Holocene microweathering rates and processes on ice-eroded bedrock, Røldal area, Hardangervidda, southern Norway

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Number of words excluding abstract, references and tables = 7,632

Number of references = sixty two (62)

Number of tables = four (4)

Number of figures = ten (10)

Abbreviated title = Holocene microweathering of bedrock

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Abstract: Postglacial weathering of ice-eroded metamorphic bedrock was investigated in the Røldal area (60^{o} N) of Hardangervidda Plateau in southern Norway. Quartz veins were used as reference surfaces to determine a mean postglacial surface lowering rate of 0.55 mm ka^{-1}. Chemical characteristics of late-season runoff were determined for one catchment (Snøskar) and obtained a chemical erosion rate of 4.9 t km^{-2} a^{-1}. Mean *in situ* fracture enlargement due to microweathering of 0.12 mm ka^{-1} was also determined. These rates are low, though comparable with similar environments in cold regions, and suggest microweathering has had relatively little impact on Holocene landscape evolution. Weathering rind thickness was found to be less on fracture walls than on exposed bedrock surfaces suggesting fractures have not played a significant role in microweathering. Observations of weathering morphology reveal a range of forms including shallow spalling, tafoni and pseudokarren, indicating locally intense weathering activity. Analysis of inter-relationships between multiple weathering indices points to the importance of bedrock microweathering as a precursor to macro-breakdown and landform evolution. The research re-asserts the importance of chemical activity in cold environments and the importance of moisture supply for effective microweathering.

Weathering of ice-eroded outcrops has the potential to significantly influence landscape evolution in periglacial environments during the postglacial period. However, there has been relatively little work to determine rates of postglacial bedrock weathering in cold environments. Traditionally, weathering in cold environments was seen to be the work of mechanical processes, specifically freeze-thaw (Tricart 1969), and low temperatures were thought to inhibit chemical weathering activity (e.g. Peltier 1950). However in recent years there has been a major change of emphasis in studies of weathering in cold environments away from traditional concepts to greater emphasis on the role of chemical (Anderson *et al.* 1997; Darmody *et al.* 2000; Thorn *et al.* 2001; Campbell *et al.* 2002; Dixon & Thorn 2005; Owen *et al.* 2006) and biological processes (André 1995, 2002; Etienne 2002; Etienne & Dupont 2002). In addition there is an increasing range of studies looking at the role of lithological and structural controls on cold environment weathering (Glasser *et al.* 1998; Olvmo & Johansson 2002; Whalley *et al.* 2004; Nicholson 2008). These changes are reflected in the call from Hall and André (2001) and Hall *et* *al.* (2002) to reconsider our assumptions about the nature of weathering in cold environments.

There have been relatively few attempts to determine rates of postglacial weathering in cold environments, with greater emphasis being placed on temperate and tropical regions. In the seminal study by Rapp (1960), transport rates determined from mass movements were compared against those for solute fluxes in Kärkevagge and he concluded that solution was of greater importance than mass movement. This pointed, implicitly, to the significance of chemical weathering in periglacial environments. More recent work in Kärkevagge (e.g. Darmody *et al.* 2000) and the neighbouring valley of Latnjavagge (Beylich *et al.* 2004) has determined weathering rates from the analysis of solutes in runoff together with studies of pedogenesis and weathering rinds and coatings. Following on from Dahl's work in 1967, André (2002) has also determined weathering rates in the Abisko-Riksgränsen area of northern Sweden from measurements of surface lowering with reference to upstanding quartz veins. Owen *et al.* (2006) also used this technique to determine rates of weathering on calcitic rocks along a lake shoreline. However, it has been difficult to determine precisely what the impact of weathering has been on landscape evolution.

It is helpful when there is some temporal constraint on landscape processes to enable an assessment of the impact of weathering and to this end studies of postglacial bedrock weathering are particularly useful. Notable examples include the attempts by André (1995, 1996a, 1996b) and Dahl (1967) to investigate postglacial bedrock microweathering in Swedish Lapland. Shakesby *et al.* (2006) and Sumner *et al.* (2002) have also used indices of postglacial weathering to provide relative ageing of glaciated bedrock surfaces in southern Norway and the sub-Antarctic.

The overall aim of the research presented here is to examine microweathering rates and processes on ice-smoothed bedrock forms in an active periglacial environment and to examine their influence on landscape denudation. A further aim is to contribute to the growing body of research on the weathering of crystalline rocks in cold environments. The objectives are threefold: (i) To synthesise weathering rates determined for the postglacial period from measurement of surface lowering, fracture enlargement and solute fluxes, and to compare these rates with those obtained for other, comparable locations. (ii) To infer weathering processes from measurement of surface weakening and weathering rind development and to make comparisons between processes at the surface and those occurring on fracture walls. (iii) To describe microweathering morphology and make further inferences about weathering processes.

Study area

Geomorphic environment

The study area is located in the Røldal region at $6^{o}58'$ E and $59^{o}53'$ N in the south west corner of Hardangervidda (the largest high mountain plateau in northern Europe) in southern Norway (Fig. 1). The plateau generally has subdued relief but the south west corner is more deeply dissected, with deep valleys and steep-sided ridges. Generally the plateau has tundra vegetation but the study area is dominated by exposed, ice-scoured bedrock. The study area comprises five sites totalling 4.5

km^2 which are characterised by roches moutonnées, whalebacks, scattered erratics and landforms of periglacial ground activity (Fig. 2). Many surfaces display glacial erosional features such as striations, grooves and P-forms. The area is believed to have been deglaciated c.10,000 years ago (Anderson 1980).

