

Origins and Scales of Compositional Variations in Crustally-derived Granitic Rocks: the Example of the Dartmoor Pluton in the Cornubian Batholith of SW Britain

*John D. Clemens^{1, *}, Paul A. Helps², Gary Stevens¹, Nick Petford³*

*1. Department of Earth Sciences, University of Stellenbosch, Private Bag XI, Matieland
7620, South Africa;*

2. Halliburton, 97 Jubilee Avenue, Milton Park, Abingdon OX14 4RW, UK;

3. University of Northampton, NN1 5PH, UK

* Author for correspondence; e-mail: jclemens@sun.ac.za.

ABSTRACT

The c. 280 Ma, post-orogenic, S-type Dartmoor pluton was assembled from numerous sheets of granitic magma, emplaced into the shallow crust. The main magma source lies in the middle crust, and is most probably Proterozoic metagreywackes, with minor metapelites and metavolcanic or plutonic rocks, possibly formed in a syn-collisional environment. Partial melting of this source may have occurred under fluid-deficient conditions, and the magmas most likely had relatively high initial H₂O contents. The pluton contains substantial, whole-rock-Sr and quartz-O isotope heterogeneities on scales down to a metre or less, and such small-scale heterogeneities are probably common in granitic intrusions derived from heterogeneous protoliths. Thus, variations in source terranes may not be fully captured with the sample numbers and scales commonly applied in studies of granitic plutons. The preservation of both large- and small-scale isotopic heterogeneities suggests that the

Dartmoor magmas were never efficiently homogenised by flow-driven mechanical mixing. This implies a source terrane with lithological variations on scales of tens of metres or less. The granitic rocks form five texturally, chemically and isotopically distinct groups, each of which had somewhat different sources or mixtures of sources. The main chemical variations cannot have been formed through fractionation of any combination of the major minerals in the rocks. Instead, entrainment of variable proportions of peritectic plagioclase, orthopyroxene and ilmenite was responsible, together with local crystal fractionation. Low-density, late-magmatic melts and aqueous fluids produced patchy enrichment in light elements, and extreme enrichment in some of the highly silicic, two-mica microgranites. However, although they are also enriched in light elements, the ‘aplites’ were not produced through fractionation, and seem to have had independent magmatic origins.

Introduction

Worldwide, and in a wide variety of tectonic settings, granitic bodies commonly contain geological and geophysical evidence for sheet-like, pulsed intrusion dynamics, together with mineralogical and chemical evidence for zoning or patchy occurrence of contrasting rock types, as well as cryptic chemical or isotopic variations (e.g., Mohr 1991; Atkins and Lineback 1992; Evans et al. 1994; Metcalfe et al. 1995; Price and Pitcher 1998/1999; Miller and Paterson 2001; Mahan et al. 2003; Archanjo and Fetter 2004; Coleman et al. 2004; Bartley et al. 2006, 2012; 2018; Kratinová et al. 2007; Paterson et al. 2008; Žák et al. 2009; Farina et al. 2010; Miller et al. 2011; Eddy et al. 2016; Clemens et al. 2017a; Wallis and Clemens 2018). Such heterogeneities evoke questions regarding the scales on which these can occur, and the mechanisms responsible.

Some silicic plutonic rocks contain feldspar crystals that record changes in isotope ratios during their growth (Gagnevin et al. 2005a, b; Siebel et al. 2005; Davidson et al. 2006; Jiang

et al. 2018). The interpretation is that the growing crystals were exposed to isotopically distinct magma fractions, and this has been ascribed mostly to magma mingling and mixing, although wall-rock assimilation has also been suggested (e.g., Knessel et al. 1999). At the other end of the scale, granitic plutons commonly contain kilometre-sized, whole-rock, isotopic heterogeneities (e.g., in initial $^{87}\text{Sr}/^{86}\text{Sr}$ and/or ϵNd), especially in cases where a large magma component was crustally derived. Well documented examples include Halliday et al. (1980), Hill et al. (1986) Deniel et al. (1987), Mohr (1991), Barbero et al. (1995), Krogstad and Walker (1996), Pressley and Brown (1999), and Tsuboi and Suzuki (2003). Such large-scale heterogeneities have been attributed to a wide variety of processes, such as source variability (Halliday et al. 1980; Mohr 1991), resetting by thermal events (Roddick and Compston 1977), hydrothermal alteration (Juteau et al. 1984; Lutz and Srogi 1986; Poitrasson et al. 1996), very prolonged fractionation (McCarthy and Cawthorn 1980), patchy contamination through local assimilation of country rocks (i.e., AFC, Taylor 1980; DePaolo 1981; Aitcheson and Forrest 1994 or FCA, Cribb and Barton 1996), mixing and mingling with mantle-derived or hybrid mafic magmas (DePaolo and Wasserburg 1979; Bergantz 2000), and/or the occurrence of isotopic disequilibrium during crustal anatexis (Ayres and Harris 1997; Knesel and Davidson 1999; Davies and Tommasini 2000). Many of these mechanisms have also been proposed to be the causes of whole-rock, major- and trace-element variations, along with additional models, such as: primary magma heterogeneity due to variable entrainment of peritectic mineral assemblages (Stevens et al. 2007; Taylor and Stevens 2010; Clemens and Stevens 2012; Bailie et al. 2020), differentiation by progressive restite unmixing (Scambos et al. 1986; Chappell et al. 1987; Williamson et al. 1997) and simple Rayleigh fractionation of a parent liquid, coupled with crystal accumulation.

In many types of crustal source rocks, partial melting takes place through reactions that involve mica decomposition. In these situations, Sr isotope disequilibrium occurs mainly

because Rb-rich micas selectively contribute much of their stored radiogenic Sr to the magmas when they break down in the melting reactions. Thus, in cases where a source has aged significantly between the time of formation of the mica crystals and the time of partial melting, the melts may inherit significantly higher $^{87}\text{Sr}/^{86}\text{Sr}$ than either the original protolith or its restitic remnant. As demonstrated experimentally by Hammouda et al. (1996), and explored further by Farina and Stevens (2011), if the relative contributions of biotite and plagioclase to the melt vary in the source (e.g., due to small-scale differences in H_2O activity, plagioclase composition or other bulk compositional variations) and, as is likely, mica and feldspar do not isotopically equilibrate prior to anatexis, then the magmas will inherit ranges of Sr isotope ratios. Similarly, Nd isotope variation can be caused by a lack of equilibration between melts and residual monazite and apatite. On the other hand, entrainment of accessory phosphate minerals can explain why some magmas might inherit Sr but not Nd isotope heterogeneity. Such decoupling of the isotope systems can occur because a significant proportion of the Sr is carried in the melt while a large fraction of the Nd enters the magmas in entrained accessory minerals. As concluded by Pichavant et al. (1996), this likelihood of disequilibrium effectively severs the direct connection between source and magma, for at least some isotope systems, and makes identification of possible protoliths for granitic plutons more complicated. Furthermore, as pointed out by Farina et al. (2010), lack of equilibration between melt and residual minerals, due to rapid melt segregation, can accentuate any primary variations in $^{87}\text{Sr}/^{86}\text{Sr}$ in an already heterogeneous protolith. Currently, we have little information on whether these kinds of heterogeneities are also present on scales between those of the individual crystal and the geological map. Among other reasons, this is important because it bears on our ability to characterise the isotopic signature of a granitic pluton using typical sample numbers and sample spacings. To examine some of these issues related to the formation of chemical and isotopic heterogeneities in granitic bodies, we focus on the

example of the Dartmoor pluton in SW Britain – part of the renowned Cornubian batholith of Devon and Cornwall.

The Cornubian Batholith and its Regional Setting

Among other movements, the Variscan Orogeny involved late Silurian to Carboniferous collision between the Armorican and Avalonian terranes, with the consumption of small oceans (Arenas et al. 2014; Franke et al. 2017; Dijkstra and Hatch 2018). Following the diachronous docking, Late Devonian to Permian, post-tectonic, granitic bodies were emplaced in SW Britain, the Armorican Massif, the Massif Central and the Vosges Mountains in France, the Black Forest in Germany, the Bohemian Massif (in the Czech Republic, eastern Germany, southern Poland and northern Austria), into the Variscan basement in the Alps, and further south in Corsica, Sardinia, and Iberia (e.g., Exley and Stone 1982; Schermaier et al. 1997; Dijkstra and Hatch 2018).

The Permian Cornubian batholith, in SW Britain, is composed of five major plutons and many smaller bodies (figs 1 and 2) that were emplaced between ca 295 and 270 Ma (Darbyshire and Shepherd 1985; Chesley et al. 1993; Moscati and Neymark 2020).

The Cornubian plutons intruded Devonian to Carboniferous, mostly marine, clastic and dominantly sub-greenschist facies, metasedimentary and mafic metavolcanic units (Exley and Stone 1982; Chen et al. 1993). Regional metamorphism predated batholith emplacement (e.g., Sanderson 1984; Leveridge et al. 1990), and the plutons carry no regional fabrics. They have sharp, discordant contacts with the country rocks and are surrounded by hornfels aureoles (Exley and Stone 1982; Chesley et al. 1993). Shallow emplacement levels are thus inferred (Darbyshire and Shepherd, 1994), probably at depths < 2 km (viz., $P \approx 50$ MPa, Chappell and Hine 2006).

The strongly peraluminous granitic rocks are typically coarse-grained, porphyritic or megacrystic, biotite and biotite-muscovite or biotite-cordierite monzogranites to syenogranites, with local occurrences of lithium- and fluorine-rich variants (Stone 1975; Exley and Stone 1982; Manning and Exley 1984; Hill and Manning 1987). The batholith would be classified as S-type, according to the criteria of Chappell and White (1974, 2001), with the implication that the majority of the magmas had aluminous, metasedimentary sources. The strongly peraluminous character of the granitic magmas resulted in the crystallisation of aluminous biotite, together with more highly aluminous minor minerals such as cordierite (in the Land's End and Dartmoor plutons), almandine garnet (in Dartmoor), tourmaline and even andalusite (in Bodmin Moor and Carnmenellis). Radioactive accessory minerals, such as apatite, monazite, xenotime, zircon and uraninite are generally relatively abundant.

The Cornubian granites have been subdivided on the basis of their mineralogy, texture and chemistry (e.g., Brammall and Harwood 1923; Brammall and Harwood 1932; Hawkes and Dangerfield 1978; Dangerfield and Hawkes 1981; Exley and Stone 1982; Exley et al. 1983; Stone and Exley 1985; Hill and Manning 1987; Simons et al. 2016, 2017). Medium- to coarse-grained granitic enclaves, present in some plutons, are interpreted as the disrupted remains (xenoliths) of earlier plutonic pulses (see, e.g., appendix 1, fig. 1.3), whereas the origins of finer-grained granodioritic enclaves are debated (e.g., Stimac et al. 1995).

An economically important belt of hydrothermal Sn-W-As-Fe-Cu-Pb-Zn mineralisation (e.g., Dines and Phemister 1956; Hosking 1964; Hawkes 1974; Exley and Stone 1982) and associated silicic dyke swarms is both temporally and spatially associated with the granitic plutons. The granites and the associated silicic dykes also exhibit a range of patchy, late- and post-magmatic alteration, including tourmalinisation, greisenisation and kaolinisation (e.g.,

Brammall and Harwood 1923; Exley 1961, 1976; Power 1968; Bristow 1968; Exley and Stone 1982).

In terms of petrogenesis, for the batholith as a whole, Darbyshire and Shepherd (1994) interpreted the depleted-mantle Nd model ages of 1.3 to 1.8 Ga to indicate a heterogeneous crustal magma source. They also showed that the radiogenic isotope characteristics of the Palaeozoic country rocks mean that these are unlikely to have formed the sources of any of the Cornubian granitic magmas, the true sources being structurally beneath the Palaeozoic rocks and older than 1.2 Ga. Hampton and Taylor (1983) took a different approach, surveying Rb-Sr isotope data together with their own data on Pb isotopes in feldspars from various granitic plutons in southern Britain. These authors concluded that the basement rocks here have ages of about 0.8, but no more than 1.2 Ga. Some of the I-type granitic rocks included in the study of Hampton and Taylor (1983) crystallised from magmas that contained significant volumes of apparently juvenile mantle components. Such components are not present in the Dartmoor pluton but are evident in the magma-mixing and mingling textures shown, for example, in the Early Devonian Shap Granite (Cox et al. 1996; Miles and Woodcock 2018). Using the methods of Hampton and Taylor (1983), this process would cause basement ages to be underestimated. Also, for the S-type magmas in particular, if some of the apatite and monazite in the metasedimentary source rocks were inherited as detrital grains from more ancient protoliths, the Nd model ages of the immediate source-rocks could represent overestimates. In any case, the inferred ages of the source rocks for the Cornubian granitic magmas are certainly Proterozoic.

