ASSESSING THE ROLE OF SILICA GEL AS A FAULT WEAKENING MECHANISM IN THE TUSCARORA SANDSTONE

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ABSTRACT

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Previous work demonstrated that microcrystalline quartz bands and cement in cataclasites related to faults in quartz-rich rocks may have derived from a silica gel parent fluid (Onasch et al., 2010). Occurrences of microcrystalline quartz on various fault surfaces of the Tuscarora Sandstone and other lithologies raised the possibility that a silica gel may have been present during faulting and significantly reduced the frictional strength. Fractured grains, trapped silica nanospheres, flow features, amorphous silica, uniform grain size distribution and recrystallized grains of low dislocation density within microcrystalline quartz on the fault surface are interpreted to have resulted from comminution and hydration of wallrock asperities along the fault surface coeval with Mode I fracturing and subsequent gel formation that resulted in dynamic weakening. The viscous gel may have promoted greater displacements that allowed migration of the gel from the fault surface into adjacent open Mode I fractures to rapidly precipitate opaline phases and upon silica depletion, quartz microveins less rapidly. Complex mutually crosscutting relations between microfractures, microcrystalline quartz and microveins as well as presence of brecciated clasts within breccia indicate multiple episodes of brittle deformation. Mutually overprinting textures between brittle and fault creep microstructures (stylolites) suggest alternating brittle and ductile episodes. As silica-rich rocks are common, understanding the mechanics associated with the formation of silica gel on fault surfaces of quartz-rich rocks may provide new insight into mechanics associated with other microcrystalline quartz-hosted faults in silica-rich rocks.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. INTRODUCTION</td>
<td>1</td>
</tr>
<tr>
<td>2. BACKGROUND</td>
<td>3</td>
</tr>
<tr>
<td>2.1. Formation of a silica gel</td>
<td>3</td>
</tr>
<tr>
<td>2.2. Silica gel in rock friction experiments</td>
<td>5</td>
</tr>
<tr>
<td>2.3. Silica gel and natural faults</td>
<td>8</td>
</tr>
<tr>
<td>3. STUDY AREA</td>
<td>16</td>
</tr>
<tr>
<td>4. METHODS</td>
<td>17</td>
</tr>
<tr>
<td>4.1. Field Methods</td>
<td>17</td>
</tr>
<tr>
<td>4.2. Laboratory Methods</td>
<td>17</td>
</tr>
<tr>
<td>5. FAULT FEATURES</td>
<td>19</td>
</tr>
<tr>
<td>5.1. Outcrop and Handsample description</td>
<td>19</td>
</tr>
<tr>
<td>5.2. Transmitted Light Microscopy</td>
<td>21</td>
</tr>
<tr>
<td>5.3. Cathodoluminescence Microscopy</td>
<td>29</td>
</tr>
<tr>
<td>5.4. Scanning Electron Microscopy</td>
<td>35</td>
</tr>
<tr>
<td>5.5. Transmission Electron Microscopy</td>
<td>38</td>
</tr>
<tr>
<td>6. DISCUSSION</td>
<td>42</td>
</tr>
<tr>
<td>6.1. Evidence for a gel origin in the formation of microcrystalline quartz on the fault surface</td>
<td>42</td>
</tr>
<tr>
<td>6.2. A model for the formation of microstructures at the fault surface</td>
<td>43</td>
</tr>
<tr>
<td>7. CONCLUSIONS</td>
<td>47</td>
</tr>
<tr>
<td>REFERENCES</td>
<td>49</td>
</tr>
</tbody>
</table>
APPENDIX A ............................................................................................................. 52
APPENDIX B ............................................................................................................. 55
# LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Geologic map of the Appalachian foreland showing distribution of sample</td>
<td>17</td>
</tr>
<tr>
<td></td>
<td>locations</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Fault surfaces in the Tuscarora Sandstone as seen in outcrop</td>
<td>20</td>
</tr>
<tr>
<td>3</td>
<td>Fault surface features in the Tuscarora Sandstone as seen in outcrop</td>
<td>20</td>
</tr>
<tr>
<td>4</td>
<td>Fault surface features in the Bloomsburg Formation as seen in outcrop</td>
<td>21</td>
</tr>
<tr>
<td>5</td>
<td>Microstructures associated with microcrystalline quartz bands in the</td>
<td>23</td>
</tr>
<tr>
<td></td>
<td>Tuscarora Sandstone</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>Microstructures associated with microcrystalline quartz bands in the Juniata</td>
<td>23</td>
</tr>
<tr>
<td></td>
<td>Formation</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>Silica layer on a fault surface in the Bloomsburg Formation</td>
<td>24</td>
</tr>
<tr>
<td>8</td>
<td>Crosscutting relationships near a fault surface in the Tuscarora Sandstone</td>
<td>24</td>
</tr>
<tr>
<td>9</td>
<td>Photomicrograph mosaic and sketch showing spatial and temporal</td>
<td>26</td>
</tr>
<tr>
<td></td>
<td>relationships of brittle microstructures with respect to fault surfaces in</td>
<td></td>
</tr>
<tr>
<td></td>
<td>the Tuscarora Sandstone</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>Crosscutting relationships between different microstructures in the</td>
<td>27</td>
</tr>
<tr>
<td></td>
<td>Tuscarora Sandstone</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>Photomicrograph mosaic showing a microcrystalline quartz band along a</td>
<td>28</td>
</tr>
<tr>
<td></td>
<td>fault surface in the Tuscarora Sandstone</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>Extent of deformation on striated and polished fault surfaces in the</td>
<td>29</td>
</tr>
<tr>
<td></td>
<td>Tuscarora Sandstone</td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>Matched cathodoluminescence and polarized light micrograph mosaics of a</td>
<td>31</td>
</tr>
<tr>
<td></td>
<td>polished fault surface in the Tuscarora Sandstone</td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>Matched cathodoluminescence and polarized light photomicrographs of</td>
<td>33</td>
</tr>
<tr>
<td></td>
<td>microstructures near a fault surface in the Tuscarora Sandstone</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>Matched cathodoluminescence and polarized light photomicrographs of</td>
<td>34</td>
</tr>
<tr>
<td></td>
<td>microstructures near a fault surface in the Tuscarora Sandstone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>---</td>
<td>-----------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>16</td>
<td>Matched cathodoluminescence and polarized light photomicrographs of microstructures near a fault surface in the Tuscarora Sandstone</td>
<td>35</td>
</tr>
<tr>
<td>17</td>
<td>SEM micrographs of fault surfaces in the Tuscarora Sandstone</td>
<td>35</td>
</tr>
<tr>
<td>18</td>
<td>SEM micrographs of a polished slickenside in the Tuscarora Sandstone</td>
<td>36</td>
</tr>
<tr>
<td>19</td>
<td>Elemental composition of polished surfaces</td>
<td>37</td>
</tr>
<tr>
<td>20</td>
<td>SEM micrograph of a polished fault surface showing uniform grain size distribution</td>
<td>38</td>
</tr>
<tr>
<td>21</td>
<td>TEM micrograph of the sample cut normal to the fault surface</td>
<td>39</td>
</tr>
<tr>
<td>22</td>
<td>TEM micrograph showing flow features and diffraction patterns</td>
<td>40</td>
</tr>
<tr>
<td>23</td>
<td>TEM micrographs showing different textures along the fault surface</td>
<td>41</td>
</tr>
<tr>
<td>24</td>
<td>TEM micrograph showing a trapped ellipsoidal inclusion in a pore space</td>
<td>41</td>
</tr>
<tr>
<td>25</td>
<td>Illustrations showing different mutually crosscutting relations between microstructures near the fault surface</td>
<td>46</td>
</tr>
</tbody>
</table>
1. INTRODUCTION

Frictional fault theory from experimental data suggests that high shear stresses are necessary to support topography and lithospheric flexure during seismic slip (Zoback et al., 1987). According to Byerlee’s law, the calculated average shear stress on a fault is 56 MPa (Sandwell, 2001) and most in situ stress measurements in the upper crust suggest high tectonic stresses (Goldsby and Tullis, 2002). However, many faults slip under much lower shear stresses than those predicted by Byerlee’s law (Noda et al., 2011), implying that strength reduction prior to slip is critical for dynamic rupture of earthquakes (Reches and Lockner, 2010). If dynamic friction at seismic slip rates (~1 m/s) is low, then either stress drops during earthquakes are large or tectonic stress is actually much lower than in situ measurements (Goldsby and Tullis, 2002). Strength reduction along faults can occur through several mechanisms such as flash heating (Goldsby and Tullis, 2011), frictional heating (Evans et al., 2014), dynamic recrystallization (Smith et al., 2012), or formation of nanoparticles (De Paola et al., 2011). Although studies indicate that faults are weakened during earthquakes regardless of rock composition, the lack of evidence of seismic slip on ancient faults limits our understanding of earthquake mechanics and assessments of seismic hazards (Fondriest et al., 2013).

