Spatial distribution of cold-ice within a temperate glacier - implications for glacier dynamics, sediment transport and foreland geomorphology

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Abstract. This study suggests that cold-ice processes may be more widespread even within temperate glacial systems, than previously assumed. We present the first direct observations of cold-ice at the snout of the temperate glacier Midtdalsbreen an outlet of the Hardangerjøkulen icefield (Norway) from 43 line-kilometres of ground penetrating radar data. Results show a 40 m-wide cold-ice zone within the majority of the glacier snout, where ice thickness is <10 m. We interpret ice to be frozen to its bed across this zone, consistent with basal freeze-on processes involved in the deposition of minor moraines. We also find at least two zones of cold-ice up to 15 m thick within the ablation area, occasionally extending to the glacier bed. There are two further zones of cold-ice up to 30 m thick in the accumulation area, also extending to the glacier bed. Cold-ice zones in the ablation area tend to correspond to areas of the glacier that are covered by late-lying seasonal snow patches that reoccur over multiple years. Parts of the glacier frozen to the bed occur up-glacier from controlled moraines, and this basal thermal state likely exert control on moraine formation and location. We hypothesise that basal freeze-on may also influence glacier dynamics via increased lateral drag, and may become particularly important in forthcoming decades with i) continued thinning of Hardangerjøkulen and ii) retreat of Midtdalsbreen to higher altitudes where basal freeze-on could become more widespread. Adversely, the number and size of late-lying snow patches in the ablation area may decrease reducing the extent of cold-ice.

1 Introduction

Glaciers are generally divided into three types of thermal regime depending on whether the ice is at or below pressure-melting point, or a combination of the two states. Thermal regime influences glacier surface mass balance and dynamics, and therefore affects processes of sediment delivery to the ice margin, as well as landform development. Increasingly, glacier thermal

5 regimes are more complex than previously assumed in this broad classification, and the actual thermal conditions of any individual glacier can be highly spatially variable (e.g. Rippin et al., 2011; Waller et al., 2012). Cold-ice (ice that does not contain any liquid water below pressure melting point) may in fact also be present in isolated pockets within glaciers that are typically classified as temperate. A wide range of thermal structures are observed in polythermal glaciers (e.g. Pettersson, 2004) varying from those containing a perennial cold surface layer overlying a temperate core (Pettersson et al., 2007) to high

10 Arctic ice caps and icefields with cold-ice margins and ice frozen to the bed (basal freeze-on) (Dowdeswell et al., 1999). Past Pleistocene ice sheets are also known to have had frozen-bed patches (Kleman and Hättestrand, 1999), and several geomorphological and sedimentological studies have suggested that some temperate glaciers may have a cold-ice margin. Basal freeze-on occurs particularly during winter when liquid water content at the bed is diminished and energy production through friction and internal deformation is reduced. Sediments are entrained at the base of the glacier snout and then melt out

15 during the ablation season (Sharp, 1984; Krüger, 1993; 1994; 1995; 1996; Matthews et al., 1995; Krüger et al., 2002; 2010; Bradwell, 2004; Evans and Hiemstra, 2005; Winkler and Matthews, 2010; Reinardy et al., 2013; Hiemstra et al., 2015). This growing body of evidence suggests that cold-ice processes within primarily temperate-glacier systems may influence marginal depositional processes and landforms. In addition to this influence, to date, the role of cold-ice in glacier mass balance and flow regime of temperate glacier systems has been viewed as a minor component. However, this view is supported by little

20 evidence from direct observations of cold-ice and its distribution within otherwise temperate glaciers.

Here, we determine, through direct measurement, the thermal regime of Midtdalsbreen, a Norwegian glacier previously described as temperate (Andersen and Sollid, 1971). Midtdalsbreen is an outlet glacier of the Hardangerjøkulen icefield in southern Norway (Fig. 1) that has been retreating from its Little Ice Age (LIA, cf. 1850) maximum position until the present

25 day, excepting a small regional readvance of glaciers across Scandinavia, including Midtdalsbreen, recorded during the 1990s (Nesje and Mathews, 2012). Detailed mapping of minor moraines (typically less than 2 m high) deposited during the 1960s and 1970s indicated that many were ice-cored (Andersen and Sollid, 1971). A second survey of the glacier foreland (Reinardy et al., 2013) indicated that minor moraines deposited in the early 2000s were no longer ice-cored. Instead, sedimentological investigation of the moraines suggested basal freeze-on of subglacial sediments and incremental sediment slab stacking

30 mimicking the underside of the glacier snout. Reinardy et al. (2013) therefore suggested that at least part of the snout of Midtdalsbreen was cold-based. If the glacier was cold-based, even in part, during the last two decades of relative warmth when these moraines were formed, then similar cold-based conditions are likely to have persisted at least through the last 40-50 years when the climate was comparatively colder and even more conducive to basal freeze-on. This is particularly significant because

the decade 2000-2010 is likely to have been the warmest since the LIA, reflected in increasing glacier retreat rates (Andresen, 2011; Andreassen et al., 2016), yet the sedimentological evidence within the moraines implies that the snout of Midtdalsbreen would have remained frozen throughout retreat.

5   We first investigate whether interpretations of basal freeze-on from sedimentological evidence (Reinardy et al., 2013) are supported by direct observations of thermal regime using ground-penetrating radar (GPR) over the glacier snout. Second, we discuss the influence that the thermal regime may have on glacier dynamics and the resulting geomorphic imprint preserved in the glacier foreland since the LIA. Determining the detailed thermal regime of Midtdalsbreen is associated with three wider hypotheses: i) Freeze-on processes can occur locally beneath glaciers that are otherwise classified as 'temperate', and can be
10   evidenced from moraine sedimentology; ii) The conventional association of cold-based glaciers with a lack of glacial landforms should be reassessed since 'temperate' glaciers with cold-based margins may leave an equally subtle geomorphic imprint; iii) cold-based processes, including basal freeze-on, may have a significant role for the dynamics of temperate glaciers that are thinning and retreating to higher altitudes (even during, e.g. the exceptional climatic warming observed since 2000). Implications from these hypothesis suggest that cold-based processes may be more widespread than previously thought, and
15   could influence future glacier behaviour in terms of retreat rate and ice flow velocity.

1.1 Study area

The Hardangerjøkulen icefield (60°55′ N, 7°43′ E) is situated in southern central Norway between a dry continental environment to the east, and the wet maritime climate of the Norwegian coast 150 km to the west (Fig. 1). The icefield covers 72.46 km² and ranges in altitude from 1020 to 1865 m a.s.l. (Andreassen and Winsvold, 2012). The icefield is located on a
20   mountain plateau with an average ice thickness of ~150 m, with ten minor and three major drainage basins where ice thickness is up to 385 m (Fig. 2a-c) (Andreassen et al., 2015). Midtdalsbreen is sourced from a major north-eastern basin, which extends back to the ice divide with the western outlet glacier Rembesdalskåka (Fig. 1b). Midtdalsbreen covers an area of 6.80 km² (Andreassen and Winsvold, 2012) and is ~4.8 km long, descending from 1862 to 1440 m a.s.l. Maximum ice thickness is 280 m in the accumulation area, although the majority of the ablation area is < 120 m thick (Fig. 2c) (Willis et al., 2012; Andreassen
25   et al., 2015). Ice flow velocity was measured as 31 m/yr in 2005-2006 (Giesen, 2009) on the relatively flat lower section of the glacier, but is likely higher in the heavily crevassed area where the glacier descends from the plateau, though no direct measurements are available here. Mass balance was +1.3 m w.e. in 2000 and -0.64 m w.e. in 2001 which equated to an upglacier shift in the equilibrium-line altitude from 1500 m to 1785 m (Krantz, 2002). Length variation measurements at Midtdalsbreen have been taken annually in the ablation season since 1982, close to the middle of the glacier snout near the centre flowline
30   (Fig. 3a and b) (Nesje et al., 2008). Andersen and Sollid (1971) reconstructed the glacier's extent for the years 1934, 1955, 1959 and 1960 to 1968 using photographs and field observations.

