Liquefaction features in Cambrian-Ordovician boundary strata of southeastern Minnesota:
Evidence for paleoseismicity in the cratonic interior?

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Senior Integrative Exercise
December 3, 2009

Submitted in partial fulfillment of the requirements for a
Bachelor of Arts degree from Carleton College, Northfield, Minnesota.
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**Liquefaction features in Cambrian-Ordovician boundary strata of southeastern Minnesota; Evidence for paleoseismicity in the cratonic interior?**

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**Abstract**

Intrastratal deformation features in Cambrian-Ordovician boundary strata in southeastern Minnesota, U.S.A., may be evidence for mid-continent paleoseismicity. Deformational features occur tens of kilometers east of the Midcontinent Rift zone, and include funnel-shaped structures, water-escape, and more subtle structures indicative of sand-on-sand density contrasts. The stratigraphic interval of interest is the uppermost Jordan Formation (Furongian), a very fine- to coarse-grained quartzose sandstone, and the basal Oneota Formation (Tremadocian), a heterolithic sandstone and dolostone that grades upward into bedded dolostone. Along the Jordan-Oneota contact, deformation features are extensive, and the result of sand liquefaction and fluidization. Upward migration of excess pore water was obstructed in places by shale drapes that locally ruptured, causing sand to be injected into overlying beds. Movement of sand in this manner created voids that were filled by a chaotic mixture of sand, shale, and pebbles that collapsed from above. Where upwardly percolating water was not confined by the permeability barrier, intrastratal flow produced water-escape pillars. Deformation along foresets is common in large-scale tidal dune cross-sets within ~3 m below the Jordan-Oneota contact. Features resulting from foreset deformation include digitate interfaces (interfingering) and *in situ* rounded forms interpreted as sand-on-sand boudinage. These features formed from density contrasts between individual foresets caused by subtle packing and grain size differences created by the avalanche process during dune migration. We interpret the trigger for deformation to be a post-depositional event because of the association of these subtle features with the more obvious liquefaction features in the immediately overlying boundary strata. Although these features cannot be unambiguously attributed to a paleoseismic event, some other common possibilities can be eliminated, including slumping and loading by sediment, tides, and waves. A rise in hydrostatic head as a result of rapidly rising sea level remains an alternate possibility, however. These intrastratal deformation features are documented in one outcrop in southeastern Minnesota. The subtlety of some of these features has led us to revisit unusual features in other localities that we previously interpreted as synsedimentary phenomena. Seismites may be difficult to generate, and to recognize, in quartzose sandstones of the mid-continent due to the lack of significant clay beds to serve as permeability barriers, and the homogeneous textural and mineralogical attributes of these units.

**Key words:** liquefaction, permeability barrier, sand intrusion, draw-in, water-escape pillar, reverse density gradient
Introduction

Liquefaction is most common in unconsolidated sands because of their high porosity (Lowe, 1975). Liquefaction is the transition from a grain-supported to fluid-supported framework and is typically caused by increased pore-fluid pressure. In extreme cases, pore-pressure can become high enough to support the weight of the overburden (Owen, 1996). This can reduce the shear strength at grain contacts to zero, so that the sand body behaves as a viscous liquid (Anketell et al., 1970; Owen, 1987). As the system stabilizes, grains attempt to achieve a more tightly packed orientation (Obermeier, 1996), which can result in compaction with pore space reduction of ~6% (Moretti et al., 1999).

During compaction of the sediment pile, excess pore water is expelled toward the surface (Lowe, 1975). However, where sand is immediately overlain by a permeability barrier, such as clay, upward migration of water may be prevented (Moretti et al., 1999). If the pressure created by the buildup of water below the barrier becomes great enough, the barrier ruptures locally and a mix of water and fluidized sediment flows through gaps in the barrier (Moretti et al., 1999).

Liquefaction does not occur spontaneously: it requires a trigger mechanism (Owen, 1996). A wide array of deformational features have been attributed to liquefaction caused by cyclic loading by waves (Dalrymple, 1979), shear exerted on the sediment bed by migration of the tidal bore in hypertidal estuaries (Greb and Archer, 2007), slope-induced slumping (Shi et al., 2007), rapid sedimentation (Lowe, 1975), groundwater fluctuation (Draganits et al., 2001; Massari et al., 2001), and seismic activity (Rodriguez-Pascua et al., 2000). However, it can be difficult to identify the mechanism responsible for liquefaction by examining features post facto (Obermeier, 1996).
The purpose of this study is to document and discuss the origin of a variety of features attributed to liquefaction in the upper Cambrian Jordan Formation and lower Ordovician Oneota Formation in southeastern Minnesota. Liquefaction in these formations has not been previously documented. The Jordan and Oneota Formations are the main aquifer for the Twin Cities region. Therefore, breaches in bedding planes and less permeable layers resulting from liquefaction may represent important conduits for groundwater.

