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11 **The paleoclimatic significance of deformation structures in Neoproterozoic successions**

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28 **Abstract**

29 This paper reviews the different types of soft sediment deformation structures that can form in
30 glacial and non-glacial settings and explores the potential use of these structures in resolving
31 long standing debates in paleoenvironmental reconstructions of Neoproterozoic glacigenic
32 successions. Soft sediment deformation structures are created when compressional, gravitational
33 or shear stress is applied to unlithified sediments during or shortly after deposition. In subglacial
34 or ice marginal glacial settings, shear and compressional stress imparted by ice moving on top of
35 a deformable substrate or advancing ice bulldozing unlithified ice marginal sediments can result
36 in a wide range of folding, faulting and shear structures. In glaciofluvial or stagnant ice marginal
37 setting, gravitational collapse and remobilization of sediments associated with the melting of
38 buried ice can result in normal faulting and broad folding. In glaciolacustrine or glaciomarine
39 settings, compressional, shear and gravitational types of deformation structures can occur as a
40 result of grounding ice or icebergs, rapid sedimentation and reworking downslope associated
41 with high sedimentation rates. In non glacial settings, similar deformation structures can form as
42 a result of slope instability and reworking of sediments downslope, rapid sedimentation, seismic
43 shaking, wave induced shearing or loading.

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45 In this context, two case studies are presented to demonstrate the type of paleoenvironmental
46 information that an analysis of deformation structures can provide. In the first case study,
47 analysis of deformation in the Port Askaig Formation (Scotland) reveals a distinctive
48 stratigraphic distribution of deformation structures. The types of deformation observed together
49 with their recurrence over several 100s of metres and their basinal context are used to infer a

seismic origin for the deformation, which in turn suggests a significant tectonic control on sedimentation atop a record of ice margin fluctuations in a glaciomarine setting. In the second example, analysis of deformation in the Smalfjord Formation (northern Norway) provides strong evidence for deformation by active ice overriding glaciofluvial deposits. The types of deformation in this example, together with its complexity, scale and associated facies, provide the strongest case for ice marginal deformation. In sum, analysis of deformation structures together with analysis of structural geology, stratigraphy, facies and facies associations can provide additional constraints on paleoenvironmental conditions at the time of deposition, which can help us refine or test paleoenvironmental models proposed for this critical time period in Earth history

Key words: brittle deformation; ductile deformation; Neoproterozoic; soft sediment deformation; glaciotectonics; seismites

1. Introduction

Deformation of unlithified sediments occurs in a wide range of depositional settings and as a result of different syn- to post-depositional triggers including tectonic processes, downslope reworking of sediment, wave loading or frictional drag associated with currents, rapid sedimentation, thermal expansion and contraction in periglacial settings and stresses related to advancing, grounding or stagnating glacier ice. In many studies of modern and recent glacial sediments, it is used to infer the presence and dynamics of active ice (e.g. Aber et al. 1989; Woodworth-Lynas and Guigne, 1990; Benn and Evans, 1996; Phillips et al., 2002; McCarroll and Rijdsdijk, 2003). In contrast, deformation in ancient glacial successions has primarily been attributed to non-glacial or periglacial processes, with much fewer studies referring to examples of ice marginal, subglacial or grounded ice deformation (Visser et al., 1984; Rocha Campos et al., 1994; Rocha Campos and Canuto, 2000; Le Heron et al., 2005; Arnaud, 2008, Domack and Hoffman, 2011). This is likely in part due to a preservational bias towards glaciomarine facies in ancient strata (Bjørlykke, 1985; Eyles, 1993). However, some Neoproterozoic successions do contain terrestrial facies (e.g. Deynoux, 1985; Rieu et al., 2006; Arnaud, 2008) and records of grounded ice or drifting icebergs should be preserved in relatively shallow marine settings, given sufficient accommodation space, and therefore preservation potential, as the basin develops over time. In many Neoproterozoic successions, the focus however remains on identifying the more typical indicators of glacial conditions such as diamictite with striated, faceted or extrabasinal clasts, and laminated sediments with outsized clasts interpreted as ice rafted debris (Arnaud et al. 2011). In some cases, excellent exposures have permitted a detailed facies approach where sedimentary facies associations are attributed to various glacial and interglacial settings (e.g. Allen et al., 2004; Rieu et al., 2006; Domack and

Hoffman, 2011; Le Heron et al., 2011). But even in these, deformation, while mentioned in some cases, has rarely featured prominently. The data presented here suggests that a detailed analysis of all types of deformation structures can yield significant paleoclimatic information by providing evidence for glacial, periglacial and/or non-glacial processes. These data can then be used together with other sedimentary characteristics in their stratigraphic context to identify periods of time where deposition is dominated by glaciogenic, periglacial and/or non-glacial processes, allowing for more robust and refined paleoclimatic reconstructions.

Deformation is defined as any change in form or shape resulting from an applied force (Twiss and Moore, 2007; Fossen, 2010). Deformation in unlithified sediments can be characterized as brittle or ductile and described using standard structural terminology to document the type, offset, strike and dip of fault planes as well as the scale and type of folding (Evans and Benn, 2004; Twiss and Moore, 2007; Benn and Evans, 2010; Fossen, 2010). Owen (1987) suggested that deformation structures in unconsolidated sand form as a result of a driving force (such as gravity, uneven confining load, tangential or vertical shear stress or reverse density gradient), a deformation mechanism (commonly liquefaction or fluidization) and a trigger (such as a seismic event, ice advance, or rapid deposition). This framework underscores the fact that deformation structures, as with other sedimentary structures, are not diagnostic of any one depositional environment or trigger. Although folds can be linked to a specific stress or driving force, they clearly occur as a result of a variety of triggers in a range of depositional settings (e.g. ice front, sediment gravity flow deposit, and seismically-active region). Similarly, ball and pillow structures are known to develop in sediments that exhibit reverse density gradients but triggers include rapid sedimentation or seismic shaking. Ultimately, the scale and types of deformation

structures and inferred paleostress orientations together with the associated undeformed facies, their stratigraphic context, and the nature of the depositional setting and sedimentary basin in which they formed must be used to infer the most plausible trigger for deformation and the most likely paleoenvironmental conditions associated with that deformation (McCarroll and Rijdsdijk, 2003; Owen et al. 2011).

In this paper, deformation structures observed in two Neoproterozoic glacigenic successions are presented to demonstrate how these can be used to refine depositional models and shed light on the environmental conditions of Neoproterozoic glaciations. The refined paleoenvironmental reconstructions can then be used more effectively to test existing global climate change models for the Neoproterozoic time period. To provide context, deformation of unlithified sediments in glacial and non-glacial environments is first reviewed.

2. Deformation in glacial environments

Deformation of unlithified sediment can occur as a result of a variety of processes in glacial environments (Hart and Roberts, 1994; McCarroll and Rijdsdijk, 2003; Evans et al., 2006; Benn and Evans, 2010) . In some cases, the applied force is shear or compressional stress from the moving ice, whereas in others, the applied force is related to gravitational instability, gravitational collapse or reverse density gradients (Fig. 1). Most research has focused on the former with many special volumes and case studies focusing on the glaciectonic forces common in subglacial and ice marginal settings (e.g. Croot, 1988; Aber et al., 1989; Benn and Evans, 1996; Maltman et al., 2000; Benediktsson et al. 2008). The following review is by no means exhaustive. Specific references have in part been chosen as they contain illustrations focused on

the sedimentary and structural characteristics of the deformation that in turn can become useful analogues for Neoproterozoic successions. This review provides a summary of the various conditions that result in deformation, the types of deformation structures found in different glacial settings and the types of facies associations in which these are commonly found. This will hopefully assist in future recognition and interpretation of deformation structures in Neoproterozoic glacigenic successions.

2.1 Subglacial settings

In subglacial settings, ice movement over an unlithified substrate creates shear stress that result in a variety of deformation structures (Figs. 1, 2 and 3). Typically, deformation occurs in the top metre under the sediment-ice interface as shown by the seminal studies at Breiðamerkerjökull of Geoffrey Boulton and others, although deformation has been documented up to 10's of metres below the interface (Rijsdijk et al., 1999; Evans et al., 2006). The style and extent of deformation will in part depend on the ice characteristics (velocity and basal shear stress), sediment characteristics (texture, bedding, and degree of heterogeneity) and subglacial hydrology (Hart, 1995; Benn and Evans, 1996; Boulton et al. 2001; Evans et al., 2006). Benn and Evans (1996) suggested that the increase in stress towards the ice-sediment interface would result in an upward increase in the severity of strain signatures. As a result, a typical vertical facies succession would include a basal zone with visible deformation structures, transitioning gradually into an upper zone of completely homogenized sediments. Although primarily affected by shear stress, subglacial sediments closer to the ice margin may experience compressional stresses (Hart and Boulton, 1991).

