Chapter 1
From finite to incremental strain: Insights into heterogeneous shear zone evolution

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1.1 INTRODUCTION

Heterogeneous ductile shear zones are very common in the Earth's lithosphere and are particularly well exposed in mountain belts (e.g. Iannace and Vitale 2004; Yonkee 2005; Vitale et al. 2007a,b; Okudaira and Beppu 2008; Alsleben et al. 2008; Sarkarinejad et al. 2010; Kuiper et al. 2011; Dasgupta et al. 2012; Zhang et al. 2013; Samani 2013; Mukherjee 2013, 2014; also see Chapter 9), where they provide useful tools for a better understanding of the processes and parameters controlling strain localization, type of deformation, and rock rheology. The occurrence of strain markers such as fossils, ooids and ellipsoidal clasts in sedimentary rocks, or equant minerals, deflected veins and dykes in igneous rocks, allows one to quantify the finite strain by means of various methods (e.g. Dunnet 1969; Fry 1979; Lisle 1985; Erslev 1988; Vitale and Mazzoli 2005, 2010).

Finite strains are all quantities, directly measured or derived, related to the final state of deformation. These finite quantities, such as strain ratio, effective shear strain (*sensu* Fossen and Tikoff 1993), and angle θ' between the shear plane and oblique foliation in heterogeneous ductile shear zones, cannot furnish unequivocal information about the temporal strain evolution (i.e. strain path; Flinn 1962). This is because there are several combinations of deformation types such as simple shear, pure shear and volume change, that can act synchronously or at different times, leading to the same final strain configuration (Tikoff and Fossen 1993; Fossen and Tikoff 1993; Vitale and Mazzoli 2008, 2009; Davis and Titus 2011). Appropriate constraints are needed to obtain a unique solution – or at least reduce the under-determination. This also implies introducing some assumptions in the definition of the strain model. The strain path may be envisaged as a temporal accumulation of small strain increments, and the final strain arrangement as the total addition (Ramsay 1967). A possible relationship between final strain configuration and temporal evolution (i.e. incremental strains) was suggested by different authors, such as Hull (1988), Mitra (1991) and Means (1995). The latter author envisaged strain softening/hardening as the main rheological control on shear zone evolution: shear zones characterized by a thickness decreasing with time (Type II) result from strain softening, whereas shear zones characterized by increasing thickness (Type I) are produced by strain hardening (Means 1995). Based on this view, each part of a heterogeneous ductile shear zone is the result of a different strain evolution, and taken all together, the various shear zone sectors may be able to record the whole strain history.

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Chapter 2
How far does a ductile shear zone permit transpression?

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2.1 INTRODUCTION

Understanding the shear zone kinematics has enormous implications in interpreting a wide variety of geological processes, ranging from the exhumation of deep crustal rocks to the formation of sedimentary basins. Kinematically, ductile shear zones are defined as regions marked by localization of intense non-coaxial deformations. Considering a homogeneous strain model, Ramberg (1975) first provided a theoretical analysis of the general non-coaxial deformations by combining pure shear and simple shear flows. Based on the kinematic vorticity number, expressed as: $W_k = \dfrac{W}{\left[2\left(\dot{\varepsilon}_1^2 + \dot{\varepsilon}_2^2 + \dot{\varepsilon}_3^2\right)\right]^{\frac{1}{2}}}$ (Truesdell 1954), where $\dot{\varepsilon}_i$ is the principal longitudinal strain rate and W is the magnitude of the vorticity vector, Ramberg (1975) has shown characteristic flow patterns in ductile shear zones. His analysis derives W_k as a function of the ratio (S_r) of pure and simple shear rates, and the orientation of the principal axes of pure shear with respect to the simple shear frame. For $W_k = 1$, shear zone deformations are characterized by shear-parallel flow paths, implying simple shear kinematics. On the other end, non-coaxial deformations in shear zones with $W_k < 1$ develop open hyperbolic particle paths, which transform into closed paths as $W_k > 1$. However, deformations with $W_k > 1$ described as a pulsating type, have been rarely reported from natural shear zones. Ramsay and his co-workers included volume strain in the kinematic analysis of ductile shear zones (Ramsay and Graham 1970; Ramsay 1980; Ramsay and Huber 1987). A range of natural structures (at both micro- and mesoscopic scales), for example, anastomose mylonitic fabrics (Gapais et al. 1987), porphyroclast tail patterns (Ghosh and Ramberg 1976; Simpson and De Paor 1997; Passchier and Simpson 1986; Mandal et al. 2000; Kurz and Northrup 2008), and instantaneous strain axis (ISA) (Passchier and Urai 1988; Tikoff and Fossen 1993; Xypolias 2010, and references therein) have been used to explain these structures and to demonstrate the effects of pure and simple shear kinematics in ductile shear zones. Similarly, a parallel line of studies has dealt with structural (e.g. stylolites; Tondi et al. 2006) and chemical criteria (e.g. enrichment of immobile elements; O'Hara and Blackburn 1989; Mohanty and Ramsay 1994; Srivastava et al. 1995; Fagereng 2013) to determine the volume loss in shear zones.