However, laboratory experiments focused on understanding earthquake mechanics have discovered a different weakening mechanism that occurs by formation of a silica gel on the fault surface that results from extreme comminution of silicates in the presence of water (Tullis and Goldsby, 2002). This, coupled with findings of a relic silica gel on a seismically active fault in California by Kirkpatrick et al. (2013), suggest that a gel may have provided the frictional instability that led to seismic slip. Additional syn-kinematic evidence for a silica gel precursor was reported by Stel and Lankreyer (1994) in fault gouges of the Sanddøla fault in Norway.
Sweetman and Tromp (1991) reported unusual textures in radiate, bladed quartz boulders found in the Lower Zambezi valley of northern Zimbabwe, which they believe to be derived from a silica gel associated with normal and strike-slip faults of the Chewore Complex.

Onasch et al. (2010) found tabular bands of microcrystalline quartz in sandstones of the central Appalachian foreland. Based on a lack of shear offset and wall rock fragments, gradational contacts with the wall rock, and a fine-grained quartz fill that has a different cathodoluminescence, water content, and oxygen isotope chemistry than the wall rocks, they argued that these bands were dilational in origin and the microcrystalline quartz was precipitated as a cement, possibly from a gel. The microcrystalline quartz is particularly well developed in the Lower Silurian Tuscarora Sandstone where it occurs not only in bands, but as cement in cataclasite related to faults. That a silica gel may have been present during deformation raises the possibility that gel formation plays an important role in faulting of these quartz-rich rocks. To evaluate this possibility, we studied the occurrence of microcrystalline quartz in the Tuscarora Sandstone and other lithologies associated with faults of different scales and displacements. As several lines of evidence point to an amorphous silica precursor to the microcrystalline quartz, we propose it originated by amorphization and hydration of wall rock asperities during comminution, and migration of the resulting silica-supersaturated fluid along the fault surface and into coeval Mode I fractures. Since the slip-weakening effects of silica gel are significant regardless of its origin, we discuss the role of silica gel as an indicator of dynamic weakening.
2. BACKGROUND

2.1. Formation of a silica gel

Silica gel (SiO$_2$.nH$_2$O) is amorphous porous silica (Shevkina et al., 2012) that forms as a result of amorphization during comminution of quartz in presence of water (Tullis and Goldsby, 2002). Sources of water are limited to adsorbed water, atmospheric water and water in pores and fluid inclusions, which may be released during sliding (Goldsby and Tullis, 2002). Silica particles within the gel consist of layers of various types of hydroxyl groups (Shevkina et al., 2012). According to de Freitas (2009), gels begin as a molecular dispersion in water forming a soft solid capable of hardening. SiO$_4$ loosely-bound with OH is the molecular dispersion of H$_4$SiO$_4$, which initiates silica gel formation. As H$_2$O molecules condense from this system, Si from the original structure bond with other Si by sharing a single O to form polymers (Si-O-Si). These polymers combine to form colloidal spheres of silanol/silicic acid in suspension which, in sufficient numbers, may produce a sol. Being an unstable and incoherent system, sol can coalesce to form a more stable and coherent system referred to as gel. Silica gel is capable of dehydtration, even in water, a process that can lead to formation of crystalline silica phases. A hydrous amorphous form of silica (SiO$_2$.nH$_2$O) that forms from a silica gel consists of silica spheres bound together to generate an orderly 3D structure of opal. Naturally, the solid generated will have pore spaces due to coalescing of spherical particles of similar sizes (~150 nm) (de Freitas, 2009).

The rate-limiting step for the precipitation of silanol is the breaking of Si-O bonds (Rimstidt and Barnes, 1980). This implies that the rate and amount of gel formation is dependent on the rate and amount of comminution (Rimstidt and Barnes, 1980). A large activation energy is required to produce silanol due to the difficulty in breaking strong Si-O bonds. However,
frictional processes promote the reaction even at moderate temperatures due to tribochemical and mechanochemical effects (Hayashi and Tsutsumi, 2010). These effects are most active at high pressure points at contacts of asperities on the fault surface (Hayashi and Tsutsumi, 2010). Shear at these points allows distortion and breaking of Si-O bonds, making them highly reactive, particularly for the OH groups. Presence of strained bonds on the surface due to mechanical abrasion results in high solubility of the comminuted particles and therefore rapid dissolution to produce silanol (Rimstidt and Barnes, 1980).

Amorphous material has been reported in silicate gouges resulting from friction sliding experiments (Yund et al., 1990; Tullis and Goldsby, 2002). According to de Freitas (2009), amorphization of a silicate particle involves breaking the lattice structure and increasing the surface area to volume ratio. Once broken, other elements hosted in the lattice structure (i.e. Al, Fe, Ca, Mg, K, Na, etc.) escape and leave behind a damaged lattice structure enriched in silica. This enrichment and lack of order in broken lattice structures provide the environment for development of a sol that coalesces to form a gel on the surface (de Freitas, 2009). Strained Si-O-Si bonds on the surface attract hydroxyl groups and preferentially react with water to form hydrated amorphous silica on silicate particles, pore spaces and collectively on friction surfaces resulting in dynamic weakening (Nakamura et al., 2012). This may explain why most of the silica precipitates in cataclasites and fractures occur as cements and overgrowths on silicate substrates that provide nucleation sites for growth. Silica phases precipitating from silica-undersaturated fluids are generally restricted to overgrowths on preexisting quartz surfaces while those precipitating at higher saturation levels are deposited on other surfaces in addition to quartz and may form one or more silica polymorphs by homogenous nucleation (Okamoto et al., 2010).
Although silica gel on sliding surfaces generated in rock friction experiments is formed by amorphization of quartz in presence of water during comminution, it is not necessarily the only origin of silica gel. Silica gel on natural fault surfaces may be derived from an external source where silica-saturated hydrothermal fluids are injected into open-space cavities. Stress drop accompanying Mode I fracture formation (Engelder, 1992) causes a sudden drop in the solubility and an already silica-saturated solution becomes supersaturated. This is a particularly favored origin in cases where the source rock is silica-depleted as reported in the Shimanto accretionary complex (Uijie et al., 2007). Although the host rock does not contain enough silica to generate a silica gel through comminution, silica-depletion is compensated by mineralization in preexisting fractures or faults (Power and Tullis, 1989; Faber et al., 2014) or as injections of silica-supersaturated fluids, providing a potential for this mechanism to occur. Hydrous silica (opal) is common in veins and breccia fills and may by formed by fault slip processes or independent fluid flow from a different source (Onasch et al., 2010; Faber et al., 2014). Regardless of the source of silica gel on faults, experiments and studies on natural faults demonstrate that once formed, slip-weakening effects are significant due to the mechanical properties of silica gel (Faber et al., 2014).

