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Geoscience Frontiers xxx (xxxx) xxx

Contents lists available at ScienceDirect

China University of Geosciences (Beijing)

Geoscience Frontiers

journal homepage: www.elsevier.com/locate/gsf

Coupling between fluids and rock deformation in the continental crust: Preface

1. Introduction

Fluids are essential and widespread components of the Earth crust. They regulate the main process of mass and energy transport, conditioning the geochemical and geophysical evolution of the crust (Bredehoeft and Norton, 1990) and, consequently, its properties and mechanical behavior. The coupling between fluids, rock deformation, and heat are crucial engines for plate tectonics, playing a leading role in the rheology and evolution of the lithosphere (*e.g.* Jamtveit et al., 2000; Kennedy and van Soest, 2007; Thompson, 2010; Miller, 2013; Yardley and Bodnar, 2014; Hirauchi et al., 2016; Menzies et al., 2016; Plümper et al., 2017; Vizán et al., 2017). They strongly contribute to the viability of the Wilson cycle, since cyclic redistribution of fluids guarantees the survival of a dynamic Earth (Fyfe, 1985, 1997; Bercovici, 1998; Ribeiro, 2002; Bercovici et al., 2015, chap. 7.07; Tajima et al., 2015). The influence exerted by fluids on rock deformation (and *vice versa*) is significant, regardless of the tectonic setting (*e.g.* Oliver, 1986; McCaig et al., 2000; Doubre and Peltzer, 2007; N ganjaneyulu and Santosh, 2011; Suppe, 2014; Beaudoin et al., 2014; Williams et al., 2015; Zheng et al., 2016), crustal level (*e.g.* Sibson, 1994; Cosgrove, 1995; Cox, 2002; Hobbs and Ord, 2018; Papeschi et al., 2018) and scale (*e.g.* Carter et al., 1990; Holyoke and Kronenberg, 2013; Tajima et al., 2015).

The source of fluids in the continental crust is variable (*i.e.* atmosphere, earth surface, continental crust, slab, mantle). Fluids, liquid and gas/vapor, can be meteoric, formational (cognate, basinal), hydrothermal, magmatic and metamorphic. Temperature, pressure, composition, and concentration of fluids govern fluid-rock interaction. Because of physical properties of fluids (i.e., low cohesive forces due to weak inter-molecule attraction forces; practically null restitution forces) they are able to freely move and to easily change their form, and they lack yield strength. In fluids, stress is a function of the strain rate: they can resist normal- but cannot sustain shear-stress because they respond to shear stress with motion. The resistance of fluids to move is related to fluid viscosity. Fluid molecules can separate from each others in gases and vapors, but they do not in liquids because liquids do not fill a volume by expanding into it. Consequently, gases are compressible and, in general, liquids are incompressible.

Fluid migration in the crust is triggered by potential energy gradients and controlled by permeability anisotropies. The main mechanisms of fluid transport are gravity-driven (Garven et al., 1993), chemically-driven (Zanella et al., 2014a,b; Cobbold and Rodrigues, 2007), thermally-driven (Norris and Henley, 1976), and pressure-driven (Oliver, 1986; Cobbold et al. 2009; Beaudoin et al., 2014; Suppe, 2014).

2. Coupling between fluid flow and deformation processes in upper crustal regimes

2.1. Changes that fluids introduce into rock deformation

Fluids exert a strong effect on crustal rock strength, influencing the mechanical behavior of the crust. In the upper crust, they can modify the seismic properties of fractured rocks. In presence of fluids, the behavior of rocks under stress can be conditioned by fluid pressure, fluid-rock physico-chemical interactions, fluid temperature, and fluid viscosity (Cox, 2010; Sibson, 2000, 2001, 2011; Rutqvist et al., 2007; Osiptsov, 2017; Cornelio et al., 2018a,b). Hydraulic, mechanical, thermal and chemical phenomena can also introduce changes in morphology of fractures modifying permeability with time, and therefore controlling both short- and long-term fault activity.

