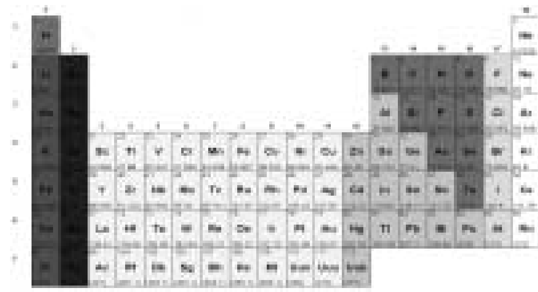
Isotopes

Reading:
Winter, Chapter 9, pp. 167-180

Isotopes

Same Z, different A (variable # of neutrons)

General notation for a nuclide: $^{14}_{6}C$



Isotopes

Have the same Z but different A (variable # of neutrons)

General notation for a nuclide: $^{14}_{6}C$

Superscript (n) specifies the different isotopes of an element

^{12}C ^{13}C ^{14}C

Stable Isotopes

- Stable means they last nearly forever
- Chemical fractionation is impossible
- Mass fractionation is the only isotopic discrimination possible

Oxygen Isotopes

^{16}O	99.756% of natural oxygen	
^{17}O	0.039%	“
^{18}O	0.205%	“

Concentrations expressed by reference to a standard

The international standard for O isotopes is:

standard mean ocean water (SMOW)

^{18}O and ^{16}O are the commonly used isotopes and their ratio is expressed as δ:

$\delta\,(^{18}O/^{16}O) =$

$$\frac{(^{18}O/^{16}O)_{sample} - (^{18}O/^{16}O)_{SMOW}}{(^{18}O/^{16}O)_{SMOW}} \times 1000$$

the result is expressed in per mille (‰)

What is δ of SMOW?

What is δ for meteoric water?

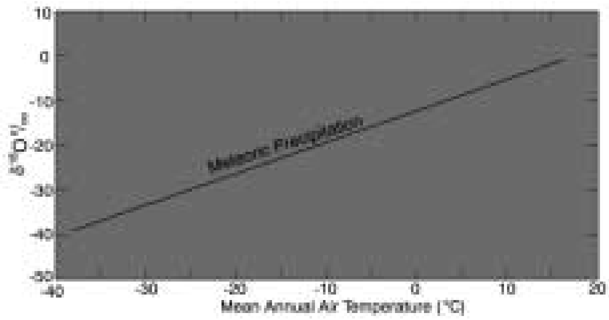
Evaporation seawater produces water vapor

– Light isotopes are enriched in vapor > liquid
– This is efficient since Δ mass = 1/8 total mass

$$\delta = \frac{(^{18}O/^{16}O)_{vapor} - (^{18}O/^{16}O)_{SMOW}}{(^{18}O/^{16}O)_{SMOW}} \times 1000$$

therefore $(^{18}O/^{16}O)_{Vapor} < (^{18}O/^{16}O)_{SMOW}$

thus δ_{clouds} is (-)



Relationship between d($^{18}O/^{16}O$) and mean annual temperature for meteoric precipitation, after Dansgaard (1964). *Tellus*, **16**, 436-468.

Stable isotopes are useful in assessing relative contribution of various reservoirs, each with a distinctive isotopic signature

- O and H isotopes
  - Juvenile vs. meteoric vs. brine water
- $\delta^{18}O$ for mantle rocks ≠ surface-reworked sediments
  - Evaluate contamination of mantle-derived magmas by crustal sediments

Radioactive Isotopes

- Unstable isotopes decay to other nuclides
- The rate of decay is constant, and not affected by P, T, X…
- Parent nuclide is the radioactive nuclide that decays
- Daughter nuclide(s) are the radiogenic *atomic* products

Isotopic variations between rocks, etc. due to:

1. Mass fractionation (as for stable isotopes)
2. Daughters produced in varying proportions resulting from previous event of chemical fractionation

$^{40}K \rightarrow {}^{40}Ar$ by radioactive decay

Basalt → rhyolite by FX (a *chemical* fractionation process)