2.2. *Silica gel in rock friction experiments*

Laboratory experiments on rock friction conducted by Tullis and Goldsby (2002) focused on understanding earthquake mechanics. According to Tullis and Goldsby (2002), dynamic strength reduction of the sliding surface of silicates may be due to formation of a weak silica gel layer. Sliding experiments were conducted on two groups of novaculite (>99% quartz) samples: air-dried samples and oven-dried samples. Air-dried samples were left in their original state without any attempts to expel water from them while the oven-dried samples were heated in a
furnace at 850°C prior to sliding to expel any intercrystalline or intracrystalline water. Each sample was slid at a normal stress of 5 MPa in a rotary shear apparatus at 1 μm/s for 2 mm before increasing the sliding velocity to 3.2 mm/s. Friction coefficient μ for the room-dry sample decreased dramatically within 25 m of displacement from ~0.7 to less than 0.2. In contrast, μ for the oven-dried sample remained high (~0.8) if the sample was kept dry during sliding. However, when the oven-dried sample was exposed to humidity, significant weakening occurred (μ < 0.2). Upon exposure to dry nitrogen once more, μ increased immediately to ~0.8. When slid again under room humidity, μ decreased significantly (~0.1). In absence of water, the sliding surface underwent extreme comminution and amorphization of quartz, but without gel formation. However, introducing the oven-dried samples to humidity resulted in gel formation by hydration of the comminuted material. This reversible strengthening and weakening though drying and wetting of samples demonstrates the role of water in silica gel formation and subsequent weakening of silica-rich rocks at intermediate slip rates. In addition to mechanical data, examination of the sliding surface of the air-dried sample revealed interesting microstructures. Elevated portions of the surface were smooth, reflective and lacked features. Depressions at the edges of the elevated areas displayed features indicating flow of a viscous material into low spots of the surface. Tullis and Goldsby (2002) attributed the smooth textures of the elevated portions to formation of a film of hydrated silica on the surface and the presence of viscous material in the depressions to dragging of silica gel to low spots by sliding.

Hayashi and Tsutsumi (2010) performed a series of similar friction experiments on chert at intermediate to high slip velocities under low normal stresses (1.5 MPa). Sliding at an initial velocity of 87 mm/s produced a gradual decrease in μ from 0.2 to 0.08 within the first 10 m of slip and maintained a steady coefficient of friction until slip velocity was decreased to 2.6 mm/s.
Following this change in velocity, $\mu$ increased to 0.3 over 1 m of slip and approached steady state friction level until slip velocity was increased once again to 87 mm/s. At this point, $\mu$ decreased to 0.1 over 20 m of slip and approached another steady state level of friction. Overall, frictional strength decreased as slip velocity increased and vice versa (Hayashi and Tsutsumi, 2010). Tullis and Goldsby (2002) attributed extraordinary weakening during gel formation at rapid slip rates to the thixotropy of the gel layer. Weakening at high velocities is caused by drop in shear resistance and viscosity due to breakdown of interparticle bonds between silica particles within the gel. When a previously sheared silica gel is allowed to rest, reconstruction of interparticle bonds within the gel increases the viscosity as a function of time (Tullis and Goldsby, 2002).

Recent friction experiments conducted at large displacements by Nakamura et al. (2012) produced sliding surfaces that revealed ductile deformation. Micro-Raman spectra of these surfaces showed broadening of quartz main band, blue shifts and new peaks, all indicative of strain-induced amorphization of quartz by developing four- and three- membered Si-O rings in six-membered quartz. The decrease in Si-O-Si angle, distortion and densification of the structure picked up by Raman spectra in combination with lack of well-defined diffraction patterns from TEM analyses restricts the mechanical process to comminution of quartz. Micro-FTIR analyses of these surfaces show broad peaks (3000-3600 cm$^{-1}$) that indicate OH symmetric stretching band of H$_2$O molecules. This suggests that hydration of quartz on the surface occurred due to friction during sliding. Although this reaction requires a large activation energy, the strained Si-O-Si bonds preferentially reacted with water molecules to form hydrated amorphous silica, which most probably caused weakening (Nakamura et al., 2012). Given the presence of asperities due to natural roughness in a rock against rock interface, it is believed that the evolution of friction is controlled by mechanical and/or chemical processes at contact asperities.
(Niemeijer et al., 2012). Since shear forces at the frictional contact distorts and breaks the SiO$_4$ network, it is possible that the hydrolysis reaction occurred at highly stresses asperities (Nakamura et al., 2012). This is consistent with microstructural observations by Tullis and Goldsby (2002) where the most elevated portions of the sliding surface were smooth and featureless and composed of hydrous amorphous silica film, suggesting that these areas were most probably former asperities where strain-induced amorphization and subsequent hydration reactions occurred to form silica gel. In real faults, it is expected that silica gel produced this way would be distributed along the fault surface with displacement (Nakamura et al., 2012) as indicated by flow structures and accumulation of viscous material in depressions (Tullis and Goldsby, 2002).

2.3. **Silica gel and natural faults**

Slickensides are smooth fault surfaces with or without linear striations parallel to the slip direction (Power and Tullis, 1989). Slickensides that form during rapid sliding between rock surfaces may be characterized by natural, highly polished surfaces referred to as fault mirrors (FM) (Siman-Tov et al., 2013). Studies that extrapolate results from friction experiments to those formed in nature attribute formation of FM’s to ancient seismic slip (Fondriest et al., 2013). Although certain slickensides are indicative of different stages of the earthquake cycle (Power and Tullis, 1989), highly reflective rock surfaces do not necessarily indicate rapid fault slip (Evans et al., 2014). Other than presence of a pseudotachylyte, there is no universally accepted indicator of ancient seismic slip along faults (Fondriest et al., 2013; Faber et al., 2014). To demonstrate a seismic origin for a fault surface, its mineralogy and microstructures must be linked to temperature, strain rate and/or equilibrium pressure relative to the setting of the fault zone (Evans et al., 2014). Microstructural characteristics of slickensides provide insight into
grain-scale deformation mechanisms operating along frictional rock surfaces (Power and Tullis, 1989). Given the wide range of strain rates that prevail along faults, microstructural characteristics are important in understanding mechanical properties of fault zones and their evolution (Power and Tullis, 1989).

Several slickensides have been documented in nature, most of which have different weakening mechanisms attributed to their formation. However, the first documented occurrence of natural silica gel coating a seismically active fault surface was reported in a study by Kirkpatrick et al. (2013) which focused on the chert-hosted Corona Heights fault slickenside in San Francisco, California. The surface is covered by a shiny layer of translucent silica displaying flow banding, armored clasts and extreme comminution compared to adjacent red and white cataclasites. Flow banding is defined by variations in silica grain-size and iron oxide content. Fine grooves on this surface form elongate troughs consisting of circular cracks that form elongate clusters parallel to the groove direction. The boundary between the silica layer and adjacent cataclasites is intercalated and embayed, suggesting local mixing. The cataclasites consist primarily of a granular matrix partially cemented with fine-grained quartz similar to the silica layer. Presence of colloidal silica particles in hydrous silica grains of the silica layer reflects nucleation of crystalline material around colloids. Iron oxide rings around the colloids indicate surrounding secondary phases. TEM diffraction patterns reveal amorphous regions amidst the crystalline material. Hexagonal crystalline material, grain boundary pores, preferred crystallographic orientation and absence of deformation-induced dislocations in all phases of the silica layer indicate that most of the crystallinity postdates slip. These lines of evidence suggest a syndeformational amorphous precursor (i.e., solidification of the silica layer from a silica-supersaturated fluid) (Kirkpatrick et al., 2013). A relict silica gel that formed during slip has the
potential to act as a fault weakening mechanism under shallow crustal conditions due to frictional instability owing to its thixotropic nature and can induce seismic slip (Goldsby and Tullis, 2002; Di Toro et al., 2004).

Fault gouge occurs along brittle faults as extremely fine-grained phases that affect the rheology of fault zones (Stel and Lankreyer, 1994). A study on the rheology of the Sanddøla fault in Norway by Stel and Lankreyer (1994) provides additional evidence for the presence of syn-kinematic viscous silica-oversaturated fluids in low-grade cataclasites. The fault is primarily characterized by extensive hydrothermal veins of quartz and chalcedony hosted by leucocratic gneisses. Clast-loaded veins display textures indicative of flow. Clasts in some veins are settled and show patterns similar to graded bedding in sedimentary rocks or crystal settling in magmas suggesting dispersion in a medium of low viscosity. Other veins consist of unsettled clasts that appear to float in a cryptocrystalline matrix suggesting settling in a highly viscous medium. The obstruction of particle settling can be explained by an increase in viscosity of the fluid with increasing polymerization, as seen in polymerization of silica sol to gel. Presence of flow textures such as pinch and swell structures and sigmoidal shear bands in the fault gouge suggest that viscosity was initially low enough to allow flow but eventually increased to higher values (~10^4 to 10^6 Poise) that preserved flow instabilities. The cryptocrystalline matrix consists of spherulitic clusters of quartz nucleated on crush fragments. In absence of crush fragments, banded chalcedony formed, possibly from a silica gel precursor where gel colloids acted as nucleation centers for subsequent growth. Disruption of chalcedony bands by wedge-shaped veins that retained parallelism of adjacent bands across the vein walls suggests that the veins represent desiccation cracks. Textures of the Sanddøla fault veins were attributed to crystallization from a silica gel. During crystallization, shrinkage cracks due to volume loss
associated with expulsion of water during dehydration results in precipitation of chalcedony. The water then migrates into these cracks to precipitate cryptocrystalline bladed quartz which assume the shape of the crack. The blades serve as substrates on which euhedral vein quartz precipitates and grows as a result of dissolution and reprecipitation of the silica gel (Stel and Lankreyer, 1994).

