Folded felsic dike and host calc-silicate from the Black Lake Shear Zone exposed on Interstate 81 south of Alexandria Bay, NY
GEOLOGY OF THE BLACK LAKE SHEAR ZONE AND
NORTHWESTERN ADIRONDAK LOWLANDS,
GRENVILLE PROVINCE, NEW YORK

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This guidebook is for the most part derived from our published research with co-workers on the NW Adirondack Lowlands (Chiarenzelli et al., 2010b, Wong et al., 2011), and our in-prep manuscripts on the igneous and metamorphic petrology of this region.
INTRODUCTION

Identifying terrane boundaries within orogenic belts is fundamental to understanding the evolution of orogenic systems because identifying these areas allows the tectonic processes involved in the assembly of mountain belts as well as the rates of growth of these systems to be assessed. However, identifying terrane boundaries between distinct crustal blocks has historically been problematic in the Grenville Province (and gneiss terranes in general) because of the lack of coherent structural markers across the shear zones that typically mark terrane boundaries, poly-phase deformation, and the large number of deformation zones to choose from as possible boundaries. Mesoproterozoic rocks of the Grenville in North America have traditionally been subdivided into a number of terranes based on broadly different tectonic histories and lithostratigraphic assemblages (Moore; 1982, Easton, 1992; Davidson 2008), and where boundaries between terranes are recognized these boundaries are in most cases ductile shear zones. A complicating factor to identifying terrane boundaries in the Adirondack Grenville Province is the recognition of orogen-scale collapse structures (e.g. Selleck et al., 2005), which raises the question as to which ‘terrane boundaries’ are tectonic assembly boundaries between exotic blocks, and which are detachment/exhumation boundaries that juxtapose different crustal levels. This Friends of the Grenville field conference focuses on new research on the Black Lake Shear Zone, which we propose to be the terrane boundary between quartzite–pelite–marble package of the Adirondack Lowlands (NY) and the marble–pelite–metavolcanic package of the Ontario Frontenac terrane, which is also present in New York along the shore of St. Lawrence river.

SOUTHWESTERN GRENVILLE PROVINCE

The Adirondack Lowlands and Frontenac terrane are part of the Allochthonous Monocyclic Belt (AMB) of Rivers et al. (1989), which comprises juvenile Mesoproterozoic rocks accreted (or formed) on the margin of Laurentia. Parts of the AMB contain thick sequences of supracrustal metasedimentary and metavolcanic rocks as well as several ca. 1.4-1.0 Ga plutonic suites. In the AMB, terranes are deformed and metamorphosed by some (or all) of a number of mountain-building events: the Elzevirian
orogeny (ca. 1.29–1.19 Ga), Shawinigan orogeny (ca. 1.19–1.14 Ga), Ottawan orogeny (ca. 1.08–1.02 Ga), and Rigolet orogeny (ca. 1.0–0.98 Ga; see Davidson 2008). The Shawinigan orogeny is the principal tectonothermal event recognized in the Adironack Lowlands and Frontenac terranes (e.g. Mezger et al., 1993). Terranes in the AMB are alternately interpreted as representing a collage of arc and continental fragments accreted to Laurentia ca. 1.2 Ga (e.g. Carr et al., 2000) and as a 1.4-1.2 Ga Andean margin and associated back-arc environments (e.g. Hanmer et al., 2000).

LITHOLOGIC EVIDENCE FOR THE BOUNDARY BETWEEN THE FRONTENAC TERRANE AND ADIRONDACK LOWLANDS

Several workers have noted that rocks typical of the Adirondack Lowlands disappear and are replaced by those typical of the Frontenac terrane close to the St. Lawrence river. This transition has generally been described as occurring at the Black Creek Fault (Carl et al., 1990, Grant, 1993; Carl and deLorraine, 1997; Wasteneys et al., 1999), which offsets the lower Paleozoic Potsdam Formation and is characterized in outcrop by low-temperature mineralized fault breccias. The Black Creek Fault is clearly a Paleozoic brittle fault and unrelated to Mesoproterozoic tectonics, but marks the transition from Adirondack Lowlands to Frontenac lithologies in the northern portion of the field area. Davidson (1995) used the term Black Lake Lineament to demark the boundary between the Frontenac terrane and the Adirondack Lowlands. This was referred to as the Black Lake Fault by Peck et al. (2004), and McLelland et al. (2010) used the term Black Creek Shear Zone in reference to a steeply dipping low-grade shear zone that parallels Black Lake. For clarity, we use Black Creek Fault to refer to the Paleozoic brittle structure in the vicinity of Black Creek and define the Black Lake Shear Zone (BLSZ) as the boundary between the Frontenac and Lowlands terranes (Wong et al., 2011).
Figure 1. Geology of the Adirondack Lowlands, compiled by Joseph Catalano and Kurt Hollocher. Note that the distribution of the Antwerp-Rossie suite, Rockport granite, and Hyde School gneiss bodies appear to be controlled by the Black Lake shear zone. E= Edwardsville Syenite.
The Black Lake Shear Zone is a 5-10 km wide northeast-trending zone of ductile deformation that extends from the vicinity of southern Wellesley Island, NY in the southwest to the northeast end of Black Lake, where it obscured by Paleozoic sedimentary rocks (Fig. 1). In the area of Black Lake the Black Lake Shear Zone is largely coincident with the Black Creek Fault.

The Frontenac terrane and Adirondack Lowlands of New York have broadly similar packages of meta-igneous and metasedimentary rocks and metamorphic timing, which have led some regional tectonic syntheses to focus on a shared geologic history (e.g. Rivers, 1997; Carr et al., 2000; Peck et al., 2004). However, differences in rock types and metamorphic histories distinguish the Frontenac terrane and Adirondack Lowlands, with the Black Lake Shear Zone representing the boundary between these two distinct terranes. In this section of the guidebook we compare and contrast the geologic character of the two terranes, which is followed by separate sections on the igneous petrology and metamorphic petrology of the region.

The amphibolite-facies Adirondack Lowlands are bounded to the southeast by the Carthage-Colton mylonite zone (Geraghty et al., 1981; Mezger et al., 1992; Selleck et al., 2005), which separates it from the Adirondack Highlands. The Lowlands are dominated by metasedimentary units that include metapelites with minor associated intrusive and volcanic rocks (‘Major Paragneiss’), calcitic and dolomitic marbles, siliceous metacarbonates, evaporates and metasediment-hosted sulfide ore deposits (Carl et al., 1990), as well as minor distal arc volcanic rocks (‘Popple Hill’ Gneiss). Quartzite is only a minor component of the Lowlands metasedimentary package. The Lowlands terrane was intruded by two metamorphosed plutonic suites that are not present in the Frontenac terrane: the bimodal ~1.20 Ga calc-alkaline Antwerp-Rossie suite (Wasteneys et al., 1999; Chiarenzelli et al., 2010b) and the ~1.18 Ga (Heumann et al., 2006) alkali- to calc-alkaline Hermon granite (Carl and deLorraine, 1997). The Lowlands also contain the distinctive ~1.17 Ga domical Hyde School leucogranite to tonalitic gneiss bodies (McLelland et al., 1991; Wasteneys et al., 1999). The last major plutonic event recorded in the Lowlands includes the ~1.16 Ga Edwardsville syenite (McLelland et al., 1993) and several correlative syenitic bodies (Buddington, 1934). These plutonic suites, which
are all penetratively deformed to some degree, were followed by younger scattered, undeformed red granites (Buddington, 1934) that are undated.

The granulite-facies Frontenac terrane is separated from the Elzevir terrane to the northwest by the Maberly shear zone (Corfu and Easton, 1997; Davidson and van Breemen, 2000). Rocks correlative to the Frontenac terrane in Ontario are recognized near the Saint Lawrence River west of the BLSZ and in islands within the river (e.g. Wellesley and Grindstone Islands). Metasedimentary packages in the terrane contain quartzite, pelitic rocks, marble, and siliceous metacarbonate but appear to lack metavolcanic rocks. Plutonic rocks in the terrane are dominated by monzonite, syenite, granite, and gabbro plutons of the A-type 1.18-1.15 Ga Frontenac suite (Marcantonio et al., 1990; Davidson and van Breemen, 2000). The Frontenac suite has unusually high magmatic oxygen isotope ratios (Peck et al., 2004), which seem to be bounded on the western side of the Frontenac terrane where this suite is deformed by the Maberly shear zone and in the east by the Edwardsville syenite, which is located adjacent to the BLSZ (Fig. 2). The heterogeneously deformed, ~1.17 Ga Rockport granite (van Breemen and Davidson, 1988; Wasteneys et al., 1999), which is found primarily near the St. Lawrence
River, is contemporaneous with the Frontenac suite, but has a somewhat distinct geochemistry (see below). Also contemporaneous with the Frontenac suite are ~1.16 Ga olivine diabase dikes of the Kingston swarm in the eastern Frontenac terrane, which are undeformed (Davidson, 1995).

