Soft-sediment deformation structures in the Late Silurian Stubdal Formation: the result of seismic triggering

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The Ringerike Group is a lower Old Red Sandstone fluvial sequence that forms part of the Late Silurian foreland basin succession of the Ringerike Group, Oslo Region, southern Norway. The Ringerike Group exhibits a number of soft-sediment deformation structures in all its constituent formations, but in the type area of the Stubdal Formation many of these can be demonstrated to show preferential distribution along the same stratigraphic horizon. It is suggested that the laterally extensive soft-sediment deformation and mass sediment flows within the braided fluvial environment are the sedimentary expressions of a single large seismic event. The soft-sediment deformation structures include abundant ball-and-pillow structures, convolute cross-stratification, mud-crack diapirism, and spectacular channelised "pillow beds" up to two metres in thickness. These are catalogued and their variety is attributed to lateral differences in sedimentary facies at the time of deposition. The seismic activity responsible for the deformation was possibly due to an event in the Late Silurian tectonic evolution of the Norwegian part of the Caledonide mountain chain and its foreland basin.

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Introduction

The term soft-sediment deformation is used to describe the plastic deformation of unconsolidated sediments caused when pore fluids are rapidly expelled from them. This expulsion results from fluidization or liquefaction of the sediment shortly after its deposition, and due to the importance of fluid for such deformation to occur it is most common in water-lain sediments (e.g. Einsele et al. 1996). However, the resultant sedimentary structures are not confined to one depositional situation, being dependent on the thixotropic properties of the sediment rather than on any environmental factor.

The genetic processes of liquefaction and fluidization are subtly different, and may result in different types of soft-sediment deformation features. A liquidized bed is one where the upward migration of fluid is caused by the downward movement of sediment grains (Middleton & Southard 1978), whilst a fluidized bed results from upward fluid flow when there is no downward movement of sediment grains (Lowe 1976). Liquefaction reflects a temporary breakdown in the strength of the affected sediment resulting from a transfer of grain support to the pore fluid. This process may occur due to the deformation itself, or to some other extrinsic factor (see review in Obermeier 1996), but requires prolonged disturbance in order to allow enough time for the bed to liquefy fully before it returns to its original rheological state (Allen 1984). This is a key difference

between liquefaction and fluidization, as a bed can only exist in a state of liquefaction for the duration of its active disturbance, whilst a fluidized bed can exist in such a state for as long as external fluid is being pumped into it (Allen 1984). Effectively, this is because fluidization occurs when interstitial fluids are moving upwards because their force is then greater than the resistance of the sediment grains, and such fluids often migrate upwards along certain pathways rather than as a uniform "front" (Middleton & Southard 1978). Fluidization may be caused by processes such as the liquefaction of a buried layer, or by water-table upwelling (Owen 1995), and both fluidization and liquefaction can result from other causes including slumping, compaction, dissolution, and shock.

Recent years have seen an increased awareness that many large-scale soft-sediment deformation structures in the rock record may have had a seismic shock origin, especially if they have a wide lateral distribution along the same stratigraphic horizon. Such features may arise from earthquake activity as the shear stresses that result from cyclic seismic wave propagation cause a build-up of porewater pressure and subsequent mobilisation of fluids (Obermeier 1996). To produce in situ deformation features it is also necessary for the seismic wave acceleration to exceed the shear strength of the sediment (Einsele et al. 1996). The subsequent liquefaction or fluidization that results from these trigger mechanisms usually occurs within 10 metres of the
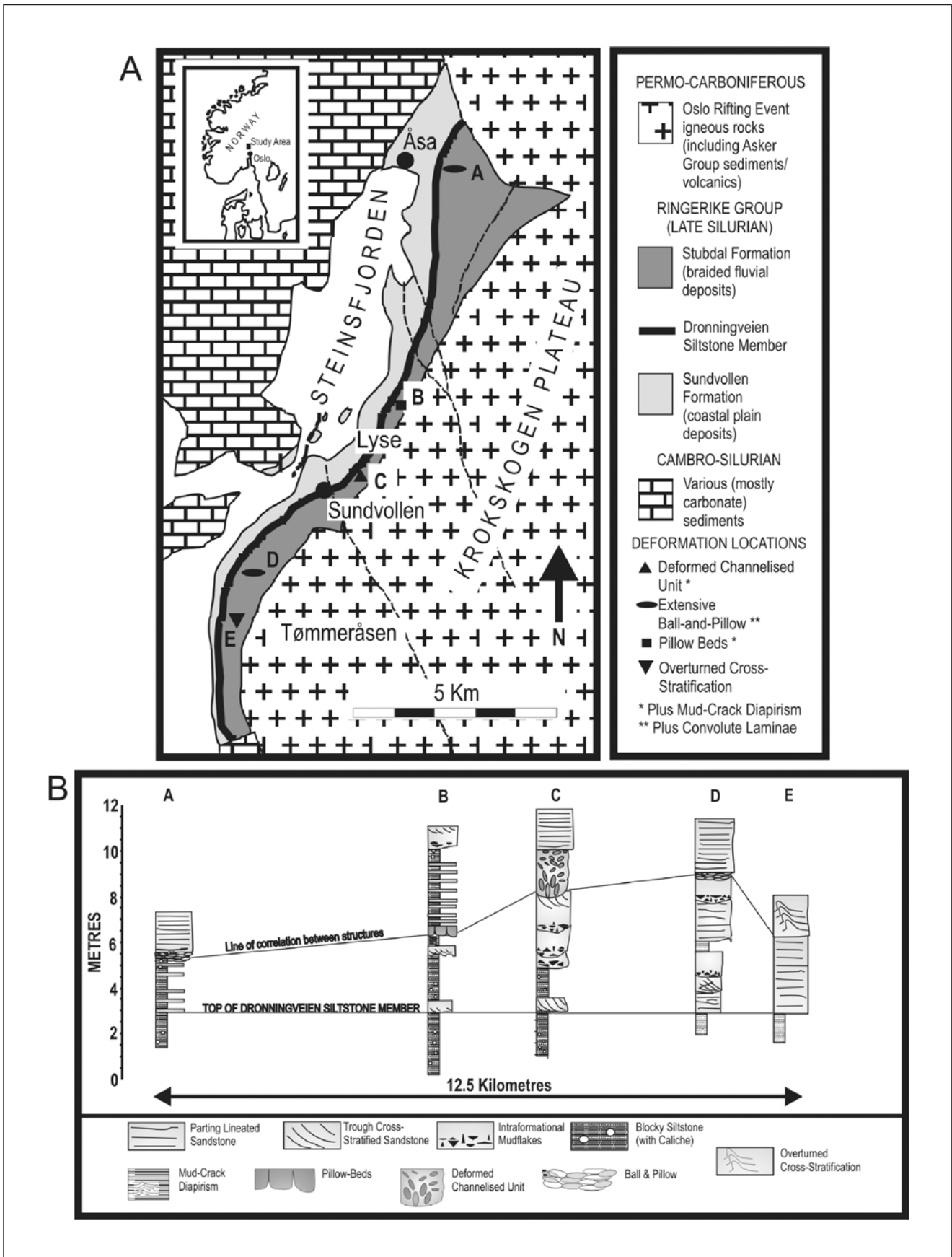
Fig.1: A: Simplified geological map of the study area, showing five seismically induced soft-sediment deformation localities situated stratigraphically above the Dronningveien Siltstone Member. B: Sedimentological context and correlation of the five localities on the map in relation to the top of the Dronningveien Siltstone Member.