**Geologic Setting**

Cambrian-Ordovician strata in southeastern Minnesota were deposited in the Hollandale Embayment, a broad, shallow depression that extended over much of the present day upper Mississippi Valley (Runkel, 1994; Runkel et al., 1999) (Fig 1). The Upper Cambrian Jordan Formation is a fine- to coarse-grained quartz arenite, and is capped by an unconformity that approximates the Cambrian-Ordovician boundary in the study area (Runkel et al., 1999). The unconformity is overlain by the Lower Ordovician Coon Valley Member of the Oneota Formation, the lower of the two formations that constitute the Prairie du Chien Group (Mossler, 1987).

During the Late Cambrian, the Jordan Formation prograded basinward (present day west-southwest) away from the Wisconsin Dome and Arch during a period of overall regression (Runkel, 1994). Regression culminated in the exposure of the Hollandale Embayment and extensive erosion of the top of the Jordan (Runkel et al., 2007). Rising sea levels resulted in deposition of the Coon Valley Member (Runkel, 1994). The landward advancing shoreline quickly covered sediment sources due to the shallow gradient of the top
Figure 1. Map of the approximate outcrop belt of the Jordan Formation (grey area) in the Hollandale Embayment. W represents the Weaver outcrop, where most of the fieldwork for this study was done. R, Red Wing; S, Stillwater: two satellite outcrops where similar features are recognized. Modified from Tape et al. (2003).
of Jordan landscape, contributing to the stratigraphically thin (~5 m) transition from sand into dolostone (Runkel, 1994).

At least four conodont subzones that represent ~3 million years are missing across the Jordan-Oneota contact (Runkel et al., 1999). A regional pebbly bed (the lowermost bed of the Coon Valley Member) truncates the Jordan and is interpreted as an erosional lag associated with the unconformity (Runkel, 1994). Jordan-Oneota boundary strata also locally contain silcrete (a silica cement), which is considered diagnostic of subaerial exposure (Smith et al., 1993).

The succession of lithofacies upward in the Jordan Formation is from low-energy offshore deposits to high-energy nearshore deposits (Runkel, 1994) (Fig 2). Extensively burrowed, very fine-grained hummocky quartz sandstone grades upward into medium to coarse-grained sandstone with swaley to trough cross-stratification that lacks bioturbation (Runkel, 1994). Tongues of hummocky sandstone that occur between trough intervals mark brief transgressive episodes that interrupt the overall regressive character of the formation (Runkel, 1994). At the top of Jordan, meter scale cross-bedding deposited in environments dominated by tides typically overlies the uppermost trough interval (Runkel, 1994). In southeastern Minnesota, the Jordan is 25-35 m thick and the pebbly bed marking the beginning of Oneota deposition truncates tidal deposits or upper shoreface deposits where tidal deposits are absent (Runkel, 1994). The Coon Valley Member contains sandy to silty dolomites deposited in an intertidal to supratidal environment (Smith et al., 1993).
Figure 2. Generalized stratigraphic column of the Jordan Formation and the base of the Coon Valley Member in southeastern Minnesota (bedforms not to scale). Modified from Runkel et al. (1994).
Study area and interval of interest

A kilometer west of Weaver, Minnesota, the top ~5 m of the Jordan Formation is an interval of medium to coarse-grained sandstone with large cross-bedding containing tidal signatures (Tape et al., 2003) referred to here as bed J6 (Fig 3). Tidal deposits of J6 consist of sets of m-scale dune cross-bedding with foresets up to ~60 cm thick (Runkel, 1994). The color of the sediment varies widely due to irregular distribution of iron oxide cement. An undulating layer of irregularly shaped shale nodules defines the base of the tidal deposits at this locality. The shale was most likely deposited as drapes at the toes of migrating dunes during slack water conditions. Tabulate intraclasts with long axes oriented subhorizontally are present among the shale nodules. Intraclasts range from very fine to coarse sand, though individual clasts contain only one size fraction. They range from 2 to 15 cm long and 1 to 5 cm thick, with one exception measuring 40 cm long and 10 cm thick.

An erosional surface caps the tidal interval, and is overlain by a sandy, cross-bedded bed ~85 cm thick referred to here as CV1 (Fig 3). A series of closely spaced shale drapes (many of which have weathered to jarosite) each no more than a few cm apart stratigraphically dominate the basal 10 to 15 cm of CV1. Individual drapes commonly taper out or, conversely, coalesce to form bands up to a few cm thick. The upper ~75 cm of bed CV1 consists of trough cross-bedded medium-to coarse-grained sandstone. Rip-up flakes of shale less than 5 cm long are common along foresets, and burrow casts are abundant locally. Compared to the Jordan Formation and the rest of the Coon Valley Member, bed CV1 is rich in pebbles ranging up to ~1 cm in diameter. Pebbles are concentrated at the base of the bed. CV1 is overlain by a series of heterolithic sandy and silty beds that grade into dolomite.
Figure 3. A generalized two-dimensional panel of the Weaver outcrop. Vertical exaggeration of X 10. Axes are measured in meters. Bedforms and deformational features are not drawn to scale. Folded strata are (see fig. 8) tabulate intraclasts Type 1 and Type 2 foreset detsm.