Deformation structures formed as a result of subglacial shear and compressional stress include attenuated bedding and boudins, shear folds and shear planes, and faulting (normal, thrust and reverse) (Figs. 2 and 3; Benn and Evans, 1996; Hart and Boulton, 1991; Boyce and Eyles, 2000; Philips et al. 2002; Evans et al., 2006). These represent a continuum of strain with rooted or open folds representing low strain and highly attenuated bedding to tectonic laminae representing high strain (Hart and Boulton, 1991). In addition, the confining pressure of the ice and a frozen substrate or foreland can lead to overpressurized subglacial conditions, hydrofracturing and injection of sediments that result in upward directed, downward directed or ‘burst out’ clastic dikes (Fig. 2; Larsen and Mangerud, 1992; Boulton and Caban, 1995; Dreimanis and Rappol, 1997; Rijdsijk et al., 1999; Le Heron and Etienne, 2005, Benn and Evans, 2010).

Subglacial deformation structures range in scale from cm to m and have been documented in subglacial tills and in a range of other sedimentary facies that have been overridden by ice such as glaciofluvial sand and gravel, variable ice marginal sediments or glaciomarine sand and mud. Micro-scale structures related to subglacial deformation have also been documented and are commonly used in the analysis of recent glacial deposits (see Menzies, 2000a for a review). However, considering the potential for post depositional overprinting by tectonic forces, micro-morphological analysis has largely not been applied to ancient glacigenic deposits, with one exception known to the author (Menzies, 2000b).

2.2 Ice marginal settings

Ice marginal settings are highly dynamic, with sediments being affected by ice margin fluctuations, ice surging, gravitational instability or collapse related to differential melting of

buried ice and topographic inversion, as well as reworking by glaciofluvial processes (Boulton, 1972; Lawson, 1982). This results in high facies variability that, in turn, responds very differently to applied driving forces, forming a wide, and at times highly complex, range of deformation structures (Figs. 1, 2 and 3). Advancing and surging ice may bulldoze ice marginal sediments resulting in longitudinal and compressional deformation structures such as broad to recumbent folding, thrust faulting, nappes and decollement surfaces (Fig. 2F); Hart and Boulton, 1991; Hambrey and Huddart, 1995; Boulton et al., 1999; Bennett, 2001; Phillips et al., 2002). These deformation structures can affect sediments up to several kilometres away from the advancing ice front (Boulton et al., 1999). Chaotic bedding, remolded intraformational clasts, flow nose or flow lobes with deformation (shear or compressional) in adjacent or underlying sediments, as well as occasional attenuated lenses of silty clay, will develop in subaerial sediment gravity flows as a result of sediment remobilization in this topographically-uneven and meltwater-rich setting, whereas normal faulting is commonly attributed to the melting of buried ice (Lawson, 1982, 1988; Benn and Evans, 2010). Clastic dikes may form in the glacier foreland under high pore water pressures as meltwater accumulates between the advancing ice and the permafrost ahead of the ice margin (Boulton and Caban, 1995). Ice marginal deformation structures range in scale from cm to 100's of metres and are found in a wide range of sedimentary facies that are often spatially highly variable.

2.3. Glaciofluvial settings

In glaciofluvial settings, deformation is a relatively minor process and is primarily related to gravitational instability or collapse as buried ice melts over time (Fig. 1; MacDonald and Shilts, 1975; Rust and Romanelli, 1975; Aitken, 1998). Interestingly, Eyles (1977) describes an

example of deformation from melting buried ice developed in subaqueous outwash. When it is well drained and cohesive, sand and gravel overlying melting ice will fail in a brittle manner, leading to extensional faulting (Fig. 4A). When it is still water saturated or relatively non-cohesive, the sediments become broadly folded as the supporting ice melts (Fig. 4B). Deformation in glaciofluvial settings range in scale from cm to m, though m-scale offsets along faults have been known to affect packages of glaciofluvial sediments 10's of metres in thickness (E. Arnaud, unpublished data).

2.4. Glaciomarine and glaciolacustrine settings

In glaciomarine and glaciolacustrine settings, deformation occurs in both ice proximal and ice distal settings. These range in scale from cm to 100's of metres and occur in predominantly fine-grained facies such as diamicton, and laminated or massive mud with or without ice rafted debris (Fig. 1). Deformation associated with seasonal ice or iceberg grounding (Fig. 5) will be found in both ice proximal and shallow shelf distal settings as boulders or ice are dragged along the substrate (Thomas and Connell, 1985; Woodworth-Lynas & Guigne, 1990; Gilbert 1990; Lønne, 1995; Dionne, 1998) and can be identified by cm - m scale plough and furrows features with faulting and folding of underlying sediment (Woodworth-Lynas and Guigne 1990; Eden and Eyles, 2001; Winsemann et al., 2003). It may be difficult to identify these in ancient strata without a cross-sectional view of the plough and furrow feature, considering the associated deformation (sub-ice scour and ice keel turbates) will be very similar to those formed as a result of subglacial or ice marginal compression and shear stresses. However, with good exposure, contextual information (facies associations and overall facies architecture) should allow

distinction between the two (see discussion of competing interpretations of deformation structures attributed to drifting ice in Eyles et al. (2005)).

In ice proximal settings, shear structures, thrust faults and folds associated with advances of the ice margin's grounding line have been documented and are similar to terrestrial glacitectonic structures (Fig. 6; Lønne, 1995; Benn, 1996). However, deformation in these subaqueous settings tend to differ from their terrestrial equivalent due to the water saturated and unfrozen nature of the sediments. In addition, glacitectonic or shear structures will be accompanied by deformation associated with abundant gravitational instability (Lønne, 1995; Plink-Björklund and Ronnert, 1999).

In ice distal marine or lacustrine settings, deformation will also occur as a result of gravitational instability and reverse density gradients (Fig. 6). Although these processes are common in non-glacial water bodies, they can be even more common in glaciated basins as a result of high sedimentation rates, oversteepened slopes and the impact of calving icebergs. Gravitational instability is often very common in steep sided fjords (e.g. Prior and Bornhold, 1990), oversteepened continental shelf edges (Weaver et al. 2000) or subaquatic fans (e.g. Lønne, 1995). In ice marginal lakes gravitational instability can also be enhanced as a result of rapid drainage associated with the failure of an ice- or moraine-dammed outlet. This instability will result in sediment gravity flow deposits with slump structures, folding, chaotic, contorted and convolute bedding and deformed inclusions of other lithofacies as heterogeneous sediments are incompletely mixed during downslope transport (Fig. 6; Eyles and Eyles, 2000). Shear planes, shear structures, flow noses and dish structures have also been reported from sediment gravity

flow deposits (Nardin et al., 1979; Mulder and Alexander, 2001), whereas clastic dikes have been reported intruding downward at the base of slump-generated megachannels (Eyles and Lagoe, 1998).

Reverse density gradients commonly occur in marine or lacustrine settings where coarse-grained sediments are rapidly deposited on finer-grained sediments (e.g. flood flow discharge events, storm events, and sediment gravity flows; Anketell et al., 1970; Allen, 1982; Owen 1987). In glaciated basins, fluctuations in discharge events associated with diurnal melting cycles and rapid changes in meltwater or lakewater discharge outlets can all increase the occurrence of rapid sedimentation events and the formation of reverse density gradients. These conditions will result in the formation of load casts, ball and pillow structures and water escape structures (Fig. 6).

2.5 Periglacial settings

Lastly, deformation in periglacial settings occurs as a result of seasonal changes in the thermal characteristics of the substrate above permanently frozen ground or permafrost. These thermal fluctuations and the resulting ground ice can lead to the formation of involutions, faulting, brecciated clasts, and ice and sand wedges (Figs. 1, and 7; French, 1986; van Vliet-Lanoe et al., 2004). Both ice and sand wedges are ultimately preserved as wedges of sand or silt with near vertical internal stratification; upturning of adjacent strata is also commonly reported (Fig. 7B). The wedges and involutions range in scale from cm to m and occur in a wide range of facies.