Darbyshire and Shepherd (1995) also noted that the Sr and Nd isotope data for the Cornubian batholith do not define a simple, binary, mixing array between a mantle-derived melt and Proterozoic crust (such as that exposed in nearby Wales and Brittany). They concluded that the magmas were most likely formed through partial melting of Proterozoic,

heterogeneous, immature, clastic, metasedimentary and metavolcanic rocks in the middle to lower crust (beneath the Palaeozoic rocks), in conjunction with some minor mixing with basaltic magmas from slightly enriched mantle sources. The notion of potential mixing with mantle magmas is based purely on Nd and Sr isotopic variations that can also be interpreted as purely crustal in origin (see e.g., Clemens et al. 2011). Since there is no geological evidence of large-scale mixing and mingling between mafic, mantle-derived magmas and granitic magmas, at least at emplacement levels in the Cornubian batholith, we apply the principle of *lex parsimoniae*, and prefer the simpler model of magma derivation by partial melting of heterogeneous crustal sources.

The Dartmoor Pluton

The Dartmoor pluton is the largest and easternmost in the Cornubian batholith (fig. 1, inset). At higher elevations it forms an often bleak and treeless landscape of moors dotted with sheep farms and rounded outcrops known locally as tors (e.g., appendix 1, figs 1.1 and 1.2). The rocks are mainly granites (monzogranites and syenogranites) that have been dated to 280 ± 1 Ma using a whole-rock Rb-Sr isochron (Darbyshire and Shepherd 1994). Chesley et al. (1993) used TIMS to produce a weighted-mean monazite U-Pb date of 279.5 ± 1.0 , and also published an $^{40}\text{Ar}/^{39}\text{Ar}$ date of 280.3 ± 1.0 Ma. Using a LA-ICP-MS, U-Pb technique, Moscati and Neymark (2020) dated two samples of cassiterite from tin deposits within the pluton, yielding ages of 286.0 ± 1.8 and 284.1 ± 1.3 Ma. Using all available Rb-Sr data, including our new analyses, for all rock types (see later), a maximum-likelihood regression, using IsoplotR (Vermeesch 2018) with the revised decay constant of Rotenberg et al. (2012) and assumed errors of 0.5 and 0.006 for $^{87}\text{Rb}/^{87}\text{Sr}$ and $^{87}\text{Sr}/^{86}\text{Sr}_0$, respectively, yields an age of 283.0 ± 0.4 (2σ) Ma. Thus, the Dartmoor granitic rocks are certainly Artinskian to Kungurian in age, and we adopt a reference date of 280 Ma.

The Dartmoor magmas were emplaced into low-grade, regionally metamorphosed, predominantly marine, Devonian and Carboniferous, sedimentary rocks (Exley and Stone 1982). Gravity surveys show that the magmas intruded from a feeder zone in the south of the pluton and flowed northward, exploiting the Devonian-Carboniferous contact (Taylor 2007). The composite character of the pluton is well documented, with map-scale variations in chemistry, mineralogy and texture (Exley and Stone 1982). Chen et al. (1993) published monazite U-Pb dates for two textural varieties of Dartmoor granite (megacrystic and poorly megacrystic; see below) and concluded that these were emplaced a few Myr apart. They, therefore, excluded any model for the evolution of these two major rock units by differentiation from a common parent magma (c.f. Ward et al. 1992 [specifically for Dartmoor] and Chappell and Hine 2006 [for the batholith in general]). With the exception of scarce igneous, microgranular enclaves (MEs), the Dartmoor granitic rocks show no physical evidence of magma mingling or mixing (e.g., in the form of ‘mafic’ enclaves, synplutonic dykes or disequilibrium textures in feldspar crystals), so it is unlikely that emplacement-level magma mixing played a part in producing the chemical variations, unless all end members in the process were silicic. Rare hornblende- and titanite-bearing microgranitic to microgranodioritic enclaves are present, and Stone (1995) suggested that these signify the presence of a mantle-derived magma component. Medium-grained granitic to granodioritic enclaves, with igneous textures, are common in granitic plutons, worldwide. These are best explained as quench-textured globules of small volumes of slightly more mafic hybrid magmas that intruded into and chilled against the host granitic magmas or the pluton’s wall rocks (e.g., Clemens et al. 2017*b, c*). To that extent, these enclaves represent a form of magma mingling, and may carry a finite mantle component, but they cannot have been responsible for any significant degree of hybridisation of their host magmas (e.g., Clemens et al. 2017*b, c*). Alternatively, these compositional features could simply reflect the heterogeneous character

of the crustal magma source regions. Thus, in the case of the Dartmoor pluton, it seems reasonable to conclude that minor volumes of I-type magmas accompanied the overwhelmingly dominant S-types. There are no compelling geological, mineralogical or isotopic reasons to hypothesise the presence of a significant juvenile mantle component in the Dartmoor magmas (see also Clemens et al. 2011).

Although scarce wall-rock xenoliths are present, models involving assimilation of the metasedimentary country rocks have been rejected, on a variety of grounds (Hall 1988; Ward et al. 1992; Boyd et al. 1993; Darbyshire and Shepherd 1994). In any case, Glazner (2007) showed that bulk assimilation processes (both reactive and mechanical) are limited strictly by the energy requirements. For relatively cool silicic magmas, these energy constraints are particularly severe and make significant amounts of wall-rock assimilation highly unlikely. The peraluminous chemistry of the rocks is judged to be a primary, source-derived characteristic. Although there are localised zones of hydrothermal alteration within the pluton, Darbyshire and Shepherd (1985) used the congruence of whole-rock Rb-Sr and biotite Ar-Ar plateau ages, of fresh rocks from the various quarries, to conclude that, in such samples, the Rb-Sr system was undisturbed by post-magmatic processes. Thus, for this granitic body, we can discount most potential causes of radiogenic isotopic variation (magmatic, hydrothermal and weathering); the heterogeneities appear to reflect near-source processes and materials.

Mappable Units

The Dartmoor pluton contains abundant evidence for pulsed intrusion dynamics (Exley and Stone 1982). Figure 1 shows the rock-units recognised here, together with the corresponding intrusion stages identified by Brammell and Harwood (1923). Figure 2 and appendix 1, figure 1.2 illustrate a magmatic sheet of poorly megacrystic microgranite that intruded and chilled against an older, coarse-grained megacrystic granite. Brammell and Harwood (1923) Stage I

produced a biotite microgranodiorite. Rocks of this stage were not sampled in the present work, and there appear to be no analyses of them in any published study, all extant analyses being monzogranitic to syenogranitic. However, this granodiorite is reported to occur as large xenoliths or rafts within Stage II, which is coarse-grained, megacrystic biotite monzogranite. Stage III is an equigranular, poorly megacrystic granite, which Brammall and Harwood (1923) considered to have intruded Stage II. However, the relatively scattered and fragmentary nature of the outcrops mean that there are no exposed contacts between Stages II and III. This led Edmonds et al. (1968) to interpret the coarse-grained megacrystic biotite monzogranite as a marginal facies, gradational into the poorly megacrystic biotite granite. Stage IV is represented by various ‘aplitic’ and microgranitic bodies, mainly dykes.

Petrographic Observations

We divided the Dartmoor granitic rocks into five textural groups, each of which also has some distinctive chemical and isotopic characteristics (see later). The groups that we recognise are:

- coarse-grained megacrystic granites (mainly biotite monzogranites) (e.g., appendix 1, fig. 1.3),
- medium- to coarse-grained poorly- to non-megacrystic granites (e.g., fig. 2),
- a garnet-bearing microgranite that formed a large xenolith exposed in Sweltor Quarry,
- fine-grained, two-mica microgranites in small late-intruded bodies and
- ‘aplitic’ bodies (mainly dykes, but including a larger body exposed in the Meldon Quarry, itself intruded by smaller ‘aplitic’ dykes). The use of quote marks is explained below.

The last two groups intrude the earlier, much larger, medium- to coarse-grained rock bodies.

Mineralogically, cordierite and especially garnet, are relatively uncommon, and the most

mafic rocks contain no more than 15% biotite and 20% plagioclase. We did not sample the garnet-bearing rock from Sweltor Quarry (Stone 1988) or the two-mica microgranites that were analysed by Darbyshire and Shepherd (1985) and Ward et al. (1992), respectively. Both these works contain detailed petrographic descriptions of some of the granitic rock textural varieties. However, the following sections briefly present our own mineralogical and textural observations, both on the Dartmoor rocks and, for comparison, our knowledge of S-type granites in general. Appendix 1 provides field photographs and photomicrographs of some of our samples. Note that the modal proportions that we provide are for our samples. We do not present these as fully characteristic of the map units.

Coarse-grained megacrystic granites. Sample DP12 (collected from a former quarry NNW of Hay Tor) is reasonably typical of this textural variety. It is a biotite syenogranite with Carlsbad-twinned K-feldspar megacrysts up to 7 cm long, set in a groundmass with crystals that typically range from 2 to 4 mm. Orientation of the K-feldspar megacrysts forms localised flow foliations. The mode is approximately 10% biotite, 35% quartz, 40% K-feldspar and 15% plagioclase. Accessory minerals are zircon, apatite, monazite, xenotime, ilmenite, pyrrhotite, tourmaline, secondary muscovite and rare secondary chlorite. Some large biotite and quartz crystals formed early in the crystallisation sequence but these minerals continued to crystallise down to near-solidus conditions. Thus, anhedral to subhedral biotite (with yellow to brown to red-brown pleochroism) and anhedral quartz are found mutually intergrown, suggesting simultaneous late crystallisation. Titanian and aluminous biotite is prevalent in S-type rocks, worldwide. In the Dartmoor pluton, much of this biotite occurs as relatively small anhedral, late-crystallised grains. This contrasts with most S-type rocks, in which the bulk of the biotite is early-formed, occurring as large subhedra or euhedra. As is usual in S-type rocks, the larger Dartmoor biotite crystals are sprinkled with inclusions of the

accessory minerals, many with intense halos of radiation damage. These halos are both more numerous and broader than is usual in S-type granites, suggesting high U and Th contents in the Dartmoor zircon and monazite. Plagioclase forms tabular to equant, subhedral to anhedral crystals, some of which have faint oscillatory zoning. This represents another contrast between the Dartmoor rocks and most other S-type rocks, which commonly contain plagioclase with complex and prominent oscillatory zoning. In DP12, plagioclase began to crystallise after biotite and the early subhedral quartz, while the groundmass K-feldspar formed late, as it is interstitial to all other minerals. This order of crystallisation is common to most S-type granitic rocks. The large sizes of the K-feldspar megacrysts suggest that this mineral also crystallised rapidly as late-nucleated phenocrysts (e.g., Vernon 1986; Holness et al. 2018). Rare pseudomorphs after orthopyroxene consist of ~ 2 mm diameter clusters of small (< 0.3 mm) anhedral biotite crystals and even smaller anhedral quartz grains. Again, the occurrence of such orthopyroxene pseudomorphs is widespread in S-type granites, worldwide, especially in the more mafic types. Photomicrographs of textural features in similar rocks to DP12 are shown in appendix 1, figures 1.5, 1.8 and 1.9.

Medium-grained non-megacrystic granites. Sample DP1, from the NW side of Eastmill Tor, in the N of the pluton (fig. 2), is a typical example. The rock is hypidiomorphic granular in texture, and lacks foliation. This syenogranite contains approximately 10% biotite, 35% K-feldspar, 35% quartz and 20% plagioclase. The biotite shows straw yellow to brown to red-brown pleochroic colours. Accessory minerals include zircon, apatite, monazite, xenotime, tourmaline, ilmenite, pyrrhotite, secondary chlorite (after biotite) and secondary muscovite. Equant, subhedral quartz crystals are up to 5 mm across, and these are set in a groundmass of anhedral to subhedral biotite, quartz and plagioclase, in which the grains range in size from < 0.5 to 3 mm. Most S-type granitic rocks contain some early-formed quartz as subhedra to

euhedra (with slightly rounded edges and corners). This is interpreted as near-liquidus quartz saturation, in magmas typically formed from quartz-bearing metasedimentary source rocks. In the groundmass of DP1, biotite, quartz and plagioclase can all be found moulded on each other, suggesting simultaneous crystallisation. K-feldspar began to crystallise later, as it poikilitically encloses smaller crystals of the other minerals. The plagioclase has some normal zoning, but oscillatory zoning is not apparent — unusual for an S-type granitic rock. Like DP12, DP1 also contains some fine-grained biotite-quartz pseudomorphs after orthopyroxene, which measure up to 2 mm across. Photomicrographs of similar rocks are shown in appendix 1, figures 1.4, 1.7 and 1.10.

