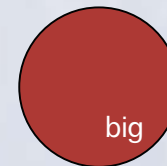


Timescales of magma chamber processes by short-lived U-series disequilibria

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University of Bristol, UK

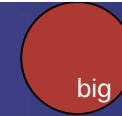


*Petrologia magmatica e Vulcanologia (SC2): Dal mantello alle camere magmatiche
Rimini, 7-8 Settembre 2009*

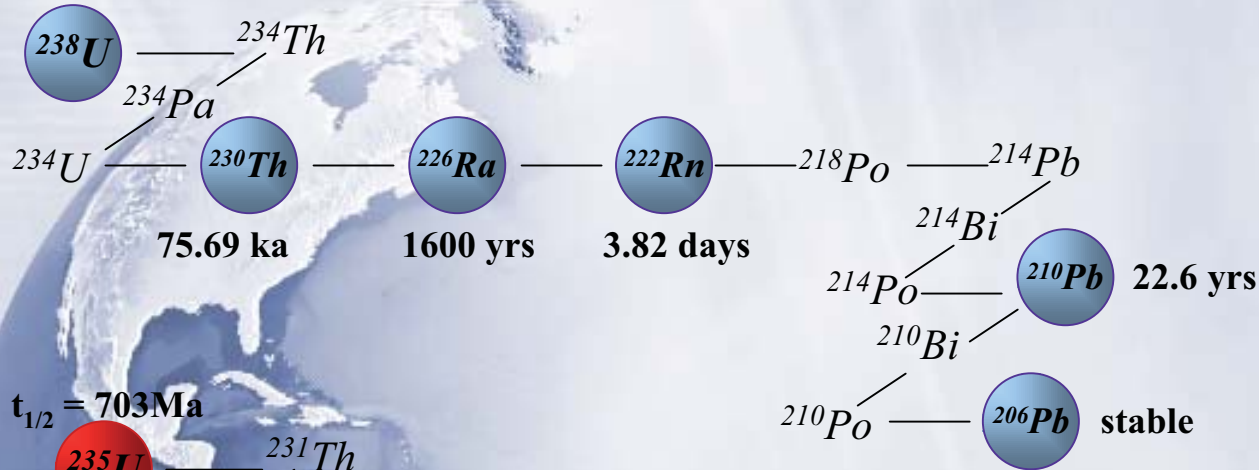


1. U-series geochemistry
2. Isochron diagrams
 - Whole rocks isochrons
 - Minerals isochrons
3. Disequilibria vs. time
 - Closed systems
 - Open systems
 - ✓ Periodic replenishment
 - ✓ Continuous replenishment
 - Examples for natural systems
4. ^{210}Pb - ^{226}Ra disequilibria: degassing

1. U-series geochemistry: Decay series



$t_{1/2} = 4.5\text{Ga}$

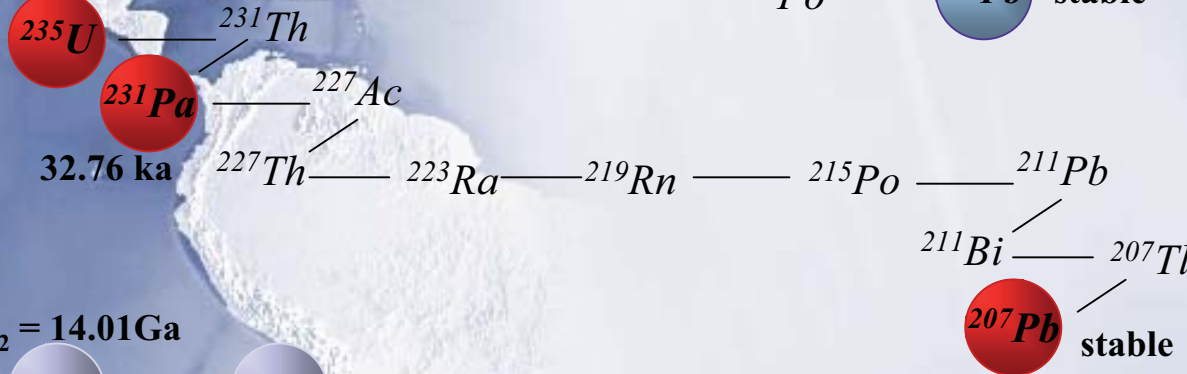


Activity = decay rate

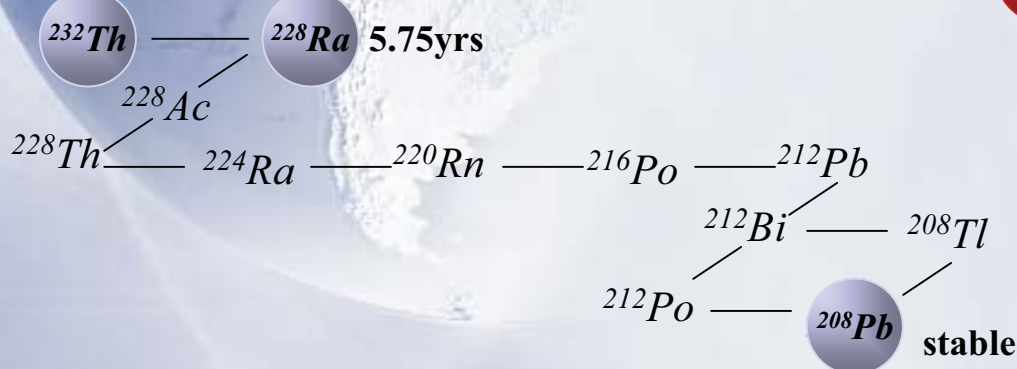
$$(N_i) = N\lambda_i$$

$$\lambda_i = \ln(2)/t_{i1/2}$$

$t_{1/2} = 703\text{Ma}$



$t_{1/2} = 14.01\text{Ga}$



1. U-series geochemistry: Radioactive decay

Concentration of any intermediate nuclide (radiogenic and radioactive) of the chain:

$$N_2 = \frac{\lambda_1}{\lambda_1 + \lambda_2} N_1^o \cdot (e^{-\lambda_1 \cdot t} - e^{-\lambda_2 \cdot t}) + N_2^o \cdot e^{-\lambda_2 \cdot t}$$

→ Daughter decay

Daughter in-growth from parent decay

Converted to activity

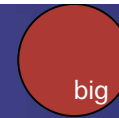
$$\lambda_2 \gg \lambda_1$$

$$({}^{230}\text{Th})_t = ({}^{238}\text{U})_o \cdot (1 - e^{-\lambda_{230\text{Th}} \cdot t}) + ({}^{230}\text{Th})_o \cdot e^{-\lambda_{230\text{Th}} \cdot t}$$

or

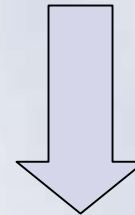
$$\left(\frac{{}^{230}\text{Th}}{{}^{238}\text{U}} \right)_t = (1 - e^{-\lambda_{230\text{Th}} \cdot t}) + \left(\frac{{}^{230}\text{Th}}{{}^{238}\text{U}} \right)_o \cdot e^{-\lambda_{230\text{Th}} \cdot t}$$

1. U-series geochemistry: secular equilibrium



$$({}^{230}\text{Th})_t = ({}^{238}\text{U})_o \cdot (1 - e^{-\lambda_{230\text{Th}} \cdot t}) + ({}^{230}\text{Th})_o \cdot e^{-\lambda_{230\text{Th}} \cdot t}$$

$$t > 350\text{ka} (\sim 5 \times t_{1/2})$$



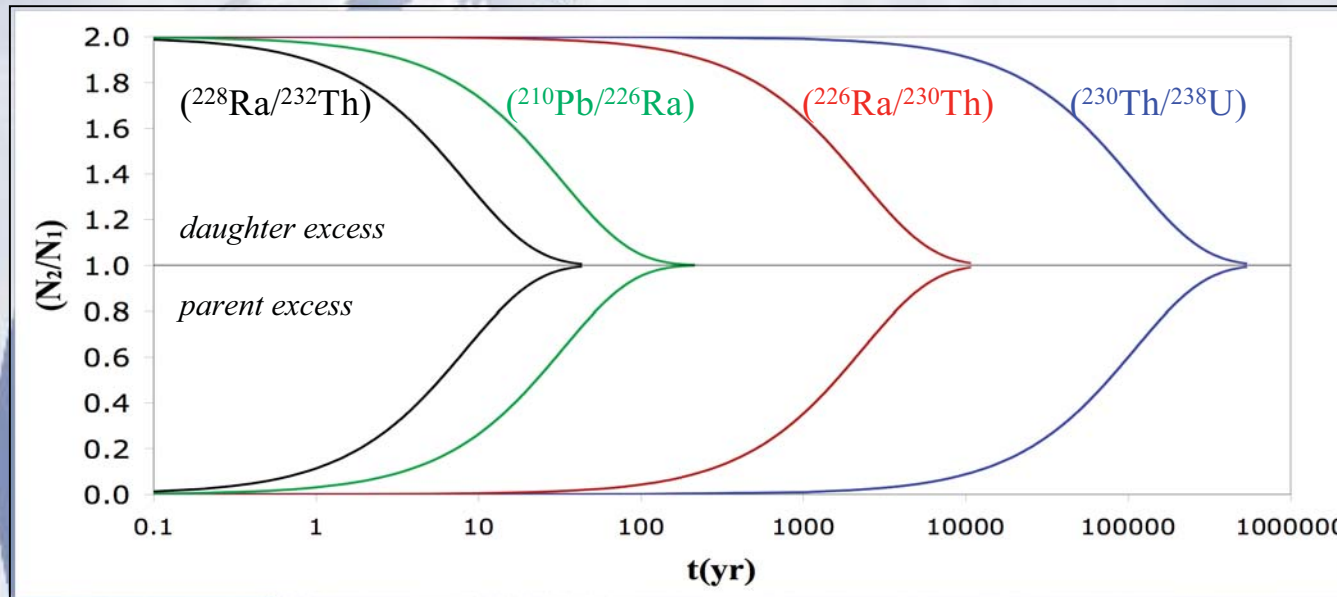
$$({}^{230}\text{Th})_t = ({}^{238}\text{U})_o$$

• **secular equilibrium**: the activity of each nuclide (i.e. the decay rate) of the radioactive chain is constant:

$$\bullet ({}^{238}\text{U}) = ({}^{230}\text{Th}) = ({}^{226}\text{Ra}) = \dots = ({}^{206}\text{Pb}) \quad \text{or} \quad ({}^{230}\text{Th}/{}^{238}\text{U}) = ({}^{226}\text{Ra}/{}^{230}\text{Th}) = 1$$

• When the *secular equilibrium* is disturbed the system reacts through in-growth or decay of the daughter nuclide to restore it

1. U-series geochemistry: Return to equilibrium



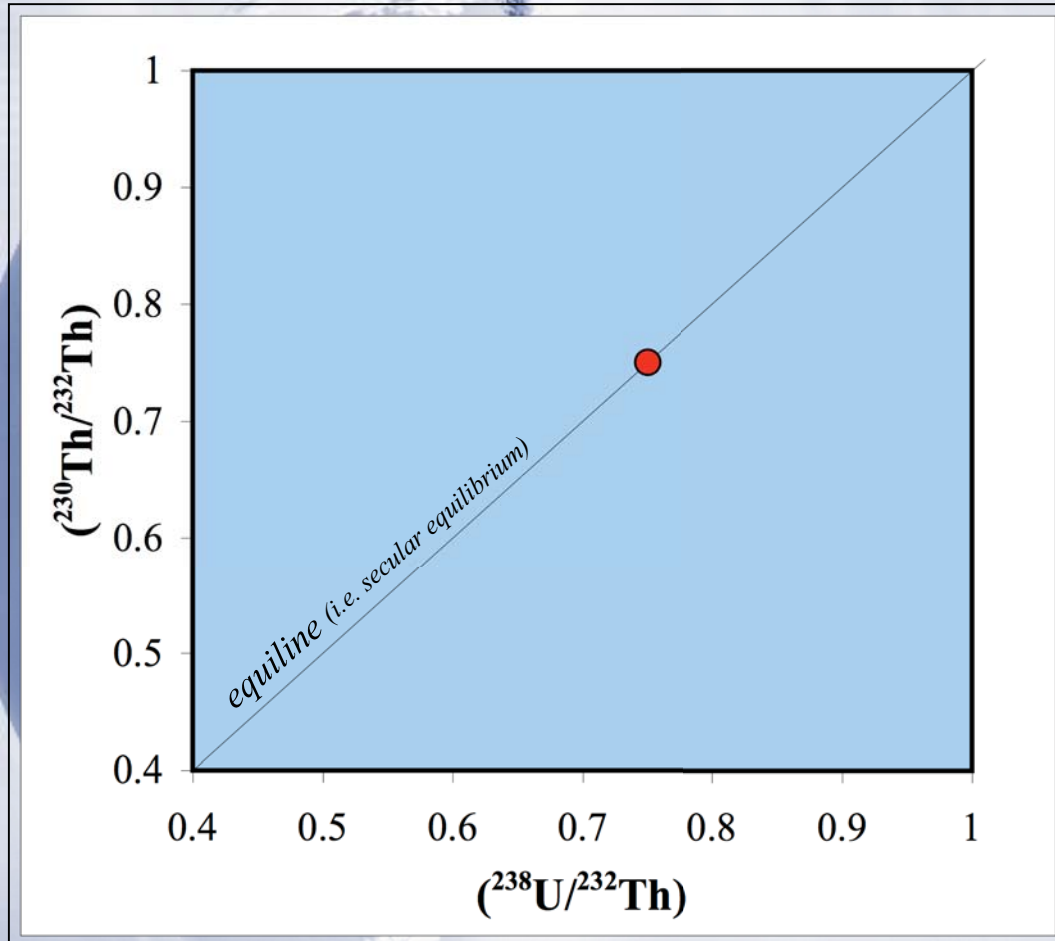
Equilibrium restored in

$\sim 5 \times t_{1/2}$:

- $(^{230}\text{Th}/^{238}\text{U}) \sim 350\text{ka}$
- $(^{231}\text{Pa}/^{235}\text{U}) \sim 175\text{ka}$
- $(^{226}\text{Ra}/^{230}\text{Th}) \sim 8000 \text{ yr}$
- $(^{210}\text{Pb}/^{226}\text{Ra}) \sim 115 \text{ yr}$
- $(^{228}\text{Ra}/^{232}\text{Th}) \sim 30 \text{ yr}$

Different parent-daughter \rightarrow different geological processes \rightarrow different timescales

