APPLICATIONS OF ISOTOPES TO IGNEOUS PETROGENESIS

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Summary

The isotopic compositions of igneous rocks can provide a wealth of information about the processes and timescales involved in their formation. Numerous isotope systems, individually or combined, have provided a detailed understanding of a range of igneous processes from the formation and early evolution of Earth's interior, to ongoing processes such as recycling of crust back into Earth's interior, and the formation and evolution of active magmatic systems. Radiogenic, cosmogenic, and light stable isotope systems each have unique characteristics that make them well suited for addressing specific aspects of igneous petrogenesis.

This review introduces the basic behavior of the different classes of isotope systems,

and applications to some of the most fundamentally important processes in igneous petrology, including constraints on the timing of crust–mantle differentiation, the evolution of the isotopically heterogeneous present-day mantle, mass fluxes between crustal and mantle reservoirs at convergent margins, timescales of magma formation and ascent, and the processes and timescales of magma evolution.

Recent advances in instrumentation for isotope ratio measurements of geologic materials are resulting in exciting new applications of isotopes to igneous petrogenesis, including the development of new isotope systems, significant improvements in the measurement precision of conventional isotope systems, and the ability to make high resolution, *in situ* isotope ratio measurements within individual crystals. These new analytical capabilities will have a fundamental impact on future investigations of igneous processes.

1. Introduction

The study of isotope systems over the past 100 years has provided a significant insight into Earth processes and timescales, starting with the application of the radiogenic U–Pb system to dating uranium ore minerals by Boltwood in 1907 (the first direct age measurement of any earth material), and the recognition by Urey in the 1930s and 1940s that stable isotopes of H and O should be fractionated in nature, and thus serve as tracers of geologic processes. Applications of isotope systems to geological problems have subsequently flourished since the early 1960s, as instrumentation and chemical separation techniques have been refined in response to lunar exploration programs and cold war activities, enabling high precision measurements of a large number of isotope systems to be made in a wide variety of geological materials. Among the most important applications of isotopes to geological problems have been those that concern igneous petrogenesis, where isotopes have been fundamental in understanding both the timescales and processes operating in igneous systems.

One of the most fundamental applications of isotopes to igneous systems is the use of radiogenic isotopes as "clocks" to date the crystallization ages of rocks. Crystallization ages of meteorites and the oldest terrestrial minerals constrain the age of the Earth and the formation and evolution of ancient continental crust; and knowledge of the crystallization ages of young volcanic rocks is necessary for determination of eruption recurrence intervals, and thus provides a critical link to volcanic hazard assessment and eruption forecasting.

Radiogenic, stable, and cosmogenic isotope systems also play fundamental roles as tracers of diverse igneous processes. The power of isotopes in this regard is in part because isotopes of an element generally undergo minor or undetectable fractionation during igneous processes. Thus, in contrast to trace element abundance ratios which can be strongly affected by melting and crystallization processes, isotope ratios more directly record the source composition of igneous rocks. In addition, significant isotopic differences exist between the major Earth reservoirs (such as crust, mantle, and core), and isotopes can thus be highly sensitive tracers of the potential sources involved in magmatic processes. Cosmogenic nuclides have provided a "smoking gun" for the recycling of crustal sediment into Earth's mantle. Stable isotopes, fractionated most

strongly at Earth's surface, also have contributed to the growing body of evidence for recycling of crustal material into the mantle. Radiogenic isotopes have constrained other deep Earth processes including melting rates and mechanisms, extent of crust extraction, and core–mantle interaction. The recent combined use of radiogenic and stable isotope systems has been fundamental in our understanding of the processes that affect shallow magmatic systems, and the importance of open-system versus closed-system fractionation.

The following discussion focuses on the behavior of radiogenic, stable, and cosmogenic isotope systems, and their strengths for addressing fundamental problems in igneous petrology.

2. Radiogenic Isotope Systems

Radiogenic isotope systems involve "parent–daughter" pairs in which radioactive parent atoms decay to form radiogenic daughter atoms. The utility of a given radiogenic isotope system depends on the relationship of the half-life of the parent atom to the timescale of the igneous process of interest, and in some cases on the extent to which parent and daughter elements are fractionated from one another during a given petrogenetic process. Table 1 lists the long-lived radiogenic isotope systems (half-lives greater than 10^8 years) that are most commonly used in igneous petrogenetic applications.