In many tectonic settings shear zones developed under general non-coaxial deformation show components of shear and shortening parallel and orthogonal to the shear zone boundaries, respectively. From a kinematic point of view, these shear zones have been

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Chapter 3
2D model for development of steady-state and oblique foliations in simple shear and more general deformations

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3.1 INTRODUCTION

The use of grain shape foliations within shear zones as a means to quantify non-coaxial progressive deformation and to determine sense of shear is now well established in the literature (see Passchier and Trouw 2005, and references therein). The development of foliations in shear zones is generally categorized into strain sensitive fabrics that record the complete finite strain history of a shear zone, and those strain insensitive fabrics that do not record the full strain history. One of the most common strain insensitive fabrics observed in low to medium grade mylonitic shear zones is a microscopic oblique foliation that is typically preserved in aggregates of small dynamically recrystallized grains (Means 1980, 1981, Lister and Snoke 1984). Typically this grain shape preferred orientation (GSPO) fabric is defined by aligned sub-grains in monomineralic aggregates or layers within mylonitic shear zones. Oblique foliations are typically developed in quartz (Law et al. 1990; Lister and Snoke 1984; Dell Angelo and Tullis 1989; Mukherjee and Koyi 2010; Mukherjee, 2013) and calcite aggregates (de Bresser 1989) but examples of oblique fabrics in olivine peridotites have also been recorded (Van der Wal et al. 1992). The angle between this fabric and the plane of shear (fabric attractor) typically ranges from 20° to 40° but can be as high as 60° and lower than 5° (Passchier and Trouw 2005). Oblique foliations are thought to represent fabrics that have reached a steady state as a consequence of two competing sets of processes; on one hand those processes that relate to grain elongation and shape fabric development that are expected in a non-coaxial strain environment competing against those dynamic recrystallization processes such as grain boundary migration that tend to counteract the development of the expected strain sensitive fabric (Means 1980, 1981; Ree 1991). Oblique fabrics are often assumed to be a product of multiple progressive deformation cycles, whereby a fabric is being continually created following a finite strain path and destroyed by subsequent or concomitant dynamic recrystallization. Thus, oblique fabrics are thought to represent a stable steady state, where the orientation and intensity of the foliation, once established, do not significantly change over the strain history of a given shear zone. As such, the orientation of these fabrics in the sense of shear (SOS) plane of a shear zone will therefore stabilize and lie in an intermediate position somewhere between the orientation of instantaneous stretching axis (ISA) of the instantaneous strain ellipse and the finite stretching axis (FSA) of the finite strain ellipse. The implication drawn from this is that the orientation of an oblique fabric with respect to the shear plane is a function of where the clock foliation destroying processes (the parameter α) and also the type of deformation (β) or kinematic vorticity number (W_k), assuming a passive response by grains to deformation.

On the other hand, if the kinematics of deformation can be confidently assumed, it is possible to estimate the relative strength of the deformation (α) and the competency (μ_r) of the inclusions.

The method and analysis presented in this chapter leads naturally to many practical applications. For example, in a tectonic terrain containing a suite of suitable shear zones occurring at different structural levels, and assuming simple shear, then it may be possible to compare the viscosity ratio of host to clast material and the strength of foliation-destroying processes between shear zones. Another possibility is that, in cases where passive behavior can be confidently assumed, oblique foliations may be used to estimate kinematic vorticity number and strength of foliation-destroying processes.

Analysis of experimental and natural examples are consistent with the developed models. In particular the measured trajectory of steady state aspect ratio and orientation in experimental data follows theoretical trajectories for estimated parameters.

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Chapter 4
Ductile deformation of single inclusions in simple shear with a finite-strain hyperelastoviscoplastic rheology

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4.1 INTRODUCTION

The behavior of deformable and rigid inclusions in shear flows poses a fundamental geological problem and has thus attracted attention in the respective literature for almost a century (for a recent review, see Marques et al. 2014). As highlighted in the literature summary tables of Jessell et al. (2009) and Griera et al. (2013), the majority of work has focused on rigid or deformable purely elastic or viscous inclusions in a purely elastic or viscous matrix (e.g. Jeffery 1922; Gay 1968; Bilby and Kolbuszewski 1977; Schmid and Podladchikov 2003). However, it seems generally accepted that the long-term, inelastic deformation of the lithosphere is not only viscous but sensitive to elastic and plastic contributions to rheology (Moresi et al. 2002; Kaus and Podladchikov 2006; Regenauer-Lieb et al. 2006, 2011; Schmalholz et al. 2009; Schrank et al. 2012). Therefore, the question arises if and how the addition of elasticity and plasticity to rheology affects the inelastic deformation behavior of inclusions in shear. This work aims to provide a systematic reference study of the large deformation of single, initially round, fully bonded, deformable inclusions in isothermal two-dimensional (2D) simple shear with Dirichlet boundary conditions (constant velocity) and a hyperelastoviscoplastic rheology (Karrech et al. 2011b, c).