A different variety of this textural type occurs as the subhorizontal sheet that intrudes the coarse-grained megacrystic granite at Hay Tor, as shown in figure 1. Sample DP7, from this locality, is an unfoliated, porphyritic biotite cordierite microsyenogranite with phenocrysts of K-feldspar, quartz and plagioclase (1.5 to 4.5 mm), set in a fine-grained (< 0.2 to 1 mm) groundmass. The modal proportions are approximately 40% quartz, 35% K-feldspar, 15% plagioclase, 6% cordierite and 4 % biotite. Accessory minerals include tourmaline, ilmenite, zircon, apatite, muscovite and secondary chlorite. The anhedral to subhedral biotite displays straw yellow to brown to red-brown pleochroism and is generally interstitial to quartz. Plagioclase is interstitial to both quartz and biotite, and shows no oscillatory zoning – a signature feature of the Dartmoor rocks, but unusual in S-types, globally. K-feldspar is mostly interstitial and encloses smaller crystals of quartz, biotite and plagioclase. The cordierite is subhedral to euhedral, which is interpreted as indicating early formation. This early saturation in cordierite suggests that the magma crystallised at low pressure and also had a relatively high H₂O content (e.g., Clemens and Wall 1981).

'Aplitic' rocks. Many of the rocks that have been grouped here (and in the past) as 'aplites' do not show the archetypal, aphyric, saccharoidal texture, which results from pressure quenching of highly evolved, eutectoid liquids. Instead, they are very felsic and fine-grained porphyritic microgranites with phenocrysts that reach only 1 to 2 mm in size. These rocks most probably crystallised from magmas that were compositionally further from the eutectoid compositions of genuine aplites. This lack of aplitic texture and inferred mode of formation is the reason for our use of single quote marks (i.e., 'aplite'). Their microporphyritic texture is exemplified by sample DP4, collected from the west slope of Eastmill Tor. This rock is a biotite-free porphyritic micromonzogranite. It contains phenocrysts (normally 1.5 to 2 mm) of subhedral, equant quartz and plagioclase, and anhedral K-feldspar set in a fine-grained (0.2 to 1 mm), hypidiomorphic-granular groundmass of quartz, K-feldspar, plagioclase and tourmaline prisms. The modal proportions are approximately 35% quartz, 30% K-feldspar, 25% plagioclase and 10% tourmaline. Accessory minerals are scarce and limited to muscovite (possibly primary) and rare apatite needles. Magmatic muscovite, possibly F- and B-bearing (e.g., Thomas et al. 2003), could occur at the inferred low emplacement pressure if the solidus were depressed due to the presence of F and B in the melt (Chorlton and Martin 1978; Manning 1981; Pichavant 1981; Pichavant and Manning 1984), also leading to the formation of suprasolidus tourmaline. The K-feldspar microphenocrysts are commonly poikilitic, enclosing tiny crystals of earlier-formed quartz, plagioclase and tourmaline; the quartz and tourmaline show mutual inclusion relations, indicating simultaneous magmatic crystallisation. A thin-section view of a very similar rock (sample DP5, a small dyke from the Meldon Quarry) is illustrated in appendix 1, figure 1.11. The main Meldon 'aplite' is texturally similar but is considerably enriched in Li, B, F, Na, Al and P, so that it contains fluorite, fluorapatite, beryl, beryllonite, datolite, herderite, topaz, cordierite, muscovite, amblygonite, montebrasite,

petalite, lepidolite and elbaite, as well as a variety of W, Nb and Ta minerals (e.g., von Knorring and Condliffe 1984).

Whole-rock Chemistry

We present 36 new analyses, including 7 from ‘aplitic’ rocks, which, apart from the Meldon ‘aplite’, appear to have escaped previous analytical attention. Additional major-oxide and trace-element data are from Darbyshire and Shepherd (1985), Ward et al. (1992) and Chappell and Hine (2006). We also include some analyses from Simons et al. (2016). In this case, we correlate their G5b with our ‘aplitic’, G3a with our coarse-grained megacrystic granites and G3c with our medium- to coarse-grained poorly to non-megacrystic granites. We have not used their G2 and G4b muscovite and tourmaline granites, as these do not appear to correspond to any of our main rock units. To maintain an internally consistent dataset, the REE analyses in Simons et al. (2016) are also not used here because there are systematic offsets and slope differences of the patterns calculated from our results, as well as from those of Darbyshire and Shepherd (1985), Ward et al. (1992) and Chappell and Hine (2006). Edmonds et al. (1968) contains an additional 12 major-oxide analyses, gleaned from publications that date from 1912 to 1965, with many carried out in the 1930s. Some of the analyses in their table 5 are of samples from outcrops, rather than quarries, and these show chemical signs of weathering and alteration. Also, even the analyses of rocks sampled in quarries appear to have been subject to analytical problems, particularly for TiO_2 , Al_2O_3 , FeO and CaO . Thus, we have excluded the analyses in Edmonds et al. (1968). Watson et al. (1984) published 5 analyses of rocks from the Dartmoor pluton, but did not provide sample locations or textural descriptions. We have also omitted these from consideration because we cannot reliably attribute them to any of our recognised rock units. Two previously unpublished analyses, of the large, garnet-bearing granite xenolith in Sweltor Quarry, were provided by F.

Darbyshire (formerly of the BGS/NERC Isotope Geosciences Facility). Analyses of three igneous microgranular enclaves are from Stone (1995), and a further two are from Stimac et al. (1995). The methods and details for our new analyses are given in appendix 2, and the compiled database is presented as appendix 3.

Major oxides. Figures 3 to 6 illustrate some of the chemical variations among the Dartmoor granitic rocks. SiO₂ contents vary from ~ 71 to 78 wt% and, as is typical in granitic suites, and particularly in S-types (e.g., Sweetman 1987; Clemens and Phillips 2014; Clemens et al. 2017c), the major-oxide variations show significant scatter (fig. 3). Essentially the whole pluton consists of strongly peraluminous (fig. 3f), high-K rocks (fig. 3d). In figure 3e, the megacrystic granites (plotted as red squares) define a crudely flat array; their Mg#s do not vary systematically, even though their SiO₂ contents range from ~ 71 to 76 wt%. During differentiation by crystal-liquid separation (or mafic-silicic magma mixing), it is expected that Mg# should steadily decrease as SiO₂ increases. For the non-megacrystic rocks (plotted as orange dots) there *is* such a rudimentary trend, though it shows increasing scatter toward the highest SiO₂ contents. Several plots in figure 3 show clearly that variations among the ‘aplitic’ rocks (plotted as green triangles) contrast with those of the other groups. They are not simply magmas selectively depleted in FM and TiO₂, leading to offsets from the main trends, but some oxides show trends that run contrary to those for the rest of the rocks. Also, in no case do the aplites form extensions of the trends for the rest of the Dartmoor rocks. These features are best seen in figures 3a-c, for TiO₂, Al₂O₃ and FM (= FeO^T+MnO+MgO). In figure 3b, it is apparent that the ‘aplitic’ rocks have higher Al₂O₃ than non-megacrystic rocks with the same SiO₂ contents. These chemical features suggest that the ‘aplites’ constitute a separate magma series, that was not produced through fractionation of any of the other Dartmoor magmas. This inference is supported by the fact that, on several of the variation diagrams

(e.g., figs 3b, d and f), the Meldon ‘aplite’ (shown as a green triangle with a broad black border) plots somewhat remote from any of the other Dartmoor rocks, including most other ‘aplites’. Taken together, these relationships suggest that there are three (or more) magma series in the pluton, produced from source rocks that may occur in the same terrane, but with significant lithological and chemical differences. These preliminary suggestions are explored further below.

Trace elements. Figure 4 shows plots illustrating the variations of Ba and Zr in the Dartmoor granitic rocks. In figure 4a, Ba concentrations are plotted against K_2O . Since Ba strongly partitions only into the K-rich minerals biotite and K-feldspar (Table 1), depletion or concentration in these minerals should result in a definite positive correlation between K_2O and Ba. It is evident that there is no such correlation among either the megacrystic or non-megacrystic rocks. Indeed, if the degree of scatter in the data is ignored, for the moment, there is a weak negative correlation between Ba and K_2O among the megacrystic rocks. Thus, K-feldspar or biotite fractionation or accumulation is highly unlikely to have been responsible for chemical variations in most of the Dartmoor rocks. A possible exception to this is the ‘aplites’, which do show a weak positive correlation between Ba and K_2O .

Variations in Zr concentration with whole-rock SiO_2 are illustrated in figure 4b. As for Ba, there is considerable scatter in the points for the megacrystic rocks, and no correlation between ~ 71 and 74 wt% SiO_2 , though the points at higher SiO_2 contents appear to show a negative correlation, a common feature in granitic series. Moreover, with one exception, the points for the non-megacrystic rocks define a rather good negative correlation ($r^2 = 0.61$, including all points), which suggests a distinct role for zircon. The ‘aplitic’ rocks too seem to form a distinct correlation, though here it is positive, which suggests increasing Zr solubility in the magmas with higher SiO_2 . Zircon was not observed as an accessory mineral in the

'aplitic' rocks examined here. In granitic melts, Zr solubility increases as the melts become less peraluminous (Watson and Harrison 1983). Figure 3f shows that the 'aplitic' rocks with the highest SiO₂ are the least peraluminous, so this is in accord with the predicted trend. In F-enriched, H₂O-saturated granitic liquids, Zr solubility is also enhanced (Keppler 1993), and this may have been an additional influence. However, since the 'aplitic' rocks appear unlikely to represent differentiates of other known Dartmoor magmas (see e.g., figs 3 and 4, and especially fig. 3b), it may be that there was a primary source-related (bulk composition, temperature or fluid-condition) control on the Zr contents of the 'aplitic' magmas.

Figure 5 shows chondrite-normalised REE spectra for the Dartmoor rocks. On average, the megacrystic rocks (fig. 5a) have slightly higher REE concentrations than the non-megacrystic rocks (fig. 5c), which also have deeper negative Eu anomalies, as shown by lower values of Eu/Eu* ($= \text{Eu}_N / \sqrt{\text{Sm}_N \cdot \text{Gd}_N}$) in figure 6. Many of the 'aplitic' rocks (fig. 5b) show shallower negative Eu anomalies (higher Eu/Eu*) than the non-megacrystic granites, and instead have values comparable to those of the megacrystic rocks. The 'aplitic' rocks also have flatter patterns than any other group. The two-mica microgranites (fig. 5d) have distinctive patterns, unlike those of any other group, and both patterns for these have unusual negative Ho anomalies. As Ward et al. (1992) concluded, these magmas cannot have been formed through fractionation of the megacrystic granite magmas. Figure 6a shows that there is a general negative correlation between Eu/Eu* and SiO₂ content, which could be interpreted as the result of feldspar fractionation or accumulation. Indeed, the plot of Eu/Eu* against CaO (fig. 6b) is compatible with such an inference. However, the Eu/Eu* v. K₂O diagram in figure 6c shows a lack of correlation between these variables. Thus, K-feldspar cannot have been significantly involved. Instead, on the basis of the REE data, plagioclase would be the candidate for an accumulating or fractionating mineral.

Igneous Microgranular Enclaves (MEs)

Especially in the area around Haytor Rocks, the coarse-grained megacrystic granites contain an assortment of enclaves. These fall into two main categories – hornfelsed xenoliths of the metasedimentary country rocks, which we shall ignore because they have been shown to play no role in the genesis of the pluton variations (Hall 1988; Ward et al. 1992; Boyd et al. 1993; Darbyshire and Shepherd 1994), and fine- to medium-grained microgranular types with igneous textures and abundant biotite, known as MEs. Analyses of some Dartmoor MEs were given in Stimac et al. (1995) and Stone (1995). As described by Stimac et al. (1995) the quench textures of these enclaves are very similar to the types studied by Clemens et al. (2017*b*). An example of a chill rind, with the same texture as many of the MEs, surrounding a granitic xenolith, is shown in appendix 1, figure 1.3. In many granitic batholiths, these minor volumes of enclave magmas have isotopic characteristics that contrast with their host rocks, usually, but not always, with more primitive signatures than their hosts. Globally, with a few notable exceptions, their chemical characteristics show that they did not form mixing series with their hosts. Thus, in general, the presence of such MEs should not be taken as a sign of any significant degree of hybridism of the host magma. The volumetric relations dictate that it is the MEs that were hybridised (further), mainly through ingestion of host-derived crystals (Clemens et al. 2017*b*). The specific case of Dartmoor is examined below.

In figure 7, we show some variation diagrams with the analysed enclaves from Dartmoor plotted together with the array of analytical points for their host rocks – the megacrystic biotite granites. These plots show that linear extrapolations of the trends shown by the host rocks do not pass through the enclave arrays, meaning that the enclaves and hosts do not form parts of a mixing array. In some cases (e.g., in fig. 7c), the host-rock trend actually runs parallel to any apparent trend formed by the points for the MEs. If the enclaves were part of a differentiation series formed by fractional crystallisation, there could be an exponential rather

than a linear fit. However, as also shown in figure 7, such a relationship is also unlikely. As indicated by Clemens et al. (2017c), for such enclaves in general, these features demonstrate two important and interrelated points — 1) that the enclaves are not parts of magma mixing series with their hosts, and 2) that the processes that formed the ME magmas, and governed their compositional variations, are not the same as those responsible for the variations in the host magmas; see also Flinders and Clemens (1996). Note that Stimac et al. (1995) produced two analyses for ME sample 222J1 — one of the bulk ME (222J1A) and the other (222J1B) for its groundmass only, i.e., without the large feldspar crystals, which were probably regarded as having been captured from the host magma. As shown in figure 7, there is a small difference in composition. Figure 3A of Stimac et al. (1995) shows a sawn and polished section through this ME, from which it is evident that some of the host-derived feldspar crystals have rounded shapes that indicate partial resorption in the ME magma. This enclave actually contains large crystals of K-feldspar, plagioclase, biotite and quartz, all of which may have been xenocrysts ingested from the host magma as the latter flowed and caused plastic deformation (stretching and folding) of the enclave magma globule. Indeed, some flow alignment of the probable xenocrysts is evident in figure 3A of Stimac et al. (1995). All these processes are common in MEs, worldwide as well as at Dartmoor. The most consistent interpretation is that such enclaves represent small volumes of hybrid magma that coexisted with their host granitic magmas and were in various states of chemical disequilibrium with that host when they were chilled against it or the country wall rocks (e.g. Vernon 1984; Clemens et al. 2017b).

