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Time scales of magmatic processes

Chris Hawkesworth^{a,*}, Rhiannon George^{a,1}, Simon Turner^{a,1},
Georg Zellmer^b

^a Department of Earth Sciences, Wills Memorial Building, Queens Road, University of Bristol, Bristol BS8 1RJ, UK

^b Lamont Doherty Earth Observatory, Geochemistry Building, 61 Route 9W, Palisades, NY 10964, USA

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Abstract

The development of improved analytical techniques has facilitated the application of short-lived isotopes to the study of magmatic processes, and resulted in a renewed interest in a number of other chronometers. Two approaches have been used to determine the time scales of magmatic processes. Isotopic dating provides absolute ages for the growth of mineral phases. This usually involves analyses of mineral separates such that the textural relations of the individual grains are difficult to establish. An exception is zircon, which can be analysed in situ. The second approach is to use relative chronometry based on major, trace element and isotope profiles in crystals that may have been modified by diffusion. These yield information on how long crystals were at a particular temperature, without indicating when this occurred. The ages are obtained on individual crystals, and so age distributions can be determined on different crystals from the same whole rock. The ages of crystals and the liquid, as represented by the groundmass, in an igneous rock can be different, and in a number of cases it has been shown that even the larger, and therefore typically older crystals formed after the fractional crystallisation responsible for the whole rock composition. One implication is that the processes of magma differentiation responsible for whole rock compositions may not necessarily be inferred from the compositional record of the larger crystals. Different approaches are therefore used to investigate the crystallisation history and the differentiation of magmatic suites. Crystallisation rates are $\sim 10^{-10}$ – 10^{-11} cm/s, whereas differentiation to high-silica magmas may take up to 2×10^5 years. The ages of crystals at the time of eruption can range back to 10^5 – 10^6 years, the older ages tend to be in the more evolved rock types, and it can take 10^5 years for high-silica magmas to be generated at individual volcanic centres. It appears that the generation of such evolved magmas is thermally controlled, for both fractional crystallisation and the generation of crustal melts, and the rates of fractional crystallisation can, for example, be linked to volcanic power outputs. If crystallisation is in response to magma degassing or decompression, it will be fast and there may be too little time for fractional crystallisation to take place.

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* Corresponding author. Tel.: +44-117-954-5425; Fax: +44-117-925-3385.

E-mail address: c.j.hawkesworth@bris.ac.uk (C. Hawkesworth).

¹ Present address: GEMOC, Department of Earth and Planetary Sciences, Macquarie University, Sydney, NSW 2109, Australia.

1. Introduction

There has been considerable interest in utilising time scales of magmatic processes to constrain models of melt generation, crystallisation and differentiation. Such information also provides insights into the causes of volcanic eruptions, which in turn allows one to develop strategies for volcano monitoring. Recent developments in analytical techniques, and in approach, have refocused attention on the time scales, and hence the rates of magmatic processes. The rates of plate tectonic movements, and the vastness of geologic time, led to a sense that many changes take place very slowly. However, many volcanic eruptions are short-lived catastrophic events, and even before eruption crystallisation can take place in hours or days. Thus, exploring links between evolving magma compositions and the style and timing of volcanic eruptions [1] presents a challenge to the acquisition of a precise chronology for the shortest possible time scales. It also is necessary to develop better models for crystallisation and magma evolution in response to cooling and decompression.

Igneous rocks are the dominant record of magmatic processes. Their minerals and textures reflect the processes of crystallisation and differentiation, once the magmas have left the sites of melt generation. These processes include magma mixing, cooling, degassing and decompression, which for volcanic rocks culminate in eruption [2] (Fig. 1). Traditionally it has been assumed that the larger minerals recorded the evolving composition of the host rock. Therefore, mineral assemblages and compositions have been widely used to explore models for the differentiation of a magma represented by the bulk rock sample [3,4]. However, such simple links may be less common than previously thought, and this is highlighted by the implications drawn from better information on the time scales of magmatic processes. For example:

- Several studies have shown that the minerals in certain volcanic suites formed *after* the differentiation of the host magma: for example, most of the crystals in the Oruanui and consanguineous rhyolites in New Zealand grew after frac-

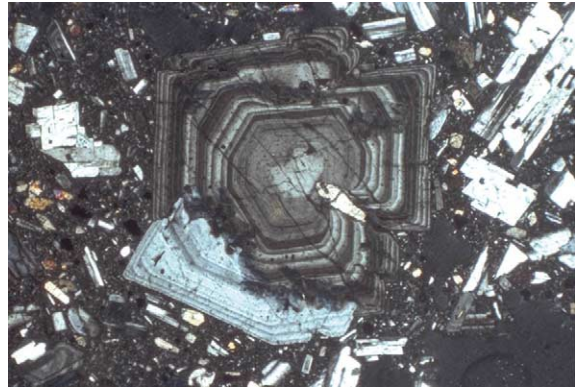


Fig. 1. A thin section of basaltic andesite highlighting the complex magmatic records preserved in plagioclase phenocrysts, and the different, much younger record in the microcrystals and groundmass. The latter will clearly yield ages much closer to that of eruption and whole rock crystallisation.

tional crystallisation had generated the whole rock compositions [5]. It is therefore important to separately determine the ages of the crystals and the time scales over which the host magma differentiated.

- Many erupted magmas represent mixtures of magmas of different compositions potentially derived from different sources, and the effects of these disequilibrium processes need to be understood before the time scales of the magmatic processes can be investigated [6,7].
- Minerals form in response to cooling [3,8], decompression and degassing [9,10], and the time scales of crystallisation associated with these are often different. If crystallisation is in response to falling temperature, the rate of crystallisation increases with the rate of cooling, and hence with decreasing depth in the crust. In some cases this may take tens to hundreds of thousands of years [8]. In contrast, crystallisation in response to decompression or magma degassing can be extremely rapid and take place over days or weeks (e.g. [11]) to a few years [12,13].
- If crystallisation is rapid, and it took place shortly before eruption, there may be too little time for separation of the crystals and the liquid, and hence too little time for magma differentiation by fractional crystallisation at that stage. Conversely, the physical separation of

crystals from the liquid may need time, which may in turn imply that the associated crystallisation is primarily in response to cooling.