Planar to low angle bedding M-scale tidal dune xp Troughy cross-strat. Swaley cross-strat. Hummocky cross-strat. Shale drapes Shale nodules Fluidization channel

Cambrian-Ordovician Boundary Cambrian-Ordovician Boundary Cambrian-Ordovician Boundary

Cover
Deformational features at Weaver are concentrated around the Jordan-Oneota contact, and commonly involve bed CV1.

**Deformational features at Weaver**

Along the Jordan Oneota contact, funnel-shaped structures occur in abundance. Fissure-like cracks and folded strata locally occur with associated funnel-shaped structures. Water-escape pillars are also present at the contact. Less pronounced water-escape pillars are found in the basal shale drapes of bed J6. In the upper ~2-3 m of J6, foresets in large-scale tidal dune cross-bedding are commonly deformed.

**Funnel-shaped structures, fissures and folded strata**

**DESCRIPTION**

Funnel-shaped structures extend downward from the Jordan-Oneota contact into bed J6. They range in size from 8 to 120 cm wide and 7 to 38 cm deep. Some are U-shaped, while others are V-shaped and terminate in a sharp point (Fig 4A-C). There is no apparent correlation between width and depth of funnels, and no clear pattern in their distribution. Funnels occur as isolated features or in small groups, and can be densely or sparsely distributed at different places along the outcrop. Funnels contain structureless medium-to coarse-grained sand, shale flakes, pebbles, and burrow casts (whole and fragmented). Shale flakes are more commonly distributed at the margins of funnels, where they are generally oriented parallel to funnel boundaries. Shale flakes within the center of funnels are oriented more horizontally.
Figure 4. Photographs (1) and line drawings (2) of funnel-shaped structures and associated fluidization channels and folded strata at the Jordan-Oneota contact. Funnels increase in scale in order of appearance. Ticks on the scale are cm and color segments on the staff are 10 cm. A-C) Funnel-shaped structures without associated features. Shale flakes near the margins of funnels are generally subvertical, while flakes within the center of funnels are more horizontally oriented. D) Funnel with a well-defined angular antiform directly above. E) Funnel with a well-defined fluidization channel extending downward from the basal tip and a rounded anticline above and slightly to the right. The anticline has ~30 cm of vertical relief, and appears to involve both CV1 and CV2.
Single fissures extend downward for >1 m from the basal tips of some funnel structures. Above larger funnel structures, bed CV1 may be shaped as an antiform with up to ~30 cm of vertical relief (Fig 4D-E). Antiforms may be rounded or angular. In one case, the fissure-like opening that extends downward from the funnel is ~15 cm thick, and folded strata above the funnel are ~1.5 m thick and include the base of bed CV2 (Fig 4E).

**INTERPRETATION**

The subvertical orientation of shale flakes near the side margins of funnels and the absence of shale and pebbles from bed J6 suggest that material within funnel structures came from bed CV1 (cf. Fortuin and Dabrio, 2008). Isolated antiforms above larger funnels, however, suggest that material from below the Jordan-Oneota contact was injected upward. This contradiction is reconciled by the draw-in process (cf. Takahama et al., 2000). In the first step of the draw-in process, increased pore pressure due to liquefaction of underlying beds causes upward intrusion of water and fluidized sediment. Intrusion is followed by the draw-in of a combination of injected sediment and overlying strata into the void created by spouting (Fig 5).

Funnel-shaped structures occur exclusively at the Jordan-Oneota contact, where clay is more concentrated than elsewhere in the Jordan Formation. This suggests that shale drapes at the base of CV1 served as a permeability barrier against water flowing upward from the Jordan. It is also possible that a permeability contrast existed across the boundary between J6 and CV1 regardless of shale content, since the two beds were deposited in different depositional settings with ~3 million years in between. Silcrete that formed as a result of subaerial exposure could conceivably be responsible for low permeability at the base of CV1 as well.
Figure 5. Illustration of the formation of funnel-shaped structures using the structure from figure 4D as an example. A) The Jordan-Oneota contact before liquefaction. B) During liquefaction, upward migrating water forms a fluidization channel and causes rupture of the shale drapes. Fluidized sand from below the contact is intruded into bed CV1, causing upward folding. C) Sediment from bed CV1 is drawn back into the void left by intrusion, creating a funnel-shaped depresssion.
The fissure-like openings that extend downward from some draw-in structures are interpreted as fluidization channels through which upward migrating water moved (cf. Lowe, 1975). The fact that fluidization channels are only found directly below draw-ins suggests that sand intrusion generally occurred where pore-pressure was greatest. Upward folding of bed CV1 above larger draw-ins indicates that relatively large amounts of sediment were injected upward during intrusion. Based on the size range and abundance of draw-ins, and the local formation of fluidization channels and folded strata, it is unlikely that these structures are the result of normal seepage processes. It is more probable that a liquefaction event resulted in the relatively rapid upward migration of pore water through the Jordan Formation, causing sudden rupture of the shale drapes at the base of the Coon Valley Member.