The prefix “hyper-” indicates that the stresses are derived from the strain energy potential and not simply assumed to be a single-valued function of strain (e.g. chapter 2 of Houlsby and Puzrin 2007). The consideration of elasticity at large strains poses a particular challenge because the mathematical treatment of large transformations requires an objective formulation of the stress rate considering both advective and corotational terms (e.g. Mühlhaus and Regenauer-Lieb 2005; Beuchert and Podladchikov 2010). An

2003), and rheologically relevant feedbacks such as shear heating (Fleitout and Froidevaux 1980; Regenauer-Lieb and Yuen 1998). Moreover, it will be worthwhile to explore the parameter space for viscosity ratios <10 more closely because Bilby and Kolbuszewski (1977) noted three characteristic regimes in AR–ϕ space for simple shear of linear viscous systems. Studies on natural (ten Grotenhuis et al. 2003; Mukherjee 2011) and modeled (ten Grotenhuis et al. 2002; Treagus and Lan 2003) mineral fish demonstrate the importance of (non-circular) shape for inclusion deformation, which highlights another research avenue.

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Chapter 5
Biviscous horizontal simple shear zones of concentric arcs (Taylor–Couette flow) with incompressible Newtonian rheology

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5.1 INTRODUCTION

Fluid caught between rotating cylinders has been intriguing physicists for over 300 years…

R.J. Donnelly (1991)

Ductile shear zones have so far been modeled mainly as zones of single lithology and with straight parallel and rigid boundaries (Ramsay 1980). Following this, thermal models of ductile shear zones were also provided (Fleitout and Froideavaux 1980). However, (i) natural shear zones can have curved boundaries in regional-scale, and (ii) may consist of more than one lithology. For example, crustal cross-sections of collisional orogens deduced from geophysical studies reveal shear zones with curved boundaries (Beaumont et al. 2001 and references therein). On the other hand, pronounced ductile shear segregates specific mineral assemblages for polymineralic rocks into zones with their interfaces parallel to the shear zone boundaries (Druguet et al. 2009). Layered shear zones have been reported/studied in granulite facies rocks (Ji et al. 1997), in models with ice (Wilson et al. 2003), from collisional terrains (Mukherjee and Koyi 2010), and in granular materials (Börzsönyi et al. 2009), besides most common cases of micaceous minerals alternating with quartzofeldspathic minerals in mylonites (Lister and Snoke 1984). Those two natural cases (i) and (ii) have recently been modeled individually (Mukherjee and Biswas 2014; Mulchrone and Mukherjee, in press) to deduce velocity profiles and shear senses. This work considers the two cases together to deduce and interpret velocity profiles of biviscous curved ductile simple shear zones. We do not address here shear zone related folds (see Mukherjee et al. 2016, Chapter 12).

5.2 THE MODEL

We use the Taylor–Couette flow model (Taylor 1923) to explain the kinematics of biviscous curved shear zone, as follows. Consider a ductile shear zone with concentric circular boundaries of radii R_1 and R_2 ($R_1 > R_2$) with two immiscible incompressible Newtonian viscous fluids within: an outer layer of fluid A with a viscosity μ_a, and an inner

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Chapter 6
Quartz-strain-rate-metry (QSR), an efficient tool to quantify strain localization in the continental crust

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6.1 INTRODUCTION

How continental crust and lithosphere absorbs ductile deformations is debated. In particular, how far deformation in the middle and deep crust localizes in narrow shear zones or is broadly distributed is discussed. Some see the continental crust as coherent blocs separated by fault zones where most of the deformation is absorbed (e.g. Tapponnier et al. 2001), while others perceive it as a continuous viscous medium where deformation is widely distributed (e.g. Beaumont et al. 2001; Mukherjee 2012). If GPS studies constrain the short-term repartition of deformation at the surface of the continents, we know less about deeper and longer-term deformations significant for the geological history of continents. This is because even if many theories and descriptions of ductile deformations exist (e.g. Ramsay 1980; Mukherjee 2012, 2013), quantification of their amount and furthermore rate are scarce. Indeed, ductile deformation rates in natural settings have been effectively measured in only three cases (Christensen et al. 1989; Müller et al. 2000; Sassier et al. 2009). However, Boutonnet et al. (2013) proposed recently a method to measure deformation rates in quartz bearing rocks deformed in the dislocation–creep regime, which could be used in numerous ductile shear zones.

This method, called quartz-strain-rate-metry (QSR), relies both on a piezometer, a flow law calibrated for quartz dislocation-creep recrystallization, and precise measurements of the temperature of deformation (Boutonnet et al. 2013). Such a method was formalized from laboratory experiments that quantitatively describe the properties of quartz at millimeter scales and at deformation rates of ~10^{-6} s^{-1}. A first set of experiments established piezometer relationships linking the size of recrystallized grains to the applied stress (e.g. Twiss 1977; Stipp and Tullis 2003) while a second set established power flow laws linking the stress to the temperature and the deformation rate (e.g. Hirth et al. 2001; Gleason and Tullis 1995; Paterson and Luan 1990; Luan and Paterson 1992). However, extrapolating from the scale of the experiment to the scale of the natural shear zones is a considerable leap across 8–10 orders of magnitude for the deformation rate, in order to reach the natural values of ~10^{-14} s^{-1}. Furthermore, for a given crystal size and a given temperature, results of the QSR vary by five orders of magnitude, depending on the piezometer and power flow law that are chosen (Jerabek et al. 2007). In resolving that problem, Boutonnet et al. (2013) performed an empiric calibration of the QSR method