Thus, the time scales of magmatic processes constrain the causes of crystallisation, when in the evolution of the host magmas crystallisation of the observed minerals took place, and even the nature of the differentiation processes responsible for the observed magma compositions. Crystallisation results in a build up of volatiles in the liquid, which may in turn trigger explosive eruptions; conversely, magma degassing can cause crystallisation. Hence, the interest in the links between crystallisation and volcanic eruptions, and in the primary controls on magma differentiation.

Much of the information on time scales has come from the application of short-lived U-series isotopes, and there have been major developments in models for partial melting and melt extraction (summarised in [14]). Such models highlight the different histories of the melt and the residual matrix, and in particular the different times they have spent in the melting zone. In this contribution we are concerned with the time scales of the subsequent magma crystallisation and differentiation processes, and the extent to which they may be different for different magmatic associations. Historically the time scales of crystallisation have been estimated from calculations of the thermal history of bodies of hot magma within the crust [8], and observed in the cooling of lava lakes, for example, on Hawaii [15]. Time scales can now be constrained using short-lived isotopes, from major and trace element profiles in crystals that have been modified by diffusion, crystal size distributions (CSD), the power output of volcanic systems, and from the rates of changes in volcanic stratigraphy. The short-lived isotope studies have highlighted that the time scales for the melt and the residual matrix in the melt zone can be very different, as can the ‘ages’ of crystals and liquid (as preserved in the matrix or groundmass) in the same igneous rock. It is this ability to separate the time scales of the different stages and components in the generation of igneous rocks that has encouraged the development of physically realistic models for melt generation, crystallisation and differentiation.

2. Measuring time scales

In view of the number of stages involved in the generation of an igneous rock body, it is important to be clear about which process or stage is being investigated, and hence about the significance of the age information from different approaches. The different aspects in the generation of an igneous rock that can in principle be dated include:

1. The eruption of individual magmatic units.
2. The age of pre-existing crystals, both major and accessory minerals, such as zircon, at the time of eruption.
3. The ‘ages’ of magma compositions at the time of eruption.
4. The average residence time of magma in the crust.
5. The rate of magma differentiation.

Eruption ages are best determined from the ages of crystals in the groundmass, which quenched during eruption, or from a decay scheme involving gases that are reset during eruption, such as Ar [16–19]. The larger and hence older crystals in volcanic rocks, as seen in Fig. 1, can also be dated, and in a number of cases these have yielded ages that are older than the age of eruption. These crystals are often chemically zoned, and those zones record the conditions of crystallisation in the growth history of the individual crystals, plus, as we shall see, any subsequent re-equilibration if they remained at high temperatures for significant periods of time. In practice it is difficult to match up the crystallisation records of these individual crystals and the evolution of the host magma [20]. However, if their ages are close to that of eruption it is likely that they reflect crystallisation from the host magma. Alternatively, older ages may reflect earlier stages of crystallisation in the same magmatic system, the presence of residual minerals from the site of melt generation, or indicate that the crystals are unrelated and just picked up by the magma en route through the crust.

Such discussion of igneous textures highlights how the histories of crystals and melt, now preserved as the groundmass, in the same rock may be different. It follows that they may have formed

over different periods of time, and so in essence have different ‘ages’. The ages of crystals simply date when the crystal formed, but the ‘age’ of the liquid component of a magma may be less straightforward. In principle it can be taken to be either the time since the magma left the melt zone, or the time since differentiation commenced. Either can be inferred from radioactive decay of an isotope system that was fractionated during melting, whereas the latter can also involve some dateable chemical change that takes place during differentiation. If the key questions are ‘how long did differentiation take for this magma’, or ‘what is the rate of magma differentiation’, it may be more useful to estimate the age of the differentiated magma than the age of the crystals.

3. What age information is available?

Dating techniques can be usefully sub-divided into those that yield absolute ages, dependent on the half-lives of the isotopes in question, and those that yield relative ages. A key difference is that most absolute ages are measured on mineral separates that may contain tens to hundreds of crystals, so there is little or no textural control, and a few old grains, for example, can bias the measured age. In contrast, relative ages are often on individual minerals and so their textural position should be clearer, and ages can be obtained on different crystals in the same rock. Absolute ages are most widely reported using the short-lived U-series isotopes of ^{230}Th and ^{226}Ra , and to a lesser extent ^{231}Pa , which have half-lives of 76 kyr, 1.6 kyr and 33 kyr respectively. These can now be determined to high precision by mass spectrometry, and their half-lives are similar to the time scales of magmatic processes. There are two other consequences of these short half-lives; these isotope systems can only be applied to young rocks (e.g. < 300 ka), and they are different from many more widely used isotope systems (K/Ar, Rb/Sr, U/Pb) in that the isotopes may have changed significantly in response to radioactive decay *while* the magmatic processes were in operation.

Relative approaches to dating include the ages of crystal populations inferred from CSD [2,21], and major, trace element and Sr isotope profiles in crystals that may have been modified by diffusion [13,22–25]. The latter rely on methods to predict the distribution of elements and isotopes in a crystal at the time of crystallisation, and then by comparison with the present day profiles it is possible to estimate the extent they have been modified by diffusion. That constrains the length of time the individual crystals were present at high temperatures after formation, and the great strength of the approach is that the histories of individual crystals can be unravelled. A weakness is that there are no absolute ages, the length of time the crystal was at high temperatures is estimated, but not when that was. Other approaches include the time scales over which isotope ratios and other chemical parameters change either within zoned eruptive units or between different units in an eruption sequence [26]. In general, these rely on models for distinguishing the emplacement of different parental magmas, and rocks that are related by fractional crystallisation from similar parental magmas.

Some general aspects of obtaining age information from variations in U and Th isotopes are summarised in Fig. 2. The techniques of absolute and relative dating have been reviewed in more detail elsewhere [18,27], and here we concentrate on the time scales inferred for magmatic processes.

