JOURNAL GEOLOGICAL SOCIETY OF INDIA
Vol.94, July 2019, pp.9-22

On Granites

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ABSTRACT

Granites are important components of the Earth's continental crust and represent the net effect of thermochemical processes that operate during partial melting, magma extraction, ascent, emplacement and crystallization. Compositional and isotopic variations in granites arise from source heterogeneities, mixing in the source, and peritectic mineral entrainment as well as crystal fractionation and assimilation. Fluid-absent hydrate-breakdown melting reactions that accompany high-temperature metamorphism are responsible for differentiating the continental crust into a granitic upper portion and a residual lower portion. In some cases, melting can proceed through the influx of a hydrous fluid, although the significance of this on the long-term compositional differentiation of the continental crust is debated. Accessory minerals in granites are important chronometers and play a primary role in transferring the radiogenic isotope signature of sources to granites. Magma ascent and emplacement are guided by deformation and pluton construction is expected to be incremental with pluton heterogeneity being related to the interaction and differentiation of individual magma batches, as well as the extent to which melt separates from solid residual source material. The nature of the source plays a crucial role in the concentration of water and volatiles that are essential for the generation of hydrothermal–magmatic mineral deposits. Future studies of granites using non-traditional stable isotope systems are expected to provide new insights into the evolution of Earth's continental crust.

INTRODUCTION

Granites make up the majority of Earth's upper continental crust and are important end products of the Earth's long-term compositional differentiation. The generation of large volumes of granites on Earth requires the cycling of water through Earth's hydrosphere and lithosphere (Campbell and Taylor, 1983). This cycling has enabled the construction of continental crust that acts as a substrate for the evolution of complex life (Campbell and Allan, 2008; Parnell et al., 2012). Earth's crust is stratified into an upper granitic portion enriched in the heat producing elements and a complementary depleted portion that is more intermediate to basic in composition and depleted in the heat producing elements (Rudnick and Fountain, 1995). Granites are also significant hosts for many mineral resources including base metals and rare metals. Therefore, understanding the petrogenesis of granites is crucial for studies of crustal evolution, the stabilization of the continental crust, the evolution of complex life, and the development of mineral deposits that are of economic interest.

Here, I briefly review granite classification, models of granite petrogenesis, and some of the tectonic and metallogenic implications of granites. This review targets graduate students and professionals that would like to understand our current ideas about granites as well as some of the routine and novel techniques used to investigate granite petrogenesis. Other reviews of granites provide a more thorough examination of the ideas that are briefly presented here (Whitney, 1988; Atherton, 1993; Clemens, 2003; Barbarin, 1999; Brown, 1994; 2010; 2013; Petford et al., 2000; Clemens and Stevens, 2016). Several books are also available that provide additional information (Clarke, 1990; Cobbing, 2000; Pitcher, 1997; Brown and Rushmer, 2006; Chen and Grapes, 2007; Nedelec and Bouchez, 2015). Because granite petrogenesis includes solid and liquid components, I use the term 'melt' to refer to a silicate liquid (including dissolved volatiles) and 'magma' to refer to a mixture of silicate liquid and solids (e.g. crystals). I conclude this review with a summary of the new research topics and techniques that will provide new insights into understanding the genesis and geological significance of granites.

CLASSIFICATION OF GRANITES

Granites (*sensu stricto*) are phaneritic (i.e. coarse-grained) igneous rocks with approximately equal volumetric proportions of quartz, alkali feldspar and plagioclase (Figure 1; Streckeisen, 1976). In addition to these three minerals, other minerals may be present that reflect bulk composition. Many granites contain a few vol% ferromagnesian minerals (e.g. biotite, hornblende) but some contain more aluminous minerals such as muscovite, garnet and cordierite. This variety has inspired different classification systems for granites based on their mineral constituents, which ultimately reflect their bulk compositions and the nature of their sources.

A popular classification system of granites based on their sources was originally developed in the Phanerozoic Lachlan Fold Belt (Eastern Australia) by Chappell and White (1974) and described in detail in Chappell and White (2001). They divided granites into I-type, inferred to be derived from anatexis (partial melting) of metaigneous rocks, and S-types, which were interpreted to be derived from partial melting of metasedimentary rocks. Clemens (2011) provided some general guidelines for distinguishing between these two types of granites based on their mineral assemblages. I-type granites usually have hornblende or clinopyroxene as the primary ferromagnesian mineral and may have magnetite or allanite. Magmatic epidote is also an indicator of I-type granites. S-type granites commonly contain biotite (with high concentrations of Al and Ti) but can also contain cordierite, monazite and pyrrhotite. However, there are always exceptions to these guidelines and other information is usually required to determine granite sources.

Major element and trace element discriminant diagrams are also commonly used to differentiate between I-type and S-type granites (Frost et al., 2001). For example, peraluminous rocks with aluminum saturation indices (molar Al / [Na + Ca + K]) >1.1 are commonly inferred to be derived from partial melting of metasedimentary rocks that contained high concentrations of aluminum (i.e. clay-rich protoliths). However, discrimination based on major and trace elements may be misleading given the variety of petrogenetic processes that can modify granite compositions (c.f. Clarke, 1992). Generally, S-type granite suites show positive covariance between ASI and maficity because the ferromagnesian minerals are commonly peraluminous (e.g. garnet, cordierite) whereas the common ferromagnesian minerals in I-type granite suites (e.g. hornblende, clinopyroxene) are subaluminous

0016-7622/2019-94-1-9/$ 1.00 © GEOL. SOC. INDIA | DOI: 10.1007/s12594-019-1261-2

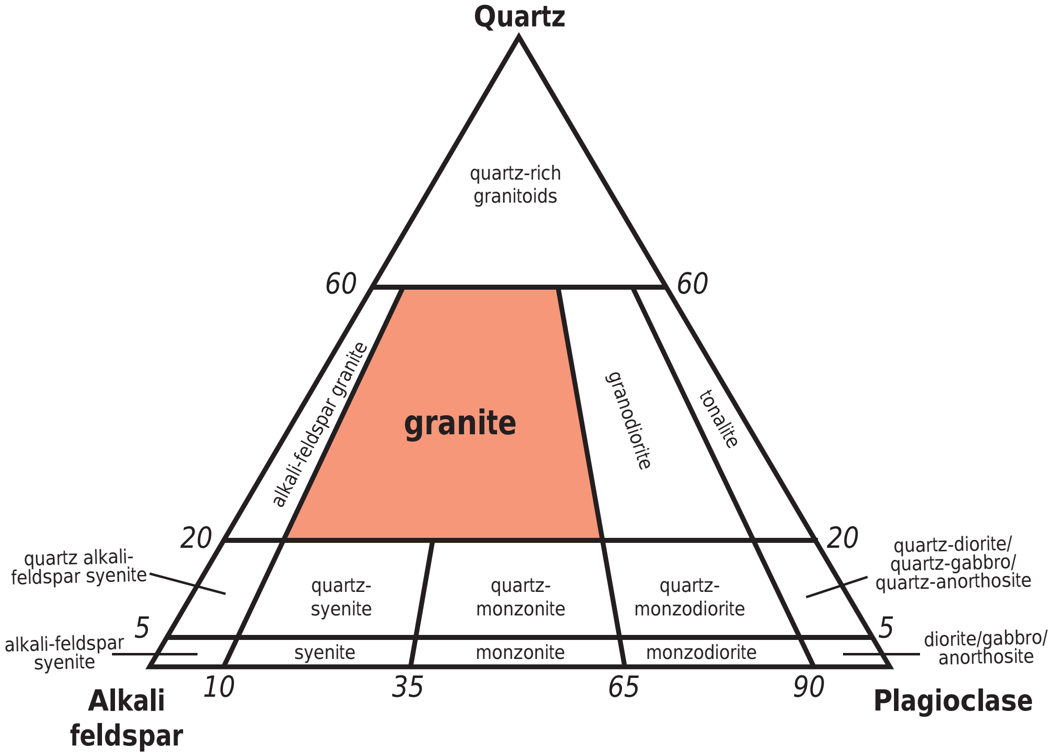

Fig.1. Classification diagram of granitoid rocks based on the volumetric proportions of quartz, alkali feldspar and plagioclase (modified from Streckeisen, 1976). Granite (*sensu stricto*) is shown by the red field.

resulting in negative correlations (e.g. Stevens et al., 2007; Clemens and Stevens, 2012, 2016). Although there are limitations of the I- and S-type granite classification system, it provides a first-order approximation of the nature of the source rocks. Stable and radiogenic isotope analysis can also be used to differentiate metaigneous from metasedimentary sources of granites (O'Neil et al., 1977; McCulloch and Chappell, 1982; Kemp et al., 2007).