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Chapter 7
Thermal structure of shear zones from Ti-in-quartz thermometry of mylonites: Methods and example from the basal shear zone, northern Scandinavian Caledonides

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7.1 INTRODUCTION

Structural evolution of collisional orogens depends closely on how major shear zones develop and evolve. These shear zones, in turn, control aspects of the geometry and thermal structure of the orogenic wedge. Consequently, analysis of the thermal structure along major shear zones can discriminate among large-scale thermal–mechanical models. The Himalayan orogen provides one example where two competing orogenic models – channel flow vs. critical taper – predict vastly different thermal gradients along the underlying master decollement. Himalayan thrusts are not generally exposed for long distances across strike, however, so accurate determination of down-dip thermal gradients is difficult. Within this context, the Caledonian orogen in Norway and Sweden, which is often compared with the Himalaya, provides an unparalleled opportunity to investigate down-dip temperatures. Due to the unusual and consistent exposure of thrust surfaces for ~140 km across strike, thermometry along the deeply eroded remains of this late Silurian-early Devonian orogen (Gee and Sturt 1985) provides a unique opportunity to test competing models of the thermal and kinematic evolution of shear zones and orogens. Here, we report temperatures of deformation in ductilely sheared rocks (mylonites) from a northern transect across the Scandinavian Caledonides using the titanium-in-quartz thermobarometer (TitaniQ; Wark and Watson 2006; Thomas et al. 2010) to investigate the dynamics of quartz recrystallization during shear and to discriminate among competing thermal models for orogenic evolution.

TitaniQ offers several advantages over other thermobarometers. First, it can be applied over a wide range of rocks because quartz, the only phase that requires analysis, is stable over a large range of temperatures and pressures. Second, domains with different Ti-content can be readily identified and targeted for analysis because cathodoluminescence (CL) intensity correlates with trace element content (Rusk et al. 2008; Spear and Wark 2009; Kohn and Northrup 2009). Last, TitaniQ is unusually precise (±3°C at a specified pressure; Wark and Watson 2006).

We evaluated TitaniQ temperatures in mylonites (T_m) from the well-exposed basal thrust zone (BTZ) of the northern Scandinavian Caledonides (Fig. 7.1) and compared them with

(Northrup 1996a) and is evidenced by spreading lineations in all directions. Therefore, we suggest that the Caledonides dominantly formed by critical wedge kinematics accompanied by gravitational spreading of the nappes, but without significant thermal contributions from channel flow.

7.5.4 Future work

Our results provide a new approach for understanding the dynamics of thrust movement. Most studies of temperatures and temperature gradients focus on (oblique) cross-sections upwards or downwards through a section. Not all such gradients help discriminate among thermal models, for example the temperature gradient across the Main Central (basal) thrust of the Himalaya can be explained by nearly any model, including channel flow or critical taper (Kohn 2008). TitaniQ in deformed quartz now allows exploration of thermal gradients *along* the transport direction of a thrust to be determined. In the case of critical taper vs. channel flow, the transport-parallel gradient is far more diagnostic than cross-structure gradients. Future work could explore this novel approach in other orogens to discriminate mechanisms of heat transport and models of crustal deformation and evolution.

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Chapter 8
Brittle-ductile shear zones along inversion-related frontal and oblique thrust ramps: Insights from the Central–Northern Apennines curved thrust system (Italy)

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8.1 INTRODUCTION

High-strain deformation within the Earth's crust often occurs in localized, narrow, and sub-parallel wall-sided zones known as shear-zones, which accommodate differential movement during the deformation of the lithosphere. They may be related to any tectonic regime (compression, extension, or strike–slip), varying in width from microns/millimeters (grain-scale) to kilometers (mega-shears). The heterogeneous character of natural deformation in shear zones produces characteristic fault rocks as mylonites and cataclasites, developed under deep-seated (10–25 km deep) ductile (viscous) or shallow-crustal (0–15 km deep) brittle–ductile (frictional–viscous) deformation regimes, respectively (e.g. Ramsay and Graham 1970; Sibson 1977, 1983; Ramsay 1980; Alsop and Holdsworth 2004).

The analysis of brittle–ductile and ductile shear zones exhumed and/or extruded and exposed at the surface through a variety of approaches and across a range of scales is essential for unraveling deformation histories. Deciphering the kinematic significance of deformation fabrics within fault rocks and reconstructing the regional tectonics contribute profoundly to understand how localized crustal deformation occurs (e.g. Casas and Sàbat 1987; Alsop et al. 2004; Carosi et al. 2004; Iacopini et al. 2008; Mukherjee 2007, 2010a,b, 2011, 2013a, b, c, 2014a, b; Mukherjee and Koyi 2010a,b; Calamita et al. 2012a; Tesei et al. 2013).

In this chapter the geometric and kinematic characteristics of shear deformation fabrics associated with frontal and oblique ramps belonging to curve-shaped thrusts are described. A detailed mesoscale structural and kinematic analysis is presented by examining some remarkable examples of brittle–ductile thrust shear zones related to regional-scale frontal and oblique thrust ramps in the Central–Northern Apennines of Italy.