**Water-escape pillars**

**DESCRIPTION**

Water-escape-pillars occur at breaks in shale drapes at the Jordan-Oneota contact (Fig 6A-D) and at the base of J6 (Fig 6E). Pillars resemble the type B pillars of Lowe (1975). They consist of vertically or subvertically oriented flame-shaped pockets or sinuous strips of structureless coarse-grained sand that is bleached white. A coarser grain size than the surrounding sand suggests elutriation of fines (Lowe, 1975). Pillars at the Jordan-Oneota contact are up to 15 cm tall, and the grain-size contrast with the surrounding sand is generally pronounced. Pillars at the base of J6 are only a few cm tall and the grain-size contrast is less apparent.
Figure 6. Photographs (1) and line drawings (2) of water escape pillars at the Jordan-Oneota contact (A-D) and in the basal shale drapes of the area enclosed in the red box in B. (A) Flame-shaped pocket of coarse-grained sand surrounded by medium-grained sand. (B) A sinuous strip of coarse-grained sand surrounded by medium-grained sand. (C) Close-up of the area enclosed in the red box in B. (D) Multiple small-scale rimples in shale drapes. The shale between rimples is synclinal, similar to local fluidization of sand below them. The shale between rimples is synclinal, similar to local fluidization of sand below them.

Shale drapes directly above the pillar sag due to local fluidization of sand below them.

Shale between rimples is synclinal, similar to local fluidization of sand below them.
INTERPRETATION

Water-escape pillars typically originate above or below breaks in layers that would restrict fluid flow, and around, above, or below less permeable sediment (Lowe, 1975). At Weaver, they are found at depositional gaps or small rupture sites in shale drapes. At the base of J6, pillars occur in the spaces between shale nodules at high points in the nodule layer (Fig 6E). Therefore, fluid escape could have been responsible for deforming the shale drapes into nodules and creating the undulating profile of the layer.

A continuum exists between water-escape pillars and sand intrusions that are the first step of the draw-in process at Weaver. Where upward flow was strong enough to induce small-scale rupturing of shale drapes and mobilize finer sand grains, water-escape pillars developed. Where flow was strong enough to cause larger scale rupturing of the permeability barrier and carry sufficient amounts of fluidized sediment through the gaps, draw-in structures resulted.

Deformation along foreset boundaries in tidal dunes

DESCRIPTION

Foresets of m-scale tidal dunes in J6 are commonly deformed such that the interface between two foresets is distorted at the cm-scale, and the foresets interpenetrate one another. Interpenetration surfaces are of two types: (Type 1) cm-scale digitate interfaces, and (Type 2) *in situ* suspended clumps of sand. To make discussion easier, let the top of a given foreset be referred to as *a* and the base of the succeeding foreset be referred to as *b*.

Type 1: The sharp interface between *a* and *b* is curviform in a regular centimetric pattern (Fig 7A, Fig 8D1). Cm-scale undulations of foreset boundaries may be superimposed over
Figure 7. Photographs (1) and line drawings (2) of deformed foresets of tidal dunes in J6. A) Type 1 interpenetration surface. The interface between foresets is defined by a train of suspended, iron-stained fragments of sand from the top of the lower foreset and the base of the succeeding foreset are intertwined in a regular centimetric pattern. The interpenetration surface displays irregular, dm-scale undulations over cm-scale undulations. See Figure 8D2 for a close-up of the area enclosed in the red box.

B) Type 2 interpenetration surface. The foreset boundary is defined by a train of suspended fragments of fine-to-medium-grained sand from the top of the lower foreset and the base of the succeeding foreset. The iron-stained fragments are darker than the surrounding medium-to-coarse-grained sand due to iron staining. Another Type 2 interpenetration surface in which the boundary is defined by a train of suspended fragments of fine-to-medium-grained sand from the top of the lower foreset are interdigitated in a regular centimetric pattern. The interpenetration surface displays irregular, dm-scale undulations over cm-scale undulations. See Figure 8D2 for a close-up of the area enclosed in the red box.

