SEDIMENTOLOGICAL ANALYSIS OF NEOPROTEROZOIC GLACIGENIC SUCCESSIONS IN NORWAY AND SCOTLAND

By

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ABSTRACT

There is much debate regarding the severity of glaciation during the Neoproterozoic, particularly in response to the snowball Earth hypothesis that suggests ice repeatedly covered the Earth from equator to poles during this time. A detailed sedimentological analysis of the Neoproterozoic Smalfjord Formation of northern Norway and the Port Askaig Formation of Scotland was conducted in order to reconstruct the nature of paleoenvironmental change and assess the evidence for such severe glacial conditions. This study focuses on identifying the depositional origin of Neoproterozoic diamictites and the assessment of the relative importance of tectonics and climate in controlling sedimentation.

Analysis of the Smalfjord Formation in two different areas of northern Norway shows that diamictites and associated facies were primarily deposited by sediment gravity flow processes, making determination of past ice mass form or distribution impossible. Sedimentological and allostratigraphic analysis of the Port Askaig Formation, showed that the thick succession of diamictites interbedded with sandstones, conglomerates and mudstones was deposited in a glacially-influenced marine setting affected at different stages of its development by both tectonic and climatic forces. Detailed analysis of the Great Breccia, a distinctive diamictite unit within the Port Askaig Formation, showed that it was the product of catastrophic subaqueous mass failure indicating substantial tectonic

activity during the early stages of Port Askaig deposition. Elsewhere in the succession, diamictites show a greater degree of glacial influence on sedimentation and record fluctuating climatic conditions.

This study shows that diamictites in Neoproterozoic successions cannot be presumed to represent glacial conditions. Sedimentation at both study sites appears to have been influenced by tectonic activity related to basin development. The Smalfjord and Port Askaig formations contain only limited evidence for glacial influence on sedimentation and do not appear to record the severe glacial conditions required by the snowball Earth hypothesis.

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PREFACE

Two chapters of this thesis consist of modified versions of manuscripts, which have been accepted for publication in refereed journals. Chapter 2 will be published in Sedimentology, and Chapter 3 will be published in Sedimentary Geology. Chapter 4 will be modified to create two manuscripts, which will be submitted at a later date to refereed journals. One will likely focus on the depositional origin of the giant cross-bedded sandstones of the Port Askaig Formation, while the other will likely focus on the sedimentology and allostratigraphy of the Port Askaig Formation in the Garvellach Islands. The introduction chapter outlines the general context in which these manuscripts were written, while the conclusion chapter highlights the major findings of this study and the implications for future research.

The manuscripts that have been accepted for publication are:

Arnaud, E. and Eyles, C. H., 2002. Glacial influence on Neoproterozoic sedimentation, the Smalfjord Formation, northern Norway. *Sedimentology*, in press. (Chapter 2)

Arnaud, E. and Eyles, C. H., 2002. Catastrophic mass failure of a Neoproterozoic glacially-influenced continental margin, the Great Breccia, Port Askaig Formation, Scotland. *Sedimentary Geology*, in press. (Chapter 3)

Both manuscripts are the result of research carried out by the first author under the supervision of Dr. Carolyn H. Eyles in partial fulfillment of a PhD degree. While Carolyn Eyles provided guidance and editorial comments, all aspects of the research including literature review, field-based data collection, data analysis and writing of these manuscripts were carried out by the first author. The arguments presented in this thesis benefited from many discussions with Carolyn Eyles. Chapter 2 (Sedimentology manuscript) benefited from formal reviews by Marc Edwards, Chris Fielding and an anonymous reviewer. Chapter 3 (Sedimentary Geology manuscript) benefited from formal reviews by Tony Prave, Jon Ineson and Keith Crook.

CHAPTER ONE

NEOPROTEROZOIC ENVIRONMENTAL CHANGE

1.1. INTRODUCTION

The Neoproterozoic is a period of geologic time (1000-543 Ma) during which significant paleoenvironmental changes are recorded. Glacial deposits accumulated in Neoproterozoic sedimentary basins on all continents and are closely associated with carbonate deposits (Harland, 1964; Fairchild, 1993, Evans, 2000). Paleomagnetic analysis of some glacial deposits suggests deposition in equatorial regions and implies that Neoproterozoic glaciation may have been severe enough to be global in scope (Schmidt and Williams, 1995). Carbonates bounding glacial deposits exhibit unprecedented negative and positive excursions in $\delta^{13}C$ and some glacigenic deposits are associated with banded iron formations, suggesting unusual geochemical conditions in Neoproterozoic oceans (Knoll *et al.*, 1986; Young, 1988; Kaufmann *et al.*, 1997).

Several models have been proposed to explain the widespread distribution of Neoproterozoic glacigenic deposits and the inferred low depositional paleolatitudes for some, including global glaciation, high obliquity of the ecliptic, rapid continental drift and tectonically-controlled adiabatic glaciation (see reviews in Meert and van der Voo, 1994; Hoffman *et al.*, 1998a; Evans, 2000). The most controversial model proposed, known as the snowball Earth hypothesis, suggests that the Neoproterozoic was characterized by extreme and rapid climate fluctuations involving global glaciation events in which both continents and oceans were ice-covered, rapidly followed by post-glacial 'hot house' conditions (Hoffman *et al.*, 1998b). The snowball Earth hypothesis attempts to account for all aspects of Neoproterozoic environmental change; it is non-uniformitarian and does not apply to any of the later Phanerozoic glacial episodes.

The snowball Earth hypothesis has renewed debates about the severity, geographic extent and timing of Neoproterozoic glaciations, the origin of the pronounced $\delta^{13}C$ excursions, and the role of atmospheric CO_2, continental configuration and tectonics in controlling climate change. Reconstructing the nature of environmental change during the Neoproterozoic has important implications for characterizing the nature of past glaciations and understanding the controls on past and future climate change. In addition, environmental changes during the Neoproterozoic may have facilitated the appearance of complex life forms including Ediacaran fauna in the latest part of the Neoproterozoic (Kaufmann *et al.*, 1997; Narbonne, 1998; Hoffman and Schrag, 2000; Runnegar, 2000), such that a better understanding of the nature of Neoproterozoic environments may help us better understand the evolution of early animals.

The sedimentary record of Neoproterozoic glaciation is commonly preserved in marine basins and contains evidence of the environmental changes that occurred at this time. Detailed sedimentological analyses of two Neoproterozoic 'glacigenic' succession in Norway and Scotland, were carried out for this study in order to reconstruct the nature of Neoproterozoic climate change preserved in these successions and to test the snowball Earth hypothesis by evaluating sedimentological evidence for severe glaciation during the Neoproterozoic.

The Smalfjord formation in northern Norway and the Port Askaig Formation in Scotland, were selected for study for several reasons. Both are attributed to the Varangian glaciation, are well exposed and are easily accessible. The Port Askaig Formation is particularly well exposed and is relatively thick, thereby allowing analysis of the nature of environmental change over time. The Smalfjord and Port Askaig formations accumulated on the margins of the future Iapetus Ocean and therefore have a similar regional setting, although local tectonic characteristics differ slightly. Lastly, both of these sites have been the subject of a number of previous studies that provide important contextual information regarding regional stratigraphic correlation, isotope geochemistry of associated carbonates and overall tectonic setting of the glacial deposits.

1.2. THE GEOLOGICAL RECORD OF NEOPROTEROZOIC ENVIRONMENTAL CHANGE

Various aspects of the Neoproterozoic geological record suggest glaciations at this time were much more extreme than their Phanerozoic counterparts. The nature of the geological record is therefore examined to discuss the nature of glacial and carbonate deposits and their paleoenvironmental significance, evidence for the paleolatitude and paleogeographic setting of Neoproterozoic glacial deposits, evidence for unusual ocean geochemistry, the interregional correlation of glacial deposits and the inferred timing of glaciations.

1.2.1. Neoproterozoic diamictites and their paleoenvironmental significance

Neoproterozoic sedimentary successions commonly contain diamictite units, poorly-sorted deposits that consist of an admixture of clay- to boulder-sized particles and that have traditionally been interpreted as recording subglacial conditions. These Neoproterozoic diamictites are widespread and can be found on every continent (Hambrey and Harland, 1985; Evans, 2000; Fig. 1.1). Neoproterozoic glacigenic successions commonly consist of multiple units of diamictite interbedded with conglomerates, sandstones and/or mudstones with abundant dropstones; these successions have previously been interpreted as recording terrestrial glacial conditions with diamictites formed by subglacial lodgement processes and associated sediments deposited by fluvial, lacustrine and marine processes (Fig. 1.2; see reviews in Hambrey, 1983; Hambrey and Harland, 1985; Evans, 2000). In other cases, glacial conditions are inferred from the occurrence of relatively thin units of diamictite in otherwise dolomitic successions (eg. correlation of 'boulder beds' in the Dalradian; Harris *et al.*, 1993; Evans and Tanner, 1996) or from the presence of an erosional unconformity and/or geochemical proxy data (Kaufman *et al.*, 1997; Halverson, 2000).

Neoproterozoic diamictites have also been correlated across different basins on the basis of their presumed glacial origin and lithological characteristics. These

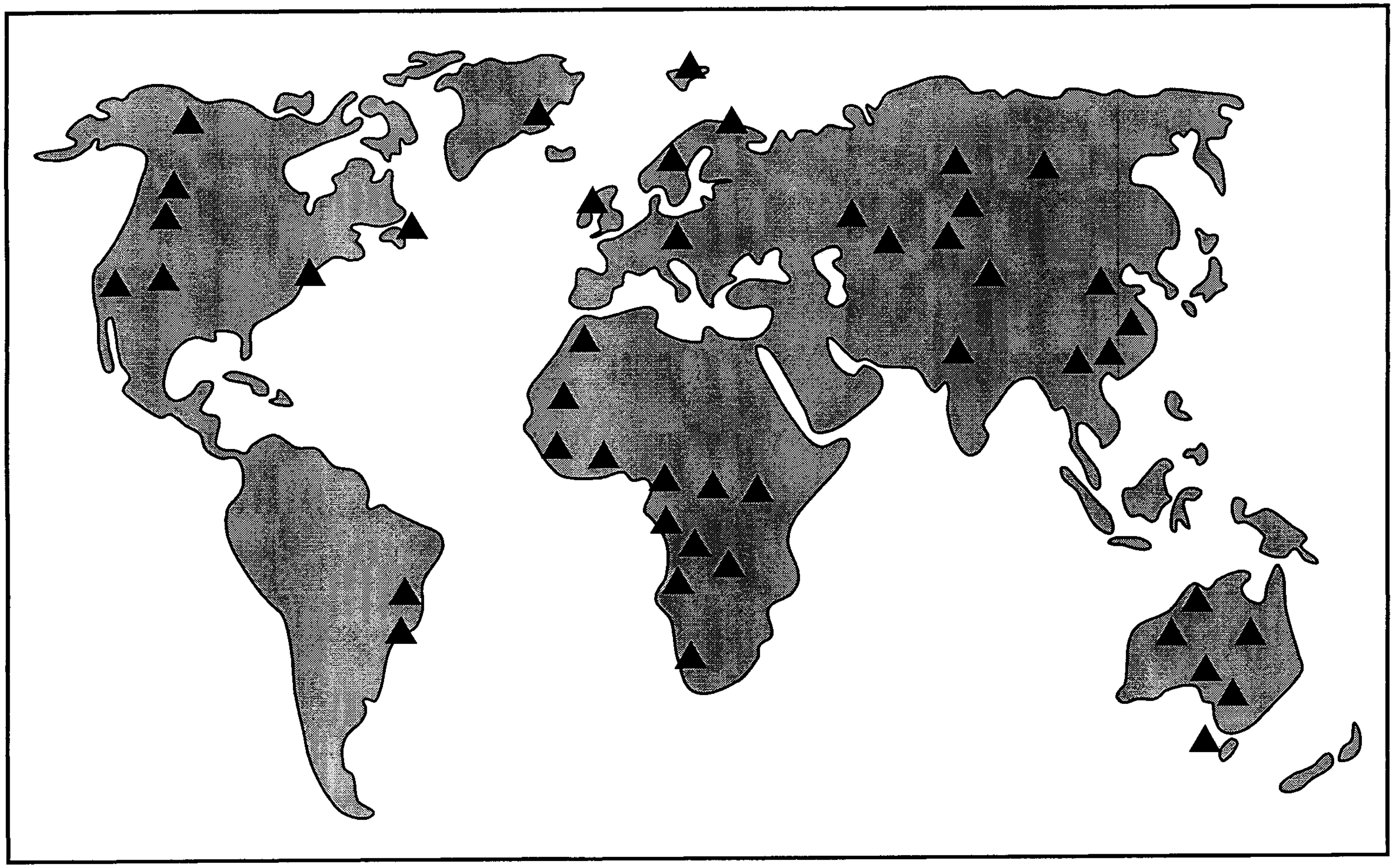

Fig. 1.1: Global distribution of Neoproterozoic diamictites (black triangles) thought to represent glacial conditions at some time between 800-550 Ma (modified from Harland, 1964).

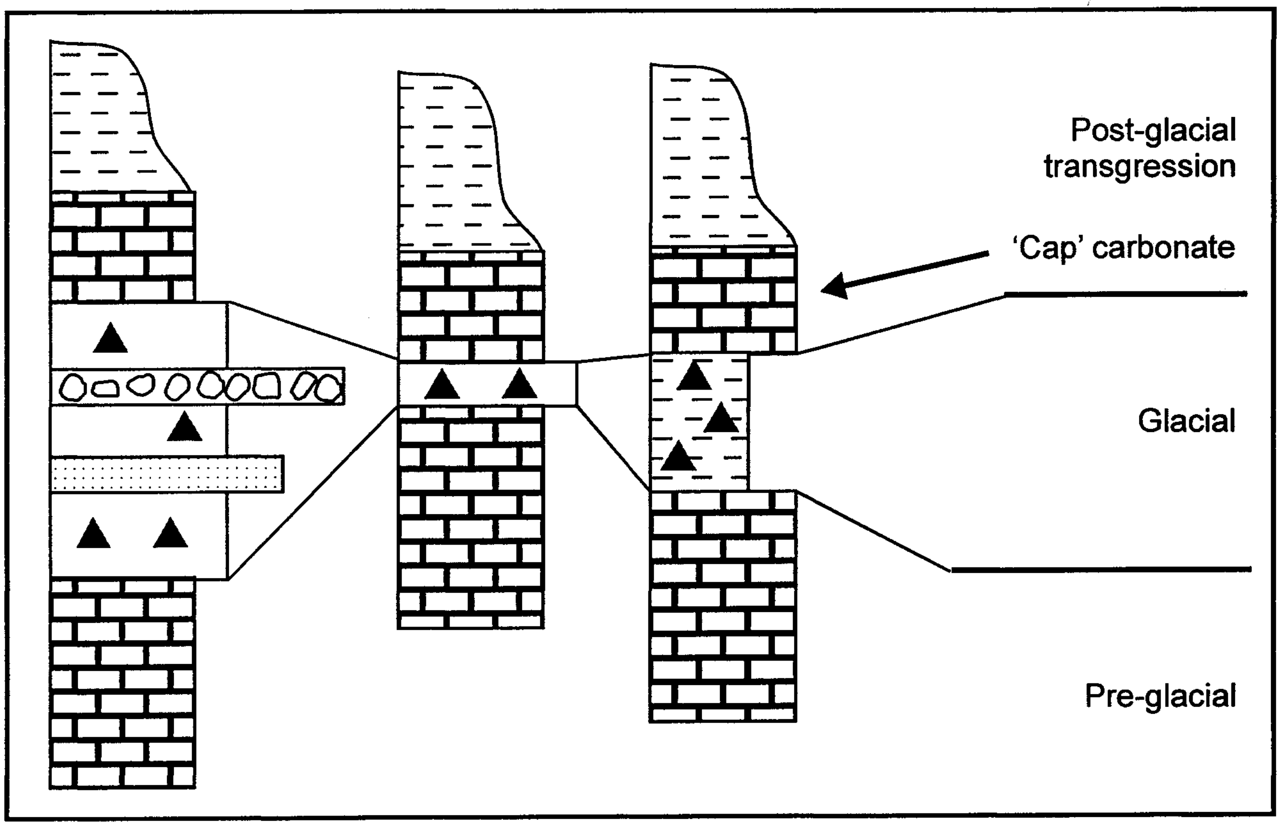

Fig. 1.2: Schematic diagram showing the various types of lithofacies found in Neoproterozoic glacigenic successions (modified from Fairchild, 1993). Glacial intervals within Neoproterozoic successions can consist of thick units of diamictites (triangles) interbedded with sandstone and conglomerates (100 m scale), relatively thin units of diamictites (1-10 m scale) or units of finely-laminated mudstones with dropstones (1-100 m scale). Glacigenic deposits are typically bounded by carbonates and facies overlying the 'cap' carbonates are thought to be post-glacial transgressive deposits.

interregional correlations have been used to infer the existence of substantial and widespread ice sheets (e.g. North Atlantic region; Spencer, 1975). However, diamictites can form in a wide variety of settings such that they may not record fully glacial conditions (Crowell, 1964; Harland, 1964; Schermerhorn, 1974; Eyles, 1993; Crowell, 1999) and their distribution alone cannot be used to determine ice sheet dimensions.

Sedimentological studies carried out since Neoproterozoic diamictites were first described have shown that diamictites can be the result of a variety of glacial processes (Boulton, 1972; Boulton and Deynoux, 1981) and also non glacial processes such as sediment gravity flows (Eyles, 1993; Crowell, 1999). Some have argued that tectonic activity associated with basin development can result in the formation of diamictites very similar in character to those deposited in glacial settings (Schermerhorn, 1974; Eyles, 1993). Recent experimental studies, field-based sedimentological analyses and offshore seismic data have documented the sedimentary characteristics and geometry of coarse-grained sediment gravity flow deposits in glacially-influenced and non-glacial marine settings (e.g. Prior *et al.*, 1984; Koster and Steel, 1984; Nemec and Steel, 1988; Pickering, 1989; Colella and Prior, 1990; Mohrig *et al.*, 1999; Sohn, 2000; Weaver *et al.*, 2000; Mulder and Alexander, 2001). These studies provide sedimentological data that may be used to discriminate subglacial from glacially-influenced or non-glacial diamictites. Studies of modern sedimentary environments and ancient sedimentary successions have also led to an increased awareness of the wide range of processes and deposits that form in glacially-influenced marine settings (e.g. Dunbar *et al.*, 1985; Eyles *et al.*, 1985; Elverhoi *et al.*, 1989; Powell and Molnia, 1989; Domack and Ishman, 1993; Hart and Roberts, 1994; Lønne, 1995; Benn, 1996; Laberg and Vorren, 2000; O'Cofaigh and Dowdeswell, 2001). Thus, Neoproterozoic diamictites that were originally interpreted as subglacial tillites may actually be the product of glacially-influenced marine processes or subaqueous sediment gravity flow processes associated with tectonic activity.

Glacial influences on Neoproterozoic sedimentation have also been inferred from the presence of dropstones in finely-laminated mudstones. However, dropstones can be released from seasonal sea ice in non-glacial settings, icebergs in very shallow marine settings close to an ice margin, or icebergs in very deep water settings far removed from an ice margin (Crowell, 1964; Gilbert, 1990; Eyles *et al.*, 1997). In short, the glacial origin of dropstone-bearing intervals within Neoproterozoic successions cannot be presumed.

Given recent appreciation for the range of possible depositional settings in which coarse-grained poorly-sorted deposits may form, reassessment of the paleoclimatic significance of Neoproterozoic diamictites seems warranted. Both the Smalfjord Formation of Norway and the Port Askaig Formation of Scotland analyzed in this study are ideal for this purpose as they are well exposed and consist of stacked diamictite units associated with sandstones, conglomerates and mudstones. Both successions have been previously interpreted as a product of subglacial and glacially-influenced marine sedimentation (Edwards, 1984; Spencer, 1971; Eyles, 1988 and references therein). The

Smalfjord Formation is up to 100 m in thickness, whereas in Scotland, the Port Askaig Formation is over 700 m in thickness.

1.2.2. Neoproterozoic carbonates and their paleoenvironmental significance

Carbonates can be found immediately below, within and/or as 'caps' above Neoproterozoic diamictite-bearing successions (Fig. 1.2; Fairchild, 1993). Carbonates are commonly assumed to have been deposited in warm tropical conditions. However, Neoproterozoic carbonates associated with diamictites record a wide variety of depositional conditions ranging from cold-water and dry, arctic conditions to warm-water and tropical conditions, and from lacustrine or supratidal to deep water marine conditions (Fairchild *et al.*, 1989; Kennedy, 1996). While bounding carbonates have been confirmed as primary in origin, those found within glacigenic successions appear to be detrital (Fairchild, 1993). Many of the primary carbonates exhibit unusual sedimentary structures such as tubes, tepees, subaqueous shrinkage cracks, roll-up structures, and large aragonite crystal fans (e.g. Kennedy, 1996; Hoffman *et al.*, 1998a; Hoffman and Schrag, 2000; James *et al.*, 2001). The origin of some of these sedimentary structures remains enigmatic and controversial suggesting, perhaps, non-actualistic processes for the formation of Neoproterozoic carbonates.

Carbonate deposits are associated with both of the diamictite-bearing successions examined in this study. The Smalfjord Formation of northern Norway unconformably overlies the stromatolitic dolomites and sabkha facies of the Grasdalen Formation (Gayer and Rice, 1989) and is overlain by very limited exposures of stromatolitic carbonate at the base of the Nyborg Formation (Edwards, 1984). Carbonates are found below (Islay limestone) and above (Bonahaven Formation) the Port Askaig Formation of Scotland (Spencer, 1971). The dolomites of the Bonahaven Formation contain abundant subaqueous shrinkage cracks and stromatolitic facies. Dolomites also occur within the Port Askaig Formation (Spencer, 1971), but these are considered to be of detrital origin (Fairchild, 1993).

1.2.3. Depositional paleolatitudes of Neoproterozoic successions

Initial paleomagnetic analyses of many Neoproterozoic glacigenic deposits suggested low depositional paleolatitudes, and this has been a key factor driving arguments for global glaciation (Harland, 1964, Kirschvink, 1992 and Hoffman *et al*,. 1998b). However, more recent appraisal of the Neoproterozoic paleomagnetic database indicates that many of the paleomagnetic data suffer from poor radiometric control and possible secondary magnetization and cannot be used to accurately determine depositional paleolatitude (see reviews in van der Voo and Meert, 1991; Meert and van der Voo, 1994; Evans, 2000). Others have suggested that the Precambrian paleomagnetic database has an apparent low latitude bias, due to assumptions made during analysis or to the erroneous assumption of an axial and dipole geomagnetic field (Morris, 1977; Lapointe *et al*., 1978; Kent and Smethurst, 1998). Meert and van der Voo (1994) showed that most Neoproterozoic glacigenic deposits most likely accumulated in latitudes >25°

and that glaciation at these latitudes could be achieved using uniformitarian climate models. The glacigenic Elatina Formation of Australia is the only deposit that has clearly been demonstrated to have formed at equatorial latitudes (Schmidt and Williams, 1995; Sohl *et al.*, 1999); most paleomagnetic data from Neoproterozoic glacigenic successions are still considered at least somewhat unreliable (Evans, 2000).

Palaeomagnetic analyses from the Nyborg Formation in northern Norway suggest depositional palaeolatitudes of 33°S for the Neoproterozoic succession that includes the Smalfjord Formation (Torsvik *et al.*, 1995), although Evans (2000) questions the primary origin of these paleomagnetic data. Early paleomagnetic analysis of the glacial deposits in Scotland suggested low depositional paleolatitudes (Tarling, 1974; Urrutia-Fucugauchi and Tarling, 1983). However, Stupavsky *et al.* (1982) showed that paleomagnetic data from the Port Askaig Formation cannot be considered reliable due to remagnetization during Ordovician times in association with Caledonian metamorphism. This conclusion has been supported by more recent paleomagnetic studies (Evans, 2000 and references therein).

1.2.4. Neoproterozoic paleogeography

Although exact paleogeographic continental configurations during the Neoproterozoic are still debated due to uncertainties in the paleomagnetic data base, most agree that a supercontinent, named Rodinia, was fragmenting during this time (Torsvik *et al.*, 1996; Dalziel, 1997, 1999; Karlstrom *et al.*, 1999). The core of Rodinia consisted of

Laurentia, which was surrounded by Siberia, Baltica, and the east and west Gondwana cratons (see review in Dalziel, 1997). The break-up of Rodinia began on the western side of Laurentia around 725 Ma, along its margin with east Gondwana, and resulted in the formation of the Pacific Ocean (Dalziel, 1992; Powell *et al.*, 1993; Torsvik *et al.*, 1995). Rifting along the Atlantic side of Laurentia occurred later around 600 Ma with the opening of the proto-Iapetus Ocean (Anderton, 1982; Torsvik *et al.*, 1995, 1996), although some have argued that rifting began earlier in southern Norway (~800-650Ma) as evidenced by dyke swarms, biostratigraphic and radiometric control (Kumpulainen and Nystuen, 1985; Torsvik *et al.*, 1995; Vidal and Moczydłowska, 1995; Drinkwater *et al.*, 1996 and references therein).

Both the Smalfjord and Port Askaig formations accumulated in marine basins of the North Atlantic region that formed as a result of fragmentation of Rodinia and creation of the proto-Iapetus Ocean (*c.* 600 Ma; Fig. 1.3; Nystuen, 1985; Eyles, 1993; Young, 1995). The Smalfjord Formation accumulated in the Gaissa Basin on the margins of Baltica, with extensional faulting related to the development of the Timanian Aulacogen and the opening of the proto-Iapetus ocean (Fig. 1.3; Siedlecka, 1975; Gayer and Rice, 1989). The Port Askaig Formation accumulated in the Dalradian Basin on the margins of Laurentia; the Dalradian Basin is considered to have developed as an ensialic basin subdivided into fault-bounded sub-basins (Harris *et al.*, 1978; Anderton, 1982, 1985).

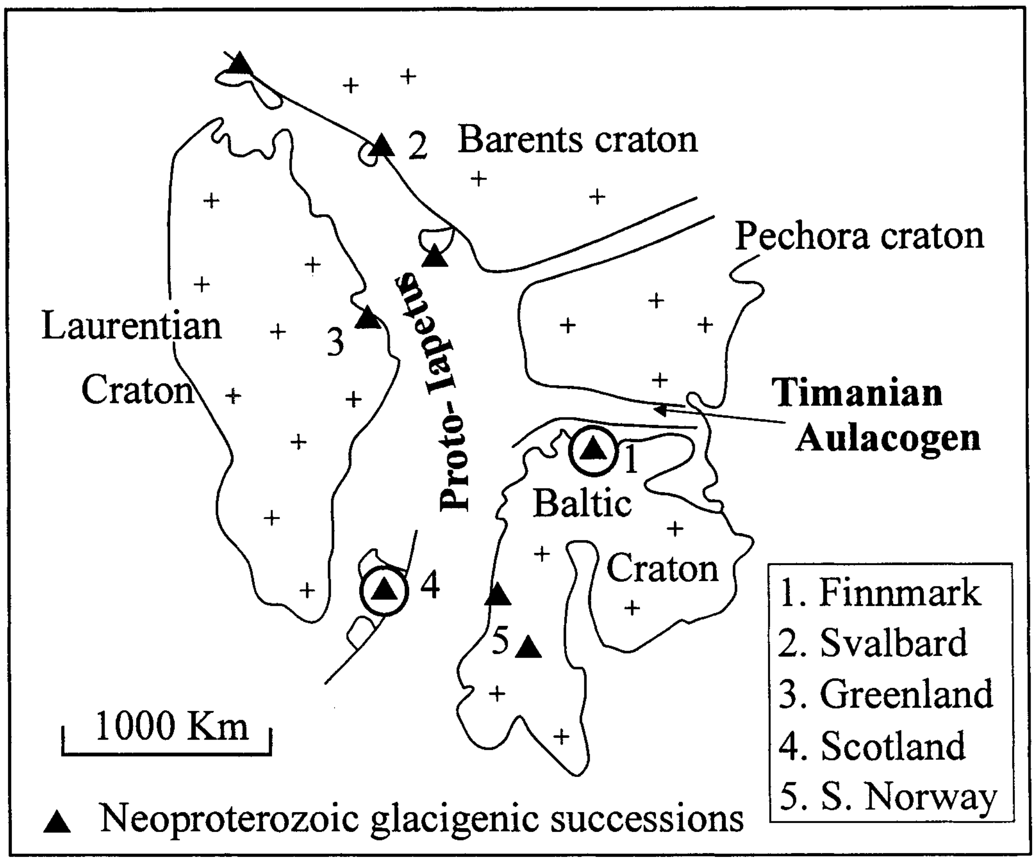

Fig. 1.3: Paleogeographic setting of Neoproterozoic glacigenic successions in the North Atlantic region (modified from Siedlecka, 1975; Nystuen, 1985 and Eyles, 1993). Successions analysed in this study are circled.

1.2.5. Neoproterozoic ocean geochemistry

Carbonates immediately overlying and underlying Neoproterozoic glacial deposits often exhibit significant $\delta^{13}C$ excursions to $\leq$- 5‰, with a subsequent shift to highly positive values in overlying transgressive carbonates (up to +7 ‰; Knoll *et al.*, 1986; Kaufman and Knoll, 1995; Kaufman *et al.*, 1997; Kennedy *et al.*, 1998). These carbon isotopic excursions are thought to represent global secular changes in seawater geochemistry; $\delta^{13}C$ values in marine carbonates are primarily controlled by fractionation of carbon during photosynthetic activity and removal of $\delta^{12}C$ from surface seawater, and by the isotopic composition of carbon coming in and out of the oceans (Knoll *et al.*, 1986; Kaufman and Knoll, 1995; Kaufman *et al.*, 1997).

However, some Neoproterozoic $\delta^{13}C$ fluctuations far exceed those seen at other times in Earth's history and suggest significant changes in seawater geochemistry. The origin of these $\delta^{13}C$ excursions is controversial; they have been attributed to overturn of stagnant glacial oceans, large-scale release of methane by destabilization of permafrost gas hydrates following glaciation or complete shutdown of marine ecosystems due to separation of ocean and atmospheric carbon reservoirs during 'snowball Earth' conditions (Knoll *et al.*, 1986; Kaufmann *et al.*, 1997; Hoffman *et al.*, 1998b; Kennedy *et al.*, 2001a).

Similar carbon isotopic excursions are also found in Neoproterozoic carbonate successions that do not contain any evidence of glacial conditions (e.g. Kaufman *et al.*, 1997). Some have suggested that the geochemical signature in these successions can be used as a proxy for glacial conditions (eg. Kaufmann *et al.*, 1997; Halverson, 2000). However, there is still much debate as to the origin of the $\delta^{13}C$ shifts and whether or not these are related to glaciation, such that this type of inference seems premature.

Banded iron formations (BIFs) are commonly associated with older (Sturtian) glacial deposits that formed predominantly in rift settings (Young, 1988; Kennedy *et al.*, 1998). Their occurrence in Neoproterozoic successions is unusual as most banded iron formations are restricted to the Archean and Early Paleoproterozoic where highly soluble iron minerals accumulated in anoxic ocean waters and were later deposited as a result of the early oxygenation of the oceans. In the Neoproterozoic, banded iron formations may have formed in extensional tectonic settings due to the combined effects of metal-rich brine production and glaciation (Young, 1988; Bühn and Sanistreet, 1997). Metal-rich brines are produced from hydrothermal vents along rift margins and thermal overturn associated with the release of cold glacial meltwater may have led to widespread precipitation of iron during the Neoproterozoic (Young, 1988).

The geochemistry of the carbonates bounding the Smalfjord Formation has not been established to date. Geochemical analyses of the Bonahaven Formation, which overlies the Port Askaig Formation, indicate $\delta^{13}C$ values that are broadly consistent with trends established from other Neoproterozoic successions, although they appear to have been affected by diagenesis in places (Brasier and Shields, 2000). $\delta^{13}C$ values at the top of the Bonahaven Formation are as high as 11.7 ‰ and are similar to those found in cap carbonates overlying glacial deposits of Sturtian age (Brasier and Shields, 2000). Neither the Smalfjord or Port Askaig formations are associated with banded iron formations.

1.2.6. Stratigraphic correlation, geochronology and timing of glaciations

There has been much debate over the correlation of diamictite-bearing units in widely-separated Neoproterozoic successions as absolute age control is generally poor. However, at least two glacial periods have been recognized (Hambrey and Harland, 1985). The older glacial period has commonly been referred to as Sturtian (*c.* 725 Ma; Brasier *et al.*, 2000), while the younger glacial period (*c.* 600 Ma; Knoll, 2000) has been referred to as the Marinoan (Australia/Canada) or the Varangian (North Atlantic; Hambrey and Harland, 1985). Each glacial period has been further subdivided into an upper and lower glacial epoch (Hambrey and Harland, 1985). Most Neoproterozoic successions contain two diamictite-bearing intervals and these have been interpreted as either representing two separate glacial periods or two glacial epochs of the same period. Successions in the North Atlantic tend to be made up of only two diamictite-bearing intervals, which have been attributed to the Varangian ice age (Hambrey, 1983).

There now appears to be growing evidence for a third, post-Marinoan glaciation. The Moelv Tillite in southern Norway was previously correlated with the upper Varangian diamictite in northern Norway. However, post-Marinoan acritarchs have been described below the Moelv Tillite and suggest another glacial interval exists between the upper Varangian glacial interval and the Precambrian-Cambrian boundary (Knoll, 2000 and references therein). In addition, the Tarim basin in China contains three

Neoproterozoic glacigenic diamictite-bearing intervals, the uppermost interval having been interpreted as post-Marinoan (Knoll, 2000 and references therein).

While Neoproterozoic glacigenic successions are generally attributed to either of the earlier (Sturtian) or later (Marinoan/Varangian) glacial period, the lack of radiometric control on the timing of deposition has made interregional correlations difficult. For example, the age of Neoproterozoic diamictite units in northern and southern Namibia and the number of glaciations represented in these successions is a matter of significant debate. Kaufman *et al.* (1997) argued for at least four glacial intervals during the Neoproterozoic; they interpret two diamictite units in northern Namibia as representative two glacial epochs of Sturtian age and they correlate two diamictite units in southern Namibia with the upper and lower glacigenic deposits of the North Atlantic Varangian (Marinoan) glaciation. In contrast, Kennedy *et al.* (1998) argued that the distinctive lithological characteristics of cap carbonates associated with the Namibian glacial deposits supported their interregional correlation and suggested that there were only two distinct glacial episodes, one Sturtian and the other Varangian.

The Smalfjord Formation is the lower of two diamictite-bearing units exposed in northern Norway and has been correlated with other Varangian glacial deposits around the North Atlantic region on the basis of lithological and biostratigraphic evidence (Hambrey, 1983). Radiometric dates (Rb-Sr whole rock isochron ages) from the overlying Nyborg Formation suggest the Smalfjord Formation was deposited prior to 654 Ma (Hambrey, 1983). However, this date is inconsistent with a suggested time interval for the Varangian of 590-610 Ma (Knoll and Walter, 1992) and is considered unreliable due to uncertainties with the radiometric dating method used (Evans, 2000).

Recently, Kennedy *et al.* (1998) suggested the Smalfjord Formation may be of Sturtian age on the basis of their analysis of the glacigenic deposits in Namibia and elsewhere, and uncertainties in the biostratigraphy of northern Norway. However, Rb-Sr analyses of diagenetic illites within associated Neoproterozoic shales suggest that the two diamictite-bearing units in Norway were deposited between 630-560 Ma (Gorokhov *et al.*, 2001), therefore confirming a Varangian age for the Smalfjord Formation.

The Port Askaig Formation is one of several diamictite-bearing units within the Dalradian Supergroup (Harris *et al.*, 1993). Structural complexities and the lack of radiometric control within the Dalradian have made correlation between individual diamictite units difficult (Harris *et al.*, 1993; Evans and Tanner, 1996; Stoker *et al.*, 1999; Condon and Prave, 2000 and references therein). The Port Askaig Formation is loosely constrained by radiometric dating to having been deposited between 595 and 806 Ma (Halliday *et al.*, 1989; Noble *et al.*, 1996) and has generally been correlated with other deposits of the upper Varangian glaciation (Hambrey, 1983). There is increasing lithostratigraphic and chemostratigraphic evidence however, to suggest that the glaciation recorded in the Port Askaig Formation likely pre-dates the Varangian glacial interval and is more likely correlative with older Sturtian deposits (Prave, 1999; Brasier and Shields, 2000; Condon and Prave, 2000).

In sum, Neoproterozoic glacigenic deposits are generally ascribed to either

Sturtian, Vendian (Marinoan/Varangian) or post-Marinoan ages on the basis of radiometric, biostratigraphic, chemostratigraphic and lithological evidence. However, radiometric control on the age of glacigenic deposits remains poor and the synchronous deposition of glacigenic deposits in widely separated Neoproterozoic basins has yet to be confirmed.

1.3. GLOBAL GLACIATION AND THE SNOWBALL EARTH HYPOTHESIS

The widespread distribution of Neoproterozoic diamictites suggests most sedimentary basins may have been affected by glacial conditions at some time during the Neoproterozoic. Furthermore, in places where diamictite and cap carbonates represent glacial and warm conditions respectively, the close association of these two lithofacies is an apparent climatic paradox and implies very rapid climate change. The inferred low depositional paleolatitude of some glaciomarine successions suggests glaciation may have been severe enough to affect equatorial landmasses. Pronounced negative and positive $\delta^{13}C$ excursions in bounding carbonates and the occurrence of BIFs also suggest significant changes in seawater chemistry. These distinctive features of the Neoproterozoic geological record are common to many successions around the world, suggesting that perhaps global-scale paleoenvironmental changes prevailed at this time.

1.3.1. Early thoughts on global glaciation

The abundance and widespread distribution of diamictites (interpreted as tillites), together with evidence that some diamictites may have been deposited at low paleolatitudes, led Harland (1964) to suggest that the Late Precambrian was characterized by global-scale glaciation with ice on continents in both polar and equatorial regions. However, the idea of global-scale glaciation was not readily accepted because the stratigraphic correlation of glacial deposits was not well constrained radiometrically and paleomagnetic data were found to be unreliable.

Based on a confirmed low paleolatitude for the Australian glacigenic Elatina Formation and improved paleogeographic reconstructions for the Neoproterozoic, Kirschvink (1992) revived and furthered Harland's idea by providing a mechanism to bring about such global glaciation. He suggested that the clustering of continents around the equator during the Neoproterozoic and the different reflective properties of land and oceans could result in a runaway 'albedo' glaciation where ice would extend over both land and oceans, and produce a reflective "snowball earth". The end of the global glaciation would depend on global warming brought about by gradual buildup of atmospheric CO_2 through volcanic emissions.

Kirschvink's (1992) model suggested that ice would extend over the oceans, although deeper waters remained ice free; ice on land would be thin and patchy due to a reduced hydrological cycle. Iron accumulated in the anoxic oceans and was later precipitated during deglaciation as ocean circulation resumed, thus explaining the common association of glacial deposits and banded iron formations (BIFs).

1.3.2. The snowball Earth hypothesis

Hoffman *et al.* (1998b) expanded on Kirschvink's (1992) idea and proposed that global-scale glaciation could single-handedly explain all the distinctive features of the Neoproterozoic geological record including the close association and widespread distribution of diamictites and carbonates, the low depositional paleolatitudes inferred for some glacigenic deposits, the geochemical signatures of carbonates bounding glacial deposits, and the occurrence of BIFs. Kirschvink's (1992) original term 'snowball Earth' was used to describe these global glaciation events (Hoffman *et al.*, 1998b). Hoffman *et al.*, 1998b) suggested that the Neoproterozoic was characterized by up to 4 snowball Earth events.

1.3.2.1. The four stages of a snowball Earth

The snowball Earth hypothesis according to Hoffman *et al.* (1998b) and Hoffman and Schrag (2000) consists of four stages (Fig. 1.4). At the beginning of a snowball Earth event (stage 1), glacial ice covers the poles and begins to accumulate in low latitudes at high altitudes. Carbonates exhibit a shift towards negative $\delta^{13}C$ values as a result of a decrease in organic productivity and a decrease in the burial rates of organic carbon. The global mean temperature during this stage is above 0 °C and conditions are much like those experienced in other Phanerozoic glaciations.

During stage 2, an increase in ice-cover occurs due to cooling caused by the high reflectivity (albedo) of land masses clustered at the equator. Increased ice cover would,

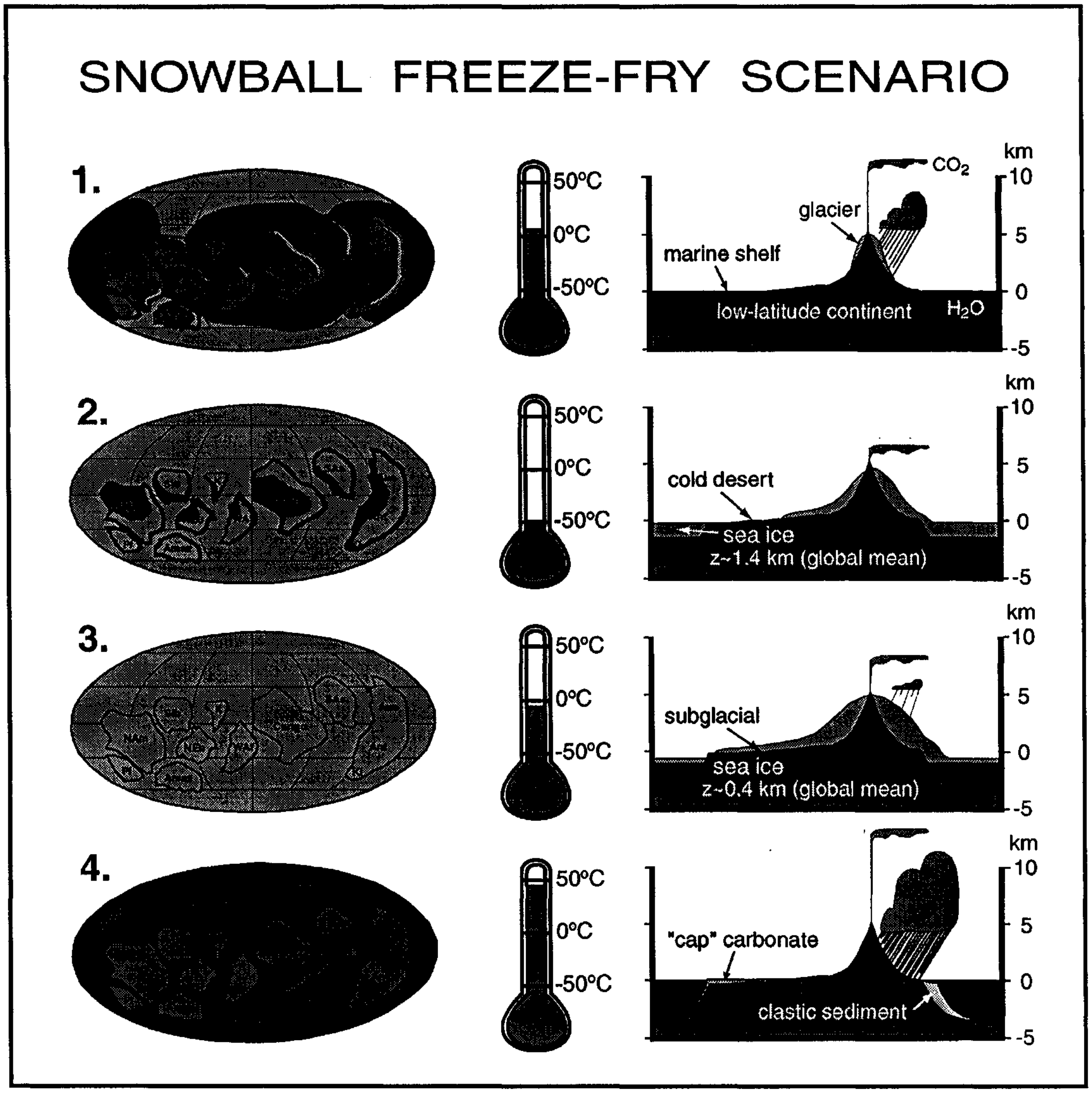

Fig. 1.4: Schematic diagram showing the four stages of a ‘snowball Earth’ global glaciation event (modified from a diagram courtesy of P. F. Hoffman). See text for explanation.

in turn, increase the earth's albedo and further promote global cooling. This feedback process known as the 'runaway albedo effect', promotes the onset of snowball Earth conditions and plunges the earth into a deep freeze. Mean temperatures would drop to -50°C, allowing the development of thick sea ice (average sea ice thickness 1.4 km). Decoupling of the atmosphere and oceans and the cessation of organic productivity in the oceans would allow the $\delta^{13}C$ geochemistry of seawater to become close to mantle values (-7‰). Hoffman *et al.* (1998b) propose that these severe glaciation events last between 4 and 30 million years and occurred several times during the Neoproterozoic.

The earth breaks out of these severely glacial conditions during stage 3, when the build-up of CO_2 in the decoupled atmosphere causes warming via the 'greenhouse effect'. Deglaciation ensues very rapidly (100 year time scale) due to reverse feedback processes. Intense evaporation and high CO_2 levels lead to the 'hothouse' conditions of Stage 4.

'Hot house' conditions of Stage 4 are characterized by mean temperatures of 50°C, intense rainfall and weathering, formation of BIFs and widespread carbonate precipitation associated with postglacial transgression of continental shelves. Carbonates precipitated in the aftermath of a snowball Earth should exhibit highly negative $\delta^{13}C$ signatures for several reasons (Hoffman *et al.*, 1998b). The sudden influx of alkalinity from intense weathering of continents is enriched in lighter isotopes as the atmospheric CO_2 reservoir approximated mantle values during snowball Earth conditions and was further enriched in lighter isotopes due to intense evaporation in the aftermath of a snowball Earth. Moreover, upwelling of anoxic bottom waters (rich in organic $\delta^{12}C$) is likely to have occurred with renewed thermohaline circulation during deglaciation, and this would have also resulted in $\delta^{13}C$-depleted surface waters.

Similar to Kirschvink's (1992) model, Hoffman's (1998b) hypothesis assumes clustering of land masses in mid to low latitudes and predicts a reduced hydrological cycle and postglacial precipitation of iron to form BIFs. However, a significant difference between the two models is the extent and severity of global glaciation over the world's oceans, and the integration of biogeochemical cycles and cap carbonate geochemistry into the 1998 model. Hoffman *et al.*'s (1998b) hypothesis proposes much more severe glacial conditions, with thick sea ice covering the entire surface of the world's oceans.

1.3.2.2. Concerns/criticisms of the snowball Earth hypothesis

The integration of a variety of geological evidence from a wide variety of places is both the strength and weakness of the snowball Earth hypothesis. On the one hand, it provides a unifying concept consistent with an increased understanding of the inter-connectedness of the earth system. On the other hand, the hypothesis postulates extreme global conditions that cannot account for important regional differences preserved in the geological record. For example, the snowball Earth hypothesis attempts to explain the diamictite/carbonate association and links the geochemistry of bounding carbonates to global glaciation. However, pronounced $\delta^{13}C$ excursions are found in some Neoproterozoic carbonates that are not closely associated with glacigenic deposits

(Kaufman *et al.*, 1997; Halverson, 2000) and such excursions may be caused by conditions other than global glaciation.

The snowball Earth model also predicts that $\delta^{13}C$ geochemistry of cap carbonates approximate near-mantle values as a result of the complete decoupling of atmospheric and oceanic carbon reservoirs during snowball Earth conditions. However, recent studies have shown that some carbonates exhibit short term fluctuations to positive $\delta^{13}C$ values both during and after the supposed glacial event (Kennedy *et al.*, 1998; Kennedy *et al.*, 2001b; and Prave *et al.*, 2001). These positive values are not consistent with completely ice-covered, anoxic oceans and suggest that perhaps other models may be more appropriate to explain the nature of Neoproterozoic seawater chemistry.

While much of the debate has focused on the geochemical evidence for snowball Earth conditions, discussion has also revolved around establishing the severity of Neoproterozoic glaciation, the lack of absolute age control on the timing of glacial events and the inability of current climate models to generate an ice-covered earth and oceans.

Many studies have presented sedimentary evidence for deposition of Neoproterozoic glacigenic sediments under temperate glaciomarine conditions typical of 'normal' Phanerozoic glaciations. Several glacigenic successions consist of ice-rafted clast layers interbedded with clast-free mudstones and suggest repeated ice margin and temperature fluctuations (Condon and Prave, 2000; Condon *et al.*, 2001). Many Neoproterozoic glacigenic successions consist of very thick accumulations of glacially-influenced marine deposits, indicative of high rates of sedimentation and temperate glacial conditions (Eyles, 1993; Christie-Blick *et al.*, 1999; Williams and Schmidt, 2000; McMechan, 2000). This sedimentological evidence is inconsistent with the severe glacial conditions predicted by snowball Earth advocates.

Although there has been much progress in better constraining the Neoproterozoic time scale through the integrative use of biostratigraphic, chemostratigraphic and lithological evidence, radiometric control is still commonly lacking in Neoproterozoic successions (cf. Knoll and Walter, 1992; Knoll, 2000). Thus, it cannot be assumed that diamictites formed in widely separate areas represent synchronous global-scale glaciation as required by the snowball Earth hypothesis; the total number of glaciations during the Neoproterozoic also remains unknown (Kennedy *et al.*, 1998; Crowell, 1999; Knoll, 2000) .

Lastly, there have been several unsuccessful attempts at modelling the climatic and geophysical conditions that would allow the development of a snowball Earth (e.g. Hyde *et al.*, 2000; Chandler and Sohl, 2000). Taking into consideration the reduced solar luminosity of Neoproterozoic times, increased or reduced ocean heat transport, increased or reduced levels of CO_2 and a higher obliquity for the earth, Chandler and Sohl (2000) were able to model snow accumulation in equatorial latitudes. However, their model could still only create 68% glacial cover on the earth. Similarly, Hyde *et al.*'s (2000) modelling of a snowball Earth event resulted in open waters at equatorial latitudes. Although these findings do not necessarily exclude the possibility of a snowball Earth event, they suggest that a less severe 'slushier' snowball Earth existed or that alternative climate change scenarios may be more appropriate to explain the Neoproterozoic geological record.

1.3.3. Alternative hypotheses

The enigmatic nature of the Neoproterozoic time period has prompted much research and discussion since the possibility of global glaciation was first proposed by Harland (1964). Much of the work carried out prior to the snowball Earth hypothesis focused on establishing the number of glaciations during this time period, the glacial origin of diamictites, the origin of the geochemical signature in carbonates, the reliability of the paleomagnetic database used to reconstruct the paleogeographic distribution of continents, and the relationship between tectonics and glaciation (e.g. Dott, 1961; Crowell, 1964; Schermerhorn, 1974; Knoll *et al.*, 1986; Meert and van der Voo, 1994; Young, 1995; Kaufman *et al.*, 1997).

As a result, several ideas have been proposed as alternatives to global glaciations to explain the unusual and paradoxical features of the Neoproterozoic (see reviews in Eyles, 1993; Meert and van der Voo, 1994; Evans, 2000). Three of these hypotheses continue to receive attention: the high obliquity hypothesis, rapid continental drift and tectonic control on glaciation.

Strong opposition to the snowball Earth hypothesis comes from those who believe that many of the features of Neoproterozoic glacial deposits can be explained by a change in the obliquity of the ecliptic. The high obliquity hypothesis suggests that earth's obliquity of the ecliptic (tilt) was greater than 54° during the Neoproterozoic (Williams, 1994). This high obliquity of the ecliptic would result in reverse climatic zonation, favouring equatorial ice sheet formation and the deposition of diamictites in low depositional paleolatitudes. It would also result in high seasonality and unstable conditions capable of causing rapid climate change and thus, could explain the diamictite-carbonate association.

However, contrary to predictions made by the high obliquity hypothesis, Neoproterozoic glacial deposits have been reported from mid-high latitudes (Meert and van der Voo, 1994; Hoffman *et al.*, 1998a; Evans, 2000), although as stated above, the reliability of the Neoproterozoic paleomagnetic database is questionable (see section 1.2.3; Williams *et al.*, 1995; Crowell, 1999; Evans, 2000). There is also much discussion about whether growth patterns of Neoproterozoic stromatolites can constitute evidence for a normal obliquity for this time period (Meert and van der Voo, 1994, Williams *et al.*, 1995).

Rapid continental drift has also been proposed to explain the juxtaposition of glacial and carbonate lithofacies and the low depositional paleolatitude of some Neoproterozoic glacigenic successions (Crowell, 1983). Rapid continental drift rates are expected during periods of continental break-up due to the nature of mantle convection (Gurnis, 1988); drift rates of up to 16 cm/yr have been calculated for Baltica following the late Precambrian to Cambrian break up of Rodinia (Gurnis and Torsvik, 1994; Torsvik *et al.*, 1996). These rates of continental movement may explain the juxtaposition of glacial and carbonate lithofacies and the low depositional paleolatitude of some glacial deposits. Testing the validity of this mechanism awaits further quantification of plate motion and a better understanding of the temporal relation between diamictites and overlying carbonate units.

Neoproterozoic glacial deposits often occur in rift basin settings; Sturtian glacial deposits (*c.* 725 Ma) occur on either side of the Pacific rifted margin of Laurentia and Varangian glacial deposits (*c.* 600) occur on either side of the proto-Iapetus ocean (Young, 1995; c.f. Torsvik *et al.*, 1995). Schermerhorn (1974) argued that Precambrian diamictites were often found in tectonically-active basins and that they may actually record sediment instability associated with tectonic activity and basin development rather than global glaciation. Eyles (1993) furthered this argument by suggesting a causal relationship between tectonics and glaciation in which uplift along either collisional or rifted margins could produce sufficient adiabatic cooling and biogeochemical change to allow the development of glaciers in tectonically-active settings of the Neoproterozoic.

While evidence for a relationship between tectonic setting and glaciation exists, the exact nature of the relationship remains controversial. Some have argued that the relationship may be merely coincidental, given the occurrence of other glacigenic deposits in tectonically-stable settings and the absence of glacigenic deposits in older extensional basins (Powell, 1995). Torsvik *et al.* (1995) also disputed a tectonically-driven model for glaciation on the basis of paleomagnetic data, suggesting that the occurrence of glacigenic deposits in the North Atlantic appears to be controlled by a shift to higher latitudes by

600 Ma rather than by tectonic setting.

1.4. RECONSTRUCTING PALEOENVIRONMENTAL CHANGE IN THE NEOPROTEROZOIC

Thick glacially-influenced marine successions preserved in Neoproterozoic sedimentary basins (e.g. Dalradian Supergroup of Scotland, Mackenzie Supergroup of Canada) contain a wealth of information about past depositional conditions. These successions can help set boundary conditions on the nature of environmental change at this time and allow realistic evaluation of the merits of various paleoclimatic models proposed. Paleoenvironmental reconstruction from sedimentary evidence in glacial settings requires discrimination of a variety of glacial and non-glacial processes responsible for the formation of diamictites and assessment of the relative contribution of tectonic factors in controlling the nature of the overall stratigraphy within a glacigenic succession.

The depositional origin of diamictites and the role of tectonics in controlling the nature of sedimentation at the time of deposition can be determined by integrating sedimentary evidence acquired at different scales (Fig. 1.5). Identifying depositional processes responsible for the formation of diamictites requires analysis of their sedimentary characteristics and geometry, as well as those of associated facies (Fig. 1.5; Boulton and Deynoux, 1981). These data can then be used to infer depositional process and depositional setting.

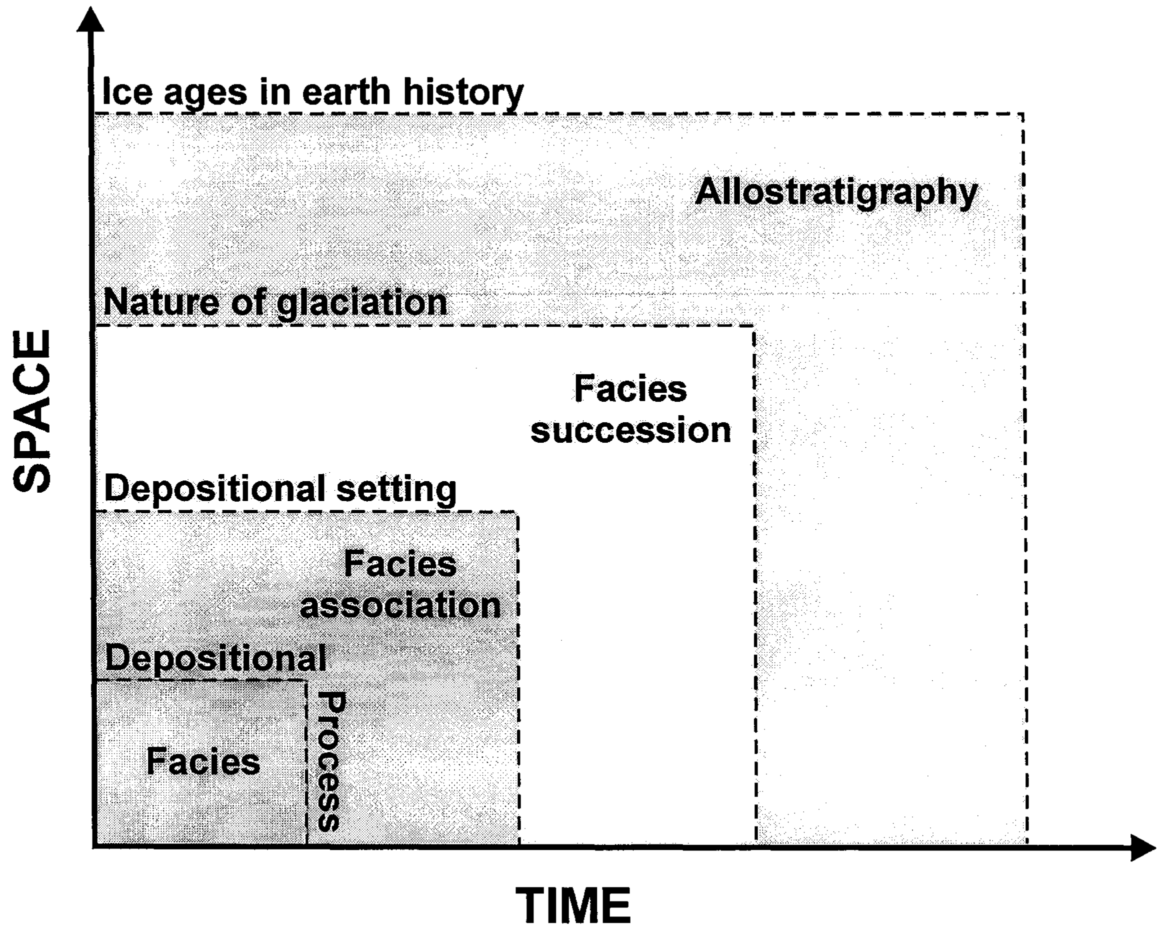

Fig. 1.5: Schematic diagram showing integrative sedimentological analysis of glacigenic sedimentary successions at different scales. The relationship between time and space may not necessarily be linear.

Analysis of the nature of facies and facies association at a larger scale, that is defining facies successions, can then be used to establish the changing nature of the depositional setting over time (Fig. 1.5). Analysis of glacigenic successions at this scale helps to characterize the environmental conditions at the onset and termination of a glaciation, as well as throughout the duration of glaciation.

At the largest scale of analysis (Fig. 1.5), allostratigraphy involves subdividing the sedimentary record into packages of genetically-related sediments bounded by significant bounding discontinuities (Walker, 1992). Allostratigraphy identifies major breaks in the nature of sedimentation and therefore aims to discriminate between larger scale controls (such as climate and tectonics) on succession development. Together, facies analysis and allostratigraphy can be applied to the study of glacigenic deposits from different periods in earth history in order to develop sophisticated depositional models that describe the nature of paleoenvironmental change over time.

Detailed sedimentological analysis carried out for this study involved describing vertical and lateral changes in the nature of sedimentary facies, clast characteristics (eg. lithology, evidence of glacial abrasion, orientation), paleocurrent indicators (e.g. cross-bedding, axis of folding structure) and the nature of bounding surfaces. This was done primarily by logging sections as well as sketching or mapping areas that were well exposed to characterize geometry of deposits, and lateral and vertical facies variability. Facies analysis of the Smalfjord Formation of northern Norway and the Port Askaig Formation of Scotland was then used to infer depositional setting and identify changes in paleoenvironmental conditions at each of these two sites. Allostratigraphic analysis was carried out on the thicker Port Askaig Formation in Scotland to more fully explore the relative importance of tectonic and climatic controls on the development of this glacigenic succession.

1.5 THESIS OVERVIEW

The overall aim of the thesis is to present an analysis of the sedimentary record of two Neoproterozoic glacial successions in order to reconstruct the nature of Neoproterozoic climate change and test the snowball Earth hypothesis. This thesis is organized as a collection of three manuscripts, which will be published in refereed journals (see preface). Chapter 1 provides background information on the nature of the Neoproterozoic geological record, the snowball Earth hypothesis, and the use of sedimentary evidence for reconstructing paleoenvironmental change.

The second chapter focuses on the sedimentology of the Smalfjord Formation in northern Norway. Analysis of the sedimentary characteristics of this glacigenic succession demonstrates that there is limited evidence for glacial conditions and that it is difficult to infer the paleoclimatic significance of diamictites as most of them appear to be the result of subaqueous sediment gravity flow processes.

The third and fourth chapter focus on the sedimentology and allostratigraphy of the Port Askaig Formation in Scotland. In Chapter 3, detailed sedimentological analysis and mapping of outcrops is presented to show that the Great Breccia, a distinctive diamictite unit within the Port Askaig Formation, records a catastrophic subaqueous mass failure associated with localized tectonic activity. This interpretation suggests a tectonic rather than climatic control on sedimentation at least in this part of the Port Askaig Formation.

Chapter 4 consists of a detailed sedimentological and allostratigraphic analysis of the Port Askaig Formation exposed on the Garvellach Islands. In this chapter, the paleoclimatic significance of diamictites is addressed, and sedimentary evidence is used to reconstruct the nature of environmental change preserved in this succession; an allostratigraphy is also developed to define major changes in paleoenvironmental conditions within this glacigenic succession and identify the relative importance of climatic and tectonic controls on succession development.

The conclusion chapter highlights the major findings of this research and comments on the similarities and differences between the two study sites. The overall implications of this research for future work on the nature of Neoproterozoic climate change are also discussed.

CHAPTER TWO

THE SMALFJORD FORMATION, NORTHERN NORWAY[1]

2.1. INTRODUCTION

The nature of Neoproterozoic glaciation is currently under debate as the interpretation of low paleolatitude depositional settings for some glacial deposits, and their close association with presumed warm water carbonates, defy current models of glaciation and climate change. Paleomagnetic, geochemical and biostratigraphic study of a number of Neoproterozoic 'glacigenic' successions has resulted in the development of several non-uniformitarian hypotheses concerning the nature, timing and palaeogeographic extent of Neoproterozoic glaciation and the controls on its initiation and development (see review in Meert & van der Voo, 1994; Hoffman *et al.*, 1998).