These geological histories document a number of important lithologic and geochemical differences between the Frontenac and Adirondack Lowlands terranes that we believe justify their separation into distinct terranes. The terranes have distinct metasedimentary packages: the Frontenac terrane is dominated by a quartzite–pelite–marble package while the Lowlands assemblage is dominated by marble–pelite–metavolcanic rock with minor quartzite (Davidson, 1995). The carbon isotope ratios of marble in the terranes are statistically distinct by t-test, suggesting different depositional environments of carbonate protoliths (Kitchen and Valley, 1995; Tortorello and Peck, 2010). The two terranes also have some unique igneous suites: the Kingston dike swarm is only present in the Frontenac terrane and the Antwerp-Rossie and Hermon granite suites are only found in the Lowlands. Contemporaneous igneous suites, such as the ca. 1.17 Ga Rockport Granite and Hyde School Gneiss bodies, have distinct emplacement styles between the Frontenac and Lowlands terranes, with the former typically expressed as large plutonic bodies or sheet-like dike intrusions and the later as isolated domical intrusions. In addition, peak metamorphic conditions drop across the BLSZ from granulite facies conditions in the Frontenac terrane to amphibolite facies conditions in the Adirondack Lowlands (>50°C difference, Tortorello and Peck, 2010), with differences in details of timing and cooling history (see below). Taken together, these data suggest that the Frontenac terrane and Adirondack Lowlands had similar yet distinct histories prior to ca. 1.17 Ga. Given that ca. 1.17 Ga plutons are present in both the Frontenac terrane and Adirondack Lowlands (Rockport Granite and Hyde School Gneiss, respectively), it seems likely that the two terranes were assembled by that time.
IGNEOUS PETROLOGY

1203 Ma Antwerp-Rossie Suite

This bimodal suite of deformed granitic and granodioritic plutonic rocks has the oldest dated rocks of the Adirondack Lowlands and intrude the supracrustal sequence, cropping out in the area of the villages of Antwerp and Rossie, southeast of the BLSZ. This summary of the geology of these rocks is from Chiarenzelli et al. (2010b), and additional geochemical analyses and discussion can be found in Carl and deLorraine (1997). The crystallization age of this suite is determined to be a 1203±13.6 Ma based on SHRIMP-RG analysis of zircon (Chiarenzelli et al. 2010b). We include in this discussion data from some mafic metamorphosed bodies (Carl, 2000) that are though to be members of this suite (Chiarenzelli et al., 2011): the Pleasant Lake Metagabbro, Metagabbros near Harrisville, Balmat Metadiorite, Balmat Metadiorite, and the Split Rock Metadiorite. These bodies generally share the same geochemistry as the rest of the Antwerp Rossie suite, although some span the 52.5–62.5 wt% SiO₂ gap recognized in other members of this suite (Chiarenzelli et al., 2010b).

Figure 3. AFM geochemistry of the Antwerp-Rossie suite, correlative metadiorites, Hermon granite, and amphibolite layers in the Popple Hill gneiss (from Carl and deLorraine, 1997, Carl, 2000, Chiarenzelli et al., 2010, and Chiarenzelli et al., unpub data).

The Antwerp-Rossie suite has high-K, mild calc-alkaline geochemistry (Fig. 3,6) and has broadly arc-like trace element contents with high Cs, Pb, La, and Nd and low Nb, Ta, P, Ti, and Zr (Fig. 4,5). Rare Earth elements show LREE enrichment (La$_n$/Sm$_n$=1.8 to 7.2) and depleted HREE (Sm$_n$/Yb$_n$=1.9 to 14.3), typical of a garnet-bearing source (Fig. 5). In general, the most felsic members of the suite (Antwerp granitoids, SiO$_2$ >62.5 wt%) have higher La$_n$/Sm$_n$ (av.=4.8) and Sm$_n$/Yb$_n$ (av.=7.0) than mafic rocks (Rossie metadiorites, SiO$_2$ <52.5 wt%), where La$_n$/Sm$_n$ and Sm$_n$/Yb$_n$ are lower yet still elevated (av.=3.0 and 4.3, respectively).
Neodymium model ages ($T_{DM}$) for the Antwerp-Rossie suite range from 1.3 to 1.6 Ga, with Epsilon Nd values of 1.5 to 5.4 at 1.2 Ga (Chiarenzelli et al. 2010b). The oldest $T_{DM}$ (~1.6 Ga) and smallest Epsilon Nd values are found in proximity to the BLSZ. Because of its location at the northwest margin of the Adirondack Lowlands and its geochemistry the Antwerp-Rossie suite is interpreted to have formed during closure of a back-arc basin between the Frontenac terrane and the Adirondack Highlands during the Shawinigan orogeny at 1.2 Ga.

**1182 Ma Hermon Granitic Gneiss**

This suite of hornblende granitic gneisses, often exhibiting distinctive K-feldspar augen (see Fig. 7), is also found southeast of the BLSZ. The Hermon granite has an age
slightly younger than the Antwerp-Rossie suite, 1182±7 Ma (U-Pb zircon by SHRIMP II, Heumann et al., 2006). The Hermon granite has several geochemical similarities to the felsic (Antwerp) members of the Antwerp-Rossie suite: it has calc-alkaline (to alkali) major element geochemistry, similar LREE enrichment (La\textsubscript{n}/Nd\textsubscript{n}=2.4 to 6.5) and depleted HREE (Sm\textsubscript{n}/Yb\textsubscript{n}=1.6 to 9.3), and like the Antwerp-Rossie suite plots with volcanic arc granites on tectonic discrimination diagrams (Fig. 4; see also Carl and deLorraine, 1997). Individual Hermon granite bodies tend to be compositionally homogeneous, but taken together the suite of plutons spans ~30 wt% SiO\textsubscript{2} (Carl and deLorraine, 1997). The Hermon granite shares the high Cs, La, and Nd and low Nb, P, and Ti with the Antwerp-Rossie suite (Fig 5). Differences include higher Ta (5.6±3.9 ppm) and Zr (often >300 ppm at intermediate SiO\textsubscript{2} contents) and lower Pb (2.3±0.9 ppm). Also, the Hermon granite suite is more potassic (K\textsubscript{2}O=4.7±1.0 wt%) and on average more felsic (SiO\textsubscript{2}=65±5 wt%) than the Antwerp-Rossie suite. The Hermon granite shares similar neodymium isotope compositions to the Antwerp-Rossie suite (Chiarenzelli et al., 2010b).

Figure 7. Typical outcrop of Hermon granitic gneiss east of the Edwardsville pluton.
1172 Ma Hyde School Granitic Gneiss

The Hyde School gneiss (HSG) bodies are perhaps the most controversial igneous suite in the Adirondack Lowlands with respect to their protolith and origin. These domical bodies range from alaskite to tonalite in composition and contain both concordant amphibolite layers and amphibolitic dikes (Fig. 8). Buddington (1929) interpreted the HCG to be plutonic bodies intruded as phacoliths, while Carl et al. (1990) and Carl and deLorraine (1997) and Carl (2000) interpreted these rocks as being metamorphosed volcanics. In this model, zoning within these bodies is interpreted to be the preserved volcanic stratigraphy and they are interpreted as representing the base of a Lowlands supracrustal sequence. The HCG bodies occur to the southeast of the Black Lake shear zone.

McLelland et al. (1991) argued that the HSG has a plutonic origin, containing magmatic assemblages and textures and that rocks in contact with the HSG show contact metamorphic temperatures higher than regional isotherms (see also Hudson, 1994). Single-grain U-Pb dating of zircon from two bodies of HSG has been interpreted as showing that these rocks are contemporaneous with the 1172 Ma Rockport granite (Wasteneys et al., 1999). This age was also seen in one multi-grain zircon population by McLelland et al. (1991), along with older inherited zircons. This 1172 Ma age is similar to metamorphic garnet and monazite ages (Mezger et al., 1991) and migmatite formation in the Lowlands (Heumann et al., 2006), indicating that the current level of exposure was at depth at this time and supporting a plutonic origin for the HSG. Because Lowlands metasediments are cross-cut by the 1203 Antwerp-Rossie suite, the ca. 1172 Ma HSG must intrude into this package as well, as opposed to being contemporaneous with sedimentation and part of the stratigraphic succession.
Figure 8. Buddington’s (1929) geologic sketch map of the Canton and Pyrites Hyde School Gneiss bodies, showing the garnet-rich contact zone and domical structure of these bodies.

The Hyde School gneiss ranges from intermediate to felsic chemistry (Fig. 9,12), but typically with SiO$_2$ >75 wt% (McLelland et al., 1991), which is higher than the Hermon granite and Antwerp-Rossie suite. The Hyde School gneiss is for the most part ferroan (Frost et al., 2001), especially at high SiO$_2$, as opposed to the magnesian character of the Hermon granite and Antwerp-Rossie suite. Mafic (amphibolitic) layers and dikes in the HSG generally plot similarly with tonalitic Hyde School gneiss, consistent with a co-magmatic origin (Fig. 9-12, McLelland et al., 1991). On tectonic discrimination diagrams (Fig. 10) the Hyde School gneiss plots transitionally between within-plates granite and volcanic-arc granites, with the tonalitic facies making up the majority of samples in the volcanic-arc fields (Wasteney et al., 1999). On spider diagrams (Fig. 11) the HSG shows similar high Cs, Pb, La, and Nd and low Nb, Ta, P, and Zr to the Antwerp-Rossie suite.

Overall, the Hyde School gneiss has high REE, mild LREE enrichment (av. La$_n$/Nd$_n$=2.5) and slightly depleted (relatively flat) HREE (av. Sm$_n$/Yb$_n$=2.5), with moderate negative europium anomalies. Tonalitic HSG is somewhat lower in REE and more depleted.


Figure 11. Spider and rare earth element plots of the Hyde School gneiss (left) and Rockport granite (right), normalized to primitive mantle and chondrite, respectively (Sun and McDonough, 1989). Also plotted is a representative Hyde School marginal gneiss for comparison to the Hyde School gneiss.
Figure 12. Major element geochemistry of the Hyde School gneiss (HSG, red circles), HSG amphibolite layers and dikes (small circles), and Rockport granite (triangles).

HREE (Sm$_n$/Yb$_n$>2.5; McLelland et al., 1993). HREE patterns are consistent with a plagioclase-bearing parent rock, and some concave-up HREE patterns indicating pyroxene or hornblende in the source as well (Catalano, 2009). Depleted mantle neodymium model ages (T$_{DM}$) for the Hyde School gneiss bodies range from 1.2 to 1.4 Ga (McLelland et al., 1993), which is similar to the Antwerp-Rossie suite and Hermon granite.