In addition to the I- and S-type granite division, granites are also sometimes divided into anorogenic within-plate (A-type) granites that are inferred to be derived from previously melt-depleted and dry sources; M-type granites that are derived directly from the mantle; and C-type (charnockitic) granites. Each of these granite types is discussed briefly below.

A-type granites are named as such due to their association with anorogenic and rift-related magmatism (Loiselle and Wones, 1979). A-type granites are Fe-rich, anhydrous and contain relatively high concentrations of halides, Zr and the REE. The high concentrations of halides (e.g. Cl, F) and trace elements makes A-type granites of particular economic interest and they are commonly associated with Iron oxide copper-gold deposits (e.g. Olympic Dam in Australia). There is disagreement over whether A-type granites are generated from anatexis of dry refractory granulite residue (Collins et al., 1982; Whalen et al, 1987; Skjerlie and Johnston, 1992), melting of a quartzo-feldspathic source (Anderson, 1983; Creaser et al., 1991), or are formed by fractionation of tholeiitic or alkali basalts (Eby 1990, 1992; Frost and Frost, 1997). Nonetheless, A-type granites have a wide variety of compositions and can occur in different tectonic settings. Therefore, granites classified as 'A-type' may represent different sources and petrogenetic processes and some workers suggest that the term should be abandoned (Bonin, 2007; Frost and Frost, 2010).

Granitoids derived directly from the mantle, from partial melting of oceanic crust or the mantle are called 'M-type' granites (White, 1979; Whalen, 1985). The term 'M-type" is somewhat misleading since they commonly include a significant crustal component. The most interesting group of 'M-type' granitoids are sanukitoids – these are mostly diorite to granodiorite but include some granites (*s.s.*). Sanukitoids are relatively rich in K_2O, MgO, Na_2O and the LREE (Martin et al., 2005) and are mostly found in Late Archean terranes (Shirey and Hanson, 1984; Martin et al., 2010; Moyen and Laurent, 2018). They are thought to be the product of interactions between mantle peridotite and a component rich in incompatible elements such as the continental crust (Laurent et al., 2014). Post-Archean equivalents of sanukitoids may be post-collisional K-rich calc-alkaline granitoids that are commonly associated with mafic magmatism (e.g. Couzinie et al., 2016).

Magmatic charnockites (C-type granitoids) are orthopyroxene-bearing (or fayalite-bearing) high-temperature granitoids (Holland, 1990; Howie 1955; Rajesh and Santosh, 2012; Frost and Frost, 2008) that are generally associated with granulite-facies metamorphic terranes (Kilpatrick and Ellis, 1992; Rajesh and Santosh, 2004; Rajesh, 2007). Note that magmatic (i.e. igneous) charnockites are different from metamorphic (i.e. incipient) charnockites (Rajesh et al., 2010; Rajesh and Santosh, 2012) whose origins are controversial (Newton et al., 2014; Blereau et al., 2016). Igneous charnockites are thought to be derived from partial melting of dry granulites in deep continental crust that contain minimal amounts of hornblende (e.g. Kilpatrick and Ellis, 1992; Rajesh and Santosh, 2004), or result from differentiation of tholeiitic melts with variable amounts of crustal involvement (e.g. Duschese and Wilmart, 1997). Regardless of the specific source rocks, charnockites require hot and dry conditions and are found in many tectonic environments, including arcs, rifts and continental collision zones (Frost and Frost, 2008).

The 'alphabet granite' (c.f. Clarke, 1992; Brown, 2013) classification described above is still routinely used, but it needs to be used with caution since it has genetic connotations that need to be tested using a variety of field, petrographic, geochemical, geochronological and isotopic methods. There is increasing isotopic evidence that granites represent mixtures of multiple components (e.g. Kemp et al., 2007) that makes their classification into igneous- or sedimentary-types not particularly useful. Without a clear genetic significance, it may be necessary to group granites by compositions or isotope ratios (e.g. peraluminous, ferroran, high-δ^{18}O).

MODELS OF GRANITE PETROGENESIS

There are two main models of granite generation, including anatexis of the crust and fractionation of magma derived from the mantle. Although granites cannot be derived directly from partial melting of the mantle, the products of which are mostly basalts and gabbros, the differentiation of mafic magma (itself generated from mantle melting) into residue and granites was proposed by Bowen (1948) and variants of this model are commonly employed to explain the petrogenesis of granitoids in oceanic crust and immature arcs (i.e. plagiogranites; Coleman and Peterman, 1975). These granitoids may relate to a single stage process (e.g. differentiation of a mantle-derived magma into granitoids) or a multi-stage process involving partial melting of basic lower crust. Granites (*s.s.*) are relatively rare in arcs (e.g. Jagoutz et al., 2009) and intermediate granitoids (e.g. tonalites and granodiorites) are more common in these settings (e.g. Moyen and Martin, 2012). While a single-stage model is not expected to generate large volumes of granite, it is expected to produce relatively juvenile additions to the continental crust that can be later reworked to generate granites either through direct partial melting or through erosion, burial and anatexis.

Anatexis of the continental crust is commonly considered to be the primary mechanism responsible for generating large volumes of granite (Brown 2010, 2013). Partial melting of the crust requires a large amount of energy due to the latent heat of melting (Bea, 2012) and can be triggered by magmatic underplating, lithospheric extension and the upwelling of hot asthenosphere, or thickening of continental crust with elevated concentrations of the heat producing elements (Clark et al., 2011). Below, I outline some of the important processes that can modify the composition of initial melt to that of a crystallized granite during anatexis, melt segregation, magma ascent and emplacement with a focus on the crustal anatexis model of granite petrogenesis. However, many of the physical processes that I discuss next (e.g. fractional crystallization, assimilation, filter pressing) are also applicable to granite generation in a single-stage system, such as proposed for the minor amount of granites in immature arcs or plagiogranites in ophiolites.

SOURCE PROCESSES

The generation of granite starts with partial melting of the crust, which usually occurs at convergent plate boundaries (e.g. Barbarin, 1999; Moyen et al., 2017), or in thickened crust (e.g. Sawyer et al., 2011) where temperatures (and pressures) are sufficient to melt crustal rocks. The temperature and depth of melting to generate granites are controlled by bulk composition and the presence or absence of a free hydrous fluid (Brown, 2013; Weinberg and Hasalová, 2015).

Partial Melting

Fluid-absent Melting

Fluid-absent hydrate-breakdown melting is generally thought to be the primary mechanism that generates granites (however see Weinberg and Hasalová, 2015). The amount of melt that can be generated from a source rock is also referred to as its *fertility*. Hydrate minerals in metamorphic rocks that break down during the melt-producing reactions in fluid-absent systems include muscovite, biotite and hornblende. Metasedimentary rocks rich in micas (e.g. muscovite, biotite) are usually more fertile than mafic rocks that contain few hydrate minerals and have relatively dry bulk compositions (e.g. granulites). However, most melt-producing reactions also consume quartz and sometimes plagioclase; if these reactants are not available, then even a mica-rich metasedimentary rock may be infertile (e.g. Tuttle and Bowen, 1958; Chappell, 2004). In addition, the progressive loss of melt from a source decreases its fertility in both metasedimentary rocks (Yakymchuk and Brown, 2014a) and metaigneous rocks (e.g. Palin et al., 2016).