One model receiving considerable attention is the 'snowball Earth' hypothesis, which suggests that Neoproterozoic glaciation was severe and global (an idea first proposed by Harland, 1964a), with ice reaching equatorial latitudes, covering both land and oceans and leaving behind widespread deposits of poorly-sorted sediments or diamictite (Hoffman *et al.*, 1998). However, there has been little sedimentological testing

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1 A modified version of this chapter will appear in *Sedimentology* as Arnaud, E. and Eyles, C. H., 2002. Glacial influence on Neoproterozoic sedimentation, the Smalfjord Formation, northern Norway. *Sedimentology*, in press.

of the 'snowball Earth' hypothesis and few Neoproterozoic glacigenic successions have been examined in sufficient detail to determine the extent and form of glacial influences on deposition. Poorly-sorted, coarse-grained deposits (diamictites) found in many Neoproterozoic successions have been ascribed a glacial origin, although diamictites can form by a variety of processes in glacial, glacially-influenced and non-glacial settings (Schermerhorn, 1974; Eyles & Eyles, 1992). Diamictites can often record tectonic events related to basin evolution and localised, adiabatic glaciation of uplifted basin margins rather than globally correlatable glaciations (see reviews in Eyles, 1993; Meert & van der Voo, 1994).

This study focuses on sedimentological analysis and palaeoenvironmental interpretation of the Smalfjord Formation of northern Norway, one of two Neoproterozoic diamictite-bearing successions in east Finnmark, which have been interpreted as glacial in origin (the Smalfjord and Mortensnes Formations; Føyn, 1937; Reading & Walker, 1966; Bjørlykke, 1967; Edwards, 1975, 1984 and references therein). The Smalfjord Formation is well-exposed along the shores of Tarmfjorden and Varangerfjorden (Fig. 2.1) and consists of interbedded diamictites, mudstones, conglomerates and sandstones (Fig. 2.2). Detailed sedimentological analysis of these exposures was conducted in order to determine the depositional origin of diamictite facies, and evaluate the 'glacial' component of sedimentation.

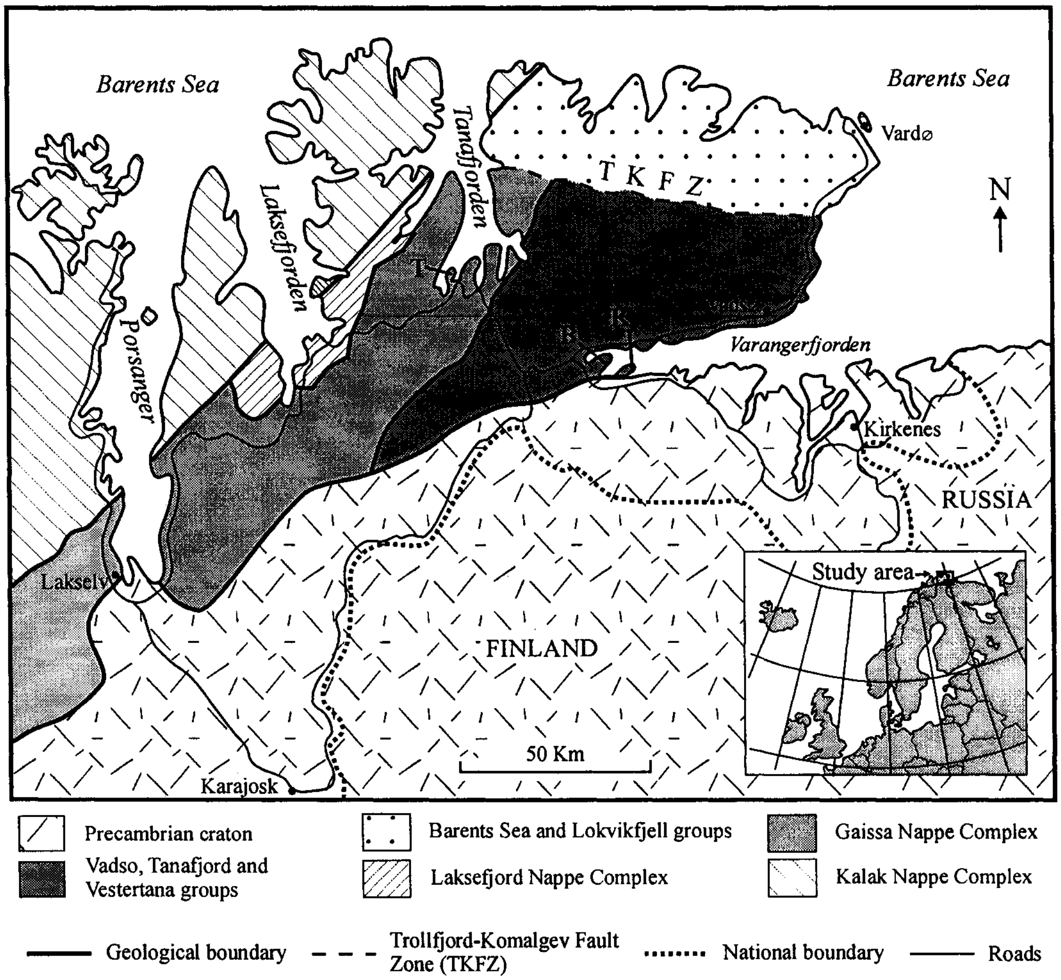


Fig. 2.1: Geology of east Finnmark and location of study sites (modified from Roberts, 1985 and Gayer & Rice, 1989). T - Tarmfjorden, B - Bigganjargga and K- Kvalnes (Varangerfjorden area).

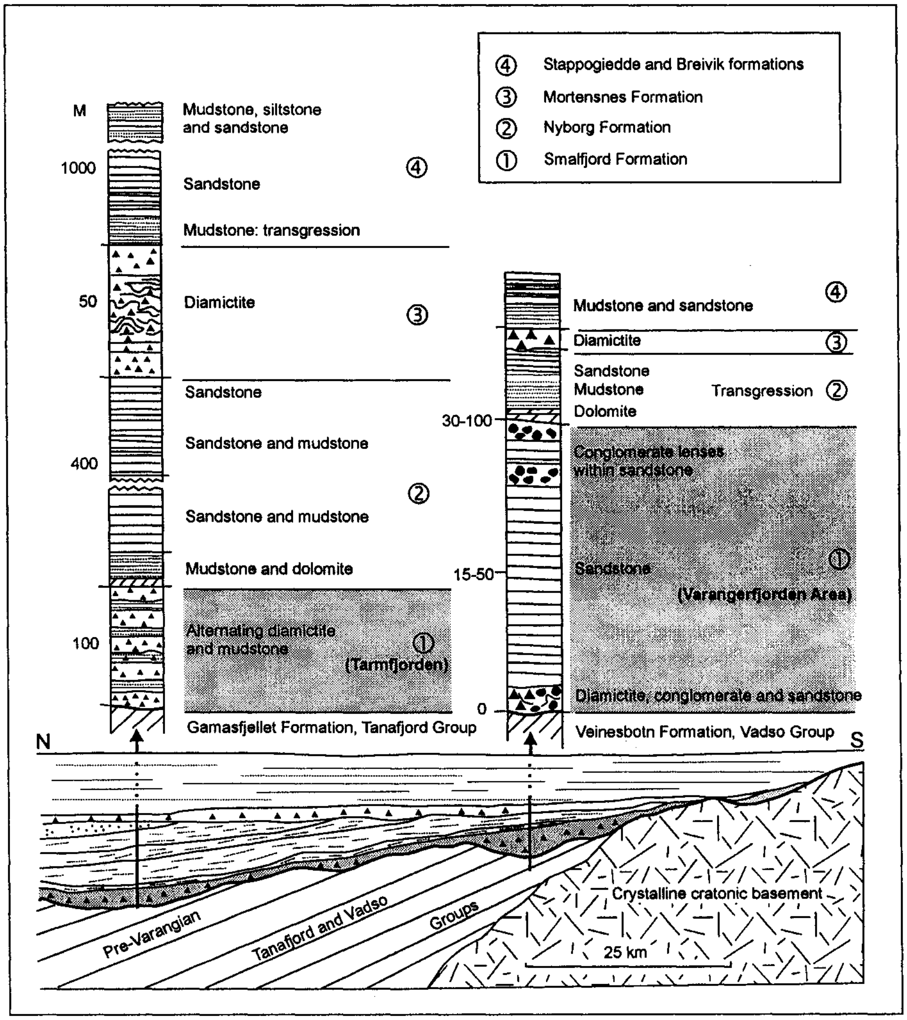


Fig. 2.2: Stratigraphic cross section through the sedimentary succession of east Finnmark, showing the geometry and stratigraphic position of the Smalfjord Formation (shaded). Note the 1-2° unconformity at the base of the Smalfjord Formation and its onlap onto crystalline basement to the south (modified from Banks *et al.*, 1971; Nystuen, 1985).

2.2. AGE AND TECTONIC SETTING

The Smalfjord Formation of the Vestertana Group unconformably overlies shallow marine and fluvial deposits of the Vadsø and Tanafjord groups, and is overlain by predominantly deep water deposits of the Nyborg Formation (Figs. 2.2, 2.3). The Smalfjord Formation onlaps the Baltoscandian craton to the south and rests on a 1-2° angular unconformity (Fig. 2.2) from which up to 2 km of section is thought to have been removed (Banks *et al.*, 1971; Siedlecka, 1975; Edwards, 1997). The formation was first described by Reusch (1891), who discovered a diamictite overlying a striated pavement at Bigganjargga (Oaibaččannjar'ga in Sami) on the shores of Varangerfjorden, and interpreted these as evidence for a Precambrian glaciation.

The Smalfjord Formation accumulated in the Gaissa Basin, which lay between the Baltoscandian craton to the south and the Timanian Aulacogen to the northwest (Siedlecka, 1975; Gayer & Rice, 1989). The extensional Gaissa Basin was related to the breakup of the supercontinent Rodinia and initial opening of the Iapetus Ocean (Gayer & Rice, 1989). The Trollfjorden-Komagelva Fault Zone (TKFZ; Fig. 2.1) north of the Gaissa Basin was the site of extensional faulting during the early Cryogenian (Siedlecka, 1975; Drinkwater *et al.*, 1996), whereas extensional stresses related to the opening of the Iapetus Ocean led to the formation of a series of NE-SW trending basement horsts on the northwestern margin of the basin during the late Cryogenian (Gayer & Rice, 1989).

The age of the Smalfjord Formation is broadly constrained to the late Cryogenian by biostratigraphic and lithological data (Fig. 2.3; Hambrey, 1983; Vidal &

Ma	Age	Lithostratigraphic unit		Thickness (m)	Environment	Tectonic events
	Cambrian	Vestertana Gp.	Breivika Fm.	600	clastic shelf	Dextral displacement along TKFZ
543			Stappogiedde Fm.	505-545	shallow marine	
c. 600?	Neoproterozoic III		Mortensnes Fm.	10-60	glacial?	Iapetus opens ~640 Ma
	Cryogenian		Nyborg Fm. 654 +/- 23	200-400	deep water	extensional faulting results in basement horsts to the NW
c. 725?			SMALFJORD FM.	2-100	glacial?	
		Tanafjord Group	Grasdalen Fm.	280	sabkha	Uplift and erosion
			Hanglecaerro Fm.	200		Extensional faulting
			Vagge Fm.	80		
			Gamasfjellet Fm.	280-300	shallow marine	
			Dakkovarre Fm.	273-350	and	
			Stangenes Fm.	205-255	intertidal	
			Gronneset Fm.	130-200		
		Vadsø Group	Ekkeroya Fm.	15-190		
			Golneselva Fm.	50-135		Fault controlled subsidence in the Timanian Aulacogen
			Paddeby Fm.	25-120	deltaic	
			Andersby Fm.	25-40	and	
			Fugleberget Fm.	125	fluvial	
			Klubbnasen Fm. 807+/- 19	50		
			Veinesbotn Fm.	300		

Fig. 2.3: Stratigraphy, depositional environment and tectonic setting of Neoproterozoic sediments in northern Norway; dashed lines are erosional unconformities (modified from Siedlecka & Roberts, 1992; Gayer & Rice, 1989). References for chronological data of Nyborg and Klubbnasen formations given in text; Neoproterozoic time scale from Knoll (2000).

Moczydłowska, 1995). Rb-Sr whole rock isochron ages have been derived for the underlying Klubbnasen Formation (807± 19 Ma; Sturt *et al.*, 1975, recalculated by Siedlecka & Roberts, 1992) and overlying Nyborg Formation (654± 23 Ma; Hambrey, 1983; Hambrey & Harland, 1985). The Smalfjord Formation has been assigned to the Varangian glacial period (Hambrey & Harland, 1985; *c.* 600 Ma), named after the striated pavement and 'Reusch's moraine' located on the shores of Varangerfjorden (Fig. 2.1). Recent chemostratigraphic and sedimentological analyses of cap carbonate deposits overlying a number of successions broadly correlative with the Smalfjord Formation, together with uncertainties regarding the biostratigraphy of the Finnmark succession, led Kennedy *et al.* (1998) to suggest that the Smalfjord Formation may be Sturtian in age (Fig. 2.3; *c.* 725 Ma). However, Rb-Sr analyses of diagenetic illites within associated Neoproterozoic shales suggest that the Smalfjord Formation was deposited between 630-560 Ma (Gorokhov *et al.*, 2001), confirming its Varangian age.

Lithological and biostratigraphic correlation of the Smalfjord Formation with other diamictite-bearing successions in the North Atlantic region, including the Elbobreen Formation of Svalbard and the Ulvesø Formation of central east Greenland (Hambrey, 1983; Harland *et al.*, 1993), has been used to infer the development of a thick and extensive Pan-European ice sheet during the Varangian (Spencer, 1975). However, poor absolute age control on these successions, and their preservation in widely separated basins, prevents confirmation of their synchronous accumulation.

2.3. TARMFJORDEN

The Smalfjord Formation is exposed over *c.* 3.5 km of the NE-SW trending shores of Tarmfjorden (Fig. 2.4). In this area, the Smalfjord Formation unconformably overlies the Gamasfjellet Formation of the Tanafjord Group (Fig. 2.3; Edwards, 1984). Five logged sections (each 50 to 100 m long) were measured in this area and show interbedded diamictite and mudstone (Figs. 2.5, 2.6).

2.3.1. Facies descriptions

Three types of diamictite constitute over 85% of the sections logged in the Tarmfjorden area; each type can be distinguished by colour, clast and matrix characteristics (Figs. 2.6, 2.7). Yellow-brown diamictites are dolomitic and have a relatively coarse-grained matrix. Purple diamictites are mudstone-rich and contain relatively few clasts. Greenish-grey diamictites contain clasts of various lithologies within a muddy, sandstone matrix (Fig. 2.7). Two types of mudstone can also be identified (Figs. 2.6, 2.8). Purple mudstones are laminated and some contain clasts, whereas grey mudstones are commonly interbedded with thin beds of laminated, rippled and massive sandstone. Diamictites dominate the lower 45 m of the Tarmfjorden succession (logs T98-1, 2, 3, 4, 5; Fig. 2.6A), whereas grey mudstones are more common in the upper 40 m (log T98-1; Fig. 2.6B).

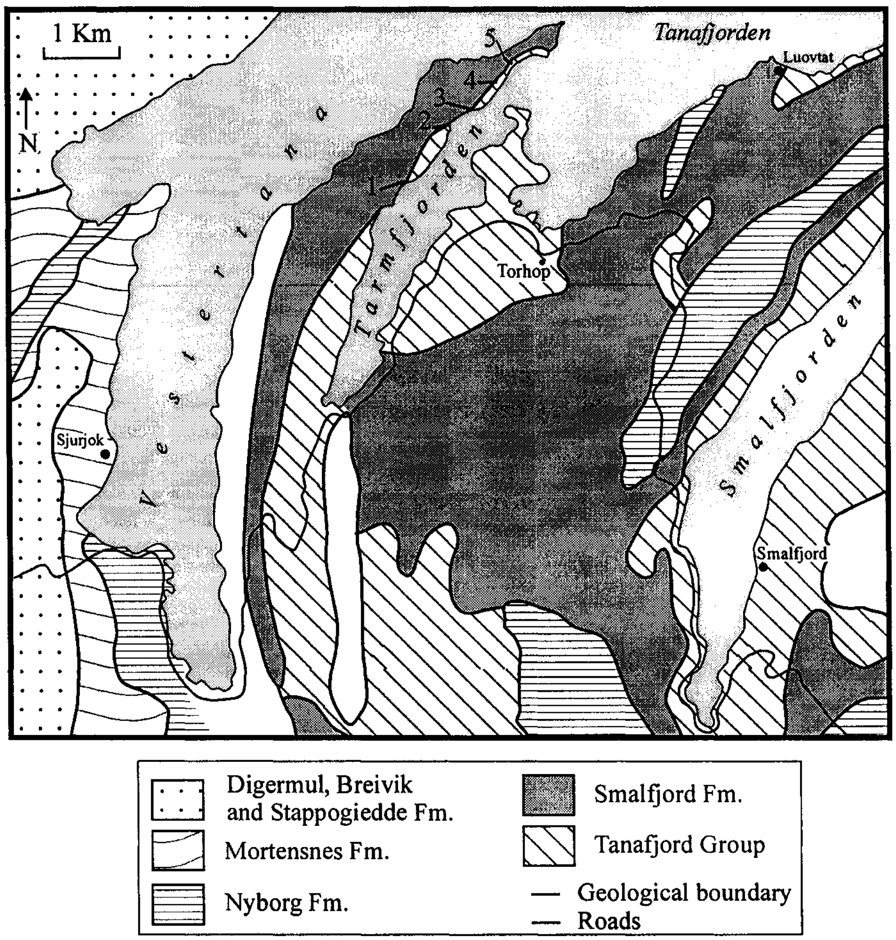


Fig. 2.4: Geological map of the Tarmfjorden area (modified from Føyn, 1937), showing location of sections measured (T98-1 to T98-5).

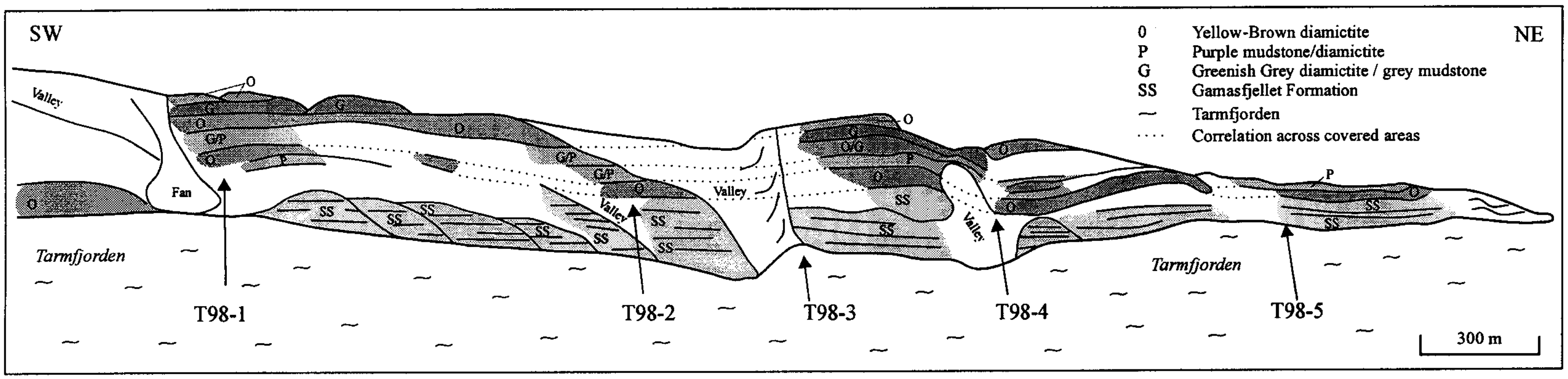
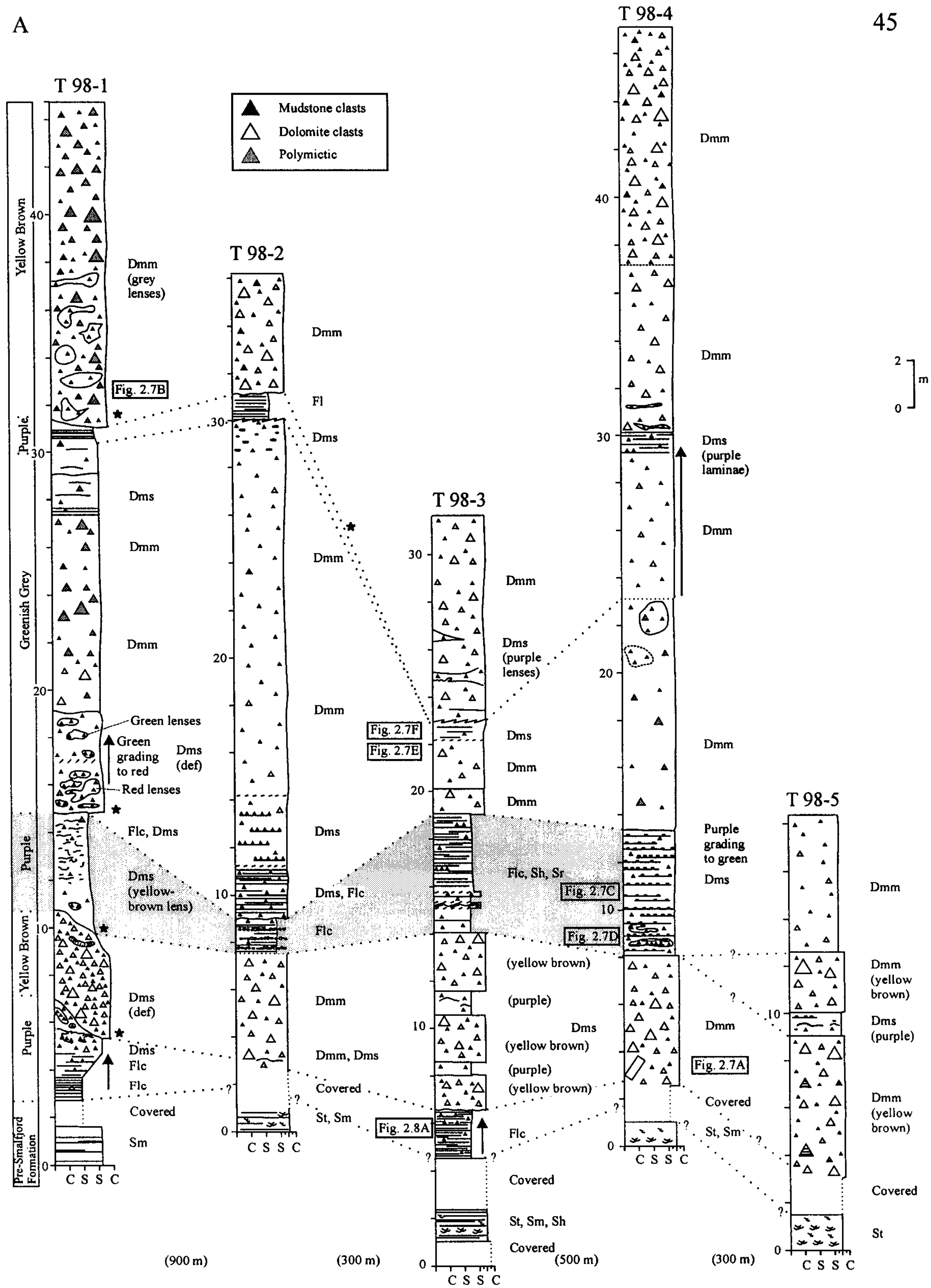


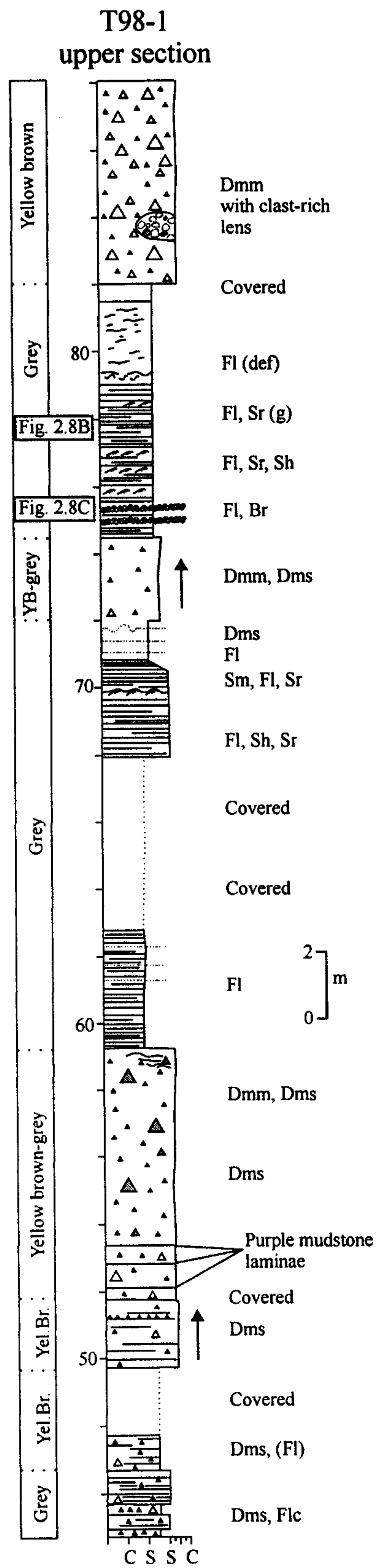
Fig. 2.5: Outcrop sketch of Tarmfjorden sections. Note how individual diamictite and mudstone units can be traced over several km of outcrop and relatively planar upper and lower contacts. White areas indicate vegetation cover.

A



T 98-1
T 98-2
T 98-3
T 98-4
T 98-5
Mudstone clasts
Dolomite clasts
Polymictic
Yellow Brown
Purple
Greenish Grey
Purple
Yellow Brown
Purple
Pre-Smalfjord Formation
Dmm (grey lenses)
Fig. 2.7B
Dms
Dmm
Dmm
Green lenses
Green grading to red
Dms (def)
Red lenses
Flc, Dms
Dms (yellow-brown lens)
Dms (def)
Dms
Flc
Flc
Covered
Sm
Dmm
Fl
Dms
Dmm
Dmm
Dms
Dms, Flc
Flc
Dmm
Dmm, Dms
Covered
St, Sm
Dmm
Dms (purple lenses)
Fig. 2.7F
Fig. 2.7E
Dms
Dmm
Dmm
Flc, Sh, Sr
Fig. 2.7C
Fig. 2.7D
(yellow brown)
(purple)
Dms
(yellow brown)
(purple)
(yellow brown)
Fig. 2.8A
Flc
Covered
St, Sm, Sh
Covered
Dmm
Dmm
Dms (purple laminae)
Dmm
Dmm
Purple grading to green
Dms
Dmm
Fig. 2.7A
Covered
St, Sm
Dmm
Dmm (yellow brown)
Dms (purple)
Dmm (yellow brown)
Covered
St
2
m
0
C S S C
(900 m)
(300 m)
(500 m)
(300 m)

B


T98-1
upper section
Yellow brown
Grey
Fig. 2.8B
Fig. 2.8C
YB-grey
Grey
Yellow brown-grey
Yel.Br.
Yel.Br.
Grey
80
70
60
50
Dmm
with clast-rich
lens
Covered
Fl (def)
Fl, Sr (g)
Fl, Sr, Sh
Fl, Br
Dmm, Dms
Dms
Fl
Sm, Fl, Sr
Fl, Sh, Sr
Covered
Covered
Fl
2
0
m
Dmm, Dms
Dms
Purple mudstone
laminae
Covered
Dms
Covered
Dms, (Fl)
Dms, Flc
C S S C

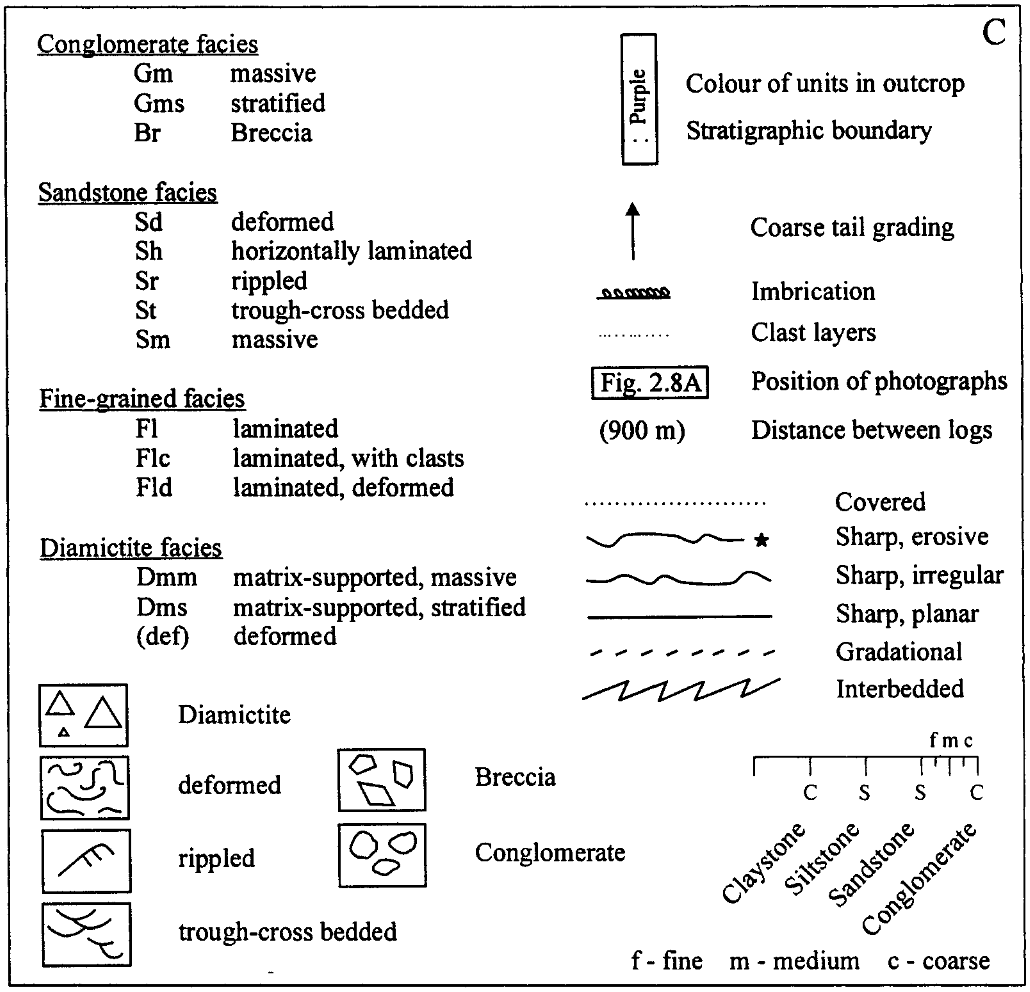


Fig. 2.6: Graphic logs of measured sections from Tarmfjorden. Vertical scale is in metres. A) Graphic logs T98-1 to T98-5. Note variable thickness of diamictite units and the lateral facies change from stratified diamictite to laminated fines with clasts over a distance of several 100 m (shaded unit); B) upper part of graphic log T98-1 in Fig. 2.6A. Note the difference in facies associations compared to the lower part of the Tarmfjorden succession, see text for discussion. The base of this log corresponds to the top of T98-1 in Fig. 2.6A; C) symbols and lithofacies codes.

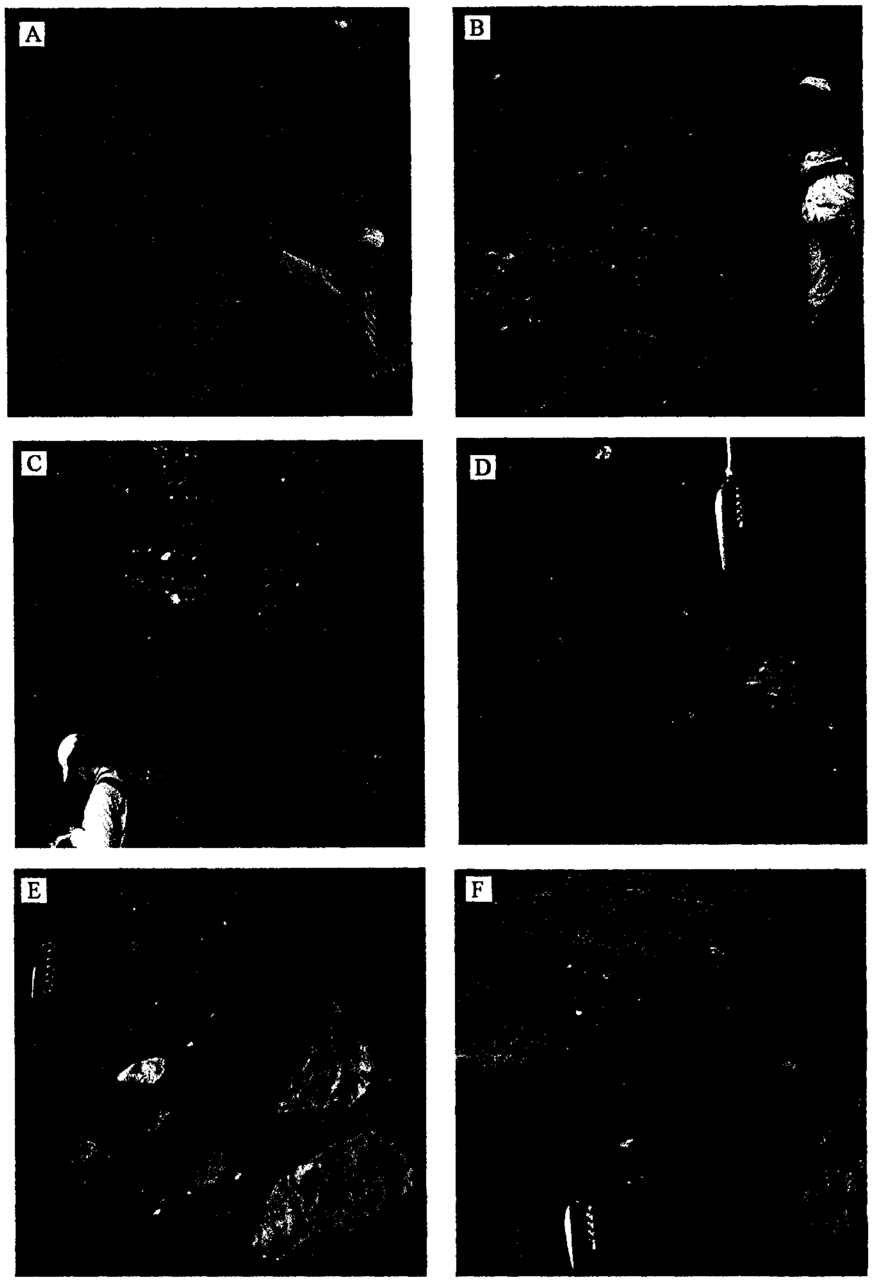


Fig. 2.7: Diamictite facies, Tarmfjorden. A) Yellow-brown diamictite, Log T98-4. Note the random orientation and angularity of dolomite clasts up to 1 metre diameter; B) Yellow-brown diamictite, Log T98-1. Note the greenish grey diamictite lenses; C) Purple diamictite, Log T98-4 showing well developed stratification and clast- and sandstone-rich layers; D) Purple diamictite, Log T98-4. Note deformed silty laminae and dolomitic clasts. E) Greenish-grey diamictite, Log T98-3. F) Greenish-grey diamictite with silty laminae, Log T98-3.

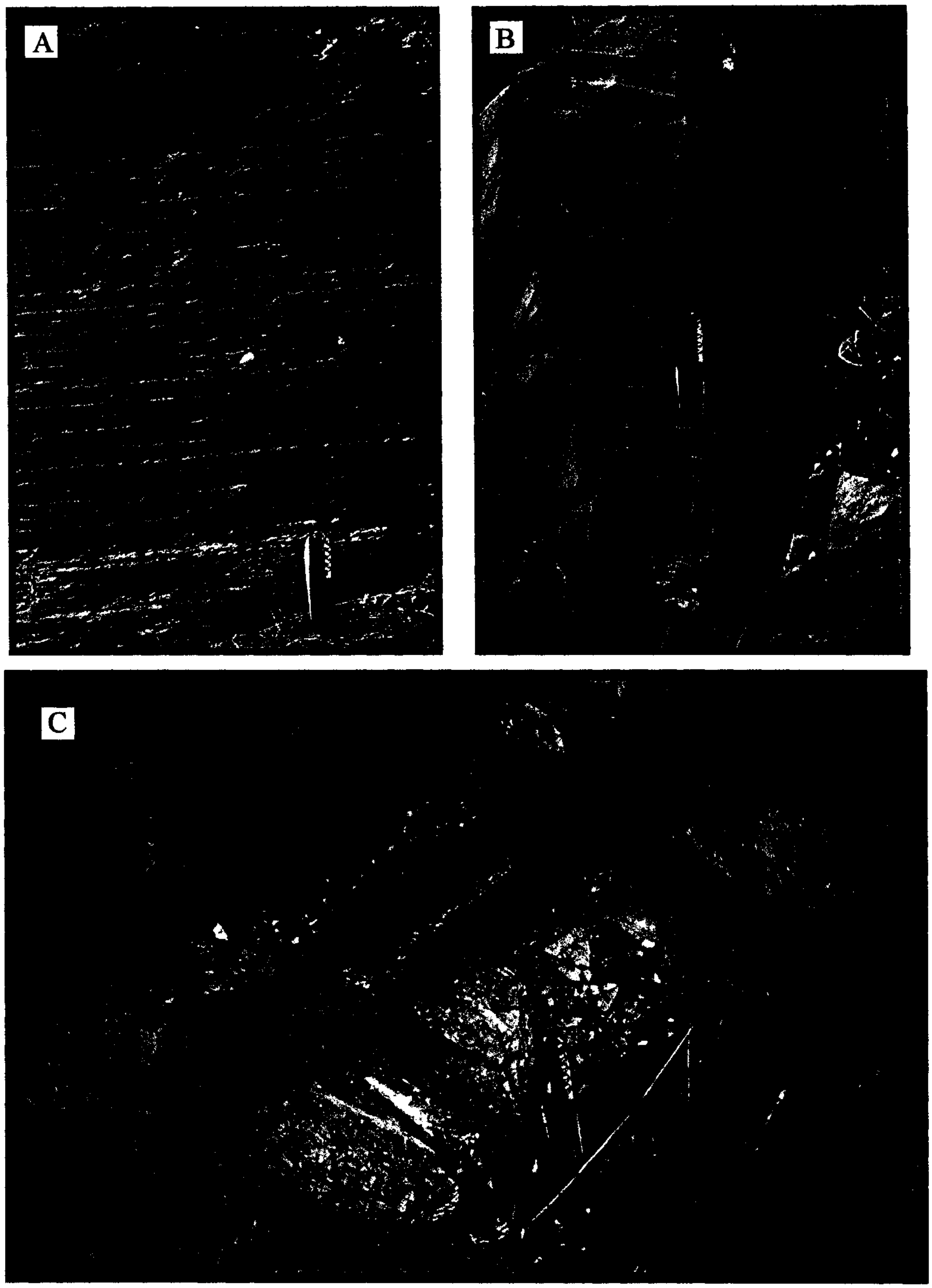


Fig. 2.8: Mudstone facies, Tarmfjorden: A) finely laminated purple mudstone with outsized dolomitic clast in centre of photo, Log T98-3. Cleavage to the upper right; B) graded beds and ripples within mudstone facies, Log T98-1; C) breccia bed containing abundant angular dolomite clasts, within grey mudstone, Log T98-1. Stratigraphic top to upper left.

2.3.1.1. Diamictites

a) Yellow-brown diamictite is predominantly massive, with a coarse- to fine-grained sandstone matrix and abundant, predominantly angular, clasts (Figs. 2.7A, 2.7B). The yellow-brown colour is attributed to the high dolomite content of the matrix (Edwards, 1984). Clasts are predominantly dolomitic but also consist of sandstone and granite. Dolomitic clasts are massive, oolitic or breccia and reach up to 1 m in diameter (most are between 1 and 35 cm). Visual estimation of clast fabric in the field indicated no apparent preferred long-axis orientation. Some beds display normal coarse-tail grading (e.g. log T98-4 between 23 and 30 m, Fig. 2.6A).

Yellow-brown diamictite beds range from 1 to 13 m thick, and are laterally continuous over several km (Fig. 2.5). Lower contacts exhibit relief of several m, and are either conformable or show evidence of erosion, such as truncation of underlying beds (Fig. 2.6).

Some units contain silty sandstone laminae and beds (1 mm-10 cm thick), whereas others contain lenses or stringers of grey or purple diamictite (eg. T98-1 at 53 and 33 m respectively; Figs. 2.6, 2.7B). Incorporated masses of diamictite are either more sandstone- or mudstone-rich than the host diamictite and contain relatively small clasts (<3 cm in diameter). Many incorporated masses have similar characteristics to underlying purple diamictites and mudstones.

b) Purple diamictite is massive or stratified, with a silty very fine sandstone to medium-grained sandstone matrix, and rare to common clasts (Figs. 2.7C, 2.7D). Clasts are predominantly dolomitic, angular to sub-rounded, up to 10 cm diameter (but commonly less than 3 cm), and in places show apparent preferred sub-horizontal long axis orientation. Clasts are either dispersed throughout the matrix or occur as discrete clast-rich layers (Figs. 2.7C, 2.7D). Purple diamictites often contain tabular or deformed units of siltstone (1-2 cm thick) or sandstone (2-10 cm thick); some diamictites contain irregular lenses of yellow-brown diamictite near their base (e.g. T98-1 at 11 m, Fig. 2.6).

Purple diamictite beds range from <1 to 5 m in thickness and are laterally continuous over several km (Fig. 2.5). Some lower bed contacts are planar and sharp, whereas others show evidence of erosion with incorporation of underlying yellow-brown diamictite (Fig. 2.6). Some purple diamictites exhibit lateral facies change from stratified diamictite to laminated mudstone with clasts, over a distance of several hundred m (Fig. 2.6A). Purple diamictites are also found as laterally discontinuous beds (tens of m) within yellow-brown diamictites (e.g. T98-3 between 24 and 26 m, Fig. 2.6A).

c) Greenish grey diamictite is massive or stratified with a medium-grained sandstone matrix, and contains rare to common clasts of dolomite and sandstone (Figs. 2.7E, 2.7F). Clasts are angular to sub-rounded and up to 40 cm in diameter. Visual estimation of clast fabric shows no apparent preferred long-axis orientation. Some greenish-grey diamictite beds contain lenses of purple diamictite with relatively abundant, small clasts (<10 cm diameter, e.g. T98-4 at 22 m, Fig. 2.6A). Subtle stratification on the scale of 0.1-10 cm is defined at the base and/or top of units by clast-rich layers, clasts with apparent sub-horizontal orientation or slight changes in matrix texture (Fig. 2.7F).

Beds of greenish-grey diamictite range from 4 to 21 m in thickness and can be traced laterally over 1800 m (Fig. 2.5). Lower bed contacts are gradational, erosional or sharp and conformable with purple mudstones (Fig. 2.6).

2.3.1.2. Mudstones

Two mudstone facies are identified, based on colour and associated facies (Figs. 2.6, 2.8). All mudstone beds are between 2 and 7 m thick and contain rare beds of dolomite breccia up to 15 cm thick (e.g. T98-2 at 8 m, Figs. 2.6, 2.8C).

Purple mudstone contains fine sandstone or siltstone units ranging from <1mm to 5 cm in thickness (Fig. 2.8A); in places, graded beds pass upwards from massive fine-grained sandstone to rippled sandstone and laminated mudstone (e.g. T98-3 at 15 m, Fig. 2.6A). Clasts are rare to common, predominantly dolomitic and occur either dispersed and increasing in abundance up-section (e.g. T98-1 and T98-3 between 4 and 6 m; Fig. 2.6A), or as a layer at the base of normally graded beds (e.g. T98-3 between 16-19 m; Fig. 2.6A). Purple mudstones are laterally gradational into purple diamictites and are most commonly associated with yellow-brown or greenish grey diamictites in vertical succession (Fig. 2.6).

Grey mudstone units (2 -7m thick) are laminated to thinly bedded (1 mm -10 cm thick) and occurs in association with rippled, horizontally-laminated, and graded sandstones (Figs. 2.6B, 2.8B). Interbedded graded sandstones are up to 10 cm thick and pass from muddy fine-grained sandstone at the base, to silt. Clasts are rare in grey mudstones and occur only as discrete single-clast layers or within interbedded breccia beds (Fig. 2.8C).

2.3.2. Facies interpretation

The depositional origin of diamictite facies is inferred by examining their sedimentary characteristics and their associated facies. Sedimentary characteristics of diamictites such as lateral continuity of units, rafts of sediments from underlying beds, sharp or erosional basal contacts, normal coarse-tail grading and subhorizontal alignments of clasts are all consistent with a subaqueous debris flow origin (Middleton and Hampton, 1976; Nardin *et al.*, 1979; Walker, 1984; Aksu and Hiscott, 1989; Laberg and Vorren, 2000).

The lateral and vertical transition from diamictites into laminated mudstones has been reported in other successions and attributed to a change from cohesive debris flows into more turbulent sediment gravity flows (Lowe, 1982; Nemec *et al.*, 1984; Postma & Roep, 1985; Eyles, 1987; Alvarenga & Trompette, 1992). This occurs by dilution and remoulding of debris during transport, by post-depositional gravity winnowing of upslope debris flow deposits, or by shearing at the upper sediment-water interface of the debris flow and generation of a low concentration turbidity current (Nardin *et al.*, 1979; Wright & Anderson, 1982; Fisher, 1983; Postma, 1984; Sohn, 2000).

A similar interpretation involving flow transformation is proposed for the lateral transition from purple diamictites to graded and laminated purple mudstones and the crude stratification and silty laminae observed at the base and/or top of thick greenish-grey or yellow-brown diamictite units (Edwards' (1984) 'banded tillites') described from Tarmfjorden outcrops. The close vertical and lateral association of diamictites and laminated mudstones at Tarmfjorden are therefore also consistent with a subaqueous debris flow origin for the diamictites.

The angularity of dolomite clasts within the diamictites indicates limited abrasion prior to or during transport and suggests a local sediment source. Dolomitic clasts probably originate from the underlying Grasdalen Formation (Fig. 2.3) which crops out to the north, but is completely eroded in the Tarmfjorden area. In the context of a mass flow origin proposed for the diamictites, colour and textural variability shown by different diamictite units (yellow-brown, purple, grey/green) is interpreted as resulting from the relative availability of coarse- and fine-grained sediment, distance travelled by the flow and the variable incorporation of fluids (Nemec *et al.*, 1984; Eyles & Eyles, 2000).

Although some of the sedimentary characteristics (e.g. erosional contact and sedimentary rafts) may also characterize diamictites formed under glacial conditions, this alternative interpretation is deemed unlikely as diamictites at Tarmfjorden show no evidence of subglacial deposition, ice-thrust deformation, blanket-like geometry resulting from glaciomarine 'rainout' processes, or rapid lateral facies changes and coarse-grained conglomerate and sandstone deposition typically associated with ice proximal conditions

(Powell, 1981; Eyles *et al.*, 1985; Powell & Molnia, 1989; Lønne, 1995). A glacially-influenced margin may be responsible for supplying large amounts of coarse-grained debris to the basin, but glacial conditions cannot be inferred based on the deposits exposed at Tarmfjorden.

Grey and purple mudstone facies are interpreted as turbidite deposits as they commonly contain graded beds with debris-rich basal layers and breccias and are closely associated with graded, horizontally laminated and rippled sandstone (i.e. partial Bouma sequences; Hill *et al.*, 1982; Walker 1984). Finely-laminated mudstones that do not exhibit these characteristics may represent deposition from suspension between mass flow depositional events. Some clasts in the purple mudstones are clearly associated with individual coarse-grained laminae and are interpreted as the coarse fraction of turbidites. The progressive upward increase in the abundance of 'lonestones' in purple mudstone units at the base of T98-1 and T98-3 (Fig. 2.6A) may indicate an increasing amount of ice-rafted debris supplied to the basin. However, none of these isolated clasts show clear evidence of rucking, bending, or penetration of underlying laminae typical of ice-rafted debris, such that the clasts may simply have been transported within turbulent flows.

2.3.3. Vertical and lateral facies distribution

In the Tarmfjorden area, the Smalfjord Formation is characterized by interbedded diamictite (yellow-brown, greenish-grey and purple) and mudstone (purple and grey, Figs. 2.5, 2.6). The lower portion of the Tarmfjorden succession is dominated by yellow-brown and greenish-grey diamictites interbedded with purple mudstones and purple diamictites (Fig. 2.6A); the upper portion is dominated by grey mudstones with occasional beds of yellow-brown diamictite (Fig. 2.6B). The ratio of diamictite to mudstone decreases from the lower to upper portion of the Tarmfjorden succession and may reflect a reduction in the amount of coarse-grained sediment available for downslope remobilization and/or increasing water depths in the basin.

2.3.4. Depositional setting of the Smalfjord Formation, Tarmfjorden area

The interbedded diamictite and mudstone facies in the Tarmfjorden outcrops indicate deposition by sediment gravity flow processes in an unstable subaqueous environment (diamictites, mudstones and sandstones) with intervals of deposition from suspension (mudstones) (Fig. 2.9). These deposits can be considered as members of a single facies association representing genetically related depositional processes and environments. The absence of deposits formed by traction current or oscillatory flows, suggests a slope setting where sediment gravity flows predominate. Sediment failure may have been triggered by rapid sedimentation and depositional oversteepening or by seismic shock resulting from tectonic activity in the basin (Eyles *et al.*, 1985; Coleman & Prior, 1988).

While it is possible that glaciers supplied heterolithic sediment to a high relief basin margin and this sediment was subsequently remobilised downslope as subaqueous

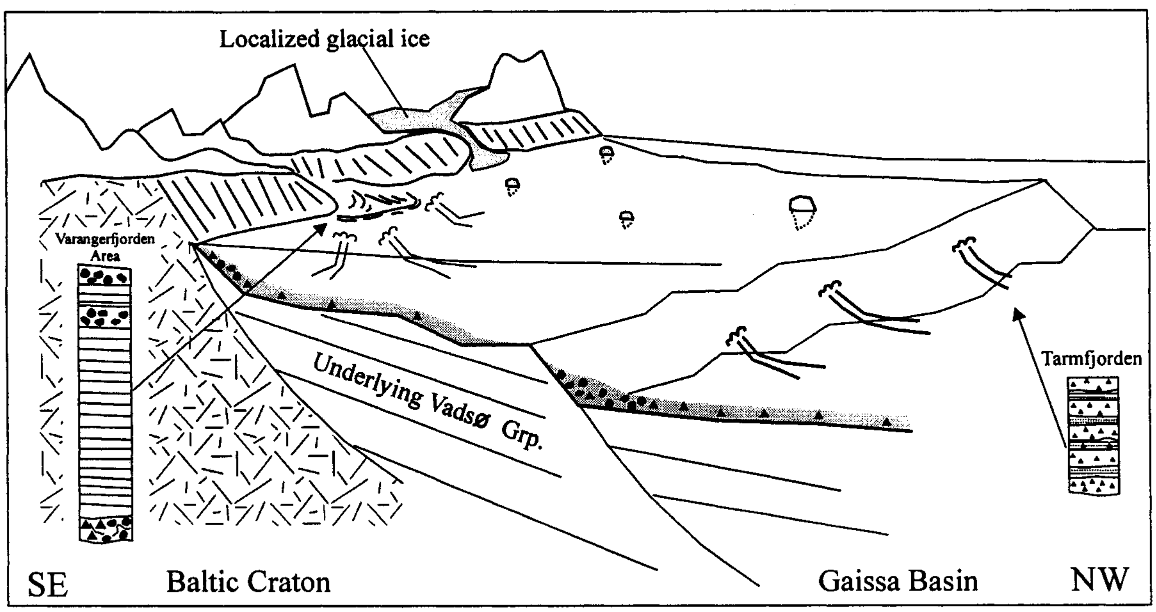


Fig. 2.9: Schematic depositional model for the Smalfjord Formation (shaded) in the Varangerfjorden and Tarmfjorden area. In the Tarmfjorden area, diamictites are interbedded with laminated mudstones; in the Varangerfjorden area, diamictites are associated with breccias, conglomerates, and sandstones.

sediment gravity flows (e.g. Howe, 1995), there is no unequivocal evidence for glacial deposition at Tarmfjorden.

2.4. VARANGERFJORDEN AREA

In the Varangerfjorden area, the Smalfjord Formation consists of sandstones, conglomerates and diamictites. These deposits were studied at Bigganjargga and along the eastern shore of the Vieranjar'ga Peninsula (Kvalnes) (Fig. 2.10). Although the section exposed at Bigganjargga is much thicker than at Kvalnes, the similarity in facies types and vertical stratigraphy (cf. K98-1B and B98-1; Figs. 2.11, 2.12) suggests that Kvalnes and Bigganjargga are correlative (Schermerhorn, 1974). The Smalfjord Formation in the Varangerfjorden area unconformably overlies the Veinesbotn Formation of the Vadsø Group (Fig. 2.3; Hobday, 1974) and at Bigganjargga, a striated pavement is exposed at this contact directly below diamictite. The unconformable contact between the two formations shows some relief (on the scale of m) at Kvalnes, whereas at Bigganjargga, it is sharp and planar (Edwards, 1975).

Diamictite units exposed in the Varangerfjorden area are thin and patchy, contain predominantly granite and sandstone clasts and are only associated with coarse-grained facies. This is in sharp contrast to the thick and laterally continuous beds of dolomite-rich diamictite interbedded with mudstones of the Tarmfjorden area.

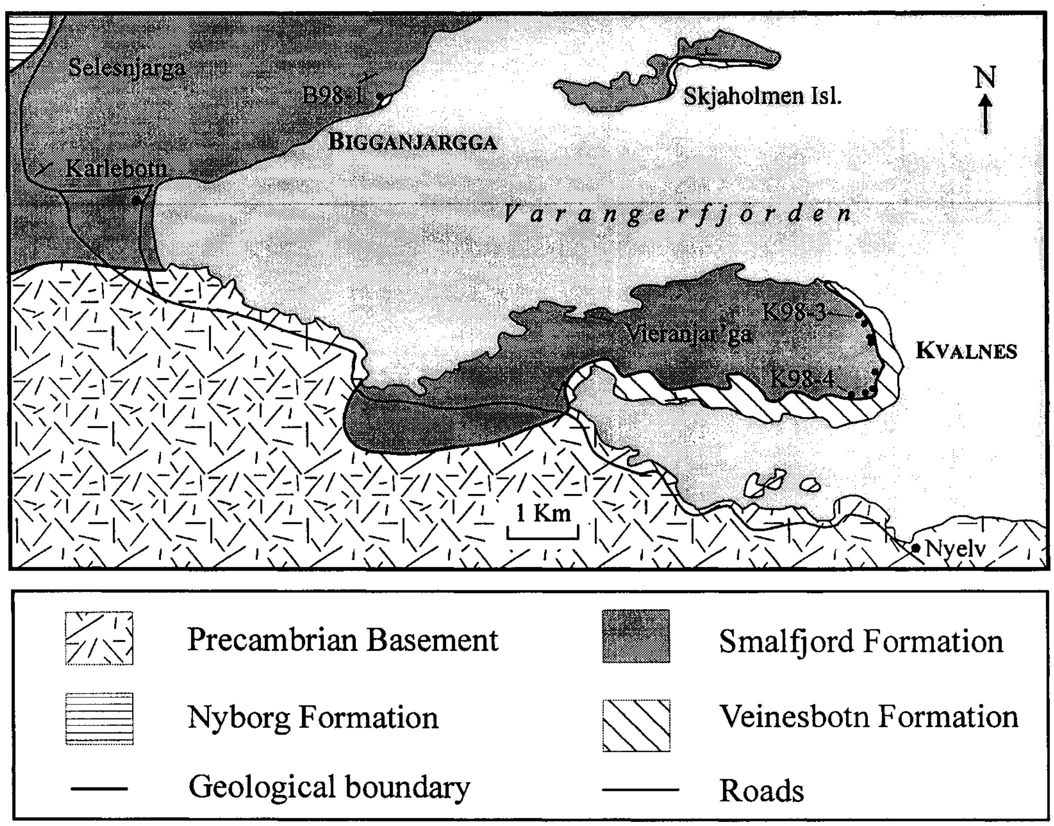


Fig 2.10: Geological map of the Varangerfjorden area (modified from Edwards, 1984), showing location of the Kvalnes and Bigganjargga sites (dots). Strike and dip measurements from Rice and Hofmann (2000).

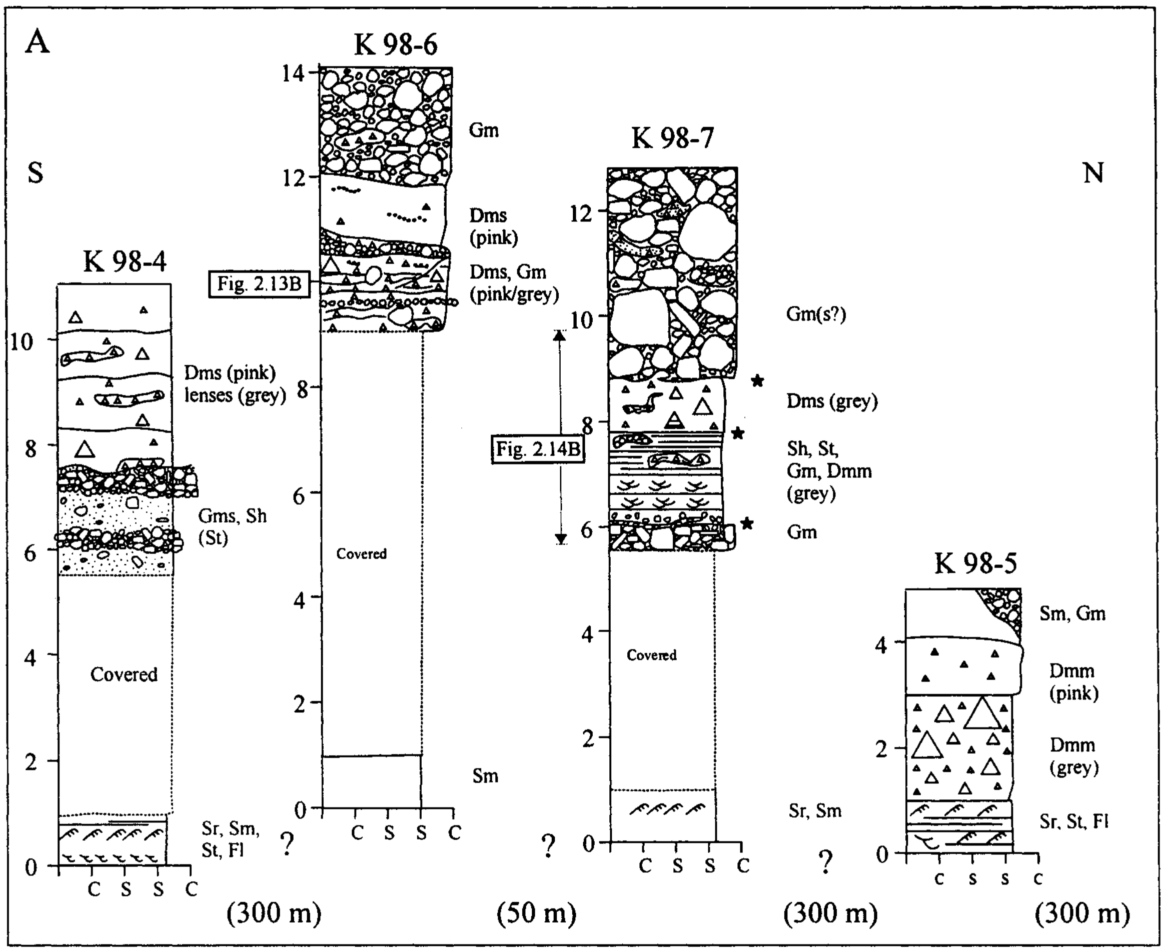
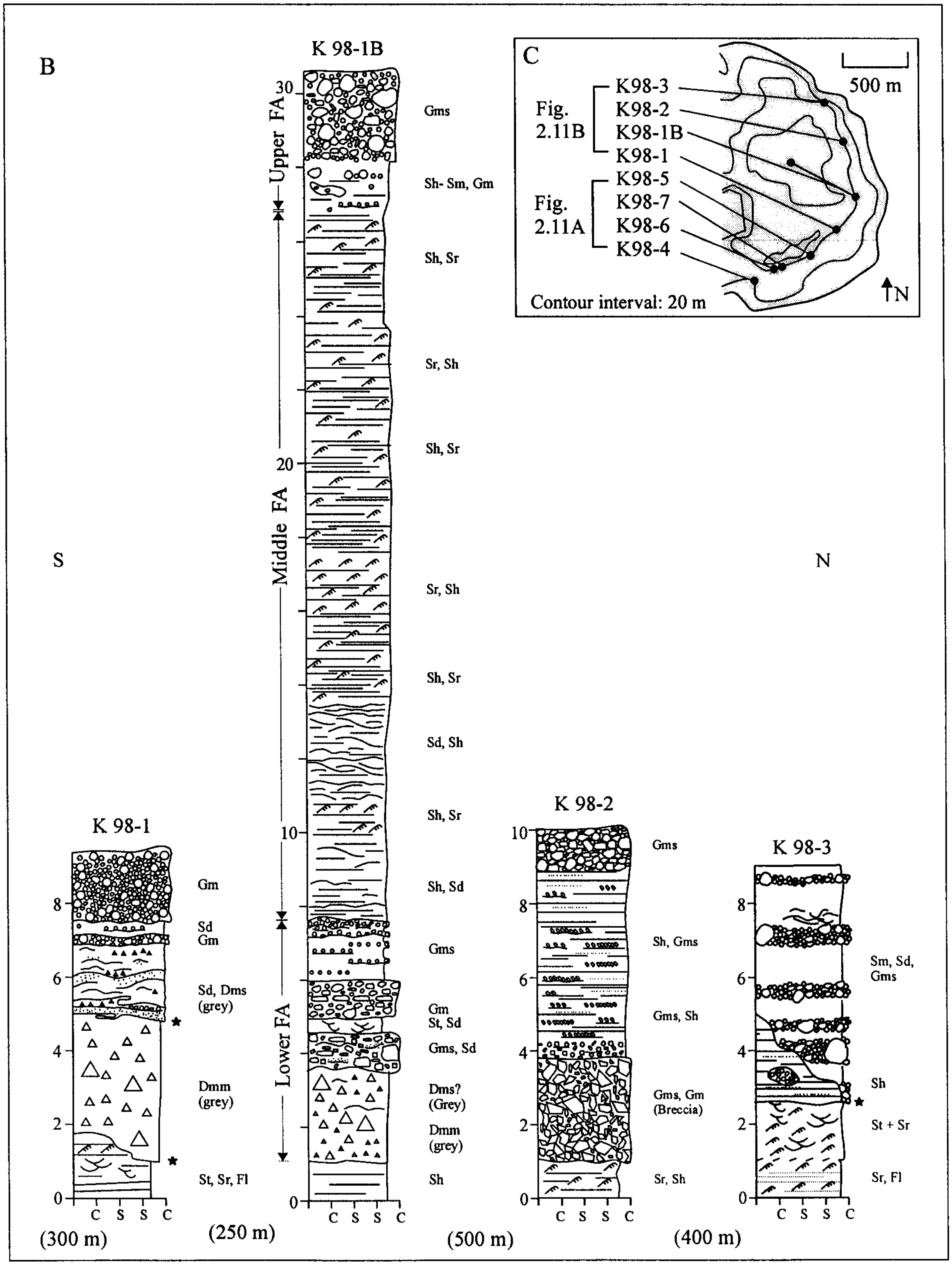


Fig. 2.11: Graphic logs of measured sections from Kvalnes. All logs start with 1-2 m of sandstone from the underlying Veinesbotn Formation. Vertical scale is in metres. A) Logs K98-4 to K98-7. Note the vertical position of logs K98-4, K98-6, and K98-7 relative to the others is unknown due to covered exposures; B) Log K98-1 to K98-3. Note breccia bed in K98-2. Log K98-1B shows all three facies associations, whereas the remainder of the logs characterize the Lower Facies Association. See Fig. 2.6C for symbols and codes; C) location of sites on the eastern shore of Vierenjarga (Kvalnes; cf. Fig. 2.10)


B
K 98-1B
Upper FA
Middle FA
Lower FA
Gms
Sh- Sm, Gm
Sh, Sr
Sr, Sh
Sh, Sr
Sr, Sh
Sh, Sr
Sd, Sh
Sh, Sr
Sh, Sd
Gms
Gm
St, Sd
Gms, Sd
Dms? (Grey)
Dmm (grey)
Sh
S
N
K 98-1
Gm
Sd
Gm
Sd, Dms (grey)
Dmm (grey)
St, Sr, Fl
K 98-2
Gms
Sh, Gms
Gms, Sh
Gms, Gm (Breccia)
Sr, Sh
K 98-3
Sm, Sd, Gms
Sh
St + Sr
Sr, Fl
C S S C
(300 m)
(250 m)
(500 m)
(400 m)
C
500 m
Fig. 2.11B
K98-3
K98-2
K98-1B
K98-1
Fig. 2.11A
K98-5
K98-7
K98-6
K98-4
Contour interval: 20 m
N

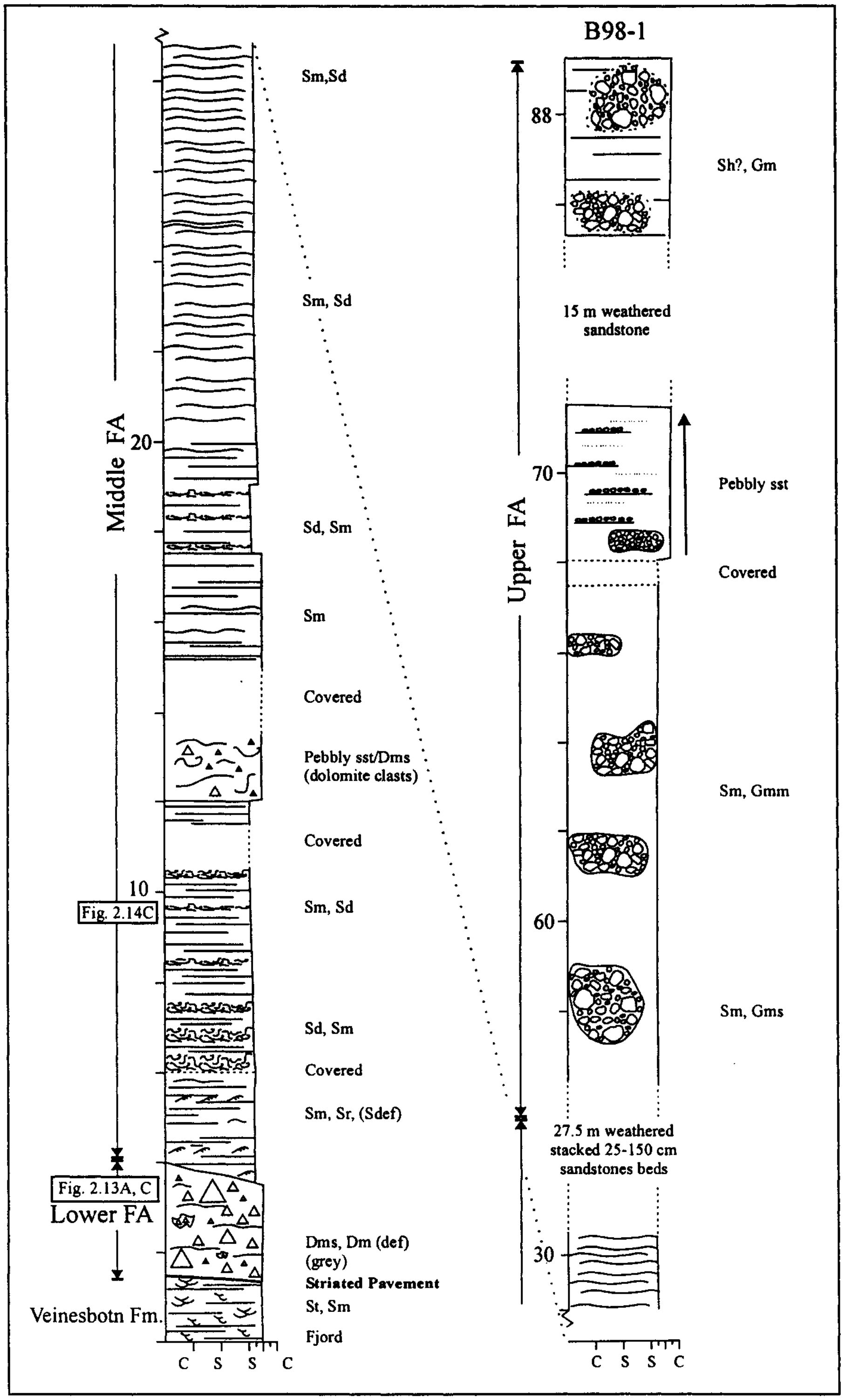


Fig. 2.12: Graphic log of measured section (B98-1) showing Lower, Middle and Upper Facies Associations exposed at Bigganjargga, Varangerfjorden area. Vertical scale in m.