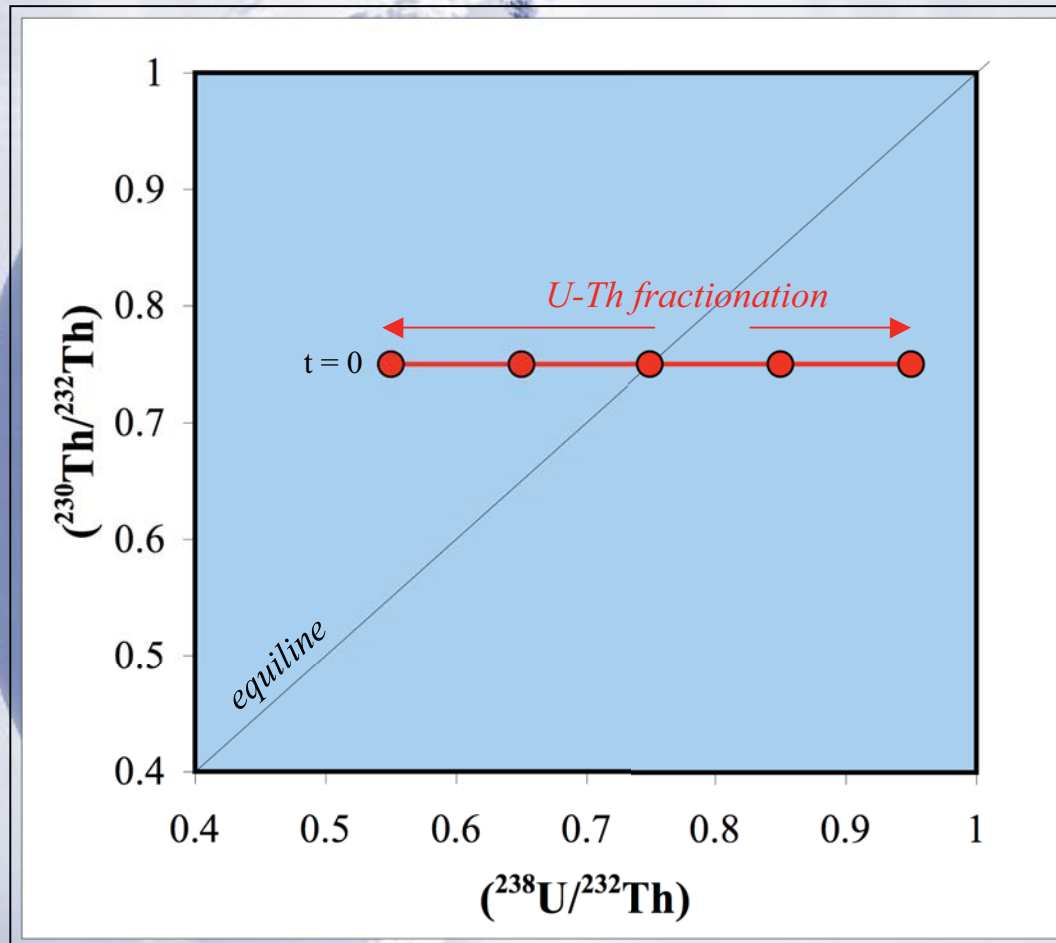
2. Isochron diagrams: U-Th



Equiline diagram:

- ✓ normalisation to a long-lived isotope of the daughter element
- ✓ equiline: $(^{230}\text{Th}/^{238}\text{U}) = 1$

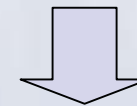
2. Isochron diagrams: U-Th



U-Th fractionation: disequilibrium

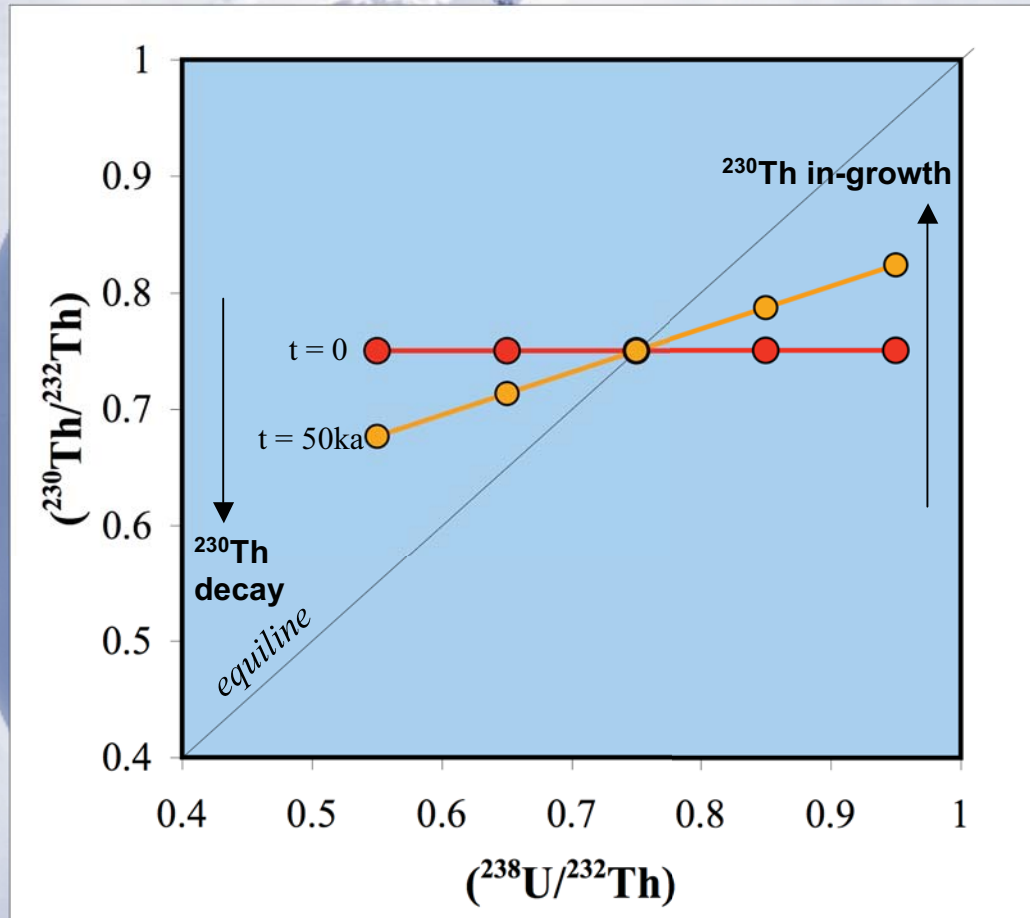
required condition 1:

- ✓ Net fractionation: constant ($^{230}\text{Th}/^{232}\text{Th}$)
- ✓ Instantaneous U-Th fractionation



Horizontal displacement

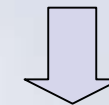
2. Isochron diagrams: U-Th



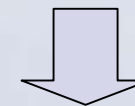
Return to equilibrium

- Constant ($^{238}\text{U}/^{232}\text{Th}$)

Required condition 2:
✓ Closed system

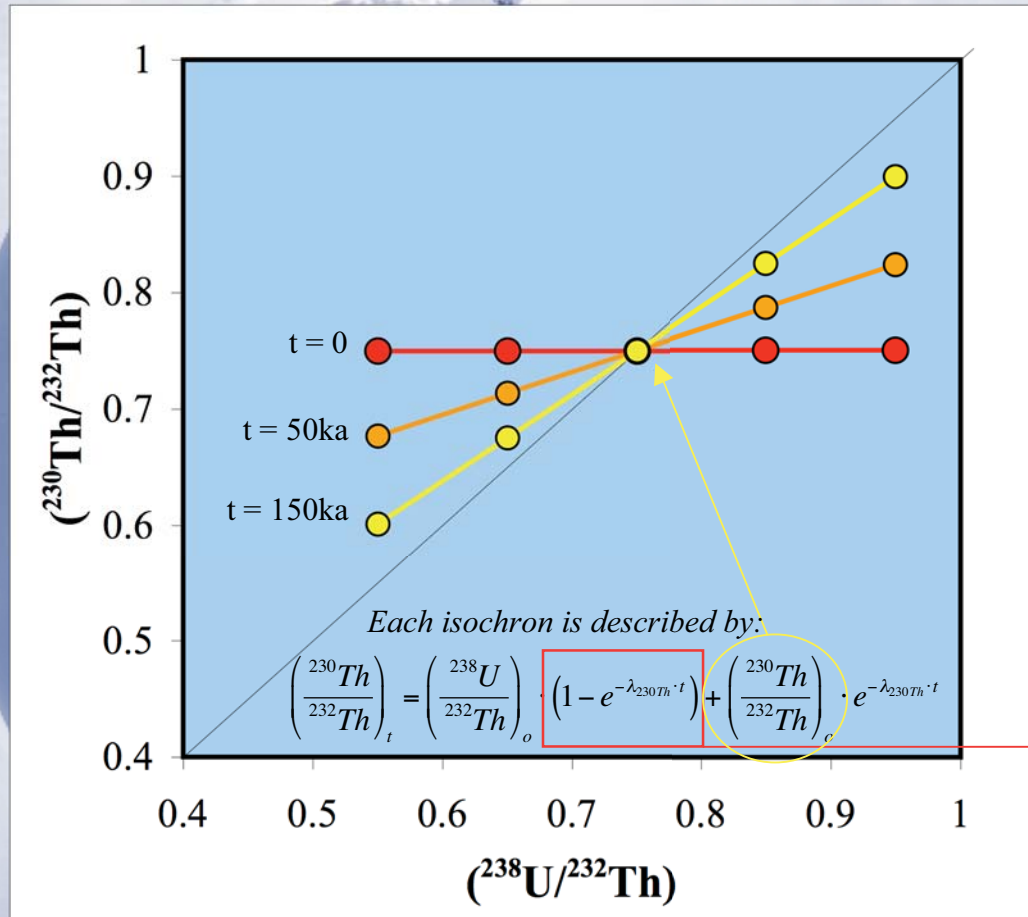


$$\left(\frac{^{230}\text{Th}}{^{232}\text{Th}}\right)_t = \left(\frac{^{238}\text{U}}{^{232}\text{Th}}\right)_o \cdot (1 - e^{-\lambda_{230\text{Th}} \cdot t}) + \left(\frac{^{230}\text{Th}}{^{232}\text{Th}}\right)_o \cdot e^{-\lambda_{230\text{Th}} \cdot t}$$



Vertical displacement

2. Isochron diagrams: U-Th



ISOCHRON:

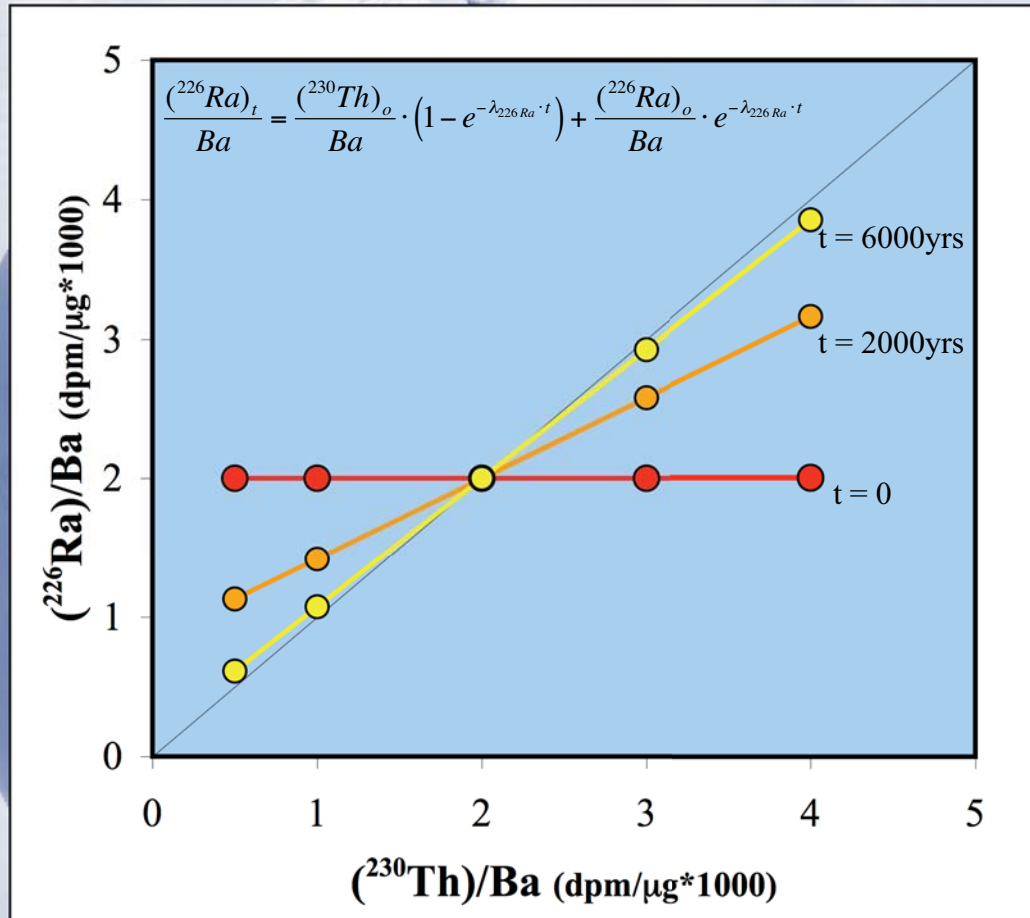
- ✓ condition of the system at a certain time after fractionation.
- ✓ The slope of the isochron is proportional to the time elapsed between the parent-daughter fractionation and the measurement

$$m = (1 - e^{-\lambda_{230\text{Th}} \cdot t})$$

$$t = -\frac{\ln(1 - m)}{\lambda_{230\text{Th}}}$$

Isochrons can be used to determine the age of the fractionation process

2. Isochron diagrams: Ra-Th



No long-lived Ra isotope

Isochron only if Ba and Ra are equally fractionated (i.e. $D_{\text{Ba}} = D_{\text{Ra}}$)

↓

$$t = -\frac{\ln(1 - m)}{\lambda_{226\text{Ra}}}$$

Natural magmatic systems

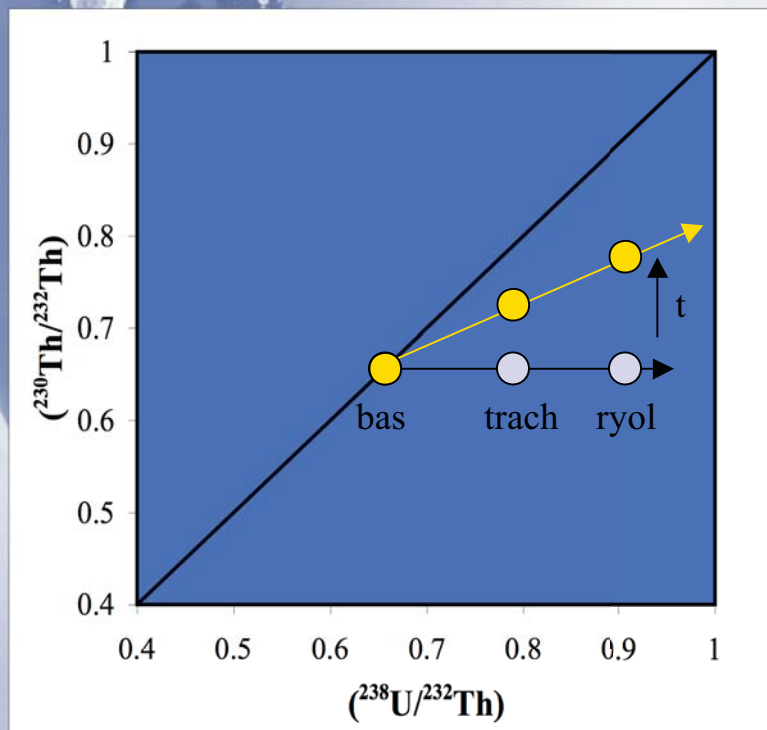
- What are the processes responsible for U-series fractionation?
- Do they produce isochrons?