Physical factors can strongly condition fracture permeability and aperture. Fluid pressure in excess of hydrostatic (supra-hydrostatic, lithostatic, supra-lithostatic) introduces significant modifications into the crust since it reduces the effective normal stress, reducing the shear resistance (shear strength) of the fracture, which results in fracturing of intact rocks and/or reactivation of old fractures (Hubbert and Rubey, 1959; Sibson, 1990, 1998, 2004; Cox, 1995; Micklethwaite, 2003; Rutqvist, 2015). In the case of pre-existing faults, both fault strength and state of stress condition the required overpressure for fault reactivation. If the crust is critically stressed, then minor overpressure perturbations will be needed to reactivate even those structures that were severely misoriented relative to the *in-situ* stress field. Overpressurized fluids enhance fracturing and dilatancy, increasing the directional permeability of rocks and facilitating their own redistribution. Induced by overpressure at low stress regimes, hydraulic fracturing reactivation could also explain the formation of dilatant shear-compressional structures (Barton, 2007; Japas et al., 2016). Another fluid-assisted mechanism that can trigger Mohr-Coulomb failure, in this case modifying the

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https://doi.org/10.1016/j.gsf.2019.07.001

differential stress, is poroelastic stressing (Rutqvist et al., 2007; Healy, 2012; Segall and Lu, 2015). The incidence of fluid pressure in reducing the effective normal stress along faults is the argument to consider pressure increase as a trigger of seismic activity in fluid-rich upper crustal settings (Healy et al., 1968; Nur and Booker, 1972; Ohtake, 1974; Raleigh et al., 1976; Sibson, 1992, 2000, 2001; Scuderi and Collettini, 2016; among many others). Fluid-induced seismicity is explained based on fluid pressure perturbations in a critically stressed upper crust (Townend and Zoback, 2000; Zoback and Zoback, 2002). The same applies to anthropogenic-induced microseismicity in the shallowest upper crust, induced by large water reservoirs (reservoir-induced seismicity, RIS), hydraulic stimulation in both enhanced oil recovery (EOR) and enhanced geothermal systems (EGS), and injection of waste fluids into deep disposal wells (*e.g.* geological long-term storage or sequestration of CO_2, oil and gas production waste, nuclear waste) (Rice and Cleary, 1976; Roeloffs, 1988; Talwani, 1997; Gupta, 2002; Rutqvist et al., 2007; Deichmann and Giardini, 2009; Baisch et al., 2010; Cappa and Rutqvist, 2011; Horton, 2012; Mazzoldi et al., 2012; Zoback and Gorelick, 2012; Keranen et al., 2014; Zang et al., 2014; Segall and Lu, 2015; Vilarrasa and Carrera, 2015; Doglioni, 2018; Foulger et al., 2018; Juncu et al., 2018; Rinaldi et al., 2019). The increase in pressure caused by the water column filling the dam reservoir or by forced fluid injection changes the pre-existing fluid pressure balance, altering the *in situ* stress along faults/fractures, leading to sudden movement along the pre-existing fracture, and triggering therefore low magnitude minor earthquakes and tremors. Furthermore, fluid-induced seismicity could be also triggered by reservoir depletion or fluid extraction, with pressure drop inducing poroelastic stressing and volumetric contraction, and, consequently, local stress perturbation (Segall, 1989; Teufel et al., 1991; Segall et al., 1994; Hillis, 2001; Streit and Hillis, 2002; Müller et al., 2010; Rozhko, 2010). These injection/depletion-induced earthquakes offer an opportunity to gain insights into earthquake physics (Galis et al., 2017). Viscosity of fluids can also affect the seismic properties of the upper crust as suggested by reservoir engineering studies, which indicate that pore fluid viscosity controls earthquake nucleation (Cornelio et al., 2018a,b).

