Due Wed. October 8 in class.

Radioactive Decay and Geochronology

Useful information for the calculations is at the end of the problem set, including a portion of the Chart of the Nuclides, and information on how to interpret isochron diagrams. Please use a spreadsheet. Make sure it's clear how you did the calculations.

1. Radioactive Decay: Nuclides (10 pts)

The Lanthanide or Rare Earth elements are those between La and Yb on the Periodic Table. Using the Chart of the Nuclides (provided):

a. List the Rare Earth elements in order of increasing Z.
b. Identifiy the naturally occurring radioactive isotopes.
c. What are their half-lives?
d. What fraction of the total element do they comprise?
e. What are their decay modes?
f. What are the decay products?

2. Radioactive Decay: Heat Production (45 pts)

This probem has a lot of parts, but it takes you through a calculation step by step. It should give some insight into heat generation within the earth by radioactive decay.

a. ^{235}U decays to ^{207}Pb. Calculate the half-life of ^{235}U from its decay constant.

b. How many half-lives has ^{235}U experienced since the Earth formed at 4.56 Ga?

c. ^{235}U decays to ^{207}Pb in a series of steps, through intermediate radioactive nuclides, via alpha decay. The *net reaction* can be written as follows:

$^{235}U \rightarrow {}^{207}Pb + n\,\alpha$

where n is the number of *alpha*-particles. What is n?

d. Using the equation for the net reaction, how much mass is converted to energy in the decay of one atom of ^{235}U to ^{207}Pb?

e. How much energy is released in the decay of one ^{235}U atom to ^{207}Pb, in MeV?

f. At 4 Ga, 8.18×10^{37} atoms of ^{235}U decay per million years. How much heat is released in calories in this radioactive decay (per million years)?

g. If all that heat was added the oceans at once, how many degrees would the oceans heat up?

h. Do you think this ever happened? If so, how would this impact the oceans? If not, where did the heat go, considering that U resides in the earth's crust.

3. Geochronology (45 pts)

Referring back to the Relative Time cartoon exercise in the lab, it was noted that geologists collected samples for radiometric dating. Sample were taken from the body labeled 'V', and the $^{206}Pb/^{238}U$ ratio of the zircon was found to be 0.03151±0.0001. Also, one was taken from the dike labeled "W" and the $^{206}Pb/^{238}U$ ratio was found to be 0.003107±0.000015. It can be assumed that there was no Pb in the zircon when it formed- it all came from radioactive decay of U.

For the metamorphosed volcanic rocks in layers A-G, geologists took four rock samples from these layers (A, B, D, G), and made measurements of the Rb and Sr concentrations and Sr isotope compositions. For two of the samples (B and G) they also separated minerals and measured their Rb and Sr concentrations and Sr isotope compositions. The Rb-Sr data are reported in the table below.

a. Plot the data on a standard isochron plot ($^{87}Rb/^{86}Sr$ on the X-axis, and $^{87}Sr/^{86}Sr$ on the Y-axis). You should use excel for this part of the exercise, and all three isochrons should be plotted on the same graph.

b. Use the data to calculate (1) a whole rock isochron age (this is based on the slope of the line) and initial $^{87}Sr/^{86}Sr$ (this is the y-intercept) for this set of rocks , (2) a mineral isochron and initial $^{87}Sr/^{86}Sr$ for rock B and (3) a mineral isochron and initial $^{87}Sr/^{86}Sr$ for rock G. The decay constant of ^{87}Rb is given in Useful Info.

c. Use the $^{206}Pb/^{238}U$ information provided in the description to calculate ages and their analytical uncertainties for the two dikes.