8.1.1 Thrust shear zone fabrics

Brittle–ductile and ductile shear zones related to thrust faults generally develop under dominant simple shear deformation (Mukherjee 2012a,b, 2014c; Mukherjee and structures developing within in-sequence thrusting. In addition, S-C fabrics associated with sub-simple shear deformation may develop in the footwall of reactivation-related oblique thrust ramps along gently-propagating splay thrusts within push-up inversion structures.

This study provides insights and potentially represents analogs when examining shear fabrics of brittle–ductile shear zones associated to frontal and oblique ramps within curved thrust systems belonging to thrust belts that enjoyed structural inheritance of extensional faults, as in the Apennines.

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Chapter 9
Microstructural variations in quartzofeldspathic mylonites and the problem of vorticity analysis using rotating porphyroclasts in the Phulad Shear Zone, Rajasthan, India

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9.1 INTRODUCTION

The Phulad Shear Zone (PSZ) is situated within the Delhi mobile belt in the state of Rajasthan in western India (Fig. 9.1). The shear zone is marked by the occurrence of mylonite, containing layers of metamorphosed siliciclastic rocks within a calcareous matrix (Ghosh et al. 1999, 2003; Golani et al. 1998; Roy and Jakhar 2002; Sengupta and Ghosh 2004, 2007). The mylonitic foliation has a general attitude of 035°/70°E. There is strong down-dip stretching and striping lineation parallel to the transport direction. Our detailed study of mesoscopic structures in this area shows the presence of at least three generations of reclined folds and a later generation of subhorizontal folds (Ghosh et al. 1999, 2003; Sengupta and Ghosh 2004, 2007). These folds developed in the mylonitic foliation during progressive ductile shearing with their axial surface parallel to the foliation. Asymmetry of mesoscopic structures indicates a thrusting sense of movement towards NW and a subhorizontal vorticity vector (Ghosh et al. 1999; Sengupta and Ghosh 2004).

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Chapter 10
Mineralogical, textural, and chemical reconstitution of granitic rock in ductile shear zones: A study from a part of the South Purulia Shear Zone, West Bengal, India

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10.1 INTRODUCTION

Ductile shear zones are storehouse of deformation fabrics that provide critical information on the mechanism of deformation operated at meso- to microscale, the rheology of rocks at different physicochemical conditions, and the processes that help exhumation of deep-seated rocks at shallower crustal levels (Vernon 2004; Passchier and Trouw 2005; Mukherjee 2012, 2013). Ductile shear zones also act as conduits for extraneous fluids and melts (Vernon 2004; reviewed in Harlov and Austrheim 2013). These fluids are not in equilibrium with the rocks they infiltrate and hence induce significant changes in chemistry (major, trace, and isotope), mineralogy, and rheology, and seismic properties of the host rocks (cf. Harlov and Austrheim 2013 and Vernon 2004). Fluid flow along crustal-scale ductile shear zones, therefore, has important consequences for thermal and chemical evolution of the continental crust (cf. Harlov and Austrheim 2013). Granitic rocks in the hydrated ductile shear zone display a plethora of microstructures and reaction textures that provide valuable insight about fluid flow, deformation kinematics, and mass transport in shear zones (cf. Vernon 2004; Passchier and Trouw 2005; Harlov and Austrheim 2013; Mukherjee 2013). The South Purulia Shear Zone (SPSZ) of the East Indian shield exposes complexly folded, sheared, and dismembered rock suites including metapelites, metabasites, granitoids, alkaline rocks, and carbonatite. No age data exist to fix the timing of deformation and metamorphism. One deformed and metamorphosed alkaline rock suite at Sushina has been dated to be 0.93 Ga (Reddy et al. 2009; Chatterjee et al. 2013).

In this chapter we present microstructures and reaction textures of a porphyritic granite that emplaced and subsequently deformed and metasomatized within the SPSZ. Deformation and petrologic attributes of the porphyritic granite are thus the subject of the present communication. Integrating the results of microstructural analyses, textural modeling, mass balance calculations, and numerical modeling of representative bulk rock composition, we tried to trace the physicochemical changes of porphyritic granite during the evolution of the SPSZ.

10.2 GEOLOGY OF THE AREA

The SPSZ, a part of the ~E W to ESE–WNW trending Tamar-Porapahar lineament of the east Indian shield occurs at the interface between two crustal blocks with contrasting

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Chapter 11
Reworking of a basement–cover interface during Terrane Boundary shearing: An example from the Khariar basin, Bastar craton, India

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11.1 INTRODUCTION

Zones of tectonic convergence or divergence may both develop sedimentary basins (Watts 1992). In convergent zones, crustal loading by thrusting commonly flexes the lithosphere and forms foreland basins (Fowler 1990; Naylor and Sinclair 2008). On the other hand, rifting is an alternative important mechanism that may form basins of extensional origin (McKenzie 1978; Roberts and Bally 2012). Since thrusting and rifting are commonly associated with processes of continental collision and break-up, large sedimentary basins may sometimes be correlated with global-scale tectonic events. In such cases, the time–space evolution of an amalgamated assembly within a Precambrian shield, or of previously adjacent, now disintegrated landmasses currently located in geographically separated continents, can be inferred from the spatial and temporal evolution of these sedimentary basins.