When fluids are of magmatic/hydrothermal origin, their input can also produce a significant modification into the thermal conditions. This thermal contribution includes partial dissolution of minerals (*e.g.* Fournier, 1999; Landtwing et al., 2005), thermal overclosure (Barton and Makurat, 2006; Barton, 2007; Rutqvist, 2015), and alteration of *in-situ* stresses (thermal stressing; Rutqvist et al., 2007; Healy, 2012; Chen, 2012; Vilarrasa et al., 2017), which introduce changes into the patterns of fracture aperture and permeability. At temperatures lower than the temperature of fracture initiation, differential contraction reduces the closure effects of opposing fracture walls (thermal over-closure; Barton, 2007), allowing open spaces to form and/or enhance, and consequently affecting fracture permeability and the hydro-mechanical behavior of rocks under stress (Barton, 2007). Depending on rock stiffness, temperature changes introduced by fluids can also induce thermal stresses (and stress redistribution), affecting fracture stability. For example, thermal stress reduction by cooling brings the state of stress closer to failure and fracture reactivation (Vilarrasa, 2016; Vilarrasa and Rutqvist, 2017; Vilarrasa et al., 2018, chap. 9, 2019).

Chemical processes linked to the presence of fluids, such as mineral dissolution (free face dissolution and pressure solution), rock alteration, volumetric expansion/contraction, hydrothermal sealing and chemical reactions by stress corrosion, can contribute to produce changes in rock-strength, fracture aperture and permeability of rocks as well as perturbations in the local stress field (Fournier, 1999; Polak et al., 2003a; Yasuhara and Elsworth, 2008; Rutqvist, 2015; Blaisonneau et al., 2016; Vilarrasa et al., 2019; among others). Also, at low stress conditions, the presence of clay alteration minerals along fracture walls (fracture skin) in some hydrothermal deposits could activate some membrane selective mechanisms, changing permeability of rocks (Japas et al., 2015; Maffini et al., 2017).

Regarding geothermal systems, permeability of fractures could be also affected by biological processes (Natarajan and Kumar, 2012). In coupled fracture and surrounded rock systems, the evolution of fracture permeability and pressure variation is influenced by bacterial growth in the fracture skins and coatings (Polak et al., 2003b; Natarajan and Kumar, 2012).

2.2. Changes that rock deformation introduces on fluid flow

Rock deformation induces pressure-driven transport of fluids, triggered by overburden (Rice and Cleary, 1976; Cobbold et al., 2009; Suppe, 2014; Frazer et al., 2014), dilatancy or faulting (Gómez-Rivas et al., 2014), and orogeny (Oliver, 1986; Cobbold et al., 2009; Ballas et al., 2014; Beaudoin et al., 2014), with suction pumping, seismic pumping, dilatancy pumping, fault-valve and mobile hydrofracturing as the most significant mechanisms (Sibson et al., 1975; Sibson, 1981, 1990, 2007; Etheridge et al., 1984; Cox, 1995; McCaig et al., 2000; Bons, 2001; Robb, 2004; Chi and Guha, 2011; Bons et al., 2012; Laurent et al., 2017).

In the upper crust, fractures represent the main permeability anisotropies exploited by fluids to migrate. Although faults mostly enhance rock permeability, they can be also barriers for fluids to flow (Sibson, 1981) depending on their orientation relative to the *in-situ* stress field (*e.g.* Jiang et al., 1997; Sibson, 1998, 2000), their morphology (*e.g.* Blaisonneau et al., 2016), and the presence of a fault seal (*e.g.* Sibson, 1990b; Cox, 1995).

Alterations into the fluid flow patterns can be introduced by *in situ* stress changes that follow a seismic event (static stresses; King et al., 1994; Micklethwaite et al., 2015), but also by deformation rates conditioning fluid viscosity, mainly in pseudoplastic non-Newtonian fluids such as colloidal fluids, since flow is favored by increasing shear stress due to a viscosity drop (Japas et al., 2016). Also, after fault slip, flow properties can change.