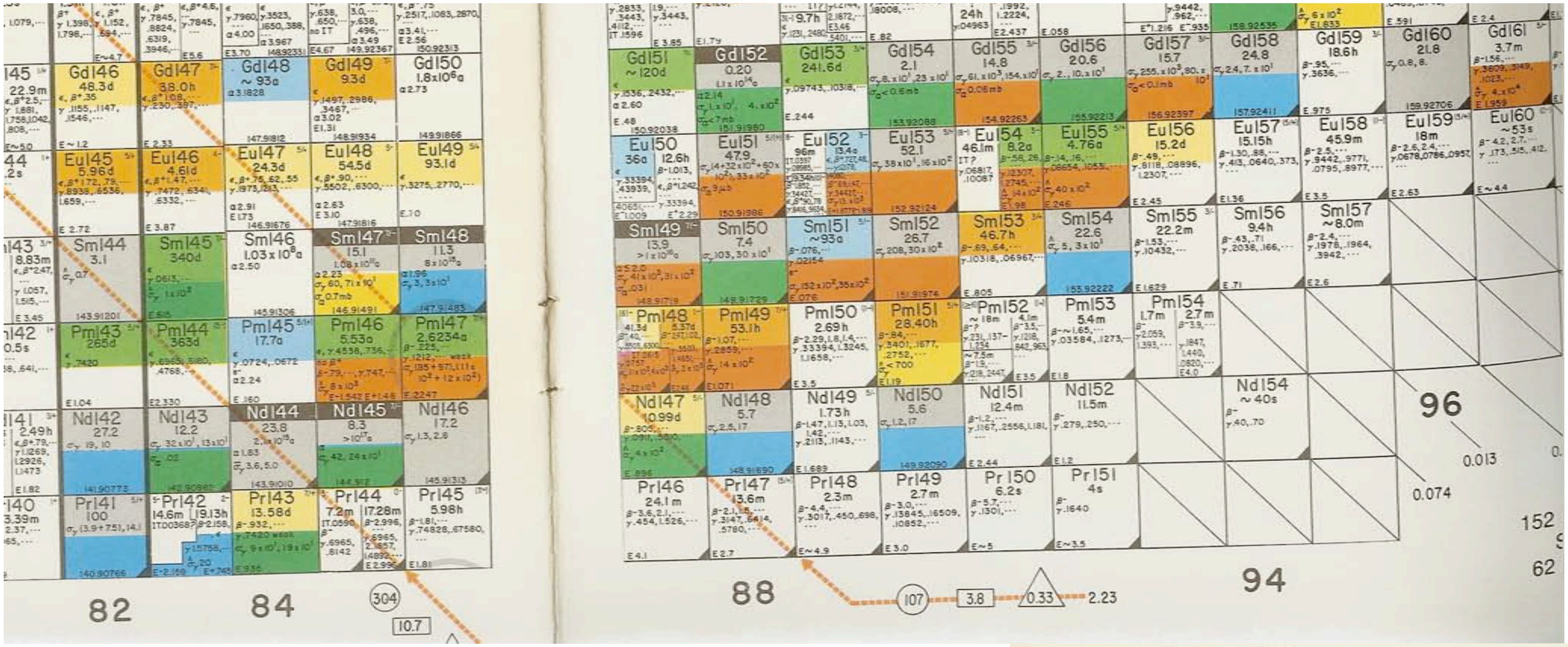
4. Write a one paragraph essay on how the calculated isotope ages constrain the geological events that you derived from the figure. You may want to consult the attached notes on the interpretation of isochron diagrams during igneous and metamorphic events.

Useful Info:

^{235}U = 235.04392525
^{238}U = 238.05078578 AMU
^{207}Pb= 206.975885 AMU
^{206}Pb = 205.974455 AMU
^{4}He = 4.0026035 AMU
Atomic wt. U = 238.029 g/mole
$^{238}U/^{235}U$ (atomic) = 137.88 present day.
Avogadro's number: 6.02×10^{23} atoms/mole
Mass of silicate earth: 4.03×10^{27} g
Mass of world ocean: 1.40×10^{24} g
$\lambda 235 = 9.8485 \times 10^{-10}$ y-1
$\lambda 238 = 1.55125 \times 10^{-10}\ y^{-1}$
1 AMU = 931.5 MeV
1 eV = 1.60207×10^{-19} joules
1 MeV = 10^{6} eV
1 cal = 4.1855 joules
$\lambda 87Rb = 1.42 \times 10^{-11}\ y^{-1}$

Table. Rb-Sr isotope data from layers A-G.