In the majority of the study area, the LIA maximum extent is ~2 km NE of the 2014 glacier front. The LIA limit is defined by a distinct boundary between fluted and non-fluted terrain and by fragments of a terminal and lateral moraine that is best preserved to the NW of the current glacier front (Andersen and Sollid, 1971; Reinardy et al., 2013). Moraine fragments, flutes, and lichenometric data provide evidence of steady retreat of Midtdalsbreen from its position at the LIA moraine until its

5   position during the 1930s, by which time the ice front had retreated ~0.5 km (Andersen and Bjørknes, 1978). Moraines formed after 1934 are limited, but glacifluvial terraces were deposited (Andersen and Sollid, 1971). Andersen and Sollid (1971) interpreted the lack of moraines during the 1930s and 1940s as indicative of rapid retreat. This may relate to the significant climatic shift that was observed in the latter part of the 1920s in western Norway, when average annual temperatures increased by 0.7°C and notably warm spring-summers occurred in 1930, 1933, 1937 and 1947 (Nordli et al., 2003). While no geomorphic

10   record of moraines deposited during this time remain, flutes are preserved along the edges of a sandur in the glacier foreland. Between 1955 and 1968 only relatively minor (< 20 m) annual retreat took place (Andersen and Sollid, 1971). The glacier margin fluctuated annually by approximately ± 10 m during the 1980s but during the 1990s a minor (< 25 m/yr) readvance of Midtdalsbreen took place (Nesje et al., 2008), which overrode (and thus destroyed) any moraines deposited during the previous decade (Fig. 3a and b). During more recent periods of warmer climate and rapid glacier retreat (up to 31 m in a single year)

15   from 2001 to 2010, minor moraines continued to be deposited at the glacier snout (Reinardy et al., 2013). Retreat of the ice margin has continued from 2011 to 2018, with the exception of 2015 when the glacier advanced 19 m. By 2016 it had however retreated 21 m behind its 2015 position (Fig. 3a and b).

2 Methods

Ground-penetrating radar (GPR) methods have been widely used to image the beds and internal structures of glaciers, and are

20   particularly useful for identifying englacial water inclusions within an otherwise frozen ice mass (Plewes and Hubbard, 2001). The englacial thermal boundary of multiple glaciers has been mapped using GPR in several previous studies (Holmlund and Eriksson, 1989; Björnsson et al., 1996; Ødegård et al., 1997; Murray et al., 2000; Copeland and Sharp, 2001; Pettersson et al., 2003; Gusmeroli et al., 2010; 2012). The technique is effective since liquid water within temperate ice causes radar energy to be backscattered, in contrast to cold-ice which has no water inclusions and therefore appears transparent (Gusmeroli et al.,

25   2012). Water inclusions scatter energy because they represent a contrast in dielectric permittivity, and thereby in GPR wave speed. The wave speed in cold-ice is typically assumed to be ~0.17 m/ns (Bælum and Benn, 2011), which can be slowed by ~4% with the introduction of 1.4% liquid water (Endres et al., 2009). Regions of warm ice can therefore be mapped by identifying regions of enhanced radar scattering, with the cold-temperate transition surface (CTS) inferred at the transition from transparent to chaotic regimes.

Forty-three kilometres of GPR line data were collected during April 2014, primarily around the Midtdalsbreen snout (Fig. 4a-c), when the glacier was snow covered. Dense coverage at the glacier snout consists of grids with 5 m line spacing, facilitating 3-D imaging across the whole snout (Fig. 4c). Away from the marginal zone, coverage is limited to sparse profiles, and a series

of isolated acquisitions were undertaken around the Midtdalsbreen/Hardangerjøkulen ice divide and two nunataks on the eastern and western side of the glacier (Figs. 4a and b).

All GPR data were acquired with a Malå Geosciences Rough Terrain Antenna (RTA) of 50 MHz centre-frequency. The RTA
5 is a bistatic system, in which transmitting and receiving antennas are mounted in parallel end-fire orientation at 3.5 m separation. A GPR trace was recorded every 0.5 s. GPS positions were recorded by a backpack-mounted antenna, offering +/- 4 m positional accuracy; a GPS position was recorded every 1 s. Throughout, the midpoint of the RTA was 6.2 m behind the operator (motivating a positional correction in later processing). Acquisition of the 3-D grids was conducted on foot, with each line separated by ~5 m and an along-profile trace sampling interval of 0.47 ± 0.12 m. For the longer lines, the RTA was towed
10 behind a snowmobile at somewhat higher speed, resulting in a trace sampling interval of 0.78 ± 0.18 m. For 50 MHz energy propagating in ice (velocity ~ 0.17 m/ns), with a wavelength of 3.4 m, our along-line sampling interval satisfies the quarter-wavelength resolution criterion of Grasmueck et al. (2005), but grids may be spatially aliased in the cross-profile direction. Our limit of vertical resolution is ~ 0.85 m, although our estimates of ice thickness are likely more accurate than this given the typically clear first-break arrival times.

Data pre-processing was conducted in Sandmeier ReflexW and comprised bandpass filtering (corner frequencies of 15-30-70-140 MHz) and static corrections to synchronise first-break travel-times to 11.7 ns (the travel-time of the direct airwave across the 3.5 m of antenna offset). From these data, regions of cold and warm ice were identified and geo-located using GPS positions corrected for the offset between the GPS receiver and the midpoint of the RTA system (in the direction of survey azimuth
20 along each profile). The CTS was identified from unmigrated data since it is most effectively defined when diffraction hyperbolae remain present in the data (e.g., Fig. 5a). Migration was applied thereafter to establish the geometries of the snow cover and glacier bed. To facilitate migration, the trace spacing in all profiles was regularised in Mathworks Matlab; traces in the marginal grids were also interpolated onto regular grids with sample dimensions of 0.2×5 m. Data were then passed back into ReflexW and Kirchhoff migrated with a constant ice velocity of 0.17 m/ns (Bælum and Benn, 2011). A variable velocity
25 field was unavailable since common midpoint (CMP) acquisition is not feasible with the RTA. Nonetheless, the majority of diffraction hyperbolae in our dataset were appropriately characterised by this velocity and became well-focused. Travel-time picks of the glacier surface (i.e., the base of the snow cover), glacier bed and CTS were gridded using a minimum curvature algorithm in Golden Software Surfer. Grids targeting the glacier snout are gridded at 2×1 m resolution, whereas those made for the whole glacier have resolution 20×20 m. Depth conversions assumed a snow velocity of 0.23 m/ns (Holbrook et al.,
30 2016) and an ice velocity of 0.17 m/ns. Considering typical variations about these ranges, we estimate that depth conversions are accurate to ± 2 m, which is small with respect to the large-scale variation we observed.

Geomorphological mapping was carried out updating and expanding on previous geomorphological maps from the area by Sollid and Bjørkenes (1978) and Reinardy et al. (2013). Sedimentological analysis, following procedures outlined by Evans

and Benn (2004), was also used alongside the geomorphological mapping to provide additional information on sediment transport and on the processes leading to landform genesis.