The breakdown of these hydrate minerals in fluid-absent systems requires incongruent reactions that produce anhydrous peritectic minerals in addition to the melt. Muscovite-breakdown melting is expected to occur at relatively low temperatures and generate peritectic K-feldspar and an aluminosilicate mineral such as sillimanite or kyanite (e.g. Thompson, 1982; Clemens, 2006). Biotite-breakdown melting occurs after muscovite exhaustion and also produces peritectic minerals such as K-feldspar, garnet, cordierite or orthopyroxene. Hornblende-breakdown melting occurs at relatively high temperatures (usually >800°C) and generates peritectic garnet, clinopyroxene or orthopyroxene depending on pressure, temperature and bulk composition (Moyen and Stevens, 2006; Palin et al., 2016).

During fluid-absent anatexis, melt and peritectic minerals are generated together and are therefore spatially associated. Melt and peritectic minerals will initially form at the grain boundary junction between the appropriate reactants (e.g. Holness and Sawyer, 2008). With increased temperature and further partial melting, melt will begin to coalesce into an interconnected network throughout the rock. Up to this point, the melt has not escaped from its host rock and is thought to be close to an initial (i.e. primary) melt composition assuming equilibrium. The major and trace element composition of this melt is a function of pressure, temperature, bulk composition, and the minerals involved in the reaction. Our knowledge of the compositions of this initial melt come from experimental studies (e.g. Tuttle and Bowen, 1958; LeBreton and Thompson, 1988; Gardien et al., 1995; PatinÞo Douce and Harris, 1998) and analyses of nanogranites and glasses in peritectic minerals (e.g. Cesare et al., 2009; Acosta-Vigil et al., 2010; Bartoli et al., 2016; Carvalho et al., 2018). These studies provide the starting point for understanding how initial melt compositions are modified during granite petrogenesis.

An alternative, but complementary, approach is to use phase equilibrium modelling to estimate the composition of melt in an equilibrated system (e.g. White et al., 2011; Palin et al., 2016). This approach uses an internally consistent database derived from experimental studies in addition to equations of state and activity–composition models to create a thermodynamic model of an equilibrated anatectic system. Model melt compositions can be investigated as a function of pressure, temperature, bulk composition, chemical potentials, molar volumes as well as other variables (Yakymchuk, 2017b).

An example of the initial melt compositions for an average metapelite (Ague, 1991) as a function of pressure and temperature is shown in Figure 2. Figure 2a shows the simplified pressure–temperature stability fields of key minerals involved in fluid-absent hydrate-breakdown partial melting reactions for the modelled composition. Muscovite-breakdown melting occurs at temperatures below the muscovite out phase boundary and biotite-breakdown melting occurs between muscovite-out and biotite-out phase boundaries (Fig. 2a). At temperatures above biotite exhaustion, melting proceeds through the consumption of anhydrous minerals such as quartz and the feldspars. Cordierite is expected to be stable at relatively low pressures and high temperatures (Fig. 2a). For more basic compositions, hornblende-breakdown melting is also an important reaction that occurs at relatively high temperatures (e.g. Clemens, 2006).

With increasing temperature, the predicted ASI of the melt in the model equilibrated metapelite increases (Fig. 2c) and SiO_2 decreases (Fig. 2d), reflecting the multivariant nature of partial melting reactions in large chemical systems (e.g. White et al., 2011). The concentration

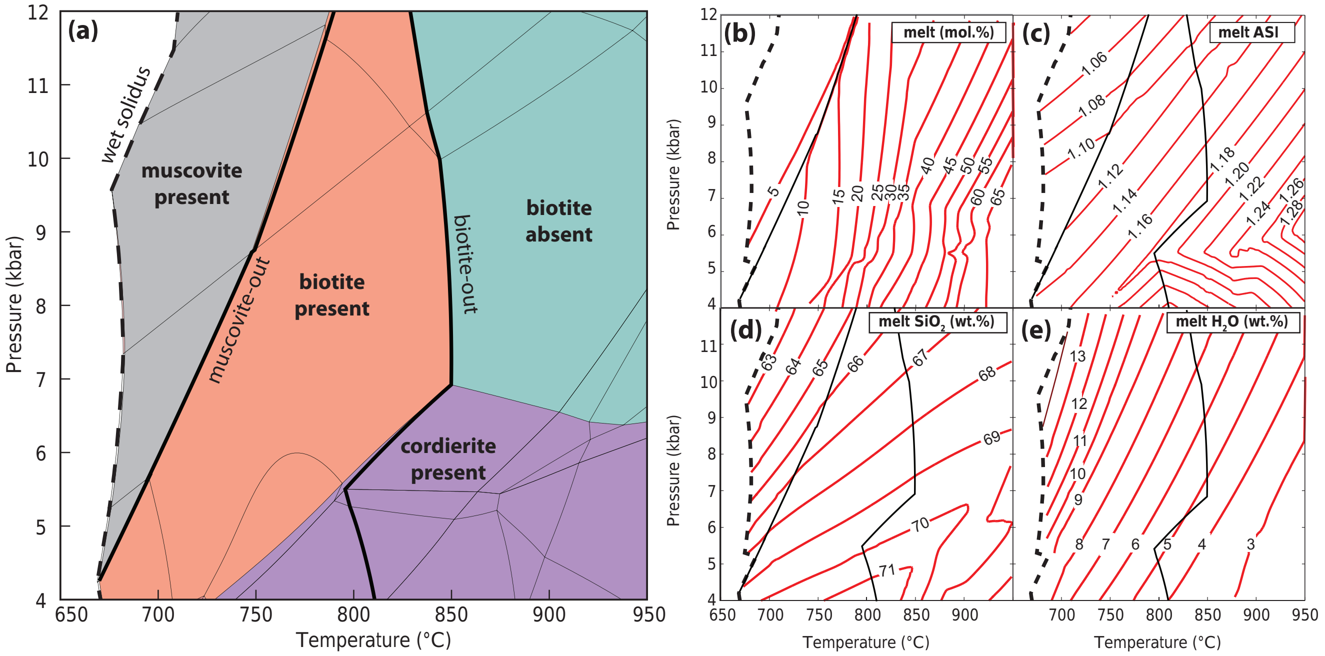

Fig. 2. Phase relations and compositional variation of melt in equilibrium with a high-temperature metapelite composition (modified from Yakymchuk, 2017a). **(a)** Isochemical pressure–temperature phase diagram for an average amphibolite-facies metapelite (Ague, 1991) showing the muscovite and biotite stability fields. **(b)** One-oxide normalized percentage of melt (approximately equivalent to volumetric percent) as a function of pressure and temperature. **(c)** Aluminium saturation index (ASI, molar Al / [Na + Ca + K]). In general, lower pressures and higher temperatures are associated with higher ASI values. This has implications for the stability of phosphate minerals in anatectic melt, such as apatite and monazite (Johnson et al., 2015; Yakymchuk et al., 2017). **(d)** Weight percentage of SiO_2 in melt. This value increases with increasing temperature. **(e)** Weight percentage of H_2O in melt. Note that equilibrated melt is expected to become depleted in H_2O with increasing temperature.

of H_2O in the silicate melt is also a function of pressure and temperature. For the average metapelite, muscovite-breakdown melting results in melt with >9 wt% H_2O whereas biotite-breakdown generates melt with 5–9 wt% H_2O. At temperatures above biotite exhaustion, the melt becomes even more depleted in H_2O, which reflects the increasing contribution of anhydrous minerals (e.g. quartz and feldspars) to the composition of the melt. The concentration of H_2O in the melt influences the pressures and temperatures of the liquidus and solidus as well as viscosity, which affect the behaviour of granitic magmas during ascent (e.g. Johannes and Holz, 1996).