3.1. *The ages of crystals at the time of eruption – radiogenic isotopes*

Bearing in mind that most crystal ages from radiogenic isotope data have been obtained from analyses of mineral separates, there have now been a number of studies that have yielded a wide range of ages on young volcanic rocks. The majority of ages have been from U–Th isotopes, but some are from Rb–Sr and Ar–Ar in high-silica rocks, and there is increasing information from ^{226}Ra – ^{230}Th isotopes [28–31]. Fig. 3 summarises some of the available data on the ages of crystal separates in recent volcanic rocks at the time of eruption, plotted against an index

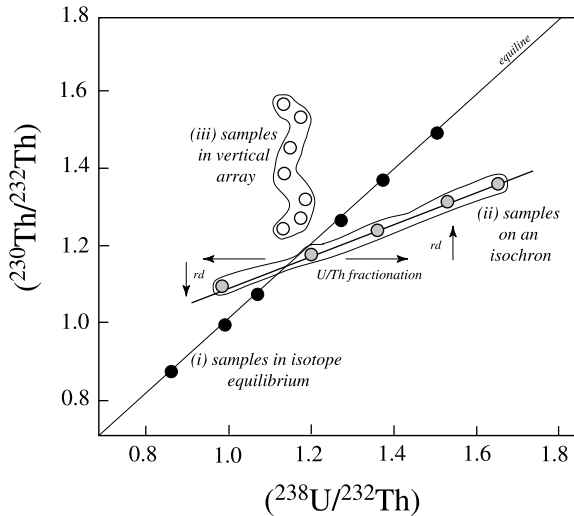


Fig. 2. A U–Th equiline, or isochron, diagram illustrating the kind of age information that can be obtained for three different sets of samples. ^{238}U decays through a number of intermediate isotopes to ^{230}Th , and ^{230}Th decays to ^{226}Ra with a half-life of 75.4 kyr. Samples that are older than ~ 5 times the half-life, i.e. ~ 350 kyr, plot on the equiline with $^{230}\text{Th}/^{232}\text{Th}$ ratios that reflect their $^{238}\text{U}/^{230}\text{Th}$ ratios (group i), and the only age information that can be obtained is that these samples are older than 350 kyr. If samples plot off the equiline and define an isochron (group ii), their age can be calculated as the slope $= 1 - e^{-\lambda t}$ (see also Fig. 3). The assumptions are that at time t all the samples had the same Th isotope ratio, and different U/Th ratios, and that they would therefore have plotted on a horizontal line on the U–Th equiline diagram. If samples plot to the right of the equiline they are said to have excess ^{238}U , and if they plot to the left they have excess ^{230}Th . Samples with excess ^{238}U and excess ^{230}Th migrate vertically towards the equiline at a rate that depends on the half-life of ^{230}Th , and hence the slope of the isochron $= 1 - e^{-\lambda t}$. The samples in group (iii) have variable $^{230}\text{Th}/^{232}\text{Th}$ and almost constant $^{238}\text{U}/^{230}\text{Th}$, and so the time needed to move from the sample with the greatest to the least ^{230}Th – ^{238}U disequilibrium can also be calculated from the decay equation [63]. (rd = radioactive decay.)

of differentiation, molar Si+Al (after [32]). These are framework-forming elements in the liquid, and there is a marked increase in the viscosity of liquids at Si+Al = ~ 66 (e.g. [33]). Critically, most of the old pre-eruption mineral ages are from rocks with relatively high Si+Al values. Thus the simplest observation is the intuitive one that the likelihood of erupted magmas containing old crystals is much greater in the more evolved, and hence higher viscosity liquids. Sim-

ple cooling and crystal settling models also imply that primitive basalts affected solely by cooling in a chamber will contain a small proportion of near-liquidus crystals whereas more evolved magmas will have an increasing proportion of older crystals inherited from earlier stages of the magma's history. As indicated earlier, it is difficult to evaluate the role of these 'old' crystals in the differentiation of the erupted magmas in that they may be unrelated crystals scavenged from the wall rocks [34]. Alternatively, they may be minerals that crystallised from earlier magma batches, and so constrain the length of time over which a particular magma system has been active.

Zircons are increasingly used for U–Th dating because they have high concentrations of these elements, they are robust, they are often zoned and so can preserve records of a number of events. Small amounts of zircon can be analysed, and U and Th isotopes can now be measured in situ by an ion probe. This has the great advantage that it is easy to see what is being dated, and it

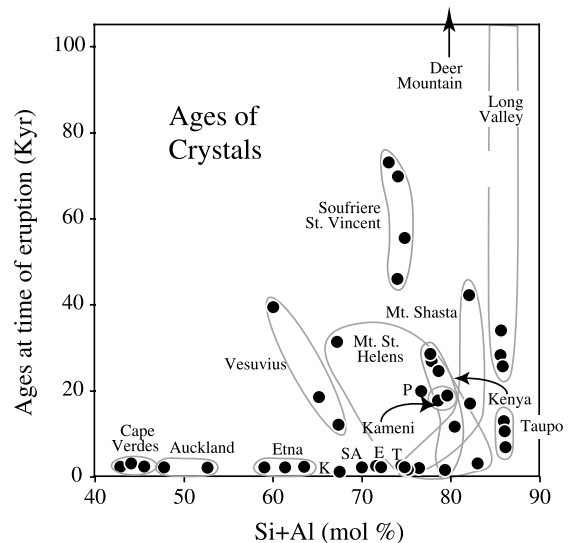


Fig. 3. A plot of age information, in many cases from internal ^{238}U – ^{230}Th mineral isochrons, from recent volcanic rocks versus the Si+Al contents of the bulk rock (after [32]). It illustrates how most of the old mineral ages have been obtained from rocks with relatively high Si+Al contents, and these will also have higher viscosities. E denotes Erebus in Antarctica, K is Kilauea, Hawaii, 1955, P is Parinacota, Chile, SA is Sangeang Api, Indonesia, and the data sources are [28,29,36,37,41,42,48,64–71].