Parent	Daughter	Measured Ratio	Half-life	λ
^{40}K	^{40}Ar	^{40}Ar/^{36}Ar	1.28×10^9y	5.54×10^{-10} y^{-1}
^{87}Rb	^{87}Sr	^{87}Sr/^{86}/Sr	4.88×10^{10}y	1.42×10^{-11} y^{-1}
^{147}Sm	^{143}Nd	^{143}Nd/^{144}Nd	1.06×10^{11}y	6.54×10^{-12} y^{-1}
^{176}Lu	^{176}Hf	^{176}Hf/^{177}Hf	3.59×10^{10}y	1.94×10^{-11} y^{-1}
^{187}Re	^{187}Os	^{187}Os/^{188}Os	4.16×10^{10}y	1.67×10^{-11} y^{-1}
^{232}Th	^{208}Pb, ^{4}He	^{208}Pb/^{204}Pb, ^{3}He/^{4}He	1.40×10^{10}y	4.95×10^{-11} y^{-1}
^{235}U	^{207}Pb, ^{4}He	^{207}Pb/^{204}Pb, ^{3}He/^{4}He	7.04×10^8y	9.85×10^{-10} y^{-1}
^{238}U	^{206}Pb, ^{4}He	^{206}Pb/^{204}Pb, ^{3}He/^{4}He	4.47×10^9y	1.55×10^{-10} y^{-1}

Table 1. Radiogenic isotope systems

2.1. Radioactive Growth and Decay

Radioactive growth and decay can be described quantitatively through the relationship between decay rate and the number of parent atoms (*N*) present at any time *t*:

$$-\frac{dN}{dt} = \lambda N \tag{1}$$

where λ is a proportionality constant that is related to the half-life ($t_{1/2}$) of the parent nuclide *N* by:

$$\lambda = \frac{\ln 2}{t_{1/2}} \tag{2}$$

Rearrangement of terms in Eq. (1) followed by integration leads ultimately to the *basic decay equation*, which relates the number of parent atoms N existing at any time t to the initial number of parent atoms N_0:

$$N = N_0 e^{-\lambda t} \tag{3}$$

In the case where a parent nuclide decays to produce a stable daughter nuclide, the number of daughter atoms (D) present at any time t can be expressed as the number present initially (D_0) plus the number produced by radioactive decay of the parent atoms (D^*). D^* can be expressed as the difference between the initial number of parent atoms (N_0), and the number of parent atoms remaining at any time t, as follows:

$$D^* = N_0 - N = (N \bullet e^{\lambda t}) - N = N\left(e^{\lambda t} - 1\right) \tag{4}$$

The *basic ingrowth equation,* which relates the number of daughter atoms to the number of parent atoms present at any time t, can therefore be expressed as:

$$D = D_0 + N\left(e^{\lambda t} - 1\right) \tag{5}$$

The *basic age equation* can be derived simply by rearranging the terms in Eq. (5) to solve for t:

$$t = \frac{1}{\lambda}\ln\left(\frac{D - D_0}{N} + 1\right) \tag{6}$$

The basic age equation can, in principle, be used to make age determinations of igneous rocks and/or minerals for any parent nuclide N that decays to form a stable, radiogenic daughter nuclide D, as long as λ is sufficiently well known and D and N can be measured accurately. However, two additional conditions must be satisfied for the basic age equation to yield a meaningful age:

- The rock or mineral must have remained a closed system with respect to the parent and daughter nuclides, such that the only changes in N and D are because of radioactive decay and ingrowth.
- The number of initial daughter atoms present (D_0) must be negligible compared to the number of radiogenic daughter atoms produced (D^*), or D^* must be well constrained.

2.2. Isochron Diagrams

In many applications to igneous petrology, D_0 is not negligible, and must be determined. This is most commonly achieved by isotopic measurements of multiple cogenetic minerals and/or whole rocks, from which D_0 can be calculated or graphically

determined.