Periglacial involutions are very similar in form to soft sediment deformation resulting from seismic shaking and rapid sedimentation (French, 1986). Likewise, in the absence of a

periglacial context, sandstone wedges can be described as clastic dikes with a number of possible trigger mechanisms due to their similarity in form (see discussion in Eyles and Clarke, 1985). As such, confident identification of a periglacial setting will require multiple lines of evidence and careful analysis (French, 1986; Van Vliet-Lanoe, 2004). In this context, the only Neoproterozoic sandstone wedges that have been convincingly shown to have formed under periglacial conditions were those documented by Williams (1986) in Australia. In that case, the sand wedges were associated with anticlines and teepee structures, earth mounds, involutions and breccia, that are typically attributed to cold and arid climatic conditions. In other cases, documented in Williams (1986), the origin of the wedges has remained contentious as these other indicators are not present (see discussion of the Port Askaig Formation wedges in section 4.1.2).

3. Deformation in non-glacial environments

Deformation of unlithified sediments also occur in non-glacial environments as a result of various triggers and driving forces, leading to the formation of very similar deformation structures as the ones described above (Figs. 1 and 8). This review is also by no means exhaustive but will alert readers to potential alternative driving forces and triggers that will need to be considered when assessing the significance of deformation structures in Neoproterozoic successions.

3.1. Deformation from seismic shaking

Seismic shaking is one of the most commonly-cited triggers for soft sediment deformation in non-glacial environments (Fig. 8; Owen et al., 2011). Deformation structures interpreted as

seismites include convolute and contorted bedding, recumbently folded cross strata, water escape structures, load casts, ball and pillow structures, micro-faulting, clastic dikes and pseudonodules (Figs. 9A and 9B; Rosetti, 1999; Jones and Omoto, 2000; Moretti and Sabato, 2007; Berra and Felletti, 2011). Although they can be larger, these are usually described as cm - 10 cm scale structures that affect different facies types within a single succession. They can occur in a wide range of settings, but they are often reported from lacustrine deposits. Several very useful and comprehensive reviews and case studies of seismites have been published in recent years providing various criteria against which to evaluate soft sediment deformation (e.g. Obermeir, 1996; Jones and Omoto, 2000; Montenat et al., 2007). Positive identification is confirmed by demonstrating the wide regional extent of the deformed stratigraphic horizon, the recurrence of deformation at distinct stratigraphic intervals, comparison with deformation structures produced using shaking tables under controlled laboratory conditions, independent evidence of tectonic activity in the region and discounting alternative triggers for deformation such as wave-induced shear, wave loading, reverse density gradients and sediment gravity flows (Rosetti, 1999; Moretti et al., 1999; Jones and Omoto, 2000). In addition to this approach, Owen et al. (2011) emphasize the need to consider the relationship between the deformation structures and facies types as this can help distinguish between an endogenic vs exogenic origin.

3.2. Deformation from current induced shear and wave loading

A variety of folding in sand has been documented in shallow marine, lacustrine and fluvial settings and attributed to current-induced shear and wave loading (Fig. 8; e.g. Owen 1987; Owen, 1995; Molina et al., 1998; Røe and Hermansen, 2006). These include broad open to recumbent folding produced by shear stresses at the upper contact of cross-bedded sand (Fig. 9C)

as well as load casts and water escape structures. These structures range in scale from cm to 10's of cm. Molina et al. (1998) attributed load casts and water escape structures to wave loading rather than seismic activity because of the consistent association between deformation structures and the deposition of tempestites. A strong association with hummocky cross stratification or the gradual lateral and vertical transition between deformed cross bedded sand to undeformed massive sand (Fig. 9C; Røe and Hermansen, 2006) would also warrant an autogenic rather than glacial or seismic origin. The high lateral facies variability of ice proximal settings and the relative complexity of glacial tectonic deformation is also notable compared to the relatively simple deformation found in non-glacial marine and fluvial deposits.

3.3. Deformation in sediment gravity flows

Sediment instability and downslope reworking of sediments can produce deformation structures within sediment gravity flow deposits (Fig. 8). This process can occur in terrestrial settings, relatively shallow subaqueous fans in marine and lacustrine settings as well as in deeper subaqueous settings. The deformation structures found in these deposits include slump structures, folding, chaotic, contorted and convolute bedding, deformed inclusions of other lithofacies, basal shear planes, shear structures, flow noses, and clastic dikes (Figs. 9D to 9F). Load casts and water escape structures form when reworked sediments are rapidly deposited and where reverse density gradients occur (Figs. 9G, 9H). A glacial influence on deposition can be discerned only when lateral association with ice proximal deposits can be demonstrated, clasts that exhibit evidence of subglacial transport (e.g. striated, faceted or bullet-shaped boulders) are found within these deformed sediments or if the deposits are closely associated with ice rafted debris in laminated facies. In addition, Plink-Björklund and Ronnert (1999) suggested that the

only difference between glacial and non-glacial coarse-grained subaquatic fans in relatively shallow water bodies is the presence of resedimented deformed diamicton originating from the ice margin.

3.4. Deformation from post depositional tectonics

Post depositional tectonic history can account for additional deformation, either as the deposits are still unlithified or much later as they have become lithified (Fig. 8). Although most geologists would recognize tectonic deformation (Fig. 10), it can sometime resemble those formed under glacial or glacially influenced conditions. Indeed much of the subglacial and ice marginal literature refers to glacitectonics, drawing upon advances made in structural geology. Common post depositional deformation structure of tectonic origin include augen structures and tectonic boudins, faulting and folding associated with collisional or extensional tectonics and shearing structures along fault planes and shear zones. It is particularly important to consider this aspect of deformation considering the active tectonic setting of many Neoproterozoic basins and their long geologic history with great potential for post-depositional tectonic overprinting.

4. The significance of deformation in Neoproterozoic successions

Data from two Neoproterozoic glacial successions-the Port Askaig Formation (Scotland) and the Smalfjord Formation (Norway) are presented here to demonstrate how the analysis of deformation structures can contribute significantly to the reconstruction of depositional conditions and the identification of larger scale controls on sedimentation. The signature of glacial processes and the relative importance of climate and tectonic controls on sedimentation has been, and continues to be, the subject of debate in these Neoproterozoic successions as

diamictites are attributed by some to sediment gravity flows related to sediment instability and tectonic activity in the basin, by others to sediment gravity flows related to glacigenic sediment instability and by others still as true tills or glacial deposits directly deposited by ice (Harland, 1964; Crowell, 1964; Schermerhorn, 1974; Hambrey and Harland, 1978; Eyles, 1993; Crowell, 1999; Arnaud and Eyles, 2002a, 2002b; Arnaud and Eyles, 2006). In this context, and as demonstrated in these two case studies, deformation can be useful in identifying evidence to support and distinguish between climatic and tectonic controls as well as glacial and non-glacial processes, so that together with all the other sedimentary evidence, a stronger case can be made for the most likely interpretation.

4.1 Port Askaig Formation

4.1.1 Geological Background

The Port Askaig Formation consists of a thick succession of diamictite, conglomerate, sandstone and mudstone thought to record glaciomarine conditions (Fig. 11; Kilburn et al., 1965; Spencer, 1971; Eyles, 1988; Arnaud and Eyles, 2006). The excellent exposures of the Garvellach Islands provide the most continuous, laterally extensive and well preserved sections of the first three members of the the Port Askaig Formation (Fig. 11). The first member is distinguished by a dominance of dolimitic diamictite, a very thick unit called the Great Breccia, and the overlying Disrupted Beds, which are most notable for their pervasive deformation. The second member is identified on the basis of the appearance of extrabasinal clasts and thick packages of diamictite interbedded with stratified sediments (conglomerate, sandstone and mudstone). The third member caps the succession at this site and consists of thick packages of cross-stratified sandstone interbedded with relatively thin diamictite units.

The Port Askaig Formation occurs at the base of the Argyll Group, within the Dalradian Supergroup (Kilburn et al., 1965; Spencer, 1971). The Argyll Group is thought to record the onset of extensional tectonic activity based on the presence of ‘slides’ interpreted as Neoproterozoic-age listric faults, increasing volcanic activity, abrupt lateral thickness changes, and seismites (Anderton, 1976; Harris et al., 1978; Soper and Anderton, 1984; Anderton, 1985; Harris et al., 1993). Most recently, Arnaud and Eyles (2006) suggested that the Great Breccia, which they interpreted as a submarine mega-slump deposit (Arnaud and Eyles, 2002a), and the numerous horizons of soft sediment deformation structures within the Port Askaig Formation provided evidence of localized faulting, consistent with the onset of extensional tectonic activity during this time period. While the tectonic setting of the Port Askaig Formation has been the subject of several studies, its paleogeography and paleolatitude remains unconstrained due to Caledonian overprinting (Stupavski et al., 1982) and the controversy related to the location of the Scottish promontory (Dalziel and Soper, 2001).