Faber et al. (2014) documented a layer of microcrystalline quartz on the carbonate-hosted Olive fault in the Naukluft Nappe Complex, Namibia. The fault separates hanging wall dolostones from footwall calcareous shales. The fault zone is composed of a mélange of shales and dolostones from the host rocks while the core consists primarily of brecciated dolomites. A distinct, semi-continuous, fine-grained quartz layer with sharp boundaries occurs between the hanging wall dolomite breccia above and fault core breccia below. The layer is composed of fine-grained wispy bands along the margins grading into bands of hexagonal to subhedral quartz crystals increasing in grain size towards the center. The fine-grained margins appear isotropic in transmitted light and luminesce white in CL, while the hexagonal to subhedral grains luminesce violet-blue. The euhedral quartz crystals have very low dislocation densities and consist of spherical to ellipsoidal inclusions believed to be opal. The fault core breccia below the quartz layer consists of mostly dolomite clasts, few shale fragments and some microcrystalline quartz clasts cemented by calcite and dolomite. Presence of microcrystalline quartz clasts, which also luminesce violet-blue and occur only in the fault core breccia near the quartz layer provides evidence for cataclastic recycling of former fault-related quartz layers. The sharp planar nature of the quartz layer indicates that it formed during the last slip event. Microstructures of the quartz layer are consistent with those formed by evolution of a silica gel. Since the host rock is not silica-rich, it is unlikely that the silica in the quartz layer was sourced by comminution of local
wall rocks. While the source for the silica layer may be different from those in high velocity friction experiments, the rheological effects would be same once formed. Based on mineralogy and textural evidence, a preferred model in this case is that the silica gel most probably formed from episodic shear of a preexisting quartz-coated fault surface (Faber et al., 2014).

Polished fault surfaces exposed along a large, seismically active normal fault zone in Dixie Valley, Nevada were examined by Power and Tullis (1989). Although the fault zone is hosted by metamorphosed gabbro, the slickensides developed only in areas enriched in silica (i.e., fine-grained hydrothermal quartz in a 1-10 m thick zone of cataclasite). The slickenside material is 98% quartz characterized by extremely fine grain size, low dislocation density, irregular grain boundaries and preferred crystallographic orientation indicative of continuous non-brittle deformation. The thickest accumulation of fine-grained quartz occurs on the smoothest and most reflective slickensides. The cataclasite consists of angular fragments of primarily quartz and some gabbroic wallrock. The quartz is most abundant closer to fault surfaces due to passage of hydrothermal fluids. The angular fragments are surrounded by a matrix of fine-grained, randomly oriented quartz crystals that is texturally similar to the strongly oriented material that constitutes the slickenside surfaces. The angular fragments most likely derived from cataclasis of earlier fault surfaces with preferred orientation, thus indicative of alternating cataclasis (high strain rate) and non-brittle continuous (low strain rate) deformation mechanisms. The slickensides are believed to have initially developed from preexisting fractures or fault surfaces. According to textural evidence, the first phase involved cataclasis and dilation between the surfaces resulting in increased porosity and formation of angular fragments due to disruption of the preexisting fault surfaces. The second phase involved healing of the fault zone by precipitation of hydrothermal quartz with preferred crystallographic orientation. It is believed
that the preferred orientation of the fine-grained material presently on slickenside surfaces resulted from precipitation of originally amorphous silica, cristobalite or chalcedony since quartz tends to grow and dissolve fastest parallel to the c-axis. Low-temperature crystallization of quartz from a fluid phase or colloid near surface hydrothermal conditions typically results in very fine-grained phases. The wide range of strain rates in the seismic cycle may be a contributing factor in the formation of well-polished, glassy slickensides in many fault zones (Power and Tullis, 1989).

Uijie et al. (2007) documented a well exposed fault zone in the Shimanto accretionary complex, Japan, consisting of ultrataclasites derived from sedimentary rocks and basalt. The ultrataclasites layer shows evidence of injection of granular material into the gouge mixture without sorting of the finer grains. Presence of a silica layer along the ultrataclasite boundary may be due to fluid-related processes such as precipitation of a silica gel. The silica layer is composed primarily of anhedral aggregates of microcrystalline quartz (1-10 μm in size) and sub-parallel, optically isotropic cryptocrystalline silica zones. Homogenous nucleation of colloidal silica particles results in spherical particles. Supersaturation of these particles in the fluid precipitates viscous amorphous silica gel at temperatures consistent with those of the fault zone (i.e., 130-150°C). TEM analysis of a silica layer revealed colloidal spherical particles consistent with precipitation of silica gel. Crystallization from silica gel may yield sphereulites of either chalcedony or poorly crystalline cristobalite. Truncation of the silica layer by injection of granular material suggests that amorphous silica gel precipitated prior to fluidization of the granular material. Injection of the granular material in the ultrataclasite layer originated from fluidization during seismic slip (Uijie et al., 2007). They also concluded that amorphous silica gel may have precipitated from fluids infiltrating the fault and resulted in dynamic weakening
that triggered an earthquake rupture. The sudden motion during rupture may have induced fluidization and injection of granular material (Uijie et al., 2007).

Buff-colored quartz boulders, up to 0.5 m in diameter, found in the Lower Zambezi valley of northern Zimbabwe are believed to have formed within extensional tension gashes in strike-slip and high angle normal faults of biotite-hornblende gneiss of the Chewore Complex basement inliers and within its exposed contact with greywacke of the Kondo Pools Formation of the Karoo Supergroup (Sweetman and Tromp, 1991). Studies on these boulders by Sweetman and Tromp (1991) revealed textures not typical of quartz developed by growth from an aqueous fluid. The boulders consist entirely of aggregates of radiating crystalline quartz plates without an intervening matrix. They are characterized by surficial lines, which represent the intersection of the plates with the surface of the boulders. The lines alternate in color between buff and brown. Each plate consists of a central zone of small randomly oriented anhedral crystals less than 10 μm in size. Elongate crystals occur adjacent to these anhedral crystals with their c-axis orthogonal to the axis of the central zone. Abrupt terminations, crosscutting relationships and vugs occur where adjacent radiating aggregates overlap. Intersection of two plates from adjacent aggregates is marked by the contact between their orthogonal crystals along lines bisecting their central zones suggesting simultaneous growth of orthogonal crystals from both plates. The brown color results from distribution of amorphous hematite and is concentrated in the central zone, intersection of orthogonal quartz crystals, vugs and late-stage crosscutting veins. The geometry of the different phases in these boulders is not typical of slow precipitation from a fluid but rather, simultaneous growth from few nucleation sites. Such textures are believed to have originated from dehydration and contraction of a silica gel injected into an open cavity in the country rock. Fractures resulting from contraction provided pathways for a silica-saturated fluid
containing suspended hematite. The quartz plates occur along these fractures. Crystallization of central zones of randomly oriented quartz provided nucleation sites for subsequent recrystallization of the remaining gel resulting in orthogonal growth of quartz crystals from these zones. Hematite phases that could not be incorporated in the quartz crystals were continuously pushed out during growth and accumulated on surfaces of crystals, in vugs or at intersection of plates (Sweetman and Tromp, 1991).

