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Rheological Bridge Zones: A New Perspective to Explore Strain Localization and Rock Rheological Weakening

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RHEOLOGICAL BRIDGE ZONES: A NEW PERSPECTIVE TO EXPLORE STRAIN LOCALIZATION AND ROCK RHEOLOGICAL WEAKENING

By

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Strain localization, a ubiquitous geological phenomenon, emerges when rock undergoes extensive weakening and accommodates disproportionate deformation. This phenomenon spans from brittle faulting to viscous flow, known as shear zones. Despite its prevalence, discussions on weakening primarily center on highly strained samples, rather than the initial deformation state. This research delves into the evolution of localization and rheology in intact rocks.

Heterogeneous mechanical properties within the lithosphere vary based on mineralogy, microstructure, and environmental conditions. Chapter 2 characterizes microscale structures termed "bridge zones." These zones, observed through optical and electron microscopy, exhibit distinctive morphologies. Comprising fine-grained aggregates, bridge zones link weak phases within shear zones and even less deformed rock margins. They result from grain size reduction, mass transfer, and reactions, ultimately weakening the rock and impacting its rheology.

Chapter 3 investigates stress field controls through finite element modeling. Spatial
distribution profoundly influences stress magnitude and gradient. Stress behavior isn't linear with applied stress; weak material configuration and orientation significantly impacts stress concentration. Notably, stress concentration areas often connect weak domains, resembling bridge zones.

Chapter 4 focuses on shallow, low-temperature deformation in natural samples. These rocks, spanning a ductile shear zone gradient, exhibit extensive bridge zones formed associated with similar mechanical and chemical processes as in Chapter 2. Numerical modeling aligns high-stress amplification areas with bridge zones, corroborating their influence on rock strength reduction. We describe a conceptual model relating far-field loading through microscale change to bridge zone development.

Bridge zones present an avenue to comprehend the inception and progression of deformation localization, crucial for plate tectonics, metamorphism, landscape evolution, seismic activity, and other lithospheric processes. This research identifies microscale mechanisms driving weak domain development, advancing the understanding of rheological change. Ultimately, these findings lay the foundation for predictive algorithms pertaining to strength evolution across the lithosphere.
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In culmination, this thesis represents the convergence of innumerable hours of toil, resolute dedication, and collaborative synergy. While the enumeration provided here is necessarily selective, the profound impact these individuals have exerted on my intellectual and personal growth is immeasurable. I extend my heartfelt gratitude to all who have played a pivotal role in shaping this transformative expedition.
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CHAPTER 1
INTRODUCTION

The study of continental lithospheric rheology plays an important role in the development of tectonics and structural geology, and is a significant factor in Earth’s internal deformation and the dynamic evolution of plate tectonics. The theory of plate tectonics reasonably explains the global tectonic framework and has become a widely accepted theory in the geological world. However, with the deepening of studies, traditional plate tectonics has encountered some new challenges in explaining the evolution of continental lithosphere: for example, the vertical stratification of continental lithosphere is not only the variation of composition, but also the difference of rheology and its role in the evolution of continental lithosphere (Rosenberg 2001; Rey et al. 2009). Strain localization, as the most common structural geological and rheological structure in continental lithosphere, is widely distributed in the whole lithosphere, in both the brittle and viscous regimes, and runs through all geological scales (e.g., Sibson 1977; Ramsay 1980; Escartin et al. 1997; Jolivet et al. 2004; Précigout & Gueydan 2009; Gueydan et al. 2014). However, although we know the general factors that influence the mechanical properties of rocks (e.g., Kirby 1985; Handy et al. 2007; Burgmann & Dresen 2008; Karato 2008; Marti et al. 2018), the question of what the physical parameters drive rheological change/weakening is still unanswered.

Arguably the dominant agent of rheological change in viscous regime is the development of new weak domains by such processes as recrystallization, fracture, neocrystallization and reactions. Interconnection of weak phases has long been recognized as a major contributor to strain localization and weakening (e.g., Cobbold 1977; Jordan 1987; Handy 1990; Johnson et al. 2004; Holyoke & Tullis 2006; Montesi 2013; Gerbi et al. 2016), yet the generation of weak phases/domains has received little attention. Work in recent years (e.g., Johnson et al. 2004, 2009, 2022; Svahnberg & Piazolo 2010, 2013; Gerbi et al. 2014, 2016; Gardner et al. 2017) reveals not only our poor understanding of the causes of weak domain development, but more importantly that these domains have an enormous potential of influence rock strength, rendering weak domain initiation a cornerstone of rheological change. In the viscous regime,
exactly when and how new weak domains develop and link with preexisting weak domains remains a sizable knowledge gap, with a particular impact on geodynamic models, many of which currently do not incorporate geologically suitable algorithms for rheological evolution. In this study, I focus on two general questions:

1) What processes accompany weak domain formation in the viscous regimes?
2) What mechanical environment facilitates the formation of those domains?

1.1. Proposed Potential Mechanisms for Local Rheological Weakening

Rocks weaken due to thermal-mechanical instability, which results in more strain accumulation than surrounding rocks (Vitale et al. 2010). In general, rocks in shear zones are usually mixed fine grain two-phase or multi-phase materials. Arguably the dominant agent of rheological change in the viscous regime is the development of new weak domains. The bulk strength decreases with break-down (e.g., grain size reduction or chemical reaction for forming weak phases, etc.) of the strong loading-bearing mineral phases in these rocks (Platt 2015). The addition of fluid phases, resulting in the pressure-dissolution and metamorphic reactions, also plays an important role (Alsop et al. 2004; Bhattacharyya & Mitra 2011, 2014).

There are two major kinds of weakening effects that contribute significantly to strain localization: chemical weakening with fluid and reactions (hydrolytic weakening, and reaction weakening, etc.) as well as physical weakening (grain size reduction, and crystallographic preferred orientation, etc.).

Hydrolytic weakening: As water fluids intervene in anhydrous minerals with water molecules, or as rocks are subjected to the influence of the fluid phases, the mechanical strengths of these rocks and minerals is prominently reduced, and dislocations are easier to form, which lead to mineral plastic deformation (Griggs & Blacic 1965; Zhou et al. 2008; Shao et al. 2013). The addition of water fluid greatly promoted the dislocation slip rate, and also accelerated the recovering of the deformed rock and recrystallization on the grain boundary.

Reaction weakening: Stünitz & Tullis (2001) suggested that changes in the strength of the deformed rock may be caused by chemical reactions that alter the mineral composition of the original
rock. Groome et al. (2008) also reported that dehydration / hydration reactions can lead to rock strengthening / weakening. In polyphase rocks, the distribution of new weaker minerals formed through metamorphic reaction tend to have banded characteristics. With the increasing degree of deformation and the continuous development of dynamic recrystallization and chemical reactions, compositional banding will develop in rocks, accompanied by layering development, thus significantly reducing rock strength (Park et al. 2006; Wintsch & Yeh 2013).

Grain size reduction: Rubie et al. (1993) indicated that grain size reduction is the most effective way of rock weakening. Via BLG or SGR recrystallization of minerals, like quartz and feldspar, the rocks from mid-crust show strong grain size reduction and the mean grain size tends to be similar to that of neo-crystals (Platt & Behr 2011; Liu 2015). Grain size reduction can switch the material creep mechanism from dislocation creep to grain size sensitive creep (GSSC), which includes diffusion creep in solid-state (Coble or Nabarro-Herring creep), dissolution – precipitation, pressure solution and superplastic grain boundary sliding, etc. A higher nucleation rate than grain growth rate, and the switch of creep mechanism, may lead to the reduction of the rock strength by several orders of magnitude, which is more likely to occur in wet rocks. However, this softening process is transient, because the grain surface energy and dislocation energy will activate the grain boundary migration processes, and eventually cause the growth of mineral grains, which will then increase the grain size and drag the creep mechanism back to the dislocation creep. Therefore, there should be a dynamic equilibrium between the grain size reduction and grain growth caused by dynamic recrystallization.

Crystallographic preferred orientation (CPO): the activation of the intracrystalline slip systems determines the strength of the crystal, while the magnitude of the shear stress is directly affected by the initial orientation of these slip systems relative to the stress applied on the crystal (Burg et al. 1986). The slip systems will activate if the resolved shear stress along that surface is greater than Critical resolved shear stress (CRSS). Some crystals are most likely to slip along slip systems because of their initial orientation, which we called “soft orientation” and could be observed as strongly elongated, orientated and even dynamically recrystallized feldspar grains in the mylonitic matrix. While the grains with “hard
orientation” show as coarser porphyroclasts, which suggest that their initial orientation minimizes the resolve shear stress along slip systems. In the process of grain size reduction, the orientation of fine grains could be with inheritance to a certain extent, so the slip system of nearly parallel direction is active during progressive deformation (Svahnberg & Piazolo 2010).

1.2. The Controls of Rheological Weakening

In line with most studies, we consider that deformation in the viscous regime follows a constitutive law of the form:

$$\dot{\varepsilon} = A \sigma^n e^{-Q/RT}$$  \hspace{1cm} (1)

where $\dot{\varepsilon}$ is strain rate, $\sigma$ is stress, $T$ is temperature, $A$, $n$, and $Q$ are material constants, and $R$ is the gas constant. Other conditions, such as water fugacity, pressure and activation volume, grain size, etc., could be explicitly included, but their presence does not change the fundamental discussion.

Necessarily, any changes in the above parameters drive rheological change. Temperature and stress are environmental factors. We consider a variation in $A$, $n$ and/or $Q$ to be a change in material properties of the constitutive law. However, we can be more specific in identifying the factors that induce the change. The environmental factors, such as stress and temperature and stress -- with no change in the constitutive law (Jessell et al. 2005; Kaus & Podlachikov 2006; Dabrowski et al. 2015) -- can be temporary. The material property changes are (quasi)permanent, which have been recorded in extensive discussion of constitutive law evolution due to the proximate mechanisms of rheological weakening (e.g., grain size reduction, metamorphism, strain, fabric evolution, and phase mixing, etc.) among other factors (White 1980; Handy & Stünitz 2002; Montesi & Hirth 2003; Holyoke & Tullis 2006; Gerbi et al. 2016; Platt 2015; de Riese et al. 2019).

The values of the material properties can vary as a result of any of a number of factors, including the phases present, the phase distribution, and the operative deformation mechanisms (e.g., Hirth et al. 2001; Marti et al. 2018). In natural systems, we can consider homogeneous deformation to operate at a sufficiently large scale, but at the mineral scale, we know brittle deformation mechanisms and cavitation
can be present in a wide range of tectonic settings (e.g., Tullis & Yund 1987; Austrheim et al. 1996; Johnson et al. 2004; Rybacki et al. 2010; Menegon et al. 2013, 2015, 2017; Hertgen et al. 2017). Thus, although Equation (1) cannot apply at all scales in the macroscopic viscous regime, it remains a useful touchpoint for the discussion. Gerbi et al. (2014, 2016) focus on weak domains distribution in viscous conditions, which point out that interconnected weak domains are more effective in leading to weakening than weak-phase geometrical changes. Such interconnected weak zones formed during deformation, rather than interconnections between weak phases that existed prior to deformation. This model is similar to the concept of interconnected weak layers proposed by Handy (1994), but the weak zones are not necessarily rheologically homogeneous nor even have a layered morphology.

This inhomogeneous phase distribution (both microscale and macroscale) is the source of lithospheric mechanical heterogeneity, which has been a subject of interest among researchers for a considerable period (Barrell 1914). This variability in mechanical properties gives rise to a non-uniform distribution of stress (Griffith 1921; Voigt 1928; Reuss 1929). Consequently, this heterogeneous stress distribution plays a crucial role in driving various geological processes, including fluid flow, diffusion, and reactions facilitated by pressure gradients (Kamb 1961; Naus-Thijssen et al. 2011; Wheeler 2018), as well as damage caused by failure and grain size reduction (Segall & Simpson 1986; Johnson et al. 2004, 2008; Bercovici & Ricard, 2014). These processes, in turn, can lead to changes in rheology (Poirier 1980; White 1980; Mancktelow 2002; Gerbi et al. 2016; Gardner et al. 2017).

1.3. Magnitude of Weakening From Phase Distribution

Researchers have continued to focus on and study the effects of the distribution of constituent phases in polyphase materials and mylonites on strength and weakening computationally (e.g., Jordan 1988; Handy 1994; Handy et al. 1999; Treagus 2002; Ji et al. 2004; Gerbi et al. 2010, 2012, 2014, 2016) and experimentally (e.g., Jordan 1987; Dell’Angelo & Tullis 1996; Ross & Wilks 1996; Rybacki et al. 2003; Dimanov & Dresen 2005; Bystricky et al. 2006; Holyoke & Tullis 2006).

These studies have concentrated on the transition from load-bearing framework to interconnected weak layers. Holyoke & Tullis (2006) evaluated the role of interconnected weak layers formed from the
load-bearing framework using deformation experiments on a quartzofeldspathic gneiss, where they observed a 30%-60% strength reduction. Between the initial material and the quasi-steady-state flow products of experiments of the orthopyroxene granulite (Ross & Wilks 1996) and the aplite (Dell’Angelo & Tullis 1996), resulting in a weakening range of 15-50%.

In a complementary study by Takeda & Griera (2006), the focus was on modeling the viscoelastic deformation of interconnected geometries comprising strong and weak phases. Their investigation centered around tracking changes in the bulk effective viscosity within a shearing model block. Interestingly, they observed that models with an initial structure featuring interconnected weak layers showed minimal weakening. However, as the strong-phase-supporting framework gradually evolved into highly elongated weak layers, a noteworthy weakening of up to 50% was observed. Park et al. (2006) also indicates significant strength drops for non-mica-bearing weak layers (25%) and mica-bearing weak layers (70%).

The modellings from Gerbi et al. (2014, 2016) reveals the inclusion of only 1% of new weak zones in a small modal structure resulted in a significant reduction in bulk strength in the viscous quartz-feldspar geometries that exceeded the expected effect of modal change alone. In synthetic geometries, the strength of rocks containing weak zones was reduced by 82% compared to rocks without new weak zones. Similarly, in the geometries based on natural microstructures, the bulk strength decreased substantially by 68%.

Nucleation and interconnection of rheologically weak components of a rock or other polyphase composite reduces the bulk strength of that composite, with the degree of weakening dependent on the strength and geometry of the domains. Rephrasing the questions from the beginning into this study, I propose to:

1) Characterize the conditions and processes associated with new weak domain formation in viscous regimes.

2) Evaluate the bulk rheological consequences of weak domain formation.
In Chapter 2, I characterize the microstructure interconnecting weaker domains that I term bridge zones in less deformed rocks in granulite and amphibolite facies. In Chapter 3, investigates the controls of phases and their distributions on stress fields within a viscous regime numerically. Chapter 4 addresses the second goal: using a series of natural samples through strain gradient that deformed at a shallower depth and lower temperature to verify the conclusions of Chapter 2 and checking the relationship between high stress amplification and bridge zones, and the effect of bridge zones on bulk strength weakening of rock with numerical modeling.
CHAPTER 2

RHEOLOGICAL BRIDGE ZONES: THE INITIALIZATION OF STRAIN LOCALIZATION

2.1. Chapter Abstract

Strain localization occurs throughout the crust, in both the brittle and viscous regimes. The causes of strain localization remain under discussion. However, realistic rock records indicate that variations of stress and temperature are not as likely as the change of phase properties (e.g., physical properties, phase distribution, grain shapes and geometry, etc.) to be the dominant factor for weakening. Determining the cause(s) of localization requires investigation of the earliest stages of strain concentration in different P-T conditions. Our study focuses on 2 different weakly deformed rocks that exhibit localization on the millimeter or smaller scale. We combine optical and electron beam petrography with chemical mapping and electron backscatter diffraction to characterize these rheologically important domains. Morphologically, these localized zones appear to mechanically link rheologically weak phases or domains. These “bridge” zones typically comprise reduced grain sizes in a narrow band and dislocation creep does not seem to be the dominant deformation mechanism. Importantly, this phenomenon occurs not only in rocks inside the shear zones, but also in the less deformed margin outside the shear zones. Bridge zones result from not only in-situ grain size reduction (such as recrystallization or cataclasis), but also chemical processes resulting in phase mixing or element mobility on a short spatial scale.

2.2. Chapter Introduction

Strain localization occurs throughout the lithosphere, in both the brittle and viscous regimes, and runs through all geological scales (Sibson 1977; Ramsay 1980; Escartin et al. 1997; Hippertt 1999; Jolivet et al. 2004; Précigout and Gueydan 2009; Gueydan et al. 2014). In the lithosphere's viscous regime, strain localization results in plate boundaries, shear zones, shear bands, and similar structures. These areas of focused strain also impact fluid flow, mineralization and metamorphic reactions, geomorphic development, and seismic hazards, among other processes.
In the viscous regime, the occurrence of strain localization nearly always requires changes in bulk strength, which is typically quantified (e.g., Bürgmann and Dresen 2008) as:

$$\dot{\varepsilon} = A\sigma^n e^{-\frac{Q}{RT}}$$

(1)

where $\dot{\varepsilon}$ is strain rate, $\sigma$ is differential or deviatoric stress, $T$ is temperature, $A$, $n$, and $Q$ are material constants, and $R$ is the gas constant. Water fugacity, pressure, activation volume, grain size, and other conditions can be explicitly expressed in the equation, but their presence does not change the primary discussion. The above equation suggests that variations of environmental conditions – stress or temperature – or material properties can alter the bulk strength, and thereby initiate strain localization. We consider that the variation in the environmental variables affects bulk strength temporarily (Kaus & Podlachikov 2006; Dabrowski et al. 2015), while changes to material properties are (quasi)permanent.

The constitutive law above (Eq. 1) represents bulk, homogeneous behavior. The bulk properties incorporate the many structural and mineralogical heterogeneities that occur at smaller spatial scales, including individual mineral properties, deformation mechanisms, phase distribution, and grain shapes (Jordan 1988; Bons & Cox 1994; Ji et al. 2003; Montesi 2007 2013; Gerbi et al. 2016).

The development of strain localization and the accompanying weakening, combined with the growth and interconnection of these localized zones, reduces the regional bulk strength at larger spatial scales. Many researchers have studied and documented the connection between weaker phases and the weakening effect of this process on strength on a meter-to-kilometer scale (Handy 1990 1994; Bons & Cox 1994; Handy et al. 1999; Fusseis et al. 2006; Klepeis & King 2009; Gerbi et al. 2010; Culshaw et al. 2010, 2011).

Similar to the m- to km-scale, weak domain interconnections are also widespread on the millimeter and even smaller scale (Handy & Stünitz 2002; Vernon et al. 2004; Johnson et al. 2004, 2008; Mukherjee 2013). Numerous studies have focused on the effects of the microstructural phase property variations and weak phase interconnection on polyphase rock bulk strength observationally, experimentally, and numerically (e.g., Jordan 1987; Handy 1990 1994; Tullis et al. 1991; Ross & Wilks
1996; Ji et al. 2003; Johnson et al. 2004; Holyoke & Tullis 2006; Park et al. 2006; Stuntz & Tullis 2001; Takeda & Griera 2006; Marsh et al. 2009; Getsinger et al. 2013; Rennie et al. 2013; Gerbi et al. 2014, 2016). Moreover, there is growing evidence that microscale rheological change is due to more than just mechanical processes such as cataclastic and viscoelastic creep. Rather, chemically related mass transfer processes such as dissolution-precipitation and metamorphic reactions also play an essential role in weakening (e.g., Wheeler 1992; FitzGerald & Stünitz 1993; Wintsch et al. 1995; Stünitz et al. 2001; Kenkmann and Dresen 2002; Johnson et al. 2004; Jessell et al. 2005; Johnson et al. 2008; Park et al. 2006; Warren and Hirth 2006; Marsh et al. 2009; Svahnberg and Piazolo 2010, 2013; Platt 2015;).

In the studies referenced above, most of the discussion and documentation of weakening has focused on high-strain samples, rather than the starting state of deformation in natural rocks. With the recognition that even a very small volume of rheological interaction of weak phases can effectively weaken the rock without significantly changing the phase geometry (Gerbi et al. 2016), in this contribution we explore how these low-strain interconnections develop in natural rocks. In rocks at the margins of shear zones, we have identified rheological domains we term "bridge zones" that link weak phases (Figure 2.1) and may hold key evidence for localization initiation. Bridge zones manifest as relatively planar domains exhibiting a finer grain size and distinct morphology than surrounding areas (Figure 2.1). The domain may include phases and/or chemical compositions absent in the surrounding microstructures (Mukherjee and Koyi 2010; Goncalves et al. 2015; Pec et al. 2016). In this contribution, we describe those zones and consider their role in macroscale localization (Jessel et al. 2005).
Figure 2.1 Conceptual sketch of the bridge zone, which links weaker phases. The bridge zones comprise an aggregation of relatively fine grains in a narrow band. Although shown here as a single phase, bridge zones can include multiple phases.

2.3. Geological Background

2.3.1. Grenville Orogeny

The Mesoproterozoic Grenville Orogeny is considered a Himalayan-scale Precambrian mountain-building event, as a response to the convergence of the Rodinia supercontinent with Laurentia (Wynne-Edwards 1972; Davidson et al. 1982). Grenville deformation trends southwest-northeast along the southeastern margin of the Canadian Shield, extending from Mexico to Greenland. In the Ontario segment of the orogen, the three major structural units include the Grenville Front Tectonic Zone, Central Gneiss Belt, and Central Metasedimentary Belt. (Figure 2.2 Wynn-Edwards 1972; Davidson 1984; Culshaw et al. 1997).

The Grenville Orogeny is narrowly defined from ~1090 Ma to ~980 Ma, which consists of two pulses: the Ottawan Pulse (1090-1020 Ma) and the Rigolet Pulse (1000-980 Ma) (Moore & Thompson 1980; Rivers et al. 1989). Northwest directional thrust and convergence of Ottawan Pulse are considered to develop a Tibet-scale plateau with regional amphibolite to granulite grade metamorphism and localized high-pressure metamorphic zone (now the Central Gneiss Belt) (Rivers 2008). Orogenic collapse (1050-
1020 Ma) ended the Ottawan event, which was followed by Rigolet Pulse (Rivers 2008). This termination of Grenville events saw the orogen advance into its foreland, as shown by a series of thrusts in the Grenville Front Tectonic Zone (Rivers 2008; Jamieson et al. 2010; Shulman 2016).

The samples we investigated derive from a segment of the Grenville Orogen in Ontario, Canada. Samples presented here include a granulite from the interior Parry Sound domain (sample GB83C) and an amphibolite facies felsic gneiss from the Grenville Front Tectonic Zone (sample GB247) (Figure 2.2). The Grenville area has experienced extensive amphibolite facies deformation (Carr et al. 2000), but both samples are from the less deformed or undeformed margins of the shear zones suffering minimal impact.

Figure 2.2 Geological map of the western Grenville Province in Ontario, Canada (Modified from Culshaw et al. 1997 & Marsh et al. 2011).
Figure 2.2 (Continued) Main litho-tectonic domains and structural boundaries are shown. The red stars are the sampling sites. GB83C is a granulite from Parry Sound Domain, Central Gneiss Belt, while GB247 is an amphibolite facies felsic gneiss from the Grenville Front tectonic zone. GFTZ, Grenville Front Tectonic Zone; CGB, Central Gneiss Belt; CMB, Central Metasedimentary Belt; CMBBZ, Central Metasedimentary Belt Boundary Zone; PSSZ, Parry Sound shear zone.

2.3.2. **Granulite from Parry Sound Domain, Central Gneiss Belt**

The Central Gneiss Belt contains two major groups, the “allochthonous rocks”, e.g., Shawanaga, Parry Sound, Upper Go Home, and Moon River domains, and the “parautochthonous rocks”, e.g., Britt domain and the Lower Go Home domain (Culshaw et al. 1997, 2002). The granulite facies Parry Sound domain is considered the highest grade of metamorphism and the oldest metamorphic record (1161 ± 3 Ma) in the Central Gneiss belt (Davidson et al. 1982; Culshaw et al. 1997; Marsh et al. 2011). Kilometers-wide shear zones, largely derived from Parry Sound domain material, separate the Parry Sound domain from adjacent amphibolite facies domains (Culshaw et al. 1989, 2010; Marsh et al. 2011, 2013). The basal Parry Sound domain, defining the northern edge of the domain (Culshaw et al. 1989; Wodicka et al. 2000; Marsh et al. 2011), experienced an amphibolite facies overprint of earlier granulite facies metamorphism. Metamorphic conditions were approximately 5 to 8.5 kbar and 615°C to 870°C (Wodicka et al. 1996, 2000). The interior Parry Sound domain, a combination of mafic, intermediate, and granitic orthogneisses, experienced metamorphic conditions of 10 to 13 kbar and 760°C to 980°C, significantly higher than those of the basal Parry Sound domain (Culshaw et al. 1989 1997; Anovitz & Essene 1990; Wodicka et al. 2000; Marsh et al. 2011).

2.3.3. **Amphibolite Facies Gneiss from Grenville Front tectonic zone**

Along the northern Georgian Bay coast, the Grenville Front Tectonic Zone contains a series of granitoids, migmatitic orthogneisses (protolith age of 1800-1700 Ma; Foley 2015), and subordinate paragneisses. However, as one of the youngest units in the Grenville Province, these rocks were modified by the metamorphic and deformational events at the termination of the Grenville orogen. NW-directed thrusting led to a series of SE-dipping mylonitic shear zones with strong foliation and down-dip lineation development (Davidson & Bethune 1988; Jamieson et al. 1995; Rivers 2008; Shulman 2016). From west to east, approaching the core of the orogenic belt, the metamorphic grade increases rapidly. West of the
Grenville Front, rocks preserve pre-Grenville history, with limited Grenvillian overprint. East of the Front, amphibolite facies metamorphism dominated, with a metamorphic peak of 6.3±0.4 kbar/594±25 °C, but some pre-Grenville conditions are preserved at 10.8 ±0.8 kbar/747±50 °C (Jamieson et al. 1995).

2.4. Analytical Methods

2.4.1. Cathodoluminescence (CL)

The description of CL features in this chapter is derived from panchromatic CL unless otherwise stated. The panchromatic CL were imaged using the Tescan panchromatic (350-650 nm) CL detector on the Tescan Vega XMU II scanning electron microscope (SEM) in the School of Earth and Climate Sciences, University of Maine with chamber pressure less than 10^-2 Pa and 20 KV acceleration voltage.

2.4.2. Electron Backscatter Diffraction (EBSD)

All EBSD data were collected at 20 kV acceleration voltage, ~6 nA beam current, 0.45-3.5 μm step size, and 70° sample tilt under high-vacuum conditions from the Tescan Vega XMU II SEM equipped EDAX-TSL EBSD system and EDAX-TSL OIM Data Collection 5.31 software in the School of Earth and Climate Sciences, University of Maine. An EDAX Genesis Energy Dispersive Spectroscopy (EDS) was also used to do simultaneous distinguish phases through chemical analysis. Both samples were polished with 0.3 μm alumina suspension and a 0.02 μm colloidal silica suspension to eliminate surface damage before the thin carbon coating to avoid electron charging.

Post-processing of the data was applied to the EDAX-TSL OIM Analysis 5.31 software. Non-indexed and low-indexed pixels (CI<0.1) were replaced with well-indexed adjacent pixels (CI>0.1). The grains in the post-processing possess an internal orientation deviation of ≥10° and a minimum grain size of 4 pixels, and the effects of twinning are subsequently absent. Crystallographic preferred orientation (CPO) pole figures, grain size distributions, grain reference orientation deviation, and kernel average misorientation parameters were calculated and presented herein as relevant. In this study, grains smaller than 5 microns are not included in the statistics, because they are unreliable at the range of analysis step size.
Grain reference orientation deviation and Kernel average misorientation are common parameters used to evaluate the plastic strain of each point (Unnikrishnan et al. 2016; Gussev & Leonard 2019). The grain reference orientation deviation is the deviation between the orientation of the grain at a certain position in the grain and the reference orientation or average orientation of the whole grain. Local misorientation values ranging from 0°-10° are shown on the grain reference orientation deviation maps. Kernel average misorientation is calculated and averaged between the kernel center data point and the surrounding data points in the kernel by custom setting (the second neighbor of hexagonal pixels kernel was used here).

2.4.3. **Electron Probe Microanalysis (EPMA)**

Electron microprobe analysis (EPMA) on the Cameca SX-100 was applied at The University of Maine, School of Earth and Climate Sciences. The target areas and grain were analyzed by mapping & spot analysis. We used such settings for major element mapping (Na, Ca, K, Si and Al): operating voltage of 15 kV, spot size of 5 μm, beam current of 100 nA with a dwell time of 200 ms. Analytical conditions included 15 kV accelerating voltage, 10 nA beam current and a 5 μm spot size with a 300 ms counting time for the elements (Na, K, Ca, Mg, Si Al, Fe, Mn, and Ba) in the feldspars.