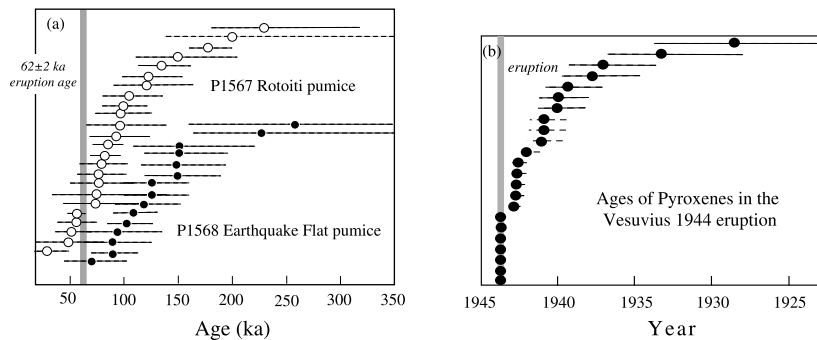


Fig. 4. (a) Summary of U–Th model ages from SIMS analyses of zircons in Rotoiti and Earthquake Flat pumices from the Taupo Volcanic Zone in New Zealand [39]. The error bars represent one sigma error, and the zircon ages are arranged by sample and relative age. (b) Diffusion-based ages on individual pyroxenes in sample DMV18 from the 1944 eruption from Vesuvius arranged by relative age [24]. The error bars correspond to an error in the temperature determination of ± 17 K combined with maximum and minimum values of diffusivity, as appropriate.

allows the age distribution of zircon populations to be unravelled (see Fig. 4). Zircons are restricted to evolved magma types, but the onset of zircon crystallisation can be inferred from bulk rock Zr trends, and the temperatures at which zircon crystallises from magmas of different composition can be calculated [35]. Thus the relative timing and temperature of zircon crystallisation can be linked to absolute ages from U–Th isotopes.

U–Th isotope studies of zircons have yielded broadly similar results from both in situ ion probe analyses, as pioneered by Mary Reid and her colleagues [36,37], and from analyses of mineral separates [38]. In Long Valley in eastern California, for example, whole rock–zircon ages clustered about 230 ka in two suites that were erupted 115 and 0.6 ka ago. Strikingly, it appears that these two suites that were erupted 115 kyr apart sampled magma that had cooled to $< 815^\circ\text{C}$, inferred for zircon saturation, more than 200 ka ago. As with a number of these studies it is more difficult to resolve whether the magma had remained as a crystal mush for 200 ka, or whether it had crystallised and was then remobilised closer to the times of the different eruptions. However, the distribution of zircon U–Th ages in different eruptive units are increasingly being used to identify suites linked by a common pre-eruption magmatic history, and those which are not. Thus, Charlier et al. [39] reported U–Th zircon ages that also ranged back from ages close to eruption

back to 300 ka before eruption in rhyolites from the Taupo Volcanic Zone in New Zealand. However, zircons from two deposits, the Rotoiti and Earthquake Flat, which were erupted within 15–20 km of each other, yielded different mean ages and different age distributions (Fig. 4a), demonstrating that they represent magmas with very different pre-eruption histories. There is now considerable potential to unravel the magmatic structure of high-silica volcanic centres, in the sense of identifying related and unrelated magmas with greater confidence, by comparing the age distributions of zircons in the different eruptive units. Moreover, in situ measurements of Hf isotopes in such zircons make it possible to determine the Hf isotope composition of the magma throughout the history of zircon crystallisation [40].

The ages discussed this far, and illustrated in Fig. 3, are all of 10^4 – 10^5 years, in large part because shorter time scales cannot be resolved without using isotopes with a shorter half-life. ^{226}Ra has a half-life of 1.6 kyr and so ^{226}Ra – ^{230}Th is used to investigate the time scales of processes that occurred in the last few 1000 years. However, their abundances are too small to be measured in situ, and those measured in mineral separates contain significant contributions from melt inclusions, so that it is necessary to estimate the distribution of Ra and Ba in the host mineral and coexisting phases. An elegant solution is to calculate how the ^{226}Ra – ^{230}Th isotope ratios of mag-

mas in equilibrium with the minerals would have changed with time, and hence the time when the magmas in equilibrium with the analysed minerals and the host magma had the same isotope composition [28]. This is then the best estimate of when the crystals formed, and reported ages range from ~ 550 years for a relatively evolved magma from Kilauea on Hawaii [28] up to 3000 years from Sangeang Api in Indonesia, the Tonga arc and Mount St. Helens [29–31], i.e. they are all much younger than those resolved from U–Th isotopes and plotted in Fig. 3.

A major difficulty with ages calculated from radiogenic isotopes is in assessing the extent that they are geologically meaningful. Isochron calculations (as in Fig. 2) rely on straight lines that could be generated by radioactive decay as well as by mixing – in which case the slopes need not have any age significance. The traditional test of the reliability of age data has been to obtain similar, and hence concordant ages from different radioactive decay schemes, and this has been undertaken in a number of studies, as described in the next section. However, it has become increasingly clear as to when concordant ages are not forthcoming, for example, ^{226}Ra – ^{230}Th ages are much less than the U–Th ages on some samples, they may provide new insights into the time scales of magmatic processes ([30,31]; see below).

3.2. Concordant ages

The old pre-eruption ages summarised in Fig. 3 offer a tantalising glimpse of relatively long-lived magmatic records. Concordance with ages from other decay schemes is the ideal test of these ages, but that is rarely possible in the period of 10^4 – 10^5 years. One exception is that U–Th ages can be compared with Rb–Sr ages in highly evolved magmas that have very high Rb/Sr ratios [41,42].

The Olkaria complex in Kenya contains peralkaline rhyolites that were erupted in the last 20 kyr, and the glasses have Rb/Sr ratios up to ~ 600 . A mineral-glass Rb–Sr isochron age of 24 ± 1 ka, and a U–Th age of 25 ± 10 ka were obtained from a rhyolite erupted 8 ka ago [41]. These concordant ages therefore provide confi-

dence in the conclusion that the crystals and the high-Rb/Sr liquids formed ~ 15 ka before eruption. However, the results also highlight the very different characteristics of the two decay systems, in that the Rb–Sr age depends on the high Rb/Sr ratios of the glasses, and the observed spread of U/Th ratios is primarily attributed to accessory phases, in this case chevkinite, as inclusions in the major minerals. Analogous results were obtained from rhyolites at Long Valley, California, where glasses yield a Rb/Sr age of 257 ± 39 ka and internal U–Th isochrons dominated by zircon and allanite yield similar and younger ages, suggesting there was a major differentiation event 150 kyr before eruption [42]. Thus, concordant ages have been obtained in these two magma systems, even though the processes responsible for the Rb–Sr and U–Th ages are very different. The small error in the Rb–Sr age can also be used to constrain magma fractionation rates, and it was concluded that at Okaria these were high at 2.5×10^{-3} km³ of magma per year [41].