Feldspar, Phosphorus and Fractionation

As shown above and, contrary to expectation and previous work on the Cornubian batholith (e.g., Ward et al. 1992; Chappell and Hine 2006; Müller et al. 2006; Simons et al.

2016, 2017), it seems highly unlikely that crystal fractionation played any large role in producing chemical variations in the voluminous, main Dartmoor granitic rocks. Note that most previous work has assumed that the entire Cornubian batholith can be treated as a single fractionation series (e.g., Chappell and Hine 2006; Simons et al. 2016). Wall et al. (1987) cautioned against using an entire granitic rock suite, in a multi-pluton batholith, as a unit for consideration of models for the chemical variations. Instead, these authors recommend that individual plutons, or even parts of plutons are the appropriate units to consider. In any case, our result for Dartmoor contrasts markedly with the conclusions of Chappell and Hine (2006), who maintained that, for the Cornubian batholith, the variations in whole-rock and feldspar P concentrations could only be explained by fractionation. These authors state, but do not illustrate, that there is a strong correlation between normative corundum and P₂O₅ contents. Chappell and Hine (1995) demonstrated that the feldspars of the batholith appear to become more P-rich as the rocks become ‘more fractionated’ (i.e., lower in CaO and FeO^T, and higher in SiO₂ and K₂O). The proposition is that, as the liquids evolve, they become more peraluminous, and this increases P solubility (e.g., Bea et al. 1992; Pichavant et al. 1992; Wolf and London 1994), with the feldspars in the more fractionated rocks becoming markedly P-enriched (London 1992). Figure 8a shows a plot of P₂O₅ against ASI (mol. Al₂O₃/[CaO-3.33P₂O₅+Na₂O+K₂O]) for the Dartmoor rocks. There is no general correlation between these two parameters. Note, however, that the two very highly silicic and strongly peraluminous ‘aplitic’ rocks have very elevated P₂O₅ contents. Thus, for some extreme compositions, both these parameters are indeed elevated, although this does not constitute evidence that these parameters are correlated within the broader sample set of the ‘aplitic’ rocks. Indeed, the other ‘aplitic’ rocks, as well as the two-mica microgranites, plot with the rest of the Dartmoor granitic rocks, despite having SiO₂ contents that range up to 77.70 wt%.

For the Cornubian batholith, it has also been suggested that fluid-driven element redistribution resulted in strong correlations between P and elements such as Rb and U, which become concentrated in evolved fluid phases (e.g., Simons et al. 2016). However, for the Dartmoor pluton, no such correlation exists (e.g., fig. 8b), and this confirms that such fluid-driven fractionation was also at a minimum in this pluton, at least for the samples and analyses that we have available for unaltered rocks.

As we have seen from the chemistry, and will be confirmed below in the section on isotopic variations, the ‘aplites’ could not have formed by fractionation of other Dartmoor granitic magmas. We therefore suggest that the behaviour of P in the ‘aplitic’ rocks probably reflects peculiarities of the magma source rocks and the conditions under which they were partially melted, a subject to which we return later.

At this point we emphasise that we fully subscribe to the occurrence of local and, in some cases, extreme fractionation to produce enrichments in elements such as Li, Be, B, F and Sn in some of the Dartmoor (and other Cornubian) rocks. This enrichment was evidently mediated by episodes of fluid flux (e.g., Simons et al. 2017). We also suggest that, during crystallisation, magmatic liquids were partially separated from their crystalline matrices to create some of the granitic rocks with SiO₂ contents > 73 wt% (see later). Nevertheless, we hold that the primary cause of heterogeneity in the Dartmoor magmas was source-related rather than caused by magmatic processes. We address these issues later, in the section on the modelling of elemental variations.

Tectonic Environment of the Magma Source

Since the 1970s, trace-element-based diagrams have been used in attempts to determine the tectonic settings of ancient igneous or meta-igneous rocks. For granites (s.l.), the most commonly used diagrams are those devised by Pearce et al. (1984) and further explained in

Pearce (1996). As demonstrated by Twist and Harmer (1987), and indeed already flagged by Pearce et al. (1984), for ancient rocks, such diagrams do not necessarily constrain the tectonic environment during the generation of crustally-derived granitic magmas, and this is especially so for those generated in post-collisional settings. Rather, these diagrams indicate the probable tectonic setting in which a magma's source rocks are likely to have formed. Figure 9 shows the Dartmoor rocks plotted on the Rb v. Y+Nb diagram of Pearce et al. (1984). Note that these authors recommended that aplitic rocks should not be used in such plots because true aplites commonly show element depletions related to hydrothermal activity. As explained earlier, all our samples were taken from unaltered rocks, remote from zones of hydrothermal mineralisation. Samples such as DP5 (from Meldon Quarry) are strongly enriched in light and LIL elements, and may thus be considered compositionally akin to pegmatites, which Pearce et al. (1984) also recommended should not be used. Nevertheless, we have included most of the Dartmoor 'aplitic' rocks in figure 9. The exception is sample DP1, which is very enriched in Ti and Nb, and evidently contained some microscopic rutile.

In many post-orogenic granitic suites that we have previously studied (S-, I- and A-type), the analyses plot in clusters that overlap the triple point between the fields for syn-collisional, volcanic-arc and within-plate rocks. However, for Dartmoor, the points for all rock groups fall unequivocally within the syn-collisional field. Considering the caveats mentioned above, we interpret this to mean that the source rocks for the Dartmoor magmas were most likely formed in a syn-collisional setting. Bearing in mind the conclusions about ages of the sources for the batholith as a whole (Darbyshire and Shepherd 1994), we suggest that this is likely to have been a Proterozoic collisional event, possibly the Mesoproterozoic Grenvillian. Alternatively, Genvillian detrital material may have been deposited in the active plate-margin environment involved in the Neoproterozoic Cadomian orogeny.

Modelling Elemental and Isotopic Variations

In this section we use simple major-oxide and elemental variations to further constrain the likely causes of the variations shown in figures 3 to 6. The two types of plots used are: molal proportions of various major elements, plotted against maficity ($M = Fe+Mn+Mg$); and log-log plots of selected trace-element concentrations.

Trace-element diagrams. Figure 10 presents log-log plots of Sr v. Ba and Sr v. Rb, in which fractionation paths appear as simple vectors. We illustrate fractionation of 25 wt% each of plagioclase, K-feldspar and biotite since, apart from quartz, these are the major minerals in the Dartmoor rocks. Table 1 provides the modelling parameters used to produce the illustrated fractionation vectors. Figure 10a shows that, if the overall variation from the megacrystic to non-megacrystic magmas were ascribed to a fractionation process, it would necessarily involve > 30% of K-feldspar and smaller amounts of plagioclase (< 10%) either as accumulated or fractionated phases. If the ‘aplites’ were to be regarded as fractionation products, very much larger K-feldspar quantities would have to have been involved. We have already demonstrated that K-feldspar fractionation is incompatible with the Ba and K_2O covariations. Thus, models with significant K-feldspar involvement are infeasible. From this figure, an alternative model would involve fractionation of around 50% of biotite together with around 30% of plagioclase but, in view of the modal data on the rocks, both of these proportions are nonsensical, biotite being present in modal amounts < 15% and plagioclase < 20%. To form the ‘aplites’ in this manner, even more extreme and improbable quantities of biotite and plagioclase would need to have been involved.

Figure 10b shows that biotite fractionation could not have been a significant process in the Dartmoor magmas, though plagioclase fractionation could fit the variations if large amounts of feldspar (about 25 %) were to have been involved. Thus, when considered together, the two

plots in figure 10 offer no support for fractionation models, and indeed are somewhat contradictory with each other. This conclusion accords with our inferences, from the variation diagrams in figures 3 to 6, that at least three magma series are represented within the pluton. Note that Simons et al. (2016) modelled parts of the Cornubian batholith as a single fractionation series involving separation of up to 40% crystals (mainly biotite, K-feldspar and plagioclase), using crystal-liquid distribution coefficients that are, in some cases, radically at variance with those that we used, from the GERM Partition Coefficient (Kd) Database for rhyolitic systems (<https://earthref.org/KDD/>). In this context, we call attention to the corroborations and extensions of our results that can be reached through consideration of the major-element modelling and isotope data that we present below.

Molal-proportion diagrams. Molal-proportion diagrams provide a simple test for the feasibilities of models in which granitic magmas are interpreted as mixtures of crystals and anatectic granitic liquids. There are two such models. The first is the restite-unmixing differentiation model, originally proposed by White and Chappell (1977) and codified in Chappell et al. (1987). In this model, partial melting of a source rock leads to the formation of a diatexite that rises as a diapir, differentiating by progressive removal of restitic minerals from the liquid. The resulting rock series is interpreted as a linear mixture, with the most mafic rocks representing restite-rich cumulates and the most silicic as pure melts. The composition of the source rock would be represented within this spectrum, and is commonly regarded as that of the most mafic exposed rock (Chappell and White 1992; Chappell et al. 1999; Chappell 2003). This model is problematic. The implicit magma ascent mechanism is mechanically implausible and lacks geological realism, the presence of significant modal restite cannot be demonstrated, and the objective existence of linear mixing and of granite

suites is debateable (e.g., Wall et al. 1987; Clemens and Mawer 1992; Clemens et al. 1997; Vernon 2007; Clemens and Phillips 2020).

An alternative model is that of peritectic assemblage entrainment (PAE), originally proposed by Stevens et al. (2007) and codified in Clemens and Stevens (2012). In this model, no primary differentiation occurs (e.g., Clemens et al. 2009). Magmas leave their sources carrying varying proportions of small crystals of the peritectic minerals formed in the melting reactions. The ratios of these crystals in the peritectic assemblage are fixed by the stoichiometries of the melting reactions, but the overall proportion of the assemblage can vary between magma batches (or pulses). This mechanism is proposed to be the primary explanation for the existence of a positive correlation between TiO_2 and FM, despite that fact that the reservoir for TiO_2 in such granites commonly includes both ilmenite and biotite, which crystallise at different times and which have markedly different size, density and shape characteristics. Along with the peritectic minerals, the magmas can entrain accessory mineral grains (e.g. zircon and monazite) that are liberated during partial melting, through decomposition of their original host crystals. This process, is seen as the main cause of intraplutonic chemical variation in many granitic magmas. Fractionation processes can form overlays on the primary PAE signals and sometimes dominate at SiO_2 contents > 73 wt%. During magma ascent and emplacement, small entrained peritectic crystals would fall out of equilibrium once P - T conditions deviated from those of the source. These crystals would be consumed by dissolution or reaction and reprecipitation (Villaros et al. 2009; Clemens and Stevens 2016; Clemens et al. 2020b).

Figure 11 shows the Dartmoor analyses plotted on molal proportion diagrams that use the parameter M ($= \text{Fe} + \text{Mn} + \text{Mg}$) in units of moles/kg of rock, as the abscissae. The plots also illustrate mixing vectors for 5 wt% each of biotite, plagioclase, K-feldspar, orthopyroxene and garnet, and 1 wt% of ilmenite. Biotite, plagioclase and K-feldspar are included because these

are all major minerals present in the rocks. Garnet and orthopyroxene (or rather its pseudomorphs) do occur, in trace amounts, in some of the Dartmoor rocks, and because partial melting reactions in metagreywackes and quartzofeldspathic rocks will commonly produce peritectic plagioclase, garnet and ilmenite, along with orthopyroxene, irrespective of the fluid conditions in the source rocks (Clemens and Stevens 2012).