The five study sites are located at Rekkingeskara, Snøskar, Middyrelva, Dyrskarnuten and Fjetlandsnuten (Fig. 1). Altitude ranges from 950m to 1520m and all of the sites lie above the upper limit of boreal forest typified by high alpine tundra with very modest vegetation and many bare bedrock surfaces. Bedrock is frequently colonised by crustose lichen including *Rhizocarpon* species, and mosses and bryophytes are common. There is some ericaceous heath with occasional coarse tussock grasses, *Vaccinium* species and *Empetrum*. Soil development is generally incipient and immature, lacking organic matter and being dominated by silt and accumulations of grus, plus limited patches of glacially-derived sediment. Mean annual temperature for the area is -2^o C and annual precipitation is about 1525 mm. Permanent and semi-permanent snowpatches are common and the Nupsfonn plateau glacier (c.3 km^2) only 5 to 10 km to the east and at an altitude of 1600 m is a shrunken remnant of a more extensive Little Ice Age glaciation which did not reach any of the study area.

The bedrock lithology consists entirely of metamorphic rocks, either Precambrian basement from the Stavsnuten and Dyrskarnuten Nappe complexes of Proterozoic age, or at Middyrelva, from the Mannevatn Nappe of Ordovician age. Bedrock at Rekkingeskara is mainly granitic gneiss and schist with limited outcrops of amphibolite. The greatest variation in bedrock occurs at Snøskar, with hornblende amphibolite and amphibolitic schist, pockets of feldspathic quartzite and quartzo-feldspathic gneiss, and various forms of mica schist (including muscovite, chlorite and biotite-epidote varieties). Bedrock at Middyrelva generally consists of chlorite-rich mica schist with numerous quartz veins while at Dyrskarnuten the bedrock is chlorite-rich amphibolite schist. Bedrock at Fjetlandsnuten consists of actinolite-amphibolite gneiss, probably derived from metamorphosed basalt.

The Rekkingeskara study site (1.5km^2) is a broad valley located from 1250 m to 1450 m between the Valldalsvatnet and Middryvatnet basins. The valley is dominated by blockstreams which appear to ‘flow’ around upstanding roches moutonnées, several of which are breaking up, and contribute to the supply of blockstream material. Occasionally there are flatter areas with plentiful moisture supply which support solifluction lobes and patterned ground. The Snøskar study site (3km^2) is an enclosed, circular basin with two central lakes and a single outlet to the south west (Fig. 3). The basin lies between 1300 m and 1520 m and is located north of the Middyrelva valley. The basin displays classic knob and basin topography with numerous roches moutonnées and whalebacks interspersed with lakes and pools. There are usually a number of late-lying snowpatches around the margins of the basin and there are numerous periglacial ground forms. The Middyrelva study site (1500m^2) is located at 1150 m in a glacially deepened valley (maximum relief c.500 m) which crosses a drainage divide falling from 1240m at Middyrvatnet to 980 m at Ulevåvatnet in the south east. The Dyrskarnuten study site (1500m^2) is similar, comprising a series of small roches moutonnées rising from 1080 m to 1160 m on a broad col between Votna and Ulevåvatnet to the east. The valleys at Middyrelva and Dyrskarnuten are both moderately vegetated with ericaceous heath. Fingers of vegetated soil spread up the lower slopes of roches moutonnées and there is also some soil development within weathering pits on the crests. The valleys both support numerous small tributaries coming off the steep valley sides. The Fjetlandsnuten study site (2000m^2) is on a small hill rising from 950 m to 1070 m at the west end of lake Votna. The roches moutonnées studied are particularly exposed and permit views across the Hardangervidda plateau for at least 8 km to the north west. Ericaceous heath is well developed around the fringes of roches moutonnées including low-growing *Salix arctica* (rock willow), *Rubus chamaemorus* (cloudberry) and *Vaccinium* species (blueberry).

Methods

With the exception of solute analysis described below, all of the other measurements were made on roches moutonnées at the five sites. At the two main sites, Snøskar and Rekkingeskara, the primary criteria for selection of roches moutonnées were bedrock lithology (it was desirable to include all of the major rock types), degree of exposure (particularly sheltered or exposed locations which might favour weathering activity were avoided) and spatial coverage. At the remaining three sites, roches moutonnées were selected where they were present. A total of 45 roches moutonnées were used in this study.

Weathering rates

Three methods were used to determine weathering rates: (a) Analysis of solutes in runoff to determine contemporary removal of material in solution from the Snøskar catchment. (b) The use of reference surfaces (Dahl 1967) together with chronological control to determine average weathering rates during the postglacial period. (c) *In situ* fracture enlargement due to weathering processes.

Determination of weathering rates from water chemistry

A total of 73 late-season water samples were collected from 64 sites (Fig. 3) at Snøskar between 23rd August and 5th September 2006. At each sampling location temperature, pH and conductivity (reference 25^{o} C) were determined. In addition, for 29 of the sampling locations, the metals Al, Ca, Fe, K, Mg, Mn, Na and Si were determined using inductively coupled plasma spectrometry and anions Cl, NO_3, SO_4 and F were determined using atomic absorption spectrometry. Water samples were taken from a range of sources including snow, snowpatch runoff, lakes, small pools, minor trickles and streams. Total dissolved solids (TDS) were estimated using a conversion factor of 0.7 for conductivity μs to TDS ppm (based on Strömquist & Rehn 1981).

Surface denudation using quartz veins as reference surfaces

Following the method of Dahl (1967), used more recently by André (1995, 1996a, 1996b, 2002), postglacial surface lowering was determined by measuring the difference in height between upstanding quartz veins and surrounding bedrock. Measurements were made only on exposed ice-scoured roches moutonnées and whaleback surfaces. The method is based on the principle that being more resistant to surface microweathering, ice-polished quartz vein surfaces represent the pre-weathered, ice-scoured surface at the end of the last glacial maximum.