Processes generating disequilibrium: do they produce isochrons?

1. Magma differentiation

Horizontal fractionation only if differentiation is instantaneous (with respect to $t_{1/2}$ of the daughter)

→ dating the differentiation event.



However:

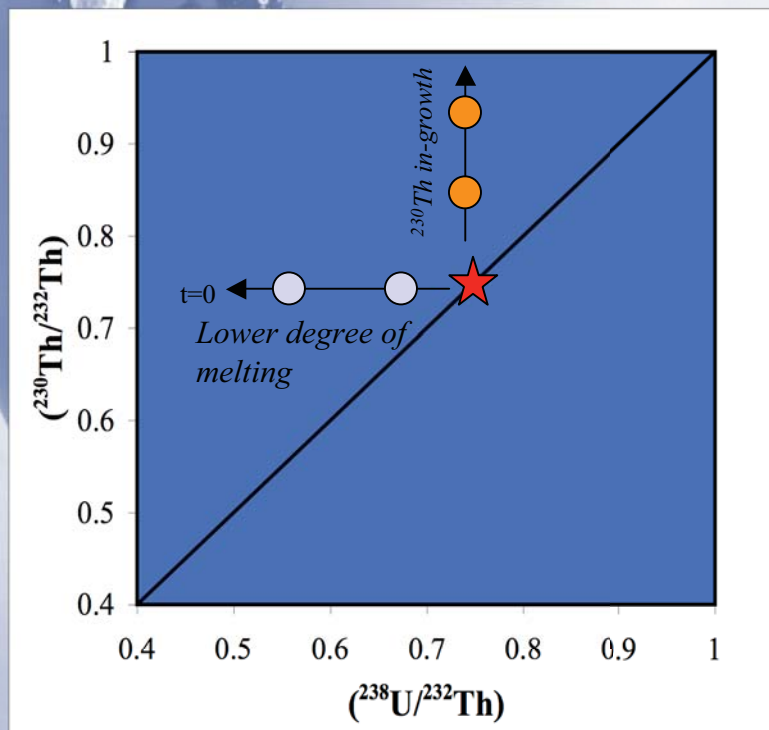
- to be an isochron fractionation should be instantaneous with respect to half life of ^{230}Th .
- U-series elements (e.g. U, Th, Ra) are highly incompatible in most of mineral phases with few exceptions (zircon, apatite)
- Fractional crystallization has little effect on the composition of the erupted volcanic rocks, especially for basaltic compositions

2.1. U-Th fractionation: Whole rocks

Processes generating disequilibrium: do they produce isochrons?

2. Source melting

Different degrees of melting could generate isochrons ($D_U > D_{Th} > D_{Ra} > D_{Pa}$ in typical source mineralogy)



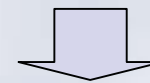
However:

- small differences between partition coefficients of parent and daughter nuclides

→ Significant disequilibrium generated only by small degree melts

- Large ^{230}Th - (^{226}Ra -, ^{231}Pa -) excesses in MORB

→ Daughter in-growth in the up-welling melting region: Dynamic melting models (e.g. McKenzie, 1985 EPSL 72, 149-157; Elliott, 1997, CG 139, 165-183) for MORB and OIB



NO ISOCHRONS

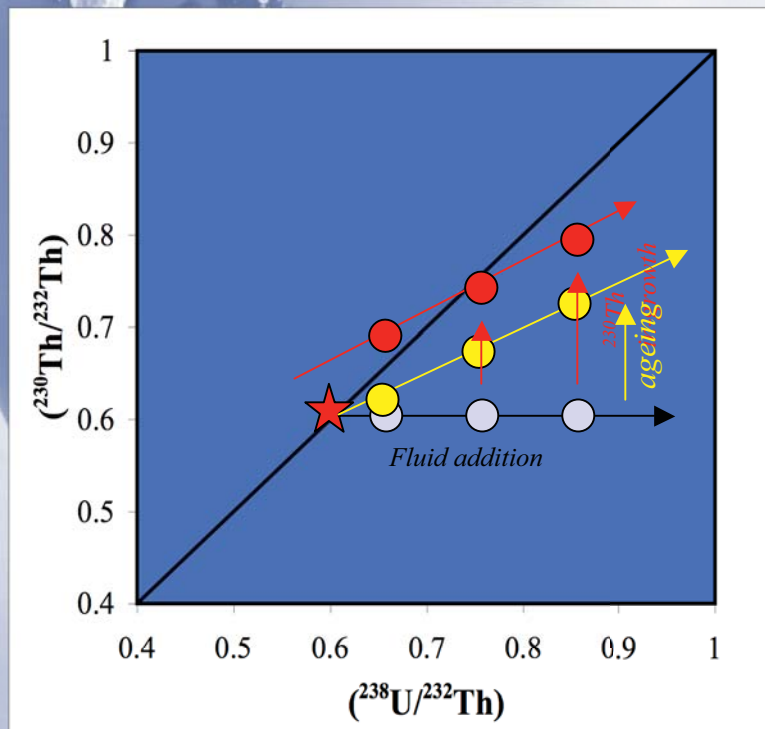
2.1. U-Th fractionation: Whole rocks

Processes generating disequilibrium: do they produce isochrons?

3. Fluid addition

Mobility in the fluid phase: $D_{Ra}^f > D_U^f > D_{Th}^f > D_{Pa}^f$

→ Many island arc magmas show ^{238}U -excess



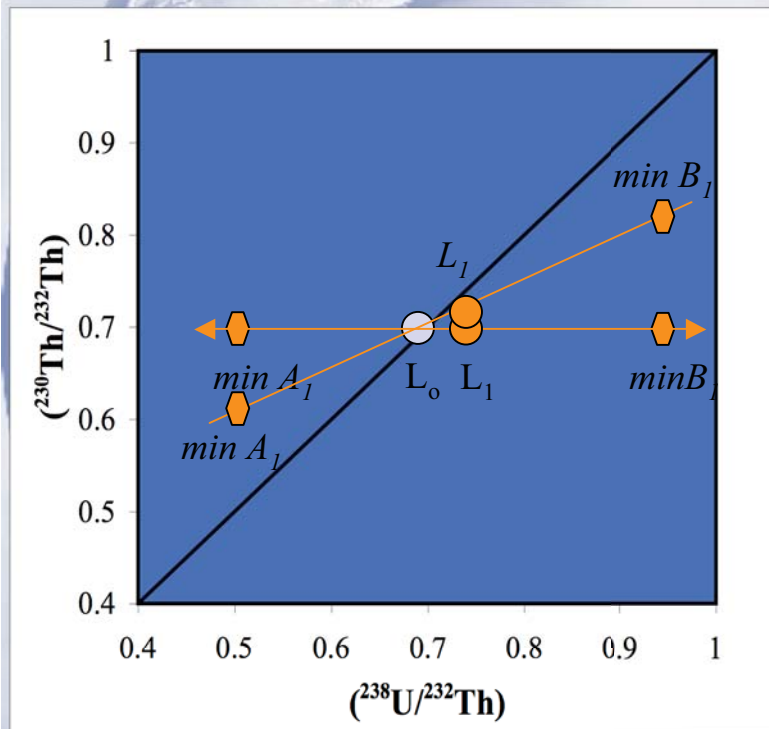
→ Age of fluid addition to the mantle wedge

However

- ^{231}Pa excesses suggest the involvement of in-growth during melting also in island arc rocks (e.g. Thomas et al., 2002, GCA 66, 4287-4309)
- dynamic models applied to U-enriched sources produce oblique trend similar to those produced by ageing.

Difficult to discriminate between the two processes

4. U-Th fractionation during mineral crystallisation



- Instantaneous and contemporaneous crystallisation
→ horizontal displacement

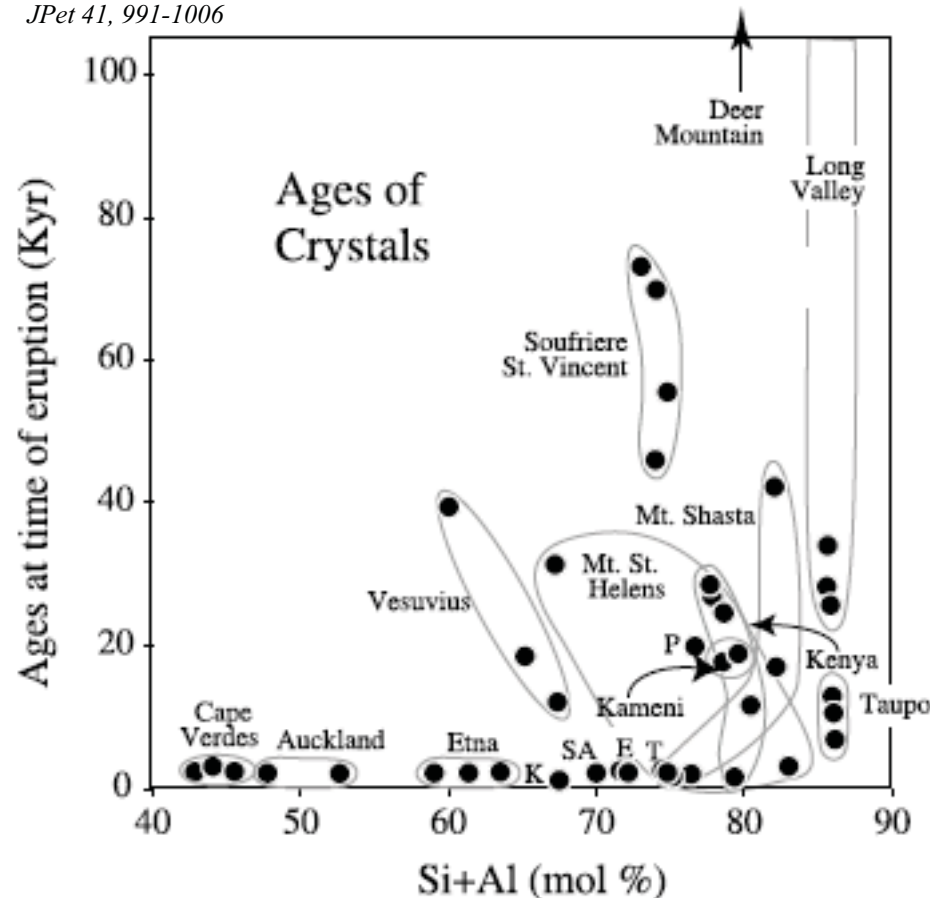
- If crystals and liquid remain isolated they evolve with time to restore equilibrium and when measured lie on an isochron.



AGE OF CRYSTALLISATION of *minA* and *minB*

2.2. Mineral Isochrons: natural systems

From Hawkesworth et al. 2004, *EPSL* 218, 1-16 after Hawkesworth et al. 2000, *JPet* 41, 991-1006



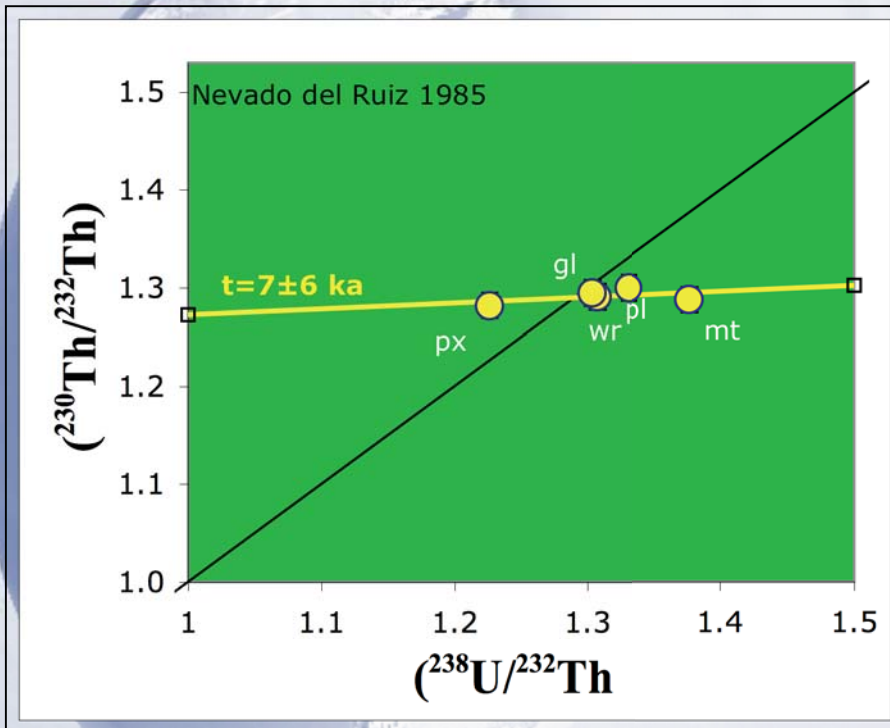
Ages of crystals at time of eruptions in different volcanoes

- crystallisation ages $>$ eruption ages
- generally older crystal ages in more differentiated products

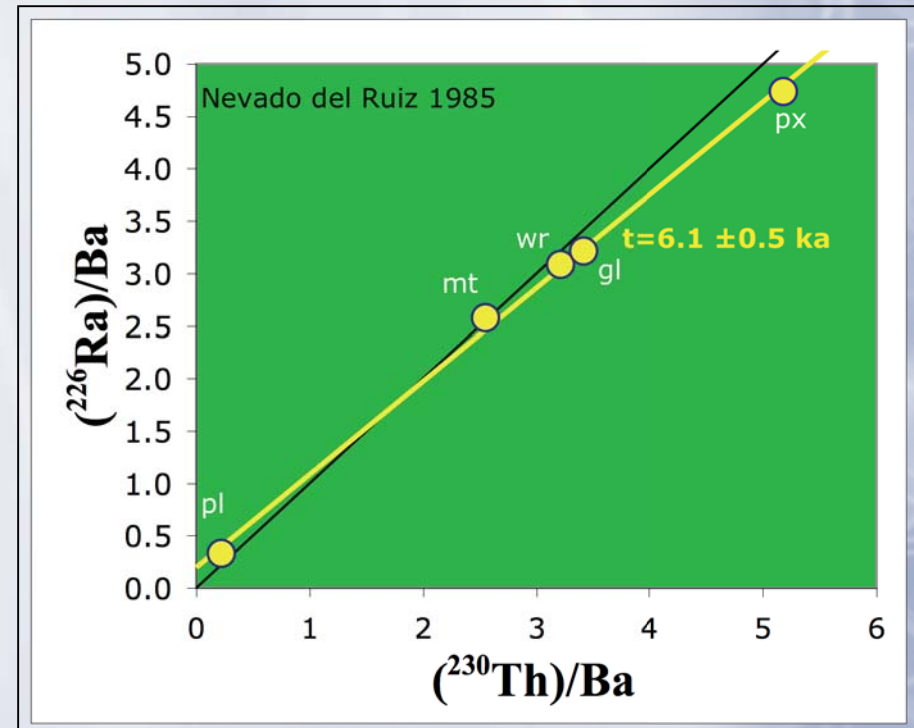
Ages $>$ residence times inferred with other methods (e.g. diffusion)