In summary, there is relatively little U–Th fractionation between the common major igneous minerals, and inclusions tend to contain U, Th and Ra contents that are much higher than the host minerals. Thus, most U-series mineral ‘isochrons’ are in effect dating the age of the inclusions, whether they are melts or accessory phases. The common exceptions are Ra–Th in feldspar where the contribution from the melt inclusions can be corrected out, and U–Th in zircons. The reasonably established ages extend back to ~ 200 ka before eruption (as illustrated in Fig. 3), and the older ages tend to be in the more evolved rock types.

3.3. Relative ages

Relative ages typically indicate how long a mineral, or glass, has been at a particular temperature, as estimated from mineral equilibria. They provide ages that are much younger, which reflect shorter periods of time, than those from isotope studies, and they are determined on individual crystals. A common approach relies on diffusive re-equilibrium across the boundaries of chemical zones that formed during crystallisation, both in

crystals and around melt inclusions. Early studies used Ni diffusion [43] and Fe–Mg interdiffusion [44] in olivine and Ti diffusion in magnetite [45] to determine the time between magma mixing and eruption, ranging from hours to years. More recently studies have used diffusion of Sr and Ba in plagioclase feldspars [13,23], and Fe and Mg in clinopyroxenes [24].

The diffusion of Ca and Na in plagioclase is slow compared to that for Sr and Ba [46]. The present day variations in An contents therefore reflect the conditions during crystallisation of the different mineral zones, and they also determine the subsequent equilibrium profile for Sr and Ba abundances. Plagioclase phenocrysts from Montserrat in the Lesser Antilles arc, for example, have present day Sr and Ba element profiles that have not reached intracrystalline diffusive equilibrium [13]. Their residence times at the estimated crystallisation temperatures of 850°C were calculated to be ~ 1220 to ~ 10 years in 12 crystals from six samples of prehistoric and very young andesites using finite difference modelling of Sr diffusion [13]. There is no link between these residence times and the age of the host magma. The complex crystal growth histories, rapid crystal growth, and variable crystal residence times can be reconciled if small volumes of andesite were intermittently intruded into the upper crust where they crystallised rapidly, e.g. due to degassing. It is inferred that differentiation to form the andesite magma took place deep in the crust, the observed minerals are younger than the differentiation process(es), crystallisation of these minerals was rapid, and the crystals may have been in the magma for too little time for further bulk rock differentiation to have taken place at shallow levels. Significantly, the residence times for the magmas at shallow levels were short compared with the longevity of the volcano, and this is consistent with generation and long-term storage of andesite deep in the crust.

In an analogous study, but on very different rocks, Morgan et al. [24] exploited the diffusional broadening of originally sharp compositional discontinuities in clinopyroxene to calculate the magma residence times of individual clinopyroxene crystals. They analysed 23 crystals from the 1944

eruption from Vesuvius, and obtained a spread of ages from one unit. The strengths of this approach are, first, that the effects of diffusion are seen in the distribution of major elements, in this case Fe and Mg, and these can be determined to high spatial resolution using back-scattered electron images. Second, the ages of different crystals have been obtained from the same rock. The magmatic temperature for this eruption was $\sim 1080^\circ\text{C}$, and the time the different crystals were at this temperature ranged from 0 to 19 years before eruption (Fig. 4b). This distribution of ages can be used to evaluate the nature of the magmatic processes. Vesuvius erupted every year from 1913 and 1944, and the erupted volumes suggest a magma chamber volume of $\sim 10^8 \text{ m}^3$. Moreover, the cumulative crystal residence curve implies that the near-surface chamber was full from 1913 to 1944, awaiting only a large enough triggering input to cause an eruption.

Finally, in this section, the ages of minerals can be inferred from diffusion profiles between the host mineral and tiny inclusions of melt. Danyushevsky et al. [47] investigated the re-equilibration of olivine crystallised from melt inclusions with the host olivine in magmas from a number of tectonic settings. As the olivine in the melt inclusion has lower Fo contents than the host olivine, the degree of re-equilibration of Fe and Mg between the inclusion and the host constrains the time the olivine crystal was at high temperatures. Up to 85% re-equilibration occurs within 3–5 months, and such short crystal residence times are consistent with efficient removal from primitive, low-viscosity magmas to cumulate zones at the base of the chamber. Olivine contains extremely small amounts of U, Th and Ra, and so their measured abundances in olivine separates will reflect the contributions from melt inclusions. This may offer a new window into the short-lived isotope disequilibria of the more primitive magma compositions preserved in melt inclusions in olivine.

4. Discordant ages

Attempts to confirm the reliability of age deter-

minations by showing that similar ages can be obtained using different techniques have often ended in failure, but discordant ages may also shed new insights into the time scales of magmatic processes [13,30,31]. On Montserrat the relative ages of the plagioclase feldspars are much younger than the age of differentiation of the host magma, and the feldspars are thought to have crystallised en route to the surface [13]. However, the striking age discordance is when rocks and minerals yield U–Th ages of tens of hundreds of years, and yet the same samples are not in ^{226}Ra – ^{230}Th equilibrium, suggesting ages of < 8000 years [30,31]. Such conflicting results are sometimes taken as reason to doubt the application of these short-lived isotopes, but recent studies have demonstrated the new insights that may result when such age differences are reconciled.