Measurements, to the nearest 0.1 mm, were usually made with a calliper though a short section of metal plate was occasionally projected from the vein surface if the surrounding bedrock topography made the use of a calliper difficult. Up to three veins were selected for each roche moutonnée. For each vein, up to 50 measurements were made, always with an approximately equal number of measurements on each side of the vein, and at intervals of 10mm. Where glacial erosion or subsequent breakdown of vein edges created an arched upper surface (in cross profile), care was taken to record the *maximum* height difference. Following the recommendations of Dahl (1967) quartz veins containing many joints were avoided, as were veins found in bedrock that contained many fractures in close proximity to the veins. Care was taken to ensure that the veins used exhibited a surface sheen, often yellowish in colour, representing glacial polish. The results are reported as a mean for each roche moutonnée. Over 1000 measurements were made at 23 sites, giving an average of 45 measurements at each site.

Fracture enlargement

A calliper was used to measure the width at the top of enlarged fractures to the nearest 0.1 mm, perpendicular to the fracture direction. Care was taken to avoid fractures which had opened due to physical displacement of the bedrock (e.g. by tectonic activity, gravitational stress or frost shattering), ensuring that measurements were only taken where enlargement was due to *in situ* microweathering of fracture walls. A total of 1244 measurements were made.

Weathering indices

Weathering rind thickness and rock surface hardness were determined to provide a relative estimate of weathering intensity over the postglacial period:

Weathering rind thickness

The use of weathering rind thickness as a weathering index is based upon the widespread assumption that rind development is dependent upon the age, or period of exposure, of the surface studied (Thorn 1975) and rind thickness has been used in a variety of ways to study postglacial bedrock weathering (e.g. Chinn 1981; Dixon *et al.* 2002b; Sumner *et al.* 2002). Weathering rind thickness was obtained for exposed bedrock surfaces and also for fracture walls. Surface weathering rind thickness was measured from 25mm diameter drilled cores sampled from smooth surfaces on the top of roches moutonnées. Fracture wall weathering rind thickness was measured from rock samples extracted from the intersections of closed joints using a geological hammer. In each case, weathering rind thickness was determined using a clear plastic ruler to the nearest 0.2 mm. A hand-held crack microscope was used to facilitate identification of the boundary between weathering rind and unweathered rock, based primarily on colour change. For drilled cores the maximum weathering rind thickness was obtained from four equally-spaced measurements around the core circumference. For rock samples several measurements were made as necessary in order to identify the maximum weathering rind thickness. At each site, between 15 and 20 measurements were taken with a total sample number of 667 measurements (525 of surface rind and 142 of fracture wall rind). The data presented represents the mean maximum weathering rind thickness for each roche moutonnée as recommended by Thorn (1975).

Rock surface weakening

Surface hardness was determined using a calibrated ‘N’ type Schmidt hammer. This portable field instrument works by measuring the rebound distance of a controlled impact by a piston on a rock surface. A full analysis of the instrument and its widespread usage in geomorphology is given by Goudie (2006). Several researchers, notably Rae *et al.* (2004), Shakesby *et al.* (2006) and Sumner (2004), have used the instrument to investigate postglacial weathering of bedrock. There has been a great deal of discussion about testing procedures (for example, see Poole & Farmer 1980; Goudie 2006) and a wide range of approaches are in use. A unique approach has been used here. If the primary purpose of using the hammer is to determine the intact strength of rock, then it is appropriate to prepare the surface well to ensure that a fresh face is presented to the hammer. The testing method of Hucka (1965) can then be adopted - which is to use multiple impacts at a single point. However, if the aim is to obtain an index that reflects the relative degree of surface weathering, then the surface, which should be lichen-free and free from loose material, should not be pre-prepared and a single impact at any point is sufficient.

Poole and Farmer (1980) conducted a statistical analysis of the consistency of repeated Schmidt hammer impacts at a series of points on four different rock types. Their results show (Fig. 3, p170) that the first rebound value is consistently lower than subsequent values. This supports the contention here, that the first rebound value (R_1) can be used to represent the hardness of the weathered surface, and that the second rebound value (R_2) is a much closer approximation (albeit there is some variability) of intact rock strength. Therefore, in this study, 25 *pairs* of readings were obtained for each roche moutonnée with each pair of impacts being obtained at a single point (i.e. without moving the hammer). A comparison between the two values allows for some relative estimation of the degree of weakening at sites with contrasting lithological characteristics. More than 2000 measurements were made and all have been adjusted for angle with respect to a horizontal surface (Day & Goudie 1977). Care was taken to avoid sites close to fractures or edges, to select even surfaces and to use the instrument only in dry conditions.

Results

Weathering rates

Chemical erosion from solute runoff

Conductivity of water and snow samples from across the Snøskar catchment yielded an overall mean of 5.5 μs or 3.9 ppm for TDS (Table 1). The highest TDS values recorded were found in some snow samples, and smaller lakes and ponds. The lowest TDS levels recorded were found in streams and small tributaries. A series of conductivity measurements were obtained through a pit dug into snow (Fig. 4). Snow above the strong discontinuity represents net accumulation during the previous twelve months and has a mean TDS concentration lower than mean solute values across the catchment. The slightly elevated solute values at the surface represent the concentration of solutes (derived mainly from precipitation) from ablation of the overlying snow and from summer precipitation. Below the discontinuity the snow is quite old (at least several years) and the higher solute values reflect concentration of solutes arriving from the surface via percolation. While it is not possible to rule out contamination of the snow by runoff of solutes derived directly from bedrock dissolution, this is unlikely, except near the snow-bedrock interface. There are low levels of Cl^-, NO_3^- and SO_4^{--} in the Snøskar catchment but F^- was absent. The dominant anion was Cl^- (mean 1.5 ppm) with concentrations being notably higher in surface snow. There are low levels of Al, Ca, Fe, K, Mg, Mn, Na and Si in the catchment. The dominant cation was K^+ (mean 0.9 ppm), also notably higher in snow. Ca (0.3 ppm) and Si (0.2 ppm) are also relatively important and appear to have slightly greater concentrations in streams and ponds. Observation of the spatial distribution of solute concentrations suggests that there are slightly elevated levels of nitrate in the central, lowest part of the basin and on a low level plateau on the east side. These areas support greater vegetation cover than is found elsewhere in the catchment and this is, therefore, an indication that nitrate has its origin in the very limited sediments and immature soils with organic matter. Silica levels are also slightly lower in the centre of the basin where the rocks are much less dominated by quartz.