	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Whole rock data		
rock A	0.25	0.710202
rock B	0.30	0.711642
rock D	0.50	0.717404
rock G	1.00	0.731807
Mineral separate data		
rock B		
apatite	0.05	0.710931
K-feldspar	0.60	1.712495
muscovite	5.00	0.725009
rock G		
apatite	0.07	0.729162
K-feldspar	1.30	0.732660
muscovite	15.0	0.771624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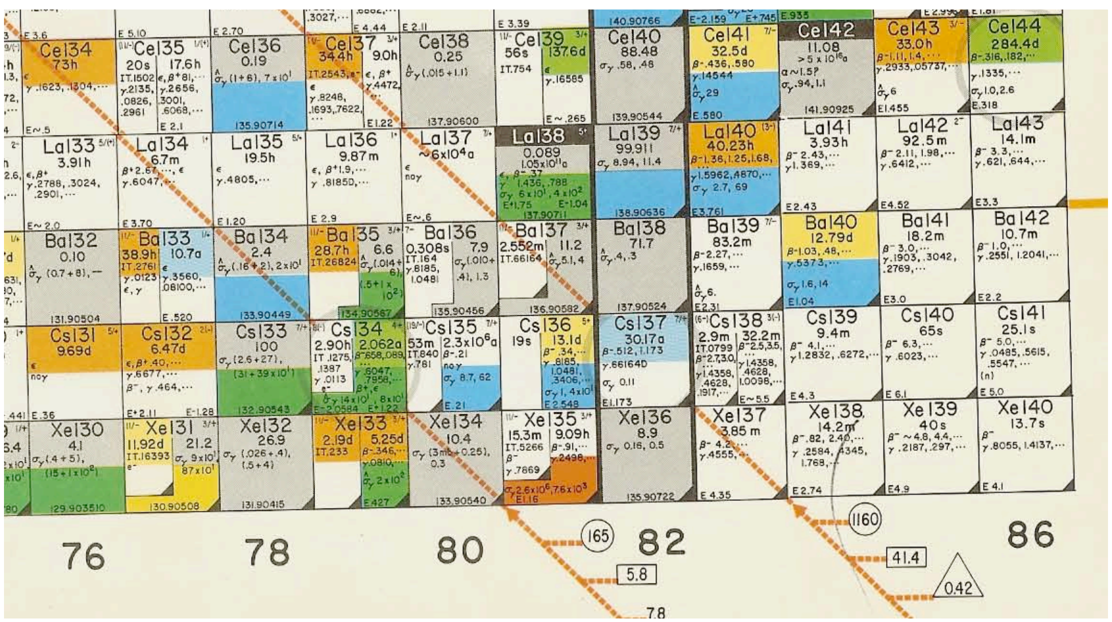
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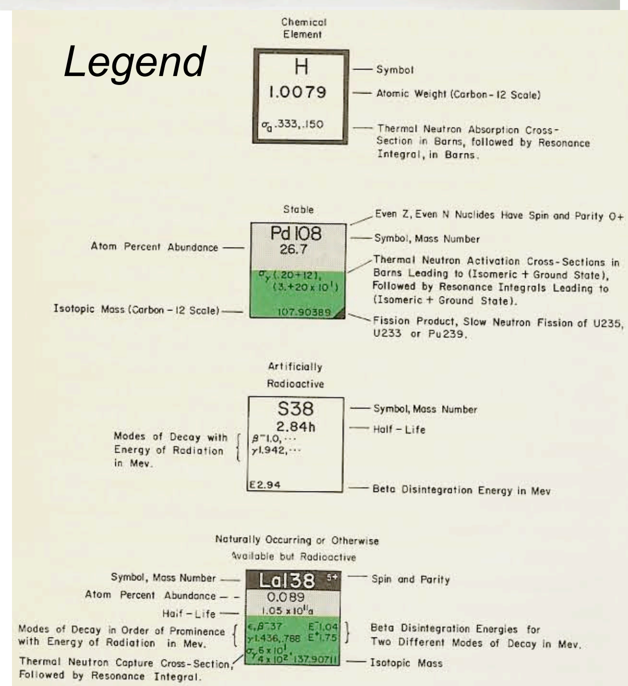
Legend
Chemical Element
H
1.0079
Symbol
Atomic Weight (Carbon-12 Scale)
Thermal Neutron Absorption Cross-Section in Barns, followed by Resonance Integral, in Barns.
Stable
Pd 108
26.7
107.90389
Even Z, Even N Nuclides Have Spin and Parity 0+
Symbol, Mass Number
Atom Percent Abundance
Thermal Neutron Activation Cross-Sections in Barns Leading to (Isomeric + Ground State), Followed by Resonance Integrals Leading to (Isomeric + Ground State).
Isotopic Mass (Carbon-12 Scale)
Fission Product, Slow Neutron Fission of U235, U233 or Pu239.
Artificially Radioactive
S38
2.84h
E2.94
Symbol, Mass Number
Half-Life
Modes of Decay with Energy of Radiation in Mev.
Beta Disintegration Energy in Mev
Naturally Occurring or Otherwise Available but Radioactive
La138
0.089
137.90711
Symbol, Mass Number
Atom Percent Abundance
Half-Life
Modes of Decay in Order of Prominence with Energy of Radiation in Mev.
Thermal Neutron Capture Cross-Section, Followed by Resonance Integral.
Spin and Parity
Beta Disintegration Energies for Two Different Modes of Decay in Mev.
Isotopic Mass