Important in this respect are the Proterozoic sedimentary successions of Peninsular India, preserved in the Vindhyanchal-, Cuddapah-, Chhattisgarh-, Khariar-, Indravati-, Pranhita-Godavari, Bhima-, and Kaladgi basins (Fig. 11.1), that are collectively referred to as “Purana basins” (Holland 1906; Ramakrishnan 1987; Kale 1991; Chaudhuri et al. 1999; Ramakrishnan and Vaidyanadhan 2008). The association of these basins with the underlying basement rocks of the Bastar, Dharwar, and Bundelkhand craton make them particularly interesting, since the cratonic rocks are in juxtaposition with the lithologic ensemble of the Proterozoic Eastern Ghats Mobile Belt (EGMB) in an intensely tectonized set up. The evolutionary history of the polychronous, multiply deformed EGMB is marked by a peak Grenvillian age, granulite facies metamorphism (ca. 1000 Ma, Kelly et al. 2002; Mezger and Cosca 1999; see Gupta 2012 for review) that affected almost the entire lithologic ensemble of the northern segment of the granulite belt: this segment is referred to as the Eastern Ghats Province (Dobmeier and Raith 2003). This was followed by upper-amphibolite to granulite facies reworking (Bhadra et al. 2003) related to the juxtaposition of the Eastern Ghats Province (EGP) with the cratonic nucleus of peninsular India during the Pan-African orogeny (ca. 550 Ma). The western boundary of the EGP has been described as a thrust (Gupta et al. 2000; Bhadra et al. 2004), although

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Chapter 12
Intrafolial folds: Review and examples from the western Indian Higher Himalaya

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12.1 INTRODUCTION

Folds are perhaps the most intensively studied structures in geology (for example Ramsay 1967; Ez 2000; Harris et al. 2002, 2003, 2012a,b; Alsop and Holdsworth 2004; Mandal et al. 2004; Carreras et al. 2005; Bell 2010; Hudleston and Treagus 2010; Godin et al. 2011). Depending on morphologies and orientations, folds can be classified using several schemes (reviews by Ghosh 1993; Davis et al. 2012, etc.). Besides their rheological aspects, deciphering whether folds inside any shear zones are produced by shear has been emphasized (e.g. Mandal et al. 2004; Carreras et al. 2005; Bell et al. 2010). A couple of shear zone models altogether neglected fold formation within them, for example Koyi et al. (2013), Mukherjee and Biswas (2016, Chapter 5), Mulchrone and Mukherjee (in press). Mukherjee (2012a, 2014a) investigated the issue in terms of deformation of inactive markers in inclined shear zones undergoing extrusion and subduction. Folds related to shear zones are broadly of two types: (i) those with low interlimb angles and with significantly curved hinge lines developed before shear, some of which are sheath folds; and (ii) flow perturbed syn-shear folds that may be overturned and "intrafolial" (Alsop and Holdsworth 2004). In shear zones, locally overturned isoclinally folded foliations bound by straight foliation planes are most commonly called "intrafolial folds" (*intra* = inside; *folia* = foliation) (Dennis 1987; Allaby 2013). Intrafolial folds are found most commonly in mylonites (Trouw et al. 2000). Such folds have also been reported from cataclasites and obsidian (Higgins 1971), deformed soft sediments (Jirsa and Green 2011), slump structures (Woodcock 1976) and debris flows (Gawthrope and Clemmey 1985). The vergence of these folds is in conformity with shear sense of the shear zones they occur in. Intrafolial folds are disrupted to rootless folds if shear is more intense than in the adjacent layers even on a local scale. The adjacent rocks might be undeformed as well (Neuendorf et al. 2005). These folds tightened as shear continued (Longridge et al. 2011). Early references to classical intrafolial folds as "drag folds" (e.g. fig. IX-38 in Hills 1965) were subsequently not followed. Carrerras et al. (2007) viewed intrafolial folds both as "syn-shear folds", and "shear-related late folds" (their fig. 1c). Depending on other mechanisms perceived for intrafolial folds, they have also been described as "intrafolial strain-slip folds" (Ratcliffe and Harwood 1975) and "intrafolial shear folds" (Keiter et al. 2011). Intrafolial folds can tear apart by pronounced shear into rootless folds showing

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Chapter 13
Structure and Variscan evolution of Malpica–Lamego ductile shear zone (NW of Iberian Peninsula)

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13.1 INTRODUCTION

The Malpica–Lamego Ductile Shear Zone (MLDSZ) is a polyphasic structure developed in the NW of the Iberian Peninsula. The northern sector (NW Spain) of this shear zone is associated with a complex geological history that ranges from earlier Variscan to the emplacement of the Variscan nappes (allochthonous). Variscan intracontinental deformation after the allochthonous nappes emplaced, recognizably strike–slip, explains its spatial and structural continuity from the NW (Malpica, Spain) to the South (Lamego, Portugal). Here we focus on the Variscan development of the MLDSZ and characterize, longitudinally, this crustal mega-structure.

At the NW tip of the MLDSZ was considered an area (segment 1) with very good exposure of the Variscan intracontinental strike–slip component on outcrops of the coastline (e.g. Llana-Fúnez 2001). In sequence, we address segment 2, with excellent left-hand kinematic markers (shear band boudins and folds), on a metapelitic rock series (Pamplona and Rodrigues 2011). In the southernmost part of the MLDSZ, strike–slip deformation, emplacement of granitic rocks (segment 3 – based on Simões 2000; and segment 5), regional structures (segment 4: based on Coke et al. 2003), and secondary manifestations of the southern tip of MLDSZ (segment 6) were studied.

