Geomorphology 306 (2018) 90–101

Contents lists available at ScienceDirect

Geomorphology

journal homepage: www.elsevier.com/locate/geomorph

Can glacial shearing of sediment reset the signal used for luminescence dating?

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ARTICLE INFO

Article history:
Received 2 August 2017
Received in revised form 19 January 2018
Accepted 19 January 2018
Available online xxxx

Keywords:
Glacial grinding
Shear stress
Luminescence

ABSTRACT

Understanding the geomorphology left by waxing and waning of former glaciers and ice sheets during the late Quaternary has been the focus of much research. This has been hampered by the difficulty in dating such features. Luminescence has the potential to be applied to glacial sediments but requires signal resetting prior to burial in order to provide accurate ages. This paper explores the possibility that, rather than relying on light to reset the luminescence signal, glacial processes underneath ice might cause resetting. Experiments were conducted on a ring-shear machine set up to replicate subglacial conditions and simulate the shearing that can occur within subglacial sediments. Luminescence measurement at the single grain level indicates that a number (albeit small) of zero-dosed grains were produced and that these increased in abundance with distance travelled within the shearing zone. Observed changes in grain shape characteristics with increasing shear distance indicate the presence of localised high pressure grain-to-grain stresses caused by grain bridges. This appears to explain why some grains became zeroed whilst others retained their palaeodose. Based on the observed experimental trend, it is thought that localised grain stress is a viable luminescence resetting mechanism. As such relatively short shearing distances might be sufficient to reset a small proportion of the luminescence signal within subglacial sediments. Dating of previously avoided subglacial sediments may therefore be possible.

1. Introduction

As the Quaternary is typified by growth and decay of ice sheets and glaciers it is hardly surprising that much research has focussed on using geomorphology to reconstruct and model these through time (e.g., Jenson et al., 1995; Dyke et al., 2001; Clark et al., 2012; Toucanne et al., 2015; Hughes et al., 2016). Unfortunately, many of the difficulties and controversies stemming from this can be traced back to uncertainties associated with age either through problematic stratigraphic correlation or through methods attempting to provide specific ages (e.g., Hamblin et al., 2005; Pawley et al., 2008; Gibbard et al., 2009; White et al., 2010, 2016; Lee et al., 2011). Radiocarbon is of limited use as it covers only part of the last glacial-interglacial cycle, and organic preservation within glacial sediments is limited and has a high potential for carbon recycling/contamination (Briant and Bateman, 2009). Uranium series dating and amino-acid racemisation often cannot be applied through lack of suitable material within glacial sequences. As a result, Quaternary scientists largely apply cosmogenic and luminescence dating. The application of cosmogenic exposure dating, although relatively new, has been making a significant contribution to the understanding of ice-sheet fluctuations (e.g., McCormack et al., 2011; Anjar et al., 2014; Davis et al., 2015). However, exposure dating is largely limited to glacially eroded boulders on, for example, moraines and crag-and-tails (e.g., Livingstone et al., 2015) and is complicated by the presence of cold-based ice (Ballantyne, 2010).

Luminescence dating has potential to date events within the last two glacial-interglacial cycles (e.g., Bateman et al., 2011) and is applicable to quartz and feldspars that are almost ubiquitous within preserved glacial landforms and sediments. As such, the method is attractive for gaining glacial chronological frameworks. However, the technique relies on the fundamental premise that at some point between erosion, transport, or deposition, glacial sediment must be exposed to sunlight for a sufficient duration to remove antecedent stored luminescence. Godfrey-Smith et al. (1988) showed that for quartz the optically stimulated luminescence (OSL) signal is reduced to <1% of its original level within 10 s of sunlight exposure. It therefore might be viewed that this is not too hard a criterion to meet, and indeed, King et al. (2014, 2014) showed that sediment redistribution in proglacial settings has a number of opportunities to reset. However, many of the events/sediment requiring dating pertain to subglacial processes and associated landforms in which light exposure is unlikely (e.g., Lamothe, 1988; Rhodes and Pownall, 1994; Fuchs and Owen, 2008). As a result, age overestimation (e.g., Duller et al., 1995; Houmark-Nielsen, 2009) and

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https://doi.org/10.1016/j.geomorph.2018.01.017
0169-555X/© 2018 Published by Elsevier B.V.

ages in saturation or highly variable ages (e.g., Thrasher et al., 2009) can occur.

It has long been established that electrons trapped in defects within the crystal lattice of quartz or feldspar can also be stimulated into releasing luminescence by heat (thermoluminescence or TL) from natural or anthropogenic fires. What is less well established is a third environmental luminescence stimulation mechanism that relies on frictional effects or pressure (McKeever, 1985), which is known to cause triboluminescence or piezoluminescence. In this, as electrons recombine and give off luminescence, so the stored charge depletes, eventually leading to resetting (see Dreimanis et al., 1978; Aitken, 1985; Lamothe, 1988; Toyoda et al., 2000). Studies of sediment found in deep faults have shown that luminescence resetting does occur during earthquake events, but in such cases the ambient temperature is elevated and pressures induced by overburden as well as during movement on the fault are high (Zöller et al., 2009; Spencer et al., 2012). Subglacial temperatures and general confining pressures are much lower than this. Nonetheless, the existence of resetting at the ice-bed has been proposed (e.g., Morozov, 1968; Dreimanis et al., 1978; Lamothe, 1988). More recently, empirical work from the Haut Glacier d'Arolla, Switzerland, by Swift et al. (2011) appeared to show a lowered luminescence signal from subglacial samples when compared to supraglacial sediments. This they suggested was caused by the resetting of sediment during subglacial crushing and grinding (specifically, bedrock erosion and debris comminution).