Overall, solute concentrations for the catchment are extremely low and do not appear to relate to spatial variations in bedrock lithology. This would be difficult to determine statistically in the absence of a much more detailed sampling regime, since there are frequent variations in rock type and mineralogical composition. Generally, there is little spatial variation in the slightly below mean solute concentrations for streams where water is moving. Where water collects, in lakes, ponds and snow, solute concentrations are a little higher.

Surface lowering

Hardangervidda is thought to have been largely ice-covered until the end of the Younger Dryas (Mangerud *et al.* 1979). Evidence presented by Anderson (1980) indicates that Hardangervidda was deglaciated about 8750±250 ^{14}C years BP. This correlates with 10,000 calendar years BP (Stuiver *et al.* 1998). A similar date was found by Dixon *et al.* (2002a) for deglaciation of the Riksgränsen area in northern Norway, which although 9^o further north in latitude, has a comparable climatic regime to Hardangervidda by virtue of its much lower elevation. Therefore, for the purposes of calculating postglacial weathering rates in this study a date of 10,000 years BP is used. The overall mean rate of postglacial lowering as determined from the measurement of quartz vein reference surfaces is 0.55 mm ka^{-1} and individual measurements for different roches moutonnées range from 0.05 to 2.2 mm ka^{-1} (Table 2). The range and mean rates of surface lowering are broadly comparable with those determined by André working on a similar range of rocks in the Abisko-Riksgränsen region of north Sweden (André 1995, 1996a, 1996b, 2002) and by Dahl working at Narvik in north Norway (Dahl 1967).

Fracture enlargement

At the surface, although many joints have been opened by postglacial weathering, there are many which remain tightly closed. The overall mean fracture enlargement is 2.4 mm (Table 2) which compares well with values obtained at Riksgränsen by André (1995) for a similar range of rocks and comparable environmental conditions. This single value masks huge variation from fractures which have barely opened up to those with 75 mm of enlargement. Most commonly, fractures are either incipient (e.g. 0.1 to 0.2 mm) or opened to c.5-20 mm. The majority of fractures contain some infilling, commonly vegetative material such as moss, and small fragments of rock that probably originate from fracture walls. Relatively few fractures contained an accumulation of fine sediment. This indicates either, that fine sediment has not been produced or, that it has been flushed away. Likewise, there is little accumulation of organic soil in fractures. Several distinctive cross profiles of enlarged fractures can be identified (Fig. 5) which may be indicative of the weathering processes at work.

Weathering indices

Weathering rind thickness

The overall mean maximum weathering rind thickness for exposed surfaces was 3.4 mm with an individual maximum value of 26.2 mm (Table 2). The overall mean maximum weathering rind thickness for fracture walls was 2.5 mm with an individual maximum of 20.7 mm (Table 2). These results show that weathering rind is generally thicker on exposed surfaces than on fracture walls (Fig.6).

Rock surface weakening

Using the Schmidt hammer the mean values for R_1 and R_2 respectively are 54 and 64 (Table 2). The range of mean values for each roche moutonnée are 38 to 71 for R_1 and 55 to 73 for R_2. The scattergraph (Fig. 7) shows the correlation between mean R_1 and R_2 site values and demonstrates (i) a significant increase in rebound value for the second measurement (R_2) and (ii) the difference between R_1 and R_2 is significantly greater at lower values of R_1. This indicates that rocks which are weaker in their fresh, unweathered state, show a greater reduction in surface hardness due to weathering, than rocks which are inherently stronger in their fresh state.

Weathering morphology

Many ice-smoothed rock surfaces display a small scale morphology indicative of weathering processes (Figs. 8, 9 and 10a-d). Shallow, small scale surface flaking or spalling is particularly ubiquitous. Occasionally the lifted flakes remain intact (Fig. 8) but mostly shallow scars, typically 1-5 mm deep, remain on the rock surface. These scars are often visible because they have a fresh, lichen-free appearance in comparison with the surrounding bedrock. Flaking is extremely common on quartzo-feldspathic bedrock and generally occurs on surfaces parallel with structural controls such as foliation and banding. A particular form of flaking very common on thinly banded gneiss is *stepped* flaking or spalling giving the effect of rounded edges. This occurs where thin foliations in the rock peel off, furthest at the surface and less at the sides of a block (Fig. 9). This form may represent weathering modification of glacially-rounded surfaces. The photograph also shows macro-breakdown of roches moutonnées via small scale fracturing. The loose blocks are slowly moving away from the main rock mass in the adjacent solifluction.