The Port Askaig Formation is a classic and well studied Neoproterozoic glacigenic succession, with J. Thomson being the first to suggest its glacial origin in 1871. It is constrained geochronologically by U-Pb dates in the underlying Grampian Shear Zone (c. 806 Ma) at the base of the Dalradian Supergroup and the overlying Tayvallich Volcanics (c. 601 Ma) at the top of the Argyll Group (Halliday et al., 1989; Noble et al. 1996; Dempster et al. 2002). Both of these dates are far removed stratigraphically from the Port Askaig Formation, and thus provide only loose constraints on its age. A recent Re-Os date on the underlying Ballachulish Slate

Formation (Appin Group) provides a maximum age of c. 650 Ma for the Port Askaig Formation (Rooney et al. 2011).

Like other Neoproterozoic successions, the Port Askaig Formation is associated with carbonates with the underlying Lossit (Islay) Limestone and the dolomitic members of the overlying Bonnahaven Formation (Spencer, 1971; Spencer and Spencer, 1972; Fairchild, 1980), although the mixed siliclastic-carbonate nature of the Bonnahaven Formation and the thick package of clastic sediments between the last evidence of glacial conditions and the carbonate strata makes it an atypical cap carbonate. Nonetheless, chemostratigraphic studies (C and Sr isotopes) have been used to try to constrain the age of the Port Askaig Formation based on comparison with geochemical signatures of other better dated successions (Brasier and Shields, 2000; Thomas *et al.*, 2004; McCay *et al.*, 2006; Prave *et al.*, 2009; Sawaki et al. 2010). Despite some concern over the primary nature of the carbon signatures, an early to Mid- Cryogenian age for the Port Askaig Formation was deemed most likely based on the strontium values and regional and global chemostratigraphic and lithostratigraphic correlations. The recent Re-Os date however, is inconsistent with these data, prompting continued debate over the exact age of this glacial successions.

The Garvellach Islands have been the subject of detailed sedimentological studies. Most of this work supports a glaciomarine setting for the accumulation of these sediments, though some emphasize fluctuations of a terrestrial grounded ice margin in this glaciomarine setting (Spencer, 1971; Benn and Prave, 2006), whereas others emphasize a component of tectonically-driven sediment instability within a glacially-influenced marine basin (Eyles, 1988; Arnaud and Eyles,

2002a; Arnaud and Eyles, 2006). In the model that emphasizes terrestrial conditions, some of the deformation structures are attributed to periglacial, subglacial or proglacial conditions, whereas others were attributed to seismic shaking and reverse density gradients (Spencer, 1971; Benn and Prave, 2006). In contrast, Arnaud and Eyles (2006) focused on the stratigraphic context, facies associations and types of deformation structures and suggested that these are primarily related to localized seismic activity.

4.1.2 Deformation structures

4.1.2.1 Description

The Port Askaig Formation is characterized by two deformed sedimentary units (the Great Breccia and the Disrupted Beds; Arnaud and Eyles, 2002a; Benn and Prave, 2006), as well as numerous horizons of deformation structures such as clastic dykes (Eyles and Clark, 1985), load casts, and contorted and convolute bedding (Fig. 12; Spencer, 1971; Arnaud and Eyles, 2006).

The Great Breccia is up to 50 m in thickness and consists primarily of diamictite with matrix supported clasts ranging up to 100 m in diameter (Spencer, 1971). It was subdivided into three distinct units based on detailed mapping of several wave-cut platforms and cliffs in the Garvellach Islands (Arnaud and Eyles, 2002a). The first unit is a thick (10's of metres) diamictite with very large (up to 100 m in diameter) clasts of variable lithologies. The second unit consists of diamictite interbedded with conglomerate, sandstone and mudstone, whereas the third unit consists of a single diamictite unit dominated by m-sized dolomite boulders. Deformation is observed at various scales (cm- 100 m) within the Great Breccia and affects the internal structure and/or outer geometry of clasts of all sizes, as well as the overall structure of

the diamictite (Arnaud and Eyles, 2002a). Although clast composition is highly variable, brittle and ductile deformation primarily affects sandstone, dolomite or interbedded sandstone and siltstone clasts-typical facies of the underlying Port Askaig and Lossit (Islay) formations. Deformed clasts are generally broadly to tightly folded (including recumbent), or have contorted and convoluted outer margins, whereas some clasts exhibit severe and complex deformation ranging from broadly folded to disaggregated and brecciated with folded and angular blocks floating in a chaotically folded matrix (Fig. 13). The diamictite in units 1 and 2 are massive or stratified, with some exhibiting chaotic bedding imparted by variable clast abundance or discontinuous folded to contorted sandstone or siltstone beds and stringers. Benn and Prave (2006) also report the presence of listric faults and shear zones associated with the rafts in the Great Breccia.

Another distinctive deformed sedimentary unit is the Disrupted Beds, which overlie the Great Breccia. This unit is approximately 25 m thick and consists of interbedded units of siltstone, thin clast horizons, conglomerate and diamictite as well as discontinuous carbonate rich sandstone beds (Figs. 14 and 15). The notable blue colour of the siltstone and diamictite matrix as well as the pervasive deformation found within the Disrupted Beds makes it a readily identifiable stratigraphic marker in the region. Deformation within the Disrupted Beds, again ranges over a variety of scales (cm - 10's of m) and includes m-scale chaotic folding of several interbeds, or cm to m scale chaotic bedding within diamictite beds, sedimentary boudinage and small scale folding (< 5cm) of carbonate-rich sandstone beds. Other notable characteristics include lateral and vertical facies changes from deformed discontinuous sandstone beds in siltstone to chaotically stratified diamictite and on to massive diamictite observed in outcrop over 100's of

metres; alternating deformed and undeformed interbeds; and first appearance of outsized extrabasinal clasts in laminated siltstone facies (Spencer, 1971; Arnaud and Eyles, 2002a; Benn and Prave, 2006). Benn and Prave (2006) also document the presence of fragmented clasts, augen like structures and bifurcating laminae around boudins.

In addition to these two thick intervals of deformed sediment, there are numerous discrete horizons of deformed sediment within the ~450 m succession exposed in the Garvellach Islands (Figs. 12, 16, 17, and 18; See Spencer, 1971; Eyles and Clark, 1985; Arnaud and Eyles, 2006 for detailed sedimentological descriptions). Some of these deformed horizons are observed within certain outcrops and only extend laterally over 10's of metres, whereas others can be traced across several of the Garvellach Islands over several kilometres (Fig. 17). The more notable deformation structures are syn-sedimentary clastic dikes (Fig. 18), including those described as sandstone wedges by Spencer (1971), and focused on by Eyles and Clark (1985). These are up to 4m in length and up to 30 cm in width and consist of sandstone or muddy sandstone with or without clasts intruding downward into finer grained facies (diamictite, mudstone). They have sharp to diffuse outer boundaries that can be relatively planar to highly contorted and exhibiting flame-like structure. Some exhibit internal stratification based on differences in texture or lithology; this internal stratification, like the outer contacts, can be relatively planar to highly contorted. Most however are relatively massive, and none of the ones that intruded interbedded sediments show the upturning of adjacent laminae. These clastic dikes are also notably found within the underlying Lossit (Islay) Formation. In three cases, the dikes are expressed as polygonal nets on bedding plane surfaces (Fig. 18B). Other common deformation structures include loading (referred to as downfold structures by Spencer (1971); cm to 10 cm scale), and

beds with convoluted, chaotic or folded internal stratification (cm to m scale; Spencer's (1971) soft sediment folds), with less common examples of ball and pillow structures (cm - 10 cm scale; Fig. 16). These deformation structures occur at over 30 different horizons within the 450 m thick succession, bounded by undeformed sediments. The deformed horizons are most common between the Disrupted Beds and the giant cross-beds (Fig. 12), a stratigraphic package characterized by variable sedimentary facies, compared to the lower diamictite-dominated member and the upper sandstone-dominated member of the Port Askaig Formation.

4.1.2.1 Interpretation

The three units within the Great Breccia were interpreted as recording the mass failure of Port Askaig and Lossit (Islay) limestone sediments along a localized fault scarp based on analogous stratigraphic and sedimentary characteristics in carbonate megabreccia (Arnaud and Eyles, 2002a). The Disrupted Beds were interpreted as continued sediment instability and gravity-driven deformation, following the initial failure and a consequent deepening of water depths. The recurring horizons of soft sediment deformation were also interpreted as signature of tectonic activity within the basin based on their similarity in form to seismites, their recurring stratigraphic distribution, their extent, and the active tectonic setting of the Port Askaig Formation (Arnaud and Eyles, 2006).