Fracture morphology (or "topography") conditions fracture permeability, and therefore plays a significant role controlling fluid circulation in the upper crust. Through-going fractures alter topography of fractures and could represent a major control of dilation in severely misoriented fractures at low-stress regimes (*e.g.* Barton, 2013; Pérez-Flores et al., 2017). Fracture roughness also influences the balance between normal stress and shear, and thus the evolution of the contact-surfaces and the friction-surfaces of opposed fracture walls (Blaisonneau et al., 2016). Fracture arrays are important as well, since the presence of meshes of fractures favors fluid flow through non-porous rocks in the seismogenic crust. Fracture meshes are common structures in those high fluid flow systems hosted by low-permeability rocks with varying tensile strength (Sibson, 1996, 2000, 2017). In subduction-related scenarios, cross-strike structures represent one of these fracture meshes, and are of special interest in mineral and energy resource exploration (Sillitoe, 1997; Sillitoe and Perelló, 2005; Oriolo et al., 2014; Galetto et al., 2018).

Fractures act as preferred fluid channel-ways and are thus potential hosts of mineral deposition, with the consequent permeability loss. Representing fracture seals, veins are outstanding witnesses of fluid-rock interactions that record physical conditions at the time of vein emplacement (Roedder, 1984; Bodnar and Vityk, 1994; Bons et al., 2012), indicate paleo-permeability and its variation with time (Stober and Bucher, 2014), paleo-stress and paleo-strain fields (Jolly and Sanderson, 1997; Micklethwaite and Silitonga, 2011; Japas et al., 2016), and paleo-microseismicity linked

Please cite this article as: Japas, S et al., Coupling between fluids and rock deformation in the continental crust: Preface, Geoscience Frontiers, https://doi.org/10.1016/j.gsf.2019.07.001

to fluid-assisted fracturing (Sibson, 1987, 1992). Fracture seals can result in overpressuring of the fluids giving rise to a pressurization-faulting-sealing cycle (fault-valve mechanism; Sibson et al., 1988; Sibson, 1989, 1990a, 1992; Cox et al., 1991; Boullier and Robert, 1992; Cox, 1995; Japas et al., 2016).

In sedimentary settings, fracturing helps the expulsion of fluids. Liquefaction of sediments, mud volcanoes, clastic dikes and bedding-parallel *beef* veins comprise particular deformational features linked to the release of overpressured fluids (Kopf, 2002; Cobbold and Rodrigues, 2007; Alavi and Gandomi, 2012; Chi et al., 2012; Novikov et al., 2013; Kansai University, 2018). Fluid expulsion is of economic significance in hydrocarbon and hydrothermal deposits, and also of tectonic consideration (*e.g.* Font et al., 2012; Vizán et al., 2017).

3. Fluids and ductile flow in the middle to lower crust

3.1. The role of metamorphic settings

In contrast to the upper crust, middle to lower crustal settings show relatively lower porosity and permeability, and higher temperatures, with fluid flow coupled to ductile deformation, metamorphic reactions and recrystallization phenomena. In general, metamorphic reaction rates depend on temperature and, therefore, are a function of heat input (*e.g.* Walther and Wood, 1984; Baxter, 2003), though fluids commonly behave as catalysts that favor faster reaction rates (Rubie, 1986; Yund, 1997; Thompson, 2010; Llana-Fúnez et al., 2012). In turn, fluid flow depends mainly on the devolatilization and the consequent fluid production rate as well as on changes in permeability arising from volumetric variations (Connolly, 2010). On the other hand, deformation also catalyses reactions, favoring the dissolution of unstable minerals and promoting the crystallization of stable phases (Williams and Jercinovic, 2012; Oriolo et al., 2018). Particularly in high-strain settings, shear heating may also contribute as a catalyser (Leloup et al., 1999; Duretz et al., 2014; Platt, 2015). Consequently, kinetics of metamorphic reactions is intimately related to fluid circulation and strain rates, which, in turn, are mutually dependant.

3.2. Coupled fluid flow, ductile deformation and metamorphic reactions

In metamorphic settings, the composition of fluids is governed by four major components: H_2O, non-polar gases (*e.g.* CO_2), salts and rock-derived components (Thompson, 2010; Yardley and Bodnar, 2014; Manning, 2018). The bulk composition of a particular fluid is thus the result of environmental conditions and the fluid source. On the other hand, deformation is governed by the rheology of mono- or polycrystalline aggregates of minerals and, consequently, by their deformation mechanisms (*e.g.* Drury and Urai, 1990; Passchier and Trouw, 2005; Bürgmann and Dresen, 2008). There is, however, a complex interaction between fluids, deformation and metamorphism.