3 Results

The most widespread glacial landform across the foreland of Midtdalsbreen are flutes, sets of which can be traced from the

5 LIA limit to the current glacier margin where they emerge from under the glacier snout (Fig. 7). Frozen flute ridges are normally found incorporated into the basal ice layer at the glacier snout. Flutes at the ice front tend to be <60 cm in height, up to 40 m in length and <1 m wide and composed of diamicton. They are observed in particularly high concentrations in the central foreland, lower concentrations at the NW glacier margin, and are absent from the SE glacier margin (Fig. 7). The minor moraines at the NW glacier margin, detailed in Reinardy et al. (2013), are considered here to highlight the contrast of the SE

10 glacier margin that consists of hummocky dead-ice topography (controlled moraine; sensu Evans, 2009) and de-iced hummocky moraine resulting from high debris concentrations on the SE glacier snout (Figs. 7 and 10c). Controlled moraines consist of sharp-peaked mounds and ridges, ranging in size from ~1 m up to 5 m in height and consist primarily of layers of sorted to well-sorted sands and gravels. Detailed sedimentological analysis of the controlled moraines at Midtdalsbreen will be the topic of an upcoming paper.

15 ### 3.1 Distribution of cold-ice

Two distinct responses, transparent and chaotic, can be seen in the GPR data (Figs. 5a, b and 6a-d), corresponding respectively to ice with and without inclusions of liquid-water. In the majority of cases, the CTS is sharply defined (Figs. 5a, b; 6a, b and d); the glacier bed appears as a highly reflective horizon until 500 m from the snout, where ice thickness exceeds 70 m and signal-to-noise ratio becomes poor (owing to a combination of signal attenuation and/or interference with scattered energy,

20 Fig. 6a and b). At the snow-covered glacier snout it was also possible to estimate where the glacier margin was located from GPR data and compare this to measured length variation data carried out in the late summer of 2014 (Andreassen et al., 2015). The distribution and thickness of cold-ice in the ablation area is shown in Figure 7 interpreted from the GPR data.

Results from the GPR data indicated that all surveyed areas had a cold upper layer 8 m thick probably caused by cold conditions

25 during the winter and then insulated by snow-covered during spring (April) when collection of the GPR data took place (Figs. 5a, b and 6a). Areas of the glacier that have excess cold-ice, or that are frozen at the glacier bed, are here termed cold-ice zones (CIZ) (Fig. 7). Three CIZ were found in the ablation area (CIZ1-3, Figs. 5a, b; 6a-d) and two CIZ in the accumulation area in proximity to the western and eastern nunataks (CIZ4-5, Figs. 4a and 8). CIZ1 is a 40-50 m wide corridor around the glacier snout where ice thickness is ≤10 m (Fig. 5a and b). The GPR data indicate that CIZ1, at the glacier snout, is frozen to the bed

30 (Figs. 5a, b and 7). CIZ2 and CIZ3 are up to 15 m thick (Fig. 6a and b). CIZ2 is located on the eastern margin of the glacier in the ablation area (Fig. 7). Ice thickness here is < 40 m and thus CIZ2 may extend to the glacier bed causing basal freeze-on, though further GPR surveys (including CMP surveys) are necessary to confirm this assumption (Fig. 6b and c). CIZ3 occurs

approximately half way up the length of the glacier and appears to extend across its full width (Fig. 7). This is above the point where the GPR is able to image the glacier bed (Fig. 6a) due to the radar frequency, but previous studies indicate total ice thickness here to be between 50-100 m (Fig. 2c) (Willis et al., 2012; Andreassen et al., 2015). We therefore assume that CIZ3 does not extend to the bed of the glacier except at its eastern margin (Fig. 6d). Lastly, two nunataks in the western and eastern

5   accumulation area of the glacier have even larger cold margins, CIZ4 and CIZ5 respectively, which extend 100 m out from the nunataks and extend to the glacier bed at ~ 30 m depth (Figs. 4a and 8).

4 Interpretation - Sediment transport and depositional processes at the glacier snout

The results from this study demonstrate how sedimentological and geophysical data can be effectively assembled to produce a detailed interpretation of contemporary and palaeoglacier dynamics. The identification of a cold-based snout and a number

10   of other CIZs in both the ablation and accumulation areas of Midtdalsbreen supports the earlier interpretation of a cold-based thermal regime at the glacier snout made from detailed examination of the sedimentary architecture of the moraines (Fig. 5a and b) (Reinardy et al., 2013). The sedimentology of the moraines indicated basal freeze-on during winter advance and melt-out of stacked sediment slabs forming minor moraines during spring-summer retreat. However, while Reinardy et al. (2013) estimated that the cold-based snout extended around 10 m up-glacier from the glacier margin at the end of the summer, this

15   study indicates that from the glacier margin at least 40 to 50 m of the glacier snout is frozen to its bed during the spring (Fig. 5a and b). This is likely facilitated by continued thinning of the glacier snout, allowing the winter cooling to penetrate through to a larger area of the bed. In addition, photographic monitoring of the ice front by Giesen (2009) during 2005-2008 shows that the glacier snout remains snow covered until late summer (end of July beginning August) while ablation areas further up glacier are snow free by the start of July (Fig. 9). Late-lying seasonal (i.e. present until the beginning of August) snow patches

20   in this location would also allow effective basal freeze-on at the glacier snout. Giesen (2009) relates this pattern of snow melt to the greater snow depths due to drifting at the relatively steep glacier snout. Similar to CIZ1, both CIZ2 and CIZ3 in the ablation area correspond to parts of the glacier that experience snow drifting facilitated by glacier topography and correspond to areas of late-lying seasonal snow cover in multiple years (Fig. 10a-c). Areas of late-lying snow and reduced ablation have previously been linked to thickening of a cold surface layer by insulating the ice from warmer air temperatures on Storglaciären

25   in northern Sweden (Gusmeroli et al., 2012). Aerial photos of Midtdalsbreen were available for 2004, 2007, 2010, 2013, 2014 and 2017 and all show late-lying snow patches in approximately the same position on Midtdalsbreen until at least the beginning of August, although in some years, such as 2007, these snow patches remained until mid-August (Fig. 10b). This suggests the late-lying seasonal snow patches have remained stable during at least the summer melt seasons of 2004, 2007, 2010, 2013, 2014 and 2017, implying that the thermal regime may also have remained relatively stable during the last decade (Fig. 10a-c).

30   By the beginning of September most of these snow patches have melted away (Fig. 4a). On Storglaciären, Pettersson et al. (2007) showed that net surface mass balance and vertical ice advection are the dominant controls on cold surface layer stability and thickness. Testing this hypothesis for Midtdalsbreen requires measured vertical ice velocity profiles, which to date are not available for this glacier.

The dense GPR survey grids over the glacier snout show that the boundary between the frozen bed and temperate bed is distinct and relatively sharp (Fig. 5a and b). This result contrasts with previous interpretations made from numerical models of ice-marginal thermal regime, which suggest that the boundary between frozen and thawed areas of the bed is likely to be diffuse

5 (Mooers, 1990). The nature of the boundary between the frozen to non-frozen bed at Midtdalsbreen is significant because it likely exerts influence over the sedimentological characteristics of moraines deposited at the glacier snout. The sharp thermal boundary could elevate subglacial debris within distinct layers of debris-rich ice which then melts out to produce the stacked sediment slabs observed within the minor moraines (Reinardy et al., 2013). If the frozen bed area at the glacier snout expands, then an increasing amount of debris within debris-rich basal ice on the underside of the glacier snout would be available for

10 the incremental thickening of the moraine ridges during spring to early summer melt. However, observations made by the authors of the glacier foreland in the summer of 2013 and 2017 show the sediment to be highly water-saturated. This likely counteracts preservation and build-up of anything larger than minor moraine ridges, considering that the moraines are composed almost exclusively of subglacial traction till. In addition to sediment availability, the construction of moraines is likely influenced by how far the ice margin advances during the winter months; minor advances probably only facilitate