C) Another Type 2 interpenetration surface in which the boundary is defined by a train of suspended, iron-stained fragments of sand from the top of the lower foreset and the base of the succeeding foreset. See Figure 8D2 for a close-up of the area enclosed in the red box.
Figure 8. Illustration of the formation of Type 1 (1) and Type 2 (2) interpenetration surfaces along foreset boundaries in m-scale tidal dunes. A) Foreset boundary before liquefaction. B1) If b is denser than a and no competency contrast exists across the foreset boundary, lobes of b begin to sink into a while lobes of a rise upward into b. C1) This process continues, resulting in a digitate interface. D1) Photograph of a Type 1 interpenetration surface. B2) If a is more competent than b, the top of a breaks into fragments analogous to boudins while b behaves like a heavy quicksand. C2) b works its way down into the cracks, leaving fragments of a suspended within it. D2) Photograph of a Type 2 interpenetration surface (close-up of the red box in figure 7C).
low amplitude dm-scale undulations (Fig 7A). In some cases, normal grading within deformed foresets is preserved such that fine- to medium-grained sand from the top of one avalanche is interfingered with the medium- to coarse-grained base of the next. Sand of a is commonly more iron stained than sand of b.

Type 2: Cm-scale clumps of fine- to medium-grained sand from a are suspended within medium- to coarse-grained sand of b. Deformed foreset boundaries are thus defined by a train of suspended sand forms rather than by a sharp, continuous interface (Fig 7B-C, Fig 8D2). Suspended forms are closely spaced within trains and may be rectangular or rounded. Most clumps of a are significantly more iron stained than the surrounding sand of b. Like some cases of Type 1 deformation, trains of suspended forms generally display decimeter-scale undulations.

Foreset boundaries are most strongly deformed at the top of J6 near the Jordan-Oneota contact. Both Type 1 and Type 2 boundaries are more prominent in the uppermost ~2 m of the Jordan Formation and are absent at depths > ~3 below the contact.

INTERPRETATION

Interpenetration surfaces are interpreted as the result of a density contrast between a and b due to both grain-size and packing differences. Though quartz sand is of uniform composition, its density decreases with grain-size due to increase in void ratio (Gray, 1968 in Owen, 1996). When the tighter packed, coarser-grained base of a foreset is deposited over the looser packed, finer grained top of the preceding foreset, a reverse density gradient is created. Loading structures resulting from density contrasts across sand-on-sand boundaries have been observed both in the field and in the lab (Anketell et al., 1970; Owen, 1996; Moretti et al., 1999; Nikolaeva, 2006).
Reverse density systems in which a higher density member overlies a lower density member have potential energy, since the high-density member will naturally sink into the low-density member. However, the system may remain static until a trigger causes the lower member to lose shear strength (Anketell et al., 1970). A common means for the release of stored energy is the weakening of grain contacts within the lower member during liquefaction (Anketell et al., 1970). Restriction of the expression of reverse density gradients to the upper ~2-3 m of J6 is consistent with a liquefaction trigger. When upward migrating pore water accumulated beneath the permeability barrier at the Jordan-Oneota contact, the shear strength of the sand decreased due to increasing pore-pressure. When shear strength became sufficiently low, deformation along foreset boundaries was able to occur.

The degree of the competency contrast across foreset boundaries determined whether interfaces between successive foresets were deformed into Type 1 or Type 2 boundaries. When little or no difference in competency existed between $a$ and $b$, lobes of $b$ sank into $a$. The result is a digitate interface with interlocking fingers of $a$ and $b$ approximately equal in size (cf. Anketell et al., 1970). Type 2 boundaries resulted when $a$ was more competent than $b$. In this case, the upper few centimeters of $a$ behaved as a brittle layer while $b$ behaved as a heavy quicksand (cf. Anketell et al., 1970). The quicksand worked its way down into cracks in the brittle layer, causing them to widen. As a result, fragments of $a$ are left suspended in $b$ (Fig 8).

The decimeter-scale undulations seen in both Type 1 and Type 2 foreset boundaries are, like the cm-scale undulations that occur in type 1 boundaries, a result of $b$ being denser than $a$. Since the wavelength of loading structures resulting from reverse density gradients is a function of the thicknesses of the involved layers (Owen, 1996), it follows that cm-scale
undulations will be superimposed over dm-scale undulations when reverse density gradients of two different scales are expressed simultaneously. Tape et al. (2003) note the presence of diffuse internal laminae within foresets of tidal dunes in the Jordan Formation, and attribute them to fallout of suspended sand during unsteady flow conditions. It is possible that cm-scale undulations are superimposed over dm-scale undulations at Weaver where foreset boundaries and the boundaries between well-defined internal laminae coincide.

**Timing of Deformation**

The preservation of isolated antiforms above draw-ins suggests that deformation was intrastratal, since the folds probably would have been leveled if they existed near the depositional surface. In one case, folded strata include the base of bed CV2 (Fig 4E), which indicates that CV2 had already been deposited at the time of deformation. Therefore, the earliest time deformation could have occurred was in the Early Ordovician after the start of Coon Valley deposition. Since deformation is assumed to be intrastratal, it is impossible to define an upper limit on its timing.