The significance of silica gel lubrication as a dynamic weakening mechanism remains unknown due to scarcity of clear natural examples of in-situ gels forming along natural fault surfaces (Kirkpatrick et al., 2013). Since opaline phases may also precipitate from hydrothermal systems (Power and Tullis, 1989), the relationship to faulting and possible mechanisms through which the gel may form from a rock, need to be shown (Kirkpatrick et al., 2013) to assess the possibility of fault weakening. Additionally, silica gel is rarely preserved on faults due to recrystallization of the gel. Viscosity increases due to increasing polymerization and water is expelled from the phase resulting in crystallization. As a result, interpretation of the weakening process is usually based on textural evidence that indicate a silica gel parent fluid. If it can be proved that crystalline silica veins in previously studied faults originated from a silica gel fluid that formed during faulting, interpretations of fault weakening processes in these cases may have to be reconsidered. Understanding the significance of silica gel lubrication may provide new insight into faults that slipped seismically in the past (Faber et al., 2009).
3. STUDY AREA

The study area is located in the Valley and Ridge province in the central Appalachian foreland. The foreland consists of a package of folded and faulted Paleozoic sedimentary rocks transported northwestward during the Late Paleozoic Alleghanian orogeny along several major detachments. The primary focus of this study is the Lower Silurian Tuscarora Sandstone. It is composed of white, well-sorted, medium-grained, silica-cemented quartz arenite 30-100 m thick. The maximum burial depth of the Tuscarora Sandstone in the Central Appalachian foreland is up to 6-7 km (Onasch et al., 2009) while conodont-alteration color studies restrict temperatures experienced during deformation between 150-250⁰C (Epstein et al., 1977). Lithostatic pressures did not exceed 200 MPa. In addition to the Tuscarora, samples were collected from several under- and overlying units, including the Ordovician Juniata Formation and Upper Silurian Bloomsburg Formation.
4. METHODS

4.1. Field Methods

Field work involved sample collection from various fault surfaces of the Tuscarora Sandstone in Maryland, Virginia and West Virginia (Fig. 1). Sampling was done to investigate the effect of possible variables, including: (1) amount of displacement, (2) direction of displacement relative to regional folding, (3) orientation of fault surface relative to bedding and folds, (4) age of fault relative to folding, and (5) presence or absence of cataclasite. For each fault surface sampled, orientations of the fault, bedding, folds, and any other faults were measured, along with the morphology of the fault (e.g., planar, undulating, etc.).

Fig. 1. Geologic map of the Appalachian foreland showing distribution of sample locations. NMT - North Mountain thrust.

4.2. Laboratory Methods

Samples were cut normal to the fault surface and parallel to any striations. Polished thin sections of the fault surface and adjacent wallrock were prepared and examined using transmitted
light (TL) microscopy and cathodoluminescence microscopy (CL) for mineralogical, textural and temporal relationships. Fault surface morphologies were examined using a Hitachi S-4800 II Scanning Electron Microscopy (SEM) at the University of Toledo. Prior to carbon coating, samples were treated with dilute nitric acid (10% HNO$_3$) for 2-3 minutes to dissolve any organic matter on the surfaces. Dislocation densities and the degree of crystallinity were examined using a Hitachi HD 2300A Transmission Electron Microscope (TEM) at the University of Toledo. TEM samples were cut normal to the fault surface by focused ion beam milling using methods described by Heaney et al. (2001) to a thickness of <100 nm. The advantage of using focused ion beam milling instead of argon milling includes preservation of spatial relationships with respect to the fault surface.
5. **FAULT FEATURES**

5.1. *Outcrop and Handsample Description*

The fault surfaces, whether parallel to or at angles to bedding, most commonly occur in outcrop as either striated in form of grooves and ridges (Figs. 2b, 4a-b), or polished (Fig. 2a). The surfaces range from planar to curviplanar with the direction of slip in most cases indicated by presence of steps. No relationship was observed between the geometry and the direction of slip. Regardless of the nature of the fault surfaces, microcrystalline quartz bands similar to those described by Onasch et al. (2010) are visible in outcrop as white bands (Fig. 3a). They occur at various angles to the fault and at distances of up to a few m to the fault. The bands vary from planar to curviplanar and occur as either single bands or a network of several subparallel anastomosing bands. Despite varying abundances of the bands, they are most concentrated closer to fault surfaces. A sharp contact is visible between the bands and the wall rock (Fig. 4d). Additionally, many fault surfaces contain colorless euhedral quartz crystals up to 1 mm in size interlayered with the white bands and vugs (Fig. 4c). Open Mode I fractures can be seen crosscutting the white bands and adjacent wallrock (Figs. 4a and 4d). Stylolites are also visible in outcrop as dark gray seams (Fig. 3b).
Fig. 2. Polished (a) and striated (b) fault surfaces in the Tuscarora Sandstone in the North Mountain thrust zone as seen in outcrop.

Fig. 3. Fault surface features in the Tuscarora Sandstone as seen in outcrop. (a) Microcrystalline quartz bands occurring as white bands. Note the subparallel and anastomosing nature of the bands. (b) Stylolites occurring as dark gray seams. Note the truncation of the stylolite by microcrystalline quartz at the fault surface.
Fig. 4. Fault surface features in the Bloomsburg Formation as seen in outcrop. (a) White silica layer coating on the fault surface of the red siltstone. Mode I fracture is crosscutting both the silica layer and siltstone underneath. (b) Grooves and ridges on the surface of the silica layer. (c) Colorless euhedral quartz crystals and vugs in the silica layer. Note the interlayering of fine-grained white bands. (d) Sharp contact between the silica layer and the bleached wall rock. A quartz vein is crosscutting the silica layer. Note the different generations of Mode I fractures - an earlier fracture in which the vein precipitated and a late-stage, open fracture crosscutting both the vein and silica layer on the surface.

5.2. Transmitted Light Microscopy

Fault surfaces in thin section consist of microcrystalline quartz occurring as a fine-grained (5 – 10 μm in size) matrix directly on the surface and as bands adjacent to the matrix (Fig. 5a). The bands vary in thickness from 50 μm to a few cm and in length from 100 μm to a few cm. Under cross-polarized light, these bands occur as either single bands or multiple subparallel anastomosing bands (Fig. 5c) most abundant near the fault surface (Figs. 6a, 7, 9a, 11
and 12b). A sharp contact is visible between the microcrystalline quartz and the adjacent wallrock, regardless of the location and orientation of the band with respect to the fault surface (Figs. 5b, 6a, 7, 9a and 11). Some band walls display wall rock geometries consistent with wall-normal displacement only while others display both wall-normal and wall-parallel displacement indicating shear displacement through offset of grains along opposite walls (Figs. 5c and d). The remaining band walls lack any geometric fit with opposite walls indicating displacement along the third dimension. Commonly, the bands are parallel to well-developed fluid inclusion planes in the wall rock (Fig. 5b). Some of these bands contain larger grains amidst the fine-grained matrix (Fig. 5b). The edges of these larger grains show irregular overgrowths marked by lower interference colors than the rest of the grain and a gradational contact with the finer grained material. Under plane-polarized light, the microcrystalline quartz bands show a lower inclusion density than the adjacent wall rock (Figs. 10c and e); hence, they appear clearer.

Other common microstructures along fault surfaces include microfractures, microveins (Figs. 10a and c) and fluid inclusion planes (Fig. 5b). The microfractures have a Mode I origin and are most abundant closest to fault surfaces. They form either very shallow angles or very steep angles to the fault surface and form conjugate sets typical of Riedel shears. Similar to the microcrystalline quartz bands, the microveins are also most abundant closest to fault surfaces and have a lower inclusion density than the adjacent wall rock under plane polarized light (Figs. 10a and c).
Fig. 5. Microstructures associated with microcrystalline quartz bands in the Tuscarora Sandstone. (a) Texture of the microcrystalline quartz band. (b) Sharp contact of the microcrystalline quartz band with the wall rock. Band is parallel to fluid inclusion planes (dark lines in the wallrock). (c) Microfractures and wall-parallel displacement across the band. Note the anastomosing nature of the bands. Wall-normal displacement is visible across some bands. (d) Open Mode I fractures crosscutting the bands and the adjacent wallrock.

Fig. 6. Microstructures associated with microcrystalline quartz bands in the Juniata formation. (a) Microcrystalline quartz bands concentrated on the fault surface. Note the sharp contact of the bands with the wall rock and microcrystalline quartz overgrowths nucleating on the vein walls. (b) Close-up of nucleating overgrowths in (a).
Complex textural relationships exist between the microcrystalline quartz bands, microfractures, microveins and stylolites near fault surfaces (Figs. 7, 8a-b, 9a-b, 10a-d).