An early study of U–Th–Ra isotopes on samples from Mount St. Helens had yielded U–Th isochron ages of up to 34 ± 16 kyr and Ra–Th ages of up to 5 kyr [48]. Following [28], Cooper and Reid [30] recalculated the Ra–Th models ages, having corrected for impurities and Ra–Ba fractionation, and the resultant plagioclase ^{226}Ra – ^{230}Th ages are typically > 2 kyr in lavas with eruption ages ranging from 2200 bp to 1982 AD. Cooper and Reid [30] suggest that the disparity between the U–Th and Ra–Th ages may reflect crystal populations of mixed ages and that the ^{226}Ra – ^{230}Th ages could be consistent either with simultaneous storage of physically distinct magma batches or with incorporation of populations of similarly aged crystals into successive magma batches. Subsequently, very different ages were obtained from plagioclase CSD and U–Th–Ra isotope data for selected lavas and cumulates from Soufriere on St. Vincent in the Lesser Antilles and from the Tonga arc [31]. The minerals yield U–Th ages of 33–50 ka, and yet the samples still preserve ^{226}Ra disequilibrium, suggesting ages of < 8000 years. For Soufriere, separated minerals, glass and whole rocks from several lavas, and minerals from a cumulate xenolith, all lie on a striking ~ 50 ka isochron (Fig. 5). Turner et al. [31] attributed their U–Th–Ra age discordance to zoning in both the composition and age of the larger crystals and on Soufriere

the U–Th isochrons have been successfully modelled as mixtures of old cores (~ 180 ka) and young crystal rims. As minerals and whole rocks from a cumulate and several lavas lie on the same 50-ka isochron, it is envisaged that the cumulate minerals also crystallised ~ 180 ka ago, and they too were remobilised recently by the emplacement of fresh magma and this resulted in new mineral growth on the crystal rims. These new hybrid minerals were collected in young cumulates that were then sampled as xenoliths, and erupted as crystals in young lavas (e.g. in 1979). The cumulate textures are young (i.e. they formed shortly before eruption) even though the mineral cores are old, ~ 180 ka in the simplest model. It also implies that at least beneath island arc volcanoes the magma differentiation time scales and eruptive periodicity are one to three orders of magnitude less than the inferred mineral storage times. In general, minerals which have cores old enough to have been in isotope equilibrium for ^{226}Ra /

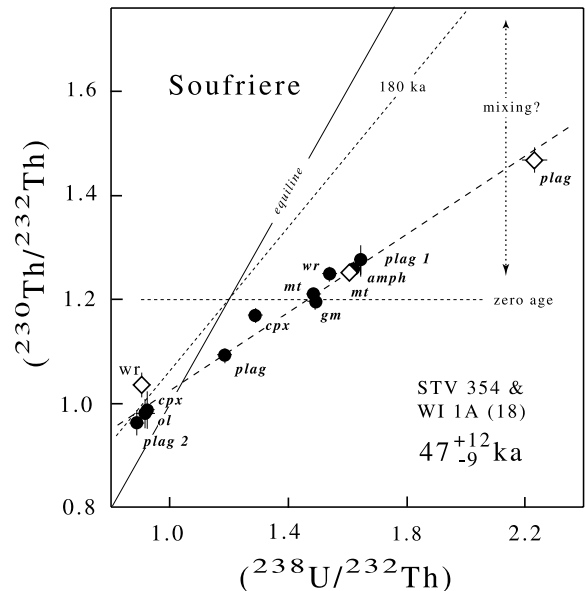


Fig. 5. U–Th equiline diagram for mineral separates from the porphyritic basalt STV 354 (filled circles) and cumulate sampled from a recent ash-flow tuff WI 1A (18) (open diamonds) from Soufriere, St. Vincent, after [31]. The mixing vector highlights how the observed mineral isochron made up from both the basalt and the cumulate can be generated as a mixture of old minerals cores (~ 180 kyr) with young, zero-age crystal rims [31].

^{230}Th (i.e. older than 8000 years) should have ^{226}Ra – ^{230}Th ages that depend on the proportion of cores to rims but be more representative of the age of the latter.

5. The time scales of magmas in the crust

An understanding of the time scales that magmas take to traverse the crust requires pre-eruption age information on the liquid component of the magmas, rather than just on the crystals. This is reinforced by studies that suggest that in some cases the larger crystals are younger than the whole rock differentiation processes [5,13]. We identified earlier that the age of the liquid could be taken to be either the time since the magma left the melt zone, or the time since differentiation commenced. The amounts of differentiation vary, and the time taken for magmas to pass through the crust varies with both tectonic setting and the degree of differentiation.

The time scales of magma differentiation have been inferred from high-Rb/Sr high-silica magmas that yield isochrons [41], where it is argued that differentiation was fast enough to have happened within the error in the isochron age, typically ~ 1 kyr. Another approach is to investigate how short-lived isotope ratios, from which age information can be obtained, vary with indices of differentiation [49–52]. Fig. 6 is a plot of $^{226}\text{Ra}/^{230}\text{Th}$ activity ratios plotted against Th abundances as an index of differentiation for a number of magmatic suites. Samples are in secular equilibrium when $(^{226}\text{Ra}/^{230}\text{Th})=1$, and higher and lower values will evolve to $(^{226}\text{Ra}/^{230}\text{Th})=1$ by radioactive decay. The degree of Ra–Th isotope disequilibria decreases with increasing Th contents in these volcanic suites, suggesting that the more evolved magmas are ‘older’ than the less evolved magmas. This parallels the data on the ages of crystals (Fig. 3) which were also more likely to be older in the more evolved magmas.