Crystals may be stored in the magma chambers for several thousand years

2.2. Mineral Isochrons: natural systems



Data from Schaefer et al., 1993, GCA 57, 1215-1219

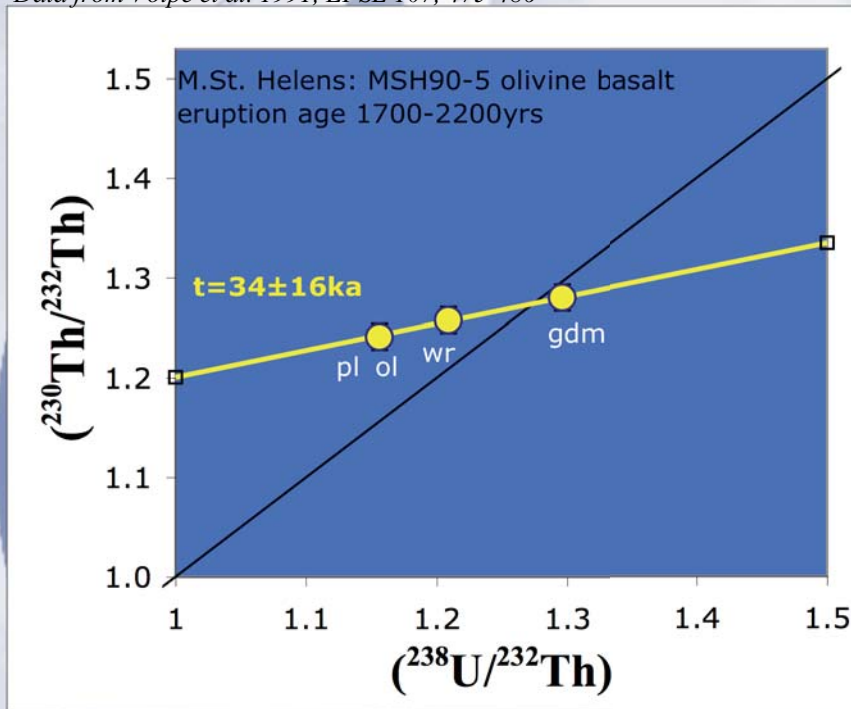


Data from Schaefer et al., 1993, GCA 57, 1215-1219

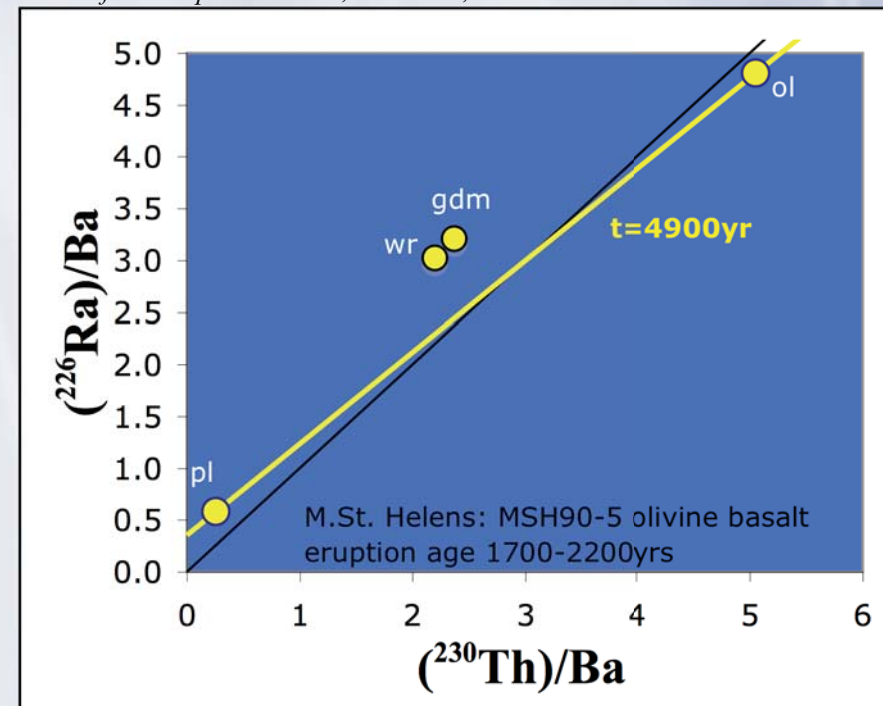
Concordant U-Th and Ra-Th ages (few cases)

2.2. Mineral Isochrons: natural systems

Data from Volpe et al. 1991, EPSL 107, 475-486



Data from Volpe et al. 1991, EPSL 107, 475-486



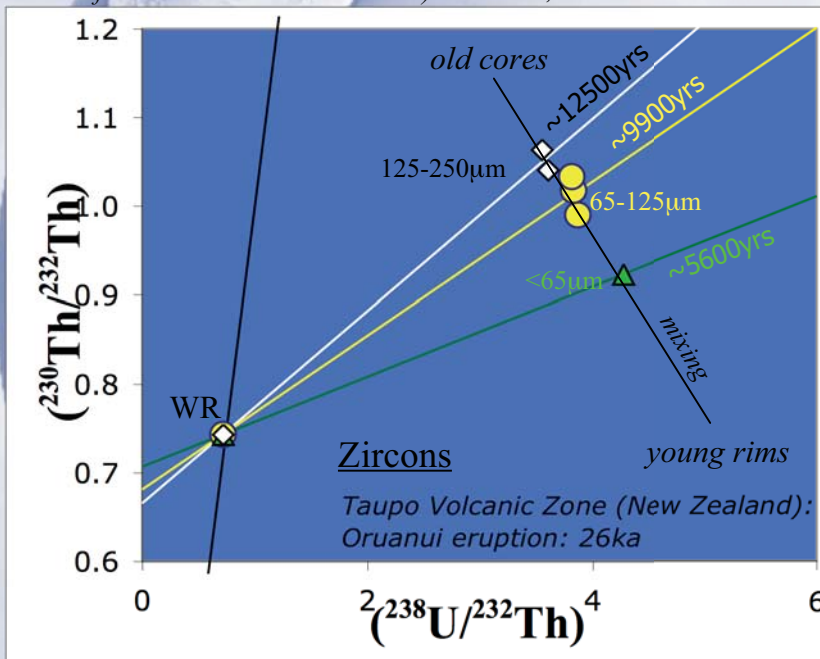
- Discordant Ra-Th and U-Th ages
- U-Th mineral isochron ages > 8 ka in rocks with Ra excess (not possible!)

Why? What ages are we measuring?

2.2. Mineral Isochrons : natural systems

Possible factors affecting mineral isochron ages:

Data from Charlier & Zellmer 2000) *EPSL* 183, 457-469



1. Mineral separates: large number of crystals required

- different populations of crystals
- recycling of old cumulates phases

2. Prolonged crystallisation history

- Crystallisation is not instantaneous: ages are averages between older core and younger rims (e.g. Charlier & Zellmer 2000, *EPSL* 183, 457-469)

Ra-Th and U-Th isochrons average different portions of the crystal population

3. Inclusions of U- or Th-rich accessory minerals

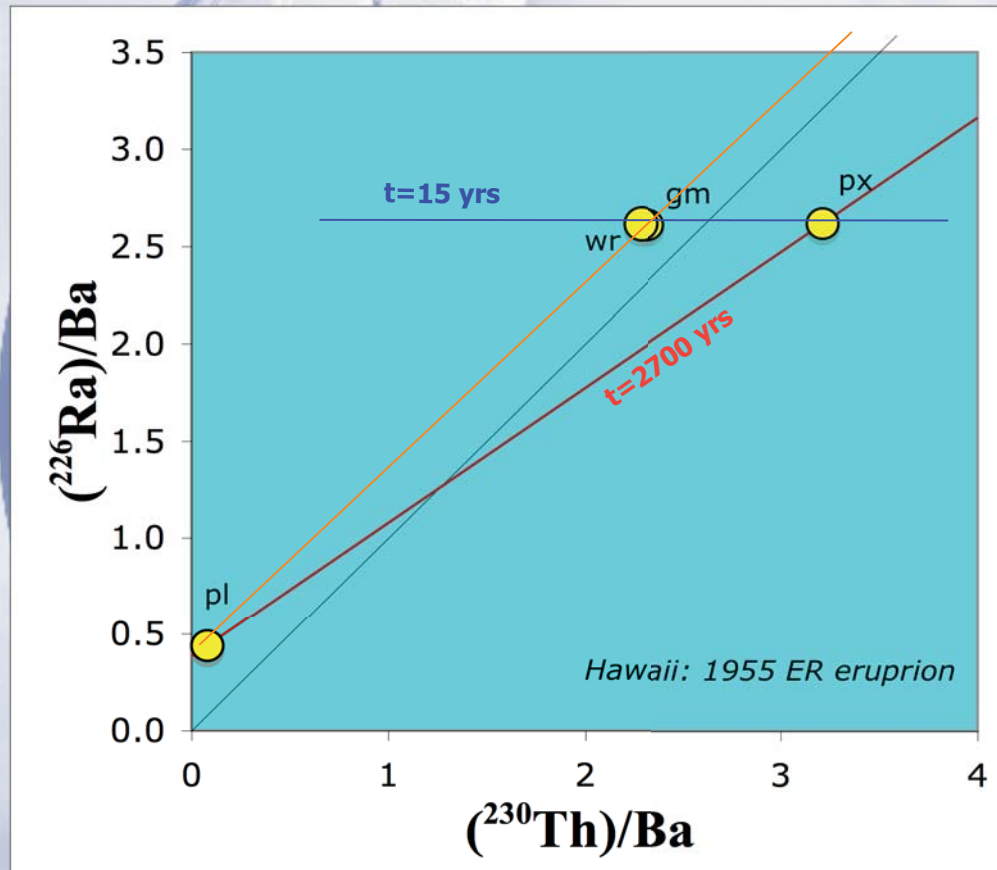
- may dictate the measured isotopic ratios of the crystal separates (e.g Heumann & Davies, 2002, *JPet* 43, 557-577)

2.2. Mineral Isochrons: ^{226}Ra - ^{230}Th ages



$$D_{\text{Ra}} = D_{\text{Ba}}?$$

Data from Volpe et al. 1991, *EPSL* 107, 475-486



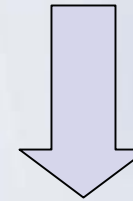
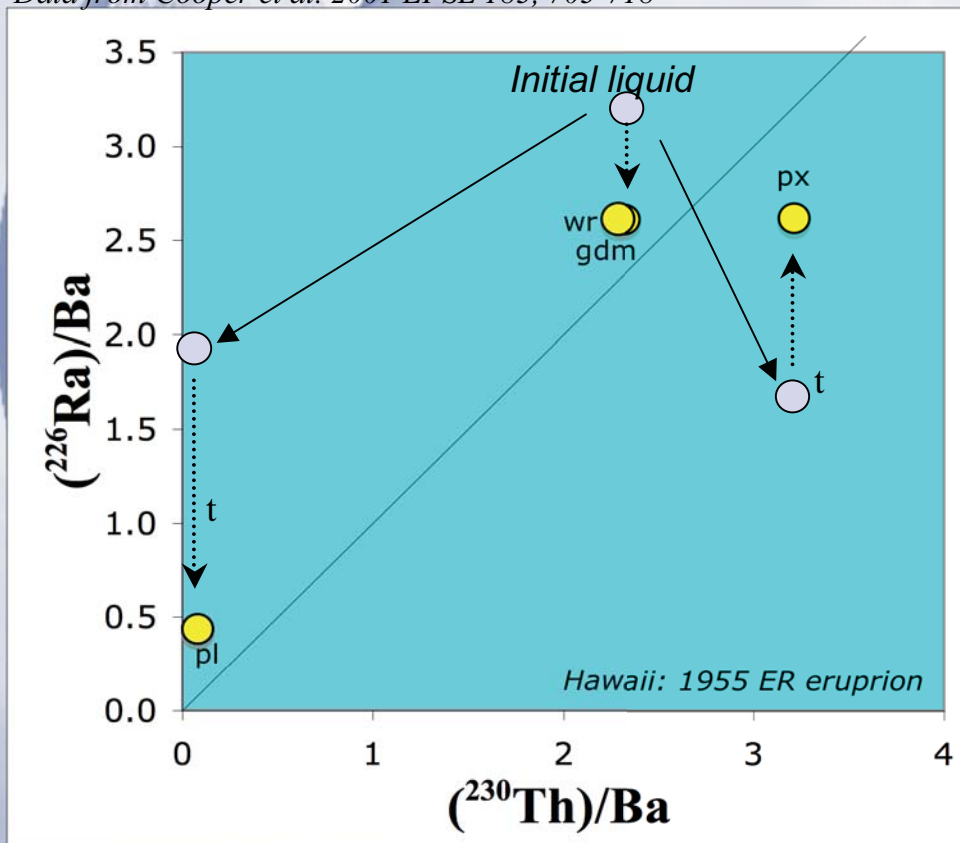
- One isochron between minerals?
- Different crystallisation ages for pyroxene and plagioclase?
- Is any of these a real isochron?

Only if $D_{\text{Ra}} = D_{\text{Ba}}$

2.2. Mineral Isochrons: ^{226}Ra - ^{230}Th ages

Blundy & Wood (1994, *Nature* 372, 452-454) demonstrated that the bigger ionic radius of Ra results in different partition coefficients for Ra and Ba in silicate minerals ($D_{\text{Ra}} < D_{\text{Ba}}$)

Data from Cooper et al. 2001 *EPSL* 183, 703-718

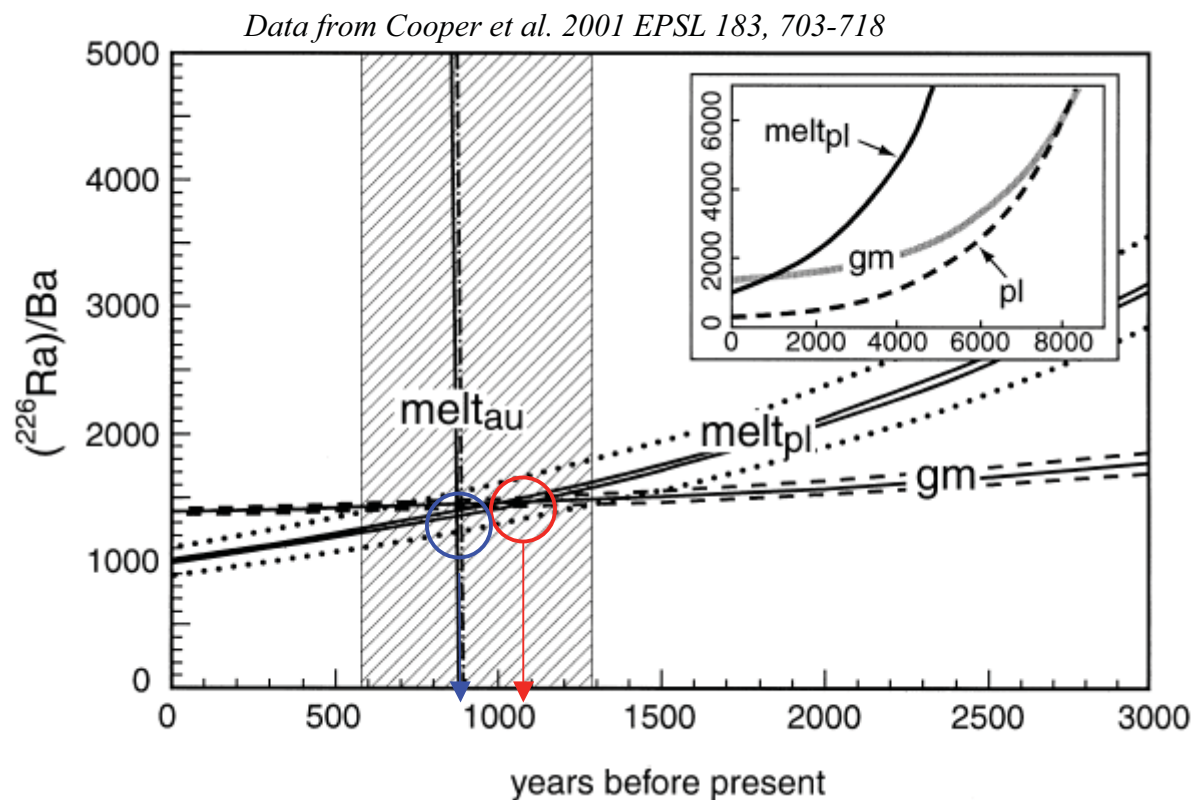


- NO ISOCHRONS
- if D_{Ra} and D_{Th} can be estimated
- Crystallisation ages can be recalculated if the initial $(^{226}\text{Ra})/\text{Ba}$ in equilibrium with the crystals is known.