15 construction of minor moraines such as the annual moraines described by Reinardy et al. (2013). Thus, the combination of limited preservation potential and only minor winter readvances indicates that evidence for extensive basal freeze-on and cold-ice processes may be difficult to identify within the glacier foreland. Flutes have also previously been described as having a limited preservation potential (Benn and Evans, 2008), but within the timeframe extending back to the LIA, they are relatively well preserved on the foreland of Midtdalsbreen. Their presence has previously been linked to glaciers with extensive cold-

20 based margins (e.g. Gordon et al., 1992; Roberson et al., 2011). The observation that flutes at Midtdalsbreen are frozen to the glacier bed at the glacier snout suggests that flutes either actively form beneath cold-based ice, or that they form beneath warm-based ice and are subsequently frozen-on. For polythermal glaciers, Eklund and Hart (1996) suggest flutes form beneath warm-based ice and are then frozen as the CTS passed over them. Thus, for flutes to form at polythermal glaciers by this mechanism, migration of the CTS is required, indicating that flutes are produced only at retreating and thinning glaciers as is the case with

25 Midtdalsbreen.

While the north-western part of the glacier foreland contains minor moraines, the south-eastern part of the glacier margin is debris-covered, resulting in an uneven ice surface formed of debris-covered ridges and hummocks (Figs. 7 and 10c). Melting at the margin results in debris-cover thickening, characterised by minor debris flows and sorting of sediment by meltwater into

30 areas of sand and finer grained material. This zone of controlled moraines (cf. Evans, 2009) is in stark contrast with the area of minor (possibly annual) moraines in the north-western part of the foreland, indicating a range of distinct processes of sediment delivery to the ice margin across the Midtdalsbreen foreland (Fig. 7). The identification of CIZ2 directly up glacier from the controlled moraines (Fig. 7), which possibly extends to the glacier bed, may contribute to the delivery of sediment to the ice surface and the development of controlled moraines in this area. Evidence of debris elevated from an englacial (and

likely subglacial) position to the ice surface can been seen in the GPR data directly up glacier from CIZ2 (Fig. 6b). The GPR data also indicate what appears to be a small mound on the surface of the glacier that occurs where the interpreted englacial debris layer reaches the ice surface. This mound may well be a small controlled moraine buried by snow at the time the GPR survey took place in April 2014. In 2013 and 2017, debris was observed within a crevasses emerging from an englacial source

5   about 200 m up glacier from the then ice margin. This emerging debris contributed directly to the zone of supraglacial debris at the south east glacier margin (Fig. 10c), but was not the only source. Previous studies have shown that debris-rich basal ice in polythermal glaciers can be elevated to englacial and supraglacial positions via compressional glacitectonics at the glacier snout at the CTS, where temperate ice is being thrusted over cold stagnant ice or as a result of recumbent folding within the glacier (Glasser et al., 2003). CIZ2 would have provided these conditions for the formation of controlled moraines at

10   Midtdalsbreen. The existence of cold-ice even during the late summer season on the south-eastern margin of Midtdalsbreen is supported by observations of frozen-on sediment slabs 10 m below the ice surface within a cavity at the glacier snout (Reinardy et al., 2013).

4.1 Cold-ice zones in the accumulation area

While several studies have indicated parts of ice sheets to be frozen to their bed (Bentley et al., 1998; Gades et al., 2000), and

15   temperate glaciers or ice caps and icefields with a frozen marginal zone (Waller et al., 2012 and references therein), here we show that nunataks in the accumulation area of temperate glaciers can also be surrounded by a cold-ice zone that extends to depths greater than the 8 m thick surface cold layer. For Midtdalsbreen, these zones are relatively large extending up to 100 m out from the nunatak margins and down to the glacier bed (CIZ4 and CIZ5, Fig. 8). The area around the eastern nunatak showed some of the highest rates of ice thickening across the whole Hardangerjøkulen icefield from 1961 until 1995

20   (Andreassen and Elvehøy, 2001). Therefore, we hypothesise that the cold-ice zones around the nunataks are relatively recent features, directly related to icefield thinning over the last two decades (Kjollmoen et al., 2016). While the GPR data did not penetrate to the bed over the rest of the plateau, the presence of temperate ice at the surface of the thickest parts of the icefield indicates that these areas are entirely temperate. However, it is possible that additional parts of the icefield also are cold-based, for example close to the ice divide where ice thickness is < 50 m (Figs. 1b; 2a and b) or where additional nunataks are present.

25   Further GPR investigations are needed to confirm this. As the icefield continues to thin (Andreassen et al., 2016), probably accompanied by a corresponding decrease in ice flow velocity, cold-ice and/or frozen bed conditions may become more widespread in the future. This process appears to already be taking place around the eastern nunatak. Thus, it is likely that cold-ice zones may become more extensive over the icefield and upper accumulation areas but conversely possibly less extensive in the ablation area where late-lying seasonal snow patches are likely to decrease in size and area in future decades.

30   5 Discussion

In the accumulation area of Midtdalsbreen, basal freeze-on may influence glacier dynamics via increased lateral drag. Annual average surface ice flow velocity was measured as 33 m/yr at the ELA during 2005-2007 (Giesen, 2009) but is likely

considerably less in areas where the ice is frozen to its bed such as CIZ4 and CIZ5 (Fig. 8). If extensive cold-ice zones exist, this could affect ice flow by means of increased shear between temperate and cold-ice zones within the glacier, and/or higher drag along cold-based margins (Åkesson, 2014). This may cause glacier slowdown and an associated reduced erosive power of the glacier (e.g. Koppes et al., 2015). Changes to glacier thermal regime and resulting altered glacier dynamics may also

5   bias glacier reconstructions based on proglacial lake sediments (Åkesson et al., 2017), which assume a relationship between eroded glacier flour and glacier variability (e.g. Hallet et al., 1996). Cold-ice is also more viscous ("stiffer") and deforms less readily than temperate ice (Cuffey and Paterson, 2010, p. 12). Therefore, cold-ice zones in predominantly temperate glaciers may act as a negative feedback on ice flow in a warming climate. If cold-ice zones become more widespread due to glacier thinning, as suggested by the likely recently formed cold-ice zones in the accumulation area at Midtdalsbreen, we may see a

10   future deceleration and further preferential thinning of previously temperate glaciers like Midtdalsbreen. Changing ice properties and its impact on large scale flow is also relevant on the ice sheet scale; variable ice rheology has been suggested to have caused deceleration of the Greenland Ice Sheet during the Holocene (MacGregor et al., 2016).

Basal freeze-on and areas of cold-ice not only affect rate and style of glacial retreat, they may also have a significant impact

15   in areas of ice-damned lakes. The outlet glacier Rembesdalskåka, draining Hardangerjøkulen to the south west (Fig. 1b), previously had an ice-dammed lake, Nedre Demmevatn. Written sources describe late summer catastrophic jökulhlaups back to the eighteenth century that likely occurred at least partially via subglacial drainage pathways (Elvehøy et al., 2002). Since 2014, Nedre Demmevatn has drained four times, once in 2014, twice in 2016 and once in 2017 (Jackson and Ragulina, 2014; Andreassen et al., 2017). These events drained the entire lake within hours (Jackson and Ragulina, 2014). With the presence

20   of a frozen glacier margin any subsequent water drainage would be prevented from draining through subglacial channels or cavities. While Nedre Demmevatn is currently completely drained, a continual source of meltwater may fill the lake again in the future and further investigation is needed to determine if basal freeze-on mechanisms are responsible for initial damming of the lake. If so, possible future changes in thermal regime at Rembesdalskåka and indeed other ice-damned lake sites needs to be considered in assessments of future risks from jökulhlaups.