Weathering pits are similar to those reported by André (2002), typically irregular but rounded in plan form and shallow with fairly smooth surfaces (Fig. 10a). Lengths range from 20 to 100 cm and widths from 10 to 50 cm. Pit depth is usually up to 5cm but depths of up to 15 cm also occur. Most pits are elongate to some degree, often oriented in alignment with fractures or geological structure. These pits occur infrequently, and generally only on the amphibolites and bedrock containing a high proportion of biotite. Pits also tend to be much more prevalent on roches moutonnées that have a moderate lichen cover. A second type of much smaller, deeper, weathering hollow occurs very commonly in the amphibolites and on the chlorite mica-schist at Middyrelva. These hollows occur particularly on vertical, rather than horizontal surfaces, and are elongated along foliation or banding. Hollows are typically up to 20 mm deep and 50mm in diameter, but there are some larger forms up to 300 mm in length. The hollows at Middyrelva (Fig. 10b) are fully developed into a dense cover of tafoni. They typically occur on lichen-free surfaces and have a rough texture indicative of granular breakdown.

A particularly unusual weathering form observed in the Røldal area are highly irregular, very deep, and often undercut, hollows (Fig. 10c). These have the appearance of karren forms produced by limestone dissolution in karst terrain and for that reason they are referred to here as pseudokarren. They are also very similar in appearance to the ‘weathering pits’ described by Dahl (1966). Pseudokarren only occur at three locations; on a group of three amphibolitic roches moutonnées at Fjetlandsnuten, on a single roche moutonnée in amphibolite at Snøskar, and on a group of granitic gneisses at Rekkingeskara. Hollows are characteristically very rough textured and pitted at a small scale. They either occur as isolated features (Fig. 10c), as coalesced hollows or as crenulations or indentations along edges (Fig. 10d). Pseudokarren vary in shape and depth but single hollows are commonly up to 80 cm in length and in the order of 10-20 cm deep.

In amphibolitic and mica-rich bedrock there are many examples of weathering-modified glacial erosional forms such as striations, crescentic gouges, P-forms and Nye channels. For example, judging from their distinctive plan form, some weathering pits have almost certainly developed from the enlargement of crescentic gouges. Striations are ubiquitous but are often difficult to locate on lichen-covered bedrock. Coarser striations can sometimes be found on exposed surfaces but well preserved fine striations are generally only found through minor excavation of sediment at the lower margins of roches moutonnées. Other indicators of weathering activity include surface discoloration, seen quite spectacularly in some of the feldspathic schist at Snøskar; fracture enlargement and rounding of fracture cross profiles (Fig. 5); weathering rind; and the presence of upstanding quartz veins, previously discussed.

Discussion

Weathering and erosion rates in the Røldal area

Using a mean TDS value for the catchment of 3.85 ppm (derived from the mean conductivity of 5.5 μs) and an effective precipitation of 1275 mm (subtracting 250 mm from mean annual precipitation for evapo-transpiration), the mean rate of chemical erosion for the Snøskar catchment equates to 4.9 t km^{-2} a^{-1}. Table 3 shows a range of erosion rates obtained for comparable environments and it can be seen that those obtained for Snøskar are similar to rates at Latnjavagge (5.4 t km^{-2} a^{-1}) in Swedish Lapland (Beylich *et al.* 2004). These rates are considerably lower than the range of values of 19.2 to 46 t km^{-2} a^{-1} determined for Kärkevagge (Darmody *et al.* 2000; Campbell *et al.* 2002). However, the erosion rate for Snøskar does not include any allowance for atmospheric inputs which were not obtained but could be in the order of the mean conductivity value for the Snøskar catchment (a mean precipitation conductivity of 9µs was found for Kärkevagge by Darmody *et al.* 2000). For this reason, the erosion rate obtained is like to be an over-estimate. Further reasons to suspect that the rate may be an over-estimate are (a) the data represent summer activity when one might expect chemical processes to be more active; (b) data were collected during a dry period following three weeks of very wet weather and thus solute concentrations may be relatively higher than if discharge rates had been greater; (c) the calculated rate assumes uniform contact between surface runoff and the bedrock from which solutes are derived. In reality, during the peak snowmelt period, meltwater moving through the snowpack will have relatively poor contact with bedrock. The calculated erosion rate may also lack rigour because of the limited temporal period over which the data were collected.

Using the overall mean surface lowering rate of 0.55 mm ka^{-1} (obtained using quartz veins as references surfaces) and assuming a mean rock density of 2650 kg/m^3, surface lowering equates to the removal of 1.5 t km^{-2} a^{-1} of material from the Røldal area. The erosion rate derived from the *maximum* of site mean values for surface lowering (2.2 mm ka^{-1}) is 5.8 t km^{-2} a^{-1}. These rates are similar to those calculated from surface lowering at Abisko-Riksgränsen in Sweden (André 1995, 1996a, 1996b, 2002). At its simplest level, surface lowering will result from a range of weathering and erosion processes, including dissolution. One might therefore expect the solutional erosion rate to be less than the denudation rate calculated from measurements of surface lowering. However, the two estimates of total denudation rate are not strictly comparable because they are derived from two independent sets of measurements (bedrock lowering and solute runoff) which reflect a very different range of micro-environments and processes: On the one hand, measurements of surface lowering reflect weathering of exposed bedrock surfaces. On the other hand, solute runoff will include weathering and erosion taking place on bedrock surfaces and beneath blockfields, within and on top of snowpatches, within rock joints and through leaching from soils. It is likely that the denudation rate derived from bedrock surface lowering is an underestimate of total denudation in the Røldal area.