2.2. Mineral Isochrons: ^{226}Ra - ^{230}Th ages

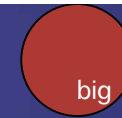
Unknown initial liquid: Cooper et al., 2001 EPSL 183, 703-718

- $(^{226}\text{Ra})/\text{Ba}$ of hypothetical melt in equilibrium with the minerals at time of eruption (from D_{Ba} and D_{Ra})
- modeling of backwards $(^{226}\text{Ra})/\text{Ba}$ evolution for calculated melt and measured groundmass.

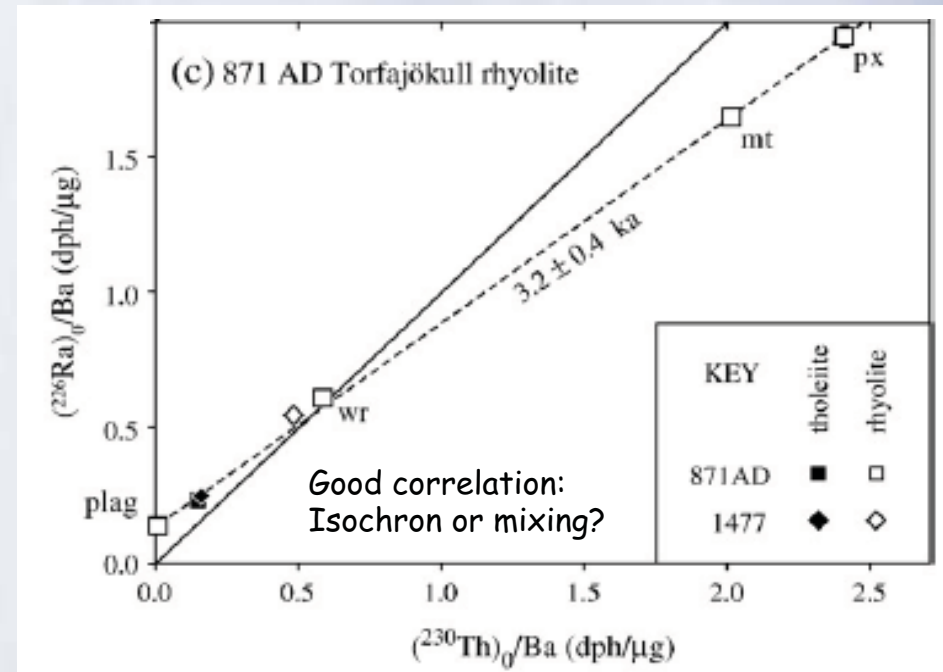
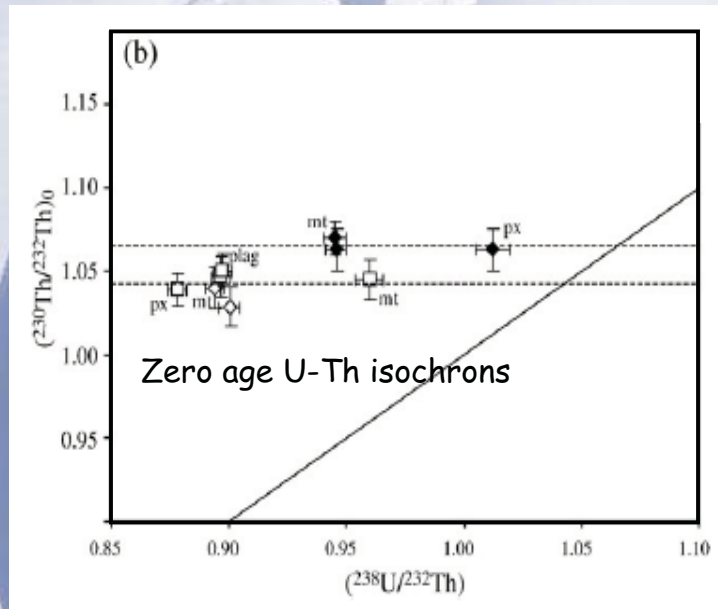


Intercept with groundmass
=
Age of crystallisation

2.2. Mineral Isochrons: ^{226}Ra - ^{230}Th ages

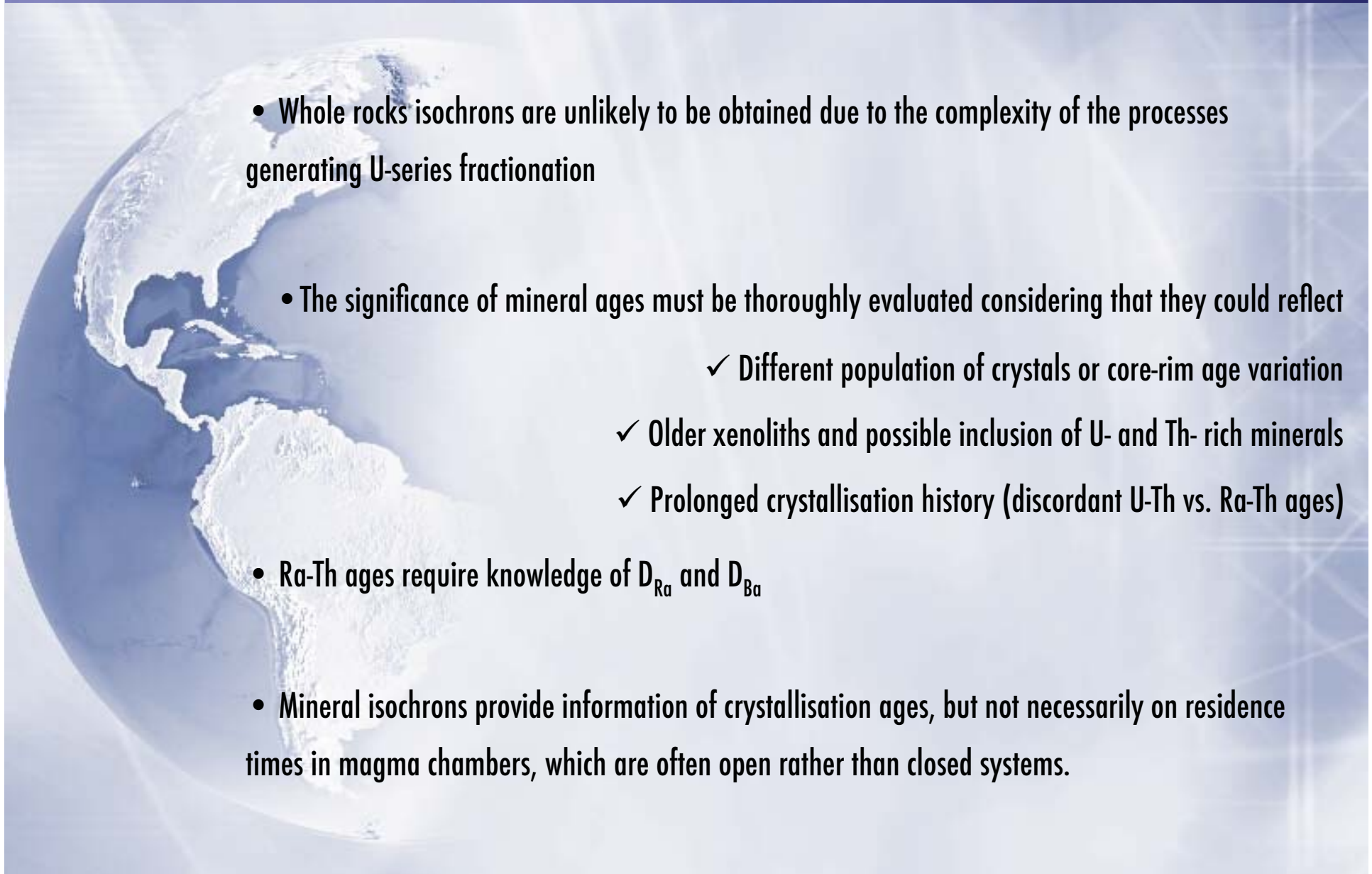


Not so simple: might $D_{\text{Ra}} = D_{\text{Ba}}$ be the best approximation?



Zellmer et al. (2008) and Rubin & Zellmer (2009) vs. Cooper, (2009), EPSL

- Only few direct D_{Ra} measurements (Fabbrizio et al., 2008, 2009)
- Large uncertainties on D_{Ra} and D_{Ba} values for different volcanic system
- Little effect of $D_{\text{Ra}} \neq D_{\text{Ba}}$ for minerals where $Ra \sim 0$ and $D_{\text{Ra}} \ll D_{\text{Th}}$ (e.g. mt, px)

- 
- Whole rocks isochrons are unlikely to be obtained due to the complexity of the processes generating U-series fractionation
 - The significance of mineral ages must be thoroughly evaluated considering that they could reflect
 - ✓ Different population of crystals or core-rim age variation
 - ✓ Older xenoliths and possible inclusion of U- and Th- rich minerals
 - ✓ Prolonged crystallisation history (discordant U-Th vs. Ra-Th ages)
 - Ra-Th ages require knowledge of D_{Ra} and D_{Ba}
 - Mineral isochrons provide information of crystallisation ages, but not necessarily on residence times in magma chambers, which are often open rather than closed systems.

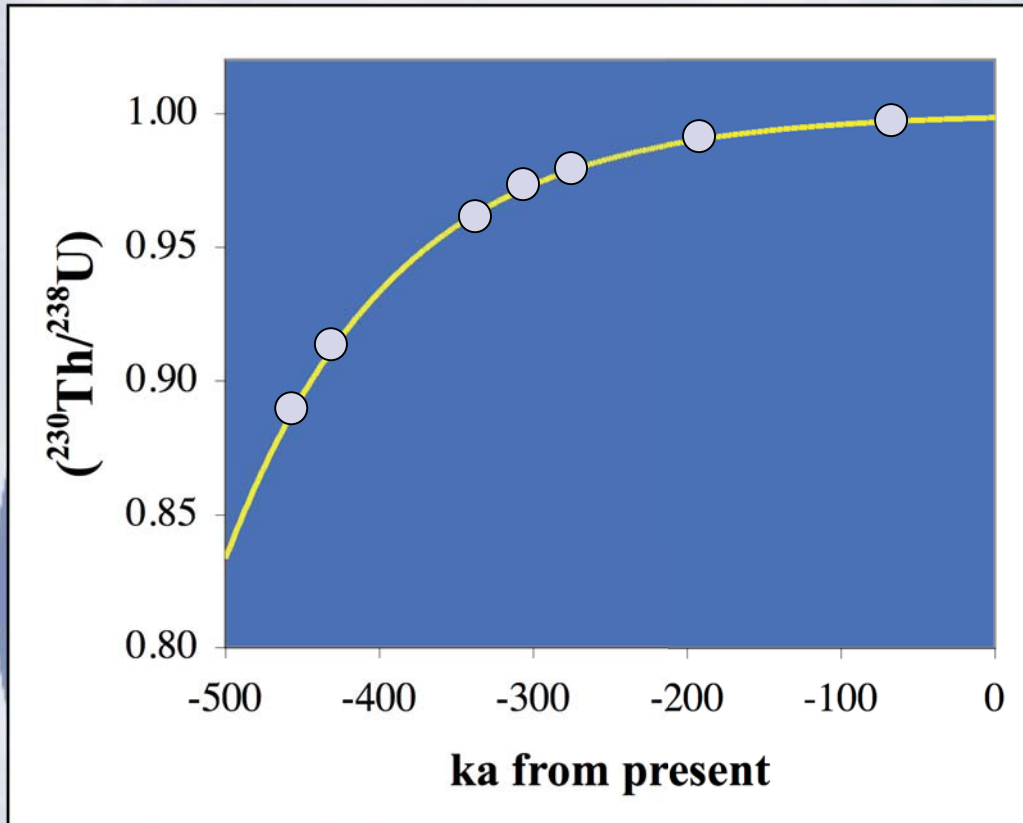
A large, semi-transparent image of the Earth's globe is positioned on the left side of the slide, showing the continents of North and South America. The globe is set against a light blue background with a faint grid pattern.