This study has wider significance in terms of the glacier thermal regime and how thermal properties relate to the glacial geomorphological record. Firstly, glaciers that are defined as temperate may contain considerable areas of cold-ice. While some studies have previously shown polythermal glaciers with a frozen margin (e.g. Pettersson et al., 2007), here we show that in addition to a cold-ice margin during spring within a temperate glacier, parts of the upper accumulation area can also

30   contain cold-ice, if ice is relatively thin, for example in proximity to nunataks. Both these characteristics apply to many plateau-type ice caps, icefields, and outlet glaciers. Thus, it is likely that cold-ice processes, including subglacial freeze-on processes, could be more widespread than previously thought. Based on numerical modelling of ice flow and surface mass balance, Giesen and Oerlemans (2010) suggest that the Hardangerjøkulen icefield could disappear by year 2100. However, it is difficult to predict the future distribution of cold-ice for both glaciers and icefields. Thinning of cold-ice layers has been linked to

regional climate change (Gusmeroli et al., 2012), and results from this study suggest that CIZ's are likely to form below late-lying seasonal snow patches on the glacier surface which will presumably also disappear under warming climate trends. However, thinning of ice, possible ice deceleration and retreat of glacierized areas to altitudes where permafrost becomes more prevalent, may increase the presence of cold-ice and therefore cold-based processes at least for a limited period during overall

5 retreat (Rippin et al., 2011). Spatial and temporal variability in ice rheology and particularly thermal regime is currently poorly resolved in numerical ice flow models. Many modelling studies currently do not account for changes to the thermal regime in their predictions, nor the potential knock-on effects on ice dynamics discussed above. Climate forcing is the main control on glacier surface mass balance but the heterogeneous thermal conditions found at Midtdalsbreen, and potential presence at other similar temperate glaciers, calls for improved model assessments of future evolution of thermal conditions and associated

10 impact on glacier dynamics and mass balance.

6 Conclusions

The potential role of cold-ice processes at the margins of palaeo-ice sheets (e.g. Boulton, 1972; Budd et al., 1976; Moran et al., 1980; Mooers, 1990) and frozen-bed patches below palaeo-ice sheets (e.g. Kleman and Borgström, 1994; Astakhov et al.,

15 1996; Kleman and Häterstrand, 1999) has long been recognised. In addition, basal freeze-on processes have also been used to interpret the stagnation of ice streams (Christoffersen and Tulaczyk, 2003), glacier surge dynamics (Murray et al., 2000) and tunnel valley infill (Kristensen et al., 2008). However, the role of cold-ice processes within temperate glacier systems is relatively unknown. The outlet glacier Midtdalsbreen is a temperate glacier system, but in this study we identify at least five significant areas of cold-ice zones. CIZ1-3 in the ablation area correspond to parts of the glacier surface covered by late-lying

20 seasonal snow patches that occur in approximately the same location during multiple years. This study demonstrates that not only the thin margins of temperate glaciers can have cold-ice frozen to the bed; more extensive and thicker cold-ice zones can exist around nunataks in the glacier accumulation area. We find that CIZ4-5 are located around nunataks on the western and eastern sides of the upper reaches of Midtdalsbreen. Both extend up to 100 m out from the nunataks, are up to 30 m thick and appear to be frozen to the bed. This is significant as it likely influences ice flow velocity by enhanced lateral drag in these

25 areas. Here we have also highlighted that the presence of CIZs may have produced a variety of characteristic landforms on or at the glacier snout. Substantial winter freeze-on across a 40 to 50 m wide zone at the glacier snout has led to the deposition of minor moraines on the north-western glacier margin while contrasting controlled moraines have been deposited at the south-eastern margin possibly in relation to a zone of thick cold-ice (CIZ2) directly up ice-flow from the moraines. This observation has significant implications for the interpretation of palaeoglacier dynamics from the geomorphological record as it suggests

30 that multiple, differing landforms can be produced at cold-based margins of temperate glaciers with multiple implications for debris entrainment, transport and deposition. In areas where extensive cold-ice has been interpreted to have been frozen to the bed, i.e. cold-based plateau icefields or ice caps and outlet glaciers, the characteristic geomorphic imprint has previously been highlighted (e.g. Dyke, 1993; Rea and Evans, 2003; Evans, 2010; Pearce et al., 2014; Boston et al., 2015; Boston and Lukas,

2016). However, considering both the minor and controlled moraines at Midtdalsbreen are, or were ice-cored, their preservation potential is very limited (e.g. Lukas et al., 2005; Reinardy et al., 2013). Thus, cold-based processes may also be far more widespread in the temperate palaeoglacier record than currently accounted for. With predicted continued warming in the upcoming decades, areas of cold-ice may decrease due to reduced or disappearing late-lying seasonal snow patches.

5 However, continued thinning of the icefield and outlet glaciers and probable reduction in ice flow velocity as well as retreat of the glacier to higher altitudes may also increase areas of cold-ice and processes such as basal freeze-on in the short to medium term at least.

Author contribution. B. Reinardy prepared the manuscript with contributions from all co-authors and all co-authors were also involved in the construction of figures and analysis of GPR and geological data sets. Fieldwork and collection of GPR data
10 was carried out by B. Reinardy, A. Booth, A. Hughes and J. Bakke. Geomorphological mapping was carried out by B. Reinardy, C. Boston and D. Pearce.

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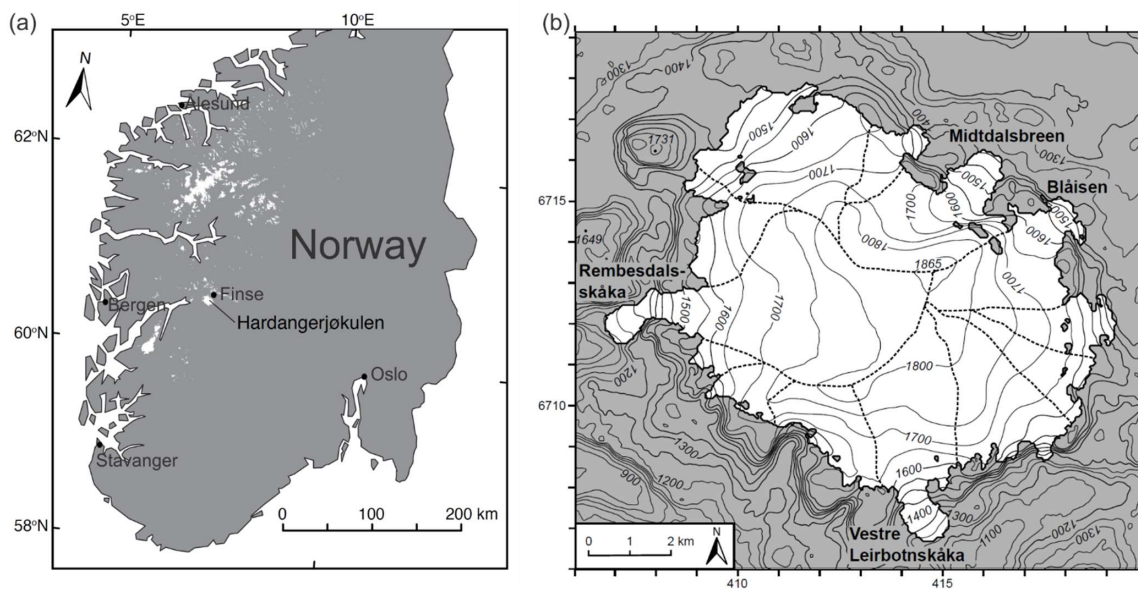