ground surface (Obermeier 1996) and ultimately forms the soft-sediment deformation features. The term "seismite" has been favoured to describe such features by some authors, but this implies process rather than form and hence terming such features as "seismically triggered soft-sediment deformation" is preferable (see also the etymological arguments in Ricci Lucchi 1995 and Shiki 1996). Soft-sediment deformation features resulting from seismically triggered liquefaction and fluidization are diverse, and include various load structures (e.g. Ringrose 1989; Dugué 1995), slump and turbidite structures (e.g. Kleverlaan 1987; Shiki et al. 2000), dewatering structures (e.g. Lowe & LoPiccolo 1974; Obermeier 1996), convoluted, looped or overturned laminae (e.g. Allen & Banks 1972; Owen 1995), and sediment injection structures (e.g. Clague et al. 1992; Jolly & Lonergan, 2002). Some seismically induced soft-sediment deformation features are so unique as to warrant more specific terminology (e.g. "Sand Blows", Rascoe 1975; or "Pillow Beds", Roep & Everts 1992), and others show deformation associations with distinct internal divisions of structures (e.g. in Matsuda 2000).

In this study an earthquake-induced shock is argued as the trigger mechanism for many of the different varieties of soft-sediment deformation structures within the Stubdal Formation. Despite the fact that the Norwegian Caledonides and their associated foreland basin were tectonically active for much of the Ordovician and Silurian (see Gee & Sturt 1985), there have been surprisingly few soft-sediment deformation horizons that have been attributed to seismic activity. Whitaker (1964) provides the only report, describing mud-crack diapirism from the Stubdal Formation and suggests that it may have a seismic origin. The mud-crack diapirism is considered here along with its relationship to laterally extensive ball-and-pillow structures, "pillow beds", unbreached cusps, convolute bedding, overturned cross-stratification, and unusual channelised soft-sediment deposits up to two metres in thickness. Many of these features occur in the same stratigraphic horizon and may be attributed to the same seismic event.

The Stubdal Formation

The Stubdal Formation is the uppermost unit of the Ringerike Group in the northern part of the Oslo Region. Originally defined by Turner (1974), its type area is along the slopes of the Krokskogen plateau to the east of Steinsfjorden in Ringerike (c. 30 km northwest of Oslo, Fig. 1) where it reaches a maximum thickness of 550 metres. Other scattered outcrops are known from the Kolsås area to the west of Oslo, and from a road section along highway 285 to the south of Holsfjorden at Sylling, although the thickness of the formation in these areas is hard to ascertain. The Stub-

dal Formation overlies the oldest unit of the Ringerike Group, the Sundvollen Formation, and together these two formations represent siliciclastic, lower Old Red Sandstone deposition at the end of the predominantly marine carbonate deposition of the underlying Cambro-Silurian strata.

The Sundvollen Formation (480 m thick) (Fig. 1A) was deposited in a prograding nearshore muddy coastal plain environment that advanced southeastwards from its Caledonide source area in the nappes of the Jotunheim region (Davies 2003; Halvorsen 2003). At around 450 m from the base of the Sundvollen Formation, in the Ringerike type area, a distinct marker horizon, the Dronningveien Siltstone Member (c. 15 m thick), occurs and consists of deep red-coloured massive siltstone. Just above the Dronningveien Siltstone Member the first beds of the Stubdal Formation are seen, and represent deposition in a braided fluvial environment, also with a northwesterly Caledonide source area (Davies et al. 2005). The dominant lithology of the Stubdal Formation is fine- or very fine-grained red- or drab-coloured sandstone (mature quartz arenite), although minor red siltstone horizons are also present. Additionally, intraformational conglomerates and breccias consisting of mudclasts within a sand matrix are common.

During the Late Silurian, the latitude of southern Norway was equatorial (Douglass 1988), and the abundance of upper flow regime sedimentary structures such as parting lineation, plane-bedding, and climbing ripples suggest that the Stubdal Formation was deposited by high energy ephemeral fluvial floods in the middle reaches of a coastal plain. Perennial channelised deposition is also represented in the succession by well developed channel and bar forms and a wide variety of cross-stratification types (Davies 2003; Halvorsen 2003; Davies et al. 2005). The siltstone horizons of the Stubdal Formation sometimes contain relatively well-developed palaeosols, including vertisols, suggestive of periods of depositional quiescence in parts of the floodplain. They are usually limited in thickness and lateral extent and from the abundance of mud clasts within the many intraformational conglomerate units it can be seen that a large proportion of the silt-grade sediment was reworked at the time of deposition.