We assume that the more silicic (lower M) rocks, such as the two-mica microgranites and the non-megacrystic granites, crystallised from magmas that originally contained the highest proportions of melt. Also, although vectors are plotted for K-feldspar and biotite, for comparison purposes, neither of these minerals is likely as a peritectic phase though, in this case, there may have been some biotite in the restitic portion. Note that K-feldspar is expected as a peritectic mineral formed by fluid-absent biotite breakdown in simple systems, but experiments on complex natural compositions have shown that it either fails to nucleate or is dissolved within a few degrees of heating above the relevant melting reaction (e.g., Stevens et al. 1997). K-feldspar is not expected as a peritectic mineral in fluid-present melting reactions involving biotite (e.g., Vielzeuf and Clemens 1992). Likewise, miniscule amounts of biotite that might be formed in some muscovite melting reactions would not survive reaction overstepping or magma ascent (e.g., Clemens et al. 1997). Before turning to the interpretation of the variations, in terms of the PAE model, it is worth noting that the ‘aplitic’ rocks do not appear to form parts of the trends shown by the rest of the rock groups in the Dartmoor pluton (see e.g., fig. 11a, b and d). From this we infer that whatever processes were responsible for the production of the ‘aplitic’ magmas, they had little to do with those that were responsible for the variations in the rest of the Dartmoor magmas. As suggested above, the ‘aplites’ seem to have formed from a discrete group of magmas, whose origins should be considered separately. In a similar line of reasoning, Ward et al. (1992) used the REE variations among the Dartmoor granitic rocks to conclude (p. 785, op. cit.) that “... the FG-granite cannot have

formed from a residual liquid left by crystallization of the CGM-granite.”. In our present terminology, this translates as saying that the magmas that formed the two-mica microgranites (blue in fig. 1) could not have formed as liquids derived through fractionation of the magmas that formed the coarse-grained megacrystic granites (pink in fig. 1). This conclusion is also evident from the concentrations of Y and Zr (e.g., fig. 4b).

From figure 11a, for Ti, it seems that the series of rocks trending from the two-mica microgranites to the megacrystic granites could have formed through entrainment of ilmenite and either orthopyroxene or garnet (e.g., about 8% garnet or 3% orthopyroxene plus < 1% ilmenite). In figure 11b, for Al, the slope of the variation is almost parallel to the garnet vector. Thus, the variation in Al content could be solely due to around 5% garnet entrainment, although a combination of plagioclase and orthopyroxene would be equally feasible. This diagram cannot be used to distinguish between these alternatives. However, in figure 11c, for Ca, it is clear that roughly 20% of peritectic plagioclase would have to have been involved, together with smaller amounts of garnet and/or orthopyroxene. The K plot in figure 11d confirms that the variation cannot have been due to any process involving biotite because there is only a slight and negative correlation between M and K in the data scatter. Modelling on this type of plot has shown that definite positive trends result from biotite fractionation or accumulation in S-type granitic magmas whereas, globally, the data for real granites define negative variations (fig. 5 c of Clemens and Stevens 2012). Furthermore, the trace-element modelling (fig. 10) showed that biotite could not have been a fractionating phase and, based on figure 11d, any kind of mixing between melt and biotite is also infeasible.

Using conventional least-squares mixing, with the phase compositions given in table 2, one of the lowest-SiO₂ megacrystic granites (DP16, with 71.31% SiO₂) could be produced through entrainment of 13.2% plagioclase, 4.0% orthopyroxene and 1.4% ilmenite as peritectic phases, assuming that one of the most silicic of the non-megacrystic granites (DP3,

with 76.46% SiO₂) represents pure melt. Note that garnet is not required. The sum of the squares of residuals is somewhat high, at 1.64%, and mostly due to a discrepancy in K₂O. In figure 3d it is apparent that the trend in K₂O for the megacrystic rocks has a different slope to that for the non-megacrystic rocks. This suggests that the two petrographic groups actually constitute separate magma series, probably derived from somewhat different rock types in a heterogeneous source terrane. If we instead assume that the megacrystic rock with 74.66% SiO₂ (sample DP8) represents the melt, the mixing calculations show that one the most mafic of the megacrystic rocks (DP16 with 71.31 % SiO₂) could be formed by entrainment of 8.1% plagioclase, 3.1% orthopyroxene and 1.2% ilmenite, again with no garnet required. In this case, the sum of the squares of residuals is 1.10.

These results show that most of the chemical variations among the Dartmoor magmas were plausibly caused by PAE, but cannot have been due to crystal fractionation of plagioclase and biotite, as suggested previously (e.g., Ward et al. 1992; Chappell and Hine 2006). This is consistent with the lack of variation in the Mg#s of the megacrystic granites. In fairness to previous authors, prior to the PAE model having been proposed by Villaros et al. (2007), it would have been difficult to consider that variations in granitic magmas might be caused by differences in the proportions of minerals that may no longer be present in the rocks.

However, it is unavoidable that the entrained peritectic assemblage will be lost or recrystallized to new minerals/compositions. It should be remembered that these peritectic crystals will be very much smaller than the magmatic crystals that are found in typical, coarse- or even medium-grained granitic rocks. The rates of dissolution and crystallisation of feldspars in silicate magmas are relatively rapid in comparison with the timescales of magmatic events. For example, in magmas of intermediate composition, feldspar crystals grow at around 10⁻⁹ to 10⁻⁷ cm/s, depending on the *T* and *P* conditions (Agostini et al. 2013 and references therein). Thus, even if small peritectic plagioclase crystals were present in a

magma for its entire evolution, from source to solidus, their compositions and textures would be likely to be reset by dissolution-precipitation processes. Similarly, ferromagnesian silicates such as orthopyroxene and garnet, entrained as part of the peritectic assemblage, are likely to be consumed by reactions that produce biotite as the magma cools. Zarrebini (2016) conducted experiments to investigate the fate of entrained peritectic plagioclase, pyroxene and garnet in an I-type granite magma undergoing cooling from 760 °C at approximately 200 MPa (2 kbar). Plagioclase crystals in excess of 200 µm in length completely recrystallised to new plagioclase within the 10-day durations of the experiments. Where pyroxene and garnet were not part of the equilibrium assemblage, these crystals were largely replaced by biotite. In the PAE model, the variations previously attributed to fractionation of accessory apatite, monazite, zircon, and xenotime (Ward et al. 1992) are viewed mainly as the products of variable entrainment of tiny crystals of these restitic accessory minerals.

The present results also suggest that the partial melting reactions that generated the Dartmoor magmas did not produce significant quantities of peritectic garnet (and see the phase-equilibrium modelling below). In agreement with the isotope-based inferences of Darbyshire and Shepherd (1994) and the chemically based arguments of Chappell and Hine (2006), this suggests that the main source rocks, though most probably metasedimentary, were not very strongly peraluminous. Rb and Ba tend to be positively correlated in clastic sedimentary rock series dominated by sandstones and shales (e.g., Baiyegunhi et al. 2017) but the main Dartmoor granitic rocks show negative correlations ($r^2 \approx 0.7$) between these elements. We suggest that this is due to the source sediments being chemically immature, feldspathic, volcanoclastic, metagreywackes, metadacites and possibly granodiorites that consist of domains that contain both biotite (Ba- and Rb-rich) and K-feldspar (Ba-rich and in low to minor proportions), as well as domains that contain only biotite. The melts of such materials would thus inherit negative correlations between Rb and Ba from this inherent

variation in source mineralogy. There may also have been some interbedded but subordinate metapelites. There is also an indication that the pressure of silicic magma formation is likely to have been < 500 MPa, since garnet would have been a stable peritectic product of greywacke partial melting at higher P (Vielzeuf and Montel 1994; Patiño Douce and Beard 1994). If we accept the PAE model, the inferred involvement of peritectic plagioclase suggests that the negative Eu anomalies in the REE spectra of figure 5 should be interpreted as inherited from source-level processes, rather than being products of post-emplacement differentiation or mixing, and that much of the map-scale heterogeneity is a primary, source-inherited feature. The more silicic and less calcic rocks have deeper Eu anomalies (figure 6a, b) because their magmas contained lower proportions of entrained peritectic plagioclase. The somewhat deeper negative Eu anomalies in the non-megacrystic to sparsely megacrystic rocks suggest that the sources of these magmas contained more restitic plagioclase and/or that the magmas entrained less of the peritectic mineral assemblages than the magmas that produced the megacrystic rocks.

We have charts of the pluton-scale heterogeneities, as in the map of Dangerfield and Hawkes (1981), modified here in figure 1, but there remains the question of the scales of more cryptic variations, such as those that might exist in isotopic systems. Some previous work has revealed map-scale isotopic zoning in granitic plutons of composite batholiths (e.g., Halliday et al. 1980; Nordgulen and Sundvoll 1992), but such zoning commonly corresponds with visible lithological variations. The scales of more cryptic whole-rock variations are largely unknown, although Bagdonas et al. (2016) documented very considerable Nd isotopic variation (-7.75 to $+1.21$ ϵ Nd) across nearly 200 km of the Neoproterozoic Wyoming batholith, a biotite granite that, apart from its leucogranitic margin, is cited as containing little or no lithological and chemical variation. The SiO_2 contents of the rocks in the Wyoming batholith

vary by about 5 wt% but there is no correlation between this parameter and ϵNd . The following paragraphs examine the isotope variations in the Dartmoor pluton.

Isotope chemistry.

Previous work. For Dartmoor, Darbyshire and Shepherd (1985) used eight poorly megacrystic and six megacrystic samples to define a whole-rock isochron age of 280 ± 1 Ma with an initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7094 ± 0.0003 . Darbyshire and Shepherd (1994) analysed two samples from Dartmoor, with indistinguishable ϵNd values of -4.6 and -4.7. Additional Sr and Nd isotope analyses of the large, garnet-bearing granite xenolith from Sweltor Quarry were furnished by F. Darbyshire, formerly of the Isotope Geoscience Facilities of the British Geological Survey. The compiled database for Dartmoor is presented in table 3, and includes our additional Rb-Sr, Sm-Nd and O data. The $\delta^{18}\text{O}$ values, for quartz separates, were kindly supplied by Prof. A. R. Fallick of the Scottish Universities Environmental Research Centre. Analytical details are given in appendix 2.

For the Cornubian batholith as a whole, the published Nd and Sr isotope data indicate substantial crustal involvement, and the N isotope study of Boyd et al. (1993) confirmed that the N had a biological, sedimentary origin. As shown in figure 12b, the $\epsilon\text{Nd}_{280\text{Ma}}$ values of the country rocks range from about -13 to -5, with most between -11 and -8 (Darbyshire and Shepherd 1994). Thus, the local country rocks are unlikely to have been the magma sources or to have been an influence on the magma compositions. The Dartmoor pluton intrudes the Devonian Kate Brook Slate and the Carboniferous Crackington formation. As shown in figure 12b, the Sr and Nd isotope characteristics of the Kate Brook Slate plot particularly remotely from those of the Dartmoor granitic rocks.

In considering the effects of alteration on the isotope systems in the Dartmoor rocks, Darbyshire and Shepherd (1985) noted the agreement between previous ^{40}Ar - ^{39}Ar studies of

biotite crystals (yielding plateau ages of 282 Ma) and the K-Ar ages of 281 ± 7 Ma, along with their own whole-rock dates for the late-stage Meldon 'aplite' (279 ± 2 Ma) and their Rb-Sr isochron age of 280 ± 1 Ma for the Dartmoor granite as a whole. As noted above, they thus considered that post-magmatic processes had not disturbed the Rb-Sr systems of fresh rock samples. This means that, given careful sampling of only fresh materials, remote from areas of hydrothermal alteration, any variations in initial $^{87}\text{Sr}/^{86}\text{Sr}$ must reflect isotopic heterogeneity inherited from magma source rocks. Such source heterogeneity might be expected in crustally-derived magmas, and particularly so if the granitic magmas under consideration were formed, at least partly, from metasedimentary source rocks that had protoliths with a range of clay contents.

Magma groups in the Dartmoor pluton. In the section on oxide and element chemistry, we demonstrated that the textural and mineralogical groups of Dartmoor granitic rocks are likely to have formed from distinct magma series, with the megacrystic and non-megacrystic granites, and the 'aplitic' rocks exhibiting different behaviours on many oxide and element variation diagrams (fig. 3a, b, c, e and fig. 4). Previous REE studies (Ward et al. 1992), and the modelling that we presented in figures 10 and 11, led to the general conclusion that the main chemical variations among the voluminous Dartmoor rock types cannot have been caused by mantle-crust magma mixing, wall-rock assimilation or fractional crystallisation from a common parent magma. Instead, the variations seem to be most compatible with a model in which the various groups of Dartmoor magmas were derived through approximately coeval partial melting events that affected different rock types, or mixtures of rock types, in a heterogeneous, mid-crustal source terrane of Proterozoic age. Figure 12a shows the Dartmoor rocks plotted on a Sr-isotope mixing diagram. Instead of defining a continuous mixing array, most texturally-identified groups form distinct clusters, although the 'aplitic' rocks from

Meldon quarry could be interpreted as forming a rudimentary mixing trend, as could the overall array for the other groups. This sort of clustered and crude positive correlation is interpreted as reflecting crustal source maturity. The rocks with lower Sr contents have higher initial Sr isotope ratios because their sources are metasediments with more of their Sr removed through weathering of feldspars in their protoliths, and higher Rb contents, hosted by more abundant clay minerals. Figure 13a shows an isotope correlation plot, essentially a magnified view of figure 12b with the fields for the Palaeozoic country rocks omitted. The important feature is that, despite a significant spread in $^{87}\text{Sr}/^{86}\text{Sr}_{280\text{Ma}}$, there is only a small variation in $\epsilon\text{Nd}_{280\text{Ma}}$ (just 1.3 epsilon units). This decoupling of the two radiogenic isotope systems is interpreted as meaning that the mainly metasedimentary source rocks for the granitic magmas varied in their original clay contents (i.e., their chemical maturity), despite having been derived through weathering and erosion of protoliths that were extracted from the mantle at approximately the same time. The few O isotope measurements for quartz in the Dartmoor rocks are plotted against $^{87}\text{Sr}/^{86}\text{Sr}_{280\text{Ma}}$ in figure 13b. This figure shows the expected positive correlation between these systems, consistent with the fact that, other things being equal, both these parameters measure the degree of chemical maturity of the sedimentary protoliths to the magma sources.