**Evolution of U-series disequilibria through time
(e.g. eruption history of a volcano)**

Closed vs. Open systems

- **Effect of episodic or continuous replenishment**

3.1. Closed system evolution



Fresh magma entering the magma chamber 500ka ago with (²³⁰Th/²³⁸U) = 0.83

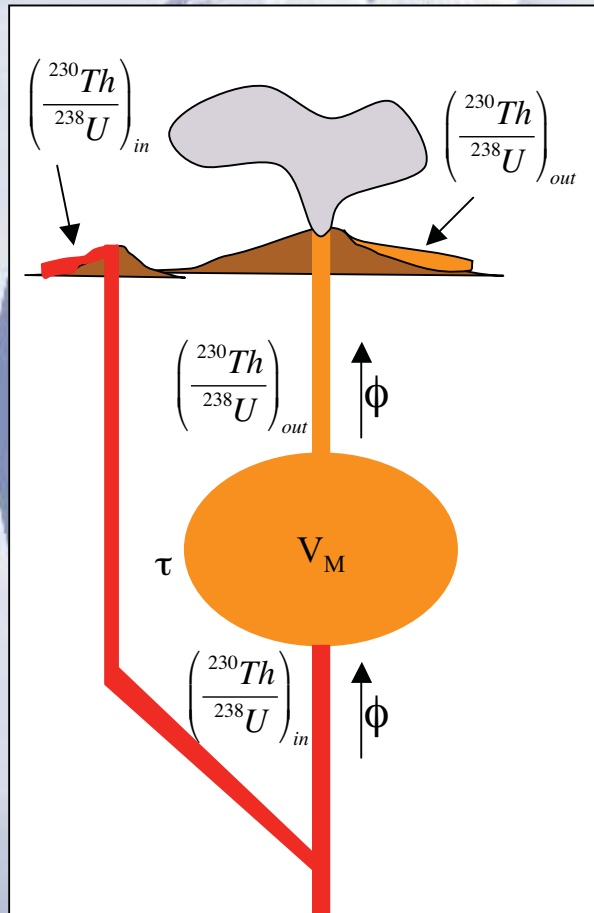
$$\left(\frac{{}^{230}\text{Th}}{{}^{238}\text{U}}\right)_t = \left(1 - e^{-\lambda_{230\text{Th}} \cdot t}\right) + \left(\frac{{}^{230}\text{Th}}{{}^{238}\text{U}}\right)_o \cdot e^{-\lambda_{230\text{Th}} \cdot t}$$

Similar evolution for (²³⁰Th/²³²Th), (²³⁸U/²³²Th) remains constant

The activity ratio increases until secular equilibrium is restored and can be tracked through time in erupted products

3.2. Open systems

Periodically or continuously replenished magma chambers:

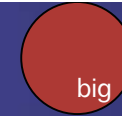


- Volume magma chamber (V_M) [*assumed constant*]
- Input volume (V_i) = output volume (V_o)
- Interval between inputs (t)
- Flux: ϕ (V_i/t)
- Residence time (turnover time): $\tau = V_M / \phi$

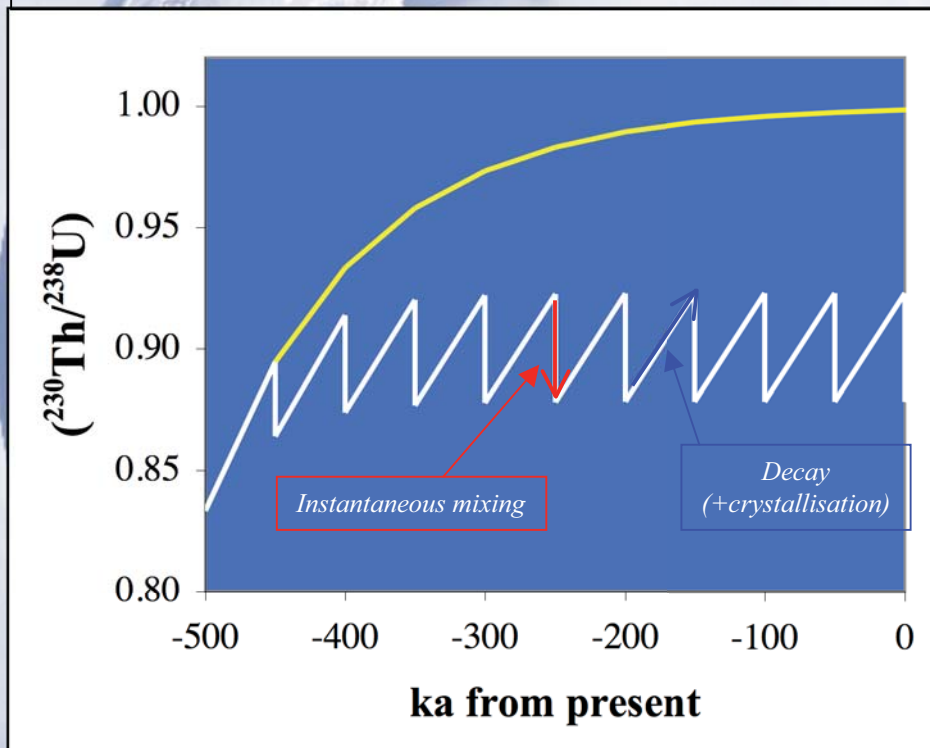
The value of the activity ratios $\left(\frac{^{230}\text{Th}}{^{238}\text{U}}\right)_{out}$ can be modeled as:

First input \rightarrow Decay {+fractional crystallisation} \rightarrow new input MIXING \rightarrow decay etc.

3.2. Open systems: discrete replenishment inputs



Vol magma chamber: $V_M = 100\text{km}^3$
Replenishment: $V_i = 50\text{km}^3$
Interval between inputs **$t = 50\text{ka}$** (i.e. comparable with $t_{1/2}$ of ^{230}Th)
Flux (average): $\phi (V_i/t) = 0.001 \text{ Km}^3/\text{yr}$
Residence time (turnover time): $\tau = V_M / \phi = 100\text{ka}$



Replenishment model from Hughes & Hawkesworth 1999 GCA 63, 4101-4110. No fractional crystallisation considered

Episodic input of fresh magma with **constant** $(^{230}\text{Th}/^{238}\text{U})_{\text{in}}$

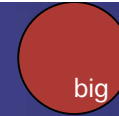
Instantaneous mixing between new input and magma in the chamber

Alternate phases of:

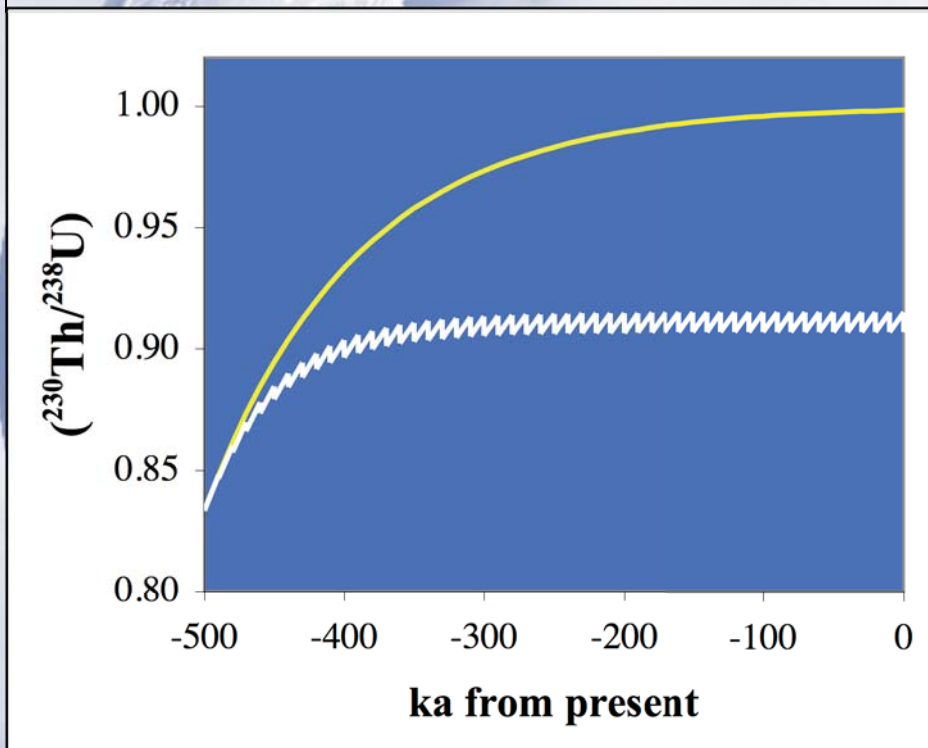
- decay {+fractional crystallisation}
- mixing

Disequilibrium can be maintained for more than 350ka

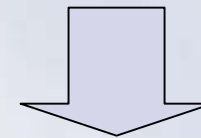
3.2. Open systems: discrete replenishment inputs



Vol magma chamber: $V_M = 100\text{km}^3$
Replenishment: $V_i = 10\text{km}^3$
Interval between inputs **$t = 10\text{ka}$** (i.e. $t < t_{1/2}$ of ^{230}Th)
Flux (average): $\phi (V_i/t) = 0.001 \text{ Km}^3/\text{yr}$
Residence time (turnover time): $\tau = V_M / \phi = 100\text{ka}$

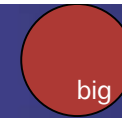


- Constant flux but **shorter interval** between the fresh inputs (i.e. also smaller V_{input})

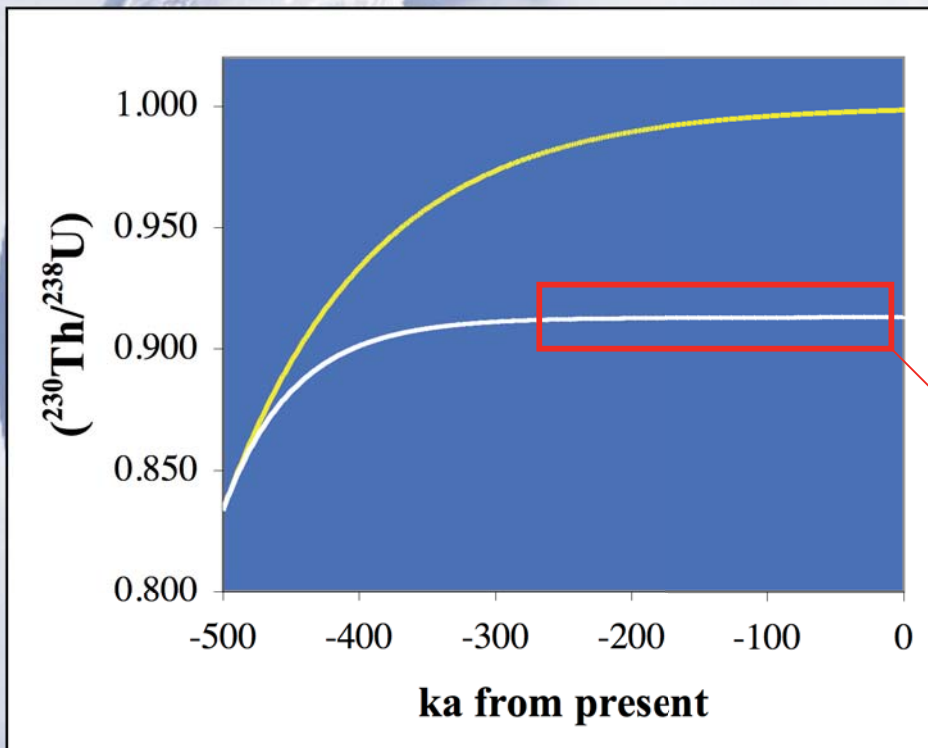


Similar behavior but **smaller fluctuation of $(^{230}\text{Th}/^{238}\text{U})$**

3.2. Open systems: continuous replenishment



Vol magma chamber: $V_M = 100\text{km}^3$
Replenishment: $V_i = 1\text{km}^3$
Interval between inputs **$t = 1\text{ka}$** (i.e. $t \ll t_{1/2}$ of ^{230}Th)
Flux (average): $\phi (V_i/t) = 0.001 \text{ Km}^3/\text{yr}$
Residence time (turnover time): $\tau = V_M / \phi = 100\text{ka}$



Smaller interval between input \rightarrow (i.e. $t \ll t_{1/2}$ of ^{230}Th)

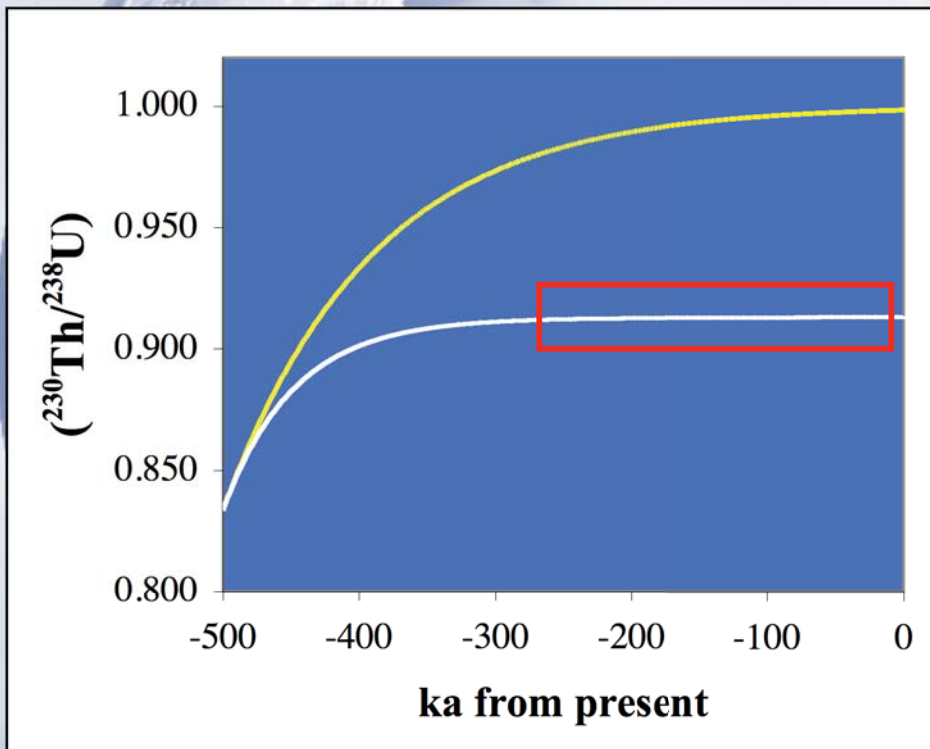
Replenishment can be considered continuous:

STEADY STATE ACTIVITY RATIO

$(^{230}\text{Th}/^{238}\text{U})_{s.s.} = \text{constant}$

3.2. Open systems: continuous replenishment

Vol magma chamber: $V_M = 100\text{km}^3$
Replenishment: $V_i = 1\text{km}^3$
Interval between inputs **$t = 1\text{ka}$**
Flux (average): $\phi (V_i/t) = 0.001 \text{ Km}^3/\text{yr}$
Residence time (turnover time): $\tau = V_M / \phi = 100\text{ka}$



$(^{230}\text{Th}/^{238}\text{U})_{s.s}$ depends only on two unknown:

1. $(^{230}\text{Th}/^{238}\text{U})_{in}$
2. $\phi/V_M = \text{Renewal rate } (\phi_R) = 1/\tau$

When both steady state and input activity ratio of a volcanic system are known, we can calculate:

- Residence time $\tau = 1 / \phi_R$

- Volume of the magma chamber

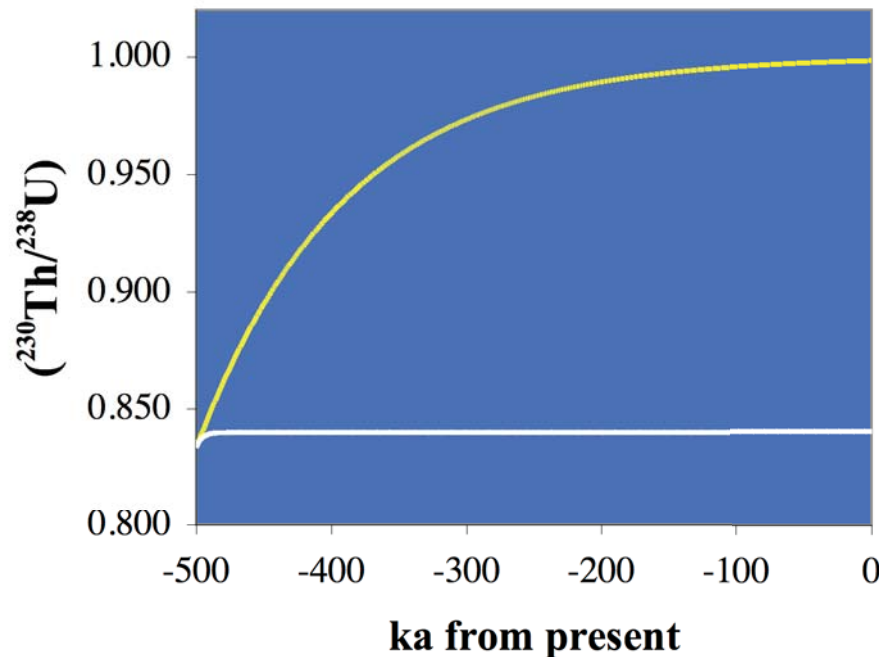
$$V_M = \phi / \phi_R = \phi \cdot \tau$$