The age of the Stubdal Formation is not clear as it is devoid of any macrofossils and no other dateable material has been discovered. Biostratigraphic dating gives a Ludlow age for the underlying Sundvollen Formation and a Pridoli age can be ascertained for Ringerike Group sediments in the Holmestrand area that are in part equivalent to the Stubdal Formation. Hence the Stubdal Formation was deposited at the earliest during the Ludlovian or at the latest during Pridolian time.
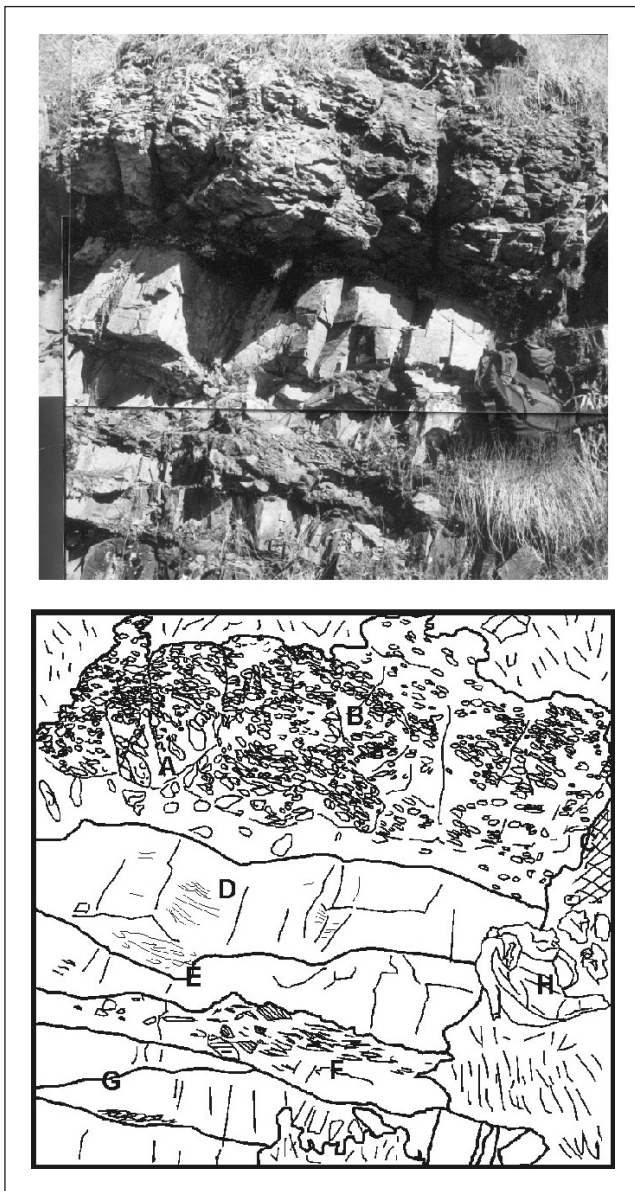
Fig. 2: Field photograph and sketch of the centre of the deformed chan-nelised unit. A: Near vertical orientation of large bulb structures (pseudonodules) in the deepest part of the channel. B: Horizontal ori-entation of smaller bulb structures in the top part of the centre of the channel. C: Horizontal orientation of smaller bulb structures towards the channel margins, note overall "mushroom" appearance when A, B, and C are considered together. D: Unbreached cusps in a channel sandstone unit underlying the deformed channelised unit (see also Figure 6). E: Isolated (non-seismic origin?) ball-and-pillow structures at the base of a channel sandstone unit. F: Brittle-fractured mudclasts. G: Intersecting fluvial channels. H: Rucksack (113 cm) for scale.

Types of soft-sediment deformation structures

A major problem in determining whether some of the soft-sediment deformation features are laterally exten-sive or merely localised features is that the type area for the Stubdal Formation is intensely faulted. As a conse-quence, and because of the scattered nature of the key
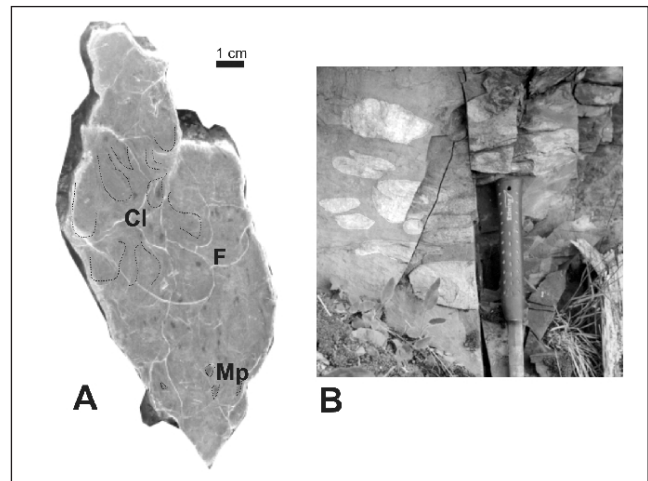

Fig. 3: Bulbous pseudonodules from the deformed channelised unit. A: Large vertical bulb from the central part of the channel (in field orien-tation) with contorted laminae (Cl), secondary fractures (F), and con-centrated iron mineral patches (Mp). B: Elongated bulbs with horizon-tal orientation from the channel margin showing both discrete and contiguous pseudonodules. Hammer shaft is 30 cm in length.

outcrops, it is often impossible to determine whether any two outcrops exhibiting soft-sediment deformation occur at the same stratigraphic horizon. However, the Dronningveien Siltstone Member provides a distinct marker horizon, and if the distance between the top of this member and the features in question is known then lateral correlation is aided considerably. All the soft-sediment deformation horizons described below (unless otherwise stated) are found at a distance of 5-10 metres above the Dronningveien Siltstone Member, and the fact that they can be correlated laterally along the same horizon (Fig. 1) is a key argument for a seismic origin.

