

Deformation processes in a mid-crustal strike-slip shear zone: Insights from the Archean Quetico Shear Zone, Superior Province, Canada

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Abstract

The Quetico Shear Zone serves as an excellent analogue for active strike-slip shear zones at seismogenic depths. The Quetico Shear Zone is a right-lateral, strike-slip shear zone located at the boundary between the Wabigoon and Quetico subprovinces and has a strike length of at least 400 km. This research focuses on the eastern extent of the shear zone, north of Thunder Bay, and aims to constrain the kinematics, structures, conditions, and timing of deformation processes within the shear zone and adjacent paleo-earthquake surfaces. We found that Quetico Shear Zone deformation is characterized by increased mylonitization and brittle deformation with increasing proximity to the shear zone trace (within 500 m) where paleo-earthquake surfaces (i.e., pseudotachylite veins) were found. Mylonitization produces recrystallized quartz ribbons and a strong foliation unique to the Quetico Shear Zone (stronger than regional Quetico subprovince transpressional structures), particularly in granitic units. Non-granitic rock units within the core of the shear zone display pervasive brittle deformation with numerous faults. The recrystallized quartz grain sizes do not correlate with increased mylonitization and proximity to the shear zone. Recrystallized quartz grain sizes remained constant within error, with a median stress value of 80 MPa. The temperatures of quartz recrystallization also show no clear evolution with increasing proximity to the Quetico Shear Zone trace, with a median value of 487°C. The quartz recrystallization temperatures correlate to retrogressive temperatures from stable mineral assemblages, which are lower than estimated peak metamorphic temperatures from stable mineral assemblages (up to 700°C). The slip systems within the samples show an overprint from a higher-temperature prism $\langle a \rangle$ slip distal to the shear zone core to a lower-temperature rhomb $\langle a \rangle$ slip within the shear zone core. The calculated strain rates for the shear zone ($3.9\text{E-}10$ to $9.5\text{E-}12 \text{ s}^{-1}$) are comparable to those in active shear zones. Apatite and titanite provided the best

ages for deformation, mainly producing interpreted ages younger than the Quetico subprovince metamorphism. The interpreted Quetico Shear Zone deformation ages are 2618 to 2596 Ma. The exhumed Quetico Shear Zone appears to consist of an increasingly localized zone of deformation with proximity to the shear zone trace, though the shear zone records constant stress conditions across the shear zone core, indicating another weakening mechanism (e.g., fluid infiltration or fabric intensity) may be localizing deformation. The timing of this event is shortly after the Kenoran orogeny and may be related to the Sacred Heart Orogeny in the Minnesota River Valley.

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Data Accessibility

Data associated with this study are within an Open Science Framework repository at DOI 10.17605/OSF.IO/YXZ9E. Field data from the field stations can be found at www.strabospot.org under the TBN Intrusive Complex project.

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1. Introduction

Most shallow, large earthquakes nucleate within the brittle-ductile transition zone of the Earth's crust, where high stresses are required to overcome the rock strength (Sibson, 1984). Shear zones are tabular, planar domains of concentrated deformation that form due to strain localization. Strain localization is thought to result from weakening of the lithosphere. There are many potential causes for weakening, such as mineral reactions, fabrics (e.g., mineral habit and crystallographic preferred orientations), presence of hydrous fluids, and/or changes in deformation mechanisms due to grain size reduction (Poirier, 1980; Rutter, 1999; Jefferies et al., 2006; Platt and Behr, 2011). Strain localization results in the formation of faults and shear zones, which allows for maintaining weakness from the same processes responsible for their formation or new processes (Holyoke and Tullis, 2006). The conditions surrounding strain localization for active shear zones are poorly constrained, especially proximal to earthquake surfaces.

The brittle-ductile transition zone, also known as the frictional-plastic transition, occurs within the crust at depths of approximately 10–15 kms for quartz-dominated rocks (Byerlee, 1968; Tullis and Yund, 1992; Gleason and Tullis, 1995). The structures produced at the brittle-ductile transition zone in active shear zones are challenging to study as they form at inaccessible depths (the deepest drill hole through an active fault intercepted the fault at a ~2.7 km depth; Zoback et al., 2010). An approach to better understanding shear zone structures at depth is to study ancient, exhumed shear zones, where uplift and erosion have brought ancient shear zone rocks to the surface. Studying exhumed shear zones allows for constraining the conditions under which large earthquakes occur (Chester and Chester, 1998; Faulkner et al., 2003, 2010).

This study focuses on the exhumed, Neoproterozoic Quetico Shear Zone as an analogue structure for active strike-slip shear zones at the brittle-ductile transition due to its erosional depths

of 10–15 km (Percival et al., 2012). The E-W striking, and subvertical Quetico Shear Zone has a strike length of ca. 400 km and a reported displacement of at least 150 km (Mackasey et al., 1974; Kennedy, 1984). The shear zone forms a localized structure exhibiting pseudotachylites (i.e., ancient, fossilized earthquakes) within the shear zone core. The Quetico Shear Zone, therefore, serves as an excellent location to test the deformation conditions surrounding the ancient pseudotachylite veins to better understand how strain localization and earthquakes were accommodated within the core of the shear zone.

Ore deposits are commonly related to shear zones, especially within the Superior Province (Richards and Tosdal, 2001). These structures can provide pathways for magmas and fluids to travel great distances through the crust (Rosenberg, 2004, and references therein; Faulkner et al., 2010). The Quetico Shear Zone is associated with several mineral deposit types, including orogenic gold, rare-element pegmatite, and mafic-ultramafic hosted Cu-Ni-PGE deposits. Some of these deposits, such as the ~1.1 Ga Thunder Bay North Ni-Cu-PGE deposits, are much younger than shear zone formation, but still utilized the structure as a pathway for magma ascension (Bleeker, 2020). Other deposits, such as orogenic gold deposits, are thought to have formed syn-kinematically, when deformation provides permeability, especially along dilational jogs, for metal-enriched fluid flow (Kerrick, 1989; Faulkner et al., 2010; Stephan et al., 2025). The timing of deformation of the Quetico Shear Zone is poorly constrained, with only one date from a late pseudotachylite vein indicating ~1.9 Ga activation (Peterman and Day, 1989). Because most orogenic gold deposits in the southwestern Superior Province formed during the Kenoran orogeny under transpressional conditions, post-Kenoran deformation along the Quetico Shear Zone potentially could have led to remobilization of gold deposits. Determining the timing of shear zone

deformation is therefore critical for constraining Archean Superior Province tectonics as well as understanding the potential for mineral deposit formation along the shear zone.

In this study, the rocks surrounding the Quetico Shear Zone were examined through detailed macro- to microstructural analysis to determine the extent of the shear zone deformation and to quantify conditions and timing of deformation. Stresses and temperatures were obtained from dynamically recrystallized quartz using electron backscatter diffraction (EBSD), pseudosection modelling, and Ti-in-quartz thermometry from samples proximal and distal to the shear zone. Empirical quartz flow laws were applied to determine strain rates from the shear zone. The timing of shear zone deformation was derived from petrochronology using scanning electron microscopy- energy-dispersive X-ray spectroscopy (SEM-EDS) imaging and laser ablation split stream inductively coupled plasma mass spectrometry (LASS-ICP-MS) geochronology. Together, these datasets provide a new framework for understanding the timing and conditions of deformation within an important but understudied shear zone within the Superior Province.

2. Regional Geology

2.1 Superior Province

The Superior Province, located within the Precambrian Canadian Shield, has been relatively stable for the last 2.6 billion years and is bound on all sides by younger orogenic events (Stott, 1997; Percival et al., 2012). It is bound to the north and west by the Trans-Hudson orogeny, to the southeast by the Grenville orogeny, and to the south by the Penokean orogeny (Card, 1990). The Superior Province is subdivided into subprovinces (Card and Ciesielski, 1986), which are generally east-west trending, belt-like terranes (excluding the few northwest-southeast trending terranes in northern Quebec) with distinct geological histories (Fig. 2.1; Stott, 1997). There are various subdivisions of the Superior Province depending on the defining criteria. Percival et al. (2012) summarized the most common terminology. Terranes refer to regions that are tectonically bound with differing characteristics to adjacent regions before the Neoarchean assembly of the Superior Province. Domains refer to regions of differing lithological characteristics with a common basement. Superterranes refer to an amalgamation event before the Neoarchean assembly, and subprovince is a general term to describe a region with similar geological and geophysical characteristics (Fig. 2.1; Percival et al., 2012).

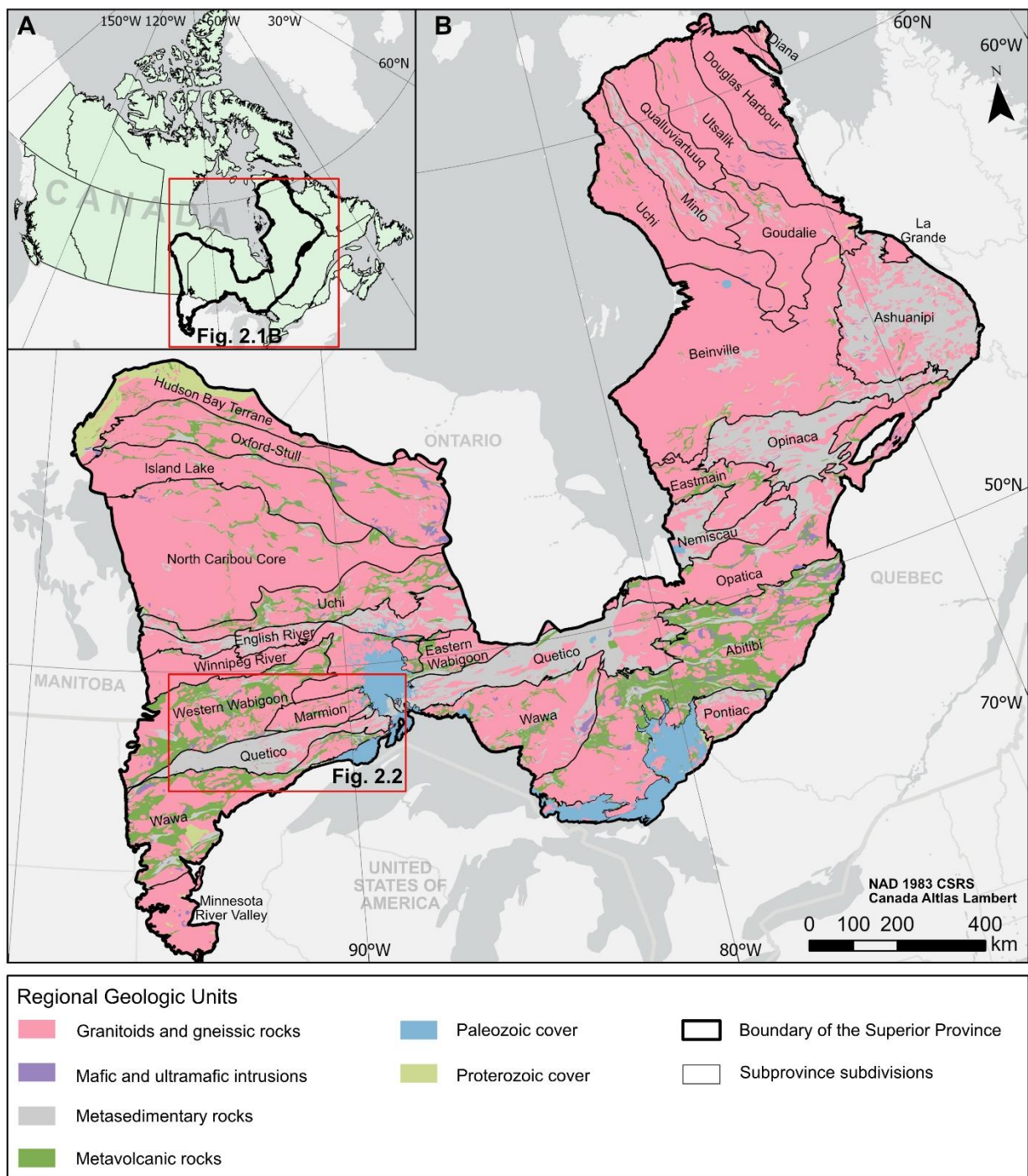


Figure 2.1: A: Location of the Superior Province within Canada and the USA. (Cartographic boundary file: Statistics Canada, 2021 Census - Boundary files) B: Geological map of the subprovinces within the Superior Province (*Geology*: after Montsion et al., 2018; *Grayscale basemap*: Esri, Tom Tom, Garmin, FAO, NOAA, USGS, © OpenStreetMap contributors, and the GIS User Community).

Subprovinces can be grouped into four main types based on their dominant lithology: volcano-plutonic, metasedimentary, plutonic, and high-grade gneiss (Card, 1990). The volcano-plutonic subprovinces consist dominantly of greenstone belts with syn-volcanic, syn-orogenic granitoid plutons (Stott, 1997). The metasedimentary subprovinces consist dominantly of greywacke, migmatite, and peraluminous granite (Stott, 1997). The plutonic subprovinces consist dominantly of granitoid batholiths derived from partial melting and lower crustal source rocks (Stott, 1997). The high-grade gneiss subprovinces consist dominantly of upper amphibolite- and granulite-facies gneisses, likely continuous with adjacent subprovinces of supracrustal and plutonic origins (Card, 1990; Stott, 1997).

2.1.1 Tectonic Framework of the Superior Province

The processes involved in the formation of the Superior Province have been debated. The two main model types are fixist or mobilist, which are related to whether the rock formed by vertical or horizontal mechanisms (Blackburn, 1980; Card, 1990; Percival et al., 2012). Fixist models (Goodwin, 1977; Young, 1978; Thurston, 1985) assume vertically dominated tectonics, mainly driven by buoyancy forces and gravitational instabilities, involving pre-existing crust and the formation of supracrustal sequences in rifts or linear downwarps triggered in the crust by mantle processes, continued downwarping (i.e., sagduction), and partial melting (Card, 1990). Mobilist models (Goodwin and Ridler, 1970; Langford and Morin, 1976; Dimroth et al., 1983a, 1983b; Ludden et al., 1986; Sylvester et al., 1987; Percival and Williams, 1989) assume horizontally dominated tectonics involving mobile, largely oceanic crust within a modern-like plate tectonic framework with subduction and horizontal accretion mechanisms in convergent plate boundary settings (Talbot, 1973; Card, 1990).

Invoking mobilist models (Card, 1990; Stott, 1997; Percival et al., 2012), the western Superior Province formed in the Archean through a series of accretional sequences of protocontinental fragments and terranes termed the Kenoran Orogeny (Percival et al., 2012). The orogeny is subdivided into the northern 2.71 Ga Uchian phase and the southern 2.69 Ga Shebandowanian phase (Stott, 1997; Stott et al., 1989). The Uchian phase begins with the juxtaposition of the Hudson Bay and Winnipeg River terranes from the north and south, respectively, to the 2.9–3.0 Ga North Caribou terrane (Stott, 1997). The North Caribou terrane consists of the Oxford-Stall, Island Lake, North Caribou, and Uchi subprovinces (Fig. 2.1) and is recognized as the core of the western Superior Province (Stott et al., 2010). The English River subprovince (2.71 Ga) formed during the collision of the North Caribou and Winnipeg River terranes as volcanism ceased; greywacke deposition occurred, forming a synorogenic flysch basin (Percival et al., 2012). The western Wabigoon and Marmion terranes are juvenile terranes that accreted to the Winnipeg River terranes shortly before 2.7 Ga (2.70 Ga; Fralick and Davis, 1999 to 2.69 Ga; Sanborn-Barrie and Skulski, 2006; Percival et al., 2012; Bjorkman et al., 2024). The Shebandowanian phase of the Kenoran orogeny began with the collision of the Wawa subprovince with the Western Wabigoon and Marmion terranes (Stott, 1997). The collision formed the Quetico subprovince (similar to the formation mechanism for the English River subprovince; Percival et al., 2012). The last part of the Shebandowanian phase is marked by the collision of the older (3.5 Ga) Minnesota River Valley terrane with the Wawa subprovince (Percival et al., 2012).

2.2 The Quetico Subprovince

The Quetico subprovince is a metasedimentary belt that varies in width from 10 to 100 km (Card, 1990). The subprovince is bound by the Wabigoon, Marmion, and Wawa subprovinces (Card, 1990). One theory for the formation of the Quetico subprovince is an accretionary prism

from the docking of the Wawa subprovince to the Wabigoon subprovince (Percival and Williams, 1989). The Quetico subprovince consists of mixed detrital zircon populations, indicating the sedimentary rocks are younger than the majority of volcanic activity in the two adjacent subprovinces (Card, 1990). Davis et al. (1990) bracket the age of sedimentation for the Quetico subprovince between 2698 and 2688 Ma, as 2698 ± 3 Ma is the maximum age of sedimentation from greywacke detrital zircon grains and 2688 ± 4 Ma is the age of a late-tectonic intrusion, the Blablock intrusion, which cuts the sedimentary rocks. The dominant lithologies consist largely of turbiditic metasedimentary rock, where protoliths include thin to medium bedded arkosic and lithic wacke, siltstone, and minor conglomerate (Wood, 1980; Ojakangas, 1985; Card, 1990; Sawyer, 1998). The turbiditic sequences experienced increasing metamorphism towards the subprovince's central axis, resulting in schists, gneisses, and migmatites reaching peak metamorphism at 2664–2657 Ma in the western part of the belt (Borradaile, 1982; Rehm et al., 2026).

The Quetico subprovince consists of low to medium metamorphic grade margins to the north and south, and a high metamorphic grade interior (Card, 1990). The metamorphic grades of the rocks consistently increase away from the margins towards the interior; this is recorded by the metamorphic index minerals within the metasedimentary rocks (Sawyer, 1983). The low-grade rocks contain chlorite and biotite, the medium-grade rocks contain staurolite and cordierite, and the high-grade rocks contain sillimanite and orthopyroxene in metapelites (Sawyer and Barnes, 1988; Rehm et al., 2026). The Quetico subprovince consists of a prevalent east-west trending structural fabric forming upright folds (Card, 1990). Early folding is thought to be overprinted by the pervasive foliation and refolding events about shallow-plunging folds during and after migmatization (Percival and Williams, 1989). The prominent north-south shortening was followed by dextral transpression that resulted in dextral motion along the major boundary faults and shear

zones (Card, 1990; Stephan et al., 2025). The most notable of these major right-lateral strike-slip boundaries is the Quetico Shear Zone (Fig. 2.2; Kennedy, 1984).

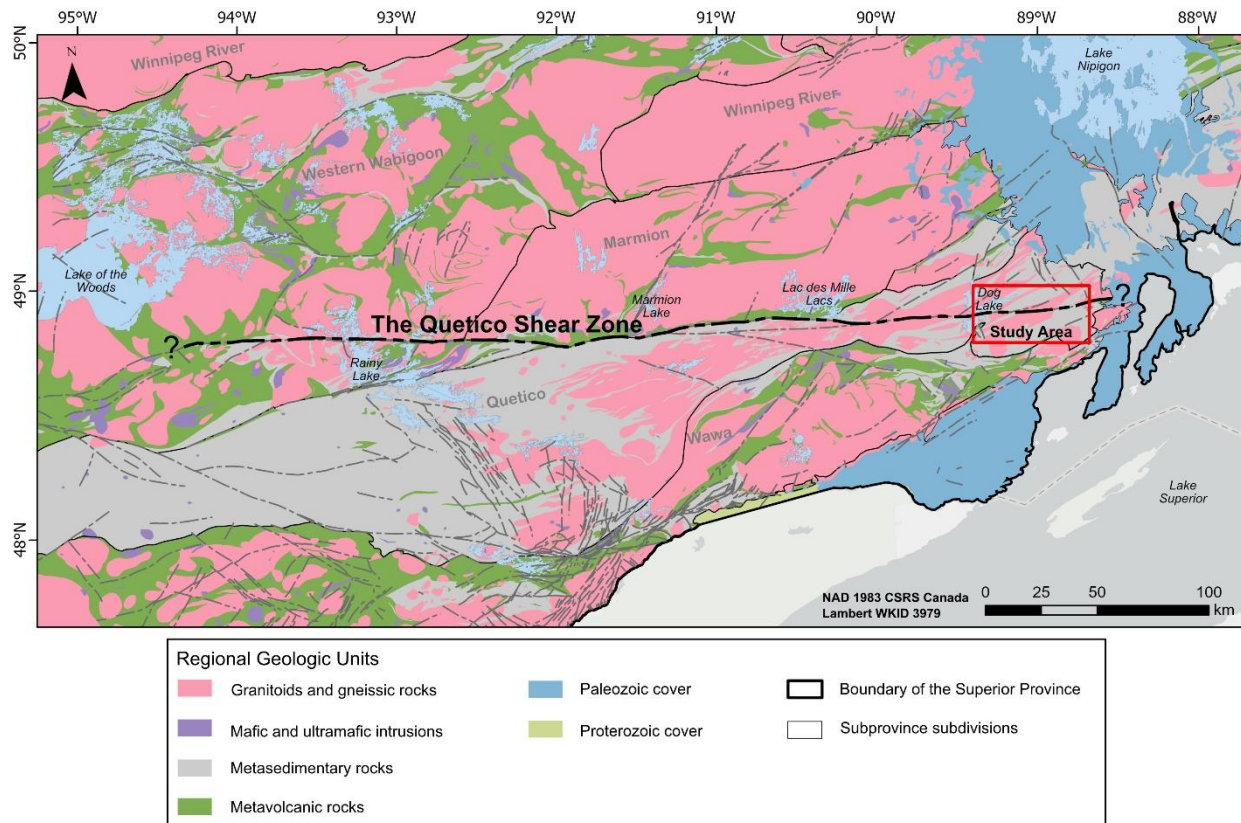


Figure 2.2: Geological map of the Quetico Shear Zone trace. The red box indicates the study area for this project. The question marks indicate the extent of the shear zone trace. The subprovince maps are outlined in grey. (*Geology*: after Montsion et al., 2018; *Lakes*: Ontario Ministry of Natural Resources and Forestry (MNRF), Ontario Hydro Network (OHN)-Watercourse; *Grayscale basemap*: Esri, Tom Tom, Garmin, FAO, NOAA, USGS, © OpenStreetMap contributors, and the GIS User Community).

2.3 The Quetico Shear Zone

The Quetico Shear Zone is a major, dextral, transcurrent shear zone located within the Wabigoon and Quetico subprovinces (Kennedy, 1984). Although the shear zone is often referred to as the subprovince boundary between the Quetico and Wabigoon subprovinces, it crosscuts the latter (Fig. 2.2), implying that the shear zone postdates the province assemblage (Borradaile, 1982; Kennedy, 1984). The Quetico Shear Zone has a strike length of at least 400 km, as younger sedimentary units cover its western and eastern extents (Kennedy, 1984). The width of the shear

zone is suggested to vary significantly along strike, with approximate widths ranging from less than 10 m to 1200 m (Kennedy, 1984). The shear zone width is marked by increasing foliation intensity with proximity to the shear zone core (Kennedy, 1984). The shear zone is delineated by linear topographic lows, often narrow lakes, and is associated with thin magnetic highs visible on aeromagnetic maps (Kennedy, 1984). The fault is exposed from Rainy River in the west to Dorion in the east (Fig. 2.2).

The first report of the Quetico Shear Zone emerged from the Rainy River area, where Lawson (1913, 1888) described a porphyroid gneiss unit with a linear belt of sheared granitic rock. Kennedy (1984) reported the most recent in-depth study of the shear zone, including a summary of previous findings as well as new results for the entire trace of the shear zone. Various reports of fault rocks have been documented along the strike of the fault, featuring cataclastic to mylonitic textures across different lithological units, indicating a strong presence of both brittle and ductile deformation related to the shear zone (Kennedy, 1984). The shear zone consists of steeply dipping fabrics and shallow plunging E-W trending lineations reflecting its strike-slip nature (Kennedy, 1984). The kinematics of the shear zone, especially the amount of slip, can be difficult to determine without clear markers (Kennedy, 1984). The sense of shear is largely dextral based on numerous studies reporting dextral shear sense indicators within the shear zone (Hawley, 1929; Shklanka, 1972; Davies, 1973; Harris, 1974; Mackasey et al., 1974; Pirie, 1978; Schwerdtner et al., 1979; Kennedy, 1984). A few studies have reported sinistral shear sense (Kaye, 1967) and vertical motion (Hawley, 1929; Mackasey et al., 1974; Pirie, 1978) within the shear zone. For the offset of the shear zone, Mackasey and others (1974) suggested that the Quetico-Wabigoon stratigraphic boundary is displaced by the shear zone from Mine Center to the Lac des Mille Lacs-Kashabowie Lake area, giving an approximate offset of 100 km.

The types of fault rocks commonly reported surrounding the Quetico Shear Zone include mylonite, phyllonite, cataclasite, and pseudotachylite-bearing rocks (Kennedy, 1984). The mylonitic rocks range from protomylonites with weakly elongated quartz and fractured feldspar porphyroclasts to ultramylonites with strong crystallographic preferred orientations (Kennedy, 1984). The phyllonites consist of localized zones of fragmentary rock with random fabrics and angular fragments of minerals (Kennedy, 1984). Pseudotachylite veins are often parallel to pre-existing mylonitic schistosity or at high angles in fractures and are often altered to sericite and chlorite (Kennedy, 1984). Some pseudotachylite occurrences show evidence of flow banding in thin section (Borradaile and Kennedy, 1982). The lithologies across the trace of the shear zone vary from metasedimentary rocks (of various metamorphic grades) to metavolcanic rocks, to granitic and gneissic rocks (Kennedy, 1984). Structures present within the shear zone consist of microfaults, small ductile shear zones/shear bands, brittle shear zones, and folding of mylonitic banding (Borradaile, 1982; Kennedy, 1984). The small shear bands and brittle shear zones often trend NE and NW, forming conjugate shear fractures (Kennedy, 1984). Subhorizontal slickensides on microfault surfaces are common, often composed of phyllosilicates, and contain slickengrooves indicating a dextral shear sense (Kennedy 1984).

2.4 Timing of shear zone deformation

The timing of deformation for the Quetico Shear Zone is poorly constrained; it is suggested to have a long, complex history with many episodes of deformation from pre-Kenoran to Proterozoic (Kennedy, 1984). The only date comes from two pseudotachylite veins hosted in biotite schist from the Quetico Shear Zone and the Seine River Fault (Peterman and Day, 1989). The pseudotachylite veins were analyzed along with the host rock using Rb-Sr ($^{87}\text{Rb}/^{86}\text{Sr}$), which resulted in an age of 1947 ± 23 Ma, which is interpreted to be a reactivation age for both fault

systems within the Early Proterozoic (Peterman and Day, 1989). The country rocks that host the pseudotachylite veins for the Seine River Fault and Quetico Shear Zone range from 2686-2728 Ma (Davis et al., 1989), with evidence for uplift at approximately 2.5 Ga (Peterman and Day, 1989). This reactivation of major transcurrent fault systems within the Superior Province is suggested to be related to far-field tectonic events (Peterman and Day, 1989; Kamineni et al., 1990; Percival and Peterman, 1994; Percival et al., 2012). The Penokean orogeny (1.89–1.83 Ga; Sims et al., 1989) to the south is often considered to be responsible. Peterman and Day (1989) ruled this event out as it is too young. The Kapuskasing uplift (1.9 Ga) is also suggested as it is thought to be responsible for rotating the western Superior Province by 20 degrees counterclockwise (Halls and Davis, 2004; Evans and Halls, 2010). Peterman and Day (1989) suggest that the most likely events (based on their ages from pseudotachylite veins) are the Trans-Hudson orogeny (1.8 Ga; Ansdell, 2005) or the New Quebec orogeny (2.2-1.77 Ga; Wardle et al., 2002).

Deformation before Proterozoic reactivation is difficult to constrain, as it is intertwined with the orogenesis of the Late Archean. As the shear zone marks the boundary between the Quetico and Wabigoon subprovinces, it has been suggested that the shear zone formed as a paleostructure during the early formation of the Superior Province (Kennedy, 1984). Much of the strike-slip activity is suggested to be either at the close of the Kenoran orogeny or to post-date it (Borradaile, 1982; Kennedy, 1984). This late orogenic activity is thought to be linked to the major transcurrent faulting of the southwestern Superior Province, ranging from 2689-2684 Ma based on U-Pb dating from the Shebandowan-Quetico subprovince boundary (Corfu and Stott, 1986; Peterman and Day, 1989). The Rb-Sr ages from biotite in the country rocks within the shear zone show slightly younger cooling ages (2482-2534 Ma; Peterman and Day, 1989). These ages

indicated the final cooling during the end of cratonization in the Late Archean-Early Proterozoic, long before the proposed reactivation event (Peterman and Day, 1989).

Table 2.1: Regional Geology and Deformation Events within the Superior Province relevant to the Quetico Shear Zone

Age (Ga)	Event
1.8 Ga	TransHudson orogeny (11)
1.89-1.83 Ga	Penokean orogeny (10)
1.9 Ga	Kapuskasing uplift (8) (9)
1.94 Ga	Quetico shear zone pseudotachylite veins (7)
2.2-1.77 Ga	New Quebec orogeny (12)
2.53-2.48 Ga	Biotite country rock cooling ages (7)
2.5 Ga	Significant southern Superior Province uplift (7)
2.68 Ga	Major transcurrent faulting in the southwestern Superior Province (13) (7)
2.68 Ga	Minnesota River terrane collides with Wawa subprovince (2)
2.69 Ga	Quetico subprovince forms (2) (2698 to 2688 Ma sedimentation (6))
2.69 Ga	Shebandowanian phase (Kenoran Orogeny) begins with collision of Wawa with Western Wabigoon and Marmion terranes (1)
2.70 Ga	Accretion of juvenile Wabigoon and Marmion terranes to Winnipeg River terrane (2) (3) (4) (5)
2.71 Ga	Uchian phase (Kenoran Orogeny) (1)
2.71 Ga	English River subprovince forms from the collision of North Caribou and Winnipeg River terranes (2)
2.9-3.0 Ga	Hudson Bay and Winnipeg River terranes juxtaposition to North Caribou terrane (1)
3.5 Ga	Minnesota River terrane (2)

References: (1) Stott, 1997; (2) Percival et al., 2012; (3) Fralick and Davis, 1999; (4) Sanborn-Barrie and Skulski, 2006; (5) Bjorkman et al., 2024; (6) Davis et al., 1990; (7) Peterman and Day, 1989; (8) Halls and Davis, 2004; (9) Evans and Halls, 2010; (10) Sims et al., 1989; (11) Ansdell, 2005; (12) Wardle et al. 2002; (13) Corfu and Stott, 1986.

3. Methodology

3.1 Geological Field Mapping

Within the study area, three main north-south trending, roughly 10 km long transects were conducted across the Quetico Shear Zone. Field station spacing was selected based on distance from the shear zone, with increasing station density within 1 km of the shear zone trace. Field mapping focused on: 1) the evolution of rock fabrics with increasing proximity to the shear zone to quantify shear zone width, 2) the orientation and kinematics of structures to quantify changes in the kinematics of deformation with proximity to the shear zone trace, and 3) the metamorphic mineral assemblages present to differentiate regional deformation and metamorphism from Quetico Shear Zone-related deformation and metamorphism. Lithological, structural, and metamorphic characteristics were recorded using StraboSpot (Walker et al., 2019). Structural orientation data were plotted and analyzed using the R package “structr: Structural Geology in R” (Stephan, 2025), <https://tobiste.github.io/structr/>. Forty-three oriented field samples were collected from selected locations along the transects for microstructural analysis.

3.2 Drill Core Logging

Diamond drill core from Clean Air Metals’ Thunder Bay North Intrusive Complex project was examined. The drill core was selected from drill holes that intersect the Quetico Shear Zone. The drill holes were drilled at a near-vertical orientation, resulting in a shallow angle between the core axis and the shear zone fabrics. The drill core was logged, focusing on lithology, structure, and metamorphism within the shear zone along small intervals, approximately every 4 m. Ninety-three drill core samples were collected, one from each 4 m interval, along with additional samples if there was a change in the structural characteristics between intervals.

3.3 Optical Microscopy

Sixty-three collected samples were prepared for polished thin sections at Lakehead University's Lapidary Facility for microstructural analysis. Forty-six of the thin sections are from oriented field samples, and the other 17 are from unoriented diamond drill core samples. The thin sections were prepared along the XZ-plane of strain, which is interpreted to be perpendicular to the foliation and parallel to the lineation. Full thin-section photomosaics were produced using a Zeiss Axioscope 5. Petrographic analysis was conducted using Olympus BX51 and BX53 microscopes.

3.4 Whole Rock Geothermobarometric Analysis

Samples were sent to ALS Canada Ltd. for whole rock geochemical analysis using the CCP-PKG01 package, where the major elements were analyzed by inductively coupled plasma atomic emission spectroscopy (ICP-AES) and trace elements were analyzed by inductively coupled plasma mass spectroscopy (ICP-MS) using lithium borate fusion digestion. The whole rock geochemistry was converted from weight percent oxides to moles.

Equilibrium phase diagrams of the samples and model compositions were calculated for the NaCaKFMASHTi system with the computer program THERIAK-DOMINO (de Capitani and Brown, 1987; de Capitani, 1994). The resulting diagrams are pseudosections, which display the stable mineral assemblages for a specific whole rock composition based on a closed system. The independent variables can be a combination of temperature, pressure, phase activity, or a compositional vector. To include highly non-ideal solution models for minerals with potential miscibility gaps, such as feldspar (Fuhrman and Lindsley, 1988), the method for computing the stable mineral assemblages must be a true Gibbs Free Energy minimization (Greenberg et al., 1985; de Capitani and Brown, 1987; Karpov et al., 1997; White et al., 2001). For details, the reader

is referred to the homepage of the THERIAK-DOMINO software at <http://titan.minpet.unibas.ch/minpet/theriak/theruser.html>.

Using the major element compositions (converted to moles), equilibrium phase diagrams were plotted for the NaCaKFMASHTi systems, using THERIAK-DOMINO to produce the stable mineral assemblages that minimize the Gibbs free energy within pressure and temperature space. Along with the major-elements geochemistry, we assumed a fluid saturated system by incorporating 80 moles of hydrogen to the starting parameters. The pseudosections were created using the `tcds55_p07` database for pelites (Holland and Powell, 1998). All generated phase equilibrium diagrams are in Appendix B.

3.5 Scanning Electron Microscopy

Scanning electron microscopy (SEM) was used for each of the following analyses:

3.5.1 Micro-orientation Analysis and Paleopiezometry

Electron backscatter diffraction (EBSD) analysis was conducted using a TESCAN Mira 4 Scanning Electron Microscope (SEM) at California State University, Northridge. EBSD was used to analyze the micro-orientations of quartz grains to compare to macro-orientations and to further characterize the shear zone deformation. EBSD was also used to measure paleo-stress of quartz grains using a method called paleopiezometry. Before analysis, the polished thin sections were additionally polished with colloidal silica using a Buehler Minimet for 2 hours. The samples were positioned at a working distance of 27 mm and a 65-degree tilt for optimal detection from the EBSD detector. The EBSD maps were collected at 1000x magnification with a step size of 1 μm . The EBSD data were post-processed using MTEX, a toolbox for MATLAB (Bachmann et al., 2010, 2011).

The microscopic orientations from the distributions of each quartz grain analyzed were plotted as orientation density functions (ODF), misorientation distribution function (MDF), and the crystallographic vorticity axis (CVA). The ODF plots display the orientation of the c [0001] and a $[11\bar{2}0]$ axes, as well as the crystallographic pole to the m plane $[10\bar{1}0]$ of the quartz grains. The MDF plots display the orientation of misorientation axes between quartz grains (10-15°) and subgrains (2-10°). The orientation of misorientation axes relates to the orientation of slip planes and slip directions where dislocations move due to high stresses (Neumann, 2000). The EBSD maps from all the domains are displayed as inverse pole figure maps with grain orientations colour coded relative to the Z-axis of the sample, or the normal to the sample surface (IPF-Z). The CVA plot displays the vorticity axis of each quartz grain. The ODF and CVA are plotted as stereographic projections into the lower hemisphere with the foliation oriented parallel to the x-direction and the lineation within the x-direction. The MDFs are plotted on an inverse pole figure.

For paleopiezometry analysis, relict and recrystallized quartz grains within EBSD maps were differentiated using the procedure outlined by Cross et al. (2017). The root mean square grain size of the recrystallized quartz grains was used for grain size piezometry using the piezometer calibrated by Cross et al. (2017).

3.5.2 Geothermometry

Ti-in-quartz thermometry using a Secondary ion mass spectrometry (SIMS) was conducted at the University of California, Los Angeles, using a CAMECA-ims 1290 (Liu et al., 2018). The analyses were conducted using a 10 kV and a 1 nA beam with a Hyperion-2 oxygen source, creating a spot size of approximately 5 μm . The analysis area was rastered with a 5x5 μm grid using a continuous beam to clean a $\sim 10 \mu\text{m}^2$ area before analysis. Mass calibration and beam centering at the 95% level were conducted on the $^{30}\text{Si}^{16}\text{O}$ (46 amu) peak before each analysis.

Twenty cycles were run for each analysis with 2 s/cycle on the $^{30}\text{Si}^{16}\text{O}$ peak (40 s total per analysis) and 6 s/cycle on the ^{49}Ti peak (120 s total per analysis). Ti-in-quartz standards consisted of various NIST glasses with Ti values of 0–1000 ppm (see Thomas et al., 2010) and three synthetic quartz samples (QTiP7, QTiP14, and QTiP38) with Ti values of 18–380 ppm (QTiP standards were synthesized by the University of Edinburgh following procedures of (Biggar and O’Hara, 1969) and Cairns (1983)). A suite of standards was run at the beginning of each day and again after every six hours of analysis (approximately) to verify the accuracy of the results and beam stability. Twenty to thirty spots were selected per sample, on dynamically recrystallized grains free of visible impurities and avoiding grain boundaries. An activity of 1 for Ti is assumed in geothermometry calculations when rutile is present (Ghent and Stout, 1984).

3.5.3 Geochronology Preparation

SEM analysis (coupled with energy-dispersive X-ray spectroscopy (EDS)) was used to select suitable, datable grains. Eighteen samples were analyzed using the automated material identification and classification system (AMICS) by Vidence Inc. across entire thin sections (Kostryzhev and Murphy, 2024). This produced phase maps to identify and locate datable minerals, such as allanite, apatite, monazite, titanite, and zircon. This method also characterized each of the grains' dimensions, areas, and shapes. Using these grain characteristics, suitable grains for each mineral were selected for geochronology based on the spot size for each phase and the grain's microstructural characteristics.

Prior to analysis, the selected grains were imaged with the SEM using a backscattered electron detector (BSE), an EDS detector, and a cathodoluminescence detector (CL) to provide detailed maps of the grains for spot selection during analysis. The additional imaging allowed for detailed characterization of entire grains to identify zonation and inclusions.

3.5.4 Geochronology

In-situ analyses of apatite, monazite, titanite, and zircon grains were conducted using laser ablation split-stream inductively coupled plasma-mass spectrometry (LASS-ICP-MS) at the University of California, Santa Barbara (Kylander-Clark et al., 2013). From polished thin sections, an ablation stream was produced by an Elemental Scientific 193 nm excimer laser and split between two separate mass spectrometers. A Nu Instruments Plasma HR-ES multicollector ICP-MS was used to analyze the U/Th-Pb isotopes, and an Agilent 8900 Triple Quad ICP-MS was used to analyze trace element concentrations. Before ablation, pre-ablation consisted of two cleaning shots with a diameter approximately 1.5 times greater than the spot size. Primary standards, 1-2 secondary standards, and NIST612 glass were run at the start and end of each run and following each 7-12 analyses during a run (see Table 3.1 for relevant setup parameters for each phase). Several (~4) NIST612 glass analyses were conducted between analyses in each phase to clean out the lines. Trace element mineral geochemistry was used to differentiate metamorphic and primary minerals. U/Th-Pb isotopes were used primarily within the shear zone in overprinting structures to target Quetico Shear Zone deformation ages.

Table 3.1: LASS-ICP-MS parameters used for analysis of each mineral, including primary and secondary standards (references for standards can be found in Appendix E).

Mineral	Spot Size (µm)	Burst (#)	Count	Frequency (Hz)	Primary Standards	Secondary Standards
Apatite	20	60		5	MAD-UCSB	Durango MRC-1 McClure
Monazite	8	45		3	Stern 44069	Trebilcock FC MAN
Titanite	25	60		5	MKED BLR	Y17 P57
Zircon	15	60		5	91500 Plešovice	OG-1 GJ-1

4. Results

4.1 Field Observations

4.1.1 Lithologies

4.1.1.1 Geological Field Mapping

The lithologies observed within the study area (Fig. 4.1) consist largely of sedimentary and granitic rocks with minor mafic intrusive and intermediate to mafic volcanic rocks. The majority of the rocks have been metamorphosed between greenschist and amphibolite facies. The lithologies have been subdivided based on protolith, then on degree of metamorphism and/or deformation. In general, the degree of metamorphism increases towards the center of the Quetico subprovince, north of the study area, and the deformation fabrics increase towards the trace of the Quetico Shear Zone. The samples are characterized by their distance from the mapped trace of the Quetico Shear Zone or the core of the shear zone (Fig. 4.1). The degree of deformation for the rocks and samples is based on foliation intensity. Foliation intensity is defined by the foliation spacing and continuity, with weak indicating poorly developed foliation (e.g., discontinuous or predominantly massive) or a wide spacing (>20 cm), and strong indicating there is a very tight foliation spacing (<5 cm to pervasive).

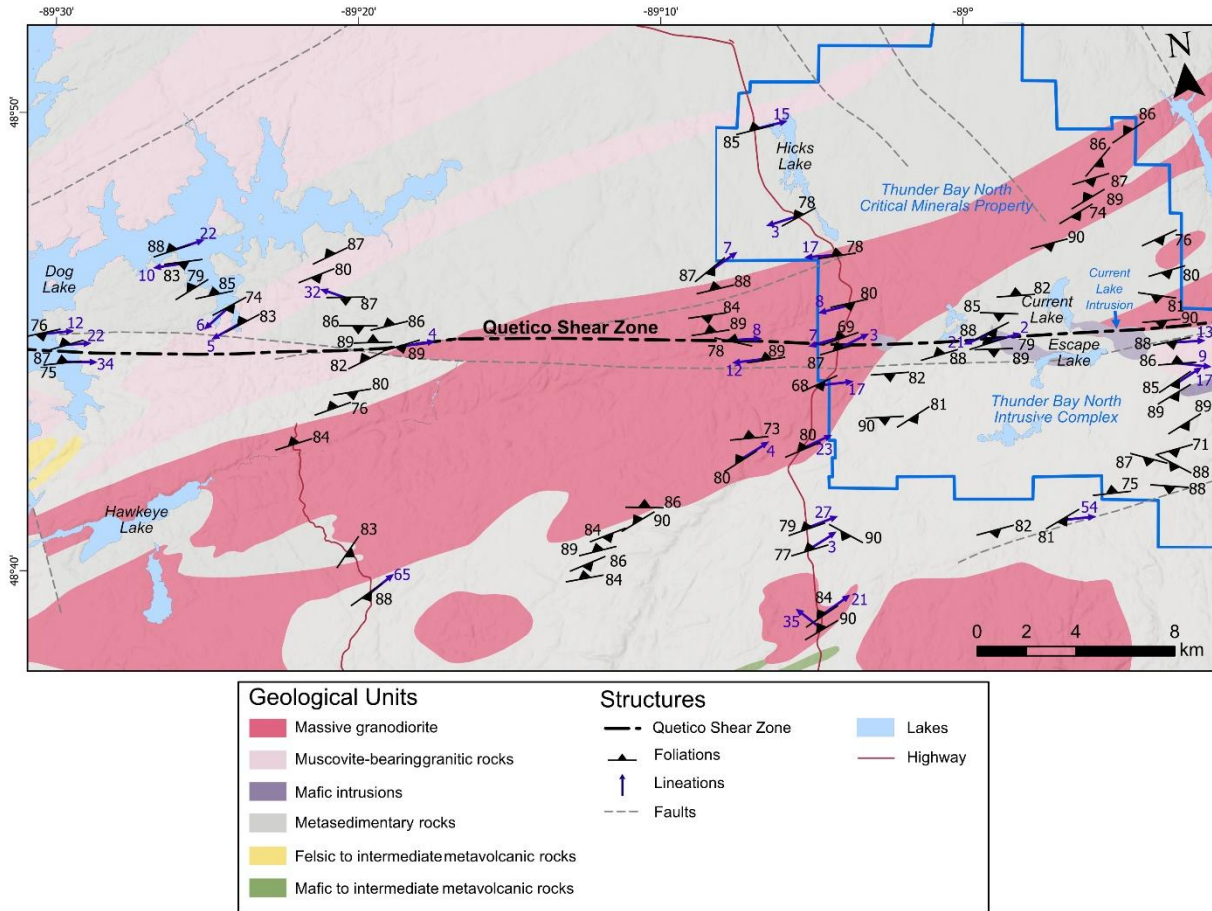


Figure 4.1: Geological map displaying the trace of the Quetico Shear Zone and mean orientation measurements within the study area. The mean foliations (the dip of foliation is indicated in black) and lineations (the plunge of the lineation is indicated in blue) for field stations with rock that display visible fabrics. (*Geology*: after Montsion et al., 2018; *Lakes and Highways*: Ontario Ministry of Natural Resources and Forestry (MNR), Ontario Hydro Network (OHN) Watercourse and MNR Road Segments).

In the study area, metasedimentary rocks are the most abundant unit. This unit consists of various rock types often interbedded with varying strain and metamorphic grades. The two main rock types are metagreywacke (i.e., metapsammites) and pelitic schists, which are largely differentiable where the metamorphism is low south of the shear zone. These two rock types occur either as isotropic and continuous foliated domains or interbedded, where the schist forms thin layers between the more isotropic metagreywacke. Both rock types were subject to higher grades of metamorphism north of the shear zone, resulting in gneissic and migmatic rocks. Granitic

intrusions, dykes, and leucosomes are also commonly interlayered in metasedimentary rocks.

The metagreywackes are either distinguished by a light grey to light orange weathered surface or a dark, moderately homogeneous, and granular fresh surface (Fig. 4.2A). The grain sizes are generally very fine-grained, creating moderately homogeneous appearances, while some occur as fine to medium-grained with feldspars up to 2 mm long. The bedding of the metagreywacke varies from uniform thin mm beds (Fig. 4.2B) to inconsistent beds ranging from thin (less than 1 mm) to thick (approximately 5 cm; Fig. 4.2C). The metagreywacke consists predominantly of quartz with minor to moderate amounts of feldspar and mica. Feldspar occurs as medium-grained clasts within a finer-grained matrix, and mica often forms thin cm bands or mm thick seams. While the metagreywackes often appear to be undeformed, many of them contain deformed carbonate or quartz-rich veins that are folded, boudinaged, or offset across microfaults (Fig. 4.2D). The less massive metagreywackes display a moderate foliation commonly defined by layers of alternating lighter and darker colours interpreted to represent bedding.

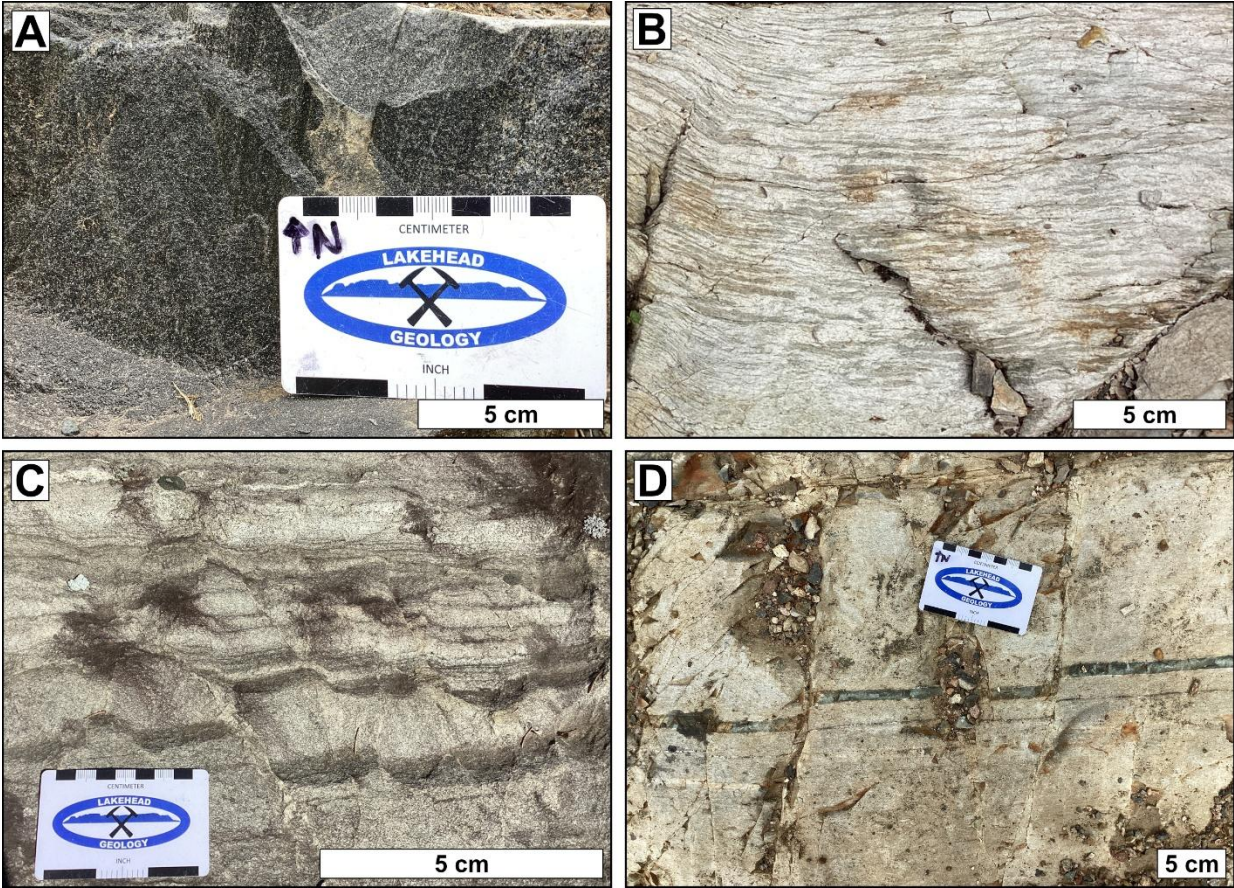


Figure 4.2: Low-grade metasedimentary rock/metagreywacke: A: Dark grey fresh granular surface of metagreywacke/sandstone (24-TBN-HT-52; 11.5 km from the shear zone trace). B: Fine-grained metasedimentary rock with discontinuous bedding with various thicknesses (24-TBN-HT-26; 230 m from the shear zone trace). C: Moderately foliated bedded metasedimentary rock (24-TBN-HT-1; 2.29 km from the shear zone trace). D: Dark green chlorite-rich vein in a fine-grained metagreywacke (24-TBN-HT-29; 245 m from the shear zone trace).

The pelitic schists predominantly have a slight grey to dark weathered surface; locations with greater iron content, such as biotite- or garnet-rich layers, have orange weathered surfaces (Fig. 4.3A). The pelitic schists predominantly consist of quartz and biotite with varying amounts of chlorite, muscovite, and minor feldspar. The fresh surfaces are usually dark unless muscovite is prevalent (Fig. 4.3B). Some pelitic schists contain a wide range of accessory minerals, either from detrital processes (e.g., zircon) or metamorphism (e.g., garnet). Accessory minerals visible on the outcrop scale are garnet, staurolite, and tourmaline. Many of the schists are inhomogeneous, with variations in the grain size and composition, which often present themselves as layering at the

outcrop scale. The quartz grains have similar grain sizes to quartz within the metagreywackes, from fine-grained to medium-grained; however, the grain size of micas varies significantly in the schists from the metagreywackes, with coarser grain sizes. Some occurrences consist of very fine-grained mica, which is difficult to see with the naked eye, while others have coarse grains up to 5 cm long. The layering is largely defined by lighter quartz / feldspar-rich bands and darker mica-rich bands (Fig. 4.3C). The schistose rocks also occur within the other units, such as the metagreywacke or intrusive rocks, as either thin (<5 cm) bands or as elongated lenses (Fig. 4.3D). Schistose textures are also associated with gneissic and migmatitic rocks.

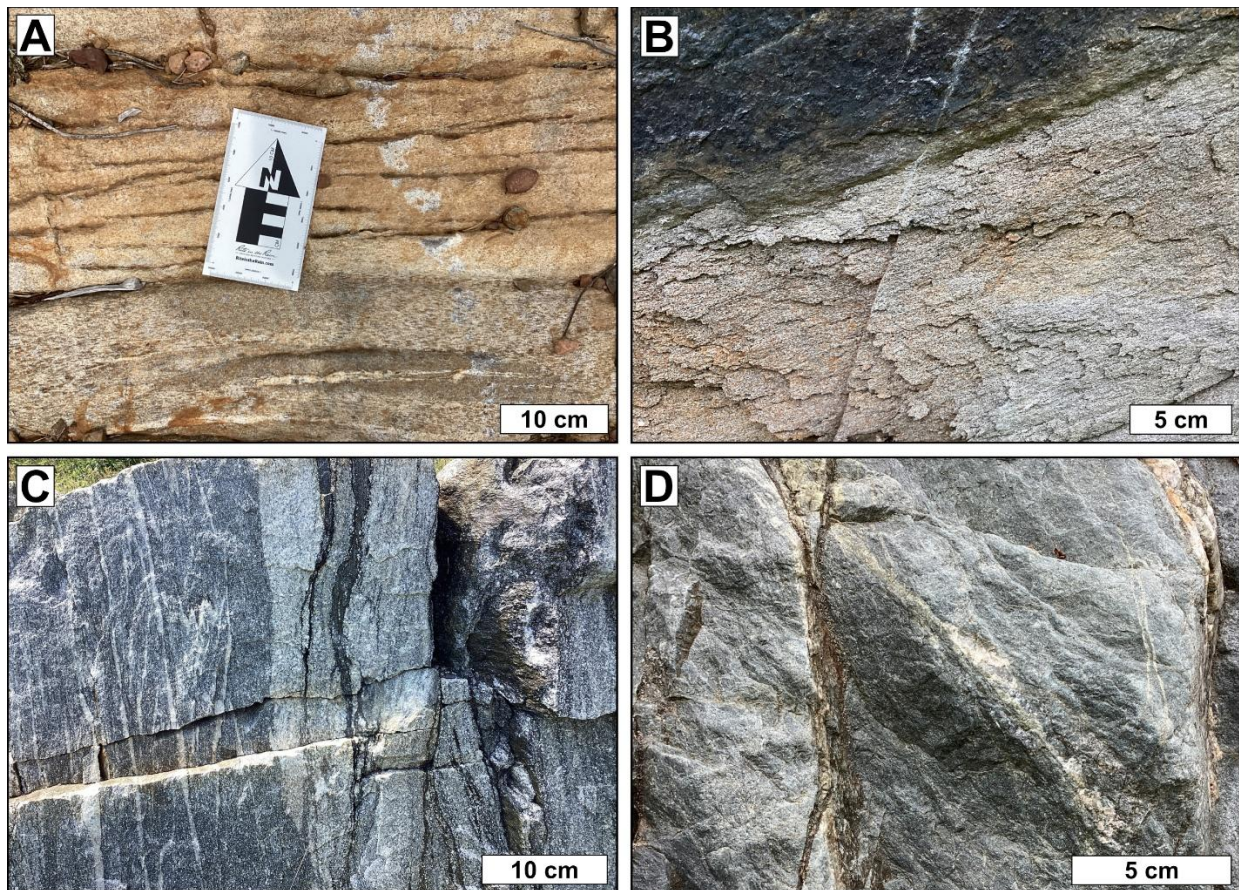


Figure 4.3: Medium-grade metasedimentary rocks: A: Orange weathered garnet-bearing schist (24-TBN-HT-19; 3.59 km from the shear zone trace). B: Light coloured muscovite-rich schist layer with dark schistose layer (24-TBN-HT-54; 1.66 km from the shear zone trace). C: Quartz-rich and dark mica-rich banding within a schistose rock (24-TBN-HT-59; 1.61 km from the

shear zone trace). D: Dark thin (<5 cm) schistose lenses within quartz-rich metasedimentary rock (24-TBN-HT-126; 5.03 km from the shear zone trace).

Towards the north of the study area, the metamorphic degree increases, forming gneisses and migmatites. High-grade metamorphic textures can either occur pervasively throughout the rock, or they form heterogeneous banded rocks with some bands resembling the lower-grade or high-grade variations. The banding consists of layers that were partitioned during metamorphism or are from magmatic dykes, which can be difficult to differentiate. Most of the gneiss appears to be paragneiss due to the abundance of micas and aluminosilicate minerals and association with the metasedimentary rocks (Fig. 4.4). The gneissic rocks are differentiated from the schist-dominated rocks due to the development of concise banding of light and dark layers. The layers are either continuous and parallel (Fig. 4.4A) or they are weakly to moderately folded and pinched out (discontinuous; Fig. 4.4B). The gneissic rocks have a coarser grain size than the schist. The more schistose or darker layers of gneiss, composed predominantly of biotite, have either a light or dark grey colour. The lighter, quartz-felsic layers have light grey, white to light pink colours (Fig. 4.4C, D). Solely in the Dog Lake area (Fig. 4.1), the metasedimentary rocks are abundant in amphibole, resulting in amphibole-rich gneissic rocks. The migmatitic rocks usually occur with gneissic and schistose metasedimentary rocks, which are characterized by the dark restites and light leucogranite melts. The leucogranites have white to light pink colours and are coarse-grained to pegmatitic (Fig. 4.4C). Some of the migmatitic gneisses are associated with larger granitic bodies that could have formed from anatexis or intruded into the sedimentary rocks. The neosome often appears undeformed internally but also occurs discontinuously within the fabric of the paleosome, making it appear to be deformed or a product of melt mixing (Fig. 4.4D). Sometimes a strong foliation is notable within the migmatitic metasedimentary rocks, while other occurrences have

chaotic textures indicating melting (Fig. 4.4D). The higher metamorphic variations of the metasedimentary rocks give evidence for the regional metamorphic trends within the study area.