As suggested by reviews of the seismite literature, alternative triggers must be considered. Indeed, the diamictite of the Great Breccia has been interpreted as recording glacitectonic deformation and the wedge-like clastic dikes as periglacial in origin (Spencer, 1971). Other interpretations of the clastic dikes include subglacial or proglacial hydrofracturing as well as

injection at the base of slump-generated megachannels (see above review). Benn and Prave (2006) interpreted the Great Breccia and overlying Disrupted Beds as recording an ice advance with proglacial (Great Breccia) and subglacial (Disrupted Beds) deformation, though they conceded that the sedimentary characteristics of the Great Breccia could support both a slump or proglacial glacitectonic interpretation.

A tectonic control on the deformation as proposed by Arnaud and Eyles (2002) is deemed most likely for the following reasons. The detailed mapping of the Great Breccia revealed it is a composite unit made up of undeformed and deformed sedimentary facies typical of sediment gravity flow deposits such as grading, lenticular geometry and load casts. The alternating deformed and undeformed character of these units also suggests intermittent syn-depositional deformation rather than deformation of all three units by an advancing ice margin. Stratigraphically, the flooding surface at the base of the Disrupted Beds is best explained by the tectonic model. This change from coarse to fine grained sediments, as well as evidence for subaqueous sedimentary facies that are undeformed, is not well explained in Benn and Prave's (2006) glacitectonic model. In addition, typical facies associations with glaciofluvial and subaqueous outwash expected in subglacial and ice marginal settings are absent. Benn and Prave (2006) suggested that the vertical change in the severity of deformation in the Disrupted Beds was typical of incremental strain profiles found in subglacial tills. However, this does not explain the same changes that occur in the lateral dimension. One possible scenario combines these two models in order to explain the shattered clasts and incremental strain profile reported by Benn and Prave (2006): the Disrupted Beds record sediment instability and reworking downslope in a marine setting as originally proposed by Spencer (1971) and these are

subsequently overridden by ice. However, on balance, a glacitectonic origin for the Great Breccia is deemed unlikely.

A stratigraphic and facies-based analysis of the deformation that occurs higher up in the succession supports a tectonic origin (Arnaud and Eyles, 2006). Loading, ball and pillow structures and various soft sediment folding can result from a wide variety of triggers and conditions. However, the horizons of deformation structures found above the Disrupted Beds affect a wide variety of facies, suggesting an external rather than endogenic trigger, unrelated to depositional processes. The origin of the sandstone wedges has been discussed at length by Spencer (1971) and Eyles and Clark (1985) and close reading of the two comprehensive descriptions and interpretations of possible trigger mechanisms have left some undecided because the depositional model proposed for the overall succession (terrestrial and glaciomarine *versus* tectonically influenced glaciomarine) seems to play a key role in weighting one interpretation over another. In view of the above review of deformation in different settings, the sandstone wedges and other clastic dikes are not associated with megachannels or other periglacial (teepee structures, breccia, involutions) or subglacial (shear structures or incremental strain profiles in associated diamictite) indicators, making a slump-generated, periglacial or hydroclastic fracturing origin unlikely. In addition, recent seismite studies have clearly demonstrated the formation of sandstone wedges and more tabular clastic dikes as a result of seismic shaking (Montenat et al., 2007). In the end, their stratigraphic recurrence and spatial extent, their similarity with other seismically-induced deformation structures, consideration of alternative triggers and the evidence for active extensional tectonic activity locally and regionally

in the Dalradian succession, makes a seismic origin and a tectonic control most likely for the sandstone wedges, clastic dykes and other deformed horizons.

4.2 Smalfjord Formation

4.2.1 Geological background

The Smalfjord Formation of the Vestertana Group is composed of diamictite interbedded with sandstone, conglomerate, and mudstone, which are best exposed along fjords in Finnmark, northern Norway (Føyn 1937; Spjeldnaes, 1964; Bjørlykke, 1967; Banks et al., 1971; Edwards 1975; Føyn and Siedlecki, 1980; Edwards, 1984). Stratigraphically, it unconformably overlies fluvial and shallow marine deposits of the Vadsø and Tanafjord groups and is overlain by the deep water clastics of the Nyborg Formation (Reading and Walker, 1966; Bjørlykke, 1967; Banks et al., 1971). Its tectonic setting has been characterized as pericratonic, and influenced by the adjacent extensional Timanian basin (Siedlecka, 1975; Røe 2003; Siedlecka et al., 2004; Nystuen et al., 2008).

The Smalfjord Formation is the oldest of two Neoproterozoic diamictite-bearing units in this region, and is constrained chronologically to the late Cryogenian using biostratigraphy (Vidal and Moczydlowska, 1995) and Rb-Sr dates of the underlying Klubbnassen (c. 807 Ma Siedlecka and Roberts, 1992) and of the overlying Nyborg Formation (c 650 Ma; Hambrey and Harland 1985; Ghorokov, 2001). Like the Port Askaig Formation, it is a classic locality with a long history of research starting with Reusch in 1891. Its most notable and famous feature is that it overlies a striated pavement at Oaibac̃c̃annjar'ga (previously referred to as Bigganjargga; Jensen and Wulff-Pedersen, 1996, 1997; Rice and Hoffmann 2000; Laajoki, 2001, 2002).

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592 The Smalfjord Formation is bounded by carbonates, as is common for diamictite-bearing units of
593 Neoproterozoic age. The Grasdalen Formation, uppermost unit of the underlying Tanafjord
594 Group, is characterized by sabkha facies, whereas a relatively thin ‘cap’ carbonate has been
595 documented at the base of the Nyborg Formation at several localities (Edwards 1984; Siedlecka
596 & Roberts, 1992; Siedlecka et al., 2004). Limited chemostratigraphic data from the Nyborg
597 Formation suggests a late Cryogenian age based on comparison with composite
598 chemostratigraphic trends developed in other dated sections (Rice, 2004; Halverson et al., 2005).

599

600 Early research on the Smalfjord Formation based its glacial origin on the striated pavement, and
601 glacial facies such as diamictite and mudstone with dropstones. Edwards (1984) proposed an
602 early phase of valley glaciation from the south-southeast followed by the advance of an ice sheet
603 from the North, leaving behind subglacial tills, glaciofluvial and glaciomarine deposits. Ice
604 margin fluctuations were inferred based on the interbedding of diamictite with other facies
605 (Edwards, 1984). However, as with many other Neoproterozoic successions, the extent of
606 glacial influence on deposition has also been, and continues to be, the subject of much
607 discussion. These debates, with specific examples taken from the Smalfjord Formation, date
608 back to Crowell (1964), during one of the earlier periods of intense Neoproterozoic research,
609 through Schemmerhorn (1974) and Eyles (1993) who drew attention to the possible confusion
610 over diamictites that formed as a result of non-glacial processes. More recently the debates have
611 focused over the origin of the diamictite that overlies the striated pavement at Oaibac’c’annjar’ga
612 and the origin of other diamictite units in the Varangerfjorden and Tarmfjorden areas (Jensen and
613 Wulff-Pedersen, 1996, 1997; Edwards, 1997; Rice and Hoffmann 2000; Laajoki, 2001, 2002;

Arnaud and Eyles, 2002b; Edwards, 2004; Rice 2004; Arnaud and Eyles, 2004). Arnaud and Eyles (2002) showed that while the sediment source may have been glacial, depositional processes recorded in the lateral and vertical distribution of facies were primarily non-glacial at several sites around Varangerfjorden and in the Tarmfjorden area. This distinction is important when considering the severity of glacial conditions proposed in some of the Neoproterozoic paleoclimate models.

4.2.2. Deformation structures

4.2.2.1 Description

Subsequent analysis of other sections along Varangerfjorden revealed extensive deformation indicative of active ice as well as facies associations typical of ice marginal settings (Arnaud, 2008). At Handelsneset, deformed conglomerate and sandstone is exposed in a series of contiguous outcrops or panels over a distance of approximately 50 m (Fig. 19). The deformation is highly variable in style, complexity and scale, with some units being extremely deformed and others completely undeformed within metres of one another. Deformation structures are primarily ductile and include contorted and convoluted bedding, broad to isoclinal folds, flow noses, attenuated bedding, tectonic laminations and shear folds (Figs. 19, 20 and 21). Several outcrops exhibit an upward decrease in deformation (Fig. 20a). These deformation structures affect two of the four stratigraphic units identified at this site (Units A and C; Fig. 19). Principal stress directions measured from the upper deformed zone suggest a paleostress applied from the NE.