Deformed channelised unit

Description – Approximately 200 metres to the north of the Dronningveien road (north of Sundvollen, Norwe-gian Grid Reference: NM 742597) is a limited exposure of a 1.2-2 metre-thick channelised siltstone deposit (Fig. 2). The unit consists entirely of oblong- or lozenge-shaped bulbous pseudonodules that are packed tightly with little or no undeformed sediment matrix. All the sediment is of the same silty grain-size, and individual pseudonodules are either pink or yellow in colour, although they also contain disrupted and inten-sely convoluted dark grey laminae. At the base of the thickest part (2 metres thick) of the channelised body, individual pseudonodules have an average dimension of 22 x 10 x 9 cm, and are oriented with their long axes near-vertical (Fig. 3A). This contrasts with the pseudo-nodules near the top and in the thinner parts of the unit, where they have much smaller dimensions (as small as 6 x 3 x 2 cm) and tend to be oriented horizon-tally or sub-horizontally. In the shallowest marginal
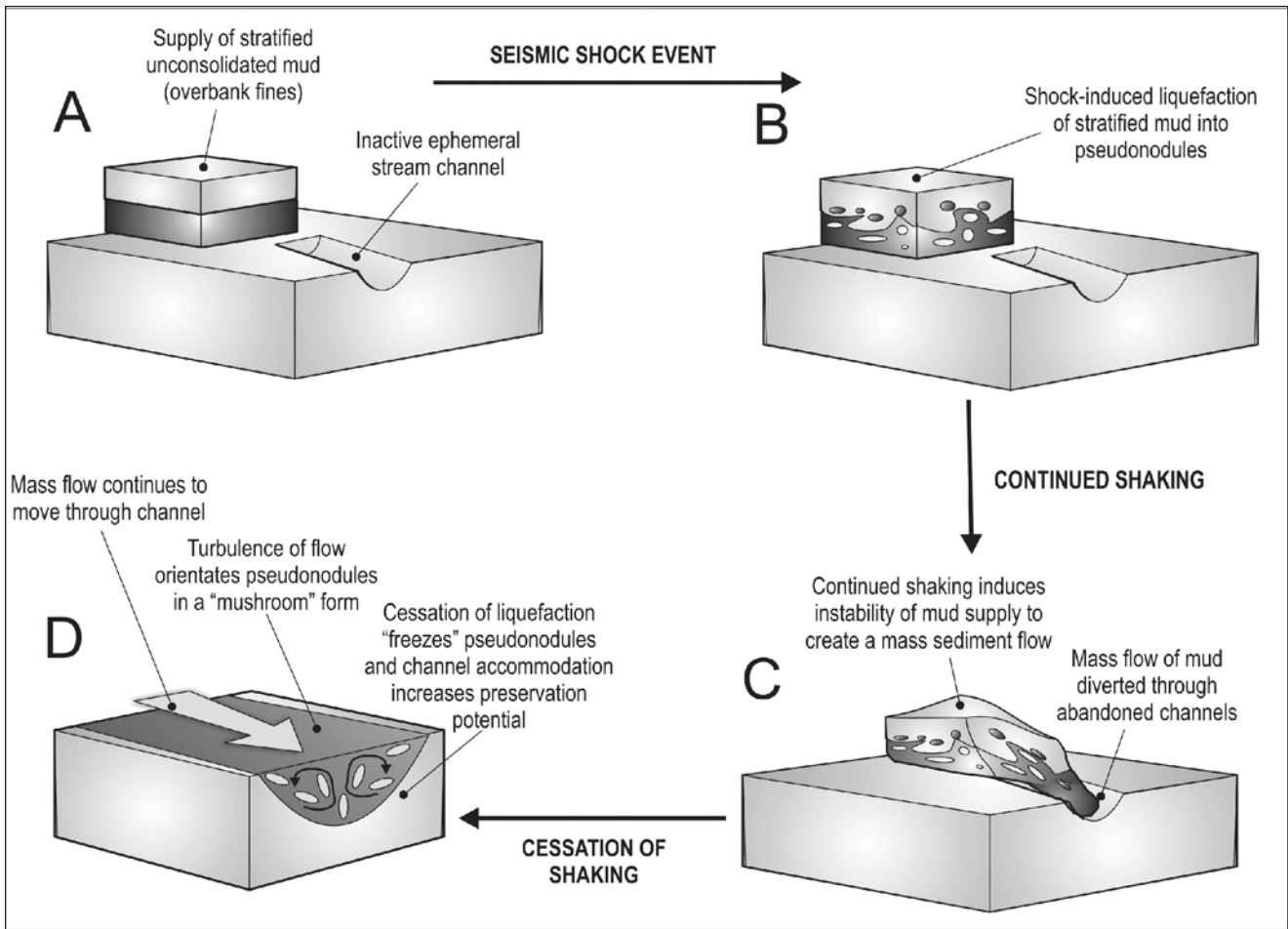
Fig. 4: Schematic cartoon showing proposed mechanism of formation of the deformed channelised unit. A: Original state. B: Liquefaction imme-diately after seismic shock event. C: Mass flow induced. D: Preserved soft-sediment deformation feature.

parts of the channel some of the pseudonodules are discrete (Fig. 3B), and superficially resemble a thick ball-and-pillow horizon. Pseudonodules with an incli-ned orientation occur between the vertical and hori-zontal forms, such that the entire complex of pseudo-nodules appears to have a mushroom shape, its central portion being at the deepest part of the channel (Fig. 2). Due to breaks in exposure, the unit can only be tra-ced laterally for 15 metres.

The unit rests on top of very fine- and fine- grained sandstones and siltstones. These exhibit little evidence of soft-sediment deformation at their upper contact with the pseudonodule-bearing unit, and resemble the typical broad channelised forms of the fluvial facies of the Stubdal Formation. The base of the channelised unit appears to cut sharply into the underlying sedi-ments, and there is no evidence to suggest that material comprising the pseudonodules was derived from the underlying beds which are still preserved. However, there are isolated unbreached cusps (see later) in some of the underlying beds. These may have been formed by fluidization at the time of formation of the channe-lised unit, or may be due to an earlier event, as both

liquefaction and fluidization have a tendency to repeat themselves at the same site in a seismically active area (Saucier 1989; Obermeier 1996). The deformed chan-nelised unit is overlain by a sediment unit that exhibits planar bedding, parting lineation and other sedimen-tary structures suggestive of high-energy fluvial deposi-tion, and has an undisturbed, non-erosional contact with the soft-sediment deformation features below. The unit has clearly a channelised geometry bounded by undeformed margins.