2.4.1. Facies descriptions

2.4.1.1. Diamictites

Although diamictites comprise only a small portion of the Smalfjord Formation exposed in the Varangerfjorden area (Figs. 2.11, 2.12), two sub-types (pink and grey) can be distinguished.

Pink diamictite is massive and matrix-supported with a medium- to coarse-grained sandstone matrix and few clasts; in places it contains lenses of clast-rich grey diamictite (e.g. K98-4 at 9 m; Fig. 2.11). Clasts up to 20 cm diameter are subrounded to subangular, composed of granite or sandstone, and show no faceting or striae. Visual estimation of clast fabric shows no apparent preferred long-axis orientation. Individual or stacked beds (between 75 cm and 1 m thick) are traceable laterally over distances of several hundred m at Kvalnes. Upper and lower bed contacts are sharp and show slight relief (on the scale of cm) at the base.

Grey diamictite is also predominantly massive and matrix-supported, with a muddy medium-grained sandstone matrix, but contains common to abundant granite and sandstone clasts (Fig. 2.13A). Clasts up to 50 cm in diameter are subrounded to angular; and some exhibit faceting. Visual estimation of clast fabric indicates no apparent preferred long-axis orientation. Grey diamictite at Bigganjargga contains a relatively high proportion of granitic clasts, whereas grey diamictite at Kvalnes contains a relatively high proportion of sandstone clasts. Granitic clasts are derived from the Baltic Shield and the sandstone clasts are typical of the Veinesbotn Formation which underlies the

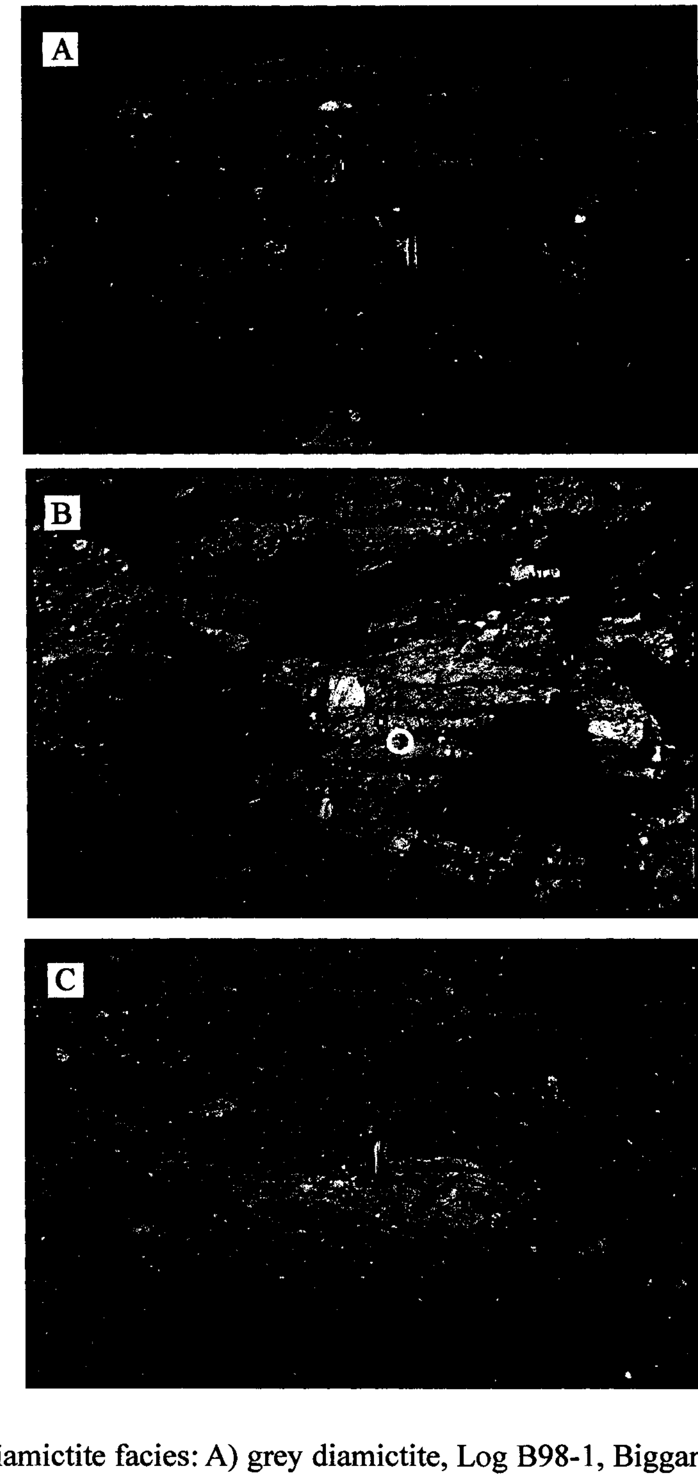


Fig. 2.13: Diamictite facies: A) grey diamictite, Log B98-1, Bigganjargga. Note random orientation and mixed lithology of clasts; B) chaotically interbedded grey and pink diamictite, Log K98-6, Kvalnes. Knife (circled) for scale; C) stringer of sandy diamictite within the grey diamictite at Bigganjargga (Log B98-1).

Smalfjord Formation (Fig. 2.3; Føyn, 1937; Edwards, 1975; Jensen & Wulff-Pederson, 1996). Bjørlykke (1967) reported striated clasts in the grey diamictite at Bigganjargga, although none was seen during this study.

Discontinuous and deformed units (10-20 cm thick) of pink diamictite occur as rafts within grey diamictites (K98-6 at 10 m; Fig. 2.13B). Grey diamictites also show localized chaotic internal structure and crude stratification defined by diamictite stringers with a relatively sandstone-rich matrix (Fig. 2.13C). Units of grey diamictite occur as distinct beds (0.3 to 4 m thick), as lenses within associated pink diamictite, conglomerate or sandstone and, at Bigganjargga, as an isolated mound within stacked sandstone beds. Lower bed contacts are irregular, sharp and conformable, or erosional with relief of between 0.5 and 2 m.

2.4.1.2. Conglomerates and breccias

Coarse-grained facies found in the Smalfjord Formation in the Varangerfjorden area include breccias and conglomerates (Figs. 2.11, 2.12). A 3 m thick breccia bed, exposed at site K98-2 (Fig. 2.11B), consists of angular sandstone clasts up to 40 cm within a coarse-grained sandstone matrix. Clasts show normal grading in places and are lithologically similar to the underlying Veinesbotn Formation. The breccia is crudely stratified with slight variations in matrix texture from fine- to coarse-grained sandstone.

Conglomerates vary from matrix- to clast-supported, massive to well stratified and poorly to very well sorted. Clasts up to 1.5 m in diameter are composed of either granite or sandstone. Granitic clasts tend to be subrounded, whereas muddy sandstone clasts are angular to sub-angular. Apparent clast imbrication was noted in some conglomerate facies (K98-1B at 30 m, and clast layers between 5 and 8 m at site K98-2; Fig. 2.11B), but no grading was observed. Massive conglomerates contain irregular lenses of diamictite and form distinct, sheet-like beds (< 3.5 m thick) with erosional basal contacts (e.g. K98-7 at 9 m, Fig. 2.11A), whereas beds of well-stratified and well-sorted conglomerates are interbedded with sandstones (e.g. K982, K98-3, Fig. 2.11B). Other conglomerates form discontinuous lenses several m thick and 10-30 m wide within massive sandstones (Fig. 2.12), some of which have a channel-like cross-sectional geometry.

2.4.1.3. Sandstones

Sandstones occur either as laterally discontinuous beds (<2 m thick) separating diamictite and conglomerate facies or as thick successions (up to 54 m) of stacked, well sorted facies (Fig. 2.11). Sandstone beds include trough cross-beds 20 to 50 cm thick, and deformed, horizontally-laminated, massive and rippled facies. Some beds contain deformed lenses of diamictite and conglomerate less than a metre wide; diamictite and conglomerate lenses also occur locally within the troughs of cross-bedded sandstones. Clasts within these conglomeratic lenses are less than 5 cm diameter and in places show apparent crude imbrication and/or inverse grading. Deformed sandstone at site K98-1 (between 5 and 7 m, Fig. 2.11B) shows a variety of synsedimentary folds and chaotic internal structure resulting from the deformation of crudely interbedded sandstone, conglomerate and diamictite.

The upper part of the Varangerfjorden succession is dominated by stacked, tabular beds of well sorted, fine- to coarse-grained, white sandstones (the 'Karlebotn Quartzite'; Banks *et al.*, 1971). Sandstone beds ranging from several cm to 2 m in thickness are stacked in a succession reaching 54 m thick (Fig. 2.12). At Bigganjargga, the sandstones are predominantly massive and deformed, whereas at Kvalnes, they are predominantly rippled and horizontally-laminated (Figs. 2.11, 2.12).

Discontinuous units (up to 4 m in thickness and several m long) of pebbly sandstone occur within the Smalfjord Formation in association with the Karlebotn Quartzites or with well sorted sandstones and conglomerates (sites K98-3, K98-2, B98-1; Figs. 2.11, 2.12). Pebbly sandstones contain subangular to subrounded clasts up to 10 cm in diameter. Clasts are either scattered in a random fashion throughout the pebbly sandstone (e.g. Log B98-1; Fig. 2.12) or occur as distinct granule layers (e.g. Log K98-3 between 3 and 5 m; Fig. 2.11); one pebbly sandstone unit displays normal coarse-tail grading (Fig. 2.12, at 70 m).

2.4.2. Facies interpretation

The sedimentological characteristics of diamictite facies, such as massive or chaotic internal structure, conformable to erosional basal contacts and common deformed diamictite rafts, are consistent with an origin as subaqueous debris flows. These diamictites were unlikely deposited at an ice margin due to the absence of ice-thrust deformation features, large depressions from grounding icebergs and ice-rafted debris in diamictites and associated facies (cf. Powell, 1981; Eyles *et al.*, 1985; Powell & Molnia, 1989; Lønne, 1995). The depositional origin of these diamictites however, is difficult to establish with certainty without first considering the depositional origin of associated facies and the lateral and vertical distribution of facies associations.

The presence of grading within the breccia unit observed at K98-2 (Fig. 2.11B) suggests a subaqueous setting, and the angularity of clasts contained within the breccia indicates relatively short distances of sediment transport. Similar breccias are reported at the base of the Smalfjord Formation elsewhere in the Varangerfjorden area (Røe, 1970; Edwards, 1984), including those on Skjahølmen Island (Fig. 2.10), which are associated with scarps (up to 4 m high) in underlying sandstones (Edwards, 1984). The breccia unit at Kvalnes is thus interpreted as resulting from localized mass failure of a relatively steep slope cut into sandstones of the underlying Veinesbotn Formation.

Conglomerates do not show any features typical of beach and fluvial gravel deposits (cf. Hein, 1984; Ethridge & Wescott, 1984; Nemec and Steel, 1984). They contain incorporated blocks of diamictite (e.g. site K98-7, Fig. 2.11), and are associated with breccias and stacked massive sandstones interpreted as turbidites. Similar associations of conglomerate and massive sandstone have been described from Pleistocene subaqueous glacial outwash deposits (e.g. Rust & Romanelli, 1975) and fan delta settings (e.g. Kleinspehn *et al.*, 1984; Martins-Neto, 1996); a similar subaqueous setting dominated by mass flow processes is proposed for the Varangerfjorden conglomerates.

Trough cross-bedded and rippled sandstones were deposited by localized traction currents. Horizontally-laminated sandstones were deposited either from suspension or as a result of upper flow regime plane bed conditions. Localized deformation of sandstone at site K98-1 was most likely caused by slumping. Deformation of these sediments by grounded ice is unlikely due to the lack of associated ice-thrust deformation features, and evidence of shearing and faulting typically associated with glacitectonized sediments. The quartzites that form the middle part of the Varangerfjorden succession at Bigganjargga (Fig. 2.12, 'Middle FA', see below) are interpreted as sediment gravity flow deposits (Edwards, 1975). Deformation within these sandstones was the result of water escape and loading caused by rapid sedimentation and/or seismic shock.

2.4.3. Vertical and lateral distribution of facies

The sections at Kvalnes and Bigganjargga record a similar vertical succession of facies; three stratigraphically distinct facies associations can be identified at both sites (Fig. 2.14). The *lower facies association* consists of chaotically interbedded diamictite, conglomerate and sandstone. The *middle facies association* consists entirely of stacked white sandstones, and the *upper facies association* consists of sandstones containing common conglomerate lenses. The depositional environments in which these deposits formed and the changing environmental conditions responsible for succession

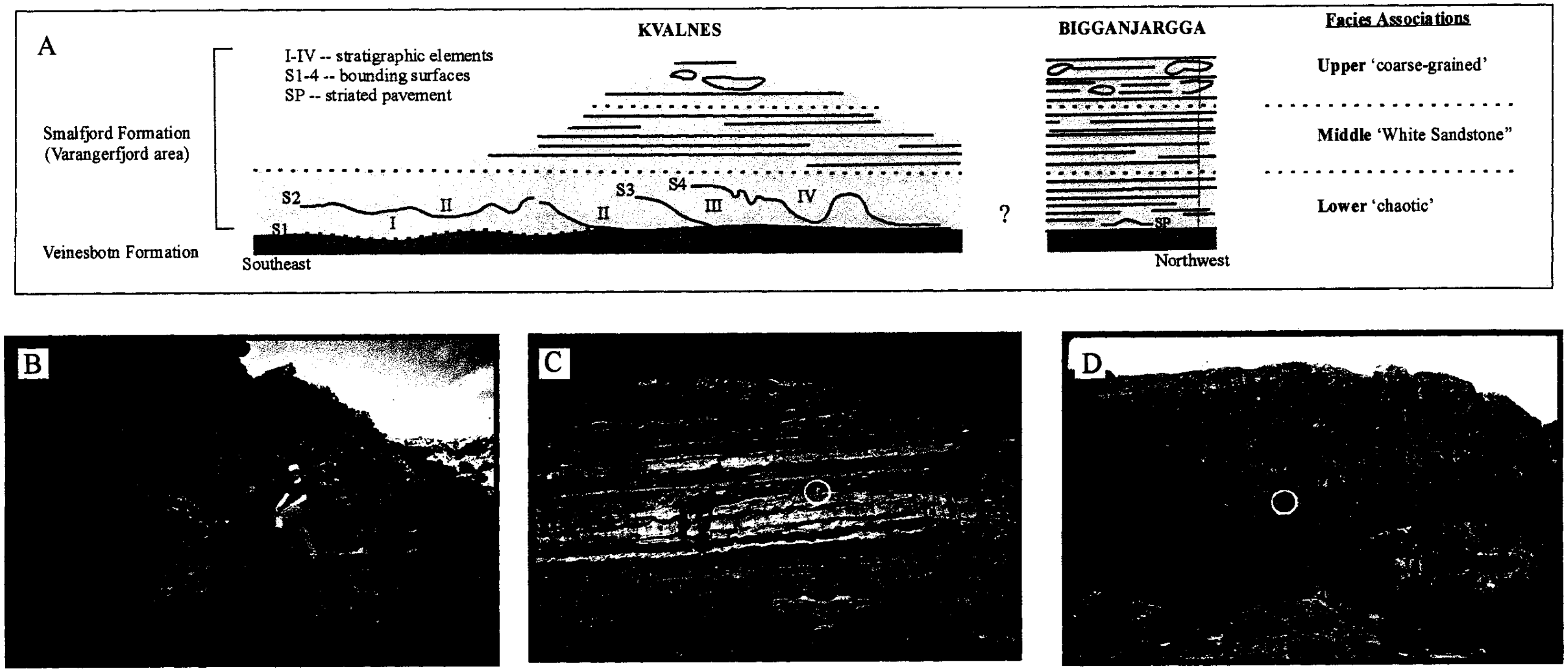


Fig. 2.14: Facies associations within the Smalfjord Formation in the Varangerfjorden area; A) schematic diagram showing stratigraphic and facies relationships between Kvalnes and Bigganjargga in the Varangerfjorden area. The deposits at Kvalnes are exposed over approximately 2 km of outcrop, while those at Bigganjargga are exposed over approximately 1.5 km of outcrop. The lower facies association is between 2 and 9 m thick, the middle facies association is between 19 and 54 m thick, and the upper facies association is between 8 and 34 m thick. The surfaces and elements that characterize the lower facies association are discussed in the text; B) conglomerate, cross-trough bedded sandstone, and diamictite of the lower facies association. Note chaotic geometry below the thick unit of conglomerate containing large boulders, Log K98-7, Kvalnes; C) stacked sandstone of the middle facies association showing soft sediment deformation structures; knife (circled) for scale, Log B98-1, Bigganjargga; D) conglomerate lens within stacked sandstones of the upper facies association, Bigganjargga. Knife (circled) for scale.

development are determined through analysis of these facies associations and their vertical and lateral distribution.

2.4.3.1. Lower Facies Association

The lower facies association forms the lowermost, diamictite-bearing part of the stratigraphy at both Kvalnes and Bigganjargga. Although the lower facies association has similar primary characteristics and stratigraphic position at each site, there are some important differences between sites (Fig. 2.14A).

Kvalnes: The lower facies association is approximately 5-10 m thick and is exposed over several km along the eastern margin of the Vieranjar'ga Peninsula (Fig. 2.10). It consists of chaotically interbedded diamictites, conglomerates and sandstones with limited lateral continuity of individual facies types (Figs. 2.14B and 2.15). Despite this lateral facies variability, four distinct sediment packages or 'elements' can be recognised within the lower facies association, each separated by a laterally extensive erosional surface. These 'elements' show a crude offlapping relationship to the north and record successive episodes of erosion, channelling and deposition (Figs. 2.14A, 2.15). They are not defined as formal 'architectural elements' (e.g. Miall, 1985; Boyce & Eyles, 2000) as they are limited to only one outcrop area, but their form and spatial distribution can be used to establish the sequence of events that created the chaotic assemblage of sediments forming the lower facies association at Kvalnes.

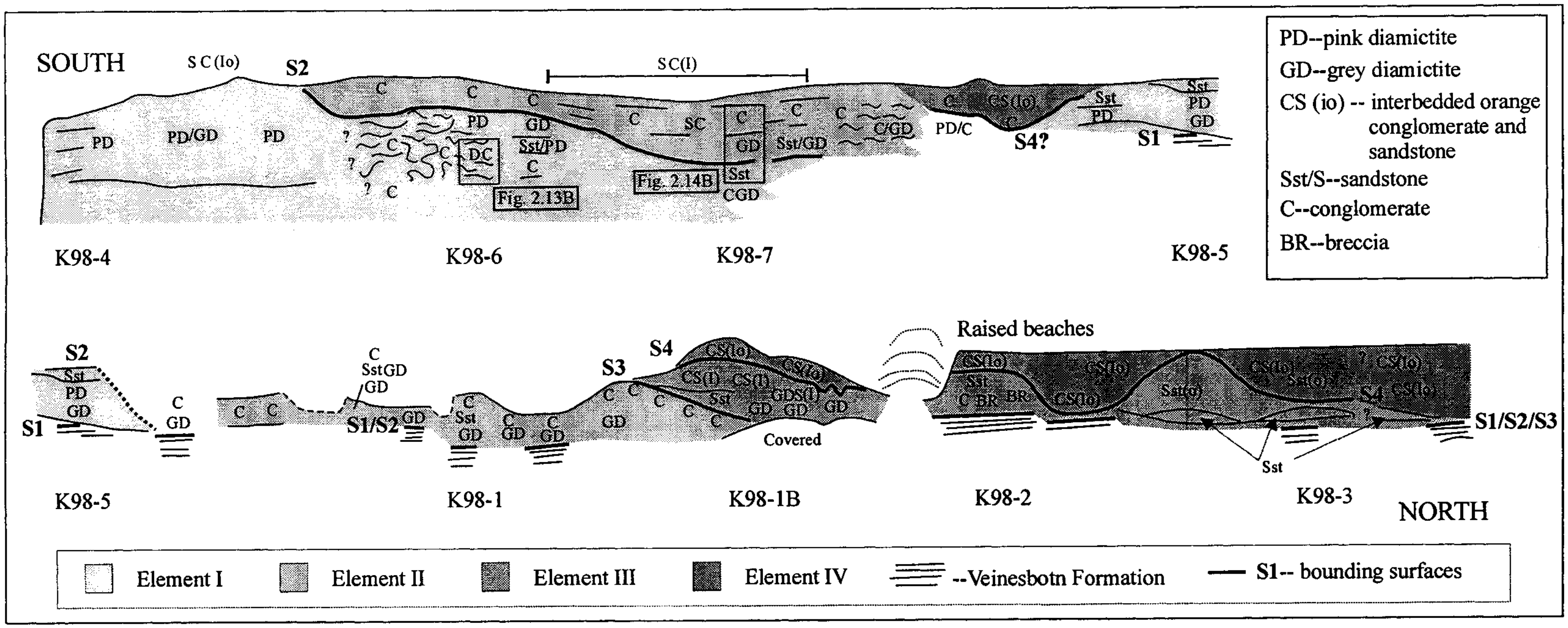


Fig. 2.15: Outcrop sketch of the lower facies association at Kvalnes. The distance from the southern end of the outcrop (upper left, upper panel) to the northern end (lower right, lower panel) is approximately 2 km and the height of outcrop ranges between 4 and 9 m. Site K98-5 is shown in both panels (c.f. Fig. 2.14; see Fig. 2.11 for detailed sedimentology at sites K98-1 to K98-7). Note offlapping geometry of stratigraphic units to the north and change from chaotic facies to the south to better organized and channelized facies to the north. Bold lines represent surfaces (S1, S2, S3, S4) that bound stratigraphic elements within the lower facies association.

The lowermost *element* (Element I, Fig. 2.15) occurs primarily at the southern end of the exposure and consists of chaotically interbedded pink diamictite with minor grey diamictite and conglomerate beds. The second *element* (Element II; Fig. 2.15) consists of grey diamictite overlain by a relatively continuous bed of massive conglomerate, and infills broad erosional lows cut into the underlying deposits of Element I (Fig. 2.15). The second *element* is distinguished from the first element by the absence of pink diamictite and its position overlying a laterally traceable erosion surface (S1; Fig. 2.15).

The third *element* (Element III; Fig. 2.15) offlaps to the north and partly overlies both the second element and the Veinesbotn Formation. It is distinguished from the second element by the presence of interbedded conglomerate and sandstone facies that onlap the conglomerate unit of Element II (surface S3; Fig. 2.15). The fourth *element* (Element IV) occurs in the northern part of the exposure and consists of interbedded massive and horizontally laminated sandstone and conglomerate; it is separated from Element III by a channelized erosion surface (S4; Fig. 2.15). Based on observation of the two-dimensional outcrop geometry, channels on this surface appear to have an east-west long axis orientation.

The depositional origin of diamictites can now be discussed in the context of the depositional origin and spatial distribution of associated facies. The close association of diamictites with breccia, conglomerates and sandstones interpreted as subaqueous mass flow deposits, the prograding geometry identified in the diamictite-bearing lower facies association together with the sedimentary characteristics of the diamictites themselves, suggest a mass flow origin for the diamictites. The diamictites are unlikely to be of subglacial origin due to the lack of association with subaqueous or terrestrial outwash sediments (Boulton & Deynoux, 1981). In addition, the prograding geometry of the elements within the lower facies association differ significantly from the tabular (subglacial till sheets) and ridge-like (grounding line fans) geometries found in ice-contact marine settings and the hummocky topography found in ice-marginal terrestrial settings.

The four depositional elements identified within the lower facies association indicate progradation of coarse-grained sediments from south to north. Conglomerate beds show a transition in character from being predominantly massive and associated with diamictites in the south (mostly poorly sorted and organized; Elements I and II), to being well stratified and associated with sandstone in the north (better organized and sorted; Elements III and IV; Figs. 2.11, 2.14A and 2.15). The increasing organization of coarse-grained sediment gravity flow deposits towards the north is consistent with deposition on a northward prograding fan delta or debris apron (Walker, 1975; Prior & Bornhold, 1990; Martins-Neto, 1996), which developed adjacent to the Baltic craton (Fig. 2.9). Although glacial ice on the basin margin may have contributed sediments to the prograding fan delta preserved at Kvalnes, an ice-contact glaciomarine setting for the fan delta is unlikely given the absence of ice-rafted debris, glacitectonic deformation, iceberg grounding features, subaqueous outwash sediments and/or large-scale ridge-like features (Powell, 1981; Eyles *et al.*, 1985; Powell & Molnia, 1989; Lønne, 1995).

The lateral variability of facies types contained within the four elements of the lower facies association at Kvalnes may be associated with changes in sediment supply and slope as documented in the evolution of Holocene fan deltas (Prior & Bornhold, 1990). Element I (Figs. 2.14, 2.15) probably represents the early stage of development of the debris apron where rapid avalanching and slumping of coarse-grained debris results in the chaotic accumulation of diamictite, conglomerate and sandstone. In contrast, Element IV represents later stages of development of the apron when decreased supply of boulder-rich gravel led to a change towards more channelized flow of coarse sands and gravels. Elements II and III represent intermediate stages between these two end members. The absence of muds in the lower facies association suggests that the depositional environment remained under relatively high energy conditions, proximal to a coarse-grained sediment supply, throughout these different stages.

Bigganjargga: At Bigganjargga, the lowermost facies association consists predominantly of stacked beds (20-60 cm thick) of massive and deformed sandstone (Fig. 2.14). Grey diamictite, similar to that found at Kvalnes, crops out as an isolated mound overlying a striated pavement on the surface of the Veinesbotn Formation (see Bjørlykke, 1967; Edwards, 1975; Jensen & Wulff-Pedersen, 1996 and Rice & Hofmann, 2000 for detailed studies of this site).

The deposits of the lower facies association at Bigganjargga were deposited by turbidity currents, which are likely to be basinward equivalents of the conglomeratic sediment gravity flows preserved at Kvalnes. The isolated mound of grey diamictite is interpreted as a relatively far-travelled subaqueous debris flow deposit, which has been partially reworked by subsequent turbidity currents.

2.4.3.2. Middle Facies Association

At both Kvalnes and Bigganjargga, the *middle facies association* consists of a thick succession (19-54 m) of stacked, well-sorted, quartzite laterally continuous over several km of outcrop (Figs. 2.14A, 2.14C). The lateral facies change from horizontally-laminated and rippled sandstones deposited in a shallow marine setting at Kvalnes to massive and deformed sandstones deposited by turbidites at Bigganjargga, suggests deepening water conditions northward toward Bigganjargga. The lower contact of the thick 'white' sandstone package is sharp to gradational on the underlying lower facies association (Figs. 2.11, 2.12).

The contact between the lower and middle facies associations most likely represents a significant change in depositional conditions where the chaotic deposition of coarse-grained sediments, characteristic of the lower facies association, is replaced by more organized deposition of medium- to fine-grained turbidite sandstones. The change in facies types and their organization may reflect a change in sediment supply or a change in water depth.

2.4.3.3. Upper Facies Association

The *uppermost facies association*, exposed along strike over 2 km at Bigganjargga, is distinguished from the middle facies association by the presence of large conglomerate lenses within the stacked massive white sandstones (Figs. 2.14A, 2.14D); in all other respects the sandstones have the same characteristics as those of the middle facies association. The appearance of conglomerate lenses within stacked massive sandstones in the upper facies association records the return of high-concentration sediment gravity flows. This change in depositional process may be the function of increased delivery of coarse-grained debris to the basin and/or a change in relative sea level creating shallower, more proximal conditions.

2.4.4. Depositional setting of the Smalfjord Formation, Varangerfjorden area

The Smalfjord Formation in the Varangerfjorden area is dominated by relatively coarse-grained facies types including conglomerate, sandstone and diamictite that indicate deposition by a combination of sediment gravity flow processes and traction currents under relatively shallow marine conditions. The facies associations and large scale facies architecture exposed at Kvalnes and Bigganjargga are not consistent with those found in ice-contact terrestrial or glaciomarine settings, but more closely resemble those of prograding fan deltas or subaqueous debris aprons. Facies characteristics and their lateral and vertical distribution indicate deposition on a northward prograding debris apron subject to changes in the supply of coarse-grained sediment and/or changes in relative sea level (Fig. 2.9).

The absence of fines in the succession indicates deposition occurred under high energy conditions, proximal to a source of coarse-grained debris. The northward 'shingling' of depositional 'elements' in the lowermost facies association at Kvalnes, the increased organisation of conglomerates and sandstones towards the north, and the presence of crystalline clasts from the adjacent craton within the grey diamictite all suggest an approximately southern sediment source (Føyn, 1937; Reading & Walker, 1966; Banks *et al.*, 1971; Edwards, 1984). Sediment was probably supplied by a fluvial system draining the craton to the south; the coarse-grained and poorly sorted nature of the sediments suggests that this may have been a glaciofluvial system.

2.5. DISCUSSION: EVIDENCE FOR GLACIATION IN THE SMALFJORD FORMATION

The Smalfjord Formation has traditionally been interpreted as the product of both terrestrial and marine glacial deposition. Glacial conditions are inferred from the interpretation of poorly sorted deposits overlying the striated pavement at Bigganjargga as terrestrial 'tillites' (Edwards, 1975), the presence of striated clasts in the grey diamictite (Bjørlykke, 1967), and of 'lonestones' in fine-grained facies (Reading & Walker, 1966; Edwards, 1984). However, sedimentological characteristics of the Smalfjord Formation exposed at both Tarmfjorden and Varangerfjorden indicate deposition primarily by sediment gravity flow processes in unstable subaqueous environments. Evidence for glacial conditions is limited to the striated pavement and the presence of ice-rafted debris reported by previous workers; the nature of this evidence and its implications for ice sheet reconstruction are discussed below.

2.5.1. Origin of the striated pavement at Bigganjargga

The striated pavement at Bigganjargga and the overlying grey diamictite have been interpreted both as glacial features (Reusch, 1891; Føyn, 1937; Bjørlykke, 1967; Banks *et al.*, 1971; Edwards, 1975, 1997; Rice & Hofmann, 2000), and as the products of subaqueous sediment gravity flow processes (Crowell, 1964; Harland, 1964b; Schermerhorn, 1974; Jensen & Wulff-Pedersen, 1996, 1997). Those that prefer a glacial origin interpret the pavement as a product of glacial erosion during ice advance and the overlying diamictite as a glacial depositional feature (meltout till or moraine) subsequently reworked during ice retreat. In contrast, those that prefer a sediment gravity flow origin believe that the pavement formed as a result of abrasion of underlying unlithified sands by clasts carried within a debris flow (overlying diamictite). A recent paper by Rice & Hofmann (2000) combines elements of both these explanations and suggests pavement formation by subglacial erosional processes, at some considerable time prior to emplacement of the overlying diamictite.

The sedimentological analysis of the Smalfjord Formation reported here does not aim to resolve the origin of the Bigganjargga striated pavement but does provide important contextual information regarding the depositional origin of associated deposits. The sedimentological characteristics of the grey diamictite at Bigganjargga and its association with sandstones interpreted as turbidites are all consistent with a debris flow origin for the diamictite (Jensen & Wulff-Pedersen, 1996). Laterally equivalent deposits at Kvalnes are interpreted as subaqueous sediment gravity flow deposits accumulating on a debris apron prograding northwards off the Baltic craton and show no evidence of ice contact glacial conditions. Thus, while the striated pavement may have a glacial origin, there is no evidence to support direct glacial deposition of the overlying diamictite.

2.5.2. Glacial influences on deposition

The occurrence of rare, isolated 'lonestones' in fine-grained deposits at Tarmfjorden provides the only direct evidence for the presence of ice in the basin during deposition of the Smalfjord Formation. However, the presence of 'lonestones' alone cannot be used to infer fully glacial conditions; icebergs can travel long distances from glaciated margins (e.g. Bond *et al.*, 1992; Eyles *et al.*, 1997) and clasts can be rafted by seasonal ice under cold, non-glacial conditions (Gilbert, 1990). In addition, outsized clasts can be transported by turbulence within turbidity currents.

The characteristics of clasts found in the diamictites of the Smalfjord Formation, such as occasional faceted or striated clasts and the presence of mixed extrabasinal lithologies, have also been used to infer a glacial origin for the deposits. These clast characteristics, however, merely reflect sediment source rather than mode of deposition

(Schermerhorn, 1974; Eyles *et al.*, 1985). Thus, although a glacial source of sediment is possible, the extent or distribution of glacial ice in the basin at the time of Smalfjord Formation deposition cannot be demonstrated. The diamictites examined in both the Tarmfjorden and Varangerfjorden areas do not appear to be the result of direct glacial deposition and their mass flow origin makes it difficult to discriminate a glacial influence on sedimentation.

2.5.3. Controls on succession development

The stratigraphic succession through the Smalfjord Formation exposed at both Tarmfjorden and Varangerfjorden shows an overall upward decrease in the amount of coarse-grained sediment. This fining-upward trend continues into the overlying Nyborg Formation (Fig. 2.2), which consists predominantly of fine-grained turbidites deposited in a deep marine setting (Reading & Walker, 1966; Edwards, 1984). Edwards (1984) interpreted the fining-upward trend from the Smalfjord Formation into the overlying Nyborg Formation as recording the loss of poorly sorted coarse-grained sediment and eustatic sea level rise associated with glacial retreat from the basin. However, similar fining-upward successions have been reported from glacial and non-glacial marine rift basin settings (Ravnås & Steel, 1998 and references therein) where tectonic evolution of the basin itself controls succession development. Given the evidence for Neoproterozoic regional extension in the Gaissa Basin and the North Atlantic region (Nystuen, 1985; Gayer & Rice, 1989; Young, 1995; Drinkwater *et al.*, 1996), the presence of breccia and scarps in the Varangerfjorden area and the predominance of mass flow facies in the Smalfjord Formation, tectonic activity may have been an important factor in controlling succession development.

2.6. CONCLUSION

The sedimentological characteristics of the Smalfjord Formation exposed at both Tarmfjorden and Varangerfjorden indicate deposition in unstable subaqueous environments dominated by sediment gravity flow processes. At Tarmfjorden, heterolithic debris flows and fine-grained turbidites were deposited on, or adjacent to, a relatively steep slope. In the Varangerfjorden area, chaotically-bedded, coarse-grained and sandy sediment gravity flow facies indicate deposition on a fan delta or debris apron proximal to a coarse-grained sediment source to the south.

There is no unequivocal evidence to suggest the direct influence of glacial ice on sedimentation in this area as previously interpreted, although the striated pavement at Bigganjargga may be interpreted as an older subglacial feature (Rice & Hofmann, 2000). Glacier ice may have contributed poorly sorted coarse-grained sediment to the Tarmfjorden and Varangerfjorden area, but facies characteristics and distribution indicate that an ice-contact terrestrial or marine depositional setting is unlikely.

The sedimentological characteristics of the Smalfjord Formation and its relationship with fine-grained deposits of the overlying Nyborg Formation are consistent with depositional models describing the infills of extensional basins. Development of the succession may have been influenced by tectonic activity associated with the break up of supercontinent Rodinia and the formation of local extensional basins rather than by globally-synchronous climate change. The predominance of mass flow facies in the Tarmfjorden and Varangerfjorden areas makes it difficult to identify any evidence for severe glacial conditions as proposed in the snowball Earth hypothesis.

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CHAPTER THREE

THE GREAT BRECCIA, PORT ASKAIG FORMATION, SCOTLAND[1]

3.1. INTRODUCTION

Megabreccias and olistostromes, resulting from large-scale failure of consolidated sediments, have been reported from deep water marine settings adjacent to carbonate platforms, active faults, and prograding deltas. These poorly sorted deposits contain large clasts, 10's of metres to kilometres in diameter, and are common in a variety of tectonically active settings such as subduction zones, clastic and carbonate slope settings, oblique-slip mobile zones, and convergent and divergent plate margins (Pickering *et al.*, 1989; Miall, 1990; Busby and Ingersoll, 1995).

Megabreccias are often used as sedimentological indicators of tectonic activity, although other factors such as gas charging, storm wave impacts, and high rates of sedimentation can also lead to their formation (Hubert *et al.*, 1977; Prior *et al.*, 1982; Labaume *et al.*, 1987; Bosellini, 1989; Hine *et al.*, 1992; Spence and Tucker, 1997; Payros *et al.*, 1999; Gardner *et al.*, 2000). In ancient sedimentary basins, where structural evidence of tectonic activity (e.g. fault traces) is often difficult to identify, megabreccias

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[1] A modified version of this chapter will appear in *Sedimentary Geology* as Arnaud, E. and Eyles, C. H., 2002. Catastrophic mass failure of a Neoproterozoic glacially-influenced continental margin, the Great Breccia, Port Askaig Formation, Scotland. *Sedimentary Geology*, in press.

provide an important clue as to the paleotectonic history of the basin. However, as megabreccias form in a number of different settings, and have similarities with poorly-sorted and coarse-grained deposits formed in glacial environments, it is important to fully understand their characteristics and mode of formation in order to evaluate their significance as paleotectonic indicators.

The Great Breccia, a 50 metre-thick dolomitic diamictite with large clasts found within the glacigenic Neoproterozoic Port Askaig Formation, was initially interpreted as a subglacial deposit similar to Pleistocene tills found in Norfolk, England (Spencer, 1971). According to Spencer (1971), the Port Askaig Formation recorded successive advance-retreat cycles of an ice margin into a marine basin and indicated repeated episodes of extensive ice cover. The diamictite of the Great Breccia, however, does not exhibit features typical of subglacial tills such as bullet-shaped boulders, striated, faceted, and oriented clasts or shear planes (Boulton and Deynoux, 1981; Eyles and Eyles, 1992). Interpretation of the Great Breccia as a subglacial till would also require its association with other terrestrial facies such as glaciofluvial deposits, which are noticeably absent in the Port Askaig Formation.

Eyles (1988) reinterpreted the Port Askaig Formation as a glacially-influenced marine succession where diamictites were formed by 'rainout' of fine-grained sediments and ice rafted debris and by subaqueous sediment gravity flow processes. In this alternative depositional model, the Great Breccia formed as a result of mass flow processes related to tectonic instability and extensional faulting in the basin

(Schermerhorn, 1974; Eyles, 1988) and may not indicate extensive ice cover in the basin at the time of its formation.

This paper presents new data on the sedimentology of the Great Breccia based on investigation of exceptional outcrops along the shores of the Garvellach Islands in southwestern Scotland (Fig. 3.1). The sedimentary characteristics of the Great Breccia are described and analysed together with associated deposits in order to establish details of its depositional origin, the setting in which it formed, and its potential tectonic and paleoenvironmental significance.

3.2. GEOLOGICAL SETTING OF THE PORT ASKAIG FORMATION

The Port Askaig Formation is found at the base of the Argyll Group of the Dalradian Supergroup and is exposed along a transect that extends from the northern coast of Ireland to the northeast coast of Scotland (Fig. 3.1). The Dalradian Supergroup contains a variety of sedimentary facies types and volcanic deposits that accumulated in predominantly marine environments during episodic rifting of Rodinia in the late Proterozoic (Fig. 3.2; Harris *et al.*, 1978; Anderton, 1982; Soper, 1994; Glover *et al.*, 1995).

The Islay Subgroup lies at the base of the Argyll Group and consists of diamictites of the Port Askaig Formation, the clastic/carbonate succession of the overlying Bonahaven Formation and a thick succession of sandstone known as the Jura Quartzites (Fig. 3.2). The Islay Subgroup marks the onset of unstable tectonic conditions in the

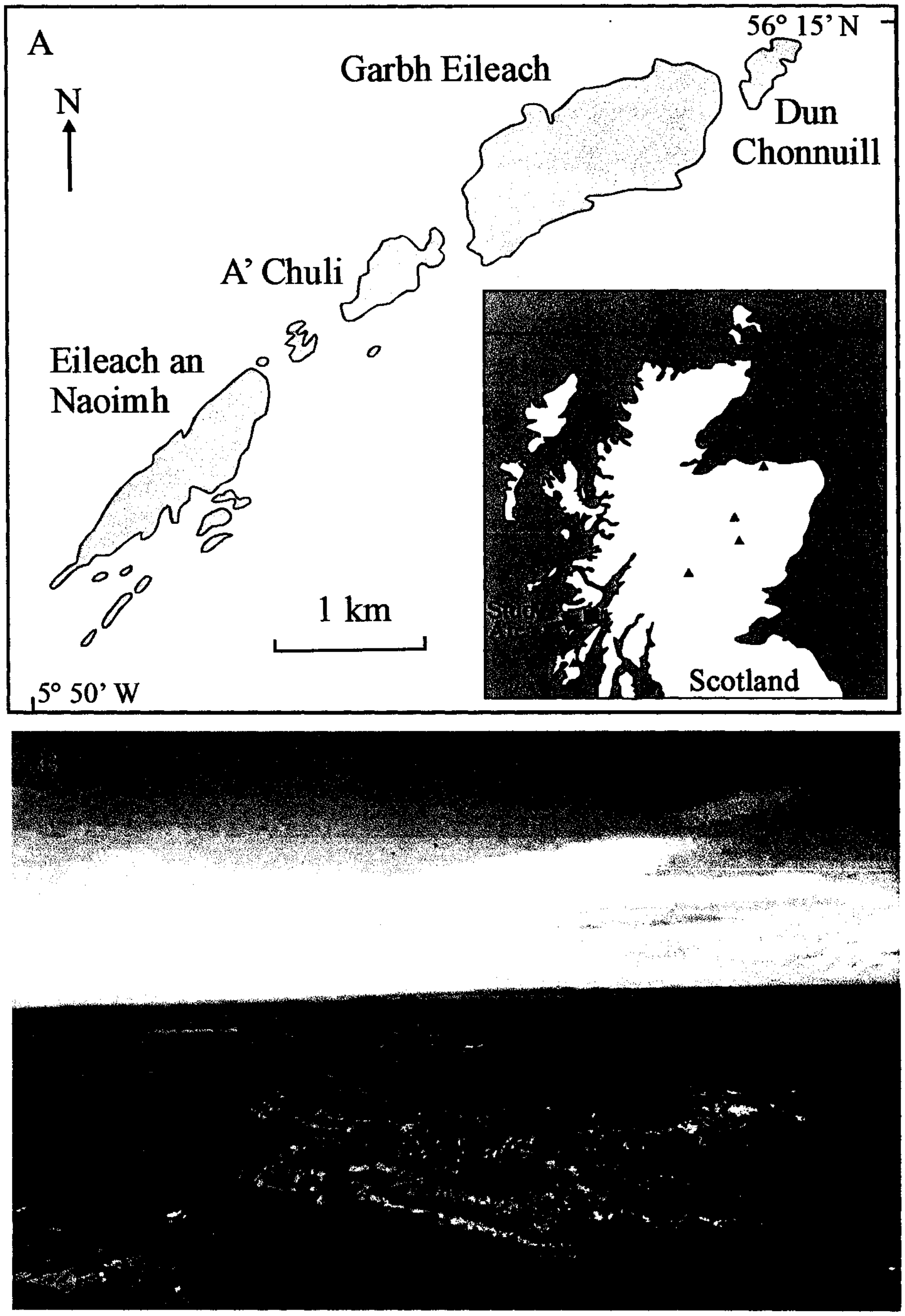


Fig. 3.1: Garvellach Islands, Scotland. A) Location map. Triangles on inset map of Scotland show other outcrops of the Port Askaig Formation (modified from Spencer, 1971); B) Photo looking from Garbh Eileach to the southwest towards A'Chuli and Eileach an Naoimh. Strata are dipping 35° to the south/southeast.

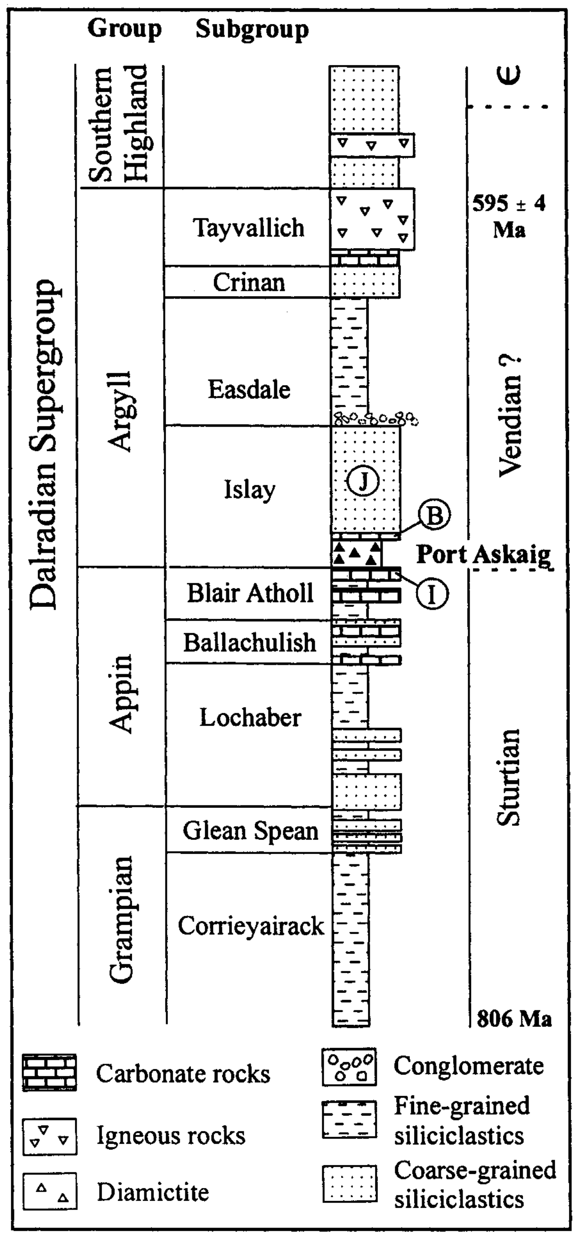


Fig. 3.2: The Dalradian Supergroup and stratigraphic position of the Port Askaig Formation. Total stratigraphic thickness is approximately 25 km (modified from Anderton, 1985; Eyles, 1988; Prave, 1999). Date for the Grampian Group is from Noble *et al.*, (1996), whereas the date for the Tayvallich Volcanics is from Halliday *et al.*, (1989). J, Jura Quartzites; B, Bonahaven Formation; I, Islay Limestone.

Dalradian basin associated with initial opening of the Iapetus or proto-Atlantic ocean (Harris *et al.*, 1978; Anderton, 1982). Tectonic influences on deposition of the Islay Subgroup have been inferred, largely on the basis of dramatic thickness changes between outcrops in northern Ireland and Scotland (Anderton, 1982); structural evidence for extensional tectonic activity is ambiguous due to subsequent deformation during the Caledonian Orogeny (Anderton, 1988). However, the stratigraphy of the overlying Easdale, Crinan and Tayvallich subgroups of the Argyll Group is consistent with depositional models of extensional basin settings (Harris *et al.*, 1978; Anderton, 1982); coarse-grained sedimentation at the base of the Easdale Subgroup is followed by accumulation of deep water, fine-grained sediments and turbidites (Easdale and Crinan subgroups), culminating in the formation of the Tayvallich Volcanics (Fig. 3.2).

The occurrence of coarse-grained and poorly sorted diamictites in the Port Askaig Formation has been attributed to glacially-influenced sedimentation on a continental margin during the Neoproterozoic Varangian glaciation (Kilburn *et al.*, 1965; Spencer, 1971; Eyles, 1988). The Port Askaig Formation, like many other Neoproterozoic glacial successions, contains and is closely associated with, carbonate deposits. It is underlain by the Islay Limestone of the Appin Group, which records carbonate sedimentation on a low energy shelf (Harris *et al.*, 1978, Anderton, 1982) and is overlain by the Bonahaven Formation, which records clastic shallow marine to carbonate lagoonal sedimentation (Spencer and Spencer, 1972; Fairchild, 1980).

The Port Askaig Formation consists of relatively planar tabular units of diamictite interbedded with detrital carbonate and siliciclastic conglomerate, sandstone and mudstone and has a stratigraphic thickness of over 700 m (Fig. 3.3; Kilburn *et al.*, 1965; Spencer, 1971; Eyles, 1988). Spencer (1971) identified 47 individual diamictite beds (D1-D47) and subdivided the Port Askaig Formation into five members defined by major changes in predominant facies types, clast lithology and abundance of sandstone interbeds. Member I of the Port Askaig Formation is characterized by stacked beds of dolomitic diamictite interbedded with carbonate-rich siliciclastic conglomerate, sandstone and mudstone and by two distinctive sedimentary units, the Great Breccia and the Disrupted Beds (Figs. 3.3 and 3.4). Members II through V contain stacked beds of sandstone and diamictite that become increasingly siliciclastic upsection (Spencer, 1971). While the type section of the Port Askaig Formation is found on the island of Islay (Spencer, 1971), the lowermost 450 m of the Port Askaig Formation (Members I through III) are best exposed on the Garvellach Islands.

There are two radiometric dates within the Dalradian Supergroup which help constrain the age of the Port Askaig Formation (Fig. 3.2). A shear zone truncating the base of the Grampian Group gives a minimum age for the onset of Dalradian sedimentation of 806 Ma (Noble *et al.*, 1996). The Tayvallich Volcanics at the top of the Argyll Group were emplaced by 595 Ma (Halliday *et al.*, 1989). Biostratigraphic evidence places the Precambrian-Cambrian boundary above the Tayvallich Volcanics and

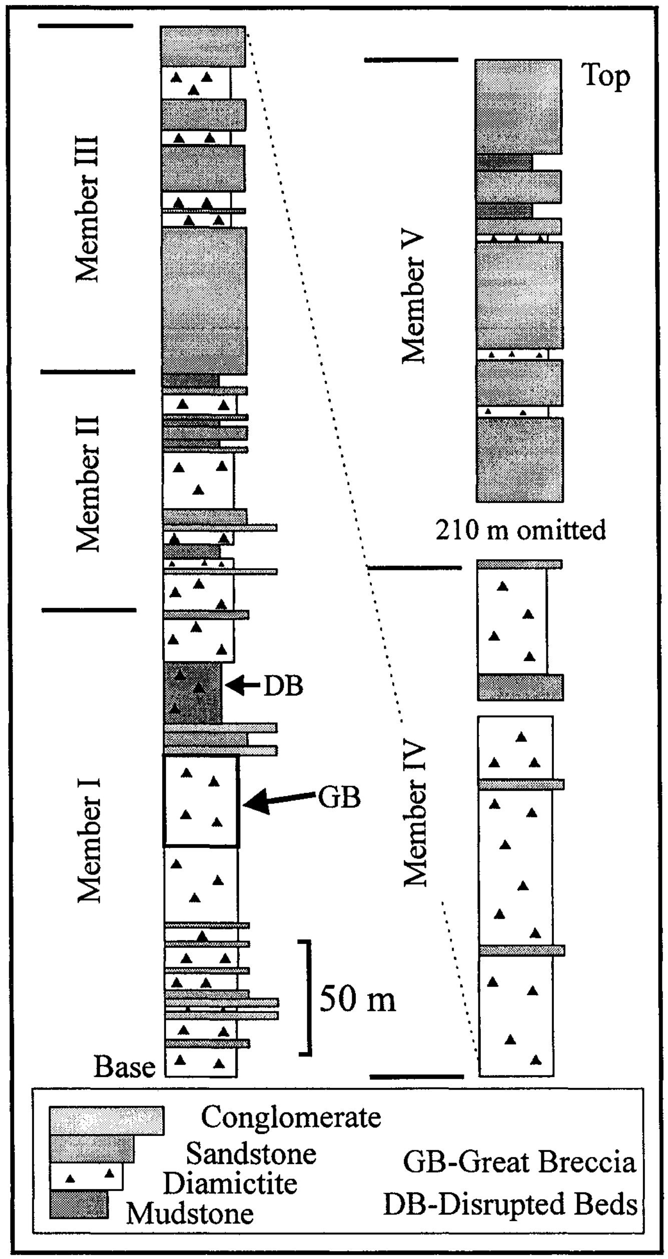


Fig 3.3: Schematic log through the Port Askaig Formation showing Members I through V (Spencer, 1971); base is at lower left, top at upper right. Logs of Members I through III are based on analysis of outcrops on Garvellach Islands; logs of Members IV and V are modified from Spencer (1971) and based on outcrops from Islay.

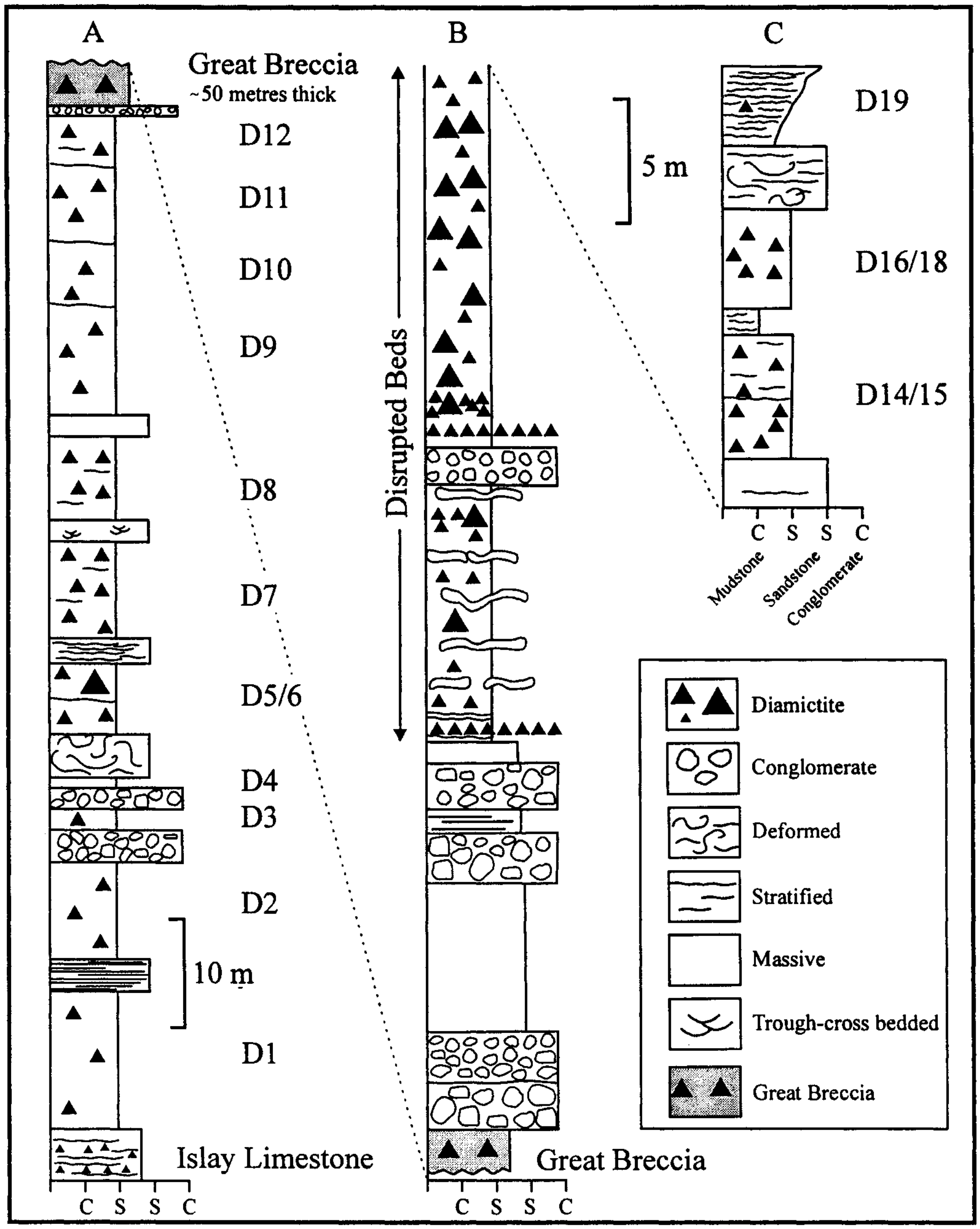


Fig. 3.4: Sedimentological log of Member I of the Port Askaig Formation showing facies associated with the Great Breccia; base at the lower left, top at upper right. Note scale change between column A and columns B/C. The full thickness of the Great Breccia (approximately 50 m) is not shown. Sediments below the Great Breccia (column A) were described from outcrops on east Garbh Eileach, whereas the log of sediments above the Great Breccia (columns B/C) is based on outcrops from Eileach an Naoimh. Numbering of diamictite units follows Kilburn *et al.* (1965) and Spencer (1971).

within the Southern Highland Group, although its exact position is still debated (see review in Prave, 1999).

The Port Askaig Formation has been correlated lithologically with other Varangian glacial deposits of the North Atlantic region dated around 590-610 Ma (Spencer, 1975; Hambrey, 1983; Harland *et al.*, 1989). However, glaciomarine deposits in Donegal, Ireland have recently been documented well above the Port Askaig Formation, suggesting that the Port Askaig Formation may actually record sedimentation during the older Sturtian glaciation (Condon and Prave, 2000). A Sturtian age (*c.*723 Ma; Brasier *et al.*, 2000) is further supported by the stratigraphic position of the Port Askaig Formation, which is approximately 8 km below the Tayvallich Volcanics (dated at 595 ± 4 Ma; Fig. 3.2; Prave, 1999) and the chemostratigraphy of carbonates associated with the Port Askaig Formation, which is similar to that found in other Sturtian successions (Brasier and Shields, 2000).