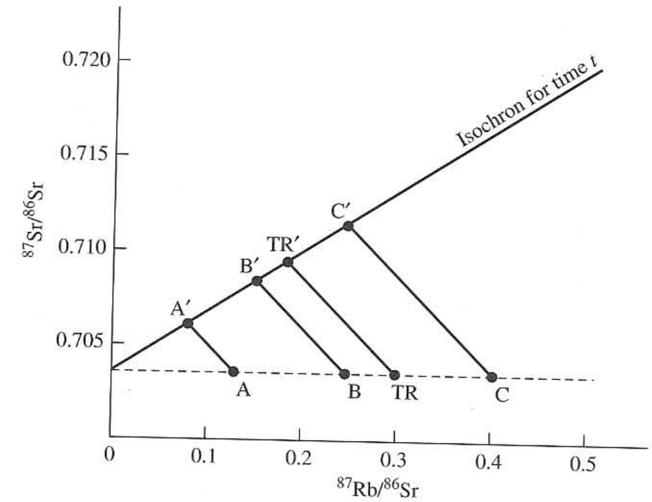

Figure 2-1 A simple isochron for an igneous rock that has not been subsequently altered. The inital composition of three different minerals of the rock is given by A, B, and C; their composition after time t is given by A′, B′, and C′. The zero intercept of the isochron gives the initial $^{87}Sr/^{86}Sr$ for the rock, and the age of the rock (t) can be calculated from the slope of the isochron, which equals $e^{\lambda t} - 1$. TR represents the bulk rock. The $^{87}Sr/^{86}Sr$ value for the bulk rock is determined by the $^{87}Sr/^{86}Sr$ values and relative proportions of the minerals in the rock. (Modified from Faul, *Ages of Rocks, Planets and Stars*, p. 34. Copyright 1966 by McGraw-Hill Book Company. Used with permission of McGraw-Hill Book Company.) After Lanphere et al. (1964).

constant, the radiometric age found for a sample may also be in error owing to gain or loss of rubidium and strontium after mineral or rock crystallization. Two types of alteration are known to affect the rubidium-strontium content of minerals: (1) a temperature effect due to metamorphism (which causes diffusion of strontium), and (2) chemical exchange with circulating water. Metamorphism can cause extensive and unequal strontium redistribution among the minerals of a rock or between different rock types. The result will be, for age measurements, a variety of "ages," none of which is a measure of a particular event (*discordant ages*).

In some cases it is possible to determine for a metamorphosed igneous rock both the time of metamorphism and the time of original crystallization. Consider an igneous rock mass that has undergone metamorphism, resulting in a redistribution of strontium among its minerals (Figure 2-2). If individual portions of the rock mass remain a closed system during metamorphism, the isochrons found for *rock* samples will still give the primary age and initial strontium composition. However, if the minerals of each rock sample have a new, homogeneous distribution of ^{87}Sr and ^{86}Sr as a result of metamorphism, their "clocks are reset" and mineral analysis will give, for a particular rock, an isochron with different slope and larger intercept than that found for rock samples. The mineral isochrons represent the time of metamorphism and the rock isochron the time of initial rock formation. This approach is not very useful for rocks that have undergone partial homogenization of isotopic ratios or that have been metamorphosed several times (see Table 2-6). Scattering of plotted points for rock samples on an isochron diagram indicates that the samples have not been a closed system.