In terms of the ^{87}Rb => ^{87}Sr isotope system, Eq. (5) can be written to express the number of ^{87}Sr atoms present at any time t as a function of the number ^{87}Sr atoms present initially (^{87}Sr$_i$), the number of ^{87}Rb parent atoms, and time t:

$$^{87}Sr = {}^{87}Sr_i + {}^{87}Rb\left(e^{\lambda t} + 1\right) \qquad (7)$$

However, because the analyses of rocks and minerals generally involve measurement of isotope ratios, it is more common to express the abundances of the nuclides in terms of ratios where the denominator is a stable, non-radiogenic isotope of the daughter nuclide. In the case of the ^{87}Rb => ^{87}Sr isotope system, both sides of the equation can be divided by ^{86}Sr, which can be considered a constant since the number of ^{86}Sr atoms does not change with time. Thus, Eq. (7) is more typically represented as:

$$\frac{^{87}Sr}{^{86}Sr} = \left(\frac{^{87}Sr}{^{86}Sr}\right)_i + \frac{^{87}Rb}{^{86}Sr}\left(e^{\lambda t} - 1\right) \qquad (8)$$
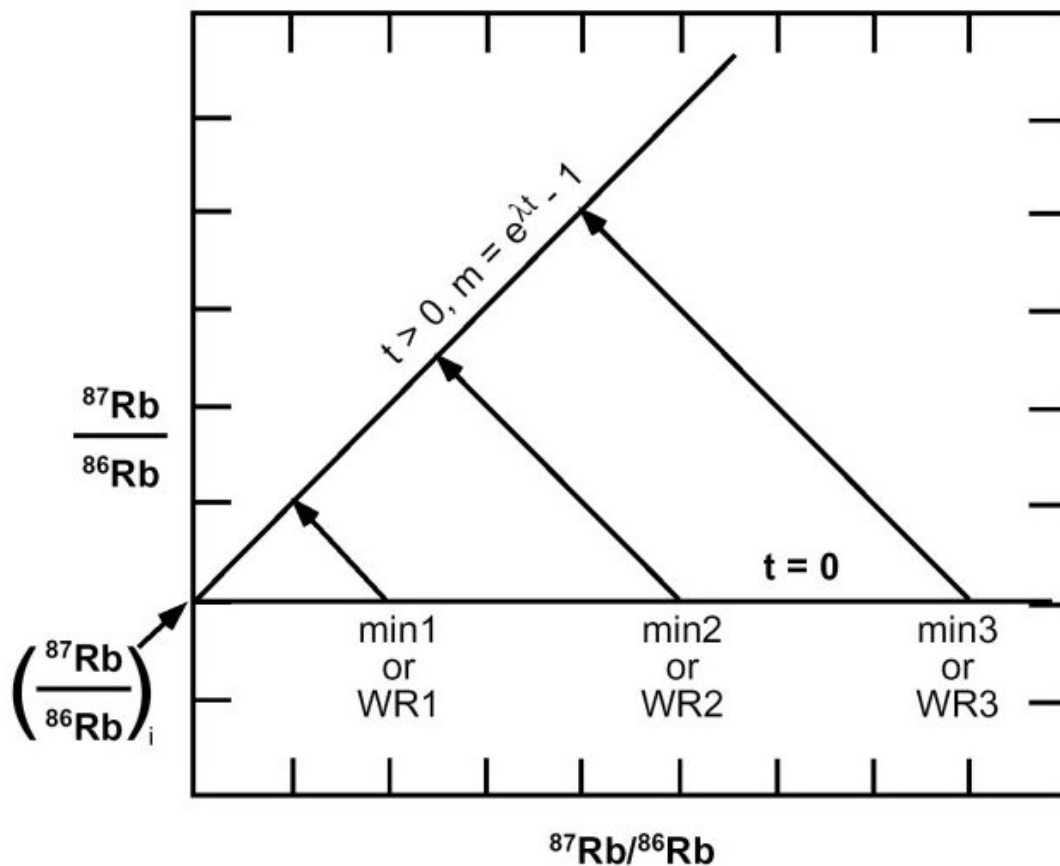

Figure 1. Rb-Sr isochron diagram. Source: after Faure, 1986.