2.4.4. **Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICP-MS)**

Polished but non-carbon-coated thin sections were analyzed for in situ trace elements in LA-ICP-MS at the University of Maine's MAGIC (Microanalytical Geochemistry and Isotope Characterization) laboratory. The method used is described in Cruz-Uribe et al. (2021). Mapping was done using an ESI NWR193UC laser ablation system coupled to an Agilent 8900 quadrupole ICP-MS. The MAGIC Lab has fast washout mapping setup with 1 mm ID tubing connected directly from the TV3 laser ablation chamber to a dual concentric injector ICP-MS torch (DCI1) from ESI. A 5x5 μm square spot with a 200 Hz laser repetition rate and 166 μm/s scan speed were applied to collect the following isotopes: $^{28}$Si, $^{39}$K, $^{41}$Ca, $^{47}$Ti, $^{56}$Fe, $^{85}$Rb, $^{137}$Ba, $^{208}$Pb.
2.5. Observations from Natural Rock Samples

2.5.1. Bridge Zone in Granulite Facies

2.5.1.1. Sample Description

The granulite sample, GB83C, derives from the interior Parry Sound domain of the Central Gneiss Belt (CGB) (45°12'52.93"N; 80° 6'48.11"W; Figure 2.3a), many meters from the nearest macroscopic, amphibolite facies shear zone (cf. Culshaw et al. 2010). Throughout the outcrop, minor mafic orthogneiss is boudinaged within dominant felsic to intermediate orthogneiss, characteristic of granulite facies deformation in the region.

The sample we investigated in detail includes a portion of a mafic boudin as well as the felsic host, the latter of which is our focus. The boudin comprises Opx+Cpx+Hbl+Qz+Bt (Figure 2.4a, mineral abbreviations from Whitney & Evans 2010). The host granoblastic rock contains coarse to medium-sized grains, with the mineral assemblage of Qz (~40%) + Pl (~55%) + Kfs (~5%) + Fe Oxide + Cpx + Opx (~20%). Marsh et al. (2011) estimated the metamorphic conditions at a nearby location as ~11.5kbar and ~850°C. Quartz grains exhibit patchy extinction or extinction bands, consistent with crystallographic misorientation <10°. Quartz clusters are irregular in shape, and elongate parallel to foliation with axial ratio of ~7-8. Plagioclase grains are subhedral, with a size of 1000-1500μm, and many display anti-perthite textures (mostly width <5 μm). Plagioclase underwent brittle-ductile deformation: uneven extinction as well as curved twinning can be observed simultaneously with cracks.
Figure 2.3 Outcrop photos of GB83C and GB247. (a) GB83C was sampled in the interior Parry Sound domain and out of any shear zones in this region. The sample, which exhibits a typical granulite characteristic containing the regional lineation and foliation, experienced only weak deformation under amphibolite facies conditions. The broader interior Parry Sound domain comprises a series of mafic, intermediate, and granitic orthogneisses, whose metamorphic conditions were ~11.5 kbar and ~850°C (Marsh et al. 2011). The sample came from the margin of a mafic boudin (Opx+Cpx+Amp+Qz+Pl+Bt) hosted in felsic granulite. (b) GB247 was sampled in a felsic composition outcrop of a km-scale strain gradient subparallel shear zones that accommodated shortening ca. 1 Ga at the Grenville Front (Davidson & Bethune 1988; Gerbi et al. 2016). This sample is at the low strain end of the gradient. The exact P-T conditions were not reported in the sampling area, but the sample shows a characteristics of amphibolite facies containing the regional lineation and foliation.
Figure 2.4 Texture and bridge zone distribution in GB83C and GB247. (a) Backscattered electron image (BSE) of GB83C. Granoblastic fabric contains coarse to medium-sized grains in the felsic part of GB83C, with the mineral assemblages of Qz (~30%) + Pl (~50%) + Fe Oxide + Cpx + Opx (<20%). (c) BSE image of GB247. As a granitoid gneiss, GB247 consists primarily of quartz, plagioclase, K-feldspar, hornblende, and biotite. All grains are elongated parallel to regional gneissic foliation. Cathodoluminescence images of GB83C (b) and GB247 (d). The red areas mark the bridge zones. Bridge zones occur in stronger mineral phases that connect weaker mineral phases and form networks within the rock. The red squares in (a) are shown in Figure 2.5 (a), (b), and (c), while the red square in (c) are shown in Figure 2.5 (d), (e), and (f). The blue square in (c) is shown in Figure 2.7a.

2.5.1.2. Distribution of Bridge Zones

Fine-grained domains are present throughout the rock, constituting <10% of the area (Figure 2.4b). These bridge zones link relatively weak quartz domains and occur in the interior or along the edges of plagioclase grains (Figure 2.4b).

2.5.1.3. Internal Bridge Zone Textures

The bridge zones in this sample contain a mixture of relatively fine-grained plagioclase (~85%), K-feldspar (~10%), and locally <5% quartz (Figure 2.5a-c; Figure 2.6a&b).
Plagioclase grains at the edge of a bridge zone exhibit subgrains developed at approximately 200 μm. In the bridge zone, the grain size of plagioclase ranges from 5 to 200 μm with most grains <120 μm. The potassium feldspar grains are generally smaller than 50 μm. The bridge zones show polygonal shapes with straight grain boundaries, many forming 120° triple junctions (Figure 2.5a&b; Figure 2.6a&b).

The cathodoluminescence intensity of fine plagioclase grains in bridge zones is brighter than in the adjacent coarser plagioclase (Figure 2.5b; Figure 2.6b). Moreover, almost all the fine grains have brighter CL cores and very thin darker CL rims (Figure 2.5b; Figure 2.6b). Some coarse plagioclase displays occasional bright cathodoluminescent rims and sometimes coarser exclusion lamellae (>10 μm) at the boundaries against quartz grains (blue arrows in Figure 2.6a&b). Color CL images from another region of the same sample show that large-grained plagioclase is generally pink, while bright fine plagioclase grains show stronger blue luminescence (see Appendix). However, the backscattered electron images of plagioclase show uniform intensity inside and outside the bridge zones (Figure 2.5a; Figure 2.6a).

In contrast to their lamellar appearance in host plagioclase grains, K-feldspar grains in the bridge zones form irregular interstitial grains among the fine plagioclase (Figures 2.5a&b; 2.6a&b). At the edges of the bridge zones or in quartz cluster embayments, the K-feldspar grains occur as granular aggregates (Figures 2.5a&b; 2.6a&b).
Figure 2.5 Bridge zone microstructures in GB83C and GB247. (a) BSE image, (b) CL image, and (c) phase map for GB83C. The bridge zone connects two quartz crystal clusters. It appears as fine-grained plagioclase and K-feldspar. Adjacent areas are coarse-grained plagioclase containing exsolution of K-feldspar lamellae. K-feldspar exhibits different morphology inside and outside bridge zone. CL images reveal differences in the brightness of fine plagioclase within the bridge zone. (d) BSE image, (e) CL image, and (f) phase map for GB247. A unique microstructure is the fine grain mixture (20-50μm) in which the tips of two quartz clusters are linked through the K-Feldspar. This fine-grained mixture is a combination of K-feldspar, plagioclase, and quartz. There is no direct contact between coarse-grained plagioclase and bridge zone. The red dotted lines represent the boundaries of the bridge zones. The square in (f) is shown in Figure 2.10.
Figure 2.6 Microstructure and major chemical element maps of a bridge zone in GB83C. In the BSE image (a) and CL image (b) of the selected region, it can be seen that the bridge zone is distributed in the center of the image. The difference in plagioclase CL brightness fine grains in the bridge zone and the coarse grains outside the bridge zone is intuitive. Some of the coarse-grained plagioclase near the edge of the quartz can be seen in some of the brighter CL rims. The colored arrows in (a) and (b) show the different forms of K-feldspar. The green arrows point to the typical morphology of K-feldspar in the bridge zone: clustered granular or fine tabular shape at the edge of other grain grains. The red arrows point to the fine exsolution lamellae K-feldspar inside the coarse-grained plagioclase (5 μm). The blue arrows show the coarser exsolution lamellae K-feldspar inside the coarse-grained plagioclase. The yellow arrows point to the accumulation of K-feldspar near quartz grains. The red boxes in (a) are the sampling locations of the EPMA map: (c) Ca distribution map, (d) K distribution map, and (e) Na distribution map. K is mainly concentrated in K-Feldspar with very low quartz and plagioclase content. K is evenly distributed in plagioclase, and no significant difference is found in and outside the bridge zone. However, Na and Ca are mainly distributed in plagioclase, and no obvious local aggregation was observed.
2.5.2.Bridge Zone in Amphibolite Facies

2.5.2.1. Sample Description

GB247 also derives from a felsic orthogneiss (45°57'29.60"N; 81°14'30.62"W), at the low-strain edge of a km-scale strain gradient that accommodated shortening ca. 1 Ga at the Grenville Front (Davidson & Bethune 1988; Jamieson et al. 1995). Though the exact P-T conditions were not reported in the sampling area, the foliated and lineated sample shows a characteristics of amphibolite facies mineral assemblage modified from an earlier granulite facies assemblage (Figure 2.3b).

GB247 primarily consists of quartz (~22%), K-feldspar (~40%), plagioclase (~23%), hornblende (~9%), and biotite (~6%) (Figure 2.4c). All minerals are elongated and oriented, defining the foliation. K-feldspar and plagioclase are subhedral (~2mm) and elongate (>1mm) polycrystalline quartz ribbons lie in the intergranular space controlled by the feldspar-defined gneissic foliation. No phase forms an anastomosing network around another phase. A thin rim is sometimes displayed in feldspar. Twinning is very rare in plagioclase and K-feldspar grains. However, undulatory extinction and fractures/cracks appear within the feldspars. K-feldspar grains are micro-perthitic textures, visible at high magnification. Myrmekite occurs at the smooth edge of some of the coarser K-feldspar with quartz.

2.5.2.2. Distribution of Bridge Zones

Bridge zones in GB247 are less prevalent than in GB83C, but still sufficiently abundant to identify characteristics (Figure 2.4d). Most GB247 bridge zones exist as narrow, fine-grained bands interconnecting quartz clusters, with the bridge zone subparallel to the regional foliation and the elongate quartz clusters. Most of the visible bridge zones are located at the edge of the K-feldspar grains, containing aggregates of K-feldspar, plagioclase, and quartz.

2.5.2.3. Internal Bridge Zone Textures

Bridge zones in GB247 were first documented by Gerbi et al. (2016). The most identifiable microstructure is a fine-grained, polyphase mixture (20-100μm) connecting the tips of two quartz clusters. This fine-grained mixture is a combination of K-feldspar (~66%), plagioclase (~27%), and quartz (~7%)
(Figure 2.5d-f). The modal ratio of plagioclase to quartz is highly variable, averaging 4:1, but can range from 9:1 to 3:1. The grain size of K-feldspar ranges from a few microns to ~100 μm, and is concentrated in the range of 5-60 μm. The fine K-feldspar grains within the Bridge zone can be divided into two groups, those with uniform grains and those with perthite texture (stars in Figure 2.7a). Plagioclase is usually a few microns to ~80 μm across, and quartz is a few microns to ~50 μm across. The decrease in grain size from over 1 mm in the parent grains to less than 100 μm in the bridge zones is dramatic. We have not observed direct contact between coarse-grained plagioclase and the bridge zone. Fine plagioclase often encloses K-feldspar of variable shape, and the boundary between them is blurred (red circles in Figure 2.7a). Some extremely fine plagioclases are usually distributed along grain boundaries where the K-feldspar grains with perthite texture are in contact with other K-feldspar grains (red circles in Figure 2.7a), especially uniform ones. Despite some similarities, the bridge zone texture differs from myrmekite texture. Quartz in myrmekite is a worm-like, or vermicular, intergrowth in feldspars, while most grains in the bridge zone are nearly equiaxial granular or fine tabular quartz close to plagioclase (red arrows in in Figure 2.7a).

The coarser K-feldspar grains outside the bridge zone contain ~5 μm wide bands that are brighter in BSE than the surrounding area (black arrows in Figure 2.7a). These linear bands intersect within the K-feldspar to form a network and connect to the edges of K-feldspar, especially into the bridge zones. The bands overlap with trails of micron-scale fluid inclusions (Figure 2.7d&e).
Figure 2.7 Chemical characteristics and fluid inclusion trails near a bridge zone in GB247. (a) BSE image revealing brighter bands in coarse-grained K-feldspar outside the bridge zone. These strips are typically 5μm wide and are connected to each other to form a network (black arrows). They are shown as a (b) Na-poor and (c) K-rich bands in the EDS element distribution maps (red box in a). In plane polarized light (d and e, blue box in a) the domains appear as tiny bands of orientated fluid inclusions. The red star in (a) marks the fine K-feldspar with the perthite texture inside a bridge zone, while the blue star marks the uniform fine K-feldspar. These uniform K-feldspars seem to contain slightly higher K. The red circles or ellipses in (a) represent some fine-grained plagioclase containing some fine K-feldspar with blurred boundaries inside. Inside the black circles or ovals are "precipitated" plagioclase (plagioclase and quartz aggregates) along the potassium feldspar boundary. The red arrows mark fine tabular quartz grains that are distributed along the grain boundaries at the edges of the fine plagioclase.
2.6. Crystallographic Orientation Analysis

2.6.1. Crystallographic Preferred Orientation (CPO) Of Fine Grains in Bridge Zones

The coarse-grained plagioclase outside the bridge zones and their associated K-feldspar lamellae have crystallographically aligned orientations (Figure 2.8c). In addition, the orientation of fine plagioclase within the bridge zones overlaps with the orientation of coarse plagioclase grains outside the bridge zones (Figure 2.8a). K-feldspar and quartz exhibit similar relationships, but with a much smaller dataset, which reduces confidence in any interpretations (Figure 2.8b).

In GB247, crystallographic orientations of K-feldspar and quartz in bridge zone coincide with those of large grains of K-feldspar and quartz outside the bridge zone (Figure 2.8e&f). The CPO of fine plagioclase grains appears to overlap with the orientation of one external K-feldspar grain (Figure 2.8a).

![Figure 2.8 Crystallographic orientation fabrics (COF) of fine grains in bridge zones in GB83C and GB247 (data selected areas shown in Figure 2.5). (a) fine-grained plagioclase in GB83C; (b) fine-grained K-feldspar in GB83C; (c) pole figures of the K-feldspar and plagioclase outside bridge zone. (d) fine-grained plagioclase in GB8247; (e) fine-grained K-feldspar in GB247; (f) fine-grained quartz in GB247. It can be seen that the orientation of large grains adjacent to the bridge zone moderately coincide with the COF of fine grains in the bridge zone. (c) illustrates that the orientations of coarse-grained plagioclase and their internal exsolution K-feldspar lamellae also have consistency.](image-url)
2.6.2. **Grain Reference Orientation Deviation and Kernel Average Misorientation**

In both GB83C and GB247, areas of high deviation from the grain reference orientation are generally distributed at the edges of large grains adjacent to the bridge zones, while the fine grains and interior of larger grains exhibit low deviations (Figure 2.9&2.10b, e &h). Similarly, kernel average misorientation values are high at edges of adjacent large grains and low in the fine grains of the bridge zone and in large grain interiors (Figure 2.9&2.10c, f & i).

![Figure 2.9](image)

**Figure 2.9** Grain orientation and misorientation distribution maps of a bridge zone in GB83C. (a), (d) and (g) Inverse pole figures, (b), (e) and (h) grain reference orientation deviation distributions and (c), (f) and (i) kernel average misorientation distributions of GB83C (the second neighbor of hexagonal pixels kernel was used here). High values of grain reference orientation deviation are generally distributed at the edges of adjacent large grains, while the fine grains and inside larger grains are low deviation. Kernel average misorientation values are high at edges of adjacent large grains suggesting a high dislocation density in these areas. The white dotted lines represent the boundaries of the bridge zone.
Figure 2.10 Grain orientation and misorientation distribution maps of a bridge zone in GB247. (a), (d) and (g) Inverse pole figures, (b), (e) and (h) grain reference orientation deviation distributions and (c), (f) and (i) kernel average misorientation distributions of GB247 (the second neighbor of hexagonal pixels kernel was used here). The white dotted lines represent the boundaries of the bridge zone.

2.7. Chemical Characteristics in Bridge Zones

The observations above – particularly the difference in CL color and intensity of fine-grained plagioclase in GB83C and the introduction of fine-grained plagioclase phase in the bridge zone of GB247
indicate that the bridge zones of both samples experienced chemical change. In this section we examine the chemical variations within and outside the bridge zones in more detail.

2.7.1. EPMA

2.7.1.1. EPMA of Granulite Facies Sample

We performed EPMA chemical mapping and spot analysis for GB83C in locations with CL variations in plagioclase and K-feldspar inside and outside bridge zones (Table 2.1). Although all the data show that plagioclase in GB83C contains ~70% Albite - ~30% Anorthite (Figure 2.11a), the finer grains with brighter CL (0.31) have a slightly higher concentration Ca than the larger grains with darker CL (0.28). The Na and K of the brighter fine grains are also slightly lower than the darker larger ones (Figure 2.11b&c). However, this slight chemical change is not evident in the EPMA maps of Ca, Na and K elements, which both show uniform gray levels inside and outside the bridge zones (Figure 2.6c-e). Due to spot size, we were only able to collect the thicker lamellae and K-feldspar inside the bridge zone. The composition of the two is effectively the same (Figure 2.11d-f).
Figure 2.11 EPMA analyses of bridge-zone finer and larger plagioclase and K-feldspar in and out bridge zones of GB83C and GB247. (a) – (c) Ca, Na and K distribution in the plagioclase grains in GB83C. The orange lines show the data from the fine plagioclase grains in the bridge zones, which almost have a brighter CL. The blue lines show the data from the larger plagioclase (darker CL). Although all the data show that plagioclase in GB83C contains ~70% Albite ~30% Anorthite, the finer grains with brighter CL (0.31 in average) have a slightly higher concentration Ca than the larger grains with darker CL (0.28 in average). The Na and K of the brighter fine grains are also slightly lower than the darker larger ones. (d) – (f) show the Ca, Na and K distribution in the K-feldspar grains in GB83C. The orange lines show the data from the K-feldspar grains in the bridge zones, while the blue lines show the data from the coarser lamellae. Ca, Na, and K barely changed in K-Feldspar inside and outside the bridge zones. (g) – (i) show the Ca, Na and K distribution in the plagioclase grains in GB247. The orange lines show the data from the plagioclase fine grains in the bridge zones, while the blue lines show the data from the coarser plagioclase. Compared with coarse plagioclase outside the bridge zone, the fine-grained plagioclase in the bridge zone is significantly K-poor and Na-rich, while the content of Ca is similar, but the fine grains have slightly lower Ca.
Table 2.1 EPMA spot analysis from plagioclase and K-feldspar in GB83C and GB247

<table>
<thead>
<tr>
<th>Sample Phase</th>
<th>GB83C</th>
<th>GB247</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>CL</td>
<td>CL</td>
</tr>
<tr>
<td></td>
<td>PL</td>
<td>Brighter Fines</td>
</tr>
<tr>
<td>wt %</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Na₂O</td>
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<td>7.66</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.43</td>
<td>0.36</td>
</tr>
<tr>
<td>MgO</td>
<td>0.00</td>
<td>0.03</td>
</tr>
<tr>
<td>CaO</td>
<td>5.85</td>
<td>6.37</td>
</tr>
<tr>
<td>MnO</td>
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<td>-0.02</td>
</tr>
<tr>
<td>FeO</td>
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</tr>
<tr>
<td>BaO</td>
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<td>0.01</td>
</tr>
<tr>
<td>SiO₂</td>
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<td>60.26</td>
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<tr>
<td>apfu</td>
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</tr>
<tr>
<td>Na</td>
<td>0.67</td>
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</tr>
<tr>
<td>K</td>
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<td>0.02</td>
</tr>
<tr>
<td>Mg</td>
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<td>0.00</td>
</tr>
<tr>
<td>Ca</td>
<td>0.28</td>
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</tr>
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<td>Mn</td>
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</tr>
<tr>
<td>Fe</td>
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<tr>
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</tr>
<tr>
<td>XCa</td>
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</tr>
<tr>
<td>XK</td>
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<td>0.02</td>
</tr>
<tr>
<td>XNa</td>
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<td>0.67</td>
</tr>
<tr>
<td>n</td>
<td>54</td>
<td>912</td>
</tr>
</tbody>
</table>
2.7.1.2. EPMA of Amphibolite Facies Sample

In a previous section, we described the fluid inclusion bands of brighter BSE in the coarse K-feldspar grains (parent grains) of GB247. These low Na/high K bands are clearly visible in the EPMA and EDS major elements maps (Figure 2.7). The maps do not show clear major element differences between the grains inside bridge zones and the grains outside bridge zones. Spot analysis was used to analyze the composition of the fluid inclusion zone and the rest of the coarse-grained K-feldspar and the composition of the fine grains with uniform and perthitic texture in the bridge zone (Table 2.1). The uniform fine-grained K-feldspar has a higher K than the grains with the perthite texture, but is not much different from the coarse-grained K-feldspar. The spot analysis also includes plagioclase within the bridge zone and not far away from (but not adjacent to) the outside. Compared with coarse plagioclase outside the bridge zone, the fine-grained plagioclase in the bridge zone is significantly K-poor and Na rich, while the content of Ca is similar (Figure 2.11g-i).

2.7.2. LA-ICP-MS

To evaluate whether minor or trace elements track any chemical processes not revealed by the major elements, we used LA-ICP-MS to map one bridge zone in each sample for Ca, Ti, Ba, Rb, Pb, Mg, and Fe (Table 2.2; Figure 2.12).
Figure 2.12 LA-ICP-MS maps of GB83C. The mapped area is similar to Figure 2.6. A detailed description can be found in Table 2.2.
Table 2.2 LA-ICP-MS maps of GB83C and GB247

<table>
<thead>
<tr>
<th>Sample</th>
<th>GB83C</th>
<th>GB247</th>
</tr>
</thead>
<tbody>
<tr>
<td>K</td>
<td>Concentrated in K-feldspar. The fine exsolved lamellae have lower K content compared to the coarse exsolved lamellae and within the bridge zone, and a small K enrichment is seen in the coarse-grained plagioclase fractures.</td>
<td>Concentrated in potassium feldspar, High-K bands within K-feldspar clearly visible.</td>
</tr>
<tr>
<td>Ca</td>
<td>Within plagioclase, high concentrations correlate with a brighter (bluer) CL</td>
<td>Mainly in plagioclase and Biotite, and more concentrated in biotite</td>
</tr>
<tr>
<td>Ti</td>
<td>Although Ti map lacks the sharp texture of the other elements, Ti concentrations seem to be mirror those of Ca, with lower abundance in the brighter region of CL in plagioclase.</td>
<td>Ti map lacks the sharp texture of the other elements,</td>
</tr>
<tr>
<td>Fe</td>
<td>Although absolute Fe content in this felsic sample is very low, Fe is enriched at grain boundaries, intra-grain fractures, and some subgrain edges.</td>
<td>Because of the appearance of Biotite, Fe is more concentrated in these minerals. The enrichment in grain boundaries and fractures is still visible, but not in the inclusion bands.</td>
</tr>
<tr>
<td>Rb</td>
<td>Concentrated in K-Feldspar, including exsolution lamellae and grains in bridge zones. Also, higher at the boundaries between plagioclase and quartz. The fine exsolved lamellae have lower K content compared to the coarse exsolved lamellae and within the bridge zone, and a small K enrichment is seen in the coarse-grained plagioclase fractures.</td>
<td>Mainly concentrated in K-Feldspar, and enriched in inclusion bands</td>
</tr>
<tr>
<td>Ba</td>
<td>Mainly concentrated in K-Feldspar, including exsolution lamellae and grains in bridge zones. The fine exsolved lamellae have lower K content compared to the coarse exsolved lamellae and within the bridge zone, and a small K enrichment is seen in the coarse-grained plagioclase fractures.</td>
<td>Mainly concentrated in K-Feldspar, and enriched in inclusion bands</td>
</tr>
<tr>
<td>Pb</td>
<td>Mainly concentrated in K-Feldspar, including exsolution lamellae and grains in bridge zones. The fine exsolved lamellae have lower K content compared to the coarse exsolved lamellae and within the bridge zone, and a small K enrichment is seen in the coarse-grained plagioclase fractures. Compared to the very low content in other minerals, it is relatively enriched in plagioclase. Enriched in inclusion bands.</td>
<td>Mainly concentrated in K-Feldspar. Compared to the very low content in other minerals, it is relatively enriched in plagioclase. Enriched in inclusion bands</td>
</tr>
</tbody>
</table>

2.7.2.1. LA-ICP-MS of Granulite Facies Sample

The Ca map clearly shows the relationship between CL brightness and Ca, where there is a brighter (bluer) CL and where there is a more abundant Ca distribution (Figure 2.12c). Even though the
results were relatively noisy, the distribution of Ti seems to be opposite to that of Ca, with a lower distribution in the brighter regions of CL in plagioclase (Figure 2.12b).

Ba, Rb and Pb are mainly concentrated in K-feldspar (Figure 2.12 e-g), which shows obvious consistency with K (Figure 2.12d). This characteristic can be seen in K-feldspar lamellae and K-feldspar grains in bridge zones. The fine exsolved lamellae have lower K content compared to the coarse exsolved lamellae and within the bridge zone, and a small K enrichment is seen in the coarse-grained plagioclase fractures, which is the same as the distribution of Ba, Rb and Pb. Although Fe content in this felsic region is very low, it is obviously enriched at the grain boundaries (Figure 2.12h). Fe enrichment can be seen not only in grain boundaries, but also in intra-grain fractures, healed fractures and even some subgrain edges.

2.7.2.2. LA-ICP-MS of Amphibolite Facies Sample

The low Na/high K inclusion bands seen in EPMA maps were still visible in LA-ICP-MS maps, and Si, Rb, Ba and Pb were also enriched in these bands. Together, they form networks inside and outside the bridge zone. Due to the appearance of dark minerals (biotite, amphibole, etc.), Fe content is more concentrated in these minerals. The enrichment in grain boundaries and fractures is still visible, but not in the inclusion bands.

2.8. Discussion

2.8.1. Detailed Definition of Rheological Bridge Zone

We base our working definition of a microstructural bridge zone on observations from the granulite and amphibolite facies gneiss samples described here. Macroscopically, both are weakly or undeformed in the latest episode to affect the region, and both lie near but outside meter-scale shear zones (cf. Davidson and Bethune 1988; Culshaw et al. 2010).

The bridge zones, which we interpret as having formed well after the macroscopic fabric and boudinage, are relatively planar domains exhibiting grain size reduction and distinct morphology compared to surrounding areas. They can either follow grain boundaries or transect large crystals. The fine grains are relatively equant and appear to have weak crystallographic inheritance from the coarser parent grains.
Morphologically, the bridge zones appear to mechanically link rheologically weak phases or domains and form networks in the rock. In the two samples described in this paper, based on standard flow laws (e.g., the plagioclase/feldspar flow laws based on Rybacki et al. 2006 and the quartz flow law based on Hirth et al. 2001) and the inferred deformation conditions, quartz is the weak phase, and feldspar is the strong phase. Mica, especially biotite, is also a weak phase but is modally minor in these samples.

Bridge zones comprise phases and/or chemical compositions absent in the surrounding areas. In our samples, the CL intensity of fine plagioclase grains in bridge zones in granulite (GB83C) is different from the parent grains, indicating chemical mobility. In addition, in GB247, fine-grained plagioclase, which is not in direct contact with coarse-grained plagioclase, requires mass transfer due to chemical processes/reactions. Chemical mobility and the introduction of new phases is consistent with chemical-mechanical coupling processes documented under all pressure-temperature conditions (e.g., Wintsch et al. 1995; Wintsch & Yi 2002; Johnson et al. 2008; Marsh et al. 2009, 2011), which is consistent with the growing recognition of chemical processes in the accommodation of strain (e.g., Stünitz et al. 2001; Menegon et al. 2013).

2.8.2. Processes Contributing to the Formation of Bridge Zone

When evaluating the processes that operated to form the bridge zones, we must satisfy the following observations:

- Grain size distribution: In the interior or edge of plagioclase in GB83C and K-feldspar in GB247, between linking quartz crystal clusters, grain-size reduction zones (<200μm) occur in bridge zones. Each phase within the bridge zones usually distributes as relatively uniform grain sizes, much finer than the host grains. The grain size reduction of feldspar minerals is the dominant process in bridge zone development. It has long been thought that the reduction in grain size of feldspar can be attributed to: dynamic recrystallization (e.g., Tullis & Yund 1985; Tullis et al. 1990; Svahnberg & Piazolo 2010); fracture (e.g., le (Paterson 1982; Goodwin and Tikoff 2002; Ree et al. 2005; Johnson et al. 2008; Okudaira et al. 2017); and dissolution-precipitation, reaction and neocrystallization (e.g., Gerald &
Grain morphology: Some of the internal phases of the bridge zones, such as quartz and K-feldspar are tabular or granular, with a different morphology from the exsolution lamellae in surrounding grains. The quartz and K-feldspar in the bridge zone may form by precipitation in the dilated zone opened by grain boundary sliding (cf. Kruse & Stünst 1999; Ree et al. 2005; Kilian et al. 2011; Platt 2015).

Mode changes: Reactions leading to modal changes can be due to closed or open system behavior and commonly involve fluid-mediated element transport. Local stress variations can drive reactions at microstructural sites (e.g., Simpson 1985; Simpson & Wintsch 1989; Menegon et al. 2008; Moore et al. 2019; Dong et al. 2019; Zhou et al. 2022).