4.2.2.2. Interpretation

These deformation structures were formed under compressional and shear stress and as a result of high pore water pressures, loss of sediment shear strength and liquefaction. Based on the styles, scale and complexity of deformation, Arnaud (2008) suggested that a combination of gravity-driven and glaciectonic reworking of these sediments in a proglacial setting was responsible for the observed deformation. Several flow noses and substrate relief were consistent with gravity-driven processes, whereas the abundance of shear structures throughout the deformed zones, the complexity and scale of the deformation, the abundance of ductile deformation in such coarse-grained materials as well as the upward increase in strain typical of subglacial deformation profiles (Benn and Evans, 1996) all pointed to active ice as an important trigger of deformation. The presence of deformed and undeformed stratigraphic units at the site suggests oscillations of the ice margin in this area.

5. Discussion

In the two case studies presented, the glacial origin of these successions remains undisputed. However, the analysis of deformation has significantly added to the understanding of the depositional processes and basin dynamic recorded in these sediments. The deformation structures within the Great Breccia and the Disrupted Beds, the stratigraphic and regional distribution of other deformation structures within the Port Askaig Formation, together with other stratigraphic, structural and sedimentary data, suggest tectonic activity within the basin played an important role in determining the nature of sedimentation (Arnaud and Eyles, 2006). The distribution of deformation combined with sedimentary indicators of glacial conditions (appearance of extrabasinal clasts, evidence for rainout processes -dropstones and clast clusters- in diamictite and laminated mudstone facies) was then used to define intervals where

sedimentation was influenced by tectonic and/or climatic controls. None of the diamictite units showed deformation characteristic of subglacial or ice marginal settings; most diamictite could be attributed to ‘rainout’ processes or sediment instability triggered by tectonic activity and/or abundant delivery of glacial sediments to the marine basin. This better understanding of larger scale controls showed that the signature of glacial conditions was most clear during tectonically quiet intervals of sedimentation and that this interval recorded fluctuations between ice-free and glacially-influenced depositional conditions (Arnaud, 2004; Arnaud and Eyles, 2006).

In Norway, deformation structures at Handelsneset together with other outcrops of the Smalfjord Formation on Skjaholmen Island clearly demonstrate the role played by active wet-based ice (Arnaud 2008). These provide details about terrestrial ice marginal processes rarely described in other Neoproterozoic successions, which tend to represent predominantly more distal, glaciomarine settings. It also provides evidence for spatially variable coverage of ice: active ice at Handelsneset to limited or no glacial influence a few kilometres away across the Varangerfjord at Kvalnes and further afield along Tarmfjorden to the west (Arnaud and Eyles, 2002).

The review of deformation structures in glacial and non-glacial settings clearly demonstrates the need to collect as much information as possible considering no single deformation structure is diagnostic of any specific setting and trigger as they often occur in a number of different glacial and non-glacial settings (e.g. contorted and convoluted bedding, clastic dikes, compressional and shear folds, and faults; Figs. 1 and 8). Analysis of deformation within the broader context of the succession is often very helpful in determining what trigger may have caused these non-unique deformation structures. In this context the depositional setting and basinal context of the

deformation must be considered through an analysis of facies associations and regional stratigraphic and structural trends indicative of the overall development of the basin. For example, are the associated facies typical of delta fronts, terrestrial ice margins or deep marine settings? Is the geometry of the deposits associated with the deformation consistent with one depositional setting over another (e.g. high facies variability in ice marginal settings vs lateral continuity of subglacial or deep marine settings)? Are the stratigraphic changes in facies associations and significant bounding surfaces recording base level or sea level changes consistent with ice margin fluctuations or tectonically active periods of basin development? And how are all these sedimentary clues connected to the nature and distribution (lateral and vertical) of the deformation structures? These lines of questioning can provide information that can then be used to distinguish the most plausible of several competing triggers. In some cases for example, deformation will be associated with specific facies suggesting that the deformation is perhaps linked with the process of deposition or depositional setting (i.e. endogenic control; e.g. Molina et al., 1998; Røe and Hermansen, 2006), whereas in others, deformation will be observed in a wide range of facies suggesting an exogenic control (e.g. seismic activity) on their formation (e.g. Pratt, 1994; Arnaud and Eyles, 2006). In terms of glaciogenic deformation, the structures tend to be associated with other sedimentary indicators of glacial conditions (e.g. faceted, striated or bullet shaped clasts, laminated mudstone with dropstones) or the associated facies and geometries are highly variable as is typical of ice marginal conditions. In addition, documenting the style, scale and complexity of the deformation (McCarroll and Rijdsdijk, 2003), carrying out a structural analysis of the succession to determine the extent of post-depositional tectonic overprinting, as well as measuring paleostress, paleoslope and paleo-ice flow directions, where possible, can also help constrain the most likely trigger of the deformation.

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707 Ideal exposures for this kind of analysis are structurally simple, with enough lateral exposure to
708 take full advantage of facies associations and to get a sense of the complexity and abundance of
709 deformation structures. This information must be considered with the possibility of polyphase
710 deformation histories in mind for three main reasons. First, oscillations of ice margins are
711 common over seasonal to millennial time scales such that sediments may be deformed by active
712 ice several times during one glacial period or ice age (Phillips et al. 2002). Second, glacial
713 settings are highly dynamic environments so that sediment may experience deformation from
714 different triggers over time. A succession may record compressional deformation related to
715 gravity-driven remobilization of sediments as abundant glacial sediments are delivered to a
716 subaqueous setting, followed by shear deformation from overriding ice and lastly, deformation
717 by melting of buried ice during the full meltback of that ice margin (e.g. Weaver and Arnaud,
718 2011). Third, many Neoproterozoic successions are preserved in active tectonic settings
719 (foreland or rift basins). As such, seismic shaking should be considered as one of several
720 possible triggers when stratigraphically-recurring deformation structures, which are similar in
721 form to seismites, are observed. With careful consideration, deformation structures can yield
722 information that can strengthen and enhance the paleoenvironmental reconstruction proposed for
723 any succession.

724
725 Refined paleoenvironmental reconstructions can then provide information on the physical
726 processes as well as the extent and nature of ice cover at the time of deposition and thereby
727 provide context and constraints for the changes observed in the biological and geochemical
728 spheres of the Earth system (e.g. the extent of ice cover and its effect on the carbon cycle and

marine life). In addition, the different global climate change models that have been proposed for the Neoproterozoic (Fairchild and Kennedy, 2007) invoke a number of different scenarios ranging from total ice cover ('hard' snowball Earth) to partial ice cover ('slushball' or Zipper-Rift model), such that the findings summarized above showing spatially and temporally variable ice cover constitute one of several similar sedimentary challenges to the snowball Earth hypothesis (Allen and Etienne, 2008). In the end, a better understanding of the nature and interconnectedness of the physical, chemical and biological processes, together with tight geochronological and paleomagnetic constraints on the timing, location and rate of changes will provide a clearer picture of global Neoproterozoic climate change and early animal evolution. In that context, clastic sedimentology plays an important part in evaluating and refining global climate change models proposed for this time period.

6. Conclusions

Analysis of soft sediment deformation structures can yield information about paleostresses that affected the sediments during or shortly after deposition. This approach has been used extensively in recent glacial deposits and in non-glaciated basins affected by seismic activity, providing important constraints on the nature of ice dynamics and the recurrence interval of tectonically-active periods. In glacial settings, shear and compressional stresses imparted by overriding, grounding or bulldozing of ice, as well as gravitational stresses related to melting of buried ice and high sedimentation rates and oversteepening affect sediments in subglacial, ice marginal, glaciofluvial, glaciolacustrine and glaciomarine settings. At the edge of glaciated basins, periglacial deformation structures develop as a result of changes in substrate thermal characteristics. In non-glacial settings, soft sediment deformation is commonly attributed to

seismic shaking, current induced shear stress, wave loading, rapid sedimentation or incomplete mixing of remobilized heterogeneous sediments. These different triggers can result in a wide range of non-unique deformation structures including various types of folding, faulting, shear or water escape structures, and clastic dikes, though careful analysis of the style, scale and complexity of deformation, the stratigraphic and facies context, as well as basin tectonic and glacial history can allow resolution of the most plausible trigger.