A suite of related minerals or whole rocks that have crystallized from magma with a uniform ^{87}Sr/^{86}Sr, will initially have an ^{87}Sr/^{86}Sr ratio equal to that of the magma from which they formed, since crystallization does not cause measurable fractionation of

"heavy" isotopes such as those of Sr, where the relative mass difference between isotopes is small. However, if the whole rocks or minerals have variable Rb/Sr ratios because of fractionation of Rb from Sr during crystallization, then with time they will develop distinct $^{87}Sr/^{86}Sr$ ratios that depend on the time elapsed since crystallization and the $^{87}Rb/^{86}Sr$ ratio, in accord with Eq. (8). Eq. (8) is a linear equation, thus any suite of cogenetic minerals or whole rocks should display a linear relationship on a plot of $^{87}Sr/^{86}Sr$ vs. $^{87}Rb/^{86}Sr$ as long as they are the same age and share a common $^{87}Sr/^{86}Sr_i$. Such a diagram (Figure 1) is referred to as an "isochron diagram." From the form of Eq. (8) it can be determined that the slope of the line on the isochron diagram is proportional to time t, and the y-intercept of the line is equal to the initial ratio $^{87}Sr/^{86}Sr_i$.

Corresponding isochron diagrams can be designed for any parent–daughter isotope system, as long as the daughter element has a stable, non-radiogenic nuclide with which to ratio the parent and daughter nuclides, and as long as the daughter isotope is itself stable. The case for parent–daughter decay where the daughter nuclide is also radioactive is discussed in the following section.

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Bibliography

Condomines M. and Sigmarsson O. (2000). $^{238}U–^{230}Th$ disequilibria and mantle melting processes: a discussion. *Chemical Geology*, **162**, 95–104. [Review of U-series disequilibria constraints on mantle melting processes.]

Dickin A.P. (1995). *Radiogenic Isotope Geology*, 452 pp. Cambridge, New York, Melbourne: Cambridge University Press. [Introduction to radiogenic isotope geochemistry with in-depth discussions of applications to igneous petrogenesis.]

Eiler J.M., Crawford A., Elliott T., Farley K.A., Valley J.W., and Stolper E.M. (2000). Oxygen isotope geochemistry of oceanic-arc lavas. *Journal of Petrology*, **41**, 229–256. [Survey of O isotope signatures in arc lavas and the implications for crustal recycling at subduction zones.]

Faure G. (1986). *Principles of Isotope Geology*, *Second Edition*, 589 pp. New York, Chichester, Brisbane, Toronto, Singapore: Wiley. [Introduction to radiogenic, cosmogenic and stable isotope geochemistry.]

Halliday A., Lee D.-C., Christensen J.N., Rehkamper M., Yi W., Luo X., Hall C.M., Ballentine C.J., Pettke T., and Stirling C. (1998). Applications of multiple collector-ICPMS to cosmochemistry, geochemistry, and paleoceanography. *Geochimica et Cosmochimica Acta*, **62**, 919–940. [Review of recent and future applications of multi-collector ICP-MS including applications to igneous petrogenesis.]

Hanan B.B. and Graham D.W. (1996). Lead and helium isotope evidence from oceanic basalts for a common deep source of mantle plumes. *Science*, **272**, 991–995. [Radiogenic isotope systematics in mid-ocean ridge basalts.]

Hart S.R., Hauri E.H., Oschmann L.A., and Whitehead J.A. (1992). Mantle plumes and entrainment: isotopic evidence. *Science*, **256**, 517–520. [Radiogenic isotope systematics in ocean island basalts with inferences about the origin of mantle plumes.]

Hawkesworth C.J., Blake S., Evans P., Hughes R., Macdonald R., Thomas L.E., Turner S.P., and Zellmer G. (2000). Time scales of crystal fractionation in magma chambers – integrating physical, isotopic and geochemical perspectives. *Journal of Petrology*, **41**, 991–1006. [Review of processes and timescales of magma fraction with a focus on U-series disequilibria evidence.]