Subgrain rotation and microfracturing at the host plagioclase boundaries: Boundary deformation of feldspar, whether core-and-mantle texture or otherwise, is common in mylonized rocks (Passchier and Trouw 2005). Boundary deformation is often due to subgrain rotation (e.g., Olsen and Kohlstedt 1984; Kruse et al. 2001; Holyoke and Tullis 2006; Svahnberg & Piazolo 2010; Hansen et al. 2013). Microfracture is a typical texture within feldspar minerals, especially at the edges during loading (Johnson et al. 2004, 2008; Vetegren et al. 2017; Svahnberg & Piazolo 2010; Menegon et al. 2013; Marti et al. 2017), and the same is true in the lower crust (Petley-Ragan et al. 2021).

Alignment of crystallographic orientations of feldspar within and outside the bridge zone: Crystallographic inheritance or control may be derived from subgrain rotation (e.g., Pennacchioni et al. 2010; Ceccato et al. 2017; Svahnberg & Piazolo 2010). Alternatively, neocrystallization may occur with the orientation of the adjacent grain lattice (e.g., Ree et al. 2005; Fukuda et al. 2012; Gardner et al. 2021).

CL pattern in plagioclase: The plagioclase in GB83C shows CL characteristics of different brightness. The most obvious is the bright CL fine-grained plagioclase inside the bridge zone and the dark CL coarse-grained plagioclase outside. However, the heterogeneity of CL is often seen in single grains of plagioclase. For example, fine grains often have bright cores and very thin dark rims, while coarse grains show bright rims and dark cores at the edge of contact with weak domains (quartz). In addition, trace
elements can affect the CL (e.g., Ba, Rb, Sr, Mn, Pb, REE, etc. substituting for Ca, K and Na, and Fe, Ti, Ga, etc. substituting for Si and Al, Götze 2012). The brightness difference may also be caused by defect density change (Watt et al. 2000; Rusk et al. 2006; Spear & Wark 2009).

GB83C: For the formation of fine plagioclase grains in GB83C, the change of K-feldspar morphology, and the change of CL brightness in plagioclase, three possible forms of formation of these fine grains are consistent with the parent grains. First, they could form from dislocation creep/dynamic recrystallization. Though the crystal orientations in the bridge zones align with the orientation of parent grains, the orientation distribution of fine plagioclase grains in GB83C is not consistent with the (010)[100] slip system (Montardi & Mainprice 1987; Rosenberg & Stünitz 2003; Satsukawa et al. 2013), which is considered to be a classical and typical pattern of plagioclase plastic deformation. Alternatively, the fine grains could form from microfracturing and cataclasis. The fine feldspar grains of these clasts may not have undergone much rotation. Or these fine grains may come from the solution-precipitation process, which also explains the core-rim texture shown in the fine-grained plagioclase CL of GB83C and their SPOs, which are weak but subparallel to bridge zone orientations (Ree et al. 2005; Fukuda et al. 2012). Plagioclase CL brightness variations are also possibly related to chemical composition. Brighter CL regions may mean richer Ca and poorer Ti. The major elements vary little but have a narrower distribution range within the brighter grains. Grain size reduction of plagioclase is often accompanied by chemical processes or chemical reactions (Hanmer 1982; Tullis 1983; Yund & Tullis 1991; Stünitz 1998), and such changes in composition during recrystallization have been observed (White 1980; Olsen & Kohlstedt 1985; Stünitz 1998; Johnson et al. 2008).

Brighter CL rims of coarse-grained plagioclase in the area adjacent to quartz may also be caused by the combination of chemical change and defect density change (Watt et al. 2000; Rusk et al. 2006; Spear & Wark 2009). The plagioclase in these areas began to lose some elements and accumulate on lamellae overgrowth or neocrystallization, which resulted in the enrichment of Ca and the loss of Na and K in the major elements. Brighter CL rims in plagioclase may be the starting of bridge zone formation before grain-size reduction occurs. After the initial grain development, they most likely underwent static
annealing to form these equant and polygonal grains with ~120° triple junctions (Passchier & Trouw 2005).

As with plagioclase, morphological changes in K-feldspar grains inside and outside the bridge zones can indicate the operation of chemical processes. Fine K-feldspar grains in bridge zones are shown as tabular interstitial grains or fine-grained K-feldspar aggregates enriched at the edges of bridge zones or in the embayment-shaped areas of quartz clusters, which are different from the exsolution of K-feldspar lamellae observed in the coarse plagioclase grains outside bridge zones. Though they share the same morphology, the coarser lamellae in plagioclase brighter CL rims are also different from the finer lamellae in terms of a very small changes in chemical composition. The fine K-feldspar grains in bridge zones preferentially occur in the triple junctions of plagioclase. The fine K-feldspar formed in the dilated zones between fine-grained plagioclase grain (Kruse & Stünitz 1999; Ree et al. 2005; Platt 2015). It appears that the K-feldspars originate by diffusion from antiperthite texture or fluid involvement, even under granulite-facies conditions. However, a fluid-mediated explanation may be limited by the slight quantitative difference in K-feldspar/plagioclase mode ratios inside and outside the bridge zone.

In summary, bridge zones in GB83C appear to have developed predominantly by chemical processes, with a minor contribution of mechanical changes. The chemical processes include migration or variation of major and trace chemical elements, solution-precipitation processes, and potential phase changes, where fluids are more likely to be involved. These chemical processes may play an important role in the formation of bridge zones. Together, these chemical-mechanical processes contribute to the formation of bridge zones.

GB247: We present a hypothesis for the formation of fine K-feldspar and the formation of new phases such as plagioclase and adaptation in GB247. The dislocation density at the edge of coarse-grained K-feldspar and the inheritance and dispersion of fine-grained K-feldspar CPO to parent grains may be the result of subgrain rotation (Menegon et al. 2008; Pennacchioni et al. 2010; Ceccato et al. 2017; Zhou et al. 2022). The fine plagioclase grains in the bridge zones are not in contact with the coarse plagioclase grains outside the zones, a relationship that requires mass transfer. This mass transfer could be catalyzed by
deformation. For example, in K-feldspar-rich deformed rocks, plagioclase in the form of myrmekite is thought to be generated in the high-stress region at grain boundaries during dynamic recrystallization by diffusion and/or reaction (e.g., Becke 1908; Hanneman & Anthony 1969; Lin & Yund 1972; Simpson 1985; Simpson & Wintsch 1989). In GB247, this displacement may begin with the enrichment of plagioclase between fine K-feldspar particles and gradually increase. In the process, the plagioclase gradually consumes K-feldspar, while the relic K-feldspar remains within the plagioclase preserving the blurred boundary. The volume change of phase transition leaves room for quartz precipitation. The replacement of K-feldspar by plagioclase (in the form of myrmekite) can be represented by the reaction equation with an aqueous solution (Becke 1908; Simpson & Wintsch 1989):

$$\text{Na}^+ + \text{Ca}^{2+} + \text{Kfs} \rightarrow \text{Pl} + \text{K}^+ + \text{SiO}_2$$

Na and K migration may be related to pressure distribution in the rock. Simpson (1985) indicates that the Na$^+$ ions have more potential to concentrate on the high-pressure margins of K-feldspar. Because of the ionic radius of Na$^+$, it could diffuse in the K-feldspar lattice. And an area of pressure increase is the site where the reaction is more favorable because there is a volume loss in the reaction from K-feldspar to plagioclase in which the "expansion area" would flavor the K$^+$-rich aqueous solution (Simpson & Wintsch 1989). The potential fracture systems in coarse feldspar provide channels for the migration of Na and K elements and other trace elements; they bring Na$^+$ into the high-stress region (bridge zones), carry K$^+$ out, enrich it into these regions. This process could lead to the observed low-Na/high-K inclusion bands.

In the fine-grained quartz in the bridge zones, the increase in the ratio of plagioclase to quartz relative to the overall ratio outside may be due to the reaction. These fine-grained plagioclase grains also inherit the lattice orientation of the parent K-feldspar grain.

Similar to GB83C, we suggest that chemical processes play a more important role in the formation of bridge zones in GB247 than mechanical processes. Our observations are consistent with some experimental results (de Ronde et al. 2005; Goncalves et al. 2015; Pec et al. 2016) and imply that the reaction products, phase mixing and concurrent mechanical processes such as grain size reduction or
grain size sensitive processes consume strong phases, eventually weakening the rock and accommodate strain.

Since the visible strain in the rock is small, we consider that the heterogeneity of the stress/pressure distribution caused by the heterogeneous distribution of the phases in the rock is responsible for this. Stress amplification (increase of stress magnitude at a point relative to the far field load) between rheological weak phases may support the bridge zone formation processes (e.g., Scholz 1968; Jordan 1987; Handy 1994; Kruse and Stüinitz 1999; Martelat et al. 1999; Stüinitz et al. 2001; Kenkmann and Dresen 2002; Johnson et al. 2004; Holyoke & Tullis. 2006). Amplification allows local stress concentrations to exceed the threshold required for mechanical or chemical action without changing loading (Feng et al., in review).

2.8.3. **Implications of Bridge Zone Formation**

We argue that bridge zones represent a rheological transition and interconnection, similar to the interconnection of weaker phases and associated weakening processes in polyphase aggregates documented by many researchers (e.g., Handy et al. 1990, 1994, 1999; FitzGerald and Stüinitz 1993; Wintsch et al. 1995; Kruse and Stüinitz 1999; Stüinitz et al. 2001; Handy and Stüinitz 2002; Kenkmann and Dresen 2002; Gueydan & Frédéric 2003; Vernon et al. 2004; Johnson et al. 2004, 2006; Wintsch & Yeh 2013; Oliot et al. 2014; Platt 2015; Smith et al. 2015; Czaplinska et al. 2015; Gerbi et al. 2014, 2016). Interconnection of the weak domain has long been recognized as a significant contributor to strain localization (Cobbold 1977; Jordan 1987; Handy 1990; Johnson et al. 2004; Holyoke & Tullis 2006; Montesi 2013). Arguably the dominant agent of rheologic change is the development of new weak domains in mechanically critical areas by processes such as recrystallization, fracture, introduction of fluids, phase changes, and reactions. Nucleation and interconnection of rheologically weaker phases in a rock or other polyphase composite can reduce the bulk strength of that composite significantly. Gerbi et al. (2014 & 2016) calculated that a ~1% modal change from strong to weak phase, in interconnected domains, can induce a nearly order-or-magnitude bulk strength reduction.
Despite the recognized importance of mechanical linking, the specific processes involved remain an area of study, being more involved than a simple phase reorganization. For example, the tips from weaker phases may produce microcracks, grain-size reduction, or chemical reactions and drive the formation of domains that link weaker phases (Handy 1994; Johnson et al. 2004; Holyoke & Tullis 2006; Svaahnberg & Piazolo 2010; Goncalves et al. 2015; Platt 2015; Gerbi et al. 2016; Hou et al. 2020) rather than simply connected weaker phases before deformation (Gerbi et al. 2016). Our conclusion is that these domains, which we label as bridge zones, incorporate both mechanical and chemical change. Mechanical change will nearly always be in the form of grain size reduction, and can also include crystallographic alignment. Chemical change, as observed in our samples, includes phase nucleation, modal or morphological alteration, and major and trace element mobility.

In low-strain rocks such as GB247 or GB83C, these fine-grained zones might not be sufficient to cause essential changes in the constitutive law of strong phases (e.g., from dislocation creep to grain-size sensitive creep). However, as strain increases (or strain rate increases), as a part of the strong phase, these existing fine-grained zones are more likely to undergo grain size-sensitive processes or interaction with other phases to help the strong phase deform to accommodate the overall strain. Additionally, if a chemical reaction brings phase change, the mechanical deformation mechanism of the dominant phase (dislocation or diffusion creep) accounts for only a portion of the deformation process, and a single mineral flow-law description is insufficient to quantify the strength change. In other words, once these connections form a network, the main loading domain in a polyphase system transition from the strong support framework domain to the new weak phase interconnection domain. In addition, with the evolution of rock strain and microstructure, the distribution relationship between strong and weak phases is constantly changing, and this change will inevitably lead to the redistribution of the deformation mechanisms that affect rock strength (Figure 2.13, Gueydan et al. 2014; Platt 2015; Gerbi et al. 2016).

Gerbi et al. (2016) also indicated that this kind of interconnection is independent of scale, ranging from microscale to meter-scale, kilometer-scale, and even tectonic scale (e.g., Tullis et al. 1991; Bons & Cox 1994; Handy et al. 1999; Ji 2004; Dayem et al. 2009; Klepeis & King 2009; Gerbi et al. 2012, 2014,
At the meter scale and larger, one example is a shear zone network, in which the existing shear zones (which can be regarded as weak domains) are interconnected by bridging shear zones and have a profound impact on the bulk strength of regional and even tectonic units with subsequent geometric evolution (e.g., Culshaw et al. 2011; Marsh et al. 2011; Kaatz et al. 2021; Zertani et al. 2022).

**Figure 2.13** Conceptual diagram of the rheological effect of bridge zone illustrating (a) how the development and condition of weak domains has the potential to reduce bulk strength more than shape change along, and (b) the potential appearance of new weak domain microstructure. After Gerbi (unpublished).

### 2.9. Chapter Summary

Strain localization occurs throughout the crust, in both the brittle and viscous regimes. Although details may differ by location, rock records indicate that variations of material properties (e.g., active
deformation mechanisms, crystallographic orientation, phase distribution, grain shapes, chemical variations, mass transfer, reactions, etc.) are likely to be the dominant factor for weakening. Our study focuses on the earliest stages of strain concentration in two rocks that experienced low macroscale strain at amphibolite facies conditions.

The bridge zones are relatively planar domains exhibiting grain-size reduction and distinct morphology compared to surrounding areas. The bridge zones appear to mechanically link rheological weak phases or domains and form networks in the rock. Bridge zones comprise phases and/or chemical compositions absent in the surrounding areas, which is consistent with a significant role for chemical processes in their formation.

The spatial distributions of bridge zones suggest that the small modal fraction of microstructural change represented by the bridge zones can lead to a high degree of bulk weakening with continued deformation, thus permitting bridge zones to serve as the precursors for large macroscale strain.
CHAPTER 3
NUMERICAL EXPLORATION OF FACTORS THAT CONTROL STRESS AMPLIFICATION
IN EARTH’S LITHOSPHERE

3.1. Chapter Abstract

The mechanical heterogeneity of Earth's lithosphere leads to significant amplification of stresses across spatial scales ranging from mineral grains to tectonic plates. These stress amplifications play a key role in mechanical and chemical processes within the rock that affect bulk rock strength. Identifying the most effective causes of stress amplification is critical for understanding processes such as strain localization and fluid transport at scales ranging from micro-shear zones to tectonic plate boundaries. However, studies quantifying and predicting stress heterogeneities and amplifications are limited. We used numerical modeling of two-phase isotropic viscous systems to explore the factors influencing and controlling stress amplification and the potential magnitude of stress amplification in viscous regimes. We found the most geologically relevant amplification factors to be weak-phase spacing, rheological contrast, and loading type. Our results indicate that stress amplification can reach a factor of ~9 under specific conditions, but most of our experiments suggest amplifications at or below a factor of 2. Pressure differences across the model domains generally do not exceed ~55 MPa, but some are as high as ~110 MPa. The stress and pressure amplifications resulting from our analyses are large enough to drive a variety of geologically important processes such as failure and strain localization, as well as transient permeability and fluid migration.

3.2. Chapter Introduction

Researchers have long been aware that Earth's lithosphere is mechanically heterogeneous (e.g., Barrell 1914), and this distribution of mechanical properties leads to a heterogeneous stress distribution (e.g., Griffith 1921; Voigt 1928; Reuss 1929). In turn, heterogeneous stress distribution drives numerous geological processes such as fluid flow, diffusion and reactions through pressure gradients (e.g., Kamb

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1961; Naus-Thijssen et al. 2011; Wheeler 2018), and damage through failure and grain size reduction (Figure 3.1, e.g., Segall & Simpson 1986; Johnson et al. 2004, 2008; Bercovici & Ricard 2014), all of which can lead to rheologic change (e.g., Poirier 1980; White 1980; Mancktelow 2002; Gerbi et al. 2016; Gardner et al. 2017).

Given the importance of microscale stress, several studies have developed analytical and numerical solutions for the stress field in a polyphase system (e.g., Eshelby 1957, 1959; Mandal et al. 2001; Johnson et al. 2004; Schmid 2005; Kenkmann & Dresen 1998; Xu & Zhang 2018; Beall et al. 2019), while others consider the whole crust or lithosphere (Mithen 1982; Höink et al. 2012; Reuber et al. 2016). Stress calculations around experimental and natural inclusions broadly agree with the modeling results (Dimanov et al. 2011; Rybacki et al. 2014; Döhmann et al. 2019). A separate class of studies considers how reactions, pressure change, phase transition, and thermal expansion can drive volume change, separate from the effects of boundary loading (Burnley & Davis 2004; Schmid et al. 2009; Kelemen & Hirth 2012; Johnson et al. 2021). A specific example is how the microscale variation in pressure and stress around a porphyroblast can control fluid movements and/or diffusion and cause "pressure shadows" (e.g., Kenkmann & Dresen 1998; McKenzie and Holness 2000; Passchier & Trouw 2005).

Several of the studies referenced above, along with many others, have explored general methods for calculating heterogeneous stress fields or applied those concepts to specific geologic and non-geologic problems. As such, our conceptual understanding of how heterogeneous stress distribution impacts important geological processes is relatively advanced (cf., Dahlen 1992; Barbe et al. 2001a & b; Ji et al. 2001, 2004; Vrijmoed et al. 2009; Griera et al. 2013; Moulas 2014; Dabrowski et al. 2015; Finch et al. 2020; Johnson et al. 2021). However, we do not have a similarly deep body of literature that records a systematic exploration of the magnitudes of stress heterogeneity and the controls thereof. This contribution is an attempt to fill at least part of that gap by seeking to identify typical magnitudes of stress heterogeneity in crustal settings, as well as the primary factors leading to those magnitudes of heterogeneity.
In this paper, we focus on the stress variations associated with weak inclusions in a strong matrix. The stress variations around such weak inclusions (Figure 3.1) may lead to a series of mechanical and chemical changes that mechanically connect the weaker phases by weakening the strong matrix, thus effectively reducing bulk strength (e.g., Handy 1994; Johnson et al. 2004; Wintsch et al. 1995; Wintsch & Yeh 2013; Fukuda & Okudaira 2013; Gerbi et al. 2016).

We use numerical modeling of two-phase systems to explore what factors lead to stress heterogeneity as well as the potential magnitude of stress amplification. We find that a typical amplification factor in geological materials can reach a factor of ~9 but is more generally in the range of 2, with weak-phase spacing, rheologic contrast, and load type being the most important factors.

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Figure 3.1 Conceptual diagram of phase distribution and its resulting effects. Weaker phases (biotite grains) in a stronger quartzofeldspathic matrix create stress concentrations resulting in damage that provides a rheological connection between the physically separate grains (Johnson et al. 2004). Such connections not only directly reduce the bulk rock strength (e.g., Handy 1994; Gerbi et al. 2016), but also provide pathways for fluid access and further petrological and rheological change.
3.3. Methodology

3.3.1. Numerical Methods

We perform the calculations described below using the Power-Law Creep (PLC) Toolbox, a MATLAB interface employing finite elements and periodic boundary conditions. The toolbox calculates bulk properties and the spatial distribution of stress and strain rate in isotropic power-law or Newtonian viscous materials (Cook et al. 2014). This approach uses a 2.5D geometry in which a 2D phase distribution is extruded into the third dimension. Details of the iterative solution, meshing scheme, and limitations are presented in Cook et al. (2014). Following the finite element calculation, we interpolate the data by converting from elements to pixels so that every pixel represents the same area, and we can perform statistical evaluations. Because all calculations iterate to a solution based on a stable geometry, they reflect an instantaneous set of parameters. These instantaneous solutions, instead of time-dependent (i.e., straining) solutions, are appropriate for the present study because our primary aim is to examine how stress distributions are impacted by idealized geometries.

Input for the PLC Toolbox includes a phase map, loading conditions, and the material constants of the standard power-law flow equation:

\[
\dot{\varepsilon} = A \sigma^n \exp \left( -\frac{Q}{RT} \right)
\]  

(1)

where \( \dot{\varepsilon} \) is strain rate, \( \sigma \) is stress, \( T \) is temperature, \( A, n, \) and \( Q \) are material constants in uniaxial form (Tullis et al. 1991), and \( R \) is the gas constant (Table 3.1). In the simple shear experiments described below, the flow law is converted from uniaxial to multiaxial form by multiplying the value of \( A \) by the term \( 1/2 \ 3^{n/((n+1)/2)} \) (Cook et al. 2014). Analysis conditions are any specified temperature and either strain-rate or stress boundary conditions. Output includes full stress and strain-rate tensors for the bulk geometry and each mesh element. The sign convention has tension and top-to-the right shear as positive, compression and top-to-the-left shear as negative. This numerical approach neglects viscous anisotropy and grain boundary effects, which may play significant roles in stress distributions in some settings (e.g., Finch et al. 2020).
Table 3.1 Rheological Parameters for numerical modeling

<table>
<thead>
<tr>
<th>Phase</th>
<th>Pre-exponential factor $A$ (MPa$^{-n}s^{-1}$)</th>
<th>$n$</th>
<th>Creep activation energy $Q$ (J/mol)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Material-F$^a$</td>
<td>4.52E+02</td>
<td>3</td>
<td>367800</td>
</tr>
<tr>
<td>Material-Q$^b$</td>
<td>1.80E-09</td>
<td>4</td>
<td>135000</td>
</tr>
<tr>
<td>Material-B$^c$</td>
<td>1.20E-30</td>
<td>18</td>
<td>51000</td>
</tr>
</tbody>
</table>

a. Based on Rybacki et al. 2006 with water fugacity and activation volume at 773 K and 600 MPa (water fugacity and activation volume from https://publish.uwo.ca/~awither5/fugacity/index.htm)

b. Based on Hirth et al. 2001 with water fugacity at 773 K and 600 MPa

c. Based on Kronenberg et al. 1990, which is specifically for biotite basal slip.

3.3.2. Experiment Sets

Except where noted, all experiment sets (Figure 3.2) use the same geometry as the reference (BG1, Figure 3.2a). This basic geometry is four square inclusions (weak phase) evenly distributed within a large square (strong phase). We run most experiment sets using flow laws that approximate biotite, quartz, and feldspar and under both simple shear and vertical axial compression conditions. We perform 5 sets of numerical experiments, varying geometry, rheological contrast, loading conditions, orientation, and spacing as detailed below. We run every experiment set at 773 °K, with a stress (2nd invariant) loading of 10 MPa. Except where noted, the modal abundance of weak-phase inclusions is controlled at ~16%±2%, in order to eliminate the effects of the fractions of phases on the bulk strength. All the geometry names are described in Table 3.2.
Table 3.2 Modeling geometry codes and detailed descriptions

<table>
<thead>
<tr>
<th>Code</th>
<th>Geometry (Set)</th>
<th>Detailed descriptiona</th>
</tr>
</thead>
<tbody>
<tr>
<td>BG1</td>
<td>The <strong>Basic Geometry</strong>/Reference model</td>
<td>Four square inclusions (weak phase) evenly distributed within a large square (strong phase)</td>
</tr>
<tr>
<td>SH1</td>
<td>Inclusions shape/circularity changes (Set #2)</td>
<td>Slightly rounded corners</td>
</tr>
<tr>
<td>SH2</td>
<td>Inclusions shape/circularity changes (Set #2)</td>
<td>More rounded corners</td>
</tr>
<tr>
<td>SH3</td>
<td>Inclusions shape/circularity changes (Set #2)</td>
<td>Circle</td>
</tr>
<tr>
<td>PO1</td>
<td>Inclusions position change (Set #3)</td>
<td>Leave the edges of BG1 unchanged and move the middle column vertically downward 0.5 times the length of the inclusion edge</td>
</tr>
<tr>
<td>PO2</td>
<td>Inclusions position change (Set #3)</td>
<td>Leave the edges of BG1 unchanged and move the middle column vertically downward 1.25 times the length of the inclusion edge</td>
</tr>
<tr>
<td>OR1</td>
<td>Inclusion orientation change (Set #4)</td>
<td>Rotate the inclusions 45 degrees to a diamond shape</td>
</tr>
<tr>
<td>SP1</td>
<td>Inclusion spacing distribution change (Set #5-1)</td>
<td>Add one column of inclusions</td>
</tr>
<tr>
<td>SP2</td>
<td>Inclusion spacing distribution change (Set #5-1)</td>
<td>Add two columns of inclusions</td>
</tr>
<tr>
<td>SP3</td>
<td>Inclusion spacing distribution change - more extreme cases (Set #5-2)</td>
<td>Rectangles with twice the length and half the height of the square</td>
</tr>
<tr>
<td>SP4</td>
<td>Inclusion spacing distribution change - more extreme cases (Set #5-2)</td>
<td>Horizontally divide the square of BG1 into 10 equal pieces</td>
</tr>
<tr>
<td>SP5</td>
<td>Inclusion spacing distribution change - more extreme cases (Set #5-2)</td>
<td>Horizontally divide the rectangle of SP3 into 10 equal pieces</td>
</tr>
<tr>
<td>SS1</td>
<td>With <strong>simple shear strain</strong></td>
<td>The end product from BG1 of sinistral simple shear with a shear strain $\gamma=1$</td>
</tr>
<tr>
<td>PS1</td>
<td>With <strong>vertical flattening strain</strong></td>
<td>The end product of vertical flattening with vertical shortening with a horizontal linear strain $\varepsilon=1$</td>
</tr>
</tbody>
</table>

a. All changes are made on the premise that the stronger and weaker phase modal remains basically constant
Figure 3.2 Experiment sets showing phase distributions. The reference geometry (BG1) consists of weak-phase squares (dark blue) evenly distributed within a large square of strong-phase material (light blue). Experiment Set 2 (SH1-3): The inclusions remain in the same locations, but the corners progressively round to form circles in SH3. Experiment Set 3 (PO1-2): The center column of inclusions is vertically shifted 0.5 (PO1) and 1.25 (PO2) times the length of the inclusion. Experiment Set 4 (OR1): The square inclusions are rotated 45 degrees, which reduces the horizontal and longitudinal distance between inclusions. Experiment Set 5-1 (SP1-2): The modal abundance of inclusions is held constant by reducing the area of the individual squares and adding more squares horizontally. This results in increased vertical spacing and decreased horizontal spacing. Experiment Set 5-2 (SP3-5): The aspect ratios of the inclusions while maintaining a constant modal abundance.
Figure 3.2 (Continued) In SP3, the squares of BG1 are split into 10 pieces and evenly distributed horizontally. In SP4, the inclusions have a 4:1 aspect ratio, and in SP5, the inclusions of SP4 are split into 10 and evenly distributed horizontally. For the selection of Phase1 and Phase2, the types of material pairs, and their rheological parameters, please refer to Table 3.1.

3.3.2.1. Set #1 Rheological Contrast

We evaluate the effects of rheological contrast by comparing results of mineral pairs with material constants approximating those of biotite, quartz and plagioclase (e.g., Kronenberg et al. 1990; Hirth et al. 2001; Rybacki et al. 2006), which we label Material B, Material Q and Material F respectively (Table 3.1). We use three material pairs: B-Q, B-F, and Q-F (Table 3.3). In all pairs, the material before the dash refers to the inclusion, while the material after the dash refers to the matrix.

Table 3.3 Material pairs and loading conditions

<table>
<thead>
<tr>
<th>Material pairs(^a)</th>
<th>Weak inclusions</th>
<th>Strong matrix</th>
<th>Loading stress (2nd invariant, MPa)</th>
<th>Temperature</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Simple shear(^b)</td>
<td>Vertical axial compression</td>
</tr>
<tr>
<td>Q-F</td>
<td>Material-Q</td>
<td>Material-F</td>
<td>10 MPa</td>
<td>773 °K</td>
</tr>
<tr>
<td>B-F</td>
<td>Material-B</td>
<td>Material-F</td>
<td>10 MPa</td>
<td></td>
</tr>
<tr>
<td>B-Q</td>
<td>Material-B</td>
<td>Material-Q</td>
<td>10 MPa</td>
<td></td>
</tr>
</tbody>
</table>

\(^a\) Constituent parameters used in numerical experiments were chosen to approximate those of quartz (Q), biotite (B) and plagioclase feldspar (F).
\(^b\) Direction of simple shear: top to the left
3.3.2.1. Set #2 Shape

In this experiment set, we keep the inclusion positions the same as in BG1 and change the roundness of the corners, from slightly rounded corners (SH1, Figure 3.2b) to more rounded corners (SH2, Figure 3.2c). This change leads to a slight decrease of weak-phase mode. Finally, we change the geometry to a circle (SH3, Figure 3.2d); for this geometry, we adjust the area of the circle to match that of the square, thereby reducing the relative distance between the weak phases slightly.