**Hyde School Marginal Gneiss (HSMG)**
The Hyde School Marginal Gneiss (HSMG) is very interesting rock type of problematic protolith that occurs discontinuously at the margin of the HSG bodies. Broadly resembling a paragneiss, these deformed rocks are for the most part quartzofeldspathic
and contain large (1-3 cm) garnets with variable amounts of sillimanite, biotite, cordierite, and spinel, with minor components containing calc-silicate minerals (and graphite) (Mclelland et al., 1991; Hudson, 1994). In the model of Carl et al. (1990) this unit represents a basal pelite to the lower marble unit in the Lowlands (and sits atop the HSG). McLelland et al. (1991) interpreted these rocks as representing pelites that had undergone melt extraction. In this model, these units could represent pelitic country rock that have been concentrated at the margin of the plutons during emplacement (e.g. Corriveau and Leblanc, 1995). Based on some similarities in chemistry, Hudson (1994) and Nowak et al. (2009) interpreted these rocks as representing hydrothermally altered marginal HSG rocks, subsequently metamorphosed.

Figure 13. Rare earth element plot of HSMG from Hudson (1994) and Nowack (2009) normalized to the Post-Archean shale composite of Taylor and McLennan (1955). The sample with the highest HREE was collected from the HSMG/HSG contact and is similar to typical HSG.

Figure 14. CaO vs FeO for HSMG (stars) compared to HSG (circles). Note that only tonalitic HSG has similar FeO contents to HSMG, and at very different CaO contents.

In general, the chemistry of these rocks is most consistent with metasediments, with some samples having undergone interaction with adjacent HSG. Most samples of HSMG have SiO$_2$ ranging from 55 to 65 wt%, Al$_2$O$_3$ ranging from 16 to 22 wt%, FeO ranging from 9 to 12 wt%, and MgO ranging from 1 to 3 wt% (Hudson 1994; Nowack 2009). Most HSMG plot as shales using SiO$_2$/Al$_2$O$_3$, and Fe$_2$O$_3$/K$_2$O as discriminants (after Herron 1988). Compared to chondrite compositions, HCMG rocks are LREE-enriched (av. Lan/Ndn=3.3) and HREE- depleted (av. Sm$_n$/Yb$_n$=2.9), also similar to shales (Fig. 13). Elemental composition of these rocks differs in detail from HSG, in
particular with the HSMG having higher Sr, CaO, FeO, Al$_2$O$_3$ and lower SiO$_2$, Na$_2$O, Zr, Hf, and HREE (Figs 14).

Stable isotopes also point towards sedimentary protoliths. Carbon isotopes of graphite-bearing HSMG are variable (δ$^{13}$C range from -7.4‰ to -17.4‰ PDB; Nowack 2009), but are broadly consistent with reduced carbon from metasediments and not with graphite from marbles. Only the Gouverneur HSG body has HSMG adjacent to country rock marbles, and its graphite has carbon isotope ratios similar to those in the marbles (δ$^{13}$C -3.9 to -4.7‰). The northern California HSG locality of McLelland et al. (1991; see McLelland 1993 and Hudson, 1994) was also analyzed for oxygen isotopes by Lancaster et al. (2009). Minerals in this migmatitic HCG correspond to δ$^{18}$O(WR)≈13‰ SMOW, a typical value for pelites and ca. 4‰ higher than igneous values for the HSG (Carl et al., 1990). REE patterns of HCMG do not have patterns clearly indicative of HREE-enrichment by melt extraction from a garnet-bearing restite (e.g. Peck and Smith, 2005), but depending on amount of melt the REE pattern may not be particularly sensitive to this process, especially in HREE-depleted protoliths, such as shales.

1172 Ma Rockport Granite and Granitic Gneiss

The Rockport granite (Fig. 15) is an igneous-textured to gneissic pink leucogranite which crops out along the shores and on islands in the St. Lawrence River in Ontario and New York, but does not extend east of the Black Lake shear zone. Rockport granite is typically emplaced as dikes and sheets into metasedimentary country rocks, and sometimes forms spectacular net-vein textures when emplaced into calc-silicate gneiss (Carl and deLorraine, 1997). Gneissic, recrystallized Rockport gneiss, massive igneous-textured Rockport granite, and isolated Rockport granite dikes share 1172 Ma crystallization ages (van Breemen and Davidson, 1988; Wasteneys et al., 1999). Single-grain U-Pb geochronology of Rockport-type granite on Wellesley Island (Wasteneys et al., 1999) showed that the age of 1416±5 Ma previously determined by multigrain techniques (McLelland et al., 1991) was probably the result of xenocrystic zircon contamination and variable U-Pb discordance. In addition, texturally early Rockport granites plot on geochemical variation diagrams similarly to rocks that cross-cut them,
supporting that the different phases of Rockport granite are comagmatic as well as contemporaneous (Carl and deLorraine, 1997).

Wasteneys et al. (1999) interpreted the Hyde School gneiss and Rockport granite as belonging to the same magmatic suite based on lithologic similarity and U-Pb zircon ages. Differences include the lack of a tonalitic facies and amphibolite in the Rockport granite and the differences in emplacement style of the two suites (HSG domes vs. Rockport dikes and sheets). These differences may reflect differences in depth of emplacement, or differences in country rock on either side of the Black Lake shear zone. It is possible that marble-rich country rocks southeast of the shear zone promoted domical emplacement of the HCG (cf. Corriveau and Leblanc, 1995), while quartzite-rich country rocks northwest of the Black Lake shear zone caused fracture-controlled emplacement of Rockport granite.

Correlation of Hyde School gneiss and Rockport granite is borne out by geochemical data. On average, Hyde School gneiss and Rockport granite have similar SiO$_2$ (~71-72 wt%), overlap on major element variation diagrams (Fig. 12), are both ferroan (Frost et al., 2001), and share similar positive Cs, Pb, La, and Nd and negative
Nb, Ta, P, and Zr anomalies (Fig 11). Included here are outcrops around Hammond NY that we correlate with Rockport granite (Wong et al., 2011), which share the same geochemistry. Rockport granite has larger LREE enrichment than HSG (av. \( \text{La}_n/\text{Nd}_n = 4.0 \)), which correlates with LREE contents. Like the HSG, Rockport granite has slightly depleted (relatively flat) HREE (av. \( \text{Sm}_n/\text{Yb}_n = 2.7 \)), but with lower overall HREE than HSG, and with similar moderate negative europium anomalies. Like the HSG, HREE for Rockport granite patterns are consistent with a plagioclase-bearing parent rocks, and with pyroxene or hornblende in the sources of some (Catalano, 2009).

1150-1180 Ma Frontenac suite

This plutonic suite is made up of granite, monzonite, syenite, and gabbro plutons emplaced into metasediments during the period 1150-1180 Ma (Marcantonio et al., 1990; Davidson and vanBreemen, 2000; Peck et al., 2004; Grammatikopoulos et al., 2007). These plutons preserve igneous textures and contact relationships except where they are deformed by the Maberly shear zone in the western Frontenac terrane and eastern Sharbot Lake domain. Undeformed Frontenac suite plutons share unusually high \( \delta^{18}\text{O} \) values, which is interpreted as indicating a hydrothermally altered ocean crust component in the basement of the Frontenac terrane (Peck et al. 2004). Many of these plutons contain different igneous compositions separated by country rock screens, and some show magma mingling textures.

Although the Frontenac suite was emplaced in part during Rockport granite magmatism, these suites are geochemically distinct. The Frontenac suite is on average not as evolved as Rockport granites and has a more continuous range of compositions on variation diagrams (Fig. 16,19). While Rockport granite and the HSG are strongly ferroan, the Frontenac suite straddles the ferroan/magnesian compositional boundary (Frost et al., 2001). In addition, while the Rockport granite for the most part is chemically alkali calcic to calc-alkalic, most Frontenac suite rocks are alkalic (Frost et al., 2001). Within the Frontenac suite there are some differences in major element geochemistry between plutons. For example, mafic rocks of the Crow Lake pluton appear initially saturated in apatite, which the mafic Leo Lake pluton was apparently apatite-undersaturated in rocks with < 50 wt% SiO._2_. Although as a suite it shows a
continuous range of compositions for major elements, this is not the case for all plutons taken alone; for example the South Lake and Crow Lake plutons are bimodal with ~10 wt% gaps in SiO₂ in-between mafic and felsic lithologies.

Figure 16. AFM geochemistry of the Frontenac suite of the Frontenac terrane and Sharbot Lake domain, and the Edwardsville syenite (data from Ermanovics, 1970; Curie and Ermanovics, 1971; Marcantonio et al., 1990; Davidson, 1996, Carl, 2000; Grammatikopoulos et al., 2007; Stocker, 2009; unpublished data).


Figure 18. Spider and rare earth element plots of the Frontenac suite (yellow) and Edwardsville pluton (blue), normalized to primitive mantle and chondrite, respectively (Sun and McDonough, 1989).
The Frontenac suite lacks or has much less prominent negative anomalies in Sr, P, and Ti than the Rockport granite, and has additional negative anomalies in Zr, Th, and U on trace element spider diagrams (Fig. 18). The Frontenac suite and Rockport granite both plot similarly as transitional between within-plate and volcanic-arc granites on tectonic discrimination diagrams (Fig. 17). The Frontenac suite shows increasing Zr with SiO$_2$ until ~60 wt%, after which Zr declines; a typical relationship for A-type granites (Whalen et al., 1987). In contrast, Rockport granites have steadily increasing Zr with SiO$_2$ to the highest Zr in the most evolved rocks.