Scales of cryptic heterogeneities. Compaction, compositional convection and partial flow segregation of minerals and magmatic liquids can lead to randomly distributed, chemical, mineralogical and textural variability, on scales of centimetres to metres (e.g., Sultan et al. 1986; Speer et al. 1989). However, due to the relatively high viscosities of granitic liquids, and the commonly sheet-like morphologies of magma additions, convection and compaction are usually quite inefficient in granitic magmas. The scarcity of monomineralic cumulate layering and intracrystalline flattening deformation in untectonised granitic rocks attest to the

generally slight effects of these processes (e.g., Petford 1993; Holness 2018). Such segregation is more likely to be due to flow and shear deformation (e.g., Petford 1993; Petford et al. 2020). All these processes can produce chemical and modal heterogeneity but should not normally result in heterogeneity in the initial ratios of isotope systems such as Sm-Nd, Lu-Hf and Rb-Sr.

As has already been established, the Dartmoor pluton is heterogeneous, both chemically and isotopically. This is quite typical of S-type granitic rocks, worldwide, but far less common in I-type suites; see the discussion in Clemens et al. (2017c). As demonstrated by Farina and Stevens (2011) and Farina et al. (2014), it is possible for the sources of S-type granitic magmas to develop extreme contrasts in their $^{87}\text{Sr}/^{86}\text{Sr}$, as rocks with different concentrations of Rb evolve isotopically over millions or tens of millions of years prior to partial melting and S-type magma formation. This behaviour stems from original variations in clay contents between different parts of heterogeneous, sedimentary rock packages. However, an important question concerns the scales on which such isotopic heterogeneities might be preserved in granitic rocks that crystallised from such magmas. This is significant because it bears on the question of the extent to which such magmas are homogenised by mixing processes. Relatively high-viscosity granitic magmas (e.g., Scaillet et al. 1998; Clemens and Petford 1999) are perhaps the least likely of magmas to undergo homogenisation by mechanical mixing and chemical diffusion. However, in the absence of other physical or chemical evidence for contrasts in rock type or mineralogy, the degree of homogeneity is very difficult to gauge, and the only sampling approach that might be followed would be a random one, since there may be few or no mineralogical guides to follow. To make a rudimentary test of this, we obtained high-precision Rb-Sr and Sm-Nd isotope analyses on large (> 3 kg), fresh, whole-rock samples of a Dartmoor granite, together with O isotope analyses of quartz

grains hand picked from coarsely crushed samples. Details of the preparation and analytical techniques can be found in appendix 2.

This part of the study involved two sets of sample pairs, each from opposite ends of two blocks of coarse-grained megacrystic granite in the Blackenstone quarry. The sample numbers are DP34A and B, and DP35A and B. Physical spacings were ~ 0.5 m for DP34A and B, and ~ 1.0 m for DP35A and B. The isotopic analyses of these paired samples are given in table 3, and petrographic descriptions, photographs of polished blocks and photomicrographs can be found in appendix 4. In figures 12a and 13, the analytical points are shown with dotted tie-lines connecting each sample to its pair. We believe that this high degree of spatial resolution is unprecedented in whole-rock studies of granitic intrusions, partly because whole-rock isotopic heterogeneity, on such a scale, would not be anticipated in what appear to be petrographically homogeneous rock units, and partly because the analytical costs would be prohibitive for any extensive programme of such sampling.

Based on the petrographic observations, the chosen samples show no mesoscopic textural or mineralogical signs of heterogeneity, and no significant contrasts in the degree of slight alteration between members of either pair. Given that the quoted analytical uncertainty of 0.2 epsilon units, the $\epsilon\text{Nd}_{280\text{Ma}}$ values of the paired samples vary only slightly (-3.8 to -3.5 for DP34A and B, respectively and -3.9 for both DP35A and B). However, there are significant differences in both $^{87}\text{Sr}/^{86}\text{Sr}_{280\text{Ma}}$ (of the order of 0.003) and for quartz $\delta^{18}\text{O}$ (up to 0.9‰) between paired samples. These differences are far larger than the propagated analytical uncertainties in both these isotope ratios (table 3).

Phase Relations, and Some Mineralogical Puzzles

In the petrography section, we noted some curious textural features of the Dartmoor granitic rocks, including the common presence of fine-grained biotite-quartz pseudomorphs

after orthopyroxene and the late crystallisation, small grain size and anhedral habit of much of the biotite. We have already flagged the latter texture as unusual for S-type granitic rocks, in which most of the biotite is present as large, early-formed euhedra or subhedra. As also noted, chemical features of the Dartmoor rocks suggest the possibility that relatively low- T partial melting reactions were responsible for magma formation. A very shallow final emplacement depth has also been inferred, consistent with the presence of a hornfels, contact aureole surrounding the pluton, as well as the Chappell and Hine (2006) conclusion that the pressure of final magma evolution was around 50 MPa. Typically, fluid-absent partial melting reactions result in granitic melts that have H₂O contents of 3 to 5 wt% (e.g., Stevens and Clemens 1993, and references therein). Thus, even if the magmas were H₂O-undersaturated at their depth of generation, by the time they had ascended to pressures \leq 100 MPa, they are likely to have become H₂O saturated, since H₂O solubility at these pressures will be $<$ 4.2 wt% (Tamic et al. 2001). The late crystallisation of K-feldspar, and its presence as flow-orientated megacrysts, is also a feature of the Dartmoor pluton. The issue of whether these are phenocrysts and whether magmas at this late stage of crystallisation are able to undergo magmatic flow has been hotly debated (see e.g., Winkler and Schultes 1982; Vernon 1986). The question then is whether near-emplacement crystallisation of H₂O-saturated granitic magmas could produce the observed textural and mineralogical peculiarities.

To throw some light on these aspects, we used Rcrust (Mayne et al. 2016, 2020) to calculate a 100 MPa, T -H₂O pseudosection (fig. 14) for the composition of sample DP12E, a relatively mafic (72.6 wt% SiO₂) granite from Haytor. Note that the composition axis here is the H₂O content of the bulk system, not the H₂O dissolved in the melt. Appendix 3 lists the composition of DP12E, and appendix 5 gives the modelling parameters. To gauge the effects of P , we also calculated sections at 50 and 150 MPa, and we comment briefly on what these reveal. From the phase relations in figure 14, it seems clear that, even under fluid-saturated

conditions, orthopyroxene is stable in a broad field above the solidus. The calculations show that, above the solidus, < 2% of this mineral is present, that its modal proportion decreases sharply within a few degrees C of the solidus, and that it is unstable at subsolidus temperatures, except at very low H₂O contents. This neatly explains its disappearance through reaction to form near-solidus biotite. The lack of mafic mineral accumulation in the Dartmoor rocks means that we see no great concentrations of orthopyroxene. Even in granites with cumulate schlieren that once contained orthopyroxene, this phase has been efficiently eliminated by reaction with low-*T* magmatic liquids, leaving only biotite-quartz pseudomorphs that can be easily overlooked (e.g., Clemens et al. 2020b).

The presence of mainly anhedral, late biotite is also consistent with the calculated pseudosections, which show that this mineral has decreasing stability above the solidus as *P* falls between 150 and 50 MPa. At 150 MPa, biotite is stable up to about 760 °C, to only about 740 °C at 100 MPa, and not stable at all above the solidus at 50 MPa. This means that any biotite crystallised at higher pressures would have progressively reacted with the melt, probably forming some cordierite, which is present in the rocks and is omnipresent, in small quantities, throughout the calculated pseudosections. As described above, biotite makes a reappearance in the phase assemblage near the solidus, emerging as orthopyroxene departs. This late, near-solidus occurrence of biotite explains the presence of the small anhedral crystals of this mineral in the groundmass of the megacrystic granites.

Garnet is only rarely encountered in Dartmoor rocks. The calculations here predict that very small amounts (< 0.1 vol.%) of this mineral would be present at conditions both just above and below the solidus. Thus, if garnet formed at all, it would have been near the solidus and as very small crystals that could easily have reacted out to biotite and cordierite with further cooling. One last puzzling textural and mineralogical feature of the Dartmoor rocks can also be addressed using figure 14. This is the much-debated meaning of the euhedral K-

feldspar megacrysts. These were present when magmatic flow occurred, since they form foliations whose local orientations vary on a scale of metres to tens of metres. The main observation that makes this difficult to understand is that K-feldspar appears late in the crystallisation history of most granitic magmas, so it is assumed that there would be insufficient melt present to permit efficient (any?) magma flow. There is also the question of why this feldspar so commonly attains such large grain sizes (up to 7 cm in the Dartmoor rocks). Holness et al. (2018) addressed both these problems, suggesting that there would commonly have been at least 40% melt present when most granitic magmas became saturated with K-feldspar and that low nucleation density and rapid, near-solidus crystal growth could explain the large grain size. In figure 14, under fluid-saturated conditions, K-feldspar does indeed only begin to crystallise very close to the solidus, but the calculated melt-proportion contours suggest that this Dartmoor magma would still have contained over 70% melt at this stage. Thus, the observed flow orientation of these feldspar crystals is fully consistent with the calculated phase relations.

Comparisons with Other S-type Granitic Rocks, and the Nature of the Source Rocks and their Melting Conditions

The chemical features of the Dartmoor granitic rocks may be compared with those of some other well-studied S-type granitic batholiths in the type locality of southeastern Australia. We chose the late Silurian Kosciusko batholith, the Early Devonian Wilsons Promontory batholith and the Late Devonian Strathbogie batholith. These three bodies present quite different structural, textural and mineralogical characteristics, covering the common range shown by S-type plutonic rocks. The very large Kosciusko batholith (Hine et al. 1978) consists of foliated rocks in separate elongate plutons that are orientated parallel to the regional fabrics in the intruded low-grade metasediments. These are what White et al. (1974) referred to as regional-

aureole granites. In the Kosciusko S-type rocks, the mafic minerals are mainly biotite, with minor cordierite in some plutons, and emplacement is inferred to have been in the upper middle crust. The Wilsons Promontory batholith (Wallis and Clemens 2018) is a twin-lobed body comprised of several contiguous, sheet-like plutons. The contacts with the low-metamorphic-grade country rocks are poorly exposed, but the batholith appears to be grossly discordant with wall-rock structures, and there is evidence for a hornfels aureole, so the emplacement level is inferred to be in the upper crust. The rocks of the Wilsons Promontory batholith are biotite-bearing, but almandine-rich garnet is relatively abundant, while cordierite is less so. The Strathbogie batholith (Clemens and Phillips 2014; Phillips 2017) is a large (100 km-long), composite body of mainly contiguous plutons that appear to be relatively thin and sheet-like. The batholith is grossly discordant with the trends of the fold axes and faults in the sedimentary country rocks; it also intrudes a silicic volcanic complex of similar age, and has an aureole of high-grade hornfels. The Strathbogie granites are relatively cordierite-rich, so much so that biotite is subordinate to cordierite in some of the rocks, and garnet is much scarcer than at Wilsons Promontory.

Figures 15 and 16 show bivariate major- and trace-element chemical plots with the analytical points for the Dartmoor rocks (in red), compared with those for the Kosciusko, Wilsons Promontory and Strathbogie batholiths (in orange, green and blue, respectively). If we take the Australian occurrences as representative, the Dartmoor granitic rocks appear to be rather atypical in several respects. Relative to most S-type granitic rocks, they have low FM, CaO, Mg# and Ba, and high K₂O, Rb, U and Zr. If the source rocks for a granitic magma are partially melted at relatively low temperatures, Ca, Fe and Mg are likely to remain behind in the melt-depleted source rocks, residing in restitic and/or peritectic minerals such as garnet, cordierite and possibly biotite, while elements such as K, Rb and U will be transported upward, dissolved in the magmatic liquids (e.g., Fyfe 1973; Stevens et al. 2007; Clemens and

Stevens 2012). The Ba contents of melts are likely to be joint functions of source-rock Ba content (or mineralogy, if the source contains a high K-feldspar content) and melting temperature (e.g., Clemens 2014). In contrast to these elements, Zr concentrations are likely to be influenced by the proportions of restitic zircon crystals entrained into the magmas (Clemens and Stevens 2012; Clemens 2014), and the Dartmoor zircons do contain inherited cores (Neace et al. 2016). This influence would be especially the case if the partial melting were at relatively low T , resulting in generally low Zr solubility in the melts (e.g., Watson and Harrison 1983). Thus, the unusual chemical characteristics of the Dartmoor S-type granitic rocks (relative to most S-type rocks that we have examined and researched over some decades) suggest that the source rocks were partially melted at relatively low temperatures. To emphasise this point, in figure 17 we illustrate some trace-element characteristics of the Dartmoor granites in comparison to other world S-type granitic plutons and batholiths. From this figure, it appears that, from a global perspective, the Dartmoor rocks are indeed compositionally atypical, which suggests either atypical source rocks or melting conditions, or both. Most S-type magmas are inferred to have formed through high- T , fluid-absent partial melting reactions (e.g., Clemens and Watkins 2001; Clemens et al. 2020a) but the Dartmoor magmas appear to have been exceptions. Many other examples of low- T , S-type magmas are diatexitic plutons, intimately associated with mid-crustal migmatites (e.g., Finger and Clemens 1995; Williams 2001), but there is no local migmatite association with Dartmoor.