3.3.2.2. Set #3 Position

To evaluate the effect of vertically offsetting the inclusions, we leave the edges of BG1 unchanged and move the middle column vertically downward 0.5 (PO1, Figure 3.2e) and 1.25 (PO2, Figure 3.2f) times the length of the inclusion edge.
3.3.2.4. **Set #4 Orientation**

To evaluate the case where inclusions are not edge-on to each other, we rotate the inclusions 45 degrees to a diamond shape (OR1, Figure 3.2g), which also reduces the horizontal and vertical separation.

3.3.2.5. **Set #5-1 & Set #5-2 Spacing**

In Set #5-1, we keep the mode in BG1 constant while reducing the area of the individual squares and adding more squares horizontally. This results in increased vertical spacing and decreased horizontal spacing. In this experiment set, we added one column (SP1, Figure 3.2h) and two columns (SP2, Figure 3.2i) of squares. With the addition of new columns, the area of a single weak square changes to 2/3 (SP1) and 1/2 (SP2) of the single weak square area in BG1. The horizontal gap between weak phases reduces from 1.5 times the length of the original inclusion to ~0.85 (SP1) and ~0.53 (SP2). Vertically, the length of the gap increases from 1.5 to about 1.68 (SP1) and 1.79 (SP2).

In Set #5-2, we evaluate a different inclusion aspect ratio with extreme scenarios. For SP3 (Figure 3.2k), rather than a square, we test a rectangle that is twice the length and half the height of the square. As a result, the horizontal gap is half the original small square length, and in the vertical direction the gap is twice the original length. To further evaluate spacing, we horizontally divide the square and rectangle of BG1 and SP3 into 10 equal pieces (SP4, Figure 3.2j, and SP5, Figure 3.2l, respectively).

3.3.3. **Data Analysis**

For each experiment set, we focus our discussion on pressure and/or the square root of the 2nd invariant of stress, which we refer to hereafter as the 2nd invariant for simplicity. Appendix describe the differential stress, which behaves similarly to that of the 2nd invariant.

In addition to maps displaying stress fields (Figures 3.3a and d), we calculate a single scalar value for the 2nd invariant (Figure 3.3b) and pressure. For the 2nd invariant, we report the stress value above which only 1% of pixel values lie (see the star in Figure 3.3c). Choosing a 1% threshold eliminates any single-pixel stress anomalies while also ensuring that the values measured represent geologically meaningful concentrations. For comparison purposes, we normalize the pixel-level stress values by the
macroscale load stress, producing an amplification factor (i.e., microscale value divided by macroscale value).

For pressure, a similar amplification factor is not relevant, so instead, we calculate the pressure variation over the system (Figure 3.3e). Like the 2nd invariant, we determine a maximum and minimum pressure concentration where 1% of the pixels are above and below the calculated values and report the difference between those threshold values (Figure 3.3f). Although we recognize some dependence of viscous flow on pressure through water fugacity and activation volume (or enthalpy), we make the simplifying assumption for this study that viscous flow is independent of pressure. To that end, we run our experiments with no confining pressure. This results in some field values being positive and some negative. Confining pressure, which is negative in our sign convention, would shift all values linearly.

3.4. Results

We report all results in Table 3.4 and describe them by experiment set below.
<table>
<thead>
<tr>
<th>EXP SET #</th>
<th>Geom.</th>
<th>Loadi ng</th>
<th>Materia ls (Weak-Strong)</th>
<th>Modal fraction strong phase</th>
<th>2nd invariant</th>
<th>Pressure</th>
<th>differ ential stress</th>
<th>Vertic al:later al spacin g</th>
<th>Surface area change</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>BG1</td>
<td>SS</td>
<td>Q-F</td>
<td>84%</td>
<td>1.94</td>
<td>0%</td>
<td>-27.63</td>
<td>27.88</td>
<td>55.50</td>
</tr>
<tr>
<td>1</td>
<td>BG1</td>
<td>SS</td>
<td>B-F</td>
<td>84%</td>
<td>1.56</td>
<td>-19%</td>
<td>-16.88</td>
<td>17.13</td>
<td>34.00</td>
</tr>
<tr>
<td>1</td>
<td>BG1</td>
<td>SS</td>
<td>B-Q</td>
<td>84%</td>
<td>1.14</td>
<td>-41%</td>
<td>-4.88</td>
<td>4.88</td>
<td>9.75</td>
</tr>
<tr>
<td>1</td>
<td>BG1</td>
<td>VAC</td>
<td>Q-F</td>
<td>84%</td>
<td>1.49</td>
<td>0%</td>
<td>-15.63</td>
<td>4.88</td>
<td>20.50</td>
</tr>
<tr>
<td>1</td>
<td>BG1</td>
<td>VAC</td>
<td>B-F</td>
<td>84%</td>
<td>1.29</td>
<td>-13%</td>
<td>-11.63</td>
<td>0.63</td>
<td>12.25</td>
</tr>
<tr>
<td>1</td>
<td>BG1</td>
<td>VAC</td>
<td>B-Q</td>
<td>84%</td>
<td>1.09</td>
<td>-27%</td>
<td>-7.38</td>
<td>-3.88</td>
<td>3.50</td>
</tr>
<tr>
<td>2</td>
<td>SH1</td>
<td>SS</td>
<td>Q-F</td>
<td>84%</td>
<td>1.91</td>
<td>-1%</td>
<td>-27.38</td>
<td>27.38</td>
<td>54.75</td>
</tr>
<tr>
<td>2</td>
<td>SH1</td>
<td>SS</td>
<td>B-F</td>
<td>84%</td>
<td>1.54</td>
<td>-2%</td>
<td>-16.88</td>
<td>16.88</td>
<td>33.75</td>
</tr>
<tr>
<td>2</td>
<td>SH1</td>
<td>SS</td>
<td>B-Q</td>
<td>84%</td>
<td>1.14</td>
<td>0%</td>
<td>-4.88</td>
<td>4.88</td>
<td>9.75</td>
</tr>
<tr>
<td>2</td>
<td>SH1</td>
<td>VAC</td>
<td>Q-F</td>
<td>84%</td>
<td>1.46</td>
<td>-2%</td>
<td>-15.88</td>
<td>5.13</td>
<td>21.00</td>
</tr>
<tr>
<td>2</td>
<td>SH1</td>
<td>VAC</td>
<td>B-F</td>
<td>84%</td>
<td>1.29</td>
<td>0%</td>
<td>-11.63</td>
<td>0.63</td>
<td>12.25</td>
</tr>
<tr>
<td>2</td>
<td>SH1</td>
<td>VAC</td>
<td>B-Q</td>
<td>84%</td>
<td>1.09</td>
<td>0%</td>
<td>-7.63</td>
<td>-3.88</td>
<td>3.75</td>
</tr>
<tr>
<td>2</td>
<td>SH2</td>
<td>SS</td>
<td>Q-F</td>
<td>85%</td>
<td>1.89</td>
<td>-3%</td>
<td>-23.13</td>
<td>23.13</td>
<td>46.25</td>
</tr>
<tr>
<td>2</td>
<td>SH2</td>
<td>SS</td>
<td>B-F</td>
<td>85%</td>
<td>1.54</td>
<td>-2%</td>
<td>-14.88</td>
<td>14.88</td>
<td>29.75</td>
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a. 2nd invariant change from reference: in BG1, these changes are compared with that of Q-F under the same loading condition, $a=X-X(QF)/X(QF)$; the rest are compared with BG1; values for the same loading BG1; values for the same loading conditions and the same

b. Pressure difference: the difference between low pressure and high pressure: $b=P_{\text{low pressure}}-P_{\text{high pressure}}$

c. Pressure difference from reference: Method is the same as 2nd invariant change from reference column, except that the contrast value is pressure difference

d. vertical/lateral: The ratio of the vertical and lateral distances of adjacent weak phases in a geometric

e. Surface area change: the change of the total length of all weak phase perimeters in several graphs relative to the total length of BG1 weak phase, $c=P-P(BG1)/P(BG1)$

f. SS: Simple Shear

g. VAC: Vertical Axial Compression
3.4.1. 2nd Invariant of Stress

3.4.1.1. Reference Model (BG1):

Under simple shear (Figures 3.4a, b and c), the maximum stress occurs in the strong phase at the sharp corners of the weak inclusions. Small regions of low stress occur in the strong phase at the straight boundaries of the inclusions, but these transition to regions of high stress between faces of the adjacent inclusions. The primary regions of low stress occur in the matrix diagonally between the inclusions.

Under vertical axial compression (Figures 3.4d, e and f), the maximum stress still occurs in the strong phase at the sharp corners of the weak inclusions. These regions of maximum stress are connected diagonally between the sharp corners of the inclusions, forming an X-shaped pattern of maximum stress.

![Figure 3.4](image)

**Figure 3.4** 2nd invariant stress field in geometry BG1 with different rheological pairs under sinistral simple shear (a-c) and vertical axial compression (d-f). The weaker inclusions outlined with dashed lines in (a) lie in the same locations in (b) through (f). A reduction in stress concentration is observed as mineral pairs change from Q-F to B-F to B-Q. Under simple shear, the high stress regions connect two adjacent inclusions in the lateral and vertical directions. Under vertical axial compression, high stress bands connect the inclusions diagonally, and an X-shaped high stress domain can be seen in the area surrounded by four adjacent inclusions.
3.4.1.2. Shape/Circularity Variations (Set #2)

In all geometries of Set #2, the maximum stresses and stress gradients occur at the phase boundaries. However, there are differences between the patterns of simple shear (Figures 3.5b, c, and d) and vertical axial compression (Figures 3.6b, c, and d).

Under simple shear (Figures 3.5b, c and d), the stress distribution is very similar to that of BG1. As the sharp corners gradually smooth to a circular shape in Figure 3.5d, the high stresses at the sharp corners become arcuate regions with increasing radius of curvature. The high-stress region between the inclusion faces gradually narrows (Figures 3.7a and b).

Under vertical axial compression (Figures 3.6b, c, and d), high stress narrow regions directly adjacent to the inclusions shift to vertical/lateral positions with increasing circularity, and the larger regions of maximum stress still form an X-shaped diagonally between inclusions, becoming more diffuse with increased circularity. The minimum stresses form an elliptical region in the vertical/lateral positions between inclusions. New high stress points appear at the junction of the arc and the straight line and finally concentrate at vertical/lateral sides of the circle (Figures 3.7e and f).

Under simple shear, shape changes result in slightly lower amplifications than those in BG1 (~1.90 for Q-F, ~1.55 for B-F and ~1.14 for B-Q). Under vertical axial compression, the maximum amplifications of all geometries are unchanged or relatively lower than those in BG1 (~1.46 for Q-F, ~1.27 for B-F and ~1.07 for B-Q).
Figure 3.5 2nd invariant stress distribution maps of the Q-F system under sinistral simple shear (a-l) and stress amplification factors for each geometry of each set (m-q). See Figure 3.2 for phase distributions used in maps (a-l). See Figure 3.3 for a graphical explanation of the calculation of amplification factor. Line colors (m-q) indicate the column in which the geometry is located. In general, the stress distribution with respect to the shape (SH) and position (PO) is consistent, with minor distortions compared to BG1. Changes in orientation (OR) and spacing (SP) show significantly higher stress amplification (hot colors in (g-l)). The lines in (m-q) indicate a gradual decreasing trend from Q-F to B-F to B-Q for all geometries. In (o), (p), and (q), the offset of each line with respect to BG1 (blue) is larger than that in (m) and (n). Note the changes in the ordinate range in (m)-(q). Amplification factors lie between 1 and 8.66.
Figure 3.6 2nd invariant stress distribution maps of the Q-F system under vertical axial compression (a-l) and stress amplification factors for each geometry of each set (m-q). See Figure 3.2 for phase distributions used in maps (a-l). See Figure 3.3 for a graphical explanation of the calculation of amplification factor. Line colors (m-q) indicate the column in which the geometry is located. The stress distribution with respect to the positions of the inclusions is consistent, with minor changes compared to BG1. The amplification factor is lower than under simple shear. Amplification factors lie between 1 and 1.6.
Figure 3.7 Detailed 2nd invariant stress distribution maps of BG1, SH3, OR1 and SP5 under sinistral simple shear (a, b, c and d) and vertical axial compression (e, f, g and h). Note changes in total range of stress in legend bars. Black circles mark the location of the maximum stress amplification. The arrows point to high stress bands linking two adjacent weak inclusions. In general, the maximum stress amplification occurs in the strong phase where the phase-boundary curvature is sharpest. Curvature affects the stress amplification patterns but does not change the magnitude significantly. In contrast, spacing does affect the stress amplification magnitude significantly.

3.4.1.3. Position Variations (Set #3):

Under simple shear (Figures 3.5 e and f), the patterns of high stress distribution are still dominated by the inclusion faces, maintaining the symmetrical pattern of BG1. In contrast, the patterns of low stress distribution change markedly, becoming distorted on PO1 and then becoming symmetrical in PO2 forming a bow-tie pattern laterally around the inclusions. Overall, these low-stress regions form a zipper-shaped pattern. Under simple shear, the amplifications within each of the three pairs in this experiment set are similar (~1.96 for Q-F, ~1.59 for B-F and ~1.16 for B-Q for both PO1 and PO2), which are only slightly higher than those of BG1.

Under vertical axial compression (Figures 3.6 e and f), the patterns of maximum stress distribution are still dominated by the inclusion corners, but progressively shift from the symmetrical X-shaped pattern of BG1 to an asymmetrical X-shape in PO1 and loss of the X-shape in PO2. Areas of low stress still occur in the vertical/lateral positions between inclusions, but the magnitudes of both the high and low stresses reduce with inclusion offset. Under vertical axial compression, the amplification factors of PO1 and PO2 for B-Q are the same (1.06) as that in BG1. For Q-F and B-F, amplification factors are
all lower than those of BG1, albeit only slightly. Amplification factors of PO1 are always higher than those of PO2.

3.4.1.4. **Orientation Variation (Set #4):**

The patterns of stress with the rotated, diamond-shaped inclusion are similar to those in BG1, with high (low) stress domains forming a symmetrical cross linking the inclusions in simple (pure) shear, and low (high) stress domains centered between the inclusions (Figures 3.5g and 3.6g). Although the pattern is similar, the high (low) stress concentration areas of OR1 are narrower (Figures 3.7a vs. c, e vs. g).

Compared to BG1, OR1 has all higher amplification factors in Q-F (2.16), B-F (1.71) and B-Q (1.21) in simple shear, while in vertical axial compression, OR1 also has similar though slightly larger factors in Q-F (1.54) and B-F (1.31), and the same factor for B-Q (1.09).

3.4.1.5. **Spacing Variations (Set #5-1 & Set #5-2):**

Set #5-1 (Figures 3.5h, i and 3.6h, i):

The patterns of stress in geometries with different spacing are similar to those in BG1. However, the relative magnitudes are different: for example, in the vertical high stress domains between inclusions, the degree of stress amplification is smaller than that in the horizontal direction under simple shear (Figures 3.5h and i). Under vertical axial compression (Figures 3.6h and i), the inclusion is connected to lower stress domains, and the stress amplifications in lateral domains are significantly lower than those in vertical domains. High-stress connections occur in regions surrounded by four inclusions. In SP1 (Figure 3.6h), the connection has a distinct X-shape, whereas in SP2 (Figure 3.6i), the stress is more concentrated near the inclusions' sharp corners.

Under simple shear, all amplification factors of SP2 are higher than in BG1 and SP1. The factors of SP1 are also higher than those of BG1. Under vertical axial compression, the factors of B-Q (1.06 for SP1 and SP2) are the same as that of BG1. B-F (1.24 for SP1 and 1.21 for SP2) and Q-F (1.39 for SP1 and 1.36 for SP2) factors are all lower than those of BG1.

Set #5-2 (Figures 3.5j, k, l and 3.6j, k, l):
Under simple shear, high-stress domains occur in SP3 laterally between the closely-spaced rectangular inclusions (Figure 3.5k) with maximum stresses at the sharp corners. Low-stress regions occur on the top and bottom faces on inclusions and nearly circular areas surrounded by four inclusions. In SP4 (Figure 3.5j) and SP5 (Figures 3.5l and 3.7d), high stresses surround the tips of the inclusions, whereas low stresses occur on their lateral faces and a broad tabular zone vertically between the inclusion trains. Most of the strong-phase domain has relatively low stress values (~12 MPa).

In the case of vertical axial compression, in SP3 (Figure 3.6k) low stresses occur in the lateral direction between inclusions, with some high stress concentrations above and below these low stress regions. SP4 (Figure 3.6j) and SP5 (Figures 3.6l and 3.7h), show the same characteristics. The stress concentration in most of the strong phase is also ~12 MPa.

The stress amplification factors in Q-F and B-F pairs are 5.56 and 3.71 (SP3), 7.74 and 4.84 (SP4), and 8.66 and 7.56 (SP5), which are all much higher than those of BG1. Though the factors of B-F (1.59 for SP3, 1.36 for SP4 and 1.34 for SP5) are still higher than those of BG1, the increase is not comparable to that of the other two mineral pairs. In the case of vertical axial compression, all amplifications in SP3, SP4 and SP5 are not significantly different than, but most are all lower than, BG1 (~1.30 for Q-F, ~1.17 for B-F, and ~1.06 for B-Q).

3.4.2. Pressure

Because this is an incompressible system, we do not incorporate confining pressure in the model calculations. As a result, the pressure output spans negative and positive values. It is important to reiterate that in our sign convention, negative pressure represents relative compression and positive pressure represents relative tension. In the text that follows, to maintain consistency with geological usage, we refer to “relative compressive pressure” or “high pressure” as the negative values, and “relative tensile pressure” or “low pressure” as the positive values.

All regions of relative compressive pressure and relative tensile pressure concentrate in the strong phase adjacent to the weaker inclusions. Relative compressive and tensile pressure under simple shear under all geometries exhibit similar magnitudes, though opposite signs. Under vertical axial compression,
the magnitude of relative tensile pressure is less than the magnitude of relative compressive pressure (Table 3.4).

3.4.2.1. Reference Model & Rheological Contrast (BG1 & Set #1, ):

Under simple shear, all relative compressive- and relative tensile-pressure domains in BG1 lie in the strong phase at the sharp corners of the weak inclusions (Figures 3.8a, b and c). The areas of relative tensile pressure are on the upper right and lower left of the inclusion, while relative compressive-pressure areas are on the upper left and lower right, which contrasts with the more commonly studied strong inclusion/weak matrix system (e.g., Kenkmann & Dresen 1998; Johnson et al. 2009). The pressures inside each weak inclusion are approximately 0 MPa, and the pressure variation concentrated in the strong phase in contact with the inclusion. The pressure distribution pattern has diagonal symmetry through inclusion centers. The difference between relative tensile- and relative compressive-pressure in the different material pairs shows the same trend as the 2nd invariant amplification factor: Q-F (55.5 MPa) > B-F (34 MPa) > B-Q (9.8 MPa).

Under vertical axial compression, relative tensile pressure regions lie above and below the inclusions, with relative compressive pressure to the sides (Figures 3.8e, f and g). This pattern is opposite to what is in a strong inclusion/weak matrix system. The pressure distributions inside each inclusion are ~5 to ~10 MPa, and the trend of differences is consistent with other experiments (Figures 3.9 and 3.10): Q-F (20.5 MPa) > B-F (12.3 MPa) > B-Q (3.5 MPa).
Figure 3.8 Pressure field of BG1 with (a) Q-F, (b) B-F, and (c) B-Q under sinistral simple shear and (d) Q-F, (e) B-F, and (f) B-Q under vertical axial compression. Pressure concentrations are reduced as mineral pairs change from Q-F to B-F to B-Q. Under simple shear, the upper left and lower right of the weak inclusions are compression domains, and the upper right and lower left are tensile domains. Under vertical axial compression, the left and right sides of an inclusion are compression domains and the top and bottom are tensile domains.
Figure 3.9 Pressure distribution maps of Q-F system under sinistral simple shear (a-l) and pressure differences for each geometry of each set (m-q). Although modified by each specific geometry, the general pattern characteristics are similar in all experiments. The magnitude of compressive and tensile pressure domains varies modestly in some experiments [(m), (n), and (p)] and more significantly in others [(o) and (q)]. See Figure 3.2 for phase distributions used for maps in (a-l). See Figure 3.3 for a graphical explanation of the calculation of pressure difference. Line colors (m-q) indicate the column in which the geometry is located.
Figure 3.10 Pressure distribution maps of Q-F system under vertical axial compression (a-l) and pressure differences for each geometry of each set (m-q). Although modified by each specific geometry, the general pattern characteristics are similar in all experiments. The magnitude of compressive and tensile pressure domains varies modestly in all experiments. See Figure 3.2 for phase distributions used for maps in (a-l). See Figure 3.3 for a graphical explanation of the calculation of the pressure difference. Line colors (m-q) indicate the column in which the geometry is located.

3.4.2.2. Shape/Circularity Variations (Set #2):

Under simple shear, as the sharp corners gradually become round, the patterns of relative compressive and relative tensile pressure become more diffuse and the difference between relative compressive and relative tensile pressure decreases (Figures 3.9b, c and d). All the differences in SH1 (54.8 MPa for Q-F, 33.8 MPa for B-F and 9.8 MPa for B-Q) are similar to those in BG1, but there is an obvious decrease in SH2 and SH3, especially in Q-F (46.3 MPa of SH2 and 40.8 MPa of SH3), followed by B-F (29.8 MPa for SH2 and 26 MPa for SH3). Values for B-Q are basically unchanged (9.3 MPa for SH2 and 8.3 MPa for SH3).
Under vertical axial compression, pressure distribution patterns remain similar to BG1 but, unlike simple shear, the pressure magnitudes increase (Figures 3.10b, c and d). The differences of SH3 are biggest for different pairs (26 MPa for Q-F, 15.5 MPa for B-F, and 4.5 MPa for B-Q). SH1 (21 MPa for Q-F, 12.3 MPa for B-F, and 3.8 MPa for B-Q) still shares similar values with BG1. Values of SH2 (22 MPa for Q-F, 13.3 MPa for B-F, and 4 MPa for B-Q) lie between those of SH1 and SH3. Q-F still generates the largest pressure variation, while B-Q generates the smallest.

3.4.2.3. Position Variations (Set #3):

The spatial distribution of simple shear (Figures 3.9e and f) and vertical axial compression (Figures 3.10e and f) is similar to that of BG1, except that the patterns are slightly distorted with the movement of the middle column. Due to position changes, the relative tensile-pressure area and relative compressive-pressure area of each inclusion are connected with adjacent areas with the same attribute. However, patterns of PO1 are more complex than those in PO2.

Under simple shear, the difference of relative compressive- and relative tensile-pressure values decrease gradually with the increase of the downward displacement of the middle column (Q-F: 51.3 MPa for PO1 and 46 MPa for PO2; B-F: 31.5 MPa for PO1 and 28.3 MPa for PO2; and B-Q: 9 MPa for PO1 and 8.3 MPa for PO2). Under vertical axial compression, the differences in relative compressive- and relative tensile-pressure values for BG1 and the two different positions are almost identical (Q-F: 21 MPa for PO1 and 21.8 MPa for PO2; B-F: 12.5 MPa for PO1 and 13 MPa for PO2; and B-Q: 3.8 MPa for PO1 and PO2).

3.4.2.4. Orientation Variation (Set #4):

Under simple shear, the regions of relative compressive and relative tensile pressure of OR1 are concentrated on the four edges (Figure 3.9g), but the patterns are the same as BG1. OR1 has lower magnitude differences: 31.8 MPa for Q-F, 20 MPa for B-F and 6.3 MPa for B-Q

Under vertical axial compression, the relative tensile-pressure areas are on the upper and lower corners, and the relative compressive-pressure areas are on the corners at the two sides (Figure 3.10g), in
an arrangement similar to that of BG1. The pressure differences follow the same pattern as BG1, with uniformly higher values: 31.3 MPa for Q-F, 17.8 MPa for B-F and 5 MPa for B-Q.

3.4.2.5. Spacing Variations (Set #5-1 & Set #5-2):

Regardless of the spatial arrangement of the weak inclusions and the loading style, all distributions of pressure are similar to that of BG1. In other words, under simple shear, the relative tensile-pressure areas lie at the upper right and lower left of each inclusion and the relative compressive-pressure areas lie at the upper left and lower right (Figures 3.9h-l). Under vertical axial compression, the relative tensile-pressure areas are on the top and bottom of the inclusion, and the relative compressive-pressure areas are on the lateral sides (Figures 3.10h-l).

Set #5-1 (Figures 3.9h, i and 3.10h, i):

Under simple shear, all pressure differences of SP2 (69.5 MPa for Q-F, 42.3 MPa for B-F and 10.8 MPa for B-Q) are higher than those of SP1 (61 MPa for Q-F, 37.3 MPa for B-F and 10.3 MPa for B-Q), which are higher than those of BG1. Vertical axial compression is the opposite, and the results of BG1 are generally greater than SP1 (18.8 MPa for Q-F, 11.5 MPa for B-F and 3.5 MPa for B-Q) and SP2 (17.8 MPa for Q-F, 10.8 MPa for B-F and 3.5 MPa for B-Q).

Set #5-2 (Figures 3.9j, k, l and 3.10j, k, l):

Under simple shear, the pressure differences of SP3, SP4 and SP5 in Q-F (75.5 MPa for SP3, 108 MPa for SP4 and 66.5 MPa for SP5) and B-F (53.5 MPa for SP3, 95.3 MPa for SP4 and 64.3 MPa for SP5) are much bigger than those of BG1. The values of B-Q are similar (11.5 MPa for SP3, 18.3 MPa for SP4 and 11.3 MPa for SP5). In contrast, the values under vertical axial compression in Q-F (14.5 MPa for SP3, 12.3 MPa for SP4 and 17.8 MPa for SP5) and B-F (9.8 MPa for SP3, 7.5 MPa for SP4 and 11.5 MPa for SP5) are all smaller than BG1, while the values of B-Q are still similar (3.3 MPa for SP3, 2.5 MPa for SP4 and 4 MPa for SP5).
3.5. Discussion

3.5.1. The Factors Controlling Stress Amplification

Our investigation focuses on stress distribution in a polyphase viscous system. The factors most influencing the magnitude of stress heterogeneity are rheological contrast, phase orientation and phase spacing, with their importance varying among models. In general, if there is a rheological contrast within the system, we can expect to see these stress amplification effects within the strong phase (e.g., Ji & Wang 1999; Johnson et al. 2004; Madi et al. 2005; Bystricky et al. 2006; Holyoke & Tullis 2006; Gerbi et al. 2014, 2016; Xu & Zhang 2018, etc.), particularly at the boundary with the weak phase. The sharper the phase boundary curvature the more focused the stress concentration (e.g., Shulman 2016). Materials with a dense distribution of isolated weak phases, and therefore small distances between weak phases, are more likely to have large stress concentrations. We did not evaluate the effect of viscous anisotropy but recognize that it may play a role in stress distribution similar to the effect of elastic anisotropy (Johnson et al. 2021), perhaps comparable to that of rheological contrast (e.g., Muto et al. 2011; Tielke et al. 2016).

Although the tested features that control stress heterogeneity are not entirely independent, some exhibit characteristics that differ markedly from others. As we explain in more detail below, the spatial distribution of weak phases (Set #4 and Set #5), loading type, and rheological contrast have the largest effect on stress magnitudes and patterns. Shape (Set #2) and position (Set #3) affect patterns, but do not appear to significantly influence the magnitude.

3.5.1.1. Loading

Different loading types - simple shear and vertical axial compression - have different effects on the same geometry. Model results of simple shear show regions of high stress around and connecting the inclusions, while a low stress domain is formed in the region surrounded by four inclusions (Figures 3.7a-d). Under vertical axial compression, inclusions are typically connected with adjacent inclusions in the diagonal direction through high stress domains, and high stress domains are formed in the regions wrapped by four adjacent inclusions (Figures 3.7e-h). For a given geometry, the 2nd invariant amplification factor under simple shear is always larger than that under vertical axial compression, with
B-Q the least sensitive rheologic pair to changes in load. Taking BG1 as an example, the stress amplification factors under vertical axial compression condition are lower than those of simple shear by 23% in Q-F, 17% in B-F and only 4% in B-Q. For SP5, which exhibits the largest heterogeneity, the vertical axial compression amplification factor in Q-F is 85% lower than that of simple shear, B-F decreases by 85%, and B-Q decreases by 21% (Table 3.4, Figure 3.12a).

The variation in pressure due to loading geometry is even more pronounced than for 2nd invariant patterns. Under simple shear conditions, tensile pressure domains surround the inclusions on the upper left and lower right and compressive pressure domains on the lower left and upper right (Figures 3.11a-d). Under vertical axial compression, the upper and lower regions of the inclusions are tensile pressure domains, and the left and right regions are compressive pressure domains (Figures 3.11e-h).

As with the 2nd invariant, simple shear produces higher magnitudes of pressure variation across the model system than does vertical axial compression. Again, taking BG1 as an example, the pressure difference of Q-F under vertical axial compression is lower than that of simple shear by 63%; B-F decreases by 64%; and B-Q decreases by 64%. In SP4, which has the largest pressure difference under simple shear, Q-F under vertical axial compression is lower by 89%; B-F decreases by 92% and B-Q decreases by 86%. Under simple shear, the absolute value of the maximum tensile pressure and maximum compressive pressure of a given geometry are typically similar. Under vertical axial compression, the magnitude of the maximum compressive pressure is typically larger than that of the maximum tensile pressure. Moreover, the increase or decrease of the pressure difference relative to BG1 is opposite under simple shear and vertical axial compression (Table 3.4, Figure 3.12b).
Figure 3.11 Detailed pressure distribution map of BG1, SH3, OR1 and SP4 under sinistral simple shear (a, b, c and d) and vertical axial compression (e, f, g and h). Under the same shear type, the basic pressure distribution pattern around each inclusion remains unchanged. The effect of phase-boundary curvature on pressure distribution is minor compared to its effect on the 2nd invariant of stress (Figure 3.7).