Such interpretations of isotope data as that in Fig. 6 imply that these magmatic systems consist of pulses of magma, some erupt after small amounts of differentiation, and others differentiate further and then erupt. Each batch of parental

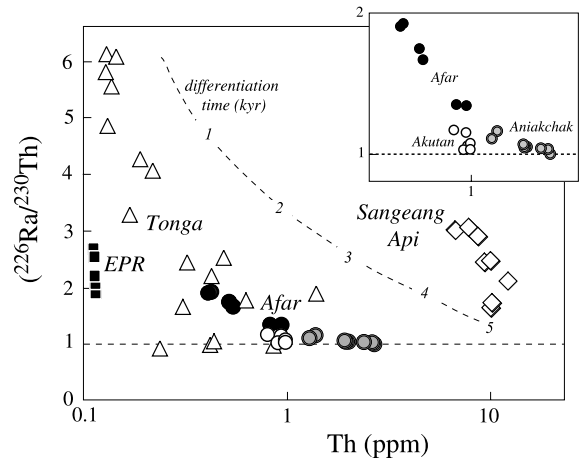


Fig. 6. A plot of $^{226}\text{Ra}/^{230}\text{Th}$ versus Th contents for selected magmatic suites, after George et al. [52]. In particular, the tholeiitic lavas from Akutan and the calc-alkaline lavas from Aniakhchak in the Aleutian–Alaskan arc preserve a similar range in ^{226}Ra – ^{230}Th disequilibria even though the Aniakhchak rocks (circles) experienced significant differentiation, and the Akutan (squares) did not. The slope of the Akutan lavas is similar to those of other tholeiitic lavas from extensional settings, e.g. the Asal Rift [50] and the East Pacific Rise [72], while the Aniakhchak lavas have shallower slopes more akin to the Sangeang Api lavas [29]. In situations where the rate of differentiation is proportional to the slope of the arrays, the shallower the slope, the more differentiation was achieved per unit of ^{226}Ra decay, or time.

magma had similar Ra–Th isotope ratios and Th contents, and different pulses of magma coexisted as separate batches at any one time. If differentiation is dominated by open system processes, the variation in Ra–Th isotopes may simply reflect mixing, and so no time information can be obtained. Nonetheless, different magmatic suites have different slopes on the $(^{226}\text{Ra}/^{230}\text{Th})$ –Th diagram, which in some cases reflect different rates of differentiation and in others are due to differences in the minerals being fractionated, because Ra/Th ratios can be fractionated by feldspar crystallisation. Where they can be isolated the time scales of differentiation range from ~ 550 years [30] up to $\sim 200\,000$ years using U–Th isotopes for phonolites differentiated from basanites at the base of the crust beneath Tenerife in the Canary Islands [51,53]. Such time scales are slow enough for differentiation to take place in response to cooling of the magma rather than by faster decompression

and degassing processes during ascent [54]. It follows that there may be a link between the rates of fractional crystallisation and the power outputs of volcanic systems, at least when they have not been modified by shallow hydrologic systems [51].

Early attempts to infer magma residence times from short-lived isotopes developed models for steady state systems in which the eruptive output balances the influx so that the magma chamber volume remains constant. These were applied to persistently active centres such as Hawaii, Etna, Stomboli and Reunion, where they yield magma residence times of tens to hundreds of years (e.g. [55,56]). The time spent by magmas in the crust can also be inferred from the shift towards isotope equilibrium in response to radioactive decay of U–Th–Ra disequilibria formed in the generation of the parental magmas, as illustrated in Fig. 6. The ages of crystals (Fig. 3) also gives some idea of how long a magmatic system has been in operation. Such age information offers different perspectives, but when they are compared with average eruption rates there is a broad negative array (Fig. 7), such that magmas with high eruption rates tend to spend less time in the crust, and vice versa. One striking exception is the Auckland volcanic field, where very small amounts of mag-

ma have been rapidly erupted from mantle depths, which is where any differentiation takes place [57]. Mid-ocean ridge basalts may take weeks to a few thousand years to traverse the crust [58], and systems dominated by evolved magmas tend to yield older pre-eruption ages (compare Figs. 3 and 7). It appears that volcanoes that erupt semi-continuously have relatively small volumes of evolved magma types. In contrast, evolved magma types tend to be more common at volcanoes where magmas spend longer in the crust, the eruptions are more episodic, and the volcanic centre has been active for $\sim 10^5$ years. In western North America, for example, there is typically 200–300 kyr of magmatic activity at any centre before caldera-related rhyolites are erupted [59,60]. Moreover, models of the thermal evolution of magmatic provinces suggest that significant amounts of crustal melting will also take place after 10^5 – 10^6 years at reasonable emplacement rates of mafic magma [61]. Another inference of the broad array in Fig. 7 is that it constrains the rate of eruption, and hence presumably the melt generation rate, of mafic magmas that traverse the crust very rapidly, which may in turn inform models of magma transport through the crust.

Steady state models for magmatic systems predict a negative relation between magma residence times in a magma chamber (or in the crust) and either the influx or outflux of magma, linked by the volume of magma in the chamber (or in the crust) (e.g. [56]). Some of the time estimates in Fig. 7 are on the basis of steady state modelling from Pyle [56], but for the others there need be no simple link between the estimated magma time scales and magma residence times. Nonetheless, it is interesting that steady state modelling of the negative array in Fig. 7 would suggest magma volumes in the crust of 10–400 km³ consistent with other estimates.

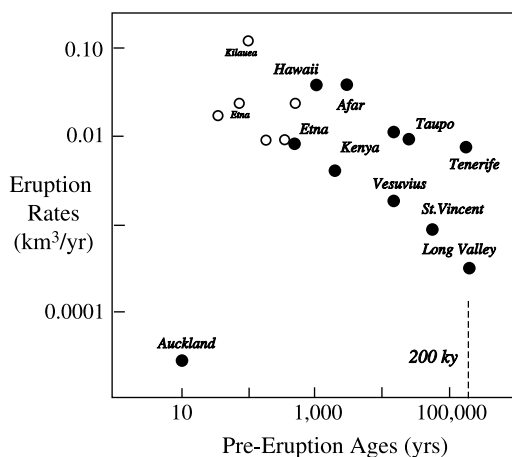


Fig. 7. A plot of the eruption rates (from [73]) against pre-eruption ages (filled dots, as in Fig. 3, and open dots from [56]) from different volcanic centres. The dashed line at 200 kyr is to mark the likely incubation time for the generation of high-silica magmas by crustal melting for reasonable rates of emplacement of mafic magma [61].