Figure 1: (a) Map of south central Norway with major ice bodies indicated in white (modified from Giesen, 2009). (b) Hardangerjøkulen ice cap created from a 1995 digital elevation model by Statens Kartverk (modified from Giesen, 2009). The reference system is UTM zone 32 (EUREF89) and the contour interval is 50 m. The dashed lines indicate the estimated extent of the
5   **drainage basins of the different outlet glaciers (Andreassen and Elvehøy, 2001).**

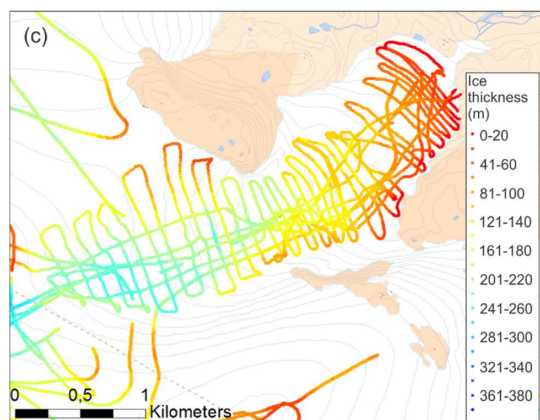
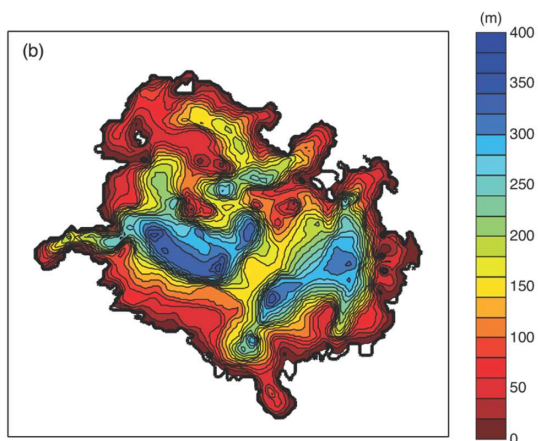
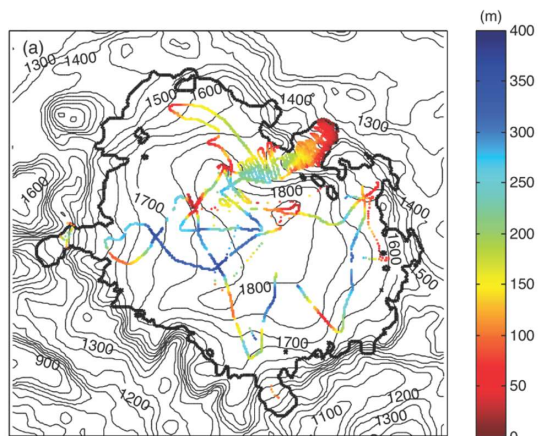



Figure 2: (a) Ice thickness data from studies 1963-2008. (b) Ice thickness calculated from the surface DEM from 1995 and thickness measurements from studies 1963-2008. (c) Detailed ice thickness data from Midtdalsbreen. Figures from Giesen (2009) and data from K. Melvold, NVE.

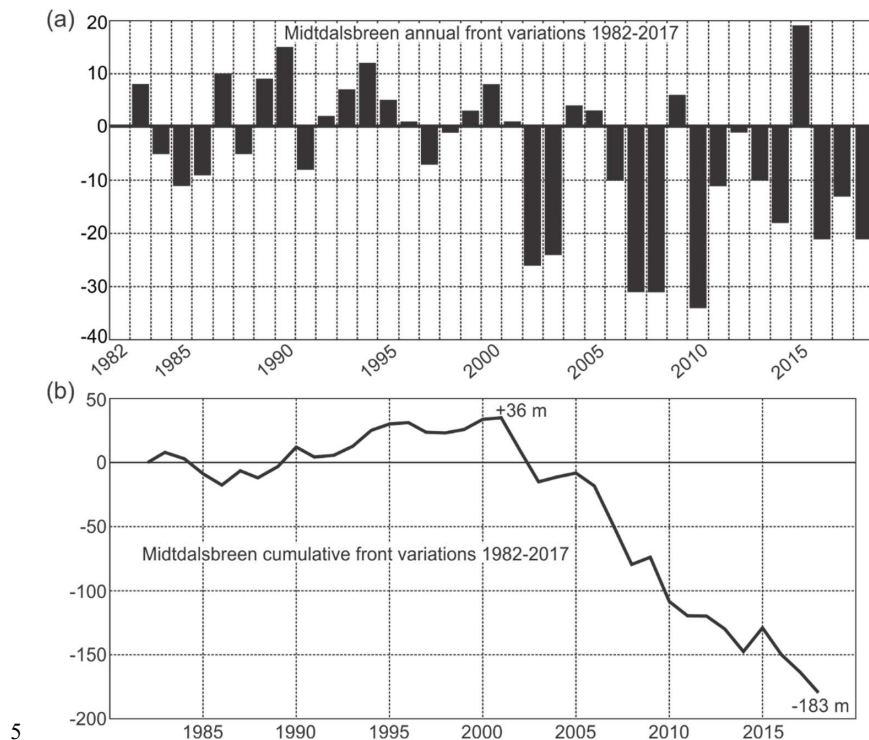



Figure 3: (a) Midtdalsbreen annual front variations from 1982-2018 and (b) cumulative front variations from the same time period. Date collected and compiled by A. Nesje.

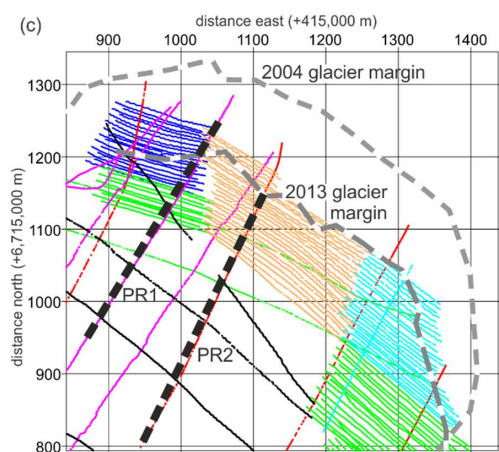
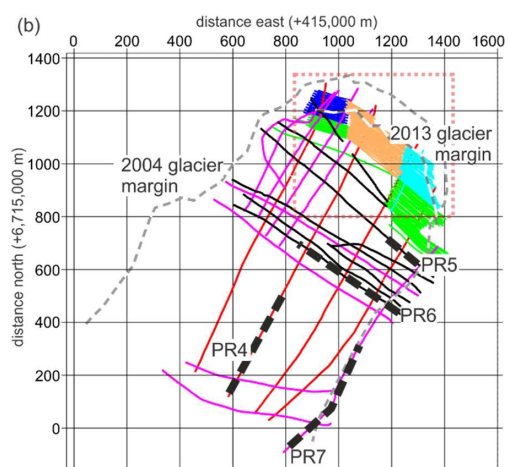
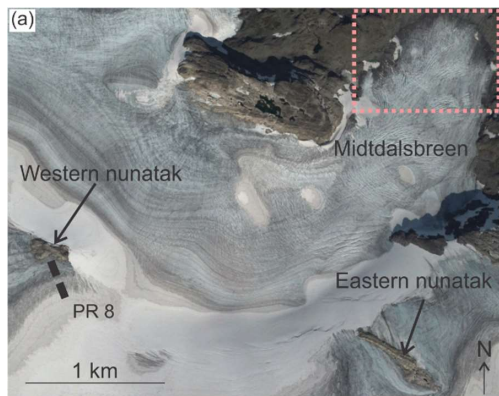
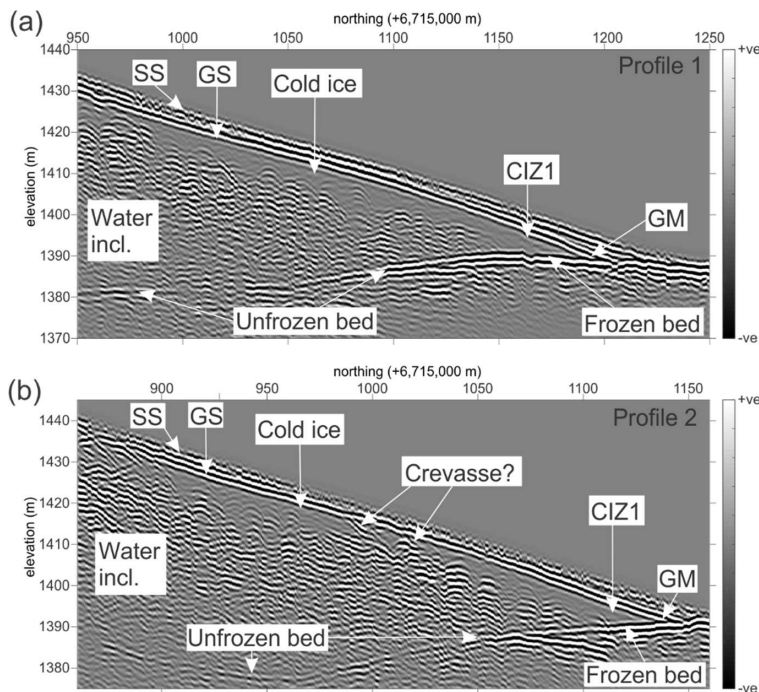