Interpretation – Attempting to find an analogy for these features in published literature is problematic. Similar large-scale pseudonodule horizons have previously been described, but differ in a number of ways, namely, they contain discrete pseudonodules with a different grain-size to the host material (e.g. McArthur & Onesti 1970; Bhattacharya & Bandyopadhyay 1998); they have either a disturbed upper or lower boundary (e.g. Rodrí-guez-Pascua et al. 2000); or the pseudonodules have a chaotic orientation within the unit (e.g. Pope et al. 1997). The unit within the Stubdal Formation displays none of these characteristics, although it does bear some resemblance to Quaternary seismogenic "sand

with clast supported conglomeratic structures" of the Ranafjorden area, as described by Olsen (1998). The undisturbed base indicates that the channel form was a predefined topographic feature at the time that it was infilled, and the undisturbed sediments on top of the unit and the planar upper contact suggest that soft-sediment deformation occurred prior to burial and a fluid upper layer existed during deformation (similar to that shown by the "flame-like structures" of Blanc et al. 1998). The uniform sediment grain-size of the unit, along with its thick and extensive nature, show that soft-sediment deformation was not caused by loading as even foundering and accumulation of pseudonodules into a bed in a quasi-liquid state usually results in discrete nodules (e.g. Bhattacharya & Bandyopadhyay 1998) and certainly would not result in a mushroom-like orientation of the pseudonodules. The mushroom-like orientation of the pseudonodules indicates a turbulent cellular motion or flow, the boundary between the two cells being at the centre of the channel. Such upward and lateral movement of intergranular flow has been documented by Middleton & Southard (1978).

It is argued that the complete unit represents a turbulent liquefied mudflow that filled a pre-existing abandoned channel within a semi-arid fluvial setting (Fig. 4). In practice, this means that soft-sediment deformation is a slumping/flowage feature. The contorted grey laminae within the nodules (Fig. 3A) represent the original laminations of a silt accumulation that was convoluted as the silt underwent liquefaction and separated into a number of pseudonodular packets (a similar gradational form from convoluted bedding to pseudonodules has been recorded by Maurya et al. 1998). Whilst the silt body was experiencing thixotropic conditions it flowed for a short distance along the abandoned fluvial channel for as long as it was liquefied, and the turbulent nature of this flow was established when the liquefying mechanism was withdrawn, the mushroom-like orientation of the clasts being preserved as a result (Fig. 4). Allen (1985) illustrated how instant removal of a medium that separates grain boundaries will immediately terminate the liquefied state of a sediment body and permits the preservation of such features. This whole process requires the sediment grains to be moved by the fluid content (rather than vice versa) and then the unit is most accurately described as a fluid gravity flow deposit. The apparent instantaneous cessation of the trigger-mechanism that led to the complex structure of the unit and the occurrence of extensive soft-sediment deformation at the same stratigraphic horizon, imply a seismogenic origin. Mass movements of this kind are relatively common features of modern earthquakes (Keefer 1984), although many ancient examples contain extraneous clasts as well as soft-sediment deformation features (Kleverlaan 1987). The absence of such clasts indicates that the deformed silts did not flow far, and were probably already present

in the channel at the time of liquefaction. The limited distance of flowage can be explained by the sediment being only in a liquefied state for the short duration of an earthquake shock.

Ball-and-pillow structures

Description – A number of ball-and-pillow structures have been found. These occur at the same distance above the Dronningveien Siltstone Member as the large deformed channelised deposit. Such features are most common in exposures in road and track cuttings to the west of Tømmeråsen, but have also been recorded in the area to the northwest of Åsa.

Interpretation – Ball-and-pillow structures are caused by packets of one sediment type loading, sinking and foundering in another type beneath it. This is due to the instability of the two sediment layers. Although some of the ball-and-pillow structures present in sediments of the Ringerike Group may be the result of "normal" sedimentary processes, those which are found in the same stratigraphic horizon together with various other soft-sediment deformation features may be of seismogenic origin. The fact that most of these neither grade into nor are laterally equivalent to sand lenses and have minimal vertical repetition, can be interpreted as the result of seismic triggering (Obermeier 1996; Olsen 1998).

Pillow beds

Description – The term "pillow bed" was introduced by Roep & Everts (1992) to describe a specific type of seismogenically formed soft-sediment deformation structure. One example from the Stubdal Formation (Fig. 5) is similar in form to the pillow beds described by both Roep & Everts (1992) and Rodríguez-Pascua et al. (2000). The outcrop in which the unit occurs is situated to the northeast of Lyse and consists of a succession of sheet sandstones and overbank deposits (Norwegian Grid Reference: NM 746605).

The pillow bed is a very fine sandstone horizon that appears to have been breached in a number of places by a series of planar elements that are partially intruded by sediments that are lithologically similar to those of the underlying unit. This unit is approximately 40 cm thick and is thought to have been formed by the upward mobilization of sand from the underlying horizon due to fluidization.