3.3. THE GREAT BRECCIA

The Great Breccia is approximately 50 metres thick and consists of dolomitic matrix-supported diamictite with large clasts up to 100's of metres across (Figs. 3.3, 3.5 and 3.6). On the Garvellach Islands, the Great Breccia is found within Member I of the Port Askaig Formation, approximately 100 metres above the underlying Islay Limestone (Figs. 3.2 and 3.3). It overlies a thick succession of diamictites (D1-D12) interbedded

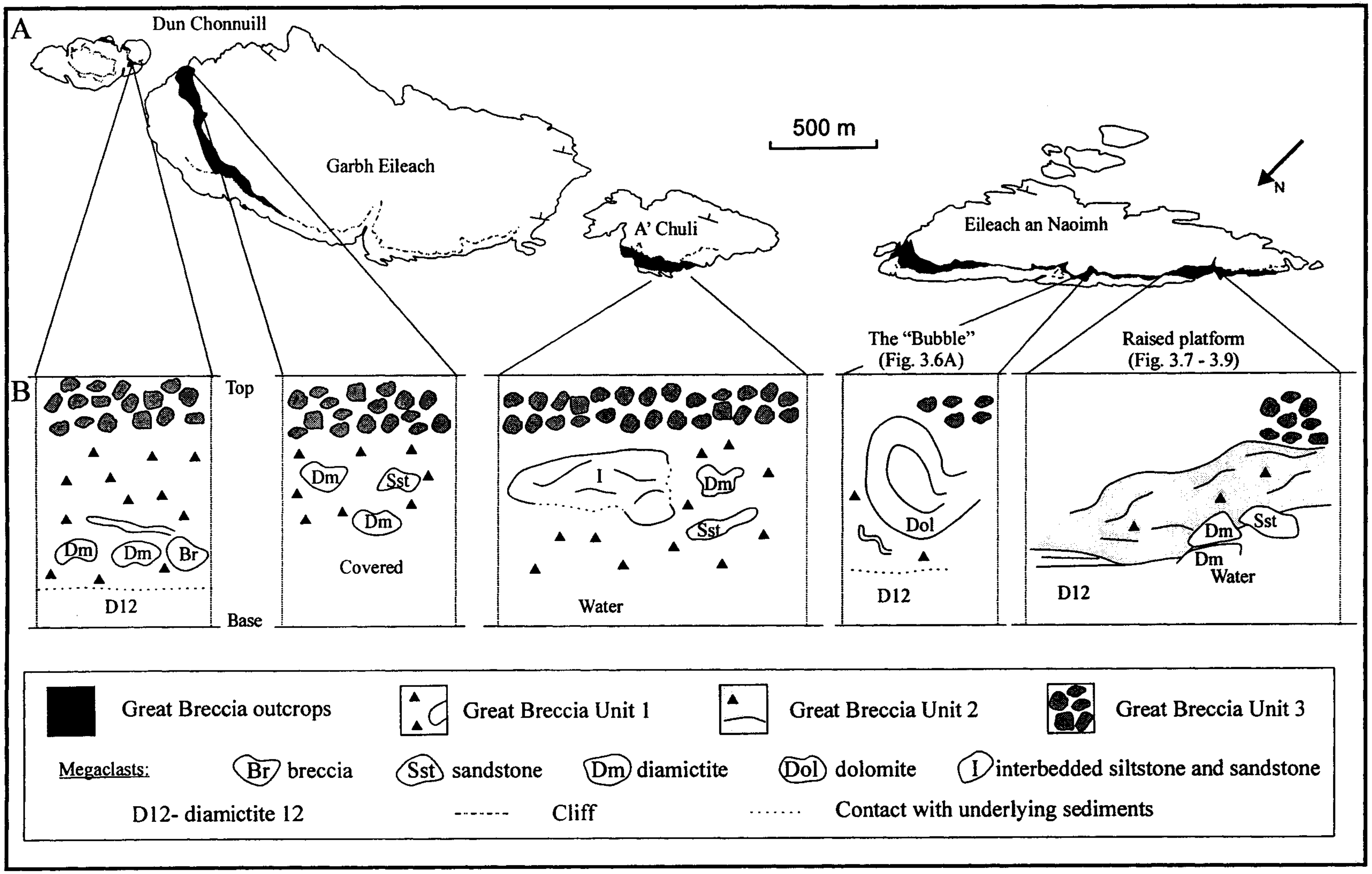


Fig. 3.5: A) Plan view of the outcrops of the Great Breccia on each of the four Garvellach Islands; B) Schematic vertical sections (each between 30 and 50 m thick) through outcrops of the Great Breccia, showing the three sedimentary units.

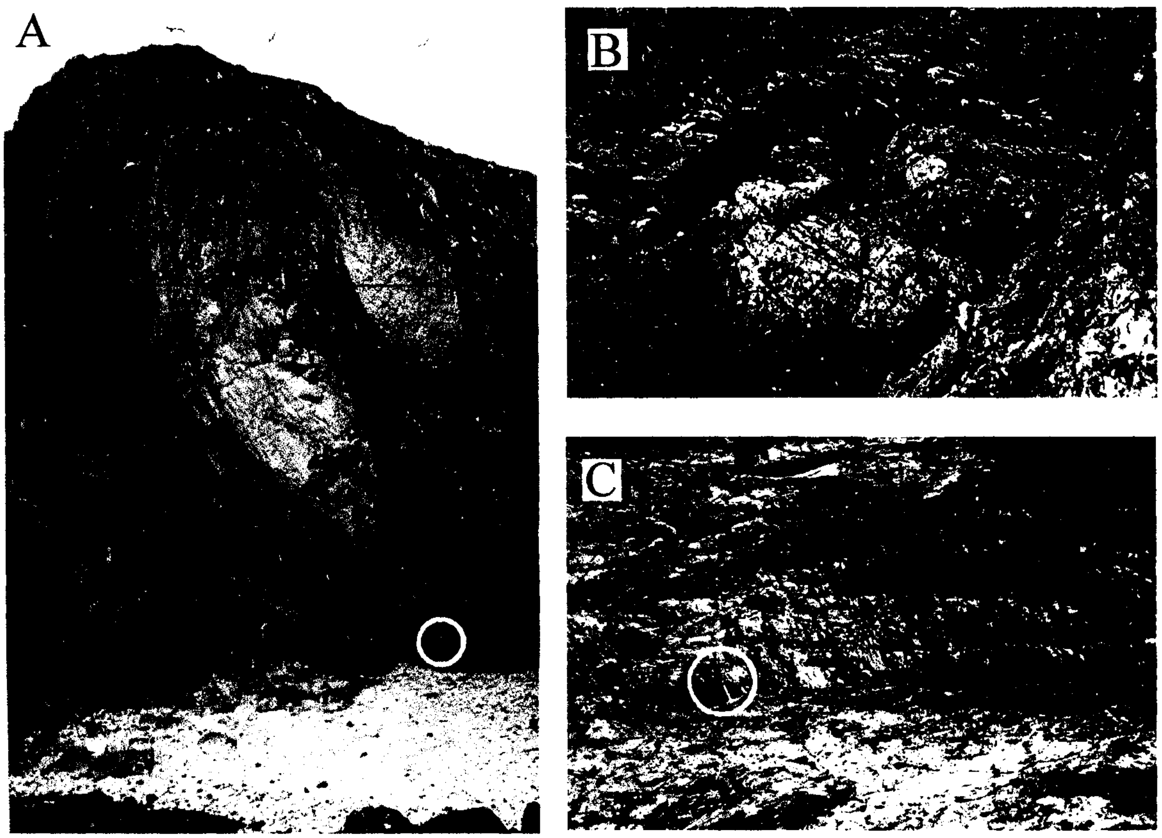


Fig. 3.6: Great Breccia, Unit 1; megaclasts within the megabreccia. A) the 'Bubble', a megaclast of dolomite exhibiting broad recumbent folding of dolomite beds, Eileach an Naoimh. Note figure (circled) for scale; B) megaclast of breccia in dolomitic diamictite, Dun Chonnuill; C) megaclast of bedded coarse sandstone at hammer (circled) in dolomitic diamictite, A' Chuli.

with conglomerates and sandstones and is overlain by the Disrupted Beds (Figs. 3.3 and 3.4).

The Great Breccia was mapped by Spencer (1971) as a distinctive lithologic unit as it differs from other diamictite units within the Port Askaig Formation in a number of ways. First, at between 30-50 m in stratigraphic thickness, the Great Breccia is much thicker than most of the 46 other diamictite beds Spencer (1971) identified in the Port Askaig Formation, which are generally less than 20 metres thick (e.g. Fig. 3.4). Second, the Great Breccia contains clasts in excess of 100 metres in long dimension; the majority of clasts in other diamictites are less than a metre across (Spencer, 1971; Eyles, 1988). In addition, some of the large clasts in the Great Breccia exhibit brittle and ductile deformation of internal structure, whereas most clasts in other diamictites are undeformed.

The regional extent of the Great Breccia is uncertain as it can only be clearly identified in exposures on the Garvellach Islands. Spencer (1971) identified a diamictite unit as the Great Breccia on Islay based on its stratigraphic position relative to the overlying Disrupted Beds. However, this diamictite is much thinner (4 m thick) and contains much smaller clasts (less than several metres across). The analysis presented here is restricted to the outcrops on the Garvellach Islands and focuses on the excellent exposures in the cliffs and wave-cut platforms of A'Chuli and Eileach an Naoimh (Figs. 3.1 and 3.5).

Detailed facies analysis and mapping of the Great Breccia on all four Garvellach Islands has revealed that it consists of three separate sedimentary units (Fig. 3.5). Unit 1 is a megabreccia comprising diamictite with matrix-supported clasts up to 100's of metres across, Unit 2 consists of matrix-supported diamictite interbedded with other sedimentary facies such as conglomerate, sandstone and mudstone, and Unit 3 is a diamictite containing size-sorted, matrix-supported, dolomite clasts. Unit 1 is distinguished from Unit 2 by the large clast sizes in Unit 1 and the presence of bedding within Unit 2. Unit 3 is distinguished from the other two units by the predominance of metre-sized dolomitic clasts and its stratigraphic position at the top of the Great Breccia (Fig. 3.5). The characteristics of each of these sedimentary units within the Great Breccia are described in detail below and are used to interpret the depositional origin of this distinctive deposit.

3.3.1. Unit 1: Megabreccia

The megabreccia unit is exposed on all four Garvellach Islands, and consists of a 30-50 m thick, massive to stratified, diamictite containing abundant matrix-supported dolomitic clasts in a dolomitic matrix (Figs. 3.5 and 3.6). Clasts within the megabreccia range in size from several centimetres to 100's of metres across. The most spectacular clast contained within the megabreccia unit is a large folded dolomite block exposed in the cliffs of Eileach an Naoimh ('the Bubble'; Fig. 3.6A). This clast consists of stacked beds of massive and horizontally laminated dolomite, which are deformed into a recumbent fold (Fig. 3.6A) closing towards the NW (Kilburn *et al.*, 1965; Spencer, 1971).

Clasts up to several metres across within Unit 1 are predominantly dolomitic, whereas clasts ranging in size from several metres to 100's of metres across (referred to as megaclasts) consist of a variety of carbonate-rich siliciclastic facies (Table 3.1). These carbonate-rich facies include deformed or bedded sandstone, conglomerate, breccia, interbedded siltstone and sandstone, and diamictites distinguished from the host diamictite by different matrix texture and/or clast size and abundance (Table 3.1, Fig. 3.6). Megaclasts within the megabreccia unit are variably consolidated and exhibit a range of brittle, plastic and soft sediment deformation, whereas smaller dolomitic clasts are massive and undeformed. Outer margins of megaclasts are predominantly sharp and well defined by differences in texture between the megaclast and the hosting diamictite. Megaclasts are randomly distributed and there does not appear to be any spatial organization according to lithology, size, clast orientation or nature of deformation.

The extent of deformation within the megaclasts does vary according to lithology; deformation is not observed in megaclasts of massive diamictite, conglomerate or breccia but is noticeable in sandstone, dolomite and interbedded silt and sandstone megaclasts (Table 3.1). Deformation most commonly takes the form of broad folding of internal structure within a megaclast over a distance of 10 to 100 metres (e.g. Fig. 3.6A), or of smaller scale (<10 cm) soft sediment deformation or contortion of the outer margin of megaclasts.

An interbedded siltstone and sandstone megaclast on A' Chuli (Fig. 3.5) shows extremely complex internal deformation, which ranges from undeformed to broadly and

Table 3.1: Characteristics of large clasts within the megabreccia (Unit 1), Great Breccia. DC, Dun Chonnuill; GE, Garbh Eileach; AC, A' Chuli; EN, Eileach an Naoimh.

Size class	Island	Clast types	Matrix between clasts	Deformation
several m's (Fig. 3.6B and 3.6C)	DC	Diamictite, breccia	Coarse- and fine-grained sandstone	Not apparent
	GE	Sandstone, dolomitic sandstone with clasts	Coarse sandstone	Crudely contorted and folded on < 1 m scale.
	AC	Very coarse dolomitic sandstone, diamictite	Sandy silt	Contorted margins and broad folding of sandstone megaclast; no apparent deformation of diamictite megaclast.
10s of metres	EN	Diamictite, diamictite overlain by laminated mudstone and sandstone, bedded sandstone	Not apparent. Megaclasts are in contact with each other; upper contacts are infilled by overlying diamictite of Unit 2.	Not apparent in diamictite; broadly folded and cm-scale contorted margins and broad folding of bedded sandstone megaclast.
100s of metres (Fig. 3.6A)	AC	Interbedded dolomitic siltstone and sandstone	Siltstone	Spatially highly variable in degree, lateral extent and nature of deformation; faulted, contorted, folded, and convoluted bedding.
	EN	Stacked beds of finely laminated to massive dolomite	Siltstone	Recumbent folding on 50- 100 m scale.

complexly folded, to highly contorted, convoluted and brecciated; in areas of intense deformation, bedding is completely disaggregated into folded or angular blocks within a chaotically folded fine-grained matrix. Both brittle and ductile deformation is evident within this megaclast over a variety of scales ranging from several centimetres to over 100 metres.

Megaclasts of similar scale have been reported from modern subaqueous environments characterized by sediment instability, such as Lake Tahoe (Gardner *et al.*, 2000) and Kitimat Arm, a fjord in British Columbia (Prior *et al.*, 1982), as well as in carbonate megabreccias found in both modern (e.g. Hine *et al.*, 1992) and ancient sedimentary successions (see review and references in Spence and Tucker, 1997; and Payros *et al.*, 1999). These megaclasts are contained in muddy matrix materials (Hubert *et al.*, 1977; Payros *et al.*, 1999) or poorly sorted coarse-grained deposits (Cook *et al.*, 1972; Srivastava *et al.*, 1972; Davies, 1977) and have been interpreted as the product of catastrophic mass failure of variably consolidated sediments and entrainment of large sediment blocks into debris flows. A similar mass flow origin is proposed for the megabreccia unit of the Great Breccia.

The largest clasts within the megabreccia unit consist of a variety of carbonate-rich siliciclastic facies (Table 3.1) and indicate catastrophic failure of a succession of interbedded dolomitic diamictite, and carbonate-rich conglomerate, breccia, sandstone and siltstone. The absence of organization or grading of clasts within the megabreccia may be the result of limited travel distance or transport by a flow with insufficient grain interaction or turbulence to allow organization or sorting (Cook *et al.*, 1972; Nardin *et al.*, 1979). The facies contained within megaclasts of the Great Breccia are consistent with those found in the underlying beds of the Port Askaig Formation and the upper part of the Islay Formation, indicating an intrabasinal source of sediment (Kilburn *et al.*, 1965). The direction of overturn in the folded megaclast of dolomite on Eileach an Naoimh suggests northwest movement (Kilburn *et al.*, 1965) and a possible southeast sediment source for the megabreccia.

Soft sediment deformation of megaclast outer margins and internal structure indicates a variably consolidated sediment source and deformation during transport. Source materials for the Great Breccia (i.e. the Islay Formation and the lowermost part of the Port Askaig Formation) are relatively undeformed when observed in their original stratigraphic positions, suggesting that deformation of the megaclasts occurred during or immediately after transport. Variability in the degree of megaclast deformation appears to be related to lithology and may also have been affected by the shear strength of the original materials. Some of the original sediments that formed the megaclasts must have been sufficiently ductile to fold and deform plastically during downslope transport, whereas others were sufficiently lithified to undergo brittle deformation or remain undeformed.

Catastrophic failure of sediments on the scale required to create the megabreccia (Unit 1) was most likely associated with large scale fault movement. The intrabasinal nature of the sediment and inclusion of megaclasts composed of Islay Formation sediments, which, in the Garvellach Islands, lie stratigraphically 95 m below the Great Breccia are consistent with substantial fault movement. The restriction of the Great Breccia to outcrops on the Garvellach Islands and Islay also suggests a localized, fault controlled origin for the megabreccia.

3.3.2. Unit 2: Diamictite with interbeds

Unit 2 of the Great Breccia consists of stacked, lenticular diamictite units interbedded with siltstone-rich facies, sandstone and conglomerate (Figs. 3.5, 3.7, 3.8 and 3.9). Outcrop exposures are restricted to a raised wave-cut platform on Eileach an Naoimh; Unit 2 lies adjacent to Unit 1 along strike, but also partially overlies megaclasts of Unit 1 (Figs. 3.5 and 3.7). Unit 2 diamictites are laterally continuous for up to 100 m along strike and several 10's of metres in thickness. Diamictites contain dolomitic matrix-supported clasts up to 40 cm across in a dolomitic silty sandstone matrix (Fig. 3.8A). Most diamictite units show chaotic internal structure defined by folded and contorted discontinuous muddy sandstone beds or siltstone stringers, and discontinuous clast-rich lenses (Fig. 3.7B). Lower contacts of diamictites in Unit 2 are commonly erosional and show soft sediment deformation (loading) and evidence of incorporation of underlying sediment. Upper contacts of these diamictites with interbedded siltstone, sandstone, and conglomerate facies are sharp to gradational and relatively planar with less than a metre of relief.

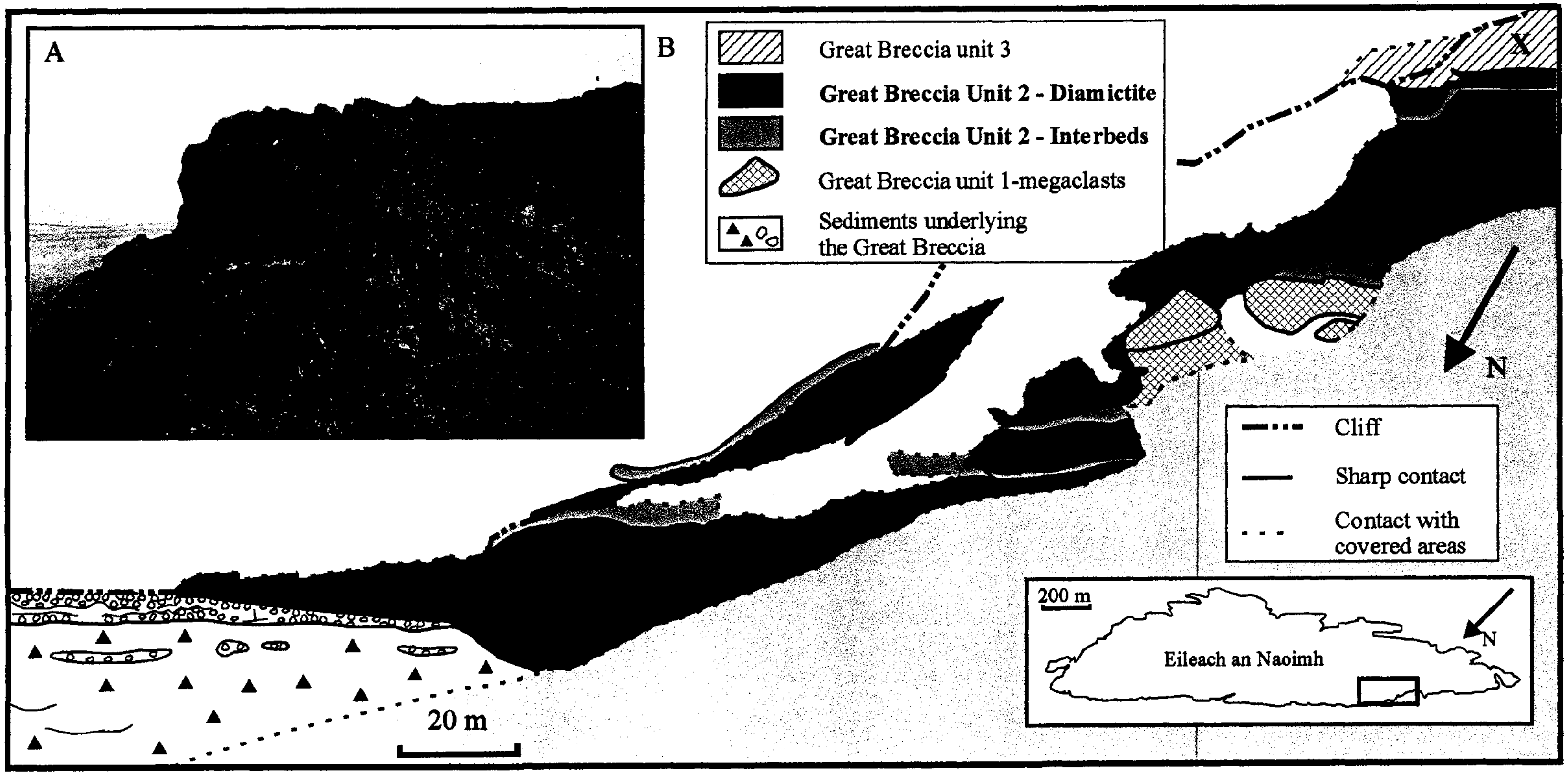


Fig. 3.7: Great Breccia, Unit 2; A) Raised wave-cut platform and cliff on northwest shore of Eileach an Naoimh where Unit 2 was mapped in detail; photograph taken looking north. B) facies map of wave-cut platform; inset shows the location of the mapped area on Eileach an Naoimh. Diamictites of Unit 2 are either massive or stratified. Large clast circled in (A) is found within unit 3 diamictite labelled with an 'X' in (B).

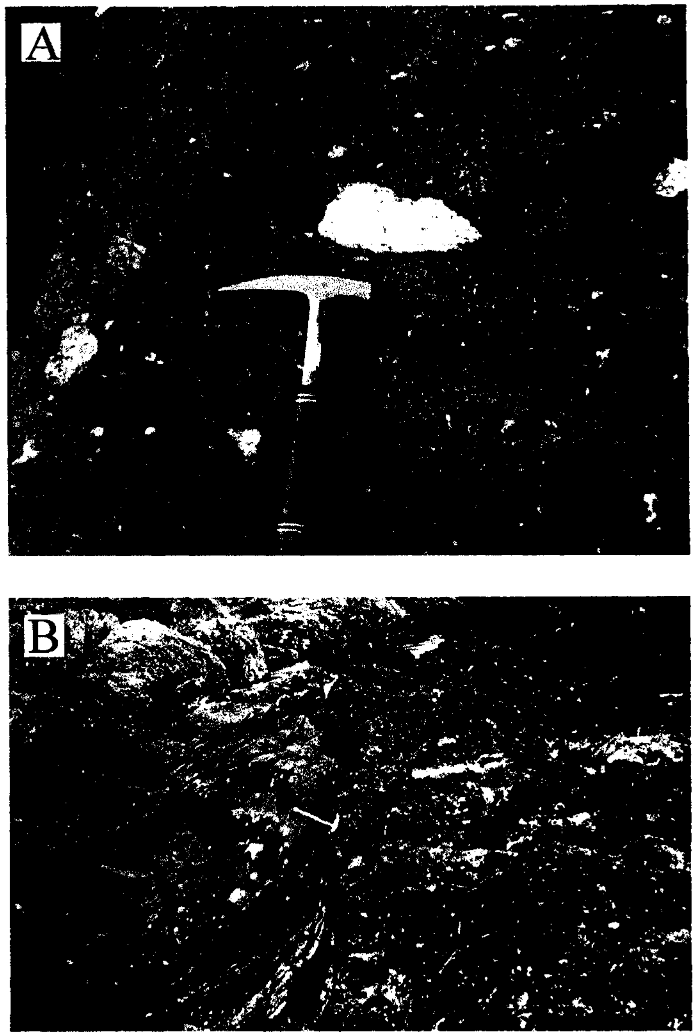
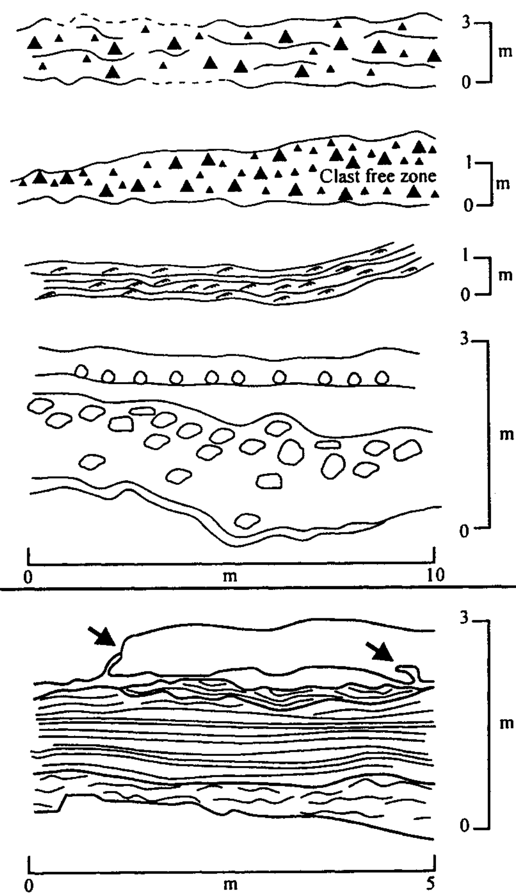


Fig. 3.8: Great Breccia, Unit 2; Dolomitic diamictite with interbeds, Eileach an Naoimh. A) Dolomitic diamictite of Unit 2. Clasts are angular to subrounded, predominantly dolomitic and up to 25 cm across and are supported by a sandy silt dolomitic matrix. B) boundary between two diamictite units (one to left of hammer, the other to right of hammer) with thin interbed of siltstone-rich diamictite at the position of the hammer. Diamictite to the left is stratified, matrix supported, contains predominantly dolomitic clasts less than 2 cm in a dolomitic matrix and contains some rafts of underlying siltstone-rich diamictite. Diamictite to the right is crudely stratified, matrix-supported and contains predominantly dolomitic clasts less than 10 cm across in a dolomitic matrix. Regional dip 35 ° from the lower right to the upper left.



	FACIES	THICKNESS
A	Stratified siltstone-rich diamictite with rare exotic clast	70 cm- 3 m
B	Stratified diamictite with clast free zones and localized inverse coarse tail grading	75 cm- 1.5 m
C	Rippled sandstone, mm-cm scale laminations	80 cm - 1 m
D	Massive sandstone, clast-rich base	0 - 70 cm
	Massive sandstone	0 cm- 1 m
	Conglomerate, inverse coarse tail grading	<1 m- 5 m
	Massive sandstone, silty	0 - 10 cm
	Overall	1-5 m
E	Massive sandstone	0 - 1 m
	Massive mudstone	10 - 60 cm
	Laminated and deformed sandstone	0 - 50 cm
	Laminated sandstone, clast layers	60 cm - 2 m, mm scale laminations
	Deformed medium sandstone beds in deformed very fine silty sandstone	60 cm - 1 m, beds within 2 - 15 cm

Fig. 3.9: Schematic diagrams showing characteristics of interbeds within Unit 2 of the Great Breccia. Note scale differences. Some interbed types exhibit significant lateral thickness changes. Flame structures and soft sediment deformation (arrowed) are found at the contact between siltstone and massive sandstone interbeds in (E).

Diamictites of Unit 2 are interbedded with siltstone-rich facies, carbonate-rich sandstone and carbonate-rich conglomerate (Figs. 3.7B and 3.9). Interbeds range from several centimetres to several metres in thickness and are exposed laterally over 10's of metres of outcrop (eg. Figs. 3.7B and 3.9). Siltstone-rich facies such as laminated siltstone and sandstone with clast layers and stratified siltstone-rich diamictites have sharp to erosional lower contacts. Medium- to coarse-grained sandstone interbeds are commonly massive and either load into underlying siltstones or infill irregular substrate topography. Conglomerate interbeds up to 5 m in thickness contain abundant dolomitic clasts up to 50 cm across. Loaded lower contacts, clast-rich bases and inverse coarse tail grading were observed in both sandstone and conglomerate interbeds.

Unit 2 of the Great Breccia is interpreted to have formed by successive debris flows (diamictites) and intervening turbidity flows (interbeds). The erosional and loaded lower contacts of diamictite units, their chaotic internal structure, and their association with other facies interpreted as mass flow deposits suggests an origin as subaqueous debris flows. Interbeds consisting of laminated sandstone and siltstone with clast layers, and massive and graded sandstone, are interpreted as turbidites (Walker, 1984). Conglomerate was deposited by high concentration turbulent flows, characterized by locally significant dispersive pressures as indicated by the presence of inverse coarse tail grading (Nardin *et al.,* 1979; Nemec *et al.*, 1984). This suggests a depositional environment in which a variety of sediment types were available for downslope transport by processes ranging from debris flows to high concentration turbidity currents. Unit 2 deposits have limited lateral extent and appear to have infilled the irregular topography on the surface of the underlying megabreccia (Unit 1).

3.3.3. Unit 3: Diamictite with large dolomite clasts

Unit 3 is found at the top of the Great Breccia overlying both Units 1 and 2 and can be identified on all four Garvellach Islands (Fig. 3.5). It consists of stratified matrix-supported diamictite containing abundant subangular to subrounded dolomite clasts in a poorly sorted dolomitic matrix. This unit is differentiated from the other two units of the Great Breccia by the abundance of dolomite clasts that are restricted in size from 50 cm to 3 m (Fig. 3.10). Most clasts are internally massive, but some exhibit contorted and disrupted internal laminations of syndepositional origin. Unit 3 also contains highly deformed clasts of siltstone and muddy sandstone (approx. 50 cm across), which impart a chaotic internal structure to the matrix. The matrix of Unit 3 varies from sandy siltstone to very coarse silty sandstone and from carbonate to mixed carbonate and siliciclastic in lithology. The lower contact of Unit 3, where exposed, is gradational to poorly defined. The occurrence of Unit 3 on all four islands (Fig. 3.5) indicates that it forms a laterally-extensive 'blanket'.

Unit 3 is also interpreted as a debris flow deposit; a mass flow origin is supported by the poor organization and chaotic internal structure of the diamictite and the close association of Unit 3 with the megabreccia (Unit 1). The lack of grading and poorly sorted nature of this unit suggests limited grain interaction and fluid turbulence within the

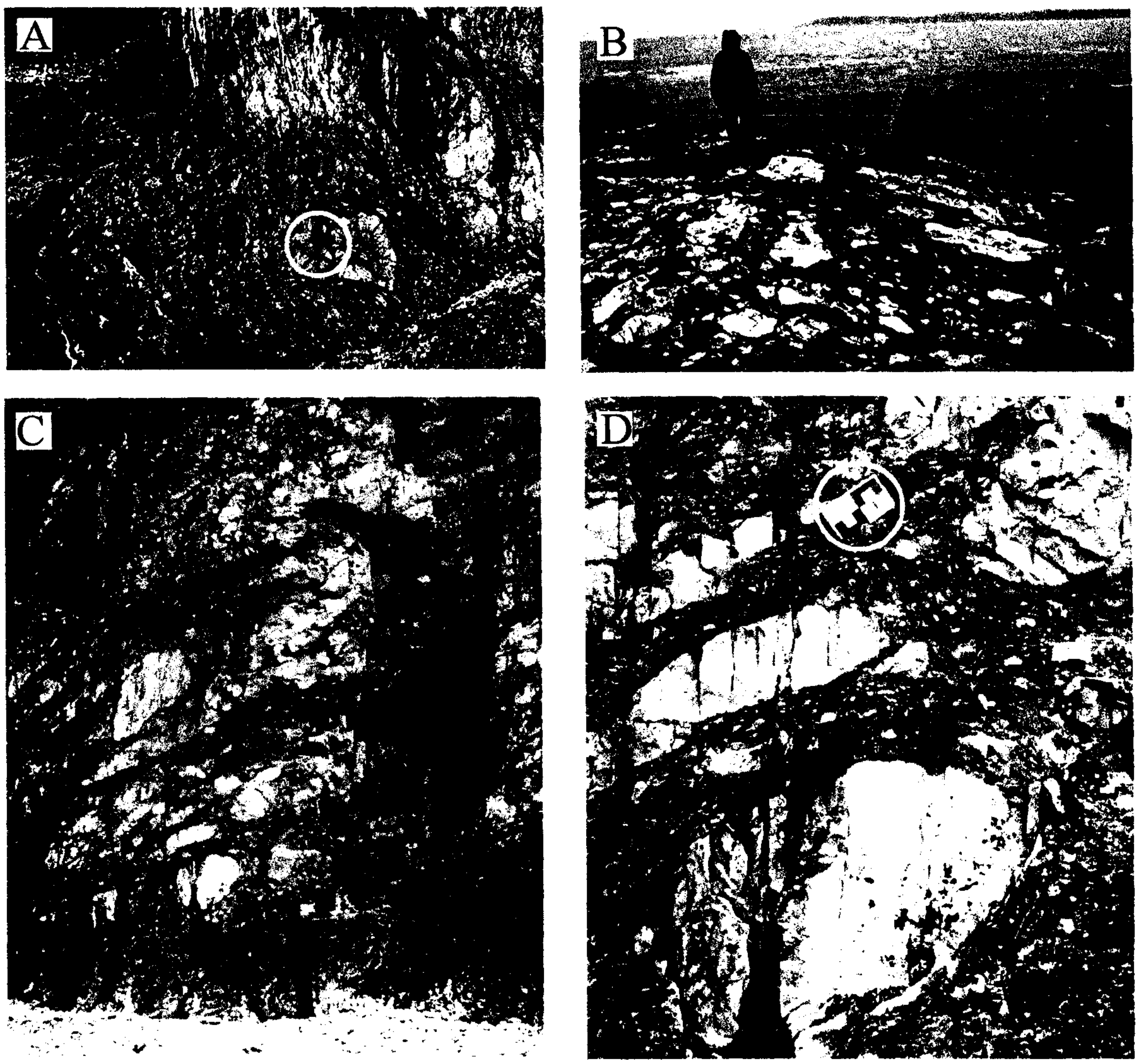


Fig. 3.10: Great Breccia, Unit 3; Dolomitic diamictite with size-sorted dolomite clasts. A) Large dolomite clast next to figure (circled) is approximately 3 metres across, Eileach an Naoimh; B) Unit 3 on A' Chuli; diamictite is clast-supported with rounded to subrounded clasts ranging from 3 cm to over 1 m; C) Cliff section on A' Chuli, general view. Note the large prominent dolomitic clasts to the left above the person; D) Cliff section on A' Chuli, close up. Note fieldbook (18 cm across-circled) for scale, abundant rounded to subangular dolomitic clasts (up to 1 m) in a coarse-grained dolomitic sandstone matrix.

flow responsible for its deposition (Nardin *et al.*, 1979). The predominance of large clasts of a specific size range (50 cm to 3 m), included in a poorly sorted matrix, suggests size sorting of clasts occurred at source prior to downslope transport. This sorting may be related to lithologically-controlled mechanical disintegration of the source dolomites. The lateral extent of Unit 3 contrasts with the restricted lateral continuity of Unit 2 and indicates that the debris flow responsible for this depositional event was relatively more mobile and/or larger than the ones responsible for the deposition of Unit 2.

In sum, all three subunits identified within the Great Breccia indicate deposition by subaqueous sediment gravity flow processes. Sedimentological differences between the three units reflect differences in the scale of the mass failure events, the nature of the materials being transported and sediment support mechanisms.

3.3.4. Depositional model for the Great Breccia

The sedimentary characteristics of the Great Breccia are very similar to carbonate megabreccias described in ancient carbonate slope deposits. Eocene megabreccias found within the SW Pyrenean Basin have been interpreted as the product of a single 'megaturbidite' (Johns *et al.*, 1981; Labaume *et al.*, 1983, 1987). The 'megaturbidite' model explains the erosional base of megabreccias, and the overlying fining-up succession of megabreccia, matrix-supported debrite, graded sands and fine-grained turbidites as the product of a single, large scale turbidity current. However, Payros *et al.* (1999) offer an alternate model for the development of such successions in their examination of 'megaturbidites' in a different part of the Pyrenean Basin and conclude that the megaturbidites record downslope evolution of a debris flow into a turbulent flow. Evidence used to support this 'flow transformation' model identifies relatively coarse-grained facies deposited by debris flows in proximity to the failed basin margin and finer-grained facies deposited by turbidity currents in more distal parts of the basin (Payros *et al.*, 1999).

The vertical succession of facies (Units 1, 2, 3) within the Great Breccia are interpreted here to represent deposition from a series of separate flows rather than from a single megaturbidite or flow transformation within one flow. All three units within the Great Breccia, together with the overlying coarse-grained facies (see column B in Fig. 3.4), more closely resemble deposits formed as 'composite-graded sequences' similar to those found within the Cambrian-Ordovician Cow Head Breccia in Newfoundland (Hubert *et al.*, 1977). 'Composite graded sequences' consist of stacked units of thinning- and fining-upward diamictites and are interpreted as the product of successive flows that record backstepping erosion of a failing basin margin. A similar model is proposed here for the Great Breccia of the Port Askaig Formation.

Lateral and vertical facies relationships between the three units identified within the Great Breccia suggest the following sequence of events (Fig. 3.11). Rapid subsidence along a fault in the basin created a scarp that exposed sediments comprising the lower section of the Port Askaig Formation and the underlying Islay Formation. Catastrophic mass failure of unstable sediments exposed along this fault scarp led to the formation of

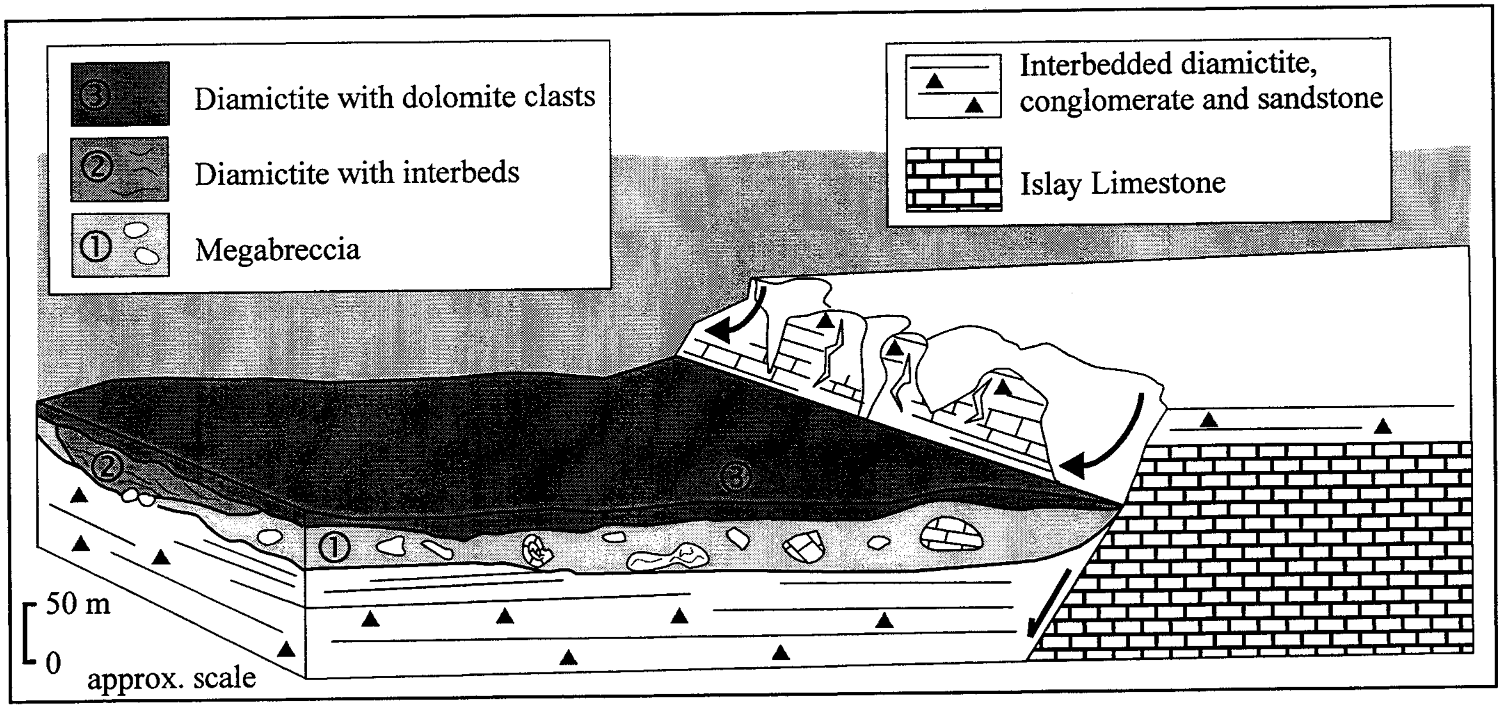


Fig. 3.11: Schematic block diagram showing depositional setting of the Great Breccia. See text for discussion. Curved arrows represent direction of failure and mass movement, whereas straight arrow represents downthrow of faulting. Relative sea level is shown to emphasize subaqueous nature of the Great Breccia; position of sea level relative to footwall is unknown.

the megabreccia (Unit 1), which because of the size of the clasts it contains, had an irregular upper surface topography. Repeated retrogressive failure of the steep and unstable scarp following the initial large scale failure event produced a series of debris flows and turbidites that infilled relief created by the megabreccia (Unit 2). Unit 3 sediments are laterally traceable over the entire study area and represent a subsequent pulse of instability and mass flow, predominantly of Islay Formation dolomite exposed along the fault scarp. The Great Breccia thus records creation of and initial catastrophic failure along a fault scarp followed by retrogressive failure of unstable sediments exposed along the scarp.

3.4. STRATIGRAPHIC CONTEXT OF THE GREAT BRECCIA

In order to determine the broader paleoenvironmental setting of the Great Breccia and establish potential controls on its development, the sedimentology and depositional origin of facies both under- and overlying the Great Breccia are discussed below. The Great Breccia occurs within Member I of the Port Askaig Formation (Spencer, 1971). The lower section of Member I is characterized by coarse-grained sediments, which include the Great Breccia, whereas the upper section of Member I is characterized by relatively fine-grained sediments, including a distinctive 25 metre thick unit known as the 'Disrupted Beds' (Spencer, 1971; Fig. 3.4, 3.12, 3.13). Both coarse-grained and fine-grained facies lying above and below the Great Breccia occur as relatively planar tabular units that can be traced over several kilometres along strike on the Garvellach Islands.

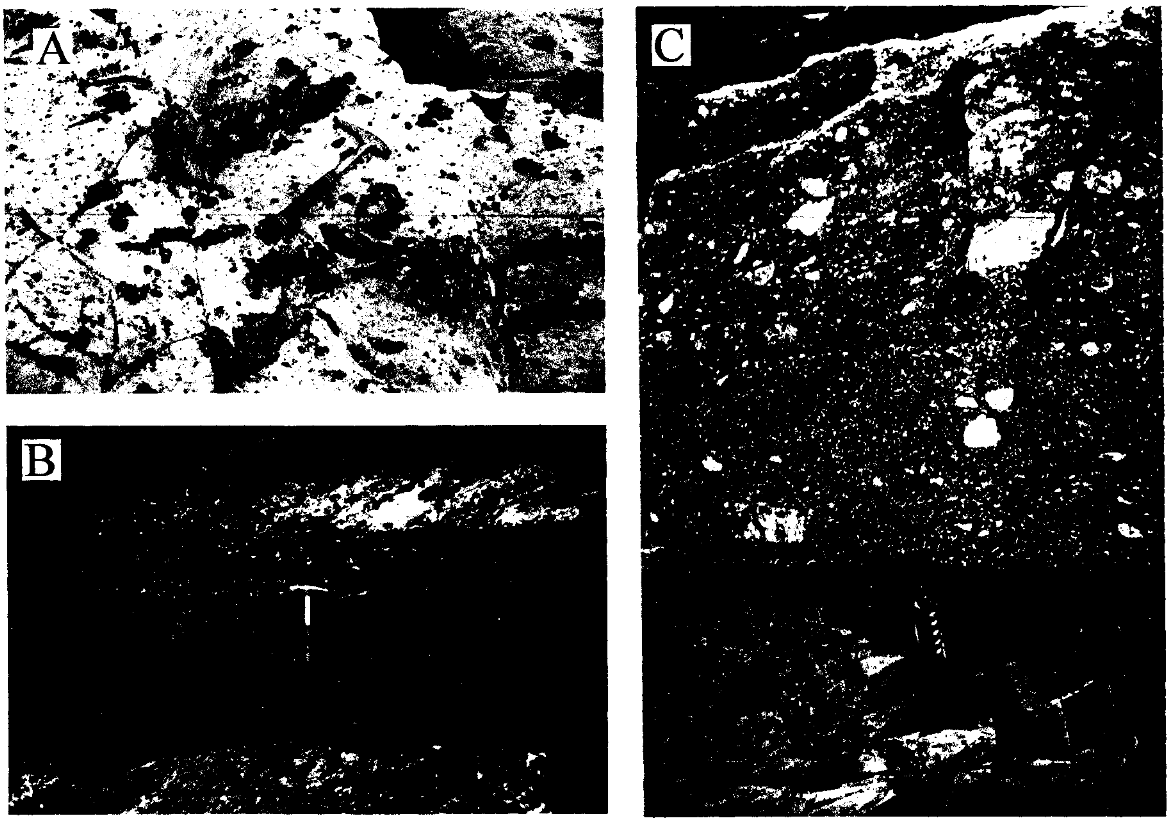


Fig. 3.12: Member I, Port Askaig Formation; coarse-grained facies underlying the Great Breccia (see Fig. 3.4). A) dolomitic diamictite (D1), Dun Chonnuill. Diamictite is massive, matrix- supported with abundant angular to subrounded clasts up to 30 cm and a siltstone matrix; B) silty sandstone interbedded with pebble layers above diamictite D8, Garbh Eileach. Sandstone beds are less than 5 cm in thickness and either horizontally-laminated, massive, rippled or graded. Pebble layers increase in abundance and thickness (from < 1 cm to < 10 cm) upsection; C) crude coarse tail inverse grading in conglomerate interbed above D2, Garbh Eileach. Conglomerate consists of abundant granules and subangular to subrounded, predominantly dolomitic clasts less than 5 cm. Contact with the underlying diamictite is sharp and/or loaded with up to 70 cm relief.

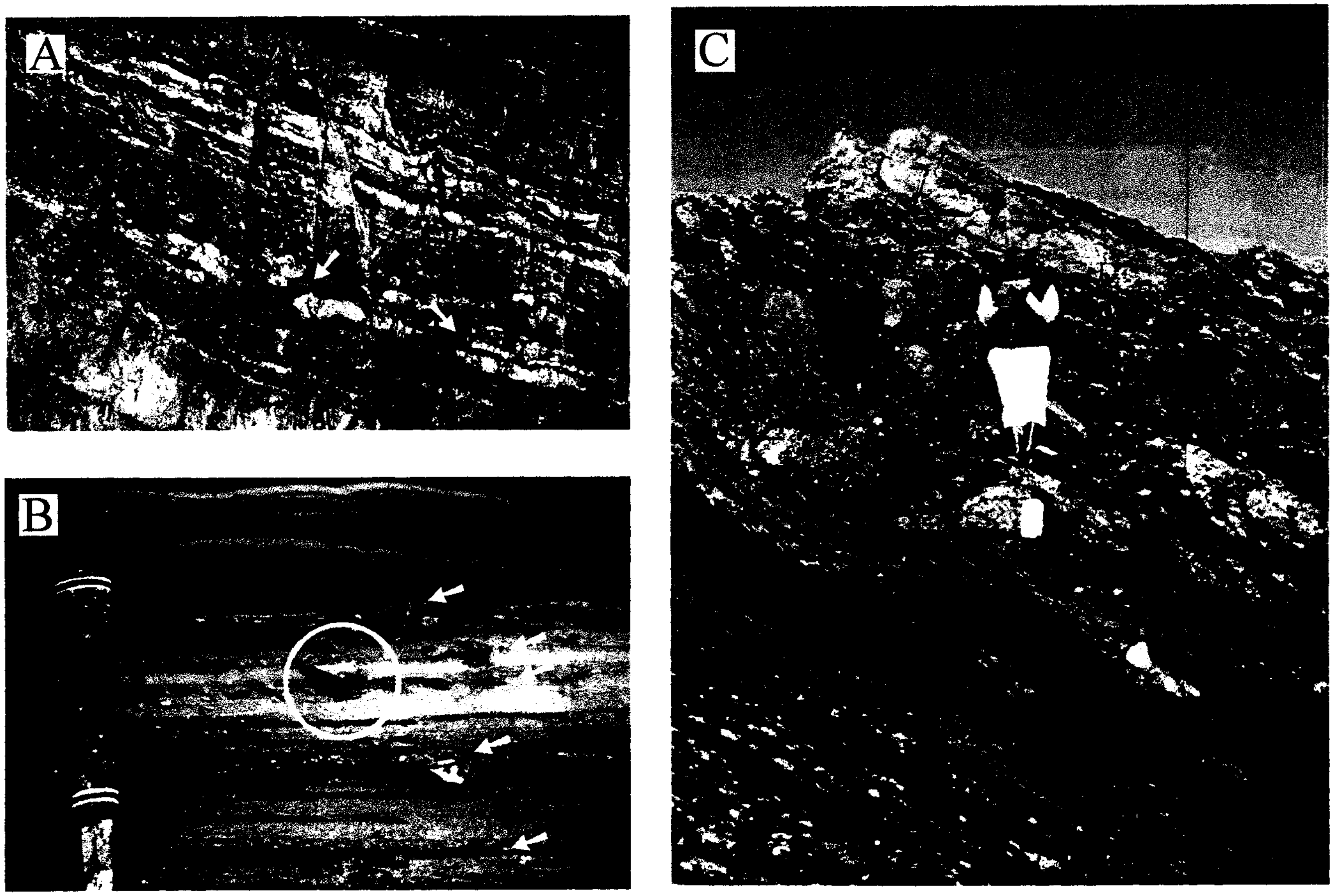


Fig. 3.13: Member I, Port Askaig Formation; the Disrupted Beds. A) General view of cliff section approximately 20 metres in thickness, Garbh Eileach, showing the sedimentary boudinage of light coloured dolomitic sandstone beds (arrowed) interbedded with dark coloured siltstone. Stratigraphic top to upper right; B) Clast layers (arrowed) in finely-laminated siltstone with outsized clast (possibly a dropstone-circled), A' Chuli; C) chaotically deformed diamictite, conglomerate, sandstone and siltstone, A' Chuli. Regional dip to the lower right at 35°; Note the relatively undeformed beds at the bottom of the photograph.

On a more regional scale, only the upper section of Member I above the Great Breccia can be correlated to outcrops outside of the Garvellach Islands: the lower part of Member I appears to be restricted to the Garvellach Islands (Spencer, 1971).

3.4.1. Sediments underlying the Great Breccia

A thick succession of dolomitic diamictites interbedded with carbonate-rich conglomerates and carbonate-rich sandstones underlies the Great Breccia on the Garvellach Islands (see column A in Fig. 3.4). Diamictite beds are each up to 12 m thick and are predominantly massive with dolomitic sandy siltstone matrix and common to abundant matrix-supported dolomitic clasts up to 70 cm across (Fig. 3.12A). Sandstone interbeds up to several metres in thickness are predominantly massive and/or deformed, although some are graded, horizontally-laminated, or rippled (Fig. 3.12B). One isolated outsized clast was found in a horizontally-laminated sandstone unit. Conglomerate interbeds are up to several metres thick and are massive to stratified, showing localized coarse tail normal and inverse grading (Fig. 3.12C).

Diamictites (D1-D12, Fig. 3.4) are interpreted as the product of glacially-influenced sediment gravity flow processes and 'rainout' of fine-grained sediments and ice rafted debris in a glaciomarine setting (Eyles, 1988). Other facies found underlying the Great Breccia within Member I show sedimentary characteristics consistent with an origin by subaqueous sediment gravity flow processes ranging from turbidites (massive and graded sandstones), to high-concentration turbulent flows (conglomerates).

Localized coarse-tail grading within coarse-grained facies indicate significant turbulence (normal grading) and dispersive pressures (inverse grading) operated within these high-concentration turbulent flows (Walker, 1975; Middleton and Hampton, 1976). Rippled sands observed in some places are indicative of deposition either by traction currents or by turbidity currents. In sum, sedimentation prior to the deposition of the Great Breccia was dominated by subaqueous sediment gravity flow processes that remobilized coarse-grained sediments (diamictites, conglomerates and sandstones).

3.4.2. Sediments overlying the Great Breccia

Facies overlying the Great Breccia in Member I of the Port Askaig Formation include dolomitic diamictites, carbonate-rich and calcareous conglomerates, carbonate-rich and calcareous sandstones, laminated siltstones and a distinctive lithologic unit, the Disrupted Beds (Figs. 3.3 and 3.4; Spencer, 1971). The diamictite, conglomerate and sandstone facies are similar to those found below the Great Breccia and are interpreted as subaqueous sediment gravity flow deposits. Finely laminated siltstones contain isolated clasts and common clast horizons and are interpreted as turbidites. The occurrence of rare isolated outsized clasts indicates the presence of floating ice in the basin.

The Disrupted Beds are distinguished from other units in the Port Askaig Formation by their dark blue siltstone matrix and by the presence of pervasive soft sediment deformation features (hence the name 'Disrupted Beds'). Facies within the Disrupted Beds include relatively planar but discontinuous carbonate-rich sandstone beds, finely-laminated mudstone with clast horizons, carbonate-rich conglomerate and matrix-supported diamictite (dolomitic clasts in a siltstone matrix; Fig. 3.13). Sandstone beds show evidence of sedimentary boudinage (Fig. 3.13A), small scale soft sediment deformation (up to 5 cm) and folding (up to 1 m), whereas coarse-grained facies have chaotic internal structure created by spatially variable clast content and matrix texture (Fig. 3.13C). Lateral and vertical facies changes from deformed carbonate-rich sandstones to stratified diamictite with chaotic internal structure were observed over outcrop lengths of several metres to 10's of metres. Almost all sedimentary facies within the Disrupted Beds show evidence of soft sediment deformation but to varying degrees and on scales ranging from centimetres to several 10's of metres.

Alternating units of deformed and undeformed sediments within the Disrupted Beds indicate that disturbance was synsedimentary. Sedimentary boudinage of relatively planar, but discontinuous beds of sandstone has been interpreted as the product of tensional stresses and associated lateral mass movement (Kilburn *et al.*, 1965; Spencer, 1971). Vertical and lateral facies changes from boudinaged sandstone beds, conglomerate and siltstone to chaotically stratified and massive diamictite are interpreted to represent increasing amounts of deformation and disruption by downslope movement of originally interbedded sandstone, conglomerate and siltstone. The Disrupted Beds therefore indicate continued slope instability following the deposition of the Great Breccia.

Sandstone beds showing boudinage structures similar to those found in the Disrupted Beds are reported in deposits associated with a Cambrian-Ordovician carbonate megabreccia in Newfoundland (Hubert *et al.*, 1977); these features are also thought to indicate continued slope instability following an initial massive failure that produced the megabreccia. The relatively fine-grained nature of the Disrupted Beds and other sediments overlying the Great Breccia suggest an increase in depositional water depths. Although other factors contributing to relative sea level change cannot be ruled out, the close stratigraphic relationship between the Great Breccia and associated coarse-grained facies, and the immediately overlying fine-grained sediments of the Disrupted Beds suggests that basin deepening was probably related to the creation of a fault scarp.

3.5. DISCUSSION

3.5.1. Sedimentological evidence for tectonic activity

The sedimentological characteristics of the Great Breccia indicate considerable localized sediment instability in this part of the Dalradian basin during Islay Subgroup times. Catastrophic slope failure recorded by the Great Breccia was most likely associated with localized faulting. The sedimentary boudinage and deformation, characteristic of the Disrupted Beds overlying the Great Breccia, have been attributed to extensional activity in other settings (e.g. Bose *et al.*, 1997; Calvo *et al.*, 1998) and may also record continued tectonically-controlled instability within the Dalradian basin. This localized sediment instability and the absence of over 90 m of Port Askaig sediments at

Islay is consistent with an extensional tectonic setting characterized by fault-bounded blocks and sub-basins (Leeder and Gawthorpe, 1987; Frostick and Steel, 1993; Ravnås and Steel, 1998); a setting previously proposed for this area (Harris *et al.*, 1978; Anderton, 1982, 1985, 1988; Harris *et al.*, 1993).

3.5.2. Tectonic versus climatic controls on sedimentation

The thick succession of sediment gravity flow facies found in the lower part of Member I of the Port Askaig Formation (D1-D12, Fig. 3.4) shows neither a fining- or coarsening-upward trend; this suggests that depositional environments remained relatively constant through time and a balance was achieved between the generation of accommodation space and the supply of sediments infilling the basin. The change to predominantly fine-grained sediments above the Great Breccia reflects an increase in accommodation space that temporarily outpaced sediment supply.

This balance between accommodation space and sediment supply may have been influenced by tectonic and/or climatic controls. Fault movement and basin subsidence associated with the formation of the Great Breccia may have lead to sediment starvation due to changes in water depth and sediment source (Leeder and Gawthorpe, 1987; Frostick and Steel, 1993; Ravnås and Steel, 1998). Eustatic sea level changes may have also been superimposed but are impossible to discriminate.

The coarse-grained nature of facies contained within Member I contrasts with those of other carbonate megabreccia successions where megabreccias are under- and overlain by fine-grained turbidites and basinal muds (Srivastava *et al.*, 1972; Cook *et al.*, 1972; Davies, 1977; Labaume *et al.*, 1987; Surlyk and Ineson, 1992; Payros *et al.*, 1999). Glaciation of the basin margin may account for the abundance and continued supply of coarse-grained, poorly sorted sediment during accumulation of Member I of the Port Askaig Formation. The presence of some dropstones in sandstone and fine-grained facies overlying the Great Breccia suggests that floating ice was present in the basin during this time. However, there is no evidence of direct glacial deposition prior to, or following, the deposition of the Great Breccia, such that a glacial influence on relative sea level change is difficult to demonstrate.

3.6. CONCLUSIONS

Detailed sedimentological analysis of the Great Breccia on the Garvellach Islands shows that it consists of three distinct sedimentary units: a megabreccia (Unit 1), a diamictite with interbeds (Unit 2) and a diamictite with metre-sized dolomite clasts (Unit 3). The characteristics and spatial organisation of these units suggests that they record initial catastrophic failure of a succession of calcareous siliciclastic sediments (Unit 1), followed by a series of debris flows and high concentration turbulent flows (Units 2 and 3) generated by backstripping of an active fault scarp. The initial catastrophic mass failure responsible for deposition of the megabreccia (Unit 1) was probably triggered by creation of a steep fault scarp. A mass failure depositional origin for the Great Breccia is very different to that made by early workers who ascribed it a subglacial origin (Kilburn *et al*., 1965; Spencer, 1971) but agrees with later interpretations of tectonically-active conditions made by Schermerhorn (1974) and Eyles (1988).

Active tectonism in the basin at this time was previously inferred from lateral thickness changes in Islay Subgroup outcrops across Scotland and Ireland (Anderton, 1982, 1985, 1988). Sedimentological evidence from the Great Breccia and the overlying Disrupted Beds provides additional evidence for localized tectonic activity in the basin, at least during the initial accumulation of the Port Askaig Formation. While glaciation on the basin margins may account for the abundance of coarse-grained sediment supplied to the basin, Member I of the Port Askaig Formation is dominated by subaqueous sediment gravity flow deposits and a distinctly glacial component of sedimentation is difficult to identify. This, in turn, makes interpretation of the form and extent of glacial ice in and around the basin at this time extremely difficult.

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CHAPTER FOUR

NEOPROTEROZOIC ENVIRONMENTAL CHANGE IN THE PORT ASKAIG FORMATION, SCOTLAND

4.1. INTRODUCTION

The geological record of the Neoproterozoic contains evidence to suggest significant paleoenvironmental changes occurred during this period that included several glaciations, profound changes in the global carbon cycle and critical steps in the evolution of biota. Neoproterozoic diamictites, poorly-sorted coarse-grained sedimentary rocks commonly interpreted as glacigenic deposits, are found on every continent; some of these glacial deposits are thought to have accumulated in equatorial regions (see reviews in Harland, 1964; Hambrey and Harland, 1985; Evans, 2000). Almost all diamictites are closely associated with carbonates, which exhibit unprecedented negative and positive excursions in $\delta^{13}C$ (Fairchild, 1993; Knoll *et al.*, 1986; Kaufman *et al.*, 1997). In addition, the apparently extreme environmental changes recorded in the sedimentological and geochemical properties of Neoproterozoic rocks are followed by the first appearance of metazoans (see review in Narbonne, 1998).

A variety of hypotheses have been proposed to account for the dramatic paleoenvironmental shifts recorded in Neoproterozoic successions including global glaciation, rapid continental drift, and high obliquity of the ecliptic (see review in Meert and van der Voo, 1994, Hoffman *et al.*, 1998a). The snowball Earth hypothesis is one of the more recent, non-uniformitarian models proposed to explain the Neoproterozoic geological record (Kirschvink, 1992; Hoffman *et al.*, 1998b). This model suggests that the earth was completely ice-covered during the Neoproterozoic for repeated periods of between 4 -30 Ma, with sea ice up to 1 km in thickness. Such global-scale glaciation is argued to explain the development of low latitude glacigenic deposits; rapid climate change in the aftermath of such episodes of severe global glaciation is proposed to explain the paradoxical association of diamictites with carbonates. In addition, the unique geochemical signatures preserved within the carbonates are explained by the geochemical effects of isolation of ocean and atmospheric carbon reservoirs during a snowball Earth and extreme weathering in its aftermath (Hoffman *et al.*, 1998b).

One of the early criticisms of global Neoproterozoic glaciation as it was first proposed by Harland (1964) was that perhaps not all diamictites recorded severely glacial conditions and that basin tectonics may have played a significant role in the development and preservation of diamictite-hosting successions (e.g. Crowell, 1964; Schermerhorn, 1974; Bjørlykke, 1985; Nystuen, 1985; Young, 1988; Eyles, 1993). While diamictites were once largely interpreted as indicators of subglacial depositional conditions, studies have shown they can form as a result of a wide variety of glacial and non-glacial processes (Crowell, 1964; Boulton, 1972; Schermerhorn, 1974; Boulton and Deynoux, 1981; Eyles *et al.*, 1985; Powell and Molnia, 1989; Eyles and Eyles, 1992; Eyles, 1993).

In glacial settings, diamictites can record deposition by subglacial lodgement and deformation processes, by meltout and slumping of supraglacial and proglacial debris at subaqueous or terrestrial ice margins and by slumping and/or 'rainout' of suspended sediments and ice-rafted debris in glacially-influenced marine settings. In non-glacial settings, diamictites have been attributed to both subaerial and subaqueous sediment gravity flows (Nardin *et al.*, 1979; Nemec and Steel, 1984; Pickering, 1989; Nemec, 1990; Eyles *et al.*, 1988). Distinguishing between these different diamictite-forming processes has important implications for reconstructing the nature and severity of past climate change and the distribution of past ice sheets.