**A Common Origin of Deformation**

Draw-in structures and associated features, water-escape pillars, and foreset deformation at Weaver are all the result of obstruction of upward migrating water by a permeability barrier at the Jordan-Oneota contact. The upward succession of features from the lowermost foreset deformation in the Jordan Formation to folded strata in the Coon Valley Member reflects the progression from increasing pore pressure to its release. Deformation is most likely absent from more than ~3 m below the contact because pore...
pressure never grew high enough there to sufficiently weaken grain contacts so that reverse density contrasts could be expressed. Increasing prominence of foreset deformation with increasing proximity to the contact therefore reflects the role of the contact as a permeability barrier. Draw-in and water-escape pillars occur at the contact itself, and were formed when accumulating pore water caused the barrier to rupture. Small water-escape pillars in the basal shale drapes of J6 are attributed to the same event responsible for deformation at the contact. These water-escapes are most likely the result of shale drapes and well-lithified intraclasts acting locally as permeability barriers.

There are other outcrops in southeastern Minnesota where the Jordan-Oneota contact is exposed, and some contain more shale at the contact than occurs at Weaver. However, no other outcrops have nearly the abundance nor as striking an assemblage of deformational features. For features such as draw-ins, fluidization channels, folded strata, and water-escape pillars to form at Weaver and nowhere else, a localized liquefaction event must have occurred that caused water to move rapidly toward the surface.

**Cause of Liquefaction**

Many triggers of liquefaction in sediments have previously been recognized, including breaking waves, tidal currents and bores, slope-induced slumping, rapid sedimentation, ground water phenomena, and seismic activity.

**Waves and tides**

Because the Hollandale Embayment was a shallow marine environment, it is possible that loading by waves and tides could be responsible for deformation. Dalrymple (1979)
found that under fair weather conditions in the Bay of Fundy, waves 10-30 cm high caused slumping down the stoss sides of dunes and foreset deformation up to 35 cm below the surface in fine sand. It is likely that larger storm waves would cause deformation further below the surface. However, large waves generally turn sediments into bulk homogenous bodies (Greb and Dever, 1998). Since Coon Valley deposition had already begun at the time of deformation, liquefication of sands below the Jordan-Oneota contact by waves would have distorted bedding immediately overlying the contact as well. Because bedding in CV1 and CV2 is undeformed, a wave mechanism for liquefaction is unlikely.

Tides are also capable of liquefying unconsolidated sands. According to Greb and Archer (2007), contorted bedding and flow rolls formed on the sediment surface in Turnagain Arm, Alaska as a result either of shear from a 1.8 m tidal bore or overpressurization of tidal flats during rapid tidal drawdown. Dewatering pipes 1-5 cm tall occurred between flow rolls. Though tides in the Alaskan estuary rank sixth highest in the world, deformation is restricted to within centimeters of the sediment-water interface. Therefore, it is not likely that tides are responsible for liquefaction that occurred meters below the sediment surface.

**Rapid sedimentation and slope-induced slumping**

Liquefaction can be induced by rapid sedimentation upon unconsolidated beds (Owen and Moretti, 2008). Since the gradient of the Hollandale Embayment was extremely shallow (~1 m/km or less) (Runkel et al., 2007), there would have been little potential for mass sedimentation (cf. Draganits et al., 2001). One mechanism that remains a possibility is slope-induced slumping due to undercutting by laterally migrating tidal channels (Owen, 1987; Greb and Dever, 1998). Runkel (1994) noted tidal channel deposits in the Jordan Formation,
so it is possible that undercutting of m-scale dunes or tidal channel walls could have resulted in the collapse of large piles of sediment and led to compaction of underlying beds.

However, features commonly associated with slumping are not seen at Weaver. According to Owen (1996), collapse of sloping heaps of sand causes strata at the toe of the slope to be folded due to basal shear, and stratification at the surface to be flattened. Since deformation is interpreted to be Early Ordovician or later, slumping would have occurred after the period of subaerial exposure that caused extensive erosion of the top of the Jordan. Therefore, if slumping involved enough sediment to cause compaction of underlying beds, features indicative of slumping would probably be at least partially preserved. Because basal shear zones and other types of foreset deformation are absent from the Coon Valley Member, compaction due to slope-induced slumping is not considered a likely mechanism for liquefaction.

**Groundwater processes**

Dragnatis et al. (2003) attribute the formation of cylindrical and cone-shaped fluidization pipes in the Muth Formation, a quartz arenite deposited in a barrier island system, to a rise in hydrostatic head due to rapidly rising sea level. Most fluidization pipes originate at interfaces between beds, are concentrically laminated, and have very little sand extruded on top of them. Dragnatis et al. (2003) interpret the size (up to 155 cm tall and 80 cm in diameter) and regularity of the pipes to mean that flow was relatively continuous over an extended time period. Because the Hollandale Embayment experienced a period of relatively rapid sea level rise starting in the Early Ordovician, it is conceivable that a rise in hydrostatic head was responsible for increased pore pressure in the Jordan Formation.
However, the irregular shapes of draw-in structures and fluidization channels suggest that deformation at Weaver was caused not by steady, continuous flow, but by brief, forceful expulsion of water (cf. Draganits et al., 2001; Gallo and Woods, 2004). Creation of low-pressure spaces that resulted in draw-in, as well as the antiforms above larger draw-ins, indicate mobilization of relatively large amounts of sediment, which is in contrast to deformation in the Muth Formation. It is possible, however, that if the Muth Formation had contained permeability barriers above fluidization pipes, high pore pressures would have caused local rupture of the barriers and more extensive sediment spouting resulting in draw-in. Therefore, although deformational features at Weaver bear little resemblance to fluidization pipes in the Muth Formation, a groundwater-related mechanism for liquefaction cannot be ruled out.