Figure 3.12 Compilation of amplification factors (a) and pressure differences (b) of selected geometries under simple shear and vertical axial compression. Line colors represent geometry, solid lines represent simple shear (SS), and dotted lines represent vertical axial compression (VAC). In general, the amplifications of stress and differences in pressure are greater in simple shear than in vertical axial compression. See text for discussion.
3.5.1.2. Shape

The effect of shape on stress heterogeneity varied with the examined stress component. Patterns in the 2nd invariant were relatively constant across geometries (see annotations in Figure 3.7). Similarly, magnitudes were similar, with amplification factors within 4% of BG1 (Table 3.3, Figure 3.12a). The effect of sharp corners on pressure distribution is small. Under vertical axial compression, there is even no "pressure concentration" near the sharp corners (Figure 3.11e). In contrast, the magnitude of pressure difference caused by shape change can be greater than 25% (Table 3.4, Figure 3.12b). As the inclusions are more rounded from BG1 to SH1, SH2, and SH3, the pressure difference decreases under simple shear and increases under vertical axial compression.

3.5.1.3. Orientation

Orientation change affects stress similarly to shape change. With respect to BG1, a change in orientation does not cause a change in the overall stress/pressure distributions, despite the sharp corner effects. Most 2nd invariant stress amplification factors of OR1 are higher than those of BG1, except for B-Q, which remains constant under vertical axial compression. Under simple shear, the factor of Q-F increased by 12%, B-F by 10%, and B-Q by 7%. The increase of the maximum value in vertical axial compression is not as high as that in simple shear, with Q-F increasing by 3% and B-F by 2% (Table 3.4, Figure 3.12a). The interconnection bands are narrower, but the relative position of high stress and low stress is the same (arrows in Figures 3.7a vs. c and e vs. g). The change in local stress is caused by the change of the position of the sharp corners (circles in Figures 3.7a vs. c and e vs. g). The change in pressure difference is also significant. Under simple shear, the values of OR1 are smaller than those of BG1. Q-F, B-F and B-Q decreased by 43%, 41% and 36%, respectively. These pressure differences increase under vertical axial compression: Q-F by 52%, B-F by 45% and B-Q by 43% (Table 3.4, Figure 3.12b).

3.5.1.4. Rheological Contrast

Different material pairs represent different degrees of rheological contrast. The stress amplification factors of the same geometry with different material pairs under the same loading condition
are different, and the sensitivities of different material pairs to other features discussed above are also different. Our results indicate that rheological contrast does not affect the stress distribution markedly, but it does affect the magnitude of stress amplification and pressure differences.

Regardless of geometry and loading conditions, the amplification factors of the same geometry are always $Q-F > B-F > B-Q$. Taking BG1 as an example, under simple shear, the amplification factors of $B-F$ and $B-Q$ are 19% and 41% lower than that of $Q-F$, and 13% and 27% lower, respectively, under vertical axial compression (Table 3.4, Figure 3.12a).

Regardless of geometry, the pressure difference magnitudes are also $Q-F > B-F > B-Q$. Taking BG1 as an example, under simple shear, compared with $Q-F$, the pressure difference of $B-F$ is 39% lower, while that of $B-Q$ is 82% lower. Under vertical axial compression, those values are 40% and 83% (Table 3.4, Figure 3.12b).

Defining rheological contrast involves subtle complexities, particularly the interplay of different components of the constitutive relationship. We explored whether effective viscosity is a sufficient metric to predict amplification magnitude (see Appendix). We found that the values of $A$ and $n$ from Equation 1 independently influence stress amplification. As such, we find it difficult to generalize at this point about the relationship between rheological parameters and stress amplification or pressure variation.

### 3.5.2. The Magnitude of Stress Amplification

Stress heterogeneity in rocks, due to rheological contrasts and mechanical anisotropy, is ubiquitous. Broadly speaking, geologic systems appear capable of stress amplification or reduction of up to a factor of 14 (Beall et al. 2018), with a lower amount relative to bulk conditions being more typical in both viscous and elastic regimes (Kenkmann & Dresen 1998; Gerbi et al. 2014, 2016; Johnson et al. 2004, 2021; Xu & Zhang 2018). Our experiments produce 2nd invariant stress with a maximum stress amplification factor of 9, though, with the exception of the spacing and orientation changes, calculated factors remain below 2 under simple shear and below 1.5 under vertical axial compression (Figure 3.13).

As mentioned in the previous section, under simple shear, the range of amplification factors of $B-Q$ is relatively small. Even considering the largest spacing change, these factors do not exceed 1.6, and
most of them are concentrated in the range of 1.1-1.3. The minimum stress amplification factor of B-F is 1.5 but reaches 7.6 in SP5 with spacing change. The Q-F pair in geometry SP5 showed the largest stress amplification factor in this experiment, 8.7. All amplification factors in Set #5 and Set #4 exceed 2 and all other amplification factors for Q-F are between 1.9 and 2.0. Under vertical axial compression, all amplification factors of B-Q are limited to a small range between 1.06-1.09. The amplification factor of B-F is no more than 1.32, but most of them are above 1.18 except SP5 (1.16). The range of amplification factors of Q-F is wider than that of the other two material pairs, but generally the factor is not more than 1.5, but above ~1.3.

The theoretical upper boundary of bulk strength is isostrain-rate bound, in which case the strong phase bears more load and the stress is enhanced in order to maintain the same strain rate in both phases. Under this condition, and for a 10 MPa load, the strong phase experiences stress amplifications of 18% (Q-F), 12% (B-F), and 4% (B-Q) relative to the macroscale load. The stress amplifications from our numerical results are much larger than these values, indicating the importance of factoring in the spatial distribution of weak phases.

In our experiments, pressure differences are generally not more than ~55 MPa, but some reach as high as ~110 MPa (Figure 3.14; Table 3.4). Pressure differences under vertical axial compression are typically half or less than those under simple shear, with B-Q displaying the lowest values (up to 18 MPa) and Q-F the highest (up to 108 MPa). Experiment sets varying shape and position yield pressure differences comparable to that of the reference geometry (BG1), whereas orientation values are ~40% lower in simple shear and similarly higher in vertical axial compression. Spacing has a large span of effect on pressure differences. All spacing simple-shear experiments have pressure differences larger than 10 MPa, with several over 60 MPa. In many vertical axial compression experiments, the pressure differences of spacing variation are less than 12 MPa.
Figure 3.13. Range of amplification factors for each experiment set under simple shear (a) and vertical axial compression (b). See Figure 3.3 for a graphical explanation of the calculation of amplification factor. Different material pairs are distinguished by different colors: blue is Q-F, red is B-F, and gray is B-Q. The description of the experimental set corresponds to the text as follows: Shape (Set #2), Position (Set #3), Orientation (Set #4), Spacing-1 (Set #5-1) and Spacing-2 (Set #5-2). Note that weak phase spacing is a strongly dominant factor in simple shear, and that amplifications in simple shear are larger than those in vertical axial compression.
Figure 3.14 Range of pressure difference values for each Experiment Set under simple shear (a) and vertical axial compression (b). See Figure 3.3 for a graphical explanation of the calculation of amplification factor. Different material pairs are distinguished by different colors: blue is Q-F, red is B-F, and gray is B-Q. The description of the experimental set corresponds to the text as follows: Shape (Set #2), Position (Set #3), Orientation (Set #4), Spacing-1 (Set #5-1) and Spacing-2 (Set #5-2). Note that weak phase spacing is a dominant factor in simple shear, and that differences in simple shear are larger than those in vertical axial compression.
3.5.3. Effect of Strain

Since the PLC toolbox calculates the instantaneous state of internal stress, strain rate, effective viscosity, and other properties of the geometry during loading, our data intentionally do not include any time-dependent factors, including strain. However, if we were to forward calculate the strain rates, area reduction in the visible plane is possible. For the stress loading, the boundary condition in the third dimension is the same as in the horizontal dimension. Accurately incorporating strain into our analysis would necessitate iterative adjustments of model node positions according to the calculated velocities and recalculation of the velocity field, steps which are not currently available. However, to coarsely evaluate the effect of strain on stress amplification and related metrics, we evaluated an experiment set as the deformed “end product”. SS1 is defined as the end product from BG1 of sinistral simple shear with a shear strain \( \gamma = 1 \) (Figure 3.15b), while PS1 is the final product of vertical flattening with a horizontal linear strain \( \varepsilon = 1 \) (50% vertical flattening, Figure 3.15g). These end products are produced with an isostrain geometry and are therefore highly idealized.

Under simple shear, the sharp corner effect is still evident, with high stress still connecting adjacent inclusions in both horizontal and vertical directions. The low-pressure area is altered by shear, with sinistral asymmetry (Figure 3.15f). The simple shear strain results in 5% less amplification for Q-F (~1.83), the same amplification for B-F (~1.56) and 4% more for B-Q (~1.19) relative to BG1.

Under vertical axial compression, the stress distribution is relatively constant, with minor distortion produced by strain. The maximum amplifications for all material pairs of Q-F, B-F and B-Q are 10%, 8% and 2% less relative to BG1, respectively (~1.34 for Q-F, ~1.19 for B-F and ~1.06 for B-Q).

The pressure distribution did not change markedly under either simple shear (Figure 3.15k, l & o) or vertical axial compression/flattening (Figure 3.15m, n & p), but the pressure difference value decreased. Under simple shear, the pressure difference decreased by about 45% compared with BG1 (28.8 MPa for Q-F, 18.5 MPa for B-F and 5.5 MPa for B-Q). Under vertical axial compression/flattening, the
pressure difference decreased by 24%, 25% and 7%, respectively, for Q-F, B-F and B-Q (15.5 MPa for Q-F, 9.5 MPa for B-F and 3.3 MPa for B-Q).

These calculations indicate that moderate, idealized strain has a relatively minor effect on stress amplification. Though most of the amplification factors decreased, the effect of both vertical flattening and simple shear strain is less than 10% relative to the reference geometry. For pressure, all values decrease moderately, with simple shear having a greater effect than vertical axial compression/flattening.
Figure 3.15. Comparison of stress distributions from reference model BG1 under sinistral simple shear (a, c, e) and vertical flattening (i, k, m) versus those from deformed by simple shear (b, d, f) and vertical flattening (j, l, n). Under simple shear (d), the 2nd invariant high stress regions still connect adjacent inclusions in both horizontal and vertical directions, but low stress regions are markedly asymmetric. Patterns of high pressure in the simple shear geometry (f) are markedly realigned to more horizontal structures. Under vertical flattening, the 2nd invariant stress and pressure distributions are relatively unchanged, though some distortion is produced by strain (l, n). Most amplification factors (g, h) and all pressure differences (o, p) decreased relative to the reference geometry.
3.5.4. The Impact of Stress Heterogeneities on Geological Processes

Based on our results from the idealized geometries described above, we performed similar calculations for two simplified "natural rock" microstructural geometries, SJ12 and SJ18 (Figure 3.16, Figure 3.17). These geometries are simplified from samples with different finite strains from the San Jose tonalite, located within the Peninsula Ranges batholith of Baja California (Johnson et al. 2004; Gerbi et al. 2014). This system is modeled as a B-F material pair under 10 MPa under sinistral simple shear (Figure 3.16) and vertical axial compression (Figure 3.17) at 773 °K. The maximum stress amplification factors for both samples are ~2.1 under simple shear (2.10 for SJ12 and 2.06 for SJ18), and ~1.7 under vertical axial compression (1.64 for SJ12 and 1.79 for SJ18) under vertical axial compression. Thus, an amplification factor of 2 appears reasonable for natural settings. The pressure differences in SJ12 and SJ18 are 36.3 MPa and 35.3 MPa for simple shear and are 22 MPa and 25.8 MPa for vertical axial compression respectively (Figure 3.16, Figure 3.17, Table 3.4). Amplification factors in the range presented herein indicate that processes controlled by stress magnitude can occur when the macroscale loading stress is well below a given threshold. This implies that there is a high possibility that numerous geological processes at all scales may not require high bulk loading to initiate.

Returning to our primary concern of stress/pressure variation causing rock weakening, the breakdown or decomposition mechanisms and processes of the strong load-bearing framework to form the interconnected weak phases play an important role in the reduction of rock bulk strength. The local microscale stress amplification appears in strong framework minerals more than sufficient to push local values across thresholds for, e.g., grain size reduction, recrystallization, fracture, or chemical reaction.
Several studies have recognized fine-grained fracture and/or recrystallized zones that bridge rheological weak domains at sites of stress amplification and/or mirror step-over structures in brittle systems (Scholz 1968; Jordan 1987; Johnson et al. 2004; Holyoke & Tullis. 2006; Gerbi et al. 2016), which could be recognized as the consequences of stress amplification and reduction in the rock record in microscale. Rheological interconnection is of course expected as a viscous system evolves with strain (e.g., Handy 1994); we have explored here the potential for mechanical - rather than phase – interconnection, via the stress field. Either mechanical or phase interconnection will reduce bulk strength markedly, leading to potentially significant rheologic change and localization.

Alternatively, the weak-in-strong scenario can represent the introduction of fluids into the dry continental crust and the subsequent shear zone formation and widening of the shear zone. In order for the shear zone to widen, an adequate fluid supply is required (Kaatz et al. 2021). Stress amplification along the phase boundaries may lead to more microfractures, thus allowing fluid-rock interactions to gradually widen the shear zone. Stress amplification due to changes in spacing (e.g., SP4 & SP5) may contribute significantly to the above process, without adding additional fluid.

The modeling results illustrate the stress deviation (i.e., amplification and reduction) from the mean values and the pressure gradients. These deviations and gradients can play a vital role in mass transfer processes (e.g., Wheeler 1992; Zhu et al. 2001; Johnson et al. 2004; Vernon et al. 2004; Naus-Thijssen et al. 2010; Zertani et al. 2022). Several studies have addressed the potential for microscale pressure variation to drive metamorphic reactions outside the stability field defined by the bulk conditions (Dahlen 1992; Dabrowski et al. 2015; Hobbs & Ord 2016; Powell et al. 2018; Wheeler 2018). Because metamorphic reactions are commonly limited by fluid availability (e.g., Yardley 2009; Jamtveit et al.
the presence of microcracks or other dilatational sites can enhance fluid transport and therefore catalyze reactions, and/or even widen the shear zone.

Stepping back to look at the lithosphere through a very coarse lens, we see that much of the mechanical behavior derives from stress gradients or limits. Earthquakes, landslides, and volcanic eruptions represent failure events, as does even grain-scale erosion, where a particle is discontinuously removed from its host (e.g., Koons et al. 2013). For several other significant processes, such as mantle convection, regional metamorphism, crustal shearing, magma transport, and the formation of eclogite, the motive force is commonly a stress-related gradient, such as pressure variation (e.g., Vrijmoed et al. 2009, Reuber et al. 2016, Yamato & Brun 2017). Moreover, initiation of many of those processes requires a threshold-crossing event, such as microstructural damage and bulk weakening leading to shear zone formation or fracturing enhancing permeability in metastable metamorphic rock (e.g., Poirier 1980; Simpson & Wintsch 1989; Chu et al. 2017; Bellas et al. 2018; Nardini et al. 2018; Moore et al. 2019; Campbell et al. 2020; Orlandini & Mahan 2020). Therefore, quantifying stress heterogeneity allows better explanation of current processes and prediction of future events.
Figure 3.16 Calculated stress distributions in natural microstructures under sinistral simple shear. Images of phase distribution in tonalites (a) SJ12 and (e) SJ18 used as the basis for the models in in Johnson et al. (2004) and Gerbi et al. (2014). Black represents biotite and minor amphibole, and white represents feldspar and minor quartz. (b) and (f) show simplified and smoothed geometries used for calculation of 2nd invariant stress and pressure distribution. In the numerical experiment, the white matrix in the two geometries is defined as Material F, while the blue inclusions are defined as Material B. Both loading conditions are sinistral simple shear of 10 MPa at 773 °K. (c) and (g) 2nd invariant stress distributions display similar patterns as the idealized geometries. The weak inclusions have very low stresses, while the surrounding strong matrix hosts high stresses. High-stress regions connect adjacent weak inclusions and form a network within the system. (d) and (h) Pressure is more variable than in the idealized geometries.
Figure 3.17 Calculated stress distributions in natural microstructures under vertical axial compression. (a) SJ12 and (e) SJ18, simplified and smoothed geometries (b) and (f) and the material properties the same as Figure 3.16. The loading conditions vertical axial compression of 10 MPa at 773 °K. (c) and (g) 2nd invariant stress distributions display similar patterns as the idealized geometries. The weak inclusions have very low stresses, while the surrounding strong matrix hosts high stresses. High-stress regions connect adjacent weak inclusions and form a network within the system. (d) and (h) Pressure is also more variable than in the idealized geometries.

3.6. Chapter Conclusion

Using a numerical approach, we evaluate factors that affect and control stress amplification in weak-in-strong two-phase systems and quantify how much local stress is amplified relative to macroscale loading. Our results show that the maximum stress amplification occurs in the strong phase where the phase-boundary curvature is sharpest and that high stress bands connect weak phases. Orientation, spacing, shear type and rheological contrast have the greatest influence on stress amplification and pressure gradient. We calculate stress amplification up to a factor of ~9, but most of our results lie at or under 2. Our calculated pressure differences are generally not more than ~55 MPa but some reach as high as ~110 MPa. Our results highlight the value in quantifying stress heterogeneity and considering its role in triggering rheologic and metamorphic change at a variety of scales.
CHAPTER 4

RHEOLOGICAL BRIDGE ZONES: WEAKENING THROUGHOUT THE CRUST & DEFORMATION GRADIENT

4.1. Chapter Abstract

In Chapter 2, the observation of less deformed rocks of granulite and amphibolite facies from the Grenville Orogeny, Canada shows that the bridge zone is a relatively platy region composed of fine grains that mechanically connects rheological weak domains. Chemical processes play an important role in the formation of the bridge zone, and there are phases and/or chemical components in the bridge zone that are not present in the surrounding area. This microstructure may play an important role in rock weakening. Chapter 2 identifies presence of bridge zones and hypothesises about their effect on rheology, while Chapter 3 shows instantaneous stress amplifications in a rheologically heterogeneous material, however, we cannot in detail assess the evolution from singular bridge zones to throughgoing zones of weakness. This chapter aims to bridge the shortcomings of the previous chapters, by a) going more natural candidates in lower temperatures (greenschist facies) and b) looking at evolution, focusing on tonalites through strain gradient from the Cerro de Costilla complex, Mexico. The observation shows that similar mechanical and chemical processes can promote the development of bridge zones in the greenschist facies and at different strain, indicating that bridge zones can be formed throughout the crust. Numerical modeling of the simplified geometries of tonalites microstructures shows that a small modal addition (~2%) of the bridge zone results in bulk strength reduction by 25%-94%. The modeling results also show that the stress amplification occurs at the phase boundary and at the strong phase connecting the weak phase. Stress amplification can increase the macroscopic load stress by more than three times, which can
lead to the rheological changes and can also drive microscale metamorphic changes of the rock, thus promoting the formation of bridge belt.

4.2. Chapter Introduction

The mechanisms of rock weakening have been a long-standing concern in structural geology. While changes in external environmental conditions such as temperature, pressure, and stress can affect rock strength, how the microstructure evolves can also play a crucial role in rock strength changes. Components of microstructure include the mode of strong and weak phases and phase morphology, both of which affect the degree of rheological interconnection. Such connections include not only weak phases themselves, but, as described in the previous chapter, are commonly associated with breakdown of the strong supporting framework phases, phase mixing, and metamorphic reactions. Several studies have proposed that the spatial and temporal distribution of active deformation mechanisms are caused by a combination of mechanical processes, chemical processes, and the coupling of these two effects (e.g., Behrmann and Mainprice 1987; Wheeler 1992; Fitzgerald & Stünitz 1993; Wintsch et al. 1995; Kruse & Stünitz 1999; Martelat et al. 1999; Stünitz et al. 2001; Kenkmann & Dresen 2002; Johnson et al. 2004; Jessell et al. 2005; Johnson et al. 2008; Park et al. 2006; Warren & Hirth 2006; Marsh et al. 2009; Smith et al. 2015; Czaplinska et al. 2015; Zhou et al. 2022).

Chapter 2 documents the characteristics of a structure we term “bridge zone” in two low-strain rocks of Grenville and the influence of the bridge zones on bulk strength. Here we extend that analysis to include a different rock type, formed at lower pressure and temperature conditions, and the effect of strain. The new samples exhibit varying degrees of deformation at the edges of the Cerro de Costilla complex in northern Baja California, Mexico. These mylonized tonalites were well documented by
Johnson et al. (2008) and constrained by hornblende -- plagioclase thermometry to have deformed at 475 ± 50°C.

4.3. Geological Setting

The Cerro de Costilla complex is located in the Northern Sierra San Pedro Mártir region of northern Baja California, Mexico, and intruded into the Mesozoic metasediments in the hanging wall of the Main Mártir thrust fault (Figure 4.1a, e.g., Gastil et al. 1975; Johnson et al. 1999, 2003, 2008; Melis 2006; Alsleben et al. 2014). The Cerro de Costilla complex is a magmatic ring complex (Johnson et al. 2002; Melis 2006) with the following composition from core to margins: a coarse-grained gabbro core, a tonalite partial inner ring, a partial ring of deformed and partially melted metamorphic and igneous rocks, and a complete tonalite outer ring (Johnson et al. 2002, 2008). The outer tonalite formed at a pressure of 3-4 kbar (Melis 2006) at 103 ± 1.0 Ma (Johnson et al. 1999, 2003). The narrow margin of the outer ring of tonalite shows a strain gradient from an igneous texture through mylonite. Johnson et al. (2008) sampled and reported in detail the deformation gradients (Figure 4.1b&c), deformation characteristics evaluation, and mineralogical composition variation, with a basis of plagioclase, quartz, hornblende, and biotite. The movement of the Main Martir Thrust caused widespread north-northeast-trending foliation in the region (Johnson et al. 1999). However, the deformation we are concerned with here originates from post-thrust extension (Simpson & Schmid 1983). The extensional structure follows the foliations parallel to the thrust-related texture, but shows a north-side-down movement. These steeply dipping structures are in the hanging wall rocks, south and southeast of the Cerro de Costilla complex, and cut into the southern margin in the tonalite outer ring.
Figure 4.1 Geological map and sampling locations of the sheared tonalites from the Cerro de Costilla complex. (a) Generalized geological map showing terranes in the southwestern United States and Mexico (modified from Campa & Coney 1983; Keppie 2004; Centeno-Garcia et al. 2008; Dickinson 2008). The red star marks the location of the Cerro de Costilla complex. (b) Map showing the southern portion of the Cerro de Costilla complex and surrounding area (modified from Johnson et al. 2004; Melis 2006; and Johnson et al. 2008). The mylonitic shear zone of this study (red line) deforms the edge of the outer tonalite ring. (c) A 28.3-m transect through the fabric gradient, perpendicular to the shear zone. Letters C, D, and H correspond to locations of oriented samples referred to in the text.
4.4. Methods

4.4.1. Cathodoluminescence (CL)

We collected panchromatic CL using the Tescan panchromatic (350-650 nm) CL detector attached on the Tescan Vega XMU II scanning electron microscope (SEM) from the School of Earth and Climate Sciences, University of Maine with chamber pressure less than $10^{-2}$ Pa and 20 kV acceleration voltage. We collected color CL images using the Tescan Vega III scanning electron microscope (SEM) with the attached Tescan Rainbow CL detector at Colby College. The chamber pressure remained less than $10^{-2}$ Pa with a 20 kV acceleration voltage.

4.4.2. Electron Backscatter Diffraction (EBSD)

All EBSD data were collected at 20 kV acceleration voltage, ~6 nA beam current, 2.5 μm step size, and 70° sample tilt under high-vacuum conditions from the Tescan Vega XMU II SEM equipped EDAX-TSL EBSD system and EDAX-TSL OIM Data Collection 5.31 software at the University of Maine. All samples were polished with 0.3 μm alumina suspension and a 0.02 μm colloidal silica suspension to eliminate surface damage before the thin carbon coating to avoid electron charging.

Post-processing of the data employed the EDAX-TSL OIM Analysis 6.1 software. Non-indexed and low-indexed pixels (CI<0.1) are replaced with well-indexed adjacent pixels (CI>0.1). The grains in the post-processing possess an internal orientation deviation of 10° and a minimum grain size of 4 pixels, and the effects of twinning are subsequently cleaned. Crystallographic preferred orientation (CPO) pole figures, grain size distributions, grain reference orientation deviation, and kernel average misorientation...
parameters were derived from the cleaned data. In this study, grains larger than 10 microns are generally considered valid data, and grains smaller than 10 microns are not included in the statistics.

Grain reference orientation deviation and kernel average misorientation are the common parameters to evaluate the plastic strain of each point (Unnikrishnan et al. 2016; Gussev & Leonard 2019). The grain reference orientation deviation is the deviation between the orientation of the grain at a certain position in the grain and the reference orientation or average orientation of the whole grain. Local misorientation values ranging from 0°-10° are shown on the Grain reference orientation deviation maps. Kernel average misorientation is calculated and averaged between the kernel center data point and the surrounding data points in the kernel by custom setting (the second neighbor of hexagonal pixels kernel).

4.4.3. **Electron Probe Microanalysis (EPMA)**

Electron probe microanalysis (EPMA) on the Cameca SX-100 was applied at The University of Maine, School of Earth and Climate Sciences. Analytical conditions included 15 kV accelerating voltage, 10 nA beam current and a 5 μm spot size with a 300 ms counting time for the elements (Na, K, Ca, Mg, Si, Al, Fe, Mn, and Ba) in the feldspars and elements (Na, K, Mg, Ca, Mn, Fe, Al, Cr, Si and Ti) in the biotites.

4.4.4. **Numerical Modeling**

The Power-Law Creep Toolbox (PLC; Cook et al. 2014; Gerbi et al. 2014, 2016) is used to calculate the stress field and strain-rate around rheological heterogeneities under temperature and macroscale stress geometries and magnitudes relevant to the samples. For calculating stress amplification factors and identifying whether a relationship exists between stress amplification and observed structures, we made calculations at 773 K, with a stress (2nd invariant) loading of 10 MPa under both shear stress
and vertical compression. To calculate bulk strength reductions after the formation of ~2% modal amount of bridge zone, we use the above conditions and a second load stress of 50 MPa. The geometric patterns used for the numerical modeling were simplified from the microstructure of three rock samples. The rheological parameters for plagioclase, quartz and biotite were derived at 773 K, 600 MPa pressure, and saturated water conditions. Since the rheological data of hornblende are very scarce, we treat hornblende as plagioclase for the present study (Table 4.1).

Table 4.1 Rheological Parameters for numerical modeling

<table>
<thead>
<tr>
<th>Phase</th>
<th>Pre-exponential factor $A$ (MPa$^n$s$^{-1}$)</th>
<th>$n$</th>
<th>Creep activation energy $Q$ (J/mol)</th>
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</thead>
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<tr>
<td>Feldspar &amp; Hornblende $^a$</td>
<td>4.52E+02</td>
<td>3</td>
<td>367800</td>
</tr>
<tr>
<td>Quartz &amp; Bridge zone $^b$</td>
<td>1.80E-09</td>
<td>4</td>
<td>135000</td>
</tr>
<tr>
<td>Biotite $^c$</td>
<td>1.20E-30</td>
<td>18</td>
<td>51000</td>
</tr>
</tbody>
</table>

a. Based on Rybacki et al. 2006 with water fugacity and activation volume at 773 K and 600 Mpa (water fugacity and activation volume from https://publish.uwo.ca/~awither5/fugacity/index.htm)
b. Based on Hirth et al. 2001 with water fugacity at 773 K and 600 MPA
c. Based on Kronenberg et al. 1990

4.5. Observations from Natural Rock Samples

4.5.1. Sample selection and preparation

We used samples from the Johnson et al. 2008, those were sampled a 28.3 m transect from the tonalite boundary with the wall rock orthogneiss, where structural gradients and mineralogical features are most variable, toward the interior of the tonalite where the rock shows no macroscopic deformation
(Figure 4.1b, Johnson et al. 2004). In the subsequent text we focus on three samples: H, the least sheared sample that contains some folia cutting the magmatic texture (Figure 4.2a); D, a moderately deformed rock showing more continuous folia (Figure 4.2d); and C, a strongly deformed rock (Figure 4.2g). Each thin section is cut perpendicular to the foliation and parallel to the lineation. All samples contain fine-grained domains, or bridge zones, throughout the areas viewed. Table 4.2 summarizes the main features of each sample, organized by mineral.