Detailed geological mapping was done. Kinematic interpretation of structures in micro- (Passchier and Trouw 2005; Mukherjee 2011) and mesoscale (e.g. Spry 1969; Goscombe and Passchier 2003; Mukherjee and Koyi 2010), quartz c-axis analysis (e.g. Schmid and Casey 1986), thermodynamic conditions inferred from textural equilibrium (e.g. Winkler 1979), magmatic fabrics (Fernández 1982), and U-Pb geochronology on zircons and monazites (Heaman and Parrish 1991), are integrated.

Along the MLDSZ lineament, and mainly on its southern tip, brittle–ductile and brittle structures (e.g. Penedono segment) were described that are not directly related with Variscan activity of MLDSZ. However, they are possibly due to the late Variscan or Alpine deformation.

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Chapter 14
Microstructural development in ductile deformed metapelitic–metapsamitic rocks: A case study from the greenschist to granulite facies megashear zone of the Pringles Metamorphic Complex, Argentina

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14.1 INTRODUCTION

Deformation microstructures in minerals represent an extremely valuable tool to determine the physical conditions of ductile deformation of rocks (e.g. Vernon 2004; Passchier and Trouw 2005; and references therein). Nevertheless, most of the existing experimental studies and natural examples are based on the rheologic behavior of monomineral aggregates, eminently of quartz (Hirth and Tullis 1992 and references therein; Stipp et al. 2002 and references therein). On the contrary, the available information for the interpretation of polymineral aggregates -where the interaction between phases with different compositional and chrystallographic characteristics can influence its rheological behavior considerably- is scarce (e.g. Renard et al. 2001; Herwegh et al. 2005, Huet et al. 2014).

Another major drawback when performing a comparative analysis between deformation microstructures observed in natural samples and those obtained in experimental studies is that most of them involve coarse-grained monomineral aggregates as starting materials. Consequently, recrystallized grains smaller than the parent crystals are produced, which increase its size with the increasing of temperature and/or fall in strain rate. This is the most frequently used reference when deformation mechanisms in minerals and associated microstructures, are used to estimate the thermal conditions of ductile deformation of natural felsic and intermediate rocks (see for example, Spear 1993; Vernon 2004; Passchier and Trouw 2005; Trouw et al. 2010, and references therein).

But what happens to the rocks generated from fine-grained polymineralic sediments undergoing prograde metamorphism? The characteristics of the microstructural development of fine-grained polymineralic sediments undergoing progressive metamorphism and deformation, is less well understood. It is well known that ductile behavior of previously formed coarse-grained quartz aggregates begins at ~250–300°C, intracrystalline deformation in the other silicates. As a result, grain boundary migration growth of quartz and metamorphic growth of plagioclase are favored, instead of quartz subgrain rotation crystallization and bulging recrystallization in feldspars. Adjacent coarse-grained rocks with granular textures (leucosome veins and bodies), show typical microstructures of recrystallization connoting conditions for ductile deformation.

3. At ~600°C typical of upper amphibolite facies, grain growth and dehydration reactions promote significant mineralogical and microstructural changes. Reduction of phyllosilicate proportions and discontinuation of micaceous cleavage domains allow mutual contact between anhydrous phases and promote strain-induced grain boundary migration recrystallization of feldspars. Free water might enhance grain boundary migration and recrystallization of quartz and feldspars.
4. At >600°C, coarse-grained metapelitic–metapsamitic rocks behave as typically observed in natural and experimental studies of ductile deformation. Quartz forms Type 3 and 4 ribbons, whereas dynamic recrystallization of feldspars through the grain boundary migration mechanism forms progressively bigger new grains. Observed microstructures point to microfracturing–microshearing combined with fluid-assisted grain boundary migration as the most suitable deformation mechanism for feldspars recrystallization.
5. At >700°C (typical of granulite facies), quartz ribbons coalesce and enclose feldspars. At these thermal conditions, first evidence of rotation recrystallization is observed in plagioclase porphyroclasts. Low-strain sectors of coarse-grained rocks show microstructures typical of ductile deformation at very-high temperature, like diffusion creep between quartz and feldspars grains.

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Chapter 15
Strike–slip ductile shear zones in Thailand

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15.1 INTRODUCTION

Mainland south-east Asia includes development of numerous intraplate strike–slip deformations. These intense deformations confine show in narrow and sub–parallel sided zones, which normally are termed as "shear zones" (Ramsay 1980). Shear zones can reveal brittle, brittle–ductile, and ductile deformations. Many brittle shear zones are also described as fault zone, where rocks in the shear planes could be brecciated. The brittle–ductile shear zones are associated with fault rocks with some ductile deformation, while the ductile shear zones are commonly in the high plastic deformed rocks or mylonites (Ramsay 1980; White et al. 1980).