Tectonics have also been shown to play a significant role in controlling the nature and preservation of successions in glaciated basins (Schermerhorn, 1974, 1975; Bjørlykke, 1985; Nystuen, 1985; Young, 1988; Eyles, 1993). Diamictite can form as a result of tectonically-generated sediment instability under glacial or non-glacial conditions making differentiation of climatic and tectonic influences on sedimentation patterns difficult (Eyles, 1993; Ravnås and Steel, 1998). Evidence for sea level change in glaciomarine successions, often previously interpreted as reflecting glacial eustacy and isostacy, may in fact be a function of basin tectonics and variable subsidence rates. Furthermore, the preservation of glacial deposits may ultimately be controlled by available accommodation space and basin development (Nystuen, 1985). Many Neoproterozoic glacigenic successions appear to have formed as infills of rift basins associated with break-up of the supercontinent Rodinia (Schermerhorn, 1974; Eyles, 1993; Young, 1995). In order to evaluate the relative importance of tectonics and climate change in influencing deposition and preservation of glacigenic succession, detailed analysis of sedimentary processes and their changes with time must be carried out.

Given the different processes that can result in the formation of diamictites and the complex relationship between tectonic and climatic controls on successions development, diamictite-bearing intervals within Neoproterozoic successions cannot be presumed to represent glacial conditions or that their formation was climatically-controlled. Integrative analysis of sedimentological properties, facies associations and the geometry of deposits, and establishment of an allostratigraphy identifying the nature and severity of environmental changes over time are required to determine the most likely controls on stratigraphy (e.g. Boulton and Deynoux, 1981; Eyles *et al.*, 1985; Hart and Roberts, 1994; Walker, 1992).

The Port Askaig Formation is a thick and well exposed Neoproterozoic glacigenic succession found in the North Atlantic region and has the potential to provide important data regarding the nature and severity of paleoenvironmental change during the Neoproterozoic. Studies of the Port Askaig Formation have to date focused on establishing its regional stratigraphy and the depositional origin of the diamictites (Kilburn *et al.*, 1965; Spencer, 1971; Eyles, 1988). The exceptional exposures of the Port Askaig Formation on the Garvellach Islands of Scotland (Fig. 4.1) provide an excellent opportunity to carry out an allostratigraphic analysis of diamictites and associated facies in order to establish the nature of Neoproterozoic paleoenvironmental change preserved

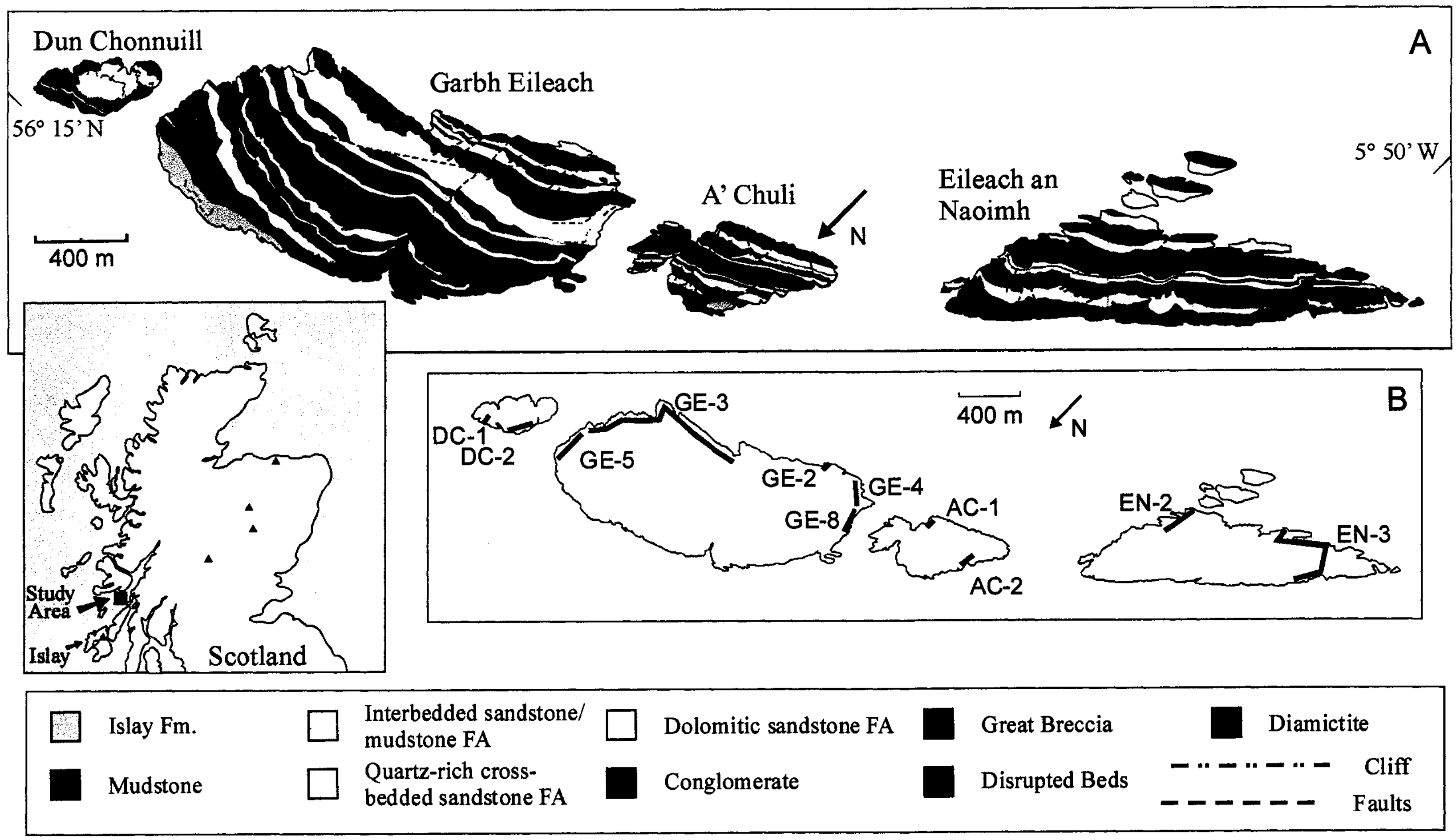


Fig. 4.1: The Garvellach Islands, Scotland. A) geological map showing Islay Formation and various facies types in the Port Askaig Formation (modified from Spencer, 1971); FA- Facies Association; B) map showing sections logged in this study. Triangles on inset map of Scotland show other outcrops of the Port Askaig Formation (modified from Spencer, 1971). The outcrop is covered in the area shaded grey on Dun Chonnuill. Strata are dipping 35° to the S/SE.

in this succession and to evaluate the validity of the snowball Earth hypothesis.

4.2. GEOLOGICAL BACKGROUND

The Port Askaig Formation is a diamictite-bearing unit within the predominantly marine succession of the Dalradian Supergroup (Fig. 4.2, Kilburn *et al.*, 1965; Spencer, 1971). Similar to many other Neoproterozoic successions, it contains and is associated with carbonates; the Port Askaig Formation is underlain by the Islay Limestone and overlain conformably by the mixed carbonate-siliciclastic Bonahaven Formation, both of which record shallow marine sedimentation (Spencer, 1971; Spencer and Spencer, 1972; Fairchild, 1980). The Bonahaven Formation is in turn overlain by a thick succession of quartzites (Jura Quartzites; Fig. 4.2) that record the migration of large tidal bedforms in a shallow marine setting (Anderton, 1976).

The Dalradian Supergroup consists of approximately 25 km of mostly marine sediments and metavolcanics and is subdivided into the Grampian, Appin, Argyll and Southern Highland groups (Harris *et al.*, 1978; Fig. 4.2). It is thought to record sedimentation in an ensialic basin prior to the opening of the proto-Iapetus Ocean (Harris *et al.*, 1978; Anderton, 1982). The onset of extensional tectonic activity is indicated by significant lateral thickness changes in the uppermost part of the Appin Group. This tectonic activity is sustained and intensifies through the Argyll Group as evidenced by rapid thickness changes in stratigraphic units, faulting and volcanic activity (Anderton, 1985). The opening of the Iapetus Ocean is recorded in the area by the formation of the

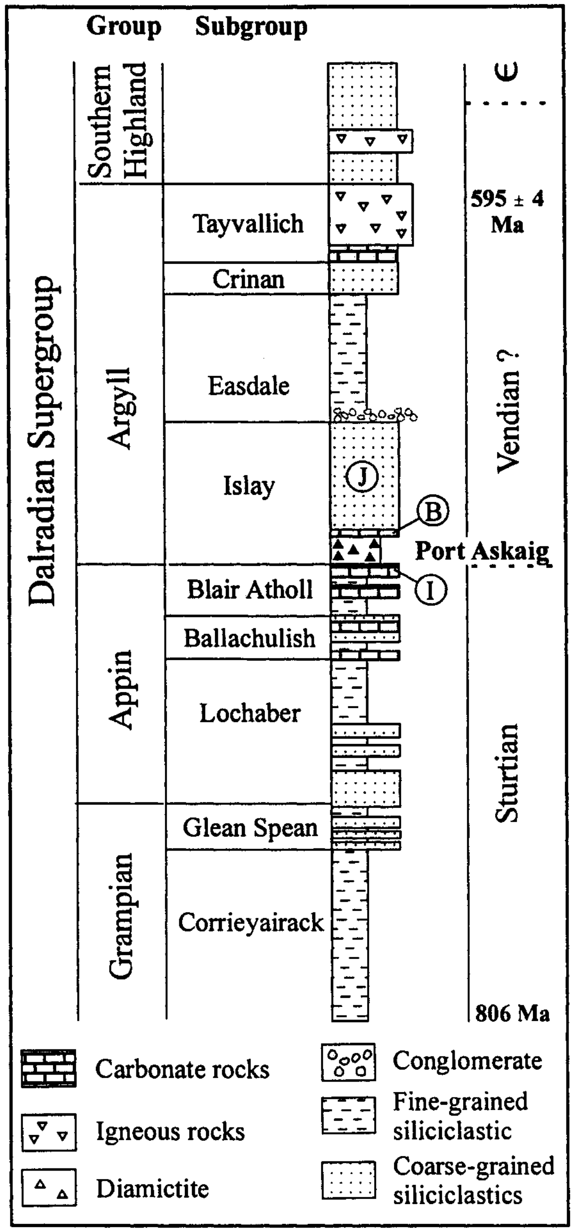


Fig. 4.2: The Dalradian Supergroup and stratigraphic position of the Port Askaig Formation. Total stratigraphic thickness is approximately 25 km (modified from Anderton, 1985; Eyles, 1988; Prave, 1999). Date for the Grampian Group is from Noble *et al.*, (1996), whereas the date for the Tayvallich Volcanics is from Halliday *et al.*, (1989). The Port Askaig Formation is likely to be Sturtian in age (*c.* 725); see text for discussion. J, Jura Quartzites; B, Bonahaven Formation; I, Islay Limestone.

Tayvallich Volcanics.

The Port Askaig Formation is only broadly constrained chronometrically by a date of 806 Ma from the underlying base of the Dalradian Supergroup and a date of 595 Ma from the overlying Tayvallich Volcanics (Fig. 4.2, Halliday *et al.*, 1989; Noble *et al.*, 1996; Prave, 1999 and references therein). It was initially correlated with other Varangian glacigenic successions in the North Atlantic region (Hambrey, 1983). However, recent discovery of a younger glacigenic unit in Ireland, the preservation of an approximately 8 km thick succession lying between the Port Askaig Formation and the overlying Tayvallichs dated at 595 Ma, as well as geochemical data in associated carbonates, all suggest that the Port Askaig Formation is more likely to be equivalent to the Sturtian or older Neoproterozoic glacial period (*c.* 725 Ma; Prave, 1999; Brasier and Shields, 2000; Condon and Prave, 2000; Knoll, 2000).

The Dalradian succession and the Port Askaig Formation outcrop in a broad belt that extends from northern Ireland to northeastern Scotland (Fig. 4.1; Kilburn *et al.*, 1965; Spencer, 1971; Harris *et al.*, 1993). The Port Askaig Formation is best exposed on the Garvellach Islands and Islay off the western coast of Scotland. This study focuses on outcrops on the Garvellach Islands where excellent exposures allow high-resolution sedimentological analysis and examination of lateral and vertical facies changes through the lower 450 m of the Port Askaig Formation (Fig. 4.1).

4.2.1. Previous sedimentological analysis of the Port Askaig Formation

While many have studied the Port Askaig Formation (references in Kilburn *et al.*, 1965; Spencer, 1971), its sedimentology and stratigraphy were most comprehensively documented by Spencer (1971). Spencer (1971) defined five stratigraphic members within the Port Askaig Formation based on changes in clast lithology and dominant facies types and identified 47 different diamictite units (D1-D47) separated by sandstones, conglomerates and mudstones (Fig. 4.3).

Member I consists of diamictites interbedded with conglomerates, sandstones and mudstones as well as two lithologically distinctive sedimentary units (the Great Breccia and the Disrupted Beds). Clasts in conglomerates and diamictites are predominantly dolomitic. Member II consists of diamictites interbedded with sandstone and mudstone and is distinguished from Member I by the appearance of significant amounts of granitic clasts within coarse-grained facies. Member III is dominated by thick successions of sandstone interbedded with relatively thin beds of diamictite and conglomerate. Member IV consists of thick diamictite units interbedded with sandstones, whereas thick successions of sandstone once again dominate in Member V.

Spencer (1971) suggested that the Port Askaig Formation recorded 17 glacial ice advances into a marine basin. Diamictites, including the Great Breccia, were interpreted as subglacial lodgement tills, while sandstones and mudstones recorded marine sedimentation between glacial periods. Spencer (1971) reported many horizons of 'sandstone wedges' penetrating finer-grained facies and interpreted these as periglacial

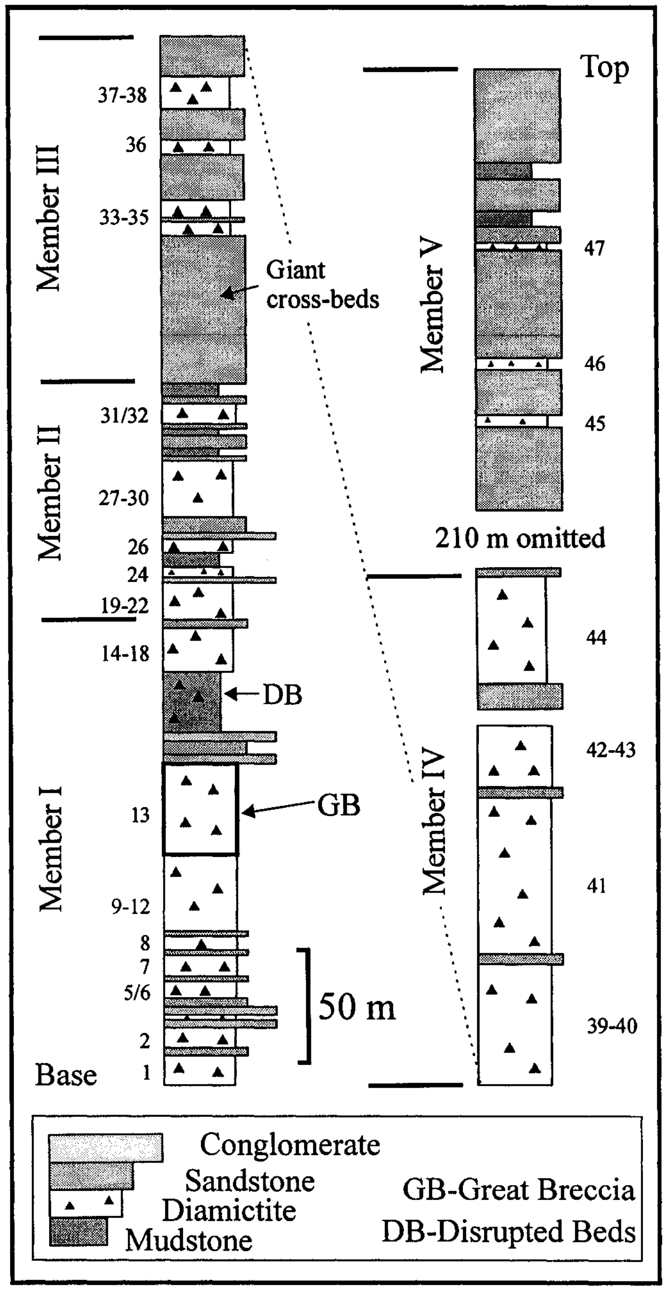


Fig 4.3: Schematic log through the Port Askaig Formation showing Members I through V (Spencer, 1971); base is at lower left, top at upper right. Logs of Members I through III are based on analysis of outcrops on Garvellach Islands; logs of Members IV and V are modified from Spencer (1971) and based on outcrops from Islay. Numbering of diamictite units first established by Kilburn *et al.*, (1965) and used later by Spencer (1971) is retained here.

features. The common association of diamictite penetrated by sandstone wedges with overlying marine sandstones was thought to record the transition between glacial terrestrial and periglacial to non-glacial marine conditions (Spencer, 1971).

More recent sedimentological work focused on sedimentary characteristics and facies associations of diamictites and marine sediments, and concluded that the Port Askaig Formation records glacially- and tidally-influenced marine sedimentation (Eyles *et al.*, 1985, Eyles and Clark, 1985; Eyles, 1988). Eyles (1988) demonstrated that diamictites most likely formed as a result of 'rainout' of fine grained sediments and ice-rafted debris and by subaqueous downslope remobilization of glacigenic sediments in a marine setting. In this alternative 'glaciomarine' model, the Great Breccia was interpreted as an olistostrome, its deposition related to the extensional tectonic setting of the Dalradian Basin (Eyles, 1988). Sandstones, conglomerates and siltstones interbedded with the diamictites were interpreted as recording shallow marine sedimentation on a tidally-influenced shelf (Eyles, 1988). Eyles and Clark (1985) challenged the periglacial origin of the sandstone wedges proposed by Spencer (1971) and suggested gravitational loading and injection of sandstone into finer grained facies as a mechanism for their formation.

Studies to date have focused on identifying the origin and depositional setting of diamictites preserved in the Port Askaig Formation based on their sedimentary characteristics and facies associations. The current study presents detailed sedimentological data from 11 sections logged on the Garvellach Islands (total thickness of 825 m; Appendices A1-A4) and focuses on the vertical succession of facies associations in order to establish an allostratigraphy (Fig. 4.1). The allostratigraphic analysis involves subdivision of the succession into packages of genetically-related facies, which highlight major changes in depositional conditions. The development of an allostratigraphy adds to the existing sedimentological studies of the Port Askaig Formation by establishing the nature of environmental change over time and by assessing the relative importance of tectonic and climatic controls on the development of this lengthy glacigenic succession.

4.3. SEDIMENTARY FACIES

The Port Askaig Formation on the Garvellach Islands consists of a thick succession (approximately 445 m) of interbedded diamictites, sandstones, mudstones and conglomerates. Diamictites (55%) and sandstones (35%) comprise the bulk of the total succession thickness, and are interbedded with relatively thin units of mudstone and conglomerate (6% and 4% respectively). Clast lithologies change from predominantly intrabasinal (dolomite and other detrital dolomitic facies) in the lower part of the Port Askaig Formation to increasingly extrabasinal up-section (granites, quartzites, schist, gneisses and igneous rocks; Fig. 4.4; Kilburn *et al.*, 1965; Spencer, 1971). The thick heterolithic succession contains no major angular unconformities (Spencer, 1971).

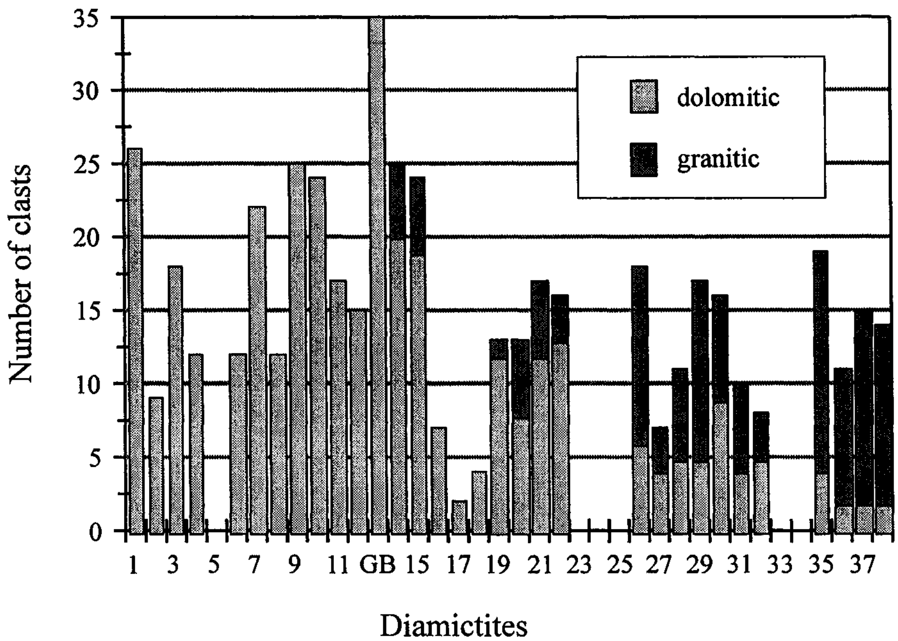


Fig. 4.4: Graph showing up-section trend in clast lithology and abundance. Clasts >1 cm in diameter were counted over 0.1 m^2 of diamictite outcrop of the Port Askaig Formation (Data from Spencer, 1971). Diamictites numbered as in Fig. 4.3.

4.3.1. Diamictites

4.3.1.1 Description

Diamictites are matrix-supported, massive or stratified and contain clasts up to several metres in diameter within a siltstone to silty sandstone matrix (Fig. 4.5, see also Figs. 3.6, 3.8, 3.10, 3.12A). Stratification is defined by slight changes in matrix texture or clast characteristics, deformed inclusions of other facies types (see below), or discontinuous sandstone stringers. Clasts are subangular to rounded and occur in variable abundance. In places, clast fabric has been affected by tectonic cleavage so that it is difficult to determine if clasts ever had a preferred orientation (Spencer, 1971). Coarse-tail inverse grading was noted in some diamictite units (e.g. D15, Eileach an Naoimh).

Diamictites often contain inclusions of other facies types and/or sandstone stringers (Fig. 4.5). Inclusions are up to several m^2 and consist of irregular masses of dolomitic conglomerate, sandstone and siltstone or diamictite with different matrix texture or clast abundance. They often show evidence of soft sediment deformation with contorted and convoluted internal structure and/or outer margins, and impart a chaotic internal structure to the hosting diamictite. Outer margins of inclusions are either well defined or poorly defined and gradational with the host diamictite. Discontinuous stringers of deformed sand (with or without clasts <3 cm in diameter), several cm thick and up to several m in length, commonly have flame-like irregular margins (Fig. 4.5C).

The *Great Breccia* was mapped by Spencer (1971) as a distinctive diamictite unit up to 50 m thick (Spencer, 1971). Recent mapping and sedimentological analysis of

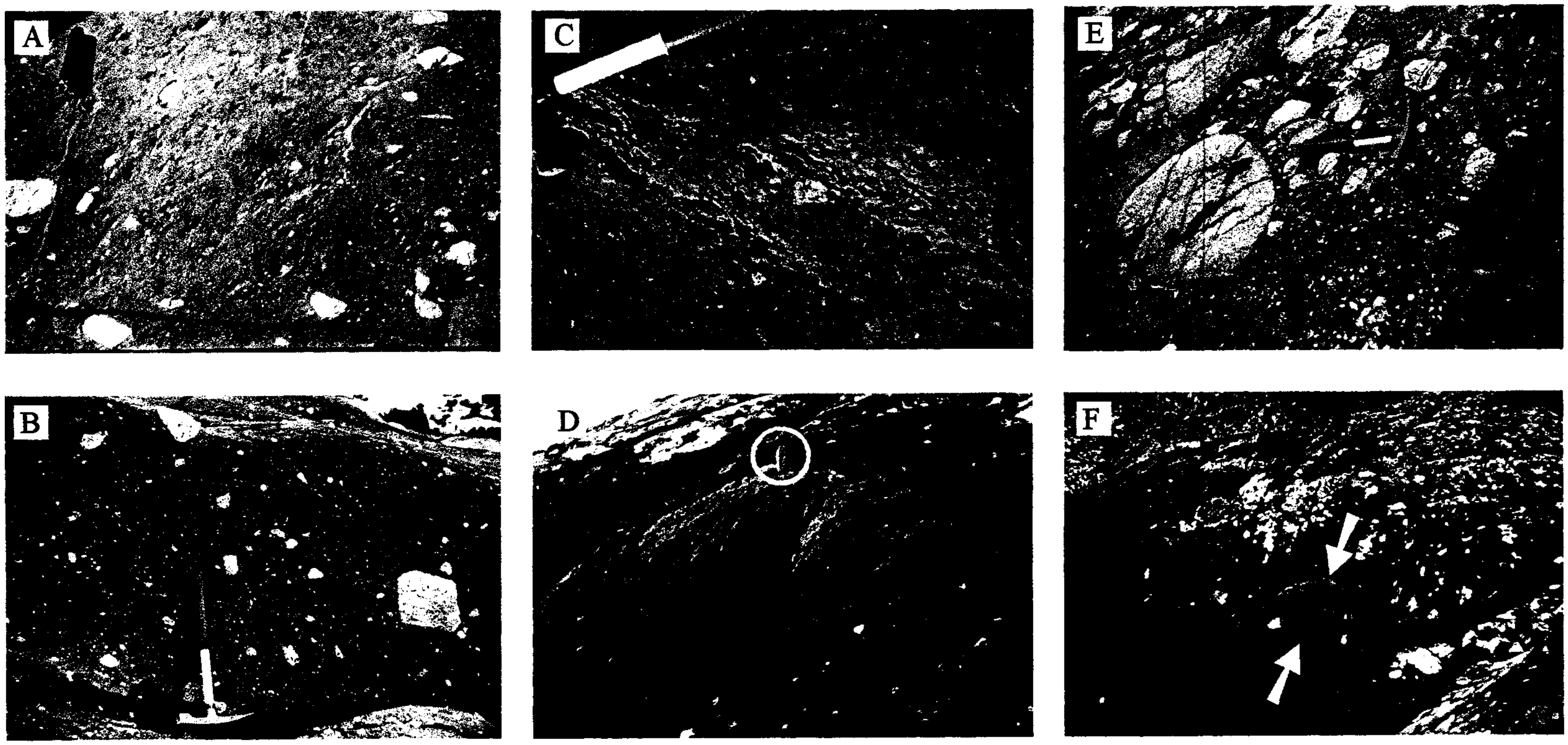


Fig. 4.5: Diamictites of the Port Askaig Formation. A) dolomitic diamictite, D19-22, A'Chuli; B) granitic diamictite typical of the uppermost diamictite units in Member III, D38, west Garbh Eileach, log GE-4; C) sandstone stringer within diamictite D26, A'Chuli, log. AC-2. Note the soft-sediment deformation of the outer margin of the sandstone stringer; D) deformed sandstone lens within diamictite D30, east Garbh Eileach, log GE-3; E) clast-rich inclusion or clast cluster within diamictite D35, east Garbh Eileach, log GE-3. Note the poorly sorted nature of the inclusion, the rounded shape of the larger granitic clasts and the poorly defined outer margin of the clast cluster; F) deformed inclusion of sandstone (arrows), diamictite D22, east Garbh Eileach, log GE-3.

outcrops on all four Garvellach Islands shows that the Great Breccia comprises three sub-units (see Figs. 3.5 to 3.10): Unit 1 is a diamictite that contains clasts up to 100s of m in diameter consisting of various detrital carbonate facies types, Unit 2 is a laterally discontinuous unit of interbedded diamictite, conglomerate, sandstone and siltstone, and Unit 3 is a diamictite unit containing predominantly metre-sized dolomitic clasts. The Great Breccia is sharply overlain by massive dolomitic sandstones and conglomerates (Fig. 4.3).

Diamictites most commonly occur as relatively tabular beds, on average up to 10 m thick, that can be traced up to several kilometres across the Garvellach Islands. Several diamictite units are laterally discontinuous over 100's of metres (e.g. D26, D31 and D32; Fig. 4.6), whereas others merge together as interbedded sediments thin out laterally (e.g. D2/3, D5-12, D16-18, D20-22, D27-29; Fig. 4.1). One diamictite unit (D36, west Garbh Eileach) exhibits rapid lateral facies change, passing abruptly into well-stratified sandstone and conglomerate over several metres. Upper and lower contacts of diamictite units are often sharply defined: lower contacts are conformable to erosional, with some gradational contacts, whereas upper contacts are loaded or conformable. Diamictites are commonly overlain by conglomerates, but are also associated with sandstones and mudstones.

4.3.1.2. Interpretation

The Port Askaig diamictites were originally interpreted as terrestrial deposits

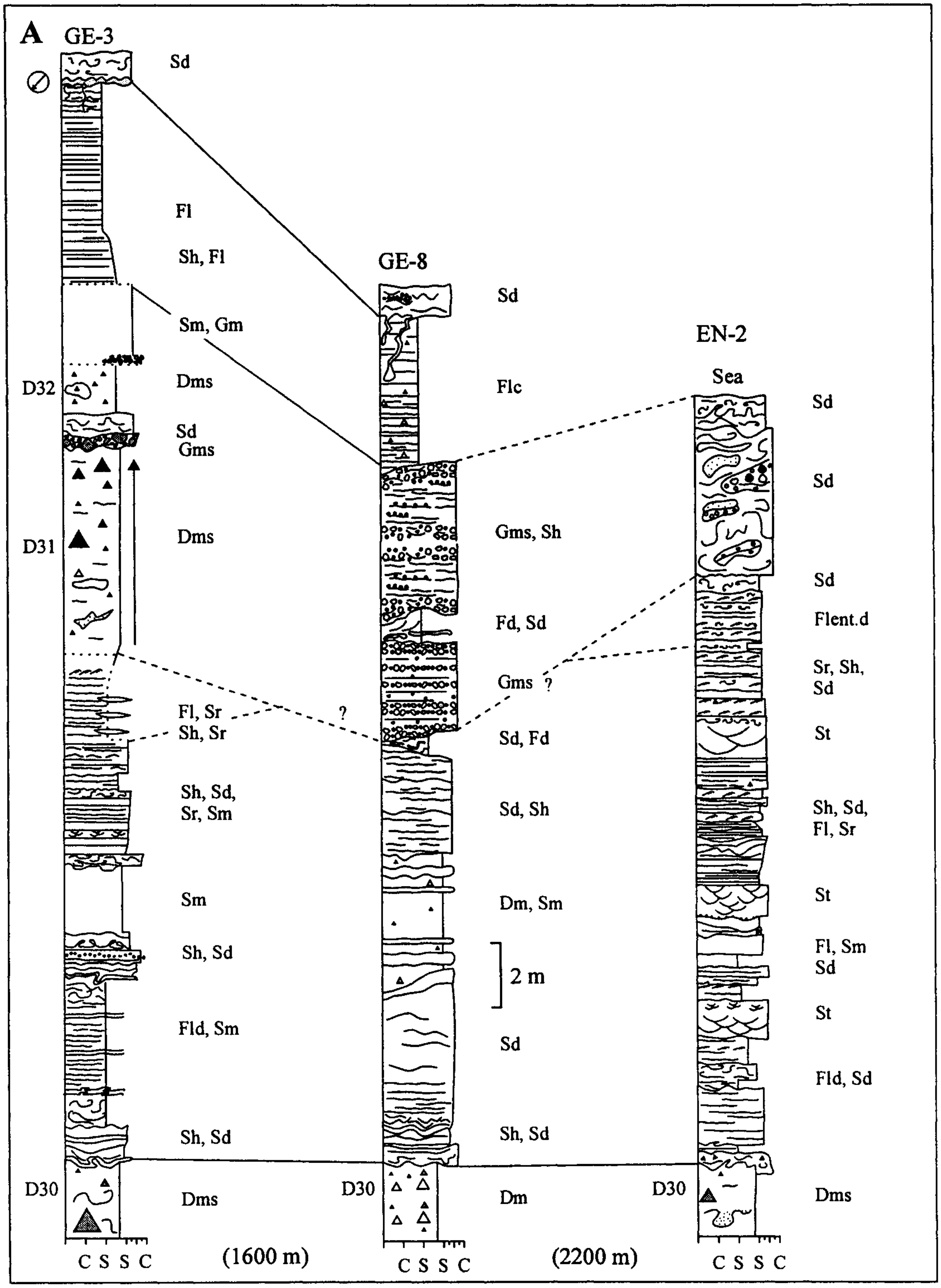


Fig. 4.6: A) Partial graphic logs at selected sites showing the interval between diamictite D30 to the base of the giant cross beds. Note the lateral facies changes. See Fig. 4.1 for location of logs.

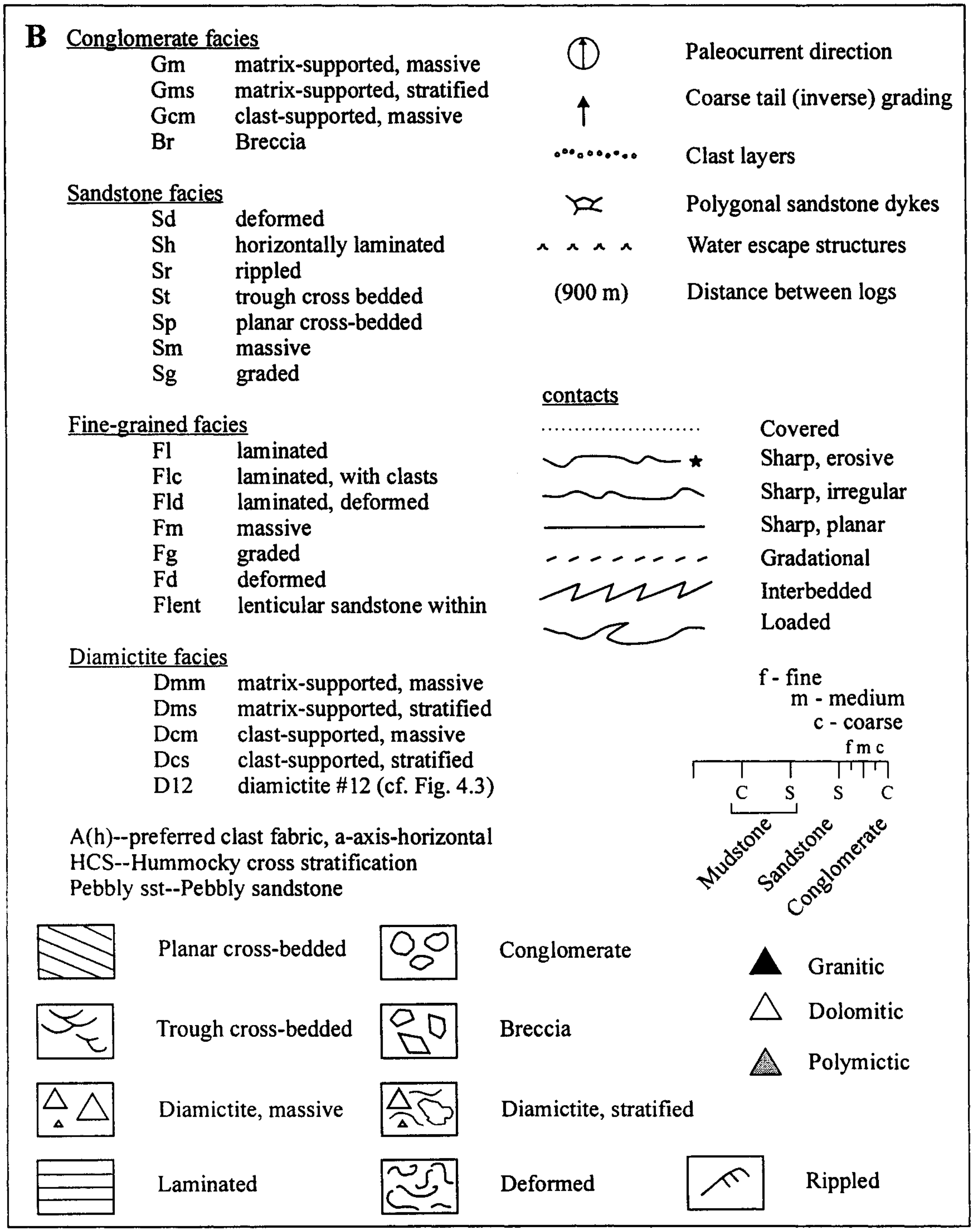


Fig. 4.6: B) Symbols and lithofacies codes used in this study.

formed subglacially (Kilburn *et al.*, 1965; Spencer, 1971), but were later interpreted as the product of glacially-influenced subaqueous sediment gravity flow processes and 'rainout' of fine-grained suspended sediments and ice-rafted debris in marine environments (Eyles, 1988). Differentiation of terrestrial and marine depositional environments is important as each reflects rather different ice cover and climatic conditions. Figure 4.7 presents a summary of the nature of diamictites produced by subaqueous subglacial lodgement and deformation, sediment gravity flow and 'rainout' processes based on information available in the literature. The diamictite characteristics identified in Figure 4.7 can be used together with associated facies and outcrop geometry to discriminate between terrestrial and marine depositional environments.

A subglacial origin for all the Port Askaig Formation diamictites is unlikely due to the absence of striated and faceted clasts, bullet-shaped boulders and boulder pavements, the lack of organized conglomerates and sandstone deposits typical of outwash fans found at ice margins and the absence of glacitectonic deformation features (Fig. 4.7; Eyles, 1988). The lateral continuity of diamictite units in the Port Askaig Formation also contrasts with the chaotic outcrop geometry and rapid lateral facies changes that characterize ice-marginal terrestrial and marine deposits (cf. Eyles and Eyles, 1984; Hart and Roberts, 1994).

In glacially-influenced marine settings, diamictites can form as a result of sediment gravity flow or 'rainout' processes or a combination of the two. Diamictites formed as a result of subaqueous sediment gravity flow may be massive, graded or

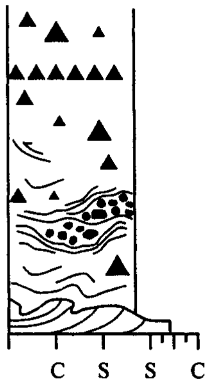
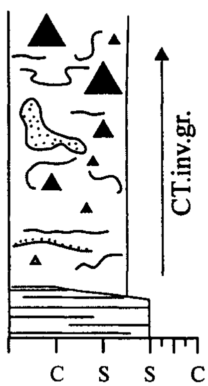
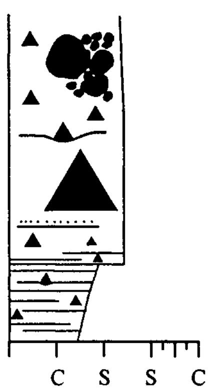
 A) Lodgement and deformation	-Massive or chaotically stratified -Inclusions (can be deformed and/or well sorted) from outwash sediments -Boulder pavements, bullet-shaped boulders -Succession of erosional to deformed basal contact, deformed and sheared glacitectonite and massive diamictite -Thin, tabular geometry, rapid lateral facies changes (m to 10's m scale) -FA: outwash conglomerates and sandstones	Powell, 1981; Eyles *et al.*, 1985; Powell and Molnia, 1989; Hart and Roberts, 1994; Lønne, 1995; Benn and Evans, 1996; Hambrey and McKelvey, 2000
 B) Sediment gravity flow	-Grading from turbulence or dispersive pressures within the flow prior to deposition -Inclusions and stringers (deformed) from incomplete mixing of sediment rafts eroded into or carried by the debris flow -Sharp, conformable or erosional contact, deformation at base of unit -Lenticular, tabular geometry -FA: other sediment gravity flow facies	Nardin *et al.*, 1979; Prior *et al.*, 1984; Eyles *et al.*, 1985; Postma *et al.*, 1988; Powell and Molnia, 1989; Prior and Bornhold, 1990; Aksu and Hiscott, 1992; Eyles, 1993; Eyles and Eyles, 2000; Hambrey and McKelvey, 2000; Lowe and Guy, 2000; Laberg and Vorren, 2000; Mulder and Alexander, 2001
 C) 'Rainout'	-Massive or stratified with horizontal stratification from winnowing (stringers) -Dropstones, inclusions, clast clusters -Variable clast content from changes in sediment accumulation rates or supply of IRD -Gradational to conformable contact -Tabular blanket-like, draped geometry -FA: outwash sediments (ice proximal), sediment gravity flow facies and laminated mustones with dropstones (ice distal)	Powell, 1981; Eyles *et al.*, 1985; Powell and Molnia, 1989; Hart and Roberts, 1994; Lønne, 1995; Crowell, 1999; Hambrey and McKelvey, 2000; O' Cofaigh and Dowdeswell, 2001

Fig. 4.7: Idealized graphic log and characteristics of diamictite units deposited by A) subglacial lodgement and deformation processes; B) sediment gravity flow processes; and C) 'rainout' processes. Lithofacies codes as in Fig. 4.6B; FA-Facies Association. No particular clast lithology is inferred. Deposition by subglacial lodgement or deformation processes can result in diamictite units up to 10's of m, whereas deposition by sediment gravity flow and 'rainout' processes can result in units up to 100's of m in thickness.

stratified, with sharp and conformable or erosional basal contacts, and inclusions of deformed sediments and sandstone stringers, which give the diamictite a chaotic internal geometry (Fig. 4.7). In addition, sediment gravity flow diamictites tend to have tabular or lenticular geometry, resulting from the onlapping accumulation of successive flow lobes (Fig. 4.7).

Several of the diamictite units on the Garvellach Islands (e.g. D15, D26a, D28/29, D30, D31) exhibit these sedimentary characteristics and are therefore attributed to sediment gravity flow depositional processes. Inclusions of sediments in these diamictites result from incomplete mixing of heterolithic source sediments during transport or incorporation of underlying materials (Nardin *et al.*, 1979; Eyles and Eyles, 2000; Lowe and Guy, 2000). Sandstone stringers form by incomplete mixing of heterolithic sediments or winnowing of the sediment gravity flow deposits by traction currents (Powell and Molnia, 1989; Eyles and Eyles, 2000). Coarse-tail inverse grading indicates significant dispersive pressures and development of a traction layer at the base of the flow prior to deposition (Postma *et al.*, 1988; Mulder and Alexander, 2001).

The diamictite unit termed the Great Breccia by Spencer (1971) exhibits similar characteristics to carbonate megabreccia deposits reported from carbonate slope settings and is interpreted as the product of catastrophic subaqueous mass failure of a fault scarp (see Chapter 3). Large blocks of diamictite, conglomerate, sandstone, mudstone and other debris were mobilized in a thick debris flow to form the megabreccia unit (Unit 1). Subsequent units (Units 2 and 3) represent deposition from additional sediment gravity flows created by progressive backstepping and stabilization of the fault scarp (Fig. 3.11).

Diamictites formed as a result of 'rainout' of suspended fine-grained sediments and ice-rafted debris may be massive or stratified, exhibiting variable clast content and gradational to conformable basal contacts (Fig. 4.7). 'Rainout' diamictites contain dropstones, clast clusters (e.g. Fig. 4.5E), inclusions of deformed sediments and sandstone stringers (Fig. 4.7). Outsized clasts should show disruption of underlying laminae (i.e. rucking, bending or piercing) in order to be identified as dropstones (Thomas and Connell, 1985). 'Rainout' diamictites are typically characterized by tabular and blanket-like geometry.

Diamictites D20 and D35 exhibit these characteristics and are therefore interpreted as the product of 'rainout' processes. Both of these diamictite units contain inclusions of deformed sediments, which may be attributed to the post-depositional gravity-induced deformation of ice-berg dump deposits (Söhnge, 1984; Lønne, 1995). Sandstone stringers in these diamictites probably result from winnowing of sediment by traction currents as they accumulated on the sea floor.

Many diamictite units within the Port Askaig Formation do not show diagnostic characteristics that allow differentiation of either 'rainout' or sediment gravity flow depositional processes. These facies may have originated as a result of one or the other process or a combination of both. Several Port Askaig diamictites show characteristics indicative of both processes suggesting 'rainout' diamicts may have been readily remobilized on subaqueous slopes.

4.3.2. Sandstones

4.3.2.1. Description

Sandstones are fine- to coarse-grained in texture and occur in three different facies associations (Fig. 4.1): 1) dolomitic sandstone association, 2) interbedded sandstone/mudstone association, and 3) quartz-rich cross-bedded sandstone association. Rare isolated outsized clasts can be found within each of these three sandstone associations but are generally less than 5 cm in diameter. All three sandstone facies associations occur as relatively tabular units with sharp, planar, undulating or loaded basal contacts.

The *dolomitic sandstone association* consists of poorly-sorted and massive, deformed or horizontally-laminated sandstones. Individual sandstone beds are generally 10's of cm in thickness, but up to several m in places. Deformed sandstones have loaded basal contacts and/or convoluted or contorted internal laminations disrupted on cm to m scales. The dolomitic sandstone association is most commonly found in units that are less than several metres thick separating thicker units of diamictite.

The *interbedded sandstone/mudstone facies association* consists of sandstones that are most commonly massive, deformed and laminated (horizontally, wavy or irregular), but can also be rippled or trough cross-bedded (Figs. 4.6, 4.8, 4.9). These sandstone facies are interbedded with < 1m thick units of finely-laminated or deformed siltstones. Cross-bed sets are up to 2 m thick in places, whereas asymmetrical ripples have heights and lengths of several cm. Lithologically, these sandstones can be either

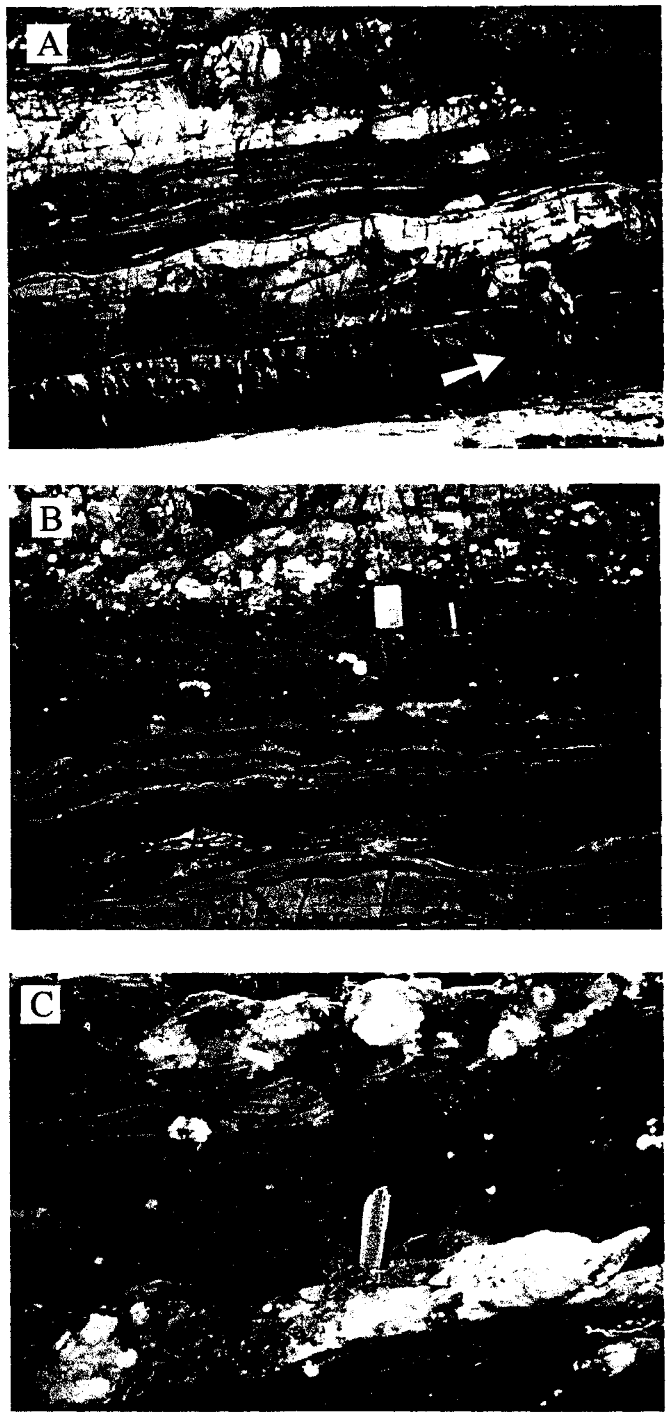


Fig. 4.8: Sandstones of the Port Askaig Formation. A) general view of interbedded sandstone/mudstone association between diamictites D30-31, east Eileach an Naoimh, log EN-2; B) close up of interbedded sandstone/mudstone association shown in (A); C) rippled sandstone facies above diamictite D38, quartz-rich cross bedded sandstone association, south Garbh Eileach, log GE-2.

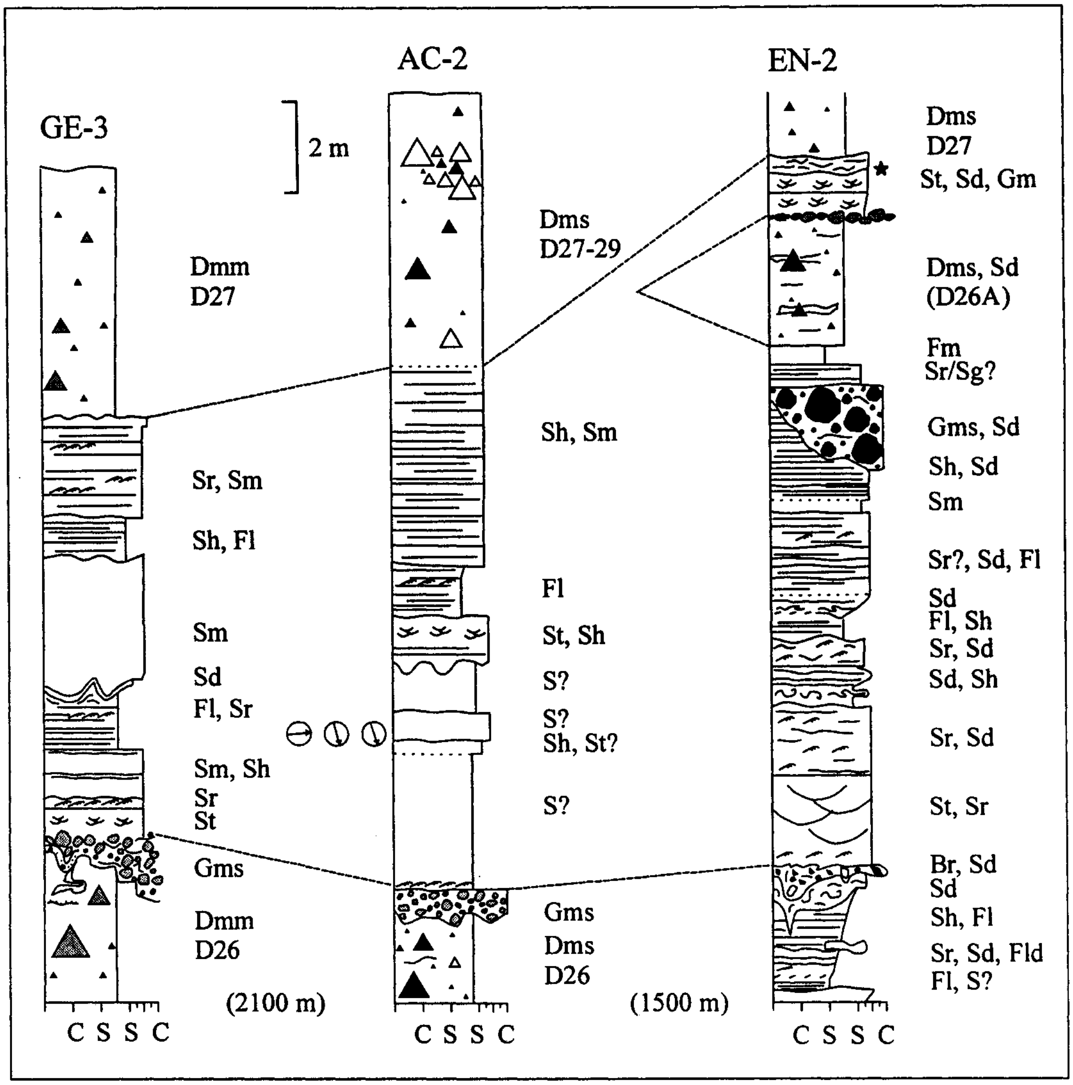


Fig. 4.9: Partial graphic logs from selected sites showing the interbedded sandstone/mudstone interval between diamictites D26 and D27. See Fig. 4.1 for location of logged sections. Note current-generated structures (Sr, St), laterally discontinuous diamictite units (e.g. D26A) and lateral facies variability within sandstones. Symbols and lithofacies codes in Fig. 4.6B.

poorly-sorted and dolomitic or well-sorted and quartz-rich. The interbedded sandstone/mudstone association is found in relatively thick units (10-20 m thick) interbedded with diamictites.

The *quartz-rich cross-bedded facies association* consist of relatively well sorted, quartz-rich, fine- to medium-grained sandstones, which are either cross-bedded, massive or horizontally-laminated. Cross-bed sets are between 0.2 and 11 m in thickness (small to very large using the terminology of Ashley, 1990); sets > 5m thick are hereafter referred to as giant cross-beds (Figs. 4.10, 4.11). The sandstones of the quartz-rich cross-bedded facies association occur in thick successions (10-80 m), and are commonly associated with relatively thin diamictite units.

Foresets within large and giant cross-bedded sandstones (>0.75 m thick) have both planar and trough-shaped lower bounding surfaces (using terminology of Ashley, 1990). Foresets dip at ~14° angle and are either preferably oriented indicating a southward paleocurrent (n= 19, vector mean of 180° with a 59% magnitude; Fig. 4.11), or are randomly distributed indicating more variable paleocurrents (Fig. 4.3, between D35 and D36, Spencer, 1971; Eyles, 1988). In places, internal bedding within foresets is apparent and is either concordant or discordant with the foreset bounding surface identifying simple (Class 3A) and simple-compound (Class 3B) second order sets respectively (Figs. 4.11, 4.12). Several cross-bed sets with multiple bounding discontinuities are also identified and these are termed compound-compound second order sets (Class 3C; Figs. 4.11, 4.12B). In contrast, foresets of relatively thin cross-

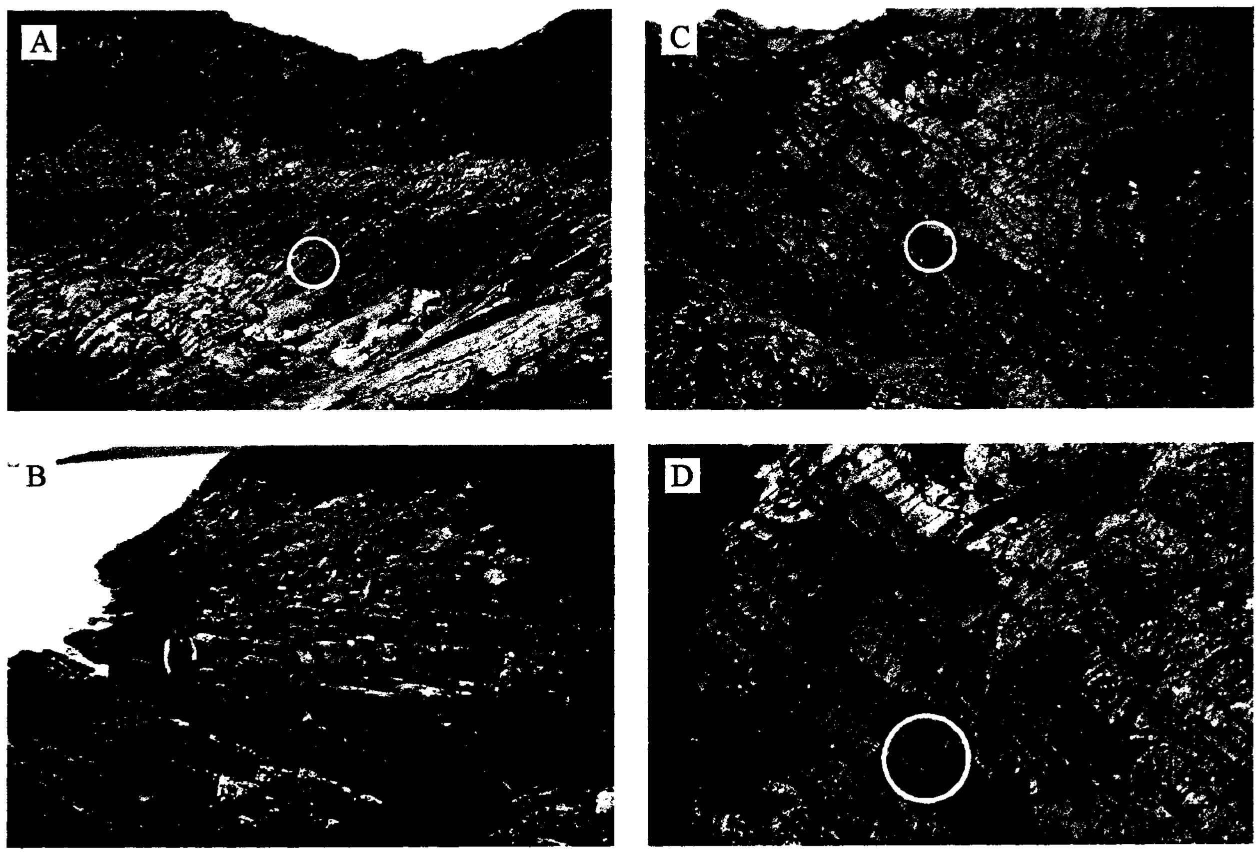


Fig. 4.10: Giant cross-bedded sandstones of the Port Askaig Formation. A) 11 m thick trough cross-bedded set on east Garbh Eileach, log GE-3. Regional dip to the upper left; B) cross-bedded sandstones, east Garbh Eileach, log GE-3. Person for scale standing at bounding surface between two cross-bed sets. Regional dip to the upper right; C) general view, cliff section west Garbh Eileach, log GE-8. Regional dip to the upper right. Person for scale standing on planar massive sandstone unit between two cross-bed sets; D) close up view of cliff section shown in (C).

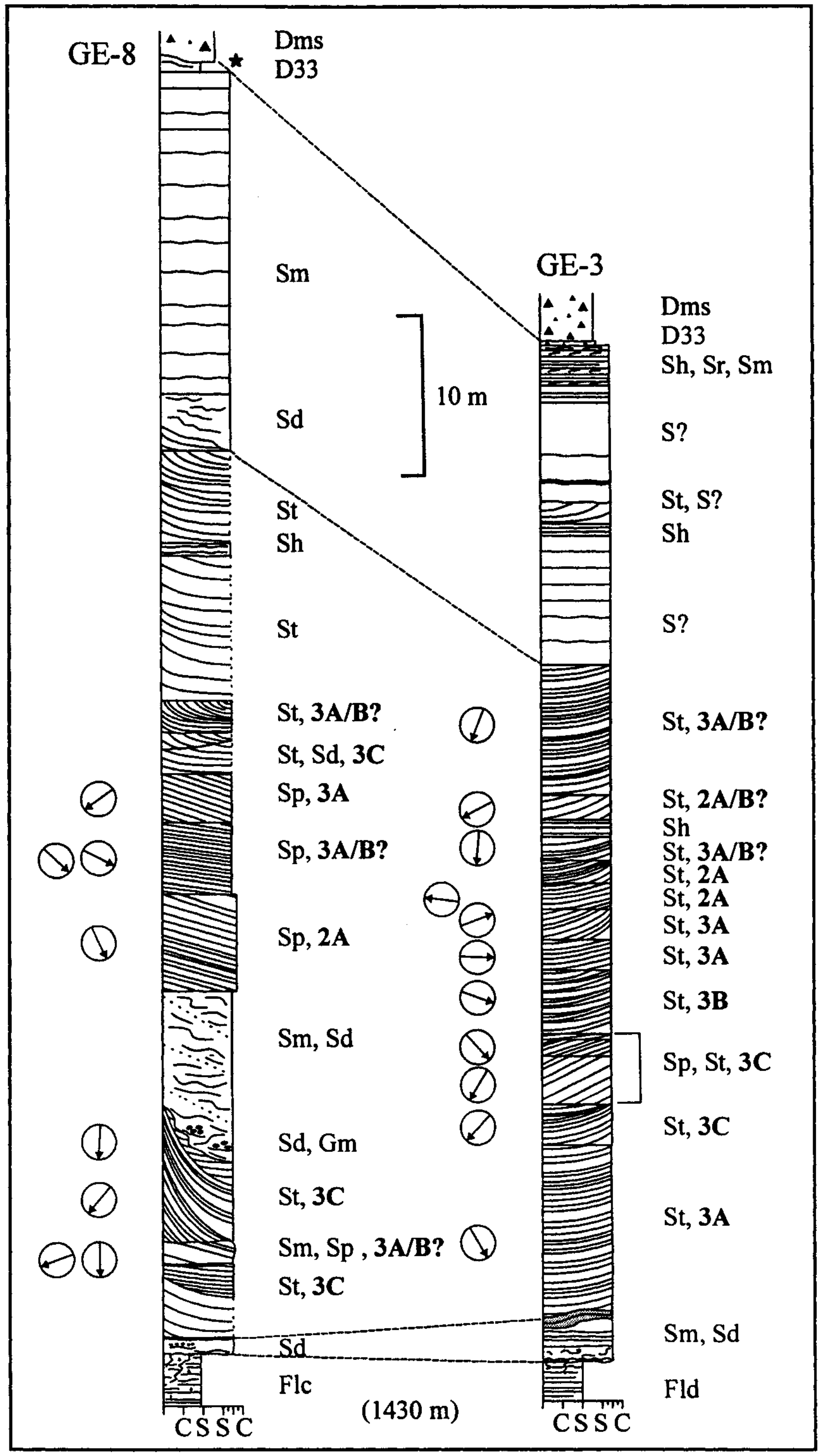


Fig. 4.11: Partial graphic logs at selected sites showing quartz-rich cross-bedded sandstone association, Garbh Eileach. Symbols and lithofacies codes in Fig. 4.6B. Bold numbers after lithofacies codes refer to cross-bed set classification as defined in Fig. 4.12B. See Fig. 4.1 for location of logged sections.