Figure 4: (a) Locations of western and eastern nunataks in the upper accumulation area of Midtdalsbreen, from Google Earth imagery taken on the 4th September 2014. Note by this time in the melt season nearly all patches of snow have melted from the glacier surface. Dashed black line shows location of profile (PR) 8 shown in figure 8. Pink dashed box shows **(b)** locations of GPR surveys on Midtdalsbreen (coloured areas and solid lines), between 20th-26th April 2014, with the 2004 and 2013 glacier margin
5   (dashed grey) digitised from Google Earth. Thick black dashed line indicates locations of profiles (PR) 4 to 7 displayed in Fig. 6a-d. Pink dashed box shows **(c)**, enlarged window of dense surveys (coloured lines) on the glacier snout. Thick black dashed line indicates locations of profiles (PR) 1 and 2 displayed in Fig. 5a and b respectively.

10   **Figure 5:** See Fig. 4c for locations of profiles. **(a)** Representative profile (Profile 1) parallel to ice flow from the GPR dataset over the snout and glacier margin with upper cold-ice layer and cold-ice zone (CIZ) 1. Also labelled is the highly reflective glacier bed with part of the glacier snout frozen to the bed labelled CIZ1 and further up-flow the glacier remains unfrozen to the bed. Temperate ice gives a chaotic response in contrast to cold-ice that gives a transparent response in the radar data. GS = glacier surface; GM = glacier margin; SS = snow surface. **(b)** Profile 2 parallel to ice flow also from the area around the glacier snout.

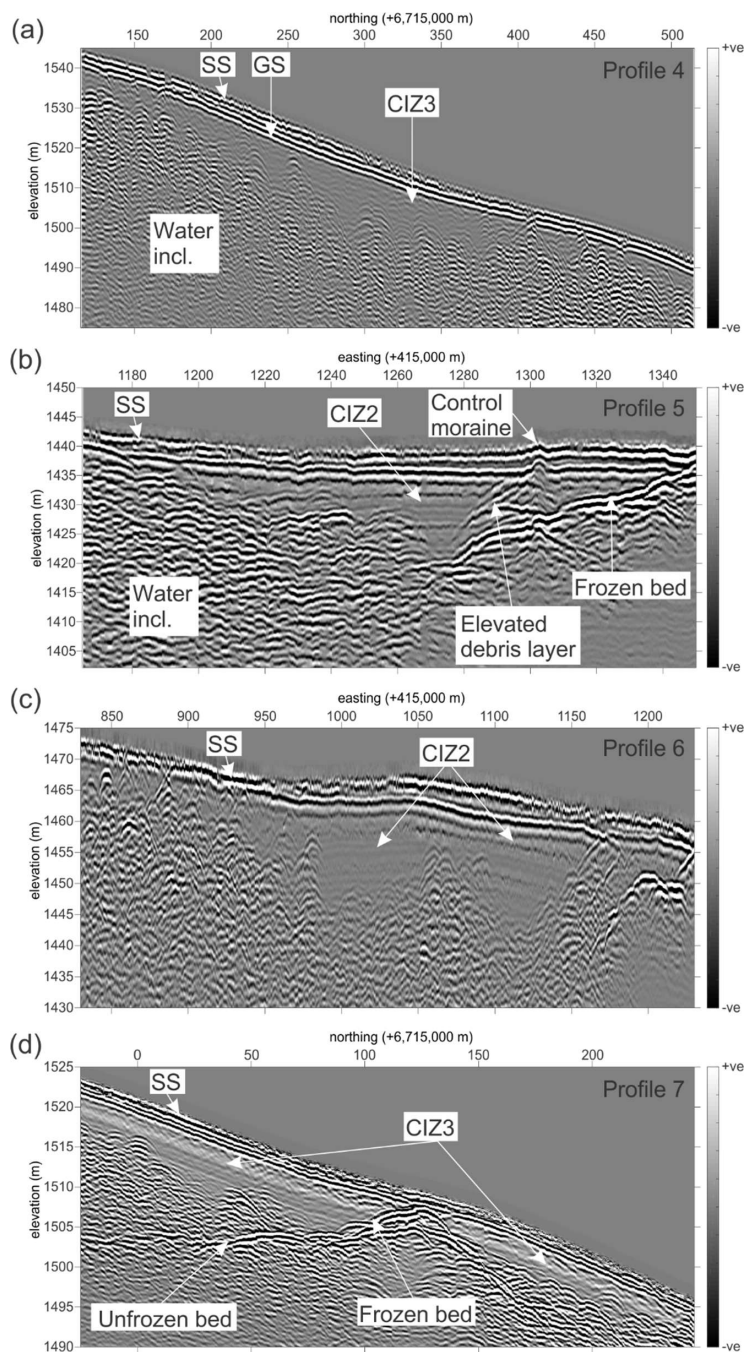



Figure 6: See Fig. 4b for locations of profiles, (a) GPR Profile 4 parallel to ice flow showing cold-ice zone (CIZ) 3 which is ~15 m thick. GS = glacier surface; SS = snow surface. (b) GPR Profile 5 perpendicular to ice flow indicating CIZ2 and a possible debris layer extending from close to the glacier bed up to the ice surface where there is also a small mound, possibly a controlled moraine. The south east lateral margin of the glacier is located to the far right hand side of the profile. (c) GPR Profile 6 perpendicular to ice flow indicating CIZ2 ~15 m thick. (d) GPR Profile 7 parallel to ice flow indicating CIZ3 ~15 m thick with well-defined cold-temperate transition surface (CTS) and highly reflective glacier bed.