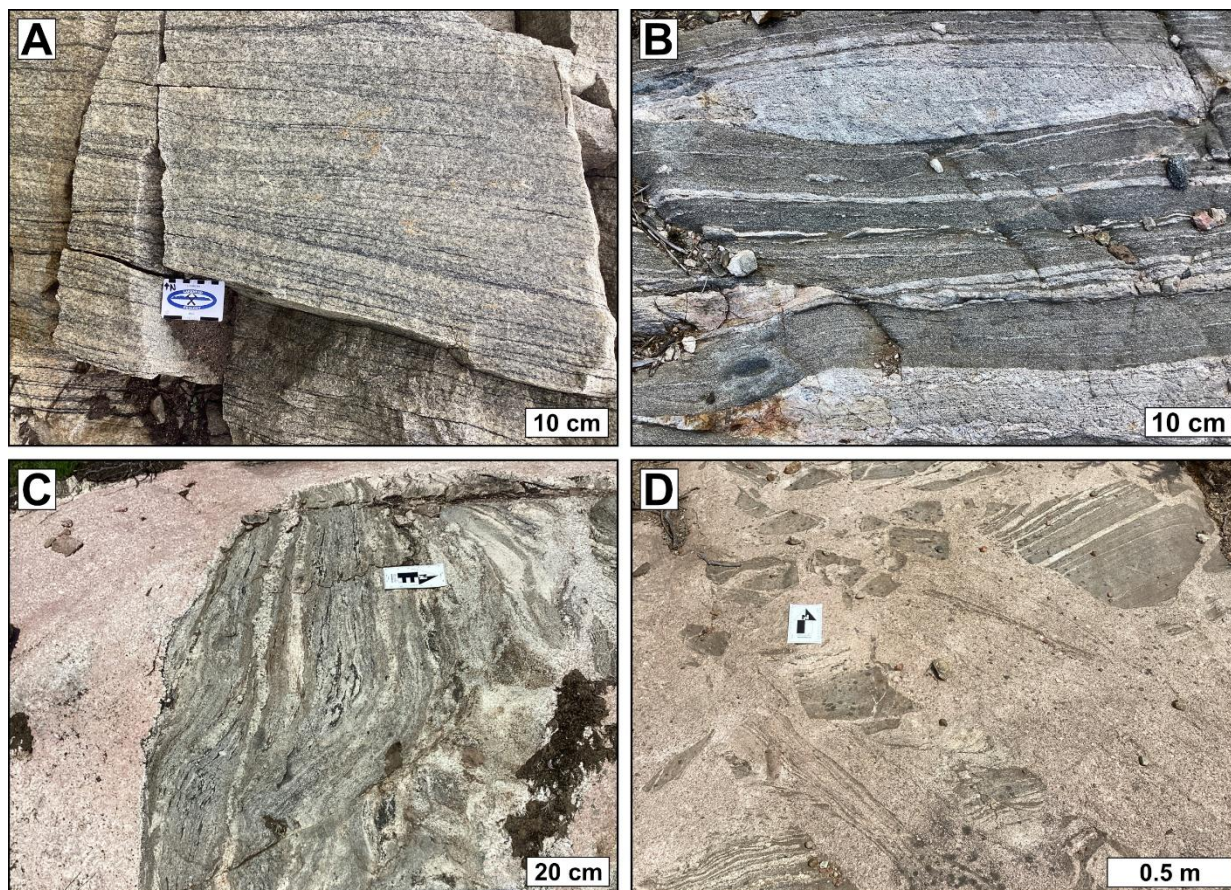


Figure 4.4: High-grade, gneissic metasedimentary rock: A: Medium-grained gneissic rock with thin (<1cm) continuous banding (24-TBN-HT-55; 3.67 km from the shear zone trace). B: Medium to coarse-grained gneissic rock with discontinuous banding that varies in thickness from <1 cm to 10 cm (24-TBN-HT-93; 955 m from the shear zone trace). High-grade, migmatitic metasedimentary rock: C: Large (>0.5 m) light pink leucosomes adjacent to strongly banded layers (24-TBN-HT-10; 8.05 km from the shear zone trace). D: Light pink leucosomes matrix with block of the metasedimentary host rock (24-TBN-HT-14; 6.45 km from the shear zone trace).

Within the study area, there are two main types of granitoids: isotropic to weakly foliated and moderately to strongly foliated (protomylonitic). There are a few locations within the study area where isotropic granitoids crop out, usually distal to the Quetico Shear Zone (Fig. 4.1). The majority of the granitoids proximal to the shear zone are foliated with mylonitic features. The granitoids occur either as large intrusions (>10 m) or as dykes (10 cm–30 cm thick) or dykelets (1-

10 cm thick) within metasedimentary rocks. The granitoids are often coarse-grained or pegmatitic, while some are medium to fine-grained and are either coloured light to dark pink or white to light grey. The main mineral assemblage of the granitoids consists of plagioclase feldspar, alkali feldspar, quartz, and minor biotite and muscovite.

The weakly foliated granitoids occur as massive bodies or dykes with weak foliations, usually only observed from a distance. There are very few occurrences, distal from the shear zone, where the granitoids truly appear to be undeformed; they have isotropic textures with very little change in colour throughout the rock (Fig. 4.5A). The grain sizes often vary, with pegmatitic lenses up to 10 cm wide occurring in finer-grained granitoids. The grain sizes of individual feldspar clasts can reach up to 5 cm. A few pegmatitic granitoids have graphic textures (Fig. 4.5B). The degree of deformation varies in these grains from undeformed and rounded to weakly deformed or elongated. There are some granitoids distal to the shear zone that are weakly foliated, likely from regional deformation (Fig. 4.5C). These appear to be undeformed but have weak fabrics, and some have weak S-C fabrics or discernible lineations. The granitoids also have varying grain sizes; the fine-grained and coarse-grained variations make the deformation difficult to discern. The grains are not strongly deformed, but there can be some elongation of weak minerals such as quartz. There are very few weakly deformed granitic dykelets; some appear undeformed internally but are often weakly to strongly deformed along their margins, often forming folds (Fig. 4.5D). Some of the weakly deformed dykes occur within migmatitic rocks, possibly representing late leucogranites.

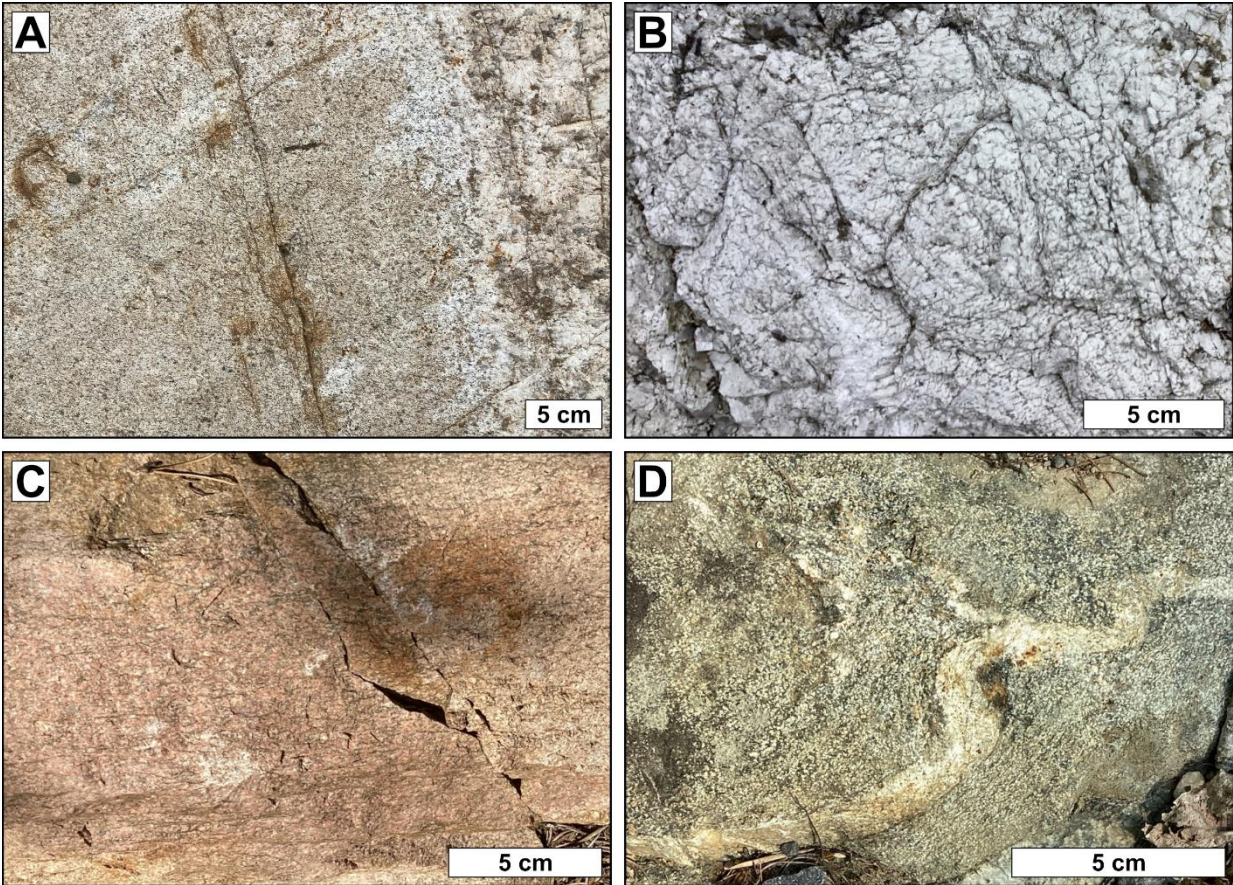


Figure 4.5: Massive to weakly foliated granitoids: A: Beige, isotropic, medium-grained granite (24-TBN-HT-136; 7.68 km from the shear zone trace). B: White, pegmatitic granite with graphic texture (24-TBN-HT-38; 630 m from the shear zone trace). C: Light pink granite with a weak foliation, distal from the shear zone (24-TBN-HT-100; 1.65 km from the shear zone trace). D: White feldspar-rich dykelet composed of weakly deformed minerals and moderately deformed margins forming open folds (24-TBN-HT-85; 3.70 m from the shear zone trace).

The moderately to strongly foliated granitoids generally occur proximal to the shear zone or within areas with strong regional deformation and/or metamorphism. The granitoids are less isotropic than the weakly foliated ones; they often have discrete layers forming due to partitioning of the deformation, causing anisotropic textures and colours. The grain sizes also vary within the granitoids due to the deformation, as some recrystallized as a result of deformation, and others did not. The granitoids have moderate, rarely strong, foliations due to the elongation of clasts or alignment of minerals like micas (Fig. 4.6A). The granitoids often have moderate to strong S-C and S-C-C' fabrics (Fig. 4.6B) from elongated quartz ribbons between weakly deformed feldspar

grains. Some of the fabrics are defined by darker micas, often as thin mm schistose mica bands acting as slip surfaces between larger clasts. With increasing proximity to the shear zone, the quartz grains become more elongated, resulting in more pronounced protomylonitic textures (Fig. 4.6C). Several of the granitic dykelets are moderately to strongly deformed. Within gneissic metasedimentary rocks, the dykelets are thin (<10 cm) and strongly folded, occurring parallel or oblique to the gneissic banding (Fig. 4.6D). Within the greywacke metasedimentary rocks, the dykes vary in size (cm-m wide), often parallel to bedding and foliation. The quartz-rich dykes are mylonitic when they are in close proximity to the shear zone.

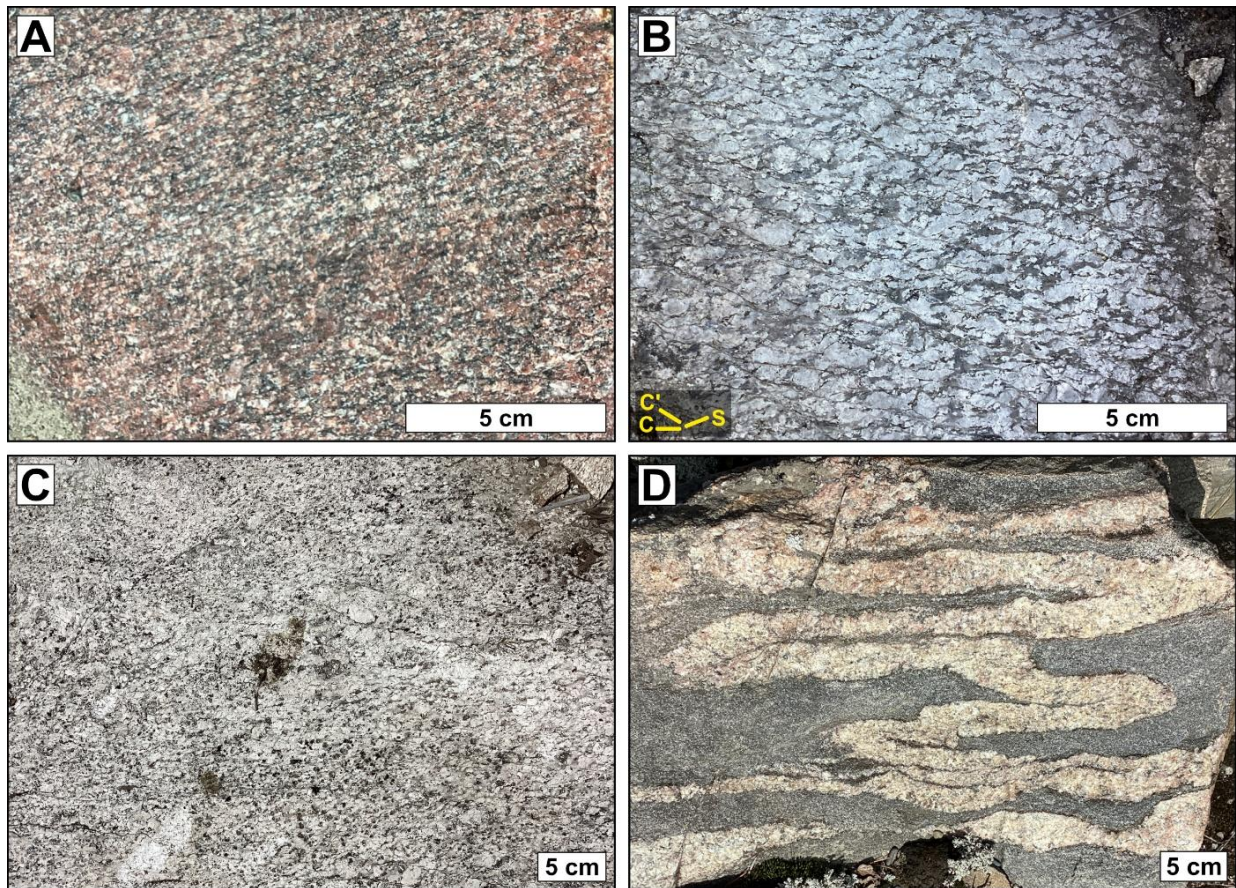


Figure 4.6: Moderately to strongly foliated granitoids: A: Strongly foliated, pink granite, foliation marked by alignment of dark biotite grains (24-TBN-HT-9; 300 m from the shear zone trace). B: Moderately foliated, white granitoid with S-C-C' structures marked by rigid white feldspars and elongated grey quartz grains (24-TBN-HT-121; 600 m from the shear zone trace). C: Recrystallized quartz elongated grains in granite (24-TBN-HT-99; 50 m from the shear zone

trace). D: Light pink, feldspar, quartz-rich dykelet that is deformed and folded in gneissic metasedimentary rock (24-TBN-HT-58; 8.55 km from the shear zone trace).

In the study area, there were two occurrences of intermediate-mafic igneous rocks within the metasedimentary belt, one metavolcanic exposure, and one metagabbroic exposure. The metavolcanic rocks are very fine-grained to fine-grained, with a light to dark grey, weathered and fresh surface (Fig. 4.7A). The metavolcanic rocks occur adjacent to the shear zone, with mylonitic textures near the shear zone, which gives it a glassy, polished-looking surface due to the abundance of quartz. There are a few lighter pegmatitic dykes associated with the metavolcanic rock. Within the metavolcanic rock, there are thin, ~1 cm thick, dark pseudotachylite veins with thin injection veins (Fig. 4.7B). The metagabbro intrusion intrudes within a very fine-grained schistose metasedimentary rock; it has a dark green weathered surface. There are ~1 cm-wide rusty shear zones within the intrusive rock and metasedimentary rock (Fig. 4.7C). The gabbroic rock is ~10 km from the shear zone, and it is only weakly deformed. There are also dark schistose lenses between the intrusive rock (Fig. 4.7D).

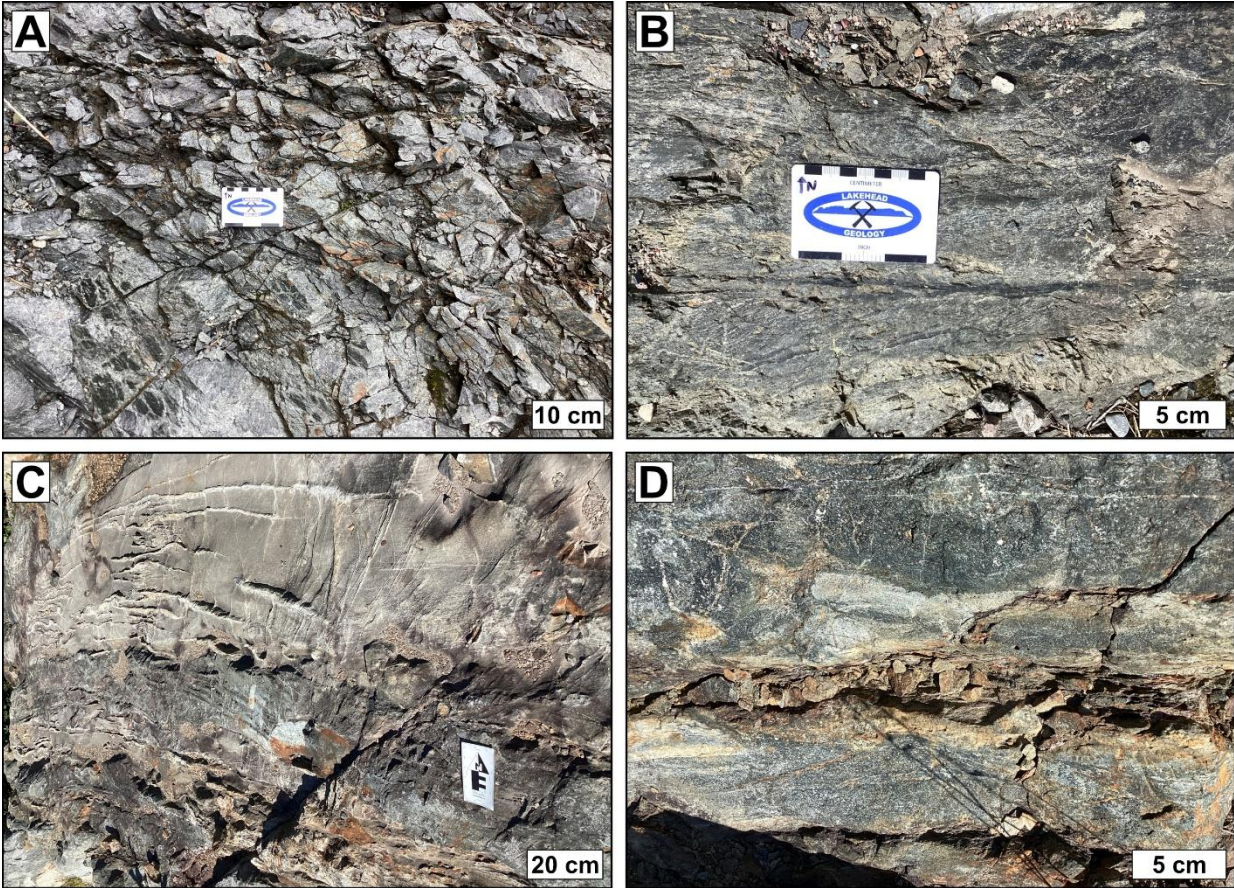


Figure 4.7: Intermediate-mafic igneous rocks: A: Fine-grained metavolcanic rock with abundant fractures (24-TBN-HT-5; 140 m from the shear zone trace). B: Fine-grained, grey metavolcanic rock containing a dark pseudotachylite vein with an injection vein branching off the main vein (24-TBN-HT-6; 110 m from the shear zone trace). C: Coarse-grained, dark green metagabbro with light grey fine-grained metasedimentary rock (24-TBN-HT-104; 8.54 km from the shear zone trace). D: Dark brown to orange, moderately sheared schistose lenses within the metagabbro (24-TBN-HT-104; 8.54 km from the shear zone trace).

4.1.1.2 Drill Core

The diamond drill core from the Thunder Bay North Intrusive Complex (Fig. 4.1) largely exhibits the same units observed from geological mapping. The main exceptions were the rock types associated with the Thunder Bay Intrusive Complex and the abundance of fault rocks due to the direct observation of the core of the shear zone at depth. The drill core that was examined intersected both the shear zone and the Current Lake intrusion (Fig. 4.1). The drill holes examined (Appendix A) consist of, from top to bottom, either/or both granitic and metasedimentary rock

from the Quetico subprovince, then the shear zone, followed by the mafic-ultramafic units. The mafic-ultramafic intrusive rock is medium-grained and dark in colour (Fig. 4.8A). The intrusive rocks do not form a sharp contact with the shear zone or other units (Fig. 4.8B). The intrusive rocks adjacent to the country rock often consist of abundant alteration or hybrid intrusive rocks (Fig. 4.8B). The hybrid rocks are fine-grained and either have a deep red or dark grey colour and sometimes consist of light-coloured quartz eyes (Fig. 4.8B). The core of the shear zone is identified by the abundance of fault gouge (Fig. 4.8C). The fault gouge varies in colour based on the rock type, and consists of a very fine-grained powder, sometimes with 5-10 cm sections of intact rock between the fault gouge (Fig. 4.8C). All units adjacent to the fault gouge are strongly deformed, either brittle or ductile or both. There is also abundant cataclasis within the shear zone (Fig. 4.8D). The apparent thickness of fault gouge/cataclasis varies from 10 cm to 5 m. The intrusive rocks closest to the shear zone are deformed brittly with lots of fractures, and the intrusive complex rocks have no fabric.

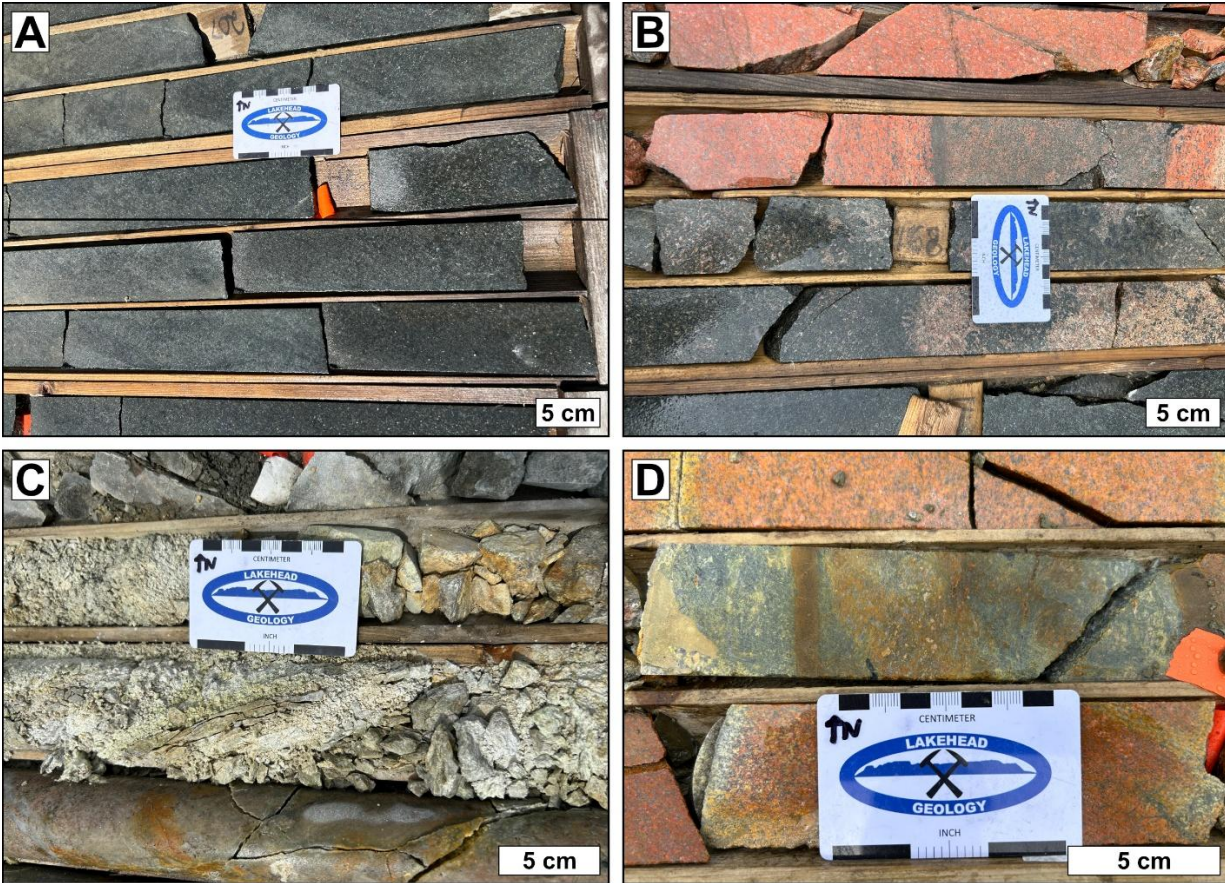


Figure 4.8: Additional rock types only observed in drill core: A: Mafic-ultramafic intrusion of the Thunder Bay Intrusive Complex, Current Lake Deposit (BL09-150). B: Deep red hybrid intrusive rock with gradual contact into the mafic-ultramafic intrusion proximal to the country rock (BL09-141). C: Light-coloured, very-fine grained fault gouge with a larger (~1cm) intact block of fault rocks (BL10-251). D: Light grey, quartz-rich zone of cataclasis within the hybrid intrusive rock (BL10-251).

4.1.2 Macroscopic Fabrics

The macroscopic fabrics remain consistent within the study area, except for minor folding and lithological differences; however, the fabrics are increasingly affected by the Quetico Shear Zone with increasing proximity. Fabric intensity increases significantly towards the shear zone, along with the abundance of both brittle and ductile structures. The brittle and ductile structures mutually crosscut each other, indicating that they occurred contemporaneously.

4.1.2.1 Orientation Analysis

A dominant steep foliation strikes ENE-WSW (Fig. 4.9). There is a some deviation in the strike of the foliations in the equal-area projection towards a more NW-SE orientation or more NE-SW orientation (Fig. 4.9). The dip of the foliation deviates from the geodesic mean up to 25 degrees (Fig. 4.9). The lineation plunges shallowly to the ENE or WSW (Fig. 4.9). There is a slight girdle in the trends of the stretching lineations towards a more NE-SW and ESE-WNW orientation (Fig. 4.9). Lineation plunge deviates significantly, up to 45 degrees from the geodesic mean (Fig. 4.9). Three of the longest transects across the fault highlight the deflection of foliations into the direction of the shear zone orientation, with proximity to the shear zone (Fig. 4.1). Due to the regional ENE-WSW orientation of foliations of the Quetico subprovince rocks and the E-W orientation of the shear zone, the degree of rotation of the foliation planes is minimal. The rotation of the foliation planes in the direction of the shear zone is highlighted by the weak girdle distribution of the poles to foliation (Fig. 4.9). The subhorizontal lineations indicate simple shear-dominated, strike-slip kinematics.

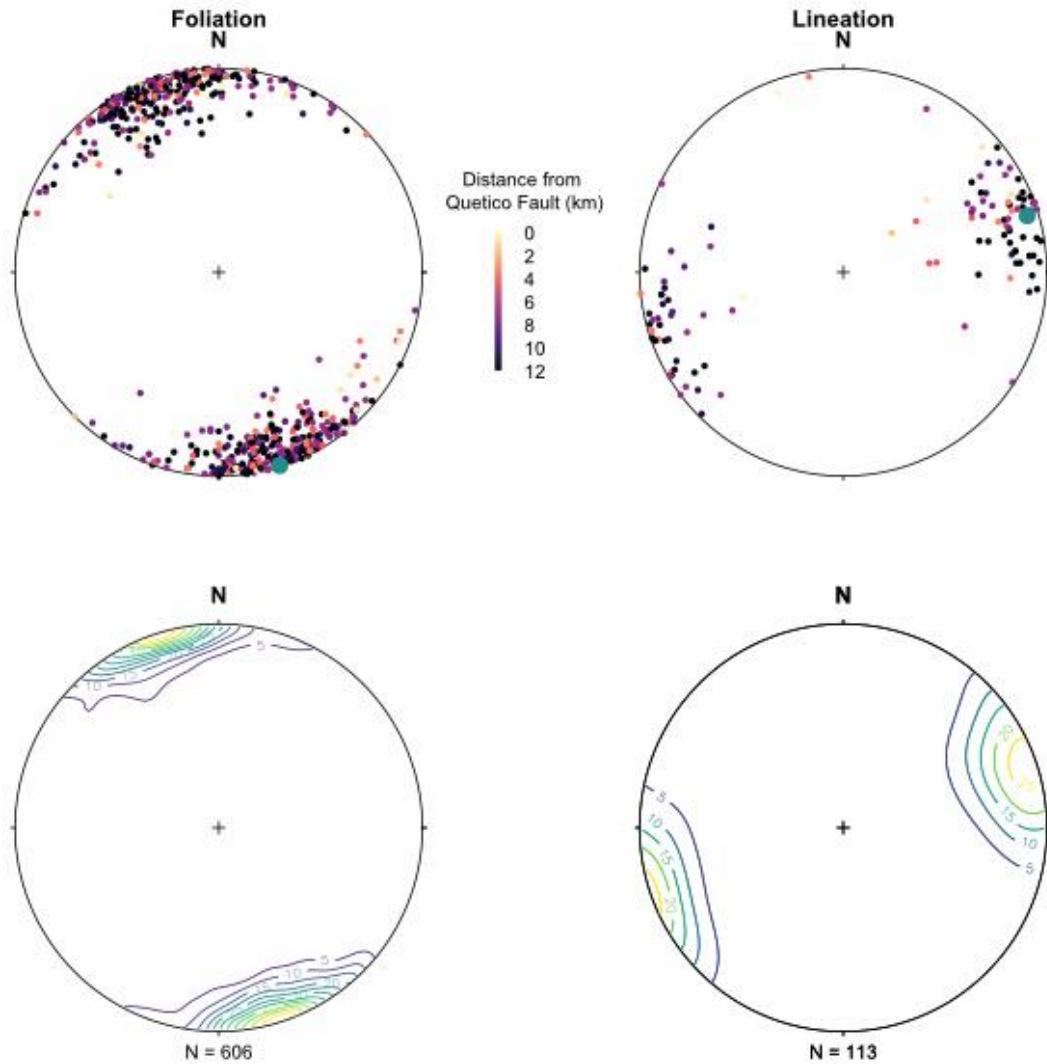


Figure 4.9: Equal-area projections (lower hemisphere, equal area) displaying foliations (left) and stretching lineations (right). The foliations are plotted as poles to planes, and the lineations are plotted as lines. The teal dot indicates the geodesic mean for both the foliation pole and lineation in their respective plots. The contouring in the bottom equal-area projections indicates the overall distribution of the measurements, with the yellow colour corresponding to the highest density of all the measurements.

4.1.2.2 Fabric Characterization

The strain intensity of the rocks notably increases approaching the Quetico Shear Zone. This is seen through an increase in mineral alignment with a similar orientation to the shear zone and decreased foliation spacing. Similar rock units at differing distances from the shear zone have strikingly different fabric characteristics. One of the most notable is granitic rocks, which display

a wide range of fabric intensities. The granitic rocks that are over 1 km from the shear zone have rounded grains with no pronounced fabrics (Fig. 4.10A). In contrast, granitic rocks within 500 m of the shear zone have strongly pronounced protomylonitic and S-C fabrics, often including banding with strongly elongated grains (greater than 10:1 axial ratio), especially quartz (Fig. 4.10B). The occurrence of quartz ribbons is used to measure the shear zone width as they indicate high-strain quartz recrystallization (see Section 4.2). The metasedimentary rocks also show distinct features in proximity to the shear zone. The low-grade metasedimentary rocks have homogeneous textures with small, rounded quartz and feldspar grains with no discernible fabrics (Fig. 4.10C). The metasedimentary rocks within the shear zone are often strongly fractured, with fractures oriented parallel to the shear zone (Fig. 4.10D). The occurrence of various fault rock types also indicates high strain conditions (see Section 4.1.1). The structures that result from strain are dependent on whether ductile or brittle deformation is dominant.

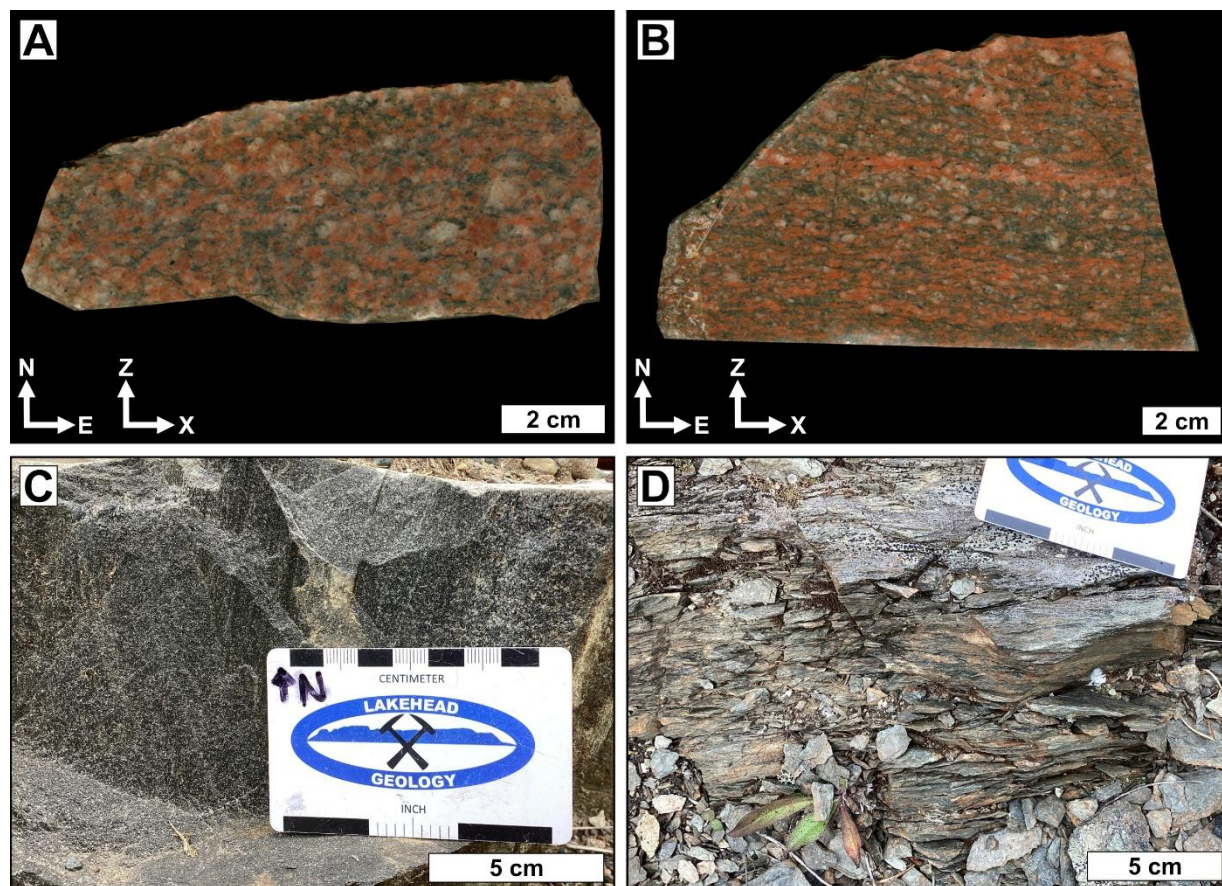


Figure 4.10: Low to high strain fabrics within similar rock types at varying distances from the shear zone: A: Hand sample scan of a weakly deformed dark pink to red granite with rounded grains (930 m from the shear zone, TBN-HT-100, 24-TBN-HT-22). B: Hand sample scan of a moderately to strongly deformed deep red granitic rock with elongated grains and compositional banding (380 m from the shear zone, TBN-HT-36, 24-TBN-HT-9). C: Weakly deformed metasedimentary rock with no visible fabric (11.6 km from the shear zone, 24-TBN-HT-52). D: Strongly deformed metasedimentary rock with abundant fractures in the same orientation as the shear zone (90 m from the shear zone, 24-TBN-HT-30).

4.1.2.3 Brittle and Ductile Deformation

The exposed rocks in proximity to the Quetico Shear Zone exhibit features formed through both brittle and ductile deformation mechanisms. Both mechanisms are observed through various structures within the study area (Fig. 4.11). Brittle structures observed include faults, fractures, and cataclasite zones (Fig. 4.11A, C). Ductile structures observed include foliation, folds, quartz ribbons, and S-C fabrics (Fig. 4.11B, D). Many of these structures are kinematic indicators, which display offset or asymmetry (see Section 4.1.2.4). Intensive brittle deformation largely occurs only

proximal to the shear zone trace, while ductile deformation is noted within the entire study area. Proximal to the shear zone, both mechanisms are observed at field stations within the same rock type and differing rock types. The metavolcanic rocks within the shear zone core display extensive brittle faults and fractures at the macroscale with a strong mylonitic fabric notable at the hand sample-scale (Fig. 4.11A, B). Other field stations display zones of pervasive ductile deformation and localized brittle deformation. One field station consisted of a granitic rock with strong S-C' fabrics adjacent to a metasedimentary rock with a localized zone of cataclasite (Fig. 4.11C, D).

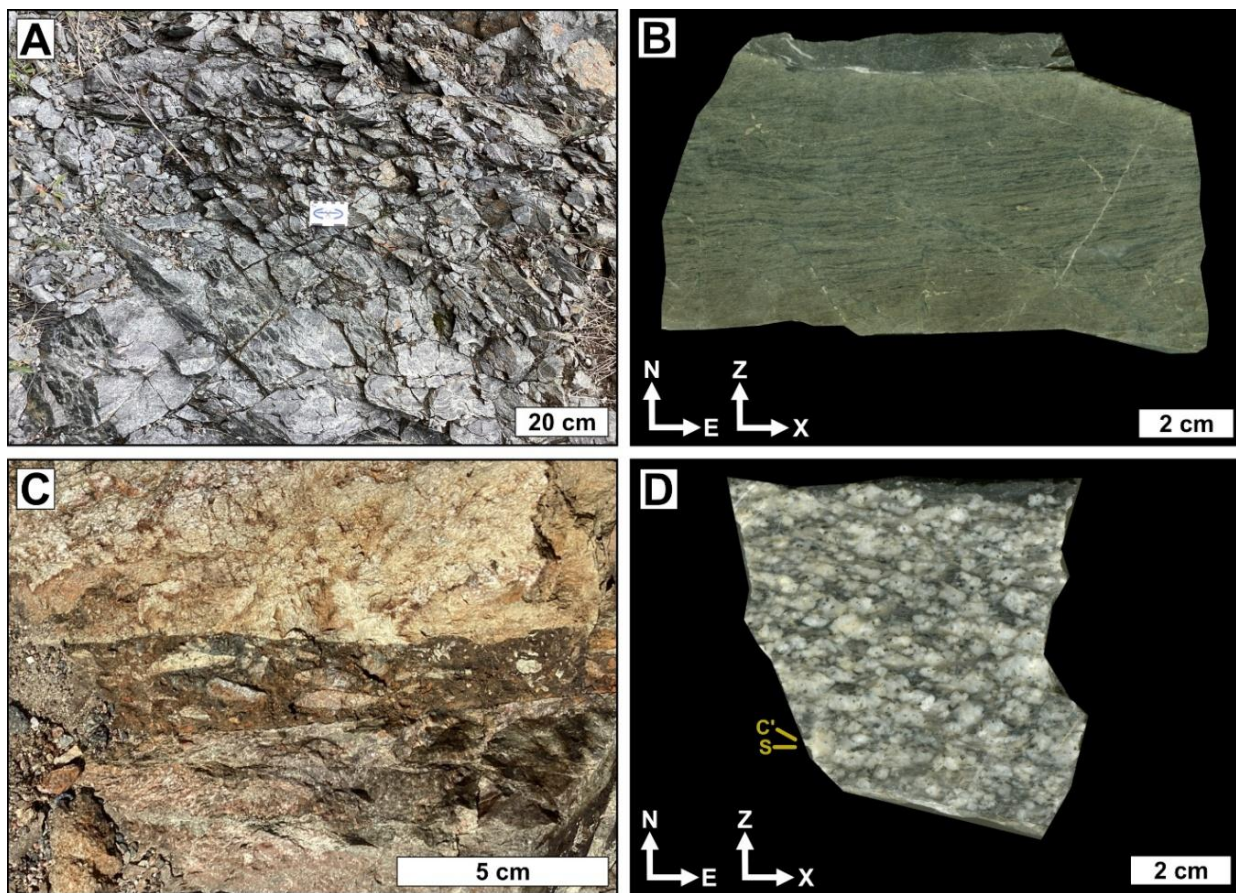


Figure 4.11: Brittle and ductile deformation dominant texture at the same field station and rock type: A: Metavolcanic rock that is abundant with brittle fractures often forming conjugate fractures (24-TBN-HT-5; 140 m from the shear zone trace). B: Hand sample scan of a metavolcanic rock with a strong ductile fabric (24-TBN-HT-5; 140 m from the shear zone trace). Brittle and ductile deformation dominant texture at the same field station within differing rock types: C: Less than 5 cm zone of cataclasite within a metasedimentary rock with a dark brown matrix of pulverized rock with light-coloured angular clasts (24-TBN-HT-59; 1.63 km from the

shear zone trace). D: Protomylonitic granite with S-C' structures from ductile elongated quartz grains and brittle feldspar grains (24-TBN-HT-59; 1.63 km from the shear zone trace).

4.1.2.4 Kinematics

The macroscopic kinematic indicators within the study area show dominant noncoaxial deformation, with a dominantly dextral shear sense. There are both brittle and ductile kinematic indicators, with the ductile kinematic indicators predominantly forming as penetrative features throughout the fabrics, while the brittle kinematic indicators are more discrete, often cross-cutting the fabrics.

Penetrative kinematics indicators include kink bands (both ductile and brittle), domino boudins, asymmetric folds, and sigma/delta porphyroclasts (Fig. 4.12). The domino boudins are observed in gneissic rocks and show an isolated offset between each of the boudins oblique to the main foliation; however, the boudin is parallel to the foliation (Fig. 4.12A). Kink bands are largely observed in metasedimentary rocks; they often display at least a 45 degrees rotation of the foliation (Fig. 4.12B). Asymmetrical folding is observed in most rock types, especially gneissic rocks, where compositional bands and veins are folded (Fig. 4.12C). Large, ~1 cm diameter, sigma and delta porphyroclasts are observed in medium- to coarse-grained granitic rocks (Fig. 4.12D).

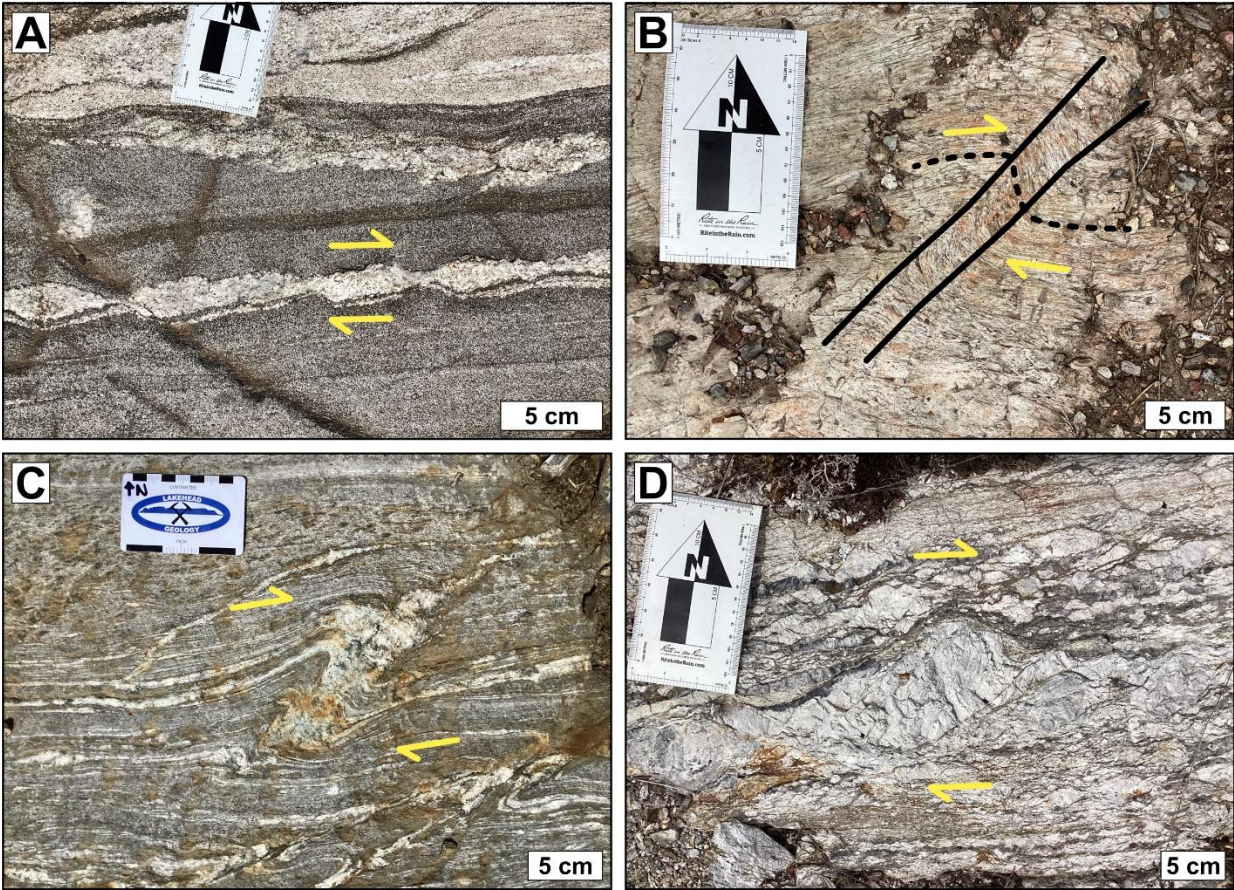


Figure 4.12: Examples of penetrative macroscopic kinematic indicators: A: Brittle domino-boudin structures with a dextral shear sense of a granitic layer within a gneissic metasedimentary rock (24-TBN-HT-15; 6.13 km from the shear zone trace). B: Approximately 1 cm wide ductile kink band with over 45 degrees of clockwise rotation of foliation within a layered metasedimentary rock (24-TBN-HT-26; 230 m from the shear zone trace). C: Asymmetrical z-folding white veins within a metasedimentary rock (24-TBN-HT-34; 510 m from the shear zone trace). D: Large (>10 cm) feldspar sigma clasts with a dextral shear sense (24-TBN-HT-20; 2.00 km from the shear zone trace).

The most common crosscutting kinematics indicator oriented oblique to the foliation are faults. Brittle faults are thin, less than 1 cm, localized zones that are commonly infilled with other material, such as country rock or dykes (Fig. 4.13). Faults observed dominantly have a dextral shear sense (Fig. 4.13A), while some structures show a sinistral shear sense (Fig. 4.13B). The sense of shear across brittle faults is observed from the offset of kinematic markers such as dykes and veins (Fig. 4.13).

The ductile kinematic indicators include shear bands, ductile kink bands, asymmetric folds, and sigma/delta porphyroclasts. The ductile kinematic indicators observed only have a dextral shear sense (Fig. 4.13). The shear bands are often less than 0.5 cm wide; they are either infilled with material or not, and are observed in most rock types (Fig. 4.13A). The shear bands are noticeable from the displacement of fabrics or layers within the rock (Fig. 4.13A). The ductile kink bands are observed largely in metasedimentary rocks; they display up to 45 degrees more rotation than their brittle counterparts (Fig. 4.13B). Asymmetrical folding is observed in most rock types, especially gneissic rocks, where compositional bands and veins are folded (Fig. 4.13C). Large, ~1 cm diameter, sigma and delta porphyroclasts are observed in medium- to coarse-grained granitic rocks (Fig. 4.13D).

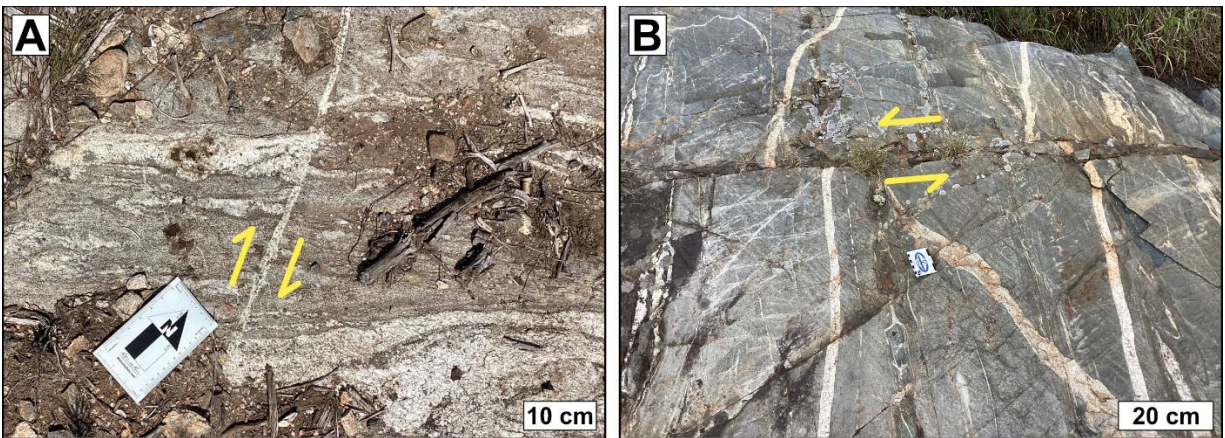


Figure 4.13: Examples of crosscutting macroscopic kinematic indicators: A: White, granitic, 5 cm dykelet offset with a dextral shear sense within a metasedimentary rock (24-TBN-HT-11; 7.88 km from the shear zone trace). B: White, quartz-rich vein offset with a sinistral shear sense in a metasedimentary rock (24-TBN-HT-53; 8.34 km from the shear zone trace).

4.2 Deformation Microstructures

4.2.1 Microscopic Fabrics

The microscopic structures observed in thin section are overall consistent with the macroscopic structures observed from field mapping. As the fabric intensity increases towards the Quetico Shear Zone, the minerals become more elongate and aligned with the foliation plane, which also becomes increasingly parallel to the shear zone trace. The crosscutting relations between brittle and ductile deformation are more evident from microstructures. The kinematics have a prevalent dextral shear sense similar to the macrostructures.

4.2.1.1 Microfabric Characterization

The increase in strain intensity is interpreted from the elongation of mineral grains, predominantly quartz, rarely feldspar, and the alignment of minerals such as micas and feldspars. The granitic samples compared in Section 4.1.2.2 (Fig. 4.1410A, B) display notably different microstructural textures, which indicate an increase in strain intensity (Fig. 4.14A, B). A granitic sample distal from the shear zone (ca. 930 m; Fig. 4.14A) displays a very weak fabric, with slight alignment of larger feldspar grains. There are 0.5 mm dark biotite and chlorite grains adjacent to the feldspar grains, which mainly follow the shape of the feldspar grains (Fig. 4.14A). The quartz grains, ranging in size from 2.5 mm to 0.5 mm, form isolated clusters between the ~5 mm feldspar grains (Fig. 4.14A). In contrast, a granitic rock 380 m from the shear zone (Fig. 4.14B) displays a moderate fabric with 5-10 mm-long quartz ribbons with high aspect ratios, and the majority of the feldspars are aligned with the foliation. The dark micaceous minerals form discrete, planar bands that are thinner than 0.5 mm and are aligned with the quartz ribbons and foliation (Fig. 4.14B). All the quartz grains are less than 1 mm (Fig. 4.14B).

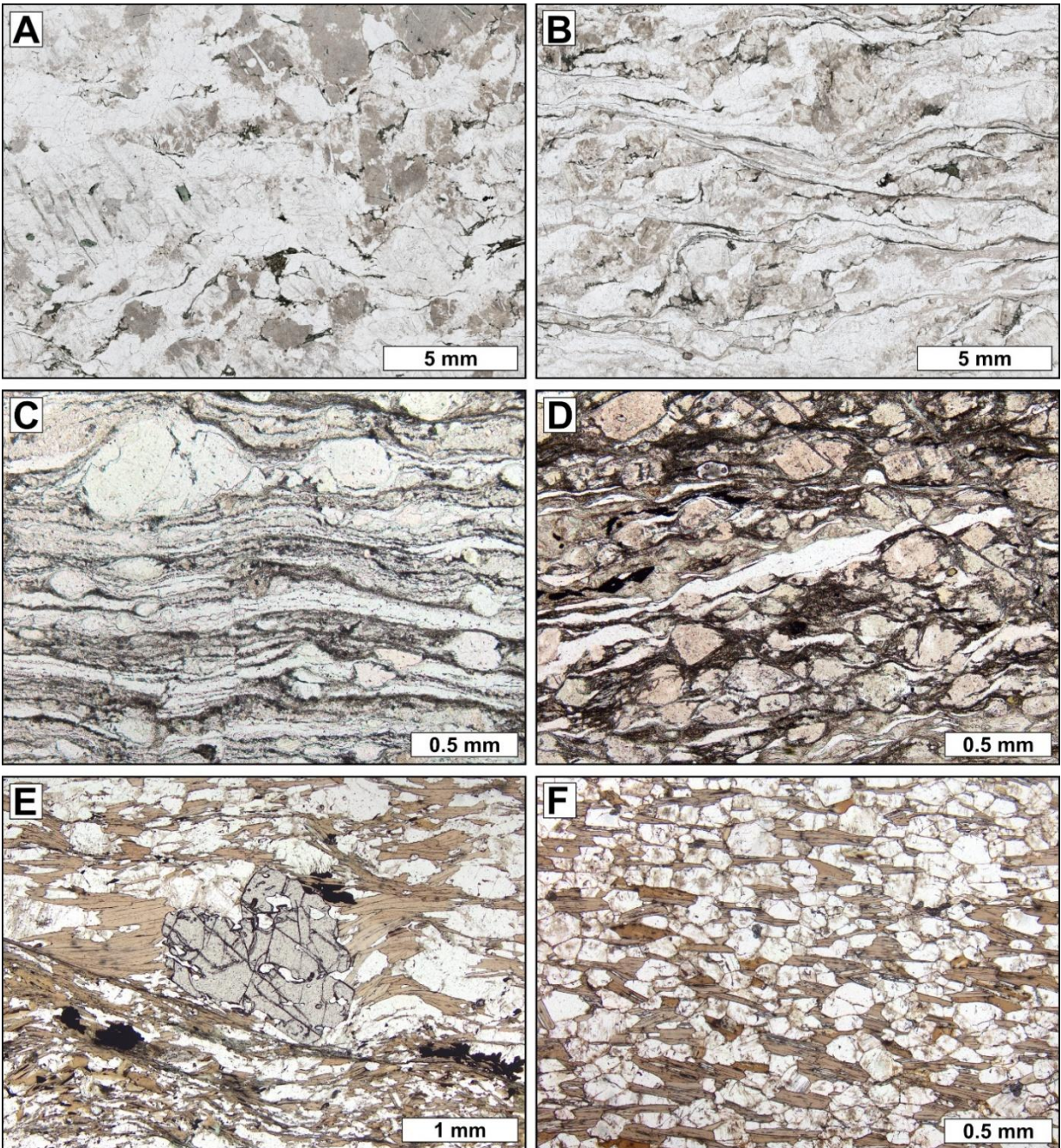


Figure 4.14: Photomicrographs (PPL) displaying examples of microfabric types within the study area (all photomicrographs are oriented with north up and east right): A: Coarse-grained granite with a weak fabric indicated by the minor alignment of the subhedral feldspars (dusty grey). The dark patches of micas are randomly oriented or envelope the feldspar grains (TBN-HT-100; 24-TBN-HT-22; 930 m from shear zone trace). B: Medium-grained protomylonitic granite with a moderate fabric indicated by the long, interconnected quartz (white) ribbons, thin, discrete, biotite bands, and weakly elongated feldspar (dusty grey) porphyroclasts (TBN-HT-36; 24-TBN-HT-9; 380 m from shear zone trace). C: Very fine-grained mylonitic to ultramylonitic granite with a strong fabric indicated by the thin (<1 mm) quartz (white) ribbons and weakly elongated feldspar (dusty grey) porphyroclasts (TBN-HT-33; 24-TBN-HT-7; 45 m

from shear zone trace). D: Fine-grained protomylonitic metavolcanic rock with an abundance of porphyroclasts, thin (<1 mm) quartz ribbons, and mafic minerals deflected around the potassium feldspar (light pink) porphyroclasts (TBN-HT-34; 24-TBN-HT-7; 45 m from shear zone trace). E: Medium-grained garnet biotite schist with moderate fabric indicated by the alignment of biotite and feldspar grains. Garnet (grey) syntectonic porphyroblasts (~1 mm) result in local deflections of the foliation (TBN-HT-115; 24-TBN-HT-55; 3.68 km from shear zone trace). F: Fine-grained biotite schist with a moderate foliation indicated by the alignment of biotite (brown) grains (TBN-HT-133; 24-TBN-HT-109; 4.62 km from shear zone trace).

Quartz-rich rocks within the core of the shear zone form strong fabrics, resulting in mylonitic to ultramylonitic textures (Fig. 4.14C). These rocks have 10-20% relict porphyroclasts, which are smaller than 1 mm (Fig. 4.14C). The rock is dominated by thin (~0.5 mm) quartz stringers with extremely fine grain sizes (down to micrometer scales) and dark, thin, discrete bands consisting of micas and oxides (Fig. 4.14C). Feldspar-dominated rocks proximal to the shear zone form porphyroclast-dominated protomylonitic-mylonitic fabrics with small stringers of quartz between porphyroclasts (Fig. 4.14D). Feldspar porphyroclasts and quartz ribbons often form at slight angles to the lineation, resulting in shear sense indicators from the asymmetry of porphyroclast tails (Fig. 4.14D). The quartz stringers and fine-grained dark micaceous minerals are deflected around porphyroclasts (Fig. 4.14D). The quartz grains within the quartz stringers are also very fine-grained (Fig. 4.14D).

The rocks distal to the shear zone display fabrics that are not as strong as the samples affected by the shear zone deformation. Some rocks consist of porphyroblasts, which are often syn-kinematic and result in deflections of the foliation surrounding them. The most common example of porphyroblasts in the distal samples is garnet, which commonly has tails from strain shadows or causes deflections of the foliation (Fig. 4.14E). Within schistose samples, the orientation of mica grains that were aligned with foliation is often deflected around a garnet grain, which can give an indication of the sense of shear (Fig. 4.14E). There is also often a C' fabric overserved adjacent to the garnet grains, which is composed of thin biotite stringers orientated at

roughly a 45-degree angle to the foliation (Fig. 4.14E). It is also common in distal schistose samples that the mica grains themselves are bent (Fig. 4.14E). In schistose to gneissic rocks with no large clasts or with a homogeneous grain size, there often is a continuous fabric throughout the rock (4.14F). In feldspar-rich schists, the feldspars rarely or slightly form alignment with the foliation, while the micas form weak to moderate alignments based on their large angles (>45 degrees) to foliation planes (Fig. 4.14F). The aligned micas either form a consistently spaced foliation (Fig. 4.14F) or, in gneissic rocks, there are often biotite-rich domains with schistose fabrics separated by feldspar-rich domains with minimal fabrics (Fig. 4.14F). Some rocks display pervasive fabrics throughout the samples (Fig. 4.14F), while the more gneissic rocks have discrete mica-rich layers with fabrics and feldspar-rich layers with minimal fabrics.