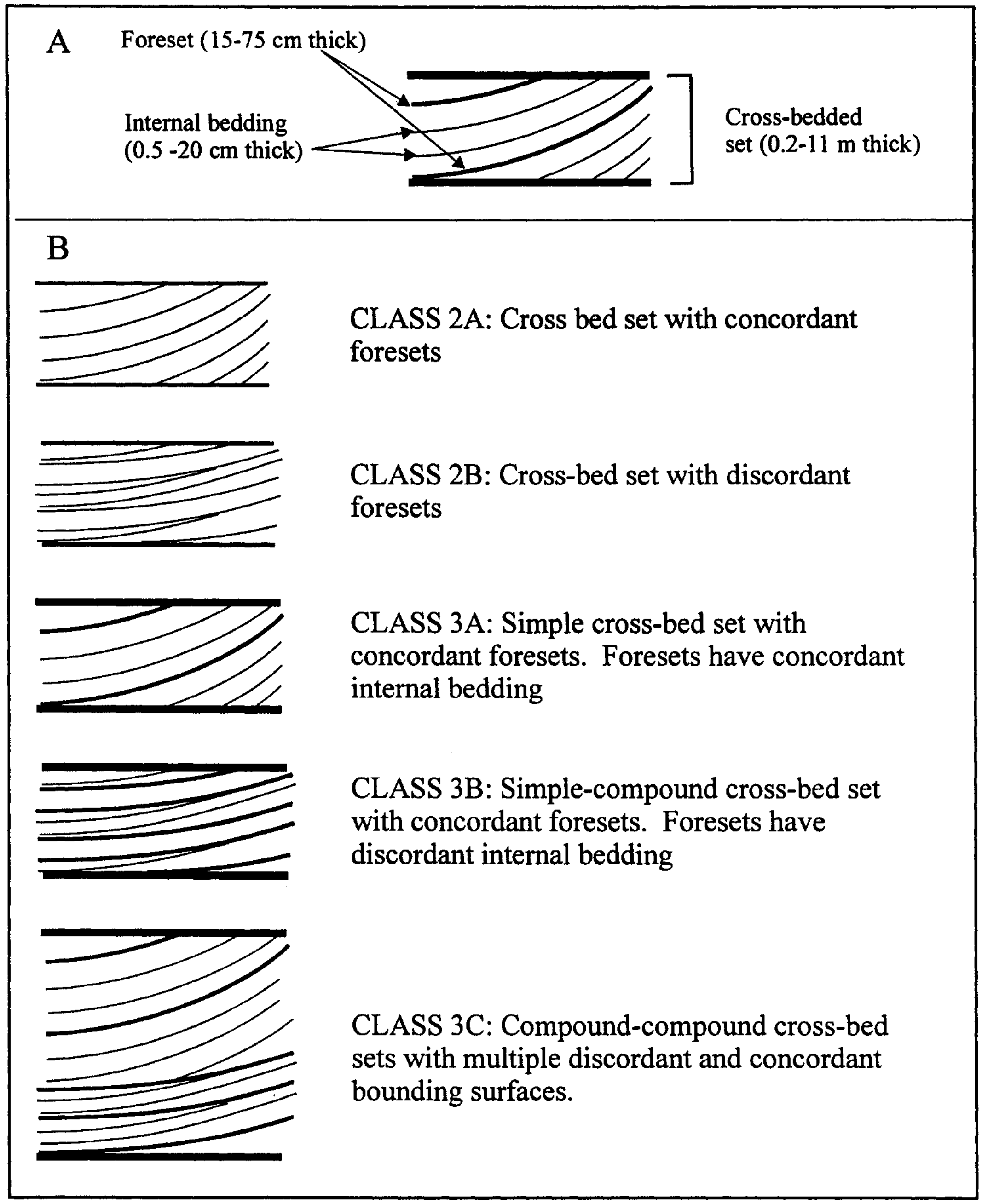


Fig. 4.12: A) Hierarchy of surfaces in the large and giant cross-bedded sandstones (>0.75 m thick) of the Port Askaig Formation and the relative thickness of sets, foresets and internal bedding; B) hierarchy of cross-bedding within giant cross-bedded sandstones (terminology from Anastas *et al.*, 1997).

bedded sandstones are generally trough-shaped, with no internal bedding and variable paleocurrent directions (Spencer, 1971; Eyles, 1988).

4.3.2.2. Interpretation

The *dolomitic sandstone facies association* records primarily rapid deposition by turbidites (massive facies), but also deposition in upper flow regime plane bed conditions and deposition from suspension (horizontally-laminated facies). Loading and contortion of structures within deformed sandstones indicates disruption by liquefaction. The *interbedded sandstone/mudstone association* records deposition by turbidites, as well as accumulation of sediments under unidirectional currents of varying velocities, including upper flow regime plane bed conditions. The *quartz-rich cross bedded facies association* records the reworking of an ample supply of quartz-rich sands by unidirectional currents of varying velocities (cross-bedded and horizontally-laminated facies) and by turbidites (massive facies).

The giant cross-bedded sandstones of the quartz-rich cross-bedded facies association form sets up to 11 m thick and imply the development of very large 2-D and 3-D subaqueous dunes under unidirectional flows (using terminology of Ashley, 1990). The hierarchy of internal laminations within foresets of the giant cross-beds suggests smaller dunes migrated on the lee face of larger dunes as a result of flow variation during bedform growth (Anastas *et al.*, 1997 and references therein).

Large sandy bedforms can develop in large rivers but the giant cross-bedded sandstones described here are not associated with channelized sandstone or floodplain deposits, nor do they contain pebble lags, and mud drapes, which are commonly associated with fluvial cross-beds (c.f. Jones and McCabe, 1980; Haszeldine, 1983; Jones and Rust, 1983; Rust and Jones, 1987; Ashley, 1990; Table 4.1). Large bedforms can also form by aeolian processes but the giant cross-bed sets of the Port Askaig Formation do not exhibit grainfall lamination, adhesion surfaces and subcritical climbing translatent strata typical of cross-beds formed by migrating aeolian sand dunes (Brookfield, 1992).

Such large dunes also develop on modern continental shelves under the influence of tidal and/or geostrophic currents (Table 4.1), which can reach speeds of between 0.5 and 3 m/s (Ashley, 1990; Ramsay *et al*., 1996). Given the sedimentary characteristics and associations of the giant cross-bedded sandstones, it is most likely that they formed as the product of migrating straight- and sinuous-crested subaqueous dunes in an open marine shelf setting (Spencer, 1971; Eyles, 1988).

Both Spencer (1971) and Eyles (1988) interpreted the giant cross-bedded sandstone on the Garvellachs as having formed under strong tidal currents on a relatively shallow marine shelf. Tidal bedforms of a similar scale have been reported in successions of various geologic ages in long narrow basins of extensional tectonic settings, restricted seaways and coastal embayments (Table 4.1; Ravnås and Steel, 1998; Gawthorpe and Leeder, 2001). In these settings, basin morphology leads to the amplification of tidal currents and allows the creation and migration of large-scale subaqueous dunes (Table 4.1).

Table 4.1: Sedimentary characteristics, facies associations and depositional settings of modern (1-9) and ancient (10-18) analogs for giant cross-bedded sandstones. Note: Associated facies uses the lithological codes established for the logs. Other abbreviations are Kinderscout and Roaches Grit (16), Seaton Sluice Sandstone (18); Geostr. is Geostrophic; Sh/Fl is interbedded sands and muds; Shb is herring-bone cross-bedded sands; HCS is hummocky cross stratified sands; SCS is swaley cross stratified sands; Drapes refer to mud drapes and lags to pebbly lags. 1) Swift and Field, 1981; 2) Harris, 1988; 3) Twichell, 1983; 4) Ramsay *et al.*, 1996; 5) Ikehara, 1998; 6) Collier and Thompson, 1991; 7) McCave, 1971; 8) Berné *et al.*, 1988; 9) Houbolt, 1968; 10) Anastas *et al.*, 1997; 11) Smith and Taverner-smith, 1988; Smith, 1992; 12) Kamp *et al.*, 1988; 13) Jones and Rust, 1983; Rust and Jones, 1987; 14) Szigeti and Fox, 1981; 15) Levell, 1980; 16) Jones and McCabe, 1988; 17) Haszeldine, 1983; 18) Anderton, 1976. Geologic ages are as follows: Holocene (1-5, 7-9), Pleistocene (6), Oligo-Miocene (10), Permian (11), Pliocene (12), Triassic (13), Jurassic (14), Precambrian (15, 18), Carboniferous (16, 17).

	LOCATION / FORMATION	SET THICKNESS (m)	SUCCESSION THICKNESS (m)	DIP ANGLE (degrees)	WATER DEPTH (m)	ASSOCIATED FACIES	PALEOGEOGRAPHY	INTERPRETATION
1.	Maryland	3-12		<2	<35	G, F	Inner shelf	Storm
2.	Torres Shelf	20, 2-6		N/A	N/A	S, F	Rimmed shelf	Tidal/wind
3.	Georges Bank	< 25		6-13	<90	G	Shallow shelf	Tidal
4.	Kwazulu-Natal Shelf	12		8	35-70	St, HCS, SCS, Sr	Outer shelf, narrow	Geostr.
5.	Bango Channel	< 4		N/A	100	N/A	Narrow channel	Tidal/Geostr.
6.	Corinth Basin	>15		<35	N/A	G, Sr, St, Shb, Sh/Fl	Fault-bounded, narrow	Tidal
7.	North Sea	2-7		7-13	<50	N/A	Shallow sea	Tidal
8.	English Channel	3.5-7.5		5-17	15-20	N/A	Restricted seaway	Tidal
9.	North Sea	27		5-18	40	Lags, drapes	Shallow sea	Tidal
10.	TeKuiti Group	<4.5	15-50	<20	40-60	Sh, Fl/Sh	Narrow seaway, inner-outer shelf	Tidal/geostr.
11.	Vryheid Formation	<40	80	<28	<90	St, HCS, Sh/Fl, Shb, drapes, lags	Intracratonic seaway	Geostr./tidal
12.	Te-Autu Limestone	10-40	40-70	7-36	N/A	St, Shb, Sh, Sr, drapes, mud clasts	Forearc seaway, narrow	Tidal/geostr.
13.	Hawkesbury Sandstone	<8	290	24	N/A	Sm, St, mud clasts, Fl, G, Sh, St	Braided river	Fluvial
14.	Unkappa Sandstone	>3	N/A	<37	N/A	Fluvial sandstones	Arid, semi-arid terrestrial	Aeolian
15.	Sandfjord Formation	<7	1500	5-15	N/A	Lags, St, Sh, Fl/Sh,Sr, Shb, HCS, drapes	Fluvial, shallow marine	Tidal
16.	K. and R. Grit	<40	400-600	30	N/A	St, Sh, drapes, plant material	Fluvial/ deltaic	Fluvial
17.	Seaton Sluice Sst.	<10	N/A	15-30	N/A	Lags, drapes, St	River bar, braided river	Fluvial
18.	Jura Quartzites	<5	5000	N/A	25-30	Fl, Sr/Fl, Fm, Sh/Fl, Shb, lags, Sd	Shallow marine shelf, extensionalbasin	Tidal/storm

A tidal origin for the giant cross-bedded sandstones is consistent with the narrow, ensialic extensional setting envisaged for the Dalradian Basin during accumulation of the Port Askaig Formation (Anderton, 1982). However, the thick successions of cross-bedded sandstones of the Port Askaig Formation lack supporting sedimentary evidence for tidal deposition such as herringbone cross-bedding, mud-drapes, reactivation surfaces and tidal bundles (c.f. Levell, 1980; Kamp *et al.*, 1988; Smith and Taverner-Smith, 1988; Ashley, 1990; Dalrymple, 1992; Smith, 1992).

Open marine shelf settings are also influenced by geostrophic currents that can produce similar large-scale bedforms to those formed under strong tidal influence (Table 4.1). Geostrophic currents evolve from pressure differences associated with onshore waves and coastal set-up during storms; the offshore-directed relaxation flow associated with storm set-up is deflected by Coriolis forces and produces an offshore current that flows shore-parallel (Fig. 4.13). The large subaqueous dunes migrating today on the southeast African KwaZulu-Natal shelf, and giant cross-bedded sandstones formed on the New Zealand continental shelf have been attributed to such geostrophic currents (Table 4.1).

The facies association of giant cross-bedded sandstones with massive and horizontally-laminated sandstones, as found in the Port Askaig Formation, is typical of sand facies accumulating in outer shelf settings (40-100 m depth) under the influence of geostrophic contour currents (Smith and Taverner-Smith, 1988; Smith, 1992; Viana *et al.*, 1998). The absence of other storm-generated facies such as hummocky cross-

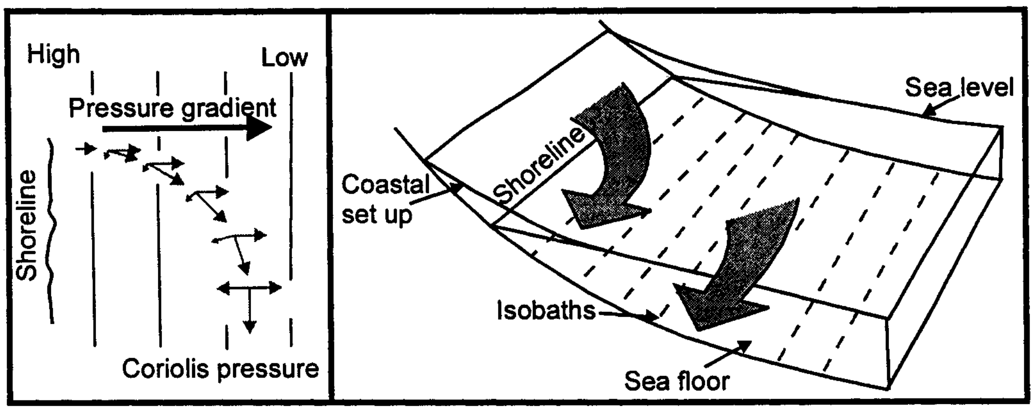


Fig. 4.13: Formation of geostrophic currents (modified from Walker and Plint, 1992). See text for explanation.

stratification (HCS) suggests accumulation of sand below storm wave base (Walker and Plint, 1992). Thus, the sedimentology of the quartz-rich cross-bedded sandstone association is most consistent with deposition by geostrophic currents in an outer shelf setting, although a tidal origin, consistent with the extensional tectonic setting of the Port Askaig Formation, cannot be completely discounted.

In sum, the sandstones of the Port Askaig Formation appear to have accumulated in open marine conditions supplied by both dolomitic and siliciclastic sediment sources. The dolomitic sandstone association is primarily the product of sediment gravity flow processes, and may record deposition on or adjacent to relatively steep slopes; the interbedded sandstone/mudstone association records the influence of unidirectional currents, possibly in shallower water depths. The quartz-rich cross-bedded sandstone association indicates migration of large dunes and smaller bedforms in an outer shelf setting under the influence of geostrophic currents.

4.3.3. Conglomerates

4.3.3.1. Description

Conglomerates are massive to crudely stratified and poorly-sorted with clasts up to 1 m in diameter in a fine- to coarse-grained sandstone matrix (Fig. 4.14). Conglomerate units range from several cm to several metres in thickness; thicker units commonly show a chaotic internal structure defined by contorted stringers of sandstone and variable clast content. Crude inverse grading is observed at the base of some

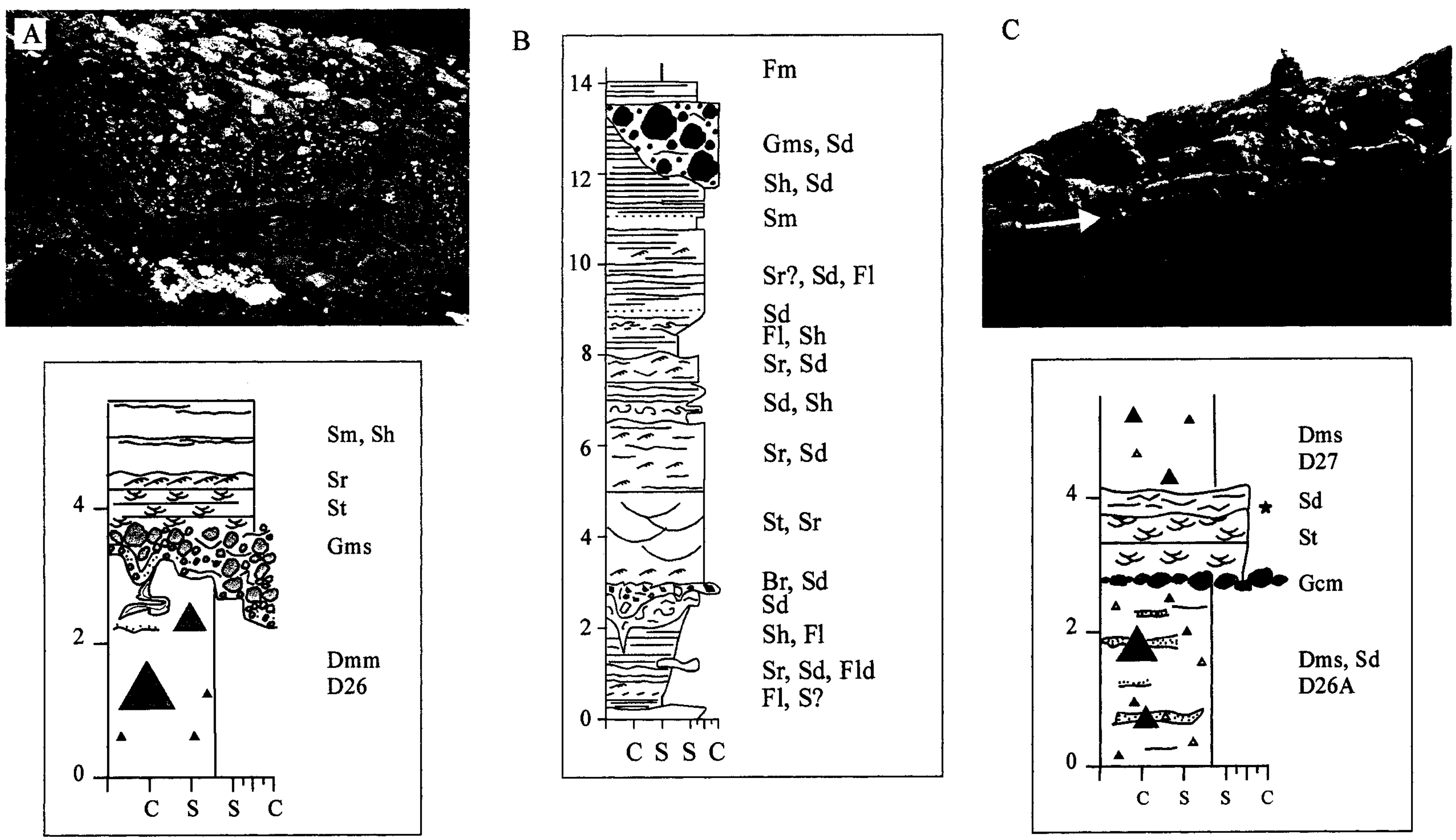


Fig. 4.14: Photographs and graphic logs showing typical stratigraphic relationships of conglomerates with other facies. A) loaded and deformed contact of conglomerate with underlying diamictite (D26), Garbh Eileach, log GE-3; B) conglomerates associated with interbedded sandstones and mudstones lying below D26A, Eileach an Naoimh, log EN-2; C) one-clast thick conglomerate bed (arrow) separating underlying diamictite (D26A) from overlying sandstones, Eileach an Naoimh, log EN-2. See Fig. 4.1 for location of logged sections, Fig. 4.6 for lithofacies codes and symbols.

conglomerates (Fig. 3.12C). Clast lithologies are similar to those of the diamictites and become progressively more granitic up-section.

Conglomerates occur most commonly as relatively tabular units with loaded (up to several m of relief), erosional and/or sharp and conformable lower contacts. Conglomerates often overlie and load into underlying diamictites, but are also associated with massive and horizontally-laminated sandstones; thin clast layers occur within mudstone facies (Figs. 4.14; 4.15).

4.3.3.2. Interpretation

Conglomerates of the Port Askaig Formation were previously interpreted as lags, recording strong current winnowing of underlying diamictites (Spencer, 1971; Eyles, 1988). Conglomerates overlying diamictites D6, D26a, and D35 form a thin surface veneer (one-clast thick) and are immediately overlain by cross bedded sandstones (Fig. 4.14C); these are likely to have originated as lag deposits. However, the thickness of conglomerates that overlie diamictites D2, D22, D26, and D34, their poorly-sorted and matrix-supported nature, chaotic internal structure, erosional and loaded lower contacts and the presence of some inverse grading is most consistent with rapid deposition by sediment gravity flow processes (Walker, 1975; Nemec *et al.*, 1984; Nemec and Steel, 1984). Other conglomerate units within the Port Askaig Formation are associated with siltstones and sandstones and show similar characteristics to those overlying diamictites; these are also interpreted as the product of high-concentration sediment gravity flows.

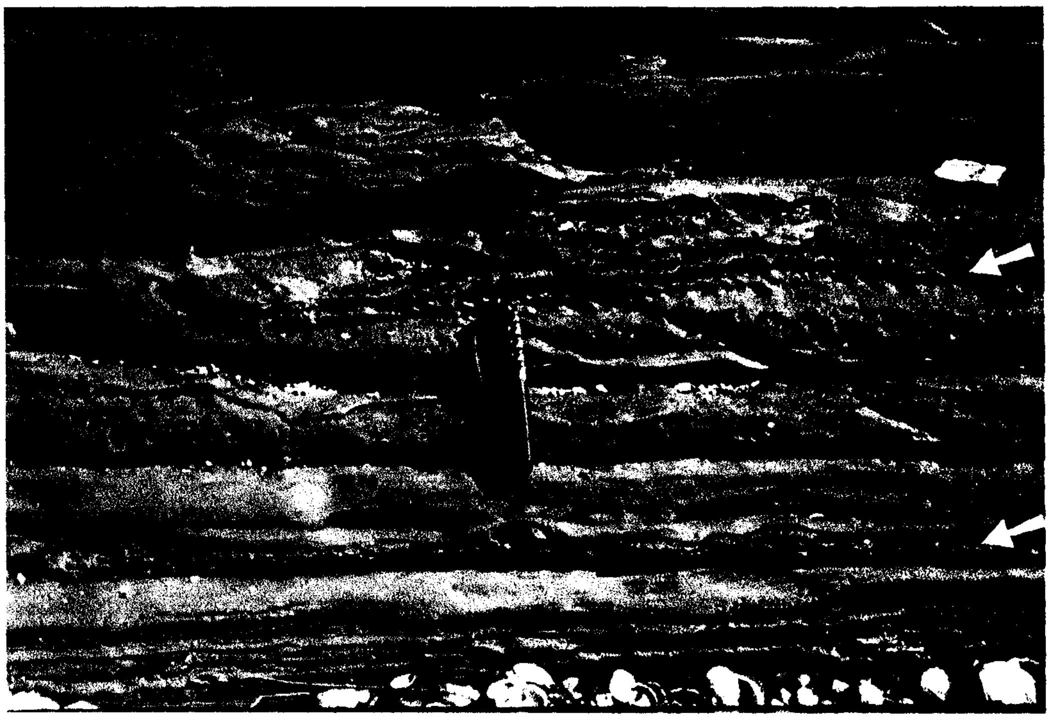


Fig. 4.15: Finely-laminated siltstone with clast layers (arrowed), below diamictite D16, Eileach an Naoimh, log EN-3.

4.3.4. Mudstones

4.3.4.1. Description

Mudstones are most commonly horizontally laminated, with laminae defined by slight changes in grain size; in places normal grading from fine sand to silt can be observed (e.g. below giant cross-bedded sandstones, Garbh Eileach, Fig. 4.11). Several mudstone units are massive or contain lenses of sandstone with weakly defined ripple structures. Although one unit of mudstone at the top of Member II appears to be rhythmically laminated, no cyclical periodicities have been identified (Spencer, 1971). Outsized clasts (< 10 cm) within finely-laminated mudstones are either rare and randomly distributed or occur concentrated in clast-rich layers (up to 5 cm thick; Fig. 3.13B and 4.15; Spencer, 1971; Eyles, 1988). Soft sediment deformation is observed in fine-grained facies and includes convoluted or contorted laminations, pseudonodules, and rare water escape structures.

Mudstone units range from 0.25 to 8 m in thickness. Lower contacts are either sharp and relatively planar or gradational. Upper contacts are commonly sharp, gradational or loaded showing flame structures. Mudstones are commonly associated with sandstone and diamictite, and are mostly found within Member II.

4.3.4.2. Interpretation

Eyles (1988) suggested that finely-laminated siltstones of the Port Askaig Formation accumulated as turbidites produced by sediment-laden underflows or sediment instability at an ice margin. The graded nature of some of the mudstones and their common association with clast layers is consistent with a turbidite origin. Laminated mudstones may also have formed by the passive settling of suspended fines under quiet water conditions (O' Cofaigh and Dowdeswell, 2001 and references therein).

Few of the rare outsized clasts found within finely-laminated mudstones (3 out of approximately 30 observed in this study) show evidence of bending, or piercing of underlying laminae, evidence which may indicate an origin as dropstones released by floating ice (Thomas and Connell, 1985). The paleoclimatic significance of dropstones remains unclear, especially considering they are small (on average less than 5 cm) and very rare (Spencer, 1971). The presence of these 'potential' dropstones may indicate proximity to a glacial ice margin, although they may also have been derived from seasonal sea ice or from icebergs far-removed from an ice margin (Crowell, 1964; Gilbert, 1990; Eyles *et al.*, 1997). Other outsized clasts found within mudstones are likely to have been transported within turbulent sediment gravity flows.

4.3.5. Soft sediment deformation features

4.3.5.1. Description

Horizons displaying intense soft sediment deformation on cm to m scale are bounded by undeformed sediments and occur at several distinct stratigraphic intervals within the Port Askaig Formation (Figs. 4.16 to 4.18). These deformation features affect all facies types and include convoluted or contorted bedding in sandstone and mudstone,

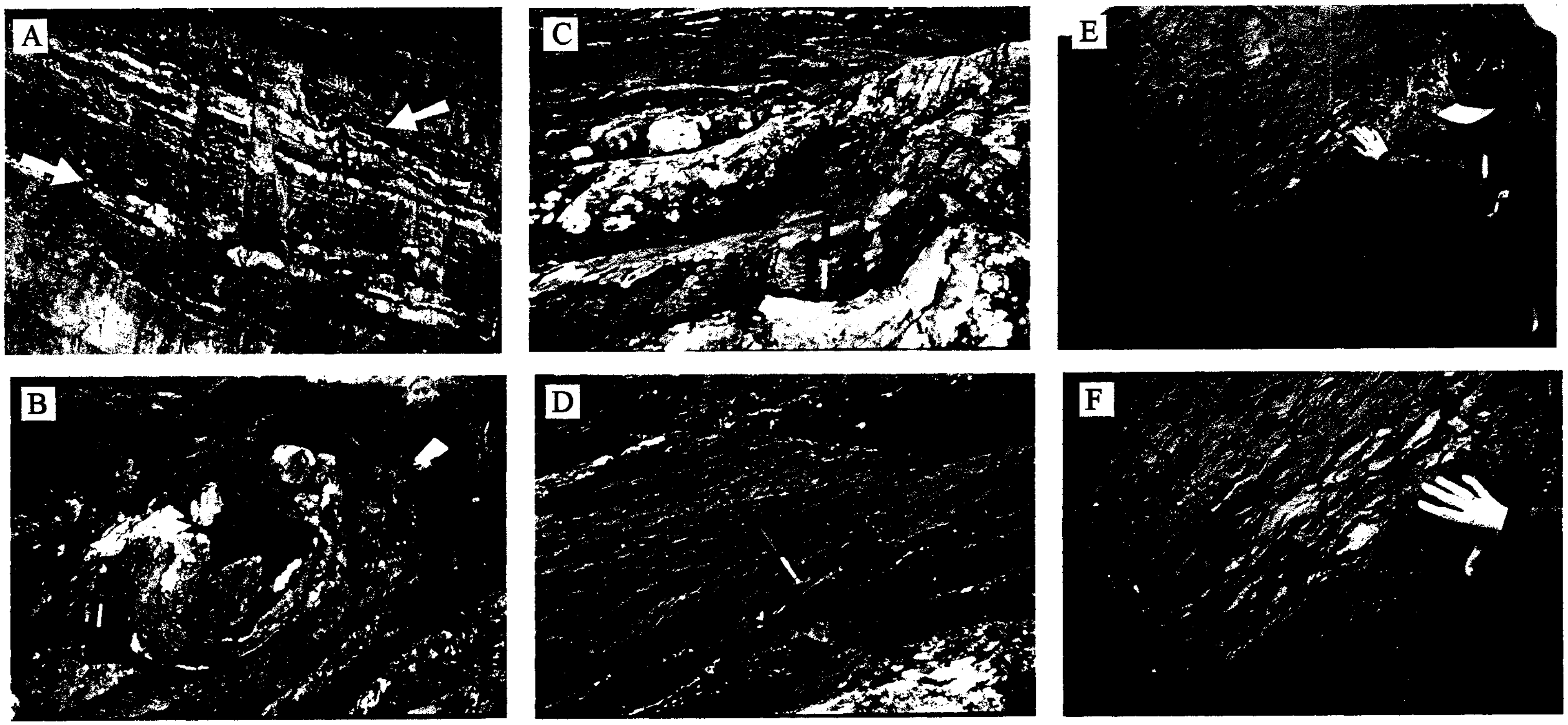


Fig. 4.16: Synsedimentary deformation within the Port Askaig Formation, Garvellach Islands. A) cliff section of Disrupted Beds, Garbh Eileach showing sedimentary boudinage of pale-coloured dolomitic sandstones. Cliff section is approximately 20 m high. Stratigraphic top to upper right; B) folded sandstone above D38, quartz-rich cross-bedded sandstone association, west Garbh Eileach, log GE-2; C) loaded lower bed contact in interbedded sandstone/mudstone association, east Garbh Eileach, log GE-3; D) folding of finely laminated siltstones and sandstones, interbedded sandstone/mudstone association between diamictites D30 and D31, east Garbh Eileach, log GE-3; E) general view of pseudonodules of sands in mudstones, Eileach an Naoimh, log EN-2. Regional dip to the upper left; F) close up of pseudonodules shown in (E).

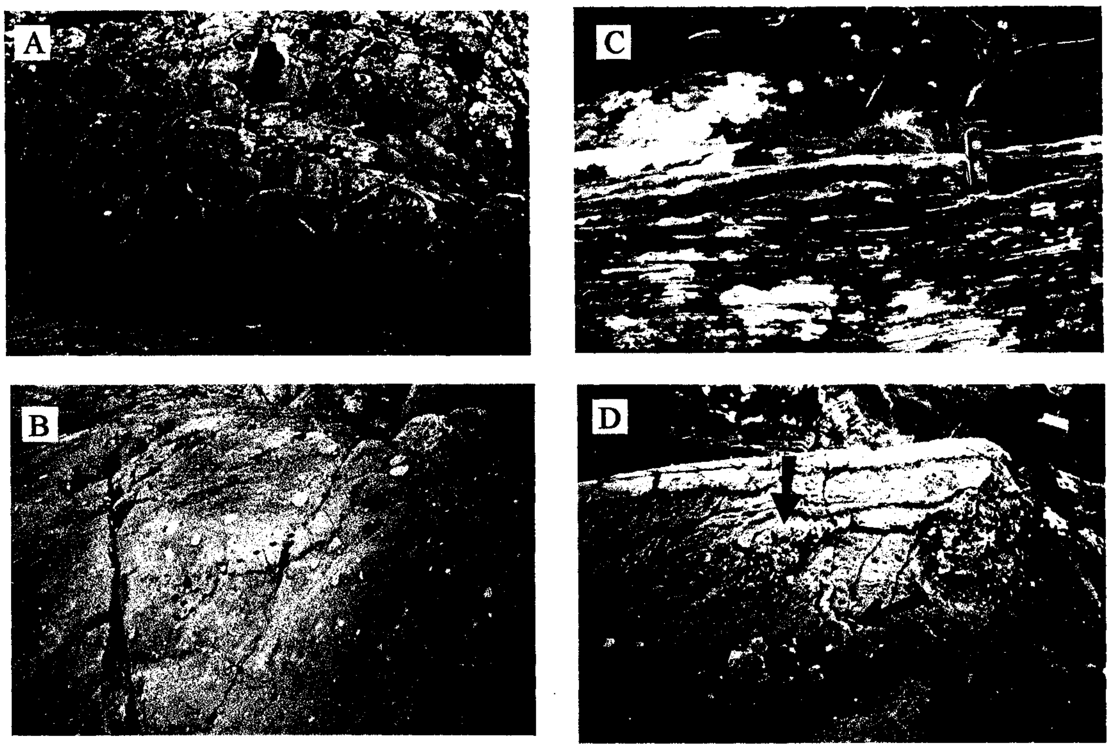


Fig. 4.17: Sandstone dykes: A) polygonal array of sandstone dykes on the surface of diamictite D22, Eileach an Naoimh, log EN-3; B) randomly-oriented sandstone dyke in diamictite D38, plan view, west Garbh Eileach, log GE-4; C) sandstone dyke in cross-section in finely laminated siltstones, east Garbh Eileach, log GE-3; D) sandstone dykes (arrowed) penetrating into diamictite D22, east Garbh Eileach, log GE-3. Note sandstone dyke that is connected to overlying deformed sandstone and conglomerate.

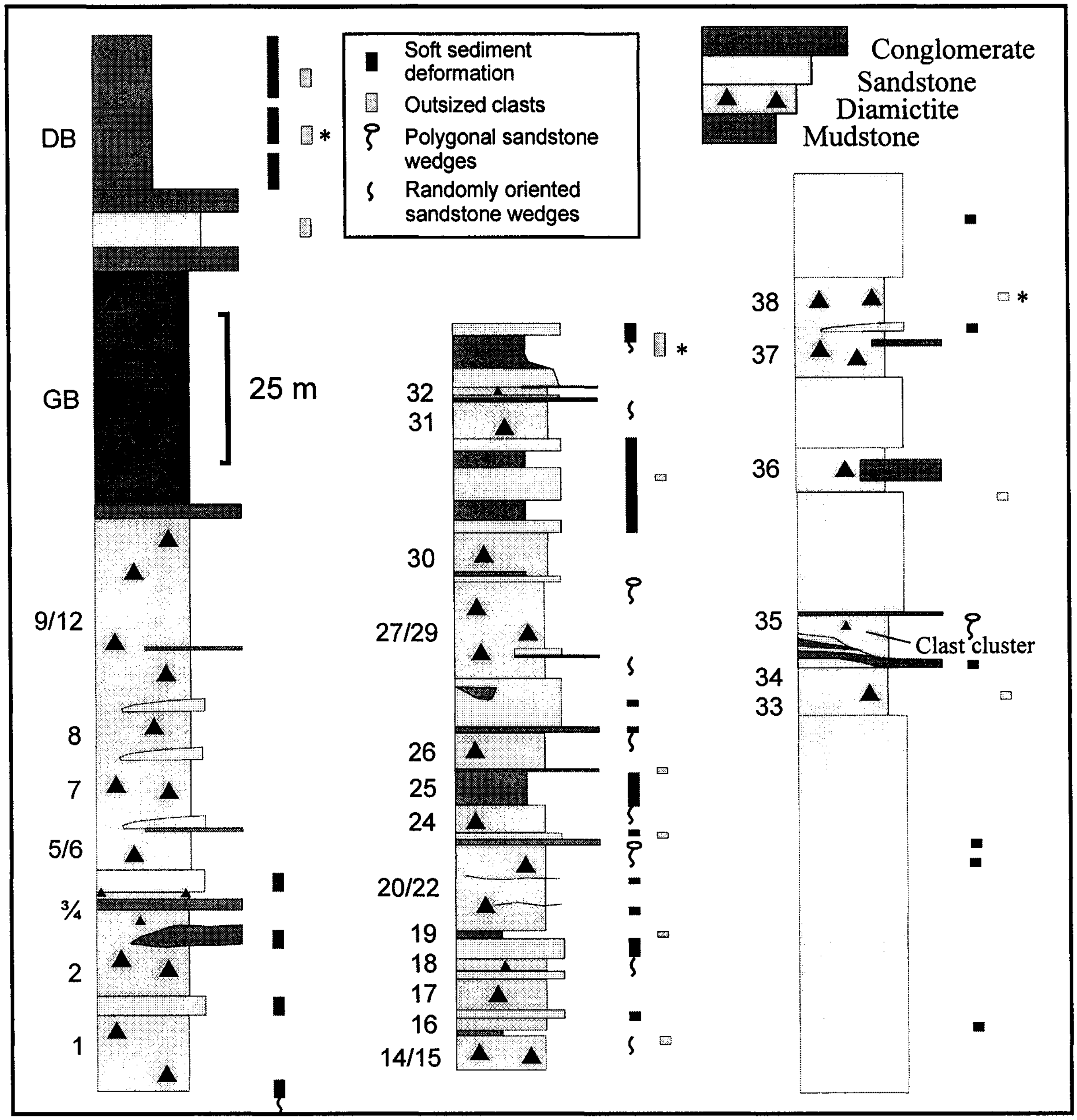


Fig. 4.18: Summary log for the Port Askaig Formation in the Garvellachs showing intervals characterized by synsedimentary deformation, horizons of sandstone dykes and stratigraphic levels where rare outsized clasts were observed (this study and Spencer, 1971; Eyles and Clark, 1985; Eyles, 1988). ' *' denotes intervals where laminae was disrupted by outsized clasts. See discussion (section 4.3.4.2) on the paleoclimatic significance of outsized clasts.

loaded bed contacts, contorted sediment inclusions and sandstone stringers within diamictites and conglomerates, pseudonodules of sandstone in mudstone, sedimentary boudinage, and sandstone dykes penetrating finer-grained facies (Fig. 4.16, 4.17).

Sandstone dykes (the 'sandstone wedges' of Spencer, 1971) are up to 30 cm wide and up to 4 m long and consist of poorly-sorted medium-grained sandstone (Fig. 4.17). They occur most commonly as randomly oriented and isolated features, but are also found arranged in polygonal nets on bedding planes (e.g. D22, D35, Fig. 4.17). These dykes most commonly penetrate into diamictite; however, similar features are found penetrating siltstones of the Port Askaig Formation and are also described from dolomites of the underlying Islay Formation, 20 to 25 m below the base of the Port Askaig Formation (Fig. 4.18; Spencer, 1971; Eyles and Clark, 1985).

4.3.5.2. Interpretation

The close stratigraphic association of horizons of deformed and undeformed sediments within the Port Askaig Formation indicates repeated synsedimentary disturbance of unlithified sediments. The occurrence of contorted and convoluted laminations, ball and pillow structures and pseudonodules are attributed to sediment liquefaction processes (deformation mechanism: *sensu* Owen, 1987) and to gravitational instabilities or unstable density gradients (driving force; *sensu* Owen, 1987; Moretti *et al.*, 1999). Deformed inclusions and stringers within diamictites and conglomerates are interpreted as structures formed by incomplete mixing of source sediment during transport by sediment gravity flow processes (see section 4.3.1.2); inclusions within some diamictites may represent deformed iceberg dump deposits (Söhnge, 1984; Lønne, 1995; Crowell, 1999).

The sandstone dykes were originally interpreted as periglacial ice-wedge casts by Spencer (1971), but Eyles and Clark (1985) suggested they formed as subaqueous deformation features resulting from reverse-density loading and intrusion of sand into finer-grained facies. This alternative model was based on the absence of other periglacial indicators in the Port Askaig Formation, the subaqueous origin of the diamictites, the close association of sandstone dykes with other soft sediment deformation structures and their occurrence in marine non-glacial sediments of the underlying Islay Formation (cf. Neoproterozoic periglacial sandstone wedges in Australia, Williams, 1994). Similar sandstone dykes to those observed within the Port Askaig Formation have been documented elsewhere and interpreted as resulting from fluid injection into seismically-generated fissures (Aspler and Donaldson, 1986; Owen, 1987).

Soft sediment deformation features can be created as a result of rapid deposition of sediment and trapping of water in pore spaces, seismic shaking and/or wave action. The deformation structures in the Port Askaig Formation are unlikely to have formed as a result of wave action as there is no evidence for large storm waves (HCS) and deformation is rarely associated with current-generated features (cf. Pratt, 1994). The abundance of sediment gravity flow facies in the succession suggests rapid deposition of sediment, which may have acted as a trigger mechanism for deformation. However, the abundance and variety of deformation features described in the Port Askaig Formation are similar to those reported in sediments of seismically-active tectonic settings (Söhnge, 1984; Aspler and Donaldson, 1986; Bose *et al.*, 1997; Lignier *et al.*, 1998; Bhattacharya and Bandyopadhyay, 1998; Jones and Omoto, 2000), and their occurrence at discrete stratigraphic intervals suggests episodic seismic shaking may have been an important trigger mechanism for soft sediment deformation during Port Askaig deposition.

4.4. ALLOSTRATIGRAPHY

Sedimentological analysis of individual facies types found within the Port Askaig Formation suggests deposition by a variety of processes in a marine environment supplied with large amounts of poorly-sorted sediments and periodically prone to seismic activity. Establishing the nature and scale of changing paleoenvironmental conditions through time requires analysis of facies types in their stratigraphic context. Seven stratigraphic units can be identified within the Port Askaig Formation on the Garvellach Islands, each consisting of a distinct assemblage of facies types (Fig. 4.19, Table 4.2). These stratigraphic units are (from base to top of the succession); dolomitic diamictites (SU1), the Great Breccia (SU2), the Disrupted Beds (SU3), diamictites and mudstones (SU4), diamictites and sandstones (SU5), cross-bedded sandstones (SU6) and sandstones and diamictites (SU7).

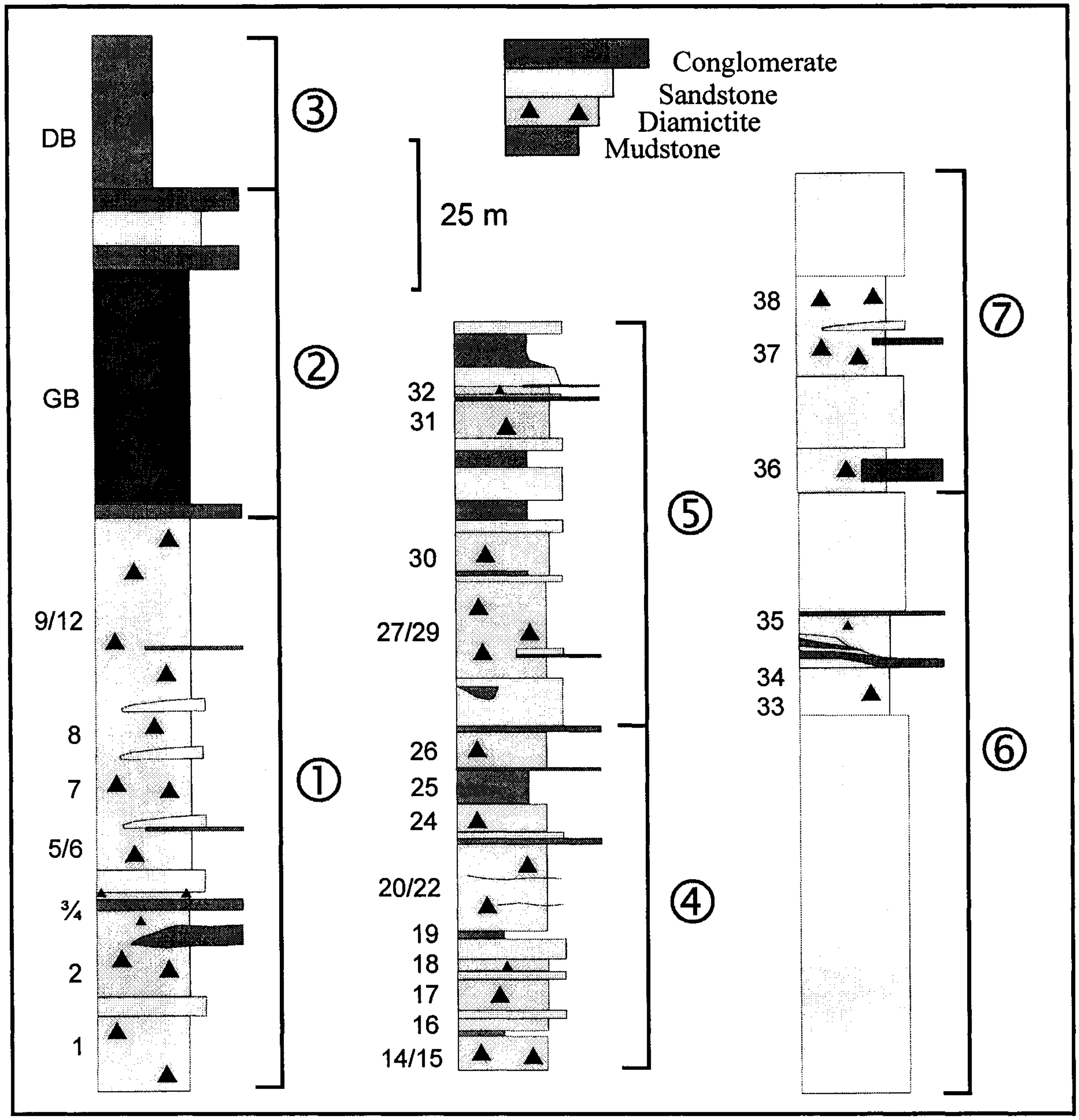


Fig. 4.19: Stratigraphic units of the Port Askaig Formation on the Garvellach Islands. See section 4.4 for explanation. GB- Great Breccia, DB- Disrupted Beds. Numbering of diamictites follows numbering in Fig. 4.3.

Phase	Stratigraphic Unit	Facies	Dominant processes	Ratio of Supply/ acommodation	Depositional Setting	Control on sedimentation Tectonic	Control on sedimentation Climatic
III	7	St, Sh, Dms	Sediment gravity flow, and/or 'rainout', unidirectional currents	N/A	Current dominated deposition alternating with glacially-influenced sedimentation		
	6	Very thick St, Sh, Dms	Geostrophic currents, sediment gravity flow and/or 'rainout'	N/A			
II	5	Sh, Sd, St, Sr, Fl Dms, (def)	Sediment gravity flow, unidirectional currents	S<A	Transitional between Phase I and Phase III		
I	4	Dms,Dmm Fl, Fd, (def)	Sediment gravity flow, and/or 'rainout'	N/A	On or adjacent to a slope, dominated by sediment instability, slumping and mass flow processes. Localized tectonic activity in Su2, episodic seismic activity in SU4		
	3	Fl, Sd, Dms, Gms (def)	Sediment gravity flow, tensional stress and downslope remobilization	S<A			
	2	Mega-Br, Dms, Gm, Fl, Sh, Sm, Sd	Catastrophic and retrogressive mass failure from fault-bounded scarp	N/A			?
	1	Dms, Dmm Gms, Sd, Sm, Sh	Sediment gravity flow and/or 'rainout' of fine sediment and ice-rafted debris	N/A		?	

Table 4.2: Summary table for stratigraphic units 1-7, Port Askaig Formation, showing dominant facies, depositional process and setting, as well as possible climatic and tectonic controls on sedimentation. Ratio between accommodation space created and sediment supplied is also inferred for some stratigraphic units. See text for explanation and see Fig. 4.19 for schematic graphic log showing stratigraphic units. Facies codes according to Fig. 4.6B, (def)-abundant soft sediment deformation features.

4.4.1. SU1: dolomitic diamictites (95 m)

This stratigraphic unit forms the base of the Port Askaig Formation and overlies the upper dolomitic member of the Islay Formation, the upper 36 m of which is characterized by interbedded dolomites, dolomitic sandstones and pelites (Spencer, 1971). The contact of SU1 with the underlying Islay Formation is gradational on the Garvellach Islands, although it is erosional on Islay (Spencer, 1971). SU1 is characterized by massive dolomitic diamictites interbedded with discontinuous units of the dolomitic sandstone association, conglomerate, and siltstone (Fig. 4.19, Table 4.2). The massive diamictites are interpreted here as sediment gravity flow deposits due to the absence of ice-rafted debris in associated sediments, although a 'rainout' origin cannot be completely ruled out. Associated conglomerates, sandstones and massive siltstones are also interpreted as the product of sediment gravity flow processes. Soft sediment deformation features are not common in SU1.

The transition from the increasingly siliciclastic sediments of the upper Islay Formation to the diamictite-bearing SU1 of the Port Askaig Formation heralds a major change in sediment supply and depositional conditions in the basin. SU1 appears to record reworking of local coarse-grained dolomitic sediments. The predominance of sediment gravity flow facies suggests an unstable depositional environment, possibly on or adjacent to a marine slope underlain by dolomite-rich sediments including conglomerate, sandstone and siltstone. This depositional environment was most likely below wave base due to the absence of wave-generated current structures. The predominance of sediment gravity flow facies and absence of ice-rafted debris make it difficult to identify a distinct glacial influence on sedimentation.

4.4.2. SU2: the Great Breccia (40-60 m)

SU2 comprises the distinctive lithological unit of the Great Breccia and immediately overlying massive to stratified conglomerates and sandstones (Fig. 4.19, Table 4.2). This stratigraphic unit is bounded by an erosional surface at its base. SU2 constitutes a crudely fining-upward succession, recording initial catastrophic failure of lowermost Port Askaig Formation and Islay Formation sediments along an active fault scarp, and subsequent retrogressive failure of unstable sediments exposed along the scarp (see Chapter 3). This stratigraphic unit indicates localized active faulting and creation of steep subaqueous slopes during this stage of Port Askaig deposition.

4.4.3. SU3: the Disrupted Beds (25 m)

This stratigraphic unit consists of fine-grained blue mudstone interbedded with dolomitic sandstone, conglomerate and diamictite and is pervasively deformed at the scale of individual beds as well as over 10's of metres of lateral and vertical exposure (Figs. 3.4, 3.13, 4.19, Table 4.2). Although the Disrupted Beds show pervasive soft sediment deformation, the alternation of deformed and undeformed units suggests synsedimentary deformation rather than post-depositional deformation of the entire stratigraphic unit (Fig. 3.4). Lateral changes from boudinaged sandstone beds indicative of tensional stresses and associated lateral mass movement (Kilburn *et al.*, 1965; Spencer, 1971), to chaotically stratified and massive diamictite are interpreted to record increasing amounts of deformation and homogenization of sediment as a result of slumping of interbedded dolomitic sandstone, conglomerate and mudstone (Table 4.2).

SU3 marks the first appearance of extrabasinal clasts in the Port Askaig Formation (10-15% of all clasts in 0.1 m^2; Fig. 4.4); rare outsized clasts in mudstone facies show disruption of underlying laminae and are interpreted as dropstones. Mudstones of the Disrupted Beds sharply overlie conglomerate of SU2 (Figs. 3.4, 4.19).

The Disrupted Beds indicate continued slope instability and slumping in the Port Askaig Basin, which may be related to continued seismic activity following deposition of the Great Breccia (see section 3.4.2). While the paleoclimatic significance of dropstones is difficult to establish (see section 4.3.4.2), their presence within the Disrupted Beds and the first appearance of extrabasinal clasts may constitute the first indication of glacial influences on sedimentation. The base of SU3, where mudstones sharply overlie conglomerates, is interpreted as a flooding surface that represents an imbalance created between accommodation space and sediment supply. The predominance of siltstone and siltstone-rich diamictites within this stratigraphic unit also suggests a change to relatively deep and quiet water depositional conditions.

4.4.4. SU4: diamictites and mudstones (55 m)

This stratigraphic unit is characterized by interbedded diamictites, finely-laminated mudstones with common clast layers, and sandstones of either the dolomitic sandstone association (base of SU4) or the interbedded sandstone/mudstone association (top of SU4; Fig. 4.19, Table 4.2). Diamictites in SU4 are interpreted as the product of sediment gravity flow processes (D15), 'rainout' processes (D20) or a combination of both, and indicate glacial influences on sedimentation. Associated fine-grained facies are interpreted as the product of turbidites and settling of fine-grained sediments out of suspension. Soft sediment deformation continues to be prevalent in this unit, although it is restricted to discrete horizons separated by undeformed sediments (Fig. 4.18). This type of deformation indicates intermittent but continued sediment instability, and may be related to episodic seismic shaking. Rare isolated outsized clasts occur within sandstones and siltstones, but only one showed disruption of underlying laminae. The lower contact of this stratigraphic unit is sharp to gradational on the underlying Disrupted Beds.

SU4 sediments record reworking of both intrabasinal and extrabasinal material and the continuation of depositional conditions dominated by sediment gravity flow processes. These unstable depositional conditions may be related to either tectonic activity and/or to high influxes of coarse-grained glacial sediments to the basin margin.

4.4.5. SU5: diamictites and sandstones (65 m)

This stratigraphic unit is characterized by stratified diamictites interbedded with sandstones of the interbedded sandstone/mudstone association and contains distinct horizons of soft sediment deformation features (Fig. 4.19, Table 4.2). Diamictites D26A, 28/29, 30 and 31 are interpreted as sediment gravity flow deposits. Sandstones record both deposition by turbidites (massive facies) and deposition under unidirectional currents (rippled, cross-bedded facies) including upper flow regime plane bed conditions (horizontally-laminated facies). Siltstones record deposition by turbidity currents. Although, the basal contact of SU5 is conformable and sediment gravity flow processes continue to dominate, SU5 is distinguished from the underlying stratigraphic units by thicker interbedded sandstone/mudstone units separating diamictites (Fig. 4.19) and the appearance of current-generated structures. This represents a significant change in the nature of sediment supply and depositional conditions within the basin.

Discrete horizons of soft sediment deformation features in sandstones and mudstones are common between diamictites D30 and D31 and, together with horizons of sandstone dykes penetrating into siltstones and diamictites (D27, 29, and 31, Figs. 4.17C, 4.18), indicate continued sediment instability, likely related to intermittent seismic shaking. Evidence for direct glacial influence on sedimentation within SU5 is very limited; only the uppermost laminated mudstone unit contains rare dropstones.

This stratigraphic unit exhibits a fining-upward trend, passing from a diamictite-sandstone association at the base, to a diamictite-siltstone association at the top (Fig. 4.19). The fining upward trend probably reflects a change in the balance between accommodation space and sediment supply, and may be a function of a variable supply of sands and/or changing water depths caused by basin subsidence. Intermittent seismic activity, recorded by the soft sediment deformation features in SU5, may have accompanied intervals of rapid basin subsidence. Given the limited evidence for glacial conditions in SU5, the fining-upward trend may also be related to ice withdrawal from the basin margin and corresponding eustatic sea level rise.

4.4.6. SU6: cross-bedded sandstones (100 m)

This stratigraphic unit consists of thick packages of quartz-rich cross bedded sandstone interbedded with thinner units of diamictite and poorly-sorted conglomerate (Fig. 4.19, Table 4.2). Cross-bedded sandstones were most likely deposited under the influence of southward-flowing geostrophic currents in an outer shelf setting. Diamictites show characteristics typical of those deposited by 'rainout' processes (D35), and by a combination of sediment gravity flow and 'rainout' processes (D33 and D34). Conglomerates are the product of both sediment gravity flow processes and winnowing of underlying diamictites. Soft sediment deformation features are a minor component of SU6 and suggest more tectonically-stable depositional conditions than existed previously. Rare outsized clasts occur in cross-bedded sandstone but show no clear evidence of being ice-rafted and may have been transported by traction currents. The base of SU6 is sharp and appears conformable with underlying deformed sandstones (Fig. 4.11).

The quartz-rich giant cross-bedded sandstones at the base of SU6 mark a significant change in depositional conditions characterized by an ample supply of well sorted quartz rich sediment and the migration of large subaqueous dunes in a current-dominated setting. The creation of shore-parallel, southward-flowing geostrophic currents requires a north-south oriented shoreline with a basin open to the east. The absence of any coarse-grained sediments or dropstones throughout the quartz-rich cross-bedded sandstone association and the development of large subaqueous bedforms suggests persistent ice-free conditions. These environmental conditions change abruptly during an interval where deposition of diamictites by mass-flow and/or 'rainout' processes indicates glacially-influenced marine sedimentation. SU6 thus records deposition in a current-dominated, ice-free marine depositional environment, which experiences an interval of glacially-influenced conditions (Fig. 4.19).

4.4.7. SU7: sandstone and diamictite (55 m)

This stratigraphic unit is characterized by interbedded sandstones and diamictites, and is distinguished from the underlying SU6 by relatively thin packages of the quartz-rich cross-bedded sandstone association, smaller cross-bed sets (<50 cm thick), variable paleocurrent directions (Spencer, 1971; Eyles, 1988) and the presence of climbing and mud-draped ripples and soft sediment deformation features (Fig. 4.19, Table 4.2). Cross-bedded sandstones of SU7 are indicative of weaker and more variable traction currents than those recorded in SU6. The presence of massive sandstone, climbing ripples and soft sediment deformation suggests rapid sedimentation in the uppermost part of SU7. Diamictites D36-38 have characteristics that indicate both sediment gravity flow and 'rainout' processes were important in their formation. The base of this stratigraphic unit is marked by an erosional surface at the base of diamictite D36.

SU7 represents deposition in open marine shelf conditions, characterized by rapid deposition of quartz-rich sandstone by variable currents (tidal and/or geostrophic?), and two intervals of glacially-influenced marine sedimentation (Fig. 4.19).

4.5. DEPOSITIONAL HISTORY

The seven stratigraphic units identified in the Port Askaig Formation on the Garvellach Islands show a progressive upward change in facies associations passing from predominantly coarse-grained associations at the base to sandstone-diamictite associations at the top of the succession. This allostratigraphy identifies changing environmental conditions over time and allows three major phases of Port Askaig deposition to be identified (Fig. 4.20): Phase I is dominated by sediment instability (SU1-SU4); Phase II (SU5) shares characteristics with both Phase I and III and constitutes a transitional phase; and Phase III (SU6 and SU7) is dominated by current-dominated sandstone deposition. Each of these phases represents a different stage of development of the sedimentary basin and records changing controls on sedimentation patterns with time.

4.5.1. Phase I (SU1-SU4)

Stratigraphic units SU1-SU4 constitute the first phase (Phase I; Fig. 4.20A) of Port Askaig deposition and consist predominantly of carbonate-rich sediment gravity flow facies that record reworking of locally-derived sediments in a tectonically-unstable depositional environment. These deposits do not contain any current-generated

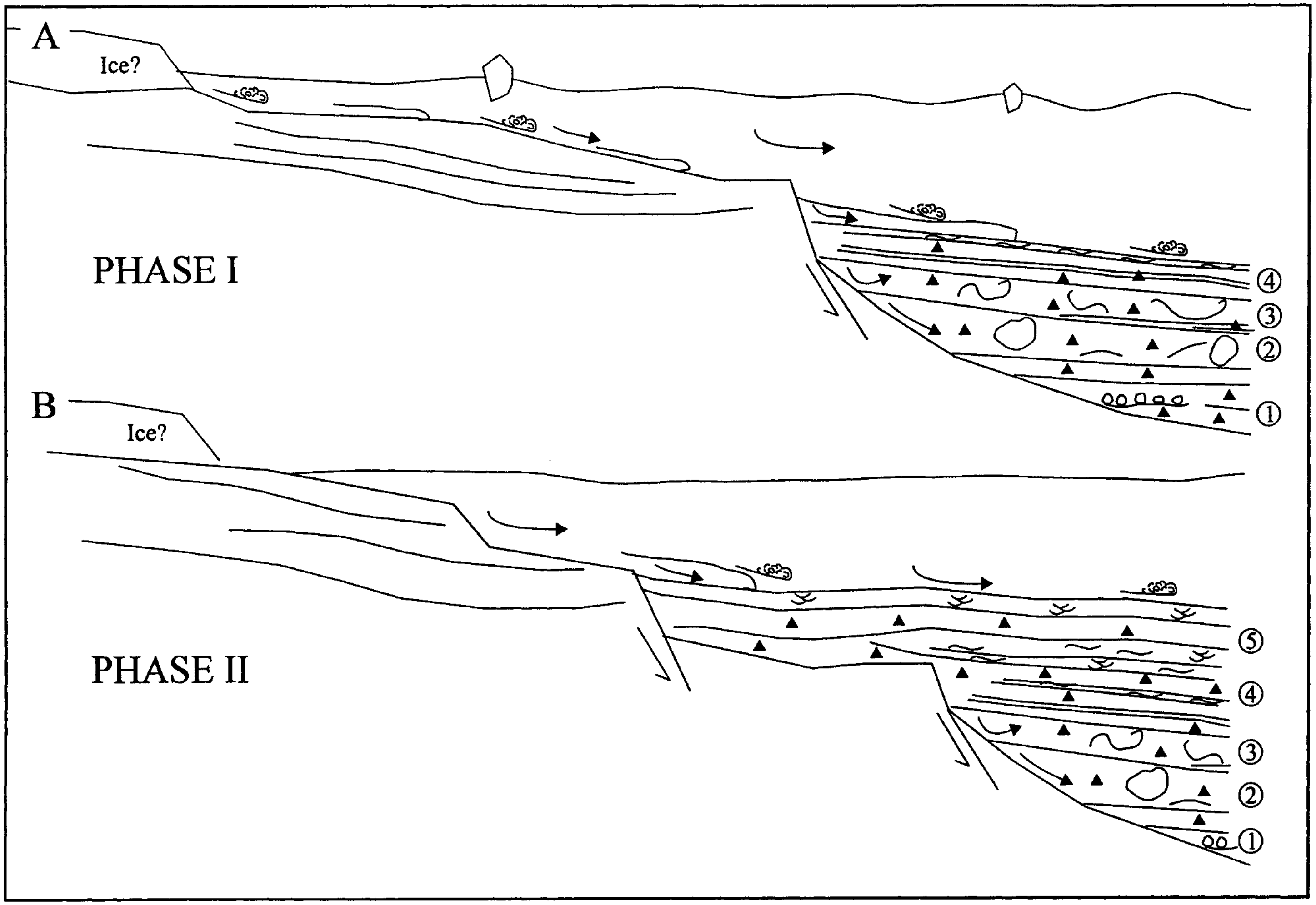


Fig. 4.20: Model showing depositional environments for different phases of Port Askaig deposition. A) Phase I; B) Phase II. Circled numbers refer to stratigraphic units. See text for explanation.

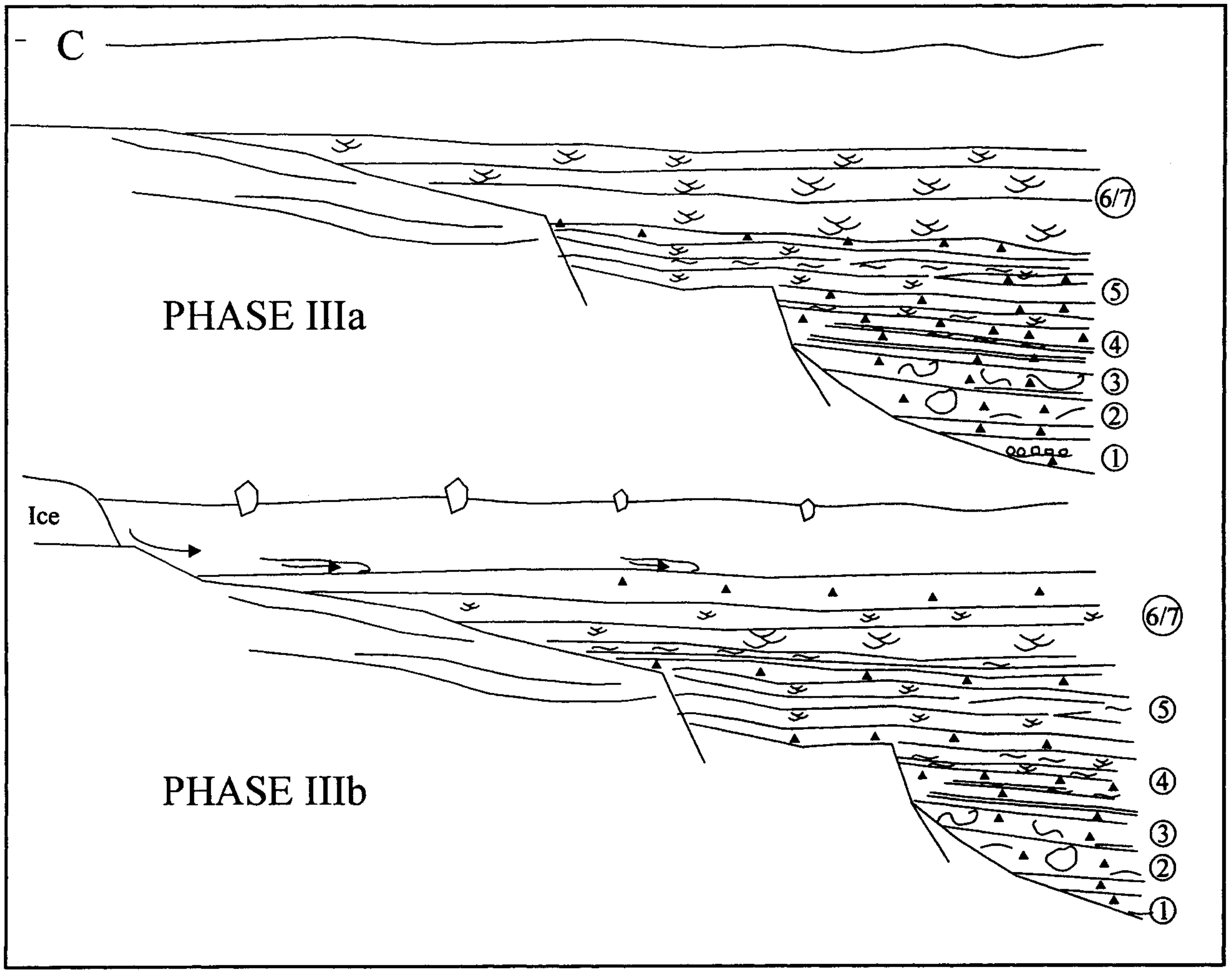


Fig. 4.20: Model showing depositional environments for different phases of Port Askaig deposition. C) Phase III. Based on modern analogues and the absence of HCS, water depths are inferred to be between 30-100 m (Table 4.1) for Phase III (a). Circled numbers refer to stratigraphic units. See text for explanation.

structures; this suggests either reworking of current-generated deposits during slumping or accumulation of sediments below wave base in relatively deep water. Repeated episodes of sediment slumping, possibly triggered by localized tectonic activity are recorded in the deposits of SU1, SU2 (the Great Breccia) and SU3 (the Disrupted Beds). Common horizons of deformation within SU4 are also attributed to seismic activity and provide evidence for continued, although intermittent, tectonic influences on sedimentation.