Rhyolite has more K than basalt

$^{40}K \rightarrow$ more ^{40}Ar over time in rhyolite than in basalt

$^{40}Ar/^{39}Ar$ ratio will be different in each

Isotopic variations between rocks, etc. due to:

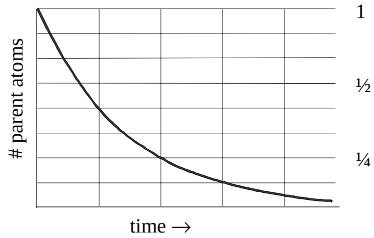
1. Mass fractionation (as for stable isotopes)
2. Daughters produced in varying proportions resulting from previous event of chemical fractionation
3. Time

The longer $^{40}K \rightarrow {}^{40}Ar$ decay takes place, the greater the difference between the basalt and rhyolite will be

Radioactive Decay

The Law of Radioactive Decay

$$-\frac{dN}{dt} \propto N \quad \text{or} \quad -\frac{dN}{dt} = \lambda N$$



$$D = Ne^{\lambda t} - N = N(e^{\lambda t} - 1)$$

To calculate the age of a sample (t) if we know:

D the amount of the daughter nuclide produced

N the amount of the original parent nuclide remaining

λ the decay constant for the system in question

The K-Ar System

^{40}K decays to either ^{40}Ca or ^{40}Ar

– ^{40}Ca is common. Cannot distinguish radiogenic ^{40}Ca from non-radiogenic ^{40}Ca

– ^{40}Ar is an inert gas which can be trapped in many solid phases as it forms in them

The appropriate decay equation is:

eq 9-16 $\quad {}^{40}Ar = {}^{40}Ar_o + \left(\frac{\lambda_e}{\lambda}\right) {}^{40}K(e^{-\lambda t} - 1)$

Where $\lambda_e = 0.581 \times 10^{-10}\ a^{-1}$ (proton capture)
and $\lambda = 5.543 \times 10^{-10}\ a^{-1}$ (whole process)

Blocking temperatures for various minerals differ

^{40}Ar-^{39}Ar technique grew from this discovery

The Sr-Rb System

On the rubidium side of the equation:

$^{87}Rb \rightarrow {}^{87}Sr$ + a beta particle ($\lambda = 1.42 \times 10^{-11}\ a^{-1}$)

Rb behaves like K, it concentrates in micas and alkali feldspar

Strontium Side

- Sr behaves like Ca
  - It concentrates in plagioclase and apatite (but not in clinopyroxene)
- $^{88}Sr : ^{87}Sr : ^{86}Sr : ^{84}Sr$
  - Average sample yields 10 : 0.7 : 1 : 0.07
- ^{86}Sr is a stable isotope
  - It is not created by breakdown of any other parent

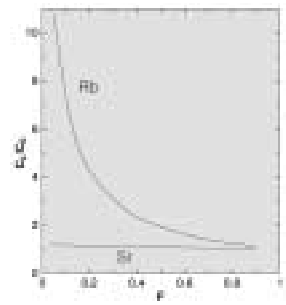
Isochron Technique

Requires 3 or more cogenetic samples with a range of Rb/Sr concentrations

3 cogenetic rocks may be derived from a single source by partial melting, FX, etc.

or

3 coexisting minerals with different K/Ca ratios in a single rock



Recast age equation by dividing through by stable ^{86}Sr

$$^{87}Sr/^{86}Sr = (^{87}Sr/^{86}Sr)_o + (^{87}Rb/^{86}Sr)(e^{\lambda t} - 1)$$

$\lambda = 1.4 \times 10^{-11}\ a^{-1}$

For values of λt less than 0.1: $e^{\lambda t} - 1 \cong \lambda t$

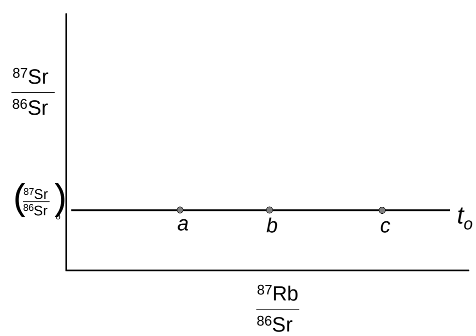
Thus eq. 9-15 for t < 70 Ga (!!) reduces to:

$$^{87}Sr/^{86}Sr = (^{87}Sr/^{86}Sr)_o + (^{87}Rb/^{86}Sr)\lambda t$$

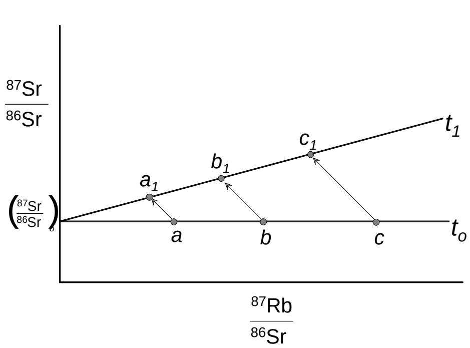
y = b + x m

=This is the equation for a line in a $^{87}Sr/^{86}Sr$ vs. $^{87}Rb/^{86}Sr$ plot

Begin with 3 rocks plotting at a b c at time t_o

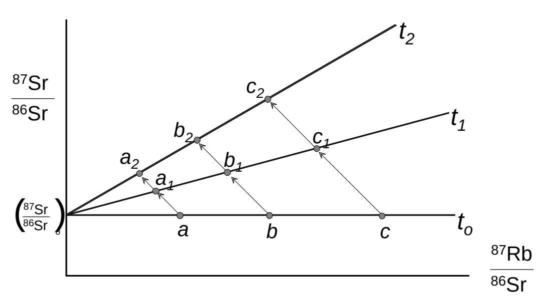


After some time increment ($t_0 \rightarrow t_1$) each sample loses some ^{87}Rb and gains an equivalent amount of ^{87}Sr



At time t_2 each rock system has evolved $\rightarrow$ new line

Again still linear and steeper line



Isochron technique produces 2 valuable things:

1. The age of the rocks (from the slope = λt)
2. $(^{87}Sr/^{86}Sr)_o$ = the initial value of $^{87}Sr/^{86}Sr$

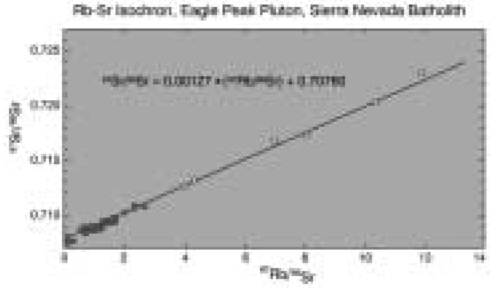


Figure 9-9. Rb-Sr isochron for the Eagle Peak Pluton, central Sierra Nevada Batholith, California, USA. Filled circles are whole-rock analyses, open circles are hornblende separates. The regression equation for the data is also given. After Hill et al. (1988). Amer. J. Sci., 288-A, 213-241.

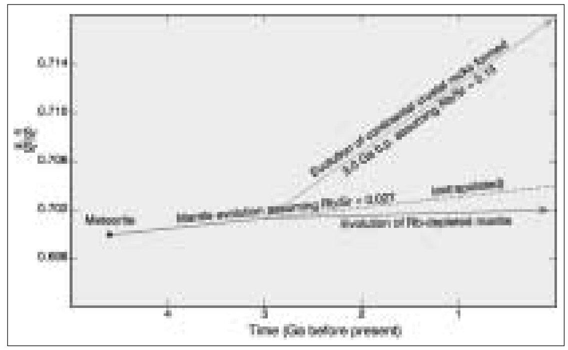


Figure 9-13. Estimated Rb and Sr isotopic evolution of the Earth's upper mantle, assuming a large-scale melting event producing granitic-type continental rocks at 3.0 Ga b.p After Wilson (1989). Igneous Petrogenesis. Unwin Hyman/Kluwer.