryas re-advance, reaching past the YD position. The Older Dryas re-advance is suggested based on Bølling ages of shell fragments in tills both outside and inside the YD moraine (Mangerud et al. 2016). Contrastingly, in the Bergen area, slightly to the north, the Older Dryas ice margin is interpreted as being located inland of the YD margin (Mangerud et al. 2017). However, when studying
Fig. 12. A. Photograph of the ice-dammed lake shoreline south of Langelivatnet (photograph, Reidun Eldegard). B. Chirp line 7 from Langelivatnet (red line in D) and lithological interpretation of reflectors. C. Bathymetric map of Langelivatnet with chirp lines and coring location. D. Sediment core with radiocarbon dates (coring location marked in C).
Fig. 13. A. $^{10}$Be ages from outside and above the Younger Dryas moraines. Ages are rounded to nearest 0.5 ka. The Younger Dryas moraines shown in (B) are marked in green. B. Topographic profile from the Norwegian channel in the SW to inner Hardangerfjorden in the NE (location of the profile shown in A). Ice-sheet profiles based on our interpretations are shown for key time slices discussed in the text.
LiDAR DTMs of areas surrounding the outer part of Hardangerfjorden, we did not find any ice-marginal landforms outside of the YD moraines, but we note that there is a shallow threshold in Hardangerfjorden between Bømlo and Stord (Fig. 13B) that potentially acted as a pinning point for the ice sheet, resulting in a temporary halt in retreat or the end of a re-advance. Unfortunately, the large spread and uncertainties of our $^{10}$Be ages from sites just outside and above the YD moraines (outside the Dyrinda moraine, Fig. 10, and on top of Solfjell, Fig. 6) span across the Older Dryas, and thus cannot provide a conclusive answer as to whether these sites were deglaciated prior to or after the Older Dryas. However, the basal dates of c. 13.5 cal. ka of the Svartatjørna sediment core (Fig. 9) indicate that this area was deglaciated after the Older Dryas, suggesting a more extensive ice cover there during the Older Dryas than the YD. Following 14 ka, the ice margin in Hardangerfjorden retreated at least another 80 km during the Allerød before it re-advanced at least 60 km to its YD maximum position at the Halsnøy moraine (Figs 2A, 13B; Mangerud et al. 2016).
The Younger Dryas ice sheet and a local ice cap between
Hardangerfjorden and Møtjøftjønn.

The culmination of the ice-sheet re-advance in
Hardangerfjorden, at the Halsnøy moraine, is dated to
the very end of the YD; 11.8 to 11.6 cal. ka BP (Lohne et al. 2012). In
earlier reconstructions (e.g. Follestad 1972) it was
assumed that the YD ice sheet inundated the entire
area between Hardangerfjorden and Møtjøftjønn
(Figs 2A, 3). However, the pre-YD 10Be ages from the
lower end of our dating transect at Ulvanosa (Figs 8, 9),
together with the undisturbed sediment successions
across the YD from Svartatjørn and Kleivahautgjøtt
(Fig. 11), show that these areas instead remained ice free
throughout the YD. The Møtjøftjønn outlet glacier
from the Scandinavian Ice Sheet must, therefore, have
been confined to lower elevations along the fjord with
only a smaller glacier tongue reaching up across Lake
Langelivatnet (Figs 10, 14).

A Younger Dryas ice cap on the Ulvanosa massif and
and the age of the Dyrrinda moraine. – The YD 10Be ages from
high elevations (~850–1100 m a.s.l.) on the Ulvanosa
mountain massif (Figs 8, 9) suggest that parts of this
mountain were covered by ice during the YD. On
the other hand, the pre-YD 10Be ages and sediment stratigraphies
from the lower areas (Fig. 11) demonstrate that
the ice sheet did not reach this high during the YD. The
interpretation must be that there was an independent ice
cap on Ulvanosa during the YD (Fig. 14) and that
the Dyrrinda moraine represents an outlet glacier from this
ice cap.

Unfortunately, given the spread and uncertainty of the
10Be ages from the Dyrrinda moraine, they span both the early
Preboreal and late YD and cannot alone provide a conclusive age of moraine formation. Likewise, the
sediment stratigraphy from Kleivahau gbjørn (Fig. 11)
can only constrain the age of the Dyrrinda moraine to
sometime between 12.1 and 11.2 cal. ka BP from
numerical dates.

The highest elevation of a lateral moraine is a
measure of the minimum height of the Equilibrium
Line Altitude (ELA) (Andersen 1954), and in practice it
is often a good approximation of the ELA (e.g. Larsen
et al. 1984). The Dyrrinda moraine can be traced to
575 m a.s.l., or ~500 m above the YD sea level, which
fits well with the tilted late YD-ELA in Mangerud et al.
(2016). In addition, a late YD age of the Dyrrinda
moraine is supported by the correlation with the
glaciolfluvial terraces in Tveitdalen (Fig. 10). A minimum
age for the terraces is the time when this part of
Møtjøftjønn was finally deglaciated, which can be
obtained by dating the marine limit terrace (78 m a.s.l.)
at the mouth of Tveitdalen. When plotting this eleva-
tion onto the master shoreline diagram for western
Norway (Mangerud et al. 2019), it yields an age of c.
11.4 cal. ka BP, consistent with ice-margin retreat from
the Halsnøy moraine at 11.6 cal. ka BP (Lohne et al.
2012). Thus, the age of the Dyrrinda moraine is bracketed between 12.1 (the Vedde Ash at Kleivahautgjøtt) and 11.4 cal. ka BP. We therefore conclude that the Ulvanosa Ice Cap reached its maximum extent in concert with the ice sheet in Hardangerfjorden
(Lohne et al. 2012), i.e. at the very end of the YD, after
which it retreated at the onset of the Holocene.

Conclusions

• The higher mountain summits in the outer
Hardangerfjorden area (~1200–1400 m a.s.l.) were ice cov-
ered during the LGM and melted out as nunataqs c.
22–18 ka, probably as a consequence of the break-up of
the Norwegian Channel Ice Stream.

• Following break-up of the Norwegian Channel Ice
Stream, the ice margin stabilized at the outermost
coast for 3500–5500 years, before the mouth of
Hardangerfjorden deglaciated at c. 14.5 ka.

• More extensive areas that were ice free during the YD,
which earlier were considered ice covered, are docu-
mented between Hardangerfjorden and Møtjøftjønn.

• We discovered that there has been a local ‘Ulvanosa
Ice Cap’ on mountains adjacent to the outer Har-
dangerfjorden during the YD. It was independent of
the Scandinavian Ice Sheet but feeding into it on the
north flank.

• The Ulvanosa Ice Cap reached its maximum extent
during late YD, simultaneously with the ice sheet in
Hardangerfjorden.

• The ELA for the Ulvanosa Ice Cap is consistent with
an inland-rising ELA values reported in prior studies.

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much of the 10Be rock sampling field work was performed prior to CR
started on his Ph.D. project. CR performed much of the subsequent
field work (all sediment coring and additional rock sampling), did the
remote sensing analyses, conducted laboratory analyses of the sediment
cores and wrote the first draft of the manuscript, but all authors
participated in formulations and modifications. JIS was project leader
and participated in all field work. JM participated in developing the
ideas and in most of the field work. HH participated in field work and
analyses concerning sediment cores and seismics. JB participated in
some field work (rock sampling) and conducted all 10Be analyses.
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