The basal contact of Phase I deposits with the underlying Islay Formation marks the onset of coarse-grained deposition in the basin. The increased delivery of coarse-grained debris to the basin margin and the abundance of sediment gravity flow facies within Phase I may have been controlled by climatic or tectonic factors, or a combination of both. Glaciation of the basin margins can contribute large volumes of coarse-grained debris to the marine environment, which may be reworked downslope as sediment gravity flows. Depositional oversteepening caused by high rates of sedimentation at the basin margin and rapid glacio-eustatic sea level changes may also have triggered resedimentation events.

However, a distinct glacial influence on sedimentation is difficult to confirm in this earliest phase of Port Askaig deposition as sedimentation is dominated by sediment gravity flow processes and evidence for the presence of ice is limited to rare dropstones and extrabasinal clasts found within SU3 (the Disrupted Beds) and SU4. Given the clear evidence for faulting and tectonic instability during Phase I, it is most likely that the onset of coarse-grained sediment deposition at the base of the Port Askaig Formation is at least in part related to local tectonic activity accompanying initial basin development.

4.5.2. Phase II (SU5)

The deposits contained within SU5 essentially record the transition from sediment gravity flow dominated deposition of locally-derived materials in SU1-SU4 (Phase I) to the predominance of current-deposited sandstone facies and farther-travelled materials in SU6-SU7 (Phase III). SU5 is therefore identified as the second phase (Phase II) of Port Askaig deposition (Fig. 4.20B). The basal contact of this phase marks the first appearance of current-generated structures in the succession and the preservation of relatively thick packages of sandstone and siltstone associated with the diamictites.

The presence of current-generated structures in sandstones of SU5 suggests that Phase II was characterized by relatively shallow-water conditions and/or more stable substrates (lower slopes?), where sandy bedforms could develop and be preserved. Continued, but intermittent tectonic activity within the basin is inferred from the abundance of sediment gravity flow facies in SU5 and occurrence of discrete horizons of soft sediment deformation features (Fig. 4.20B). Evidence for glacial influence on sedimentation is limited to rare dropstones in the uppermost mudstone unit of SU5, which may suggest minimal ice contact at the basin margin (Fig. 4.20B). The fining upward trend observed within SU5 records a change in the balance between accommodation space and sediment supply and may reflect either increased basin subsidence during

Phase II, eustatic sea level rise and/or a reduced supply of sediments to the basin.

4.5.3. Phase III (SU6-SU7)

The uppermost stratigraphic units of the Port Askaig Formation on the Garvellach Islands (SU6 and SU7) consist of thick successions of quartz-rich cross bedded sandstones interbedded with diamictites; these record the migration of large subaqueous dunes and glacially-influenced deposition by 'rainout' or sediment gravity flow processes on a relatively stable marine shelf (Fig. 4.20C). SU6 and SU7 characterize the third and final phase of Port Askaig deposition recorded on the Garvellach Islands. The conformable basal contact of Phase III identifies a shift in the nature and amount of sand being supplied to the basin as well as the development of relatively stable tectonic conditions on a current-dominated open marine shelf.

Soft sediment deformation features are rare in SU6 and SU7 and are associated with indicators of rapid sedimentation such as climbing ripples and massive sandstones; the sediment disturbance by faulting and seismic shaking that predominates in the previous two phases appears to have ceased by Phase III. Given that sediments deposited during Phase III appear to have accumulated in relatively tectonically stable marine depositional settings and that the sedimentary record is no longer 'overwhelmed' by sediment gravity flow processes, the lithological changes recorded in SU6/SU7 are more likely to record climatic influences on sedimentation.

The interbedding of thick units of current-deposited sandstone with glacially-influenced diamictite units is interpreted to reflect ice margin fluctuations during deposition of Phase III sediments (Fig. 4.20C). Thick packages of cross-bedded sandstones in SU6 and SU7 contain no coarse-grained or ice-rafted debris, suggesting ice-free depositional conditions (Phase IIIa; Fig. 4.20C). The sudden, and repeated lithological change from well-sorted sandstone to poorly-sorted coarse-grained diamictites and conglomerates most likely records ice advances into the marine basin (Phase IIIb; Fig. 4.20C). Several relatively thin diamictite units in SU6 and SU7 have been interpreted as the product of 'rainout' processes. Given that tectonic activity is not likely responsible for deposition of sediment gravity flow diamictites, these are attributed to downslope remobilization of coarse-grained debris delivered to the basin margin by ice and/or remobilization of 'rainout' diamict. Thus, unlike the diamictites deposited during Phase I and II, the delivery of coarse-grained sediments and formation of diamictites in SU6 and SU7 may indeed record a climatic signal.

The abundant supply of well sorted sands during Phase III probably reflects development of a major fluvial feeder system on the basin margin. Large quantities of sands were probably delivered onto the shelf by seaward-returning relaxation flows, flood discharge jets or rip currents; these land-derived sands were then reworked by geostrophic currents on the outer shelf (e.g. south African shelf, Johnson and Baldwin, 1996).

4.6. DISCUSSION: PALEOGEOGRAPHIC IMPLICATIONS

Allostratigraphic analysis of the Port Askaig Formation exposed on the

Garvellach Islands reveals several major shifts in sediment supply and depositional conditions during accumulation of the succession. Locally-derived coarse-grained materials deposited by sediment gravity flow processes dominate at the base of the succession and are replaced up section by farther-travelled materials and thick sandstones that accumulated under current-dominated or glacially-influenced marine conditions. In addition, a shift in the predominant controls on succession development can be identified that shows a change from tectonically-dominated controls in the earliest phase of Port Askaig deposition to more stable tectonic conditions and intermittently glacially-influenced controls in the later phase (Table 4.2).

Assessing the sedimentological evidence for a glaciation as severe as that proposed by the snowball Earth hypothesis requires reconstruction of the nature of glacial influences preserved in the Neoproterozoic sedimentary record and an understanding of the severity and extent of glaciation at this time. In order to do this, an assessment of the relative importance of climate and tectonic controls on regional sedimentation patterns must be made.

4.6.1. Glacial *versus* tectonic controls on regional sedimentation patterns

Phase I of the Port Askaig Formation on the Garvellach Islands is dominated by deposition by sediment gravity flow processes and tectonic instability. No faults or fault traces are preserved due to the limited outcrop; however, sedimentary characteristics indicate existence of local faults. The characteristics of the Great Breccia, the Disrupted

Beds and the repeated intervals of synsedimentary soft sediment deformation, strongly suggest localized tectonic activity has played a significant role in controlling the nature of sedimentation.

On a regional scale, there are also significant thickness changes between various outcrops of the Port Askaig Formation in Scotland, which provide further evidence for tectonic controls on stratigraphy (Anderton, 1982, 1985). The thickness of the Great Breccia varies from 50 m on the Garvellach Islands to only 4 m on the island of Islay, which lies 50 km to the southwest (see Fig. 4.1 for location, Spencer, 1971). In addition, over 100 m of Port Askaig and Islay Formation sediments exposed on the Garvellach Islands below the Great Breccia are absent on Islay, where the Great Breccia unconformably overlies Islay Formation carbonates (c.f. Fig. 4.3; Spencer, 1971).

This sedimentological and stratigraphic evidence is consistent with existing models for the Dalradian Basin, which identify an ensialic basin, made up of fault-bounded sub-basins affected by localized extensional tectonics (Harris *et al.,* 1978; Anderton, 1982; Harris *et al.,* 1993). The changing tectonic conditions recorded by the allostratigraphy of the Port Askaig Formation are also similar to those resulting from initial basin formation in other glacial and non-glacial extensional settings (Schermerhorn, 1974; Nystuen, 1985, Young, 1988; Eyles, 1993; Ravnås and Steel, 1998; Gawthorpe and Leeder, 2000); initial fault activity and basin differentiation is recorded by abundant coarse-grained material followed by tectonic quiescence and infilling with relatively fine-grained facies.

A distinct glacial signature on sedimentation is obscured by the predominance of sediment gravity flow deposits in most of Phase I. However, the occurrence of several diamictites on the Garvellach Islands interpreted as the product of 'rainout' processes, rare dropstones and the presence of far-travelled clasts in several stratigraphic units suggests glacial conditions did have at least some influence on sedimentation during Port Askaig times. Glaciation may also explain the delivery of abundant coarse-grained sediments to the basin margin, thus allowing sediment supply to keep up with local subsidence rates.

Schermerhorn (1974) argued that tectonic activity was much more important than climate change in controlling the nature of Neoproterozoic successions; he attributed evidence for limited glacial influence in some successions to localized mountain glaciation of uplifted basin margins. Given the evidence for tectonic activity and the limited evidence for glacial influence on sedimentation in the Port Askaig Formation, this study appears to support Schermerhorn's (1974) contentions. The diamictites of the Port Askaig Formation most likely record sediment instability related to tectonic activity and basin development, and localized glaciation of basin margins, rather than severe regional or global glaciation.

4.6.2. The Port Askaig Formation and the snowball Earth hypothesis

While glaciation may have had an influence on sedimentation of the Port Askaig Formation, there does not appear to be any evidence for a glaciation as severe as the one proposed by the snowball Earth hypothesis. The snowball Earth hypothesis, as recently proposed by Hoffman *et al.* (1998b), suggests Neoproterozoic glaciations were global, lasting between 4-30 My, with thick ice covering the oceans. The proposed severity of any one of these glaciations would lead to an almost non-existent hydrological cycle, thin, patchy and slow-moving ice on land, and consequently, sediment starvation in marine basins (Hoffman *et al.*, 1998b, Hoffman and Schrag, 2000). Deglaciation following snowball Earth conditions is predicted to be extremely rapid (century scale) due to feedback processes associated with strong evaporation and high CO_2 levels, such that very little sediment is expected to be deposited prior to ensuing hot house conditions and cap carbonate deposition on marine shelves (Christie-Blick *et al.*, 1999).

The Port Askaig Formation consists of a thick succession of 47 diamictites interbedded with sandstones, conglomerates and mudstones, which record a wide variety of depositional conditions. Deposition and preservation of thick and variable sediment types in a relatively shallow marine setting, as well as the alternating ice-free and ice-influenced depositional conditions inferred for SU6/SU7 sediments, are incompatible with a completely frozen earth, a reduced hydrological cycle and sediment starvation of marine basins proposed by the snowball Earth hypothesis (Hoffman *et al.*, 1998b; Hoffman and Schrag, 2000).

The delivery of abundant coarse-grained material to the basin suggests erosion of dolomitic and granitic substrates and an active hydrological cycle. In addition, the absence of subglacial deposits or erosion surfaces within the succession on the Garvellach

Islands indicates that glaciation was unlikely to have been severe. The nature of the Port Askaig Formation sediments suggests restricted ice volumes contributing sediment and occasional icebergs to the basin margin. This, together with evidence for high rates of sediment delivery, is most consistent with temperate glacially-influenced or non-glacial marine settings; these conditions are at odds with the frozen earth of the snowball Earth hypothesis, which is more likely to show similarities to modern polar glacial settings characterized by repeated erosion of the continental shelf and low sedimentation rates (Anderson, 1999).

The formation of large subaqueous dunes under the influence of geostrophic currents is also inconsistent with a frozen earth scenario. Geostrophic currents require open water conditions under the influence of storm winds and storm set up. Geostrophic currents would likely have been suppressed under 'normal' Phanerozoic glacial conditions (e.g Howe, 1995); they are highly unlikely to occur during 'snowball Earth' conditions when the oceans are predicted to have a 1 km thick cover of sea ice.

Lastly, the nature of the transition from glacially-influenced conditions of the Port Askaig Formation to postglacial cap carbonates of the Bonahaven Formation is also inconsistent with the snowball Earth hypothesis. The uppermost member of the Port Askaig Formation exposed on Islay and described by Spencer (1971) is characterized by thick sandstones and three very thin diamictites. The uppermost diamictite (D47) lies approximately 150 m below the contact with the 'cap' carbonate (Fig. 4.3), not immediately below the carbonate as predicted by the snowball Earth hypothesis (Hoffman *et al.*, 1998b). Although the sedimentation rate is unknown, the 150 m of sandstones separating the diamictite and overlying dolomites are unlikely to represent the rapid postglacial meltback phase proposed in the snowball Earth hypothesis.

4.7. CONCLUSIONS

The Neoproterozoic time period appears to record significant paleoenvironmental changes, which have been considered paradoxical and enigmatic. The snowball Earth hypothesis is a non-uniformitarian explanation for the various unusual features of the Neoproterozoic geological record. The Neoproterozoic Port Askaig Formation exposed on the Garvellach Islands consists of a thick succession of diamictite units interbedded with conglomerates, sandstones and mudstones, which accumulated by various processes in a glacially-influenced marine setting. This lengthy glacigenic succession contains a wealth of information about the nature of paleoenvironmental change at this time, and can be used to test the snowball Earth hypothesis.

Detailed sedimentological analysis has shown that the majority of diamictites on the Garvellach Islands were formed by sediment gravity flow processes and a distinct glacial influence on their formation cannot be discerned. Relatively few of the diamictites are interpreted as 'rainout' deposits, which together with rare dropstones, record the presence of floating ice in the basin. Associated non-diamictite facies were deposited as sediment gravity flows or under current-dominated conditions. Localized tectonic activity is inferred from the Great Breccia, and the Disrupted Beds and the intervals of abundant soft sediment deformation features, which are interpreted to record episodic seismic activity. Giant cross bedded sandstones in the upper part of the succession formed as a result of the migration of large subaqueous dunes; these most likely accumulated under the influence of geostrophic currents in an outer shelf setting.

The nature of paleoenvironmental change identified through allostratigraphic analysis allowed 3 main phases of Port Askaig deposition to be established. The Port Askaig Formation records a change from an early stage of sediment gravity flow-dominated depositional conditions (Phase I and II) to a later stage where glacially-influenced sedimentation alternates with ice-free current-dominated depositional conditions (Phase III).

Diamictites of the Port Askaig Formation do not necessarily record severely glacial conditions, and their deposition is more likely related to tectonic activity and basin development. The three phases of Port Askaig deposition are broadly similar to those identified from other extensional margin settings in their early stages of basin development (Råvnas and Steel, 1998). Glacial influence on sedimentation is relatively limited and can only be unambiguously identified in Phase III, where ice margin fluctuations are recorded by interbedded cross-bedded sandstones and diamictites.

The snowball Earth hypothesis suggests the Neoproterozoic earth was completely ice-covered, implying a reduced hydrological cycle and sediment-starved marine basins. Thick successions of cross-bedded sandstones interbedded with diamictites, indicate high rates of sediment delivery, deposition under geostrophic currents and limited glacial influence, and do not support the snowball Earth hypothesis.

CHAPTER FIVE

SUMMARY AND CONCLUSIONS

5.1. INTRODUCTION

The Neoproterozoic glacigenic record is considered to be characterized by a number of enigmatic and paradoxical features such as the close association of diamictites (glacial?) and carbonates (tropical?) facies and extreme fluctuations in $\delta^{13}C$ values. The non-uniformitarian snowball Earth hypothesis was most recently proposed to explain these distinctive features and proposes that the Neoproterozoic Earth underwent severe glaciation events followed by rapid climate change to 'hot house' conditions (Hoffman *et al.*, 1998b). Snowball Earth glaciation events are hypothesized to have been much more severe than later Phanerozoic glaciations and involve glaciation of landmasses and oceans from polar to equatorial latitudes.

The snowball Earth hypothesis has generated much debate regarding the severity of Neoproterozoic glaciation and the nature and causes of climate change at this time in earth history (e.g. Williams and Schmidt, 2000; McMechan, 2000, Hoffman, 2000). In particular, there is concern as to whether diamictite-bearing intervals found in many Neoproterozoic successions can be interpreted as representing globally synchronous and/or severe glaciation events. While Neoproterozoic diamictites have often been attributed to deposition under glacial conditions, these deposits can also form in a variety of non-glacial and glacially-influenced depositional settings and cannot be axiomatically ascribed to severely cold climate settings (Eyles, 1993; Crowell, 1999). In many instances, diamictites actually record sediment instability related to tectonic activity and basin development rather than glacial conditions and climatically-controlled environmental change (Schermerhorn, 1974).

Sedimentological and allostratigraphic analysis of lengthy and well exposed Neoproterozoic successions in northern Norway and Scotland were carried out in order to better define the paleoclimatic significance of diamictites and the nature of Neoproterozoic paleoenvironmental change preserved in these successions.

5.2. SUMMARY AND SIGNIFICANT FINDINGS

5.2.1. Smalfjord Formation, northern Norway

Sedimentological analysis of the Smalfjord Formation in two areas of northern Norway (Tarmfjorden and Varangerfjorden; Fig. 2.1) focused on characterizing the lateral and vertical distribution of facies types, establishing the sedimentary characteristics of diamictites, the nature of facies with which diamictites are associated, and the outcrop geometry of different sedimentary units.

The sedimentary characteristics of diamictites and their close association with turbidites suggest that diamictites in the Tarmfjorden area formed as a result of sediment gravity flow processes; the presence of rare lonestones in fine-grained facies indicates the presence of some form of floating ice in the basin. The succession exposed in the Varangerfjorden area indicates accumulation of coarse-grained debris by sediment gravity flow processes on a prograding subaqueous debris apron adjacent to the margin of the Baltic craton. An isolated diamictite mound overlying a striated pavement at Bigganjargga, which was previously interpreted as a subglacial 'tillite', is interpreted here as a far-travelled subaqueous debris flow based on its sedimentary characteristics and facies associations. Deposition of the diamictite may not be related to formation of the underlying striated pavement.

Sedimentary evidence for glacially-influenced conditions at the time of Smalfjord deposition is limited to the occurrence of rare lonestones at Tarmfjorden, rare striated and faceted clasts in the diamictites of Varangerfjorden (Bjørlykke, 1967) and the striated pavement underlying the Smalfjord Formation at Bigganjargga. The paleoclimatic significance of rare lonestones in the succession at Tarmfjorden is difficult to establish as these may have been rafted by seasonal or glacial ice; similarly, clast characteristics reflect sediment source rather than depositional conditions and cannot be used to identify glacial depositional processes. While it seems likely that the striated pavement at Bigganjargga was formed by subglacial erosion of relatively lithified sediments, subglacial conditions could have existed a considerable time prior to the deposition of the overlying diamictite by subaqueous sediment gravity flow processes (Rice and Hofmann, 2000).

Thus, while Smalfjord Formation deposition may have been influenced by glacial conditions at the basin margin, there is no unambiguous evidence for direct deposition by glacial ice at Tarmfjorden or Varangerfjorden as originally proposed by previous authors (e.g. Edwards, 1984 and references therein). The role of tectonic activity in controlling sedimentation during Smalfjord deposition remains unclear due to limited outcrop, but the extensional tectonic setting of the Gaissa basin, the predominance of sediment gravity flow facies in the succession and the local presence of breccias and fault scarps, all hint at a probable tectonic influence on sedimentation.

5.2.2. The Great Breccia, Port Askaig Formation, Scotland

The Great Breccia is a distinctive diamictite unit within the Neoproterozoic Port Askaig Formation of Scotland that contains matrix-supported megaclasts up to 100's m in diameter; the Great Breccia was previously interpreted as a subglacial terrestrial deposit and used to infer the presence of substantial ice sheets in the region during the Neoproterozoic (Spencer, 1971). New sedimentological data presented in this study has resolved the depositional origin of the Great Breccia as a sediment gravity flow deposit formed by catastrophic failure of sediments along a subaqueous fault scarp.

Detailed sedimentological analysis and mapping of outcrops on the Garvellach Islands of Scotland was used to demonstrate that the Great Breccia consists of three separate sedimentary units: a megabreccia, a unit of diamictites interbedded with other sedimentary facies, and a diamictite unit characterized by the predominance of m-sized dolomitic clasts. The characteristics and distribution of these three units suggest that they formed by successive subaqueous sediment gravity flows as a result of localized faulting and catastrophic mass failure of previously deposited calcareous and siliciclastic deposits.

These findings provide strong evidence for tectonic controls on sedimentation, at least in this part of the Port Askaig Formation. The sediment gravity flow origin of the Great Breccia makes it difficult to identify any glacial influences on sedimentation, which may have coincided with this depositional event.

5.2.3. The Port Askaig Formation, Garvellach Islands, Scotland

Sedimentological analysis of outcrops of the Port Askaig Formation on the Garvellach Islands has helped constrain the depositional origin of diamictites and associated conglomerate, sandstone and fine-grained facies. The most significant contribution of this study is the use of these sedimentological data to develop an allostratigraphy for the Port Askaig Formation, which in turn can be used to assess the relative importance of tectonic and climatic influences on sedimentation.

Port Askaig Formation diamictites are attributed to deposition by sediment gravity flow processes or 'rainout' of fine-grained sediments and ice-rafted debris in a glacially-influenced marine setting. Associated facies represent deposition by various depositional processes ranging from sediment gravity flows (conglomerates, massive sandstones and laminated mudstones) to deposition under unidirectional currents (cross-bedded sandstones). In particular, giant cross-bedded sandstones in the upper part of the Port Askaig Formation are interpreted as recording the migration of large subaqueous dunes under the influence of geostrophic currents in an outer shelf setting. This study also documents abundant soft sediment deformation features at discrete intervals throughout the Port Askaig Formation and it is argued that these deformation features represent disturbance by episodic seismic activity.

Allostratigraphic analysis reveals three phases of Port Askaig deposition. In Phase I, deposition is dominated by sediment gravity flow processes and sedimentation is primarily tectonically-controlled. Evidence for glacial influence on sedimentation is limited; rare dropstones, extrabasinal clasts and the delivery of abundant coarse-grained sediment in this early phase may be attributed to the onset of glaciation on the basin margin. However, the predominance of sediment gravity flow facies makes it difficult to confirm a glacial influence on sedimentation.

Phase II of Port Askaig deposition is characterized by continued sediment gravity flow processes, an increased supply of sands to the basin and the preservation of facies formed under the influence of unidirectional currents. The abundance of soft sediment deformation features in Phase II deposits suggests continued episodic seismic activity.

In the third and final phase of Port Askaig deposition on the Garvellach Islands, the formation of thick units of cross-bedded sandstones with very few soft sediment deformation features suggests development and accumulation of large sandy bedforms under the influence of geostrophic currents in an ice-free, tectonically-stable marine setting. Diamictites deposited during Phase III are interpreted as having formed by glacially-influenced sediment gravity flow and 'rainout' processes; their occurrence as interbeds within thick sandstone deposits is argued to reflect ice margin fluctuations in a tectonically-stable basin.

5.2.4. Glacial influences on sedimentation and the snowball Earth hypothesis

Sedimentological analyses of both the Smalfjord and Port Askaig formations emphasize the ambiguous paleoclimatic significance of Neoproterozoic diamictites and the need for careful consideration of the depositional origin of associated facies when attempting to reconstruct past climate change. Deposition of similar diamictite facies by glacial, non-glacial or glacially-influenced marine processes often makes it difficult to discriminate a glacial origin or infer ice mass form or distribution. Whereas both the Smalfjord and Port Askaig formations are well exposed, the much thicker succession in Scotland allows more detailed analysis of the nature of paleoenvironmental changes over time and of the likely controls on these changes.

Allostratigraphic analysis of the Port Askaig Formation demonstrates that tectonic activity particularly during the early phases of basin development, had a significant influence on facies types and the nature of facies successions preserved on the Garvellach Islands. Sediment instability associated with tectonic activity can generate deposits very similar to those accumulating in glacial settings and may obscure or make it difficult to identify a glacial signal. In contrast, a climatic signal can be more easily discerned during periods of tectonic quiescence (e.g. Phase III of the Port Askaig Formation). The findings presented in this study underscore the need to examine glacigenic deposits in their stratigraphic context in order to determine the degree to which paleoclimatic influences may be obscured by basin tectonics.

The predominance of sediment gravity flow facies in the Smalfjord Formation makes it difficult to reconstruct paleoclimatic conditions sufficiently to test the validity of the snowball Earth hypothesis. However, no characteristics of the Smalfjord Formation in the Tarmfjorden or Varangerfjorden area suggest the existence of severely cold conditions (e.g. subglacial tillites) or a reduced hydrological cycle. In the Port Askaig Formation, the thick succession of diamictites interbedded with current-deposited sandstones is not consistent with the low sedimentation rates, reduced hydrological cycle and severe glacial conditions envisioned by the snowball Earth hypothesis. The Smalfjord and Port Askaig formations are not unlike other Phanerozoic glacially-influenced marine successions, suggesting that uniformitarian paleoclimatic models may be more appropriate to account for their development.

5.3. IMPLICATIONS FOR FUTURE WORK

The findings presented in this study have important implications for our understanding of Neoproterozoic climate change and for future research carried out on Neoproterozoic glacigenic successions.

5.3.1. Neoproterozoic climate change

This study suggests that relatively 'normal' glacially-influenced marine conditions prevailed in the Gaissa and Dalradian basins during the Neoproterozoic. This raises the question of how 'unusual' and distinctive Neoproterozoic glaciations really were. If they were similar to Phanerozoic glaciations, then similar climate change models could apply and non-uniformitarian models such as the snowball Earth hypothesis, would no longer be required.

In terms of establishing the paleoenvironmental significance of Neoproterozoic diamictites, this study and others have shown that many Neoproterozoic diamictites may not actually record fully glacial conditions and cannot be used to map past ice sheet configurations; the widespread distribution of Neoproterozoic diamictites may therefore not imply global or even regional glaciation. The age of Neoproterozoic diamictites is also very poorly constrained despite advances in isotope geochronology and chronostratigraphy (Kaufmann *et al.*, 1997; Kennedy *et al.*, 1998; Knoll, 2000; Gorokhov *et al.*, 2001). The temporal correlation of diamictites preserved in different Neoproterozoic basins is therefore unwarranted; diamictites could represent diachronous localized glaciations similar to those preserved in Permo-Carboniferous basins of Gondwanaland (Eyles, 1993; Crowell, 1999), rather than global glacial events.

There is still much uncertainty about the paleoenvironmental significance of carbonates associated with diamictites. Not all carbonates record deposition under warm, tropical conditions and, similarly, not all diamictites record deposition under glacial conditions; hence, the common association of carbonates and diamictites in Neoproterozoic successions may not necessarily represent a climatic paradox.

Furthermore, arguments for low latitude glaciation during the Neoproterozoic are based on paleomagnetic data that are widely considered to be unreliable; the few well documented cases for low latitude glaciation may be explained by recent climate models that show that ice can form at equatorial latitudes without having recourse to a completely frozen earth (Chandler and Sohl, 2000; Hyde *et al.*, 2000).

The Neoproterozoic geochemical record contains some 'unusual' features, such as extreme excursions in $\delta^{13}C$ values and the deposition of BIFs that suggest significant fluctuations in seawater geochemistry. However there are several competing hypotheses which explain their occurrence without requiring snowball Earth conditions, including overturn of stagnant glacial oceans, release of methane and precipitation of iron due to the combined effect of thermohaline circulation and metal-rich brine production at hydrothermal vents (Knoll *et al.*, 1986; Kennedy *et al.*, 2001a; Young, 1988).

The distinctive nature of the Neoproterozoic geological record may not require extreme climate change scenarios such as the one proposed in the snowball Earth hypothesis. It is likely that a number of interconnected factors were responsible for the nature of Neoproterozoic environmental change including the break up of supercontinent Rodinia, changes in biogeochemical conditions, lower solar luminosity, and other factors particular to a vegetation-free Neoproterozoic Earth. However, climate change itself may have been very similar to Phanerozoic climate change.

Although there is still some debate concerning the causal relationship between tectonics and glaciation (Powell, 1995; Torsvik *et al.*, 1995), the break-up of the supercontinent Rodinia likely had a significant influence on the nature and preservation of Neoproterozoic sedimentary successions (Nystuen, 1985; Young, 1988; Eyles, 1993). The widespread distribution of Neoproterozoic diamictites may record sediment instability created by local tectonic activity and early basin development associated with the break up of the supercontinent Rodinia (Schermerhorn, 1974; Eyles, 1993). Hence, future work on Neoproterozoic glacigenic successions, should focus on establishing the influence of basin tectonics on sedimentation and the paleoclimatic significance of diamictite-bearing intervals in order to refine the depositional models proposed.

5.3.2. Neoproterozoic glacigenic successions in the North Atlantic region

In the North Atlantic region, the Smalfjord and Port Askaig formations have been broadly correlated with other Neoproterozoic glacigenic deposits in southern Norway, Greenland and Svalbard on the basis of lithological, biostratigraphic and radiometric data (Spencer, 1975; Hambrey, 1983). Similar to the Smalfjord and Port Askaig formations, these glacigenic successions range in thickness from several 10's of m to 100's of m, and consist of diamictites interbedded with conglomerates, sandstones and/or mudstones; these successions also accumulated in sedimentary basins on the margins of the proto-Iapetus Ocean (Eyles, 1993).

The Moelv Tillite of southern Norway consists of interbedded diamictites and mudstones with dropstones, which are interpreted as subglacial tillites and glaciomarine deposits respectively (Bjørlykke *et al.*, 1976); these sediments are closely associated with coarse-grained sediment gravity flow facies and accumulated in a fault-bounded rift basin (the Hedmark Basin; Bjørlykke *et al.*, 1976). The Moelv Formation was inferred to record fully glacial conditions based on its regional extent, and the presence of glacially-striated or faceted clasts and dropstones (Bjørlykke *et al.*, 1976).

The Neoproterozoic Ulvesø and Storeelv formations of Greenland consist of glacigenic diamictites interbedded with lenticular sandstones and conglomerates and are interpreted as subglacial, glaciomarine, glaciofluvial, glaciolacustrine, subaqueous and subaerial mass flow deposits (Hambrey and Spencer, 1987). The Ulvesø Formation also contains sandstone dykes or wedges similar to those in the Port Askaig Formation, which Spencer (1985) interpreted as periglacial indicators. Spencer (1985) suggested that the Ulvesø, Storeelv and Port Askaig formations accumulated under very similar depositional conditions of repeated advances of terrestrial ice sheets into a marine basin; however the succession in Greenland records only two glacial advances compared to 17 inferred for the Port Askaig Formation of Scotland.

Both the Elbobreen and Wilsonbreen/Sveanor formations of northeast Svalbard are thought to be broadly equivalent to the Port Askaig Formation, whereas the Smalfjord Formation has been directly correlated with the Elbobreen Formation (Hambrey,1983). The Petrovbreen Member (E2) of the Elbobreen Formation is of highly variable thickness (<2 to 42 m) and consists of diamictites, tabular and lens-shaped conglomerate deposits, and mudstones with abundant dropstones (Harland *et al.*, 1993). The Wilsonbreen/Sveanor formations attain much greater thickness (up to 150 m), and consist of diamictite, conglomerate, breccia, sandstone, siltstones, sandstone/breccia 'wedges' interpreted as periglacial features and cold-water stromatolites (Harland *et al.*, 1993). Both diamictite-bearing intervals in the Neoproterozoic succession of northeast Svalbard were interpreted as representing subglacial and glacially-influenced marine deposits formed in a tectonically stable basin (Harland *et al.*, 1993). The broadly equivalent Neoproterozoic succession in western Svalbard is dominated by glaciomarine mudstones with dropstones and subaqueous sediment gravity flow facies, which accumulated in an active extensional basin; post-depositional tectonic deformation has made paleoenvironmental reconstruction in this area difficult (Hambrey, 1983; Harland *et al.*, 1993).

Based on the lithological similarities of Neoproterozoic glacigenic successions throughout the North Atlantic region, a large ice sheet was inferred to have covered the area during the Varangian glacial period (Spencer, 1975; Hambrey and Spencer,1987). However, the sedimentological analysis of the Smalfjord and Port Askaig formations presented here shows that diamictites do not necessarily represent glacial conditions and that tectonic activity may have played a significant role in the deposition and preservation of these two glacigenic successions. Neoproterozoic diamictite-bearing successions in other parts of the North Atlantic region should therefore be re-examined in light of these findings to reassess their paleoclimatic significance and more realistically assess the past distribution of ice sheets. Allostratigraphic analysis of the thick successions preserved in Greenland and northeast Svalbard could help further refine the nature of environmental change during this time. The glacigenic successions of southern Norway and western Svalbard that accumulated in tectonically-active settings could be analysed to further explore the relationship between basin tectonics, climate change and sedimentation.

In sum, future work in Neoproterozoic glacial geology should explore the following two issues: 1) the timing of environmental change, with particular emphasis on the timing of glaciation in different basins and the temporal relationship between diamictites and associated carbonates, and 2) the paleoenvironmental significance of diamictites and the role of tectonics in controlling the nature and preservation of glacigenic successions. Constraining the timing of Neoproterozoic climate change depends on future developments in isotope geochronology, biostratigraphy, and chemostratigraphy. Improved geochronological control together with sedimentological and allostratigraphic analyses of other lengthy Neoproterozoic succession in the North Atlantic region and elsewhere will contribute to a better understanding of the nature and causes of Neoproterozoic glaciations.

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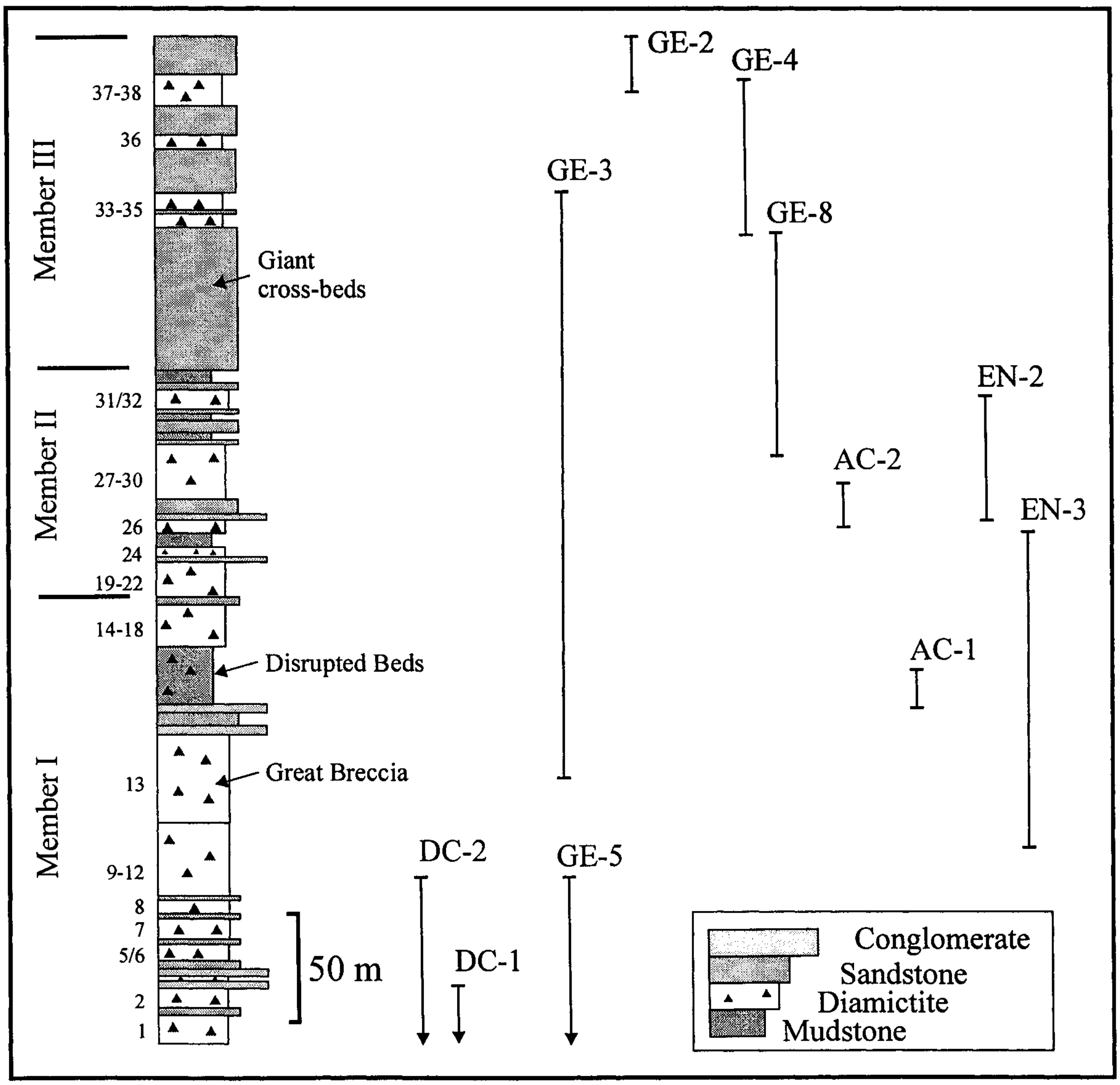
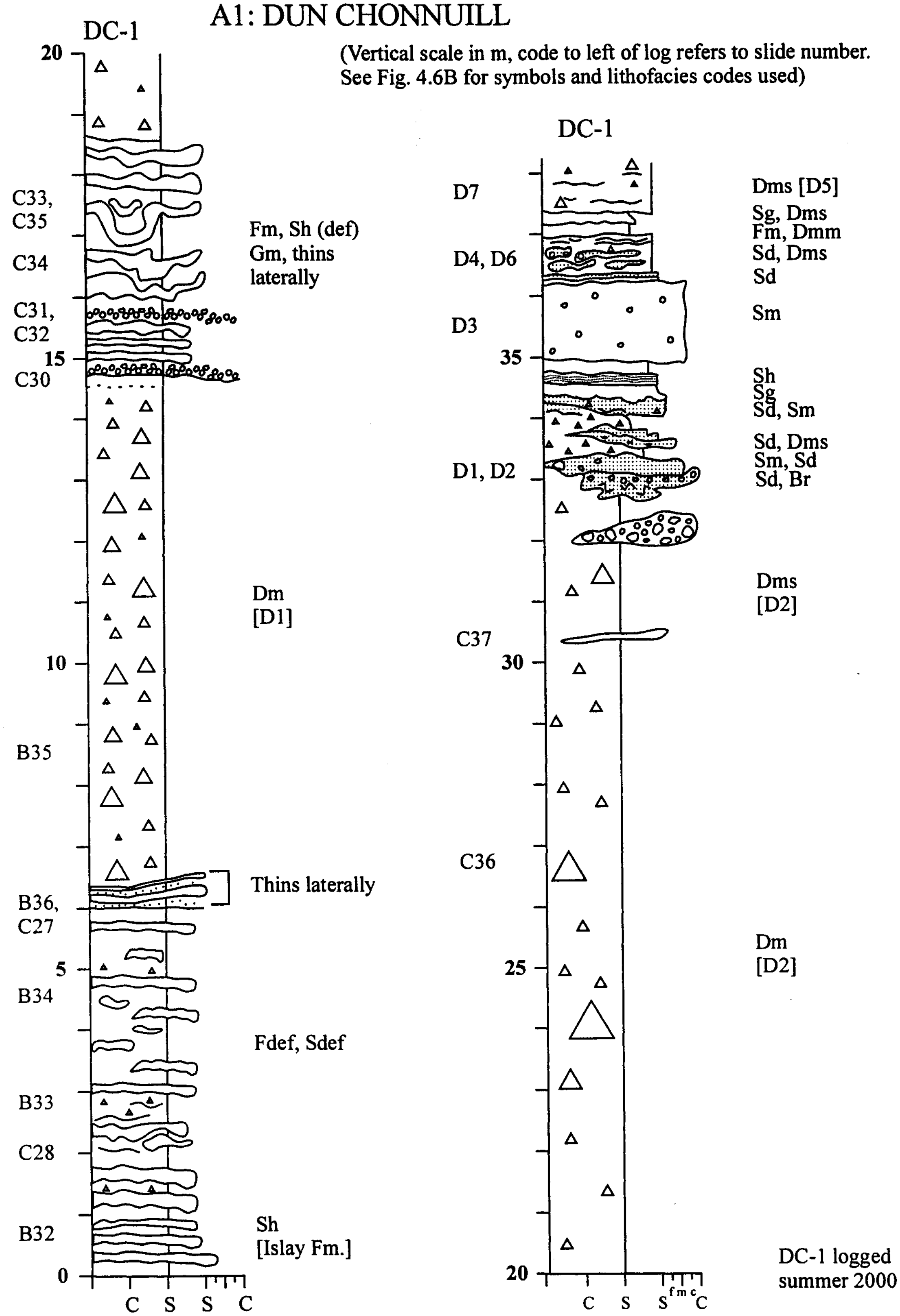
APPENDIX

GRAPHIC LOGS OF THE PORT ASKAIG FORMATION, GARVELLACH ISLANDS

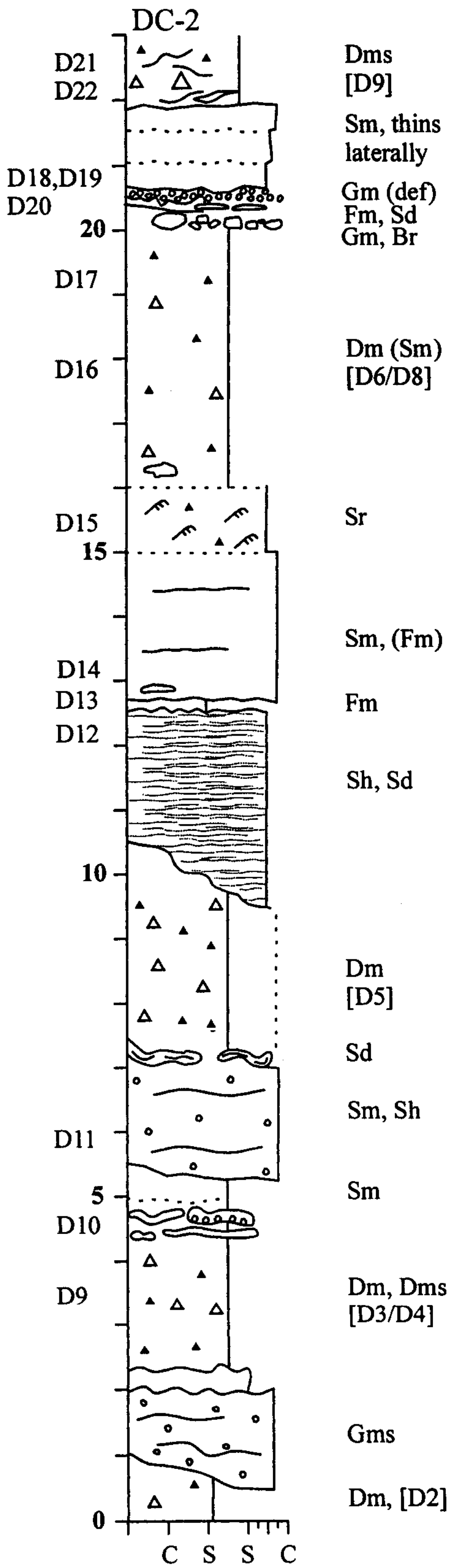


A1: DUN CHONNUILL

(Vertical scale in m, code to left of log refers to slide number. See Fig. 4.6B for symbols and lithofacies codes used)



DC-1 logged
summer 2000

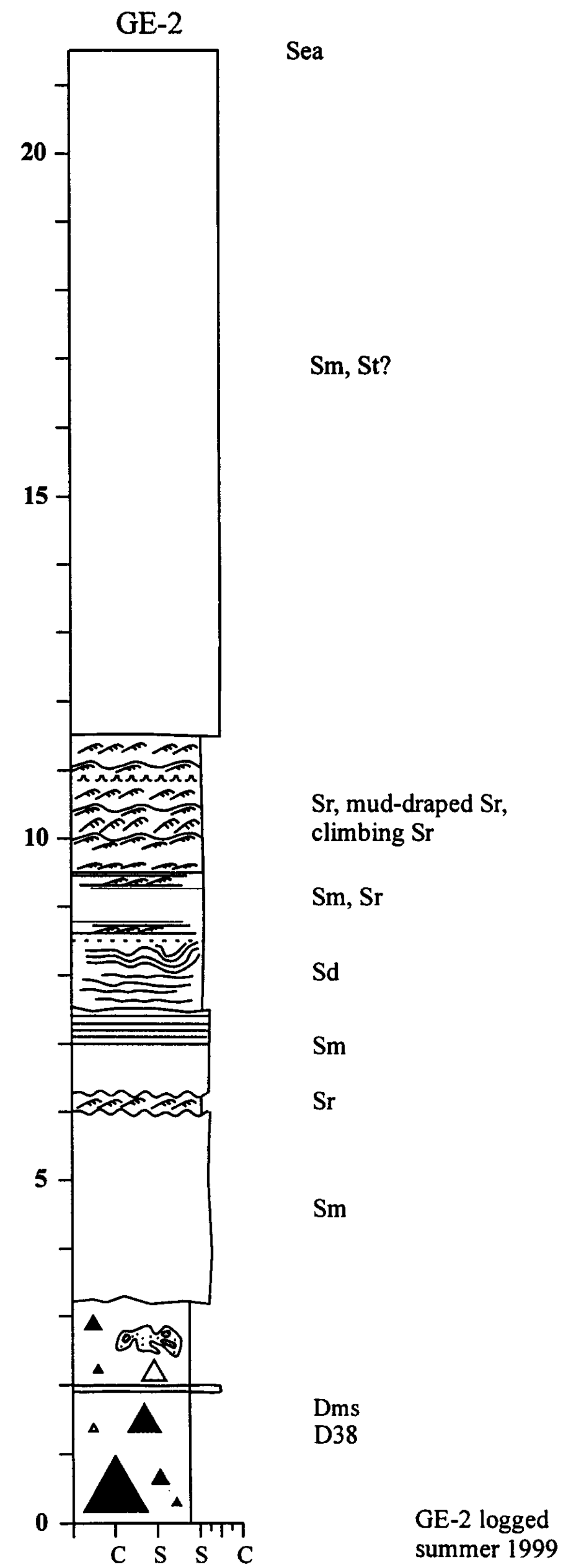


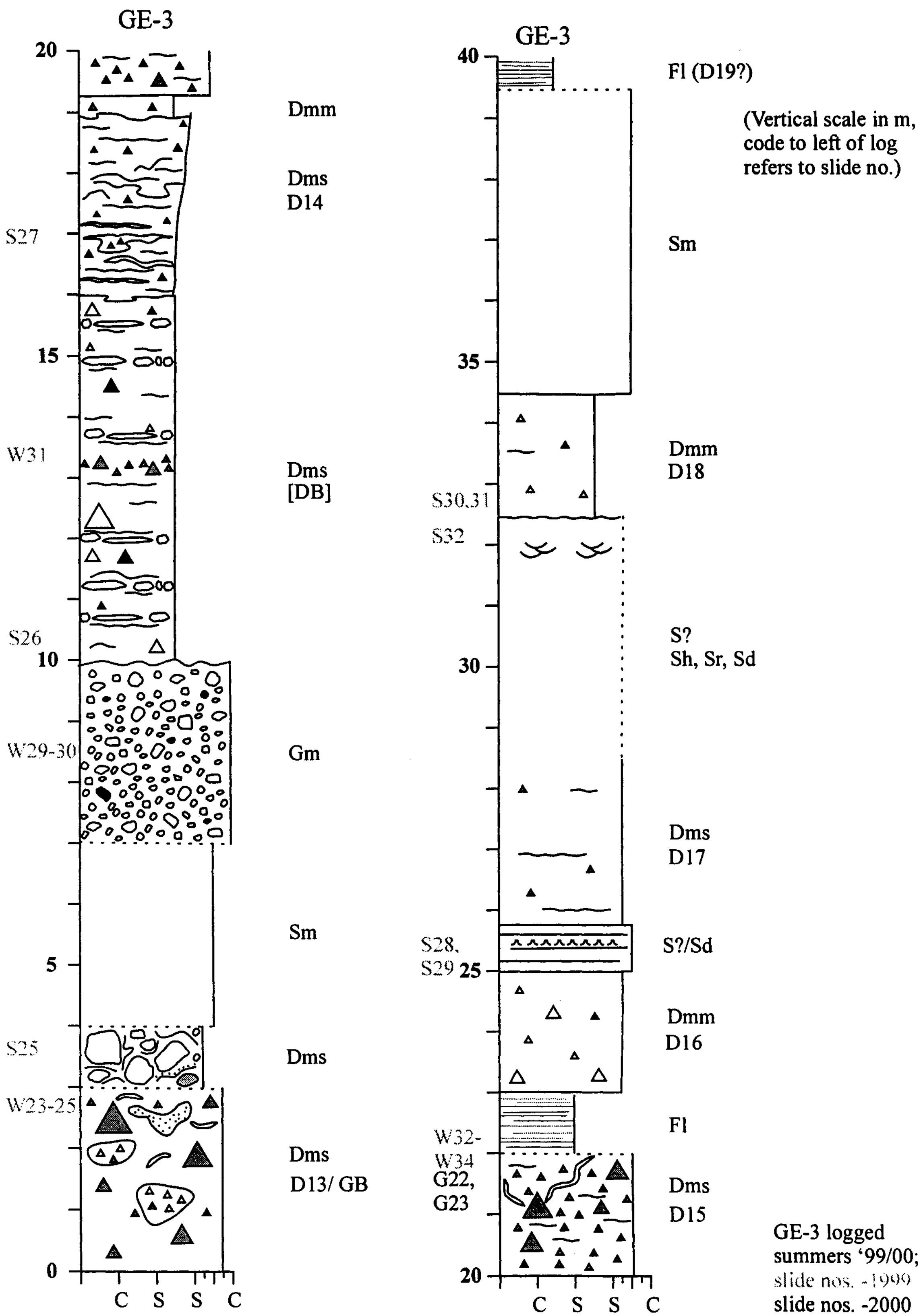
(Vertical scale in m,
code to left of log
refers to slide no.)

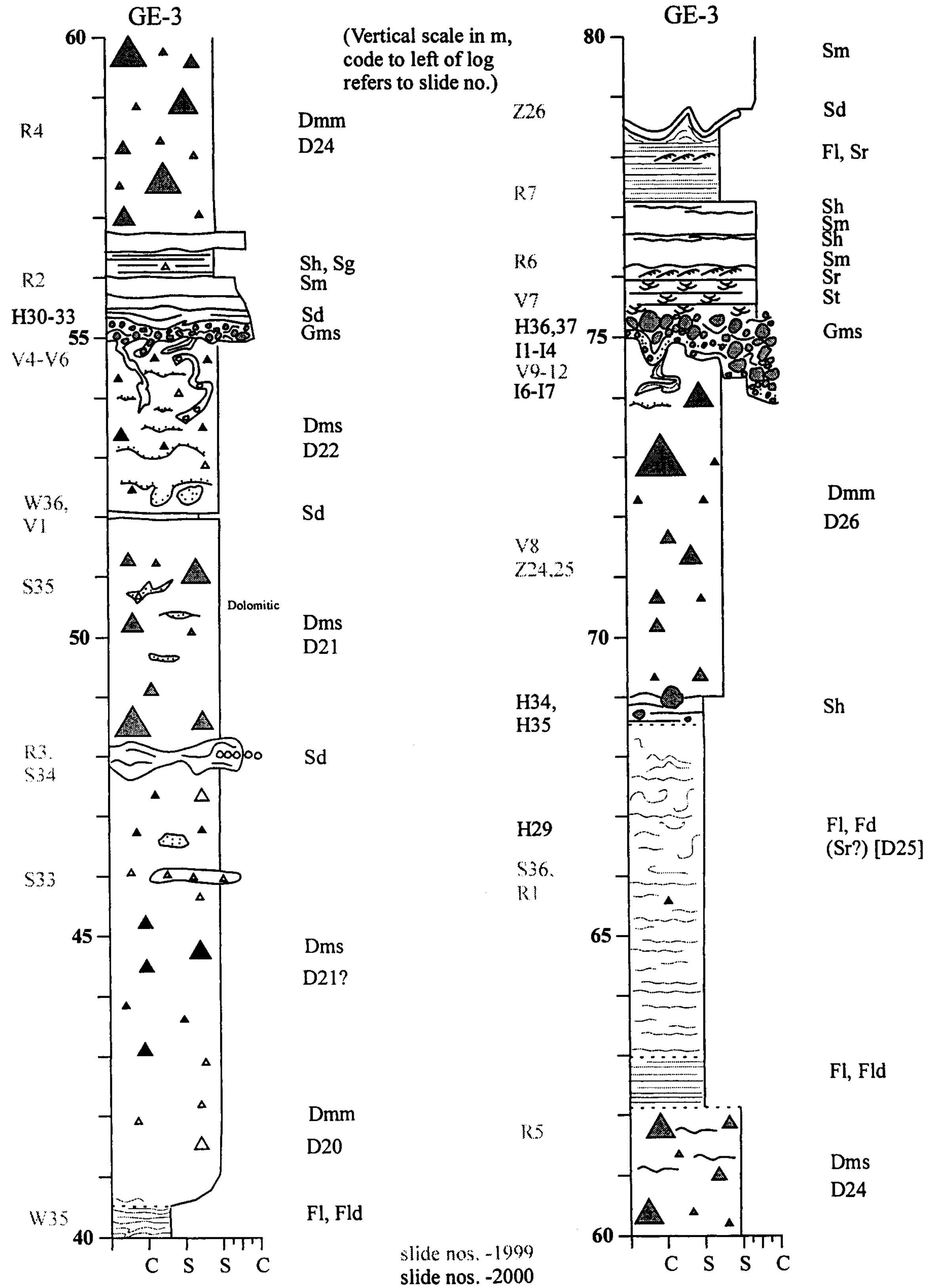
DC-2 logged
summer 2000

A2: GARBH EILEACH

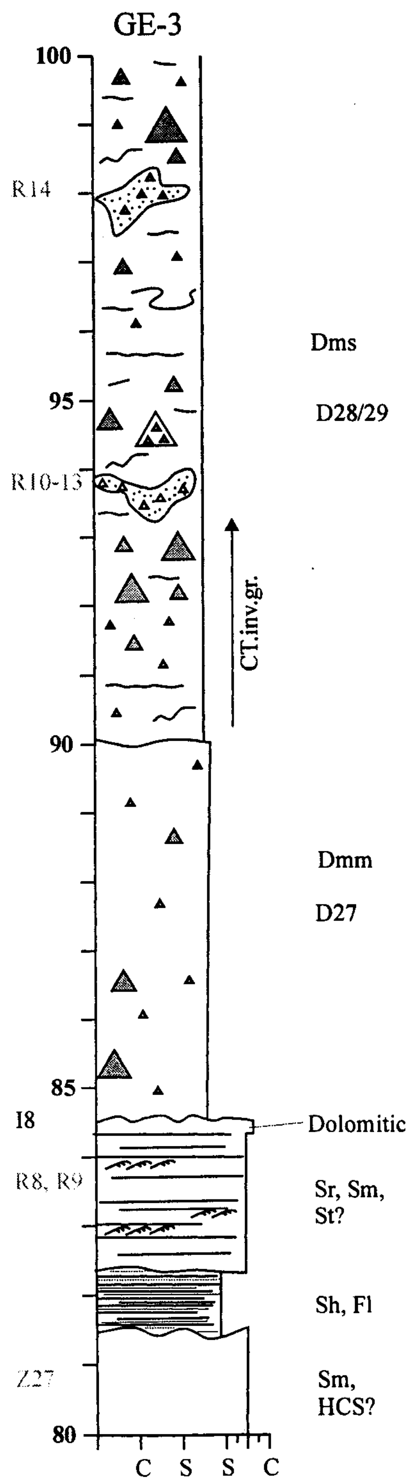
(Vertical scale in m, code to left of log refers to slide number. See Fig. 4.6B for symbols and lithofacies codes used)



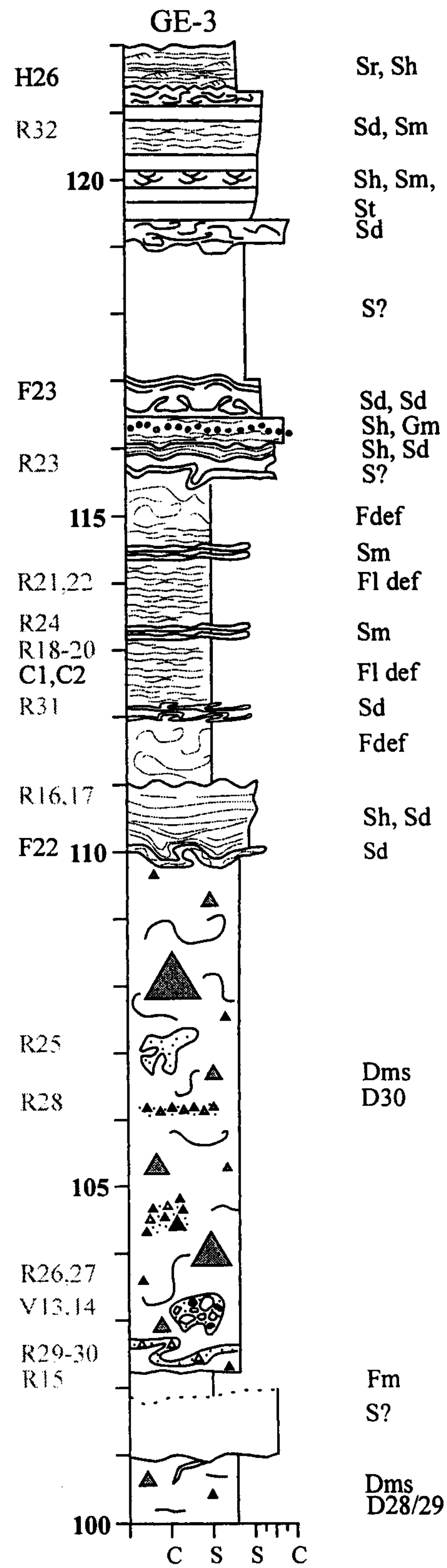

GE-3
20
15
10
5
0
S27
W31
S26
W29-30
S25
W23-25
Dmm
Dms
D14
Dms
[DB]
Gm
Sm
Dms
Dms
D13/ GB
C S S C
GE-3
40
35
30
25
20
Fl (D19?)
(Vertical scale in m, code to left of log refers to slide no.)
Sm
Dmm
D18
S30,31
S32
S?
Sh, Sr, Sd
Dms
D17
S28,
S29
S?/Sd
Dmm
D16
Fl
W32-
W34
G22,
G23
Dms
D15
C S S C
GE-3 logged summers '99/00;
slide nos. -1999
slide nos. -2000

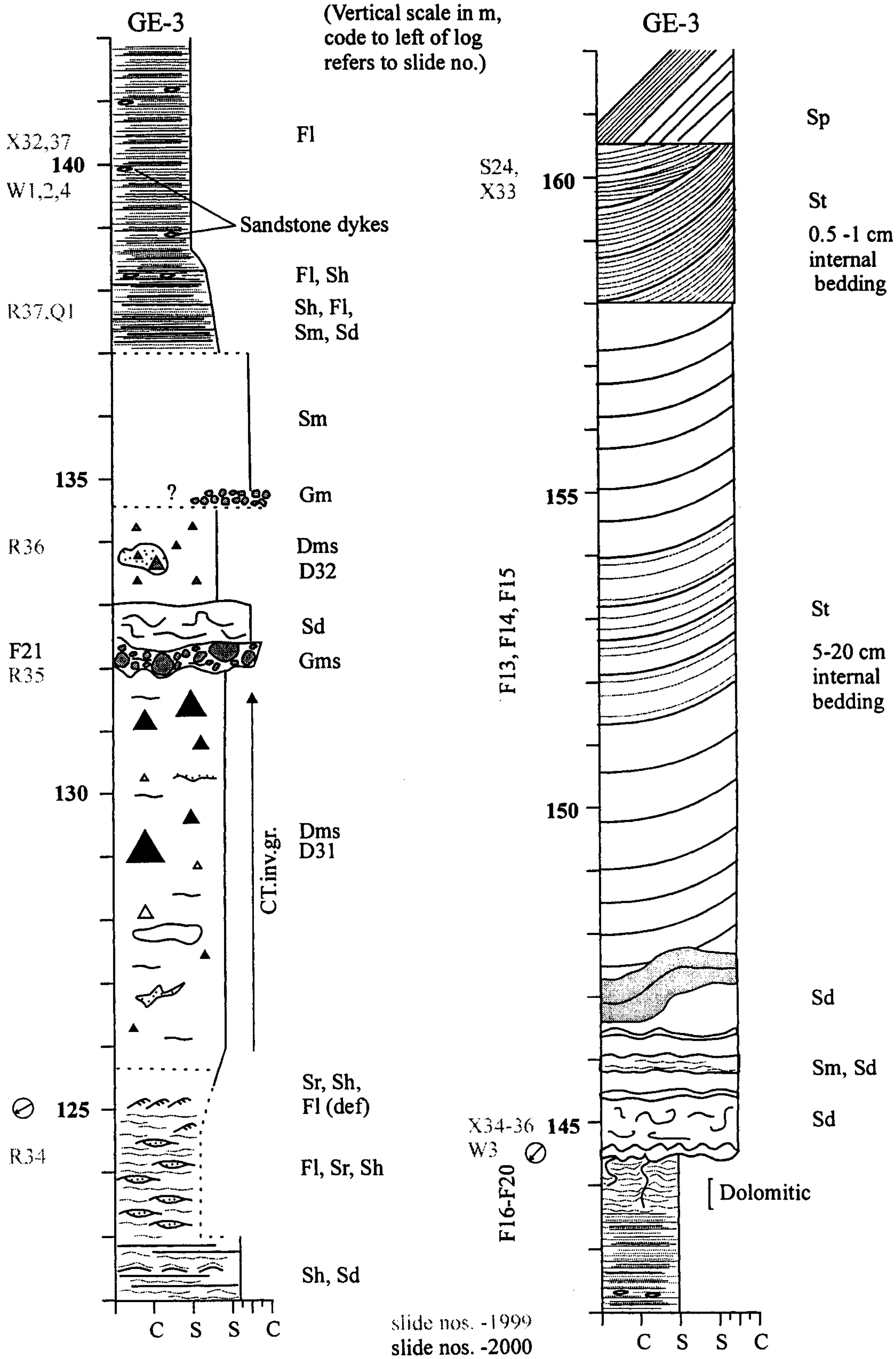

GE-3
(Vertical scale in m,
code to left of log
refers to slide no.)
60
R4
Dmm
D24
Sh, Sg
Sm
R2
H30-33
Sd
Gms
55
V4-V6
Dms
D22
W36,
V1
Sd
S35
Dolomitic
Dms
D21
50
R3,
S34
Sd
S33
45
Dms
D21?
Dmm
D20
W35
Fl, Fld
40
C S S C
GE-3
80
Sm
Z26
Sd
Fl, Sr
R7
Sh
Sm
Sh
R6
Sm
Sr
V7
St
H36,37
Gms
75
I1-I4
V9-12
I6-I7
Dmm
D26
V8
Z24,25
70
H34,
H35
Sh
H29
Fl, Fd
(Sr?) [D25]
S36,
R1
65
Fl, Fld
R5
Dms
D24
60
C S S C
slide nos. -1999
slide nos. -2000

(Vertical scale in m,
code to left of log
refers to slide no.)