4.2.1.2 Microscale Brittle and Ductile Deformation

The coexistence of brittle and ductile deformation types is more easily observed in microstructures and in rocks proximal to the shear zone. It is common for ductile foliations to be offset by brittle features such as shear fractures. There are brittle structures near the shear zone that display evidence of a ductile overprint. A good example of the crosscutting nature is the microstructures adjacent to the pseudotachylite veins (Fig. 4.15). The mineralogy of the metavolcanic host rock is quartz, plagioclase, muscovite, potassium feldspar, chlorite, and biotite within a fine-grained groundmass (Fig. 4.15A). The primary fabric is a schistose, anastomosing foliation that is marked by the banding of micas grains aligned parallel to the schistosity (Fig. 4.15A, B). The microlithons between the micas are composed of quartz and plagioclase feldspar; the quartz grains are moderately elongated up to 1 mm, while the feldspars are subhedral (Fig. 4.15A, B). Shear bands (C-planes) are common fabric features and form at an angle of ~45 degrees to schistosity, producing S-C structures (Fig. 4.15B). The pseudotachylite veins are composed of

a dark brown, very fine-grained mineralogy comparable to the host rock and form sharp contacts with the host rock (Fig. 4.15A, C, D). One of the pseudotachylite veins displays injection veins orthogonal to the main veins, which form small ~1 mm triangular shapes filled with an even finer grained material as well as internal flow structures (Fig. 4.15A). The fabrics, pseudotachylite veins, and injection veins are crosscut by dextral shear fractures (Fig. 4.15A, B, C). Some of the shear fractures show little to no filled material, while others are infilled (Fig. 4.15A, C), forming veins consisting of quartz and calcite (Fig. 4.15A). The pseudotachylite veins are overprinted by ductile fabrics; these are largely observed from quartz stringers within the veins that show a weak ductile fabric slightly oblique to the orientation of the pseudotachylite vein (Fig. 4.15D). There are some larger ~5 mm pyrite clusters within the rock that are either oriented with the foliation or are oriented along the shear bands (Fig. 4.15A). At the microscale, many of the minerals display brittle or ductile deformation, or a combination of both; this is also observed in many microstructures that display kinematic criteria.

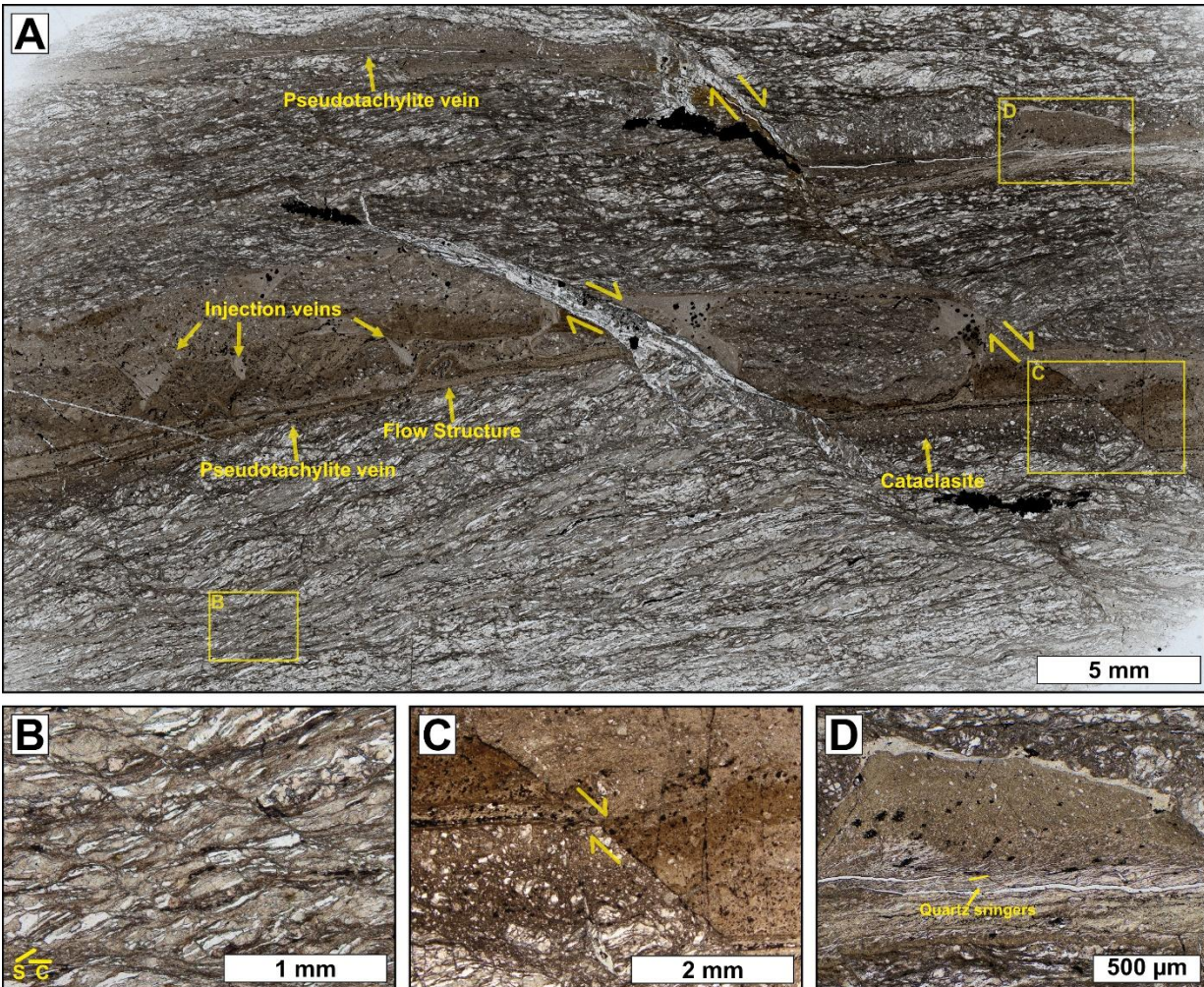


Figure 4.15: Crosscutting of brittle and ductile deformation with a metavolcanic rock (TBN-HT-114a; 24-TBN-HT-6; 100 m from shear zone trace; Thin section scan and photomicrographs are oriented with north up and east right): A: Full thin section scan (PPL) displaying the ductile fabric throughout the thin section, two thin (<5 mm) extremely fine-grained pseudotachylite veins with injection veins and flow structures. Additional structures observed include cataclasite and shear fractures. The central, middle, and top faults are infilled with quartz and calcite. The large cluster of opaques is pyrite. B: Photomicrograph (PPL) displaying the orientation of the intersecting fabrics within the host rock. C: Photomicrograph (PPL) displaying a dextral shear fracture offsetting the pseudotachylite veins and cataclasite. D: Photomicrograph (PPL) displaying a fabric marked by thin (<1 mm) quartz stringers within the pseudotachylite vein, slightly oblique to the veins.

4.2.1.3 Micro-kinematics

Contrary to the macroscale, all the kinematic indicators observed display a dextral shear sense, especially the shear fractures, which only display an observable dextral shear sense (Fig. 4.16A). The shear fractures are common kinematic indicators; they form very thin (~5 μm) discrete

features (Fig. 4.16A). The shear fractures crosscut the foliation as well as individual grains (Fig. 4.14D) and veins within the rocks (Figs. 4.16A & 4.15A, C). The majority of the porphyroclasts with distinct tails or pressure shadows are asymmetric sigma-type, while no delta- or theta-type clasts have been observed. Porphyroclasts, often plagioclase, potassium feldspar, or mica, range in size from 0.5 mm to 1 cm (Fig. 4.16B-D). The tails or strain shadows are often composed of fine-grained minerals, usually recrystallized quartz as well as feldspars and micas or other very fine-grained trace minerals (Fig. 4.16B). Some of the porphyroclasts are partially to completely enveloped by other minerals (Fig. 4.16C, D), such as quartz with varying amounts of feldspar (Fig. 4.16C) or micas (Figs. 4.14E & 4.16D). The mica porphyroclasts (mica fish) are abundant within schistose rocks and are composed of biotite or muscovite, ranging in size from 1-5 mm (Fig. 4.16D). The mica fish are sometimes mantled by very fine-grained micas (Fig. 4.16D).

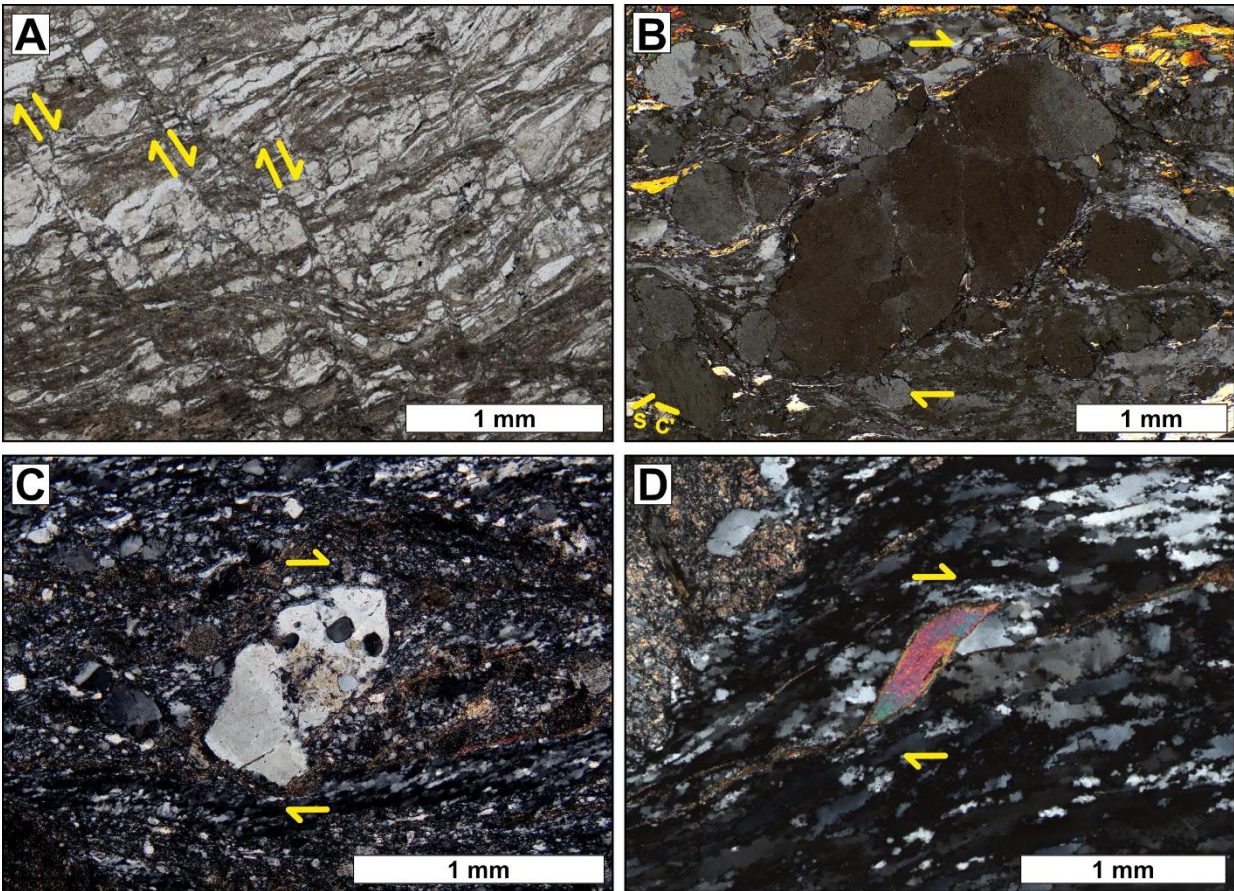


Figure 4.16: Example of kinematic indicators at the microscale displaying a dextral shear sense (all photomicrographs are oriented with north up and east right): A: Photomicrograph (PPL) of a fine-grained metavolcanic rock displaying multiple micro-faults offsetting the fabric (TBN-HT-32; 24-TBN-HT-6; 110 m from shear zone trace). B: Photomicrograph (XPL) of a gneissic metasedimentary rock displaying a ~2 mm plagioclase feldspar sigma porphyroblast with asymmetric strain shadows composed of fine-grained quartz (TBN-HT-119a; 24-TBN-HT-68; 40 m from shear zone trace). C: Photomicrograph (XPL) of a mylonitic granite displaying a ~0.5 mm potassium feldspar sigma porphyroblast with asymmetric strain shadows composed of very fine-grained quartz (TB-HT-33; 24-TBN-HT-7; 45 m from shear zone trace). D: Photomicrograph (XPL) of a quartz vein within a gneissic metasedimentary rock displaying a biotite mica fish completely enveloped by very fine-grained biotite, which extends away from the grain parallel to the foliation (TBN-HT-121; 24-TBN-HT-75; 620 m from shear zone trace).

4.2.1.4 Quartz Recrystallization

The greatest indicator of an increase in strain intensity towards the Quetico Shear Zone is an increase in recrystallization of minerals. The mineral for which recrystallization textures are largely observed is quartz. There are various textures and grain sizes of recrystallized quartz observed within the rocks (Fig. 4.17). Most of the samples have recrystallization textures

consistent with dynamic recrystallization (Fig. 4.17A-C), while a few samples display static recrystallization textures (Fig. 4.17D). Within granites distal to the shear zone, the coarse-grained quartz displays lobate interfingering of the quartz-quartz grain boundaries with weak patchy undulatory extinction (Fig. 4.17A). Most of the quartz within the sample is fine-grained, with serrated boundaries with minor coarser grains enveloped by the finer grains, along with moderate patchy undulatory extinction (Fig. 4.17B). Subgrain rotation is largely notable within rocks proximal to the shear zone; greater proximity is correlated to a finer grain size. A few quartz veins distal to the shear zone within the center of the Quetico subprovince display coarse-grained, long (~5 mm) quartz ribbons with undulatory extinction (Fig. 4.17C). The quartz ribbons are often separated by clusters of fine-grained quartz with serrated boundaries (Fig. 4.17C). The coarse-grained interfingering texture of the quartz is consistent with recrystallization by bulging recrystallization and is rarely observed within the rocks (Fig. 4.17A). The fine-grained serrated textures are indicative of subgrain rotation recrystallization (Fig. 4.17B). Static recrystallization of quartz is only observed from drill core samples at the core of the shear zone, which are adjacent to the Current Lake Intrusion (Figs. 4.1 & 4.17D). These samples display relict structures similar to samples distal to the intrusion, such as long quartz ribbons and kinematics indicators, although the quartz textures indicate minimal strain (Fig. 4.17D). The quartz grains within the drill core rocks have granoblastic polygonal textures indicated by the approximately 120-degree angle triple junction boundaries between quartz grains and their homogeneous grain sizes (Fig. 4.17D).

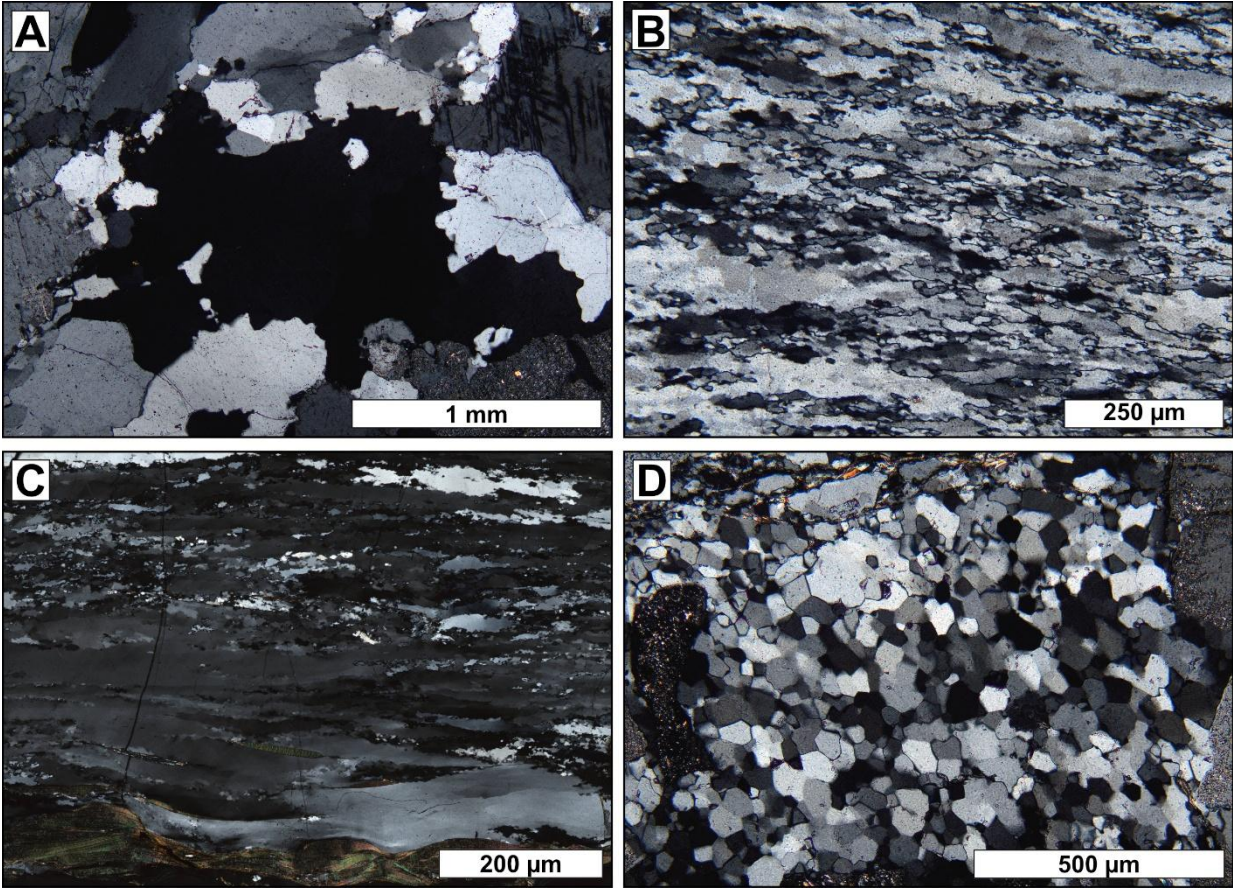


Figure 4.17: Examples of quartz recrystallization texture (all photomicrographs are oriented with north up and east right): A: Photomicrograph (XPL) of coarse-grained quartz displaying medium-grained lobate interfingering of quartz recrystallization along quartz-quartz grain boundaries with some weak patchy undulatory extinction within the quartz (TBN-HT-126; 24-TBN-HT-89; 1.52 km from the shear zone trace). B: Photomicrograph (XPL) of a quartz vein within a gneissic metasedimentary rock displaying very fine-grained quartz with serrated grain boundaries adjacent to medium-grained quartz with patchy undulatory extinction (TBN-HT-119a; 24-TBN-HT-68; 40 m from the shear zone trace). C: Photomicrograph (XPL) of a quartz vein within a gneissic metasedimentary rock displaying very fine-grained quartz grains adjacent to long quartz ribbons with sweeping undulatory extinction (TBN-HT-121; 24-TBN-HT-75; 630 m from the shear zone trace). D: Photomicrograph (XPL) of quartz within a gneissic granitic rock displaying granoblastic polygonal quartz-quartz grain boundaries with no undulatory extinction (TBN-HT-73; BL09-141; 0 m from shear zone trace).

4.2.2 Crystallographic Orientation and Vorticity of Quartz

The microscopic orientation of quartz within the rocks differs from sample to sample as well as from domains within the eight samples analyzed by electron backscatter diffraction (EBSD). There are three main microfabric types throughout the samples based on orientation distributions, which largely correspond to strain intensity as well as rock type. Representative

inverse pole figure maps for each of the three fabric types are displayed to show the fabric within the rock and the orientation of the quartz grains with respect to the sample. All microfabric plots can be found in Appendix C.

The most distal sample, TBN-HT-121 (620 m from the shear zone), displays the strongest fabric of all samples based on the density values within the ODF plot (Fig. 4.18). The rock analyzed within the sample TBN-HT-121 is a ~1 cm quartz vein within a biotite gneiss (Fig. 4.18A). Within the analyzed area (Fig. 4.18A), the quartz vein consists of long quartz ribbons that are ~0.5 mm wide (Fig. 4.18A). There are many fine-grained quartz grains adjacent to the grain boundaries of the larger quartz ribbons as well as within the quartz ribbons (Fig. 4.18A). The quartz crystallographic preferred orientations (CPO) within the sample only display slight variations across the three areas analyzed (Appendix C). The c-axes display a strong maximum within the foliation plane and perpendicular to the lineation (i.e., parallel to the z-direction), with a weak girdle perpendicular to the foliation (Fig. 4.18B). The a-axes display strong point maxima in two directions sub-parallel to the lineation direction and a weak point maximum perpendicular to the foliation and lineation (Fig. 4.18B). The m-axes display one strong point maximum parallel to the lineation with two weak point maxima sub-perpendicular to the foliation and lineation (Fig. 4.18B). The subgrain boundary misorientation axes display two strong clusters around the c- and π/r -axis, indicating prism $\langle a \rangle$ or rhomb $\langle a \rangle$ slip (Fig. 4.18C; Neumann, 2000). The grain boundary misorientation axes display a moderate cluster around the c-axis and a weak cluster around the m-axis, indicating prism $\langle a \rangle$ slip and very weak basal $\langle a \rangle$ slip (Fig. 4.18C; Neumann, 2000). The CVA displays a strong point maximum within the foliation plane perpendicular to the lineation (Fig. 4.18D).

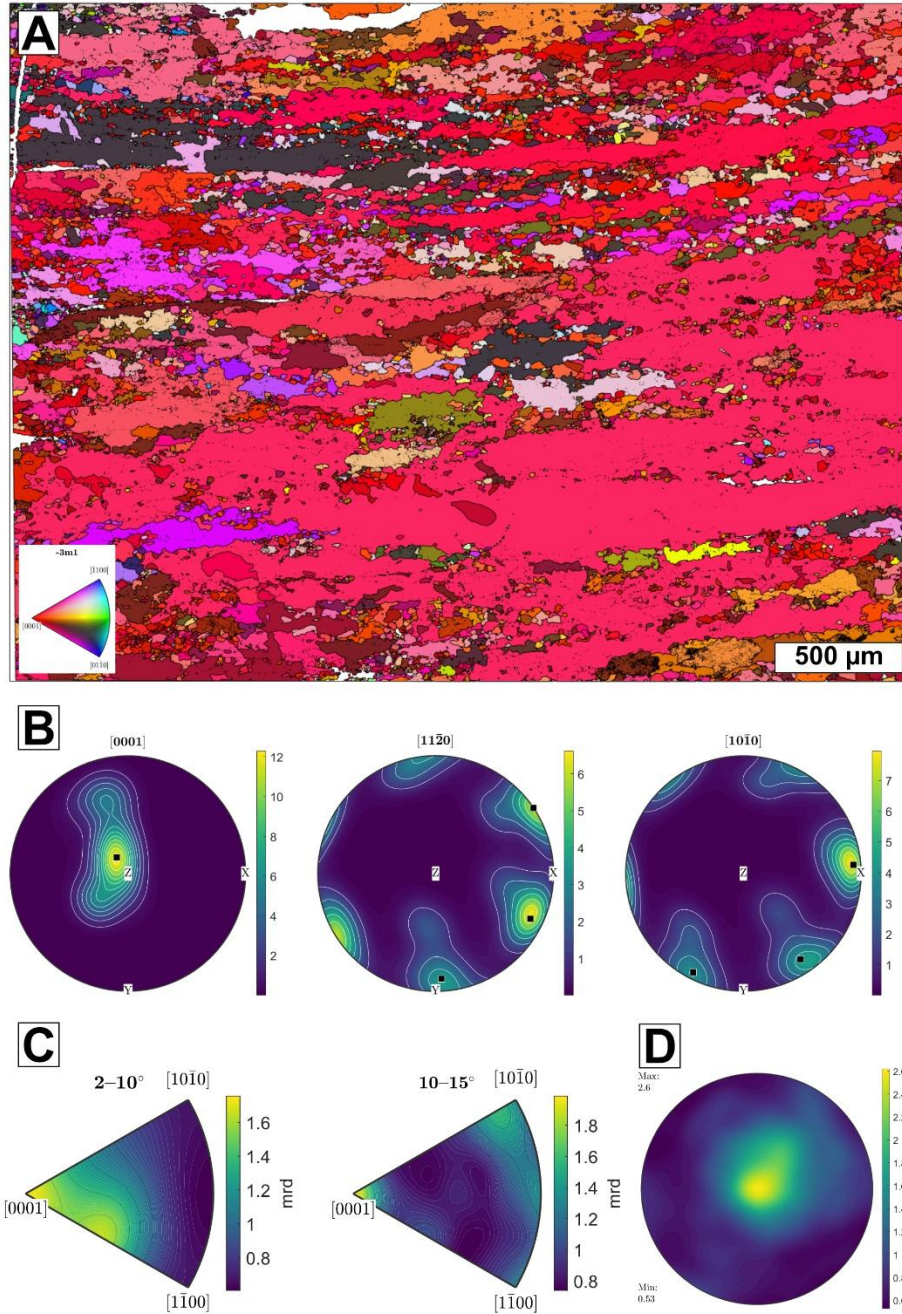


Figure 4.18: Micro-orientations from the most distal sample from the shear zone analyzed with EBSD (TBN-HT-121; 24-TBN-HT-75; 620 m from shear zone trace): A: IPF-Z map displaying the fabrics of quartz grains within the quartz vein in a biotite gneiss (EBSD maps are oriented with north up and east right). The IPF-Z key in the lower left corner indicates the orientation of the quartz grains with respect to the sample in the Z-direction. B: Equal-area projections (lower hemisphere) plots of ODF displaying the CPO of quartz grains for the c-axis [0001] (left), a-axis [11 $\bar{2}$ 0] (center), and the m-axis [10 $\bar{1}$ 0] (right). C: MDF inverse pole figure displaying the misorientation axes of quartz subgrain boundaries (left) and grain boundaries (right). D: Equal-area projection (lower hemisphere) plot of CVA for quartz grains.

The next representative sample, TBN-HT-35 (240 m from the shear zone), displays a moderate fabric, based on the density values within the ODF, with varying orientations throughout the sample (Fig. 4.19). The rock analyzed within the sample is a protomylonitic granite (Fig. 4.19A). The fabric within the protomylonitic granite is mainly defined by isolated quartz ribbons between the coarser-grained feldspar grains, while the non-ribbon quartz grains themselves are not strongly elongated (Fig. 4.19A). The CPO within the sample varies significantly across the four areas analyzed within the sample (Appendix C). The c-axes display a point maximum slightly oblique to the foliation and perpendicular to the lineation (Fig. 4.19B). There are two other weaker point maxima, both of which are orthogonal to the strongest maximum; one is within the foliation plane, and the other is perpendicular to the foliation plane (Fig. 4.19B). The a-axes display many weak maxima with one moderate maximum along the foliation plane oblique to the lineation (Fig. 4.19B). Similar to the a-axes, the m-axes are spread out with two weak to moderate maxima slightly oblique to foliation (Fig. 4.19B). The subgrain boundary misorientation axes display one strong cluster around the π -r-axis, indicating dominant rhomb $\langle a \rangle$ slip with very weak distribution around the c-axis (Fig. 4.19c; Neumann, 2000). The grain boundary misorientation axes display as a moderate cluster around the c-axis, indicating prism $\langle a \rangle$ slip (Fig. 4.19C; Neumann, 2000). The CVA displays a weak point maximum within the foliation plane perpendicular to the lineation, while the overall distribution is uniform (Fig. 4.19D).

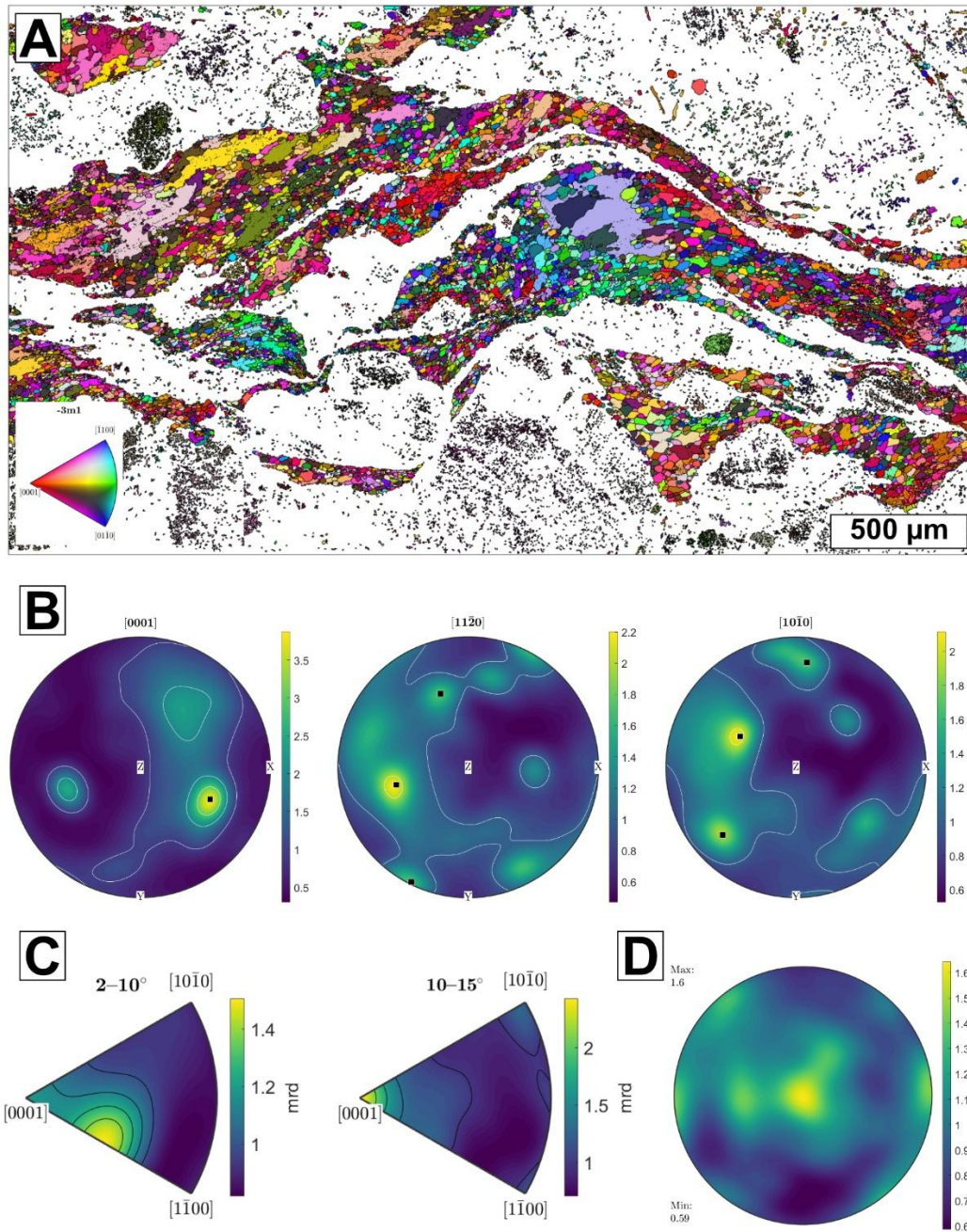


Figure 4.19: Micro-orientations from the moderately proximal sample from the shear zone analyzed with EBSD (TBN-HT-35; 24-TBN-HT-8; 240 m from shear zone trace): A: IPF-Z map displaying the fabrics of quartz grains within the quartz clusters and ribbons within a granite (EBSD maps are oriented with north up and east right). The IPF-Z key in the lower left corner indicates the orientation of the quartz grains with respect to the sample in the Z-direction. B: Equal-area projections of ODF displaying the CPO of quartz grains for the c-axis [0001] (left), a-axis [1120] (center), and the m-axis [1010] (right). C: MDF inverse pole figure displaying the misorientation axes of quartz subgrain boundaries (left) and grain boundaries (right). D: Equal-area projection of CVA for quartz grains.

The most proximal sample, TBN-HT-33 (45 m from the shear zone), displays a moderate fabric, based on the density values within the ODF, with very minor variation in orientation throughout the sample (Fig. 4.20). The rock analyzed within the sample is a quartz vein within a mylonitic granite (Fig. 4.20A). The fabric within the mylonitic granite is formed by thin quartz ribbons between the elongated feldspar porphyroclasts (Fig. 4.20A). Within the quartz veins, the quartz grains have a fabric that is oblique to the main lineation (Fig. 4.20A). The CPO within the sample only displays a slight variation across the two areas analyzed (Appendix C). The c-axis displays two strong maxima that form a small girdle perpendicular to the foliation plane (Fig. 4.20B). The a-axes display one moderate point maximum subparallel to the lineation, with the axes spread around the outside of the pole figure (Fig. 4.20B). The m-axes display one moderate point maximum parallel to the lineation, with the rest of the axes spread around the outside of the pole figure (Fig. 4.20B). The subgrain boundary misorientation axes display one strong cluster around the π -r-axis, indicating rhomb $\langle a \rangle$ slip (Fig. 4.20C; Neumann, 2000). The grain boundary misorientation axes display a moderate cluster around the c-axis, indicating prism $\langle a \rangle$ slip (Fig. 4.18C; Neumann, 2000). The CVA displays a strong point maximum within the foliation plane perpendicular to the lineation (Fig. 4.18D).

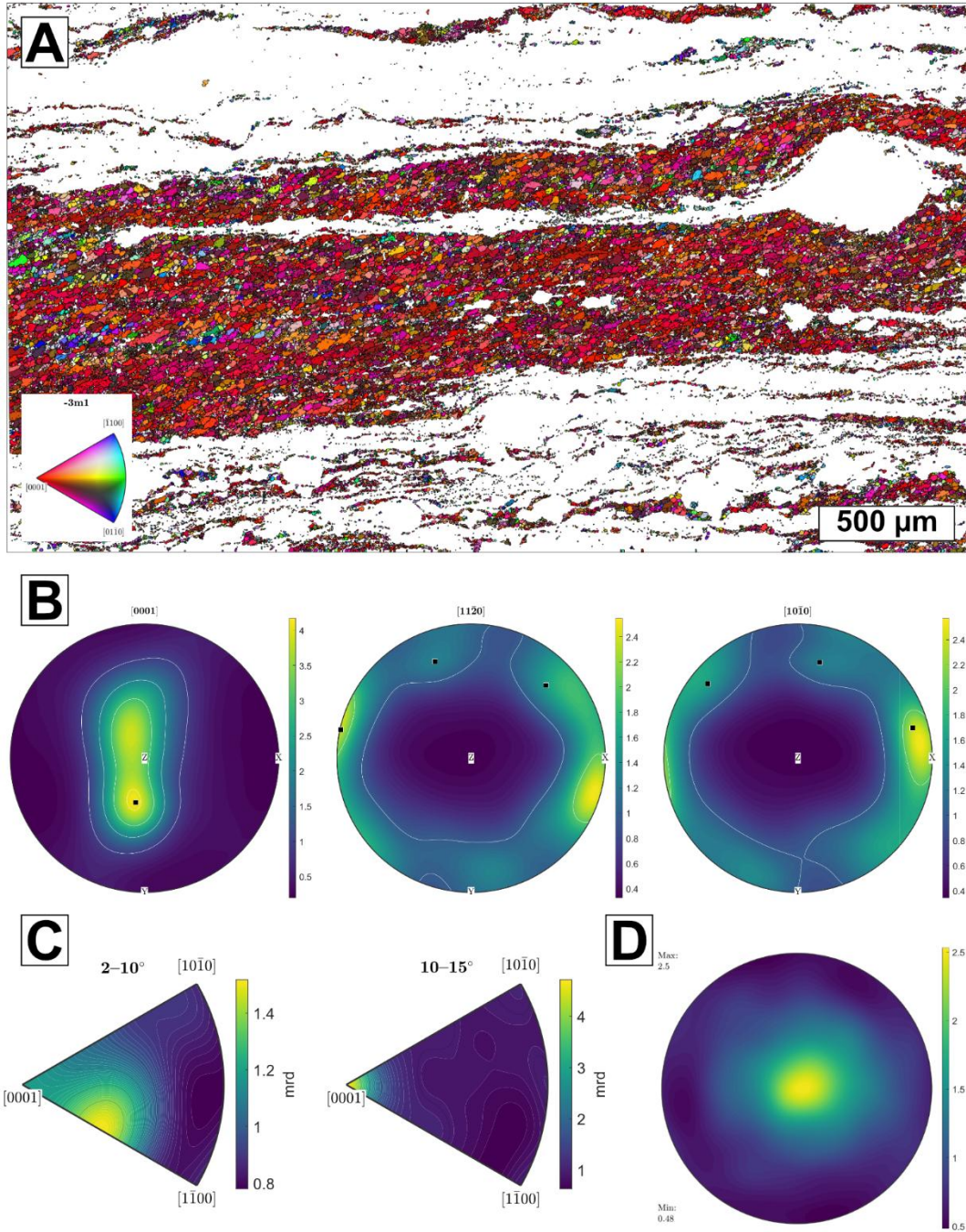


Figure 4.20: Micro-orientations from the most proximal sample from the shear zone analyzed with EBSD (TBN-HT-33; 24-TBN-HT-7; 45 m from shear zone trace): A: IPF-Z map displaying the fabrics of quartz grains within a quartz vein within a mylonitic granite (EBSD maps are oriented with north up and east right). The IPF-Z key in the lower left corner indicates the orientation of the quartz grains with respect to the sample in the Z-direction. B: Equal-area projections of ODF displaying the CPO of quartz grains for the c-axis [0001] (left), a-axis [11 $\bar{2}$ 0] (center), and the m-axis [10 $\bar{1}$ 0] (right). C: MDF inverse pole figure displaying the misorientation axes of quartz subgrain boundaries (left) and grain boundaries (right). D: Equal-area projection of CVA for quartz grains.

The majority of the CPOs for all samples display c-axes with a maximum around the center of the pole figure, oriented within the foliation plane (parallel to the z-direction) and perpendicular to the lineation (Appendix C). The samples with CPO that differ in orientation are largely sample areas where isolated quartz veins were not analyzed; these samples contain abundant porphyroclasts or areas affected by many larger grains. These CPOs display varying orientations with more random distributions. The subgrain boundary misorientation axes for all the samples plot within the c- and pi-/r-axis except for one sample (TBN-HT-31, area 2), which plots in the m-axis (Appendix C). These indicate that the primary slip systems within the shear zone are prism $\langle a \rangle$ and rhomb $\langle a \rangle$. The majority of the grain boundary misorientation axes plot within the c-axis, with a few within the a and m axes, indicating dominantly prism $\langle a \rangle$ slip (Appendix C). The majority of the CVAs plot within the center of the pole figures parallel to the foliation and perpendicular to the lineation, with a few samples plotting as maxima oblique to the dominant CPO or as random distributions (Appendix C).

4.3 Conditions of Deformation

4.3.1 Stress estimated from Quartz grain size piezometry

The decreasing grain size with increasing proximity to the Quetico Shear Zone noted at both the macro- and micro-scales (see Sections 4.1.2 & 4.2.1) corresponds to an increase in deformation towards the shear zone. With quartz recrystallization evident from the mylonitic fabrics, there is an increasing degree of mylonitization with increasing proximity to the shear zone (Fig. 4.21). The degree of mylonitization is defined by pervasive fine-grained quartz, which tends to form ribbons within the protomylonitic rocks (Fig. 4.21A, B) and to form a groundmass in the mylonitic texture rocks (Fig. 4.21C). Samples occurring at a distance greater than 1 km from the shear zone have weak fabrics, except for gneissic textured rocks, which have a moderate fabric from gneissic banding, but no mylonitization of grains. The granitic rock samples usually have weak fabrics with minimal evidence of recrystallization along the boundaries of quartz grains (Fig. 4.21A). Metasedimentary rocks further from the shear zone often have either coarse-grained gneissic textures or fine-grained schistose fabrics with no clear evidence of mylonitization.

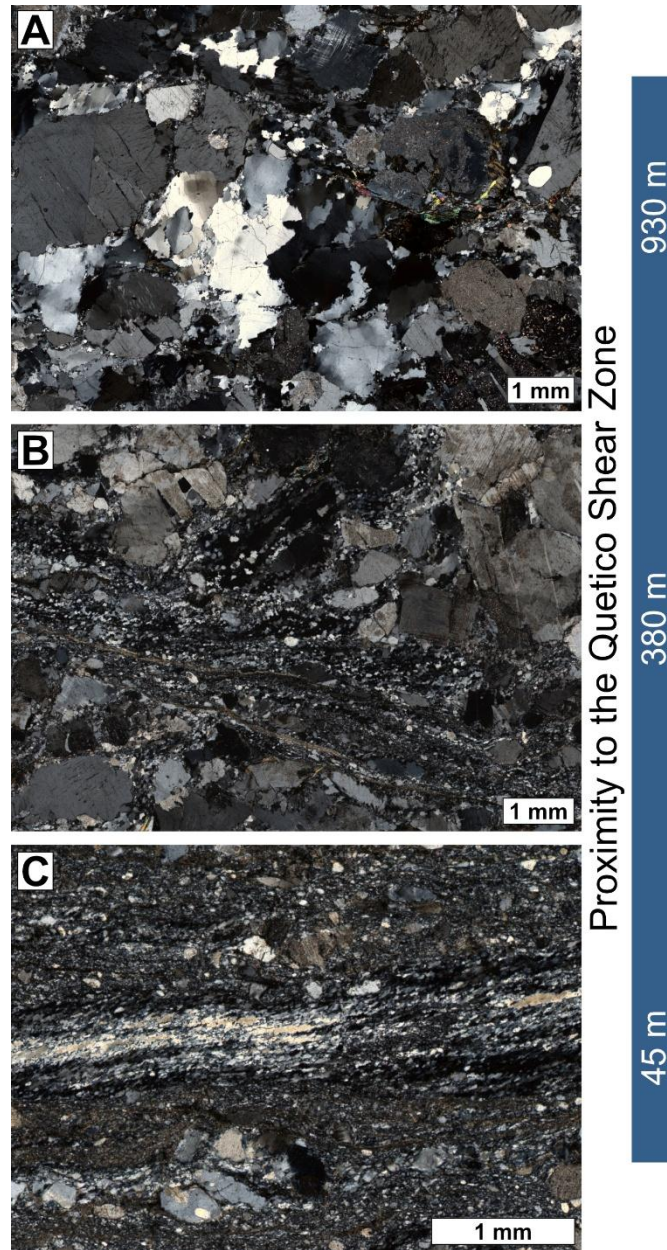


Figure 4.21: Mylonitization development (all photomicrographs are oriented with north up and east right): A: Weakly foliated granite with quartz grain size ranging from $\sim 1 \mu\text{m}$ to 3 mm. The finer-grained quartz is located along the edges of the coarser quartz grains. The quartz grains show weak patchy undulatory extinction. The deformed quartz grains concentrated in isolated patches adjacent to larger ($\sim 1\text{cm}$) feldspar grains (TBN-HT-100; 24-TBN-HT-22). B: Moderately foliated protomylonitic granite with quartz grain sizes ranging from $\sim 1 \mu\text{m}$ to 500 μm . Fine-grained recrystallized quartz is located within shear bands adjacent to larger feldspar grains (TBN-HT-36; 24-TBN-HT-9). C: Strongly foliated mylonitic granitic rock with quartz grain sizes ranging from $<1 \mu\text{m}$ to 200 μm . A 1 mm quartz vein displays recrystallized quartz grains with similar orientations at an angle to the main foliation (TBN-HT-33; 24-TBN-HT-7).

Protomylonitic fabrics are often seen within granitic rocks at distances of 100–500 m from the shear zone. The protomylonitic rocks contain very fine quartz grains confined to quartz ribbons within shear bands between coarser-grained feldspars (Fig. 4.21B). Medium-grained quartz also occurs within the protomylonitic samples; they are usually adjacent to similar-sized feldspar grains and absent in the shear bands (Fig. 4.21B). The mylonitic rocks are dominated by the presence of quartz. One of the granitic samples is completely mylonitized (Fig. 4.21C). The mylonitic granite contains ~10 % K-feldspar porphyroclasts, within a very fine-grained groundmass composed of quartz and feldspars (Fig. 4.21C). The majority of the mylonitic textures of the granitic samples located within the shear zone core occur within quartz ribbons. The quartz ribbons are characterized by very fine-grained quartz, generally with a consistent grain size and usually less than 500 μm in size. Metasedimentary rocks that occur at a similar distance (relative to the shear zone) as the protomylonitic and mylonitic granitic rocks similarly contain fine-grained quartz. However, the metasedimentary rocks contain an abundance of mica-group minerals, making the degree of quartz recrystallization difficult to characterize and differentiate. The conditions at which quartz recrystallization took place will be investigated in this section.

To investigate the stress conditions under which quartz recrystallization occurred, a subset of samples proximal to the shear zone was analyzed using EBSD. The subset of samples consists of various rock types with distances less than 1 km from the shear zone, with an increasing number of samples within 100 m of the shear zone. Prior to constraining the stress conditions during dynamic recrystallization, the quartz grains within each sample were differentiated into relict and recrystallized grains. To characterize the quartz grains as relict or recrystallized, the degree of misorientation of each grain was calculated in MATLAB following the procedures of Cross et al. (2017), which defines an orientation cutoff value for each individual sample (Fig. 4.22). The grains

with a higher degree of internal misorientation (Fig. 4.22A) were characterized as relict (Fig. 4.22B). In contrast, quartz grains with a lower degree of misorientation (Fig. 4.22A) were characterized as recrystallized (Fig. 4.22B). In most of the analyzed samples, the recrystallized quartz grains are finer in grain size relative to the relict grains (Fig. 4.22). Some relict grains display regions with a very large degree of misorientation, corresponding to areas proximal to, and often pinched between fine-grained recrystallized quartz grains (Fig. 4.22A).

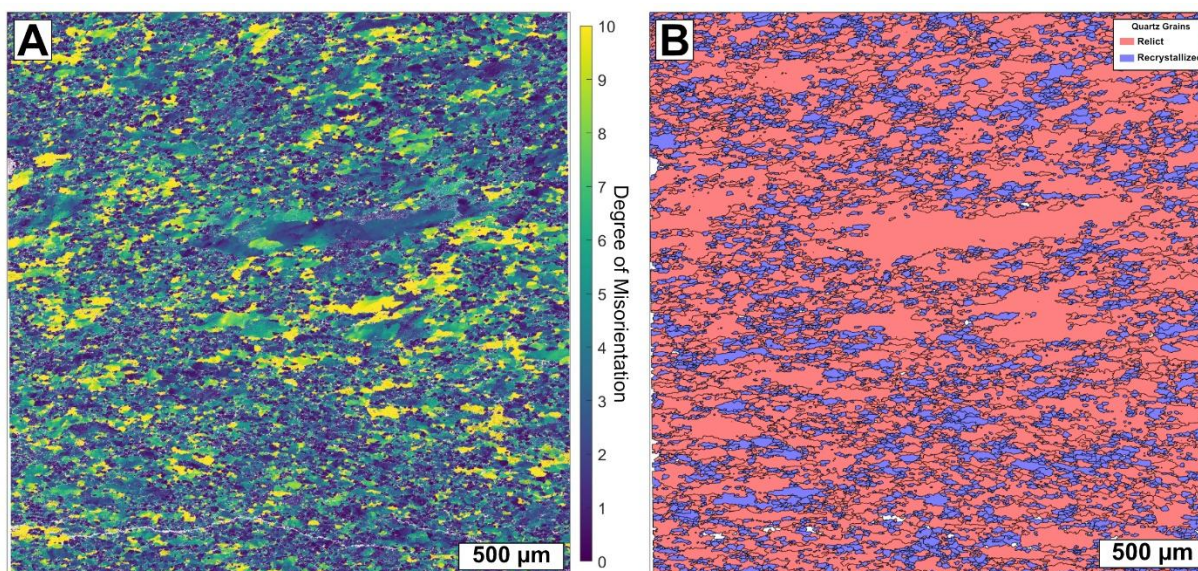


Figure 4.22: Recrystallized versus relict quartz grains within a quartz vein in a metasedimentary rock (TBN-HT-119a; 24-TBN-HT-68; 40 m from the shear zone trace): A: Electron backscatter diffraction map displaying the degree of misorientation (from 0-10 degrees) within grains with respect to the grain's average orientation. The warmer (yellow/green) colours indicate areas within a grain with a significant difference in orientation compared to the grain's average orientation. Grains with darker (purple/blue) colours indicate there is minimal difference in orientation throughout the grain. B: EBSD map displaying the differentiated relict (red) and recrystallized (blue) grains. The black lines indicate grain boundaries. (EBSD maps are oriented with north up and east right).

The recrystallized grain size for the samples was derived from the MATLAB code, which provided a root-mean-square grain size for each EBSD map area in each sample (Appendix D). To mitigate the interfering effects of quartz veins and/or ribbons, as well as smaller amounts of misclassified phases, each EBSD map was subdivided into smaller regions of interest. For the regions within one sample, the mean grain size was calculated, resulting in one average root mean

square grain size for each sample (Appendix D). Between the regions, the grain size variation was less than 15 μm , with the majority of the regions varying by less than 5 μm . The stress values for the regions were then obtained using the equation of Cross et al. (2017):

$$D = 10^{3.91 \pm 0.41} \sigma^{1.41 \pm 0.21},$$

where D is the root-mean-square diameter of recrystallized quartz grains in μm and σ is the differential stress in MPa for that grain size. In Figure 4.23, the stress values for the samples are displayed with their respective distances from the shear zone. The stress values range from 68 MPa (TBN-HT-36) to 115 MPa (TBN-HT-32; Fig. 4.32). The sample closest to the shear zone (TBN-HT-119a) has a stress value of 75 MPa, and the sample furthest from the shear zone (TBN-HT-121) has a stress value of 73 MPa (Fig. 4.23; Appendix D). The errors based on the min-max spread of the calculated stress values range from 43–70MPa for the upper errors and 18–28MPa for the lower errors, based on the errors (68% interpercentile range) for the grain size calculation (Appendix D). Overall, the samples have a consistent stress value with a median of ~80 MPa. Sample TBN-HT-32 has a higher stress value than the median; however, it is within error of the other samples (Fig. 4.23). There is no systematic increase or decrease observed in the stress values with increasing proximity to the Quetico Shear Zone, as the stresses are constant within error (Fig. 4.23).

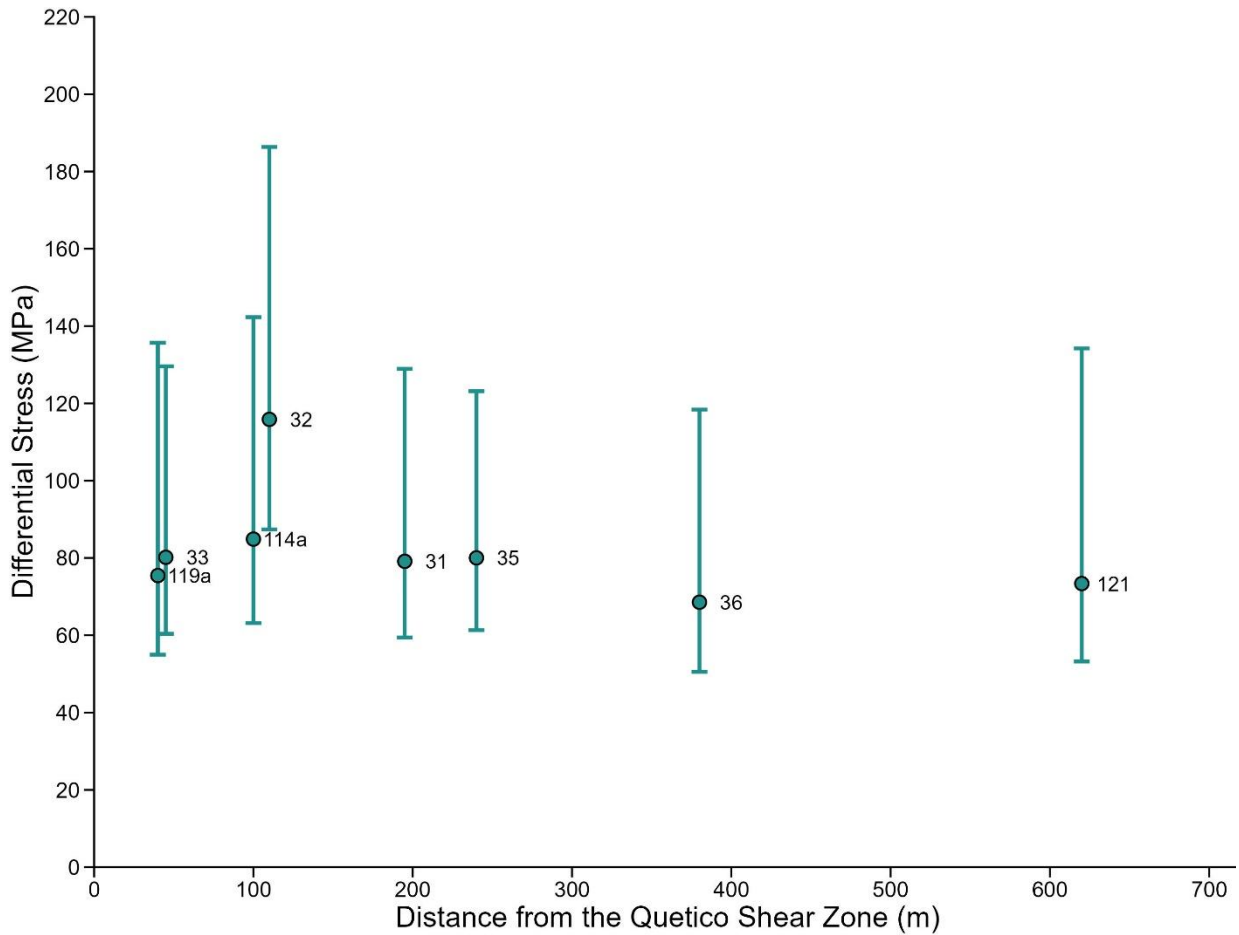


Figure 4.23: Differential stress versus distance from the Quetico Shear Zone plot: The derived differential stress value using the average recrystallized grain size of the samples is represented by the teal dots. The respective errors for each of the samples are represented by the teal error bars, which correlate to the 68% interpercentile range of recrystallized grain size values. The sample numbers (TBN-HT-XX) are located to the right of the data points.

4.3.2 Metamorphic condition constraints on Temperature and Pressure

Given the pelitic and psammitic protolith of the rocks, the metamorphic grade of the samples was determined using Barrovian and Buchan index minerals, as well as identified recrystallization textures. Overall, the metamorphic index minerals observed in thin section are andalusite, biotite, chlorite, cordierite, feldspar, garnet, muscovite, and sillimanite. In general, the index minerals correspond to greenschist to amphibolite facies metamorphic conditions. The most encountered recrystallization type was subgrain rotation, which is consistent with greenschist to amphibolite facies metamorphic conditions. The metamorphic conditions of ten samples were

quantified using phase equilibrium diagrams, modelled via THERIAK-DOMINO (de Capitani and Brown, 1987; de Capitani, 1994), using the stable metamorphic assemblages observed from petrographic analysis (Table 4.1). The stable metamorphic mineral assemblages were only differentiated from the observed mineral assemblages if a phase was not indicated to be stable within the phase equilibrium diagram (Table 4.1). For example, sample TBN-HT-36 has approximately 1 % chlorite in thin section; however, it is not indicated to be a part of the stable mineral assemblage of the phase diagram, as it likely postdates peak metamorphism (Table 4.1). Scanning electron microscopy and EDX analysis of the chlorite indicate that many of the chlorite grains show retrograde replacement of biotite (Fig. 4.24). The stable mineral assemblage for TBN-HT-36, including biotite, corresponds to a stable mineral assemblage prior to chlorite retrograde metamorphism. The stable metamorphic mineral assemblages obtained constraints of metamorphic temperature conditions for all samples, while pressure could only be determined from two samples (Table 4.1; Appendix B). Only some of the observed index minerals provided significant constraints for the temperature and pressure of metamorphism for the samples. Many of the pelitic rocks are mica schists, which cannot constrain either the temperature or pressure of the samples largely due to biotite and quartz being stable for all temperatures within the phase equilibrium diagrams (Figs. 4.25 & 4.26; Appendix B). In addition, muscovite and feldspar are stable for most common temperatures (Figs. 4.25 & 4.26; Appendix B).