During prograde metamorphism, fluids released by metamorphic dehydration reactions are expulsed mainly due to overpressure (Lyubetskaya and Ague, 2009; Connolly, 2010; Thompson, 2010; Yardley and Bodnar, 2014). Fluids may thus either migrate towards shallower crustal levels or remain stored in rock porosity (Ernst, 1990; Lyubetskaya and Ague, 2009; Connolly, 2010). In the presence of low-permeability layers, fluids may alternatively migrate laterally, being controlled by temperature gradients (Lyubetskaya and Ague, 2009). When migrating upwards, fluids resulting from devolatization reactions commonly trigger retrograde metamorphism in upper crustal levels (Jamtveit and Austrheim, 2010), though retrograde reactions may also imply fluid infiltration from the surface (Yardley et al., 2000). In the case of storage, fluids may trigger metasomatism leading to metamorphic reactions such as hydration and carbonation, which may in turn modify the rock volume and porosity and, thus, fluid flow (Jamtveit and Austrheim, 2010; Bedford et al., 2017).

A critical part of the equation is to consider deformation in the scenario of coupled fluid flow and metamorphic processes. In the first place, changes in deformation mechanisms and rheology, which may occur at short temporal and spatial scales (Bürgmann and Dresen, 2008), may change rock porosity and permeability (Violay et al., 2017), thus affecting fluid flow. Deformation-induced features ranging from intracrystalline lattice defects to crustal shear zones also promote anisotropic diffusion, providing effective pathways for fluid flow (Carter and Dworkin, 1990; Cox, 2002; Essaifi et al., 2004; Bakker, 2009; Sosa et al., 2018). Likewise, different deformation mechanisms such as dissolution-precipitation and dynamic recrystallization may significantly modify grain boundaries, thus affecting fluid circulation and resulting fluid-mineral reactions that take place at crystal interfaces (Walther and Wood, 1984; Lasaga, 1989; Drury and Urai, 1990; Lasaga and Rye, 1993; Schmatz and Urai, 2011; Kruhl et al., 2013; Bedford et al., 2017).

At regional scale, high-strain brittle-ductile and ductile shear zones are one of the most important crustal features for fluid flow, behaving as preferred channels for fluid circulation (*e.g.* McCaig, 1988; Carter and Dworkin, 1990; Ridley, 1993; Pili et al., 1997; Cox, 2002; Fossen and Cavalcante, 2017; Maffini et al., 2017; Oriolo et al., 2018). Since plastic behavior may actually increase fluid pressure and thus promote upward flow (McCaig, 1988), microcracking resulting from the pressure-dependent visco-plastic rheology of ductile shear zones seems to be a key process for channelized fluid circulation (Mancktelow, 2006), though syndeformational porosity resulting from viscous grain-boundary sliding, creep cavitation and dissolution-precipitation may contribute as well (Fusseis et al., 2009). Once in the shear zone, fluids may in turn favor hydrolytic weakening, mineral reactions, and changes in deformation mechanisms and grain size (*e.g.* Griggs, 1967; Carter et al., 1990; Yund, 1997; Kolb, 2008; Lagoeiro and Fueten, 2008; Fossen and Cavalcante, 2017; Sosa et al., 2018). This coupled fluid-shearing effect may induce volume changes (Essaifi et al., 2004) and, thus, further modifications in porosity and permeability. In addition, mylonite development due to fluid-assisted deformation triggers a chemical potential gradient between the shear zone rocks and the metastable mineral association of the undeformed wallrock, which may lead to further metamorphic reactions (Goncalves et al., 2012). One of the most common reactions that results from fluid-assisted metasomatism in shear zones is the breakdown of K-feldspar and plagioclase to white mica (Wibberley, 1999). The development of mica-rich layers, which can culminate with phyllonite development, promotes dramatic changes in the rheology of shear zones, mostly due to strain-hardening/softening reactions that result from grain-size reduction and modification of deformation mechanisms (Goodwin and Wenk, 1995; Wibberley, 1999; Jefferies et al., 2006).