The Frontenac suite has overall high REE except for a mafic sample from the Leo Lake pluton and three felsic samples from South Lake plutons, which have LREE ≈ 10-80x chondrite. All Frontenac suite samples have similarly mild LREE enrichment (av.
Lan/Ndₐ=2.7) and HREE depletion (av. Smn/Ybn=3.4), with negligible europium anomalies. Depleted mantle neodymium model ages (TDM) for the Frontenac suite range from 1.3 to 1.5 Ga (Marcantonio et al., 1990), which is similar to Antwerp-Rossie, Hermon granite, and Hyde-school magmatism.

Ca. 1160 Ma Edwardsville syenite

The Edwardsville pluton is a composite body of broadly syenitic composition that sits adjacent and to the east of the Black Lake shear zone (Fig. 20). This body has been called the Pope Mills mass (Buddington, 1934), Edwardsville syenite (Mclelland et al., 1993; Peck et al., 2004), and Pope Mills metagabbro (Carl, 2000). Buddington (1934) describes this body as “a belt of pink to red syenite…intrusive into a belt of pyroxenic amphibolite with which it forms a migmatite”, and correlates it with other mapped (but essentially unstudied) syenite bodies in the Lowlands. The Edwardsville pluton’s northern lithology is a (leuco-) syenite containing occasional mafic enclaves and country rock xenoliths. Mafic lithologies making up the southern portion of the pluton include a fine-grained equigranular melanocratic syenite that is cross-cut by porphyritic melocractic syenite (Fig. 21). Major-element variation diagrams show smooth trends leading from fine-grained melanocratic syenite (SiO₂≤ 49 wt%) to porphyritic melocractic syenite (SiO₂≤ 53 wt%) to leucocratic syenite (SiO₂≥ 62 wt%), with a gap at ca. 54-60 wt% SiO₂, and a single wispy sample of 57wt% that may represent magma mingling. Major elements and similar trace elements compositions (Figs. 16-19) are
taken as indicating that the Edwardsville syenite pluton is an eastern member of the Frontenac suite, as may be some other bodies in the area mapped as syenite by Buddington (1934). Like other members of the Frontenac suite, Edwardsville pluton samples have mild LREE enrichment (av. La$_{n}$/Nd$_{n}$=2.3) and HREE depletion (av. Sm$_{n}$/Yb$_{n}$=3.2), with small or no europium anomalies.

Figure 21. Leucocratic syenite (A) and porphyritic melanocratic syenite (B) of the Edwardsville Pluton (Stocker, 2009).

Geochrononology of the Edwardsville Pluton also correlates it with the Frontenac suite. ID-TIMS dating of zircon from a felsic lithology (McLelland et al., 1993) yielded 1164±4 Ma. This is identical to the 1149±21 Ma determined by SHRIMP-RG for a mafic phase of this pluton. Our SHRIMP-RG dating of another felsic lithology yielded 1187±19 Ma, which probably reflects some degree of inheritance from source rocks (see Fig. 22).

Oxygen isotopes of this pluton also tie it to the Frontenac suite. The δ$^{18}$O of zircon from the dated melanosyenite sample is 9.06‰, which is high for a mafic igneous rock. Two felsic syenites have δ$^{18}$O values of 11.15 (this study) and 11.11‰ (Peck et al., 2004). The Edwardsville pluton is on the eastern side of the Black Lake Shear Zone, but the δ$^{18}$O of these zircons may indicate that the Edwardsville pluton is still derived from Frontenac terrane basement in the subsurface (Peck et al., 2004).
Figure 22. New SHRIMP-RG U-Pb geochronology of leucocratic syenite (left) and melanocratic syenite (right) and of the Edwardsville Pluton (from Stocker, 2009). We interpret 1149±21 Ma to be a more accurate estimate of the age of this pluton, given the mafic character of this melanocratic syenite and lack of zircon inheritance (contrast to the abundant cores in the leucocratic syenite).

Discussion of Plutonic Geochemistry

The magmatic geochemistry in the Adirondack Lowlands progresses from bimodal calc-alkaline (Antwerp-Rossie suite) to potassic calc-alkaline (Hermon granite), through alkali calcic/calc-alkalic (Hyde School and Rockport granites), and finally to strongly alkalic (Frontenac suite) during a period of ca. 50 m.y. Transition from calc-alkaline subduction-related magmatism to alkalic compositions is reported from other collisional orogens. Subduction-related Antwerp-Rossie suite magmatism is interpreted as being formed during closure of a back-arc basin between the Frontenac terrane and the Adirondack Highlands (Chiarenzelli et al., 2010b). The calc-alkaline Hermon granite may record the last melting related to subduction during docking of these terranes, or perhaps subsequent collisional melting of mantle previously metasomatized during Antwerp-Rossie subduction, a feature seen in other collisional settings (e.g. Pearce et al., 1990). Syn-collisional Hyde School and Rockport magmatism and post-collisional
Frontenac-suite magmatism are more alkalic than earlier magmas, and they have trace elements similar to other alkalic post-collisional suites elsewhere (i.e. They are A2-type granites after Eby, 1992). Post-collisional magmatism with similar compositions has been ascribed to crustal delamination, creating a regional area of plutonism or alternatively slab break-off, resulting in a structurally controlled and relatively narrow zone of melting (e.g. Tang et al., 2010). The former seems to be the case in the Adirondack Lowlands, and post-collisional collapse and crustal delamination has also been called upon as the cause of voluminous ~1.15 Ga anorthosite-mangerite-charnockite-granite (AMCG) magmatism in the Adirondack Highlands and in Quebec (McLelland et al., 1996; Corrigan and Hanmer, 1997). The Frontenac suite is correlative to the 1.15 Ga AMCG suite (Peck et al., 2004), and the Hyde School and Rockport granites may represent slightly earlier lower-crustal melting during this event.

**METAMORPHIC PETROLOGY**

Figure 23. Calcite-graphite fractionations from the western Adirondack Lowlands. Samples from Kitchen & Valley (1995) are shown with *s, and a grey dashed 675 °C isotherm. Newer samples from Will (2009) are bold, with a revised 675 °C isotherm shown in black. Figure from Will (2009).

The Adirondack Lowlands experienced amphibolite-facies metamorphism (640–680°C, 6–7 kbar; Bohlen et al., 1985; Kitchen and Valley, 1995; Fig. 23). U-Pb ages for
metamorphic minerals with high closure temperatures overlap with the ages of plutonic suites, and are 1.17-1.13 Ga for garnet, 1.16 Ga for monazite, and 1.16-1.10 Ga for sphene (Mezger et al., 1991). $^{40}$Ar-$^{39}$Ar ages of hornblende and micas increase between the Carthage-Colton mylonite zone and the Black Lake shear zone, which is interpreted as indicating slow cooling between 1.1 and 0.9 Ga and subsequent tilting of the Lowlands accommodated by movement on the Carthage-Colton mylonite zone (Fig 24, Dahl et al., 2004).

Figure 24. Summary of geochronology of metamorphic minerals and minerals with closure temperatures below peak metamorphism in the Adirondacks and adjacent Ontario, from Dahl et al. (2004; see also references therein). The Black Lake shear zone marks discontinuities in titanite and hornblende ages, and most notably a break in slope of hornblende ages. Note that more geochronology in the eastern Frontenac terrane would help constrain these relationships.

Metamorphic conditions in the Frontenac terrane reached granulite facies and have been estimated at 700-800°C and ~6 kbar (Streepey et al., 1997; Tortorello and Peck, 2010). Timing of metamorphism is constrained by U-Pb dating of sphene at 1.18-1.15 Ga and monazite at 1.17-1.14 Ga (McLelland et al., 1993; Mezger et al., 1993; Corfu and Easton, 1997). $^{40}$Ar-$^{39}$Ar cooling ages of hornblende are constrained to 1.12-
1.10 Ga (Cosca et al., 1992). The majority of metamorphic sphene in the Frontenac terrane are up to 50 m.y. older than in the Adirondack Lowlands, indicating cooling below ~650°C was diachronous across this boundary (Mezger et al., 1993). ⁴⁰Ar-³⁹Ar cooling ages in the Adirondack Lowlands suggest tilting after 0.9 Ga (Dahl et al., 2004), which is not seen in the Frontenac terrane (Cosca et al., 1992), suggesting that the 1.12-1.10 Ga paleoisotherms in the Frontenac terrane are still broadly horizontal (Fig. 24).

**STRUCTURAL GEOLOGY**

The following summary of the structural geology of the BLSZ region is based on results reported in Wong et al. (2011). Rocks within the proposed Black Lake shear zone have penetrative fabrics that vary in intensity and orientation. Metasedimentary rocks such as calc-silicate and schist are typically strongly foliated whereas fabric development in granitoid rocks is more variable and can range from weakly to highly deformed (Figure 25). Foliation in granitoid rocks is typically defined by both aligned mafic minerals such as micas as well as by highly elongate quartz and feldspar grains. As a result, it is clear that fabrics are not magmatic but instead largely record solid–state deformation. The dominant foliation in both metasedimentary and granitoid units, where present, strikes NE and dips steeply (60-70°) NW or SE, although other foliation orientations are present locally (Figure 26).

Although foliation is the most common tectonic fabric, penetrative lineations are present in some areas, including some outcrops along the I-81 highway locality and limited areas of Wellesley Island (Fig. 26). The lineations are present in both metasedimentary units and various granitoids and dikes, and are defined by elongate clinopyroxene grains in the calc-silicate unit and by elongate quartz and feldspar grains in the granitoids. The lineations in the calc-silicate unit typically occur in 5–10 meter wide zones and are weakly to moderately defined, whereas lineations in the granitoids are well defined where they are present. The lineations define two nearly orthogonal sets; one that is sub-horizontal and another that is nearly down-dip of the foliation. Sub-horizontal lineations are most common in the metasedimentary rocks and have a mean trend and plunge of 065°/9° (n=14) or 240°/9° (n=14) for the northeast and southwest plunging
lineations respectively (Fig. 26). Less commonly, metasedimentary units contain a down-dip lineation set that has a trend and plunge of 115°/56° for SE plunging lineations. Granitic rocks also locally contain these lineation sets, with the down-dip lineation mean trend and plunge of 117°/47° (n=4) or 302°/73° (n=3), depending on whether the foliation dips southeast or northwest. Anisotropy of magnetic susceptibility (AMS) measurements of granite indicate that K_max axes, a proxy for the lineation direction in strained rocks, have orientations that match the observed lineations, even in rocks where no obvious lineation was visible. As a result, these lineation orientations are a likely an important part of the overall strain, although they can be difficult to see in the field.