**Table 4.2 Microstructure evolution throughout the gradients in the Cerro de Costilla complex.** (Modified from Johnson et al. 2008)

<table>
<thead>
<tr>
<th></th>
<th>Least deformed tonalite (protolith)</th>
<th>Discrete foliae</th>
<th>More continuous foliae</th>
<th>Strongly deformed rocks</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Plagioclase (Pl)</strong></td>
<td>1. oscillatory zoning</td>
<td>1. fine-grained aggregates within domains that link biotite-rich areas</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>2. scattered, round qz + hbl inclusions</td>
<td>2. primary recrystallization + ‘sintering’ of small fragments→locally small new marginal grains and aggregates of grains;</td>
<td>3. fragments→internal fractures + grain boundaries: local minute subgrains + new grains (Temperature too low to dominate);</td>
<td></td>
</tr>
<tr>
<td></td>
<td>3. Growth twins are commonly bent, with local lenticular deformation twins</td>
<td>4. fractured and pulled apart region: undeformed/recrystallized qz + beard qz, bt + hbl + pl fragments</td>
<td>4. fractured and pulled apart region: undeformed/recrystallized qz + beard qz, bt + hbl + pl fragments</td>
<td></td>
</tr>
<tr>
<td>Table 4.2 Continued</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>---------------------</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Hornblende (Hbl)</strong></td>
<td>1. inclusions of qz + biotite + pl; 2. brown-green subhedral grains and aggregates; 3. simple or multiple twinning; 4. zoned, with brown-green cores and blue-green rims.</td>
<td>1. bt replace hbl: <strong>hbl relics</strong> into biotite-rich foliae + (uncommon) <strong>neck down fragments</strong> into foliation; 2. fracturing→slightly misoriented ‘subgrains’; 3. apart hbl fragments: beards of blue-green hbl + qz + bt;</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Quartz (Qz)</strong></td>
<td>1. subgrains parallel to the c-axis 2. subgrain rotation recrystallization + some grain boundary migration</td>
<td>1. forms incipient foliae (with Bt ± Ttn + Apt); 2. recrystallization increases with increasing fabric intensity; 3. large recrystallized grainsize; 4. slightly elongated</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 4.2 Continued

| Biotite (Bt) | 1. greenish brown, subhedral; 2. ‘decussate’ aggregates locally pass into incipient foliae (some deformation and recrystallization) 3. fracturing → locally bent + slightly misoriented → ragged ‘subgrains’ 4. adjacent to pl: replaced by fine-grained, symplectic aggregates of ttn and pl, the main contributor to foliation initiation 2. small biotite grains: grow into fractures → separate pl; 3. biotite aggregates: cutting through + replacing hbl; 4. necking down to form incipient foliae: lenticular biotite grains and aggregates splintering 5. sub-parallel splinters at bends in biotite grains; 6. bending into an alignment by subgrains and recrystallized aggregates; 7. fragment + replaced by biotite carry bt-rich foliae + small hbl away from hbl; 8. initiation by fracturing (grain boundaries + locally cutting through grains): thin veinlets of very fine-grained bt + qz + others 9. beard of bt + qz 10. foliae cut biotite splays or fans; |

| 1. main contributor to foliation initiation 2. small biotite grains: grow into fractures → separate pl; 3. biotite aggregates: cutting through + replacing hbl; 4. necking down to form incipient foliae: lenticular biotite grains and aggregates splintering 5. sub-parallel splinters at bends in biotite grains; 6. bending into an alignment by subgrains and recrystallized aggregates; 7. fragment + replaced by biotite carry bt-rich foliae + small hbl away from hbl; 8. initiation by fracturing (grain boundaries + locally cutting through grains): thin veinlets of very fine-grained bt + qz + others 9. beard of bt + qz 10. foliae cut biotite splays or fans; |

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### Table 4.2 Continued

<table>
<thead>
<tr>
<th><strong>K-feldspar (Kfs)</strong></th>
<th>small interstitial grains marginally replaced by fine-grained myrmekite</th>
<th>Myrmekite→fine pl + qz: contribute to the initiation of foliae.</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>others</strong></td>
<td>Titanite (Ttn): scattered large grains + small grains, aligned aggregates and veinlets in bt. Tourmaline (Tur) + Apatite (Ap): rare pink to blue-black Allanite (Aln): Brown</td>
<td>Ttn: minor contribution to the initiation of foliae: fragments + from bt(ttn + pl);</td>
<td></td>
</tr>
</tbody>
</table>
| **General**          | weak magmatic foliation by biotite aggregates and aligned pl grains | **S-C type foliation pattern**: new biotite-rich foliae follow grain boundaries of strong minerals (pl + hbl) | 1. S-C  
2. bt foliae: well aligned, locally ‘decussate’  
3. continuous foliae: recrystallized biotite ± quartz, with or without irregularly distributed ttn, pl, epidote and hbl  
1. continuous aggregates rich in biotite,  
2. elongate aggregates of recrystallized quartz  
3. between clasts: qz + bt beards |
| **Geochemistry**     | uniform tonalitic bulk composition |  |  |
| **Bulk**             |  |  | Lower in Na₂O, MgO, CaO, Fe₂O₃, Al₂O₃  
Higher: SiO₂, K₂O, LOI |  |
| Pl: | coarser: XCa=0.37  
finer: XCa=0.30 | $\text{Bt}_1 + \text{Pl}_1 \rightarrow \text{Ttn} + \text{Pl}_2 + \text{Bt}_2$ (release Ca)  
recrystallization  
myrmekite Pl: more Ca  
replaced Pl: less Ca | coarser: More calc  
finer matrix: XCa=0.47  
(breakdown of Ca-Amp)  
Recrystallization + neocryotallization: leaching Na |
|---|---|---|---|
| Bt | (Greenish-brown) | $\text{Hbl} + \text{K} + \text{H}_2\text{O} = \text{Bt}_2 + \text{Ca}$  
Al increases + Fe decreases | redish-brown  
1. recrystallization  
2. replace Hbl |
| Hbl | 1. Core→rim: 
brown-green→blue-green  
2. beard: (Na + Al rich)  
edenite (Ed)→actinolite (Act)  
3. Ti decreases | | |
Figure 4.2 Distributions and orientations of bridge zones in Sample H (a-c), D (d-f), and C (g-i).
Figure 4.2 (Continued) The left column includes full thin section backscattered electron (BSE) images of the samples, and the middle column contains the same background image with the locations of bridge zones marked in red. The right column contains rose diagrams illustrating the orientation of the long axis of the bridge zones, with the numbers inside the diagram representing the percentage of measurements in each angular bin. The horizontal direction of each circle is parallel to the horizontal direction of each thin section. The radial variable represents the frequency, while the black dot represents the average of all orientations. They clearly show the mutual parallel/subparallel relationship between the bridge zone orientation and the thin section foliation orientation.
4.5.2. **Description of the Deformation Gradient**

The following sections describe, by sample from least to most deformed, the distribution of bridge zones then their internal structure. The relative strain increase is mentioned in Johnson et al. (2008).

**4.5.2.1. Bridge Zone in the Tonalite with the Initiation of Folia (Least Deformed)**

**4.5.2.1.1. Overall Microstructure and Distribution of Bridge Zones**

Generally, the least deformed end of the strain gradient (sample H) consists of Qz(20%)+Pl(55%)+Hbl(10%)+Bt(10%)+Ttn+Kfs. Large-grained quartz clusters show undulatory extinction, with some 100–300-micron subgrains visible and some exhibiting a foam texture. Large biotite grains are up to 3 millimeters in size, with some fan-shaped or lenticular. Primary biotite is nearly always bent, and sometimes is cut by secondary biotite grains. Large grains of plagioclase are typically in the range of 2mm across, with oscillating compositional zoning. Some plagioclase grains have hornblende inclusions in the inner core, while others have rounded quartz inclusions in the outer ring zone. Healed fractures are more common in the outer ring zone than in the inner core. Plagioclase growth twinning is well-developed. The boundary between two adjacent large plagioclase crystals can be irregular. Internal fractures are common in plagioclase, and misorientation or twinning bending can be seen in some of the grains. Hornblende grains are commonly 1-3 mm with some quartz and biotite inclusions. Some hornblende zoning is visible under cross polarized light with brown-green cores and blue-green rims. Some hornblende grains are pulled off or necked down into fragments at the grain edges.
Bridge zones, defined here as in the previous chapter, are present throughout the rock, constituting 1.5%-3% of the area (Figure 4.2b). These bridge zones primarily occur along the edges of plagioclase and hornblende grains, and appear as narrow areas connecting the biotite and/or quartz domains. The long axes of the bridge zones are weakly aligned (Figure 4.2c), with an average orientation close to the weak macroscale foliation.

4.5.2.1.2. Internal Bridge Zone Structure

Most of the bridge zones connect biotite-rich domains, with others connecting quartz or mixed quartz and biotite (Figure 4.2a, b). The assemblages of bridge zones typically includes the phases that surround the bridge zone. Bridge zones that cut hornblende tend to contain fine-grained biotite, hornblende fragments, and fine-grained quartz and plagioclase. Bridge zones that cut plagioclase contain mostly fine-grained biotite, plagioclase, and titanite.

As one example of a typical structure, here we focus on a bridge zone that cuts through a plagioclase domain (Figure 4.3). The bridge zone connects two regions containing quartz clusters and biotite grains—the quartz clusters outside the bridge zone show a foam texture. Neither quartz nor biotite outside the bridge zone show evidence of internal deformation leading to significant optical misorientation (Figure 4.3a & b). Plagioclase grains on both sides of this bridge zone also do not exhibit optical distortion (Figure 4.3 a & b). Even with the development of cracks, the plagioclase twinning is not affected. The bridge zone appears as fine-grained plagioclase (80%), biotite (10%), and titanite (10%). Fine grains of plagioclase are generally 20-70 microns across. Finer grains are relatively equant, while some larger grains are elongated slightly along the bridge zone. These fine plagioclase grains have flat boundaries and form 120° triple junctions (Figure 4.3a-c). CL imaging (Figure 4.3d) reveals differences in
the brightness of fine plagioclase inside and outside the bridge zone that is only weakly visible in BSE imaging (Figure 4.3c). This difference is evident in color CL images (Figure 4.3e-h), especially in the blue channel (Figure 4.3f). The coarse plagioclase grains outside the bridge zone show bright CL and compositional zoning. In contrast, the finer grains show a very dark CL and a more homogeneous composition (Figure 4.3d).

Although the fracture outside the upper right corner of this bridge zone (Figure 4.3) also shows darker CL, the lighter dark CL differs from the one inside the bridge zone. Only in areas where coarse-grained plagioclase is in contact with a weak phase (quartz or biotite) or in fractures connecting weak phases does the same CL dark color appear as inside the bridge zone. Most of the fine-grained biotite has a long axis (parallel to the basal direction) aligned with the orientation of the bridge zone and is generally 50-100 µm long. Some tiny biotite grains (10 µm or finer) exist in the fine-grained plagioclase rims or 120° triple junctions. Fine-grained titanite (20 microns or less) fills the grain gaps of fine-grained plagioclase in a worm-like form.
Figure 4.3 Example bridge zone in sample H. (a) and (b) optical photomicrographs, in which (a) is under plane-polarized light and (b) is under cross-polarized light. (c) BSE image, and (d) panchromatic Cathodoluminescence (CL) image.
Figure 4.3 (Continued) The red box in (d) is the location of (e)-(h), which are the color CL (e) and the single channels of blue (f), green (g), and red (h). The bridge zone connects two weak phases containing quartz clusters and biotite grains. It appears as fine-grained (tens of microns) plagioclase, biotite flakes, and titanite. Adjacent areas are coarse plagioclase grains (a few millimeters across). CL images, particularly the blue channel, reveal differences in the brightness of fine plagioclase within the bridge zone that are not visible in the BSE image. The coarse plagioclase grains outside the bridge zone show a very high brightness with visible zoning texture. In contrast, the finer grains show a very low brightness and a more homogeneous intracrystalline character. Although the fracture in the upper left corner of this bridge zone also shows some dark CL (green dash circle in d), it is clearly different from the one inside the bridge zone. The blue dashed ovals in d show a thin dark CL band at the edge of the coarse-grained plagioclase. These dark bands generally appear at the edges of the plagioclase in contact with the weak phase and have the same CL brightness as the grains in the bridge zone.

4.5.2.2. Bridge Zone in Moderately Deformed Rock

4.5.2.2.1. Overall Microstructure and Distribution of Bridge Zones

The studied sample with moderate deformation (D) contains the assemblage Qz(20%)+Pl(50%)+Hbl(10%)+Bt(12%)+Ttn. The deformation characteristics within the larger grains are similar to those of sample H. Hornblende does not show significant intracrystalline optical inhomogeneity/misorientation, while some plagioclase grains show undulatory extinction and exhibit some “subgrains” at their edges. Foam-textured quartz clusters are common in the thin section. Biotite grains also commonly exhibit undulatory extinction.

The modal amount of the bridge zone in D is similar to that in H, 1.5%-3%, but it is more aligned and parallel to the macroscopic foliation (Figure 4.2e). These bridge zones connect biotite and/or quartz domains. Bridge zones are located on the edges of plagioclase (or hornblende) and seem to follow grain boundaries. Most bridge zones form part of the S-C fabric around strong primary igneous plagioclase and hornblende (Figure 4.2f).
4.5.2.2.2. Internal Bridge Zone Structure

Many bridge zones consist of fine-grained biotite and plagioclase with or without irregularly distributed titanite, quartz, epidote, and hornblende (Figure 4.2 d & e). For a detailed description of a typical bridge zone, we focus on one that cuts through a plagioclase domain (Figure 4.4). The bridge zone connects two weak phases containing quartz clusters, biotite grains, and an “island” biotite grain in the middle. This bridge zone forms a dogleg shape, contains two subzones and an “island” of biotite fragments occurs at the inflection point. The coarse-grained plagioclase boundaries around the bridge zone are not straight but show a sinuous wavy shape. The bridge zone appears as fine-grained (tens of microns) plagioclase (80%), titanite (15%), and biotite (5%).

Adjacent areas comprise coarse plagioclase grains (a few millimeters). CL images (Figure 4.4d-h) also show differences in the brightness of fine plagioclase within the bridge zone that are not visible in the BSE images (Figure 4.4c). This difference is evident in color CL images (Figure 4.4d-h), especially in the blue channel (Figure 4.4h). The coarse plagioclase grains outside the bridge zone show very bright CL and compositional zoning texture. In contrast, the finer grains show very dark CL and a more homogeneous composition. The orientations of the bridge zone are controlled by plagioclase grain boundaries. Compared to the CL brightness shown near zoning, cracks and inclusions in coarse plagioclase grains, the narrow bands near the tips of the weak phases (e.g., biotite tips, shown by the yellow arrows in Figure 4.4d-h) show the same CL characteristics as fine-grained plagioclase in the bridge zone. These fine plagioclase grains have flat boundaries and form 120° triple junctions. Fine-grained biotite and titanite grains fill the gaps at the edges of the plagioclase grains. The long-axis direction of fine-grained biotite is often parallel to the boundary direction of the coarse-grained
plagioclase on both sides of the bridge zone. In contrast, the distribution of fine-grained titanite is more random.

**Figure 4.4** Example bridge zone in sample D. (a) and (b) optical photomicrographs, in which (a) is under plane-polarized light and (b) is under cross-polarized light. (c) is the BSE image, and (d) is the panchromatic Cathodoluminescence (CL) image.
**Figure 4.4** (Continued) While (e) is the color CL image with the single channels of blue (f), green (g), and red (h). The bridge zone connects two weak phases containing quartz clusters and biotite grains. It appears as fine-grained (tens of microns) plagioclase, biotite flakes, and titanite. Adjacent areas are coarse plagioclase grains (a few millimeters across). CL images also showed differences in the brightness of fine plagioclase within the bridge zone, that are not visible in the BSE image, especially in the blue channel. The coarse plagioclase grains outside the bridge zone show a very high brightness and visible zoning texture. In contrast, the finer grains show a very low brightness and a more homogeneous intracrystalline character. The orientations of the bridge zone are controlled by grain boundary. Compared to the CL brightness shown near zoning, cracks and inclusions in coarse plagioclase grains, the narrow regions of feldspar near the tips of the weak phases (e.g., biotite tips, shown by the yellow arrows in d-h) show the same CL characteristics as fine-grained plagioclase in the bridge zone.

4.5.2.3. Bridge Zone in the Most Sheared Mylonite

4.5.2.3.1. Overall Microstructure and Distribution of Bridge Zones

The most sheared sample in our study (C) has Qz(40%) + Pl(45%) + Bt(15%). Ribbons of recrystallized quartz alternating with biotite folia and biotite-quartz folia connect in the thin section to form a mature S-C fabric. The sample contains no hornblende. The fractures in the plagioclase are more developed, and some possible subgrains can be seen in the boundary. The sample exhibits two types of fine-grained areas: (1) biotite and quartz perpendicular to the plagioclase boundary appear as beard - similar to pressure shadows; (2) Fine-grained biotite and plagioclase aggregates are distributed parallel to the plagioclase grains. The second is the bridge zone, which is what we focus on.

The bridge zones in C are more abundant than those in H and D, reaching 2%-4% of the total area of the thin section (Figure 4.2h). The bridge zones in C form at the boundaries of the plagioclase grains. Bridge zone generally lie parallel or subparallel to the macroscale foliation ((Figure 4.2i).

4.5.2.3.2. Internal Bridge Zone Structure

The bridge zone we consider in detail in this sample connects two weak domains containing quartz clusters and biotite grains (Figure 4.5). The bridge zone contains fine-grained (tens of microns)
plagioclase (40%), biotite (50%), quartz (5%) and minor titanite. The orientations of the bridge zone are along grain boundaries. The fine-grained plagioclase is relatively equant among the fine-grained biotite, while the quartz and titanite grains are interstitial between plagioclase and biotite. Surrounding the bridge zone are coarse plagioclase grains (a few millimeters). CL images (Figure 4.5d-h) reveal differences in the brightness of fine plagioclase within the bridge zone not visible in BSE imaging. This difference is evident in color CL images (Figure 4.5e-h), especially in the blue channel (Figure 4.5f). The coarse plagioclase grains outside the bridge zone show a high brightness and visible zoning texture. In contrast, the finer grains show a very dark CL and a more homogeneous internal character. In addition, dark CL bands are also visible inside the coarse-grained plagioclase. These dark bands are usually in fractured areas around quartz or biotite. Some plagioclase shows fractures and dilatant zones that have been rehealed by plagioclase and quartz (Figure 4.5c-e).
Figure 4.5 Example bridge zone in sample C. (a) and (b) optical photomicrographs, in which (a) is under plane-polarized light and (b) is under cross-polarized light (green box in c). (c) is the BSE image, and (d) is the panchromatic Cathodoluminescence (CL) image.
Figure 4.5 (Continued) The red box in (d) is the location of (e)-(h), which are the color CL (e) and the single channels of blue (f), green (g), and red (h). The bridge zone connects two weak phases containing quartz clusters and biotite grains. It appears as fine-grained (tens of microns) plagioclase and biotite flakes. Adjacent areas are coarse plagioclase grains (a few millimeters across). CL images, particularly the blue channel, reveal differences in the brightness of fine plagioclase within the bridge zone that are not visible in the BSE image. The coarse plagioclase grains outside the bridge zone show a very high brightness and visible zoning texture. In contrast, the finer grains show a very low brightness and a more homogeneous intracrystalline character. The orientations of the bridge zone are along grain boundary. Areas where the coarse-grained plagioclase is in contact with the weak phase usually show the same brightness as the bridge zone (blue dotted circle in d).

4.5.3. Crystallographic characteristics

4.5.3.1. Inverse Pole Figure, Grain Reference Orientation Deviation and Kernel Average

Misorientation Maps

In this section, we investigate the crystallographic orientations of plagioclase grains and misorientation distribution in single grains. Based on these data, we explore the relationship between the particle size variation of plagioclase and its misorientation in and outside the studies bridge zone in each sample.

4.5.3.1.1. Bridge Zone in the Least Deformed Rock

The coarse-grained plagioclase crystals that bound the bridge zone do not show strong evidence of internal deviations from the mean grain orientation (Figure 4.6b) or kernel misorientation (Figure 4.6c), even at grain boundaries. One grain (labeled #2 in Figure 4.6a) internally shows parallel high deviation/high kernel misorientation bands, reflecting twinning lamellae (Figure 4.6a). In contrast, the orientation deviation (Figure 4.6h) and kernel misorientation (Figure 4.6i) of coarse-grained quartz grains are more pronounced.
In the inverse pole figure map, fine plagioclase grains inside the bridge zone show orientations similar to coarse grains on both sides (Figure 4.6a). We can divide fine grains into different groups (yellowish and purplish) according to their orientation. The line between the two different groups is not straight. Grains with a similar orientation to grain 2 (purplish) form a “lobster claw” shape, surrounding grains with a similar orientation to grain 1 (yellowish).

**Figure 4.6** Crystallographic data for plagioclase (Pl) and quartz (Qz) from the bridge zone in Sample H. (a) and (g) Inverse pole figure maps. (b) and (h) Grain reference orientation deviation map. (c) and (i) Kernel average misorientation map. (d) Pole figure of plagioclase fine grains (colored) and parent/host grains (dots). (e) and (f) are plagioclase Misorientation Angle distributions of fine grains and all grains.
Figure 4.6 (Continued) The “hot spots” of quartz grains in grain reference orientation deviation are generally distributed at the edges of adjacent large grains, while the coarse plagioclase have low deviation with mainly twining showing. Kernel average misorientation values of quartz are high at edges of adjacent large grains consistent with the presence of grain boundaries. (d) The strong CPO maxima of fine-grained plagioclase overlaps with the orientations of the two parent/host grains, which shows strong parent/host grain control/inheritance. As can be seen from the all-plagioclase-grain misorientation angle distribution (f), the uncorrelated distribution is similar to the theoretical random distribution, but with a slight deviation. The 70°-90° and 150°-170° ranges account for a large proportion. However, (e) the misorientation profile of fine grains shows a zigzag distribution. The correlated distribution of all grains (f) shows peaks at 180° (twinning) in addition to the presence of peaks at low angles (<20°). The correlated distribution of fine grains shows subpeaks at low angles (<20°) in addition to 180°, with peaks at 50°-100° and 130° (e).

4.5.3.1.2. Bridge Zones in the Moderately Deformed Rock

For the purposes of crystallographic analysis, we divide the bridge zone into two subzones, with Subzone 1 being sub-horizontal in Figure 4.4 and Subzone 2 being subvertical.

Subzone 1

Two coarse-grained plagioclases on either side of the bridge zone do not show strong evidence of internal deviation from the mean grain orientation (Figure 4.7b) or kernel misorientation (Figure 4.7c). Grain 1 (labeled in Figure 4.7a) internally shows parallel high deviation/high kernel misorientation lines, reflecting twinning lamellae. Though Grain 2 has a very high deviation but lower in kernel misorientation. The boundary of the coarse-grained plagioclase in contact with the bridge zone is curved rather than straight (Figure 4.3).

Subzone 2

Three coarse-grained plagioclases on either side of the bridge zone do not show strong evidence of internal deviations from the mean grain orientation (Figure 4.8b) or kernel misorientation (Figure 4.8c). Some high deviations and kernel misorientation areas are located at the boundaries. The orientation
deviation (Figure 4.8h) and kernel misorientation (Figure 4.8i) of coarse-grained quartz grains are more obvious.

**Figure 4.7** Crystallographic data for plagioclase (Pl) from the bridge zone in Subzone 1 of Sample D. (a) Inverse pole figure map. (b) Grain reference orientation deviation distribution. (c) Kernel average misorientation distribution. (d) Pole figure of plagioclase fine grains (colored) and parent/host grains (dots). (e) and (f) are plagioclase Misorientation Angle distributions of fine grains and all grains. The lower coarse plagioclase are larger grains are low deviation with only twining showing in the Grain Reference Orientation Deviation maps, but the upper grain shows pretty higher deviate. Kernel average misorientation (KAM) values of quartz are high at edges of adjacent large grains suggesting a high dislocation density in these areas. The CPO maxima of fine-grained plagioclase overlaps with the orientations of the two parent/host grains, which shows strong parent/host grain control/inheritance (d). From all and fine plagioclase misorientation angle distribution (e & f), the uncorrelated distribution has a high peak at low angles (<20°). The correlated distribution of all grains (f) shows peaks at 180° (twinning) in addition to the presence of peaks at low angles (<20°). The correlated distribution of fine grains shows subpeaks at low angles (<20°) in addition to 180°, with peaks at 140° (e).
Figure 4.8 Crystallographic data for plagioclase (Pl) and quartz (Qz) from the bridge zone in Subzone 2 of Sample D. (a) and (g) Inverse pole figure maps. (b) and (h) Grain reference orientation deviation map. (c) and (i) Kernel average misorientation map. (d) Pole figure of plagioclase fine grains (colored) and parent/host grains (dots). (e) and (f) plagioclase Misorientation Angle distributions of fine grains and all grains. (j) quartz Misorientation Angle distributions all grains.

Three coarse-grained plagioclases on either side of the bridge zone show some evidence of internal deviations from the mean grain orientation (b) or kernel misorientation (c), though the deviations are very low. The grain reference orientation deviation (h) and kernel misorientation (i) of coarse-grained quartz grains have higher values than in feldspar. (d) The pole figures of fine-grained plagioclase have maximum values at (010), (001), and [100]. Although slightly offset, their maximum positions overlap with the outer three large grains. As can be seen from the all-plagioclase-grains (f) misorientation angle distribution, the uncorrelated distribution is similar to the theoretical random distribution, but with a slight deviation. The correlated distribution of all grains (f) shows peaks at 180° (twinning) in addition to the presence of peaks at low angles (<20°).
Figure 4.8 (Continued) The correlated distribution of fine plagioclase grains (e) shows subpeaks at low angles (<20°) in addition to 180°, with peaks at 70° and 90°. The misorientation profile of fine grains shows close to (e) a random distribution except at higher angles. The misorientation angle distribution of all quartz grains (j) tracks the random distribution with slightly high distribution at lower angle (<20°). The correlated distribution of all quartz grains (j) has a high concentration of misorientations at low angles (<20°) in addition to 60° (Dauphiné twin).

4.5.3.1.3. Bridge Zone in the Strongly Deformed Rock

The coarse plagioclase grains outside the bridge zone show some evidence of internal deviations from the mean grain orientation (Figure 4.9b) and kernel misorientation (Figure 4.9c) except the location of twinning lamellae. Most fine plagioclase grains in bridge zone are low in internal deviation and kernel misorientation. The orientation deviation and kernel misorientation of coarse-grained quartz grains are more pronounced (Figure 4.9k & l).
Figure 4.9 Crystallographic data for plagioclase (Pl) and quartz (Qz) from the bridge zone in Sample C. (a) and (g) Inverse pole figure maps. (b) and (h) Grain reference orientation deviation map. (c) and (i) Kernel average misorientation map. (d) Pole figure of plagioclase fine grains (colored) and parent/host grains (dots). (e) and (f) are plagioclase Misorientation Angle distributions of fine grains and all grains. (k) and (l) are plagioclase Misorientation Angle distributions of fine grains and all grains. The coarse plagioclase grains outside the bridge zone show some evidence of internal deviations from the mean grain orientation (b) or kernel misorientation (c) except the location of twinning lamellae. Most fine plagioclase grains in bridge zone are low in internal deviations and kernel misorientations. And the orientation deviation (k) and kernel misorientation (l) of coarse-grained quartz grains are more obvious. The pole figures of fine-grained plagioclase (d) have maximum values at (010), (001), and [100]. However, the maximum value is quite low at 2.86. The pole figures do not conform to most slip systems in plagioclase (e.g., [100] (010)).
Figure 4.9 (Continued) The fine-grained quartz CPO (j) shows a weak pattern of combination of rhomb <a> and prism <a> slip. The outside quartz grains’ CPO is still weak but stronger than that of the fine grains (j). The outside quartz grains’ CPO fits the rhomb <a>. From the all-plagioclase-grain misorientation angle distribution, the uncorrelated distribution (f) is basically similar to the theoretical random distribution, but with a peak at 20°. However, the MAD of fine plagioclase grains shows a zigzag distribution, but tracks the random distribution line (e). The misorientation angle distributions of all quartz grains (l) and fine grains fit the random distribution with slightly high distribution at lower angle (<20°). The correlated distribution of all plagioclase grains (f) shows peaks at 180° (twinning) in addition to the presence of peaks at low angles (<20°). The correlated distribution of plagioclase fine grains (Fig. 9e) shows subpeaks at low angles (<20°) in addition to 180°, with peaks at 50°-100° and 130°. The correlated distribution of all quartz grains and fine quartz grains ((k & l)) both have high distribution at low angle (<20°) in addition to 60° (Dauphiné twin).

4.5.3.2. Crystallographic Preferred Orientation (CPO) and Misorientation Angle Distribution (MAD) of Fine Grains in Bridge Zones

In this section, we explore how the crystallographic relationships in plagioclase change across the strain gradient (Figures 4.6-4.9). Other minerals are not present in sufficient abundance to make statistical evaluations feasible. Even for plagioclase, the smaller data sets of Misorientation Angle Distribution angles within a single bridge zone reduces the confidence of any interpretation.

4.5.3.2.1. Bridge Zone in the Least Deformed Rock

The principal axes of the fine-grained plagioclase have maximum concentrations that align with the two coarse-grained feldspar grains outside (Figure 4.6d).