Fluid-present Melting

The presence of a free H_2O-rich fluid can flux melting at the wet solidus or temperatures just above it. The reactants are typically quartz and the feldspars and the fertility of a particular rock will depend on the availability of these minerals. Some fluid-present reactions are congruent where no peritectic mineral is produced (e.g. Sawyer, 1998; Watkins et al., 2007), whereas others involve the generation of hydrate peritectic minerals such as hornblende (Lappin and Hollister, 1980; Berger et al., 2008; Slagstad et al., 2005; Lee and Cho, 2013) or anhydrous peritectic minerals such as garnet (e.g. Stevens and Clemens, 1993; Jung et al., 2009).

Whereas fluid-present melting is commonly proposed in migmatite terranes, its ability to generate large amounts of granite in the crust is debated (Weinberg and Hasalová, 2015; Clemens and Stevens, 2015). The movement of large quantities of fluid necessary for inducing extensive fluid-present melting is not compatible with the limited permeability of the deep crust (Yardley and Valley, 1997). Granites are also expected to be undersaturated in H_2O during their generation in the deep crust (e.g. Clemens and Watkins, 2001). Nonetheless, fluid-present melting can occur in different rock types and tectonic regimes and recent studies demonstrate that it may be a more common process than previously acknowledged.

Accessory Minerals

Accessory minerals act as chronometers to date the timing and duration of anatexis in source rocks (Taylor et al., 2016; Rubatto, 2017) and the crystallization of granites during emplacement (Hoskin and Schaltegger, 2003). Accessory minerals are also important repositories for key trace elements used in isotopic fingerprinting of granites. Therefore, their behaviour during partial melting is particularly important in studies of granite petrogenesis. The most common accessory minerals in granites are zircon and apatite. In addition, monazite is commonly found in peraluminous granites whereas allanite and titanite are important accessory minerals in metaluminous and Ca-rich granites although there are rare exceptions. Accessory mineral behaviour during partial melting has been determined from experiments (e.g. Watson and Harrison, 1983; Harrison and Watson, 1983, 1984; Klimm et al., 2008; Stepanov et al., 2012; Boehnke et al., 2013) and empirical studies (e.g. Williams, 2001; Rubatto, 2006; Bingen et al., 1996; Cottle et al., 2018; Kunz et al., 2018). Building on these studies, quantitative phase equilibrium modelling predicts that zircon, monazite and apatite are expected to dissolve during partial melting in order to saturate the melt in their essential structural constituents (e.g. Zr, LREE, P); new growth is only predicted to occur during cooling and melt crystallization (Kelsey et al., 2008; Kelsey and Powell, 2011; Yakymchuk and Brown, 2014b; Wang et al., 2014; Yakymchuk et al., 2017, 2018; Yakymchuk, 2017a). Mass balance calculations also predict that zircon will dissolve during partial melting due to the predicted increased amount of Zr accommodated by garnet and amphibole with increasing temperature (Kohn et al., 2015). However, kinetic restrictions on accessory mineral dissolution and their entrapment in stable or growing major minerals may retain accessory minerals in the source. This also leads to the retention of the essential structural constituents (i.e. Zr, LREE, P) of accessory minerals in the source (e.g. Williams et al., 2018; Wolfram et al., 2018).

Because the major hosts of the rare earth elements in most high-temperature metamorphic rocks are the accessory minerals (e.g. Bea and Montero, 1999), the radiogenic isotopic composition of the melt is strongly affected by the behaviour these minerals (e.g. Ayres and Harris, 1997; Zeng et al., 2005; Taylor et al., 2016). For example, zircon that crystallized from anatectic melt may develop a Hf isotope composition that is similar to that of detrital zircon in the host metasedimentary migmatites (Flowerdew et al., 2006; Wu et al., 2007; Li et al., 2018). This implies that the Hf composition of the melt is dominated by that of inherited zircon, but radiogenic Hf liberated from the breakdown of Lu-rich minerals such as garnet (e.g. Taylor et al., 2016) and apatite (e.g. Yakymchuk et al., 2015a) is also expected to influence the isotope composition of the melt. This is significant because the Hf isotope compositions of zircon are commonly used to understand the source of granites and ultimately the timescales of generation of the continental crust (e.g. Dhuime et al., 2011). Neodymium isotopes are also used to model crustal evolution and isotope ratios are strongly influenced by the behaviour of monazite and apatite in melt (Watt and Harely, 1993; Ayres and Harris, 1997; Zeng et al., 2005; Villaros et al., 2009; Macera et al., 2011; Korhonen et al., 2010b; Hammerli et al., 2014). Again, isotopic equilibration between apatite (and monazite) and melt control whether extracted melt has the same Nd isotope ratio as its source.

EXTRACTION

The geometry of the interconnected melt network ultimately responsible for melt extraction will be a function of imposed deviatoric stress (e.g. Sawyer, 2014) as well as rheological and compositional heterogeneity (Yakymchuk et al., 2013a; Diener and Fagereng, 2014). In a static system, an interconnected melt network is predicted to occur when ~7 vol.% melt is present (Vigneresse et al. 1996; Rosenberg and Handy, 2005). This threshold may be even lower for dynamic systems (e.g. Brown et al., 1995). Nonetheless, once a permeable melt network is developed, magma exfiltration will be promoted by positive volume change during fluid-absent partial melting reactions (e.g. Clemens and Droop, 1998), the difference in viscosity between the magma and the residue (e.g. Rutter and Mecklenburgh, 2006), and shear-enhanced matrix collapse and filter pressing (Brown et al. 1995). With an adequate amount of melting as dictated by these factors, melt will be expelled from the source rock and begin ascent, which is aided by its high buoyancy and low viscosity relative to the surrounding rocks (e.g. Rutter and Mecklenburgh, 2006).

An important consideration here is that melt extracted may not represent initial unmodified melt compositions due to fractional crystallization and entrainment of residue. A common observation is that leucosomes in migmatites do not record initial melt compositions, but have compositions consistent with the accumulation of early crystallized cumulate feldspar (Sawyer, 1987; Milord et al., 2001; Solar and Brown, 2001; Morfin et al., 2014; Carvalho et al., 2016) and the drainage of complementary fractionated melt possibly due to filter pressing (Brown et al., 1995; Brown et al., 2016). This compositional variation may be due to a change in *P–T* conditions and crystal fractionation during the protracted melting process (e.g. Koblinger and Pattison, 2017) or through the intrusion of magma from elsewhere that is variably fractionated (e.g. Sawyer, 1987). Therefore, the melt extracted from a source may be more fractionated than an initial melt composition.

The deviation of granite from primary melt compositions may be due in part to the physical entrainment of residue or peritectic minerals in melt during extraction. The entrainment of various proportions of restite in extracted magma was proposed to account for the major element variations relative to minimum melt composition in granitic magmas formed at low temperature – this was termed 'restite unmixing' (e.g. Chappell et al., 1987). However, it is unclear if this is a major contributor to the composition of granites due to the absence of single linear compositional arrays connecting restite from primary melt in many granite suites (e.g. Holtz and Barbey, 1991) and many of the compositional trends, particularly in high-temperature magmas, can be alternatively explained by crystal fractionation, magma mixing and assimilation (Wall et al., 1987). The entrainment of the peritectic minerals (e.g. garnet, orthopyroxene) that are spatially associated with partial melting reactions may be reflected by high ferromagnesian contents of granite relative to primary melt compositions (Stevens et al., 2007; Clemens et al., 2011; Taylor and Stevens, 2010; Garcia-Aris and Stevens, 2017). The presence of inherited zircon in granites attests to the entrainment of residuum being common, particularly in peraluminous magmas derived from metasedimentary sources, however, the proportions of residuum to melt are not always clear (e.g. Jeon and Williams, 2018).