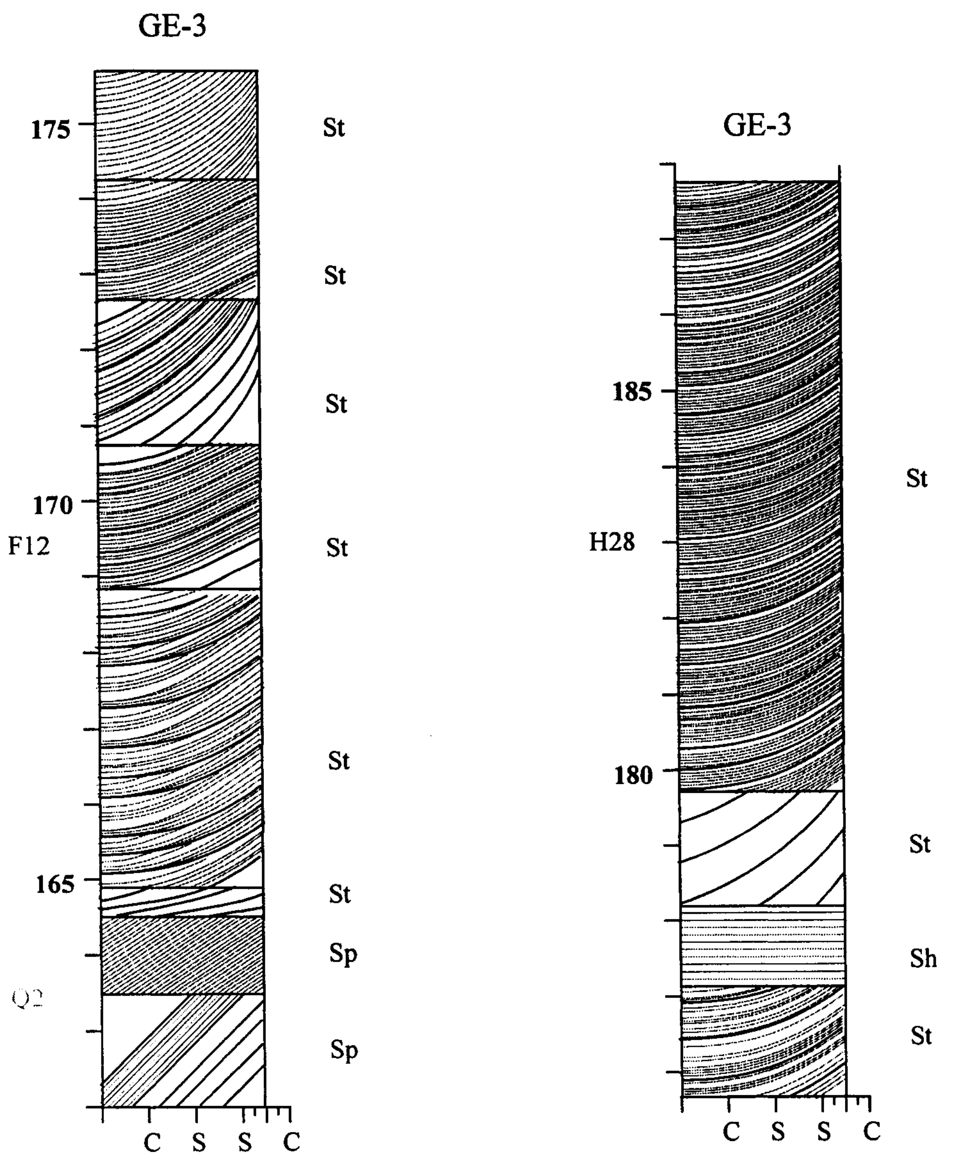


slide nos. -1999
slide nos. -2000



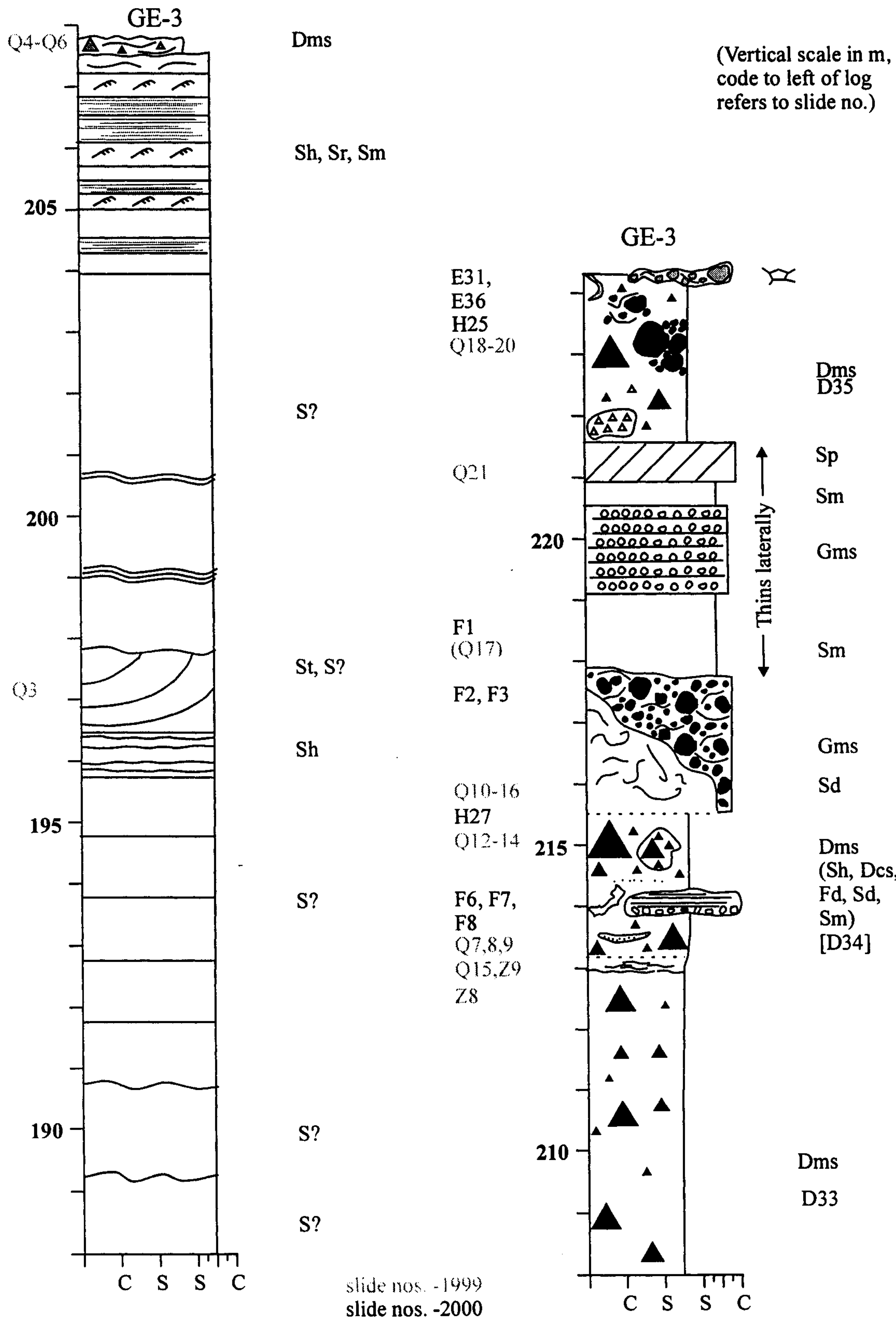

GE-3
(Vertical scale in m, code to left of log refers to slide no.)
X32,37
140
W1,2,4
Fl
Sandstone dykes
Fl, Sh
R37,Q1
Sh, Fl, Sm, Sd
Sm
135
?
Gm
R36
Dms D32
Sd
F21
R35
Gms
130
Dms D31
CT.inv.gr.
Sr, Sh, Fl (def)
125
R34
Fl, Sr, Sh
Sh, Sd
C S S C
GE-3
Sp
S24, X33
160
St
0.5 -1 cm internal bedding
155
F13, F14, F15
St
5-20 cm internal bedding
150
Sd
Sm, Sd
X34-36
W3
145
Sd
F16-F20
Dolomitic
C S S C
slide nos. -1999
slide nos. -2000

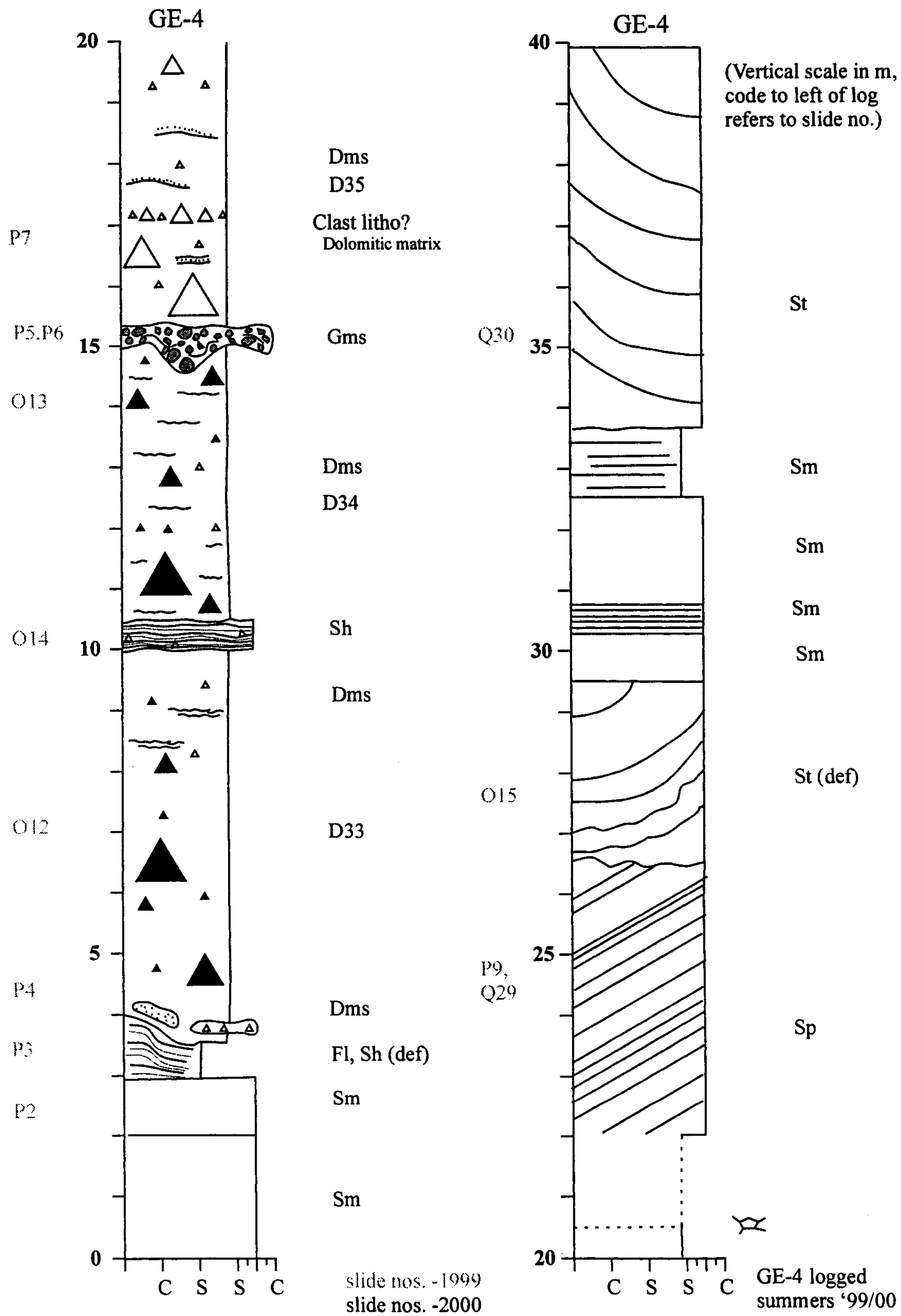
(Vertical scale in m,
code to left of log
refers to slide no.)



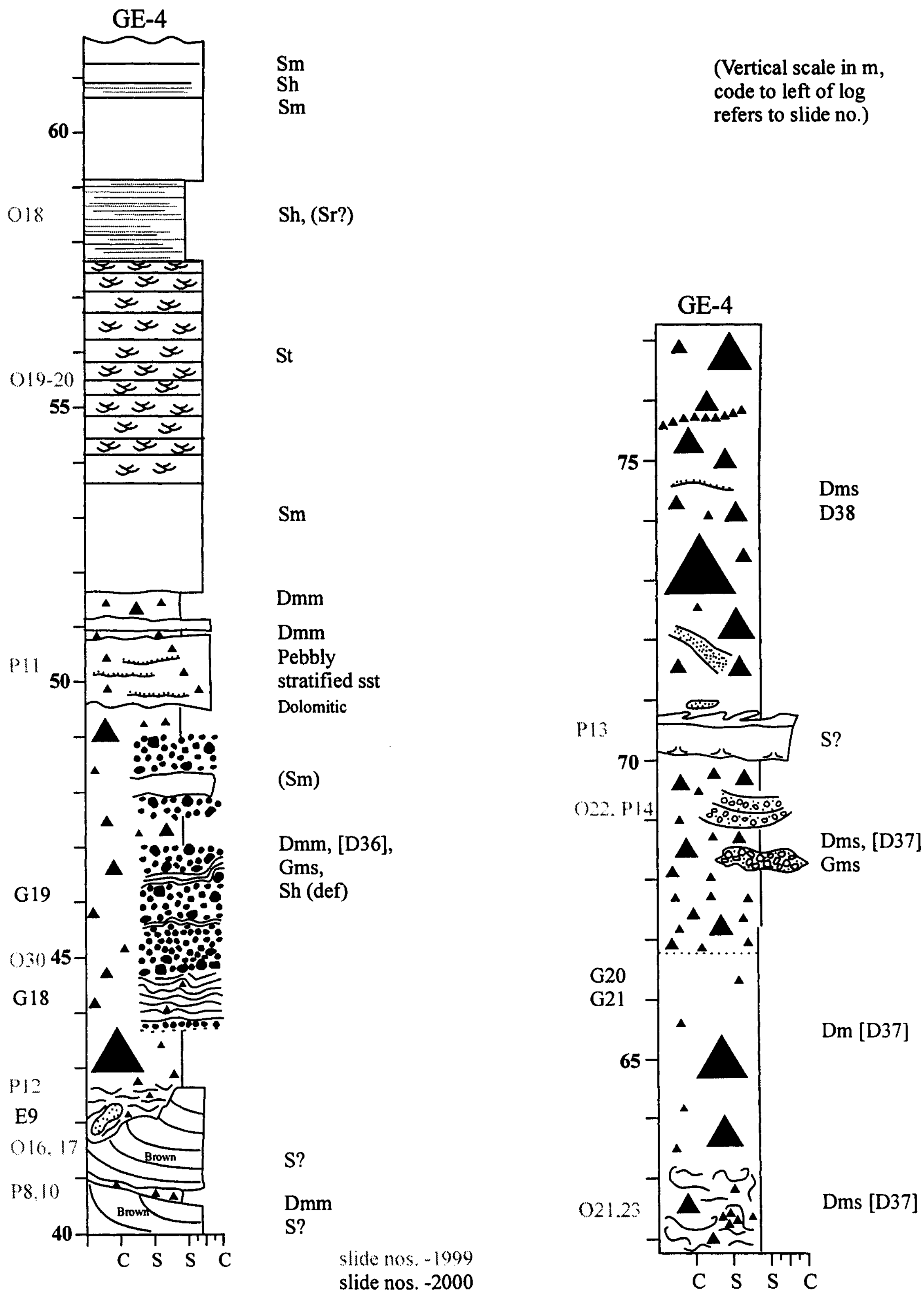
Note: paleocurrent directions for giant cross-bedded sandstones reported in Fig. 4.11.

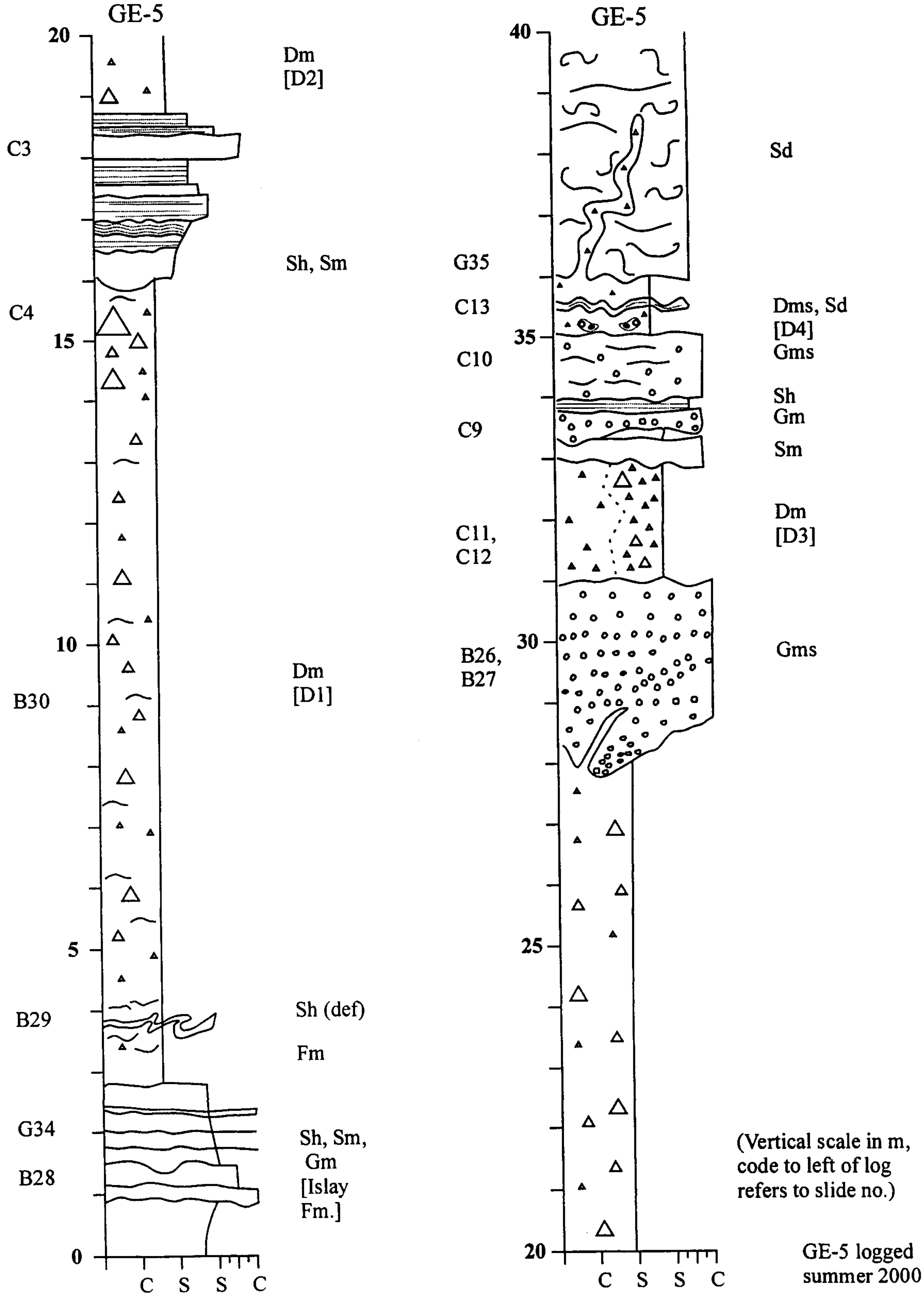
slide nos. -1999
slide nos. -2000

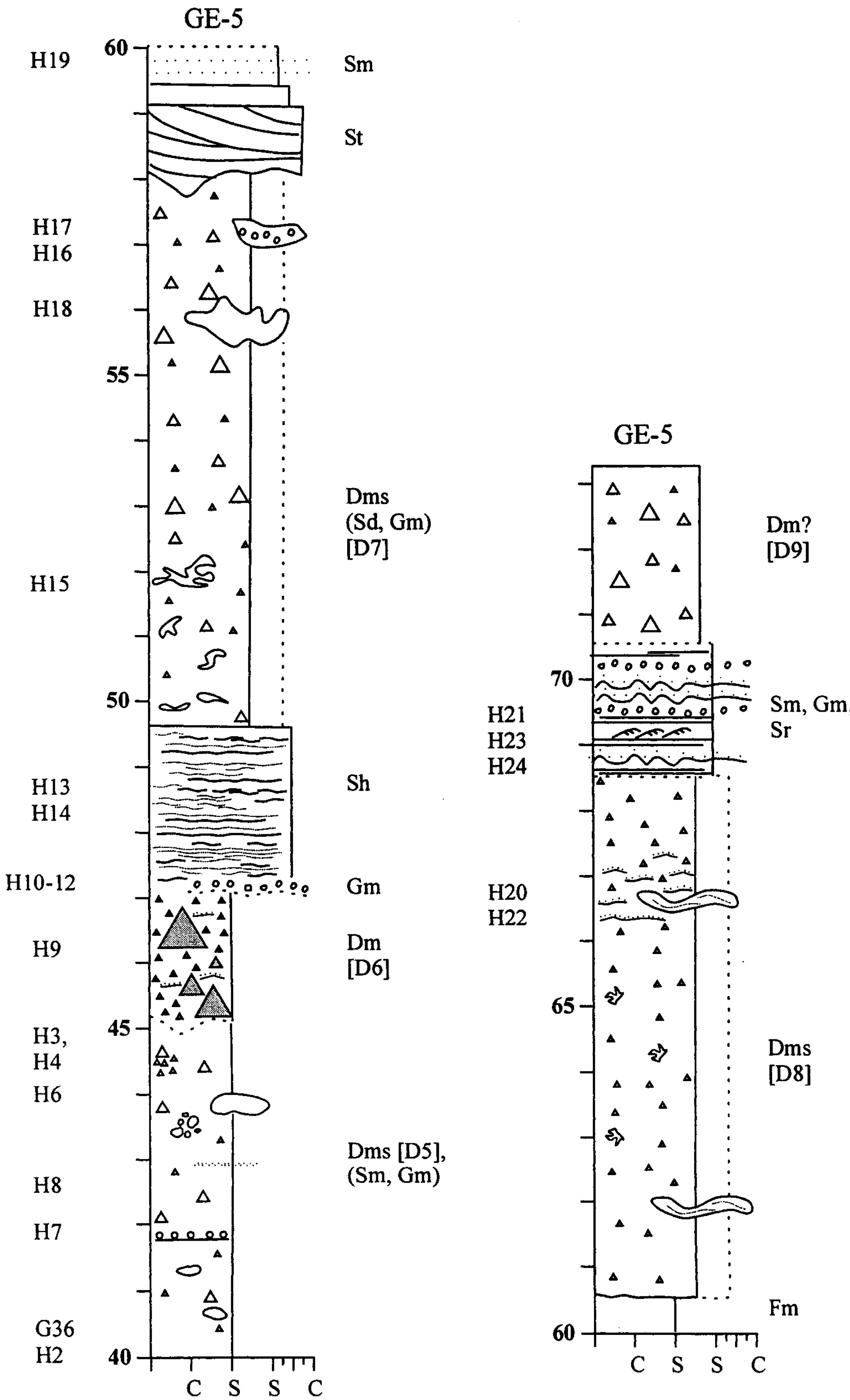

GE-3
Q4-Q6
Dms
Sh, Sr, Sm
205
S?
200
St, S?
Q3
Sh
195
S?
190
S?
S?
C S S C
slide nos. -1999
slide nos. -2000
(Vertical scale in m, code to left of log refers to slide no.)
GE-3
E31,
E36
H25
Q18-20
Dms
D35
Sp
Q21
Sm
220
Gms
Thins laterally
F1
(Q17)
Sm
F2, F3
Gms
Sd
Q10-16
H27
Q12-14
215
Dms
(Sh, Dcs,
Fd, Sd,
Sm)
[D34]
F6, F7,
F8
Q7,8,9
Q15,Z9
Z8
210
Dms
D33
C S S C

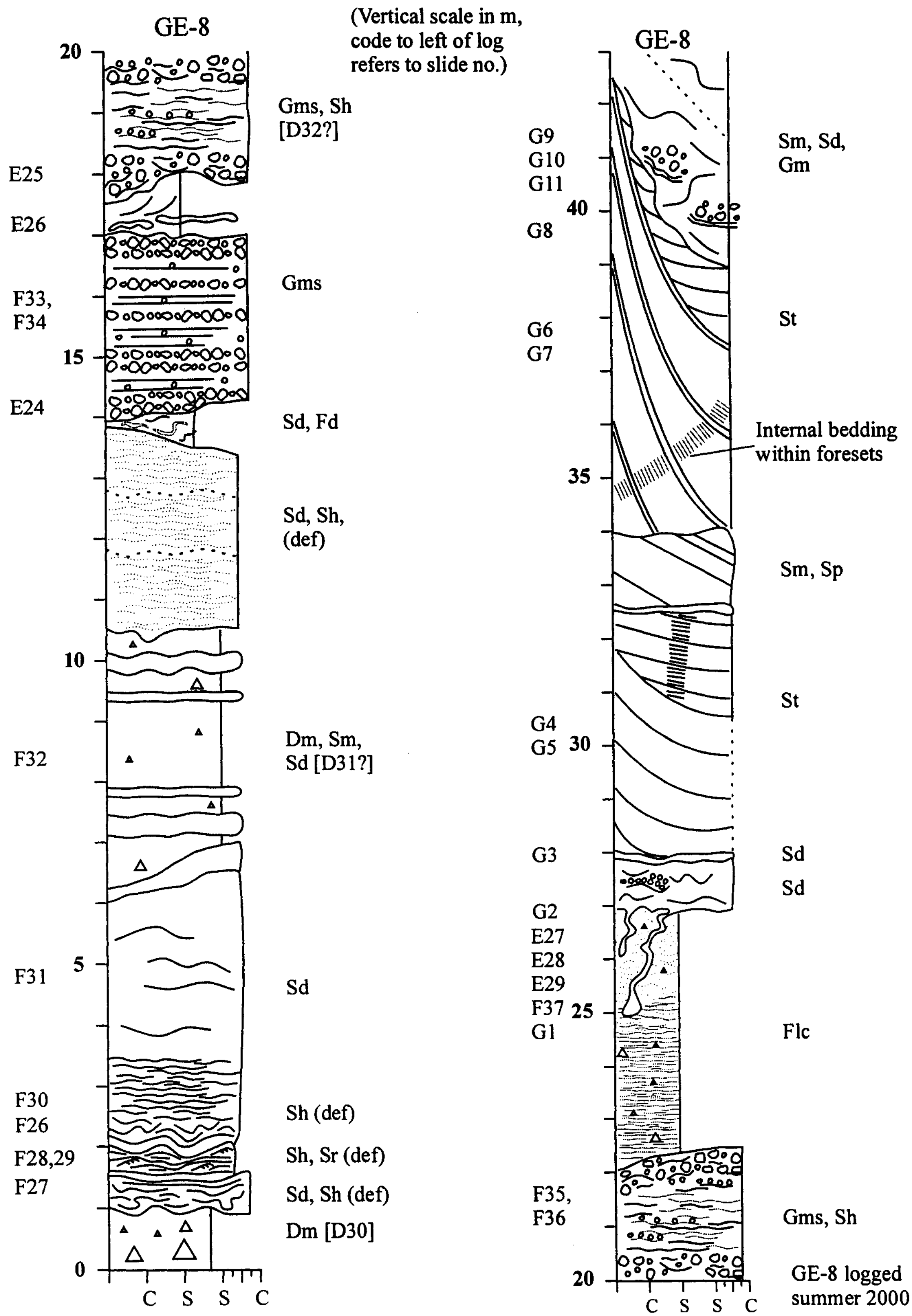


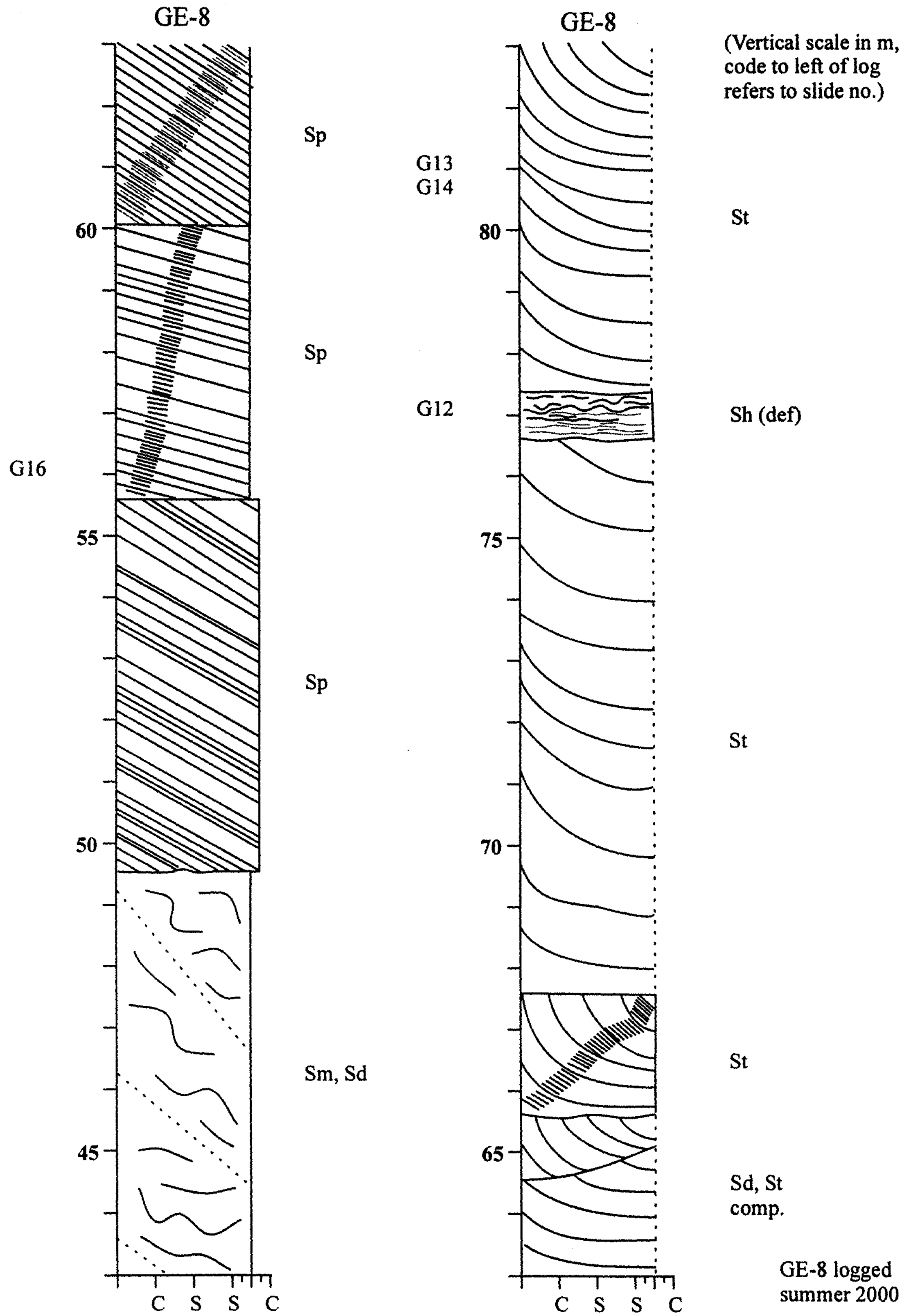
GE-4
20
15
10
5
0
P7
P5.P6
O13
O14
O12
P4
P3
P2
Dms
D35
Clast litho?
Dolomitic matrix
Gms
Dms
D34
Sh
Dms
D33
Dms
Fl, Sh (def)
Sm
Sm
C S S C
slide nos. -1999
slide nos. -2000
GE-4
40
35
30
25
20
(Vertical scale in m, code to left of log refers to slide no.)
Q30
O15
P9, Q29
St
Sm
Sm
Sm
Sm
St (def)
Sp
C S S C
GE-4 logged summers '99/00


GE-4
Sm
Sh
Sm
60
O18
Sh, (Sr?)
St
O19-20
55
Sm
Dmm
Dmm
Pebbly
stratified sst
Dolomitic
P11
50
(Sm)
Dmm, [D36],
Gms,
Sh (def)
G19
O30
45
G18
P12
E9
O16, 17
Brown
S?
P8,10
Brown
Dmm
S?
40
C S S C
slide nos. -1999
slide nos. -2000
(Vertical scale in m,
code to left of log
refers to slide no.)
GE-4
75
Dms
D38
P13
S?
70
O22, P14
Dms, [D37]
Gms
G20
G21
Dm [D37]
65
O21,23
Dms [D37]
C S S C

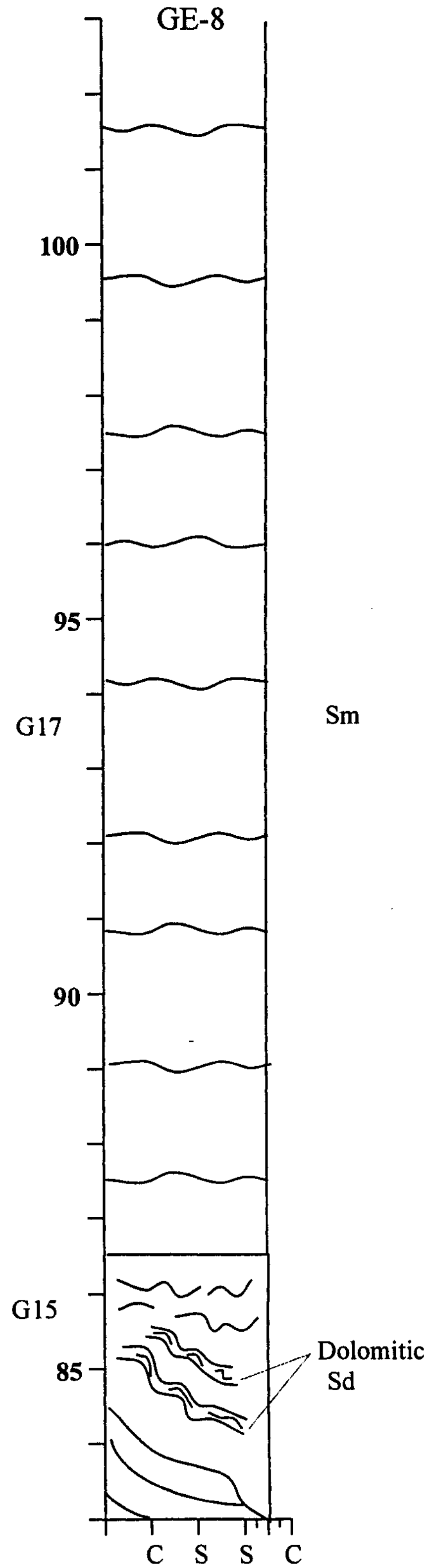

GE-5
20
15
10
5
0
C3
C4
B30
B29
G34
B28
Dm
[D2]
Sh, Sm
Dm
[D1]
Sh (def)
Fm
Sh, Sm,
Gm
[Islay
Fm.]
C S S C
GE-5
40
35
30
25
20
G35
C13
C10
C9
C11,
C12
B26,
B27
Sd
Dms, Sd
[D4]
Gms
Sh
Gm
Sm
Dm
[D3]
Gms
C S S C
(Vertical scale in m,
code to left of log
refers to slide no.)
GE-5 logged
summer 2000


GE-5
H19
60
Sm
St
H17
H16
H18
55
Dms
(Sd, Gm)
[D7]
H15
50
H13
H14
Sh
H10-12
Gm
H9
Dm
[D6]
H3,
H4
45
H6
Dms [D5],
(Sm, Gm)
H8
H7
G36
H2
40
C S S C
GE-5
Dm?
[D9]
70
H21
H23
H24
Sm, Gm,
Sr
H20
H22
65
Dms
[D8]
Fm
60
C S S C

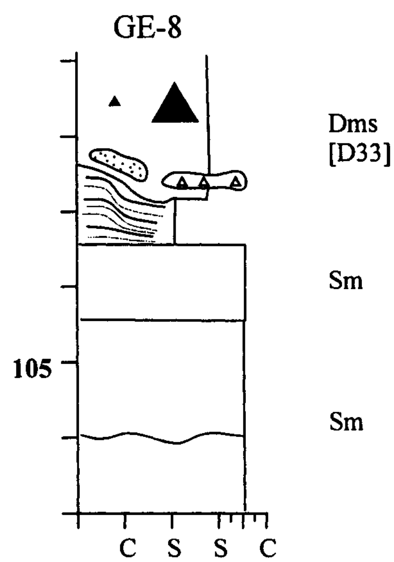

GE-8
(Vertical scale in m, code to left of log refers to slide no.)
20
15
10
5
0
E25
E26
F33, F34
E24
F32
F31
F30
F26
F28,29
F27
Gms, Sh [D32?]
Gms
Sd, Fd
Sd, Sh, (def)
Dm, Sm, Sd [D31?]
Sd
Sh (def)
Sh, Sr (def)
Sd, Sh (def)
Dm [D30]
C S S C
GE-8
40
35
30
25
20
G9
G10
G11
G8
G6
G7
G4
G5
G3
G2
E27
E28
E29
F37
G1
F35, F36
Sm, Sd, Gm
St
Internal bedding within foresets
Sm, Sp
St
Sd
Sd
Flc
Gms, Sh
GE-8 logged summer 2000
C S S C



GE-8
GE-8
(Vertical scale in m, code to left of log refers to slide no.)
Sp
Sp
Sp
Sm, Sd
G16
60
55
50
45
C S S C
G13
G14
G12
80
75
70
65
St
Sh (def)
St
St
Sd, St comp.
C S S C
GE-8 logged summer 2000



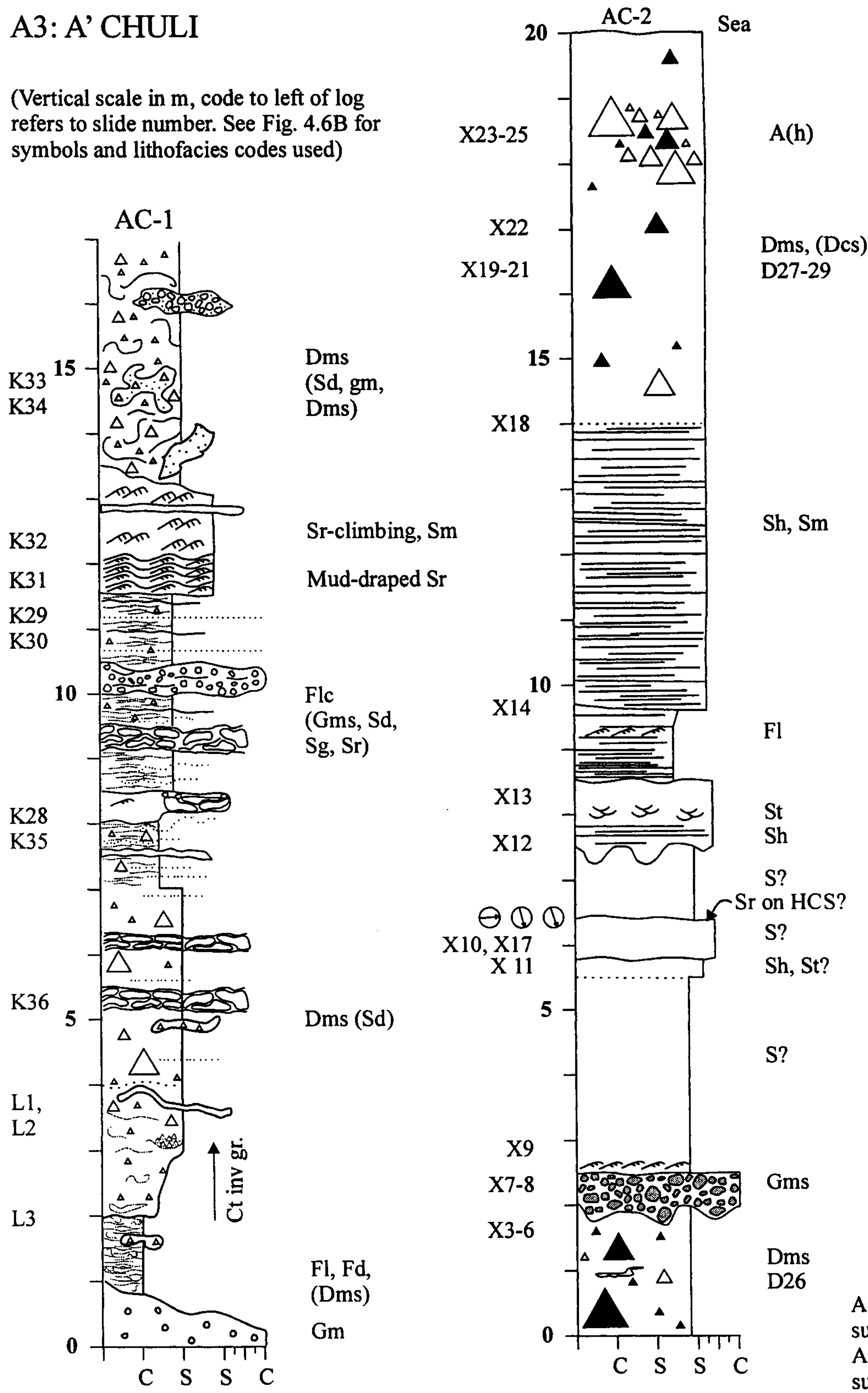
(Vertical scale in m, code to left of log refers to slide no.)



GE-8 logged summer 2000

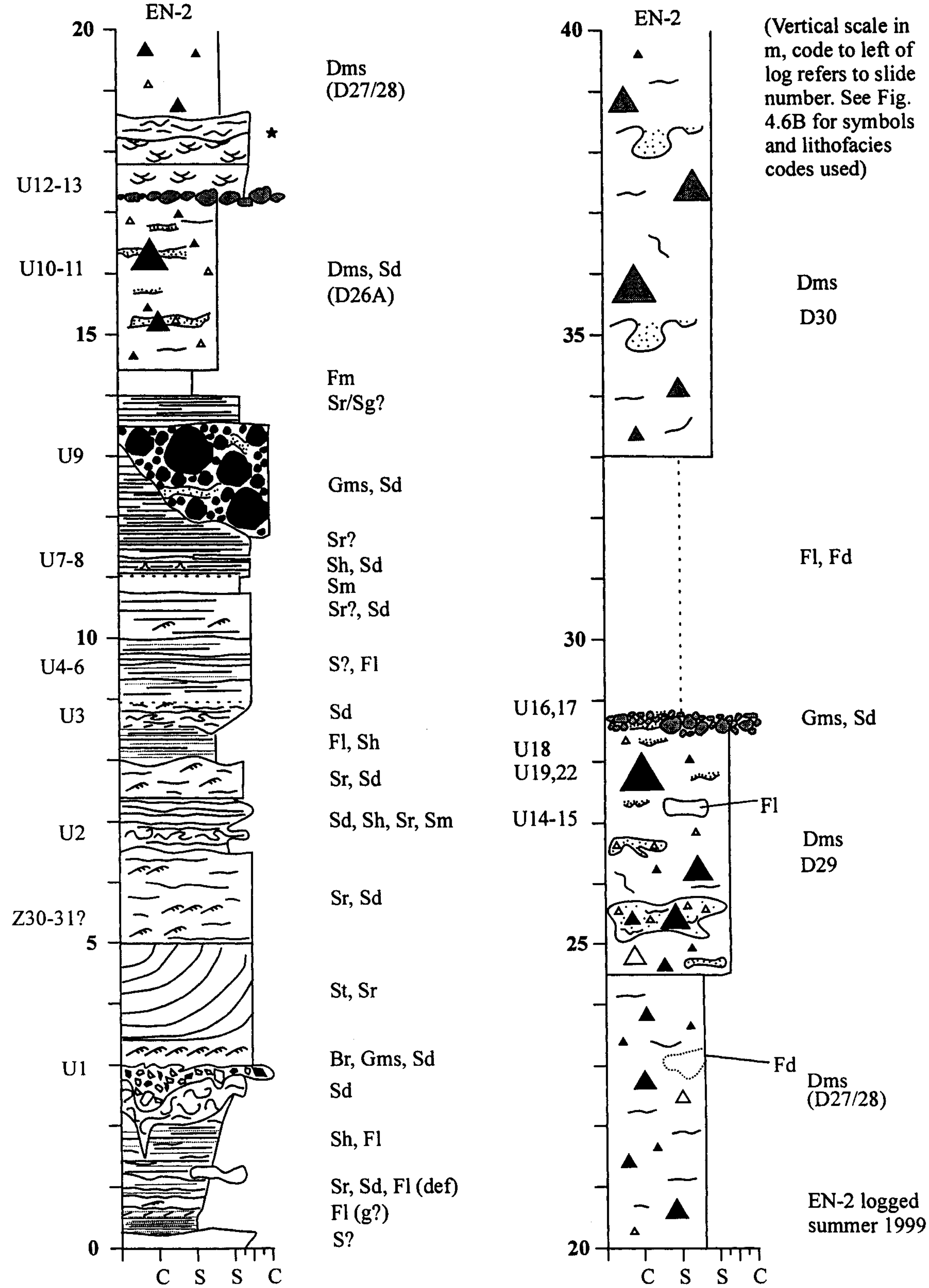
A3: A' CHULI

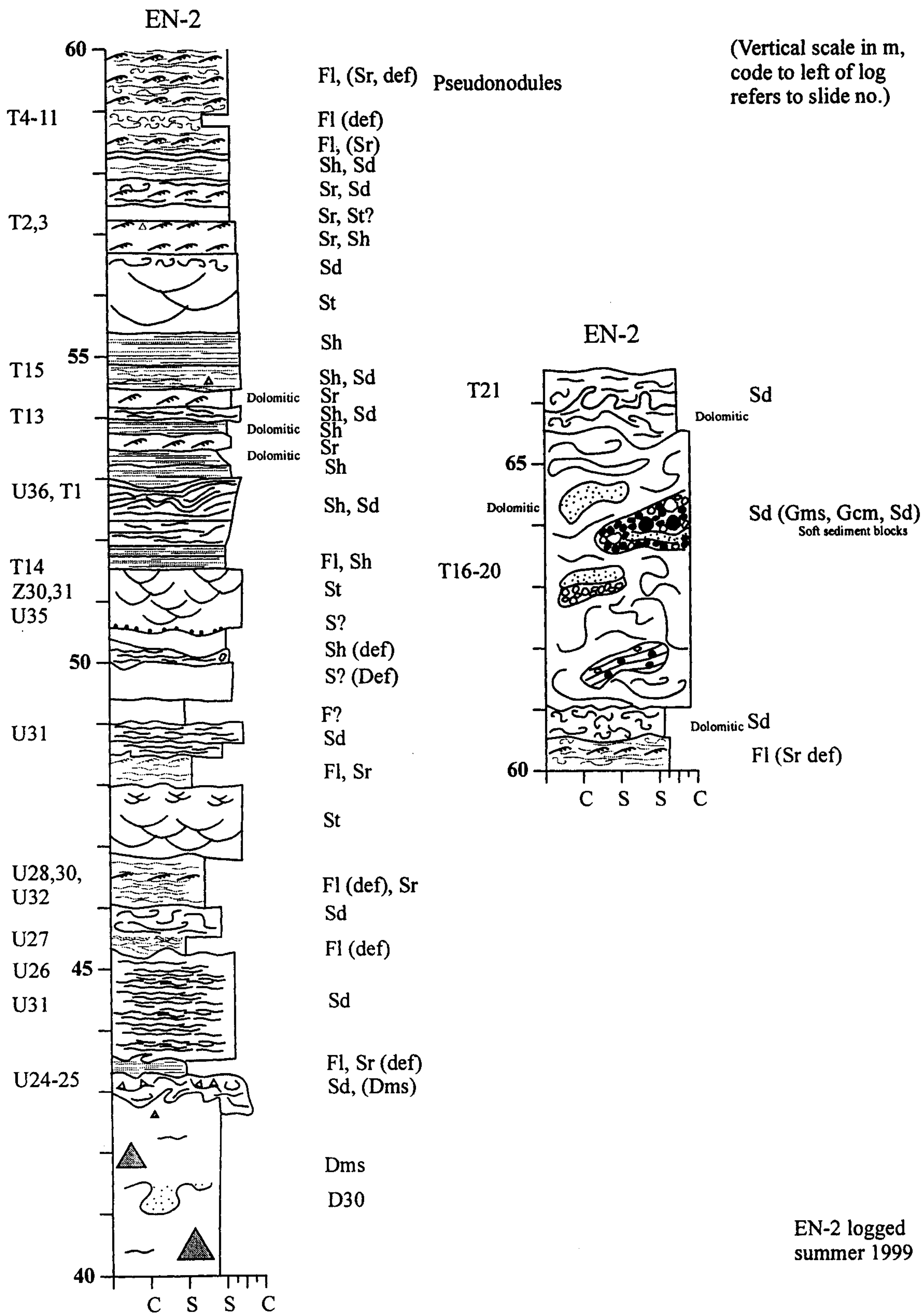
(Vertical scale in m, code to left of log refers to slide number. See Fig. 4.6B for symbols and lithofacies codes used)



AC-1 logged summers 99/00,
AC-2 logged summers 99/00

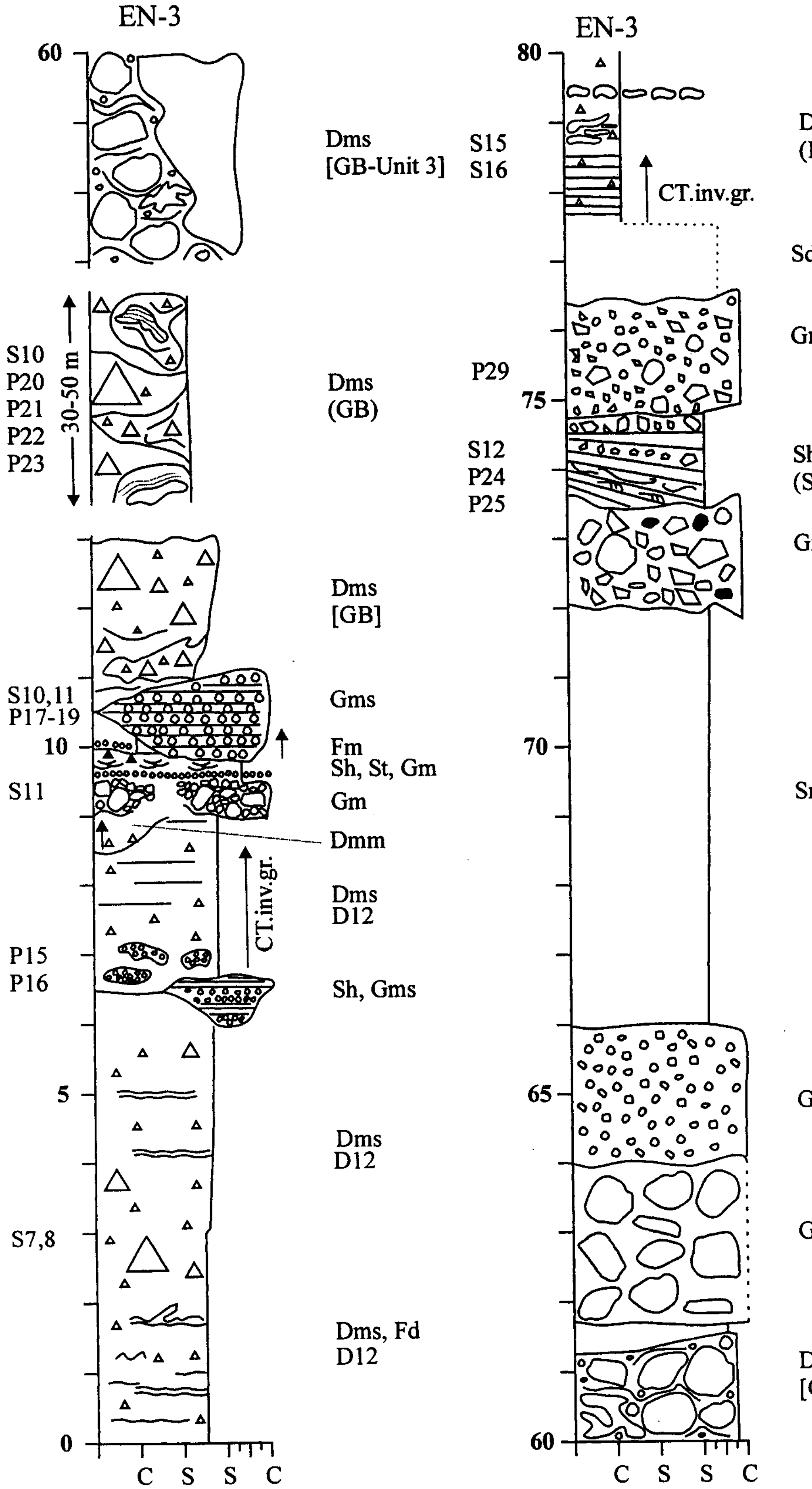
A4: EILEACH AN NAOIMH





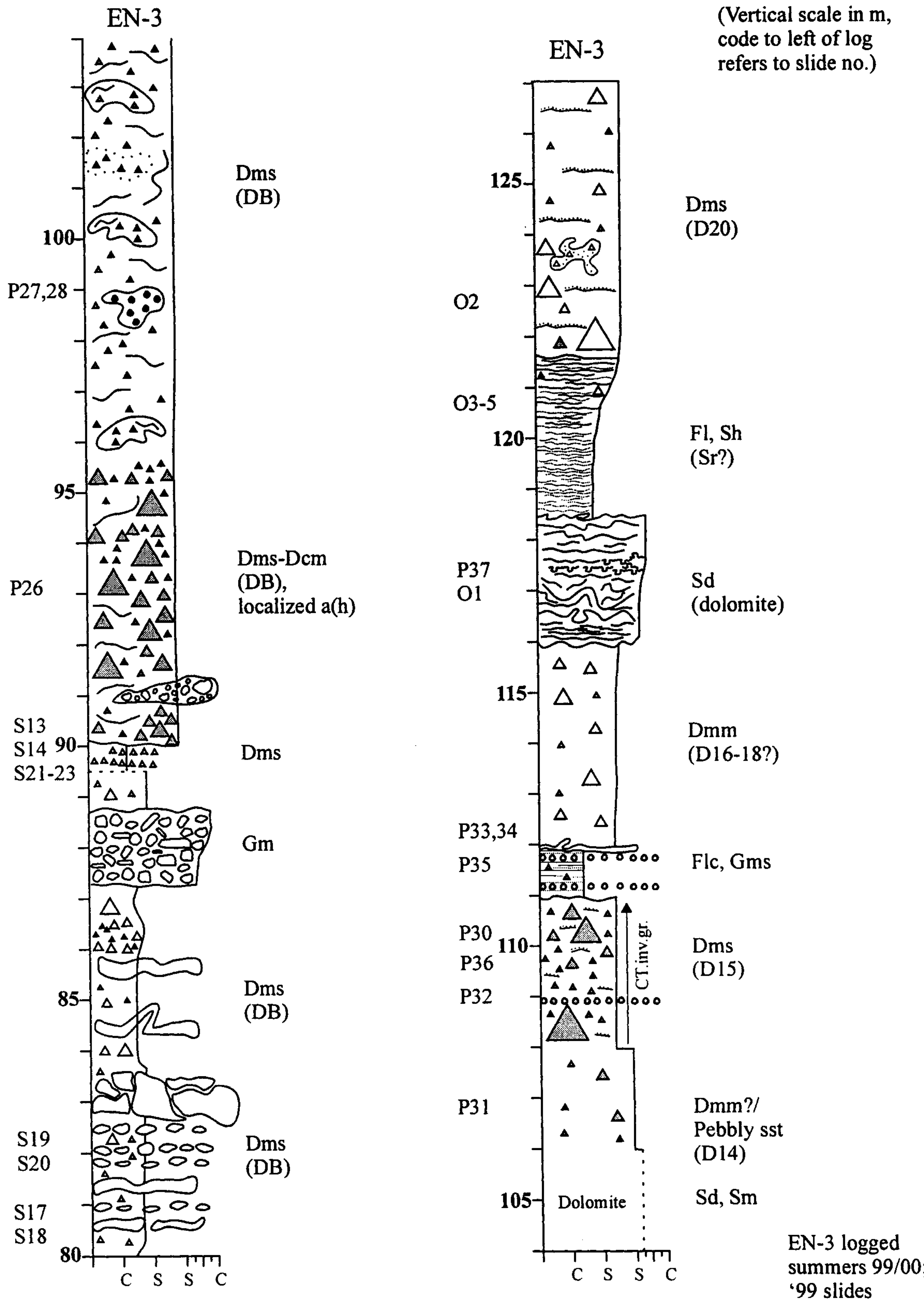
(Vertical scale in m, code to left of log refers to slide no.)

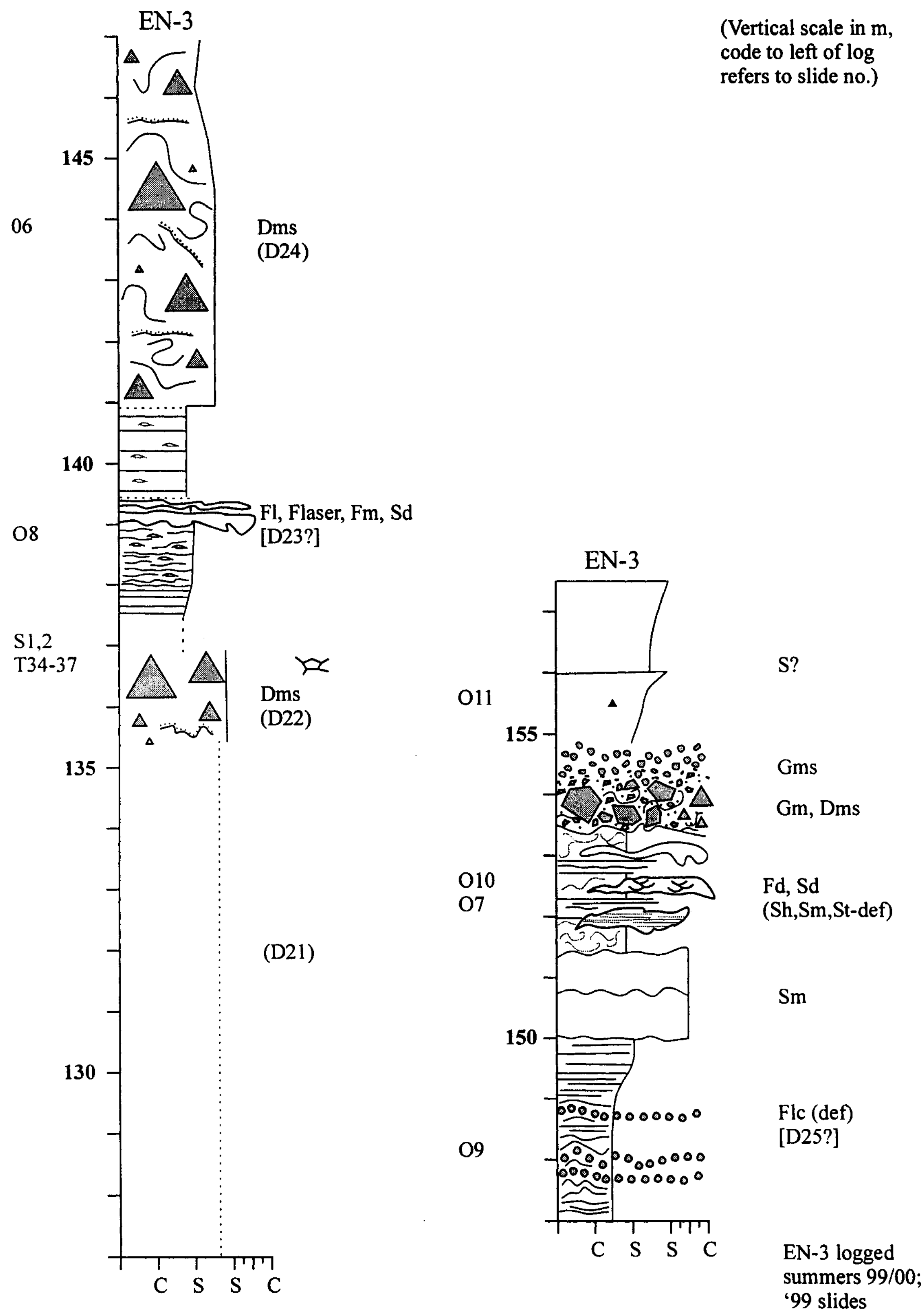
EN-2 logged summer 1999



(Vertical scale in m, code to left of log refers to slide no.)

EN-3 logged summers 99/00; '99 slides


EN-3
(Vertical scale in m, code to left of log refers to slide no.)
Dms
(DB)
100
P27,28
95
Dms-Dcm
(DB),
localized a(h)
P26
S13
S14 90
S21-23
Dms
Gm
85
Dms
(DB)
S19
S20
Dms
(DB)
S17
S18
80
C S S C
EN-3
125
Dms
(D20)
O2
O3-5
120
Fl, Sh
(Sr?)
P37
O1
Sd
(dolomite)
115
Dmm
(D16-18?)
P33,34
P35
Flc, Gms
P30
110
P36
P32
CT.inv.gr.
Dms
(D15)
P31
Dmm?/
Pebbly sst
(D14)
105
Dolomite
Sd, Sm
C S S C
EN-3 logged summers 99/00;
'99 slides


EN-3
(Vertical scale in m, code to left of log refers to slide no.)
145
06
Dms
(D24)
140
Fl, Flaser, Fm, Sd
O8
[D23?]
S1,2
T34-37
Dms
(D22)
135
(D21)
130
C S S C
EN-3
S?
O11
155
Gms
Gm, Dms
O10
O7
Fd, Sd
(Sh,Sm,St-def)
Sm
150
Flc (def)
[D25?]
O9
C S S C
EN-3 logged summers 99/00; '99 slides