The Misorientation Angle Distribution of fine grains shows a zigzag distribution (Figure 4.6e). Both show subpeaks at low angles (<20°). However, as can be seen from the all-plagioclase-grain Misorientation Angle Distribution (Figure 4.6f), the uncorrelated distribution is similar to the theoretical random distribution, but with high deviation in the 70°-90° and 150°-170° ranges.
The correlated distribution of fine grains shows subpeaks at low angles (<20°) in addition to 180°, with peaks at 50°-100° and 130° (Figure 4.6e). The correlated distribution of all grains (Figure 4.6f), which considers only adjacent grains, shows peaks at 180° (twinning) in addition to the presence of peaks at low angles (<20°).

4.5.3.2.2. Bridge Zones in the Moderately Deformed Rock

Subzone1

The principal axes of the fine-grained plagioclase have maximum concentrations aligned with the two coarse-grained feldspar grains outside (Figure 4.7d).

The fine plagioclase Misorientation Angle Distribution (Figure 4.7e & f), the uncorrelated distribution has a high peak at low angles (<20°). The correlated distribution of fine grains shows subpeaks at low angles (<20°) in addition to 180°, with peaks at 140° (Figure 4.7e). The correlated distribution of all grains (Figure 4.7f) shows peaks at 180° (twinning) in addition to the presence of peaks at low angles (<20°).

Subzone2

Although slightly offset, the fine-grained plagioclase maximum positions overlap with the outer three large grains (Figure 4.8d).

The Misorientation Angle Distribution of fine grains shows (Figure 4.8e) a random distribution with a small subpeak at low angles (<20°). The all-plagioclase-grains (Figure 4.8f) Misorientation Angle Distribution, the uncorrelated distribution is basically similar to the theoretical random distribution, but with a slight deviation. The 0°-20° and 180° ranges account for a large proportion. The Misorientation
Angle Distribution of all quartz grains (Figure 4.8j) basically fits the random distribution with slightly high distribution at lower angle (<20°).

The correlated distribution of all grains (Figure 4.8f) shows peaks at 180° (twinning) in addition to the presence of peaks at low angles (<20°). The correlated distribution of fine plagioclase grains (Figure 4.8e) shows subpeaks at low angles (<20°) in addition to 180°, with peaks at 70° and 90°. The correlated distribution of all quartz grains (Figure 4.8j) both have high distribution at lower angle (<20°) in addition to the 60° (Dauphiné twin).

4.5.3.2.3. Bridge Zone in the Strongly Deformed Rock

The pole figures of fine-grained plagioclase (Figure 4.9d) have principal axes maximum values. The fine-grained quartz CPO (Figure 4.9j) shows a weak pattern of combination of rhomb <a> and prism <a> slip. The outside quartz grain CPO is still weak but stronger than that of the fine grains (Figure 4.9j) and fits the rhomb <a>, but with a much smaller dataset, which reduces confidence in any interpretations.

The Misorientation Angle Distribution of fine plagioclase grains shows a zigzag distribution, but basically fits the random distribution line (Figure 4.9e). The Misorientation Angle Distributions of all quartz grains (Figure 4.9l) and fine grains basically fit the random distribution with slightly high distribution at lower angle (<20°). The all-plagioclase-grain Misorientation Angle Distribution, the uncorrelated distribution (Figure 4.9f) is similar to the theoretical random distribution, but with a peak at 20°.

The correlated distribution of all plagioclase grains (Figure 4.9f) shows peaks at 180° (twinning) in addition to the presence of peaks at low angles (<20°). The correlated distribution of plagioclase fine grains (Figure 4.9e) shows subpeaks at low angles (<20°) in addition to 180°, with peaks at 50°-100° and
130°. The correlated distribution of all quartz grains and fine quartz grains ((Figure 4.9k&l)) both have high distribution at lower angle (<20°) in addition to the 60° (Dauphiné twin).

4.5.4. Chemical Characteristics

The observations above – particularly the difference in CL color and intensity of fine-grained plagioclase and the introduction of fine-grained titanite in the bridge zones – indicate that all three bridge zones experienced chemical change. In this section we examine the chemical variations within and outside the bridge zones in more detail. To do so, we performed EPMA chemical spot analysis for the three samples in locations with CL variations in plagioclase as well as the large and fine-grained biotite inside and outside bridge zones (Table 4.3).

Overall, the anorthite content of both coarse- and fine-grained plagioclase increased from the lowest to the highest deformation degree samples (Figure 4.10). Coarse-grained plagioclase anorthite content is 0.36 in the least deformed (H), 0.38 in the moderately deformed (D), 0.40 in the highest deformed sample (C). The anorthite content of fine-grained plagioclase in the bridge zone is reduced in the lowest and moderately deformed samples relative to their coarse-grained plagioclase anorthite content (0.30 in H and 0.36 in D). Sample C, with the highest degree of deformation, differs from the plagioclase variation exhibited by H and D, where the anorthite content within the bridge zone is 0.43 elevated compared to the coarse-grained plagioclase.

In general, with increasing of deformation, Al increases, while Fe and Mg gradually decrease in biotite. The content of Ti was the highest in the most deformed C (~0.3 apfu), while the content of Ti was the same in the lowest deformed H and the moderate deformed D (~2.3 apfu). However, the content of Mg/(Fe+Mg) in biotite remained unchanged in the three samples (~0.55). Fine-grained biotite also had
higher Al and lower Mg and Fe per sample than coarse-grained biotite. Other elements remain in roughly the same amount.

Figure 4.10 EPMA analyses of bridge-zone finer and larger/coarser plagioclase grains. (a), (b) and (c) anorthite content of finer (orange) and larger/coarser (blue) plagioclase grains in H, D and C. (d) the variation of plagioclase anorthite average content in three samples. Overall, the anorthite content of both coarse- and fine-grained plagioclase increased from the lowest to the highest deformation degree samples. Coarse-grained plagioclase anorthite content is 0.36 in the lowest deformed H, moderately deformed rock, and 0.38 in D, while the highest deformed sample C has the highest anorthite content, 0.40. The anorthite content of fine-grained plagioclase in the bridge zone is reduced in the lowest and moderately deformed samples, D and H, relative to their coarse-grained plagioclase anorthite content (0.30 in H and 0.36 in D), while the reduction is smaller in D than in H. C, as the sample with the highest degree of deformation, differs from the plagioclase variation exhibited by H and D, where the anorthite content within the bridge zone is elevated compared to the coarse-grained plagioclase.
Table 4.3 EPMA spot analysis from plagioclase and biotite in H, D and C

| Sample Phase | Plagioclase | | | Biotite | | | Biotite | | | Biotite | | | Biotite |
|--------------|-------------|---|---|-------------|---|---|-------------|---|---|-------------|---|---|-------------|---|---|
|              | Large | Fine | Large | Fine | Large | Fine | Large | Fine | Large | Fine | Large | Fine | Large | Fine |
| H            | | | | | | | | | | | | | | |
| Na₂O         | 7.13 | 7.91 | 0.08 | 0.18 | 6.88 | 7.20 | 0.13 | 0.17 | 6.71 | 6.36 | 0.08 | 0.12 | | |
| K₂O          | 0.23 | 0.15 | 9.52 | 9.29 | 0.33 | 0.20 | 8.53 | 9.22 | 0.15 | 0.17 | 9.38 | 9.40 | | |
| MgO          | 0.00 | 0.00 | 12.04 | 11.91 | 0.00 | 0.01 | 11.93 | 11.69 | 0.00 | 0.05 | 11.57 | 11.50 | | |
| CaO          | 7.31 | 6.12 | -0.13 | 0.32 | 7.70 | 7.39 | 0.82 | 0.06 | 8.15 | 8.73 | -0.09 | -0.07 | | |
| MnO          | -0.01 | -0.01 | 0.25 | 0.25 | -0.02 | -0.02 | 0.21 | 0.19 | 0.02 | 0.02 | 0.30 | 0.26 | | |
| FeO          | 0.07 | 0.13 | 18.10 | 17.22 | 0.01 | 0.09 | 18.46 | 17.89 | 0.03 | 0.27 | 17.39 | 16.47 | | |
| BaO          | -0.02 | -0.02 | -0.01 | -0.01 | -0.01 | -0.01 | | | | | | | | |
| Al₂O₃        | 26.33 | 25.51 | 16.40 | 16.60 | 26.65 | 26.54 | 16.68 | 17.76 | 27.08 | 27.86 | 18.03 | 18.26 | | |
| Cr₂O₃        | 0.05 | 0.23 | 0.02 | 0.17 | | | | | | | 0.07 | 0.18 | | |
| SiO₂         | 59.23 | 60.91 | 36.85 | 37.03 | 58.54 | 59.40 | 37.38 | 36.71 | 58.29 | 57.41 | 36.54 | 36.67 | | |
| TiO₂         | 2.12 | 2.14 | 1.97 | 1.85 | | | | | | | 2.52 | 2.76 | | |
| Total        | 100.30 | 100.73 | 99.37 | 99.21 | 100.13 | 100.84 | 100.23 | 99.72 | 100.43 | 100.87 | 99.89 | 99.64 | | |
| apfu         | | | | | | | | | | | | | | |
| OH           | | | | | | 4.00 | 4.00 | 4.00 | 4.00 | 4.00 | 4.00 | | |
| Na           | 0.61 | 0.68 | 0.02 | 0.05 | 0.59 | 0.62 | 0.04 | 0.05 | 0.58 | 0.55 | 0.02 | 0.03 | | |
| K            | 0.01 | 0.01 | 1.84 | 1.79 | 0.02 | 0.01 | 1.63 | 1.77 | 0.01 | 0.01 | 1.79 | 1.79 | | |
| Mg           | 0.00 | 0.00 | 2.72 | 2.68 | 0.00 | 0.00 | 2.66 | 2.62 | 0.00 | 0.00 | 2.58 | 2.56 | | |
| Ca           | 0.35 | 0.29 | 0.00 | 0.06 | 0.37 | 0.35 | 0.14 | 0.01 | 0.39 | 0.42 | 0.00 | 0.00 | | |
| Mn           | 0.00 | 0.00 | 0.03 | 0.03 | 0.00 | 0.00 | 0.03 | 0.02 | 0.00 | 0.00 | 0.04 | 0.03 | | |
| Fe           | 0.00 | 0.00 | 2.29 | 2.17 | 0.00 | 0.00 | 2.31 | 2.25 | 0.00 | 0.01 | 2.18 | 2.06 | | |
| Ba           | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | | | | | | 0.00 | 0.00 | |
| Al           | 1.38 | 1.33 | 2.93 | 2.95 | 1.40 | 1.38 | 2.94 | 3.14 | 1.42 | 1.46 | 3.18 | 3.21 | | |
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4.5.5. Numerical Modeling

4.5.5.1. Stress Amplification

Bridge zones are not all randomly distributed, and Gerbi et al. (2016) point out that recrystallization zones may be consistent with zones of stress amplification. The stress amplification in these domains may be sufficient to cause the activation of grain size reduction and/or reaction mechanisms. We therefore used the PLC toolbox (Cook et al. 2014) to explore the possible instantaneous stress distribution in the three rocks under bulk loading conditions of simple shear and vertical axial compression.

The presence of high stress regions that could lead to localized microstructure formation is shown by microscopic stress field distribution maps in the geometries drawn without bridge zones (Figure 4.11, Table 4.4). These high stress regions are in the strong phase and connect the weak phases. The results at a 10 MPa load demonstrate that significant stress amplification (commonly 4-7 times the applied stress, Table 4.5) occurs in dispersed locations in the strong phases under shear loading. For vertical axial compression, the amplification was ~3.3-4.2 (Table 4.5). The spatial correlation between sites of stress amplification and bridge zone formation is moderate (Figure 4.11), with some high stress sites not containing bridge zones and vice versa.
Figure 4.11 Geometries with and without bridge zones for modeling and stress distributions calculated from bridge-zone-free geometries under shearing and vertical compression. (a), (f) and (k) are BSE images of local regions in H, D and C, respectively. (b), (g) and (l) are their simplified geometries, including observed bridge zones. (c), (h) and (m) are the geometries with the bridge zones removed. (d), (i) and (n) are stress maps in simple shear loading. (e), (j) and (o) are stress maps in axial loading. High stress regions are in the strong phase and connect the weak phases/domains. The results demonstrate that significant stress amplification (commonly 4-7 times the applied stress) occurs in dispersed locations in the strong phases under shear loading. The correlation between sites of stress amplification and bridge zone formation is moderate, with some high stress sites containing bridge zones and some not.
Table 4.4 Phase mode in modeling geometries

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<th>Geometry</th>
<th>Strong Phases</th>
<th>Qz</th>
<th>Bt</th>
<th>Fraction Bridge zone</th>
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<td>Ph</td>
<td>Hbl</td>
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<tr>
<td>H</td>
<td>47.86%</td>
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<td>55.64%</td>
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<td>7.78%</td>
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<td>60.05%</td>
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</tr>
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<td>C-bz</td>
<td>58.22%</td>
<td>0.00%</td>
<td>58.22%</td>
<td>26.86%</td>
</tr>
</tbody>
</table>

4.5.5.2. Bulk Strength Reduction

Using additional output derived from the PLC Toolbox, we calculate the effect of the presence of bridge zones between the weak phases of these three geometries on the bulk rock strength. Based on observations of the natural samples, we added ~2% modal bridge zones to each of the three geometries (Table 4.4), mostly replacing stronger phases. The flow law parameters of quartz were used to calculate the bridge zone rheology (Table 4.1).

Under shear stress, for the least sheared rock, H, the strength dropped 88% with the 2.4% bridge zones added under a 10 MPa shear load, and 94% weaker than the original geometry under 50 MPa. D, which experienced moderate deformation, is 86% and 93% weaker with 2.5% bridge zones under 10 MPa and 50 MPa respectively. For the most sheared sample, C, the strength dropped 79% and 91% respectively under 10 MPa and 50 MPa (Table 4.5).
Under vertical compression, the strength drops for H, D, and C are 68%, 25%, and 64% respectively under 10 MPa. Under 50MPa, the samples with bridge zones are 89%, 45%, and 75% weaker than the geometries without bridge zones (Table 4.5).

Under different stress conditions of the two loads, the three samples showed significant weakening with the addition of bridge zones. In general, for the same loading condition, the effect of weakening of 50 MPa is more pronounced than that of 10 MPa. In all calculations, the weakening effect caused by bridge zone addition is greater under shear than it is under axial compression.
### Table 4.5 Results of modeling experiments

<table>
<thead>
<tr>
<th>Geometry</th>
<th>Strain-rate under 10MPa</th>
<th>Strain-rate under 50MPa</th>
<th>Magnitude of bulk weakening 10MPa</th>
<th>Magnitude of bulk weakening 50MPa</th>
<th>Stress amplification factor 10MPa</th>
<th>Stress amplification factor 50MPa</th>
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<td>Axial Compression (E-14)</td>
<td>Axial Compression (E-15)</td>
<td>Shearing (E-14)</td>
<td>Shearing (E-15)</td>
<td>Shearing</td>
<td>Shearing</td>
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<tr>
<td>H</td>
<td>8.69</td>
<td>12.44</td>
<td>8.09</td>
<td>5.96</td>
<td>88%</td>
<td>68%</td>
</tr>
<tr>
<td>H-bz</td>
<td>72.84</td>
<td>39.41</td>
<td>145.21</td>
<td>55.55</td>
<td>86%</td>
<td>25%</td>
</tr>
<tr>
<td>D</td>
<td>1.53</td>
<td>5.34</td>
<td>0.58</td>
<td>1.71</td>
<td>86%</td>
<td>93%</td>
</tr>
<tr>
<td>D-bz</td>
<td>10.97</td>
<td>7.14</td>
<td>8.35</td>
<td>3.10</td>
<td>86%</td>
<td>25%</td>
</tr>
<tr>
<td>C</td>
<td>7.66</td>
<td>17.27</td>
<td>4.38</td>
<td>8.36</td>
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<td>64%</td>
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<tr>
<td>C-bz</td>
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<td>49.85</td>
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</tr>
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</table>

(See Fig. 11)
4.6. Discussion

4.6.1. Candidate Microstructural Processes

As mentioned in Chapter 2, chemical processes appear to contribute more than mechanical processes to bridge zone formation in the samples we have investigated. The additional samples described in this chapter provide us an opportunity to evaluate the generality of that interpretation. To that end, we explore the possible formative mechanisms for the major microstructural features associated with bridge zones.

• Finer grains in the bridge zones: A defining feature of the microstructure we term bridge zones is the finer grain size compared to the host microstructure. These fine grains in the bridge zone consist mainly of plagioclase, quartz, biotite and titanite. Each phase within the bridge zones exhibits a relatively uniform grain size (commonly <150 microns). It has long been thought that the formation of these finer plagioclase can be attributed to either: dynamic recrystallization (e.g., Tullis & Yund 1985; Tullis et al. 1990; Piazolo et al. 2002; Svaahber & Piazolo 2010); fracture (e.g., Goodwin and Tikoff 2002; Okudaira et al. 2017); or dissolution-precipitation, reaction and neocrystallization (e.g., FitzGerald & Stünitz 1993; Svaahber & Piazolo 2013; Mukai et al. 2014; Petley-Ragan et al. 2018; Soda & Okudaira 2018). Biotite recrystallization can occur by dissolution-precipitation and fracturing (Kronenberg et al. 1990; Shea & Kronenberg 1992; Mares & Kronenberg 1993; Wintsch & Yi 2002) in addition to basal slip. Grain boundary migration of biotite is also abundant at higher conditions (Bell 1998). Fine quartz is commonly attributed to grain-size reduction through dynamic recrystallization (e.g., Jessell 1987; Wu & Groshong 1991; Hirth & Tullis 1992; Stipp et al. 2002). However, some studies (e.g., den Brok 1992; Vernooij et al.
also suggest the possibility of dissolution-precipitation processes and neocrystallization (e.g., FitzGerald & Stünitz 1993; Giuntoli et al. 2018). Johnson et al., (2008) pointed out that fine-grained titanite grains were likely generated by the reaction of coarse-grained plagioclase with biotite.

• Alignment of crystallographic orientations of feldspar within and outside the bridge zone:

Crystallographic inheritance or control may be derived from subgrain rotation (e.g., Pennacchioni et al. 2010; Ceccato et al. 2017; Svahnberg & Piazolo 2010). Alternatively, neocrystallization may occur with the orientation of the adjacent grain lattice (e.g., Ree et al. 2005; Fukuda et al. 2012; Gardner et al. 2021).

• Misorientation distributions: Dislocation creep often brings about concentration of correlated misorientation magnitudes at low angles (refs), which has also been extensively demonstrated in mechanical experiments with other minerals such as quartz (e.g., Trimby 1998) and olivine (e.g., Fliervoet et al. 1999). In our samples, the crystallographic fabric of the fine grains aligns with the coarse grain orientations, but the uncorrelated misorientations do not dominate at low angle peaks (or sub-peaks). This pattern is commonly ascribed to grain size-sensitive process (e.g., Fliervoet et al. 1997, 1999; Jiang et al. 2000; Law 1990) or dissolution-precipitation, reaction and neocrystallization (more random Misorientation Angle Distribution, e.g., Hippertt & Egydio-Silva 1996; Ree et al. 2005; Vernooij et al. 2006; Fukuda et al. 2012; Gardner et al. 2021).

• CL pattern in plagioclase: In contrast to the high signal and igneous zoning in large-grained plagioclase outside the bridge zone, fine-grained plagioclase grains in the bridge zone have much lower panchromatic CL brightness. These dark fine grains contain minimal internal structure visible in either the BSE or CL images. This phenomenon is also visible in color CL and is particularly evident in the blue channel. Trace
elements (e.g., Ba, Rb, Sr, Mn, Pb, REE, Fe, Ti, Ga, etc. on, Götze 2012) and defect density changes (Watt et al. 2000; Rusk et al. 2006; Spear & Wark 2009) can influence the CL intensity.

4.6.2. The Processes Associated with Bridge Zone Formation in Different Samples

H: The plagioclase and quartz outside the bridge area show no significant plastic strain or deformation. The twining lamellae of plagioclase are straight, while the quartz shows a foam texture. The boundaries between the plagioclase and other phases, including the bridge zone, do not show bulging or curved migration boundaries (as shown in Gates & Glover 1989; Tullis & Yund 1991; Ji & Mainprice 1990). Grain reference orientation deviation, kernel average misorientation, and misorientation angle distributions of plagioclase do not have a high peak of low angles. However, the boundaries between the biotite grains and the bridge zone are not straight. The fine grains inside the bridge zone seem to be embedded in the biotite interior and form sinuous boundaries. In addition to fine-grained plagioclase, fine biotite and titanite grains are also present in the bridge zone. Fine- and large-grained biotite are optically consistent with orientation while the fine-grained titanite is located at the edge of other minerals or in the biotite cleavages. The distribution of fine-grained titanite grains and their fine tabular morphology suggest that they formed as reaction products, as interpreted by Johnson et al. (2008), rather than from the grain size reduction of large titanites.

Fine-grained plagioclase may be derived from the growth of precipitated grains, which react with biotite, gradually consume dissolved biotite, and generate titanite at the edges of biotite and plagioclase grains, in a manner similar to the precipitation processes reported by Stüinitz & Tullis (2000), and Imon et al. (2002). This neocrystallized plagioclase inherited the lattice orientation of one or the other of the two large plagioclase grains adjacent to the bridge zone, a common feature of dissolution-precipitation (e.g.,
Ree et al. 2005; Fukuda et al. 2012; Gardner et al. 2021). Since the bridge zone is a low-strain region, grain rotation /grain-size sensitive process has not begun, and the fabric has not been modified, so the CPO is not weakened. However, this does not explain the lack of low angle uncorrelated misorientations (Figure 4.6). The Ca loss caused by this reaction was reported by Johnson et al. (2008), and may be the cause of CL darkening.

D: Feldspar around Subzone1 shows some degree of bulges and subgrains, combined with low angle correlated and uncorrelated misorientations (Figure 4.7) and a weak but present CPO, suggesting that dislocation creep may be a possible contributor. Above 30˚, the congruence of correlated and uncorrelated misorientations with the theoretical random distribution may indicate chemical reactions, dissolution-precipitation mechanisms, or grain size sensitive processes, which is consistent with a weak CPO (cf. Trimby 1998; Fliervoet et al. 1999; Wheeler et al. 2001; Price et al. 2016). Both quartz and plagioclase outside Subzone 2 show some degree of subgrains, with weak misorientation at the edges (Figure 4.8). Quartz and feldspar both exhibit low angle misorientations. The low angle peak of feldspar is not as high as in Subzone 1. However, the uncorrelated Misorientation Angle Distributions of both quartz and plagioclase are close to random, indicating that dislocation creep is not dominant even though dislocation creep is a possible general mechanism for grain size reduction (Fliervoet et al. 1997; Jiang et al. 2000) and does happen in these rocks (Jonson et al. 2008). Fluid induced mass transfer or response in both subzones was involved, especially in promoting the reaction of fine plagioclase and biotite. The neocrystallization through fluid also generated parent-controlled CPO (Ree et al. 2005; Fukuda et al. 2012). This is consistent with the CL signal in the plagioclase fracture area connecting the weak phase or surrounding the weak phase, which is identical with that of fine-grained plagioclase. The contributors
differ in the two subzones suggest that the bridge zone formation mechanism is variable, even in such adjacent areas within the same rock.

C: The bridge zone is in the S-component of the S-C fabric, and the external quartz and biotite exhibit plastic deformation. Inhomogeneous quartz extinction and subgrains correlate with the grain misorientation orientation deviation and kernel average misorientation maps. The combination of low-angle misorientations and a certain degree of <40° distribution, suggests that subgrain rotation recrystallization may have led to grain size reduction and further progressive rotation (e.g., Lloyd 2004; Lloyd & Freeman 1994; Neumann 2000; Stipp & Kunze 2008; Trimby et al. 1998; Wheeler et al. 2001; Price et al. 2016). Within the bridge zone, fine-grained quartz may represent grains that have experienced grain-size sensitive chemical and mechanical processes after dislocation recrystallization or formed by dissolution-precipitation between fine-grained plagioclase grains (Fliervoet et al. 1997, 1999; Jiang et al. 2000; Law 1990; Platt 2015). The edges of plagioclase are sinuous, possibly caused by bulging as well as possible “subgrain” development, but there is no evidence showing dislocation creep dominate the processes. The correlated Misorientation Angle Distribution also has too low dataset to support the dislocation domination, though shows subpeaks at low angles. Thus, the dominant deformation processes are grain size sensitive processes or dissolution-precipitation mechanisms, which lead to a random distribution of uncorrelated Misorientation Angle Distribution. Migrating fluids may have played a significant role, delivering material to support neocrystallization of plagioclase, biotite, and titanite, and to heal fractured feldspar.

Due to the potential impact of bridge zones on the weakening of rocks, we are primarily concerned with the damage (fracturing and grain size reduction) of strong framework phases such as
plagioclase. As discussed in Section 4.5.1, feldspar grain size reduction processes have been extensively reported and discussed from experimental and natural rock samples. Johnson et al. (2008) suggested a deformation temperature of 475 ± 50 °C by microstructural observations and hornblende-plagioclase thermometry. This temperature appears to be inadequate for the dislocation creep process of feldspar generally supported by experiments (e.g., Gleason & Tullis 1993; Kruse & Stünitz 2001; Stünitz et al., 2003; Rybacki & Dresen 2004). It is important to note that this does not imply that ductile deformation of feldspar at such low temperatures is impossible (e.g., Tullis 1983; Stünitz & Fitz Gerald 1993; Prior & Wheeler 1999; Ishii et al. 2017; Hentschel et al. 2019). However, numerous researchers have demonstrated that fracturing, dissolution-precipitation, early-stage crystal plasticity, and chemical reactions facilitated by fluids are significant and considerably variable factors (e.g., Tullis & Yund 1987; Fitz Gerald & Stünitz 1993; Stünitz & Fitz Gerald 1993; Tullis et al. 1996; Prior & Wheeler 1999; Kruse & Stünitz 2001; Ree et al. 2005; Menegon et al. 2008; Kilian et al. 2011; Brander et al. 2011; Svaehnberg & Piazolo 2013; Mukai et al. 2014; Eberlei et al. 2014). The misorientation deviate in the coarse plagioclase of samples D and C and some visible "subgrains" may come from before or during initial cooling of tonalite or healing microfractures cause minor dislocations (the similar processes as mentioned in Tullis & Yund 1987 and Johnson et al. 2004), or the microfracturing driven dislocation density increase (Tullis & Yund 1985). The bulk strains of these three rocks differ greatly, and the mechanisms of bridge zone formation also vary, even in the same rocks. However, despite demonstrating the common mechanical processes such as microfracturing (e.g., Pryer, 1993; Atenberger & Wilhelm 2000), dislocation processes (e.g., Hentschel et al. 2019) and grain-size sensitive processes (e.g., Tullis et al. 1990) are expected to play a role in making incipient damage of the plagioclase framework and also increasing
permeability. Fluid-participated diffusion, precipitation and even chemical processes and reactions can make a greater contribution to form the bridge zones with the characteristics observed.

4.6.3. Bridge Zone, Stress Amplification & Modeling Limitations

Using natural geometry as the basis for calculating the stress distribution (Figure 4.11), we compare the locations of bridge zones and stress concentrations. At a shear stress of 10 MPa, the bridge zones and areas of amplified second invariant of stress have a moderate degree of colocation in either shear loading, axial loading, or both. The calculations are necessarily simplified due to the geometry reduced in complexity, the lack of robust constitutive laws for some phases, and the lack of full anisotropic flow laws for all phases. In addition, we do not explicitly take temporal evolution into account. As such, even if stress concentrations were the sole reason for bridge zone formation, the presence of bridge zones in locations that do not correlate with modeled stress concentrations does not indicate a faulty concept. Rather, we feel that the degree of correlation is a strong indicator of the viability of stress concentrations to drive bridge zone formation and subsequent rheologic change. The degree of stress amplification, at more than double the macroscale load stress, is more than sufficient to catalyze local mechanical and chemical processes that create bridge zones (cf. Naus-Thijssen et al. 2011; Johnson et al. 2004; Bercovici & Ricard 2014; Wheeler 2018).

In the modeling, we assigned the rheology of quartz to the bridge zone and also assigned the rheology of feldspar to hornblende. This may be an overestimation or underestimation of the natural rock strength. Numerically, dislocation creep-deformed quartz and diffusion creep-deformed wet plagioclase (in fine grained plagioclase grain size in the bridge zone as described in this paper) have almost the same
strength, but the bridge zones are not monophase regions and any phase mixing would complicate the estimation, especially as it is still only considering viscous regime mechanical processes. Therefore, the use of quartz constitutive law is a reasonable attempt. Hornblende is a mineral for which rheological data are extremely scarce and where natural rock environments or even numerical conditions may be dominated by brittle fractures, sliding along cleavage or even dissolution-precipitation mechanisms (e.g., Stünitz 1993; Berger & Stünitz 1996; Wintsch & Yi 2002; Lafrance & Vernon 1993; Vernon et al. 2004; Imon et al. 2004). Moreover, the anisotropy of hornblende is more pronounced than that of quartz and feldspar (Ko & Jung 2015). Our simulations reflect only isotropic materials in the viscous regime, and there are no boundary effects at different grain boundaries in the same phase.