Table 4.1: Stable Mineral Assemblages for suite of analyzed samples

Sample	Distance (m)*	Rock Type	Observed Stable Mineral Assemblage	Interpreted Stable Mineral Assemblage	T (°C)	P (kbar)
TBN-HT-119a	40	Metasedimentary rock	Qz, Pl, Kfs, Ms, Bt, Chl, Ttn	Qz, Pl, Kfs, Ms, Bt, Chl, Ttn	415-477	<3 - 6
TBN-HT-105	75	Metasedimentary rock	Qz, Ms, Bt, Chl, Pl, And, Ilm, Tur, Kfs, Py	Qz, Ms, Bt, Chl, Pl/Ab, And, Ilm, Kfs	538-555	<3 - 3.7
TBN-HT-114a	100	Metavolcanic rock	Qz, Pl, Ms, Kfs, Chl, Bt, Py, Cal, Rt	Qz, Pl, Ms, Kfs, Chl, Bt, Rt	430-520	—
TBN-HT-32	110	Metavolcanic rock	Pl, Qz, Bt, Ms, Chl, Py, Cal, Kfs, Rt,	Pl/Ab, Qz, Bt, Ms, Chl, Kfs, Rt	430-553	—
TBN-HT-106	150	Metasedimentary rock	Pl, Qz, Bt, Crd, Tur, Ms, Chl, Grt(Alm), Kfs, Sil, Ilm, Rt	Pl, Qz, Bt, Crd, Tur, Ms, Chl, Grt(Alm), Kfs, Sil, Ilm	640- >700	4.75 - >6
TBN-HT-31	195	Metavolcanic rock	Pl, Qz, Chl, Py, Tur, Ms, Bt, Kfs, Rt, Epi, Ttn, Prh	Pl/Ab, Qz, Chl, Ms, Bt, Kfs, Rt, Ttn	415-508	—
TBN-HT-36	380	Granitic rock	Qz, Pl, Kfs, Ms, Chl, Bt, Cal, Rt, Ilm	Qz, Pl/Ab, Kfs, Ms, Bt, Rt, Ilm	460-590	—
TBN-HT-121	630	Metasedimentary rock	Qz, Pl, Bt, Ms, Chl, Kfs, Hem	Qz, Pl, Bt, Ms, Chl, Kfs	415-570	—
TBN-HT-134	680	Metasedimentary rock	Pl, Bt, Qz, Ms, Kfs, Sil, Sil, Tur, Chl,	Pl, Bt, Qz, Ms, Kfs, Sil	615-668	3.1 - >6
TBN-HT-115	3680	Metasedimentary rock	Pl, Qz, Bt, Amp(Ath), Grt(Alm), Chl, Kfs, Ilm, Ms	Pl, Qz, Bt, Amp(Ath), Grt(Alm), Chl, Kfs, Ilm, Ms	560-605	—

*Distance from the Quetico Shear Zone

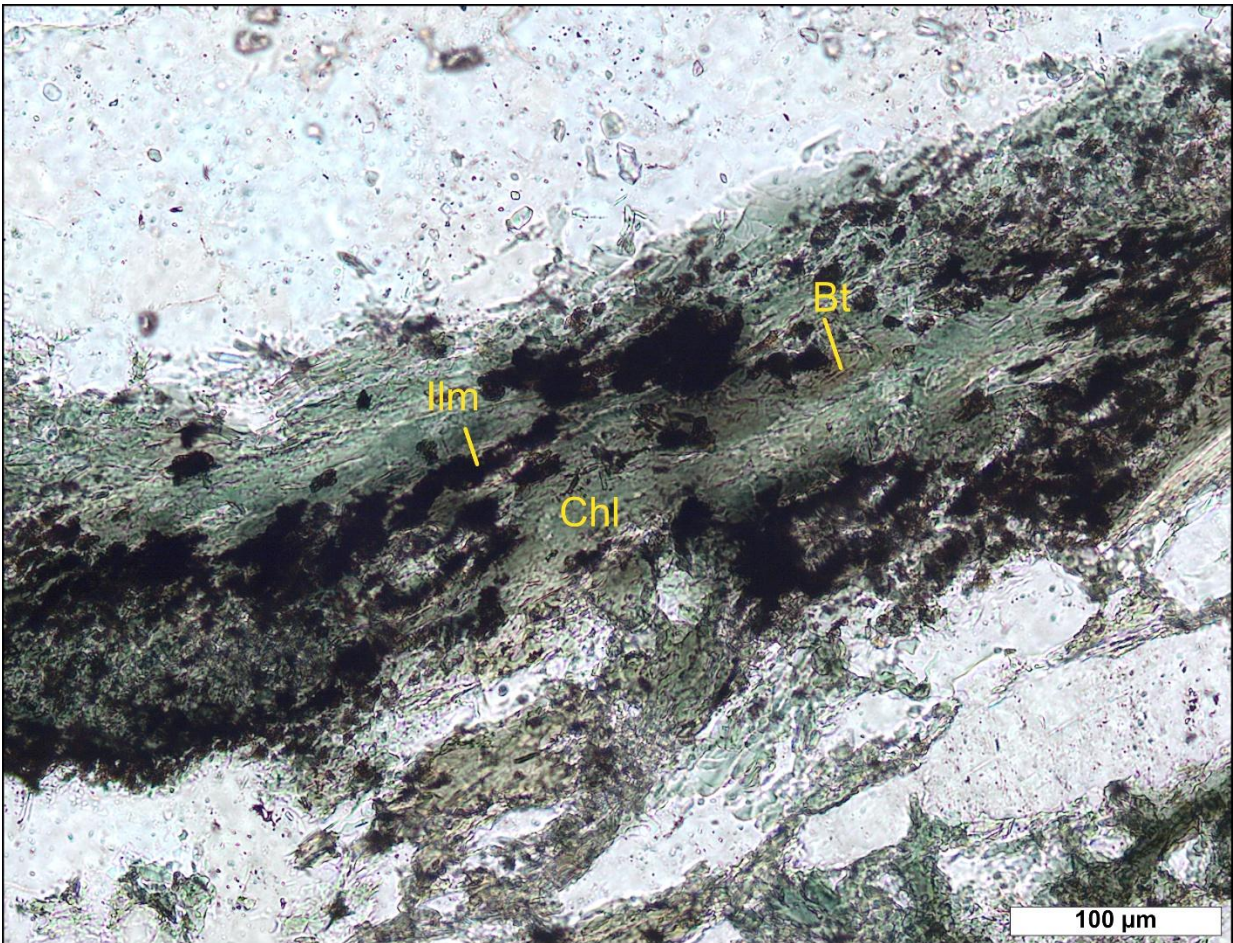


Figure 4.24: Chlorite replacement of biotite in a protomylonitic granite (photomicrograph oriented with north up and east right): Photomicrograph (PPL) displaying green chlorite with relict brown biotite patches as well as abundant oxides (ilmenite) within the chlorite (24-TBN-HT-09, TBN-HT-36, 380 m from shear zone trace).

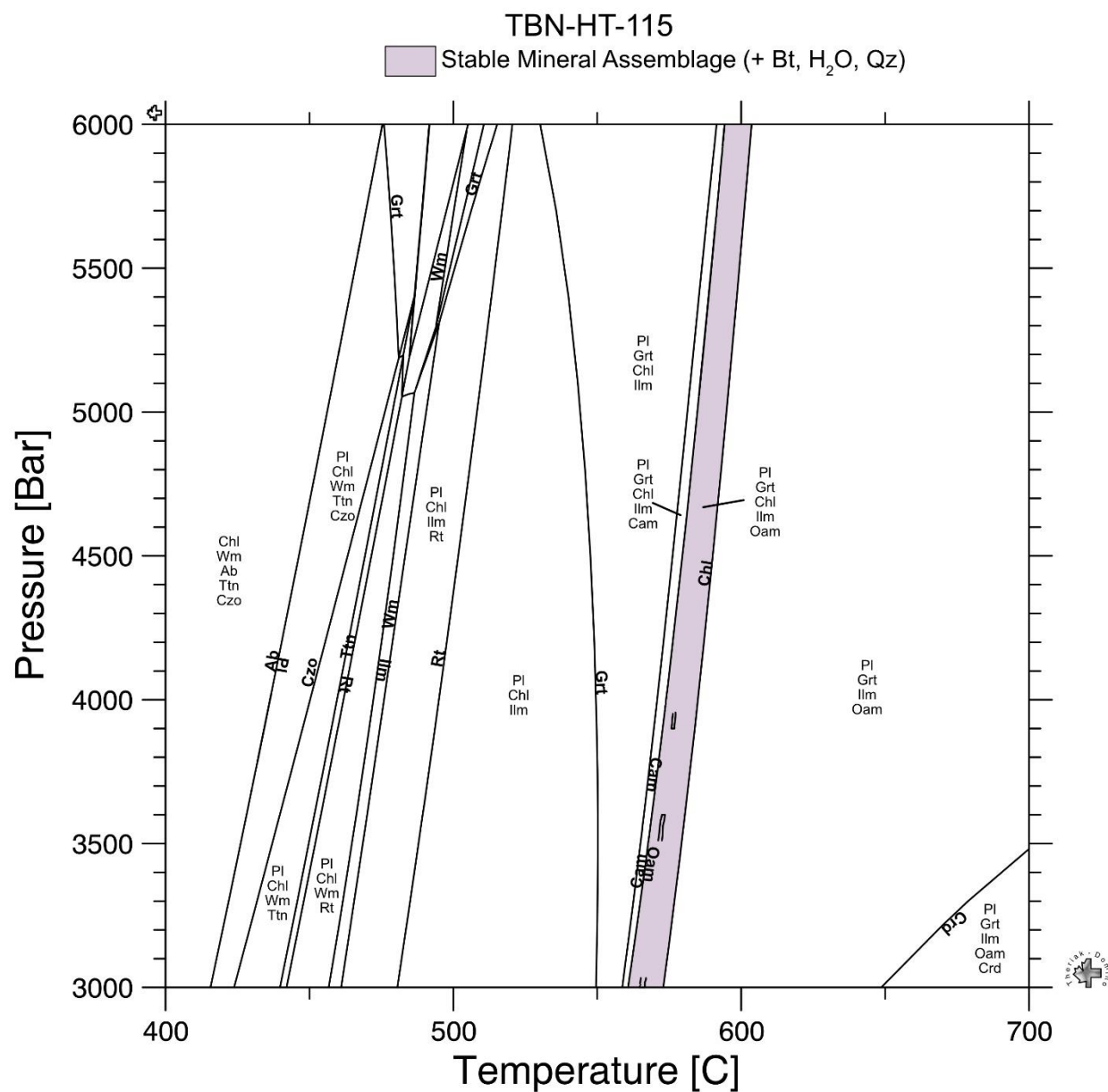


Figure 4.25: Phase equilibrium diagram for sample TBN-HT-115 (Metasedimentary rock, 3.68 km from shear zone trace). The purple polygon indicated the stable mineral assemblage (including biotite, water, and quartz) for the sample based on petrography and the bulk geochemistry of the sample.

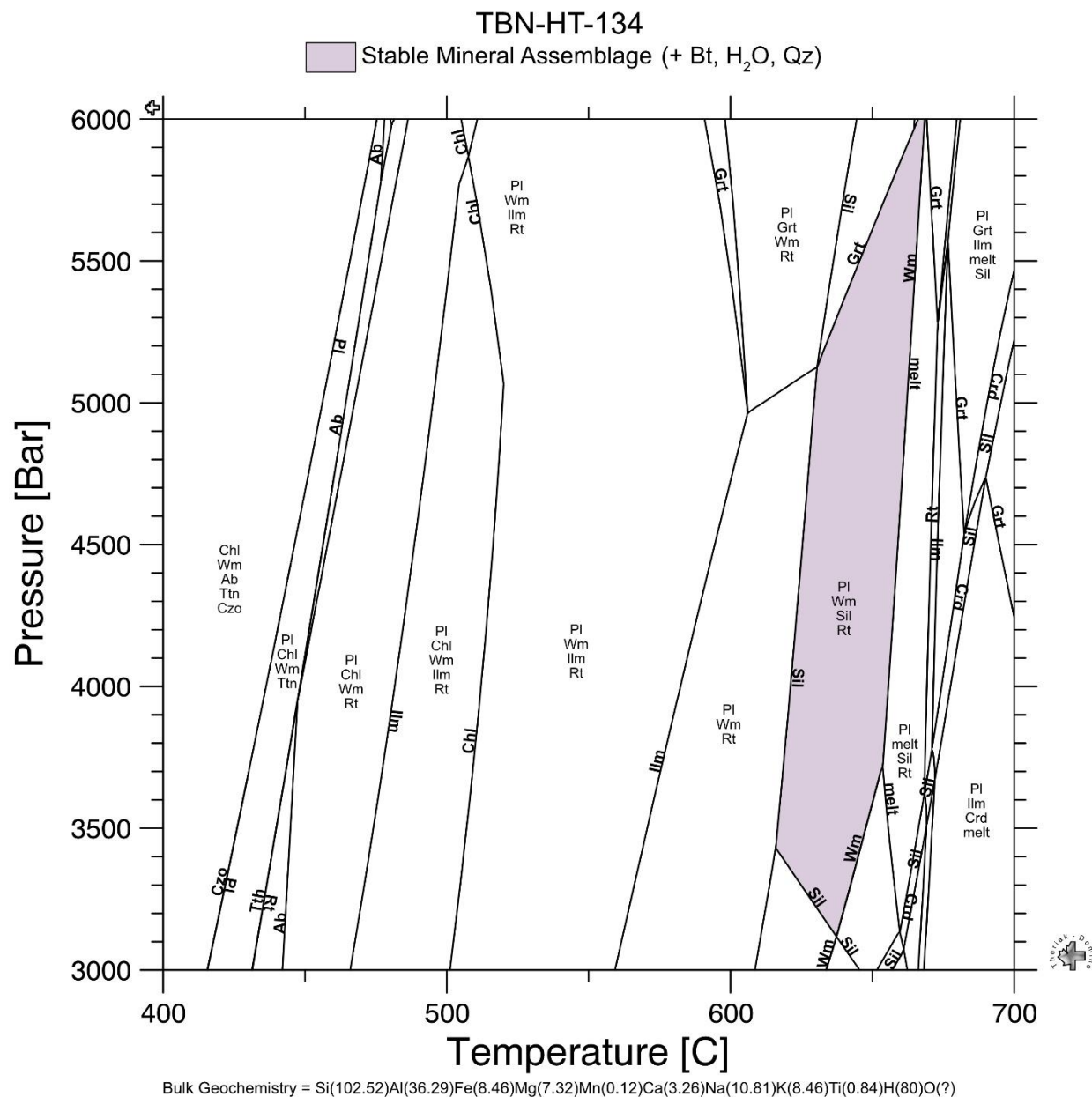


Figure 4.26: Phase equilibrium diagram for sample TBN-HT-134 (Metasedimentary rock, 680 m from shear zone trace). The purple polygon indicated the stable mineral assemblage for the sample (including biotite, water, and quartz) based on petrography and the bulk geochemistry of the sample.

Many of the reaction lines produced in the phase equilibrium diagrams are oriented parallel to the y-axis (pressure), and the stable mineral assemblages between reactions lack pressure constraints (Fig. 4.25). The resulting pressure constraints within the stable field of the phase equilibrium diagrams are less than 3 kbar or greater than 6 kbar (Fig. 4.25). For two samples (TBN-HT-119a and TBN-HT-134; Fig. 4.26), the reaction lines start to taper out with other reactions near the end of the diagrams, yielding pressures of ~3-6 kbar (Table 4.1; Fig. 4.26). In samples TBN-HT-105 and TBN-HT-106, pressures were constrained by andalusite-white mica and cordierite-sillimanite assemblages, respectively (Table 4.1; Appendix B). The abundance of index minerals in these two samples provides the upper pressure limit for sample TBN-HT-105 and the lower pressure limit for TBN-HT-106 (Table 4.1; Appendix B). In sample TBN-HT-105, the pressure range is between less than 3 kbar and ~3.7 kbar (Table 4.1; Appendix B). In samples TBN-HT-106, the pressure ranges between 4.8 kbar and greater than 6 kbar (Table 4.1; Appendix B).

The metamorphic temperatures for the samples range from 415–700 °C, with the majority of the samples plotting around 500 °C (Fig. 4.27). There is no increase or decrease in temperatures observed with respect to distance from the Quetico Shear Zone (Fig. 4.27). The closest sample to the shear zone (TBN-HT-119a) has a temperature range of 415–477 °C (Fig. 4.27). The furthest sample from the shear zone (TBN-HT-115) has a temperature range of 560–605 °C (Figs. 4.25 & 4.27). There is no systematic trend in the temperature of samples with distance to the shear zone, as the temperatures vary from high to low at all distances (Fig. 4.27). Many of the samples with moderate temperatures (400–600 °C) have weak constraints, providing a wide temperature range (Fig. 4.27). Six of the poorly constrained samples were further analyzed to constrain the temperature proximal to the shear zone.

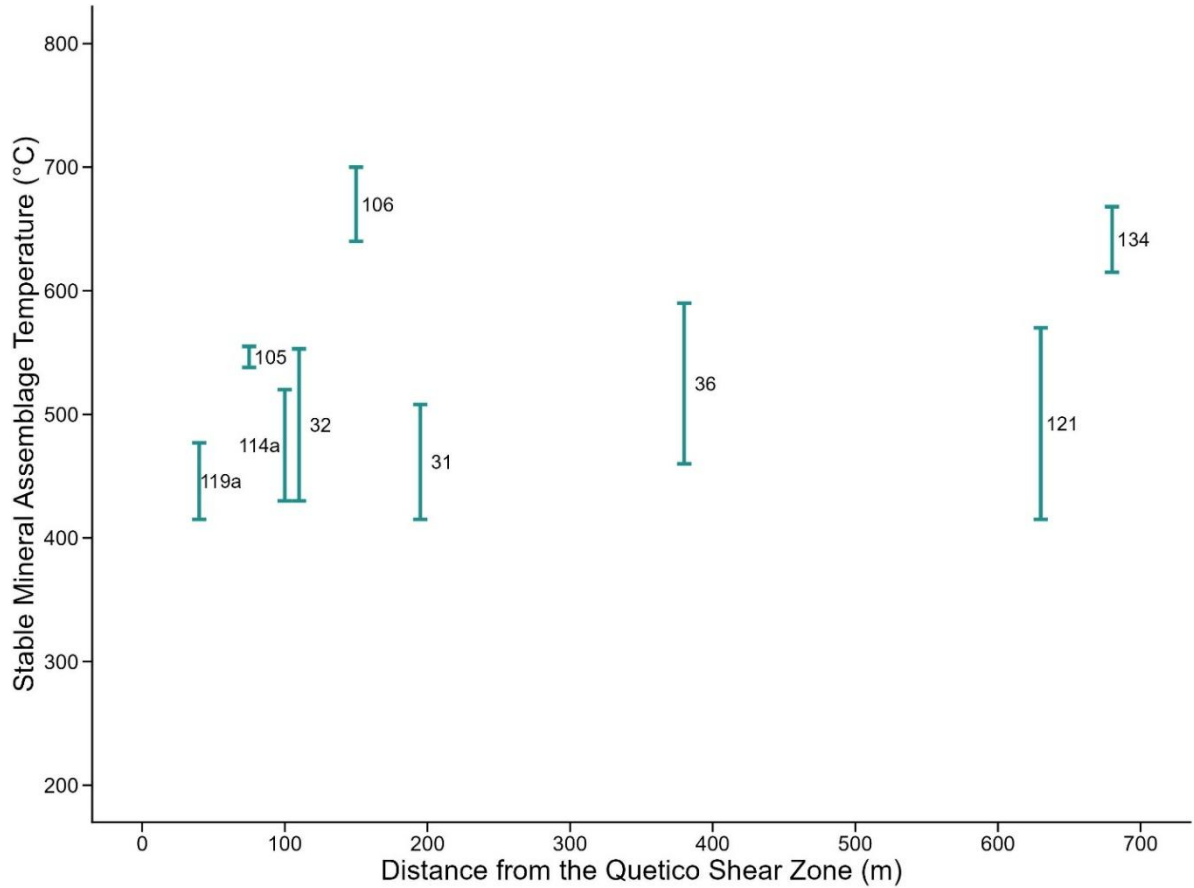


Figure 4.27: Temperature versus distance plot: The obtained temperature range of the samples from the stable mineral assemblages from petrography and the phase equilibrium diagrams is indicated by the teal error bars. The sample numbers (TBN-HT-XX) are located to the right of the temperature ranges.

4.3.3 Quartz recrystallization temperatures

The six samples proximal to the Quetico Shear Zone were analyzed using SIMS to obtain the concentration of Ti within recrystallized quartz grains, which is exponentially correlated with temperature. Areas of fine-grained quartz within quartz ribbons and shear bands were selected, usually two areas within each sample. The selected areas were analyzed petrographically using transmitted and reflected light prior to analysis to show the quartz grain boundaries in cross polarized light and to highlight pits and inclusions to avoid (Fig. 4.28). The Ti concentration was analyzed for approximately 20–30 spots within each area (Fig. 4.28). The Ti value for all the samples ranges from 3 to 20 ppm (Appendix D). The Ti values do not vary significantly within a

single sample, with values deviating by less than 10 ppm (Fig. 4.28). The core and rims of the grains were not differentiated due to the fine-grained nature of the quartz that was targeted for analysis.

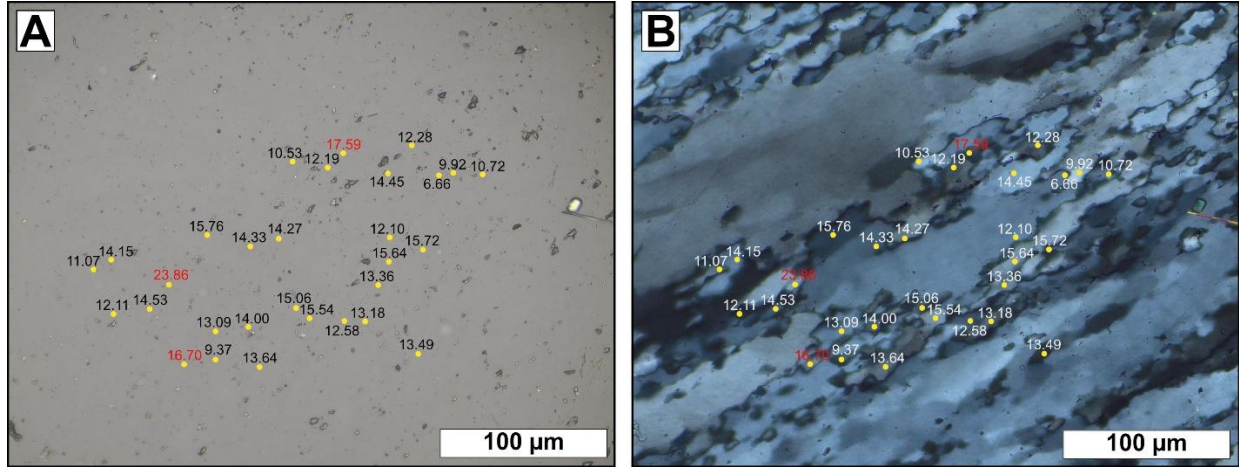


Figure 4.28: Photomicrographs of a quartz vein within a metasedimentary rock displaying SIMS spots (yellow) with their corresponding obtained Ti concentration values in ppm. The red values indicate spots that obtained erroneous values during analysis (TBN-HT-119a): A: Reflected light (PPL) displaying the abundant pits and inclusions within the quartz grains. B: Transmitted light (XPL) photomicrograph displaying the grain boundaries of quartz with SIMS spots targeted preferentially within the finer-grained recrystallized quartz.

Temperatures were obtained following the outlined procedures of Wark and Watson (2006) using the following equation:

$$T(^{\circ}\text{C}) = \frac{-3765}{\log\left[\frac{X_{Ti}^{qtz}}{a_{TiO_2}}\right] - 5.69} - 273,$$

where the temperature ($^{\circ}\text{C}$) of quartz recrystallization is exponentially correlated to Ti concentrations in quartz (X_{Ti}^{qtz}) in ppm by weight for rutile saturated rocks implying a Ti activity of 1 (Ghent and Stout, 1984). For sample TBN-HT-119a, which is rutile-undersaturated but contains titanite, a Ti activity (a_{TiO_2}) of 0.6 was used (see Behr and Platt, 2011). The obtained temperatures of the samples range from 457 to 589 $^{\circ}\text{C}$ (Fig. 4.29). There is no systematic increase or decrease in temperature with distance from the Quetico Shear Zone (Fig. 4.29). The sample closest to the shear zone has a temperature of 549 $^{\circ}\text{C}$, and the most distal sample has a temperature

of 584 °C (Fig. 4.29). The remainder of the samples between the most proximal and most distal samples have obtained temperatures that range from 410 to 515 °C (Fig. 4.29). The temperatures were also calculated using an equation from Thomas et al. (2010), indicating a pressure dependence:

$$T(^{\circ}\text{C}) = \frac{a+cP}{b-R \times \ln X_{\text{TiO}_2}^{\text{qtz}} + R \times \ln a_{\text{TiO}_2}} - 273.15,$$

where a, b, and c are the least-squares fit parameters derived in Thomas et al (2010; a is $60,952 \pm 3,177$, b is 1.52 ± 0.39 , and c is $1,741 \pm 63$). The temperature range from the pressure dependence is indicated within the phase equilibrium diagrams (Appendix B).

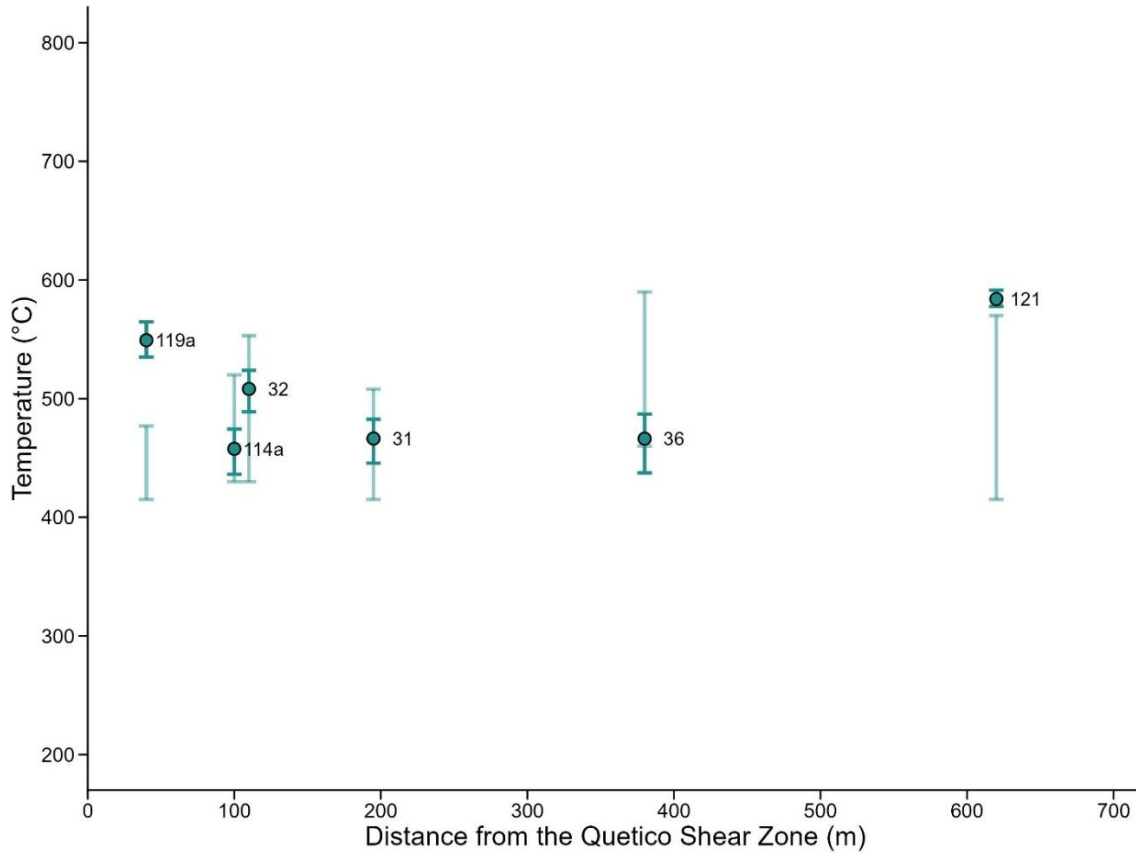


Figure 4.29: Temperature versus distance plot: The obtained temperatures from SIMS Ti-in-quartz thermometry for each sample are indicated by the teal dots. The respective errors for each of the samples are represented by the teal error bars, which correlate to the range in Ti values based on their 2-sigma standard deviation. The stable mineral assemblage temperatures for the samples are indicated by the translucent temperature ranges. The sample numbers (TBN-HT-XX) are located to the right of the data points.

The Ti-in-quartz temperatures were compared to the temperature ranges from the phase equilibrium diagrams (Fig. 4.29; Appendix B). The Ti-in-quartz temperatures plot within the same range as the stable mineral assemblages (Fig. 4.29; Appendix B). For example, sample TBN-HT-114a, the pseudotachylite bearing sample has a stable mineral assemblage of quartz, plagioclase $\pm$ albite, muscovite, potassium feldspar, chlorite, biotite and rutile (Table 4.1; Fig. 4.30). The Ti-in-quartz temperatures correlate to the temperature range from the stable mineral assemblages until 4.3 kbar (Fig. 4.30). The temperatures using the Thomas et al. (2010) thermometer plot up to 50 °C below Wark and Watson (2006) Ti-in-quartz temperature (Fig. 4.30). The only two samples that do not align with their stable mineral assemblages are samples TBN-HT-119a and TBN-HT-121 (Fig. 4.29). For sample TBN-HT-119a, the Ti-in-quartz was constrained using a Ti activity of 0.6 (Behr and Platt, 2011) due to the presence of titanite and the absence of rutile. For sample TBN-HT-121, there was no Ti phase observed within the sample, so a Ti activity of 1 was used, indicating that the resultant Ti-in-quartz temperature is a minimum estimate (Wark and Watson 2006). The stable mineral assemblage for sample TBN-HT-121 consists of quartz, plagioclase, biotite, muscovite, chlorite, and potassium feldspar (Table 4.1; Fig. 4.31). The phase equilibrium diagram is constrained by plagioclase feldspar and white mica; however, it indicates that titanite, rutile, or ilmenite should be stable, which was not observed (Fig. 4.31). The temperatures using pressure dependence for samples TBN-HT-119a and TBN-HT-121 plot within the temperature range of the stable mineral assemblages (Fig. 4.31; Appendix B). The temperatures from Thomas et al. (2010) do not correlate with either the Wark and Watson (2006) or the phase equilibrium diagrams for most of the samples (Figs. 4.30 & 4.31; Appendix B).

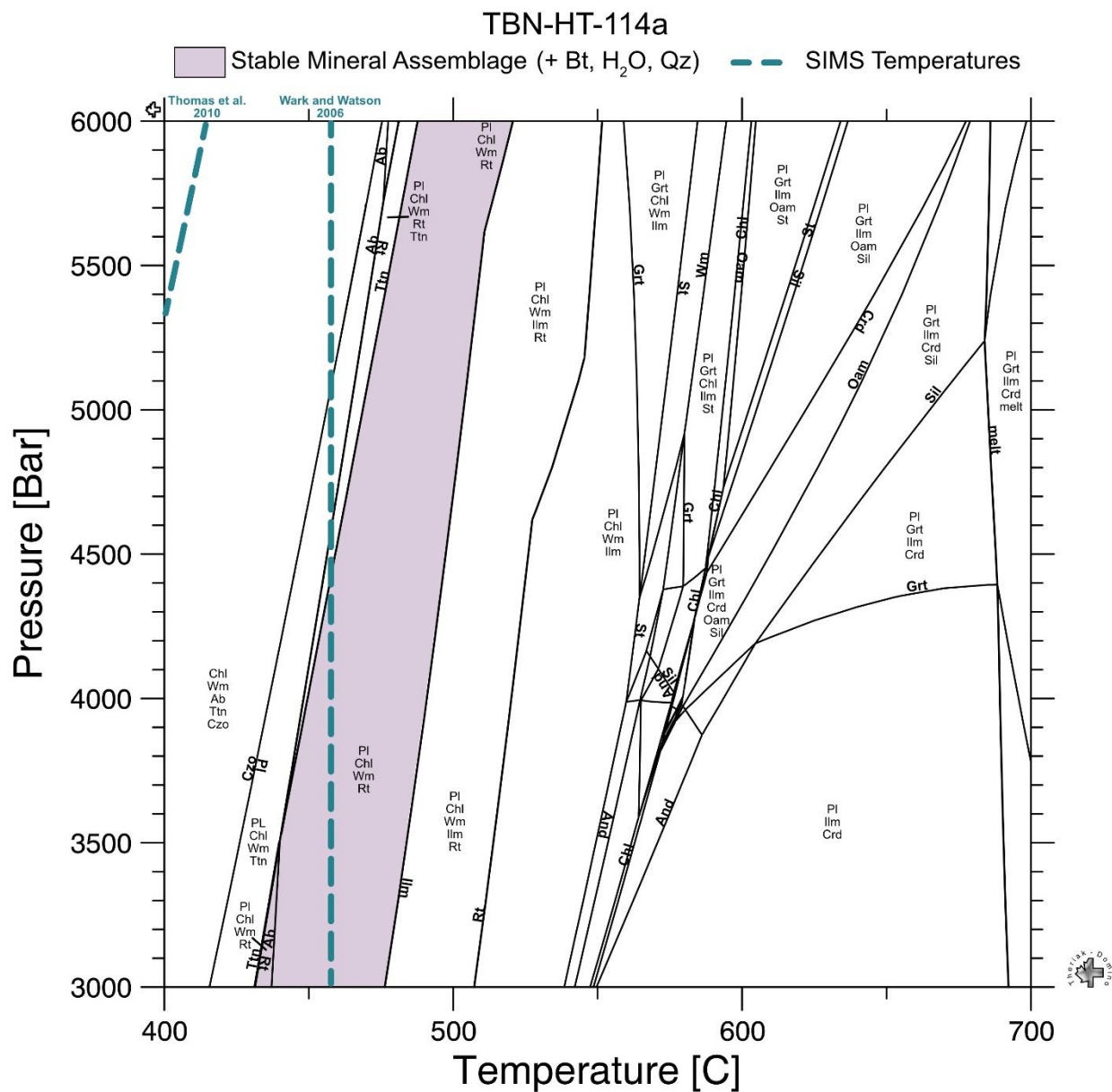


Figure 4.30: Phase equilibrium diagram for sample TBN-HT-114a (Metavolcanic rock, 100 m from shear zone trace). The purple polygon indicated the stable mineral assemblage for the sample (including biotite, water, and quartz) based on petrography and the bulk geochemistry of the sample. The blue dotted lines indicate the temperature values obtained from Ti-in-quartz thermometry using equations from Wark and Watson (2006) and Thomas et al. (2010).

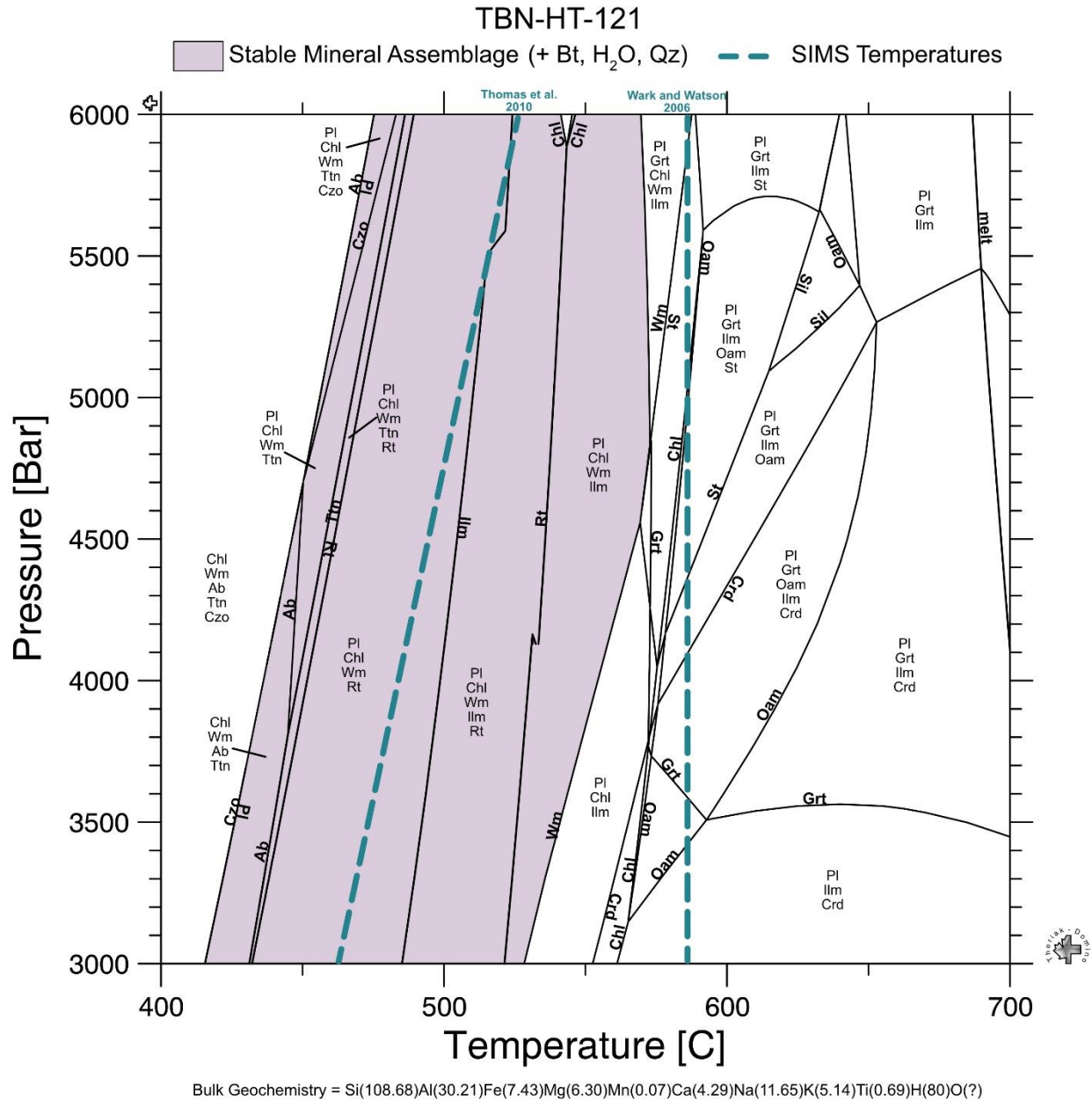


Figure 4.31: Phase equilibrium diagram for sample TBN-HT-121 (Metasedimentary rock, 620 m from shear zone trace). The purple polygon indicated the stable mineral assemblage for the sample (including biotite, water, and quartz) based on petrography and the bulk geochemistry of the sample. The blue dotted lines indicate the temperature values obtained from Ti-in-quartz thermometry using equations from Wark and Watson (2006) and Thomas et al. (2010).

4.4 Timing of Deformation

The timing of deformation within the study area was investigated using in situ laser ablation split stream (LASS) U-Th-Pb (through MC-ICP) and trace-element (through SC-ICP) analyses for various datable minerals. The datable minerals that were analyzed include apatite, monazite, titanite, and zircon. The seven samples analyzed comprise various rock types, including metasedimentary, metavolcanic, and granitic rocks. The samples varied in distance from the Quetico Shear Zone trace, ranging from 40 to 435 m. Each mineral was characterized, and its relation to microstructures was recorded, with emphasis on grains that lay within foliation planes/sheared regions. Backscattered electron imaging was used to further characterize each grain in terms of zoning and to determine locations for spot analyses. Selected grains were analyzed with a selected number of spots based on the grain's size and zonation. The U-Pb data were plotted using Concordia diagrams, while the trace elements were plotted as chondrite-normalized rare earth element (REE) plots.

4.4.1 Microstructures of datable minerals

4.4.1.1 Apatite

Apatite commonly occurs within all the samples. The apatite is fine-grained, with most grains being approximately 250 μm , with a rounded or, rarely, a subhedral prismatic habit. The analyzed apatite is located within the foliated groundmass of samples, shear bands, and cataclasite, and occurs as inclusions within other minerals (Fig. 4.32A, B). The groundmass apatite is commonly adjacent to feldspar, biotite, or quartz (Fig. 4.32A, B). The apatite is usually aligned with the foliation, depending on its extent of elongation. The grains either are parallel to the main foliation of the sample or follow the orientation of smaller domains, usually similar to adjacent larger grains (Fig. 4.32A, B). Some apatite are also located within, or associated with, thin shear

bands, usually consisting of mica or recrystallized quartz. Some apatite grains are found within cataclasite zones, similar to the shear bands; they form larger zones of localized deformation (Fig. 4.32A). In a sample 40 m from the shear zone (TBN-HT-119a), a 2 mm cataclasite zone contains abundant fine-grained apatite (Fig. 4.32A). The apatite can also commonly form inclusions within other minerals, usually feldspar and hornblende (Fig. 4.32B). The apatite commonly has minimal inclusions, with only a few grains containing abundant inclusions (Fig. 4.32C). The grains overall are intact, but a few are significantly broken down. Most grains have no zonation discernible in the SEM, but a few display weak zonation representing changing conditions during growth (e.g., Fig. 4.32D).

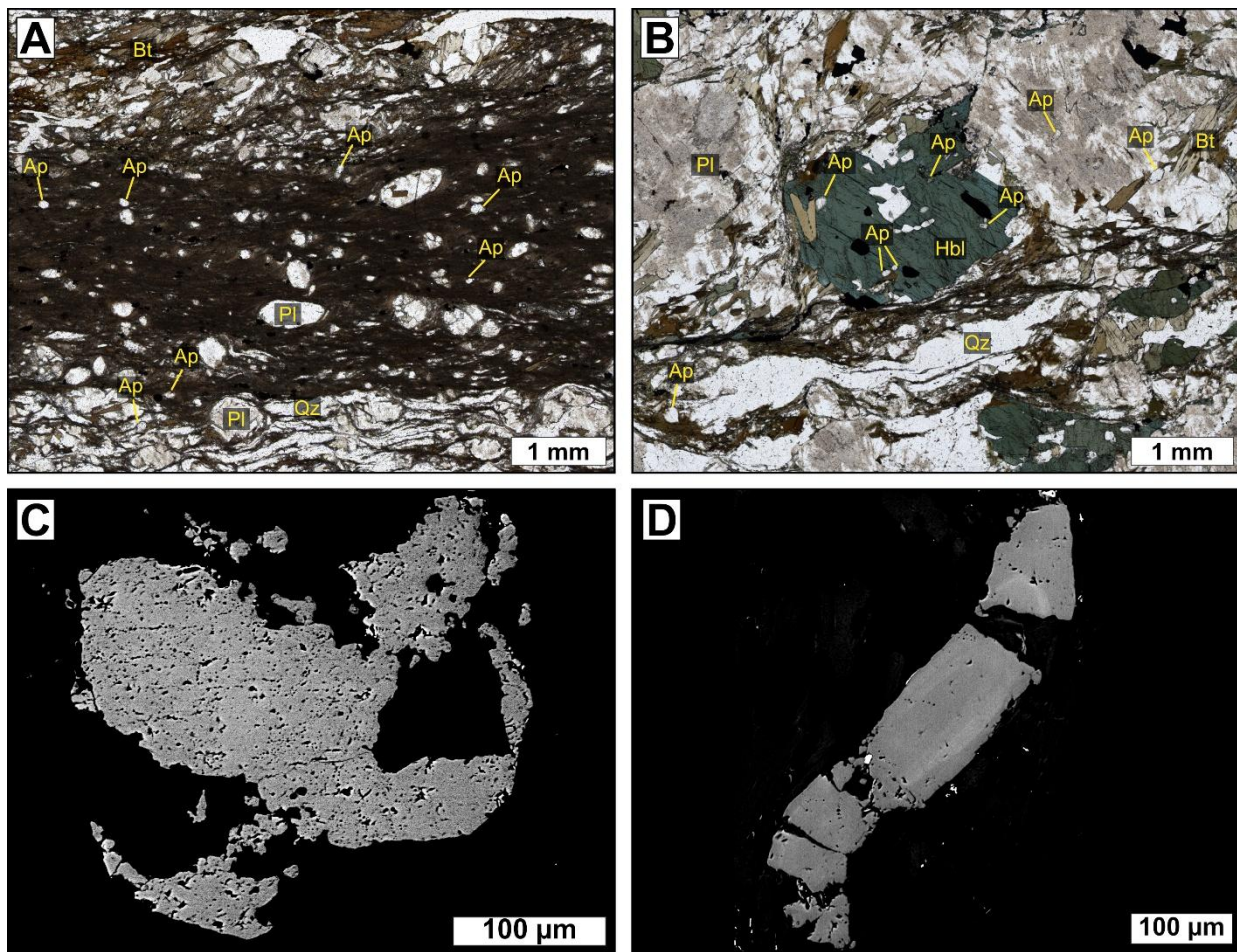


Figure 4.32: Photomicrographs (PPL) displaying the textures of apatite grains within metasedimentary rocks (all photomicrographs are oriented with north up and east to the right): A: Numerous apatite grains within a dark, 2 mm wide zone of cataclasite. The oblate to elongate apatite grains are oriented parallel with the foliation (TBN-HT-119c; 24-TBN-HT-68). B: Apatite occurring as inclusions within hornblende and feldspar also within quartz ribbons (TBN-HT-117; 24-TBN-HT-66). Backscatter electron images of apatite: C: An anhedral apatite grain riddled with extremely fine-grained inclusions (TBN-HT-119c; 24-TBN-HT-68). D: An elongate apatite grain with weak zonation (TBN-HT-117; 24-TBN-HT-66).

4.4.1.2 Monazite

Monazite occurs within most samples and is most abundant within the granitic samples. The grains are usually very fine-grained, approximately 50 μm , with some grains being up to 200 μm in size (Fig. 4.33). They occur as anhedral, rounded grains or as subhedral, tabular to prismatic grains. The elongate grains are oriented parallel to the foliation (Fig. 4.33). The analyzed monazite grains are located within shear bands, often associated with biotite and feldspars or enclosed within minerals such as feldspar or quartz ribbons (Fig. 4.33A, B). The grains are commonly associated with micas, especially muscovite, either located adjacent to muscovite or enclosed in muscovite (Fig. 4.33A). Monazite is also commonly associated with feldspar, again either adjacent to or as inclusions within feldspar (Fig. 4.33A). Monazite also occurs within recrystallized quartz ribbons (Fig. 4.33B). Many of the monazite grains are strongly broken down (Fig. 4.33C D), with only very, usually smaller grains remaining intact. The grains either start to break down at the rims with intact cores or intact rims with broken down cores (Fig. 4.33C, D). The grains commonly have inclusions (Fig. 4.33C, D) and are frequently weakly zoned with at least one notable rim (Fig. 4.33C, D).

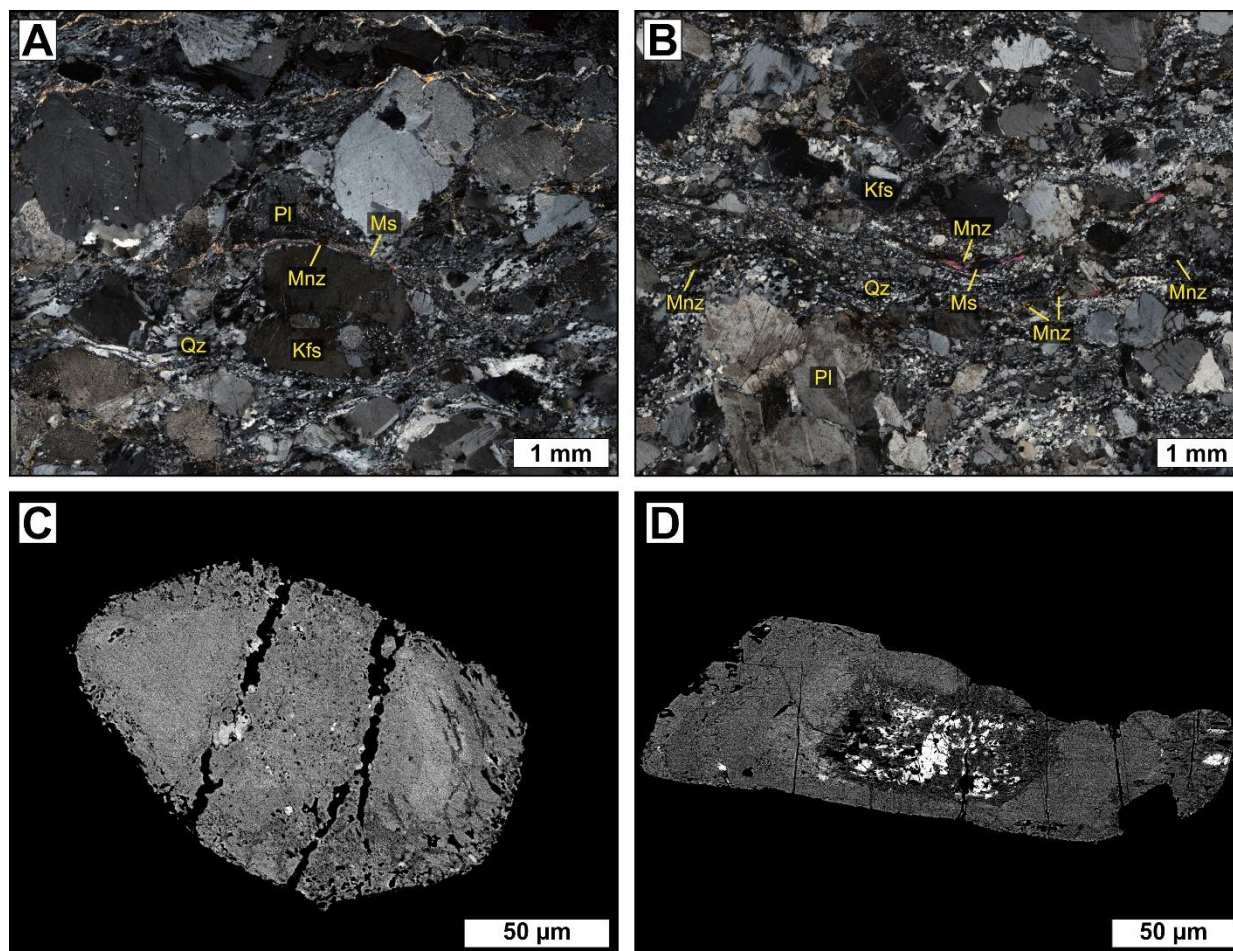


Figure 4.33: Photomicrographs (XPL) displaying the textures of monazite within protomylonitic granitic rocks (all photomicrographs are oriented with north up and east to the right): A: A monazite grain within a thin muscovite band adjacent to plagioclase and potassium feldspar grains (TBN-HT-35; 24-TBN-HT-8). B: Monazite grains within a quartz ribbon, the grains are associated with the quartz directly or within muscovite bands (TBN-HT-36; 24TBN-HT-9). Backscatter electron images of monazite grains: C: A subhedral prismatic monazite grain with weak zonation proximal to the rim and significant decay along the rim (TBN-HT-36; 24TBN-HT-9). D: A subhedral elongate monazite grain with significant break down of the core and some zonation (TBN-HT-35; 24-TBN-HT-8).

4.4.1.3 Titanite

Only one sample has a significant amount of titanite, an amphibole-rich metasedimentary rock (TBN-HT-117; Fig. 4.34). The titanite grains are usually fine-grained, with most grains being approximately 150 μm (Fig. 4.34). The grains are commonly anhedral, with some grains having a euhedral wedge habit, and do not commonly follow the overall foliation of the sample but adjacent minerals follow the overall foliation (Fig. 4.34A, B). The titanite grains are commonly associated

with hornblende or with dark, very fine-grained biotite and trace minerals (Fig. 34A, B). Some analyzed titanite grains are inclusions within hornblende grains, usually near the grain boundary, or are adjacent to hornblende grains, sometimes in strain shadows (Fig. 4.34A). Titanite grains are sometimes located within/adjacent to feldspar grains and quartz ribbons (Fig. 4.34A, B). Many of the grains are associated with dark, very fine-grained biotite and trace minerals, sometimes forming shear bands between coarse-grained minerals (Fig. 4.34A, B). The grains are commonly strongly deformed, with many of the grains being broken apart (Fig. 4.34C). The grains also have a moderate number of inclusions (Fig. 4.34C, D). The grains are commonly zoned with a very patchy zonation (i.e., the zonation is not continuous or parallel to the grain boundaries; Fig. 4.34C, D).

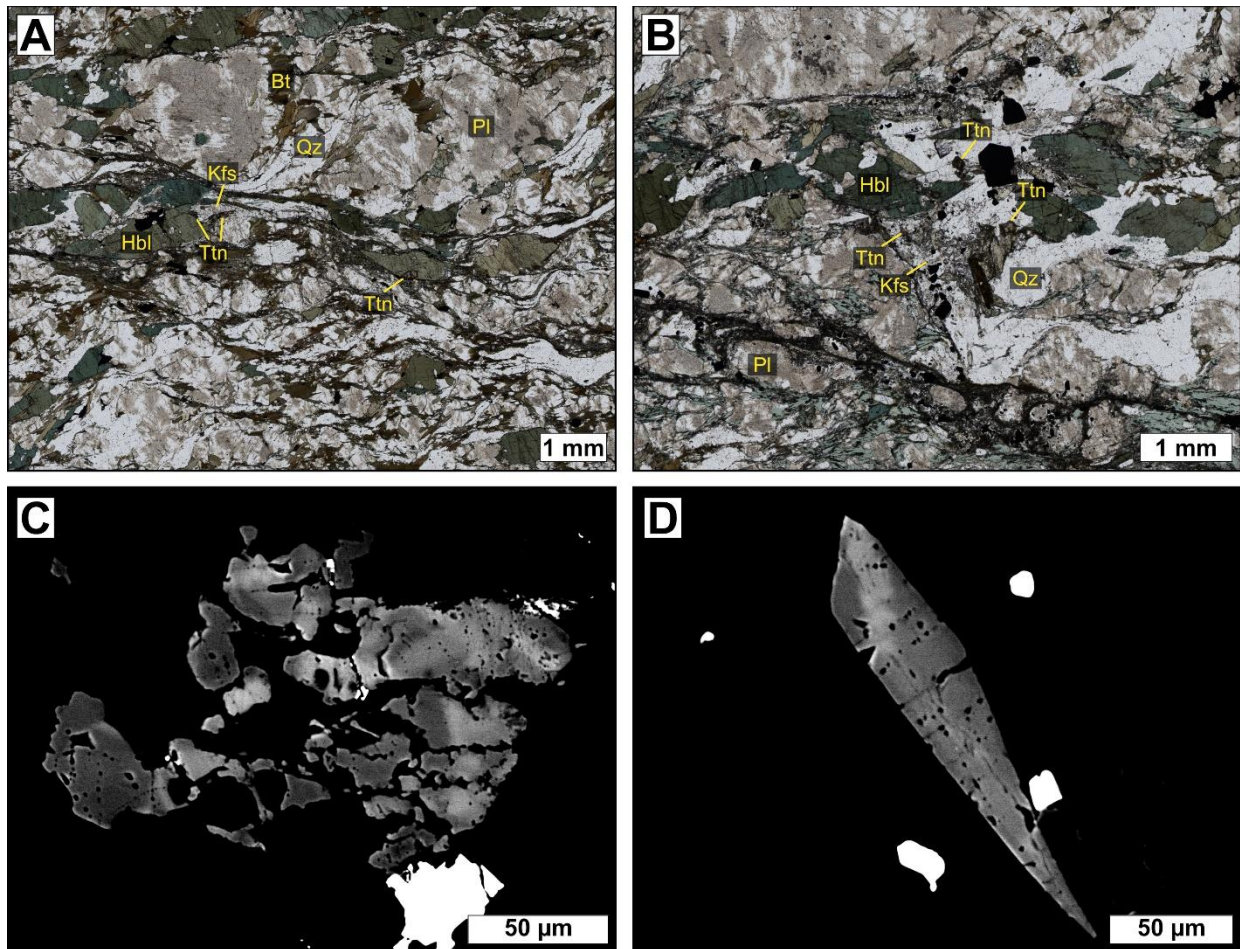


Figure 4.34: Photomicrographs (PPL) displaying the textures of titanite within a metasedimentary rock (TBN-HT-117; 24TBN-HT-66; all photomicrographs are oriented with north up and east to the right): A: Grains adjacent to deformed hornblende grains proximal to biotite shear bands and quartz ribbons. B: Grains within quartz displaying various orientations often oblique to the foliation. Backscattered electron images of titanite grains: C: Fragmented titanite grain with inclusions. D: Euhedral wedge shaped titanite grain with inclusions and cracks.

4.4.1.4 Zircon

Zircon occurs within all samples; however, the number of suitable zircons in the samples was strongly limited by their size. The zircon are usually very fine-grained, with most of the zircons being approximately 50 μm and only a few grains occurring up to 200 μm (Fig. 4.35), especially the smaller grains usually have an anhedral, rounded habit, with some grains having a subhedral to euhedral prismatic habit (Fig. 4.35). The elongate zircons do align with the orientation of the foliation within the samples. The analyzed zircons are located within recrystallized quartz ribbons, within thin mica bands, and associated with feldspars (Fig. 4.35A, B). Zircon grains are commonly associated with recrystallized quartz ribbons proximal to the shear zone, either adjacent to the edge of the ribbon or within the ribbons, usually forming very fine-grained zircons (Fig. 4.35A, B). The zircon are also located within thin mica bands, usually biotite; these mica bands are sometimes along the boundary of a quartz ribbons or within the quartz ribbons (Fig. 4.35A). In some samples, the zircon are associated with feldspars, either as inclusions or along grain boundaries (Fig. 4.35A). Zircon rarely contain inclusions, but are often fractured (Fig. 4.35C, D). Some of the zircons have broken down, often with the core being the most broken down (Fig. 4.35D). Most of the zircon exhibit strong zonation, with only a few displaying weak to no internal zonation (Fig. 4.35C, D). Some zircons show strong oscillatory zonation (Fig. 4.35C), while a few grains have a more distinct core and rim (Fig. 4.35D).

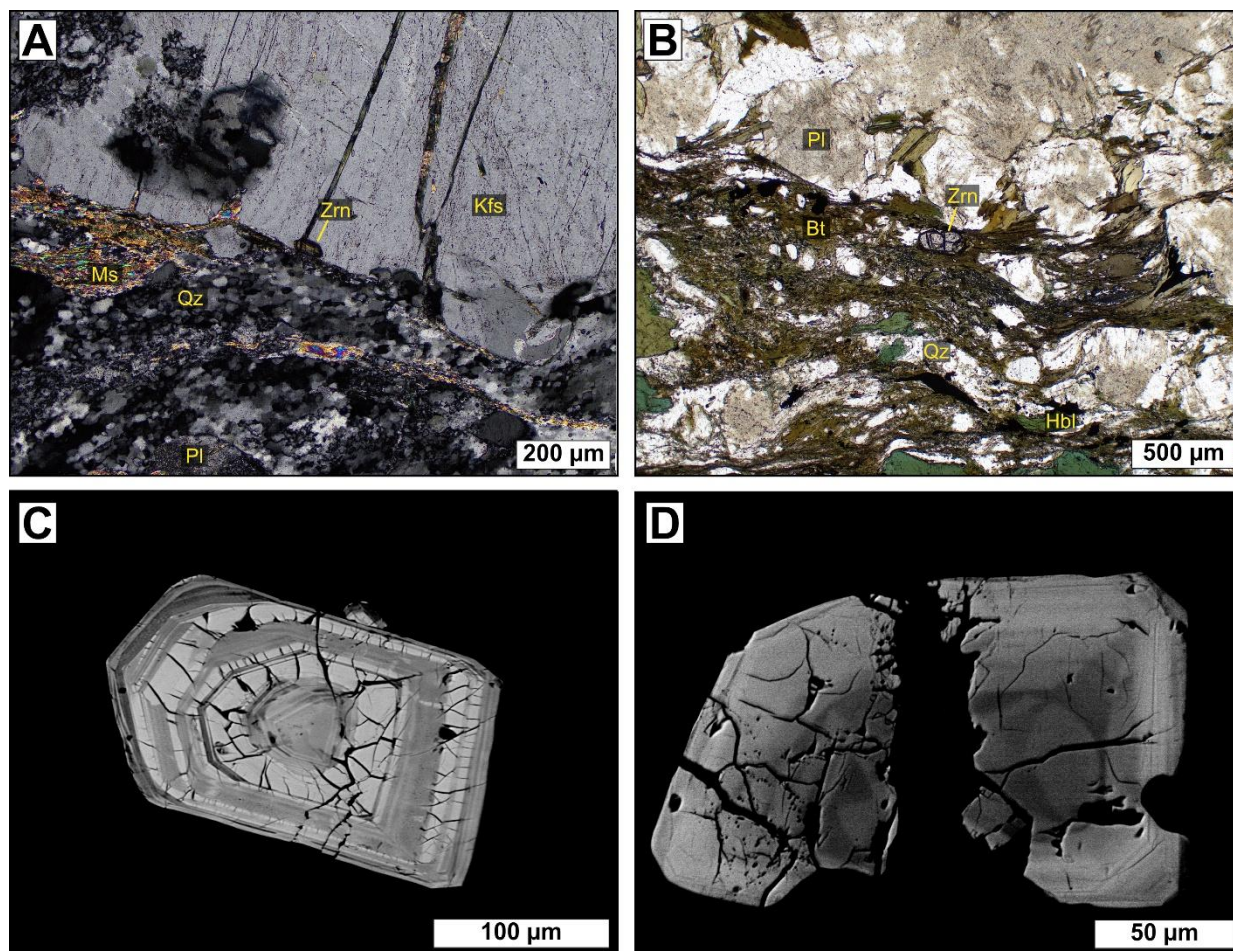


Figure 4.35: Photomicrographs displaying the textures of zircon (all photomicrographs are oriented with north up and east to the right): A: Zircon along the grain boundary of a coarse-grained potassium feldspar, adjacent to a quartz ribbon within a mylonitic granitic rock (XPL; TBN-HT-35; 24TBN-HT-8). B: Zircon within a biotite band, oriented parallel with the foliation (PPL; TBN-HT-117; 24-TBN-HT-66). Backscattered electron images of zircon grains: C: Euhedral prismatic zircon grain with prominent oscillatory zonation and numerous cracks (TBN-HT-118; 24TBN-HT-66). D: A fragmented subhedral prismatic zircon grains with moderate zonation, a distinct core and numerous cracks (TBN-HT-117; 24-TBN-HT-66).