Interpretation – These features are similar to "sand blows" and "sand rolls" caused by modern earthquakes (Saucier 1989). Whilst some of these features are thought to have a purely sedimentary origin, particularly in fluvial settings (Coleman 1969), others can be
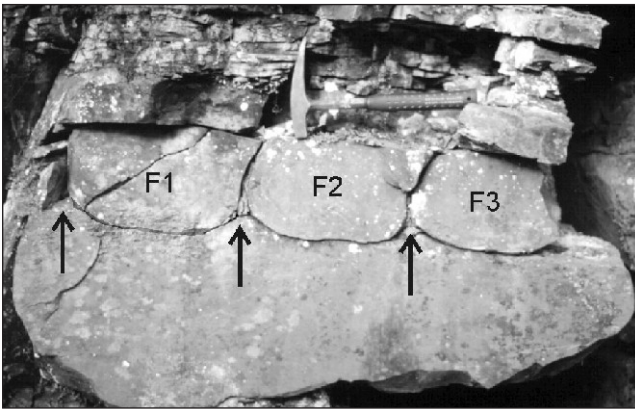
Fig. 5: Pillow-bed from channel overbank deposits (mixed fine sand and silt) near Lyse. Photograph shows section of three pillows of rolled sediment (F1-F3) with long axes orientated into the rock face. Arrows indicate where lower part of the pillow-bed (same lithology) is intruded upwards between the pillows. Hammer shaft is 30 cm in length.

seismogenic (Rascoe 1975; Roep & Everts 1992). Li et al. (1996) distinguish between the two causal mechanisms, the seismogenic examples having planar "dykes" or breached portions, as opposed to the tubular fluid escape structures found in fluvial flood deposits. The lateral, planar continuation of the breaches in the Stubdal Formation pillow bed therefore suggests the operation of a seismic trigger mechanism.

Mud-crack diapirism and unbreached cusps

Description – The presence of mud-crack diapirism in the Ringerike Group has previously been recorded and described by Whitaker (1964). Mud-crack diapirism has been noted above the Dronningveien Siltstone Member along the Dronningveien road, and such features are also common at the same locality as the pillow-beds (see above). At the latter locality, the typical mud-crack diapirism recorded by Whitaker (1964) actually appears to grade into connected, dish-like structures. Isolated examples of these structures are extremely common at the horizon 5-10 metres above

the Dronningveien Siltstone Member, particularly in beds immediately underlying the deformed channelised deposit, and in very fine-grained silty sandstones to the west of Tømmeråsen.

True mud-crack diapirs consist of desiccation cracks in siltstone or mudstone that have been forcefully injected by sand or silt. In effect, therefore, the mud-cracks taper upwards rather than downwards, being formed by the injection of fluidized sediment from below. They often grade vertically into smaller-scale dewatering structures of mud and clay laminae within gradationally overlying, very fine-grained sandstones, without injected sediment (Fig. 6). They are similar to the "unbreached cusps" described from experimental studies by Owen (1996). In effect they are unbreached dish-and-pillar structures formed when semi-permeable laminae act as barriers to upward moving fluids (Lowe & LoPiccolo 1974). They represent relatively weak incidents of fluidization and dewatering. This fluidization causes an upward movement of agitated sediment (forming the mud-crack diapirism) and which faces increased resistance as it moves vertically and becomes less powerful. Thus the unbreached cusps can either represent the diminished later stages of a fluidization event (also responsible for mud-crack diapirism), or they can be the sole soft-sediment structures that occurred during a period of minor fluidization.

Interpretation – Both types of mud-crack diapirism present within the Ringerike Group are argued to be seismogenically induced based upon their lateral persistence at the suspected seismically deformed horizon, and they tend to occur primarily in the overbank deposits present in the Stubdal Formation.

Convolute laminae

Description – Convolute laminae are relatively rare, but are occasionally seen in conjunction with ball-and-pillow structures to the west of Tømmeråsen and north-
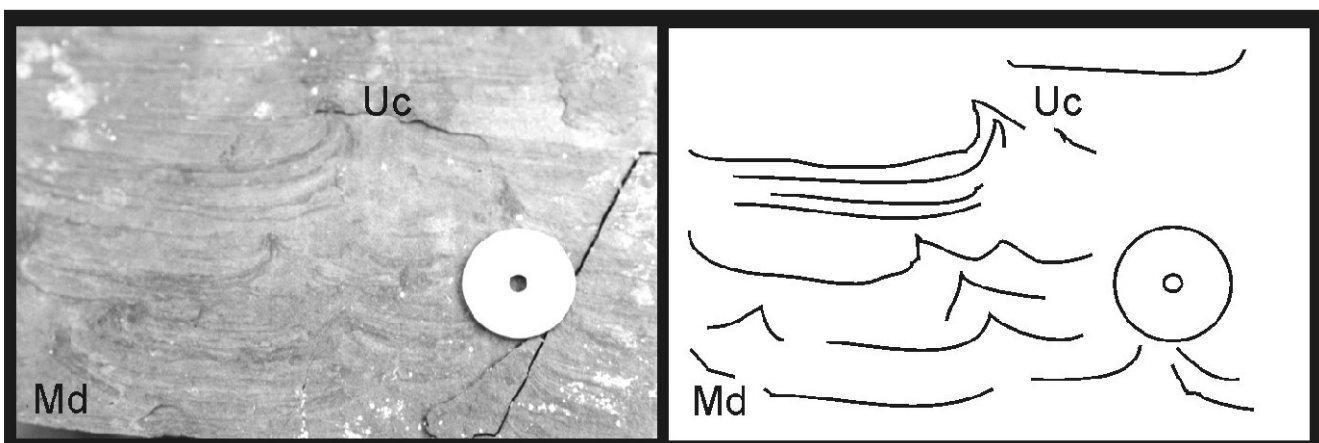

Fig. 6: Field photograph and sketch of unbreached cusps (Uc) and mud-crack diapirism (Md) in a very fine-grained channel sandstone at the deformed channel unit location. Coin is 2.3 cm in diameter.
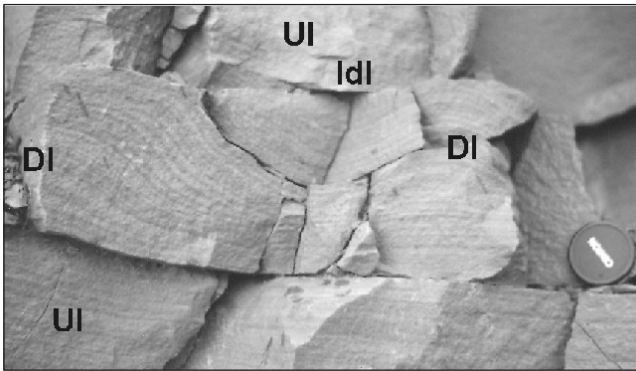
Fig. 7: Convolute laminae in sheet sandstones near Tømmeråsen showing domed laminae (Dl), and inverted domed laminae (Idl) bounded by beds above and below with undisturbed planar lamination. Camera lens cap is 5.3 cm in diameter.

west of Åsa, and often grade into slumped and contorted ball-and-pillow structures themselves. Other examples of convoluted laminae are seen to affect an entire bed and form series of domed and inversely domed structures (Fig. 7).