Our numerical results show that the addition of bridge zone, even by only 2%, has a strong weakening effect on body strength by breaking the framework strong phases. We argue that heterogeneous stress distributions drive numerous geological processes such as fluid flow and reactions through pressure gradients (e.g., Kamb 1961; Naus-Thijssen et al. 2011; Wheeler 2018), and damage through failure and grain size reduction (e.g., Segall & Simpson 1986; Johnson et al. 2004, 2008; Bercovici & Ricard 2014). These variations may lead to a series of mechanical and chemical changes that mechanically connect the weaker phases by weakening the strong matrix separating them, which is the bridge zone, thus effectively reducing bulk strength (e.g., Poirier 1980; White et al. 1980; Handy 1994; Mancktelow 2002; Johnson et al. 2004; Wintsch et al. 1995; Wintsch & Yeh 2013; Fukuda & Okudaira 2013; Gerbi et al. 2016; Gardner et al. 2017).
4.6.4. Rheological Bridge Zones through the Whole Lithosphere

In Chapter 2, we described bridge zones in amphibolite facies and granulite facies rocks. In this chapter, we describe the existence of bridge zones in the greenschist facies, across a strain gradient. The bridge zone consists of phases and/or chemical compositions that are not present in the surrounding area, suggesting that not only mechanical but also chemical processes play an important role in their formation. The small modal fraction of microstructural changes represented by the bridge zone can lead to a high degree of bulk weakening and sustained deformation, which reduces the rock mass strength to trigger greater macroscopic strain.

The bridge zone concept is similar to the concept of interconnected weak phases reducing the bulk strength of rocks (e.g., Handy 1994; Holyoke and Tullis 2006; Hou et al. 2022). Bridge zones represent a rheological rather than strictly phase-based switch from a load-bearing framework to interconnected weak components. This is consistent with the absence of weak phase interconnections in the microstructure of some experimental deformation products (e.g., Jordan 1987; Bons & Urai 1994; Dell'Angelo & Tullis 1996). Natural (e.g., Wintsch et al. 1995; Wintsch & Yi 2002; Johnson et al. 2008; Marsh et al. 2009, 2011) and experimental (e.g., de Ronde et al. 2005; Gonçalves et al. 2015; Pec et al. 2016) results suggest that chemical processes as well as mechanical-chemical coupling may play an important role. Mechanisms such as grain-size reduction, metamorphism, and phase mixing are more likely to interconnect the weak phases in rocks and to be adapted to mylonitic matrix with strain (Vissers et al. 1995; Martel et al. 1999; Johnson et al. 2004; Warren & Hirth 2006; Herwegh et al. 2011; Kilian et al. 2011; Culshaw et al. 2015; Gerbi et al. 2016). These bridge zones occur in low-strain regions of even high-strain rocks, eventually consuming all load-bearing frameworks. The bridge zone model does
not rule out weak phase interconnections. Instead, we recognize that phase mixing around stronger phases (e.g., Gonçalves et al. 2015; Pec et al. 2016) and spatially variable deformation mechanisms are important for the role of new microscopic rheological domains (Jessell et al. 2005). Stress amplification between rheologically weak phases may help support bridge zone formation processes, which are sufficient to drive a variety of geologically important processes such as damage and strain localization, as well as transient permeability and fluid migration (e.g., Scholz 1968; Jordan 1987; Handy 1994; Johnson et al. 2004; Holyoke & Tullis 2006; Feng et al., in review).

4.7. Chapter Conclusion

The bridge zone acts as a relatively tabular region of fine grains that mechanically connects rheologically weak phases or domains. The important role of chemical processes in its formation results in the presence of phases and/or chemical compositions inside the bridge zone that are not present in the surrounding region. Numerical modeling demonstrates that a small modal addition of bridge zone leads to a high degree of bulk weakening for sustained deformation and that the weakening effect increases with increasing loading stress. Modeling results also show that stress amplification occurs at phase boundaries and in strong phases connecting weak phases. The stress amplification can more than triple the macroscale load stress, which could trigger rheological and metamorphic changes in the rock, thus facilitating the formation of the bridge zone.

As discussed in the previous chapter, variations of material properties can assume an important role in rock weakening. The tonalites from the Cerro de Costilla complex with different strain levels indicate that the bridge zone can exist at lower temperatures (greenschist facies). Combined with rocks
from the granulite facies and amphibolite facies of the Grenville Orogenic belt (Chapter 2), this study demonstrates that bridge zones can form throughout the crust.
CHAPTER 5

DISCUSSION AND FUTURE WORK

5.1. Characteristics and Definition of Rheological Bridge Zone

Rheological bridge zones are found under a wide range of geologic temperature and pressure conditions, arguably at all levels of the Earth's crust, based on the range of samples described in this study. More importantly, the bridge zone is found not only in undeformed/weakly deformed rocks, but also across the strain gradient.

In terms of morphology, the rheological bridge zones are relatively flat areas of reduced grain size that can either follow grain boundaries or cut through large crystals. The finer grains within these zones appear to be more equant and appear to have inherited some crystallographic properties from the coarser parent grains, albeit weakly.

Mechanically, bridge zones appear to act as links between rheologically weak phases or domains, forming interconnected networks within the rock. In the samples described here, based on standard flow laws (e.g., the plagioclase/feldspar flow laws of Rybacki et al. 2006, the quartz flow law of Hirth et al. 2001, and the biotite flow law of Kronenberg et al. 1990) and inferred deformation conditions (Canadian samples in Wodicka et al. 1996, 2000 and Jamieson et al. 1995, and the Cerro de Costilla complex samples in Johnson et al. 2008), quartz and biotite are considered the weak phases, while feldspar is considered the strong phase. Although the rheology of the bridge zone will depend on multiple factors, such as local conditions and mineral combinations, since these zones need to accommodate faster strain rates than possible without them, they must be weaker than the materials they replaced.
Chemically, the bridge zones consist of phases and/or chemical compositions that are not present in the surrounding areas. For example, in our samples, the chemical mobility of fine plagioclase grains in bridge zones in granulite (GB83C) differs from that of the parent grains, indicating possible chemical interactions. Additionally, in GB247, the presence of fine-grained plagioclase not in direct contact with coarse-grained plagioclase suggests mass transfer due to chemical processes or reactions. These observations are consistent with chemical-mechanical coupling processes documented under a variety of pressure-temperature conditions (e.g., Wintsch et al. 1995; Wintsch & Yi 2002; Johnson et al. 2008; Marsh et al. 2009, 2011) and support the growing recognition that chemical processes play a role in strain accommodation (e.g., Stünitz et al. 2001; Menegon et al. 2013).

5.2. Processes Associated with Rheological Bridge Zone Formation

The development of bridge zones at the extremities of weaker phases might instigate phenomena such as microcracks, reduction in grain size, or chemical reactions, thereby propelling the creation of linkages between these weaker phases (Handy 1994; Johnson et al. 2004; Holyoke & Tullis 2006; Svaehnberg & Piazolo 2010; Goncalves et al. 2015; Platt 2015; Gerbi et al. 2016; Hou et al. 2020) rather than a simple connectivity observed before deformation (Gerbi et al. 2016). Our deduction is that these zones, termed as bridging regions, encompass both mechanical and chemical modifications. The mechanical alterations predominantly manifest as a reduction in grain size and can also encompass alignment in crystallographic structure. Meanwhile, chemical modifications, as witnessed in our specimens, involve the inception of new phases, changes in modal or morphological attributes, and the mobility of major and trace elements. Regardless of the strain-temperature-pressure conditions of the samples in this study, the main physical and chemical processes associated with the formation of the
bridge zone are essentially similar: (1) damage and grain-size reduction of strong domains such as plagioclase, K-feldspar, and hornblende; (2) recrystallization and dislocation creep of quartz, and fragmentation and cleavage slip of biotite; (3) dissolution-precipitation/recrystallization/neocrystallization of feldspar and quartz; (4) alterations in major and trace elements within plagioclase both within and beyond the bridge zone, along with variations in the brightness under cathodoluminescence; (5) generation of new phases due to chemical reactions, such as plagioclase appearing within K-feldspar domains in the amphibolite facies sample, and the occurrence of titanite in the greenschist facies samples.

A comprehensive analysis of the potential mechanisms and processes leading to the formation of bridge zones has been extensively deliberated upon in both sections 2.8.2 and 4.5.2. Mechanical processes, such as destruction of the strong framework mineral feldspar, are certainly evident in the bridging zones developed in our samples. The high misorientation deviation of the coarse-grained feldspar boundary in the granulite facies and amphibolite facies samples in Canada and the microfractures produced by the greenschist-facies deformation of the Cerro de Costilla complex (e.g., Johnson et al. 2004, 2008) indicate that the mechanical process of cataclasis may be the dominant grain size reduction process. However, the crystallographic orientations of the fine and coarse grains outside the bridge zone are consistent, and there is no evidence to support that they experienced significant plastic deformation. This inheritance combined with the core-rim CL differences of fine-grained plagioclase shown in some samples and the presence of other phases (e.g., K-feldspar, quartz, titanite, etc.) in the intergranular or triple-junction voids of fine-grained plagioclase suggests that dissolution-precipitation or neocrystallization plays an important role in the formation of fine grains in the bridge zone (e.g., Ree et al. 2005; Fukuda et al. 2012; Gardner et al. 2021).
neocrystallization undoubtedly necessitates the participation of fluid phases. While the extent of open-system behavior within the examined specimens remains unclear, the alterations in major and trace elements within the inner and outer phases of the bridge zone, induced by diffusive and mass transfer phenomena facilitated by fluid presence, persist. This phenomenon is notably evident, as exemplified by fluid inclusion trails facilitating the mass transfer into and out of the bridge zone, particularly observable within samples of amphibolite facies. Likewise, fluids can facilitate chemical reactions. Chemical reactions are quite common in the samples involved in this study, which introduced new minerals in the bridge zone, such as plagioclase in the amphibolite-facies samples and titanite in the greenschist-facies samples.

Hence, it becomes evident that the participation of fluids or chemical factors assumes a more influential role in the mechanisms of bridge zones formation, surpassing the influence of purely mechanical processes. Naturally, the damage of the strong framework phases induced by mechanical processes concurrently serves to augment both permeability and the available surface area, thereby enhancing the efficacy of fluid-mediated interactions (e.g., Carlson and Gordon 2004). The conceptual diagram of fluid-chemical-mechanical coupling in Johnson et al. (2008) is a good description of this mechanism (Fig. 39 in Johnson et al. 2008). This model is not only for the formation of bridge zone, but also for the entire process of microstructure evolution.

Phase arrangement induced microscale stress field, including pressure distribution and stress amplification between rheologically weak phases may facilitate the formation of bridge zones, which can drive various geologically significant processes such as damage and strain localization, transient permeability, and fluid migration (Scholz 1968; Jordan 1987; Handy 1994; Johnson et al. 2004; Holyoke
& Tullis. 2006; Feng et al. under review). A conceptual illustration of weakening from stress/phase arrangement/bridge zone formation is shown in Figure 5.1.

Figure 5.1 Model diagram of rock weakening caused by phase arrangement, stress amplification and bridge zone formation. The formation of bridge zone framed as a three-way coupling among mechanical processes, diffusion/advection, and reaction (modified from Johnson et al., 2008).
5.3. Bulk Rheological Consequences of Rheological Bridge Zone Formation

We argue that bridge zones represent a rheological transition and interconnection, akin to the linking of less robust phases and the ensuing weakening mechanisms observed in multi-phase aggregates by numerous researchers (e.g., Handy et al. 1990, 1994, 1999; FitzGerald & Stünitz 1993; Wintsch et al. 1995; Kruse & Stünitz 1999; Stünitz et al. 2001; Handy & Stünitz 2002; Kenkmann & Dresen 2002; Gueydan & Frédéric 2003; Vernon et al. 2004; Johnson et al. 2004, 2006; Wintsch & Yeh 2013; Oliot et al. 2014; Platt 2015; Smith et al. 2015; Czaplinska et al. 2015; Gerbi et al. 2014, 2016). The interconnection of the weaker realm has long been acknowledged as a noteworthy contributor to strain localization (Cobbold 1977; Jordan 1987; Handy 1990; Johnson et al. 2004; Holyoke & Tullis 2006; Montesi 2013). Bridge zones represent a rheological rather than strictly phase-based switch from a load-bearing framework to interconnected weak components. This aligns with the absence of interconnections between weak phases in the microstructure of certain experimental deformation outcomes (Jordan 1987; Bons & Urai 1994; Dell’Angelo & Tullis 1996).

Both natural (Wintsch et al. 1995; Wintsch & Yi 2002; Johnson et al. 2008; Marsh et al. 2009, 2011) and experimental (de Ronde et al. 2005; Gonçalves et al. 2015; Pec et al. 2016) findings suggest that chemical processes and mechanical-chemical interactions may play crucial roles. Processes like grain-size reduction, metamorphism, and phase mixing are more likely to interconnect weak phases in rocks and adapt to mylonitic matrices with strain (Vissers et al. 1995; Martel et al. 1999; Johnson et al. 2004; Warren & Hirth 2006; Herwegh et al. 2011; Kilian et al. 2011; Culshaw et al. 2015; Gerbi et al. 2016).
Arguably the dominant agent of rheologic change is the development of new weak domains in mechanically critical areas by processes such as recrystallization, fracture, introduction of fluids, phase changes, and reactions. Nucleation and interconnection of rheologically weaker phases in a rock or other polyphase composite can reduce the bulk strength of that composite significantly. Gerbi et al. (2014 & 2016) calculated that a ~1% modal change from strong to weak phase, in interconnected domains, can induce a nearly order-or-magnitude bulk strength reduction. The calculations in this study similarly demonstrate that ~2% weak phase incorporation achieves upwards of 90% weakening.

In low-strain rocks or low-strain parts in highly deformed samples, these fine-grained zones might not be sufficient to cause essential changes in the constitutive law of strong phases (e.g., from dislocation creep to grain-size sensitive creep). However, as strain increases (or strain rate increases), as a part of the strong phase, these existing fine-grained zones are more likely to undergo grain size-sensitive processes or interaction with other phases to help the strong phase deform to accommodate the overall strain. Additionally, if a chemical reaction brings phase change, the mechanical deformation mechanism of the dominant phase (dislocation or diffusion creep) accounts for only a portion of the deformation process, and a single mineral flow-law description is insufficient to quantify the strength change. In other words, once these connections form a network, the main loading domain in a polyphase system transition from the strong support framework domain to the new weak phase interconnection domain. In addition, with the evolution of rock strain and microstructure, the distribution relationship between strong and weak phases is constantly changing, and this change will inevitably lead to the redistribution of the deformation mechanisms that affect rock strength (Figure 2.13, Gueydan et al. 2014; Platt 2015; Gerbi et al. 2016).
Gerbi et al. (2016) also indicated that this kind of interconnection is independent of scale, ranging from microscale to meter-scale, kilometer-scale, and even tectonic scale (e.g., Tullis et al. 1991; Bons & Cox 1994; Handy et al. 1999; Ji 2004; Dayem et al. 2009; Klepeis & King 2009; Gerbi et al. 2012, 2014, 2016; Culshaw et al. 2010, 2011; Marsh et al. 2011; Betka & Klepeis 2013). At the meter scale and larger, one example is a shear zone network, in which the existing shear zones (which can be regarded as weak domains) are interconnected by bridging shear zones and have a profound impact on the bulk strength of regional and even tectonic units with subsequent geometric evolution (e.g., Culshaw et al. 2011; Marsh et al. 2011, Kaatz et al. 2021; Zertani et al. 2022).

5.4. Possibilities for Future Expansion

This study has demonstrated that bridge zones can arise and develop at most levels of continental crust. We have defined the bridge zone and pointed out its potential rheological implications. This study has illustrated that the bridge zone can be generated and developed in most layers of the continental crust. We have defined the bridge zone and indicated its potential rheological effects. However, this study is only the beginning of the exploration of rock weakening from this rheologically important domains, and I believe there is potential for expansion in the following areas.

5.4.1. More Natural Candidates

To evaluate the relative impact of different weakening processes leading to strain localization and, as far as possible, establish the causative chain of events leading to the rheological change, we compiled studies from dozens of shear zones worldwide (Table 5.1, which is representative but not nearly exhaustive). As the previous references indicate, the common drivers of permanent rheological change in the viscous regime include grain size reduction, phase rearrangement, and metamorphism, with or without
hydration. From these studies, which include mm- to km-scale shear zones, we recognize that weakening leading to localization has its roots at the grain-scale, meaning the nucleation and recrystallization within local domains initiate more widespread rheologic change. The commonality is apparent, but it is better to focus on either the low strain region of the low strain sample or the high strain sample. The observation and discussion should focus on (1) verifying the existence of a bridge zone in these samples, (2) determining whether the same or similar mechanical and chemical processes mentioned in this study have occurred, (3) illustrating the rheological significance of bridge zone throughout the whole crust or even the whole lithosphere, and (4) establish quantification of different rock types mechanically and rheologically to better explain the crustal and smaller scale strain patterns throughout the lithosphere. The samples (or their protoliths) involved in this study are granite and tonalite. I am interested in exploring more naturally deformed oceanic crust or mantle rock samples. Examination of bridge zones in individual samples will be conducted through a comprehensive utilization of all available analytical methodologies, including but not limited to, optical and electron microscopy, cathodoluminescence (CL) imaging, electron backscatter diffraction (EBSD), electron probe microanalysis (EPMA), and laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS). Systematically analyzing each sample yields diverse datasets, eventually integrated into tables and diagrams, documenting distinct characteristics and disparities.

Secondly, more shear zones with sufficient knowledge of temperature-pressure-fluid-deformation histories can be added to the study to detect the evolution and development of the weak domain with strain. I propose that attention be focused on shear zone margins and shear zone networks. In order to track the initiation and development of the weak domain, we will be looking for strain gradients on deci-
to decameter scale scales across a range of lithologies and deformation conditions. Fieldwork will consist of (1) mapping outcrop-scale patterns in the shear zone network, with particular emphasis on nascent shear zones; and (2) collecting samples across strain gradients for laboratory microscale analysis. Mapping will document the geometric distribution of strain intensity, lithology, and structure. From the fieldwork, we should obtain spatial maps of outcrop-scale structural features in preparation for relating these features to the simulated stress field.

### Table 5.1 Examples of mechanisms initiating shear zones in a variety of settings and scales

<table>
<thead>
<tr>
<th>Setting</th>
<th>Causative mechanism</th>
<th>Selected References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Granitoids</td>
<td>Fracturing, localized reduction</td>
<td>Segall &amp; Simpson (1986); Tourigny &amp; Tremblay (1997); Johnson et al. (2004, 2008); Mancktelow &amp; Pennacchioni (2005); Pennacchioni (2005); Oliot et al. (2014); Smith et al. (2015); Condit &amp; Mahan (2018); This study</td>
</tr>
<tr>
<td>Retrogressed high-grade rocks</td>
<td>Fractures allowed fluid access and metamorphism to weaker phases and/or smaller grain size</td>
<td>Austrheim &amp; Boundy (1994); Jolivet et al. (2005); Jamveit et al. (2008); Marsh et al. (2011); This study</td>
</tr>
<tr>
<td>Oceanic crust</td>
<td>gran-size reduction</td>
<td>Idlefonse et al. (2007); Harigane et al. (2008); Mehl &amp; Hirth (2008)</td>
</tr>
<tr>
<td>Mantle</td>
<td>gran-size reduction, variation, and/or phase mixing; fracturing and reaction weakening</td>
<td>Vissers et al. (1995); Newman &amp; Drury (2000); Handy &amp; Stünitz (2002); Djikstra et al. (2004); Warren &amp; Hirth (2006); Czertowitz et al. (2016)</td>
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</table>

#### 5.4.2. Experimental Deformation

By applying physical experiments, we can quantitatively compare internal structural changes against modeled stress fields while also providing sound comparators for the structures in the naturally deformed rocks. One of the benefits of the experimental deformation is the ability to measure the boundary conditions, which we are rarely able to do in the natural Earth. A second benefit is the ability to
stop the deformation at low strain, prior to major structural change that might overprint features related to
the initiation of, for example, localization.

The experiments, in which we know the deformation conditions, provide an important calibration
point for the naturally deformed samples, and also a chance to directly compare the microstructures and
calculated stress-strain enhancement in the experimental samples with those from the nature samples.
Because of the interest in weak domain initiation, the deformation of the should be small, i.e., up to 10%
strain, the level at which Dell’Angelo & Tullis (1996), among many others, reported weakening in
experiments. For example, to complement the natural samples in this study, aplites (because of their
quartz-feldspar ratio, grain morphology, simple mineralogy, and grain size) will be great candidates for
experiments in a Griggs-type apparatus at confining pressure of 1.4 GPa and at a temperature of 800 °C,
with varied loading stress and/or strain rate, respectively, between 300 and 800 MPa and $1\times10^6s^{-1}$ and
$2\times10^6s^{-1}$. Post-deformation laboratory analysis will proceed as with the naturally deformed samples,
providing the datasets for explicit comparisons.
CHAPTER 6

CONCLUSIONS

Earth's lithosphere is mechanically heterogeneous, and this distribution of mechanical properties leads to a heterogeneous stress distribution. Heterogeneous mechanical properties within the lithosphere vary based on mineralogy, microstructure, and environmental conditions and have a major influence on deformation and corollary geologic processes at all scales.

We used numerical modeling of two-phase isotropic viscous systems to explore the factors influencing and controlling stress amplification and the potential magnitude of stress amplification in viscous regimes. We found the most geologically relevant amplification factors to be weak-phase spacing, rheological contrast, and loading type. Our results indicate that stress amplification can reach a factor of ~9 under specific conditions, but most of our experiments suggest amplifications at or below a factor of 2. Pressure differences across the model domains generally do not exceed ~55 MPa, but some are as high as ~110 MPa. The stress and pressure amplifications resulting from our analyses are large enough to drive not only a variety of geologically important processes such as fluid migration and strain localization, but also a series of microscale physical and chemical processes associated with rheological change.

One aspect of rheologic change is the development of microscale structures termed “bridge zones”, which mechanically link weak phases or domains and form networks in the rock. These bridge zones typically comprise an aggregation of relatively fine grains in a planar narrow band. Bridge zones result from not only in-situ grain size reduction (such as recrystallization or cataclasis), but also chemical mass transfer and reactions on a short spatial scale that introduce phases and/or chemical compositions absent in the surrounding areas. From the microstructural and chemical analysis on the natural samples
studied here, we found the participation of fluids or chemical factors assumes a more influential role in the mechanisms of bridge zones formation, surpassing the influence of purely mechanical processes. The same microstructural and associated processes can span larger temperature-pressure environments, such as greenschist facies (tonalites from Cerro de Costilla complex, Mexico), amphibolite facies (amphibolite facies gneiss from the Grenville Front Tectonic Zone, Canada), and granulite facies (granulite from the Parry Sound Domain, Central Gneiss Belt, Canada). Importantly, this phenomenon occurs not only outside the shear zone where the rock is just beginning to deform at the less deformed edges (samples in Chapter 2 and H in Chapter 4), but also in low-strain regions of highly deformed rocks (C and D in Chapter 4).

Although the rheology of the bridge zone will depend on multiple factors, such as local conditions and mineral combinations, since these zones need to accommodate faster strain rates than possible without them, they must be weaker than the materials they replaced. The spatial distribution of bridge zones suggests that the small modal fraction of microstructural change represented by the bridge zones can lead to a high degree of bulk weakening with continued deformation, thus allowing bridge zones to serve as precursors to large macroscale strain. Numerical modeling results calculated from simplified geometries of tonalites from the Cerro de Costilla complex show that a small modal addition (~2%) of the bridge zone results in a 25%-94% reduction in bulk strength. Modeling results also show that stress amplification occurs at phase boundaries and in strong phases connecting weak phases. The calculated stress amplification from the tonalites can reach more than a factor of 3 of the macroscale loading stress, which could induce rheological and metamorphic changes in the rock, facilitating the formation of the bridge zone.
Bridge zones provide an opportunity to study how globally significant geologic structures related to strain localization start and develop. This concept underlies major geologic phenomena such as plate tectonics, fluid flow, metamorphic reactions, landscape development, seismic hazards, and other important geologic processes in the lithosphere. By identifying the microscale processes facilitating weak domain development, en route to providing a more thorough evaluation of the significance of this mechanism for rheologic change, I am breaking important new ground towards developing predictive algorithms for strength evolution throughout the lithosphere.
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APPENDIX: DIFFERENTIAL STRESS AMPLIFICATION RESULTS AND DISCUSSION OF
PARAMETERS RELATED TO STRESS AMPLIFICATION IN THE PLC MODELS

A.1 Introduction

The appendix information is for supporting Chapter 3, which contains the Differential stress field
maps of BG1 with different material pairs under different shear types (Figure A.1), Differential stress
distribution maps of the Q-F system under sinistral simple shear (Figure A.2) and vertical axial
compression (Figure A.3) stress amplification factors for each geometry of each set, comparison of stress
amplification factors and pressure differences with the area of strong and weak phase contact and the ratio
of vertical to horizontal spacing of adjacent weak phases (Figure A.4), the evaluation of the independence
of n and the pre-exponential factor A (from Equation 1) on stress amplification (Figure A.5 and Table
A.1), and the comparison of stress amplification under constant strain-rate and under constant stress
(Table A.2).
Figure A.1 Differential stress field of BG1 with (a) Q-F, (b) B-F, and (c) B-Q under simple shear and (d) Q-F, (e) B-F, and (f) B-Q under vertical axial compression. The stress distribution is similar to that of second invariant stress (Figure 3.4). A reduction in stress concentration occurs as mineral pairs change from Q-F to B-F and to B-Q. Under simple shear (a, b, and c), the high stress regions connect two adjacent inclusions in the lateral and vertical directions. Under vertical axial compression, high stress bands connect the inclusions diagonally, and an X-shaped high stress domain occurs in the area surrounded by four adjacent inclusions.
Figure A.2 Differential stress distribution maps of the Q-F system under sinistral simple shear (a-l) and stress amplification factors for each geometry of each set (m-q). See Figure 3.2 for phase distributions used in maps (a-l). See Figure 3.3 for a graphical explanation of the calculation of amplification factor (same as 2nd invariant stress). Line colors (m-q) indicate the column in which the geometry is located. In general, the stress distribution with respect to the shape (SH) and position (PO) is consistent, with minor distortions compared to BG1. Changes in orientation (OR) and spacing (SP) show significantly higher stress amplification (hot colors in (g-l)). The lines in (m-q) indicate a gradual decreasing trend from Q-F to B-F to B-Q for all geometries. In (o), (p), and (q), the offset of each line with respect to BG1 (blue) is larger than that in (m) and (n). Note the changes in the ordinate range in (m)-(q). Amplification factors lie between 1 and 8.66.
Figure A.3 Differential stress distribution maps of the Q-F system under vertical axial compression (a-l) and stress amplification factors for each geometry of each set (m-q). See Figure 3.2 for phase distributions used in maps (a-l). See Figure 3.3 for a graphical explanation of the calculation of amplification factor (same as 2nd invariant stress). Line colors (m-q) indicate the column in which the geometry is located. The stress distribution with respect to the positions of the inclusions is consistent, with minor changes compared to BG1. The amplification factor is lower than under simple shear. Amplification factors lie between 1 and 1.8.
Figure A.4 Comparison of stress amplification factors (a-d) and pressure differences (e-h) with the area of strong and weak phase contact (orange lines) and the ratio of vertical to horizontal spacing of adjacent weak phases (blue lines). See Table 3.4 for data. The contact area and the ratio of vertical to horizontal spacing don’t have a strong correlation with stress amplification or pressure difference.
Figure A.5 Exploration of part of generic parameter space to evaluate the independence of $n$ and the pre-exponential factor $A$ (MPa$^n$s$^{-1}$) on stress amplification (using Eq.1). We find that stress amplification increases with increasing $A$ for a given value of $n$. Varying $n$ with a constant $A$, the stress amplification does not follow a uniformly predictable pattern. For the values we tested, $A$>~100 produced increasing amplification with decreasing $n$, whereas we observed the opposite for $A$<~100 (Table A.1).
Table A.1 2nd invariant stress amplification for different combinations of $A$ and $n$

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Note: results of 10MPa sinistral simple shear loading for BG1 at 773°K
Table A.2 Stress amplification under constant strain-rate vs. under constant stress

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a. All results are at 773\(^\circ\)K;
b. Changing strain rate to control the 2nd invariant of bulk stress around 10 MPa for each geometry;
c. Constant strain-rate is pure shear strain-rate, and constant stress is vertical axial compression.
BIOGRAPHY OF THE AUTHOR

He Feng was born and raised in Tanggu, Tianjin, China, graduating from Tanggu No.1 High School, Tianjin, China. He attended the China University of Geosciences, Beijing, and graduated in 2015 with a Bachelor’s degree in Geology. He entered the Structural Geology graduate program at the China University of Geosciences, Beijing, in 2015 and received a Master of Science degree in 2018. He Feng is a candidate for the Doctor of Philosophy degree in Earth and Climate Sciences from the University of Maine in December 2023.