While both lineation sets are found sporadically with the study area as a whole, the lineation sets appear to be broadly partitioned into specific domains, with sub-horizontal lineations dominating the Wellesley Island and Saint Lawrence River region and down-dip lineations largely occurring further southeast (Figure 27), although both lineations are found in some areas. This transition generally occurs along the Black Lake Shear Zone.

In addition to penetrative fabrics, deformation in the study area was also accommodated by folding. One of the clearest markers of the folding are dikes, which are often tightly to isoclinally folded at outcrop scale (Fig. 26). Poles to all folded dikes in the study area define a clear northwest girdle and a best fit cylindrical fold axis with a mean trend and plunge of 048°/23° (Fig. 26). Although commonly less obvious, the foliation is also folded in some areas. Small–scale (cm-scale) folds in the calc-silicate foliation have a mean trend and plunge of 256°/8° (Fig. 26). The foliation is also folded on larger scales (100s of meters), although the hinge regions of these larger folds are narrow and are often difficult to locate in outcrop. Although we were unable to directly measure the fold axes of larger folds, poles to the metasedimentary foliation lie along a northwest trending girdle on a stereonet (Fig. 26) and a cylindrical best fit yields a mean fold axis with a trend and plunge of 62°/2°, similar to the other folds. The axial plane of these folds generally strikes northeast and is nearly vertical.
Figure 25. Photographs of key outcrops. (A) Macroscopic kinematic indicator in the calc-silicate unit along I-81 outcrop. Asymmetric tails on a large sigma-type pyroxene grain (arrow) suggest left-lateral shear in an area with sub-horizontal lineations. Image looking down with the top of image northwest. (B) Highly strained dike (arrows) that locally cross-cuts the foliation in the host calc-silicate along the I-81 outcrop. (C) Highly strained calc-silicate rock with a strong foliation defined by alternating pyroxene and plagioclase–calcite layers. (D) Highly folded felsic dike within host calc-silicate along the I-81 outcrop. (E) Foliation in moderately strained dike rotates and merges with higher strain foliation along its margins and in host rock, suggesting thrust sense shear. Top of image is up, right is northwest, Rt. 12 outcrop between Alexandria Bay and North Hammond. (F) Moderately deformed granitic dike with a west trending foliation that rotates towards a northeast orientation along its higher strain margins and in the host quartzite. Lineations in the host rock are sub-horizontal, suggesting left-lateral shear. Outcrop is located on Wellesley Island. From Wong et al., 2011.
Figure 26. Structural measurements from within the Black Lake shear zone plotted on equal area stereonets, sub-divided by rock type. (A) Foliation and lineation measurements of metasedimentary rocks. Foliations typically dip steeply northwest to southeast (n=67, only mean of southeast dipping set shown for clarity) with minor sets that dip steeply west-northwest (n=11) and moderately northeast (n=3). Lineations, where present, are typically sub–horizontal and plunge gently northeast to southwest (n=28, mean of northeast trending group shown), with a minor set that is down–dip. (B) Poles to foliation lie along a northwest trending girdle, suggesting that the foliation is folded into upright, gently northeast plunging folds, matching the orientation of measured small–scale folds. (C) Foliations in granitic rocks are similar to the dominant set in the metasedimentary rocks. Lineations are rare but more commonly trend down–dip. (D) Measured foliations and lineations in strained dikes are similar to those in other rock types. Both down–dip and sub–horizontal lineations are present in dikes. (E) Poles to dike margins of folded dikes, defining a gently northeast trending fold axis. (F) Poles to undeformed dikes, which largely match the orientation of foliation in the metasedimentary rocks they intrude. From Wong et al., 2011.

Several other structural features are present in the calc-silicate unit. Locally, pyroxene–rich layers define boudins surrounded by less competent plagioclase and calcite layers. Commonly, areas between the boudins are filled with quartz, feldspar and pyroxene pegmatite, likely derived from local partial melting, suggesting deformation during peak metamorphic conditions. The calc-silicate also has a number of discrete narrow shear zones, sometimes localized along dikes, which typically dip moderately to steeply northwest. One of these shear zones has a gently west plunging lineation along its
margin defined by elongate amphiboles. Drag folding of the adjacent foliation along this and other discrete shear zones suggests left-lateral shear.

A number of dikes are also undeformed. These dikes both cross-cut the metamorphic foliation of the host metasedimentary rock as well as parallel it. Most undeformed dikes have orientations similar to that of the dominant foliation in the study area, with a mean strike and dip of 048°/65° northwest, although there is a significant amount of scatter in these data (Fig. 26). The mean strike and dip for southeast dipping dikes is 064°/55° southeast, but this average also hides significant scatter. In general, undeformed dikes typically dip moderately to steeply and range in strike from north to east. There are few if any dikes that dip to the northeast or southwest.

**Kinematic indicators**

We observed a number of kinematic indicators, both in the field and petrographically. In the field, fabric development in deformed dikes of the Rockport/Hyde School suite provides some of the most compelling kinematic indicators. For example, on Wellesley Island, a northeast trending vertical dike intrudes host quartzite that contains a weak sub-horizontal lineation. The dike has a moderate east trending foliation that becomes more intense and rotates towards the margins of the dike, ultimately merging with the strong northeast trending foliation in the quartzite, suggesting left-lateral shear (Fig. 25) assuming non-coaxial strain. Similar relationships along a steeply dipping dike at an outcrop at the I-81 site suggest right lateral shear, although our confidence in this indicator was lower than the previous example. In an area with a down-dip lineation along Rt. 12, a moderately south–southeast dipping dike has similar foliation relationships, in this case suggesting top–north–northwest or thrust sense of shear (Fig. 25).

The calc-silicate unit also contains a number of kinematic indicators in the form of asymmetric sigma-type pyroxene porphyroclasts (Figure 25). While individual porphyroclasts often yield convincing kinematic indicators and the sense of shear is often consistent within a compositional layer, the sense of shear is not usually consistent within an outcrop as a whole, with some indicating left–lateral and others right-lateral shear. The most convincing macroscopic kinematic indicators in the calc-silicate are associated with
discrete, small-scale shear zones along which drag folding of the foliation indicated left-lateral shear. However, there were only two such shear zones, so their significance is somewhat unclear. Petrographic observations also provided compelling kinematic indicators, especially for samples with down-dip lineations. S-C fabrics and an oblique grain shape foliation in dynamically recrystallized quartz and feldspar indicate a top-NW sense of shear in mylonitic granite along Rt. 37. The foliation here dips southeast with a down-dip lineation, indicating a thrust sense of shear.

**Strain patterns**

In order to quantify strain variations in the study area, we measured elongation of individual mineral grains in thin section. We chose to measure quartz grains, as the grain boundaries of quartz in these samples were typically easiest to identify. This generally limited our strain measurements to granitoid units. Thin sections were cut perpendicular to the foliation and parallel to the lineation where present and therefore should represent the X-Z plane in the rock. We measured the entire quartz ribbon (typically in plane polarized light) rather than the polycrystalline grains within the ribbon that were likely formed during static recrystallization. Reported X:Z ratios reflect an average of all measured grains within a given sample. The X:Z ratios within the study area vary from as low as 1 in undeformed samples to as high as ~50 in the most highly strained samples. The degree of strain varies spatially within the study area, with the lowest strain values in the northwestern study area on Wellesley Island (Fig. 27). Strain values generally increase towards the southeast towards Black Lake and then appear to decrease in the easternmost part of the study area towards Gouverneur, although our data from this area is fairly limited. This northeast trending belt of high strain corresponds to the location of the BLSZ.
Figure 27. Representative structural measurements, strain measurements (X:Z ratio of strained grains) and degree of magnetic anisotropy (P') results from within the study area. The structural data from field measurements and AMS represent an average of many measurements from that general area. All of these results vary throughout the study area. When projected into the northwest transect A-A' strain measurements are generally lowest in the northwest (Wellesley Island), increase to the southeast towards Black Lake (inset) and decrease further southeast of Black Lake to Gouveneur, corresponding with the location of the Black Lake Shear Zone (BLSZ). One location in the southern study area shows anomalously high strain values with are likely associated with the Pleasant Lake shear zone. P’ values follow a similar pattern, but deviate somewhat from strain measurements on a site by site basis. Sub-horizontal lineations dominate the northwestern part of the study area near the Saint Lawrence River, while down-dip lineations dominate areas further to the southeast, with the BLSZ generally marking the transition between the lineation directions. From Wong et al., 2011.

Discussion

Previous regional compilations have identified a northeast trending structure called the Black Lake lineament (Davidson, 1995) in the Black Lake area. Taken together, our AMS and structural data suggest that a northeast trending zone of high strain exists roughly between Black Lake and the edge of the Saint Lawrence River on the U.S. side (Fig. 27), which we interpret as the location of the BLSZ. Strain in this zone was accommodated by the development of a strong foliation and to a lesser extent lineations as well as tight to isoclinal folding. The intensity of the foliation, degree of folding and aspect ratio of strained grains indicates that strain is generally high.
throughout the zone, although strain is heterogeneous in detail. Due to the heterogeneity of strain and the extent of Paleozoic sedimentary cover, it is unclear where the precise boundaries of the high strain zone are. Strain generally decreases towards the northwest of the zone in Wellesley Island as well as to the southeast towards Gouverneur, consistent with a northeast trend of the BLSZ. High strain rocks in the Honey Hill area do not match this general pattern, although they likely related to deformation on the nearby Pleasant Lake shear zone (Fig. 27; Carl et al., 1990). Regardless of the exact geometry, our results confirm the presence of high strain rocks within the BLSZ.