Hofmann A.W. (1997). Mantle geochemistry: the message from oceanic volcanism. *Nature*, **385**, 219–228. [Review of isotope systematics in oceanic island basalts and mid-ocean ridge basalts with implications for chemical geodynamics.]

Ishikawa T. and Nakamura E. (1994). Origin of the slab component in arc lavas from across-arc variation of B and Pb isotopes. *Nature*, **370**, 205–208. [Application of combined stable and radiogenic to island arc lavas and implications for origin of slab-derived fluid at convergent boundaries.]

James D.E. (1981). The combined use of oxygen and radiogenic isotopes as indicators of crustal contamination. *Annual Reviews of Earth and Planetary Science*, **9**, 311–344. [Combined Sr–O isotope mixing models for discrimination between crustal recycling versus crustal assimilation.]

Kellogg L.H., Hager B.H., and van der Hilst R.D. (1999). Compositional stratification in the deep mantle. *Science*, **283**, 1881–1884. [Model for the structure of the Earth's interior based on seismic, heat budget and geochemical data.]

Morris J. and Tera F. (1989). 10Be and 9Be in mineral separates and whole rocks from volcanic arcs: implications for sediment subduction. *Geochimica et Cosmochimica Acta*, **53**, 3197–3206. [Application of cosmogenic nuclides to island arc lavas and implications for sediment recycling at convergent boundaries.]

Thirlwall M.F., Jenkins C., Vroon P.Z., and Mattey D.P. (1997). Crustal interaction during construction of ocean islands: Pb–Sr–Nd–O isotope geochemistry of the shield basalts of Gran Canaria, Canary Islands. *Chemical Geology*, **135**, 233–262. [Oxygen and radiogenic isotope evidence and two-component mixing models for crustal assimilation by ocean island mafic basalts.]

Turner S., Evans P., and Hawkesworth C. (2001). Ultrafast source-to-surface movement of melt at island arcs from [226]Ra-[230]Th systematics. *Science*, **292**, 1363–1366. [Magma ascent rates based on U-series disequilibria data for arc lavas.]

Vervoort J.D. and Blichert-Toft J. (1999). Evolution of the depleted mantle: Hf isotopic evidence from juvenile rocks through time. *Geochimica et Cosmochimica Acta*, **63**, 533–556. [Hf isotope constraints on the evolution of mantle and crust from Archaean to present.]

Wilson M. (1989). *Igneous Petrogenesis*, 466 pp. London, Boston, Sydney, Wellington: Unwin Hyman. [Introduction to igneous petrogenesis including trace element and isotope geochemistry.]

Zindler A. and Hart S.R. (1986). Chemical geodynamics. *Annual Reviews of Earth and Planetary Science*, **14**, 493–571. [Radiogenic isotope systematics in oceanic basalts with applications to chemical geodynamics.]

Biographical Sketch

Elisabeth Widom received a Ph.D. in Geology from the University of California Santa Cruz in 1991. From 1991–1994 she held a Carnegie Postdoctoral Fellowship at the Department of Terrestrial Magnetism of the Carnegie Institution of Washington, and from 1994–1996 a National Research Council fellowship at the National Institute of Standards and Technology. From 1997 to the present she has been an Assistant Professor in the Department of Geology at Miami University.

Her research involves the application of trace elements and radiogenic and stable isotope systems to address a wide variety of geologic problems, including the processes and timescales operating in active magmatic systems, and the composition and evolution of the Earth's interior. Elisabeth Widom has ongoing projects in a variety of ocean island and arc settings, including the Azores, the Canary Islands, Kamchatka, and Japan.

Elisabeth Widom has authored more than ten papers published in international peer-reviewed scientific journals, has served as a panel member for the National Science Foundation, and serves as a reviewer for numerous international journals including *Science, Nature, Earth and Planetary Science Letters, Geochimica Cosmochimica Acta, Journal of Petrology*, and *Journal of Volcanology and Geothermal Research*. She is currently a member of several national and international scientific organizations including the American Geophysical Union, the American Geological Institute, the Geological Society of America, the Geochemical Society, Sigma Xi, and the International Association for Chemistry and Volcanology of Earth's Interior.