Controls on cold environment weathering processes

Analysis of relationships between variables (Table 4) indicates that there is a statistically significant inverse correlation between surface lowering and surface hardness (R_1 and R_2). This demonstrates, not unexpectedly, that weaker bedrock is more susceptible to surface lowering than stronger bedrock. That the inverse correlation with surface lowering is stronger for R_1 than for R_2 may also be an indication that a weakened bedrock surface is a pre-requisite for surface lowering. These relationships also point to the fact that surface lowering is achieved through some mechanism that results in mechanical weakening of the rock. Previous studies indicate that this could be through biophysical disruption of bedrock by lichen thalli (e.g. André 1995; Carter & Viles 2004) or fungal activity (e.g. Etienne & Dupont 2002; Arocena *et al.* 2003), biochemical dissolution and alteration by lichen and fungi (e.g. Etienne & Dupont 2002; Arocena *et al.* 2003; Hall *et al.* 2005), development of mineral grain porosity due to chemical dissolution (e.g. Dixon *et al.* 2002a), or through entirely physical processes such as thermal shock (e.g. Hall *et al.* 2002). In the Røldal area, the very low levels of chemical erosion derived from analysis of solute runoff suggest very limited chemical weathering in this high mountain plateau environment. However, this is not supported by the ubiquitous presence of weathering rind or from observations of weathering morphology.

Weathering rind is ubiquitous on bedrock surfaces in the Røldal area, although its thickness varies. This is clear evidence that chemical and/or biochemical processes are an important component of microweathering in this area. There is increasing evidence to suggest that rind formation is strongly dependent upon moisture availability (Etienne 2002; Dixon *et al.* 2006). There is also evidence that rind formation is associated with an increase in porosity. For example, field experiments at Kärkevagge using buried granite disks, demonstrated that early development of rind coincides with a significant increase in rock porosity (Dixon *et al.* 2006). Oguchi and Matsukura (2000) also noted a coincidence of higher porosity in andesites with greater rind thickness. The fundamental cause of the increase in porosity associated with rind formation is chemical or biochemical in nature. Dixon *et al.* (2006) argue that intra- and inter-grain dissolution and the formation of microcracks bring about these changes at the rock surface. Etienne (2002) also argues the case for the role of organic acids from fungal growth. The common presence of iron oxides in weathering rind (e.g. Dixon *et al.* 2006) indicates the role of oxidation of ferromagnesian minerals. Dixon *et al.* (2006) indicate that the incidence of increased porosity was greater in feldspar and quartz and this concurs with the finding in the Røldal area that weathering rind is thickest in the quartzo-feldspathic rocks (Nicholson 2008). It is suggested that the widespread occurrence of weathering rind on bedrock in the Røldal area is also indicative that chemical weathering processes are active.

It is useful to consider the relationship between weathering rind thickness and rock surface hardness. Several authors have noted the importance of micro-erosion in the formation and evolution of weathering rind (Etienne 2002; Gordon & Dorn 2005). It is highly probable that susceptibility of rind to micro-erosion is related to rock surface hardness, particularly since the latter is partially dependent upon rock porosity (e.g. Nicholson 2001). However, the relationship between the two indices is not straightforward. In this study, there is a weak positive correlation between surface rind thickness and surface hardness (Table 4). The coincidence of greater porosity with rind formation found in other studies would lead one to expect the opposite (i.e. that thicker rinds would be found on weaker bedrock). However, it is suggested that in relatively weak rocks or rocks with relatively high porosity, the rate of rind formation may be outpaced by rind removal due to micro-erosion, thus giving the false impression that there is little absolute rind development.

The common coincidence of shallow surface flaking with quartzo-feldspathic rocks and their occurrence parallel with metamorphic foliation and banding suggest there might be some lithological control on their formation. However, an alternative explanation is possible, analogous to the cause of exfoliation envisaged by Etienne (2002) in relation to the evolution of weathering rind. Etienne observed microcracks parallel to the surface at depths of 1-3mm, which broadly coincide with the thickness of flakes observed in the Røldal area. Etienne (2002) proposed that these microcracks, representing structural heterogeneity between weathered rind and the unaltered bedrock beneath, might coalesce to form larger flakes. Furthermore, he envisaged that lifting, or exfoliation of flakes, could come about through biological activity (e.g. growth of fungi or lichen) or due to physical stresses (e.g. ice lens growth or thermal stress). It seems reasonable that this range of processes (Etienne 2002) could be responsible for the ubiquitous presence of shallow surface flakes in the Røldal area.

That weathering pits are often oriented in alignment with metamorphic structure or local fractures, and occur primarily on biotite-rich bedrock, is an indication that their development is at least partially geologically controlled. Observations also suggest a strong coincidence with lichen-covered bedrock, supporting the contention of previous studies (André 1995, 2002) that biological weathering (e.g. of biotite crystals) could be key. However, given that late-lying snow inhibits lichen colonisation, it could be argued that weathering pits favour snow-free sites where frequent fluctuations in temperature and moisture conditions are more likely to occur.

The honeycomb-like weathering tafoni observed are similar to those observed by French and Guglielmin (1999) on meta-granites in Antarctica which were ascribed to granular disintegration associated with frost action in the presence of salts. French and Guglielmin (1999) believe these features to have developed rapidly over a period of 2000-3000 years. The cause of the deep pitting found in the Røldal area is unknown. However, given the predominance of these features in the amphibolitic and chlorite mica-schist, it seems likely that selective and differential weathering of ferromagnesian minerals leading to granular disintegration could be key (Campbell & Claridge 1987).

The deepest of the pseudokarren observed represents a weathering rate at least two orders of magnitude greater than that obtained from measurements of bedrock surface lowering and so it is clear that they represent locally intense weathering conditions. One possibility that needs to be considered is that pseudokarren represent a pre-glacial weathering surface which has survived glaciation. However, given that two of the three sites where they are found display numerous striations, and at Fjetlandsnuten there are also Nye channels and other P-forms of glacial meltwater erosion origin, this explanation seems highly unlikely. A more promising explanation is that pseudokarren are sites favouring intense biotic and/or chemical activity concentrated around flaws such as cracks or shallow exfoliation scars. The former might explain the elongate nature of most of these features. Pseudokarren have the appearance of having been formed through dissolution though it is quite likely that vegetative material has contributed to dissolution through the provision of organic acids (Dahl 1966; Dixon *et al.* 2002a).