The dominance of S-tectonites in the BLSZ may indicate that strain was primarily northwest–directed shortening accommodated by flattening. The fold geometry of both dikes and foliation (Fig. 26) is also consistent with substantial northwest–directed shortening. However, the presence of lineations in some parts of the study area suggests that plane (2D) strain may have occurred in at least some areas. In addition, the presence of two orthogonal lineation sets also suggests a more complex strain history.

Kinematic indicators provide some insight into the significance of these lineations. Areas with down-dip lineations typically contain convincing kinematic indicators suggestive of top-NW thrust motion, which is consistent with the shortening direction inferred by the foliation and folds, suggesting these structures are part of the same deformational event. Given the strength of foliation development, the strain recorded by individual grains and the degree of folding, the magnitude of northwest shortening accommodated in this area was substantial. It remains unclear why shortening was accommodated by flattening and folding in some areas and by ductile thrusting in others.

Although evidence for northwest directed shortening dominates the study area, samples from one area (Honey Hill site, Fig. 27) suggest a top-northwest normal sense of shear. This is probably a reflection of movement on the Pleasant Lake shear zone (Carl et al., 1990) rather than the BLSZ. However, Brown (1989) interpreted the Pleasant Lake shear zone as a right-lateral structure that evolve from ductile to brittle. Our results suggest that the Pleasant Lake shear zone may have had a more complex kinematic history than previously thought. Normal-sense shear zones have been identified in other areas of the Central Metasedimentary Belt–Adirondack Lowlands, including the Carthage
Colton mylonite zone (e.g. Isachsen and Geraghty, 1986; Streepey et al., 2000; Selleck et al., 2005) the Bancroft (Carlson et al., 1990; Mezger et al., 1991) and Robertson Lake shear zones (e.g. Busch and van der Pluijm, 1996). As a result, it is possible that the Pleasant Lake shear zone represents a previously unrecognized extensional structure in the Adirondack Lowlands. However, these results were obtained from only one locality and additional work is needed to determine if this result is significant.

The kinematic significance of areas with a sub-horizontal lineation is less clear. The sub-horizontal lineation set is limited to narrow (5-10 meter wide) zones and the lineations are often only weakly to moderately defined. Kinematic indicators in these areas tend to be mixed, although a left-lateral sense of shear was more prevalent, especially among high confidence indicators. However, Reitz and Valentino (2006) reported evidence for dextral shear for sheared rocks on Wellesley Island.

There are several possible interpretations of these results. One is that the deformation was simply dominated by flattening and the lack of ubiquitous sub-horizontal lineations indicates only a modest role for localized plane strain in these areas. Alternatively, plane strain via sub-horizontal shearing may have been a significant aspect of the deformation but widespread lineations did not develop or have not been well preserved due to fine grain sizes, lithology, or poly-phase deformation. In this case, AMS results may be more sensitive to this stretching direction while field observations are not. A final possibility is that this area may have subjected to several phases of deformation, leaving behind a composite fabric that only appears to be dominated by flattening. In this case, areas that contain the sub-horizontal lineation may have escaped significant poly-phase deformation, thus preserving one phase of the bulk strain history. Given the complex history of the Grenville region, such a poly-phase deformational history is certainly possible, especially in the older metasedimentary rocks. The simplest explanation is that deformation was predominately characterized by northwest-directed flattening with narrower zones of sub-horizontal shear, perhaps dominated by left-lateral shear, although we cannot rule out a more substantial role for sub-horizontal shear.

The presence of two lineation orientations also raises the question of how these fabrics are related in time. One possibility is that the lineations developed synchronously during transpression, which would argue for an oblique convergent setting during
deformation. Transpressional deformation has been proposed for other shear zones in the region (e.g. Streepey et al., 2001; Gates et al., 2004; Mezger et al., 1993; Busch et al., 1997; Martignole and Friedman, 1998), and also has been suggested for the areas near the BLSZ (Baird and Shrady, 2009) so our results may lend further support to transpressional models for the region.

However, several lines of evidence suggest that these two lineation sets represent discrete deformational events. Although there are exceptions, down-dip and sub-horizontal lineations appear to be generally partitioned spatially, with the BLSZ marking the transition between these lineation sets (Fig. 27). In addition, there are subtle differences in the deformational style in areas with different lineation directions, as areas with the sub-horizontal lineation generally appear more equigranular, polygonal, and coarser grained, while areas with the down-dip lineation tend to be finer grained and preserve distinct core and mantle textures in feldspar. These differences suggest that the sub-horizontal lineation formed at higher temperatures and were statically recrystallized to a greater extent than areas with the down-dip lineation. If correct, the sub-horizontal lineation may have formed early and thrusting associated with the down-dip lineation occurred later. Alternatively, it is possible that these differences are a function of varying strain rates/magnitudes and/or compositional or initial grain size differences in the protolith. We hypothesize that these lineation orientations represent distinct deformational events, with an early flattening/sub-horizontal shearing event followed by late-stage thrusting.

**Timing of deformation**

The results presented in this study, when combined with previous geochronologic work, provide constraints on the timing of deformation in the BLSZ. Little is known about the depositional age of the metasedimentary rocks the Adirondack Lowlands, other than they predate intrusion of the Antwerp–Rossie suite at ca. 1203 Ma (Chiarenzelli et al., 2010b). Thus, we cannot put good upper bounds on the timing of deformation in the metasedimentary units. However, the Rockport Granite and age equivalent Hyde School gneiss (ca. 1170 Ma; Wasteneys et al., 1999) are both strongly deformed locally. U-Pb SHRIMP titanite dates from deformed dikes show that they were intruded and
metamorphosed at ca. 1150–1160 ± 30 Ma (Wong et al., 2011). These ages overlap with the Rockport/Hyde School granitoid suite within error and compositional similarities suggest the dikes are most likely related to that suite. Metasedimentary units are typically highly strained, whereas strain in granitoid rocks is variable, suggesting that some deformation predated ca. 1170 Ma, although significant deformation clearly occurred during or after ca. 1170 Ma as well.

Evidence for ongoing deformation at ca. 1170 Ma comes from geochronology on the Rockport suite by van Breemen and Davidson (1988) and Wastenays et al. (1999), which yields nearly identical ages of ca. 1172 Ma on both highly and weakly deformed samples. This suggests that the Rockport was likely intruded syn-tectonically, perhaps progressively during deformation over a period of several million years. Thus, some Rockport granite is highly strained while other exposures are only weakly deformed. Strain was also likely to have been heterogeneous, which may also have contributed to differences in the degree of strain within the same unit.

This interpretation is also consistent with petrographic observations, which document deformation at amphibolite grade temperatures, consistent with amphibolite facies metamorphic conditions in the Lowlands (Kitchen and Valley, 1995). These peak conditions were likely achieved at ca. 1170 ± 10 Ma (Heumann et al., 2006), further supporting our interpretation that the Rockport suite is syn-tectonic. Given that quartz is statically recrystallized and that feldspar shows no evidence for brittle deformation, deformation must have ceased prior to cooling below ca. 500°C. Although undeformed dikes in the region are undated, ⁴⁰Ar/³⁹Ar ages on hornblende indicate that this part of the Lowlands cooled below ca. 500°C by ca. 1100 (Dahl et al., 2004), indicating that deformation was complete by that time, although we believe that most deformation along the BLSZ was likely complete well before this. Just northwest of the BLSZ, undeformed dikes of the ca. 1160 Ma Kingston swarm (Davidson, 1995) cut Rockport granite, especially on Grindstone Island (Cushing et al., 1910). If this relationship is representative of the area as a whole, then deformation along the BLSZ may have been largely complete by ca. 1160 Ma. Thus, our results suggest that deformation within the BLSZ was ongoing by ca. 1170 Ma and complete no later than ca. 1100 Ma and likely earlier, perhaps by as early as 1160 Ma.
Regional implications

A number of models have been proposed for the tectonic evolution of the Adirondack Highlands/Lowlands and Frontenac terrane region. Most of these models emphasize a poly-phase tectono–magmatic history including the early Elzevirian orogeny (ca. 1245–1220 Ma), AMCG magmatism (ca. 1160–1140 Ma), and the later Ottawan (ca. 1090–1020 Ma) and Rigolet (ca. 1010–980 Ma) orogenies (e.g. McLelland et al., 1996; Rivers, 1997; Wastenays et al., 1999; McLelland et al., 2010). Recent work has emphasized the significance of the contractional Shawinigan orogeny from ca. 1190–1140 Ma (Rivers, 1997; Rivers and Corrigan, 2000), which preceded or overlapped with AMCG magmatism. The Shawinigan is typically attributed to the collision of continental fragments, likely the Adirondack Highlands and associated blocks, to the Laurentian margin (e.g. McLelland et al., 2010). Heumann et al. (2006) documented widespread partial melting in the Adirondack Lowlands during Shawinigan time, highlighting the thermal significance of the Shawinigan to the Lowlands. Although significant Ottawan deformation has been proposed for the Lowlands (e.g. McLelland et al., 1996), our results indicate that deformation in the BLSZ occurred during the late stages of Shawinigan, thus highlighting the role of the Shawinigan in producing not only metamorphic events in the Lowlands but also deformation as well. The northwest directed shortening found in this study correlates well to other studies of Shawinigan deformation (e.g. Carr et al., 2000; Rivers and Corrigan, 2000; McLelland et al., 2010) and are consistent with a collision in that direction, perhaps with a component of horizontal shear. Our results highlight the role of the BLSZ as a likely Shawinigan suture between the Frontenac and Adirondack Lowlands terranes.