An additional mechanism that can explain the isotopic and chemical variability of granites is the interactions in the source that include melting, assimilation, storage and homogenization (e.g. Annen et al., 2006). This model has been used to explain isotope arrays in granite suites that record variable input of crustal and mantle sources (Collins, 1996; Hawkesworth and Kemp, 2006; Kemp et al., 2008). The mixing of isotopically distinct crustal sources can also generate similar apparent mixing arrays (Fisher et al., 2017) and field-based studies suggest that mixing and hybridization of crust-derived magmas occur near their sources (e.g. Schwindinger and Weinberg, 2017). One complication in assessing the mixing of magmas derived from multiple isotopically distinct sources is that disequilibrium behaviour of accessory minerals may result in variable isotope ratios of Sr, Nd and Hf from a single isotopically homogenous source (Iles et al., 2018).

The timescales of partial melting and melt extraction are commonly inferred using accessory mineral geochronology and saturation concentrations of the extracted melt. The preservation of relict accessory minerals and low concentrations of LREE and Zr in granites have been used to argue for short timescales of magma extraction of decades to centuries (Sawyer, 1991; Harris et al., 2000; Villaros et al., 2009). These timescales may be too short for isotopic equilibration between accessory minerals and melt; this may result in magmas with different isotope ratios from that of the source (e.g. Zeng et al., 2005; Tang et al., 2014; Gao et al., 2017) although some of this variability may relate to open system behaviour during anatexis (e.g. Hammerli et al., 2014).

ASCENT

Coagulation of melt (with any entrained components) and redistribution in the source (i.e. Diener et al., 2014) may eventually result in an extracted magma that ascends via fracture-controlled conduits (i.e. dikes) or shear zones towards the shallower crust due to its buoyancy relative to the surrounding rock (e.g. Petford et al., 1994; Brown and Solar, 1998, 1999; Cruden 2006; Hall and Kisters, 2016). The timescales of magma ascent are expected to be as fast as days to weeks if magma transport is via diking (e.g. Petford et al., 1993). Ascent of a granitic magma involves a reduction in pressure with variable amounts of cooling, which leads to physical and chemical changes in the magma. Magma ascent may result in the dissolution of entrained peritectic minerals into melt due to changes in the stability of these minerals as a function of pressure and temperature. The variations of maficity (molar Fe + Mg) with SiO_2 in granites have been related to the proportion of entrained peritectic material (Stevens et al., 2007; Clemens et al., 2011; Clemens and Stevens, 2012; Clemens and Stevens, 2016). The implication is that although most upper-crustal granites do not contain garnet or clinopyroxene, the whole-rock compositions reflect the entrainment of peritectic minerals from the source and their dissolution and/or reaction during magma ascent.

The ascent and cooling of magma can also result in fractional

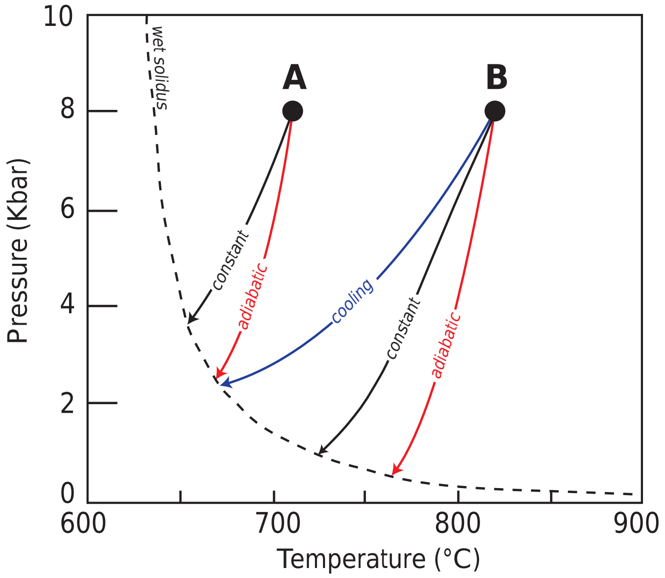

Fig. 3. Pressure–temperature diagram showing three end-member ascent paths for two haplogranites that each begin with 50 wt% melt and 50 wt% crystals (modified from Johannes and Holtz, 1996). Adiabatic ascent results in the consumption of crystals during ascent whereas cooling causes crystallization. The 'constant' ascent path represent no change in the proportion of melt and crystals during ascent. In all cases, final crystallization of the granites is expected to occur at the fluid-saturated solidus (dashed line).

crystallization *en route* to its sink (e.g. Johnson et al., 2003; Tartèse and Boulvais, 2010; Yamato et al., 2015), although the minerals that crystallize will be a function of magma composition, pressure, temperature and if the magma remains superheated (e.g. Clemens, 2003). Nonetheless, crystal fractionation and the physical separation of these from the ascending magma will result in more evolved melt further along the direction of flow; this has also been documented in leucosomes in migmatites (e.g. Sawyer, 1987). The importance of fractional crystallization to the compositional variation in large granite suites is unclear, but it may be small for S-type granites (e.g. Garcia-Arias and Stevens, 2017), which have relatively high viscosities and small density differences between the melt and silicate minerals that may hinder crystal–melt separation. Because viscosity is temperature dependent, crystal fractionation may be more efficient in higher-temperature granites (e.g. some I-type granites). However, viscosity also decreases with the concentration of H_2O; these factors are codependent and high-temperature granites may not have significantly lower viscosities than lower temperature granites (e.g. Clemens and Petford, 1999).

The amount of granite crystallization or crystal dissolution during ascent is a function of the water content of the magma as well as pressure and temperature. Figure 3 shows modelled haplogranite ascent paths for two magmas at 8 kbar that initially contain 50 wt% melt and 50 wt% crystals (from Johannes and Holtz, 1996). The granite at A contains 4 wt% dissolved H_2O and the granite at B contains 2 wt% H_2O. Magmas that ascend adiabatically (i.e. no heat loss or gain) cool slightly but generate melt during ascent ('adiabatic' in Figure 3). These magmas are expected to have an increasing melt/crystal ratio with ascent and early crystallized or residual minerals may be partly to completely consumed. This effect is most pronounced at low pressures due to the stronger curvature of the liquidus (see Johannes and Holtz, 1996). Magmas that ascend with a constant crystal/melt ratio ('constant' in Figure 3) cannot ascend to as shallow depths as granites undergoing adiabatic ascent due to the higher pressure of the solidus. Finally, magmas that cool during ascent ('cooling' in Figure 3) are expected to crystallize minerals during decompression and these magmas are more likely to be trapped even deeper in the crust.

EMPLACEMENT

The final emplacement of granites reflects the transition from upward to horizontal flow of a magma (e.g. Petford et al., 2000) and results in granite plutons of variable shapes and dimensions although most can be approximated by tabular or wedge-shaped geometries when emplaced in a brittle regime (Cruden, 2006). Although the mechanisms of granite emplacement are not always clear and sometimes controversial, deformation of the host rock plays a critical role (Brown, 2013). Pluton growth is expected to be incremental as suggested by the heterogeneous composition of granites (e.g. Pressley and Brown, 1999; Bartley et al., 2008; Farina et al., 2012) and the timescales of pluton growth range from thousands (Michel et al., 2008; Matzel et al., 2006; Schoene et al., 2012; Samperton et al., 2017) to millions of years (Coleman et al., 2004; Schaltegger et al., 2009).

The integrated composition of a granite pluton is derived from the composition of the individual feeder magmas (and their fractionated derivatives), the interaction between these magmas, and possibly the assimilation of wall rock. However, the assimilation of wall rock is energy intensive and self limiting (e.g. Glazner, 2007); it is not expected to be an efficient process during pluton growth. Heterogeneity in granites is primarily inherited from the source (Chappell and White, 1974; Clemens et al., 2010); this includes the entrainment of peritectic and restitic minerals that are also inherited from the source (e.g. Clemens and Stevens, 2012). In addition, the role of fractional crystallization as a primary driver of compositional differentiation of plutons may be relatively minor compared with heterogeneities inherited from the source (Clemens et al., 2010; Clemens and Stevens, 2016; Garcia-Arias and Stevens, 2017).