6. Concluding comments and ways forward

The recent advances in the study of the time scales of magmatic processes are linked to the development of improved techniques for the anal-

Table 1
Summary of time scales

Fluids from the subducted slab	100–1000 m/yr [49]
Solid diffusion rates	10^{-10} – 10^{-15} cm ² /s [75]
Rates of crystal growth	10^{-10} – 10^{-11} cm/s
Rates of crystal dissolution	5–20 mol/cm ² /s [75]
Ages of crystals at eruption	up to ~ 1.5 Ma (e.g. [34])
Rates of magma differentiation:	
Rb/Sr isochrons	2.5×10^{-3} km ³ of magma/yr [41]
U–Th–Ra	fraction of magma crystallised $\sim 3.5 \times 10^{-4}$ /yr [50]
Incubation period for crustal melting	10^5 – 10^6 years [61]
Magma ascent rates	26 km/day from xenolith-bearing magmas [74] 10 km/yr from U-series isotopes [76]
Rates of eruption:	
Plinian	m/s [77]
Sub-plinian	< cm/s [78]

yses of short-lived U-series isotopes and renewed interest in the information available from diffusion modified element profiles in igneous crystals. The time scales of different aspects of magmatic processes are summarised in Table 1 (see also [75]), and crustal residence times, and degrees of differentiation, vary with tectonic setting, and perhaps with rates of magma supply. The striking observation is that the time scales of partial melt and residual matrix in the sites of melt generation, and the ages of crystals and the liquid component in an igneous rock, are often different, and these should strongly influence models of magmatic processes. A common test of the significance of ages on rocks and minerals is whether similar, or concordant, ages are obtained using different geochronologic techniques [41,42]. However, it is now clear that even when different approaches yield different ages, these may offer additional constraints on the time scale of magmatic history [30,13]. In situ U–Th ages on zircons and relative ages are obtained on individual crystals and so the age distributions of different crystals in a single rock can now be determined [24,39]. Many crystals appear to have formed within tens to hundreds of years prior to eruption, but the more evolved rock types can preserve crystallisation histories of up to 10^3 – 10^5 years. It appears that the generation of such evolved magmas is thermally controlled, irrespective of whether that involves crustal melting or closed system fractional crystallisation, and the rates of fractional crys-

tallisation has been linked to volcanic power outputs [51].

Any discussion of possible ways forward inevitably reflects what is becoming feasible analytically, and here we highlight some possibilities, in no particular order.

- To extend the approach of applying a number of chronometers to the same rocks to build on the recent insights into the time scales of magmatic processes obtained from such studies [30,31]. Resolving the implications of discordant U–Th and Ra–Th ages has isolated different stages in the evolution of a magmatic system, and how, for example, cumulates may be young rocks made up of minerals with old cores.
- Olivine contains such small amounts of U, Th, Ra and Pa that bulk analyses of these isotopes in olivine separates will primarily reflect those of the melt inclusions. Melt inclusions in olivine tend to be of relatively primitive magmas, and they may offer a way to obtain the U–Th–Ra–Pa isotope composition of primitive mantle-derived melts to inform models of melt generation.
- To explore the magmatic structure of individual centres by characterising the pre-eruption history of different units, and identifying those with similar and different pre-eruption histories, from the age distribution in zircons from high-silica centres [39].
- High-silica magmas offer the clearest evidence

for lengthy magmatic histories in the crust (Fig. 3) [36–39,41,42,60]. It can be difficult to determine whether such relatively low-temperature magmas crystallised and were then remelted, or remained as crystal mushes over such time periods, and either can have happened at broadly similar temperature. However, studies that document old pre-eruption ages from Sr isotopes in high-Rb/Sr glass samples strongly suggest that they were derived from long-lived magma bodies, rather than crystal mushes that had been reheated (e.g. [41,42]).

- Models have recently been presented for the thermal history of magmatic systems [61], but the controls on the distribution of crystals of different sizes and ages in a volcanic rock are less well understood. CSD studies rely on independent evidence for the rates of crystallisation, or their age, and models are required that break out of this circle and can build on the new age information from individual crystals.
- The links between crystallisation history and eruption style can now be investigated using the age distributions in crystals from different eruptive products. Moreover, because these ages are relative ages, they can be determined in rocks that are too old for U-series studies.
- To explore the potential of the Hf isotope record of zircons to unravel magmatic histories given that they crystallise over a restricted range in compositions, they can be readily dated by U/Pb and U/Th isotopes and they preserve robust Hf isotope signatures [40].
- In view of the evidence that crystals form in response to decompression and degassing, there is an increasing role for short-lived isotopes that have a gaseous parent and are therefore influenced by magma degassing histories (e.g. [62]). These include ^{210}Pb , ^{210}Bi and ^{210}Po , and they are typically determined by counting rather than mass spectrometric techniques.

There is considerable current interest in the relation between the mineral record in volcanic rocks, in terms of texture, age and size distributions, and composition, the evolution of their host rocks and eruption style. These have long been the goals of studies of magmatic processes, but there is now much more of a sense that excit-

ing new developments are happening in this area.

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Chris Hawkesworth is Professor of Earth Sciences at the University of Bristol, UK. He obtained his D.Phil from the University of Oxford in 1974 having worked on the tectonics of the Eastern Alps. He has studied flood basalt provinces associated with the break-up of Gondwana, subduction-related magmatism, crustal evolution, active volcanoes and changes in palaeoclimate. He has been interested in ways to constrain natural processes with better information on their time scales.



Rhiannon George completed her doctoral research at the Open University in 1997 on the timing and petrogenesis of flood basalts in Ethiopia. She has since worked on the time scales of magmatic processes in island arc volcanic suites using uranium-series and beryllium isotopes, and is currently a Research Fellow in the GEMOC Key Centre, Macquarie University, Sydney, Australia.



Simon Turner obtained his PhD at the University of Adelaide in 1991. Currently he is a Federation Fellow and Professor of Geology in the Department of Earth and Planetary Sciences at Macquarie University, Sydney, Australia, where he specializes in the application of U-series isotopes to constraining the time scales of Earth processes with particular emphasis on subduction zone magmatism.



Georg Zellmer is a postdoctoral fellow at the Lamont-Doherty Earth Observatory of Columbia University, New York, USA. He has worked on the rates and chronologies of magma production, ascent and volcanic eruption throughout his career, notably in Greece, the West Indies and Chile. He has focussed on developing geochronological techniques to elucidate the links between petrogenetic processes and volcanic eruptions, primarily at subduction zone volcanoes.