Relatively unaltered microcrystalline quartz bands and microveins can be seen crosscutting darker (i.e., more inclusions) bands and microveins. Microfractures can be seen crosscutting microcrystalline quartz bands and microveins (Figs. 9a-b, 10e-f, 11). Although microcrystalline quartz is most concentrated at the fault surface, several bands and microveins within the adjacent wall rock occur at varying angles to the fault surface (Fig. 7), similar to the orientation of microfractures. The bands and microveins are linked to the microcrystalline quartz matrix at the
fault surface indicating continuity of the bands with the matrix. Mutually crosscutting relations occur between some of these bands, microveins and microfractures. Similarly, mutually crosscutting relations also exist between brittle microstructures and those associated with fault creep. Some stylolites offset by brittle microstructures (Figs. 8a-b, 10c-d) are visible near the fault surface alongside other stylolites that appear to crosscut the brittle microstructures (Figs. 10c and d).
Fig. 9. (a) Photomicrograph mosaic and (b) sketch showing spatial and temporal relationships of brittle microstructures with respect to fault surfaces in the Tuscarora Sandstone. Microcrystalline quartz matrix is most concentrated at the fault surface. Complex mutually crosscutting relations can be seen between the bands and healed microfractures. Different generations of microcrystalline quartz are determined from crosscutting relations between microstructures and the degree of alteration.
Fig. 10. Crosscutting relationships between different microstructures in the Tuscarora Sandstone. (a) Microvein crosscutting a preexisting microvein. Note the significantly less alteration of the microveins. (b) Same as (a) as in cross-polarized light. (c) Microcrystalline quartz band bounded by stylolites. Microveins crosscutting the band and stylolites. Note the significantly less alteration of the bands and microveins. (d) Same as (c) as in cross-polarized light. (e) A fracture crosscutting a microcrystalline quartz band with significantly less alteration than the adjacent wallrock. (f) Same as (e) as in cross-polarized light.
Fig. 11. Photomicrograph mosaic showing a microcrystalline quartz band (double-sided white arrow) along a fault surface in the Tuscarora Sandstone. Note the sharp contact of the band with the wall rock and decreasing abundance of microcrystalline quartz away from the surface. An open fracture can be seen crosscutting the band. White dashed line highlights exposed slip surface.

Microcrystalline quartz is more abundant along polished fault surfaces than striated surfaces (Figs. 12a and b). The extent of brecciation is also much higher along polished surfaces
as noted from the higher density of angular clasts and several generations of bands, microveins and fractures.

![Fig. 12. Extent of deformation on (a) striated and (b) polished fault surfaces in the Tuscarora Sandstone. Note the significantly higher density of brecciated fragments and abundance of microcrystalline quartz near the polished fault surface compared to striated surface.](image)

5.3. **Cathodoluminescence Microscopy**

CL is an invaluable tool for investigating brittly deformed rocks as it clearly delineates generations of fractures and fill materials (Marshall, 1988; Onasch et al., 2010). The difference in luminescence between the quartz of the wall rock and that of the filling reveals the boundary between the two regions in details that cannot be discerned from transmitted light microscopy (Anders et al., 2014) due to growth of the fill in optical continuity with the host rock grains. CL
imaging of many samples reveals a sharp boundary between the microcrystalline quartz bands, which luminesce dark red, and the wall rock which luminesces a brighter red. The difference in luminescence arises from differences in trace element chemistry or defect structures (Marshall, 1988). Larger grains in the microcrystalline quartz bands that resemble wall rock fragments show the same dull luminescence as the fine-grained regions within the bands.

Several generations of microcrystalline quartz bands, microveins and microfractures are visible near the fault surface in CL. Crosscutting relations between microcrystalline quartz bands of different generations can be seen by differences in luminescence (Fig. 13a). Abrupt terminations of microveins and microcrystalline quartz occur within breccia clasts as cement for earlier formed breccia (Figs. 13a, 14a, 15a and 16a). Late-stage, open fractures crosscut microcrystalline quartz and microveins (Fig. 15a). Crosscutting relationships also occur between microveins that luminesce differently from each other (Figs. 14c and 15c).
Fig. 13. Matched (a) cathodoluminescence (CL) and (b) polarized light micrograph mosaics of a polished fault surface in the Tuscarora Sandstone. Yellow dashed polygon delineates a breccia fragment consisting of smaller breccia fragments. Note the different luminescence and abrupt terminations of microveins in the breccia. Microcrystalline quartz bands near the fault surface luminesce differently than the adjacent matrix.
Fig. 14. (a-b and c-d) Matched cathodoluminescence and polarized light photomicrographs of microstructures near a fault surface in the Tuscarora Sandstone. (a-b) White arrow indicates offset microveins and microveins with abrupt terminations within breccia clasts of different luminescence. (c-d) White arrows show crosscutting relations between two microveins of different luminescence.
Fig. 15. (a-b and c-d) Matched cathodoluminescence and polarized light photomicrographs of microstructures near a fault surface in the Tuscarora Sandstone. (a-b) A late-stage open fracture (white arrow) crosscutting a microvein (black arrow), which crosscuts a microcrystalline quartz band (between yellow dashed lines). (c-d) White arrows indicate at least three different generations of microveins delineated by different luminescence and crosscutting relations between at least two.
Fig. 16. Matched (a) cathodoluminescence and (b) polarized light photomicrographs of microstructures near a fault surface in the Tuscarora Sandstone. Note the differences in luminescence of fill materials. White arrow indicates interruption of the microvein by a breccia clast.

5.4. **Scanning Electron Microscopy**

While smooth in hand sample, most fault surfaces were found to be microscopically rough consisting of abundant euhedral quartz crystals up to 1 mm in length on the surface and in vugs (Figs. 17a and b).

Fig. 17. SEM micrographs of fault surfaces in the Tuscarora Sandstone. (a) Striated fault surface consisting of euhedral quartz crystals. (b) Vug on the fault surface containing euhedral quartz crystals.
The more highly polished fault surfaces are much smoother on scales of μm to nm. Just below the smooth surface, as can be seen in vugs and irregularities, are spherical to ellipsoidal particles that range in diameter from <50 nm up to >1 μm (Figs. 18a-f). They occur either individually in vugs or as coalesced aggregates. Energy dispersive spectroscopy (EDS) shows that the spherical particles have high Si and O content (Figs. 19a-c) and are therefore assumed to be silica. Uniform grain size distribution of the spheres is apparent (Fig. 20).

Fig. 18. SEM micrographs of a polished slickenside in the Tuscarora Sandstone. (a) Smoother regions (left side of micrograph) represent polished surfaces in handsample. (b) Vugs on the slickenside. (c) Close-up of (b) showing spherical particles inside a vug. (d) Close-up of (c) showing smaller spherical particles comprising the smoother regions around the edges of the vug. (e) and (f) show more spheres on different locations on polished surfaces surrounding the vugs.
Fig. 19. Elemental composition of polished surfaces. (a) EDS spectra of a portion of the fault surface. Note the high Si and O peaks. (b) Field of view of the fault surface for which EDS spectra in (a) was collected. (c) Element concentrations are in atomic% and weight%.
5.5. Transmission Electron Microscopy