**Seismic activity**

During earthquakes, upward propagation of cyclic shear waves can cause liquefaction of unconsolidated, water-saturated sands up to 10 (though sometimes as much as 20) m below the sediment surface (Obermeier, 1996). Previous studies have attributed a variety of soft sediment deformation to earthquake-induced liquefaction, including sand intrusions (Greb and Dever, 1998), water-escape structures (Lowe, 1975), and deformation due to reverse density gradients (Nikolaeva, 2006). In each example of the draw-in process referred to by Takahama et al. (2000), earthquakes were the cause of liquefaction. In a lab experiment using a shaking table to simulate the Loma Prieta earthquake (10/17/1989, Los Angeles), liquefaction-induced sand volcanoes and water-escape pillars formed in an artificial sand bed capped by a thin clay layer (Moretti et al., 1999). When fine sand was overlain by coarser
sand in the absence of a permeability barrier, liquefaction induced by shaking resulted in an undulating interface similar to Type 1 interpenetration surfaces in tidal dunes at Weaver. While these experiments do not suggest that liquefaction by any other mechanism is incapable of producing comparable features, it is notable that every deformational feature seen at Weaver is consistent with earthquake-induced liquefaction. In addition, the large size of some draw-ins and the folded strata above them are indicative of the sudden application of a large hydraulic force, as would result from an earthquake (cf. Obermeier, 1996).

**Discussion**

One reason Dragnatis et al. (2003) attribute fluidization pipes to rising groundwater rather than seismicity is that compaction of sediment during an earthquake is only ~3-6% of the thickness of the liquefied strata. Therefore, a single earthquake would not expel sufficient amounts of water for flow to persist over a long enough period to produce abundant meter scale fluidization pipes. Furthermore, Dragnatis et al. (2003) attribute the internal lamination within pipes to flow having been less vigorous than would generally be expected of short-lived liquefaction events such as earthquakes.

At Weaver, there is no evidence of steady, persistent flow. A combination of liquefaction and shaking could have caused interpenetration surfaces to form over a relatively short time period (cf. Moretti et al., 1999), and draw-ins and folded strata are indicative of forceful expulsion of water (cf. Obermeier, 1996). In addition, the permeability barrier at Weaver varies laterally in strength, since shale drapes at the Jordan-Oneota contact are very thin at places along the outcrop. Had increased pore-pressure that led to deformation been the result of slow accumulation of water due to rising sea level, water probably would have
escaped through preexisting weaknesses in the barrier rather than causing its rupture against meters of confining pressure. On the other hand, if pore-pressure rose relatively quickly during vigorous expulsion, water would have been less likely to migrate laterally beneath the permeability barrier until it found an opening. Therefore, seismic activity seems to be a more likely mechanism than a rise in hydrostatic head for liquefaction at Weaver.

While deformatonal features at Weaver bear strong resemblance to seismites identified in other formations and to structures created in the lab by shaking, two important criteria that are necessary to pin deformation on a seismic event must be discussed (cf. Sims, 1975). These two criteria relate to the tectonic setting at the time of deformation and the distribution of deformatonal features.

Though initial spreading associated with the Midcontinental Rift began ~600 Ma before the Jordan and Oneota were deposited, faults along the western boundary of the Rift were apparently reactivated during the early Paleozoic era. Cross-sections of bedrock maps of the Twin Cities area issued by the Minnesota Geological Survey show thinning of the Tunnel City Group (Late Cambrian) on up-thrown sides of faults and thickening on the downthrown sides, indicating syndepositional movement of faults (Mossler, 2009). The same trend is evident in the Prairie due Chien Group (Mossler, 2006), which suggests that movement along faults was taking place around the time that deformation at Weaver occurred. While thickening and thinning of Paleozoic units across fault zones has not been documented within less than ~50 km of Weaver, multiple faults do exist within ~20-30 km to the east of Weaver (Mossler, 2001). Motion along any of these faults could theoretically be responsible for an earthquake capable of causing liquefaction.
Deformational features caused by earthquakes generally increase in scale and number toward the paleoepicenter (Greb and Dever, 1998). An earthquake of magnitude 6.6 causes deformation within a 50 km radius (van Vliet-Lanoe et al., 2004). Since a magnitude of 5.5 is required for liquefaction of sand (Obermeier, 1996), features similar to those at Weaver should span across a significant area if liquefaction was earthquake-induced. Distribution of deformational features is relevant with respect to groundwater processes as well. If deformation at Weaver is in fact linked to a regional sea level rise, then comparable deformation should be seen at other locations where permeability barriers exist.