Analysis of deformation structures in the Neoproterozoic Port Askaig Formation revealed they most likely recorded periods of tectonic activity within the basin based on their stratigraphic distribution, the basin's tectonic setting, the facies in which they are found as well as the nature of the associated undeformed facies. The Great Breccia shows strong similarities with carbonate megabreccia and most likely record subaqueous slumping of underlying Port Askaig and Lossit (Islay) formation sediments along an active fault. The overlying Disrupted Beds and recurring horizons of soft sediment deformation up to the base of Member III are interpreted as continued tectonic activity in a glacially influenced marine setting. Alternative interpretations (subglacial, ice marginal, periglacial) are deemed unlikely due to the stratigraphic recurrence of these deformed intervals regardless of facies type, the absence of facies associations typical of these settings and the nature of the deformation structures themselves. In contrast, deformation of conglomerate in the Neoproterozoic Smalfjord Formation can clearly be attributed to gravitational, compressional and shear stresses in an ice marginal setting based on the complexity and scale of the deformation, as well as the associated undeformed facies.

This type of approach is best suited in sections where outcrop exposure allows the style and scale of the deformation to be well established as well as a comprehensive facies and stratigraphic analysis to be carried out. Post depositional tectonic overprinting or polyphase deformation from different sources and over time can make interpretations difficult. However, useful paleoenvironmental information can be extracted from the analysis of deformation structures when combined with strong stratigraphic control and facies analysis of Neoproterozoic glacigenic successions. In successions where diamictite and sea level changes can be interpreted in the context of both tectonic activity or climatic changes and ice margin fluctuations, the analysis of deformation structures can provide additional information to better constrain their paleoclimatic significance.

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1176

Figure Captions

Fig. 1: Two dimensional models showing nature and distribution of deformation in glacial settings and associated facies (modified from Leeder, 1999). A) terrestrial temperate glacial setting, B) temperate grounded ice margin in a glaciomarine setting, C) polar floating ice shelf in a glaciomarine setting.

Fig. 2: Deformation related to ice marginal or subglacial conditions. A) large scale shear structures of fine sand underlying Quaternary till, Waterloo, Canada; B) sedimentary dike underlying a Quaternary till, Waterloo, Canada; C) general view of Pleistocene glacitected sediments, Tierra del Fuego, Argentina. Cliff is less than 10 m in height. Note the inclusions of coarser grained material (arrows) and preferred orientation of some of them to the upper right consistent with the overall large scale shear structures; D) close up of Pleistocene glacitected sediments, Tierra del Fuego, Argentina. Note the shear structures and chaotic stratification; E) complex glacitectonic deformation in Quaternary diamicton, sand and gravel, Drumbeg Quarry, Scotland (after Benn and Evans, 1996; reproduced with permission from Elsevier); F) ice marginal or proglacial deformation in recent glaciofluvial sand and gravel, Tungnaárjökull, Iceland (after Andrzejewski, 2002; reproduced with permission from Elsevier). Photos (C) and (D) courtesy of I. Peter Martini.

Fig. 3: Field sketch showing scale and style of deformation attributed to overriding by ice (shear structures throughout) and gravitational collapse (graben-like structure in panel 7) associated with melting of buried ice in an ice marginal setting (after Weaver and Arnaud, 2011, reproduced with permission from Elsevier).

1200 SUI–SUV represent stratigraphic units at the site.

1201

1202 Fig. 4: Normal faulting (A) and folding (B) in Quaternary glaciofluvial sediments, Ontario,
1203 Canada. Photo (B) courtesy of I. Peter Martini.

1204

1205 Fig. 5: Deformation by sea ice in glaciomarine deposits. A) boulder pushed along the recent
1206 coast of James Bay, Ontario, Canada. Note furrow, levee and plough morphology. Photo
1207 courtesy of I. Peter Martini; B) large-scale furrow (circled) and deformed bedding (arrow)
1208 created by iceberg grounding, glaciomarine setting; Late Miocene Yakataga Formation, Alaska;
1209 Cliff is approximately 450 m in height. Similar deformation is found in glaciolacustrine settings.

1210

1211 Fig. 6: Deformation in glaciomarine or glaciolacustrine sediments. A) complex deformation in
1212 diamictite including tectonic laminae (arrow) and fault (left of hammer); B) attenuated bedding
1213 within diamictite; C) reverse faulting (arrows) in interbedded sand and mud underlying a
1214 diamictite; D) normal faulting (arrows) in interbedded sand and mud underlying a diamictite; E)
1215 clastic dikes penetrating downward at the contact between two diamictite units; dikes consist of
1216 coarser-grained sandstone that is more lithified than surrounding matrix of diamictite; F) chaotic
1217 bedding in diamictite; G) glaciomarine diamictite interpreted as a debris flow based on clast
1218 protruding at the upper contact and draped overlying fine grained sediments, early Proterozoic
1219 Gowganda Formation, Ontario, Canada; H) loading and ball and pillow structures developed in
1220 Quaternary glaciolacustrine sediments, Bowmanville Bluffs, Ontario, Canada; Structures in
1221 Photo A to D are stratigraphically above and below a boulder pavement, which was interpreted

by Eyles 1984 as evidence of ice advance onto the continental shelf. Photos (A) to (F) - Late Miocene Yakataga Formation, Middleton Island, Alaska. Photo (H) courtesy of I. Peter Martini.

Fig. 7: Deformation related to periglacial conditions. A) convolution in fine grained organic rich marsh deposit uplifted inland, discontinuous permafrost area, recent coast of James Bay, Ontario Canada. Note peat above and metre stick with 10 cm increments; B) sand-filled wedge, Pleistocene sediments, Tierra del Fuego, Argentina. Note the upturned adjacent bedding to the left of the wedge (arrow). Photos courtesy of I. Peter Martini.

Fig. 8: Diagrams showing nature and distribution of deformation unrelated to glacial activity. A) syn-depositional deformation resulting from (i) frictional drag associated with currents (recumbent folding) and wave loading (load casts; tens of cm scale); (ii) seismic shaking (cm to 10's of cm scale); (iii) rapid sedimentation and reverse density loading (cm to 10's of cm scale); (iv) slumping and gravitational instability (cm – 10's m scale), which may include megaclasts from large scale failure of clastic (fault scarp generated) or carbonate (sea level change and/or tectonics) sediments. Note that the latter can also occur in relatively shallow depths. B) syn- to post- depositional deformation associated with an extensional tectonic regime (modified from Fossen, 2010); C) syn- to post- depositional deformation associated with a compressional tectonic regime (modified from Fossen, 2010). See Fig. 1 for key to symbols used, fn-flow nose, rf-recumbent fold.

Fig. 9: Deformation unrelated to glacial activity. A) micro-faulting attributed to seismic activity, Mesoproterozoic Altyn Formation (Pratt, 1994); B) convolute bedding attributed to seismic

activity, Plio-Pleistocene, Taranto, southern Italy; C) recumbent fold in Precambrian fluvial cross-bedded sandstone of the Fugleberget Formation, northern Norway (Røe and Hermansen, 2006). Scale shows increments in cm (left) and inches (right); D) chaotic bedding in marine sediments, Silurian Grimsby Formation, Lake Erie, Canada. Length of core in photo is approximately 15.5 cm; E) geomorphic expression of debris flow nose (arrow), Farnham Creek, British Columbia, Canada; F) cross-sectional view of Quaternary gravel debris flow nose (arrow), Waterloo Moraine, Canada; G) ball and pillow structures, Silurian Thorold Formation, Jolly Cut, Hamilton, Canada; metre stick with 10 cm increments for scale; H) loading in Quaternary marine sediments, Vancouver, Canada. Photos courtesy of Brian Pratt (A), Geraint Owen (B) and I. Peter Martini (G).

Fig. 10: Post-depositional deformation associated with tectonic forces. A) clast flattening and development of cleavage in dolomitic diamictite associated with Mesozoic-age compression, Neoproterozoic Toby Formation, Jumbo Creek, British Columbia, Canada; B) Large scale folding of Precambrian Tanafjord Group, Giemaš anticline, Tanafjorden area, northern Norway.

Fig.11: A) Composite graphic log of the Port Askaig Formation exposed on the Garvellach Islands (modified from Arnaud and Eyles, 2006) showing changing nature of sedimentary facies within the succession from a diamictite-dominated first member (0-180 m), through a transitional member of diamictite interbedded with various facies (180-290 m) and the sandstone-dominated uppermost member (290-445m). Bold number with prefix D refer to numbering of diamictite units as done by Spencer (1971); GB-Great Breccia; DB-Disrupted Beds; XB-giant cross-beds; B) symbols and lithofacies codes used in this paper.

1268

1269 Fig. 12: Generalized composite log showing stratigraphic distribution of soft sediment
1270 deformation horizons and clastic dikes within the Port Askaig Formation, Garvellach Islands,
1271 (modified from Arnaud and Eyles, 2006). Abbreviations and diamictite numbering as in Fig. 9.