Interpretation – Convolute laminae may be formed by both liquefaction and fluid escape and do not necessarily have a seismogenic origin. Many examples are known from fluvial sedimentary processes, e.g. Picard & High 1973; Ray 1976; Eriksson & Vos 1979. However, examples of domed deformation similar to that seen in Figure 7 have been attributed to a seismic origin (e.g. Friend et al. 1976). Such features are due to the creation of three-dimensional anticlinal and synclinal features at the top of a liquidized bed. The lack of broken peaks to the domed features in the Stubdal Formation

examples contrasts with convoluted laminae formed by many normal fluvial processes (e.g. compare with Selley et al. 1963; Chakrabarti 1977).

In the Stubdal Formation the convoluted laminae occur within very fine-grained sands (usually with a high silt or clay content), and form the tops of channel-fill beds.

Overturned cross-stratification
Description – Two localities where overturned cross-stratification is visible have been discovered along the E16 highway to the west of Tømmeråsen. In each case the apex of the overturned fold is concordant with the foreset dip, and the features occur in 30-40 centimetre-thick horizons (Fig. 8).

Overturned cross-stratification was argued to have a seismic origin by Allen & Banks (1972), and more recent seismogenic examples have been described by Owen (1995), although some overturned cross-stratification is likely to have resulted from normal sedimentary processes such as loading, drag, and shearing (see Doe & Dott 1980).

Interpretation – The stratigraphic position (c. 5 metres above the Dronningveien Siltstone Member), and the distortion of bedding in a direction perpendicular to the normal palaeocurrent direction (Fig. 8), suggest that the examples from the Stubdal Formation are seismogenic. The feature was formed by distortion of liquefied sand that was cross-stratified at the time of deformation. Examples are known from cross-stratified sandstones in both channelised and point bar sand-bodies.
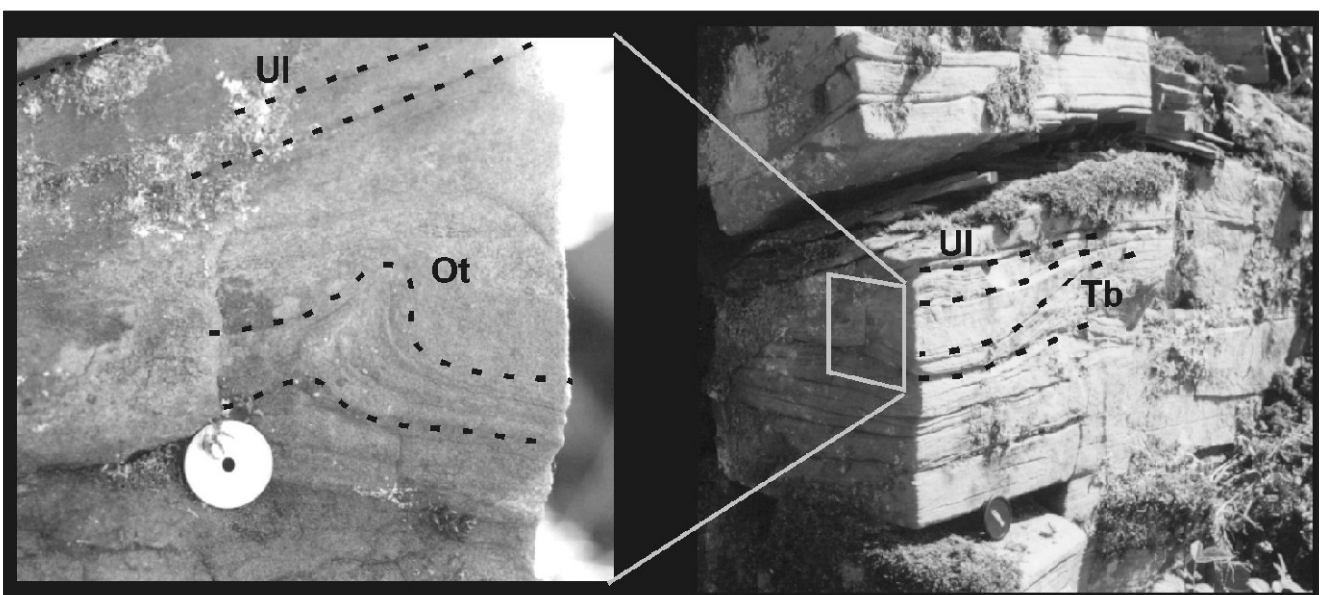

Fig. 8: Deformed trough cross-stratification (Tb) in point bar facies near Tømmeråsen, showing detail of where the cross-stratification is overturned perpendicular to palaeocurrent direction (Ot). Note undeformed cross-stratification (Ul) higher up in the same bed. Coin is 2.3 cm in diameter.

Discussion

The ability to distinguish between soft-sediment deformation features created by seismic activity and those created by normal sedimentary processes is problematic and controversial. This challenge arises because the structures are formed by liquefaction and fluidization processes that may have a number of different triggers, and it is the trigger mechanism which is sought rather than the formational process. Shiki (1996) argued that the most effective and strategic way to prove earthquake-induced deformation in the geological record is to perform well-focused investigations of trigger-known soft-sediment deformation. However, studies which have actually done this have discovered that soft-sediment deformation features arising from recent seismic activity are extremely similar to those which can be caused by normal sedimentary processes (e.g. Sims 1973). Other researchers have emphasized the importance of eliminating the possibility of the soft-sediment deformation features being attributable to normal sedimentary processes (Morretti 2000), but the wide variety of non-seismic processes which can cause soft-sediment deformation in a fluvial setting in which event flood deposition and rapid crevassing or bank collapse are common (e.g. Coleman 1969; Li et al. 1996), means that it is almost impossible to exclude such influences. Yet seismically induced soft-sediment deformation would be preserved in such an environment as there are numerous situations where there can be a protected environment containing alternating beds of wet, porous and well-sorted fine sands and muds favourable to thixotropic processes. Such situations are of key importance for the preservation of seismically induced soft-sediment deformation (Dugué 1995). Therefore, whilst it is impossible to say with absolute certainty that specific deformation features are the result of seismic activity, it is possible to present a series of features that collectively provide strong evidence for a seismogenic origin (e.g. Kleverlaan 1987; Obermeier 1996; Pope et al. 1997; Rossetti & Góes, 2000).