**Deformation at other outcrops**

Although deformation at Weaver is far more pronounced, less distinct and sparser deformational features are present in outcrops at Red Wing and Stillwater, Minnesota. At Red Wing, shale drapes up to ~5 cm thick on foresets of m-scale dunes are commonly deformed into twisted convolutions (Fig 9A). In one instance, the boundary between successive foresets is deformed in a manner reminiscent of type 1 interpenetration surfaces at Weaver (Fig 9B). Deformation at Red Wing could be the result of loading by waves and tides, groundwater processes, seismic activity, and possibly even synsedimentary processes. Since shale is more abundant at the Jordan-Oneota contact at Red Wing than it is at Weaver and deformation is restricted to within ~2 m of the contact, Red Wing deformation may also be a combined result of shaking and increased pore-pressure due to obstruction of upward migrating water. However, a mechanism responsible for Red Wing deformation is not determined here.
Figure 9. Photographs (1) and line drawings (2) of deformation in the Jordan Formation in Red Wing (A-B) and Stillwater (C-D), Minnesota. A) A ~5cm thick shale drape on a foreset in a tidal dune ~1 m below the Jordan-Oneota contact that is deformed into twisted convolutions. B) The boundary between successive foresets in a tidal dune ~1.2 m below the Jordan-Oneota contact displays cm-scale undulations similar to Type 1 foreset deformation at Weaver. C) A sand intrusion in offshore deposits at least ~12 m below the Jordan-Oneota contact. There is no shale within or near the intrusion. D) A sand dike ~5 m away from and along the same horizon as C.
At Stillwater, sand intrusions occur sporadically along a single stratigraphic horizon at least ~12 m below the Jordan-Oneota contact for ~20 lateral meters. Most intrusions are rounded pockets of sand not more than ~10-15 cm in dimension (Fig 9C). The only exception is a sand dike ~15 cm tall that partially fills a vertical crack ~50 cm long in the overlying bed (Fig 9D). Since the crack grows thinner upwards and fades out, and since there are no other cracks in the intruded bed, the crack is interpreted to have formed with the dike. Because it would have required a considerable hydraulic force to cause fracturing and upward intrusion into a well-lithified bed, earthquake-induced liquefaction is the most obvious choice for a mechanism.

Red Wing is ~50 km northwest of Weaver, so deformation at these two locations could theoretically have been caused by the same seismic event. However, Stillwater is ~100 km northeast of Weaver, and sand intrusions at Stillwater occur more than 12 m below the Jordan-Oneota contact. Therefore, it is unlikely that a single earthquake is responsible for sand intrusion at both localities (cf. Sims, 1975). While it is possible that multiple seismic events occurred in the Hollandale Embayment over millions of years, this may support a rise in hydrostatic head due to rising sea level as a mechanism for liquefaction in the Jordan Formation. However, we may then expect to see features somewhere in the Jordan Formation that resemble those in the Muth Formation.

**Conclusions**

Draw-in structures, fluidization channels, folded strata, and water-escape pillars occur at the Jordan-Oneota contact at a single outcrop near Weaver, Minnesota. Foreset deformation in m-scale tidal dunes also occurs within ~3 m below the contact, and is an
expression of reverse density gradients across boundaries between successive foresets. All
deformation was caused by increased pore pressure due to obstruction of upward migrating
water by a permeability barrier at the Jordan-Oneota contact. Upward migrating water is
interpreted as the result of liquefaction of underlying beds triggered by either seismic activity
or a rise in hydrostatic head due to a regional sea level rise. Because draw-in structures and
folded strata suggest forceful expulsion of water, earthquake-induced liquefaction is favored
here.

Foreset deformation in the same stratigraphic position ~50 km to the northeast may
have formed by similar processes as foreset deformation at Weaver. Sand intrusions at least
~12 m below the Jordan-Oneota contact ~100 km to the northeast may also have formed by
the same mechanism as draw-in structures at Weaver, though it is unlikely that a single
seismic event is responsible for deformation across such a long distance and stratigraphic
range. While this may support a rise in hydrostatic head as the mechanism for liquefaction in
the Jordan Formation, there is no evidence of steady, persistent flow associated with changes
in hydrostatic head.

Acknowledgements

This study was enthusiastically funded by the Carleton College Geology department.
The guidance of Clint Cowan and the advice of Tony Runkel were invaluable through the
course of this project, and without them there would be nothing else. Thanks to Tony and
James Miller for conducting conodont biostratigraphy. We got that unconformity. Thanks
also to Nate Ryan and Sarah Meller for assistance with photography, which was quite
necessary. Lastly, thanks to the Carleton Geology faculty and staff for correction and resurrection, and to my fellow majors for commiseration and elevation.

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