Besides high-strain settings (Mancktelow, 2006), fracturing is a key process that enables fluid circulation during regional metamorphism, particularly at the microscale (Ferry, 1994; Barker and Zhang, 1998). Though fracturing may result from deformation, it may alternatively arise from volume changes triggered by metamorphic reactions (*e.g.* Connolly et al., 1997; Jamtveit et al., 2009; Jamtveit and Austrheim, 2010). For instance, the serpentinization of mantle rocks, which occurs under low-grade metamorphic conditions, allows for a volume increase of ca. 50%, thus favoring fracturing development and, consequently, an increase in rock porosity and permeability (Iyer et al., 2008; Jamtveit et al., 2008; Jamtveit and Austrheim, 2010; Malvoisin et al., 2017). Moreover, the coupling between fluids, deformation and metamorphic processes related to serpentinization has major implications for the rheology of the lithosphere and regional geodynamic processes (*e.g.* Escartín et al., 2001; Rüpke et al., 2004).

4. Fluids and rock deformation: what is next?

The study of fluid and rock deformation interactions is a continuously growing area. Though physical effects of fluids on rock deformation were classically considered, more recent approaches (fluid-injection linked to geothermal energy, CO_2 sequestration, oil recovery, etc) have introduced aspects related to the chemical and thermal contribution of fluids. The incorporation of experimental work and numerical modelling (Izadi and Elsworth, 2013; Rutqvist, 2015; Blaisonneau et al., 2016; Nermoen et al., 2016) have contributed significantly to explain changes in the mechanical behavior of rocks as a consequence of physico-chemical interactions. Also, numerical modelling of hydraulic fracturing (Sánchez et al., 2015; Jeanne et al., 2018; Jin and Zoback, 2018; Rutqvist et al., 2018; Tang et al., 2019), analysis of shear stimulation in EGS (Ucar et al., 2018; Rinaldi and Rutqvist, 2019) and fault geometry affecting fluid flow (Micklethwaite et al., 2015; Zenghua et al., 2017) represent new perspectives in the analysis of the coupling between fluid and rock deformation.

In middle to lower crustal settings, fluid inclusion studies combined with microstructural and petrological data may hold the key to disentangle complex fluid-deformation interactions (*e.g.* van den Kerkhof et al., 2014; Sosa et al., 2018), providing details on fluid activity in the reconstruction of P-T-A-X-t paths (pressure-temperature-fluid activity-composition-time). In this context, there is growing evidence of the role of fluid-assisted processes as main factors controlling metamorphic reactions (*e.g.* Putnis, 2009) and, when present, such reactions may not be necessarily isochemical (Putnis and John, 2010). Future petrological modelling thus needs to consider processes that are beyond the mineral reactant-product equilibrium state, since equilibrium may be reached between interfacial fluids and product phases in fluid-assisted systems (Putnis and John, 2010). In such settings, the “t” of P-T-A-X-t paths can thus only be constrained by hygrochronology, since fluids rather than temperature control isotopic diffusion (*e.g.* Villa, 2016; Oriolo et al., 2018; Smye et al., 2018; Bosse and Villa, 2019).

5. Introduction to this special issue

This thematic issue of *Geoscience Frontiers* collects a set of research papers from world-known experts, presenting diverse aspects of the coupling between fluids and rock deformation at different crustal levels, geological times and scales. This thematic issue includes theoretical aspects and field-based case studies. A valuable multidisciplinary approach is considered in most contributions, focusing in structural, geophysical and petrological data, among others.

Sibson (2019, this issue) introduces the concept of *Arterial Faults* to refer to those shallow structures connecting the hydrostatically pressurized upper crust with the overpressured lower seismogenic zone. Rooted in deeper overpressured reservoirs, these structures are efficient to distribute fluids from the deep crust to the Earth surface. The most conspicuous settings for arterial faults as well as the arterial potential for such faults are also presented and discussed in this inciting *Focus Article*.