Accessory Mineral Geochronology

Accessory mineral geochronology is routinely used to determine the absolute timing of granite crystallization. By far the most common accessory mineral chronometer used is zircon, however, monazite is also routinely used to determine the timing of crystallization of peraluminous granites. Zircon is expected to grow once zirconium saturation (relative to zircon) is reached in silicate melt. Zircon may grow on pre-existing (i.e. inherited or early-crystallized) zircon, nucleate spontaneously in zirconium oversaturated liquid (Schaltegger and Davies, 2017), or grow during an Ostwald ripening process where smaller grains are preferentially consumed during the growth of larger grains. New zircon (i.e. autocryst) crystallization in granites is driven by changing conditions that saturate the melt in zirconium with respect to zircon. Experimental studies have demonstrated that the main controlling factors on zirconium saturation in felsic melts are temperature and the ratio of melt-forming network elements (Al and Si) to the elements (Ca, K and Na) that modify this melt network (Watson, 1979; Watson and Harrison, 1983; Boehnke et al., 2013). This ratio has been called the 'M' value of the melt and represents normalized (to a value of 1) molar concentrations of these elements ([Na + K + 2Ca] / [Al x Si]). In general, more peraluminous granites have lower M values than metaluminous ones. Figure 4a shows the results of the zircon saturation equation of Boehnke et al. (2013) using an assumed value of 500,000 ppm Zr in zircon (an approximately stoichiometric amount). Zirconium saturation concentrations are higher at elevated temperatures and larger M values (Fig. 4a).

When the concentration of Zr in the melt is less than the modelled saturation concentration, then the melt is undersaturated in Zr (with respect to zircon) and zircon is not expected to crystallize. When the two values are equal, zircon will grow as long as there are no substantial kinetic or nucleation restrictions (e.g. Harrison and Watson, 1983; Watson, 1996). Zircon growth occurs during a 'crystallization window'

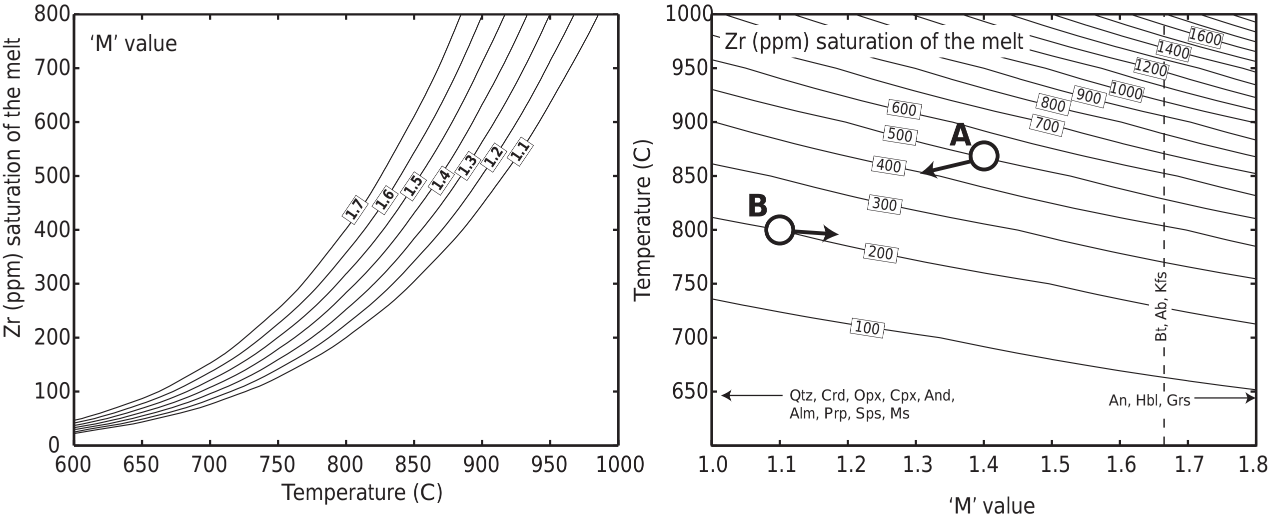

Fig. 4. Zirconium saturation related to the behaviour of zircon in granitic melt. **(a)** Zirconium saturation in melt plotted as a function of temperature and contoured for 'M', which is the molar concentrations of these elements ([Na + K + 2Ca] / [Al x Si]) in the melt using the equations of Boehnke et al. (2013). Zirconium saturation increases with temperature and with the M value. Therefore, heating during partial melting is expected to consume zircon and cooling and melt crystallization is expected to grow zircon. In the absence of reaction overstepping, the temperature at which zircon begins to crystallize is when concentration of Zr in the melt is equivalent to the concentration of Zr in zircon-saturated melt at a given 'M' value. **(b)** Relationship between 'M' value and temperature contoured for zirconium saturation. Two example granitic melts are shown by 'A' (a typical S-type granite) and 'B' (a typical I-type granite). Cooling results in a decreasing saturation temperature for both compositions. For 'A', early crystallization of anorthite-rich plagioclase, clinopyroxene or hornblende is expected to drive the melt composition towards the left on the diagram and to lower M values and may promote zircon crystallization. For melt composition 'B', early crystallization of peraluminous minerals (e.g. garnet, cordierite, sillimanite) would drive the composition of the melt towards the right, which has the opposite affect and may result in zircon dissolution. Mineral abbreviations from Kretz (1983).

that extends from zirconium saturation of the melt at high temperatures down to the solidus. This is a protracted process (e.g. Schaltegger and Davies, 2017) that can last for up to thousands or millions of years (e.g. Schoene, 2014; Samperton et al., 2015; 2017). High-precision U–Pb zircon geochronology and thermal modelling by Samperton et al. (2017) demonstrate that most zircon crystallizes early in the cooling history when temperatures are highest.

The absolute ages of granitic magma emplacement and zircon crystallization may be significantly different depending on cooling rates, nucleation delays, and the amount of zirconium undersaturation when the granitic magma was emplaced. Because granites are dynamic systems, the composition of the melt can change during fractional crystallization, which can impact the timing of zirconium saturation in a crystallizing granite. Figure 4b shows the compositions of typical granitic melt and minerals as a function of their M values. Melt composition 'A' is a metaluminous melt that may represent an I-type granite (M ~ 1.4) and 'B' is a more peraluminous composition that may represent an 'S-type' granite (M ~ 1.1). For both compositions, a decrease in temperature at a constant M value decreases the zirconium saturation value and is expected to result in new zircon crystallization (Figure 4b).

Fractional crystallization may have variable effects on zircon behaviour in granitic magmas. For 'A', early crystallization of anorthite-rich plagioclase, clinopyroxene or hornblende is expected to drive the melt composition towards the left on the diagram and to lower M values. Because zirconium saturation values are lower for lower M values (Figure 4a), the removal of these minerals from the melt may promote zircon crystallization. For melt composition 'B', early crystallization of peraluminous minerals (e.g. garnet, cordierite, sillimanite) would drive the composition of the melt towards the right (Figure 4b). This has the opposite influence and increases the saturation concentration of Zr in the melt (Figure 4a), which may promote dissolution of inherited or early crystallized autocrystic zircon during protracted melt crystallization. Indeed, some S-type granites contain zircon with complex dissolution–growth textures that may be reflecting periods of zircon undersaturation and zircon saturation during melt crystallization (e.g. Vavra, 1994; Zeck and Williams, 2002; Hoskin and Schaltegger, 2003). Undercooling may also modify the predicted crystallization sequence (e.g. London, 2014) and this may be particularly important for studies of zircon in pegmatites.