The common preservation of glacial striations in the Røldal area is a clear indication that weathering has had relatively limited impact on landform denudation since the last period of glaciation. Although glacial striations were found on all rock types, they are almost an order of magnitude more abundant on amphibolite despite surface lowering, and several examples were observed of their presence in close proximity to upstanding quartz bands. This may indicate that the original striae were larger and persist despite bedrock surface lowering. Alternately, it may simply indicate that surface lowering has perpetuated the original shape and size of striations.

The role of fractures in weathering

Much of the literature on the role of fractures in cold environment weathering has focused on the traditional concept of freeze-thaw as the dominant process (e.g. Walder & Hallett 1985; Murton *et al.* 2006). Field-based studies have included the role of fractures and microcracks in the formation of blockfields (e.g. Boelhouwers 2004; Whalley *et al.* 2004), the macro-breakdown of rockwalls (e.g. Matsuoka & Sakai 1999) and fractures as structural controls on macro-breakdown (e.g. Gordon 1981; Glasser *et al.* 1998; Olvmo & Johansson 2002). However, there has been relatively little field-based study of the direct role of fractures in cold environment microweathering. One exception is the work of André (e.g. 2002) who measured fracture enlargement and obtained similar values to those obtained in the present study. If the mean fracture enlargement of 2.4 mm obtained in this study is assumed to reflect bi-directional recession of the fracture walls, then uni-directional recession can be assumed to be 1.2 mm. Given that measurements of fracture enlargement were made at the intersection of fractures with the surface, for which a chronological reference exists, it is reasonable to accept that this value represents uni-directional widening over a period of 10,000 years. The rate per thousand years is therefore 0.12 mm ka^{-1}. Given a mean surface lowering rate of 0.55 ka^{-1}, this value represents only one fifth of the equivalent lowering achieved at the surface over the same period of time.

There is a highly significant inverse correlation between fracture enlargement and R_1, and an equally significant positive correlation with surface lowering (Table 4). Given the correlation between surface lowering and surface hardness (see above), this suggests that although fracture enlargement occurs at a much slower rate than removal of material from exposed bedrock surfaces, the processes and controls involved are very similar. In previous studies (André 2002) it has been proposed that postglacial granular disintegration is the primary cause of fracture enlargement. However, a two-stage process of fracture enlargement is envisaged here, in which initial modification of fractures occurs entirely *in situ* by chemical and physical microweathering. This is clearly indicated by the presence of weathering rind on fracture walls and results in weakening and micro-erosion (e.g. through transport of solutes and fine particles), leading to the initial enlargement of fractures. With void space now created between fracture walls there is scope for further enlargement through multiple processes operating at a larger scale. Evidence from fracture cross profiles (Fig. 5) indicates that there are three main processes involved: (a) granular disintegration producing rounding of fracture edges; (b) spalling of fracture walls producing thin flakes or wedge-shaped rock fragments; and (c) more general surface break-up in association with multiple fracture intersections and shattered zones. Some fracture cross profiles (e.g. overhanging fractures, Fig. 5) reflect the interaction of weathering processes with rock structure.

The fact that weathering rind is thicker on exposed surfaces than on fracture walls (Fig. 6) is perhaps contrary to what might have been expected. That there is no correlation between fracture wall rind thickness and any of the other variables measured (Table 4) also indicates that there are different controls on the development of fracture rind and a different set of processes involved. One might have expected fracture wall rind to develop well, given that fractures are sites of potential moisture accumulation, thus aiding chemical activity. Furthermore, it is unlikely that significant rind has been removed from fracture walls since samples were taken from closed fractures. This is an interesting finding and there are several possible explanations. It is possible that in reality, moisture does not easily penetrate closed fractures and thus weathering activity is inhibited by the limited quantity of moisture available. Alternately, moisture penetrates but travels downwards rapidly and thus its residence time, and the opportunities for fracture wall weathering, are limited. It is also likely that biotic weathering processes are inhibited in cracks by the lack of photic activity. This would lend support to the view that bacteria and fungi are important in rind development (Etienne 2002). It may also help to explain why rind thickness is greater in pale, quartz-rich rocks (Nicholson 2008) which probably allow greater penetration of solar radiation. A further explanation for the contrast between surface and fracture wall rind thickness is that development of rind at the surface is particularly enhanced by the presence of moisture at the interface between bedrock and late-lying snow, a view shared by others (Thorn 1975; Ballantyne *et al.* 1989; Nyberg 1991; Thorn & Hall 2002). An alternate view is that fractures have only opened during the Holocene and therefore there has simply been less time for fracture wall rind to develop.

The finding that fracture enlargement is much less than lowering of the surface broadly concurs with the observation that weathering rind found on fracture walls is thinner than that on exposed surfaces and suggests that fractures have less influence on microweathering than might have been expected. However, it is important to note that there is substantial observational evidence from Rekkingeskara and Snøskar that fractures play a significant role in large-scale landform development. In particular, there are numerous examples of roches moutonnées which have experienced significant breakdown, the products of which are being assimilated into surrounding blockstreams.

Conclusions

In this study, multiple weathering indices demonstrate that for the Røldal area, rates of periglacial microweathering are generally low, though surface morphology indicates there is locally intense weathering activity. Rates are similar to those obtained for comparable regions, indicating that at a regional level, climatic conditions are a major controlling factor. Interactions between indices (e.g. surface lowering, weathering rind thickness, surface hardness) have been evaluated to make inferences about the nature and efficacy of microweathering processes. There is evidence from analysis of surface lowering and fracture enlargement that denudation is achieved through a two-phase process in which initial *in situ* microweathering produces rind and weakening of the bedrock surface. Subsequently, as indicated by surface spalling and fragmentation in relation to fracture cross profiles, larger scale breakdown and erosion is promoted. The ubiquitous presence of weathering rind in this region is clear evidence of the importance of chemical and/or biochemical processes in microweathering.