Druguet (2019, this issue) reviews mechanics and kinematics related to the occurrence of veins and dikes in deformed rocks, providing insights into progressive vs polyphase deformation. Based on a large database of field evidence, a novel classification of mechanisms of axial-planar vein formation under progressive deformation is presented, which has major implications for the occurrence of these structures in magmatic, metamorphic and hydrothermal systems.

Giambiagi et al. (2019, this issue) analyze fluid migration around seismically active faults in a geothermal field prospect in the southern Central Andes to assess the role of structures (main faults and fracture meshes) in controlling crustal permeability. Based on field data (balanced cross-sections, fault-slip data, etc), estimations of different stress parameters (stress tensor, slip and dilation tendencies, Coulomb stress changes, etc), and geophysical information (aeromagnetic, gravity and MT/TDEM), they present a geomechanical model for the area.

Afşar and Luijendijk (2019, this issue) present a numerical modelling based on field data of the Liassic layered limestone-marl/shale sequence exposed in the Bristol Channel Basin (UK), to explore the incidence of marl thickness in controlling the vertical extension and propagation of fractures. The obtained critical marl thickness constraining fracture propagation represents a valuable contribution to evaluate the potential of unconventional reservoirs.

We anticipate that these articles will contribute to a better understanding of the fluid-rock deformation coupled effect and its implication to the short- and long-term rheological properties of the crust.

The Guest Editors would like to thank all the contributors to this thematic issue as well as the reviewers John W. Cosgrove, Stephen F. Cox, and five anonymous ones whose expertise contributed to the high level of quality of the manuscripts. Dr. M. Santosh and Dr. Lily Wang are also thanked for their kind and valuable editorial support.

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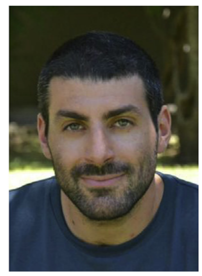
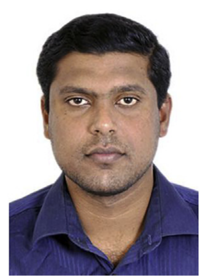
Silvia Japas is a researcher of the CONICET national research council at the Universidad de Buenos Aires (Argentina). Her main interest comprises rock deformation studies in transpressional orogens combining structural, geophysical, kinematic and strain analysis. Currently she is focusing on the interactions between rock deformation and hydrothermal fluids.



Sebastián Oriolo is a researcher at the National Scientific and Research Council (CONICET) and the University of Buenos Aires, Argentina. He works on the coupling between deformation, metamorphism, magmatism and fluids in the continental crust, combining structural geology, geochronology, petrology and geochemistry. Of particular interest are studies focused on the tectonometamorphic and magmatic evolution of Precambrian to Paleozoic orogens in Western Gondwana.



Vinod O. Samuel is a researcher at the Structural Geology and Tectonics Lab, Department of Earth System Sciences, Yonsei University, Seoul, The Republic of Korea (South Korea). He also serves as an associate editor for Geoscience Frontiers. His main research topics include igneous and metamorphic petrology, geochemistry and tectonics. He has been working on magmatism, metamorphism and tectonics in a wide variety of tectonic settings in Korea, China and India. His recent works describe the stability of granitoids in the Early Earth and role of fluids during granulite facies metamorphism/metasomatism.

Silvia Japas*
Instituto de Geociencias Básicas, Aplicadas y Ambientales (IGeBA), CONICET, Universidad de Buenos Aires, Ciudad Universitaria, Buenos Aires, Argentina

Sebastián Oriolo
Instituto de Geociencias Básicas, Aplicadas y Ambientales (IGeBA), CONICET, Universidad de Buenos Aires, Ciudad Universitaria, Buenos Aires, Argentina
E-mail addresses: soriolo@gl.fcen.uba.ar, seba.oriolo@gmail.com

Vinod Samuel
Department of Earth System Sciences, Yonsei University, Seoul, Republic of Korea
E-mail addresses: vosamuel@yonsei.ac.kr, vosamuel@gmail.com

* Corresponding author.
E-mail address: msjapas@gl.fcen.uba.ar (S. Japas)

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