While the diagram in Figure 4 provides simplified models of how changing bulk composition may influence zircon saturation, fractional crystallization during granite emplacement will also impact the heat budget of the rock through the latent heat of crystallization, which will, in turn, impact the temperature and zircon saturation. In addition, the effectiveness of crystal fractionation in relatively viscous granites is not clear (e.g. Clemens and Stevens, 2016). Nonetheless, changing melt composition during granite crystallization is worth considering especially in studies that use high-precision U–Pb zircon geochronology in slowly cooled terranes.

Mineral Deposits

Granite-related mineralization is a key source of base and rare metals. High concentrations of minerals of economic interest in granites are a function of magma composition (usually controlled by source composition), oxygen fugacity (f_{O2}) and the process of metal enrichment. Metal enrichment can occur in granites due to the crystallization sequence (e.g. rare metal granites), extreme fractionation (greisen-type mineralization and rare metal pegmatites) or late-stage hydrothermal–magmatic activity associated with granite crystallization and fluid–rock interaction (Cernyì et al., 2005).

Porphyry deposits are associated with Cu, Mo, W mineralization and are the result of magmatic–hydrothermal processes that require the favorable interaction between magma crystallization, structural conduits, element scavenging by migrating hydrothermal fluids, and the composition and oxidation state of the magma (Blevin and Chappell 1995; Blevin et al., 1996; Candela and Piccoli, 2005). Greisen and vein-related mineralization of Sn, W and Mo are usually associated with hydrothermal mobilization of these elements in fluids derived from shallowly-emplaced metaluminous to peraluminous granites (Cernyì et al., 2005). Rare metal granites are enriched in the lithophile elements (e.g. Li, Sn, W, Nb, Ta; Linnen, 1998) and may result from extreme fractional crystallization (Lehmann, 1990) or from very low degrees of partial melting of rare-metal-rich sources (Cuney et al., 1992; Cuney and Kyser, 2008). Granitic pegmatites are important sources of the rare earth elements as well as Li, Rb, Cs, Be, Ga, Sn, Ta and Nb (Cernyì, 1991) and can also be generated from extreme fractionation or low degrees of partial melting (Cernyì et al., 2005).

In many cases, granites represent the direct source of the metals of economic interest although leaching of metals from the wallrock during hydrothermal alteration can also contribute. Therefore, the source of the granites plays an important role in the concentration of base and rare metals in granites and their fractionated derivatives (Blevin and Chappell, 1992; Cernyì et al., 2005; Gardiner et al., 2017). Two of the most important variables in granite composition for the generation of granite-related mineral deposits are oxidation state (e.g. Einaudi et al., 2003) and the water content (Strong, 1981).

Variable oxidation states of granitic magmas control their metallogenic potential (Carmichael et al., 1991). In general, S-type granites are commonly reduced relative to I-type granites and this may reflect the presence of reduced organic material in the source rocks (i.e. metasedimentary rocks) of S-type granites (Blevin and Chappell, 1992). However, granites form a continuum of oxidation states that are associated with different elements of economic interest (e.g. Barton, 1996). Relatively reduced granites (ilmenite series of Ishihara, 1977) are commonly associated with Sn and W deposits whereas oxidized granites (magnetite series of Ishihara, 1977) are linked to Cu–Au±Mo deposit systems. Reduced magmas generally crystallize sulfide minerals early in their crystallization sequences and this prevents the later mobilization of metals that require sulfide speciation. Relatively oxidized magmas retain sulfur in the melt until later in their crystallization sequences and are more likely to generate sulfide-rich fluids during the final stages of granite crystallization that can mobilize and ultimately precipitate important base metals. In addition, element speciation can control whether elements of economic interest remain in the melt until late in the crystallization history and can be scavenged by hydrothermal fluids or substitute into early-crystallized minerals.

The volatile content of granites has also been linked with their emplacement depths and their metallogenic potential (Strong, 1981). Relatively H_2O-rich magmas (e.g. S-type granites generated from muscovite-breakdown melting reactions) may arrest deeper in the crust than H_2O-poor magmas (e.g. I-type granites generated from hornblende-breakdown melting reactions). Concentrations of chlorine and fluorine are important as they act as ligands for metals that are important in granite-related mineral deposits (Candela and Holland, 1984). Magmas generated in arc systems generally have high concentrations of chlorine, which is thought to derive from subcontinental sources such as dewatering of subducted oceanic lithosphere (Piccoli and Candela, 1994; Candela and Piccoli, 2005). Chlorine is particularly important for complexing base and precious metals in I-type magmas due to its low solubility in melt relative to aqueous fluids (Cernyì et al., 2005).

LINKING SOURCE TO SINK

A key goal of studying granites is to determine the nature of their sources that can then be used to test tectonic models of orogenesis and to evaluate contrasting models of crustal evolution and differentiation. There are several techniques routinely used to evaluate the source of granites, including major and trace element compositions, stable and radiogenic isotopes, and inheritance of accessory minerals with complex zoning.

Major and trace elements are routinely used to examine the compositional trends of granite suites. However, the major element composition of granites is usually quite different from that of the source. Some major elements (e.g. K) and major element ratios (e.g. ASI) are fairly useful for inferring the nature of the source rock, but many processes modify the major element compositions of granites from that of the initial anatectic melt. Similarly, granites inherit their trace element compositions from their sources, but major and accessory mineral behaviour can modify these values in the source, during ascent, emplacement and crystallization.

U–Pb geochronology of complexly-zoned zircon in granites can give indications of the nature of the source. Inherited cores in zircon may define a single population or a range of ages. Inherited cores of a single age are suggestive of an igneous source. However, this may not be distinguishable from a sedimentary source that contains a single population of detrital zircon. A range of ages retrieved from zircon cores may be more suggestive of a sedimentary source. But, this may also represent inheritance from multiple possible sources or scavenging of zircon from partially molten wall rock during magma extraction, ascent and emplacement. Nonetheless, the age of relict cores in complex-zoned zircon can help determine granite sources.

Isotope ratios between mantle and crustal reservoirs can be quite different due to the long-term compositional evolution of the Earth's crust. Differentiating the sources of granites has been successful through the use of whole-rock Nd ($^{143}Nd/^{144}Nd$) isotope ratios, Sr ($^{87}Sr/^{86}Sr$) isotope ratios (e.g. Compston & Chappell, 1979), O ($^{18}O/^{16}O$) isotope ratios (O'Neil & Chappell, 1977) as well as Hf ($^{176}Hf/^{177}Hf$) and O isotope ratios in zircon (e.g. Kemp et al., 2007). Neodymium and hafnium isotopes are commonly used to infer whether granites are derived from relatively radiogenic (i.e. juvenile) or non-radiogenic (i.e. evolved) sources because the parent isotopes (^{147}Sm and ^{176}Lu) are slightly more enriched in mantle reservoirs relative to the crust. Strontium is also used for this purpose, but more radiogenic Sr isotope ratios are suggestive of a crustal component because the parent isotope (^{87}Rb) is preferentially concentrated in the crust. Oxygen isotopes are used to determine if granite sources are made up of material that experienced a weathering cycle at the surface; sedimentary rocks usually have elevated oxygen isotope ratios relative to the mantle (Valley et al., 2005). Oxygen isotope ratios of whole-rock samples can be analyzed (e.g. Taylor, 1968) although *in situ* analysis of zircon is more common.

An example of these techniques applied to a suite of granites from western Marie Byrd Land in West Antarctica is shown in Figure 5. The region is dominated by a Cambrian metasedimentary sequence and Devonian metaigneous granodiorite suite and both show evidence for *in situ* partial melting in a rare exposure of the deep crust in a migmatite–granite complex in the Fosdick Mountains (Siddoway et al., 2004; Korhonen et al., 2010a; Yakymchuk et al., 2015a). Late-Devonian to early Carboniferous granites in the region yield ages and isotope ratios similar to these two potential sources (Pankhurst et al., 1998; Korhonen et al., 2010a, b, 2012; Yakymchuk et al., 2013b, 2015b; Brown et al., 2016). Together, the rocks in western Marie Byrd Land provide a good natural laboratory to investigate the petrogenesis of I- and S-type granites and interaction between them.