Exploring the fault rocks at a nanoscale with TEM was used to: (i) determine grain size in ultrafine-grained surfaces, (ii) detect and characterize the degree of crystallinity of phases in the fault rock, and (iii) identify deformation-induced features (e.g., Viti, 2011). Because amorphous silica is the key to documenting the presence of a silica gel, SAED (selected-area electron diffraction) patterns were used to distinguish amorphous from poorly or highly crystalline material (Viti, 2011). Unlike crystalline phases, which produce sharp and distinct SAED diffraction patterns, amorphous phases do not produce diffraction patterns (Viti, 2011; Kirkpatrick et al., 2013). TEM images also reveal textural relations at the nanoscale not possible from other forms of microscopy.
Contrast in TEM images is a function of the dose of transmitted unscattered electrons such that brighter regions correspond to higher doses (Viti, 2011). Regions that appear dark are either: (1) thicker, (2) higher in density, and/or (3) heavier in atomic weight. A lighter featureless region is visible at the fault surface (Fig. 21). SAED patterns of this region produce diffuse rings characteristic of amorphous phases whereas sharp, intense diffraction spots on SAED patterns away from the featureless region indicate high crystallinity and structural order (Figs. 22a-c).
Distinct flow features are clearly visible at the fault surface (Fig. 22). Some grains with fractured surfaces (Fig. 23a) and/or high dislocation density can be seen (Fig. 23b). Numerous recrystallized grains are characterized by euhedral grain boundaries and/or very low dislocation density (Fig. 23a). Hydrous silica phases occur as cement around clasts (Fig. 23a). Abundant pore spaces are visible between grains. Sub-angular 100-200 nm pores occur at triple junctions of recrystallized grains while elongate 10-30 nm pores occur along grain boundaries. Spherical to ellipsoidal inclusions occur as trapped phases in the surrounding matrix and pore spaces (Fig. 24).
Fig. 23. TEM micrographs showing different textures along the fault surface. (P = pore spaces, X = crystalline silica, D = dislocations). (a) Hydrous silica phases cementing fractured clasts and evidence for recrystallization. (b) Dislocations in some grains.

Fig. 24. TEM micrograph showing a trapped ellipsoidal inclusion in a pore space (P).
6. DISCUSSION

Fault surfaces of the Tuscarora Sandstone and other lithologies consist of high concentrations of microcrystalline quartz that decrease in abundance away from the surface. High concentrations of microfractures, microcrystalline quartz bands and microveins at the fault surface suggest that their formation is spatially related. Mutually crosscutting relationships exist between brittle and ductile microstructures near the fault surface indicating a complex deformational history surrounding these faults. The microcrystalline quartz bands are parallel to well-developed fluid inclusion planes in the wall rock suggesting that their formation may have been the latest deformational event. A study by Onasch et al. (2010) documented similar veins and breccia fill that they believe formed by precipitation of a silica gel and not cataclasis, owing to a low dislocation density. In light of their findings, occurrences of microcrystalline quartz along fault surfaces and as cement in cataclasites of the Tuscarora Sandstone raised the possibility that a silica gel may have been present during faulting and significantly reduced strength. In this study, several lines of evidence suggest a gel origin in the development of microcrystalline quartz on the fault surfaces.

6.1. Evidence for a gel origin in the formation of microcrystalline quartz on the fault surface

Several lines of evidence suggest presence of a silica gel parent fluid for the microcrystalline quartz on and adjacent to the fault surfaces:

1. Opaline phases
2. Polygonal network of grains with a uniform size distribution
3. Vugs with euhedral quartz crystals
4. Flow features
5. Amorphous silica

6. Hydrous silica nanofilms surrounding clasts

   Evidence suggests that an amorphous, viscous, supersaturated silica gel precipitated rapidly into opal nanospheres or hydrous silica nanofilms around clasts. Upon silica-depletion of the remaining fluid, it was expelled from the precipitating phases. Dehydration of these phases resulted in shrinkage cracks into which the remaining fluid migrated and precipitated euhedral quartz, leaving behind vugs.

6.2. A model for the formation of microstructures at the fault surface

   Complex mutually crosscutting relationships exist between different brittle microstructures near the fault surface indicating multiple deformational events. Textural evidence suggests that the following sequence of events were involved in the formation of brittle microstructures during a single faulting event:

1. Open Mode I fractures are generated along the fault surface and accompanied by formation of Riedel shears resulting in brecciation and an increase in porosity.

2. Silica gel is frictionally generated along the fault surface via comminution of quartz at asperities and subsequent hydration.

3. The viscous gel lowers strength of fault zone thereby facilitating slip.

4. Formation of additional fractures results in pressure drop, which decreases quartz solubility further increasing degree of saturation/supersaturation.

5. With displacement, the gel flows from the fault surface into open fractures and Riedel shears and precipitates rapidly as opaline phases to form microcrystalline quartz bands and cement between breccia fragments.
6. With silica-depletion of the gel, quartz precipitates in open fractures to form microveins and cement between remaining breccia fragments.

As silica gel is an unstable fluid, it cannot survive transportation from distal sources over a long period of time in its suspended state. Since opaline phases precipitate rapidly, the source (i.e., site of generation of the gel) has to be close to the site of precipitation, if not the same. Although direct evidence that the gel originated at the fault surface has yet to be found, other evidence suggest a gel origin for the microcrystalline quartz. Close proximity and high concentrations of microcrystalline quartz along the fault surface and in bands adjacent to the fault surface suggest that its formation is associated with faulting and therefore most likely derived at that surface. Several of the fine-grained bands appear to be in contact with the fine-grained matrix at the surface suggesting a link/path between the microcrystalline quartz at the fault surface and the bands, possibly the same path that allowed flow of the gel from the surface to open fractures.

A greater abundance of brecciated clasts and microcrystalline quartz along polished surfaces indicates a greater extent of deformation and possibly greater displacement than striated surfaces. In general, faults that consist of microcrystalline quartz along the entire surface demonstrate more brecciation than faults with sparse, discontinuous/discrete portions of microcrystalline quartz suggesting that the rate of comminution was greater than the rate of aggregation in the former. Higher rates of comminution may indicate more rapid slip rates. Presence of microcrystalline quartz covering larger areas of the fault surfaces may suggest greater displacement that allowed distribution of the gel along the entire surface than faults with only discrete areas of microcrystalline quartz.
As mentioned earlier, complex mutually crosscutting relationships between brittle microstructures at the fault surface indicate that their formation was episodic (i.e., multiple slip events). Earlier generations of microcrystalline quartz bands, microfractures and microveins crosscut and offset later generations of the same features. Microcrystalline quartz bands interpreted as filling in open fractures formed during faulting also consist of other generations of open fractures. Breccia cemented by microcrystalline quartz and microveins occur within single breccia fragments. This could not have been derived from the same faulting event as each event experiences a single episode of pressure drop associated with formation of Mode I fractures. Similarly, bands and microveins crosscutting other bands and microveins could not result from the same deformational event as the fractures in which they precipitate are all generated at the same time in a faulting event. More than one episode of brittle deformation is required to produce crosscutting relations between similar features.

The crosscutting relations between brittle microstructures and stylolites indicate that periods of brittle faulting alternated with ductile deformation by pressure solution (Fig. 25). These temporal changes are believed to form from alternations of slip followed by periods of fault creep indicating a wide range of strain rates within the fault zone.
Fig. 25. Illustrations showing different mutually crosscutting relations between microstructures near the fault surface. (a) A network of open Mode I fractures form. (b) Microcrystalline quartz precipitates first in the fractures. (c) Microveins precipitate later in the some of the remaining fractures. (d) Fault creep microstructures (stylolites) form and offset earlier brittle microstructures. (e) New generation of open Mode I fractures form and crosscut all of the earlier microstructures. (f) Microcrystalline quartz bands and microveins precipitate in the latest generation of open fractures.
7. CONCLUSIONS

Evidence suggests that for the faults studied, microcrystalline quartz along the fault surfaces and in bands adjacent to the faults originated from a silica gel. Its abundance along fault surfaces and as cement between breccia suggests considerable dilation and weakening in fault zones. The presence of a gel has been shown experimentally to reduce the shear strength of a fault zone to nearly zero. Given that silica-rich rocks are common in the Earth’s crust, this mechanism of strength reduction may be more widespread than initially thought.

We suggest that open Mode I fractures created porosity for a silica gel that formed by comminution and hydration of quartz-rich asperities. The pressure drop that resulted in displacement on a non-planar fault surface further increased the state of supersaturation of the silica gel. With displacement, the viscous gel flowed into fractures in the wallrock and rapidly precipitated opal. Upon silica-depletion of the gel and subsequent increase in water content, progressively slow precipitation of quartz from an aqueous fluid produced microveins. The brittle events were episodic and alternated with fault creep processes to form stylolites, hence the mutually overprinting textures.