Arguments for the Stubdal Formation deformation having been triggered by seismic activity at the time of deposition are as follows: (1) The Stubdal Formation was deposited in the foreland of the Caledonian mountain chain, where seismic activity would be expected in connection with the Late Silurian Scandian Phase of tectonic evolution (Roberts & Sturt 1980; Roberts & Gee 1985; Dallmeyer 1988). (2) Within the Ringerike area, a wide variety of soft-sediment deformation features can be traced laterally (at intervals) along the same stratigraphic horizon for up to 16 kilometres. (3) The soft-sediment deformation features occur between undeformed beds and most of the features show no vertical repetition, suggesting they were not formed by normal sedimentary processes. (4) The features in the sediments attest to deformation in an unconsolidated and semi-consolidated state, showing that the deformation occurred soon after deposition. (5) Some of the larger scale features (e.g. the deformed channelised unit) cannot be accounted for by ordinary depositional processes. (6) The features occur randomly across a number of different fluvial sub-environments (i.e. channels, overbank deposits, crevasse splays, point bars). (7) There appears to be a "core area" (north of Dronningveien) where soft-sediment deformation is most intense. (8) Vertically extensive sequences of undeformed sediments within the Stubdal Formation suggest long periods of quiescence without any activation of the deformation trigger mechanism. (9) The sedimentary features are similar to those documented as having been by earthquakes in Recent historical times.

When these factors are considered collectively, it can be argued that it is likely that the soft-sediment deformation features that occur 5-10 metres above the Dronningveien Siltstone Member had a seismic origin.

One unusual feature, on a regional scale, is the apparent absence of any brittle deformation, or of large-scale sediment intrusions such as sand dykes. The latter are common in many seismically deformed sediment sequences, as are complex associations of brittle and ductile deformation, and have been considered to be key features of some sequences (e.g. Rossetti & Góes 2000). Yet the absence of such features in the Stubdal Formation does not negate the interpretation of the features as seismogenic. Unknown factors such as consolidation rates and earthquake magnitude and duration may also have influenced the form of the Stubdal Formation structures. Brittle deformation may be seen in association with soft-sediment deformation if there is variation in fluid content and plasticity within unconsolidated sediments (Matsuda 2000), but such a variation is not a necessary feature of all such sediments. Owen (1995) suggested that if an unconsolidated mass was saturated at the time of deformation there would be no fracturing to allow the formation of dykes. Although semi-arid, the palaeogeographic location of the Stubdal Formation (equatorial palaeolatitude, proximity to Caledonide mountain belt) would have meant that seasonal flooding (monsoonal?) conditions were likely, and there is no reason to suppose that the unconsolidated sediments were not saturated at the time of deformation. The nature of the sediments in question may also have been a factor in influencing the subsequent form of soft-sediment deformation, as the form varies according to the type of facies of the fluvial regime (Fig. 1). In each case the factors that determine the end form would have included sediment grain-size, susceptibility of local conditions to either fluidization or liquefaction, and whether the unconsolidated sediments at specific sites were heterolithic or not. These factors were primarily controlled by local environmental conditions.

A final test as to whether or not such features were seismically generated is to consider the characteristics of the suspected earthquake. If these can be ascertained are they relevant for the tectonic setting in question? Pratt (1994) argued that it is not possible to determine the exact tectonic evolution of an area from the seismogenic, soft-sediment deformation features alone, but some general observations can be made. Ambraseys (1988) suggested that large earthquakes may result in liquefaction features up to 500 km from the earthquake epicentre, although other authors have argued that most liquefaction sites lie within 40-100 km of the epicentre (Galli & Meloni 1993, quoted in Morretti 2000). Both estimates are potentially feasible for the Stubdal Formation liquefaction, although the exact position in which the Stubdal Formation deposition took place within the Caledonide foreland basin is unclear, nor is it possible to pin-point exactly where the epicentre was. The intensity of the earthquake is easier to determine: Keefer (1984) suggested that earthquakes with magnitudes as low as V or VI on the Mercalli Scale could result in the preservation of liquefaction features, and Tuttle et al. (2002) argued that a VI magnitude would be the minimum required for the preservation of such features over a wide area. Therefore it is suggested that the Stubdal Formation features were the result of an earthquake of at least magnitude VI and that this earthquake occurred somewhere within the active foreland basin setting of the present Oslo Region during the Late Silurian. A number of the soft-sediment deformation features described from the Stubdal Formation have also previously been suggested as those to be expected as commonly occurring features in a foreland basin setting (see Pope et al. 1997).

Conclusions

The Late Silurian Stubdal Formation of the Ringerike area, southern Norway, is a braided fluvial succession that contains at least one extensive horizon showing soft-sediment deformation features formed shortly after deposition. These features include deposits due to liquefied mass fluid flows, extensive ball-and-pillow structures, small sediment-fluid injection features such as pillow-beds and mud-crack diapirism, and the more common effects of liquefaction and fluidization such as convoluted bedding and overturned cross-stratification. The lateral persistency of such features, along with other characteristics of the succession, can be used as arguments for a seismically triggered origin for the genetic liquefaction and fluidization processes. This trigger effect is believed to be due to a relatively large earthquake that had its epicentre somewhere in the evolving foreland basin in the forefront of the Caledonides in southern Norway.

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