There are two main scenarios for relatively low- T partial melting. The first is fluid-present melting, and the second is low- T fluid-absent melting involving muscovite breakdown (see e.g., Clemens et al. 2020a). As recognised by Simons et al. (2016), metagreywackes are typically muscovite-poor, leading to relatively low melt productivity through fluid-absent muscovite breakdown. At some places in the Cornubian middle crust, there could have been small amounts of aqueous fluid present, or some rock types in the source region could have

partially melted by reactions involving combined muscovite-biotite breakdown. Such reactions occur at relatively low temperatures and the melts produced have low Ca, Fe and Mg contents (e.g. Huang and Wyllie 1973, Patiño Douce and Harris 1998) and could have produced the ‘aplitic’ Dartmoor magmas that formed the dykes of porphyritic micromonzogranite and two-mica microgranite. However, for the voluminous and generally earlier megacrystic and non-megacrystic Dartmoor granites, the presence of biotite-quartz pseudomorphs after orthopyroxene means that the formation of these magmas must have involved higher- T biotite breakdown. In quartzofeldspathic crustal rocks, peritectic orthopyroxene is produced only in cases in which biotite is the main hydrous mineral breaking down in the melting reactions (e.g. Hewitt and Wones 1984), and it will only crystallise from granitic magmas formed by biotite breakdown reactions (e.g. Clemens and Wall 1981, 1988).

As summarised in Clemens et al. (2020a), relatively low- T partial melting in granite source terranes can occur through either fluid-present or fluid-absent reactions. In the present case, with the chemistry-based inference of relatively low- T magma formation, this suggests fluid-deficient partial melting, a process in which there is a limited budget of aqueous fluid. This fluid can cause melting and partial biotite breakdown in the source rocks, but the system would then evolve toward a fluid-absent condition, since all the initially free H₂O would be dissolved in the melt phase. In the case of Dartmoor, the melting T must have considerably exceeded that of the ‘wet’ solidus but not have been high enough to cause significant fluid-absent biotite breakdown. Note here that the occurrence of orthopyroxene in a granitic magma cannot be taken as firm evidence of either high temperature or formation by fluid-absent partial melting. Fluid-present partial melting involving biotite breakdown also produces peritectic orthopyroxenes (e.g., Clemens 1990, 1995). As shown by Clemens and Droop (1998), the topologies of the reactions mean that fluid-saturated granitic magmas can ascend

in the crust, provided that they are formed at temperatures above the ‘wet’ solidus and that the melts can efficiently segregate from the residual solids in the source rocks. In this connexion, Glazner (2019) pointed out that crystallisation during adiabatic ascent, hydrous magmas will release heat and enhance their ascent own capacity. Whether formed by fluid-present or fluid-absent reactions, the ascending magmas that made up the main biotite-bearing Dartmoor granites are likely to have entrained variable proportions of the peritectic assemblage that formed in the melting reactions. These minerals include garnet, orthopyroxene and ilmenite, formed through breakdown of Ti-bearing biotite, and plagioclase formed through preferential dissolution of the sodic component into the melt (Clemens and Stevens 2012). Biotite-breakdown reactions similar to $Bt + Pl_1 + Qz \pm \text{fluid} = Opx \pm Grt + Ilm + Pl_2 + \text{melt}$ would be responsible. Grains of accessory minerals, such as zircon, that were hosted by biotite crystals would also be released and potentially entrained in the ascending magmas (Clemens and Stevens 2012). This explains the observed negative correlations between SiO_2 (mainly in the melt phase) and FM, TiO_2 , CaO and Zr (mainly in the entrained crystals) See figures 3a and c, and 4b. Later, we model these variations, expressing the variables in molal quantities, with the mafic component plotted as the abscissae.

S-type granitic magmas commonly contain inherited zircon (e.g. Williams 1995, 2001; Appleby et al. 2009; Gao et al. 2016), and the anomalously old U-Pb bulk zircon age (~310 Ma) published by Neace et al. (2016) demonstrate that the Dartmoor granitic magmas did contain such inherited zircon. Many of the discordant zircon crystals studied by Neace et al. (2016) have Grenvillian (Mesoproterozoic) ages, though, as these authors point out, their discordancy precludes them from being used to definitively identify the source rocks for the magmas. In any case, for Dartmoor, the likely overall picture is one of mid-crustal fluid-deficient partial melting involving muscovite and some biotite breakdown, leading to the

formation of a variety of granitic magmas, the more mafic of which contained significant but varying proportions of entrained peritectic minerals, as well as some inherited zircon.

The SiO₂ content above which the Dartmoor magmas may have contained very little or no entrained peritectic crystals is probably around 73 wt%, as this appears to be the minimum SiO₂ content for pure melts formed in experiments on crustal melting at temperatures up to 800 °C (e.g., Patiño Douce and Harris 1998). As we have seen, the Dartmoor ‘aplitic’ magmas cannot have been formed through differentiation of the more mafic main granitic magmas. We assume that crustal temperatures increased with depth and that the light-element-enriched ‘aplitic’ magmas are likely to have formed at lower temperatures than the main Dartmoor magmas. Thus, the late generation of the ‘aplitic’ magmas may be connected with fluid-present partial melting of source rocks at higher levels in the crust than for the formation of the main magmas, the sources of which are likely to have attained higher temperatures. This would be consistent with upward migration (and weakening) of the responsible thermal anomaly and or with the suggestion that the crust in the region was undergoing extension throughout the period of silicic magma generation (e.g., Simons et al. 2016). Boron and fluorine lower the solidi of crustal rocks, especially under fluid-present regimes (e.g., Manning 1981; Pichavant 1981). Thus, given the evident enrichment of the source rocks in B and F, it seems possible that the melting epoch here began or continued to lower temperatures than is usually the case at these crustal levels.

Discussion and Conclusions

The results of chemical and isotopic analyses have shown that the Dartmoor magmas can be grouped into four separate groups (coarse-grained megacrystic granites, coarse-grained poorly to non-megacrystic granites, two-mica microgranites and ‘aplites’). These data also demonstrate that the compositions of most of the magmas were not controlled mainly by

processes such as mafic-silicic magma mixing, upper crustal assimilation or crystal fractionation. The main variations in the most voluminous Dartmoor magmas appear to have been inherited from heterogeneous, crustal magma sources at mid-crustal depths, at pressures < 500 MPa. Mid-crustal source depths were also advocated by Simons et al. (2016), who pointed out that since pre-Devonian basement is not exposed, we cannot identify a specific rock package as the source of the Dartmoor magmas. However, this unexposed source terrane must contain mildly aluminous metasedimentary rocks, with minor metapelitic and metavolcanic components, all of which are likely to be of Proterozoic age, and probably formed in a collisional tectonic setting.

Modelling of the chemical variations shows that the various Dartmoor magmas attained their parental compositions through the mechanisms of H₂O-undersaturated crustal melting and peritectic assemblage entrainment, and that the entrained minerals would have been plagioclase, orthopyroxene and ilmenite, along with accessory crystals such as zircon and monazite. Local fractional crystallisation modified the magma compositions, but the late, 'aplitic' magmas were not differentiates of either of the two most voluminous magma groups (represented by the megacrystic and non-megacrystic granites). Given that the 'aplitic' magmas were enriched in light elements, this feature is probably a source-derived characteristic that was intensified by internal fractionation. Production of these 'aplitic' must have involved somewhat different processes and source materials from the rest of the Dartmoor magmas. Since some of the 'aplitic' rocks have SiO₂ contents similar to the non-megacrystic granites, perhaps these late silicic magmas were products of lower-*T* processes at higher levels in the crust. The present data do not allow us to be more explicit, but it does appear that these highly silicic magmas, which might be presumed to have originated through crystal fractionation, could not have attained their compositions through fractionation from other magmas that formed the exposed parts of the pluton.

Globally, some silicic volcanic and plutonic rocks contain plagioclase crystals with complex zoning patterns, as well as evidence for partial resorption and overgrowth, in some cases with multiple events. Cashman and Blundy (2013) note that mineral textures in some igneous rocks imply considerable complexity in their genetic processes, and that this is inconsistent with simple chemical models such as fractional crystallisation. They point out that, for large plagioclase crystals, sluggish diffusion Si and Al in plagioclase growth can result in preservation of textural evidence for complex magmatic evolution. However, consideration of plagioclase dissolution and crystallization rates, and the experimentally documented rapid recrystallization of out-of-equilibrium plagioclase in melt-dominated magmas, as discussed above, indicates that the sorts of features documented by Cashman and Blundy (2013) are most likely to be recorded in magmas that form by mingling between magmas containing calcic plagioclase phenocrysts and cooler, crystal-rich magmas or mushes with substantially more sodic plagioclase. It is thus noteworthy that the Dartmoor granitic rocks contain plagioclase crystals that are only faintly zoned and only rarely show signs of chemical or physical shifts that could have resulted from corrosion and overgrowth. We take this as evidence for the general simplicity of the mechanisms responsible for construction of the pluton, with no evidence for mafic-silicic or intermediate-silicic magma mingling or, for most of the SiO₂ range, fractional crystallisation, and cumulate mush formation and disruption. There was evidently minimal chemical interaction between compositionally similar, individual magma additions, and this is consistent with the preservation of the small-scale isotopic heterogeneities that we have documented, as well as our inference that the magmas with SiO₂ < 73 wt% were formed by PAE and subsequently underwent little or no chemical evolution following emplacement.

In closely-spaced sample pairs, measured differences in Sr and O isotope ratios are substantial. The whole-rock samples used for Sr isotope analysis were relatively unaltered

(see appendix 4), and Darbyshire and Shepherd (1985) argued that the Dartmoor Rb-Sr system is generally undisturbed. Furthermore, since it is the quartz crystals (rather than whole-rock data) that show considerable variation in $\delta^{18}\text{O}$, these isotopic contrasts are unlikely to have been caused by alteration, which affects mainly feldspars and micas. Other features, as described above, rule out essentially all sources of isotope disequilibrium except those operating at the level of the source rocks, during magma genesis. At Dartmoor, this significant whole-rock isotope heterogeneity appears to have been preserved on quite small scales. The similarity of the mineralogy and major-element chemistry of the rocks, despite the Sr and O isotopic differences, is most probably due to the fixed stoichiometries of the melting reactions that produced the magmas. However, many of the trace-element concentrations are likely to have been controlled by disequilibrium between melt and solid residue (restite). Specifically, the lack of variation in $\epsilon\text{Nd}_{280\text{Ma}}$ suggests that restitic apatite and monazite were entrained into the magmas, which thereby inherited the Nd isotope signatures of those source rocks.

It is worth considering possible relationships between the sizes of heterogeneities in magma sources and the sizes of heterogeneities in granitic plutons. For a granitic magma system, Clemens and Mawer (1992) calculated that a vertical, magma-filled feeder fracture of up to 6 km height could potentially remain open to the passage of magma at any instant in time. Assuming this fracture height, a width of 3 m, the minimum needed to prevent magma freezing (Clemens and Mawer 1992; Petford et al. 1994), and an arbitrary plan length of 1 km, this corresponds to a magma pulse (solitary wave) with a volume of $1.8 \times 10^7 \text{ m}^3$. Based on experiments on partial melting of metasedimentary rocks at crustal pressures (LeBreton and Thompson 1988; Vielzeuf and Holloway 1988; Vielzeuf and Montel 1994; Stevens et al. 1997; Pickering and Johnston 1998) it seems reasonable to model an anatectic system that would generate around 30 vol.% melt. Note here that discussions commonly revolve around

the attainment of rheologically critical melt fractions, which are assumed to be somewhere near 50 vol.%. These figures are derived through misconstrual of the experiments of Arzi (1978). As shown by Rosenberg and Handy (2005), the important rheological step is actually attainment of the melt connectivity transition, at about 7 vol.% melt. Crustal melts segregate from their sources efficiently and at low melt fractions (e.g., Clemens and Stevens 2016). Using the above considerations, our calculated pulse volume could be extracted from a sphere of source rock with a radius of 22 m. If we assume that individual magma pulses might be derived from different mixtures of end members in a heterogeneous source, granitic rocks might contain source-inherited heterogeneities on a scale of a few metres. Intraplutonic, chaotic magma flow might reduce the sizes of these inherited heterogeneities (e.g., Flinders and Clemens 1996). Thus, it seems that rock packages undergoing partial melting could easily produce groups of magma pulses, with a variety of different isotopic characteristics, simply by the partial melting of rocks that contain heterogeneities on the order of tens of metres in size.