(ϕ estimated from eruption volumes over time)

3.2. Open systems: continuous replenishment



Vol magma chamber: $V_M = 100 \text{ km}^3$
Replenishment: $V_i = 20 \text{ km}^3$
Interval between inputs $t = 1 \text{ ka}$
Flux (average): $\phi (V_i/t) = 0.02 \text{ km}^3/\text{yr}$
Residence time (turnover time): $\tau = V_M / \phi = 5 \text{ ka}$



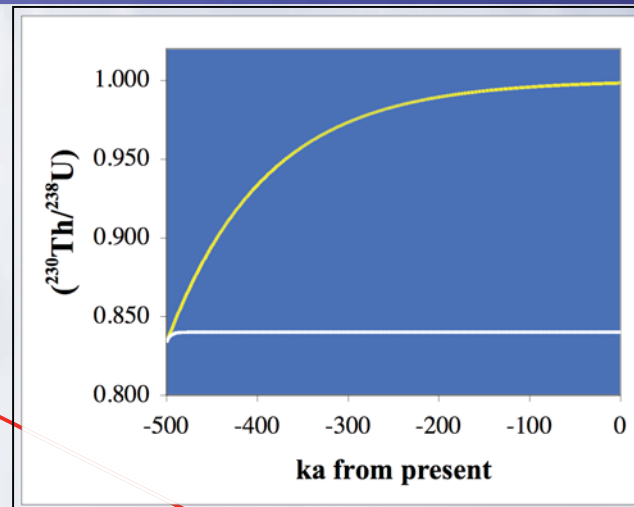
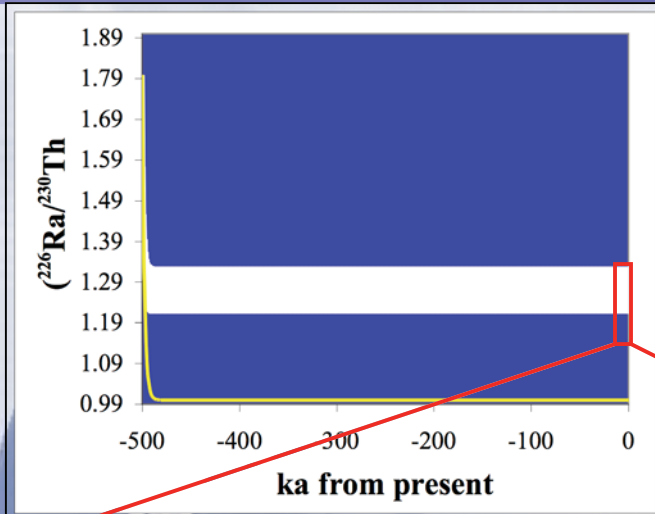
Changing V_{input} (i.e. changing input flux)

Increased flux of magma:

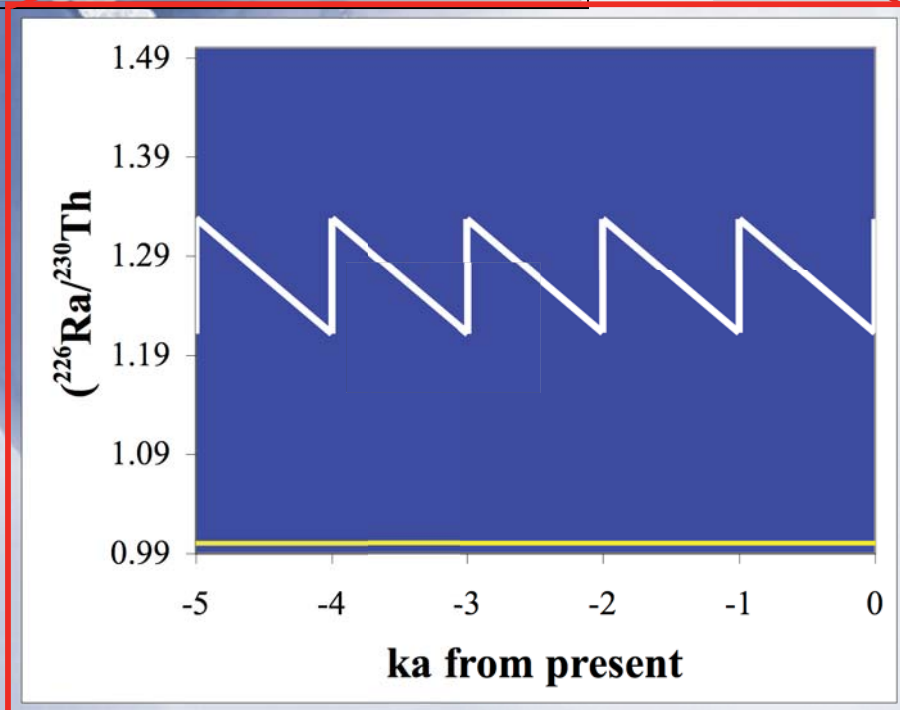
- $(^{230}\text{Th}/^{238}\text{U})_{\text{ss}}$ closer to input value
- shorter residence time
- high renewal rate: $\tau \ll \lambda$

$$\rightarrow (^{230}\text{Th}/^{238}\text{U})_{\text{ss}} \sim (^{230}\text{Th}/^{238}\text{U})_{\text{in}}$$

3.2. Changing isotopic system



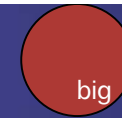
$V_M = 100\text{km}^3$
 $V_i = 20\text{ km}^3$
 $t = 1\text{ka}$
 $\phi (V_i/t) = 0.02\text{km}^3/\text{yr}$
 $\tau = V_M / \phi = 5\text{ka}$



- The interval between inputs is significant for Ra-Th but not for Th-U system: different parent-daughter pairs may behave differently
- Steady state can be reached for shorter intervals between inputs

Similar evolution for $(^{226}\text{Ra})/\text{Ba}$

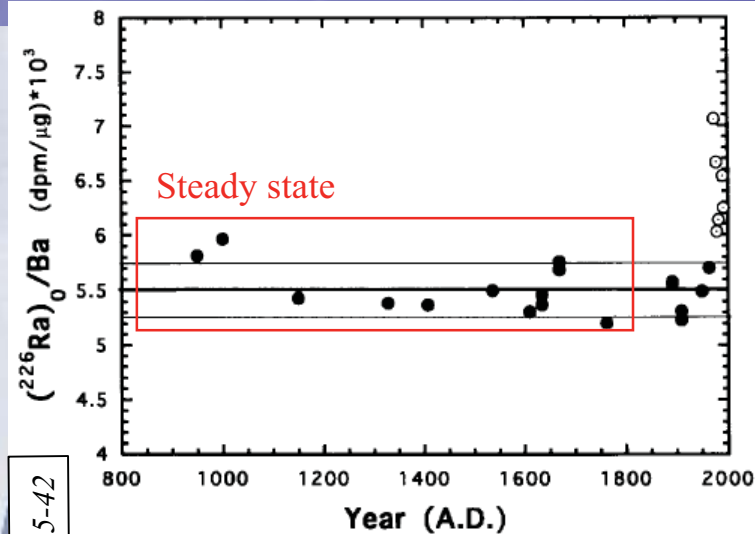
3.3. Examples from natural systems



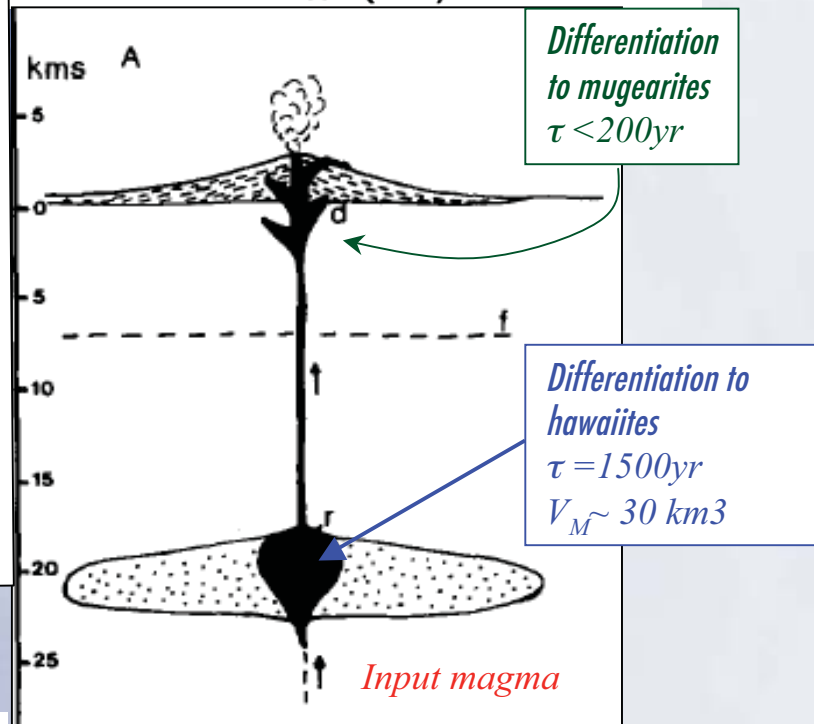
Real volcanoes are much more complicated than the modeled one;

- Fractional crystallisation likely affects the activity ratios
 - ✓ use of $(^{230}\text{Th}/^{232}\text{Th})$ and $(^{226}\text{Ra})/\text{Ba}$ to avoid this effect or addition of crystallisation to the model
- Magma input might not be constant through time: activity ratios variation due to source processes
 - ✓ Combined use of different isotopic systematic permit to distinguish source from magma chamber processes
- Mixing is not an instantaneous process (magmas erupted over a short period relative to $t_{1/2}$)
 - ✓ can be modeled in relationship with the renewal rate and the activity ratio of the input magma (e.g., Condomines et al 1982, GCA 46, 1397-1416)
- Zoned magma chamber: the new input of fresh magma mixes only with a minor part of the melt in the chamber.

3.3. Examples from natural systems



From Condomines et al. 1995, EPSL 132, 25-42



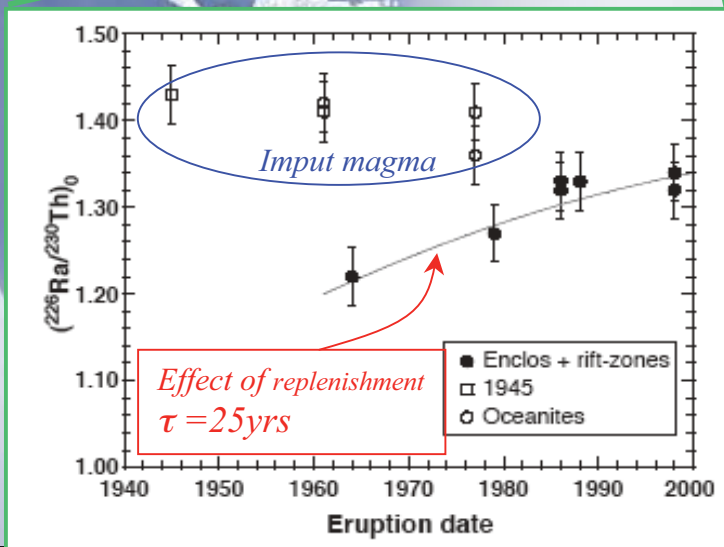
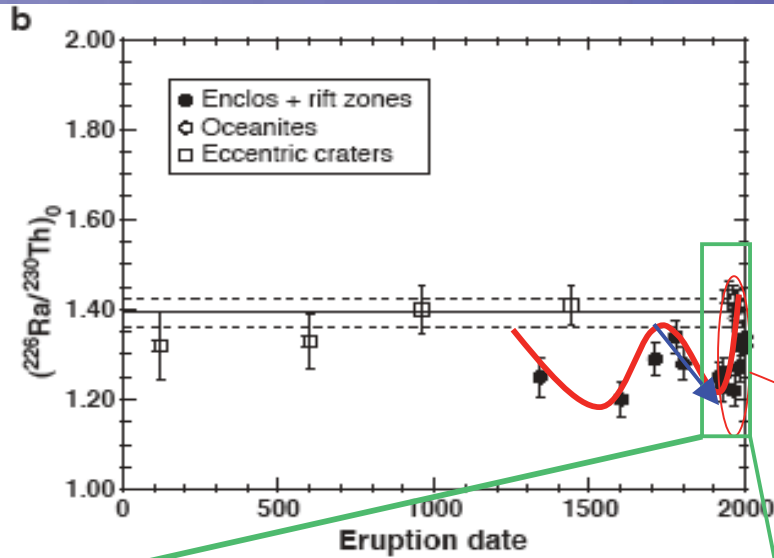
1. Etna (Condomines et al, 1995 EPSL 132, 25-41)

Continuously replenished magma chamber

Steady state activity in the past 1000 yrs (before 1970) from $(^{226}\text{Ra})/\text{Ba}$

- Constant $(^{226}\text{Ra}/\text{Ba})$: no difference between hawaiites and mugearites
- No Ra-Ba fractionation during differentiation
- Constant input isotopic composition: small variation in $(^{230}\text{Th}/^{238}\text{U})$, $(^{230}\text{Th}/^{232}\text{Th})$ and $^{87}\text{Sr}/^{86}\text{Sr}$
- $(^{226}\text{Ra}/^{230}\text{Th})_{\text{in}}$ from eccentric eruption

3.3. Examples from natural systems



From Sigmarsson et al. 2005, EPSL 234, 223-234

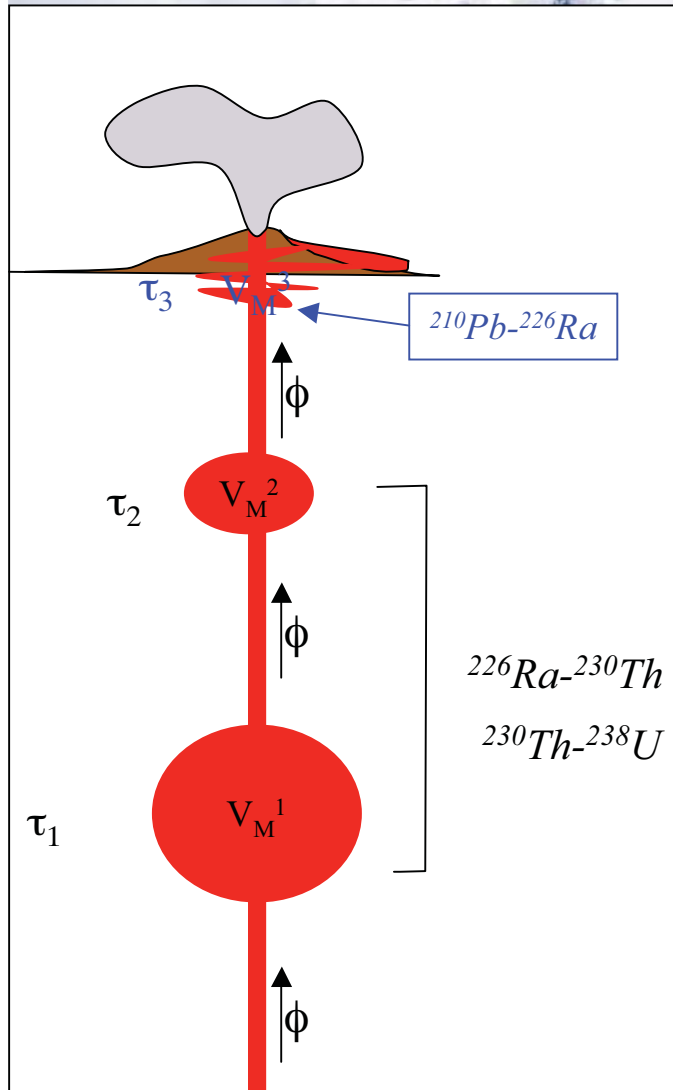
2. Piton de la Fournaise (Sigmarsson et al, 2005, EPSL 234, 223-234)

Periodically replenished magma chamber.