While the nature of this study does not allow the determination of the precise weathering mechanisms at work, it has been possible to recognise three primary controls on microweathering rates and processes; moisture availability, biochemical activity and bedrock characteristics. It is clear that moisture availability plays an important role in most microweathering processes and is fundamental for dissolution and for the transport of solutes from catchments. Moisture is also essential for biotic processes, for the formation of weathering rind, the enlargement of fractures and in the development of a range of weathering forms (e.g. weathering pits, tafoni and pseudokarren). The findings of this study concur with the sentiment from Hall *et al.* (2002) that the role of moisture in periglacial weathering has hitherto been underestimated in comparison with the stronger focus on temperature-related factors.

That chemical and biotic processes are a further controlling factor in microweathering is evident from the presence of a number of morphological forms including pseudokarren, tafoni and weathering rind. With respect to the potential role of lichen, there is an apparent dichotomy between the coincidence of several enhanced weathering forms with lichen-covered sites, which generally preclude late-lying snow, and the concept that weathering is enhanced beneath snowpatches (e.g. Thorn 1975; Berrisford 1991; Ballantyne *et al.* 1989). A growing body of literature points to the importance of lichen and fungi in cold environment weathering (e.g. André 1995; Etienne and Dupont 2002; Arocena *et al.* 2003) but further investigation of the relationships between lichen-cover and late-lying snow and their role in bedrock microweathering would be helpful. Moreover, it would be useful for predicting the nature, intensity and spatial distribution of microweathering if there were data comparing moisture availability at the snow-bedrock interface with moisture availability in snow-free areas.

The third controlling factor in microweathering rates and processes is bedrock characteristics. The strength of intact, unweathered bedrock is a key control in the response of exposed bedrock surfaces to microweathering processes including surface lowering, weathering rind formation and the evolution of rind. Moreover, geological structure interacts with the mechanisms of weathering including spalling, the orientation of weathering forms and the distribution of fractures. Data are presented here that show microweathering is less intensive in fractures than at the exposed bedrock surface. There also appear to be similar controls on enlargement of fractures as for denudation of exposed bedrock surfaces. Observations suggest that fractures substantially contribute to large scale breakdown of roches moutonnées and the development of blockstreams. However, repeating the general call from Viles (2001), further work is needed to improve our knowledge at all scales, of the interactions between fractures and weathering rates and processes, if the evolution of periglacial landscapes is to be truly understood.

This study reinforces the belief that microweathering processes are apparently insignificant in terms of postglacial landform evolution in cold environments. Nevertheless, there is substantial macro-breakdown of *in situ* bedrock in the Røldal area. The author contends that the mechanisms responsible for landform modification at these two contrasting scales lie at the opposite ends of a continuous, inter-related spectrum of weathering processes. The two-phase nature of fracture development interpreted here, is perhaps one demonstration of scale linkage in the micro- to meso- range of the spectrum. The challenge for periglacial geomorphologists is to test this assertion by obtaining further evidence of scale linkages in weathering processes.

Acknowledgements

Fieldwork was undertaken on the MMU-LJMU Joint Norex Research Expeditions of 2003-6 and the author thanks student members for their invaluable field assistance. Grateful thanks are due to Frank Nicholson for field assistance and valuable advice and support throughout. This work was partly funded by the Nuffield Foundation (Grant NAL/00698/G). Valuable and constructive comments received from Jan Boelhouwers and an anonymous referee greatly improved the first version of the manuscript and to them I am very grateful.

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Figure captions

Fig. 1. Map of general area and five study locations.

Fig. 2. Typical landscape of the Røldal study area.

Fig. 3. Solute sampling locations at Snøskar (note some seepages and streams are too small to be shown at this scale).

Fig. 4. Conductivity depth profile for one snowpatch (all snow above dashed line is from last winter).

Fig. 5. Observed cross profiles of enlarged fractures.

Fig. 6. Box plots comparing weathering rind thickness for exposed surfaces and fracture walls.

Fig. 7. Correlation between mean R_1 and R_2 site values.

Fig. 8. Shallow surface spalling in quartzitic schist.

Fig. 9. Fracturing and downslope movement of loose blocks. Stepped spalling (see text) associated with foliation in granitic schist.

Fig. 10(a). Typical shallow weathering pit in actinolite amphibolite.

Fig. 10(b). Honeycomb weathering pits in chlorite mica-schist.

Fig. 10(c). An isolated pseudokarren in actinolite amphibolite.

Fig. 10(d). Crenulated and undercut edges of pseudokarren ‘solution’ forms in amphibolite.

Table captions

Table 1. Solute data for the Snøskar catchment. [1]Value indicates number of samples used to obtain conductivity, pH and temperature (value in parenthesis indicates number of samples used in elemental analysis); [2]conductivity in μS; [3]temperature ^{o}C; [4]all elements in ppm; [5]mean given in bold, other values give range; [6]bd=measurement below instrument detection levels

Table 2. Summary of microweathering data for the Røldal area. Mean given in bold, other values give range of individual measurements; [*]WRT = Weathering rind thickness

Table 3. Rates of chemical erosion and surface lowering for the Røldal area. Values in parentheses are ranges, other values are means. Values in italics are erosion rates (t km^{-2} a^{-1}) calculated from direct measurement of surface lowering (mm ka^{-1}).

Table 4: Correlation matrix of weathering variables for the Røldal area. Values in bold indicate that the correlation is significant at the 99% confidence level.