Figure 5 explores the elemental and isotopic variation in a granite suite and its putative sources from western Marie Byrd Land. Major elements for seven investigated granites mostly plot in the field defined by the composition of the metaigneous source rocks (Fig. 5a) and some plot in between the two potential sources. The concentrations of rare earth elements in both source rocks are similar, and the granites generally overlap the values of both sources, but some have lower concentrations of the LREE (Fig. 5b). This probably reflects the behaviour of monazite and apatite in the source (e.g. Korhonen et al., 2010a). Figure 5c shows the Nd and Sr isotope values of the two sources and the granites. While some granites plot in the field for the

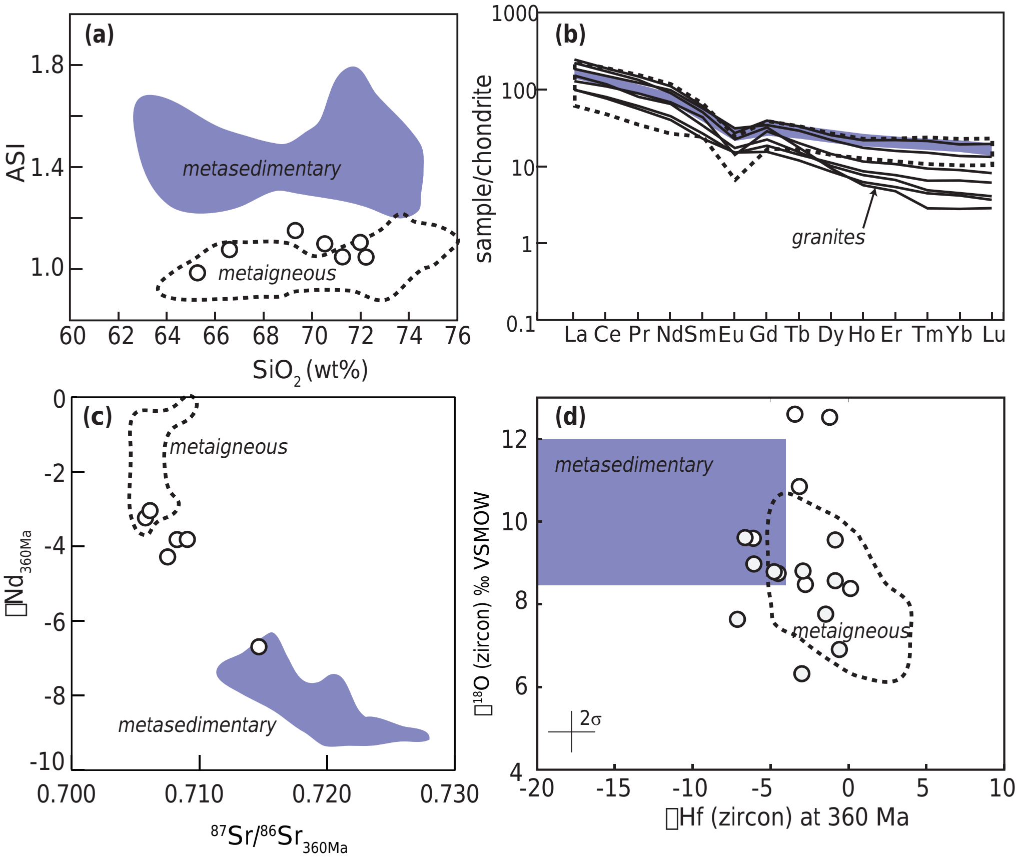

Fig. 5. Example of using element and isotope compositions of granites to infer their sources from western Marie Byrd Land in West Antarctica (modified from Yakymchuk et al., 2015b). The blue fields represent the range of measured compositions of a potential metasedimentary source and the dashed line outlines the values for the metaigneous source. The open circles represent granites that are inferred to be the result of anatexis of these two sources in the late Devonian and early Carboniferous. (a) Aluminum saturation index (ASI) of granites and the two potential sources. (b) Chondrite-normalized rare earth element concentrations. (c) Whole-rock Sr and Nd isotope ratios calculated at 360 Ma. (d) Zircon Hf and O isotope ratios from two samples. The cross in the lower left represents the average uncertainty on each data point. Note the large compositional heterogeneity of both the metaigneous and metasedimentary source rocks.

metaigneous source and one plots in the field for the metasedimentary source, most plot in between the two sources (Fig. 5c). This could be interpreted as mixing of magma derived from the two sources or isotope disequilibrium. The intimate layering of both source rocks at depth (e.g. Korhonen et al., 2010a, b) and the presence of melt networks that transverse both rock types in the Fosdick migmatite–granite complex (Yakymchuk et al., 2013b, 2015a, b) suggest that mixing of magmas in the source is the most likely explanation although source heterogeneity probably also played an important role. Hafnium and oxygen isotope ratios of zircon from two Carboniferous granites and from the two potential sources are shown in Figure 5d. Again, most zircon analyses of the Devonian granites plot in the metaigneous source field but some plot in the metasedimentary source field. A key point of this is that although the two potential sources are quite different (e.g. sedimentary versus igneous), they are themselves quite heterogeneous. In particular, the Hf and O isotope ratios of some zircon from the Devonian metaigneous rocks overlap with those from the metasedimentary rocks. Some of the zircon values plot in the region where the values of the two sources overlap; it is unclear if these represent derivation from a less radiogenic (i.e. lower εHf) portion of the Devonian metaigneous rocks, a more radiogenic portion of the metasedimentary rocks, or a mixture of the two different rock types. Therefore, when possible, it is important to characterize and consider the heterogeneity of the potential source rocks in a granite suite. Indeed, much of the compositional and isotopic variation in granite suites can be explained mostly by source heterogeneity (Clemens and Stevens, 2016).

FUTURE DIRECTIONS

Studies of granite petrogenesis will continue to rely on field-based, laboratory-based and modelling-based approaches. Many of the petrogenetic models developed for mafic systems (i.e. crystal fractionation, magma mixing) that were applied to granitic systems are now being abandoned in favour of models that consider the heterogeneous composition of magmas extracted from their sources. A major step forward in our understanding of the variable compositions of magmas that can be extracted from different source rocks has come from thermodynamic modelling of anatectic processes in addition to experimental petrology. Only recently has phase equilibrium modelling been used to test different models of granite heterogeneity for S-type granites (Garcia-Aris and Stevens, 2017) and similar studies can explore compositional variations in I-type granites. The development of a new thermodynamic model for silicate melts (Holland et al., 2018) builds on previous work and now allows quantitative modelling from the liquidus to the solidus for granites in equilibrium with biotite and hornblende.

While radiogenic isotopes are still routinely used to understand the sources and evolution of granite suites, non-traditional stable isotope systems have begun to provide new insights into granite petrogenesis. These techniques can also be applied to mineralized granites to understand redox processes associated with granite petrogenesis using Fe isotopes (Foden et al., 2015; Wawryk and Foden, 2015) and evaluate the proportion of juvenile to evolved crust using Si isotopes (Savage et al., 2012), Mg isotopes (Shen et al., 2009) and Cu isotopes (Li et al., 2009).

Finally, the recent recognition of evolved silicic rocks on other planets such as Mars (e.g. Wray et al., 2013; Sautter et al., 2015) provides an opportunity to link our understanding of granite petrogenesis on Earth to that on other planets with different thermotectonic histories and crustal compositions. This may provide an analogue to crustal growth and reworking processes that operated in the early Earth (e.g. Harris and Bédard, 2015).

Acknowledgements: This admittedly biased review has benefited from numerous discussions with colleagues and students. Nonetheless, any errors or omissions are my own. I thank Sandeep Singh for the invitation to write this review and an anonymous reviewer for constructive comments.

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(Received: 13 October 2018; Revised form accepted: 30 January 2019)