The initial $^{87}Sr/^{86}Sr$ ratios found for oceanic and continental rocks have been used to interpret the history of the Earth's crust and to define the source region of igneous rocks. To do this,

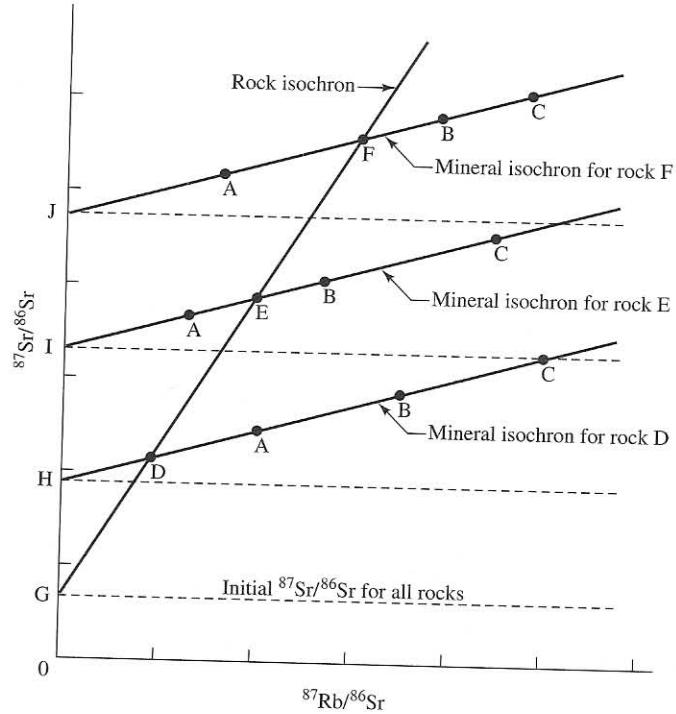

Figure 2-2 Isochron diagram for three related igneous rocks of similar age that have undergone subsequent metamorphism. The rocks D, E, and F define an isochron representing the age of the rocks. The minerals A, B, and C in each rock define individual isochrons with parallel slopes determined by the time of a later metamorphism. At the time of metamorphism the rocks have differing ratios of $^{87}Sr/^{86}Sr$ (because of differing original Rb/Sr ratios). G is the initial $^{87}Sr/^{86}Sr$ for all rocks; H, I, and J are the $^{87}Sr/^{86}Sr$ ratios for rocks D, E, and F at the end of the metamorphism. (Modified from H. Faul, 1966, *Ages of Rocks, Planets and Stars*, McGraw-Hill, New York, p. 37. Copyright 1966 by McGraw-Hill Book Company. Used with permission of McGraw-Hill Book Company.) After Lanphere et al. (1964).

assumptions have to be made about the primeval $^{87}Sr/^{86}Sr$ ratio and about the extent of fractionation of ^{87}Rb between crust and mantle (Figure 2-3). Because of the geochemical character of rubidium (see Chapter One), the crust has a higher Rb/Sr value than the mantle. Thus a magma from the mantle that forms new igneous rock in the crust will have a different initial $^{87}Sr/^{86}Sr$ ratio than will a magma formed in the crust. This ratio will depend on the original ratio in the source area and on the time of melting. Although igneous rocks have a restricted range of variation for their initial $^{87}Sr/^{86}Sr$ ratio, it is possible to characterize individual bodies by their ratio. For example, oceanic basalts have an initial $^{87}Sr/^{86}Sr$ ratio of 0.702 to 0.707. This has been interpreted as indicating a source in the mantle. The wider variation in ratios found for continental basalts is believed to be due to contamination of their mantle-derived magmas by crustal material. Most magmas formed entirely in the crust should produce rocks with a higher (greater than about 0.707) initial $^{87}Sr/^{86}Sr$ ratio. This type of approach has been used to study the question of whether the various continents have grown continuously throughout geologic time or if, instead, an initial crust has been periodically regenerated. Either process would explain the decrease in ages found toward continental margins.