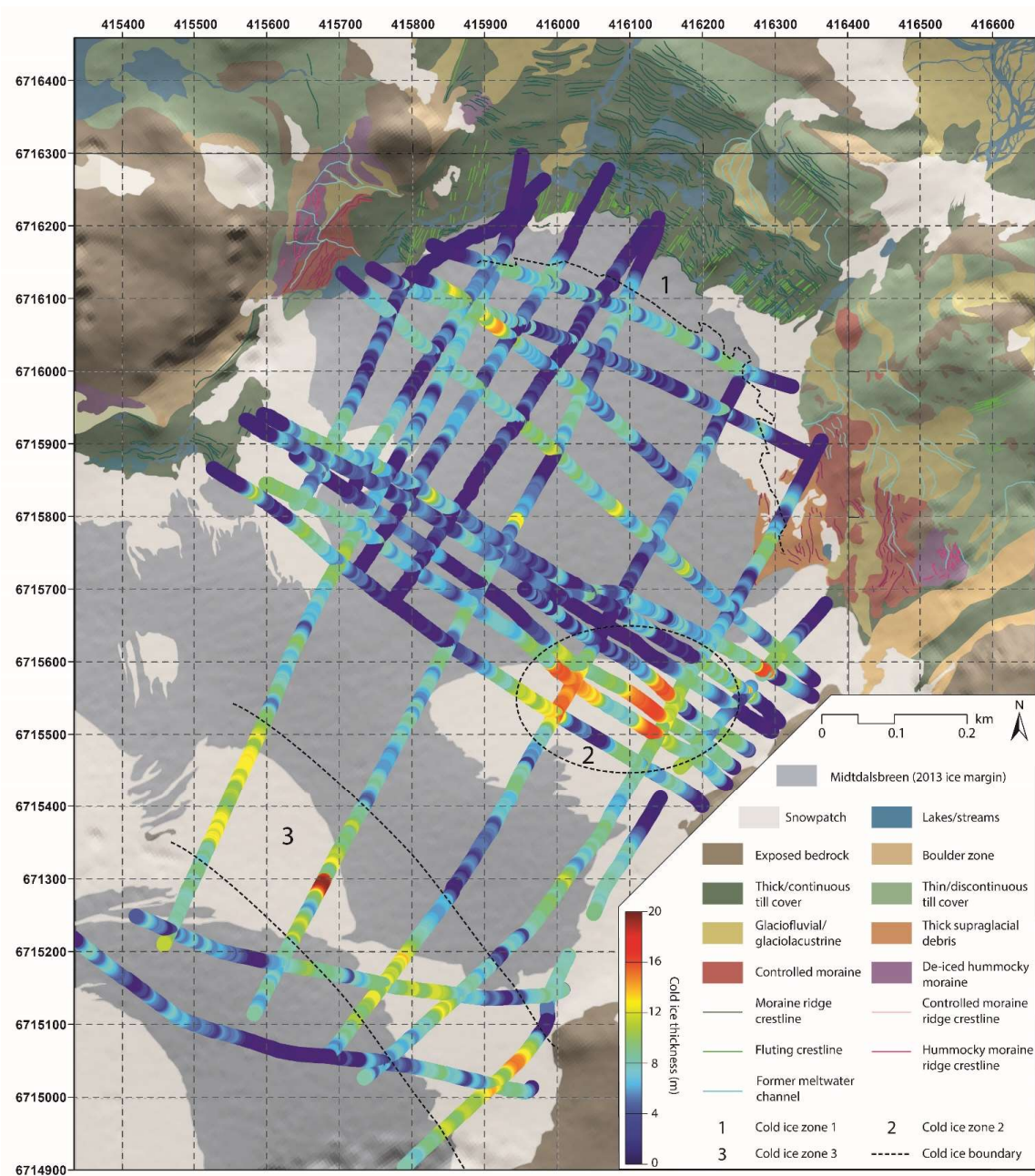



Figure 7: Estimates of cold-ice thickness on Midtdalsbreen from selected GPR tracks. Three cold-ice zones, CIZ1-CIZ3 are indicated. A corridor of 40 to 50 m of frozen basal ice is present at the glacier snout (CIZ1). Also included is a detailed geomorphological map of the glacier foreland which highlights the distinct difference between the NW and SE part of the foreland linked to the varying thermal regime measured in the GPR data.

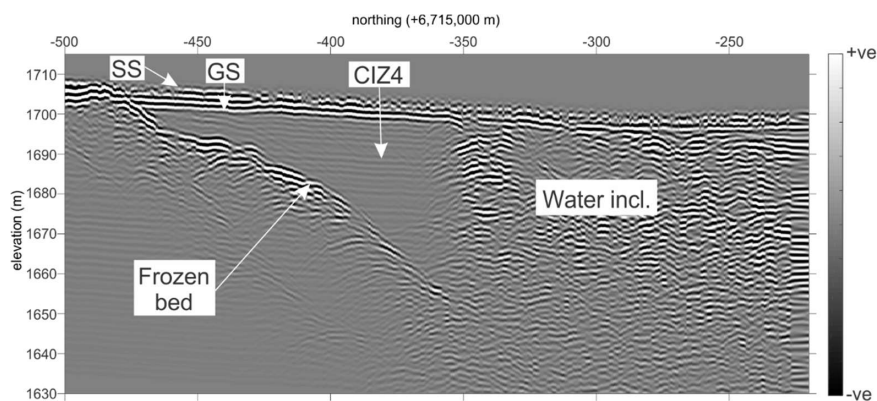



Figure 8: Radar profile 8, see Fig 4a for location. The transparent response of CIZ4 up to 30 m thick and frozen to the glacier bed can clearly be seen before the entire ice column becomes temperate. The strong dipping reflector is the glacier bed, not imaged below the temperate ice. GS = glacier surface; SS = snow surface.

Figure 9: A selection of pictures taken by a camera set up by Giesen (2009) overlooking the Midtdalsbreen ablation area. An automatic weather station where Giesen (2009) collected ice velocity and ablation data mentioned in the text is indicated in the pictures.

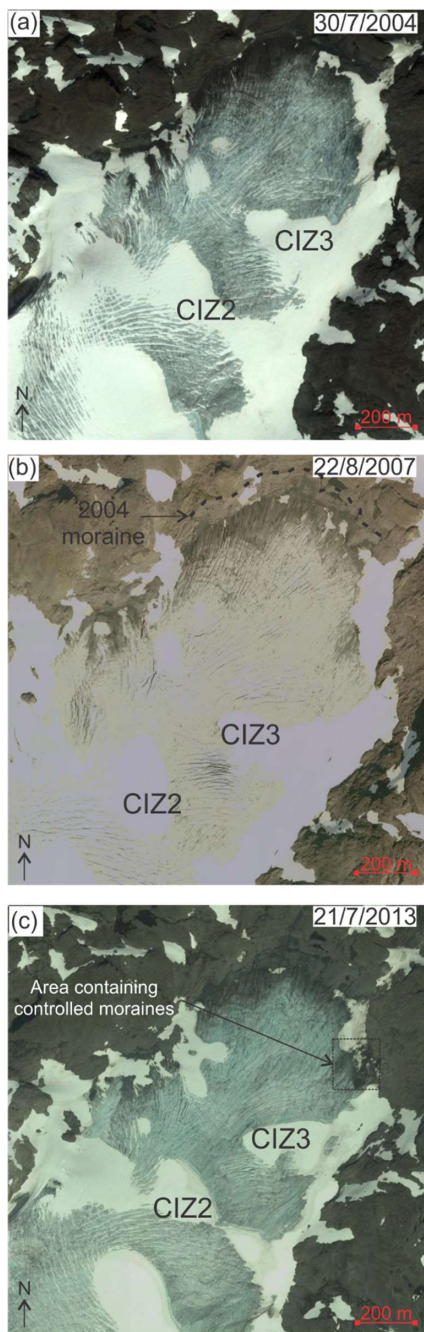



Figure 10: Selection of satellite images during the late summer of 2004, 2007 and 2013 from the ablation area of Midtdalsbreen showing areas of late-lying seasonal snow patches that also correspond to CIZ2 and CIZ3. Sources (a) Google Earth, (b) Norge I Bilder, (c) Google Earth, area of controlled moraines discussed in text also indicated.