Strike–slip ductile shear zones in Thailand are the Mae Ping shear zone, the Three Pagodas shear zone, the Ranong shear zone, and the Khlong Marui shear zone (Fig. 15.1). In the last few years, geological studies on the strike–slip shear zones in Thailand have been carried out by a number of workers (Watkinson et al. 2008, 2011; Morley et al. 2011; Kanjanapayont et al. 2012a, b; Nantasin et al. 2012; Palin et al. 2013) to understand their genesis and kinematics. The whole system of strike–slip shear zones in Thailand requires a review. First, I present an overview tectonics of Thailand. I also present the meso- and microstructural analyses, plus previous geochronological data in the best exposed mylonite shear zones in Thailand.

Fig. 15.15. Summary of age constraints for periods of shear along the Mae Ping, Three Pagodas, Ranong, and Khlong Marui shear zones.

The data are taken from Table 15.1.

15.6 CONCLUSIONS

The meso- and microstructures of the mylonites within the strike–slip shear zones in Thailand include asymmetric folds, σ-objects, δ-objects, strain shadow, S–C and S–C' fabrics, shear band types, domino types, mica fish, stair stepping, asymmetric mymerkites, and "V"-pull-apart structures. The dynamic recrystallization in the shear zones were expressed by the undulose extinction, basal gliding, bulging, subgrain rotation, and grain boundary migration. Rocks deformedunder the ductile strike–slip motion. The NW–SE Mae Ping and Three Pagodas shear zones deformed by sinistral movement, while the NNE–SSW Ranong and Khlong Marui shear zones have undergone dextral motion. Geochronology data indicates Late Cretaceous timing for the Ranong shear zone, and Eocene–Oligocenefor all four strike–slip zones. These ages suggest the timing of shear.

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Chapter 16
Geotectonic evolution of the Nihonkoku Mylonite Zone of north central Japan based on geology, geochemistry, and radiometric ages of the Nihonkoku Mylonites: Implications for Cretaceous to Paleogene tectonics of the Japanese Islands

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16.1 INTRODUCTION

The Japanese Islands are situated on the eastern margin of the Asian Continent before opening of the Sea of Japan during Miocene time (Otofuji and Matsuda 1984; Maruyama et al. 1989). In the Japanese Islands, Cretaceous to Paleogene sinistral shear zones are widely distributed along the Median Tectonic Line (MTL; e.g. Takagi 1986; Takagi et al. 1989; Shimada et al. 1998), the Tanagura Tectonic Line (TTL; Koshiya 1986), the Hatakawa Tectonic Line (Sasada 1988; Takagi et al. 2000; Shigematsu and Yamagishi 2002) and in the Kitakami Mountains (Sasaki and Otoh 2000; Sasaki 2001) (Fig. 16.1). Therefore, studies on the geological and temporal relations among these shear zones are important to understand the pre-Neogene tectonics of the eastern margin of the Asian Continent.

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Chapter 17
Flanking structures as shear sense indicators in the Higher Himalayan gneisses near Tato, West Siang District, Arunachal Pradesh, India

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17.1 INTRODUCTION

Flanking structures (FS)/flanking folds are deflections of a planar layer or host fabric elements (HE) such as bedding, foliation or compositional layering around a cross cutting element (CE) such as a fault, vein, joint, a patch of melt, a mineral grain, or even a boudin (Passchier 2001; Coelho et al. 2005; Mulchrone 2007; Mukherjee and Koyi 2009; Mukherjee 2014a). The first descriptions of flanking structures were based mainly on sub-meter scale (Passchier 2001). Microscale examples of the structures came later (Mukherjee 2007, 2010a, b, 2011, 2014b; Mukherjee and Koyi 2009; Grasemann et al. 2011). Description of flanking structures is an expansion of the concept of fault drag-the deflection of layers in the vicinity of the fault (Gayer et al. 1978; Hudleston 1989; Druguet et al. 1997). Between the two end members of simple shear and pure shear, the resulting flanking structures are classified as s-type flanking folds, a-type flanking folds and shear bands (Grasemann et al. 2003). As contractional or extensional offset of central markers, both s-type and a-type FS may exhibit normal or reverse drag (Hamblin 1965) of the central markers in reference to the shear sense along the CE (Wiesmayr and Grasemann 2005). Two new schemes of classification of flanking structures have recently been proposed (Mukherjee 2014b). First, whether (i) the CE is a sharp plane of discontinuity or, (ii) it consists of rock(s)/mineral(s). In case (ii), FS is again divided in to whether HE penetrates CE or not. The second classification is based on drag and slip along the CE margins. It is already defined that along the direction of shear if a convex HE is reached, the sense of drag is “normal” and in the opposite case is “reverse” (Grasemann et al. 2003). What controls the sense of drag is a long studied issue. The latest view is that it depends on (i) angle between the HE and the CE before deformation started; and (ii) relative values of vertical separation and throw of faulting along the CE (reviewed in Mukherjee 2014b). A clear convex/concave drag does not always develop in all the mesoscopic FS. Also, the HE could be thicker near the CE. The two latest classification schemes consider all possible slip and drag of the HE.

s-type flanking structures are regarded as reliable shear sense indicators (Exner et al. 2004) and are used here for that purpose under contractional tectonic setting.

Numerous flanking structures developed in a ductile shear zone affecting the Higher Himalayan migmatitic gneisses around the Tato area of the West Siang District,

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