Laboratory studies of the effects of mechanical crushing on sediment luminescence have generally failed to see an effect (e.g., Sohbati et al., 2011; Rittenour et al., 2012). However, Bateman et al. (2012) reported initial results from a ring-shear experiment in which changes to palaeodose (D_e) were monitored as shearing distance increased. This demonstrated for the first time that changes in stored palaeodose are possible when sediment was placed under a modest pressure (100 kPa) and sheared. They suggested that the average confining pressure applied within the ring-shear apparatus was insufficient alone to cause these changes. Instead, they concluded that stress induced during grain bridging (grain stacks or forced chains consisting of several aligned grains) events was important. They therefore suggested that geomechanical luminescence signal reduction may be a viable alternative mechanism for resetting (referred to as 'bleaching' when performed by light) of glacial sediments. However, the experiment on its own was not conclusive as it was hampered by low quantities of grains showing signs of resetting and high levels of palaeodose scatter. It was also impossible to discern, because of the low palaeodose (~4 Gy) of the sediment used, whether grains were being fully reset or their stored dose just depleted. Finally, the experiment was unable conclusively proved whether the observed changes in palaeodose were caused by pressure (normal stress), shear stress, or other mechanical changes such as localised recrystallization (or the causation and migration of defects within grains).

The aims of this present study were twofold. First, to test the results of Bateman et al. (2012) using an annealed gamma irradiated sample with much higher dose, increased sensitivity, and lower initial D_e scatter. It was hoped such an approach would provide the opportunity to see more effectively whether OSL signal resetting is actually taking place or just that palaeodose is being reduced. Second, using new surface texture and shape data from the Bateman et al. (2012) experiment and the new experiment to better understand the potential mechanisms causing any signal removal.

2. Experimental details

2.1. Sample preparation

The experiment of Bateman et al. (2012; experiment 1) and the new experiment (experiment 2) were based on sediment sampled from a dune field at Lodbjerg, Denmark, studied by Murray and Clemmensen (2001) and Clemmensen et al. (2009). This sediment was originally derived from local, sand-rich glacial till. Actual glacial sediment was not used because of its complex transport history (e.g., Fuchs and Owen, 2008), sometimes poor OSL sensitivity (e.g., Preusser et al., 2007), and mixed lithologies that may be associated with different luminescence properties and behaviour (Rhodes and Bailey, 1997; Rhodes, 2000). Sampling consisted of driving 50+ opaque 20-cm-diameter PVC tubes into the exposed dune face (Fig. S1). The tubes were transported to the laboratory, where the outer 2–3 cm of sand from each tube-end was discarded (thus excluding any grains that may have been exposed to light). Sand was then sieved through a 500 μm sieve to remove extraneous organic material (mostly small rootlets) and homogenised by mixing. Mineralogy was confirmed to be dominantly quartz by mineral-mapping ~100 grains using a Zeiss Sigma field emission analytical SEM equipped with an Oxford Instruments INCAWave detector. Further, laser granulometry confirmed the size distribution to be well-sorted medium sand (M_d = 295 μm, d_{10} = 197 μm, d_{50} = 319 μm, d_{90} = 543 μm).

For the new experiment (experiment 2), sediment was additionally annealed to 500 °C for 1 h to remove any naturally acquired palaeodose and to improve the quartz sensitivity to dose. The sediment was then given a 38.1 ± 1.2 Gy dose using the $Cobalt^{60}$ gamma source at Risø, Denmark. This dose was selected to be of a similar magnitude to what would be expected for a relict glacial deposit from the Last Glacial Maximum (~21 ka). As the annealing and gamma dosing was undertaken in batches, all were thoroughly remixed prior to ring-shear experimentation.

2.2. Shearing in the ring-shear

For both experiments, sediment was loaded under dark room conditions into the Aarhus University ring-shear apparatus (Fig. 1A). The ring-shear consists of a large (sample surface of 1800 cm^2) circular shearing chamber with a trough for the sediment 120 mm wide and depth of 80 mm (see Larsen et al., 2006, for further details). It has two plates between which the shearing gap in the sample is located. Ribs 6 mm in length are attached to both plates to fix the sample, and shearing is created by rotating the lower plate at a constant velocity (Fig. 1B). A uniform normal stress is applied hydraulically to the sample through the normal-load plate, which is free to move vertically according to sample compaction or expansion during shearing. Shear stress is measured by two sensors mounted on the normal-load plate, and sediment compaction is monitored by three sensors attached to the normal-load plate at equal distances around the shearing chamber whereby average data recorded by each group of sensors are considered further. The approximate shearing zone position was determined during test runs conducted using glass beads as strain markers, which showed the zone of deformation to be around 2.5 cm thick (Fig. S3). During the shearing, the sand had a preexisting moisture content making it cohesive but not saturated.

For experiment 1, the ring shear apparatus was run at a uniform normal stress of 100 kPa and a shearing velocity of 1 mm min^{-1} (i.e. parameters that are in the range of typical conditions beneath glaciers and ice sheets; Paterson, 1994) to a distance of 1280 cm. During the experiment, sediment compaction, shear stress, and normal stress were recorded in 30 s intervals. Experiment 1 was periodically paused to allow sampling after shearing displacements of 10, 20, 40, 80, 160, 320, 640, and 1280 cm. At each pause, two opaque 20-mm-diameter tubes were inserted vertically into the sand in the middle of the shearing chamber, marked at the level of the shearing chamber sand, and then slowly pulled out and sealed. The mark was subsequently used to infer the location of the shearing zone in each sample. The space in the shearing chamber left after sampling was naturally backfilled by lateral sediment creep while the tube was removed so that the original stratification was reinstated as closely as possible. This formed the basis of the samples used for sediment and for OSL characterisation. Before