1272

1273 Fig. 13: Photographs of deformation within the Great Breccia, Port Askaig Formation. A)
1274 recumbent folding of mega-clast of dolomite (cliff is approx. 50 m in height), locally referred to
1275 as the ‘Bubble’, Elieach an Naoimh; B) complex deformation of megaclast with disaggregated
1276 blocks floating in a chaotically bedded matrix, A’ Chuli.

1277

1278 Fig. 14: Partial graphic logs through the Disrupted Beds on the islands of A’Chuli (A) and
1279 Elieach an Naoimh (B). See Fig. 9B for symbols and lithofacies codes.

1280

1281 Fig. 15: Facies and deformation structures within the Disrupted Beds. A) Cliff exposure (~20 m
1282 thick) of sedimentary boudinage and folding of dolomitic sandstone beds, Garbh Eileach (after
1283 Arnaud and Eyles, 2002a, reproduced with permission from Elsevier); B) diamictite with
1284 discontinuous sandstone bed overlying mudstone, Eileach an Naoimh. Note the dark blue fine
1285 grained matrix; C) general view of cliff exposure (~ 17 m thick) of the Disrupted beds showing
1286 broad folding of bedding, A’ Chuli. Note person at lower left for scale; D) close up view of cliff
1287 exposure in (C). Note perched gulls for scale (circled); E) chaotically deformed interbeds of
1288 diamictite, conglomerate, sandstone and mudstone, A’Chuli; lowermost beds are undeformed
1289 and dipping regionally towards the lower right (after Arnaud and Eyles, 2002a, reproduced with
1290 permission from Elsevier); F) Disrupted Beds on the island of Islay, 48 km away from the

exposures on the Garvellach Islands-note the distinctive blue grey matrix. Exposed bedrock is approximately 1.5 m high; G) undeformed siltstone with clast horizons overlain by dolomitic conglomerate, Eileach an Naoimh.

Fig. 16: Photographs of deformed strata in the Port Askaig Formation. A) contorted and convoluted dolomitic sandstone (arrow); B) loaded basal contact (arrows) in interbedded sandstone; C) chaotic bedding in interbedded sandstone and siltstone above D30, west Garbh Eileach. Person for scale (circled) in lower right; D) contorted and convoluted finely laminated dolomitic siltstone; E) horizon of ball and pillow structures (arrows).

Fig. 17: Partial graphic logs from the islands of Garbh Eileach (A) and Eileach an Naoimh (B) showing deformed horizons within sandstone and mudstone facies occurring between diamictite D30 and the base of the giant cross-beds, and clastic dikes penetrating mudstone in the uppermost part of log in (A). See Fig. 9 for stratigraphic overview log, as well as symbols and lithofacies codes. Number in bracket represents distance between the two sections (modified from Arnaud and Eyles, 2006).

Fig. 18: Photographs of sedimentary clastic dikes within the Port Askaig Formation. A) relatively tabular dike intruding diamictite; B) dikes expressed as polygonal net on bedding plane surface.

Fig. 19: Field sketches showing rapid lateral facies changes and variability of deformation in deformed (Units A and C) and undeformed zones (Units B and D) of conglomerate and

sandstone, Smalfjord Formation, Handelsneset (After Arnaud, 2008; reproduced with permission from John Wiley and Sons Inc.). Shading is used to highlight colour differences attributable to variable matrix lithology. Red arrows highlight apparent direction of shear structures. Inset plan view map shows relative position of panels or outcrop faces.

Fig. 20: Photographs of deformed conglomerate, Neoproterozoic Smalfjord Formation, Handelsneset. A) General view of panel 3, Handelsneset; B) close up view of shear structures visible in (A); C) close up view of chaotic bedding and shear structures visible in (A). Photo B and C after Arnaud (2008); reproduced with permission from John Wiley and Sons Inc.

Fig. 21: Field sketches and photographs of deformation in conglomerate, Neoproterozoic Smalfjord Formation; Handelsneset. A) field sketch of outcrops (panels 7 and 8) showing large scale geometry of units, shear structures at the base and within the middle unit, and overall direction of stress (red arrows). Note that the two outcrops are almost perpendicular to one another. Colour of deformed and undeformed stratigraphic units as defined in Fig. 17; B) photograph of the outcrop on the left in (A); C) close-up photograph of the flow nose visible in the outcrop on the right in (A). Photo (C) after Arnaud (2008), reproduced with permission from John Wiley and Sons Inc.

Table 1: summary of stress types and deformation structures in glacial settings

Setting	Stress type	Deformation structure	Scale of deformation	Facies Association
subglacial	shear stress with compressional stress near ice margin	attenuated beds, boudins, rooted and unrooted shear folds, shear plane, faulting (normal, reverse, thrust), clastic dikes	cm - m	diamictite with other variable facies depending on what substrate the ice overrides
ice marginal	compressional, gravitational and shear stress	nappes, decollement surfaces, folding (chaotic, isoclinal to recumbent and shear), faulting (normal, reverse, thrust), clastic dikes	cm - 100's of m	diamictite, conglomerate, sandstone, mudstone; high lateral facies variability
glaciofluvial	gravitational collapse of buried ice	normal faulting, open folding	cm - m	conglomerate and sandstone
glaciolacustrine and glaciomarine	gravitational (slumping, sediment gravity flows), reverse density gradients and loading related to rapid sedimentation	chaotic bedding, ball and pillow structures, load casts, flow noses, convolute bedding, water escape structures, shear structures at base of sediment gravity flow deposits, clastic dikes	cm - 10's of m	diamictite, massive or laminated mudstone with or without outsized clasts
	compressional and shear stress from ice push and iceberg grounding	folding, faulting, shear structures, plough and furrow structures	cm- 10's of m	same as above
periglacial	thermal expansion and contraction	convolutions, ice-or sand wedge casts, faulting	cm - m	variable with breccia from frost shattering of clasts

Table 2: Summary of stress type and deformation structures in non-glacial settings.

Setting	Stress type/Trigger	Deformation structure	Scale of deformation	Facies Association
variable but best expressed in lacustrine or marine settings	seismic shaking	load casts, ball and pillow, pseudo-nodules, contorted and convolute bedding	cm - 10's cm	variable but often in interbedded sandstone and mudstone
shallow water (marine or fluvial)	frictional drag associated with high speed currents; wave loading	recumbent folding; convolute bedding	tens of cm	cross bedded sand; sand
unstable subaqueous settings	gravitational instability, slumping	chaotic bedding, flow noses recumbent folding, contorted to convolute bedding, shear structures at base of sediment gravity flow deposits	cm - m	diamicite, interbedded sandstone and mudstone
subaquatic fans; delta front; pro-delta	reverse density loading, rapid sedimentation	ball and pillow, load casts	cm - tens of cm	interbedded sandstone and mudstone
active tectonic setting	post-depositional compressional, shear and extensional stresses	nappes, decollement surfaces, stretched out or elongated clasts, normal and reverse faulting, folding	cm - km	variable

Figure 1

A **Figure 2**

E

Panel 5

Panel 6

Panel 7

Figure 3

Panel 8

Panel 9

W

A

B

Figure 4

A

Figure 5

B

A

Figure 7

B

A

Figure 8

B

C

Figure 9

A Figure 11

B Conglomerate facies

Gm matrix-supported, massive
Gms matrix-supported, stratified

Coarse tail inverse
grading

°°°°°°°°°°

Clast layers

(900 m)

Distance between logs

Sandstone facies

Sd deformed
Sh horizontally laminated
Sr rippled
St trough cross-bedded
Sp planar cross-bedded
Sm massive
Sg graded

Covered

Sharp, erosive

Sharp, irregular

Sharp, planar

Loaded

Fine-grained facies

Fl laminated
Flc laminated, with clasts
Fld laminated, deformed
Fd deformed
Fm massive
F lent. lenticular

f - fine
m - medium
c - coarse

Claystone
Siltstone
Sandstone
Conglomerate

Diamictite facies

Dmm matrix-supported, massive
Dcm clast-supported, massive
Dms matrix-supported, stratified
D12 diamictite #12

Diamictite,
stratified, w/
inclusions

Laminated

Deformed

Diamictite,
massive

Conglomerate

Rippled

Cross-bedded,
planar foreset

Cross-bedded,
tangential foreset

Breccia

Figure 12

A

Figure 13

B

Figure 14

Figure 16

Figure 17

A

Figure 18

B

(A) **Figure 20**