Presence of a silica gel in a fault surface is important in understanding fault history and assessing the state of stress of the fault. The thixotropic nature of the gel provides the potential for this mechanism to not only significantly reduce the strength of the rocks but also to promote seismic slip. Since crystallization products of silica gel fill microcracks, the fault zones associated with these mechanisms are completely healed and strength recovery is rapid (Stel and Lankreijer, 1994). Strength recovery and fault rock restoration provides an environment for
reactcumulation of stress which may eventually lead to another phase of brittle failure (i.e.,
periodicity of seismic events) (Stel, 1981; Stel and Lankreyer, 1994).
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APPENDIX A: (sample locations and descriptions)

FS-1: Tuscarora Sandstone (Latitude: 39.6848550539934; Longitude: -77.9655066364579)

   Bedding: 024/64 SE
   Fault: 314/82 NE
   Rake of striations: 18° NW
   Notes: Left lateral, hanging rock.

FS-2: Tuscarora Sandstone (Latitude: 39.6848550539934; Longitude: -77.9655066364579)

   Bedding: 005/60 SE
   Fault: 015/74 NW
   Notes: Right lateral? Faint striations, 50 m north of FS-1.

FS-3: Rosehill Shale

   Bedding: 030/33 SE
   Fault: 090/81 NE
   Notes: Normal with some oblique slip, overturned beds, sandstone interbedded with shale, trace fossils, 1/10th mile west of intersection near FS-1.

FS-4: Juniata Formation (Latitude: 39.65416666666667; Longitude: -78.78916666666666)

   Bedding: 045/33 SE
   Fault: 312/78 NE
   Notes: Oblique strike-slip.

FS-5: Tuscarora Sandstone (Latitude: 39.609200; Longitude: -77.967972)

   Bedding: 050/75 SE

FS-6: Tuscarora Sandstone (Latitude: 39.609200; Longitude: -77.967972)

   Fault: 020/70 NE

FS-7: Mahantango Formation (Latitude: 39.609200; Longitude: -77.967972)

   Bedding: 013/63 SE
   Fault: 280/20 NE
   Notes: Low-dipping reverse fault (thrust fault), sample collected from footwall, minor faults, internal shearing.

FS-8: Tuscarora Sandstone (Latitude: 39.609200; Longitude: -77.967972)

   Bedding: 354/68 SW
Notes: Polished fault surface, very steep, almost vertical.

FS-9: Oriskany Sandstone (Latitude: 39.0652608429626; Longitude: -78.6613497106376)

Bedding: 050/47 SE
Fault: 060/60 SE
Notes: Multiple fault orientations, several silica-cemented casts, corner of Sauerkraut Rd and Olde 55.

FS-10a: Tuscarora Sandstone (Latitude: 39.0426968802766; Longitude: -78.7124962345295)

Bedding: 030/27 SE
Fault: 330/80 N
Rake: 5° N
Notes: Almost pure strike-slip on some surfaces and dip-slip on others, SE limb of Baker Anticline on Olde 55.

FS-10b: Tuscarora Sandstone (Latitude: 39.0426968802766; Longitude: -78.7124962345295)

Bedding: 324/58 NE
Notes: Many fault surfaces, subhorizontal striations.

FS-11: Tuscarora Sandstone (Latitude: 38.9958175752131; Longitude: -79.1928228739072)

Bedding: 052/25 NW
Fault1: 320/69 W
Rake1: 22° NW
Fault2: 280/76 SW
Rake2: 30° SW
Notes: Left lateral.

FS-12a: Tuscarora Sandstone (Latitude: 39.0682192490007; Longitude: -79.2082425125981)

Bedding: 018/32 SE
Fault: 024/50 W
Notes: Klines Gap.

FS-12b: Tuscarora Sandstone (Latitude: 39.1283378112116; Longitude: -79.1922803010478)

Bedding: 018/32 SE
Fault: 020/60 W
Notes: Fault is approximately 30° to the bedding. Joints perpendicular to fold axis. Cross fold joints that are very closely spaced. SE limb of the anticline in Maysville Gap. Minor folding.
FS-13: Tuscarora Sandstone

  Bedding: 028/66 NW
  Fault: 015/80 NW
  Notes: Striations directly down dip.

FS-14a: Tuscarora Sandstone (Latitude: 39.4280508233861; Longitude: -78.9586004626143)

  Notes: Curved fault surface, striations down dip. Limestone Road, Keyser. SE limb.

FS-14b: Tuscarora Sandstone (Latitude: 39.4280508233861; Longitude: -78.9586004626143)

  Fault: 036/46 NW
  Notes: SE fold.

FS-14c: Tuscarora Sandstone (Latitude: 39.4280508233861; Longitude: -78.9586004626143)

FS-14d: Tuscarora Sandstone (Latitude: 39.4280508233861; Longitude: -78.9586004626143)

  Notes: NW fold.

FS-15: ? (Latitude: 38.9455677951402; Longitude: -78.3049558750168)

  Bedding: 325/16 NE
  Fault: 020/35 W
  Notes: Striations directly down dip, white coating on slightly curved fault surface. Reverse fault.

FS-16: ? (Latitude: 38.8927850875007; Longitude: -78.3709009801028)

  Bedding: 280/30 SW
  Fault: 280/30 SW
  Notes: Fault surface parallel to bedding. Tight fold. Striations down dip. Intersection of Boyer Road and Fort Valley Road.

FS-17: Bloomsburg Formation (Latitude: 38.784048411593; Longitude: -78.508815312166)

  Bedding: 060/58 W
  Fault: 000/76 W
  Notes: Red siltstone, white coating on fault surface.

FS-18: Tuscarora Sandstone (Latitude: 38.784048411593; Longitude: -78.508815312166)

  Bedding: 045/79 W
  Fault: 005/89 W
  Notes: White coating on fault surface.
APPENDIX B:

Fig. 1B. SEM micrograph of the Tuscarora Sandstone.

Fig. 2B. SEM micrograph of the Tuscarora Sandstone.
Fig. 3B. SEM micrograph of the Tuscarora Sandstone.

Fig. 4B. SEM micrograph of the Tuscarora Sandstone.
Fig. 5B. SEM micrograph of the Tuscarora Sandstone.

Fig. 6B. SEM micrograph of the Tuscarora Sandstone.
Fig. 7B. SEM micrograph of the Tuscarora Sandstone.

Fig. 8B. SEM micrograph of the Tuscarora Sandstone.
Fig. 9B. SEM micrograph of the Tuscarora Sandstone.

Fig. 10B. SEM micrograph of the Tuscarora Sandstone.
Fig. 11B. SEM micrograph of the Tuscarora Sandstone.

Fig. 12B. SEM micrograph of the Bloomsburg Formation.
Fig. 13B. SEM micrograph of the Tuscarora Sandstone.

Fig. 14B. SEM micrograph of the Tuscarora Sandstone.
Fig. 15B. SEM micrograph of the Tuscarora Sandstone.

Fig. 16B. SEM micrograph of the Tuscarora Sandstone.
Fig. 17B. SEM micrograph of the Tuscarora Sandstone.

Fig. 18B. SEM micrograph of the Tuscarora Sandstone.
Fig. 19B. SEM micrograph of the Tuscarora Sandstone.

Fig. 20B. SEM micrograph of the Tuscarora Sandstone.
Fig. 21B. SEM micrograph of the Tuscarora Sandstone.

Fig. 22B. SEM micrograph of the Tuscarora Sandstone.
Fig. 23B. SEM micrograph of the Tuscarora Sandstone.
Fig. 24B. TEM micrograph of the Tuscarora Sandstone.

Fig. 25B. TEM micrograph of the Tuscarora Sandstone.
Fig. 26B. TEM micrograph of the Tuscarora Sandstone.

Fig. 27B. TEM micrograph of the Tuscarora Sandstone.
Fig. 28B. TEM micrograph of the Tuscarora Sandstone.

Fig. 29B. TEM micrograph of the Tuscarora Sandstone.
Fig. 30B. TEM micrograph of the Tuscarora Sandstone.

Fig. 31B. TEM micrograph of the Tuscarora Sandstone.
Fig. 32B. TEM micrograph of the Tuscarora Sandstone.

Fig. 33B. Z-contrast image of TEM micrograph in Fig. 32B.
Fig. 34B. TEM micrograph of the Tuscarora Sandstone.

Fig. 35B. Z-contrast image of TEM micrograph in Fig. 34B.