Given that ca. 1170 Ma granites of the Rockport and Hyde School suites are present in both the Frontenac Terrane and the Adirondack Lowlands respectively and are adjacent to the BLSZ, we interpret that these plutonic units define a stitching suite between the two terranes, assuming that these units are correlative and not simply coeval. This interpretation would require that the two terranes were juxtaposed by ca. 1170 Ma. It is important to note that the correlation of these suites is geochronologic, not geochemical: the Hyde School bodies have a wider range in SiO₂ and distinct trace
element compositions from the Rockport granite (Carl and deLorraine, 1997).
Geochemical differences and distinct emplacement styles (in this case dikes vs. domical intrusions) are expected for a stitching suite derived from different basement rocks and emplaced into different country rocks.

Our results on the BLSZ may have implications for the different tectonic models that have been proposed for the southern Grenville Province. Many of these models tend to interpret the Frontenac terrane and Adirondack Lowlands metasedimentary rocks as representing passive margin environments, possibly located on the trailing margin of an arc represented by volcanic rocks of the Elzevir terrane (Wasteneys et al., 1999; Peck et al., 2004). Other reconstructions place the Frontenac terrane and Adirondack Lowlands at the margin of the Adirondack block offshore of Laurentia (e.g. Carr et al., 2000). In these scenarios the Black Lake shear zone represents the juxtaposition of related but distinct tectonic elements during the Shawinigan terrane assembly. Recognition of ultramafic rocks with enriched upper mantle affinities and associated mafic rocks interpreted as ocean crust in the Adirondack Lowlands (Chiarenzelli et al. 2010a) has led to a new tectonic model that interprets the Lowlands as a back-arc basin (Chiarenzelli et al., 2010b). In this model, rifting separated parts of the Adirondack Highlands from the Frontenac terrane, and the Black Lake shear zone marks the rifted pre-1170 Ma margin of Laurentia that was later closed by Shawinigan contraction.

Our results may also have broader implications for the correlation of regional terranes and shear zones in the Grenville. North of the Ottawa River, the eastern Central Metasedimentary Belt of Quebec (CMB-Q) is likely the extension of the Frontenac terrane, and its eastern margin is marked by the Labelle Shear Zone (LSZ; Fig. 1). This high-strain zone separates rocks of the amphibolite-facies CMB-Q from the granulite-facies Morin terrane (Indares and Martignole, 1990), which is geologically similar to the Adirondack Highlands (Peck et al., 2005). The LSZ has a width of ~10 km and is made up of highly strained ortho- and paragneisses from the adjacent terranes (Martignole and Corriveau, 1991, Peck et al., 2005). The LSZ contains a gneissic foliation with an early subhorizontal mineral lineation with kinematic indicators showing sinistral sense of shear (Indares and Martignole, 1990) and a later down–dip lineation is defined by amphibolite-facies minerals and suggests downfaulting of the CMB-Q (Zhao et al., 1997). In the
easternmost LSZ south–dipping mylonites show evidence for top to the north thrusting of the Morin terrane (Martignole and Reynolds, 1997). The age of LSZ deformation is poorly constrained, but the zone seems to have been active into the Ottawan Orogeny. One granitic dike that crosscuts mylonitic rocks of the LSZ has been dated, and it constrains deformation to >1079 Ma (Martignole and Reynolds, 1997). This dike is part of an en echelon group that is interpreted to be late in the deformation history. In addition, the LSZ also deforms the margin of the 1076 Ma Loranger pluton (Corriveau et al., 1990; Martignole and Reynolds, 1997). One interpretation of relationships described above requires that the Labelle Shear Zone acts as the continuation of both the BLSZ and the Carthage-Colton mylonite zone (CCMZ), accommodating both Shawinigan contraction/transpression of the BLSZ and Ottawan collapse of the Carthage Colton mylonite zone. If this is the case, the Adirondack Lowlands may represent a terrane that does not continue to the north into Quebec, at least as a discrete tectonically bounded block, and instead may pinch out beneath the Paleozoic sedimentary cover in the intervening Saint Lawrence basin.
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Starting Point: 44.3279°, -75.9189°
Big M Parking Lot, Route 12 just south of Alexandria Bay, NY.

Stop 1: 44.3729°, -75.7037°
Lineated eye-schist on South Hammond Road

This outcrop is part of a belt of biotite-sillimanite schists mapped in the area by Buddington (1934) as commonly containing white aggregates (‘eyes’) containing sillimanite, quartz, and feldspars, sometimes with magnetite cores. A variety of eye mineralogies are observed: 1) quartz cored by sillimanite, 2) quartz cored by sillimanite +chlorite, 3) quartz +sillimanite, 4) quartz +kspar +plagioclase surrounded by sillimanite, 5) sillimanite+plagioclase, 6) sillimanite +plagioclase with a plagioclase rim, 7) Kspar +plagioclase +magnetite, and 8) quartz +kspar +plagioclase +sillimanite (Buddington, 1934). In this outcrop the eyes define a strong downdip lineation to the south.
In general, mineral assemblages in this unit are consistent with upper amphibolite-facies conditions. Ti-in-biotite thermometry in these rocks tends to overestimate metamorphic conditions (~750°C) while two feldspar thermometry seems to be reset, and yields apparent temperatures of ~500°C (Russell, 2009; Russell et al., 2009). Calcite-graphite carbon isotope thermometry from regionally distributed marble units, part of the Lowlands metasedimentary package, yield more petrologically realistic temperature estimates of ~675°C (see Fig. 23; Kitchen and Valley, 1995; Will 2008; Russell et al., 2009).

Portion of the Hammond 15’ quadrangle (Buddington, 1934) showing field trip stops 1 to 3 from Day 1. Gpbg= Biotite gneisses and associated metasediments, including eyeschists. dqd= diorite and quartz diorite of the Antwerp-Rossie suite.

Optional Stop A: 44.4051°, -75.6953°
Antwerp-Rossie suite diorite at Split Rock road.
This outcrop is typical of the more mafic phases of the Antwerp-Rossie suite, is foliated, and is cross-cut by felsic pegmatites. This outcrop rock was dated using SHRIMP-RG at 1203±13 Ma.
Stop 2: 44.4169°, -75.6865°
Biotite-sillimanite schist cut by leucosome on County Road 3.

This pavement outcrop is located just to the east of CR 3 on the dirt drive that parallels the county road (but is erroneously marked as CR 3 on Google Maps). Foliation in this area strikes north to northwest and this outcrop is on the western limb of a large-scale fold (the Alamgin Antiform of Baird and Shready, 2011). Tourmaline-bearing leucosomes here and elsewhere are seen to clearly cross-cut layering in biotite-sillimanite schists. Baird and Shready (2011) present U-Pb SHRIMP-RG concordia upper intercepts of ca. 1160 Ma for a similar tourmaline-bearing leucosome nearby that cross-cuts layering and also is involved in early folding, and interpret the leucosomes as being contemporaneous with the earliest regional deformation event (their D1).

Portion of the Hammond 15’ quadrangle (Buddington, 1934) showing field trip stops 4 to 6 from Day 1. Gra= Granitic gneisses of the Hyde School Gneiss. Sy and Syam= Syenitic rocks of the Edwardsville Pluton.

Stop 3: 44.3822°, -75.6529°
Antwerp-Rossie suite granodiorite north of Rossie.

This outcrop of foliated granodiorite has been dated by ID-TIMS at 1183 ± 7 Ma (McLelland et al., 1992) and by SHRIMP at 1207±26/−11 Ma (Wasteneys et al., 1999). This granodiorite is typical of this bimodal suite of deformed granitic and granodioritic plutonic rocks are the oldest dated rocks of the Adirondack Lowlands (ca 1203±13 Ma by SHRIMP-RG).
Stop 4: 44.4429°, -75.546°  
**Hyde School Gneiss on California Road**  
Folded and boudinaged amphibolite layering is well exposed in this locality of alaskitic Hyde School Gneiss (HSG), and this field trip description is taken from deLorraine and Carl (1993). This body is zoned both compositionally and with respect to abundance of biotite amphibolite layers, which was taken by deLorraine, Carl, and others are representing metamorphosed stratigraphy in supracrustal volcanic unit. These authors also argue that the stratigraphic position of the Hyde School Gneiss bodies within Lowlands stratigraphy and the general lack of xenoliths are evidence against a plutonic origin.

![Structural map of the Hyde School body from Carl and Van Diver (1975), showing orientation and character of amphibolite layers.](image)
Stop 5. 44.4882° -75.5811°
**Mafic lithologies of the Edwardsville Pluton on Route 58.**

A variety of gabbroic to syenitic lithologies of the Edwardsville Pluton are well exposed here, showing mafic enclaves and country rock xenoliths. At this locality fine-grained equigranular melanocratic syenite is cross-cut by coarser and locally porphyritic melanocratic syenite. These rocks appear to be related on Harker diagrams (Fig. 19) and yield a SHRIMP-RG age of 1149±21 Ma, which we interpret to be the age of the Edwardsville Pluton. The δ¹⁸O of these zircons is 9.06‰, which is high given the mafic character of these rocks and is more akin to the high oxygen isotope ratios of the 1.15 Ga Frontenac suite in the Frontenac terrane than the lower values in the Adirondacks (Fig. 2).

Stop 6. 44.4981°, -75.5825°
**Contact of the Edwardsville Pluton on Route 58.**

This outcrop exposes the contact between pink syenite of the Edwardsville Pluton and the metasedimentary screen which separates the mafic and felsic phases of the pluton. ID-TIMS dating of zircon from this locality (McLelland et al., 1993) yielded 1164±4 Ma, within error of our new age of the mafic phase of the pluton. We interpret the geochemistry of this unit as being comagmatic with the mafic phases of the pluton and correlative with the Frontenac suite in the Frontenac terrane (Figs. 16-19). The Edwardsville pluton is on the eastern side of the Black Lake Shear Zone, but the δ¹⁸O of these zircons is 11.11‰ which links it to the high-δ¹⁸O Frontenac suite plutons in the Frontenac terrane to the west. This may indicate that the Edwardsville pluton is still derived from the same basement as the high-δ¹⁸O plutons in Ontario in the subsurface (Peck et al., 2004).
Stop 6. 44.5063°, -75.5864°
Contact of the Fish Creek body of the Hyde School gneiss on Route 58.

This outcrop description is from McLelland (1993). Roadcuts along Rt. 58 expose subequal granitic and tonalitic facies of the Fish Creek pluton of Hyde School Gneiss and its contacts with marble, calcsilicate, and amphibolite. Amphibolite layers (commonly discontinuous are common with HSG), and the larger, coarser examples show subophitic relationships with plagioclase attesting to a plutonic origin. Lowlands metasediments are cross-cut by the 1203 Antwerp-Rossie suite; this indicates that the ca 1172 Ma HSG is intrusive into this package as well. Titanite in calcsilicate at this stop yields ages of 1150-1130 Ma (Mezger et al., 1991) and recent Ar-Ar dating demonstrates that the Lowlands remained below 500°C since 1150 Ma. The SE portion of the roadcut is dominated by tonalite and a variety of intrusive features can be found. Also visible is the spalling off of layers and slices from amphibolite as the tonalite intruded its way up along magma conduits. The NW section of the roadcut consists mainly of hypersolvus granite and alaskite that form a deformed intrusion breccia with amphibolite and calcsilicate.

Portion of the Brier Hill 15’ quadrangle (Dietrich, 1957) showing the last field trip stop of Day 1. Pa= Alaskitic gneisses of the Fish Creek body of Hyde School Gneiss. Ppb is the tonalitic border facies of the Fish Creek body, and red bars are amphibolite layers.
Stop 1. 44.4751°, -75.7236°
**Deformed Rockport granite and metasediments on County Road 3**

At this outcrop, Rockport granite has intruded into quartzite country rock. Although the Rockport granite here is only weakly deformed, granitic dikes of similar composition that intrude the quartzite are tightly to isoclinally folded in this area, documenting intense shortening. Although fold hinges themselves are rarely exposed and are difficult to measure, stereonet plots of poles to dikes fall along a well-defined girdle that defines an estimated fold axis trend and plunge of N45E/23°. Together with the dominant steeply NW/SE dipping foliation in the area, the folds suggest that the finite horizontal shortening direction was NW/SE in this area. Folds exhibit thinning in the limbs and thickening in the hinges, which indicates deformation and folding at relatively high temperatures. Folds are often asymmetric along this outcrop and the sense of asymmetry typically reverses around fold hinges, possibly due to flexural flow folding. Similar folded dikes are common in and around the Black Lake shear zone. Zircon in granite from this outcrop was dated by ID-TIMS at 1155±15 Ma, while monazite (interpreted as metamorphic) has an age of 1137±1 Ma (McLelland et al., 1993).
Poles to dikes in the North Hammond region. The poles fall along a well-defined girdle that define an estimated fold axis with a trend and plunge of N45E/23°.

Portion of the Hammond 15’ quadrangle (Buddington, 1934) showing field trip stops 1 and 2 from Day 2. Gra= Rockport Granitic gneisses. Country rock to the granites are not shown in this area.

Stop 2. 44.4614°, -75.7108°
Deformed Rockport granite dikes on County Road 3

This outcrop is similar in basic structure to the previous one in that dikes likely associated with the Rockport granite are variably deformed, although here, dikes intrude into calc-silicate wall rock. Most of the dikes here are strongly deformed with a clear foliation that strikes NE and is sub-vertical. Some dikes exhibit pronounced strain
gradients on the outcrop scale, suggesting that strain was highly heterogenous within the area.

Although the foliation in both host calc-silicate and the dike are obvious, it is difficult to see a lineation, suggesting that strain was dominated by flattening. However, anisotropy of magnetic susceptibility (AMS) analyses from this and other areas within the Black Lake shear zone show that the axis of maximum susceptibility, which should be a proxy for the lineation direction in strained rocks, is consistently sub-horizontal and trends NE, which is consistent with rare lineations found at other outcrops in the BLSZ. These results suggest that sub-horizontal shear may be an important but cryptic aspect of strain in this area.

U-Pb SHRIMP dating of titanite from one highly strained dike at this locality yielded concordant to very slightly discordant ages with a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date of 1163 ± 26 Ma (Wong et al., 2011). Although the uncertainties are large, this mean age overlaps with the age of the Rockport Granite/Hyde School Gneiss suite within error and therefore suggests that these dikes are related to the Rockport Granite. This result demonstrates that significant deformation post-dated intrusion of the Rockport suite.

AMS and geochronologic data from a strained dike at Stop 2 in the North Hammond area. AMS data (left) shows that the axis of maximum susceptibility ($K_{\text{max}}$, shown as white boxes) is sub-horizontal, suggesting that there is a sub-horizontal lineation here that is difficult to observe in outcrop. Geochronologic data (right) on titanite from this strained dike yields concordant to slightly discordant ages that yield a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1163 ± 26 Ma.

**Stop 3. 44.4061°, -75.7933°**

**Deformed Rockport granite dikes and intrusive relationships on Route 12**

This outcrop has a similar setting to the previous outcrops in North Hammond, except that clear lineations and kinematic indicators are visible in outcrop within dikes at this location. Lineations that are obvious in the field are relatively rare within the Black Lake shear zone, so this is an important locality for understanding the kinematic history
of deformation in this area. The country rock includes both mica schist and calc-silicate, which are intruded by Rockport granite and associated dikes. Several highly deformed dikes dip moderately southeast to south and have clear down-dip lineations. Kinematic indicators within these dikes include sigma-type asymmetric porphyroclasts and S-C fabrics and consistently indicate a top-NW or thrust sense of deformation. This deformation was probably not confined only to the dikes but was likely also accommodated throughout the country rock. However, the metasedimentary country rock is older and probably has experienced multiple phases of deformation. As a result of this poly-phase strain history in the metasedimentary units, it is likely that lineations formed during this event may only be preserved in the younger dikes that experienced a simpler strain history.

It is important to note that these lineations are nearly orthogonal to sub-horizontal lineations implied by AMS results at the previous stop and observed in other outcrops. This raises important questions regarding the kinematic history of the region. It may be that this suggests a transpressional setting or alternatively, two distinct events that include NW-directed shortening and sub-horizontal shear may have occurred. Both lineation directions seem to affect dikes associated with Rockport granite, suggesting that some of this strain post-dates intrusion of the Rockport at ca. 1170 Ma.
Portion of the Alexandria Bay 15’ quadrangle (Cushing et al., 1910) showing field trip stops 3 to 5 from Day 2. Yellow= Older phases of Rockport Granitic gneisses, Green= younger phases of Rockport granite. Map units with lens pattern denotes abundant inclusions.

**Stop 4. 44.3451°, -75.8766°**

**Unconformity on Route 12**

This outcrop displays a major unconformity between the Proterozoic basement rocks and the overlying Cambrian Potsdam sandstone. The unconformity surface here undulates subtly, indicating some topography on the Cambrian surface during Potsdam deposition. The basement rocks here include calc-silicate and schist and are intruded by variably deformed dikes that are likely of Rockport age. Foliation in both the metasedimentary rocks and the dikes dips steeply northwest.

**Stop 5. 44.3124°, -75.9978°**

**Rockport granite dikes intruding quartzite on Wellesley Island**

At this location, granite intrudes host quartzite. The age of the granite was initially determined to be 1415 ± 6 Ma using multi-grain ID-TIMS (McLelland et al., 1991), which was thought to be the oldest known age in the Adirondacks. This old age had important implications for both the age of the metasediments, which must have been older than 1415 Ma as well as the crustal structure of the Adirondack Lowlands, as this
would have required a major age discontinuity across the Saint Lawrence. Subsequent single grain TIMS analyses revealed that the granite here was \(1172 \pm 5\) Ma, and that the older age was the result of detrital and discordant juvenile zircons (Wasteneys et al., 1999). Therefore, this granite is coeval with the Rockport Granite/Hyde School gneiss suite.

In addition, this outcrop contains some interesting structural relationships. At least one of these Rockport granite dikes contains a moderately developed foliation in the dike interior that rotates towards the dike margin and merges with a high-strain south-southeast dipping foliation in the host quartzite (see figure). The host quartzite contains a weakly developed lineation along the margin of the dike that is sub-horizontal, again supporting a component of sub-horizontal shear that matches that AMS results. Given this, the rotation of the foliation suggests a left-lateral sense of shear, assuming a component of non-coaxial strain, although similar relationships could be produced by coaxial strain acting on a pre-existing fabric.

Foliation in the dike (left) gradually rotates and merges with a higher strain foliation along the dike margin and in the host quartzite (right) on Wellesley Island. The sense of this rotation suggests a left-lateral shear.
Appendix: Additional Geochemical Data

Figure A1. Major element geochemistry of pelitic gneisses and schists from the Hammond area, Adirondack Lowlands (Russell, 2009). Samples showing eye-schist textures (Buddington, 1934) are filled circles, samples lacking this texture are open circles. From this limited sample it appears that eye-schists are distinguished by low Ca and high Si.

Figure A2. Spider and rare earth element plots of the eye schists (filled circles) and other pelitic gneisses and schists (open) normalized to North American Shale Composite (Gromet et al., 1984).
Figure A3. Major element geochemistry of Rockport granite (triangles) and dikes interpreted as being related to Rockport granite in the structural study by Wong et al. (2011). Geochemical data from Carl and deLorraine (1997), Catalano (2009) and Hochman (2009).