- variable $(^{226}\text{Ra}/^{230}\text{Th})_0$: alternating episodes of closed system and replenishment
- constant input isotopic composition: small variation in $(^{230}\text{Th}/^{238}\text{U})$, $(^{230}\text{Th}/^{232}\text{Th})$
- $(^{226}\text{Ra}/^{230}\text{Th})_{\text{in}}$ from eccentric eruption

- Closed system periods:
Difference between input magmas and lowest erupted $(^{226}\text{Ra}/^{230}\text{Th})$
→ storage time $\sim 1600\text{yr}$ (maximum value)

- Replenishment periods:
Increasing $(^{226}\text{Ra}/^{230}\text{Th})$ in a 40yr time span: mixing not instantaneous
effect of decay can be neglected, activity ratio depends on the input and chamber composition, and residence time (Condomines et al., 1982)
→ $\tau = 25\text{yrs}$
→ $V_M = 0.35 \text{ km}^3$ (from estimated output rate)



- Magmatic system might behave as closed, periodically or continuously (steady state) replenished system with respect to different U-series pairs: e.g. $^{230}\text{Th}-^{238}\text{U}$, $^{226}\text{Ra}-^{230}\text{Th}$, $^{210}\text{Pb}-^{226}\text{Ra}$

- The evolution of U-series isotopes through time in such systems can be modeled providing important information

- ✓ storage and differentiation times (closed system)

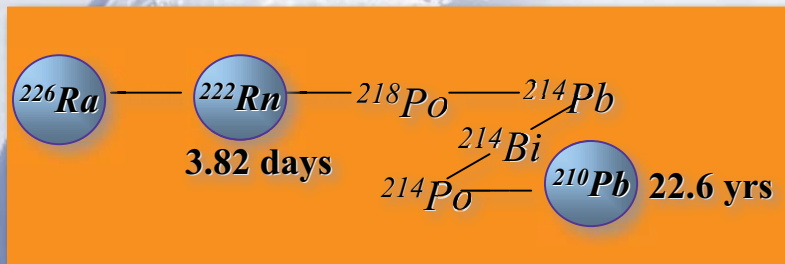
- ✓ residence times and magma chambers volumes (open system)

- Different U-series parent-daughter pairs provide information on the timescales of different processes operating in different times and at different depth in the magmatic systems.

4. ^{210}Pb - ^{226}Ra disequilibria

Magma degassing in shallow reservoir

Recent fractionation events < 100yrs à Shallow magmatic processes

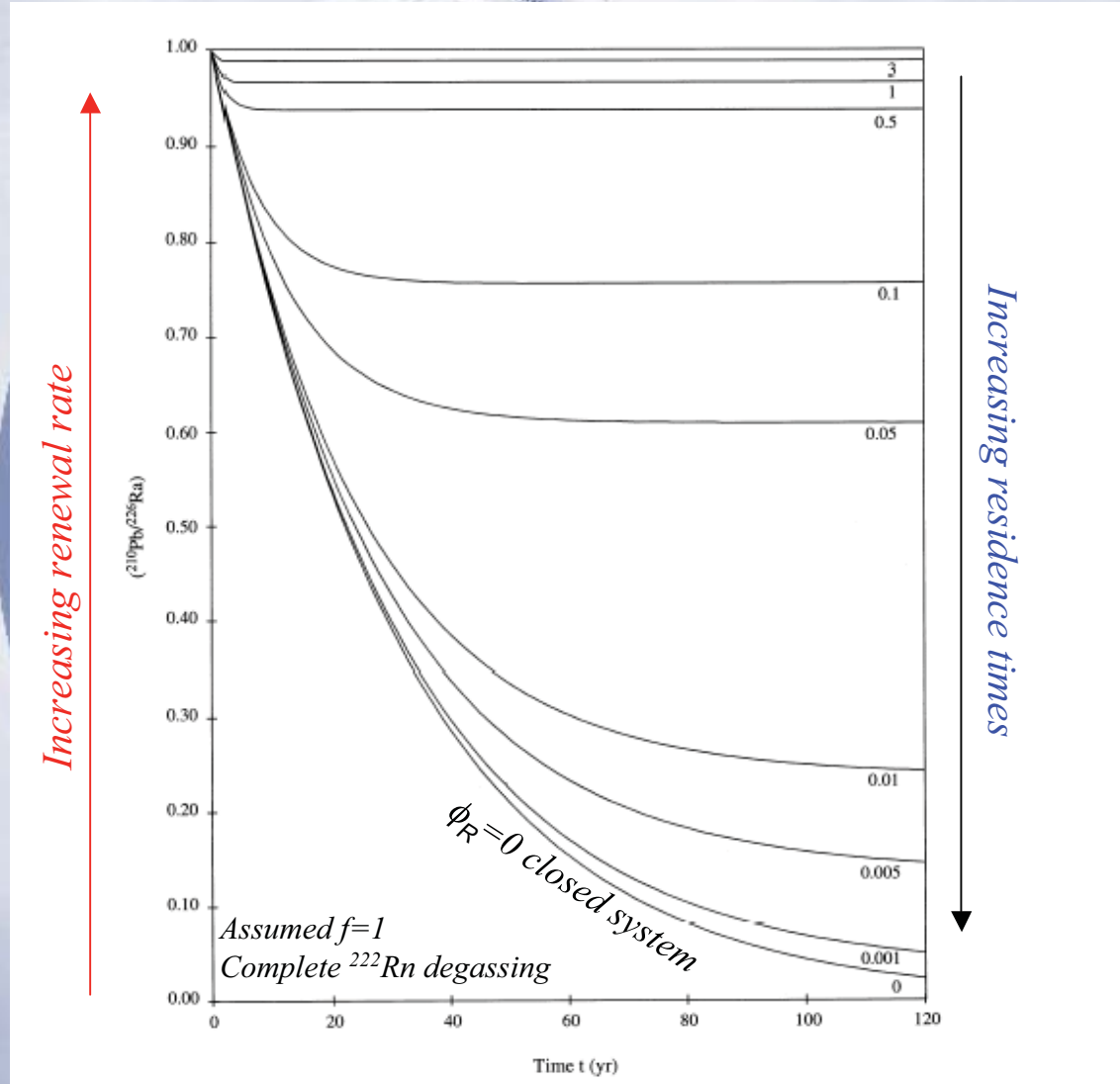


Nuclides between ^{222}Rn and ^{210}Pb have $t_{1/2} < 1\text{h}$ and can be neglected

- ^{222}Rn is highly volatile and is continuously degassed in the shallower reservoirs (open conduit) and its concentration in the magma can be modeled.
- $(^{210}\text{Pb}/^{226}\text{Ra}) < 1$ in the erupted degassed lavas can be related to degassing of ^{222}Rn (Gauthier & Condomines, 1999) and depends on:
 - renewal rate ϕ/M (ϕ_R) = $1/\tau$ [unknown]
 - duration of degassing t [unknown]
 - loss of ^{222}Rn during degassing: f [can be estimated by monitoring ^{222}Rn in-growth in freshly erupted lavas]
 - initial volatile content: α [can be determined from melt inclusions contents]

4. ^{210}Pb - ^{226}Ra disequilibria

From Gauthier & Condomines, 1999 *EPSL* 172, 111-126

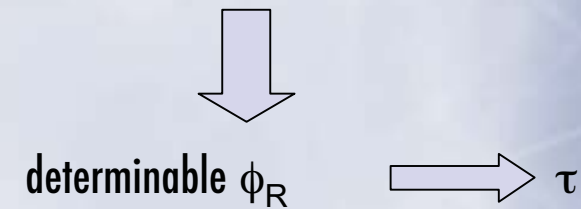


Closed system (but open conduit)

- $\phi_R = 0$
- ^{222}Rn rapidly lost ($^{210}\text{Pb}/^{226}\text{Ra}$) $\rightarrow 0$

Continuously replenished system

STEADY STATE ($^{210}\text{Pb}/^{226}\text{Ra}$)



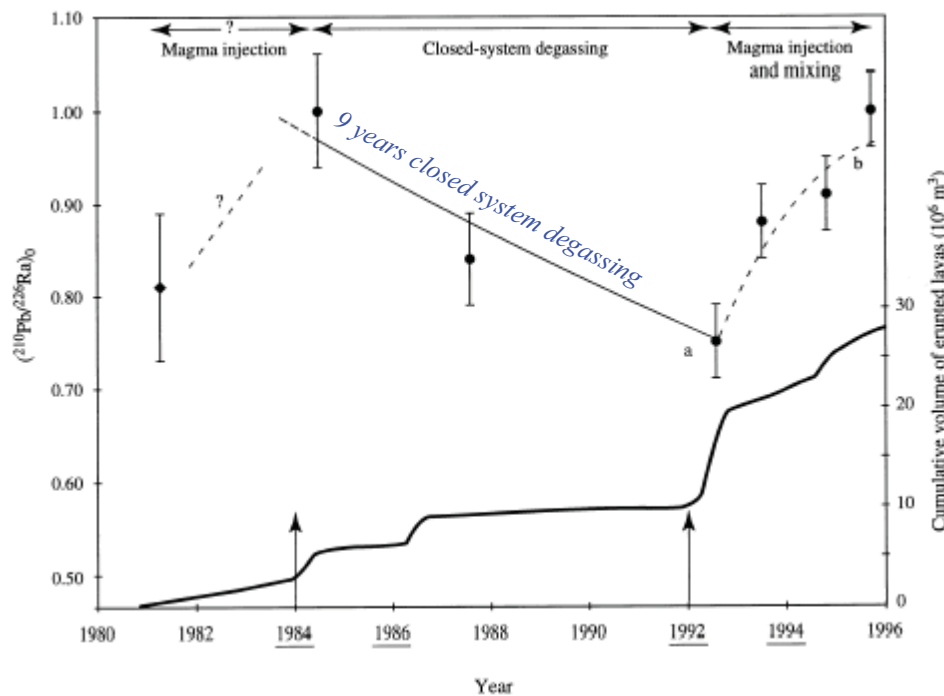
4. ^{210}Pb - ^{226}Ra disequilibria

Natural systems

1. **Stromboli** (Gauthier & Condomines, 1999 EPSL 172, 111-126)

- Constant ($^{210}\text{Pb}/^{226}\text{Ra}$) ~ 0.96 - 0.99 : STEADY STATE
- $f = 0.9$ - 1.0 (estimated)
- Residence time in the shallow reservoir $\tau = 110$ - 520 days

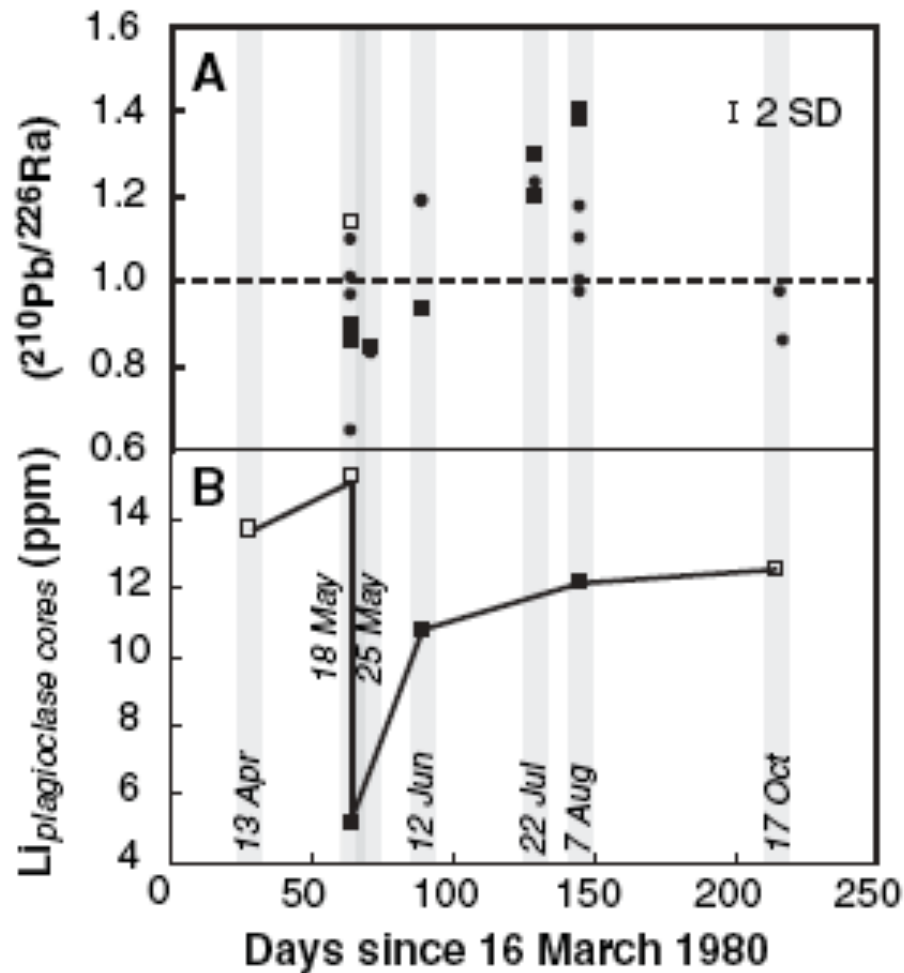
From Gauthier & Condomines, 1999 EPSL 172, 111-126



2. **Merapi** (Gauthier & Condomines, 1999, EPSL 172, 111-126)

- variable ($^{210}\text{Pb}/^{226}\text{Ra}$): alternate episodes of closed system and replenishment
- $f = 1.0$ (estimated)
- Both episodes can be modeled:
 - 1984-92: closed system degassing (9 years)
 - 1992-1996: re-injection
 - $\tau \sim 2$ yrs
 - $V_M \sim 1.6 \cdot 10^7 \text{ m}^3$

4. ^{210}Pb - ^{226}Ra disequilibria



Berlo et al., (2005) Science 306, 1167-1171:

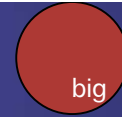
Mount S. Helens

$(^{210}\text{Pb}/^{226}\text{Ra}) > 1$

Capping of the system: shallow accumulation of volatiles (^{222}Rn decaying to ^{210}Pb) exsolved from lower magma bodies:

^{210}Pb - ^{226}Ra as precursor to explosive activity?

From Berlo et al. 2005, Science 309, 1167-1169



Thanks...