4.4.2 Trace Element Chemistry and Geochronology

The trace element mineral chemistry and geochronology data of the samples are discussed by mineral type. The trace elements were normalized to chondrite values obtained from Sun and McDonough (1989) (Figs. 4.36, 4.38, 4.40, & 4.42). The $\text{La}/\text{Sm}_{(n)}$ vs. $\text{Gd}/\text{Yb}_{(n)}$ ratios for each spot were also plotted on bivariate plots (Appendix E). The corresponding $^{207}\text{Pb}/^{206}\text{Pb}$ date of each spot is noted within the trace element plots. The obtained radiogenic isotope data were plotted on Tera-

Wasserburg concordia diagrams (Figs. 4.37, 4.39, 4.41, & 4.43). The interpreted ages for each of the samples are reported in Table 4.2. Outliers were removed from the age calculations (outliers are visible in the concordia plots) based on three criteria: laser spot accuracy, concordance values, and statistical outlier detection. The accuracy of the laser spots was analyzed based on post-laser documentation; laser spots that drifted off the grain were removed. For minerals where spots are predominantly concordant, analyses with concordance values not within the range of 0.95-1.05 were removed. Using the modified Chauvenet outlier detection criterion built into IsoplotR, outliers are detected based on analytical error and geological scatter based on the weighted mean ages; outliers were removed only for minerals where analyses are predominantly concordant.

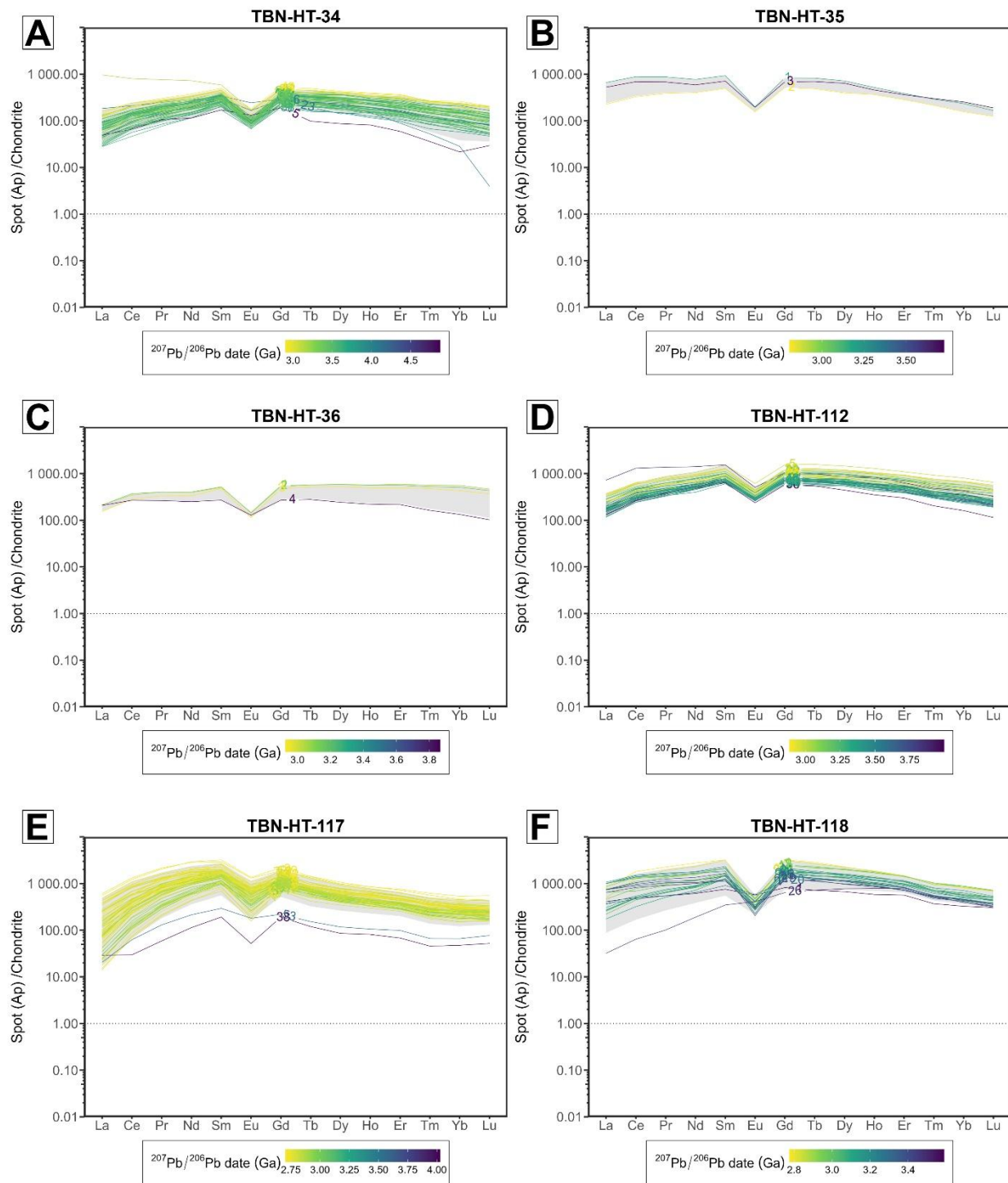
Table 4.2: LASS U-Pb ages

Sample	Rock Type	Distance (m)*	Mineral # grains	Age	$\pm 2\sigma$	$\sqrt{\text{MSWD}} \pm 2\sigma$	Age Type
TBN-HT-119c	Metasedimentary	40	Ap	2598	16	26	Lower Intercept
			22 grains	3245	105		Mean
			Mnz	2676	17	33	Concordia
			2 grains	2649	25		Mean
			Zrn	2707.5	8.1	47.0	Concordia
			10 grains	2748	64		Mean
TBN-HT-34	Metavolcanic	45	Ap	2601	28	64	Lower Intercept
			22 grains	3431	70		Mean
			Mnz	2662	22	35	Concordia
			1 grain	2650	24		Mean
			Zrn	2690	11	58	Concordia
			5 grains	2708	67		Mean
TBN-HT-35	Granite	240	Ap	2707	61		Lower Intercept
			3 grains	3209	430		Mean
			Mnz	2640.1	3.6	5.5	Concordia
			15 grains	2642.0	4.2		Mean
			Zrn	2374	11	77	Concordia
			7 grains	2421	64		Mean
TBN-HT-112	Metasedimentary	250	Ap	2647	24	63	Lower Intercept
			21 grains	3301	84		Mean
			Mnz	2659.4	5.3	11.3	Concordia
			21 grains	2638.0	6.4		Mean
			Zrn	2770.8	8.8	23.9	Concordia
			10 grains	2777	30		Mean
TBN-HT-36	Granite	380	Ap	2744	51	127	Lower Intercept
			2 grains	3218	349		Mean
			Mnz	2648.5	3.8	6.7	Concordia
			20 grains	2648.1	4.6		Mean
TBN-HT-117	Metasedimentary	435	Ap	2650	12	17	Lower Intercept
			26 grains	2803	25		Mean
			Ttn	2620	17	24	Lower Intercept
			12 grains	2945	125		Mean
			Zrn	2692.2	6.5		Concordia
			5 grains	2690.3	7.6		Mean
TBN-HT-118	Metasedimentary	435	Ap	2626	30		Lower Intercept
			12 grains	3122	85		Mean
			Zrn	1670.3	5.4	94.0	Concordia
			23 grains	1928	107		Mean

*Distance from the Quetico Shear Zone

4.4.2.1 Apatite

The majority of the trace element patterns for apatite are characterized by consistent, enriched trace element patterns, with the metasedimentary rocks being the most enriched relative to chondrite (Fig. 4.36). The majority of the analyzed grains are characterized by a pronounced negative Eu anomaly; however, three analyses display positive Eu anomalies in sample TBN-HT-119c, which correspond to analyses with the oldest $^{207}\text{Pb}/^{206}\text{Pb}$ dates within the sample (dates greater than 3.5 Ga; Fig. 4.36G). Three analyses within samples TBN-HT-34 and TBN-HT-118 display weak negative Eu anomalies, which also correspond to analyses with older $^{207}\text{Pb}/^{206}\text{Pb}$ dates (Fig. 4.36A, E). The Eu/Eu* values within all the samples range from 0.11 to 5.81 with a median of 0.32. The apatite displays fractionated light rare earth element (LREE) patterns with weak to moderately positive slopes, with two analyses displaying weak negative slopes in samples TBN-HT-34 and TBN-HT-119c (Fig. 4.36A, F). The La/Sm_(n) values for all the samples range between 0.02 and 1.64, with a median of 0.376 (Fig. 4.36). The heavy rare earth element (HREE) patterns are weakly fractionated with a weak negative slope and Gd/Yb_(n) values between 0.98 and 10.31 with a median of 2.46 (Fig. 4.36; Appendix E). The La/Sm_(n) vs Gd/Yb_(n) ratios for most of the samples are well clustered within a sample, with the La/Sm_(n) range deviating by less than 0.5 and the Gd/Yb_(n) range deviating by less than 3 (Fig. 4.36; Appendix E). Two samples, TBN-HT-34 and TBN-HT-119c, have a greater spread of La/Sm_(n) values up to 1.52, and TBN-HT-34 has a greater spread of Gd/Yb_(n) values up to 8.47 (Fig. 4.36A, F; Appendix E).



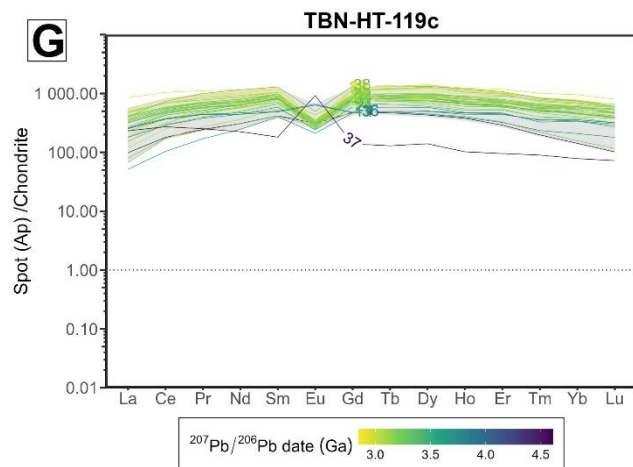


Figure 4.36: LASS chondrite-normalized REE concentrations for apatite spot, coloured by the $^{207}\text{Pb}/^{206}\text{Pb}$ date (Ga). The grey band represents the composition range for each sample based on the 95% interquartile range of data. Chondrite values are from Sun and McDonough (1989).

An apatite U-Pb age for all samples was obtained, with the majority of the samples containing 12-26 analyzed grains, while samples TBN-HT-35 and TBN-HT-36 only contain 3 and 2 analyzed grains, respectively (Table 4.2). The analyses are variable, ranging from discordant to concordant, with concordance values of 0.83-1.03, with a median of 0.93. Some samples display higher concordance values, such as TBN-HT-117, with concordance values of 0.88-1.01 with a median of 0.98. All the samples, including TBN-HT-117, display significant discordant trends away from the concordia line, allowing for well-constrained discordia lines to be interpreted (Fig. 4.37). The least constrained discordia lines correspond to samples TBN-HT-35 and TBN-HT-36, with the fewest number of analyses, and TBN-HT-117, where the majority of the analyses are clustered near the lower intercept (Fig. 4.37B, C, E). Based on discordia values and well-constrained discordia lines, a lower intercept age was selected for each sample (Fig. 4.37; Table 4.2). The lower intercept ages for all the samples range from 2598 ± 16 Ma to 2744 ± 51 Ma (Fig. 4.37; Table 4.2).

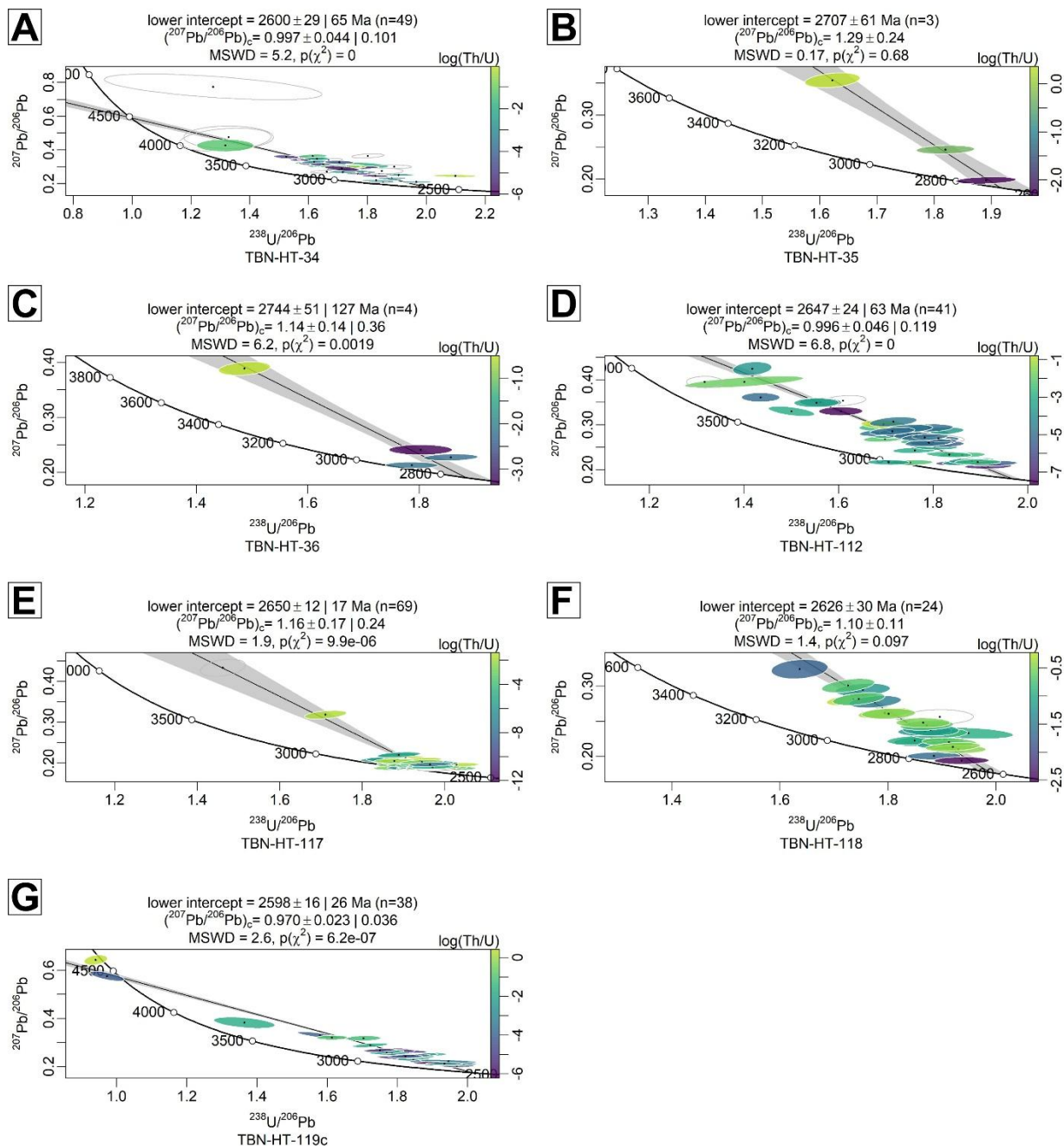


Figure 4.37: LASS U-Pb apatite geochronology for seven samples. Tera-Wasserberg concordia diagram showing all analyses within each sample, coloured by the $\log(\text{Th}/\text{U})$. Age uncertainties are given by the 2-sigma standard deviation and the MSWD-corrected standard deviation.

4.4.2.2 Monazite

The trace element patterns for monazite are characterized by consistent, enriched trace element patterns with similar enrichment values relative to the chondrite across all samples (Fig.

4.38). All the analyzed grains are characterized by a pronounced negative Eu anomaly (Fig. 4.38). The Eu anomalies are the strongest within the granitic samples, TBN-HT-35 and TBN-HT-36 (Fig. 4.38B, C). The Eu/Eu^* values within all the samples range from 0.09 to 0.55 with a median of 0.40. The monazite display fractionated LREE patterns with consistent weak negative slopes in all samples (Fig. 4.38). The $\text{La}/\text{Sm}_{(n)}$ values for all samples range between 3.06 and 8.14 with a median of 3.91 (Fig. 4.38; Appendix E). The HREE patterns are fractionated with a moderate slope, with $\text{Gd}/\text{Yb}_{(n)}$ values for all the samples ranging between 7.17 and 879.92, with a median of 40.12 (Fig. 4.38; Appendix E). The $\text{La}/\text{Sm}_{(n)}$ vs $\text{Gd}/\text{Yb}_{(n)}$ ratios within two of the samples, TBN-HT-34 and TBN-HT-119c, are well clustered within each sample, with the $\text{La}/\text{Sm}_{(n)}$ range deviating by less than 1 and $\text{Gd}/\text{Yb}_{(n)}$ deviating by less than 3 (Fig. 4.38A, E; Appendix E). The other three samples (TBN-HT-35, TBN-HT-36, and TBN-HT-112) have a greater spread of ratios, with $\text{La}/\text{Sm}_{(n)}$ deviating up to 4.64 and a large spread of $\text{Gd}/\text{Yb}_{(n)}$ deviating up to 863.46 (Fig. 4.38B, C, D; Appendix E).

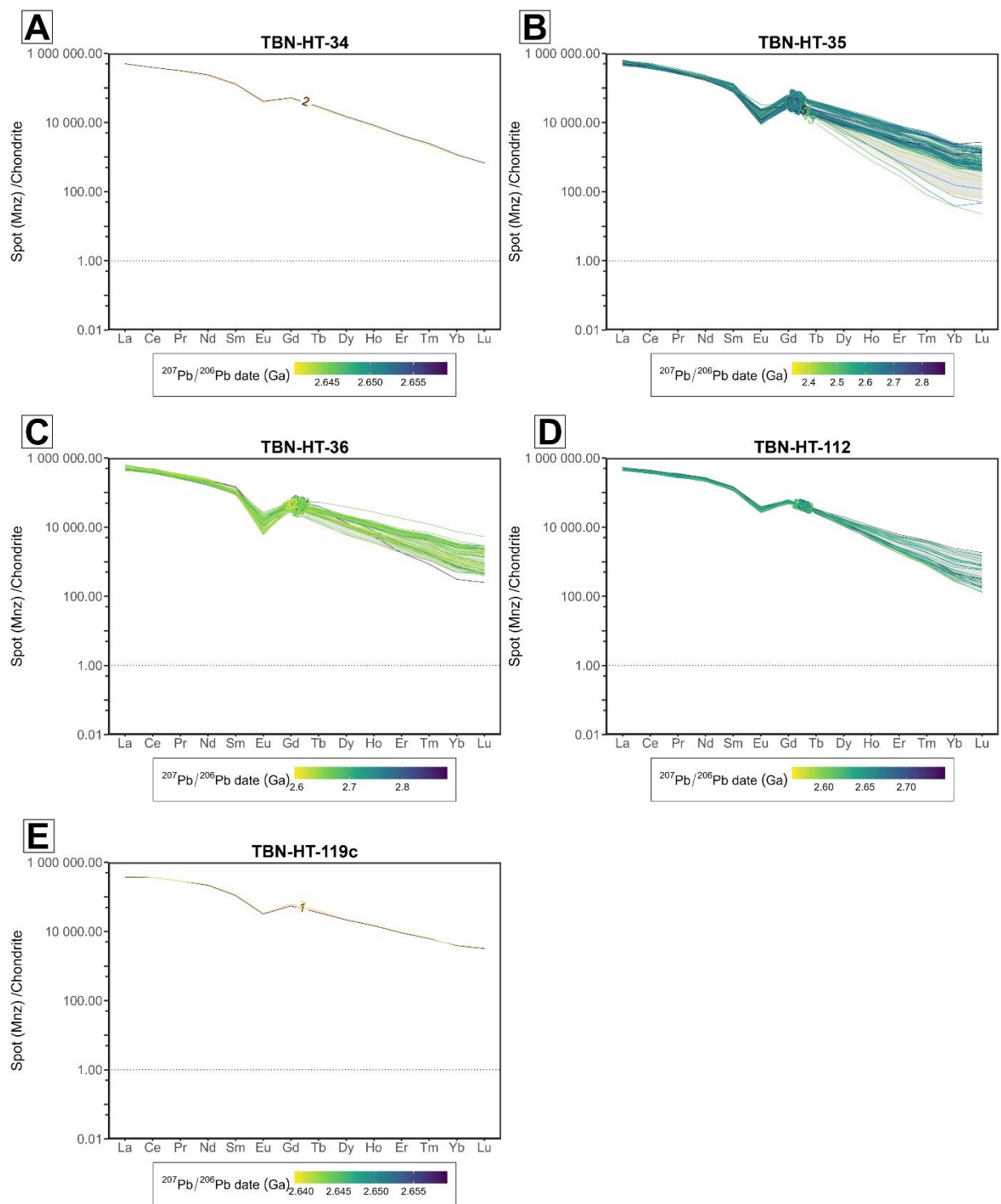
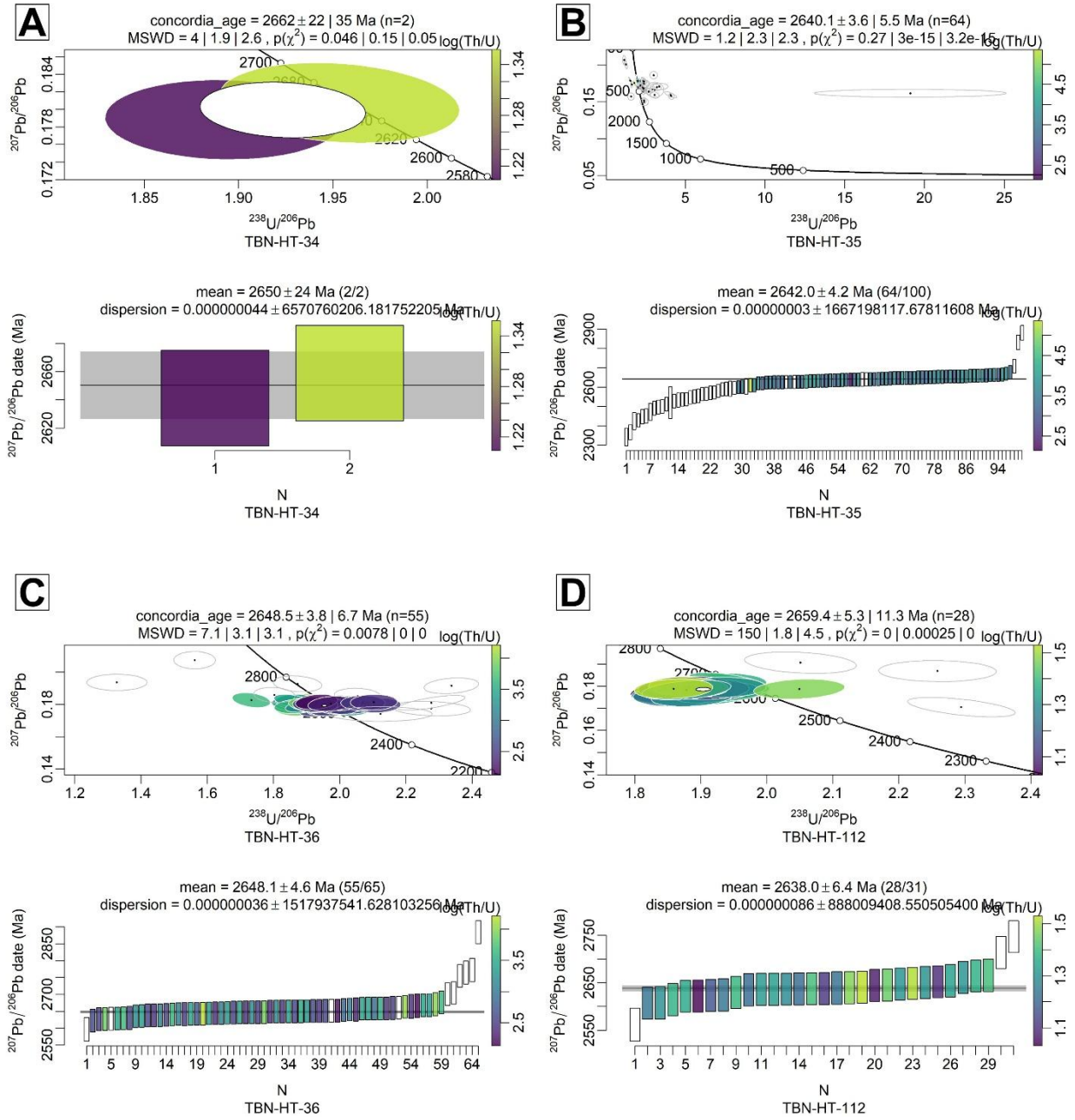


Figure 4.38: LASS chondrite-normalized REE concentrations for monazite, coloured by the $^{207}\text{Pb}/^{206}\text{Pb}$ date (Ga). The grey band represents the composition range for each sample based on the 95% interquartile range of data. Chondrite values are from Sun and McDonough (1989).

Monazite U-Pb ages were obtained for five of the samples, with most samples containing 15-21 analyzed grains, while samples TBN-HT-34 and TBN-HT-119c only contain 1 and 2 analyzed grains, respectively (Fig. 4.39; Table 4.2). The majority of the samples consists of analyses that are strongly concordant and well constrained within each sample, with concordance values for these samples ranging from 0.90 to 1.05, with a median of 1.02. Sample TBN-HT-35 contains the analyses with the highest spread of all the samples, with concordance values between 0.75 and 1.18, with a concordant median of 1.00. Based on the concordance values, a concordia age was selected for each sample only incorporating analyses with concordance values between 0.95 and 1.05 (Fig. 4.39). The concordia ages for all the samples range between 2640.1 ± 3.6 Ma to 2676 ± 17 Ma, and mean ages range from 2638.0 ± 6.4 Ma to 2650 ± 24 Ma (Fig. 4.39).



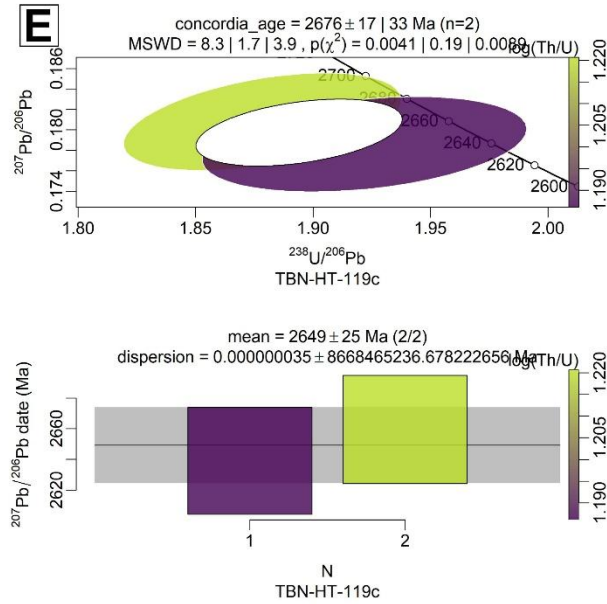


Figure 4.39: LASS U-Pb monazite geochronology for five samples. Top: Tera-Wasserberg concordia diagram showing all analyses within each sample, coloured by the $\log(\text{Th}/\text{U})$. Bottom: Weighted mean plot displaying the $^{207}\text{Pb}/^{206}\text{Pb}$ dates of each grain. Age uncertainties are given by the 2-sigma standard deviation and the MSWD-corrected standard deviation.

4.4.2.3 Titanite

The majority of the trace element patterns for titanite within sample TBN-HT-117 are characterized by consistent, enriched trace element patterns relative to chondrite, with four analyses displaying less enrichment than the majority (Fig. 4.40). Many of the analyzed grains are characterized by a weak positive Eu anomaly while others, especially the least enriched analyses, display negative Eu anomalies or strong positive Eu anomalies (Fig. 4.40). The Eu/Eu^* values within the sample ranges from 0.44 to 5.14 with a median of 1.33 (Fig. 4.40). The titanite grains display fractionated LREE patterns with moderately positive slopes with $\text{La}/\text{Sm}_{(n)}$ values between 0.02 and 0.15 with a median of 0.04 (Fig. 4.40; Appendix E). The HREE are weakly fractionated with a weak negative slope and $\text{Gd}/\text{Yb}_{(n)}$ values between 1.13 and 4.26 with a median value of 2.19 (Fig. 4.40; Appendix E). The $\text{La}/\text{Sm}_{(n)}$ vs $\text{Gd}/\text{Yb}_{(n)}$ ratios of the analyses within the samples are well clustered, with the $\text{La}/\text{Sm}_{(n)}$ having a deviation of less than 0.1 and the majority of the

Gd/Yb_(n) ratios falling between 2 and 3 (Fig. 4.40; Appendix E). The analyses that deviate the most from the median correspond to the least enriched analyses within the sample (Fig. 4.40).

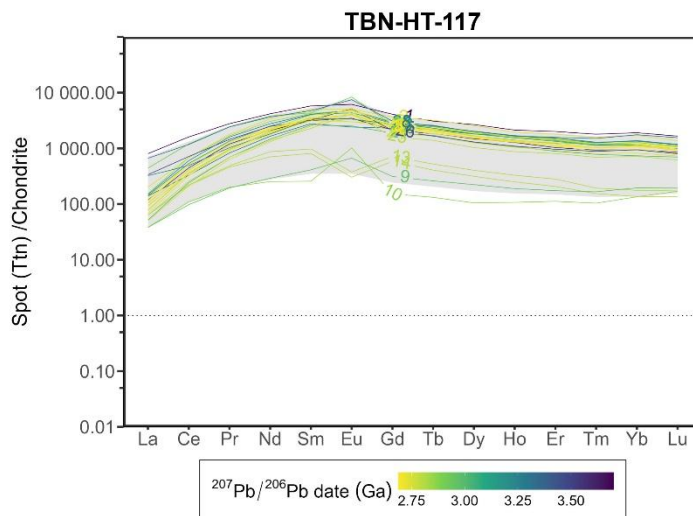


Figure 4.40: Chondrite-normalized REE concentrations for titanite spot analyses within one sample, coloured by the $^{207}\text{Pb}/^{206}\text{Pb}$ date (Ga). The grey band represents the composition range for each sample based on the interquartile range of data. Chondrite values are from Sun and McDonough (1989).

A titanite U-Pb age for sample TBN-HT-117 was obtained from 5 analyzed grains with 22 spot analyses (Fig. 4.41; Table 4.2). The majority of the analyses have high concordance values of 0.90-1.00, with a median of 0.97. Despite the high concordance values, the analyses display systematic deviation away from the concordia line, indicating Pb-loss (Fig. 4.41). Based on the well-constrained discordia line, a lower intercept age was selected (Fig. 4.41). The lower intercept age for the sample is 2620 ± 16 Ma (Fig. 4.41).

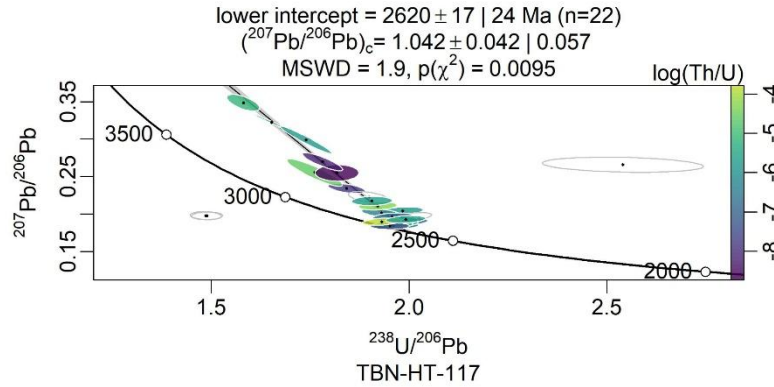
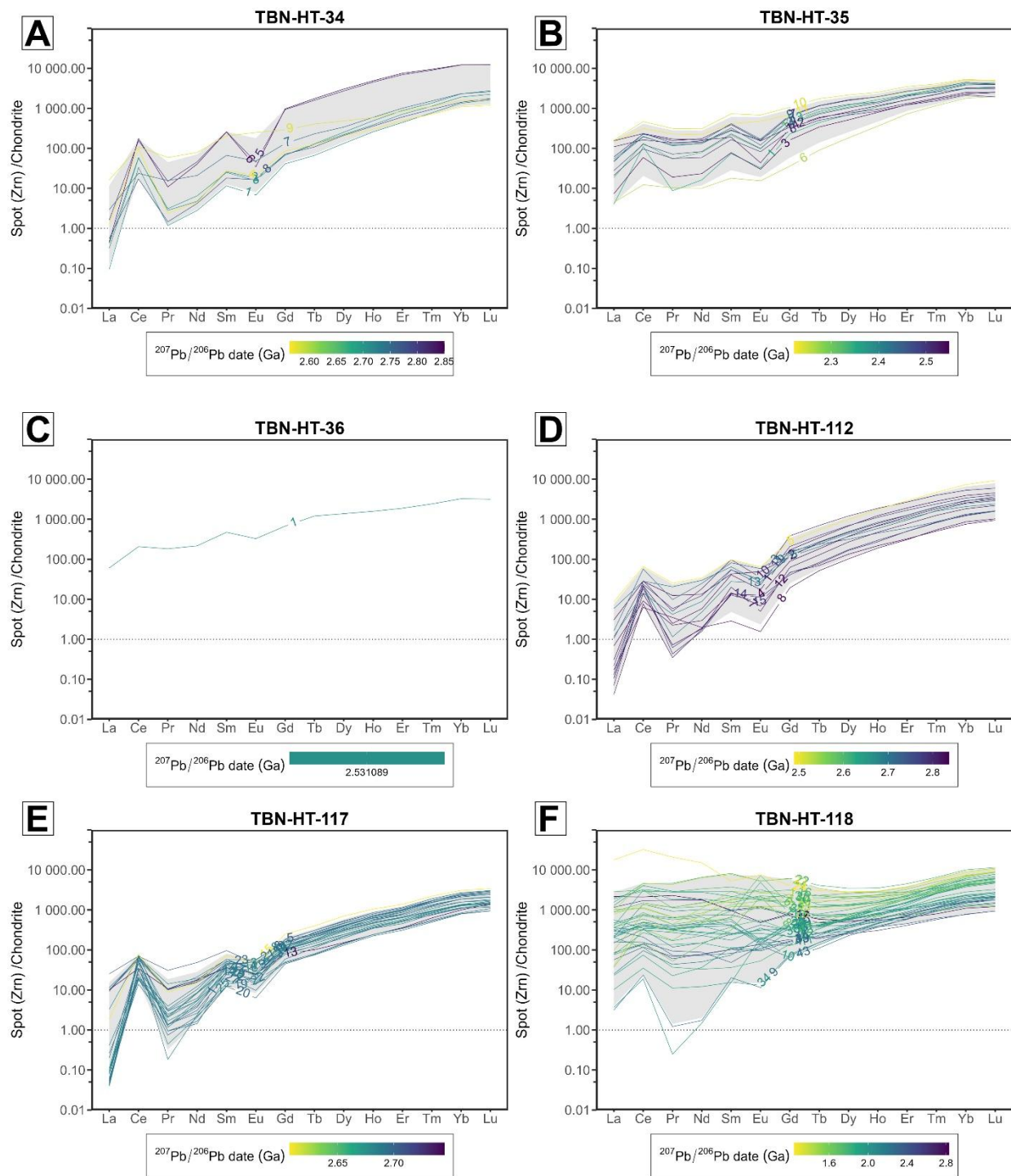


Figure 4.41: LASS U-Pb titanite geochronology. Tera-Wasserberg concordia diagram showing all analyses within each sample, coloured by the $\log(\text{Th}/\text{U})$. Age uncertainties are given by the 2-sigma standard deviation and the MSWD-corrected standard deviation.

4.4.2.4 Zircon

The majority of the trace element patterns for zircon are characterized by depleted to weakly enriched LREE and strongly enriched HREE trace element patterns relative to chondrite (Fig. 4.42). Sample TBN-HT-118, which corresponds to the sample with some of the lowest and highest range of $^{207}\text{Pb}/^{206}\text{Pb}$ dates, has the greatest enrichment relative to chondrite and the least consistent patterns compared to the rest of the samples (Fig. 4.42f). Many of the samples display some analyses with depleted trace elements relative to chondrite; samples TBN-HT-112 and TBN-HT-117 display depletion in both La and Pr, sample TBN-HT-34 displays depletion in La, and sample TBN-HT-118 displays depletion in Pr (Fig. 4.42A, D, E, F, G). While the granitic samples, TBN-HT-35 and TBN-HT-36, display no depletion relative to chondrite (Fig. 4.42B, C). All the samples display Ce anomalies, with the majority of the samples displaying strongly positive anomalies and only a few analyses within sample TBN-HT-118 displaying weakly negative Ce anomalies (Fig. 4.42F). The Ce/Ce^* values within all the samples range from 0.81 to 607.00 with a median of 8.98 (Fig. 4.42). The majority of the samples display negative Pr anomalies, with samples TBN-HT-35, TBN-HT-36, and TBN-HT-118 displaying the weakest negative to slightly positive Pr anomalies (Fig. 4.42B, C, F). The Pr/Pr^* values within all the samples range from -

0.06 to 1.17 with a median of 0.36 (Fig. 4.42). Most of the samples display a weak positive Sm anomaly, with analyses in sample TBN-HT-118 displaying moderate positive and weak negative Sm anomalies (Fig. 4.42F). The Sm/Sm^* values within all the samples range from 0.31 to 7.63 with a median of 2.40 (Fig. 4.42). All the samples display negative Eu anomalies, with samples TBN-HT-34, TBN-HT-118, and TBN-HT-119c displaying positive Eu anomalies as well (Fig. 4.42A, F, G). The Eu/Eu^* values within all the samples range from 0.07 to 6.71 with a median of 0.42 (Fig. 4.42). The zircon grains display fractionated LREE patterns with weak to moderately positive slopes, with some analyses within sample TBN-HT-118 displaying negative slopes (Fig. 4.42). The $\text{La}/\text{Sm}_{(n)}$ values for all the samples range between 0.00 and 5.63, with a median of 0.08 (Fig. 4.42; Appendix E). The HREE patterns are fractionated with weak to moderately positive slopes and $\text{Gd}/\text{Yb}_{(n)}$ values between 0.02 and 0.73, with a median of 0.07 (Fig. 4.42; Appendix E). The $\text{La}/\text{Sm}_{(n)}$ vs $\text{Gd}/\text{Yb}_{(n)}$ ratios for most of the samples are well clustered within a sample, with the $\text{La}/\text{Sm}_{(n)}$ range deviating by less than 1 and the $\text{Gd}/\text{Yb}_{(n)}$ range deviating by less than 0.2 (Fig. 4.38; Appendix E). Sample TBN-HT-118 has the greatest range in $\text{La}/\text{Sm}_{(n)}$ vs $\text{Gd}/\text{Yb}_{(n)}$ ratios, with the $\text{La}/\text{Sm}_{(n)}$ range deviating by 5.61 and the $\text{Gd}/\text{Yb}_{(n)}$ range deviating by 0.7 (Fig. 4.38; Appendix E).



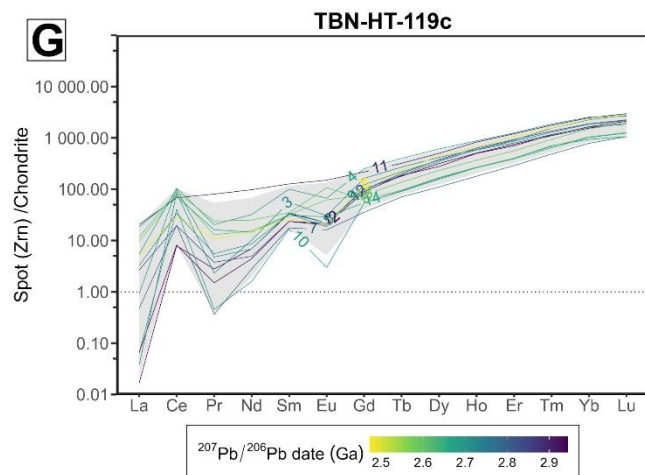
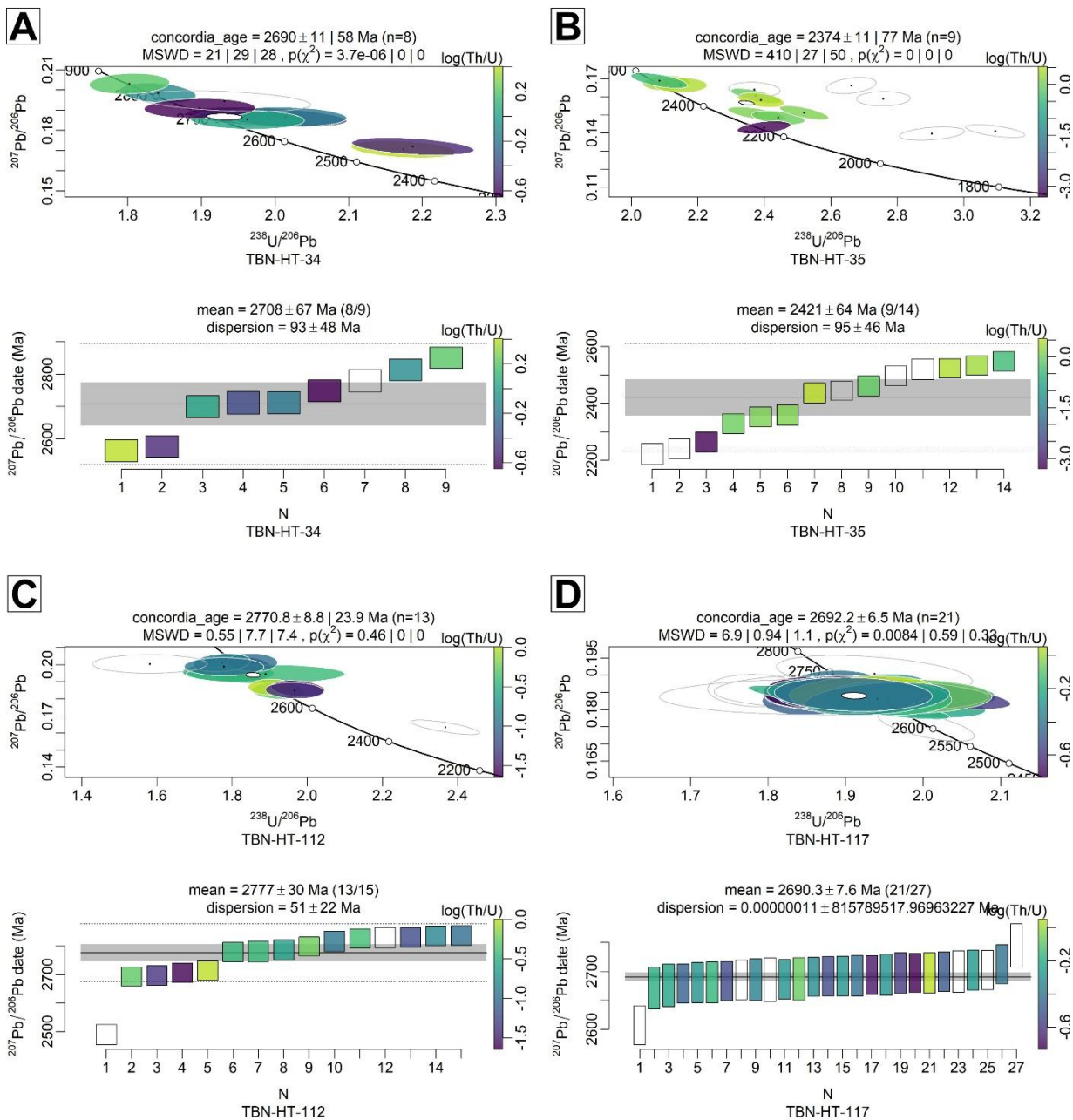


Figure 4.42: Chondrite-normalized REE concentrations for zircon, coloured by the $^{207}\text{Pb}/^{206}\text{Pb}$ date (Ga). The grey band represents the composition range for each sample based on the interquartile range of data. Chondrite values are from Sun and McDonough (1989).

Zircon U-Pb ages for six of the samples were interpreted, with the majority of the samples containing 7-23 grains, with samples TBN-HT-34 and TBN-HT-117 containing only 5 analyzed grains (Fig. 4.43; Table 4.2). The majority of the samples consists of analyses that are strongly concordant and well constrained within each sample, with concordance values for these samples ranging from 0.95 to 1.05, with a median of 1.02. Sample TBN-HT-36 consists of weakly concordant analyses, with concordance values of 0.89-1.00 with a median of 0.95. Sample TBN-HT-118 consists of analyses with the highest spread of all the samples, with concordance values of 0.77 to 1.03, with a concordant median of 0.97. Based on the concordance values, a concordia age was selected for each sample only incorporating analyses with concordance values between 0.95 and 1.05 (Fig. 4.43). The concordia ages for all the samples range from 1670.3 ± 5.4 Ma to 2770.8 ± 8.8 Ma, and mean ages range from 1928 ± 107 Ma to 2777 ± 30 Ma (Fig. 4.43).



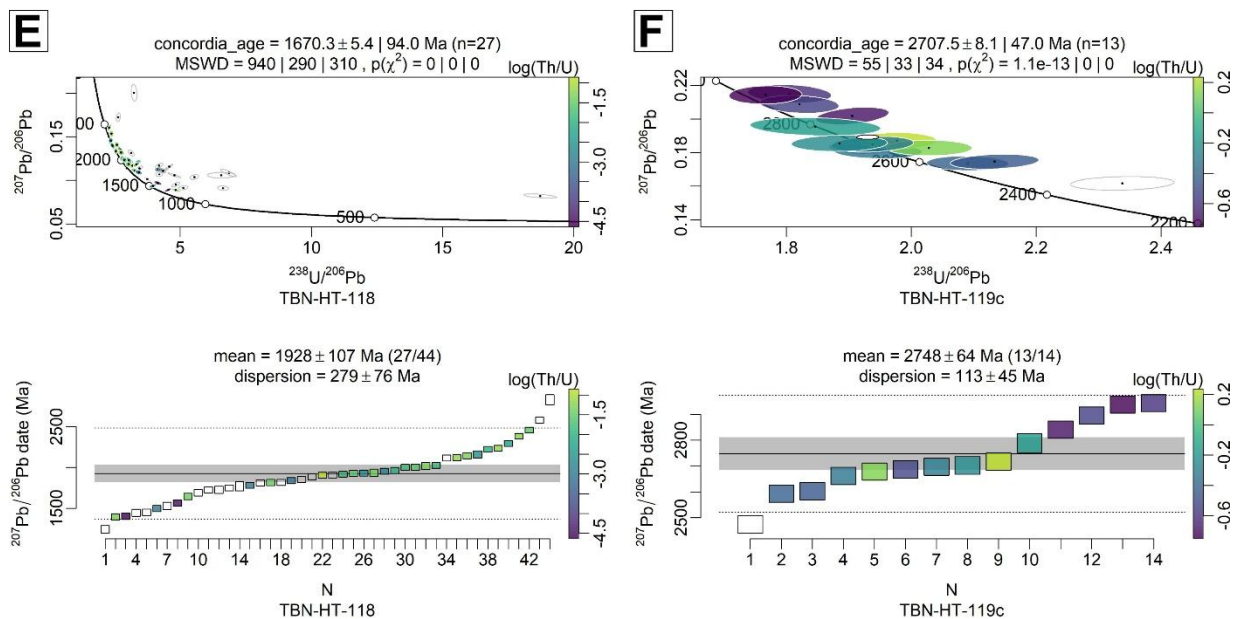


Figure 4.43: LASS U-Pb zircon geochronology. Top: Tera-Wasserberg concordia diagram showing all analyses within each sample, coloured by the log(Th/U). Bottom: Weighted mean plot displaying the $^{207}\text{Pb}/^{206}\text{Pb}$ dates of each grain. Age uncertainties are given by the 2-sigma standard deviation and the MSWD-corrected standard deviation.

5. Discussion

5.1 Extent of Quetico Shear Zone Deformation

The width of deformation associated with the Quetico Shear Zone is clearly noted from regional deformation within the study area; however, specific structures are not always easily identified as products of the shear zone outside of the mylonitic core. The interpreted width of the shear zone (D) within the study area, approximately 1 km based on the degree of mylonitization, is comparable to the previous width estimates in the area (1–1.2 km; Kennedy, 1984). Mylonitization was the best feature to distinguish shear zone deformation from preexisting structures within the Quetico subprovince. Brittle fault rocks parallel to the shear zone trace were not commonly observed, and foliation intensity was not easily distinguishable for the shear zone due to strong preexisting fabrics. As well in thin section, recrystallized quartz and quartz ribbons were unique to rocks within the shear zone. The core of the Quetico Shear Zone has a width of approximately 200 m, which is evident from the presence of fault rocks, including ultramylonites and pseudotachylite. The eastern section of the Quetico Shear Zone has the greatest width estimates of the entire trace of the shear zone (Kennedy, 1984). The large width of the shear zone within the eastern section could be due to rock types, as this is the only section of the shear zone that is solely within the Quetico Subprovince. Within this section of the shear zone, the way strain is accommodated differs based on rock type, with granitic rocks displaying dominantly ductile structures and metasedimentary and metavolcanic rocks displaying dominantly brittle structures within the core of the shear zone. Granitic rocks within the shear zone have varying strain intensities, with a few occurrences displaying no deformation; these are suggested to have been emplaced after Quetico Shear Zone-related deformation (see Section 5.3; Percival and Sullivan,

1988). The mafic-ultramafic intrusions, excluding the gabbro field exposure, are also suggested to have been emplaced after the shear zone deformation (Bleeker et al., 2020).

5.1.1 Brittle vs. Ductile Deformation

The abundance of interrelated brittle and ductile structures implies that the exposed surface of the shear zone represents the Archean brittle-ductile transition zone. This is supported by estimated erosional depths in the region of approximately 10-15 km (Percival et al., 2012). While there are multiple brittle-ductile transition zones within the crust based on rock type, the presence of strongly recrystallized quartz supports that the exposed surface is below the quartz brittle-dominated regime (~15 km; Byerlee, 1968; Tullis and Yund, 1992; Gleason and Tullis, 1995; Kohlstedt et al., 1995). Additionally, the presence of weakly deformed brittle feldspar supports that the rocks were above the feldspar ductile-dominated regime (~20 km; Byerlee, 1968; Tullis and Yund, 1992; Gleason and Tullis, 1995; Kohlstedt et al., 1995). It is interpreted that the exposed rocks show evidence of being at depths of 10-15 km; as discussed above, the rock type largely determines whether structures are brittle or ductile. The abundance of brittle and ductile kinematic structures with consistent dextral shear senses and the presence of all fault rock types (including fault gouge, pseudotachylite, cataclasite, and mylonites) support the interpretation that shear zone deformation occurred near the quartz brittle-ductile transition, resulting in both incohesive and cohesive fault rocks (Sibson, 1977).

5.1.2 Spatial patterns of macroscopic and crystallographic fabrics

The macro-and microscopic trends of fabrics and minerals within the rocks differ with distance to the shear zone. On a regional scale, the foliation orientations are mainly consistent and subparallel to the E-W strike of the core of the shear zone. However, macroscopic orientations display a slight deflection of the foliation planes towards the orientation of the center of the Quetico

Shear Zone. Notable deflection of the fabric in the surrounding rocks of the Quetico subprovince occurs over 1 km from the shear zone center (Fig. 4.1). Nevertheless, the degree of the regional fabric rotation is rather small ($<10-15^\circ$), i.e., smaller than the variation of orientation in most outcrops (Fig. 4.9). Given the width of the shear zone and the amount of displacement, a larger angular shear strain and thus, stronger deflections should be expected (Ramsay and Graham, 1970). The small deflection of the fabric underlines that a preexisting fabric was overprinted by the later simple shear-dominated Quetico Shear Zone deformation (though the instantaneous strain axes did not need to be similarly oriented).

The crystallographic orientations of quartz are generally consistent within the analyzed samples, with differing orientations with proximity to the shear zone. The orientation of the c-axes and CVAs perpendicular to the lineation and within the foliation plane implies the quartz crystals rotated about the c-axis during deformation, with a girdle distribution of the c-axis proximal to the shear zone. The prevalent rhomb $\langle a \rangle$ slip near the shear zone, indicated from the subgrain misorientation axes, indicates relatively low-temperature conditions, which is in agreement with the distribution of c-axes, while prism $\langle a \rangle$ slip becomes dominant with distance from the shear zone, indicating higher-temperature conditions (Parsons et al., 2016; Passchier and Trouw, 2005). The slip systems are supported by the majority of the samples having at least one a-axis plot parallel to the lineation, indicating that it represents the slip direction. Based on proximity to the shear zone, the rhomb $\langle a \rangle$ slip patterns are interpreted to represent Quetico Shear Zone deformation, while the most distal prism $\langle a \rangle$ slip patterns are inherited from the earlier regional deformation. The samples with the weakest fabric usually correspond to rock types where quartz deformation was impeded by rigid minerals or was subject to microfabrics. For example, the quartz ribbons within granitic samples vary in orientation based on deflections

around surrounding rigid feldspar grains throughout the sample causing a wide range of orientations. As well, samples with microfolds within thin sections result in differing orientations from the main foliation. As many of the samples were from quartz veins, the crystallographic preferred orientations may also be influenced by vein growth mechanics; however, this is not observed within the results (Pennacchioni et al., 2010; Appendix C3). There is an evolution in slip system with distance from the shear zone trace irrespective of whether the quartz is in veins or matrix.

5.2 Quartz Strain Rate Estimates

Empirical flow laws for quartz deforming by dislocation creep provide estimates for strain rate (Hirth et al., 2001; Lusk et al., 2021; Tokle et al., 2019). Grain size is not required in flow laws for quartz deforming by dislocation creep, as dislocation creep is not sensitive to grain size (Lusk et al., 2021). The most abundant recrystallization textures of quartz within the study area represent bulging and subgrain rotation dynamic recrystallization processes (Fig. 4.17). Static recrystallization evident in drill core samples proximal to the Thunder Bay Intrusive Complex indicates annealing processes due to heat from the Current Lake intrusion, resetting/obliterating dynamic recrystallization textures (Figs. 4.1 & 4.17; Ree and Park, 1997).

Using the stress and post-peak metamorphism temperature values (Ti-in-Quartz) obtained in this study, strain rate values were estimated/calculated using the following flow law equation from Hirth et al. (2001):

$$\dot{\epsilon} = A f_{H_2O}^m \sigma^n e^{-\frac{Q}{RT}},$$

where $\dot{\epsilon}$ is the strain rate, A is a prefactor, f_{H_2O} is water fugacity, m is the water fugacity exponent, σ is differential stress, n is the stress exponent, Q is the activation energy, R is the ideal gas constant, and T is absolute temperature. The empirical values for the prefactor, water fugacity

exponent, stress exponent, and activation energy values are derived from (Hirth et al., 2001; Appendix D). Water fugacity was calculated for each of the samples based on temperature and the confining pressure estimated from geothermometry and pseudosection work using the Water Fugacity Calculator by Aaron Springstroh (see Section 4.3; Appendix D; <https://aaron-sp.github.io/fugacity/index.html>). The strain rate values for the six samples range from 9.5×10^{-12} to $3.9 \times 10^{-10} \text{ s}^{-1}$ with high uncertainties due to the exponential relationship between stress and strain rate (Fig. 5.1; Appendix D). The strain rate values for the samples show no clear trend with distance from the shear zone as two of the proximal samples have high strain rates and the most distal sample has the highest strain rate (Fig. 5.1). The other three intermediate samples show a constant trend in strain rate values (Fig. 5.1). These rates are comparable to strain rates in modern shear zones ranging from 10^{-15} to 10^{-9} s^{-1} , such as the San Andreas Fault (Thatcher, 1975; Spinler et al., 2010).

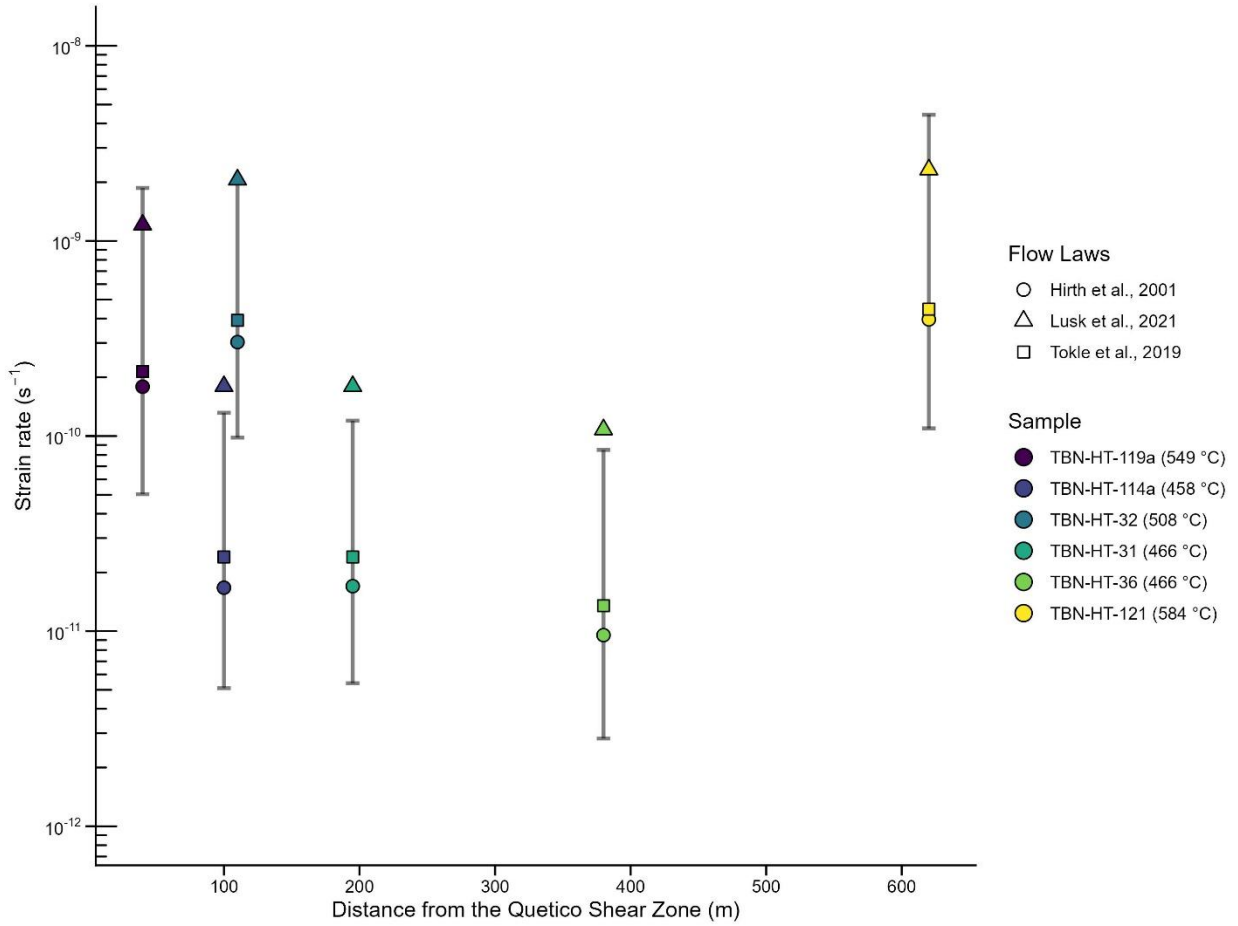


Figure 5.1: Strain rate vs. distance plot: The calculated strain rate values using three different flow laws for each sample are displayed based on distance from the Quetico Shear Zone. The grey error bars display the effect of the 68% interpercentile range of stress values, which are the least constrained variable within the flow laws.