Using the same assumptions regarding pulse volume, a dyke-fed batholith of say 2000 km³ would represent the aggregation of around 10⁵ such magma pulses. Chaotic stretching and folding in the flowing magma could reduce the sizes of the heterogeneities and produce the kinds of apparently continuous variations in rock composition that we observe in many granitic bodies. A probable example of magma heterogeneity and more efficient mingling of batches is given by Schoene et al. (2012), who showed that the Adamello batholith (in Italy) exhibits ID-TIMS, U-Pb, zircon crystallisation dates that vary by as much as 200 ka, within single hand specimens. They attribute this to the mingling of numerous pulses of silicic magma introduced into the growing pluton. This may be a case of either very small magma batches or of a high degree of liquid magma stirring in this I-type body. In less well-stirred granitic magmas, such primary heterogeneity might also be the origin of the small (1 to 3

wt%) SiO₂ compositional gaps that, in our experience, characterise many granitic series. McCaffrey and Petford (1997) studied width to thickness relations in plutons, concluding that large granitic intrusions are generally tabular in shape. For an assumed volume of 2000 km³, we can therefore imagine a disk-like pluton 20 km across and 5 km thick. In such a growing intrusion, an individual magma pulse, if it were to extend across the whole width of a pluton, would have an approximate thickness of only a few centimetres. If contrasting packages of such thin magma units were stretched and dismembered by chaotic magma flow, this suggests that complex, small-scale heterogeneities should be the rule rather than the exception in granitic intrusions that were derived from heterogeneous protoliths such as metasedimentary packages.

The apparent preservation of such heterogeneities (on a variety of scales) suggests that, at emplacement level, the aggregated Dartmoor magma was never efficiently homogenised by mechanical mixing (due to forced flow or convection), although some flow and mixing did occur, that solidification occurred on a time scale shorter than that necessary for slow diffusional homogenisation (e.g., Johannes 1978; Villa 2006), and that the magma batches or pulses escaped their partially molten sources quickly, limiting homogenisation and preserving isotopic disequilibrium. In the granitic products, this confirms previous studies that have suggested rapid melting and magma withdrawal from source terranes, resulting in isotope and trace-element disequilibrium (Sawyer 1991; Bea 1996; Ayres and Harris 1997; Davies and Tommasini 2000; Villaros et al. 2009; Clemens and Stevens 2016; Etheridge et al. 2020). Although there are exceptions, such as the Harney Peak leucogranite and the Wyoming batholith in the USA (e.g., Nabelek et al. 1992a, b; Krogstad and Walker 1996; Bagdonas et al. 2016), in sampling a granitic pluton, it is not particularly common to take and analyse multiple samples from apparently homogeneous tracts of rock. This is especially so at the small spatial scales of the present work. Thus, significant degrees of isotopic heterogeneity

could be overlooked, and inferences on the character and homogeneity of a magma source could therefore be incomplete.

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Figure Captions

Figure 1. Geological sketch map of the Dartmoor pluton, based on original mapping by Brammell and Harwood (1923) and figure 1 of Dangerfield and Hawkes (1981), showing the mappable rock types recognised here, together with the intrusive stage (roman numerals) that each represents, as recognised by Brammell and Harwood (1923). The positions of sample localities (quarries) referred to in the text are indicated, and the inset shows the location of the pluton within the Cornubian batholith of SW Britain.

Figure 2. Outcrop on the western flank of Haytor rocks, on the southeastern margin of the Dartmoor pluton; see figure 2. The lower part of this outcrop shows a roughly flat-lying sheet of poorly megacrystic microgranite (with blocky weathering) that intruded into and chilled against a thicker sheet of much coarser-grained, and more mafic, megacrystic monzogranite. Within a few centimetres of the upper contact of the younger microgranite sheet, its grain size decreases, indicating that the coarser-grained megacrystic monzogranite was cooler than the microgranite magma when it intruded. See also appendix 1, figure 1.2.

Figure 3. Harker-style plots of some major-element parameters plotted against SiO₂ for the analysed rocks of the Dartmoor pluton – (a) TiO₂, (b) Al₂O₃, (c) FM (= FeO^T+MnO+MgO), (d) K₂O (with fields shown for different magma series), (e) Mg# (= 100Mg/[Mg+Fe] mol. ratio), (f) ASI (P-corrected alumina saturation index = Al₂O₃/[CaO-3.33P₂O₅+Na₂O+K₂O] mol. ratio). In (f) the solid blue horizontal line separates the fields for peraluminous (ASI < 1.0) and metaluminous (ASI > 1.0) rocks, and the dotted blue line separates weakly from strongly peraluminous (ASI > 1.1) rocks. The samples are colour-coded as shown in the legend.

Figure 4. Two trace-element Harker-type plots – (a) Ba and (b) Zr. Symbols and colour codings are given in the legend.

Figure 5. Chondrite-normalised rare-earth-element spectra for the Dartmoor granitic rocks – (a) for the megacrystic granites, (b) for the ‘aplitic’ rocks, with the Meldon ‘aplite’ distinguished as a thicker line, (c) for poorly to non-megacrystic granites, and (d) for two-mica microgranites. The normalising factors were from Anders and Grevesse (1989). Note that Ward et al. (1992) and Chappell and Hine (2006) did not analyse for the whole suite of REE, so the resulting gaps in the patterns have been bridged with interpolated line segments. This affects 25 out of the 61 patterns – 14 in (a), 9 in (c), and both in (d).

Figure 6. Eu anomaly ($\text{Eu}/\text{Eu}^* = \text{Eu}_N/\sqrt{\text{Sm}_N \cdot \text{Gd}_N}$) plotted against various other chemical parameters – (a) SiO_2 , (b) CaO and (c) K_2O contents. Symbols and colour codings are given in the legend.

Figure 7. Harker-type plots illustrating chemical relationships between megacrystic Dartmoor granites (red squares) and their igneous-textured microgranular enclaves (black dots, and one white dot), for (a) TiO_2 , (b) Al_2O_3 , (c) Sr and (d) V. The red lines and blue curves are, respectively, extrapolated linear and exponential regressions through the points for the megacrystic granites, with r^2 correlation coefficients given. Two of the analytical points (from Stimac et al. 1995) are for the same enclave – 222J1A being the whole-rock analysis, including large feldspar crystals, and 222J1B being the groundmass, plotted as the white dot. See the text for further information and interpretations.

Figure 8. Plot of P₂O₅ against ASI (P-corrected alumina saturation index = Al₂O₃/[CaO-3.33P₂O₅+Na₂O+K₂O] mol. ratio) for the rocks of the Dartmoor pluton. Symbols and colour codings are given in the legend.

Figure 9. Rb v. Y+Nb source tectonic discrimination diagram for granitic rocks, after Pearce et al. (1984), showing the various fields for tectonic settings: VA = volcanic arc, syn COL = syn-collisional, WP = within-plate and OR = ocean-ridge. The data points for the Dartmoor granitic rocks are plotted, with colours and symbols given in the legend.

Figure 10. Log-log plots of some trace-element variations in the Dartmoor granitic rocks – (a) Sr v. Ba and (b) Sr v. Rb. In each plot, vectors are shown to indicate the compositional effects of Rayleigh fractionation of 25 wt% each of plagioclase, K-feldspar and biotite. Symbols and colour codings are given in the legend, and mineral abbreviations are as recommended in Whitney and Evans (2010).

Figure 11. Element concentrations (moles/kg) in the Dartmoor granitic rocks plotted against the parameter M (maficity = mol. Fe + Mn + Mg). Vectors are shown for addition of various mineral species that are either present in the rocks (5 wt% each of plagioclase, K-feldspar and biotite, and 1 wt% of ilmenite) or that might have been present as peritectic minerals in the magma source, during partial melting to form the magmas (5 wt% each of orthopyroxene and garnet). Note that the use of maficity (M) avoids problems that arise due to variations in Mg#s of the rocks or their component minerals, and allows linear vectors to be constructed (Clemens et al. 2017b). Symbols and colour codings are given in the legend, and mineral abbreviations are as recommended by Whitney and Evans (2010).

Figure 12. Plots of radiogenic isotope variations in the Dartmoor granitic rocks – (a) Sr isotope mixing diagram and (b) isotope-correlation diagram, with initial $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵNd both calculated at a reference age of 280 Ma. Symbols and colour codings are given in the legend. In a, paired samples, taken at spacings of 0.5 and 1.0 m in Blackenstone Quarry are joined by dotted tie lines; see also figure 17. In (b), for comparison, coloured fields are shown for various Palaeozoic country-rock formations that were intruded by the magmas of the Cornubian batholith, and blue horizontal and vertical lines indicate the composition of bulk Earth at 280 Ma. Data for the country rocks were taken from Darbyshire and Shepherd (1994).

Figure 13. Plots of isotopic variations in the Dartmoor pluton – (a) is a magnified version of the isotope correlation diagram in figure 16b, focusing on variations within the pluton, rather than relationships with wall rocks, and (b) shows the relationships between whole-rock $^{87}\text{Sr}/^{86}\text{Sr}_{280\text{Ma}}$ and $\delta^{18}\text{O}$ in quartz for the granitic rocks. Dotted tie-lines connect the points for the two sets of paired samples (DP34A and B, and DP35A and B, taken at spacings of 0.5 and 1.0 m, respectively). Symbols and colour codings are given in the legend.

Figure 14. Calculated isobaric (100 MPa) T - H_2O content pseudosection for Dartmoor granite DP12E. Note that the composition axis here shows bulk H_2O content, not H_2O content of the liquid. The diagram was calculated using Rcrust (Mayne et al. 2016) in the system NCKFMASHTMn, with all Fe as Fe^{2+} and Ca corrected for Ca in apatite. Appendix 5 shows the complete details, including the phase-solution models used. The diagram shows phase-saturation boundaries. In general, a given phase is stable on the low- T side of its saturation curve. There are two exceptions; ilmenite is stable everywhere, except within the small area to the left of its phase boundary, and orthopyroxene is stable between its upper- and lower- T

saturation boundaries, though always in proportions of < 2 vol.%. There is a small field for subsolidus orthopyroxene at very low bulk H₂O contents, at H₂O-undersaturated conditions. Here, the pyroxene is stable above this curve; this is a subsolidus extension of Opx-out on cooling. Note that cordierite is stable throughout the diagram, and is therefore a near-liquidus phase. In some cases, phase boundaries almost exactly coincide and are shown dashed, with the colours for the two minerals alternating. From lower to higher bulk H₂O contents this occurs for biotite and the solidus, for orthopyroxene and biotite, and for orthopyroxene and K-feldspar. Mineral abbreviations are as recommended in Whitney and Evans (2010). The red contours are for wt% melt present. See text for discussion.

Figure 15. A selection of major-element parameters plotted for the Dartmoor rocks, together with points for analyses from the Kosciusko, Strathbogie and Wilsons Promontory S-type batholiths in southeastern Australia; (a) FM (= FeO^T+MnO+MgO), (b) CaO and (c) Mg# (= 100Mg/[Mg+Fe] mol. ratio) vs SiO₂, and d K₂O vs Na₂O. Kosciusko data are from Hine et al. (1978), Strathbogie from Clemens and Phillips (2014) and Wilsons Promontory are from an unpublished dataset compiled by J. Clemens and G. Wallis. See the text for further details.

Figure 16. Selected trace-element parameters plotted for the Dartmoor rocks, together with points for analyses from the S-type Kosciusko, Strathbogie and Wilsons Promontory batholiths in southeastern Australia; (a) Zr vs Ba and (b) Sr vs Rb. In (a), linear fits are shown to each of the data arrays. Data sources are given in the caption to figure 9, and further details can be found in the main text.

Figure 17. Selected trace-element parameters plotted for the Dartmoor rocks, together with points for analyses of S-type granitic rocks (from Brittany, the French Massif Central, the

Harney Peak Granite in South Dakota, the Cape Granite Suite of South Africa, and the Himalayan granites); (a) Zr vs Ba and (b) Sr vs Rb. Data for Brittany are from Georget (1986) and Euzen (1993), those for the Massif Central are from Downes and Duthou (1988), Downes et al. (1997), Williamson et al. (1997) and Solgadi (2007), those from Harney Peak are from Nabelek et al. (1992a), those for the Cape Granite Suite are from Scheepers and Armstrong (2002), and those for the Himalayan granites are from Ayres and Harris (1997) and Scaillet et al. (1990).