Strain rate values were also calculated using two other flow laws to compare strain rate values for flow laws that use different parameters than Hirth et al. (2001) (Fig. 5.1; Appendix D).

These include the flow law from Lusk et al. (2021):

$$\dot{\epsilon} = A\sigma^n d^m f_{H_2O}^r e^{\left(-\frac{Q+VP}{RT}\right)},$$

where d is the grain size, m is the grain size exponent, r is the water fugacity exponent, V is the activation volume, and P is the confining pressure, and the flow law from Tople et al. (2019):

$$\dot{\epsilon} = A f_{H_2O}^r \sigma^n e^{-\frac{Q}{RT}},$$

where r is the water fugacity exponent. The strain rate values from Tokle et al. (2019) are more consistent with those of Hirth et al. (2001), whereas the values from Lusk et al. (2021) are an order of magnitude greater (Fig. 5.1). The strain rate values for each of the flow laws follow the same trend with distance from the shear zone (Fig. 5.1). Because the obtained temperature and stress values are relatively constant across the area, neither of these factors had a major control on localizing the deformation.

The estimated strain rate values for the Quetico Shear are Zone are interpreted to represent interseismic periods where ductile deformation produced quartz recrystallized quartz. The strain rates during formation of pseudotachylites would be much faster, although these strain rates are not recorded within the quartz grains within the study area because of the switch from ductile to brittle deformation mechanisms with increasing strain rate (Rowe and Griffith, 2015).

5.2.1 Temperature variations

The Ti-in-quartz temperature values obtained for the samples through SIMS span from upper greenschist facies to lower amphibolite facies metamorphism, while the temperatures obtained through the stable mineral assemblages in pseudosection modelling range from upper greenschist facies to upper amphibolite facies metamorphism. The higher temperatures indicated by the stable mineral assemblages, especially for samples TBN-HT-106, 115, and 134, are suggested to represent earlier peak metamorphism conditions rather than Quetico Shear Zone deformation conditions. This is indicated by high temperature index minerals present in the stable mineral assemblages, such as garnet and sillimanite, which would have formed during regional metamorphism when the pressures and temperatures were higher. For samples where temperatures were constrained through both methods, the Ti-in-quartz temperatures were within the range of the pseudosection temperatures for most of the samples, except for the two quartz veins, samples TBN-

HT-119a and TBN-HT-121, which give higher values from Ti-in-quartz thermometry. For sample TBN-HT-121, which is dominated by large relict quartz ribbons, SIMS spots were selected based on the finest grains; some of these were likely not recrystallized grains, but rather fine relict grains based on the EBSD map (Fig. 4.18; Appendix C). This suggests that the temperature obtained for TBN-HT-121 is impacted by temperatures obtained from relict quartz grains (which record higher temperature regional deformation). Similarly, the sample TBN-HT-119a consists of large relict quartz ribbons and fine relict grains with significantly more recrystallized grain; the temperature is also likely impacted by temperatures associated with relict quartz grains. Sample TBN-HT-36 with significant chlorite retrogression has the lowest Ti-in-quartz temperature compared to the pseudosection results, as the stable mineral assemblage likely records higher metamorphic conditions before chlorite retrogression.

Higher regional metamorphic temperatures (upper amphibolite facies) are suggested to represent peak metamorphic conditions, while the lower temperatures (lower amphibolite facies) are suggested to be associated with retrograde metamorphism. The temperatures associated with Quetico Shear Zone deformation reflect greenschist facies up to lower amphibolite facies conditions based on the majority of the Ti-in-quartz temperatures. This temperature range is also consistent with the rhomb $\langle a \rangle$ slip near the shear zone core (Parsons et al., 2016; Passchier and Trouw, 2005) and the recrystallization types within the shear zone core, which correlate with low to medium temperatures (Stipp et al., 2002) and low to moderate strain rates (Passchier and Trouw, 2005). Based on the lower temperature conditions obtained, it is interpreted that the strain localization of the Quetico Shear Zone occurred alongside the retrograde metamorphism. Additionally, the localized Quetico Shear Zone deformation produced structures and temperatures

unique to only the shear zone within the area indicating that strain was localized which is more common at lower temperature conditions (e.g., Platt and Behr, 2011).

5.2.2 Constant stress conditions

The constant stress values within the core of the shear zone and outside of the estimated shear zone width indicate that the increasing intensity of quartz mylonitization is not due to high stresses within the core of the shear zone (Fig. 5.2). Constant stress values within shear zones have been hypothesized to represent the yield stress at which weakening in the shear zone core causes an increase in strain rate (Ranalli, 1987; Platt and Behr, 2011). However, the strain rate values within the Quetico Shear Zone obtained in this study do not show an obvious increase towards the shear zone core. The grain size reduction at constant stresses could also indicate there is a switch to a deformation mechanism that involves diffusion creep (Rutter and Brodie, 2004; Platt and Behr, 2011). Changes in fabrics and orientations towards the shear zone core indicate that the regional deformation was overprinted by the Quetico Shear Zone deformation (Figs. 4.1 & 5.2). Therefore, the strain was partitioned into the shear zone core based on other factors, which resulted in the deflection of fabrics into the shear zone and the change in slip systems within the core of the shear zone. Based on the lower temperatures within the shear zone core and evidence for retrograde reactions, the presence of water was likely a contributing factor in shear zone deformation (Jefferies et al., 2006; Platt and Behr, 2011). However, structural inheritance from the earlier Kenoran transpressional tectonics is suggested to be the largest factor for localizing the deformation (Jones and Tanner, 1995; Holyoke and Tullis, 2006). While the majority of the shear zone's current extent is interpreted to be active after peak metamorphism/regional deformation, the shear zone may have used older structures from the regional deformation.

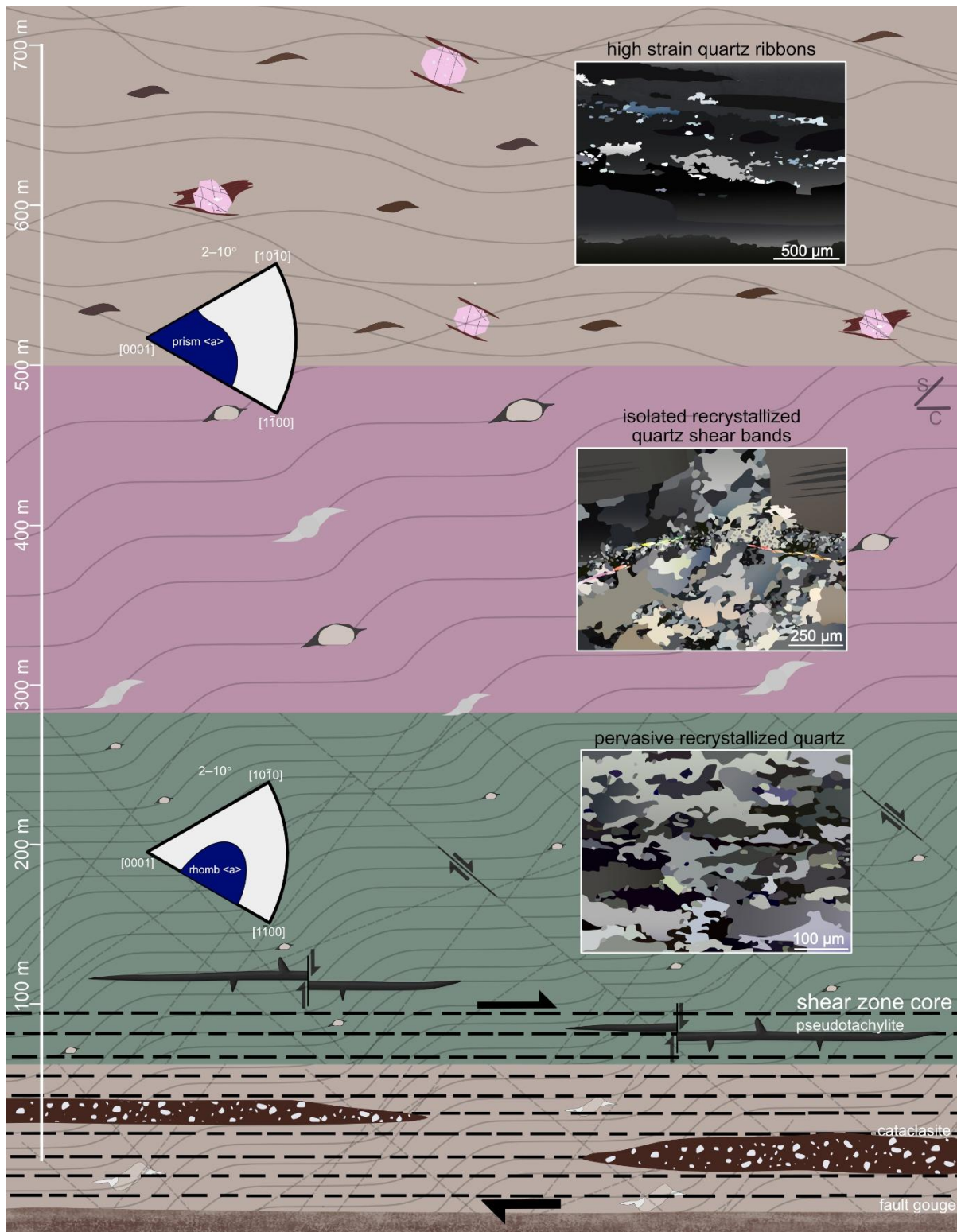


Figure 5.2: Schematic cartoon of the Quetico Shear Zone displaying key macro- and microstructures within rock types and with proximity to the shear zone core. The white rectangles display the changing quartz microstructures observed with proximity to the shear zone.

5.3 Quetico Shear Zone History

5.3.1 Timing of the Quetico Shear Zone deformation

The preferred ages for each mineral analyzed within the samples were the mean ages for concordant minerals (monazite and zircon) and lower intercept ages for non-concordant minerals (apatite and titanite) (Fig. 5.3). Zircon ages are the oldest, with these ages corresponding to igneous crystallization ages. Two of the zircon ages in two samples yield very young ages; these are suggested to be correlated with Pb loss, causing discordant ages (Fig. 4.43). The other zircon ages are not younger than the maximum deposition age for the Quetico subprovince (~2698 Ma; Davis et al., 1990), supporting that they represent crystallization ages prior to deposition of the subprovince (Fig. 5.3). Monazite ages are the second oldest and are interpreted to constrain the timing of high-temperature metamorphism in the Quetico subprovince. Most of the monazite ages (2640 to 2676 Ma) fall within the estimated ages for peak metamorphism in the Quetico subprovince, except for the two granites which show younger ages (2665–2671 Ma; Pan et al., 1994, 1998; 2657–2664 Ma; Rehm et al. 2026; Fig. 5.3). This would correspond to timing of the high temperatures obtained likely correlating to earlier transpressional deformation phases of the Superior Province (Percival, 1989; Pan et al., 1994, 1998). The apatite and titanite ages are the youngest and mostly yield similar ages, except for two granitic samples, where the apatite ages are much older (Fig. 5.3). The youngest three mineral ages (in samples TBN-HT-119a, TBN-HT-34, and TBN-HT-117) show notable resetting of apatite and titanite compared to the older monazite and apatite ages within the same sample. Using these three youngest ages and their weighted mean age and its 2-sigma standard deviation error, the interpreted Quetico Shear Zone age is 2618–2596 Ma (Fig. 5.3).

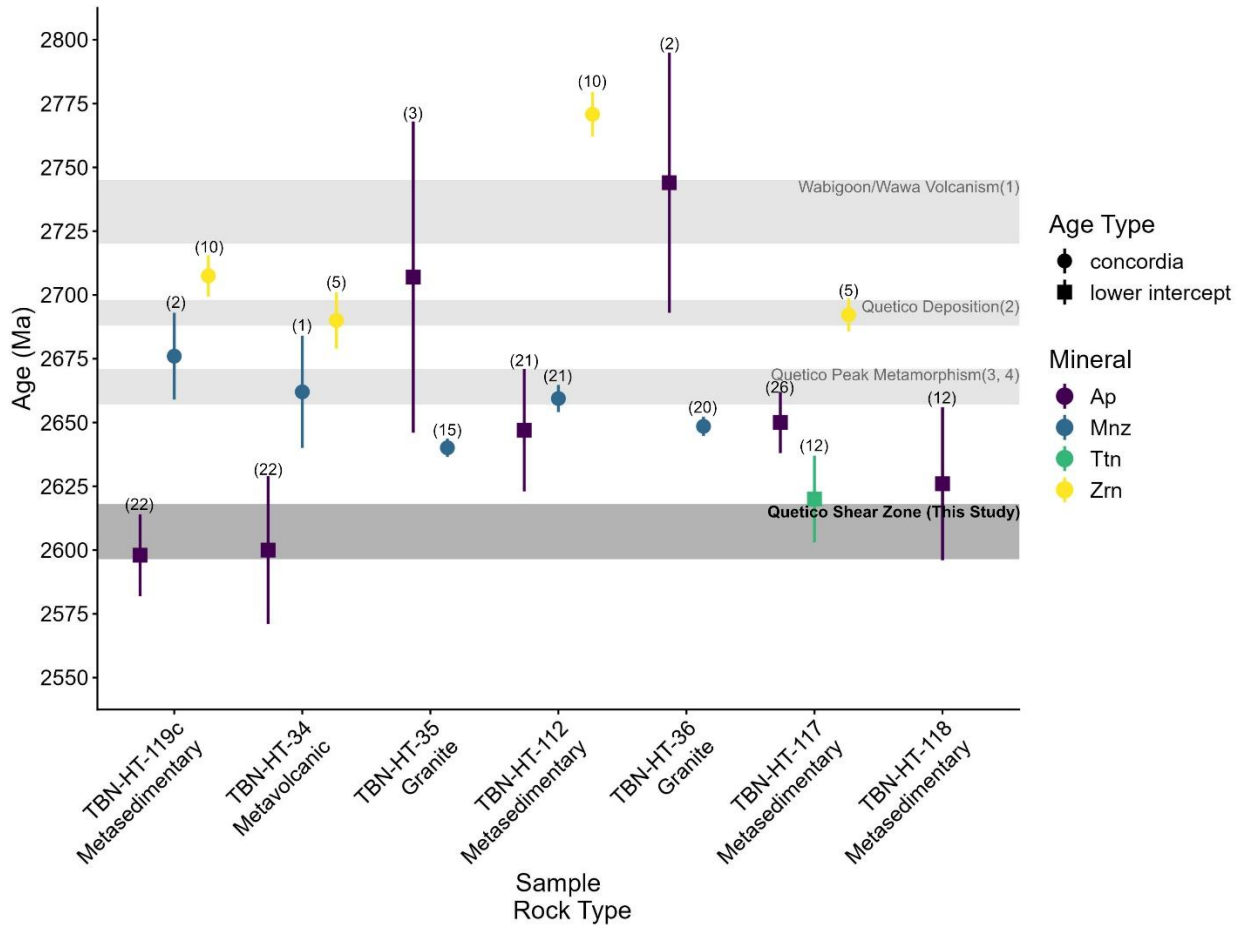


Figure 5.3: Preferred Mineral Ages: Plot displaying the preferred ages for all mineral types analyzed within a sample with respect to the distance from the shear zone trace for each sample. The error bars indicate the 2σ standard deviation. The number of grains analyzed per mineral type is noted above the data points. The ages from this study are compared to bracketed ages from the literature for adjacent subprovinces and the Quetico subprovince: (1) Percival et al., 2012, (2) Davis et al., 1990, (3) Pan et al., 1994, 1998, (4) Rehm et al., 2026.

Trace element mineral geochemistry was used to differentiate metamorphic minerals from primary minerals, especially monazite and apatite; however the REE ratios for the minerals are consistent (Cruft, 1966; Spear and Pyle, 2002; Appendix E). Based on the REE trace element patterns, there is no discernible difference between apatite and monazites that produced younger ages corresponding to shear zone deformation or older ages similar to peak metamorphism.

5.3.2 Upper age limit of the deformation

The earlier transpressional events formed structures that were overprinted by deformation associated with the Quetico Shear Zone. After formation of the weakened zone, deformation localized in today's shear zone core rather than forming a diffuse deformation zone by affecting a wider surrounding and stronger area. As mentioned above, preexisting structures from the complex regional transpressional deformation were likely exploited by the Quetico Shear Zone; however, the full extent of the shear zone was likely not active during this time. The geochronology results show there are two distinct metamorphic phases up to 40 Ma apart, supporting the interpretation that there was another significant deformation event following regional transpressional deformation.

There is a lack of deformation ages within the southern Superior Province that coincide with our interpreted shear zone age for this localized deformation. There are late-kinematic granites and $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende and biotite ages (~ 2.60 Ga) that are similar within the Minnesota River Valley subprovince (Schmitz et al., 2018; Hamilton et al., 2025). These ages have been interpreted to be related to the Scared Heart orogeny, which is interpreted to represent accretion and end-stage suturing of the Minnesota River Valley subprovince with the rest of the Superior Province (Schmitz et al., 2006). This interpreted ~ 2.60 Ga orogenesis may have a genetic link with our observed shear zone deformation (Schmitz et al., 2018).

5.3.3 Lower age limit of the deformation

The termination of Quetico Shear Zone deformation is not well constrained. To assess whether the deformation continued beyond ~ 2600 Ma would require more sensitive techniques to isolate later events from earlier higher-temperature events. However, as mentioned, there are undeformed granitic bodies within the shear zone, which could provide a minimum age for

deformation if they are postkinematic intrusions. The undeformed granitic intrusions within the study area have not been dated, although the Blablock pluton (Fig. 2.2; Blablock pluton is located south of Marmion Lake) has an age of 2688 Ma (Davis et al., 1990), is the youngest dated intrusion within the Quetico subprovince, indicating that the deformation was localized during the later stages of deformation. The Current Lake Intrusion (Fig. 4.1) within the Quetico Shear Zone core is undeformed, indicating that deformation ceased by ~ 1.1 Ga (Bleeker et al., 2020). To the east of the study area, the shear zone is overlain by sedimentary rocks of the Sibley Group, which have a maximum deposition age of ~ 1770 Ma, suggesting that shear zone deformation ceased prior to sedimentation (Rogala et al., 2007).

The only previous absolute age for the timing of Quetico Shear Zone deformation is a younger, ~ 1947 Ma Rb-Sr age from pseudotachylite veins of the shear zone's westernmost extent (Peterman and Day, 1989). Peterman and Day (1989) suggest that the Quetico Shear Zone deformation was reactivated as a far-field response to Early Proterozoic tectonic events such as the Trans-Hudson orogeny (1.8 Ga; Ansdell, 2005) or the New Quebec orogeny (2.2-1.77 Ga; Wardle et al., 2002). Using Rb-Sr to target active deformation along the shear zone has its own shortcomings, as both isotopes within the system can be easily mobilized with hydrothermal fluid interaction (Magloughlin, 2003; Stein, 2014; Tollefson et al., 2025). Significant deformation in the study area is not observed for these times, and therefore, we hypothesize that this Proterozoic age reflects extremely late localized deformation overprinting the main fabrics in the shear zone core.

While this study focused on characterizing the deformation of the Quetico Shear Zone apart from the complex preexisting regional deformation, future research should examine the deformation mechanisms in the shear zone itself further. Knowing the strain rates of the shear zone from this project provides a foundation to constrain Archean deformation mechanisms in greater

detail. The timing of deformation should also be studied with more sensitive methods to target the lower age limit of deformation. Studies along other areas of the trace of the shear zone should be conducted to compare the deformation along the entire trace of the shear zone.

6. Conclusions

This study characterized deformation fabrics associated with the Quetico Shear Zone and constrained the conditions and timing of deformation. It has been assumed that the Superior Craton was been relatively tectonically stable for 2.6 Ga (Percival et al., 2012) however, this study shows that a significant structure was active within the late Archean. The Quetico Shear Zone is characterized by both mylonitization and brittle deformation, overprinting regional transpressional structures in the Quetico subprovince (which formed during the Kenoran orogeny). Deformation associated with the Kenoran orogeny was widespread and affected many subprovinces, while deformation associated with the Quetico shear zone was localized like active shear zones today. The shear zone core consists of abundant fault rock types of both brittle and ductile nature, which together do not occur outside of the shear zone. Preexisting fabrics are slightly deflected into parallelism with the shear zone core. The overprinting deformation is also evident from patterns in the regional metamorphism. Temperatures derived from quartz microstructures, mineral assemblages, and pseudosections indicates a decrease in temperature from up to 700 degrees Celsius during peak metamorphism to retrograde conditions under 500 degrees Celsius within the shear zone core. Ti-in-quartz thermometry estimates agree with such medium and lower temperature conditions for deformation (mean temperature of ~490 degrees Celsius). Our new deformation ages (2618–2596 Ma) indicate that localized Quetico Shear Zone-related deformation postdates the Kenoran orogeny and predates Proterozoic reactivation. Together, these data show that Archean strike-slip boundaries localized deformation similar to active shear zones today.

Within this study, we show that:

- The Quetico Shear Zone has unique deformation conditions and structures when compared to the regional transpressional structures formed during the Kenora Orogeny.

- The degree of mylonitization increases towards the shear zone core. We hypothesized that the substantially increasing mylonitization implied changing stresses; however, we show that the Quetico Shear Zone has constant stresses, indicating that some other mechanism controlled strain localization.
- Archean tectonics are largely thought to have ceased by 2.6 Ga within the Western Superior Province; however, there was a major deformation event occurring at this time.
- There was substantial horizontal tectonics occurring within the southern Superior Province at the end of the Archean.

References

- Ansdell, K.M., 2005. Tectonic evolution of the Manitoba-Saskatchewan segment of the Paleoproterozoic Trans-Hudson Orogen, Canada. *Can. J. Earth Sci.* 42, 741–759. <https://doi.org/10.1139/e05-035>
- Bachmann, F., Hielscher, R., Schaeben, H., 2011. Grain detection from 2d and 3d EBSD data—Specification of the MTEX algorithm. *Ultramicroscopy* 111, 1720–1733. <https://doi.org/10.1016/j.ultramic.2011.08.002>
- Bachmann, F., Hielscher, R., Schaeben, H., 2010. Texture Analysis with MTEX – Free and Open Source Software Toolbox. *Solid State Phenomena* 160, 63–68. <https://doi.org/10.4028/www.scientific.net/SSP.160.63>
- Behr, W.M., Platt, J.P., 2011. A naturally constrained stress profile through the middle crust in an extensional terrane. *Earth and Planetary Science Letters* 303, 181–192. <https://doi.org/10.1016/j.epsl.2010.11.044>
- Biggar, G.M., O'Hara, M.J., 1969. A comparison of gel and glass starting materials for phase equilibrium studies. *Mineralogical Magazine* 37, 198–205. <https://doi.org/10.1180/minmag.1969.037.286.06>
- Bjorkman, K.E., Lu, Y., Campbell McCuaig, T., Kemp, A.I.S., Hollings, P., 2024. Linked evolution and in situ growth of the Wabigoon superterrane, Superior Craton: evidence from zircon U-Pb isotopes and whole-rock geochemistry. *Precambrian Research* 404, 107341. <https://doi.org/10.1016/j.precamres.2024.107341>
- Blackburn, C.E., 1980. Towards a Mobilist Tectonic Model for Part of the Archean of Northwestern Ontario. *Geoscience Canada*.
- Bleeker, W., 2020. The Midcontinent Rift and Its Mineral Systems: Overview and Temporal Constraints of Ni-Cu-PGE Mineralized Intrusions. Presented at the North-Central Section - 54th Annual Meeting - 2020, GSA.
- Borradaile, G., Kennedy, M., C., 1982. Microfaults and Mylonitic Schistosity, in: *Pseudotachylite*. Springer-Verlag, Berlin, Heidelberg, New York.
- Borradaile, G.J., 1982. Comparison of Archean structural styles in two belts of the Canadian Superior Province. *Precambrian Research* 19, 179–189. [https://doi.org/10.1016/0301-9268\(82\)90058-4](https://doi.org/10.1016/0301-9268(82)90058-4)
- Byerlee, J.D., 1968. Brittle-ductile transition in rocks. *Journal of Geophysical Research* (1896-1977) 73, 4741–4750. <https://doi.org/10.1029/JB073i014p04741>
- Card, K.D., 1990. A review of the Superior Province of the Canadian Shield, a product of Archean accretion. *Precambrian Research* 48, 99–156. [https://doi.org/10.1016/0301-9268\(90\)90059-Y](https://doi.org/10.1016/0301-9268(90)90059-Y)
- Card, K.D., Ciesielski, A., 1986. DNAG #1. Subdivisions of the Superior Province of the Canadian Shield. *GS*, 1 13, 5–13.
- Chester, F.M., Chester, J.S., 1998. Ultracataclasite structure and friction processes of the Punchbowl fault, San Andreas system, California, in: *Tectonophysics*. pp. 199–221. [https://doi.org/10.1016/S0040-1951\(98\)00121-8](https://doi.org/10.1016/S0040-1951(98)00121-8)
- Corfu, F., Stott, G.M., 1986. U–Pb ages for late magmatism and regional deformation in the Shebandowan Belt, Superior Province, Canada. *Can. J. Earth Sci.* 23, 1075–1082. <https://doi.org/10.1139/e86-108>

- Cross, A.J., Prior, D.J., Stipp, M., Kidder, S., 2017. The recrystallized grain size piezometer for quartz: An EBSD-based calibration. *Geophysical Research Letters* 44, 6667–6674. <https://doi.org/10.1002/2017GL073836>
- Cruft, E.F., 1966. Minor elements in igneous and metamorphic apatite. *Geochimica et Cosmochimica Acta* 30, 375–398. [https://doi.org/10.1016/0016-7037\(66\)90068-8](https://doi.org/10.1016/0016-7037(66)90068-8)
- Davies, J.C., 1973. Geology of the Fort Frances area, District of Rainy River, Ontario Division of Mines Geological Report 107. Ontario Geological Survey.
- Davis, D.W., Pezzutto, F., Ojakangas, R.W., 1990. The age and provenance of metasedimentary rocks in the Quetico Subprovince, Ontario, from single zircon analyses: implications for Archean sedimentation and tectonics in the Superior Province. *Earth and Planetary Science Letters* 99, 195–205. [https://doi.org/10.1016/0012-821X\(90\)90110-J](https://doi.org/10.1016/0012-821X(90)90110-J)
- Davis, D.W., Poulsen, K.H., Kamo, S.L., 1989. New Insights into Archean Crustal Development from Geochronology in the Rainy Lake Area, Superior Province, Canada. *The Journal of Geology* 97.
- de Capitani, C., 1994. Gleichgewichts-Phasendiagramme: Theorie und Software. Beihefte zum European Journal of Mineralogy, Jahrestagung der Deutschen Mineralogischen Gesellschaft 6, 48.
- de Capitani, C., Brown, T.H., 1987. The computation of chemical equilibrium in complex systems containing non-ideal solutions. *Geochimica et Cosmochimica Acta* 51, 2639–2652. [https://doi.org/10.1016/0016-7037\(87\)90145-1](https://doi.org/10.1016/0016-7037(87)90145-1)
- Dimroth, E., Imreh, L., Goulet, N., Rocheleau, M., 1983a. Evolution of the south-central segment of the Archean Abitibi Belt, Quebec. Part II: Tectonic evolution and geomechanical model. *Can. J. Earth Sci.* 20, 1355–1373. <https://doi.org/10.1139/e83-124>
- Dimroth, E., Imreh, L., Goulet, N., Rocheleau, M., 1983b. Evolution of the south-central segment of the Archean Abitibi Belt, Quebec. Part III: Plutonic and metamorphic evolution and geotectonic model. *Canadian Journal of Earth Sciences* 20, 1374–1388. <https://doi.org/10.1139/e83-125>
- Evans, D.A.D., Halls, H.C., 2010. Restoring Proterozoic deformation within the Superior craton. *Precambrian Research, Precambrian Large Igneous Provinces (LIPs) and their dyke swarms: New insights from high-precision geochronology integrated with paleomagnetism and geochemistry* 183, 474–489. <https://doi.org/10.1016/j.precamres.2010.02.007>
- Faulkner, D.R., Jackson, C.A.L., Lunn, R.J., Schlische, R.W., Shipton, Z.K., Wibberley, C.A.J., Withjack, M.O., 2010. A review of recent developments concerning the structure, mechanics and fluid flow properties of fault zones. *Journal of Structural Geology* 32, 1557–1575. <https://doi.org/10.1016/j.jsg.2010.06.009>
- Faulkner, D.R., Lewis, A.C., Rutter, E.H., 2003. On the internal structure and mechanics of large strike-slip fault zones: field observations of the Carboneras fault in southeastern Spain. *Tectonophysics* 367, 235–251. [https://doi.org/10.1016/S0040-1951\(03\)00134-3](https://doi.org/10.1016/S0040-1951(03)00134-3)
- Fralick, P., Davis, D., 1999. The Seine-Couchiching problem revisited: sedimentology, geochronology and geochemistry of sedimentary units in the Rainy Lake and Sioux Lookout areas 66–75.
- Fuhrman, M.L., Lindsley, D.H., 1988. Ternary-feldspar modeling and thermometry. *American Mineralogist* 73, 201–215.

- Ghent, E.D., Stout, M.Z., 1984. TiO₂ activity in metamorphosed pelitic and basic rocks: principles and applications to metamorphism in southeastern Canadian Cordillera. *Contr. Mineral. and Petrol.* 86, 248–255. <https://doi.org/10.1007/BF00373670>
- Gleason, G.C., Tullis, J., 1995. A flow law for dislocation creep of quartz aggregates determined with the molten salt cell. *Tectonophysics* 247, 1–23. [https://doi.org/10.1016/0040-1951\(95\)00011-B](https://doi.org/10.1016/0040-1951(95)00011-B)
- Goodwin, A.M., 1977. Archean basin–craton complexes and the growth of Precambrian shields. *Can. J. Earth Sci.* 14, 2737–2759. <https://doi.org/10.1139/e77-241>
- Goodwin, A.M., Ridler, R.H., 1970. The Abitibi orogenic belt In *Symposium on Basins and Geosynclines of the Canadian Shield*. 70-40 1–24.
- Halls, H., Davis, D.W., 2004. Paleomagnetism and U-Pb geochronology of the 2.17 Ga Biscotasing dyke swarm, Ontario, Canada: Evidence for vertical-axis crustal rotation across the Kapuskasing Zone. *Canadian Journal of Earth Sciences - CAN J EARTH SCI* 41, 255–269. <https://doi.org/10.1139/e03-093>
- Hamilton, J.D., Mayer, J.A., Hirsch, A.C., Severson, A.R., Retzler, A.J., Chandler, V.W., Radakovich, A.L., Radke, A.J., 2025. C-70, *Geologic Atlas of Yellow Medicine County, Minnesota*.
- Harris, F.R., 1974. *Geology of the Rainy Lake area, District of Rainy River, Ontario Division of Mines Geological Report 115*.
- Hawley, J.E., 1929. *Geology of the Sapawe Lake Area, with notes on some iron and gold deposits of Rainy River District, Ontario Division of Mines Geological Report*.
- Hirth, G., Teyssier, C., Dunlap, J.W., 2001. An evaluation of quartzite flow laws based on comparisons between experimentally and naturally deformed rocks. *Int J Earth Sci* 90, 77–87. <https://doi.org/10.1007/s005310000152>
- Holland, T.J.B., Powell, R., 1998. An internally consistent thermodynamic data set for phases of petrological interest. *Journal of Metamorphic Geology* 16, 309–343. <https://doi.org/10.1111/j.1525-1314.1998.00140.x>
- Hollings, P., Smyk, M., Heaman, L.M., Halls, H., 2010. The geochemistry, geochronology and paleomagnetism of dikes and sills associated with the Mesoproterozoic Midcontinent Rift near Thunder Bay, Ontario, Canada. *Precambrian Research, Precambrian Large Igneous Provinces (LIPs) and their dyke swarms: New insights from high-precision geochronology integrated with paleomagnetism and geochemistry* 183, 553–571. <https://doi.org/10.1016/j.precamres.2010.01.012>
- Holyoke, C.W., Tullis, J., 2006. Formation and maintenance of shear zones. *Geol* 34, 105. <https://doi.org/10.1130/G22116.1>
- Jefferies, S.P., Holdsworth, R.E., Shimamoto, T., Takagi, H., Lloyd, G.E., Spiers, C.J., 2006. Origin and mechanical significance of foliated cataclastic rocks in the cores of crustal-scale faults: Examples from the Median Tectonic Line, Japan. *Journal of Geophysical Research: Solid Earth* 111. <https://doi.org/10.1029/2005JB004205>
- Kamineni, D.C., Stone, D., Peterman, Z.E., 1990. Early Proterozoic deformation in the western Superior province, Canadian Shield. *Geological Society of America Bulletin* 102, 1623–1634. [https://doi.org/10.1130/0016-7606\(1990\)102%3C1623:EPDITW%3E2.3.CO;2](https://doi.org/10.1130/0016-7606(1990)102%3C1623:EPDITW%3E2.3.CO;2)
- Kaye, L., 1967. *Geology of Eastern Lac des Mille Lacs Area, District of Thunder Bay. Ontario Department of Mines Report 48*.
- Kennedy, M., C., 1984. *The Quetico Fault in the Superior Province of the Southern Canadian Shield (MSc)*. Lakehead University.

- Kohlstedt, D.L., Evans, B., Mackwell, S.J., 1995. Strength of the lithosphere: Constraints imposed by laboratory experiments. *Journal of Geophysical Research: Solid Earth*. <https://doi.org/10.1029/95jb01460>
- Kostrzyzhev, A., Murphy, T., 2024. Phase characterisation in minerals and metals using an SEM-EDS based automated mineralogy system. *Methods in Microscopy* 1, 163–175. <https://doi.org/10.1515/mim-2024-0015>
- Kylander-Clark, A.R.C., Hacker, B.R., Cottle, J.M., 2013. Laser-ablation split-stream ICP petrochronology. *Chemical Geology* 345, 99–112. <https://doi.org/10.1016/j.chemgeo.2013.02.019>
- Langford, F.F., Morin, J.A., 1976. The development of the Superior Province of northwestern Ontario by merging island arcs. *American Journal of Science* 276, 1023–1034. <https://doi.org/10.2475/ajs.276.9.1023>
- Lawson, A.C., 1913. The Archean Geology of Rainy Lake Restudied 115.
- Lawson, A.C., 1888. Report on the Geology of the Rainy Lake Region. Vol. 3, part 1, Report F 183.
- Liu, M.-C., McKeegan, K.D., Harrison, T.M., Jarzebinski, G., Vltava, L., 2018. The Hyperion-II radio-frequency oxygen ion source on the UCLA ims1290 ion microprobe: Beam characterization and applications in geochemistry and cosmochemistry. *International Journal of Mass Spectrometry* 424, 1–9. <https://doi.org/10.1016/j.ijms.2017.11.007>
- Ludden, J., Hubert, C., Gariépy, C., 1986. The tectonic evolution of the Abitibi greenstone belt of Canada. *Geological Magazine* 123, 153–166. <https://doi.org/10.1017/S0016756800029800>
- Lusk, A.D.J., Platt, J.P., Platt, J.A., 2021. Natural and Experimental Constraints on a Flow Law for Dislocation-Dominated Creep in Wet Quartz. *Journal of Geophysical Research: Solid Earth* 126, e2020JB021302. <https://doi.org/10.1029/2020JB021302>
- Mackasey, W.O., Blackburn, C.E., Trowell, N.F., 1974. A regional approach to the Wabigoon-Quetico belts and its bearing on exploration in northwestern Ontario, Paper 58. Ontario Division of Mines, Misc.
- Magloughlin, J.F., 2003. An evaluation of Rb-Sr dating of pseudotachylyte: Structural-chemical models and the role of fluids. *Geochem. J.* 37, 21–33. <https://doi.org/10.2343/geochemj.37.21>
- Montsion, R.M., Thurston, P., Ayer, J., 2018. 1:2 000 000 Scale Geological Compilation of the Superior Craton. MERC-ME-2018-017.
- Neumann, B., 2000. Texture development of recrystallised quartz polycrystals unravelled by orientation and misorientation characteristics. *Journal of Structural Geology* 22, 1695–1711. [https://doi.org/10.1016/S0191-8141\(00\)00060-2](https://doi.org/10.1016/S0191-8141(00)00060-2)
- Ojakangas, R.W., 1985. Review of Archean clastic sedimentation, Canadian Shield; Major felsic volcanic contributions to turbidite and alluvial fan-fluvial facies associations Evolution of Archean Supracrustal Sequences, 23–47.
- Pan, Y., E. Fleet, M., Heaman, L., 1998. Thermo-tectonic evolution of an Archean accretionary complex: U–Pb geochronological constraints on granulites from the Quetico Subprovince, Ontario, Canada. *Precambrian Research* 92, 117–128. [https://doi.org/10.1016/S0301-9268\(98\)00070-9](https://doi.org/10.1016/S0301-9268(98)00070-9)
- Pan, Y., Fleet, M.E., Williams, H.R., 1994. Granulite-facies metamorphism in the Quetico Subprovince, north of Manitouwadge, Ontario. *Can. J. Earth Sci.* 31, 1427–1439. <https://doi.org/10.1139/e94-126>

- Parsons, A.J., Law, R.D., Lloyd, G.E., Phillips, R.J., Searle, M.P., 2016. Thermo-kinematic evolution of the Annapurna-Dhaulagiri Himalaya, central Nepal: The Composite Orogenic System. *Geochemistry, Geophysics, Geosystems* 17, 1511–1539. <https://doi.org/10.1002/2015GC006184>
- Passchier, C.W., Trouw, R.A.J., 2005. *Microtectonics*, 2nd, rev.enl. ed ed. Springer, Berlin ; New York.
- Pennacchioni, G., Menegon, L., Leiss, B., Nestola, F., Bromiley, G., 2010. Development of crystallographic preferred orientation and microstructure during plastic deformation of natural coarse-grained quartz veins. *Journal of Geophysical Research: Solid Earth* 115. <https://doi.org/10.1029/2010JB007674>
- Percival, J., Skulski, T., Sanborn-Barrie, M., Stott, G., Leclair, A.D., Corkery, M.T., Boily, M., 2012. Geology and tectonic evolution of the Superior Province, Canada. *Geological Association of Canada Special Paper* 49, 321–378.
- Percival, J.A., Peterman, Z.E., 1994. Rb–Sr biotite and whole-rock data from the Kapuskasing uplift and their bearing on the cooling and exhumation history. *Can. J. Earth Sci.* 31, 1172–1181. <https://doi.org/10.1139/e94-103>
- Percival, J.A., Sullivan, R.W., 1988. Age constraints on the evolution of the Quetico Belt, Superior Province, Ontario (No. 88–2). <https://doi.org/10.4095/126607>
- Percival, J.A., Williams, H.R., 1989. Late Archean Quetico accretionary complex, Superior province, Canada. *Geol* 17, 23. [https://doi.org/10.1130/0091-7613\(1989\)017%3C0023:LAQACS%3E2.3.CO;2](https://doi.org/10.1130/0091-7613(1989)017%3C0023:LAQACS%3E2.3.CO;2)
- Peterman, Z.E., Day, W., 1989. Early Proterozoic activity on Archean faults in the western Superior province— Evidence from pseudotachylite. *Geol* 17, 1089. [https://doi.org/10.1130/0091-7613\(1989\)017%3C1089:EPAOAF%3E2.3.CO;2](https://doi.org/10.1130/0091-7613(1989)017%3C1089:EPAOAF%3E2.3.CO;2)
- Pirie, J., 1978. *Geology of the Crooked Pine Lake Area, District of Rainy River. Ontario Geological Survey Report* 179.
- Platt, J.P., Behr, W.M., 2011. Lithospheric shear zones as constant stress experiments. *Geology* 39, 127–130. <https://doi.org/10.1130/G31561.1>
- Poirier, J.P., 1980. Shear localization and shear instability in materials in the ductile field. *Journal of Structural Geology*. [https://doi.org/10.1016/0191-8141\(80\)90043-7](https://doi.org/10.1016/0191-8141(80)90043-7)
- R. Jones, R., Geoff Tanner, P.W., 1995. Strain partitioning in transpression zones. *Journal of Structural Geology* 17, 793–802. [https://doi.org/10.1016/0191-8141\(94\)00102-6](https://doi.org/10.1016/0191-8141(94)00102-6)
- Ramsay, J.G., Graham, R.H., 1970. Strain variation in shear belts. *Can. J. Earth Sci.* 7, 786–813. <https://doi.org/10.1139/e70-078>
- Ranalli, G., 1987. *Rheology of the Earth: Deformation and Flow Processes in Geophysics and Geodynamics - Ranalli, Giorgio: 9780045511112 - AbeBooks [WWW Document]. URL https://www.abebooks.com/9780045511112/Rheology-Earth-Deformation-Flow-Processes-004551111X/plp (accessed 4.8.26).*
- Ree, J.-H., Park, Y., 1997. Static recovery and recrystallization microstructures in sheared octachloropropane. *Journal of Structural Geology* 19, 1521–1526. [https://doi.org/10.1016/S0191-8141\(97\)00067-9](https://doi.org/10.1016/S0191-8141(97)00067-9)
- Rehm, A., Godet, A., Tinkham, D., Guilmette, C., Smit, M., 2026. Timing of Metamorphism and Deformation in the Neoarchean Quetico Metasedimentary Belt, Superior Craton, Constrained by Garnet Trace-Element Mapping and Lu–Hf Geochronology. *Journal of Metamorphic Geology* 44. <https://doi.org/10.1111/jmg.70034>

- Richards, J.P., Tosdal, R.M. (Eds.), 2001. Structural controls on ore genesis, Reviews in economic geology. Soc. of Economic Geologists, Knoxville, Tenn.
- Rogala, B., Fralick, P.W., Heaman, L.M., Metsaranta, R., 2007. Lithostratigraphy and chemostratigraphy of the Mesoproterozoic Sibley Group, northwestern Ontario, Canada. *Canadian Journal of Earth Sciences = Revue Canadienne des Sciences de la Terre* 44, 1131–1149. <https://doi.org/10.1139/e07-027>
- Rosenberg, C.L., 2004. Shear zones and magma ascent: A model based on a review of the Tertiary magmatism in the Alps. *Tectonics* 23. <https://doi.org/10.1029/2003TC001526>
- Rowe, C.D., Griffith, W.A., 2015. Do faults preserve a record of seismic slip: A second opinion. *Journal of Structural Geology* 78, 1–26. <https://doi.org/10.1016/j.jsg.2015.06.006>
- Rutter, E.H., 1999. On the relationship between the formation of shear zones and the form of the flow law for rocks undergoing dynamic recrystallization. *Tectonophysics* 303, 147–158. [https://doi.org/10.1016/S0040-1951\(98\)00261-3](https://doi.org/10.1016/S0040-1951(98)00261-3)
- Rutter, E.H., Brodie, K.H., 2004. Experimental grain size-sensitive flow of hot-pressed Brazilian quartz aggregates. *Journal of Structural Geology* 26, 2011–2023. <https://doi.org/10.1016/j.jsg.2004.04.006>
- Sanborn-Barrie, M., Skulski, T., 2006. Sedimentary and structural evidence for 2.7 Ga continental arc–oceanic-arc collision in the Savant–Sturgeon greenstone belt, western Superior Province, Canada. *Can. J. Earth Sci.* 43, 995–1030. <https://doi.org/10.1139/e06-060>
- Sawyer, E.W., 1998. Formation and Evolution of Granite Magmas During Crustal Reworking: the Significance of Diatexites. *J Petrology* 39, 1147–1167. <https://doi.org/10.1093/petroj/39.6.1147>
- Sawyer, E.W., 1983. The structural history of a part of the Archaean Quetico Metasedimentary Belt, superior province, Canada. *Precambrian Research* 22, 271–294. [https://doi.org/10.1016/0301-9268\(83\)90052-9](https://doi.org/10.1016/0301-9268(83)90052-9)
- Sawyer, E.W., Barnes, S. -J., 1988. Temporal and compositional differences between subsolidus and anatectic migmatite leucosomes from the Quetico metasedimentary belt, Canada. *Journal Metamorphic Geology* 6, 437–450. <https://doi.org/10.1111/j.1525-1314.1988.tb00432.x>
- Schmitz, M.D., Bowring, S.A., Southwick, D.L., Boerboom, T., Wirth, K.R., 2006. High-precision U-Pb geochronology in the Minnesota River Valley subprovince and its bearing on the Neoarchean to Paleoproterozoic evolution of the southern Superior Province. *Geological Society of America Bulletin* 118, 82–93. <https://doi.org/10.1130/B25725.1>
- Schmitz, M.D., Southwick, D.L., Bickford, M.E., Mueller, P.A., Samson, S.D., 2018. Neoarchean and Paleoproterozoic events in the Minnesota River Valley subprovince, with implications for southern Superior craton evolution and correlation. *Precambrian Research* 316, 206–226. <https://doi.org/10.1016/j.precamres.2018.08.010>
- Schwerdtner, W.M., Stone, D., Osadetz, K., Morgan, J., Stott, G.M., 1979. Granitoid complexes and the Archean tectonic record in the southern part of northwestern Ontario. *Canadian Journal of Earth Sciences* 16, 1965–1977. <https://doi.org/10.1139/e79-183>
- Shklanka, R., 1972. Geology of the Steep Rock Lake Area, District of Rainy River, Geological Report 93 pt. 1. Ontario Department of Mines.
- Sibson, R., 1977. Fault rocks and fault mechanics. *Journal of the Geological Society* 133, 191–213. <https://doi.org/10.1144/gsjgs.133.3.0191>

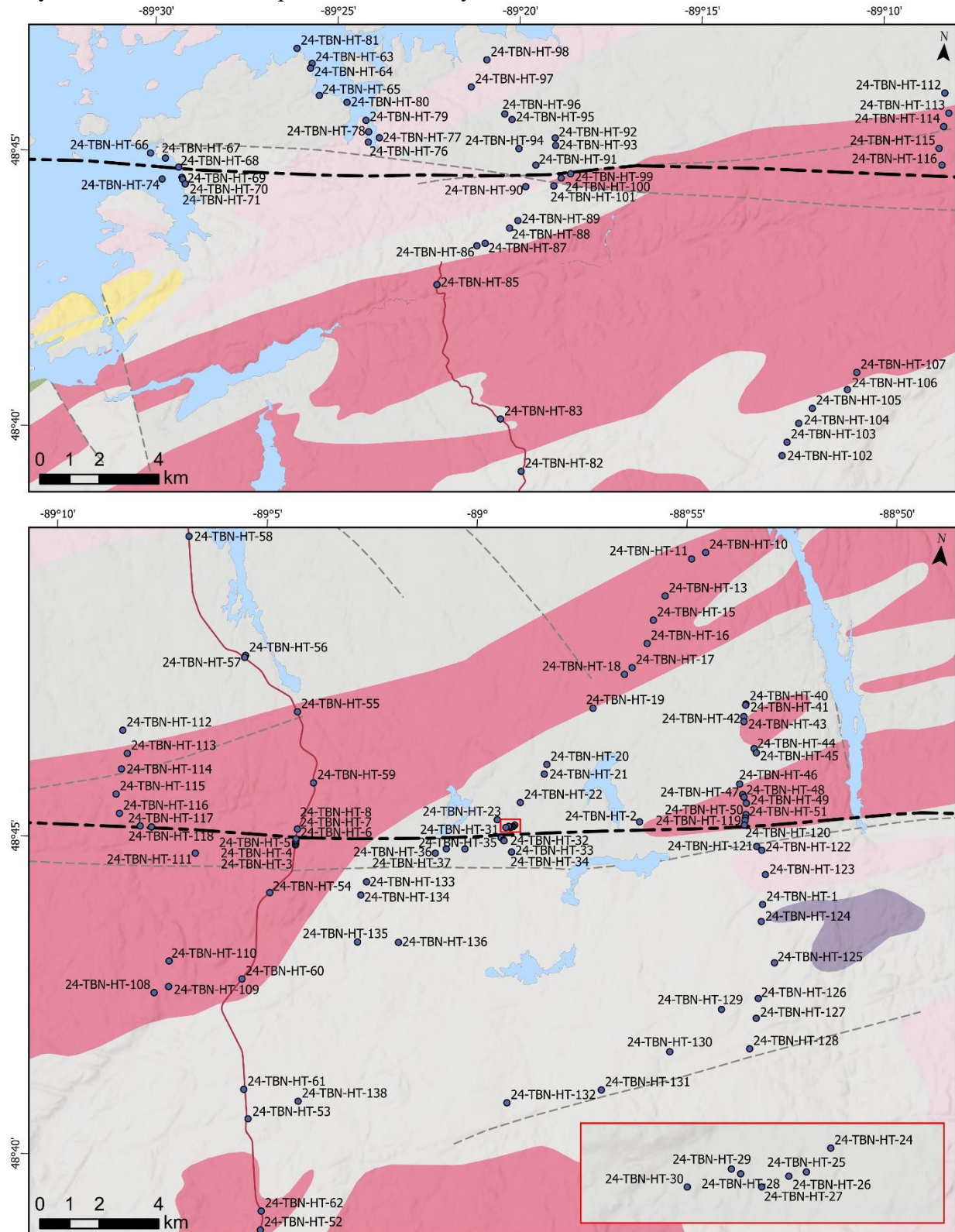
- Sibson, R.H., 1984. Roughness at the base of the seismogenic zone: Contributing factors. *Journal of Geophysical Research: Solid Earth* 89, 5791–5799. <https://doi.org/10.1029/JB089iB07p05791>
- Sims, P., Schmus, W.V., Schulz, K., Peterman, Z., 1989. Tectono-stratigraphic evolution of the Early Proterozoic Wisconsin magmatic terranes of the Penokean Orogen. *Can. J. Earth Sci.* 26, 2145–2158. <https://doi.org/10.1139/e89-180>
- Spear, F.S., Pyle, J.M., 2002. Apatite, Monazite, and Xenotime in Metamorphic Rocks. *Reviews in Mineralogy and Geochemistry* 48, 293–335. <https://doi.org/10.2138/rmg.2002.48.7>
- Spinler, J.C., Bennett, R.A., Anderson, M.L., McGill, S.F., Hreinsdóttir, S., McCallister, A., 2010. Present-day strain accumulation and slip rates associated with southern San Andreas and eastern California shear zone faults. <https://doi.org/10.1029/2010JB007424>
- Stein, H.J., 2014. Dating and Tracing the History of Ore Formation, in: *Treatise on Geochemistry*. Elsevier, pp. 87–118. <https://doi.org/10.1016/B978-0-08-095975-7.01104-9>
- Stephan, T., 2025. *structr: Structural Geology in R*.
- Stephan, T., Phillips, N.J., Tiitto, H., Perez, A., Nwakanma, M., Creaser, R., Hollings, P., 2025. Going with the flow — Changes of vorticity control gold enrichment in Archean shear zones (Shebandowan Greenstone Belt, Superior Province, Canada). *Journal of Structural Geology* 201, 105542. <https://doi.org/10.1016/j.jsg.2025.105542>
- Stipp, M., Stunitz, H., Heilbronner, R., Schmid, S., 2002. Dynamic recrystallization of quartz: correlation between natural and experimental conditions. *Geological Society of London, Special Publications* 200, 171–190.
- Stott, G.M., 1997. The Superior Province, Canada., in: de Wit, M.J., Ashwal, L.D. (Eds.), *Greenstone Belts*, Oxford Monograph on Geology and Geophysics. Oxford Clarendon Press, pp. 480–507.
- Stott, G.M., Corfu, F., Breaks, F.W., Thurston, P.C., 1989. Multiple orogenesis in northwestern Superior Province, in: *Program with Abstracts*. Presented at the Annual meeting of the Geological Association of Canada, Annual meeting of the Geological Association of Canada, p. A56.
- Stott, G.M., Corkery, M.T., Percival, J.A., Simard, M., Goutier, J., 2010. A Revised Terrane Subdivision of the Superior Province in Ontario (Open File Report No. 6260), Summary of Field Work and Other Activities 2010. Ontario Geological Survey.
- Sun, McDonough, W., 1989. Chemical and isotopic systematics of oceanic basalts: Implications for mantle composition and processes, in: *Geological Society, London, Special Publications*. <https://doi.org/10.1144/GSL.SP.1989.042.01.19>
- Sylvester, P.J., Attoh, K., Schulz, K.J., 1987. Tectonic setting of late Archean bimodal volcanism in the Michipicoten (Wawa) greenstone belt, Ontario. *Can. J. Earth Sci.* 24, 1120–1134. <https://doi.org/10.1139/e87-109>
- Talbot, C., 1973. A Plate Tectonic Model for the Archaean Crust. *Philosophical Transactions of The Royal Society B: Biological Sciences* 273, 413–427. <https://doi.org/10.1098/rsta.1973.0008>
- Thatcher, W., 1975. Strain accumulation on the northern San Andreas Fault Zone since 1906. *Journal of Geophysical Research* 80, 4873–4880. <https://doi.org/10.1029/JB080i035p04873>

- Thomas, J.B., Bruce Watson, E., Spear, F.S., Shemella, P.T., Nayak, S.K., Lanzirotti, A., 2010. TitaniQ under pressure: the effect of pressure and temperature on the solubility of Ti in quartz. *Contrib Mineral Petrol* 160, 743–759. <https://doi.org/10.1007/s00410-010-0505-3>
- Thurston, P.C., 1985. Geology of the Earngey-Costello Area; District of Kenora, Patricia Portion.
- Tokle, L., Hirth, G., Behr, W.M., 2019. Flow laws and fabric transitions in wet quartzite. *Earth and Planetary Science Letters* 505, 152–161. <https://doi.org/10.1016/j.epsl.2018.10.017>
- Tollefson, K.T., Larson, K.P., Moynihan, D.P., Cleven, N.R., 2025. In situ mica Rb–Sr geochronology and microstructural analysis to evaluate the timing and sense of shear zone movement: Shannon and d’Abbadie Faults, Yukon, Canada. *Can. J. Earth Sci.* 62, 1320–1341. <https://doi.org/10.1139/cjes-2024-0144>
- Tullis, J., Yund, R., 1992. Chapter 4 The Brittle-Ductile Transition in Feldspar Aggregates: An Experimental Study, in: *International Geophysics*. Elsevier, pp. 89–117. [https://doi.org/10.1016/S0074-6142\(08\)62816-8](https://doi.org/10.1016/S0074-6142(08)62816-8)
- Walker, J.D., Tikoff, B., Newman, J., Clark, R., Ash, J., Good, J., Bunse, E.G., Möller, A., Kahn, M., Williams, R.T., Michels, Z., Andrew, J.E., Rufledt, C., 2019. StraboSpot data system for structural geology. *Geosphere* 15, 533–547. <https://doi.org/10.1130/GES02039.1>
- Wardle, R.J., James, D.T., Scott, D.J., Hall, J., 2002. The southeastern Churchill Province: synthesis of a Paleoproterozoic transpressional orogen1. *Canadian Journal of Earth Sciences* 39, 639–663. <https://doi.org/10.1139/e02-004>
- Wark, D.A., Watson, E.B., 2006. TitaniQ: a titanium-in-quartz geothermometer. *Contrib Mineral Petrol* 152, 743–754. <https://doi.org/10.1007/s00410-006-0132-1>
- White, R.W., Powell, R., Holland, T.J.B., 2001. Calculation of partial melting equilibria in the system Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O (NCKFMASH). *Journal of Metamorphic Geology* 19, 139–153. <https://doi.org/10.1046/j.0263-4929.2000.00303.x>
- Wood, J., 1980. Epiclastic sedimentation and stratigraphy in the North Spirit Lake and Rainy Lake areas: A comparison. *Precambrian Research* 12, 227–255. [https://doi.org/10.1016/0301-9268\(80\)90030-3](https://doi.org/10.1016/0301-9268(80)90030-3)
- Young, G.M., 1978. Some Aspects of the Evolution of the Archean Crust. *Geoscience Canada*.
- Zoback, M., Hickman, S., Ellsworth, W., 2010. Scientific Drilling Into the San Andreas Fault Zone. *Eos, Transactions American Geophysical Union* 91, 197–199. <https://doi.org/10.1029/2010EO220001>

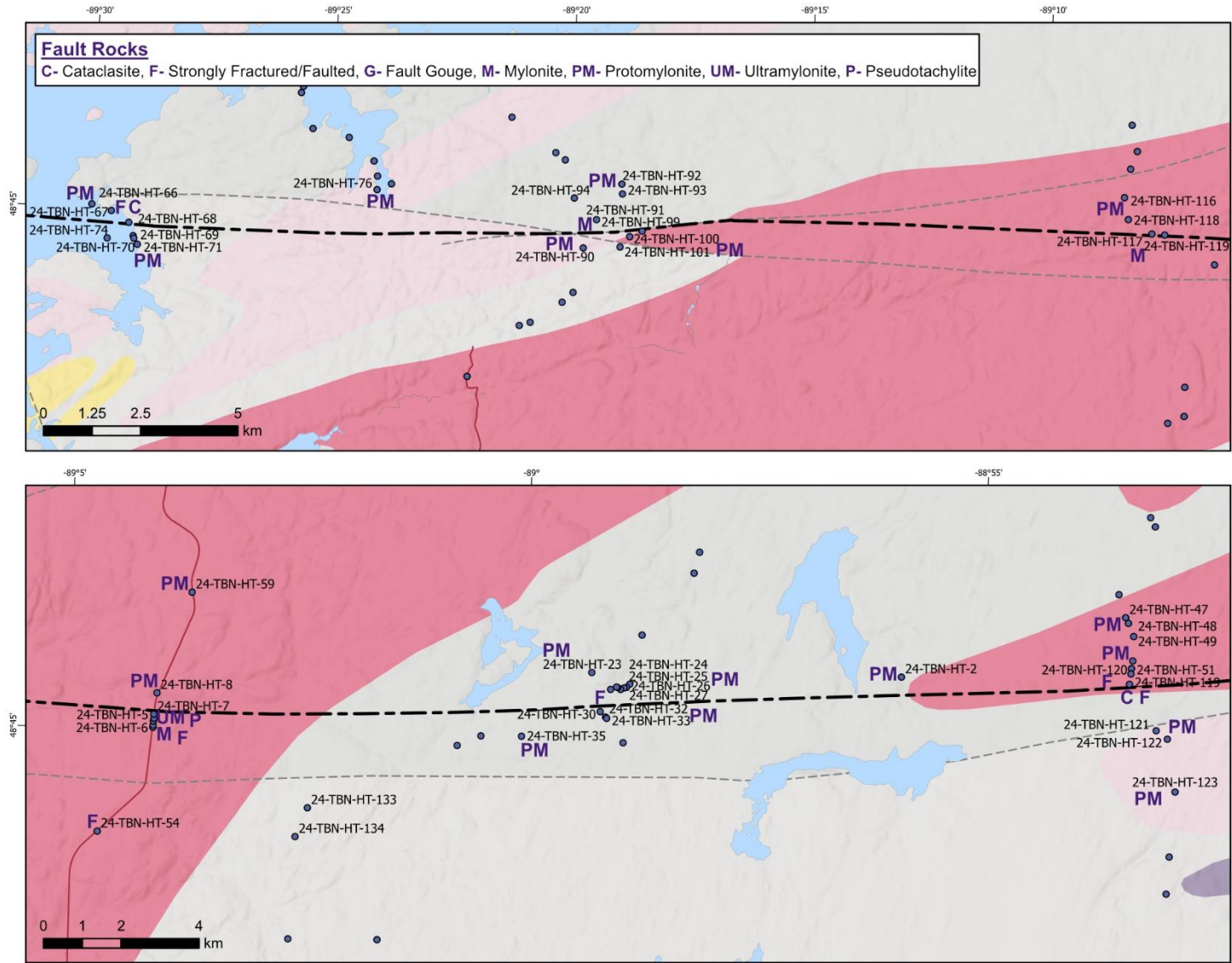
Appendices

Appendix A: Field Station and Sample Information

A1: Maps displaying field station locations within the study area. Top: Western portion of the study area. Bottom: Eastern portion of the study area:



A2: Maps displaying the location of fault rock types at field station within the shear zone within the study area. Top: Western portion of the study area. Bottom: Eastern portion of the study area:



A3: Table of field samples and the analyses applied to each sample:

Sample	Field Station	Lithology	Distance from the Quetico Shear Zone	Whole-rock Geochemistry	Polished Thin Sections	AMICS Scans	EBSD	SIMS	LASS-ICP-MS
TBN-HT-31	24-TBN-HT-4	Metavolcanic rock	195 m south	*	*		*	*	
TBN-HT-32	24-TBN-HT-6	Metavolcanic rock	110 m south	*	*		*	*	
TBN-HT-33	24-TBN-HT-7	Mylonitic granitic rock	45 m south	*	*	*	*		
TBN-HT-34	24-TBN-HT-7	Metavolcanic rock	45 m south	*	*	*			*
TBN-HT-35	24-TBN-HT-8	Granitic gneiss	240 m north	*	*	*	*		*
TBN-HT-36	24-TBN-HT-9	Granitic gneiss	380 m north	*	*		*	*	*
TBN-HT-100	24-TBN-HT-22	Granitoid	930 m north		*				
TBN-HT-101	24-TBN-HT-23	Granitic gneiss	455 m north	*	*				
TBN-HT-102	24-TBN-HT-25	Metasedimentary rock	230 m north	*	*	*			
TBN-HT-103	24-TBN-HT-26	Metasedimentary rock	225 m north	*	*				
TBN-HT-104	24-TBN-HT-30	Metasedimentary rock	85 m north	*	*				
TBN-HT-105	24-TBN-HT-31	Metasedimentary rock	75 m south	*	*				
TBN-HT-106	24-TBN-HT-32	Metasedimentary rock	150 m south	*	*	*			
TBN-HT-107	24-TBN-HT-36	Metasedimentary rock	325 m south		*				
TBN-HT-108	24-TBN-HT-41	Granitic gneiss	3.55 km north		*				
TBN-HT-109	24-TBN-HT-47	Granitic gneiss	940 m north		*				
TBN-HT-110	24-TBN-HT-49	Granitic gneiss	680 m north	*	*				
TBN-HT-111	24-TBN-HT-50	Granitic gneiss	350 m north	*	*				
TBN-HT-112	24-TBN-HT-51	Granitic gneiss	250 m north		*	*			*
TBN-HT-113	24-TBN-HT-6	Metavolcanic rock	100 m south	*					
TBN-HT-114	24-TBN-HT-6	Metavolcanic rock	100 m south		*	*	*	*	
TBN-HT-115	24-TBN-HT-55	Metasedimentary rock	3.68 km north	*	*	*			
TBN-HT-116	24-TBN-HT-59	Metasedimentary rock	1.61 km north	*	*	*			
TBN-HT-117	24-TBN-HT-66	Metasedimentary gneiss	435 m north	*	*	*			*
TBN-HT-118	24-TBN-HT-66	Metasedimentary gneiss	435 m north	*	*	*			*
TBN-HT-119	24-TBN-HT-68	Metasedimentary gneiss	40 m north	*	*	*	*	*	*
TBN-HT-120	24-TBN-HT-69	Granitic gneiss	290 m south		*	*			

TBN-HT-121	24-TBN-HT-75	Metasedimentary rock	620 m north	*	*	*	*	*
TBN-HT-122	24-TBN-HT-81	Metasedimentary rock	4.25 km north	*	*			
TBN-HT-123	24-TBN-HT-76	Metasedimentary gneiss	1.14 km north		*			
TBN-HT-124	24-TBN-HT-76	Metasedimentary gneiss	1.14 km north	*	*			
TBN-HT-125	24-TBN-HT-83	Granitoid	8.21 km south	*	*			
TBN-HT-126	24-TBN-HT-89	Granitoid	1.51 km south	*	*			
TBN-HT-127	24-TBN-HT-92	Metasedimentary rock	1.21 km north		*			
TBN-HT-128	24-TBN-HT-96	Metasedimentary rock	2.07 km north		*			
TBN-HT-129	24-TBN-HT-98	Metasedimentary gneiss	3.90 km north	*	*			
TBN-HT-130	24-TBN-HT-99	Granitic gneiss	50 m north	*	*			
TBN-HT-131	24-TBN-HT-103	Metasedimentary rock	9.17 km south	*	*			
TBN-HT-132	24-TBN-HT-104	Metagabbro	8.53 km south	*	*			
TBN-HT-133	24-TBN-HT-109	Metasedimentary rock	4.62 km south	*	*			
TBN-HT-134	24-TBN-HT-111	Metasedimentary rock	680 m south	*	*	*		
TBN-HT-135	24-TBN-HT-124	Metasedimentary rock	2.79 km south	*	*			
TBN-HT-136	24-TBN-HT-126	Metasedimentary rock	5.02 km south	*	*			

A4: Table of latitude and longitude locations for each field sample:

Sample	Field Station	Latitude	Longitude
TBN-HT-31	24-TBN-HT-4	-89.0673582	48.7506442
TBN-HT-32	24-TBN-HT-6	-89.0672644	48.7514836
TBN-HT-33	24-TBN-HT-7	-89.0673133	48.7519677
TBN-HT-34	24-TBN-HT-7	-89.0673133	48.7519677
TBN-HT-35	24-TBN-HT-8	-89.0668569	48.7545609
TBN-HT-36	24-TBN-HT-9	-89.0659706	48.7557898
TBN-HT-100	24-TBN-HT-22	-88.9788190	48.7636830
TBN-HT-101	24-TBN-HT-23	-88.9877311	48.7589568
TBN-HT-102	24-TBN-HT-25	-88.9813031	48.7573156
TBN-HT-103	24-TBN-HT-26	-88.9817280	48.7572374
TBN-HT-104	24-TBN-HT-30	-88.9841960	48.7542759
TBN-HT-105	24-TBN-HT-31	-88.9859441	48.7542759
TBN-HT-106	24-TBN-HT-32	-88.9849444	48.7536512
TBN-HT-107	24-TBN-HT-36	-89.0075230	48.7508320
TBN-HT-108	24-TBN-HT-41	-88.8908294	48.7913675
TBN-HT-109	24-TBN-HT-47	-88.8907460	48.7678790
TBN-HT-110	24-TBN-HT-49	-88.8891565	48.7656477
TBN-HT-111	24-TBN-HT-50	-88.8891739	48.7626960
TBN-HT-112	24-TBN-HT-51	-88.8893590	48.7617680
TBN-HT-113	24-TBN-HT-6	-89.0672644	48.7514836
TBN-HT-114	24-TBN-HT-6	-89.0672644	48.7514836
TBN-HT-115	24-TBN-HT-55	-89.0685708	48.7853285
TBN-HT-116	24-TBN-HT-59	-89.0611556	48.7668359
TBN-HT-117	24-TBN-HT-66	-89.4999500	48.7507065
TBN-HT-118	24-TBN-HT-66	-89.4999500	48.7507065
TBN-HT-119	24-TBN-HT-68	-89.4867883	48.7468171
TBN-HT-120	24-TBN-HT-69	-89.4850760	48.7437660
TBN-HT-121	24-TBN-HT-75	-89.5054577	48.7525168
TBN-HT-122	24-TBN-HT-81	-89.4349608	48.7842847
TBN-HT-123	24-TBN-HT-76	-89.4005440	48.7568090
TBN-HT-124	24-TBN-HT-76	-89.4005443	48.7568088
TBN-HT-125	24-TBN-HT-83	-89.3348437	48.6747091
TBN-HT-126	24-TBN-HT-89	-89.3306246	48.7349697
TBN-HT-127	24-TBN-HT-92	-89.3234630	48.7519830
TBN-HT-128	24-TBN-HT-96	-88.8844150	48.7012450
TBN-HT-129	24-TBN-HT-98	-89.3478552	48.7832309
TBN-HT-130	24-TBN-HT-99	-89.3073645	48.7497815
TBN-HT-131	24-TBN-HT-103	-89.2035296	48.6711835
TBN-HT-132	24-TBN-HT-104	-89.1985462	48.6770429
TBN-HT-133	24-TBN-HT-109	-89.1155260	48.7119443

TBN-HT-134	24-TBN-HT-111	-89.1069664	48.7472126
TBN-HT-135	24-TBN-HT-124	-88.8815856	48.7347469
TBN-HT-136	24-TBN-HT-126	-88.8816939	48.7145351

A5: Table of drill core samples and the analyses applied to each sample:

Sample	Drill Hole	Lithology	Whole-rock Geochemistry	Polished Thin Sections
TBN-HT-1	BL09-153	Granitic gneiss		
TBN-HT-2	BL09-153	Granitic gneiss		
TBN-HT-3	BL09-153	Granitic gneiss		
TBN-HT-4	BL09-153	Granitic gneiss		
TBN-HT-5	BL09-153	Granitic gneiss		
TBN-HT-6	BL09-153	Granitic gneiss		
TBN-HT-7	BL09-153	Granitic gneiss		
TBN-HT-8	BL09-153	Granitic gneiss		
TBN-HT-9	BL09-153	Granitic gneiss		
TBN-HT-10	BL09-153	Granitic gneiss		
TBN-HT-11	BL09-153	Granitic gneiss		
TBN-HT-12	BL09-153	Granitic gneiss		
TBN-HT-13	BL09-153	Granitic gneiss		
TBN-HT-14	BL09-153	Granitic gneiss		
TBN-HT-15	BL09-153	Granitic gneiss		
TBN-HT-16	BL09-153	Granitic gneiss		
TBN-HT-17	BL09-153	Granitic gneiss		
TBN-HT-18	BL09-153	Granitic gneiss		
TBN-HT-19	BL09-153	Metasedimentary rock		
TBN-HT-20	BL09-153	Metasedimentary rock		
TBN-HT-21	BL09-153	Metasedimentary rock		
TBN-HT-22	BL09-153	Metasedimentary rock		
TBN-HT-23	BL09-153	Mafic intrusive rock		
TBN-HT-24	BL09-153	Mafic intrusive rock		
TBN-HT-25	BL09-153	Mafic intrusive rock		
TBN-HT-26	BL09-153	Mafic intrusive rock		
TBN-HT-27	BL09-153	Mafic intrusive rock		
TBN-HT-28	BL09-153	Mafic intrusive rock		
TBN-HT-29	BL09-153	Mafic intrusive rock		
TBN-HT-30	BL09-153	Mafic intrusive rock		
TBN-HT-37	BL10-251	Dioritic rock		
TBN-HT-38	BL10-251	Dioritic rock		
TBN-HT-39	BL10-251	Dioritic rock		
TBN-HT-40	BL10-251	Dioritic rock		
TBN-HT-41	BL10-251	Dioritic rock		

TBN-HT-42	BL10-251	Dioritic rock		
TBN-HT-43	BL10-251	Dioritic rock		
TBN-HT-44	BL10-251	Dioritic rock		
TBN-HT-45	BL10-251	Dioritic rock		
TBN-HT-46	BL10-251	Dioritic rock		
TBN-HT-47	BL10-251	Dioritic rock		
TBN-HT-48	BL10-251	Dioritic rock		
TBN-HT-49	BL10-251	Dioritic rock		
TBN-HT-50	BL10-251	Red intrusive rock		
TBN-HT-51	BL10-251	Red intrusive rock		
TBN-HT-52	BL10-251	Red intrusive rock		
TBN-HT-53	BL10-251	Red intrusive rock		
TBN-HT-54	BL10-251	Mafic intrusive rock		
TBN-HT-55	BL10-251	Mafic intrusive rock		
TBN-HT-56	BL09-150	Dioritic rock		
TBN-HT-57	BL09-150	Dioritic rock		
TBN-HT-58	BL09-150	Dioritic rock		
TBN-HT-59	BL09-150	Dioritic rock		
TBN-HT-60	BL09-150	Dioritic rock		
TBN-HT-61	BL09-150	Dioritic rock		
TBN-HT-62	BL09-150	Dioritic rock		
TBN-HT-63	BL09-150	Dioritic rock		
TBN-HT-64	BL09-150	Dioritic rock		
TBN-HT-65	BL09-150	Dioritic rock		
TBN-HT-66	BL09-150	Dioritic rock		
TBN-HT-67	BL09-150	Mafic intrusive rock		
TBN-HT-68	BL09-150	Mafic intrusive rock		
TBN-HT-69	BL09-150	Mafic intrusive rock		
TBN-HT-70	BL09-150	Mafic intrusive rock		
TBN-HT-71	BL09-141	Dioritic rock	*	*
TBN-HT-72	BL09-141	Dioritic rock	*	*
TBN-HT-73	BL09-141	Dioritic rock	*	*
TBN-HT-74	BL09-141	Dioritic rock	*	*
TBN-HT-75	BL09-141	Dioritic rock	*	*
TBN-HT-76	BL09-141	Dioritic rock	*	*
TBN-HT-77	BL09-141	Dioritic rock	*	*
TBN-HT-78	BL09-141	Dioritic rock	*	*
TBN-HT-79	BL09-141	Metasedimentary rock	*	*
TBN-HT-80	BL09-141	Metasedimentary rock		*
TBN-HT-81	BL09-141	Metasedimentary rock	*	*
TBN-HT-82	BL09-141	Metasedimentary rock	*	*
TBN-HT-83	BL09-141	Red intrusive rock	*	*
TBN-HT-84	BL09-141	Mafic intrusive rock	*	*

TBN-HT-85	BL09-141	Mafic intrusive rock	*	*
TBN-HT-86	BL09-141	Mafic intrusive rock	*	*
TBN-HT-87	BL09-141	Mafic intrusive rock	*	*
TBN-HT-88	BL09-141	Mafic intrusive rock		
TBN-HT-89	BL09-141	Mafic intrusive rock		
TBN-HT-90	BL10-299	Metasedimentary rock		
TBN-HT-91	BL10-299	Metasedimentary rock		
TBN-HT-92	BL10-299	Granitic gneiss		
TBN-HT-93	BL10-299	Granitic gneiss		
TBN-HT-94	BL10-299	Metasedimentary rock		
TBN-HT-95	BL10-299	Granitic gneiss		
TBN-HT-96	BL10-299	Granitic gneiss		
TBN-HT-97	BL10-299	Metasedimentary rock		
TBN-HT-98	BL10-299	Metasedimentary rock		
TBN-HT-99	BL10-299	Mafic intrusive rock		

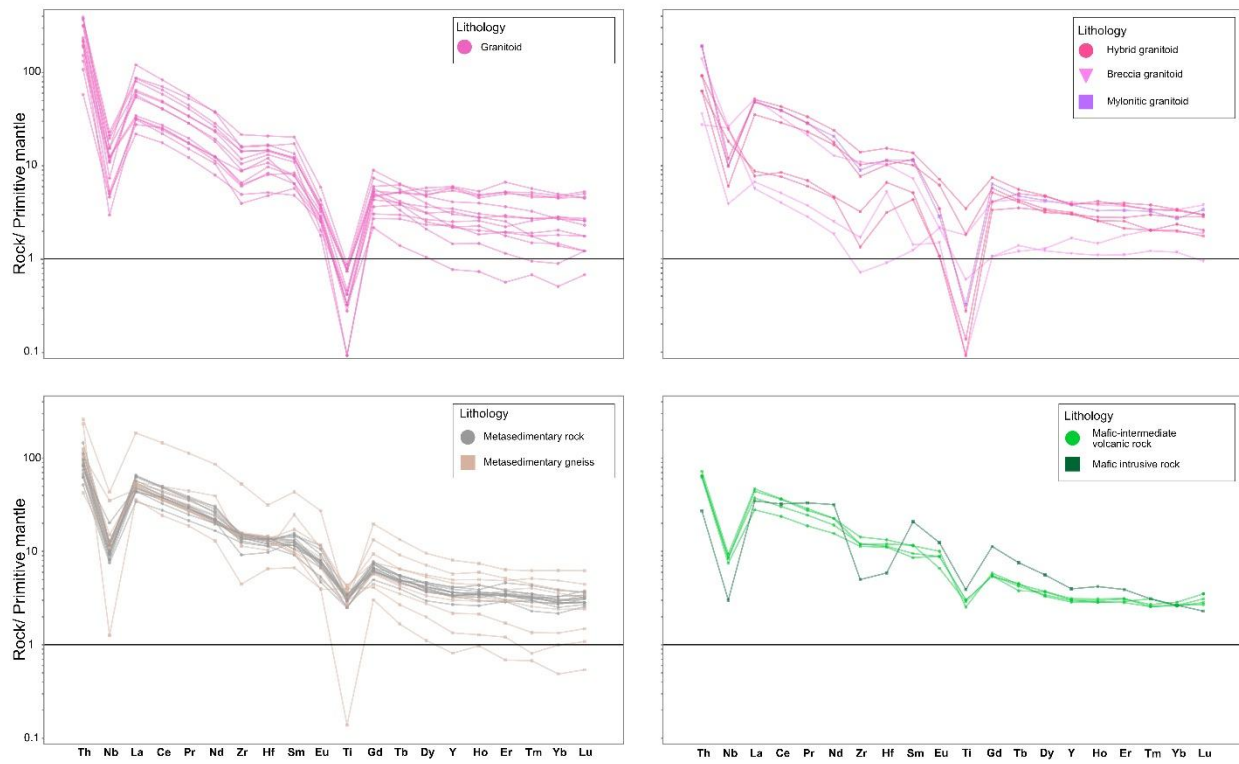
Appendix B: THERIAK-DOMINO Data

B1: Whole-rock geochemistry spider plots:

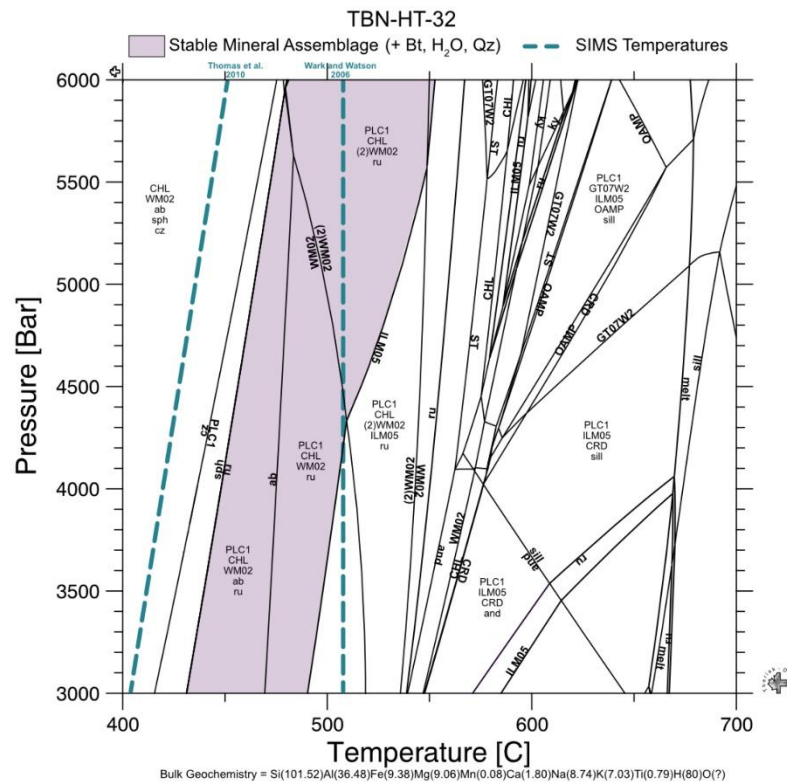
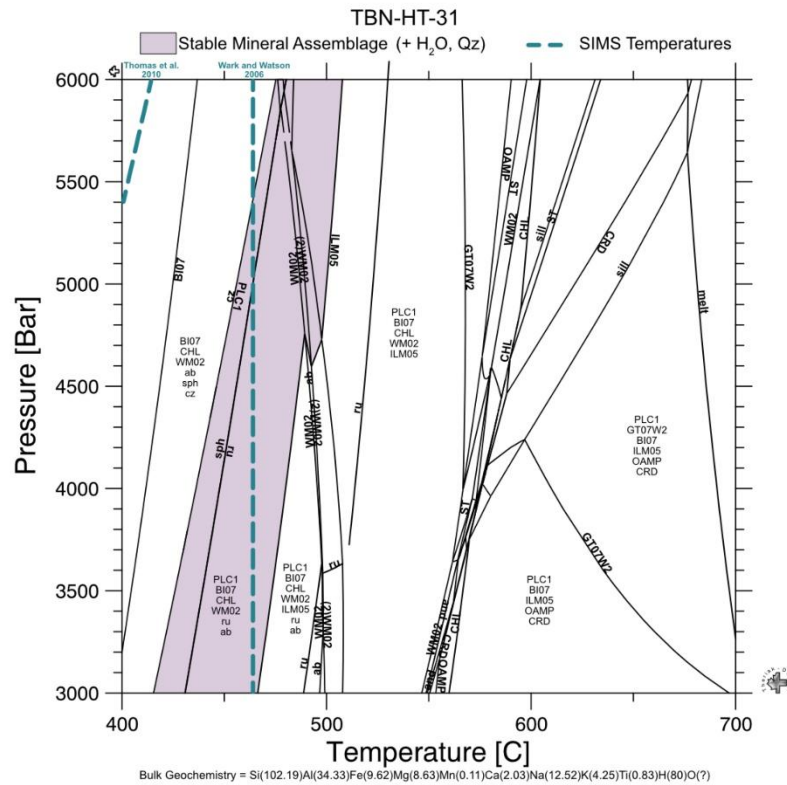
MORB-normalized (Pearce, 1983)

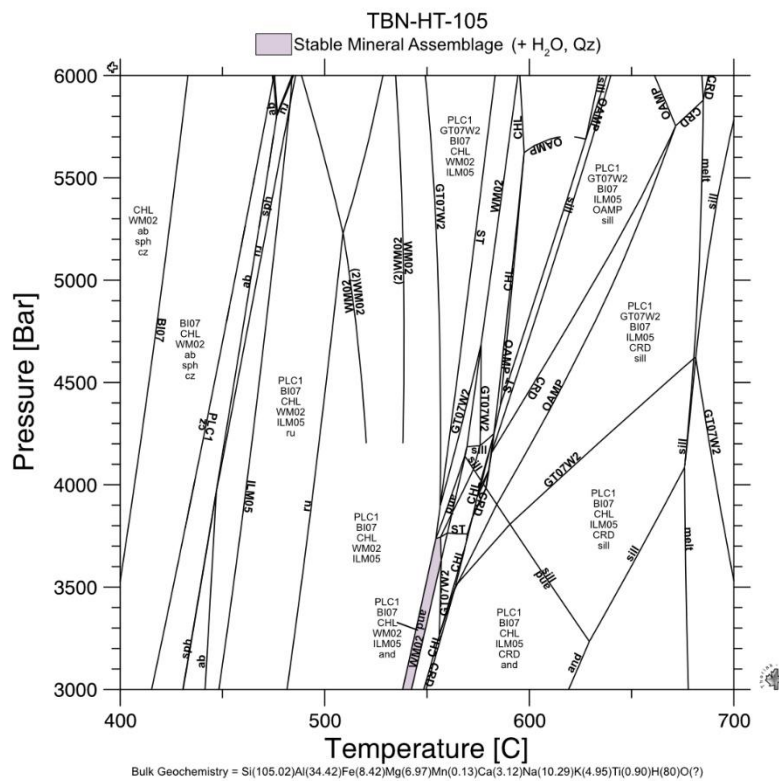
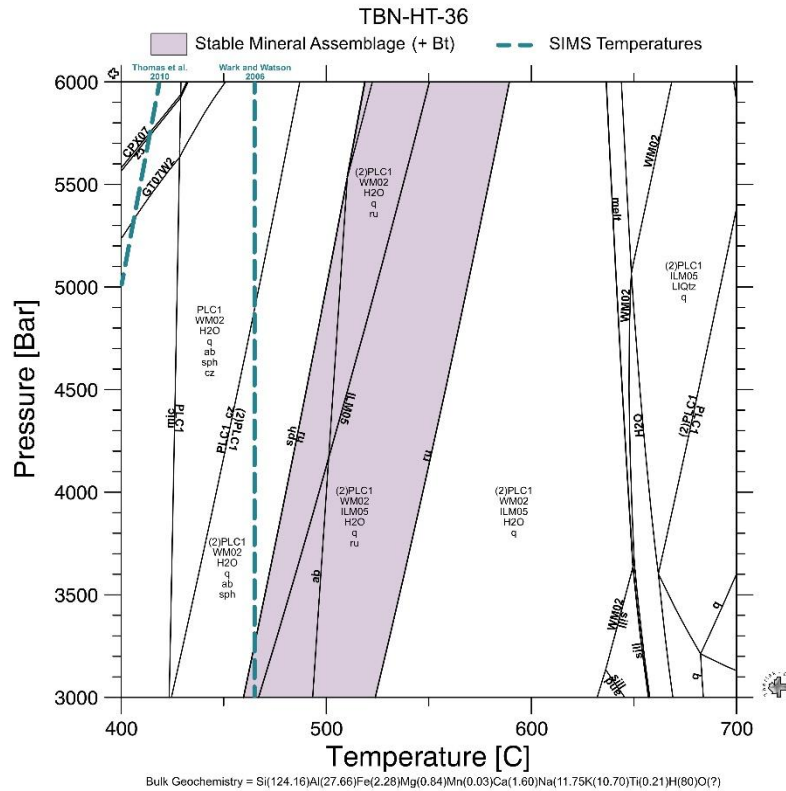


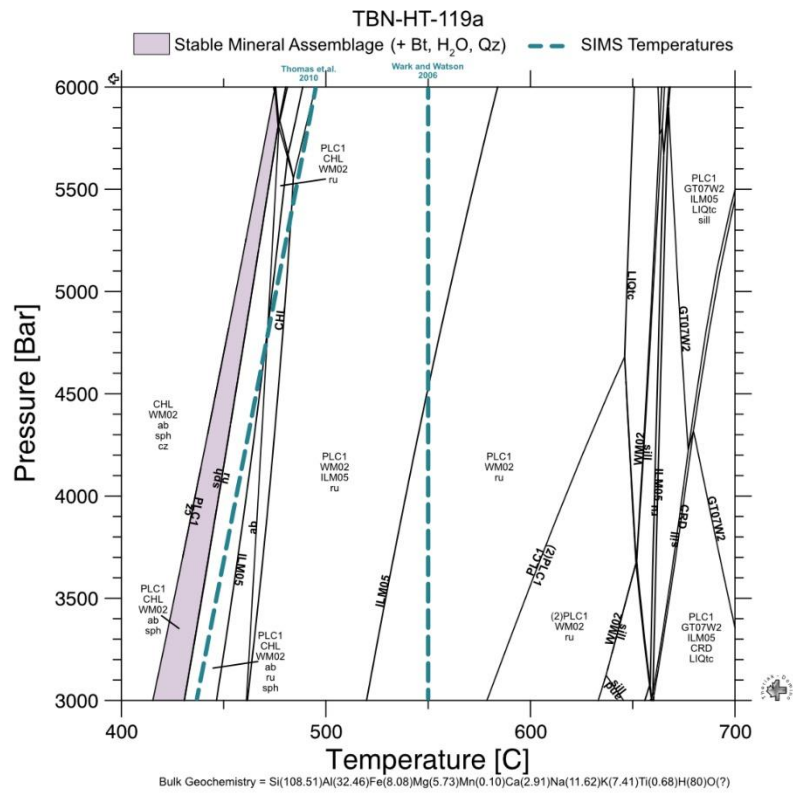
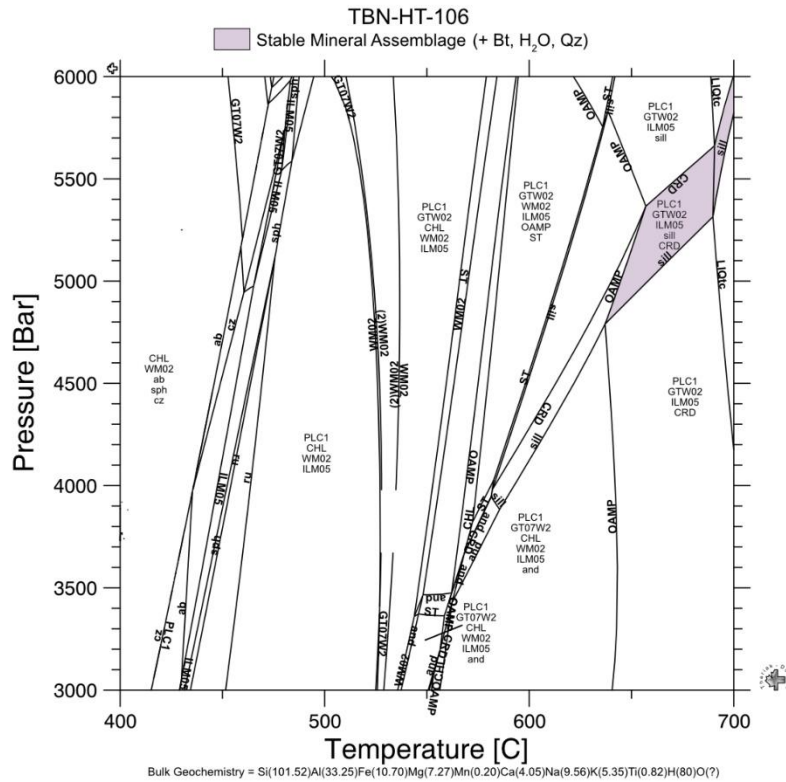
Primitive Mantle-normalized (Sun and McDonough, 1989)



B2: Remaining equilibrium phase diagrams used to estimate stable mineral assemblage temperatures:

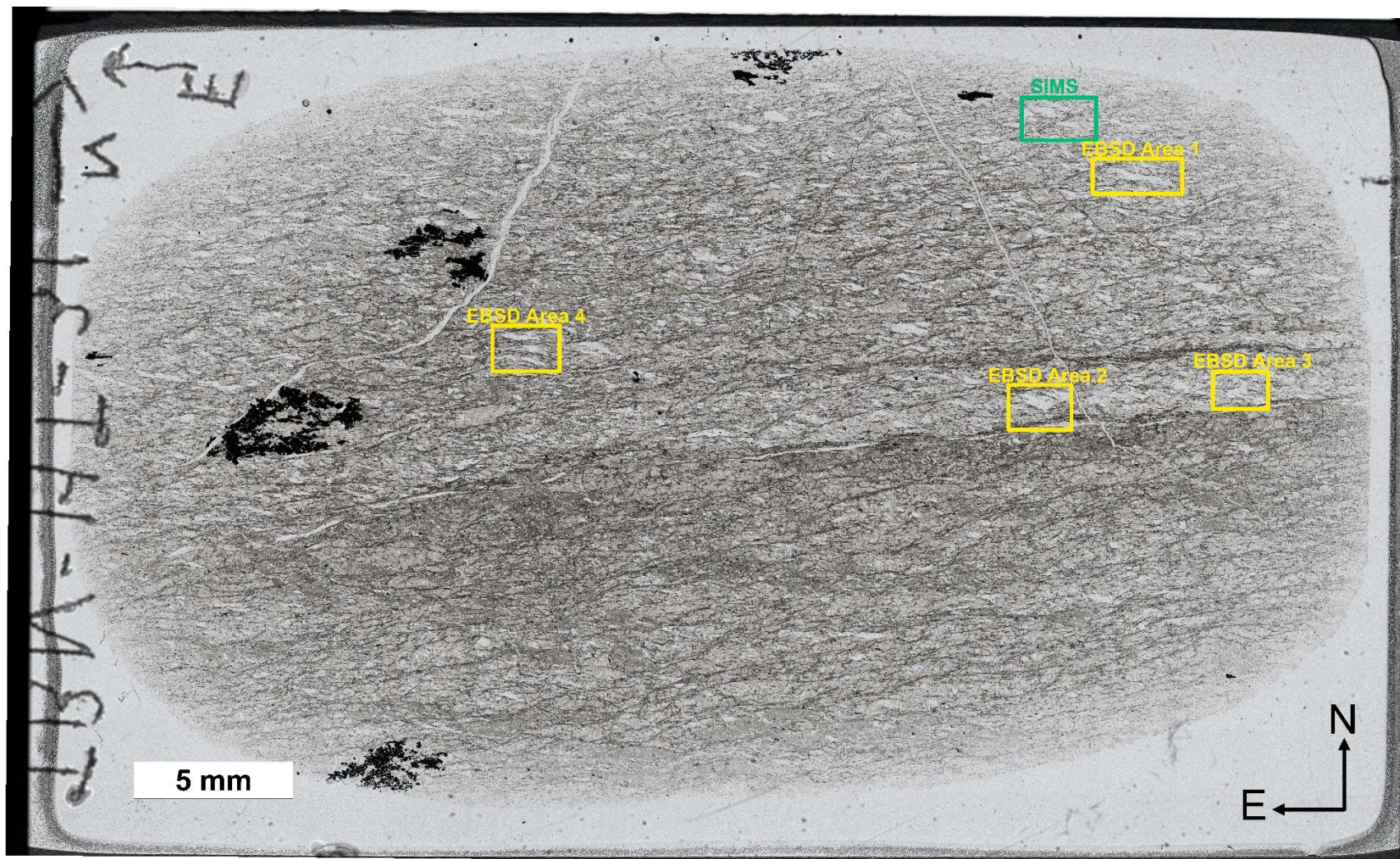




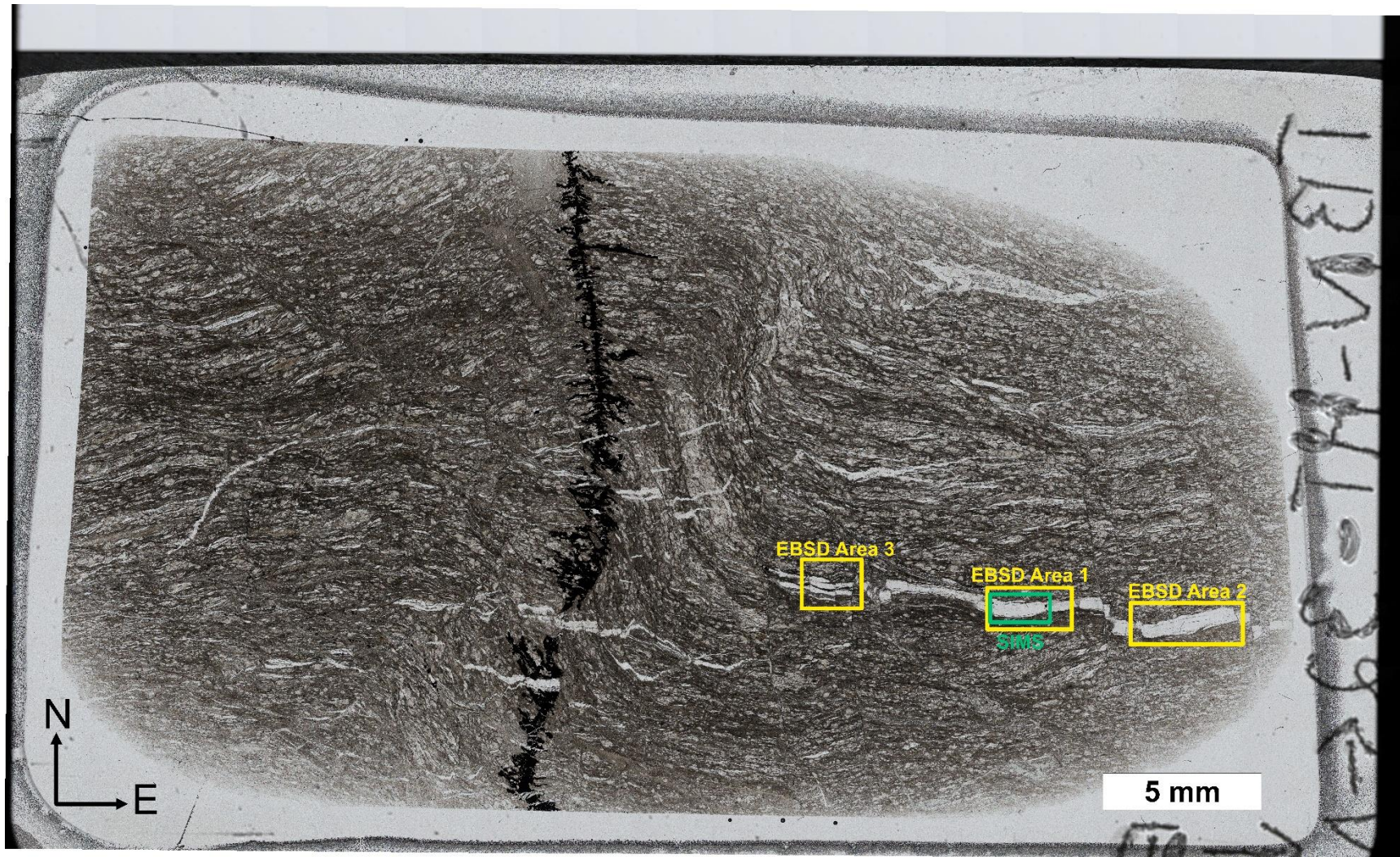


Appendix C: Electron Backscatter Diffraction Data

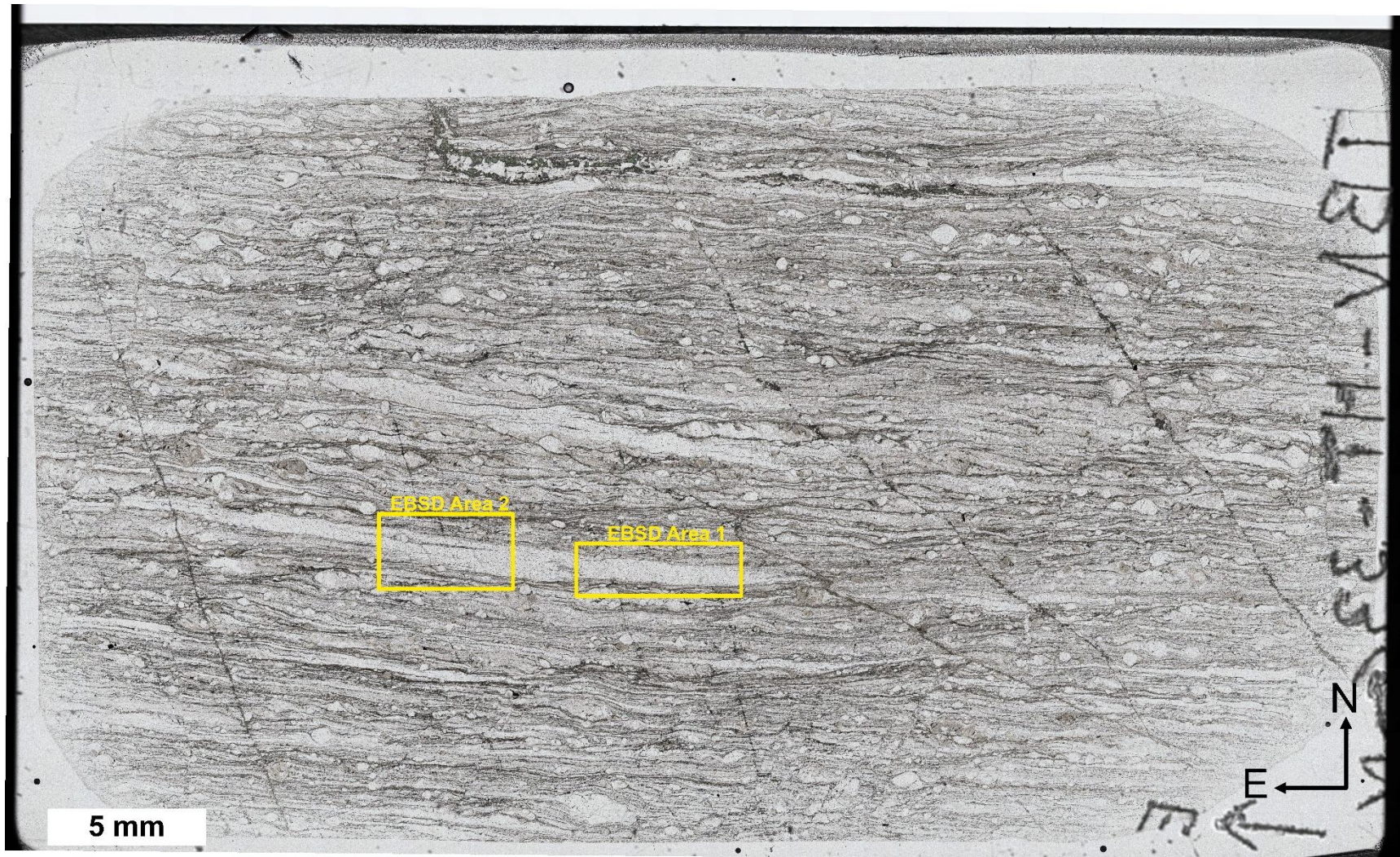
C1: Full thin section scans of samples with the locations of analyzed EBSD and SIMS areas:
TBN-HT-31:



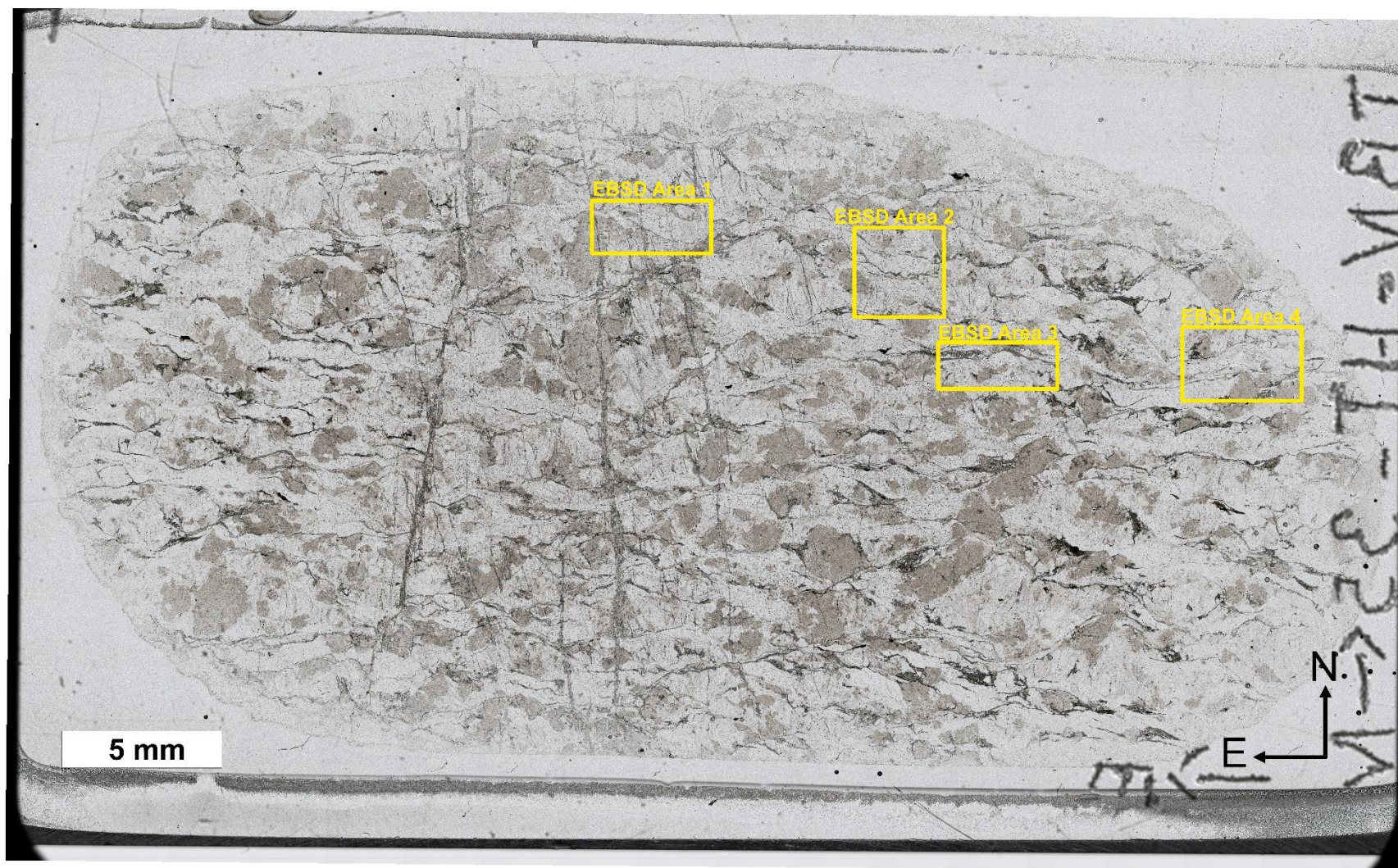
Sample TBN-HT-32:



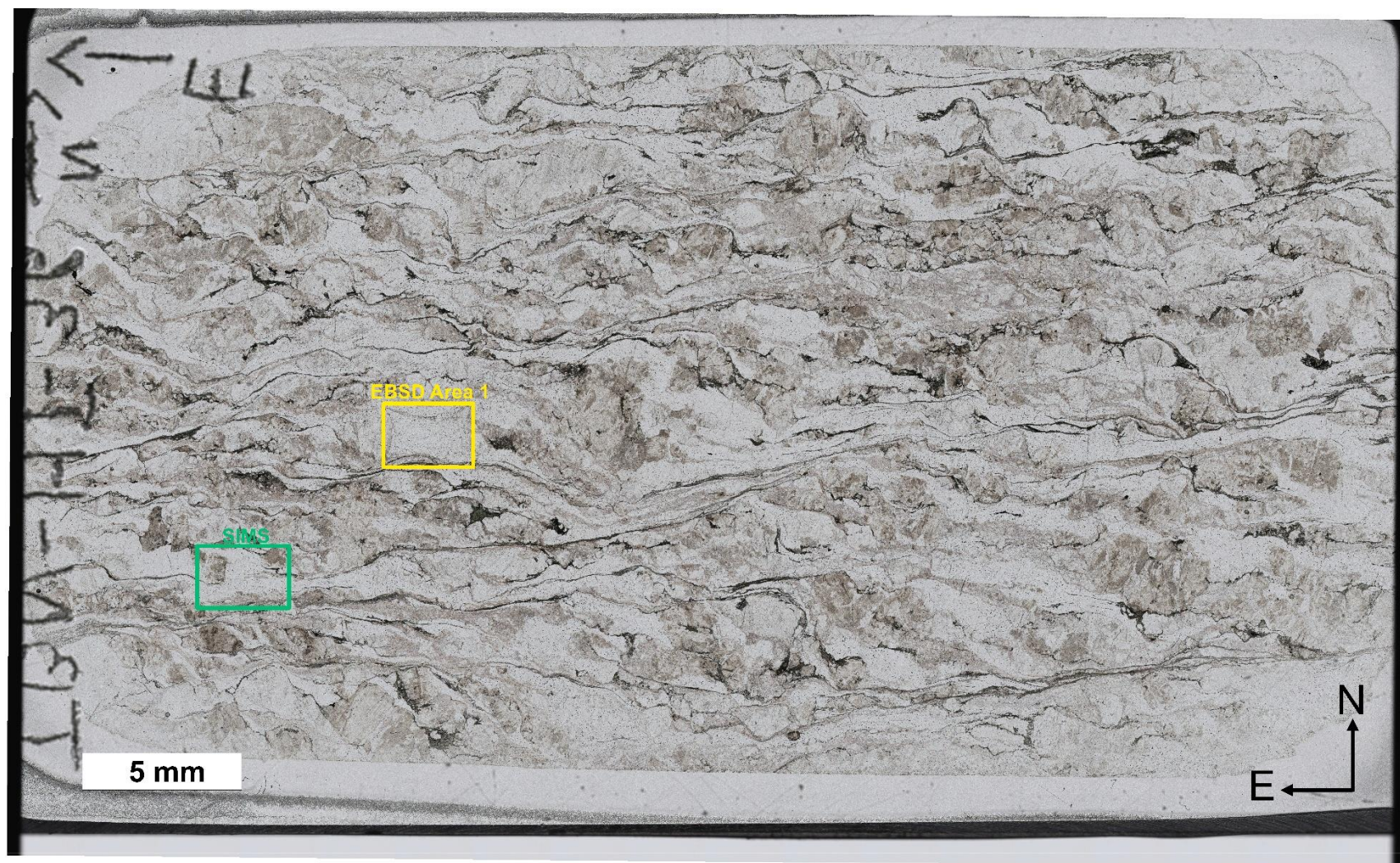
Sample TBN-HT-33:



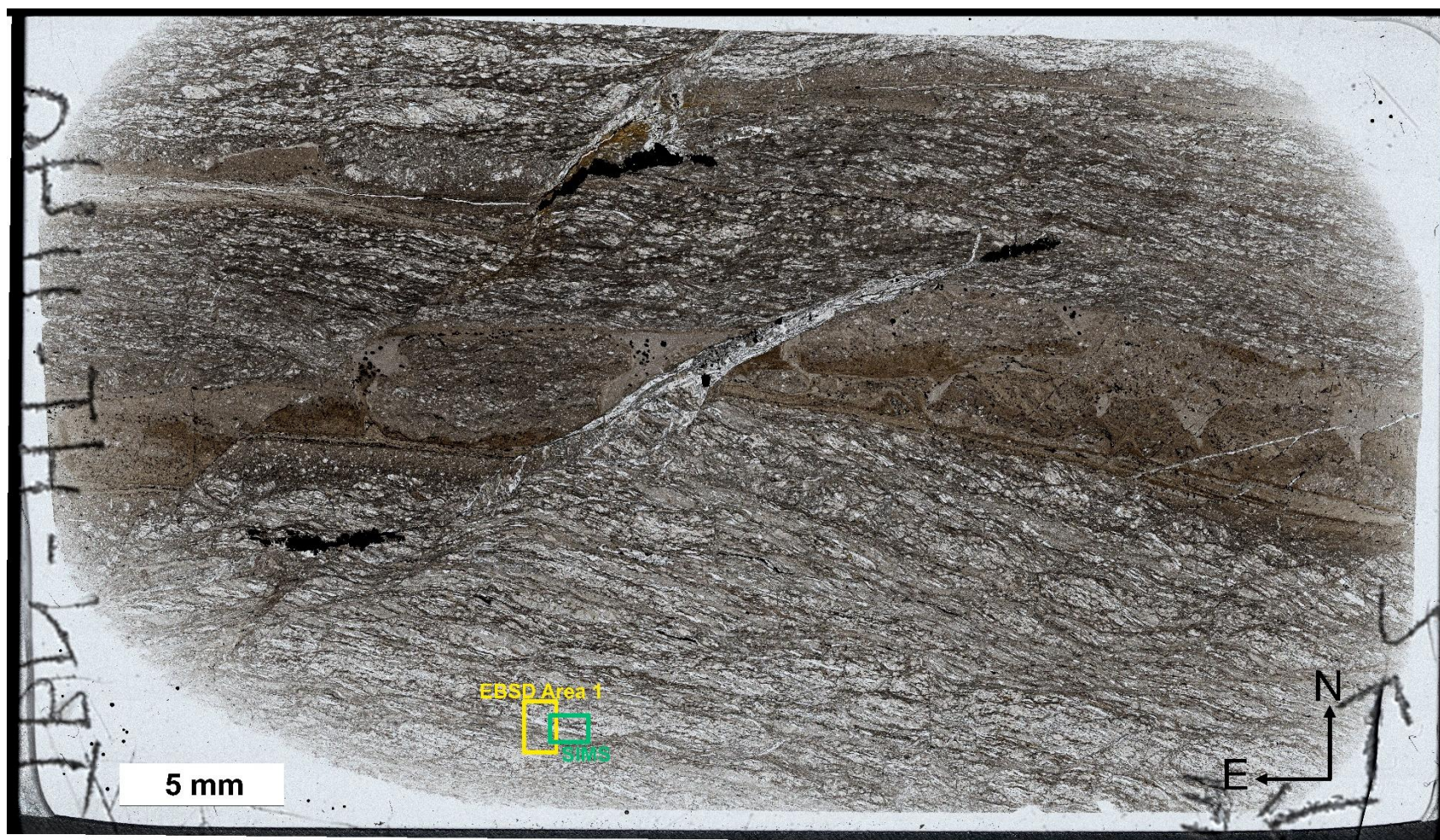
Sample TBN-HT-35:



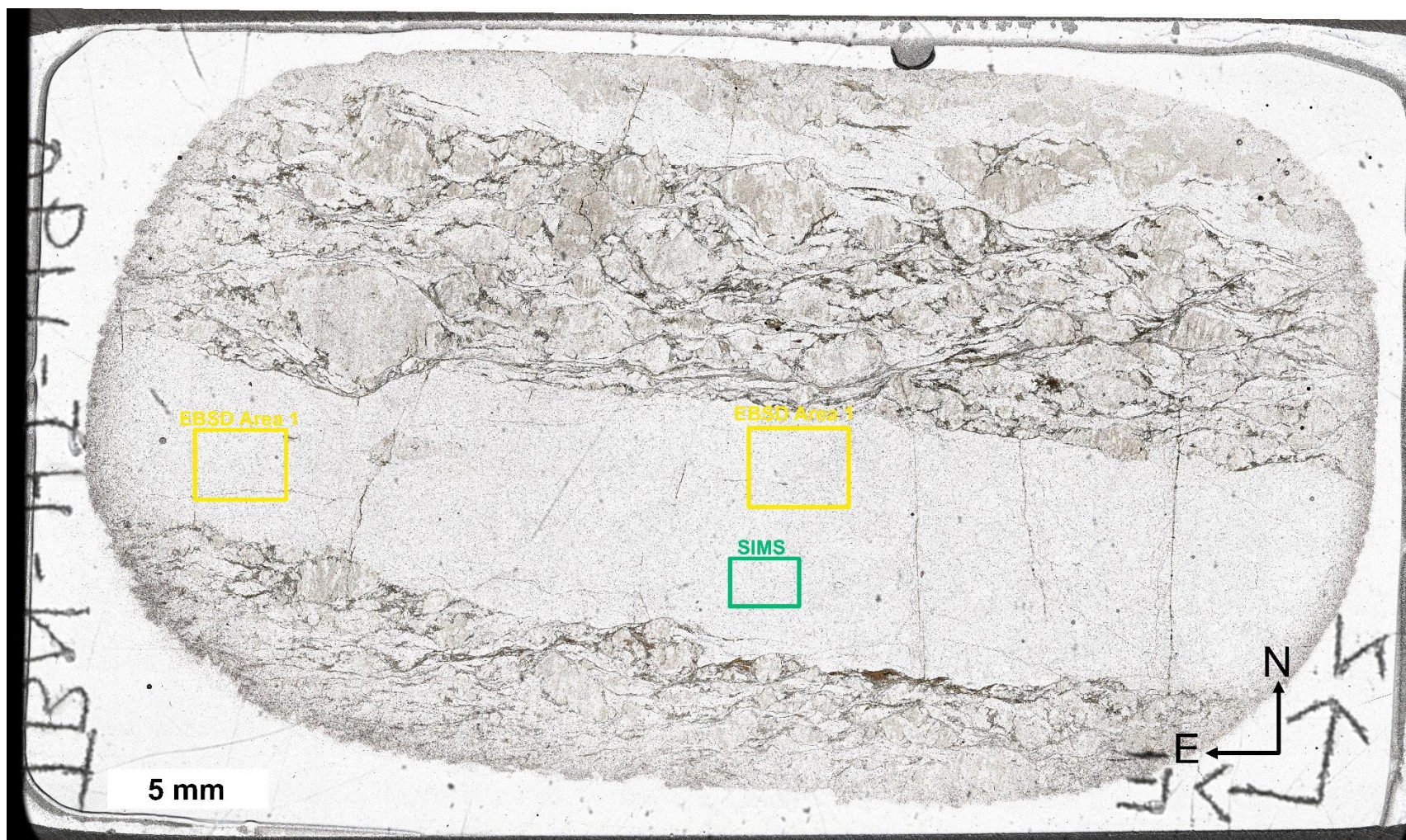
Sample TBN-HT-36:



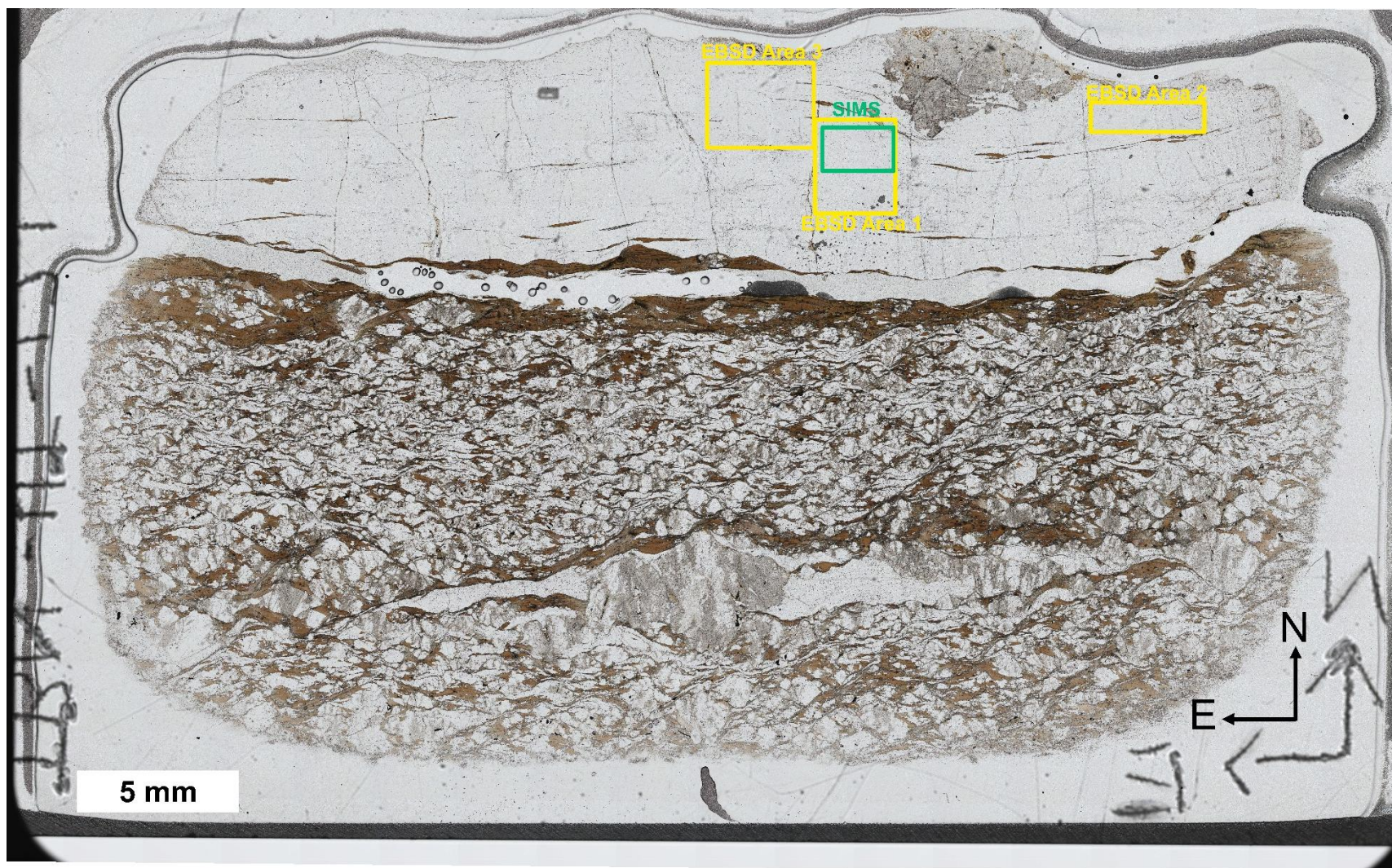
Sample TBN-HT-114a:



Sample TBN-HT-119a:

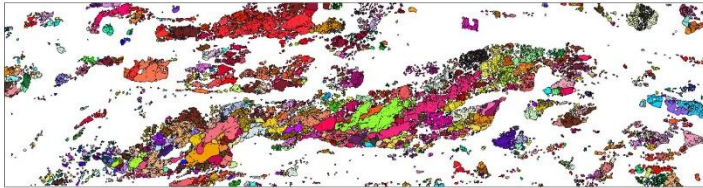


Sample TBN-HT-121:

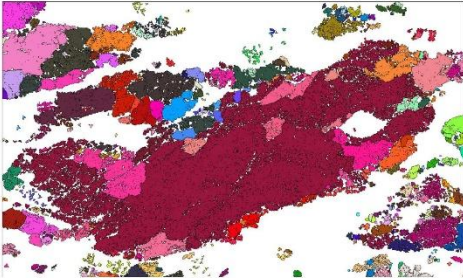


C2: Inverse pole figure (IPF) maps of the analyzed EBSD areas:

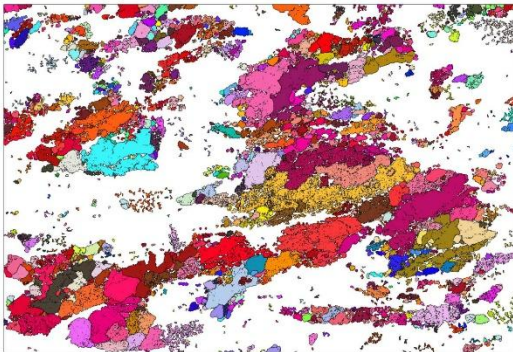
TBN-HT-31_Area1



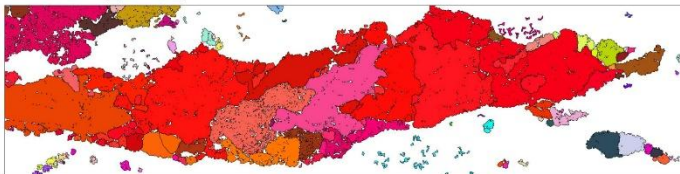
TBN-HT-31_Area2



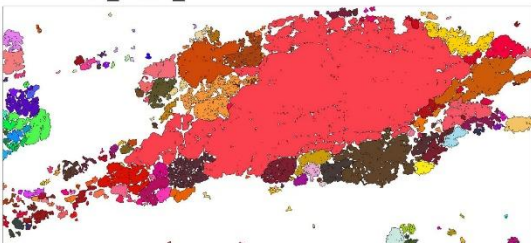
TBN-HT-31_Area3



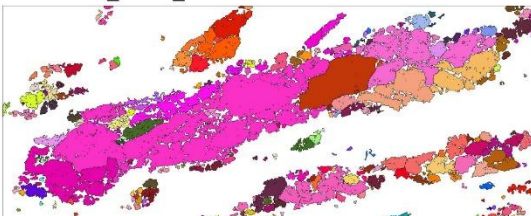
TBN-HT-31_Area4_1



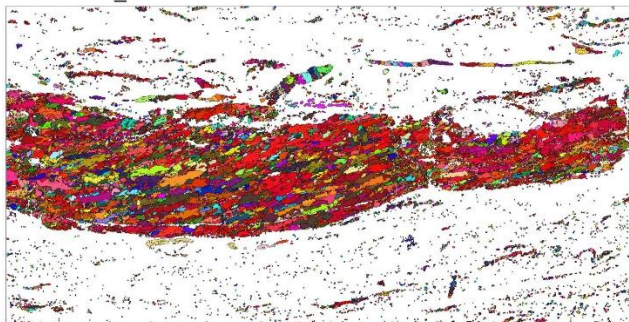
TBN-HT-31_Area4_2



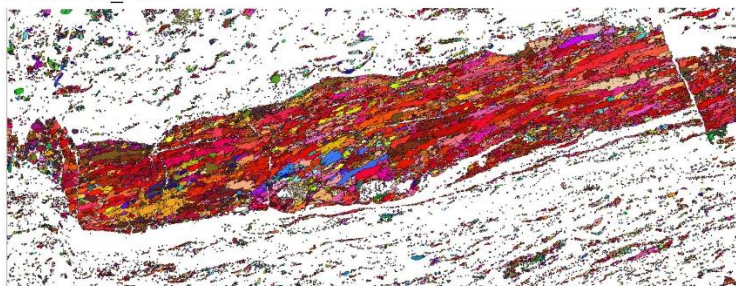
TBN-HT-31_Area4_3



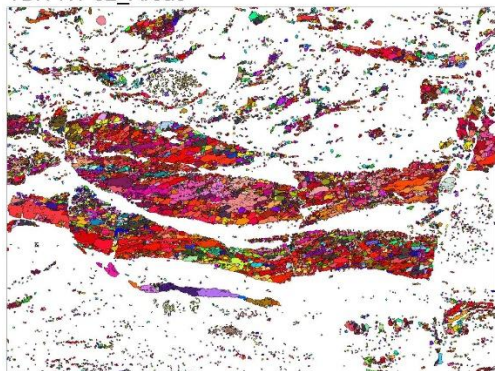
TBN-HT-32_Area1



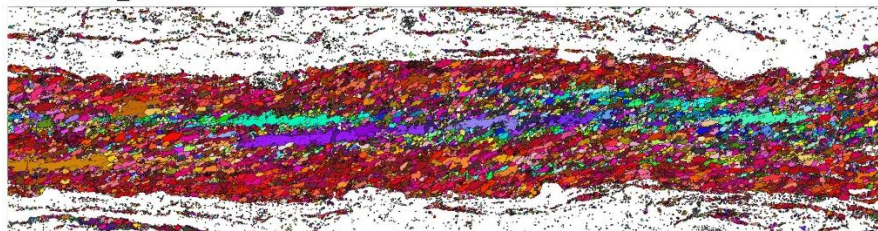
TBN-HT-32_Area2



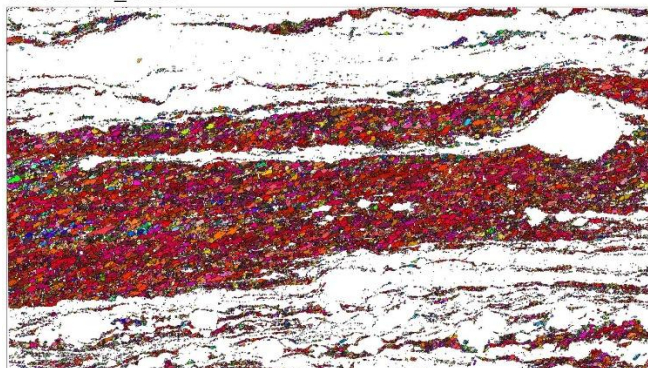
TBN-HT-32_Area3



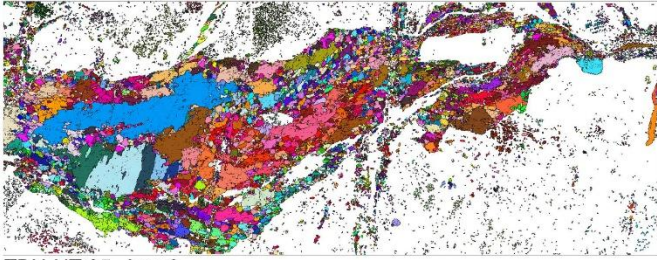
TBN-HT-33_Area1



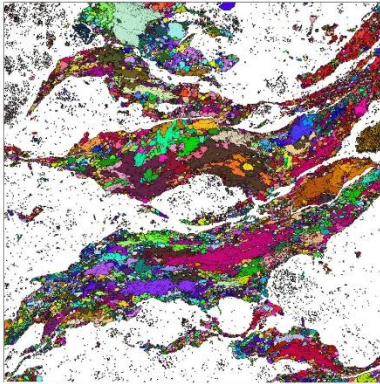
TBN-HT-33_Area2



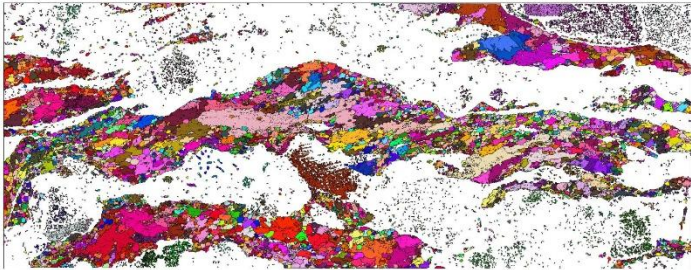
TBN-HT-35_Area1



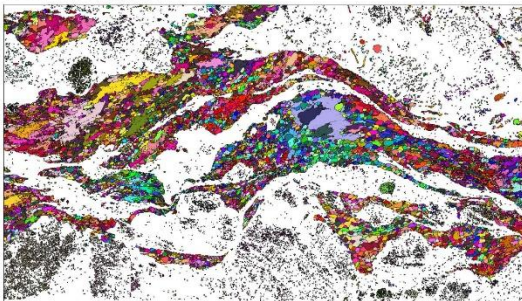
TBN-HT-35_Area2



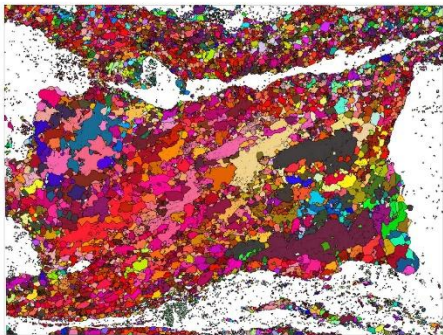
TBN-HT-35_Area3



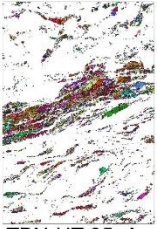
TBN-HT-35_Area4



TBN-HT-36_Area1



TBN-HT-35_Area1



TBN-HT-35_Area1



TBN-HT-35_Area1



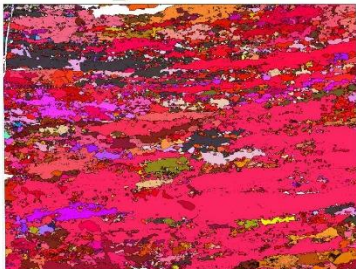
TBN-HT-35_Area1



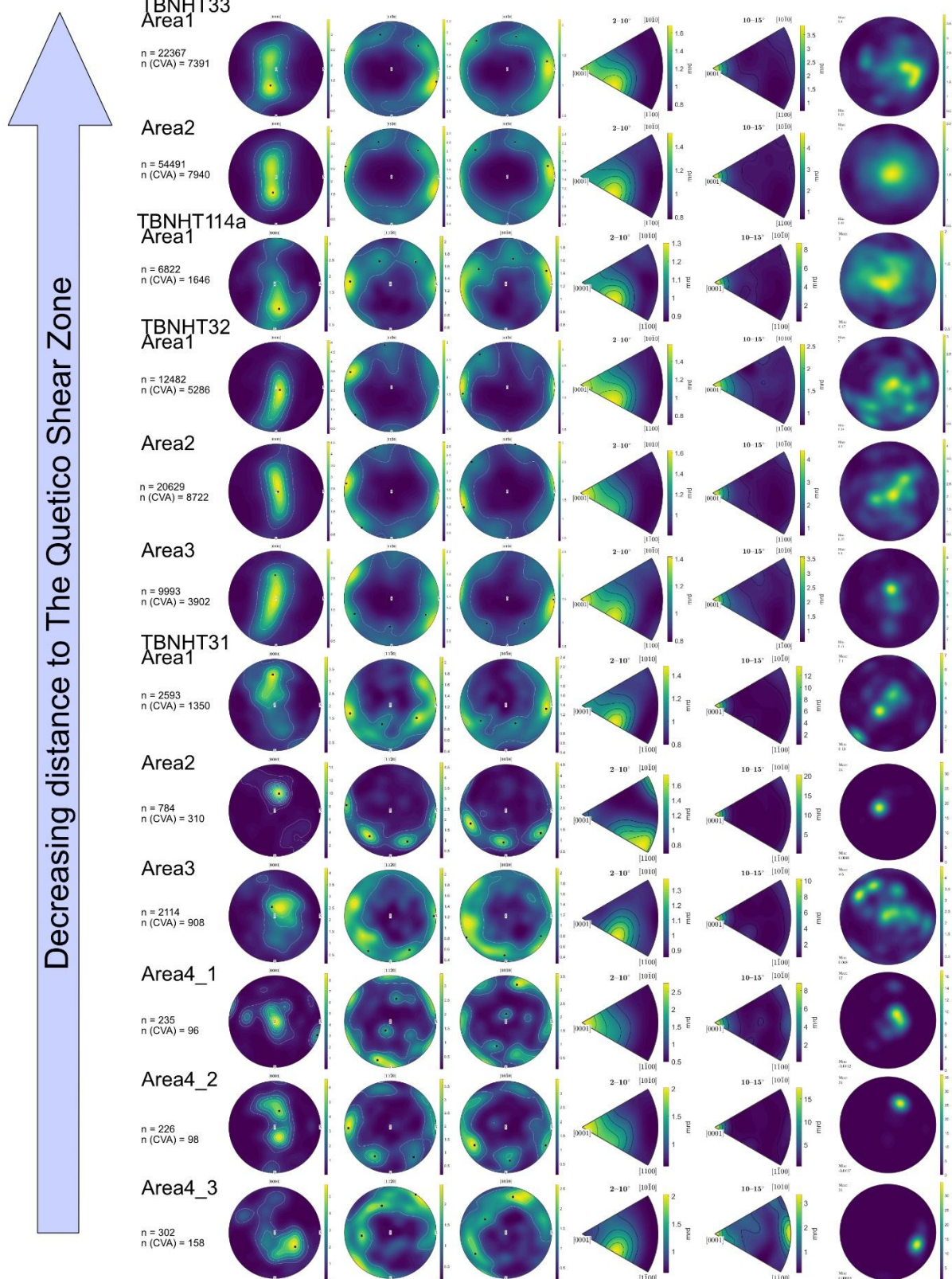
TBN-HT-35_Area1



TBN-HT-35_Area1



C3: Micro-orientation figures produced for all samples (In order of left to right the plots for each sample are the ODF for c, a, and m axes, the MDF for subgrain and grain boundaries and the CVA):



TBNHT119a Area1

n = 16452
n (CVA) = 8057

Area2

n = 45726
n (CVA) = 20802

TBNHT35 Area1

n = 12941
n (CVA) = 4780

Area2

n = 16646
n (CVA) = 5801

Area3

n = 13843
n (CVA) = 3648

Area4

n = 23497
n (CVA) = 5822

TBNHT36 Area1

n = 12080
n (CVA) = 4151

TBNHT121 Area1

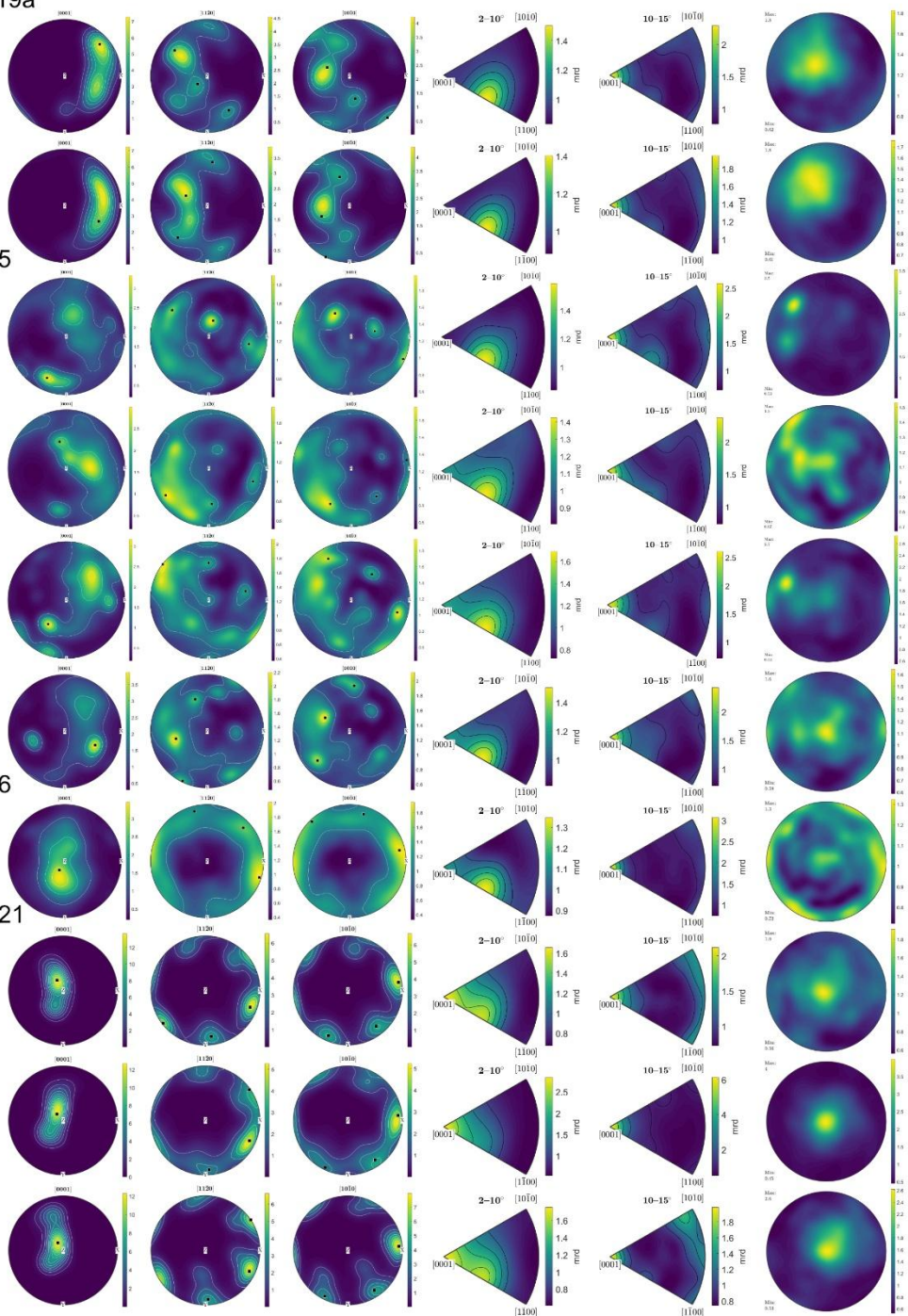
n = 7363
n (CVA) = 2727

Area2

n = 6027
n (CVA) = 2448

Area3

n = 10168
n (CVA) = 3625



Decreasing distance to The Quetico Shear Zone

Appendix D: Flow Law Data

D1: Table displaying the obtained grain size and Ti-in-Qtz values and calculated stress, temperature and water fugacity values (at 450 MPa confining pressures):

Sample	Distance (m)	Grain Size (μm)	Stress (MPa)	Ti-in-Qtz (ppm)	Temperature (K)	Water fugacity (MPa)
TBN-HT-31	195	17.11	79.13	3.96	739.49	236.09
TBN-HT-32	110	10.00	115.86	7.42	781.34	282.58
TBN-HT-33	45	16.80	80.17		739.37	
TBN-HT-35	240	16.84	80.03			
TBN-HT-36	380	20.95	68.55	3.95	739.37	235.96
TBN-HT-114a	100	15.50	84.88	3.45	730.84	226.53
TBN-HT-119a	40	18.30	75.45	12.92	862.69	327.76
TBN-HT-121	620	19.03	73.38	19.99	857.15	364.87

D2: Table displaying the three flow laws with their respective constant parameters and the calculated strain rates for the samples:

Flow Law	Constants			Sample	Strain rate
Hirth et al. 2001	Material parameter	A (Mpa ^{-n/s})	6.31E-12	TBN-HT-31	1.70E-11
$\dot{\epsilon} = Af_{H_2O}^m \sigma^n \exp(-\frac{Q}{RT})$	Water fugacity exponent	m	1	TBN-HT-32	3.03E-10
	Stress exponent	n	4	TBN-HT-36	9.54E-12
	Activation energy	Q (J/mol)	135000	TBN-HT-114a	1.67E-11
	Ideal gas constant	R (J/mol)	8.31446	TBN-HT-119a	1.79E-10
				TBN-HT-121	3.96E-10
Tokle et al. 2018	Material parameter	A	5.01E-10	TBN-HT-31	2.40E-11
$\dot{\epsilon} = Af_{H_2O}^r \sigma^n \exp(-\frac{Q}{RT})$	Water fugacity exponent	r	1	TBN-HT-32	3.92E-10
	Stress exponent	n	4	TBN-HT-36	1.35E-11
	Activation energy	Q (J/mol)	118000	TBN-HT-114a	2.40E-11
	Ideal gas constant	R (J/mol)	8.31446	TBN-HT-119a	2.14E-10
				TBN-HT-121	4.47E-10
Lusk et al. 2021	Material parameter	A (Mpa ^{-n·s⁻¹})	1.75E-12	TBN-HT-31	1.80E-10
$\dot{\epsilon} = A\sigma^n d^m f_{H_2O}^r e^{(-\frac{Q+VP}{RT})}$	Water fugacity exponent	r	0.49	TBN-HT-32	2.06E-09
	Stress exponent	n	3.5	TBN-HT-36	1.08E-10
	Activation energy	Q (J/mol)	125000	TBN-HT-114a	1.80E-10
	Ideal gas constant	R (J/mol)	8.31446	TBN-HT-119a	1.21E-09
	Grain size exponent	m	0	TBN-HT-121	2.32E-09
	Activation volume	V (cm ³ mol ⁻¹)	2.59		
	Confining pressure	P (Mpa)	450		

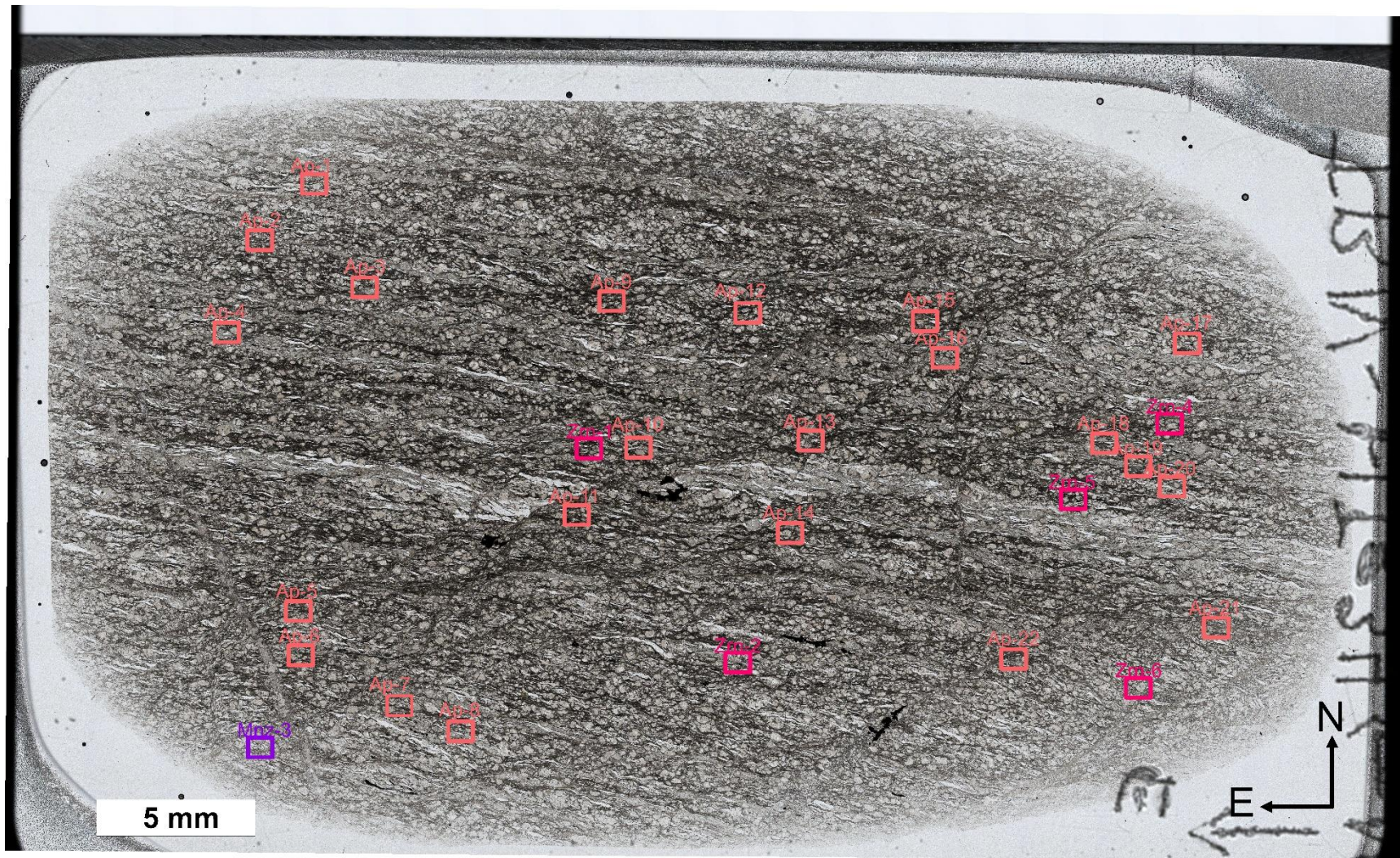
Appendix E: LASS-ICP-MS Data

E1: Table of references for LASS-ICP-MS standards used for each mineral type:

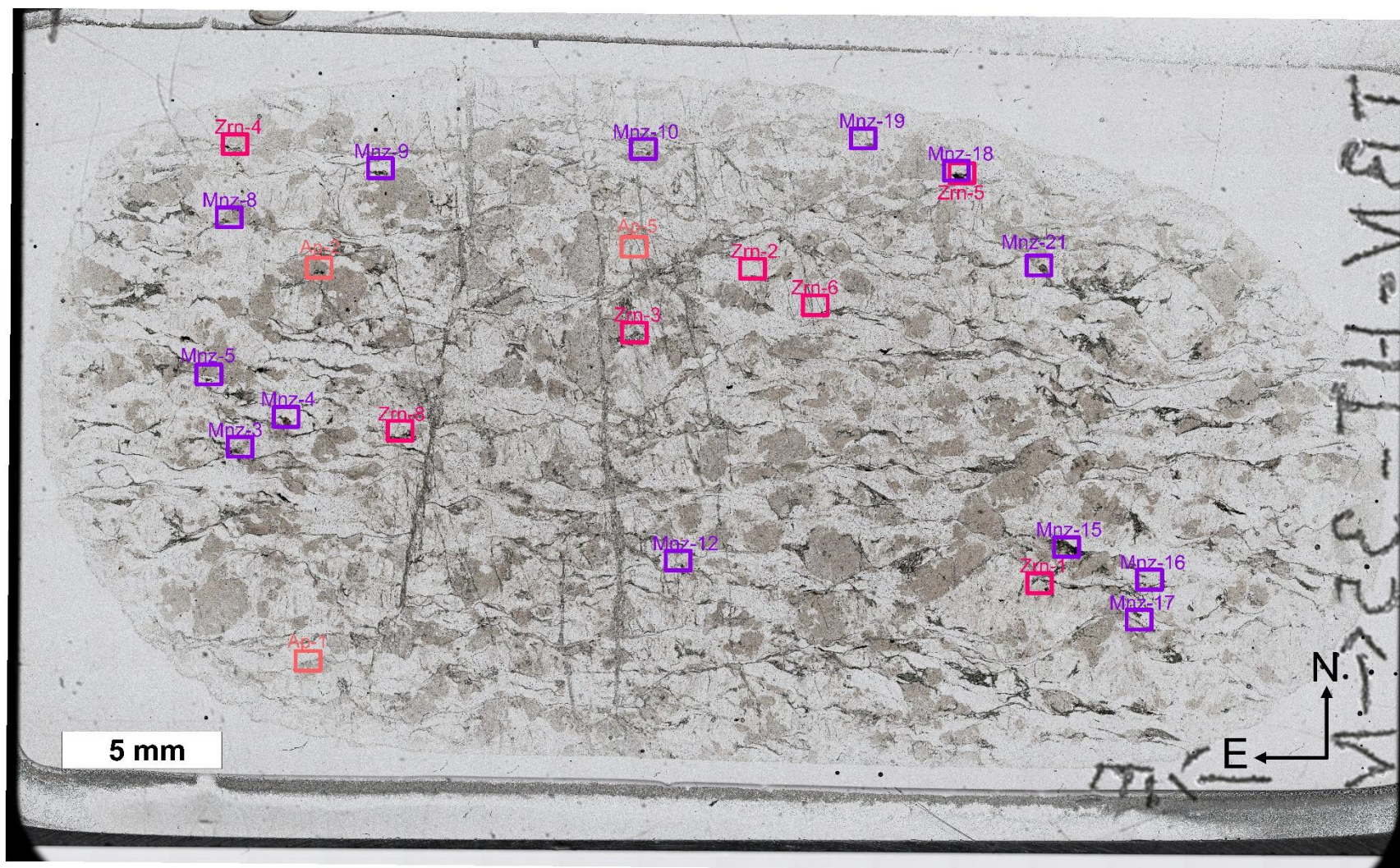
Mineral	Primary Standard	Secondary Standard	Reference(s)
Apatite	MAD-UCSB		Thomson et al., 2012
		Durango	McDowell et al., 2005
		MRC-1	Apen et al., 2022
		McClure	Schoene and Bowring, 2006
Monazite	Stern 44069		Stern and Sanborn, 1998
			Aleinikoff et al., 2006
		Trebilcock	Tomascak et al., 1996
		FC	Horstwood et al., 2003
		<u>MAN</u>	Horstwood et al., 2003
Titanite	MKED BLR		Spander et al., 2016
			Aleinikoff et al., 2006
		Y17	Spencer et al., 2013
		P57	Spencer et al., 2012
Zircon	91500 Plešovice		Wiedenbeck et al., 1995
			Slama et al., 2008
		OG-1	Stern et al., 2009
		GJ-1	Jackson et al., 2004

E2: Full thin section scan of samples analyzed with LASS with locations of analyzed grains highlighted by mineral type:

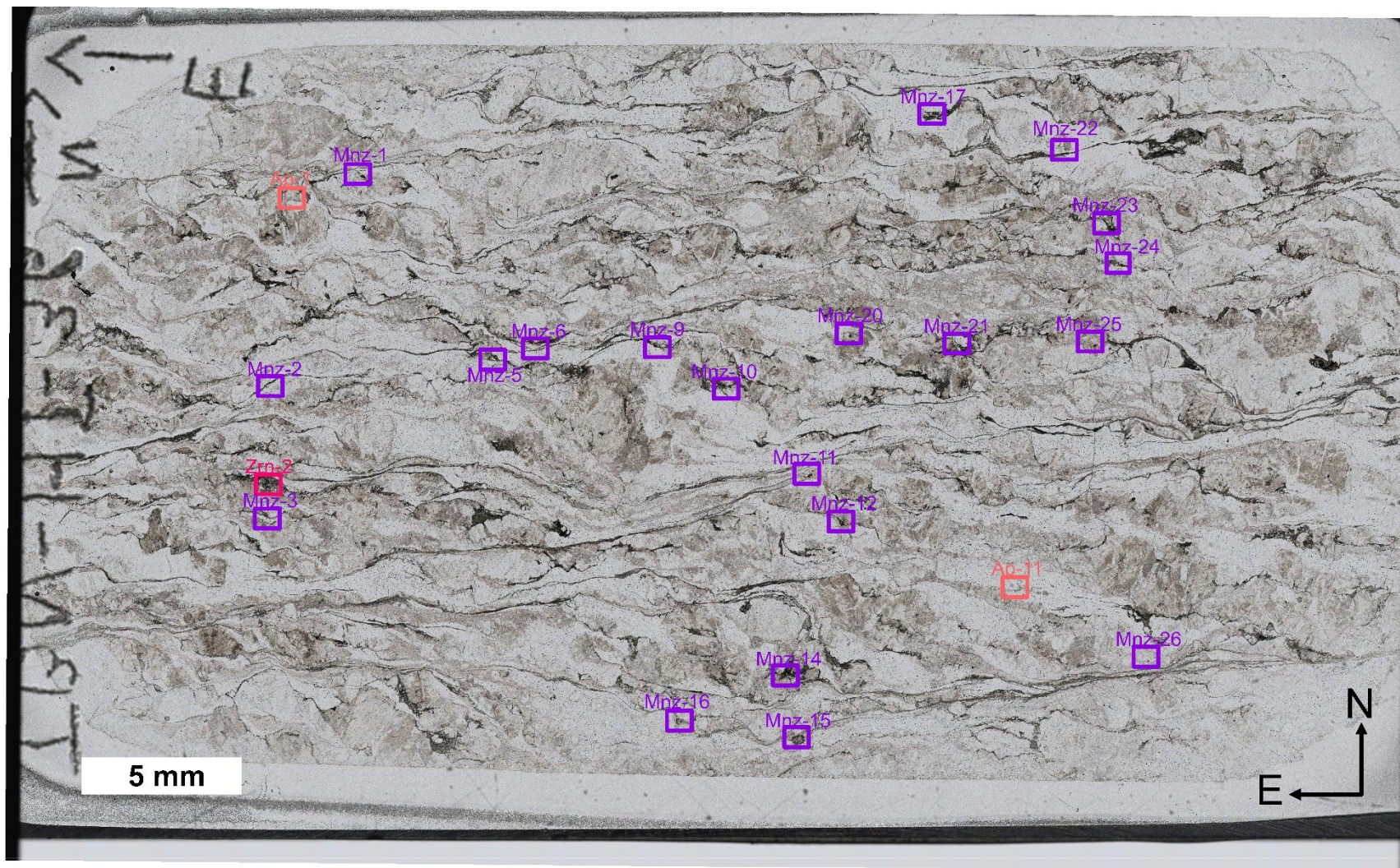
Sample TBN-HT-34 with the locations of analyzed apatite, monazite, and zircon grains:



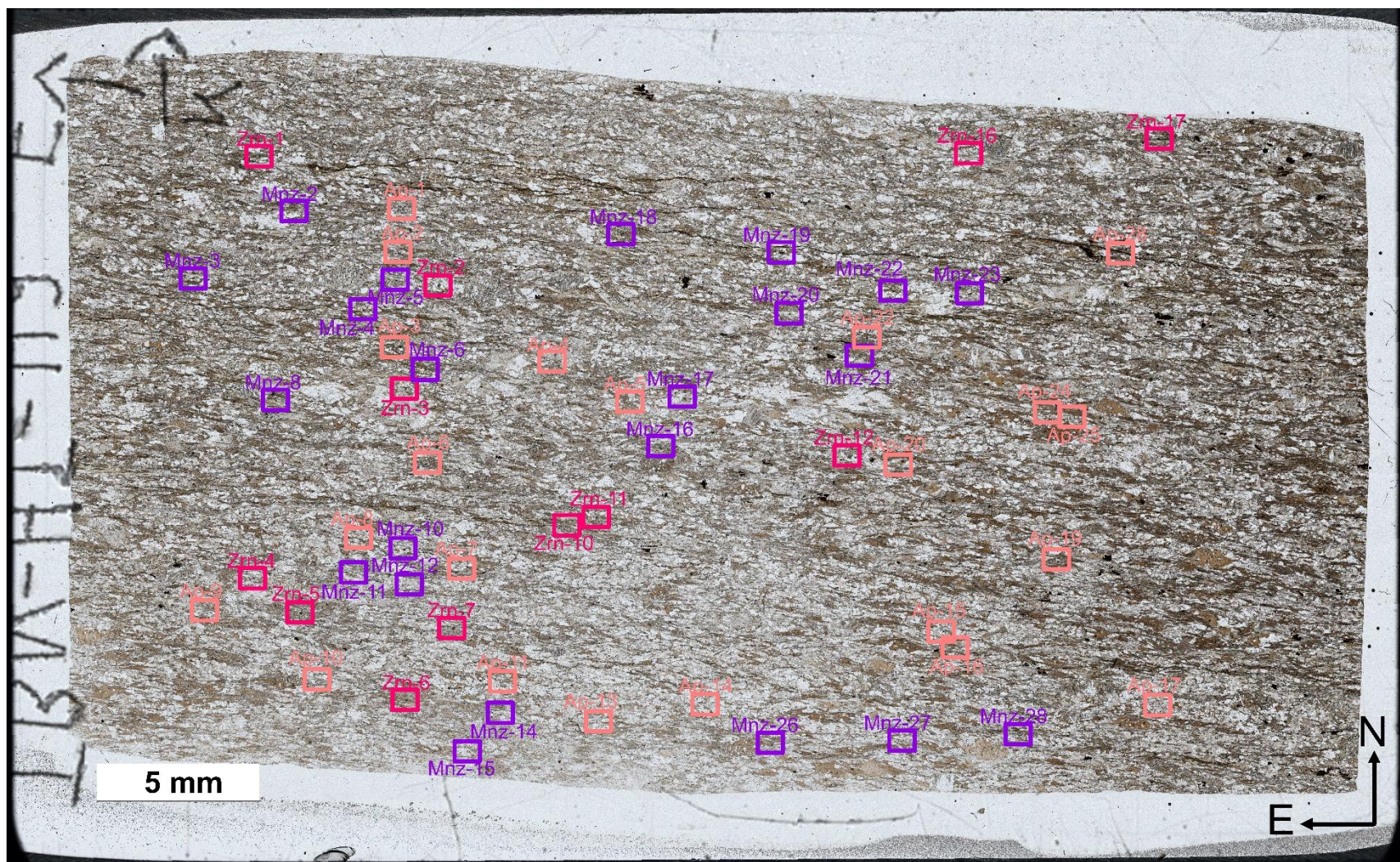
Sample TBN-HT-35 with the locations of analyzed apatite, monazite, and zircon grains:



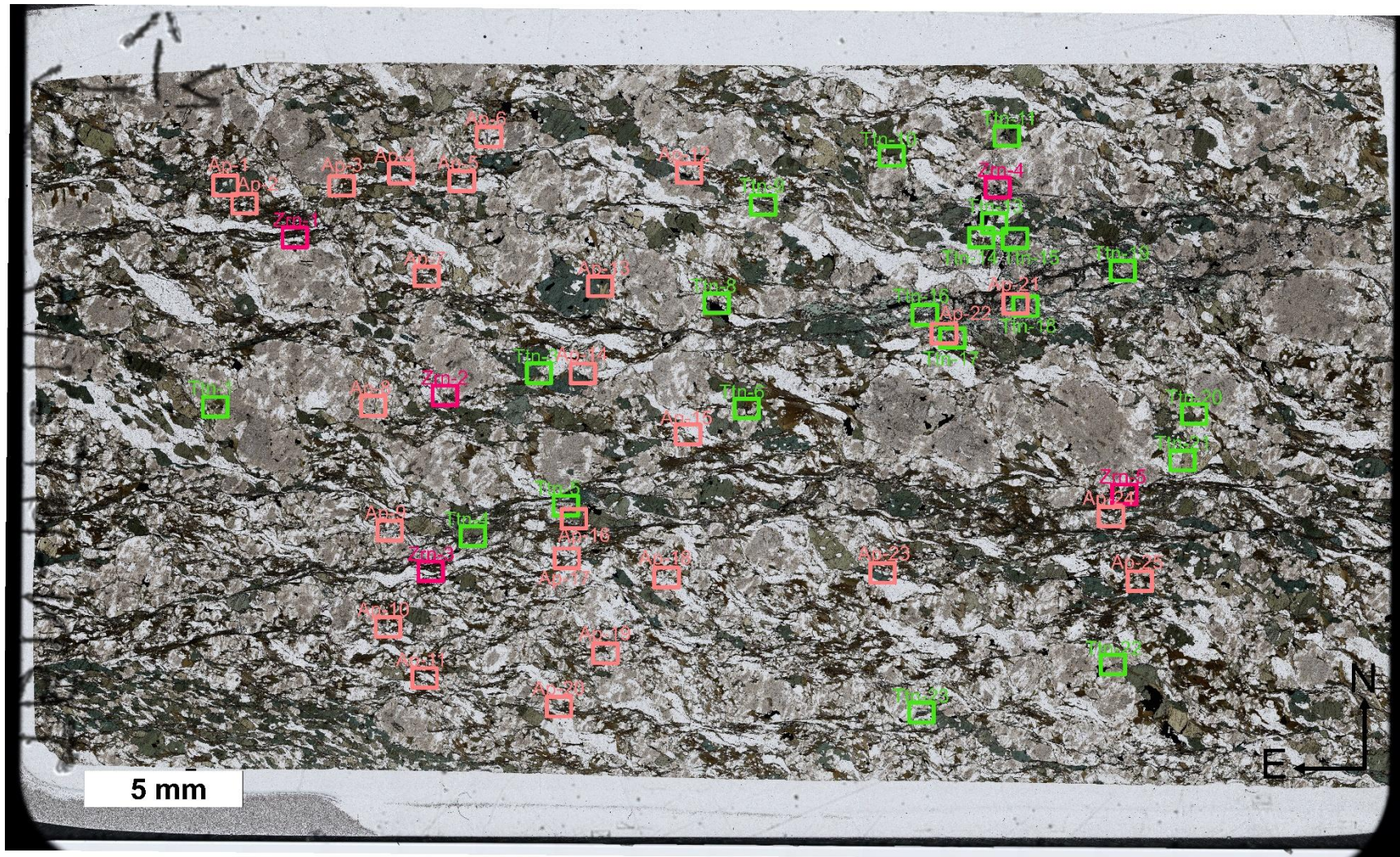
Sample TBN-HT-36 with the locations of analyzed apatite, monazite, and zircon grains:



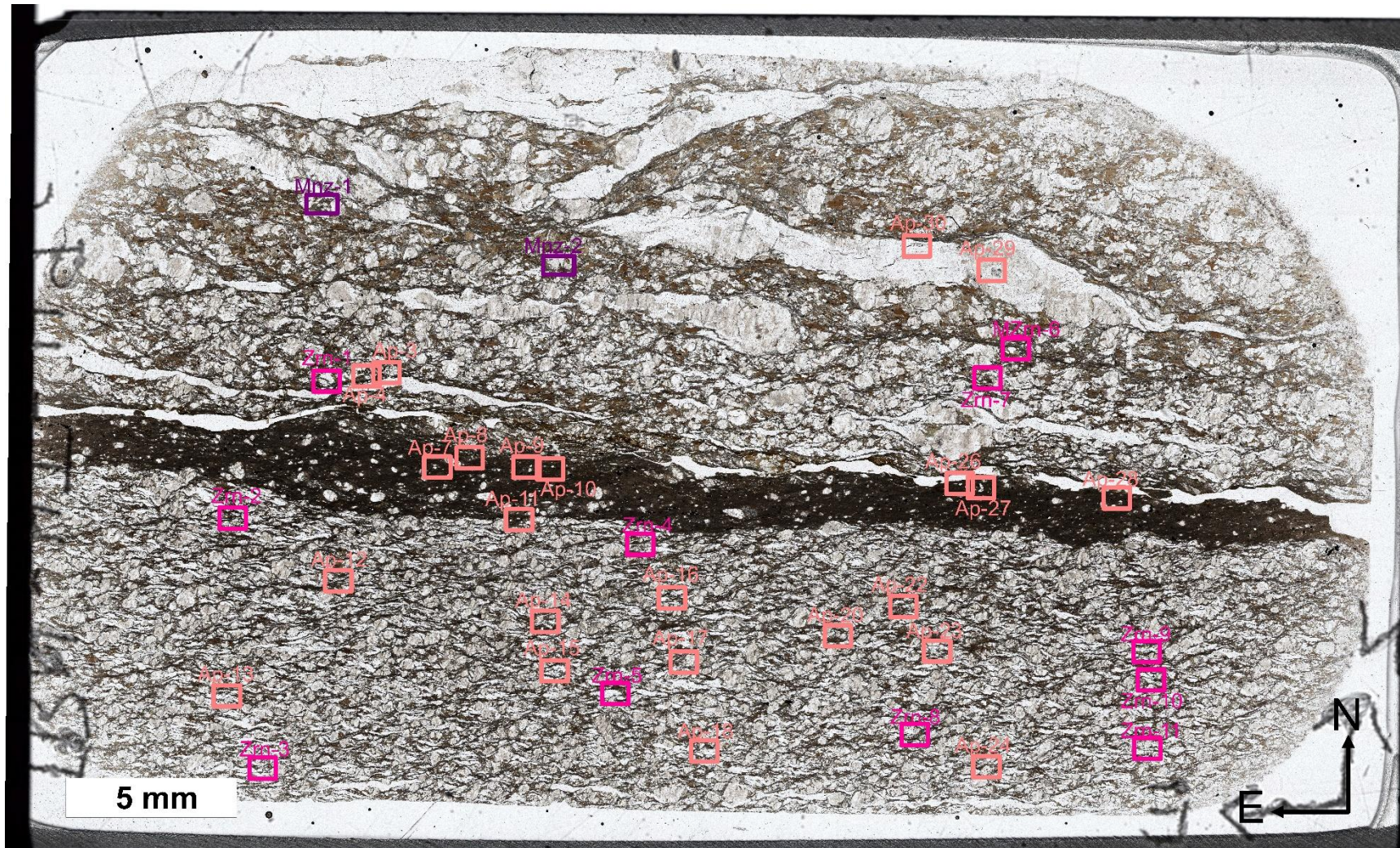
Sample TBN-HT-112 with the locations of analyzed apatite, monazite, and zircon grains:



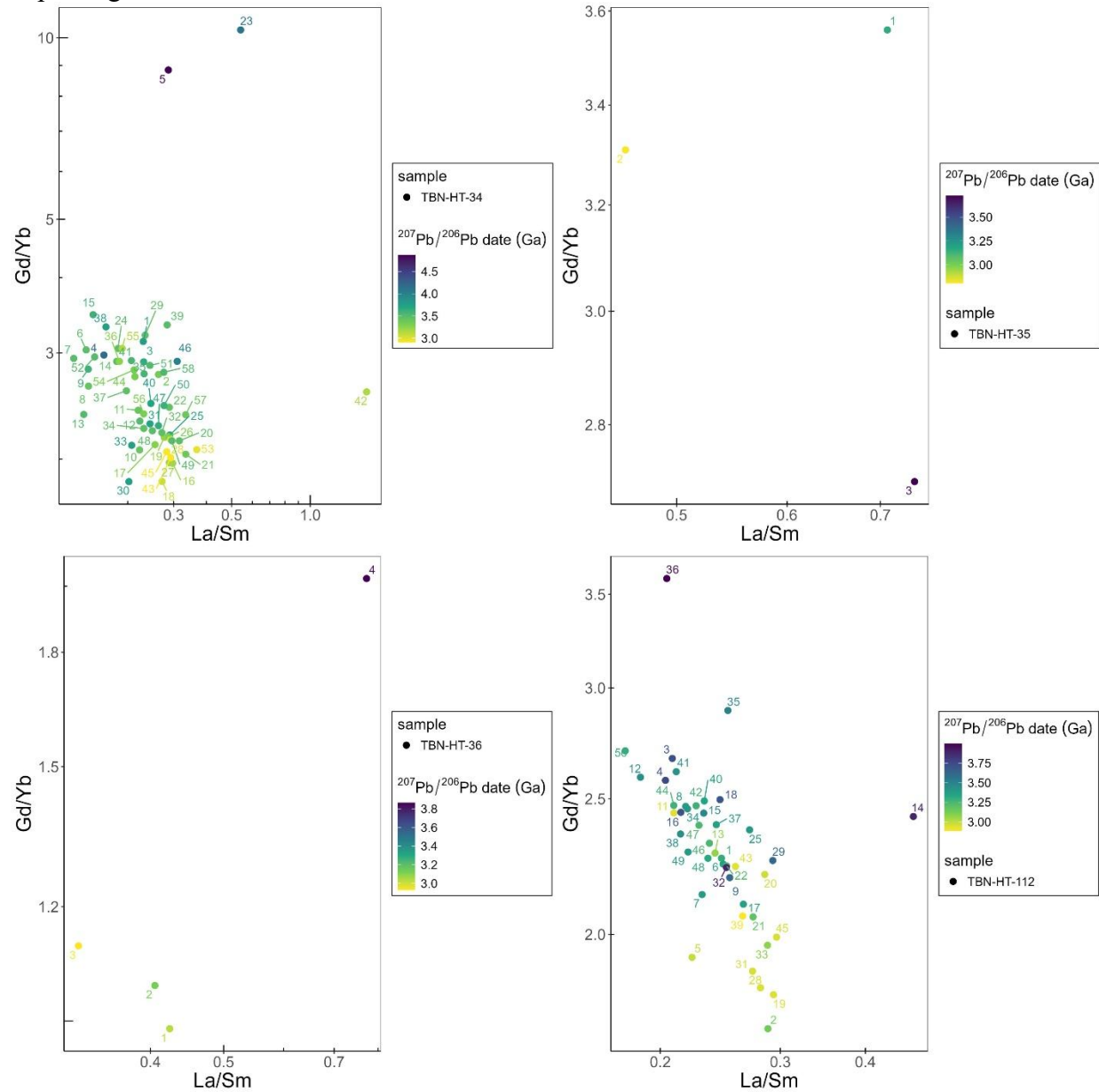
Sample TBN-HT-117 with the locations of analyzed apatite, monazite, titanite, and zircon grains:

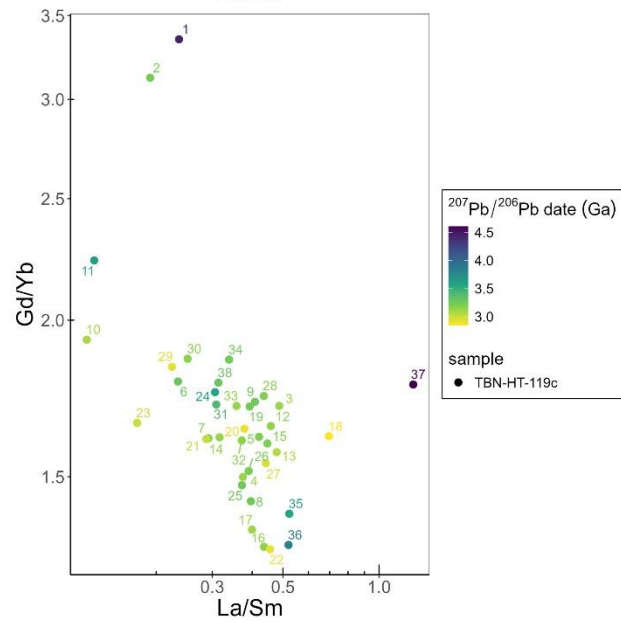
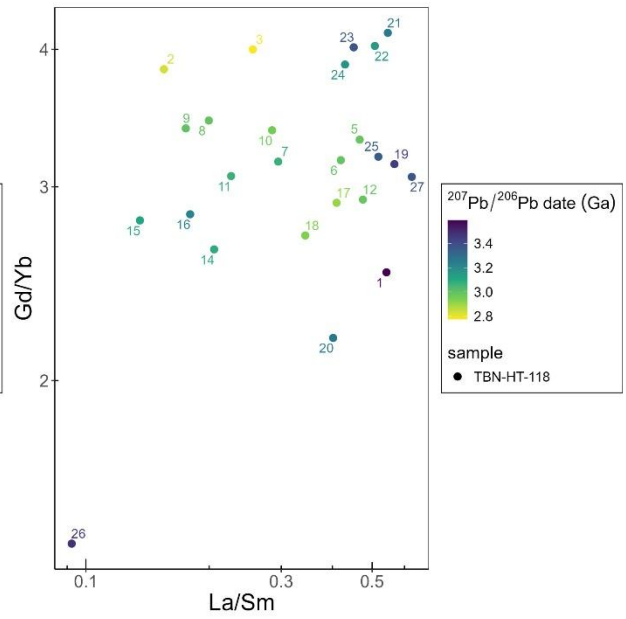
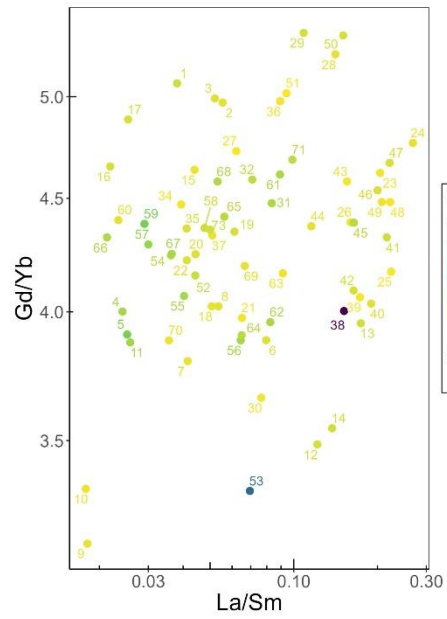


Sample TBN-HT-119c with the locations of analyzed apatite, monazite, and zircon grains:

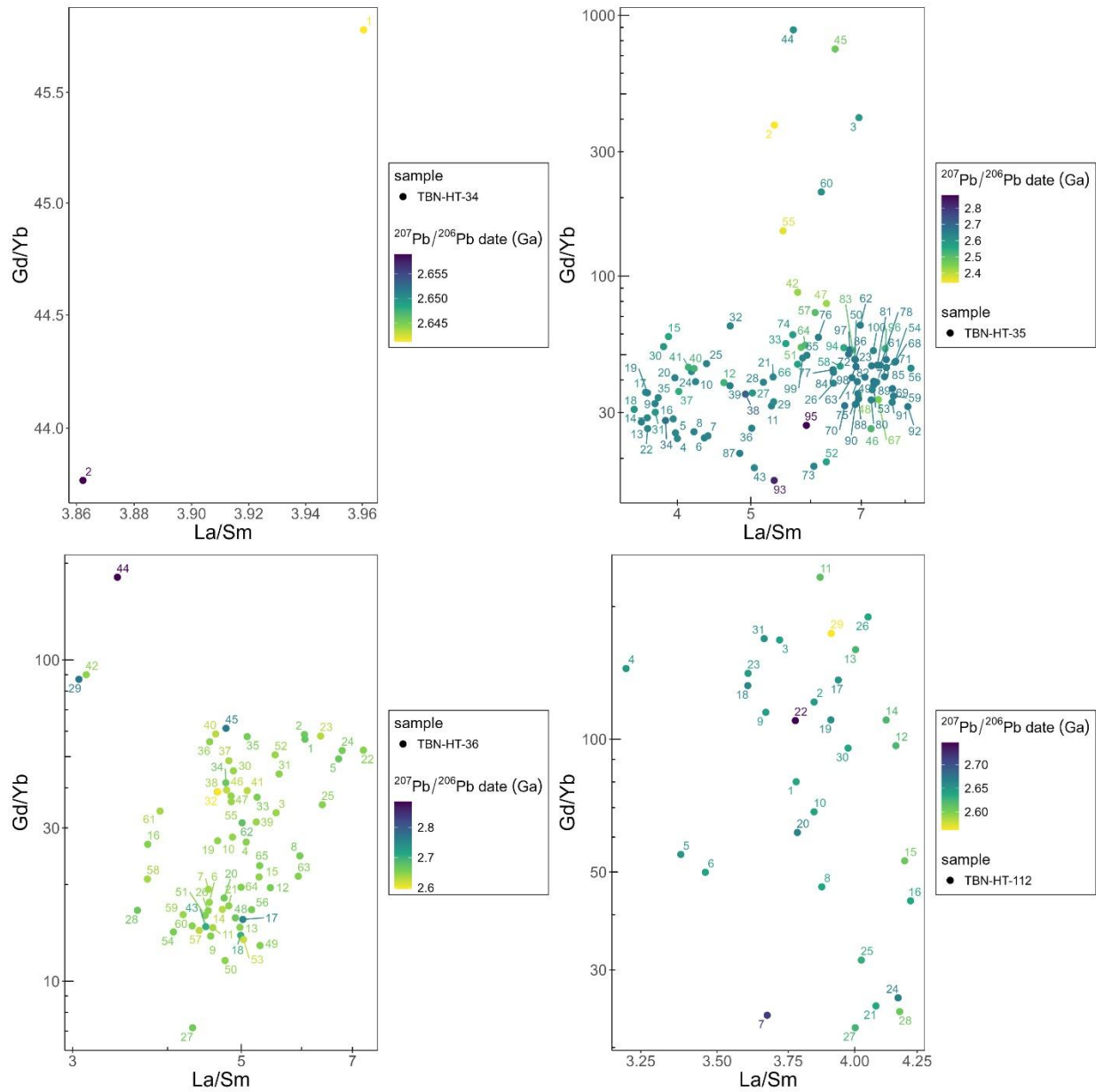


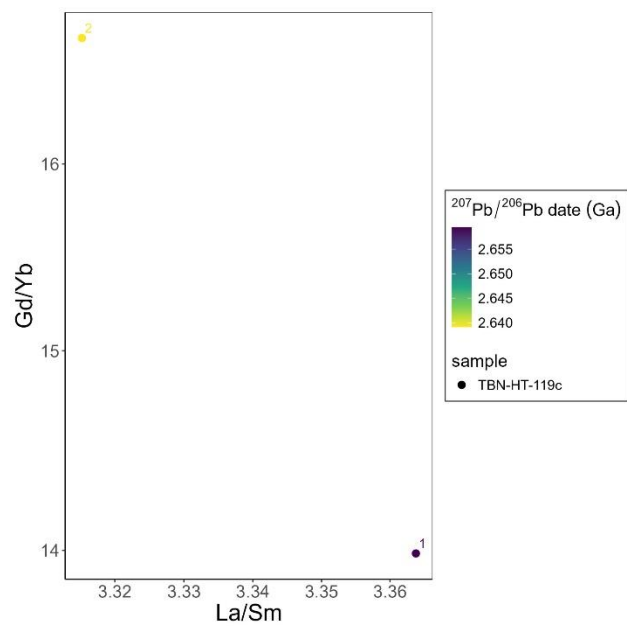
E3: Trace element La/Sm_(n) vs. Gd/Y_(n) plots for all samples analyzed by mineral type:
Apatite grains:





Monazite grains:





Titanite (bottom right plot) and zircon grains:

