



Indian Institute of Technology, Kanpur

Department of Earth Sciences

ESO213A: Fundamentals of Earth Sciences

Lecture 24. History of the Earth

Santanu Misra

Course Instructor

smisra@iitk.ac.in • <http://home.iitk.ac.in/~smisra/>



Aims of this lecture

- Methods of Studying the Past
- Reconstruction of history from stratigraphic records

Reading:

Marshak's Book (Part-IV)
Grotzinger & Jordan's book (Chapter 8)
[for the entire week]

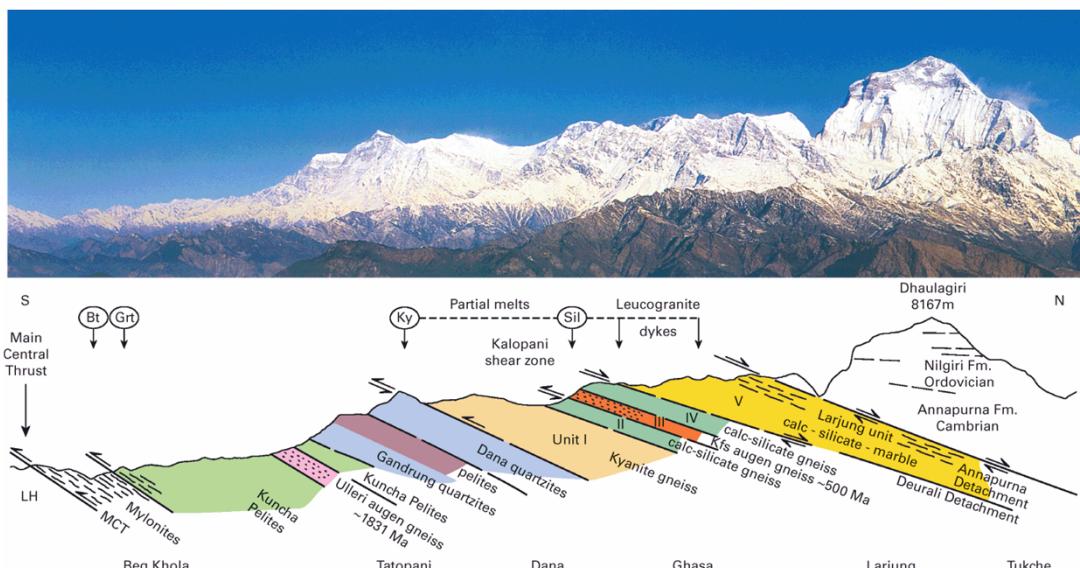
Earth has a History



- Geologic materials record enormous changes.
- Earth is a complex, evolving system.

- Physical and biological systems continuously interact.
- Earth constantly changes and has done so through time.
 - Species arise, flourish, and disappear forever.
 - Continents rift, drift, and collide.
 - Ocean basins open and close.
 - Sea level rises and falls.

Earth has a History



Searle, 2017

Methods of studying the past



- Historic Earth changes are measured by...

- Orogenic events.
- Sea level.
- Climates.
- Living organisms.
- Continental positions.
- Plate boundaries.
- Chemistry.
 - Atmosphere.
 - Ocean.
- Depositional environments.

Challenges:

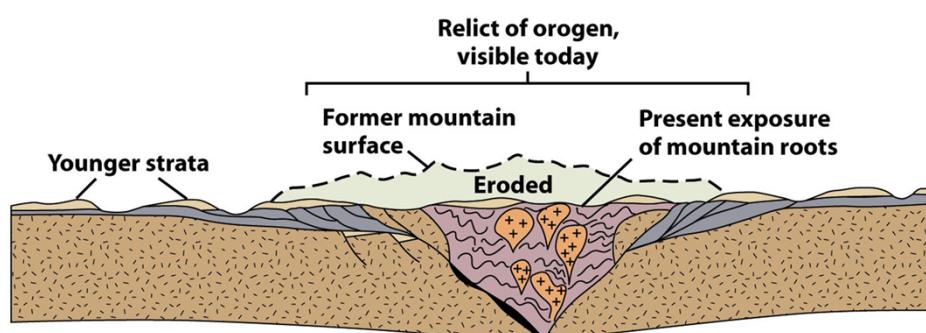
- Discontinuous record
 - Erosion
 - Younger rocks are better preserved; older rocks less
- There are still abundance of evidences**

- These changes are recorded in rocks.

Ancient Orogens



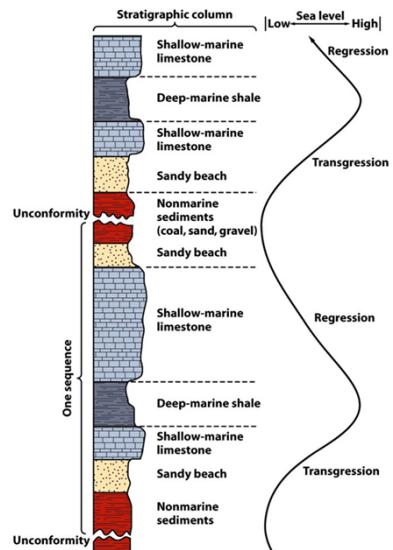
- Ancient orogens – Former mountain belts.
 - Igneous activity, deformation, and metamorphism.
 - Thick sedimentary deposits filling foreland basins.
 - Ancient orogenic belts expose deeply buried rocks.



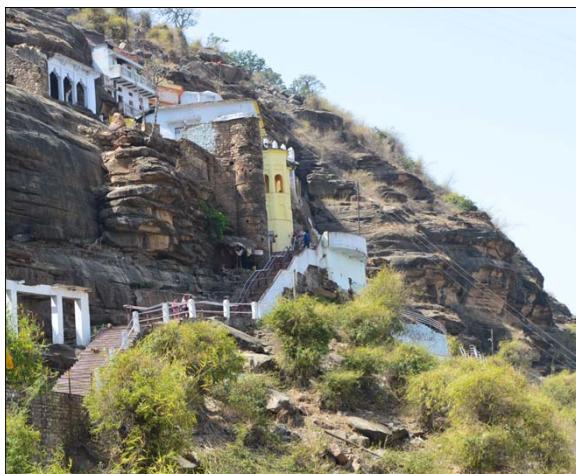
Depositional Environments



- Recognizing depositional environments.
 - Successions of strata record changes in depositional settings.
- Recognizing sea level changes.
 - Sediments record sea level flux.
 - Shallow and deep environments create distinctive sediments.



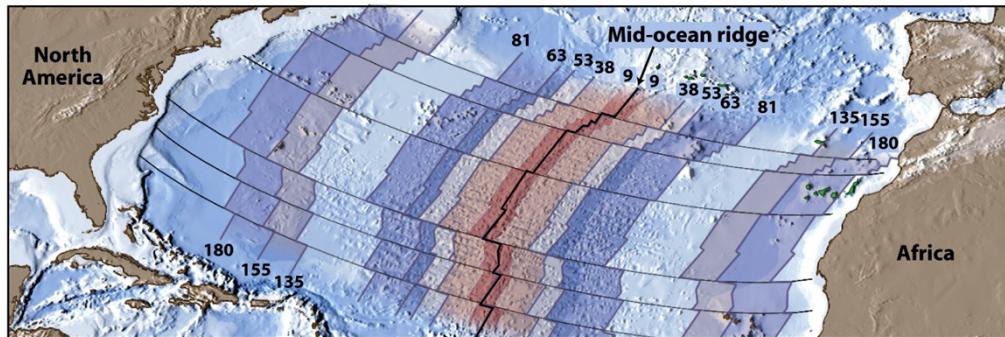
Depositional Environments



Changing Continental Positions



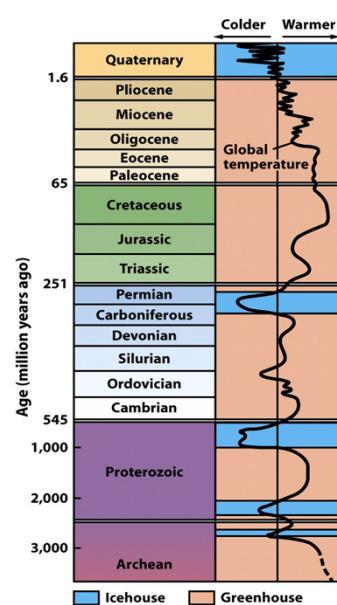
- Changing continental positions are preserved.
 - Paleomagnetism captures paleolatitude.
 - Ocean width changes by reversing sea-floor anomalies.
 - Rock and fossil distributions compare across oceans.



Palaeoclimate



- Paleoclimates – Rocks preserve ancient climates.
 - Tropical – Extensive coral reefs.
 - Subtropical – Extensive deserts.
 - Polar – Extensive glacial deposits.
- Climatic belts expand and contract.
 - Greenhouse Earth.
 - Snowball Earth.
- $^{18}\text{O}/^{16}\text{O}$ isotopic ratios preserve ancient temperatures

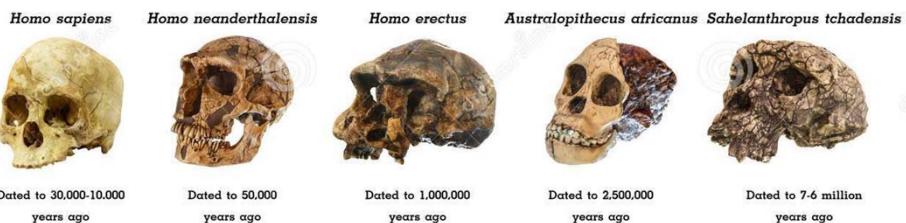


Evolution of Life



- Evolution – Fossils preserve changes in Earth's life.
 - Sedimentary rocks preserve fossil ecosystems.
 - Organisms inhabiting Earth have obviously changed.
 - Over geologic time, most species have exhibited both...
 - Trends toward specialization.
 - Catastrophic extinctions.

EVOLUTION OF HUMAN



Evolution of Life



Archaeopteryx fossil



20,000-Year-Old Human Footprints

History from Stratigraphic Records



Stratigraphy is a branch of geology concerned with the study of rock layers (strata) and layering (stratification). It is primarily used in the study of sedimentary and layered volcanic rocks.

Original horizontality: all rock layers were originally horizontal (*Nicolaus Steno*).

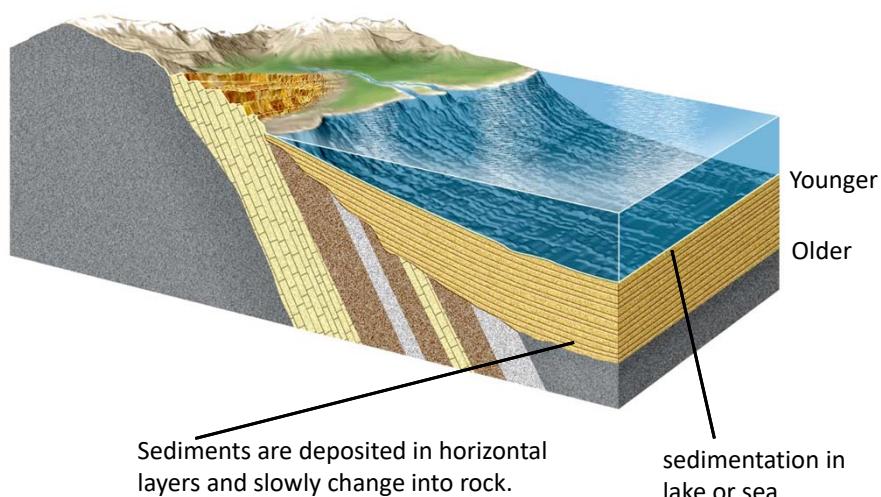
Superposition: in a sequence of sedimentary rock layers, each layer of rock is older than the layer above it and younger than the rock layer below it (*Nicolaus Steno*).

Faunal succession: fossils occur in a definite, invariable sequence in the geologic record (*William Smith*).

Cross-cutting relationship: if a fault or other body of rock cuts through another body of rock then it must be younger in age than the rock through which it cuts and displaces (*James Hutton*).

Laws of inclusions: if a rock body contained fragments of another rock body, it must be younger than the fragments of rock it contained. The intruding rock must have been there first to provide the fragments. (*James Hutton*).

Original Horizontality and Superposition



Faunal Succession



Outcrop A



Outcrop B



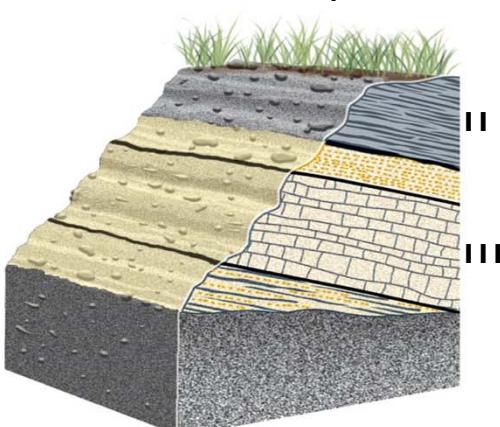
Faunal Succession



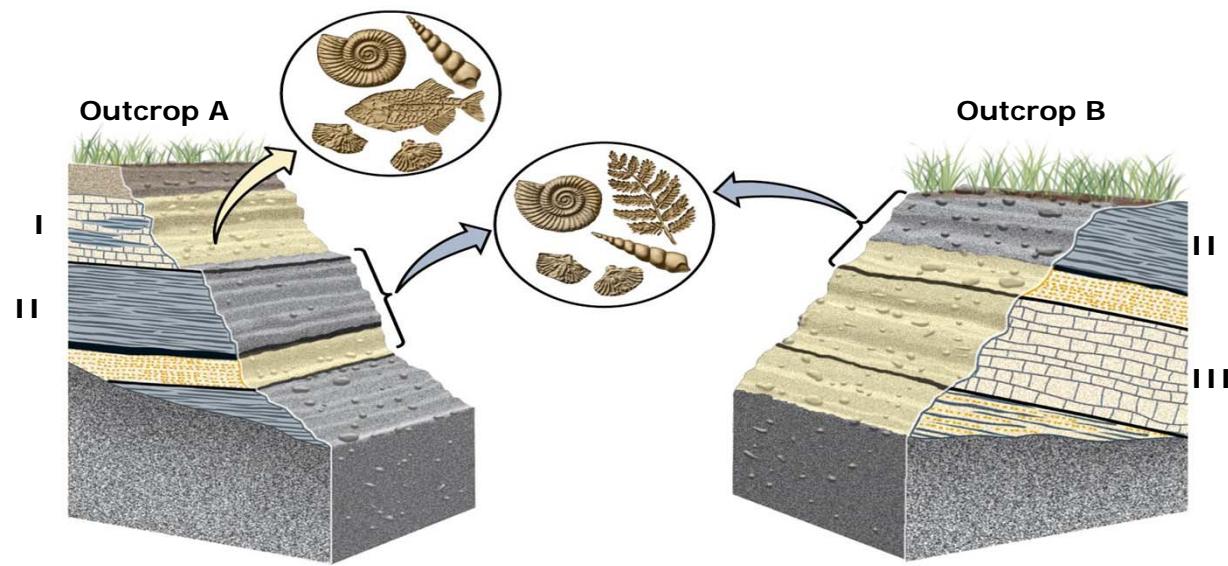
Outcrop A



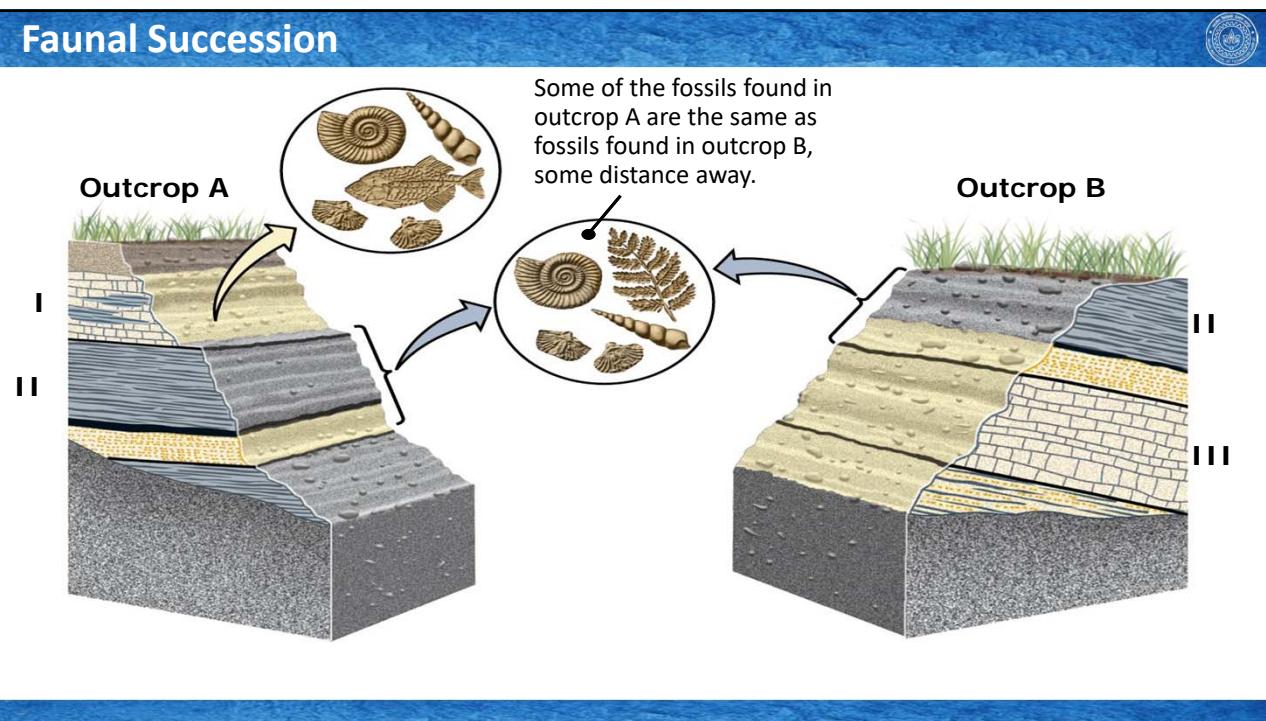
Outcrop B



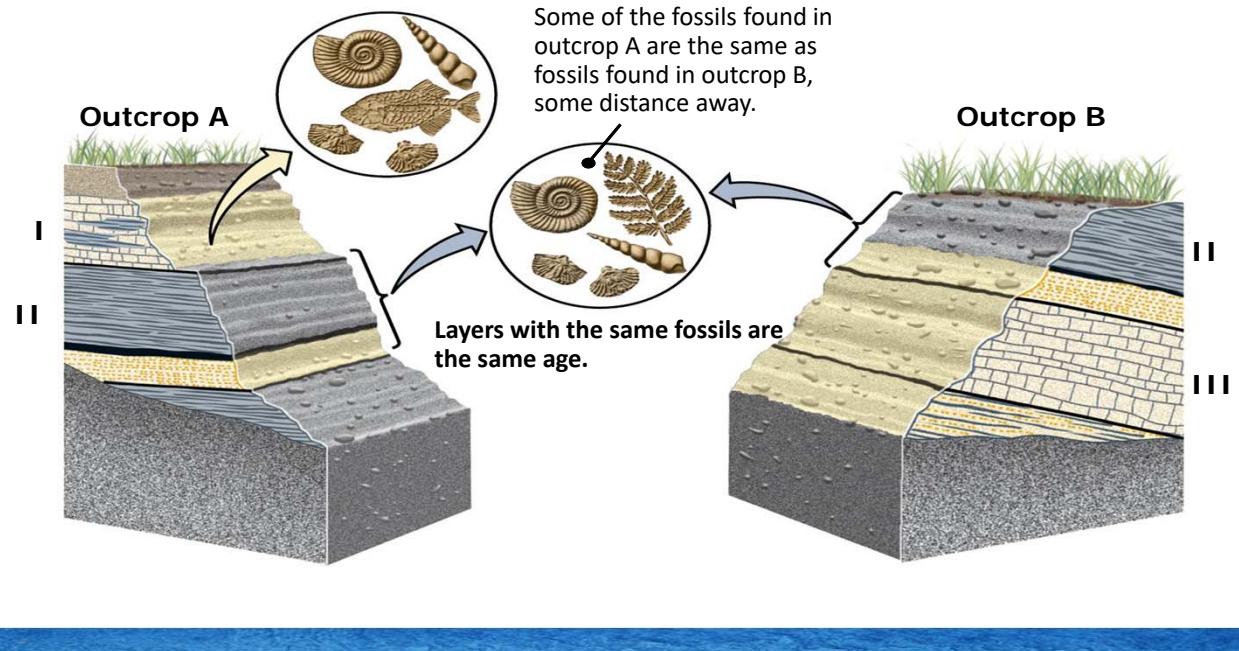
Faunal Succession



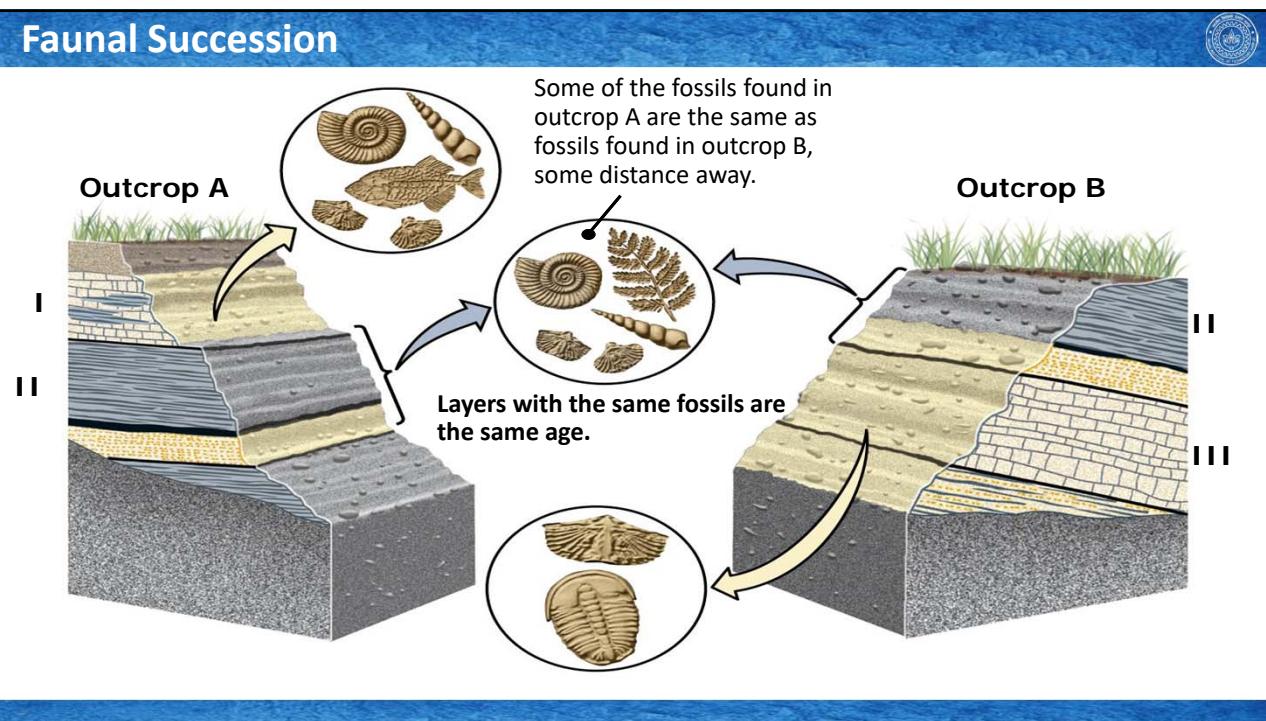
Faunal Succession



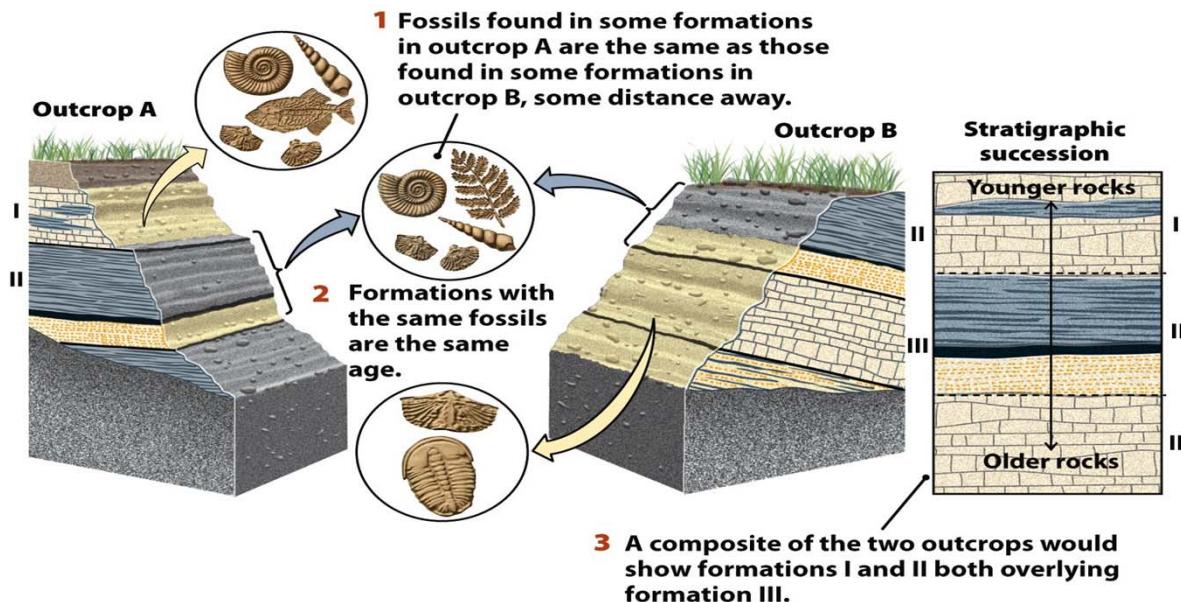
Faunal Succession



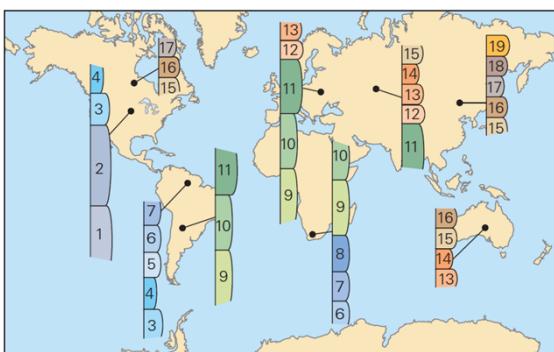
Faunal Succession



Faunal Succession



Faunal & Stratigraphic Succession



By correlation, the strata from localities around the world can be stacked in a chart representing geologic time to create the geologic column. Geologists assigned names to time intervals, but since the column was built without knowledge of numerical ages, it does not depict the duration of these intervals.

Eon	Era	Period	Epoch
	Cenozoic	Quaternary	Holocene Pleistocene
		Neogene	Pliocene Miocene
	Tertiary		Oligocene Eocene Paleocene
	Paleogene		
	Mesozoic	Cretaceous	
		Jurassic	
		Triassic	
Phanerozoic			
	Paleozoic	Permian	Pennsylvanian Carboniferous
		Devonian	Mississippian
		Silurian	
		Ordovician	
		Camrian	
Proterozoic			
Archean			

Stratigraphic Records



- Unconformities – gaps in the record

Disconformity: an unconformity between parallel layers of sedimentary rocks which represents a period of erosion or non-deposition. Disconformities are marked by features of subaerial erosion. This type of erosion can leave channels and paleosols in the rock record

Nonconformity: A nonconformity exists between sedimentary rocks and metamorphic or igneous rocks when the sedimentary rock lies above and was deposited on the pre-existing and eroded metamorphic or igneous rock.

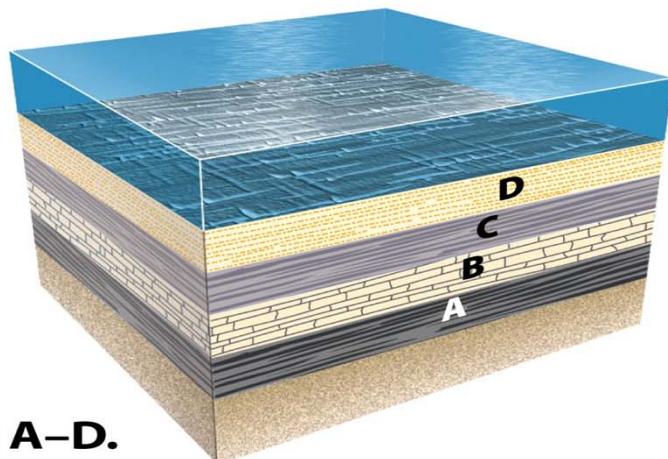
Angular Unconformity: an unconformity where horizontally parallel strata of sedimentary rock are deposited on tilted and eroded layers, producing an angular discordance with the overlying horizontal layers.

Paraconformity: a type of unconformity in which strata are parallel; there is no apparent erosion and the unconformity surface resembles a simple bedding plane. It is also called nondepositional unconformity or pseudoconformity. A paraconformity of shorter time period is known as Diastem.

Disconformity



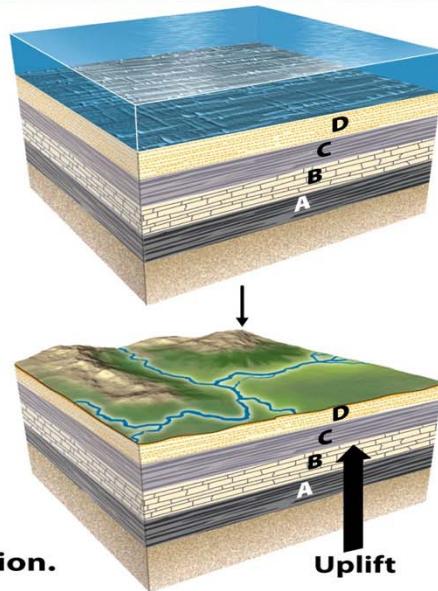
TIME 1
Beneath the ocean,
sedimentary beds
accumulate in layers A–D.



Disconformity



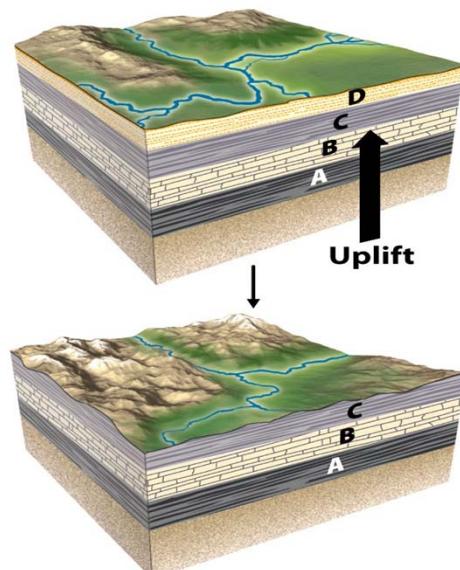
TIME 2
**Later, tectonic forces cause
uplift of the beds above sea
level, exposing them to erosion.**



Disconformity



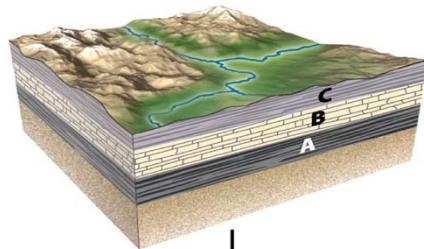
TIME 3
**Erosion strips away layer D
and part of C, leaving an
irregular surface of hills and
valleys.**



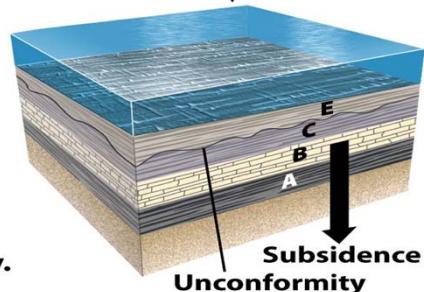
Disconformity



TIME 3
Erosion strips away layer D and part of C, leaving an irregular surface of hills and valleys.



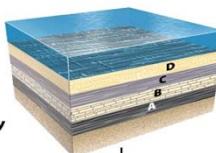
TIME 4
Subsidence below sea level allows a new layer, E, to be deposited over C. The irregular surface of C is preserved as an unconformity.



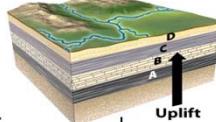
Disconformity



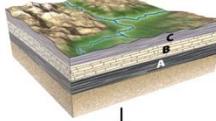
TIME 1
Beneath the ocean, sedimentary beds accumulate in layers A-D.



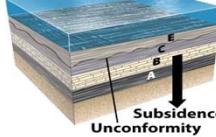
TIME 2
Later, tectonic forces cause uplift of the beds above sea level, exposing them to erosion.



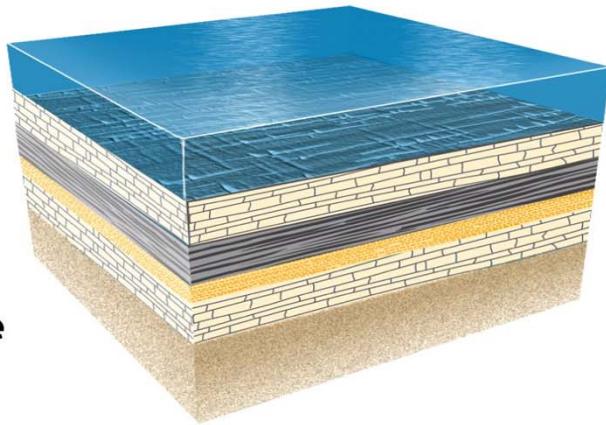
TIME 3
Erosion strips away layer D and part of C, leaving an irregular surface of hills and valleys.



TIME 4
Subsidence below sea level allows a new layer, E, to be deposited over C. The irregular surface of C is preserved as an unconformity.

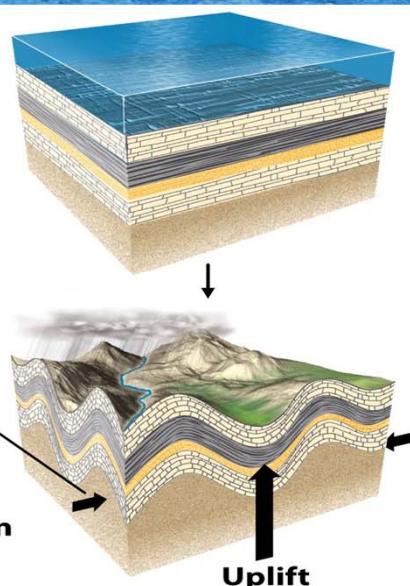


Angular Unconformity



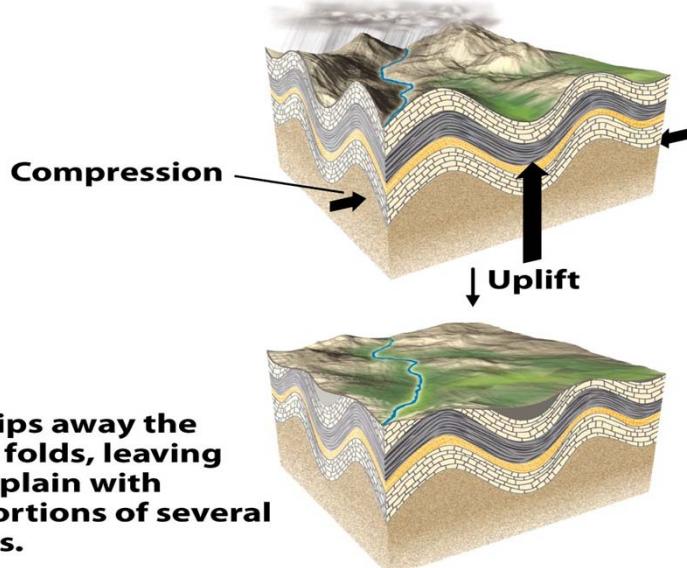
TIME 1
**Beneath the ocean,
sediments accumulate
in beds.**

Angular Unconformity



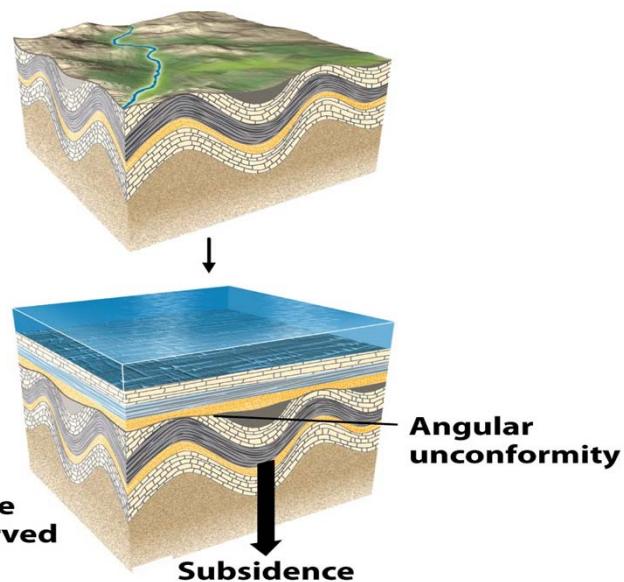
TIME 2
**Later, tectonic forces cause
uplift, folding, and deformation
of the sedimentary beds.**

Angular Unconformity



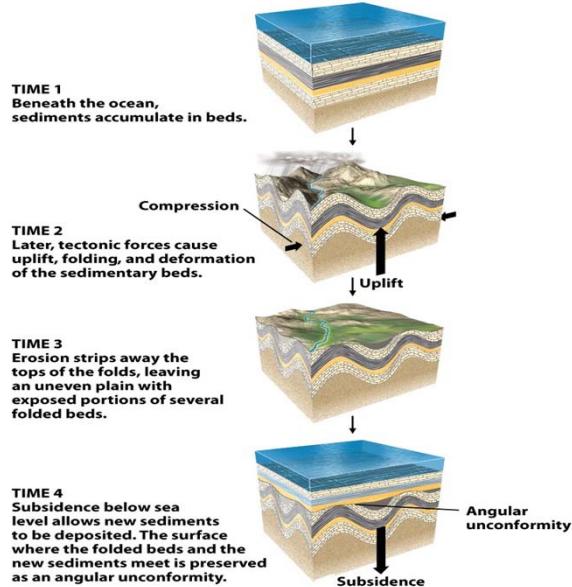
TIME 3
Erosion strips away the
tops of the folds, leaving
an uneven plain with
exposed portions of several
folded beds.

Angular Unconformity

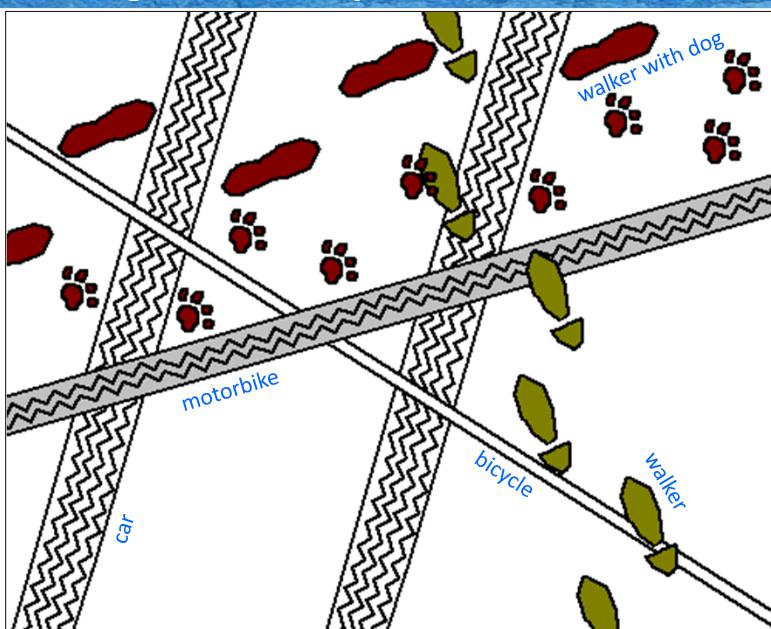


TIME 4
Subsidence below sea
level allows new sediments
to be deposited. The surface
where the folded beds and the
new sediments meet is preserved
as an angular unconformity.

Angular Unconformity



Cross-Cutting Relationship



walker with dog

walker

motorbike

bicycle

car

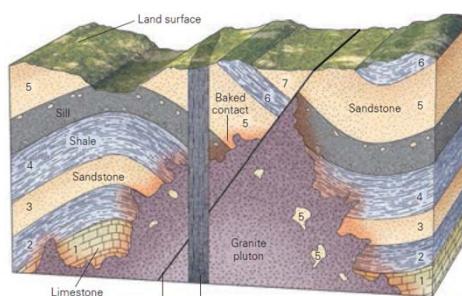
Cross-Cutting Relationship



- The latest fracture movement should displace previous ones in a terrain with multiple generation of fractures.
- If the latest fracture involve shear component (Type -II), the previous fractures should slip accordingly along latest fracture.



Cross-Cutting Relationship



(a) Geologic principles help us unravel the sequence of events leading to the development of the features shown above. Layers 1 to 7 were deposited first. Intrusion of the sill came next, followed by folding, intrusion of the granite pluton, faulting, intrusion of the dike, and erosion.

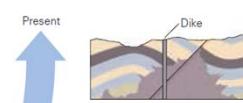
(b) The sequence of geologic events leading to the geology shown above.

Past

A sequence of strata accumulates.



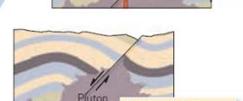
An igneous sill intrudes.



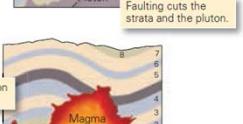
Erosion forms the present land surface.



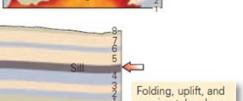
A dike intrudes.



Faulting cuts the strata and the pluton.



An igneous pluton cuts older rock.



Folding, uplift, and erosion take place.



Geological Time Scale (relative and absolute)



Indian Institute of Technology, Kanpur

Department of Earth Sciences

ESO213A: Fundamentals of Earth Sciences

Lecture 25. Geological Time Scale

Santanu Misra

Department of Earth Sciences
Indian Institute of Technology, Kanpur
smisra@iitk.ac.in • <http://home.iitk.ac.in/~smisra/>



Aims of this lecture

- Geological Time Scale: relative and absolute (numerical) ages
- Biography of the Earth

Reading:

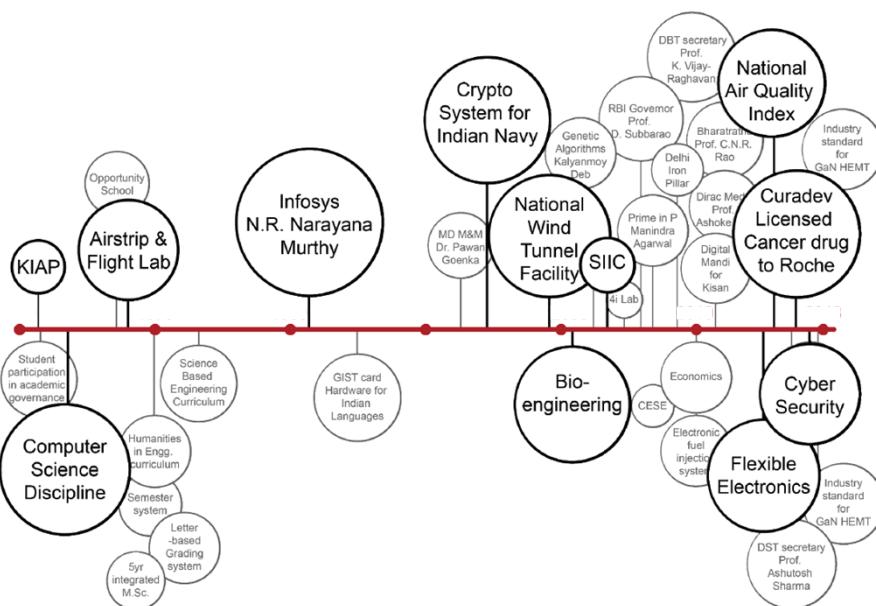
Marshak's Book (Part-IV)
Grotzinger & Jordan's book (Chapter 8)
[for the entire week]

Relative and absolute (numerical) ages

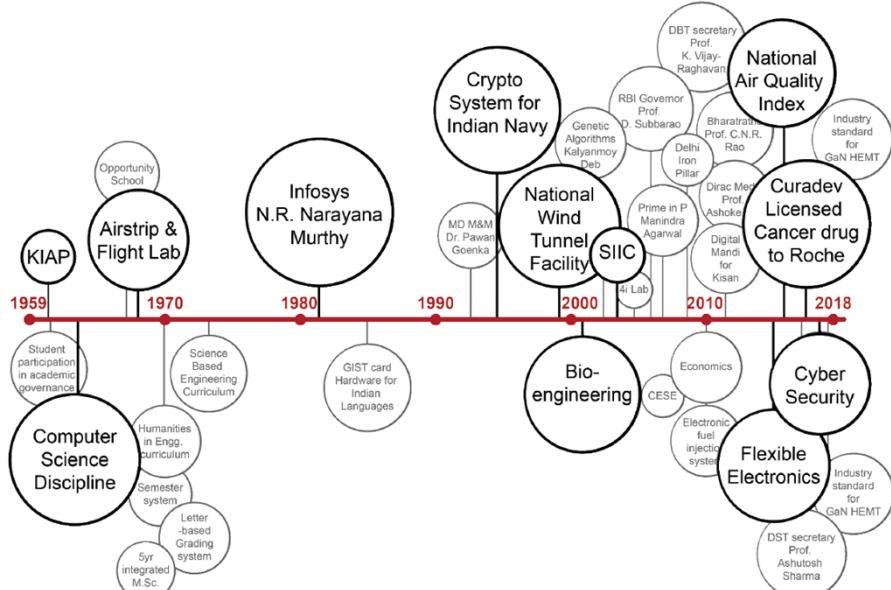


- Geologists studied the principles stratigraphic successions (previous lecture) and the Theory of Uniformitarianism (*physical processes we observe operating today also operated in the past, at roughly comparable rates OR present is the key to the past*) to determine the **relative ages** of the rocks structures, and other geologic features at a given location (Geologic Events).
- The relative ages cannot suggest time-line in the Earth history. of the rocks structures, and other geologic features at a given location (Geologic Events). This situation changed with the discovery of radioactivity. The overall determination and interpretation of numerical ages as **geochronology**. The technique of isotopic dating has been vastly improved over the years and we can now know more or less precisely the time-lines of the past geological events (**the Numerical Age**).

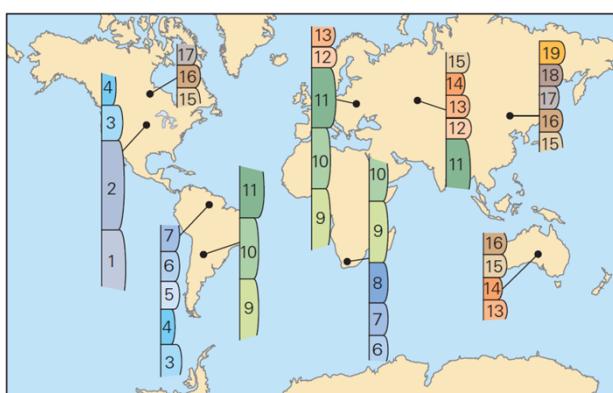
Relative timeline of IIT Kanpur



Absolute (numerical) timeline of IIT Kanpur



Relative Geological Time Scale



By correlating strata from locality to locality at millions of places around the world, geologists have pieced together a composite stratigraphic column, called the **geologic column**, that represents the entirety of Earth history.

Eon	Era	Period	Epoch
	Cenozoic	Quaternary	Holocene Pleistocene
		Neogene	Pliocene Miocene
		Tertiary	Oligocene Eocene Paleocene
	Phanerozoic	Paleogene	
		Cretaceous	
	Mesozoic	Jurassic	
		Triassic	
	Paleozoic	Permian	
		Carboniferous	Pennsylvanian Mississippian
		Devonian	
		Silurian	
		Ordovician	
		Camrian	
2	Proterozoic		
1	Archean		

Relative Geological Time Scale



EONS: The largest subdivisions break Earth's history into the *Hadean*, *Archean*, *Proterozoic*, and *Phanerozoic*—the first three of these, together, constitute the **Precambrian**.

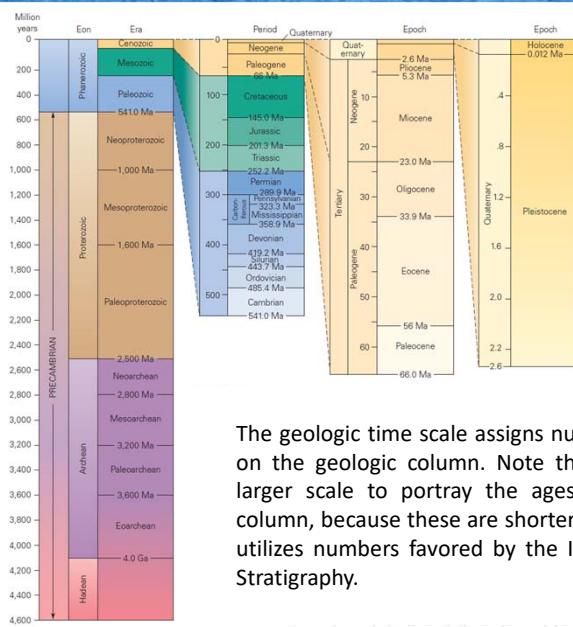
- there is no rock record from Hadean
- *zoic* means LIFE, Proterozoic → first life; Phanerozoic → visible life
- because decades after the EONS had been named, geologists discovered that the earliest life, cells of bacteria and archaea, actually appeared during the Archean Eon.

ERAS: Eons are subdivided into **eras**. The Phanerozoic Eon, for example includes, in order from oldest to youngest, the Paleozoic (*ancient life*), Mesozoic (*middle life*), and Cenozoic (*recent life*) Eras.

Eras are further subdivided to **PERIODS** and each periods into **EPOCHS**.

- The names of the Periods are mostly from places where the signature rocks are best exposed

Absolute (Numerical) Geological Time Scale



The geologic time scale assigns numerical ages to the intervals on the geologic column. Note that we have to change to a larger scale to portray the ages of intervals higher in the column, because these are shorter subdivisions. This time scale utilizes numbers favored by the International Commission on Stratigraphy.

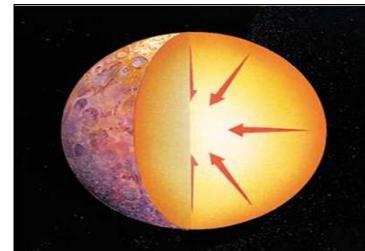
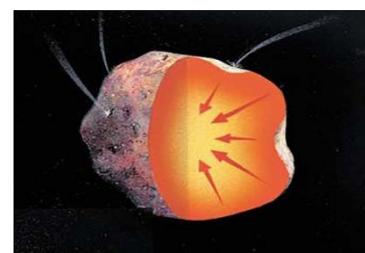


Biography of the Earth

The Hadean Eon



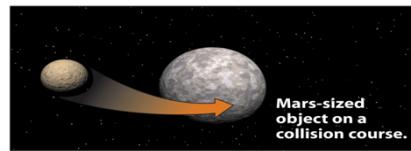
- The time between 4.57 and 3.8 Ga is the Hadean Eon.
 - Named for Hades, the Greek god of the underworld.
 - Began with formation of Earth by planetesimal accretion.
- Earth was heated by impacts and radioactive decay.
- Earth was hot enough to partially melt by ~ 4.5 Ga.
 - The molten Earth underwent chemical differentiation.
 - Gravity pulled molten iron into the center.
 - The ultramafic mantle remained as a thick shell.



The Hadean Eon



- After differentiation, Earth smashed a protoplanet.
- The size of Mars, this planet blasted...
 - A sizeable chunk of Earth's mantle.
 - Much of the protoplanet's mantle.
- Debris from the collision formed a ring around Earth.
- This debris coalesced to form the moon.
- When 1st formed, moon was much closer (20,000 km).
- Today it is 19x farther away (384,000 km).



The Hadean Eon



- Earth was inhospitable; a molten surface.
 - Evidence of solidified igneous rock dates from 4.4 Ga.
 - This evidence is from zircon grains, not a whole rock.
- Volcanic outgassing created a deadly atmosphere.
 - N₂, NH₃, CH₄, H₂O, CO, CO₂, and SO₄²⁻ were components.
 - This atmosphere had a greater density than today's.
- Early formed crust was bombarded by meteorites.
 - Meteorite impacts were abundant between 4.0 and 3.9 Ga.
 - This would have destroyed early formed crust.
 - Oldest evidence of crust is 4.03 Ga.
- The first oceans formed as rain from the skies.
 - Liquid water required cooling of the surface.
 - First evidence of oceans from marine sediments ~ 3.85 Ga.

The Archean Eon

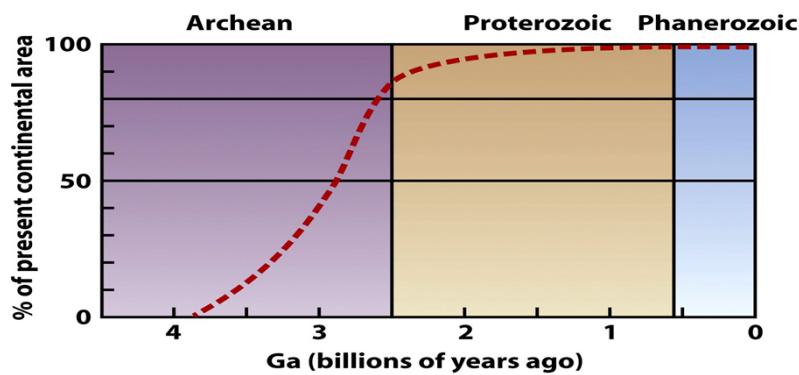


- Time of significant change to planet Earth.
- ~3.8 Ga, Earth had cooled to form lithosphere.
 - Intense meteorite bombardment ceased.
 - Portions of the rock record begin to survive.
- Had plate tectonics started yet? Two models:
 - Many small microplates, island arcs, and hot spot volcanoes rapidly formed and subducted crust.
 - Archean lithosphere was too hot to subduct; hot spot plume volcanics dominated formation of crust.

The Archean Eon



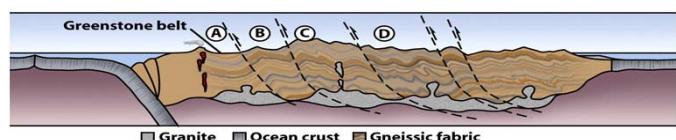
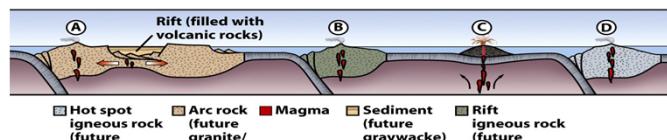
- Volume of continental crust increased dramatically.
 - 85% of modern continental area present by end of Archean.
 - Signals full development of plate tectonic processes.



The Archean Eon



- How did continental crust form?
 - Low-density felsic rocks formed above subduction zones.
 - Felsic crustal blocks grew via continental collision.
 - Felsic sediments accumulated near continental crust.
 - Mantle hot spots built mafic volcanoes.
 - Continental crust is too buoyant to subduct.
 - Jams subduction trenches.
 - Shuts off subduction.
 - Creates thickened, uplifted continental crust.
 - Frequent collisions sutured volcanic arcs, hot spots, and sedimentary debris together as protocontinents.



The Archean Eon



- Archean cratons consist of 5 principal rock types.
 - Gneiss – Hi-grade metamorphics from Archean collisions.
 - Greenstone – Metamorphosed fragments of mafic rocks.
 - Granite – Magmas from partial melting of the crust.
 - Graywacke – Sedimentary debris derived from arcs.
 - Chert – Silica precipitated in the deep sea.
- Archean shallow sediments are poorly known.
 - There were few shallow depositional settings, or...
 - Few examples have survived destruction by erosion.
- Sedimentary processes were clearly operating.
 - Transport rounded sediment grains.



The Archean Eon



- Life first appeared during the Archean. Evidence?
 - Biomarker molecules; Isotopic signatures; Fossil cells.
- Clear evidence of life in rocks dated to 3.5 Ga.
 - Life may have started earlier.
- Oldest undisputed bacteria fossils ~ 3.2 Ga.
- Rocks after 3.2 Ga contain stromatolites.
 - Layered mats of cyanobacteria (blue-green algae).
 - Sediments stuck to mucous coatings on algal filaments.
- Photosynthesis changed Earth's atmosphere.
 - Converted CO₂ and H₂O to organic matter and free oxygen.

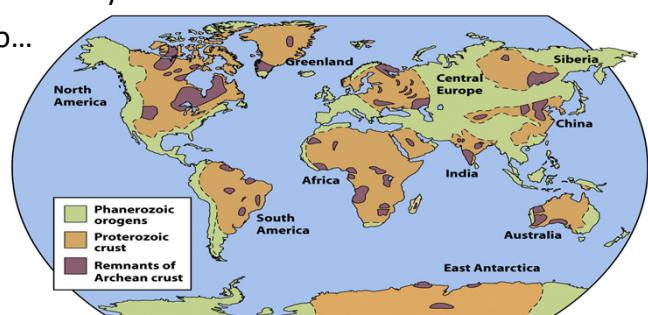


Khelen et al., 2017

The Proterozoic Eon



- Protero = first; zoic = life.
 - Named before Archean life was discovered.
- ~ 2 Ga (2.5 to 0.542 Ga); almost ½ of Earth history.
- The unfamiliar Archean world changed to...
 - Fewer, larger lithospheric plates.
 - Larger continental landmasses.
 - An oxygenated atmosphere.
- New continental crust formed, but at slower rates.
 - 90% of Earth's continental crust by the middle Proterozoic.
 - Continents grew by addition of volcanic arcs.
 - Continents cooled and strengthened to become cratons.



The Proterozoic Eon



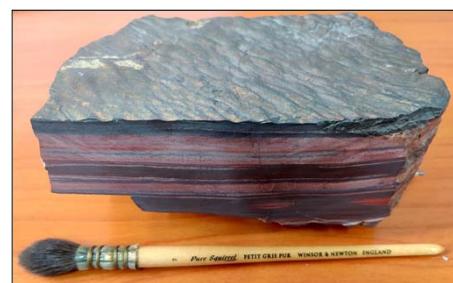
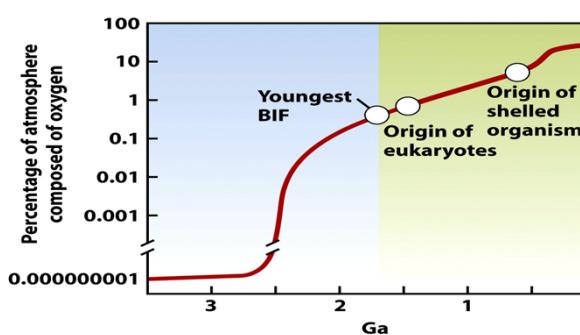
- Continental collision created Precambrian supercontinents.
 - Rodinia – Formed ~ 1 Ga.
 - The Grenville orogeny formed an extensive mountain belt.
- Pannotia – A short-lived supercontinent ~ 570 Ma.



The Proterozoic Eon



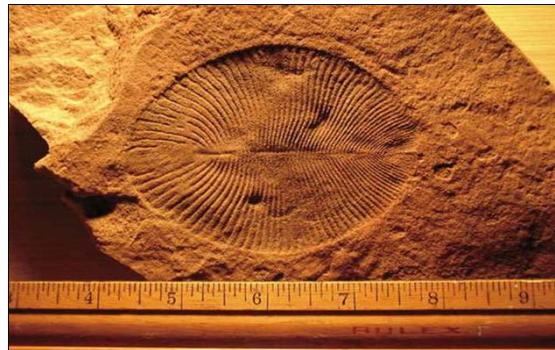
- Atmospheric oxygen (O_2) skyrocketed 2.4 to 2.2 Ga.
 - Currently, O_2 is 21% of the atmosphere.
 - Before 2.2 Ga, detrital pyrite in sediments indicated no O_2 .
 - Redbeds (red from Fe-oxides) don't appear before 2.2 Ga.
 - Banded iron formations (BIFs) – Fe dissolved in the ocean reacted with O_2 , forming worldwide iron oxide deposits.



The Proterozoic Eon



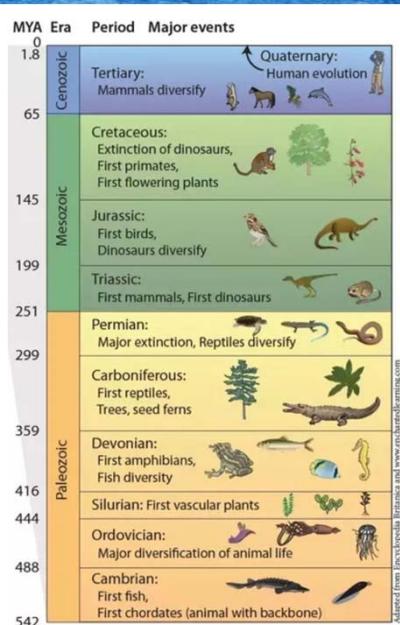
- Atmospheric O₂ permitted diversification and multicellular life (~500 Ma).
- Ediacaran fauna – Unusual soft-bodied fossils.
 - Preserved in end Proterozoic sediments; Multicellular invertebrates resembling worms and jellyfish.
 - Two events: assembly and break-up of Pannotia; Global cooling



The Phanerozoic Eon

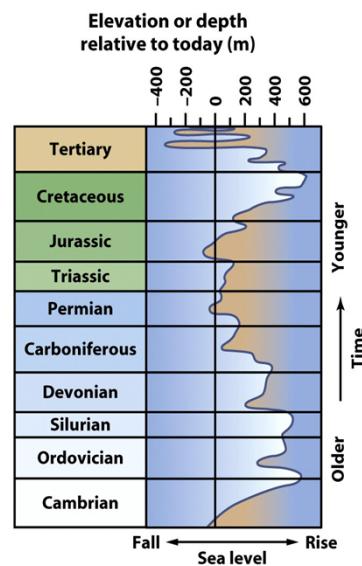


- Phaneros = visible; zoic = life.
- The most recent 542 Ma of Earth history.
 - Began with appearance of diverse hard-shelled organisms.
 - Hardshells vastly increased fossil preservation.
 - Made possible a more complete archive of life on Earth.
- The Phanerozoic is divided into 3 eras.
 - Paleozoic – Ancient life (Trilobites, Nautiloids).
 - Mesozoic – Middle life (Brachiopods, Dinosaurs).
 - Cenozoic – Recent life (early Horses, Mammoths..).
- Eras emphasize changes in Earth's biota.



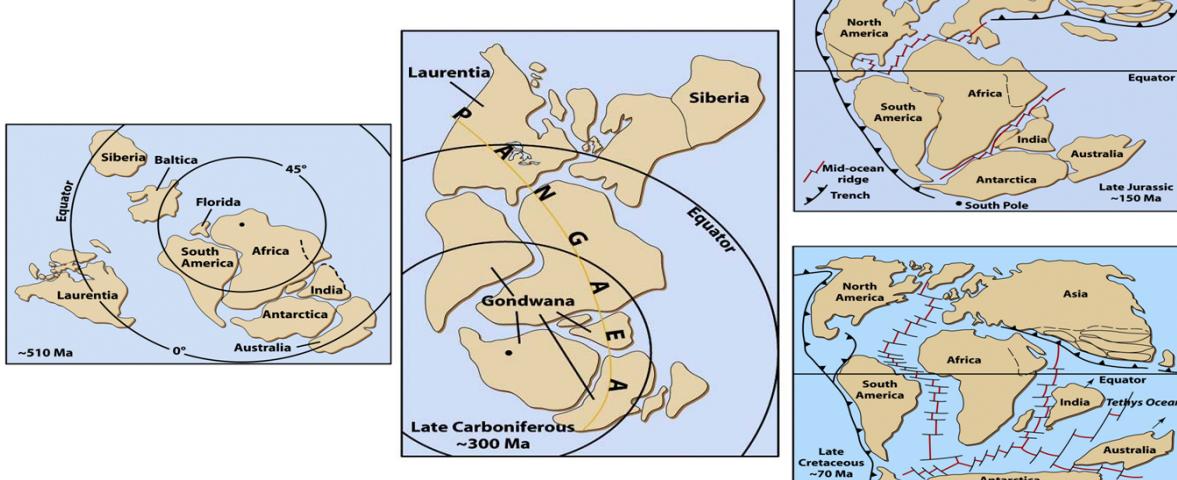
The Proterozoic Eon

- Tectonic plates and continental blocks rearranged.
- New supercontinents formed and rifted apart.
- Numerous orogenic belts were created and eroded.
- Phanerozoic sea level (SL) has changed often.
- High sea level flooded continental interiors (sediment deposition).
- Low sea level exposed continental margins (erosion or nondeposition.)



The Proterozoic Eon

- Rearrangements of the continents

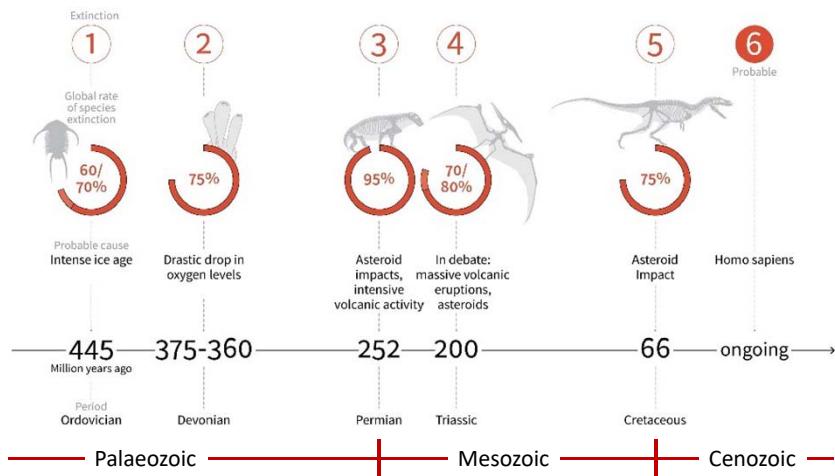


The Proterozoic Eon



Global Mass extinctions

More than 99 % of all organisms that have ever lived on Earth are extinct.

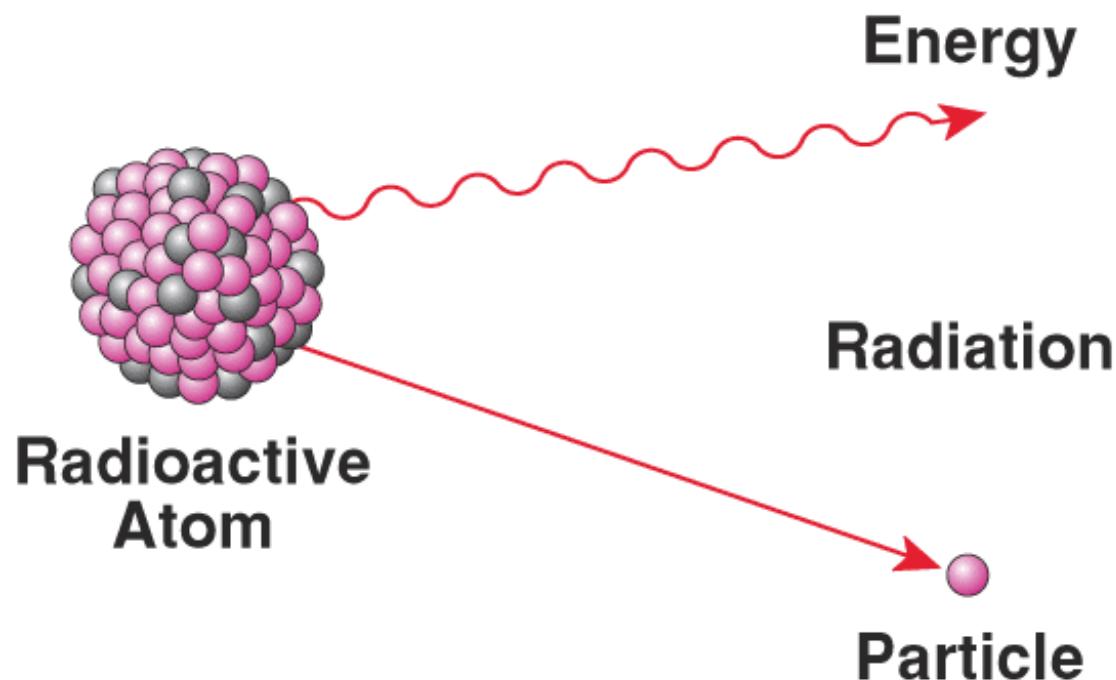


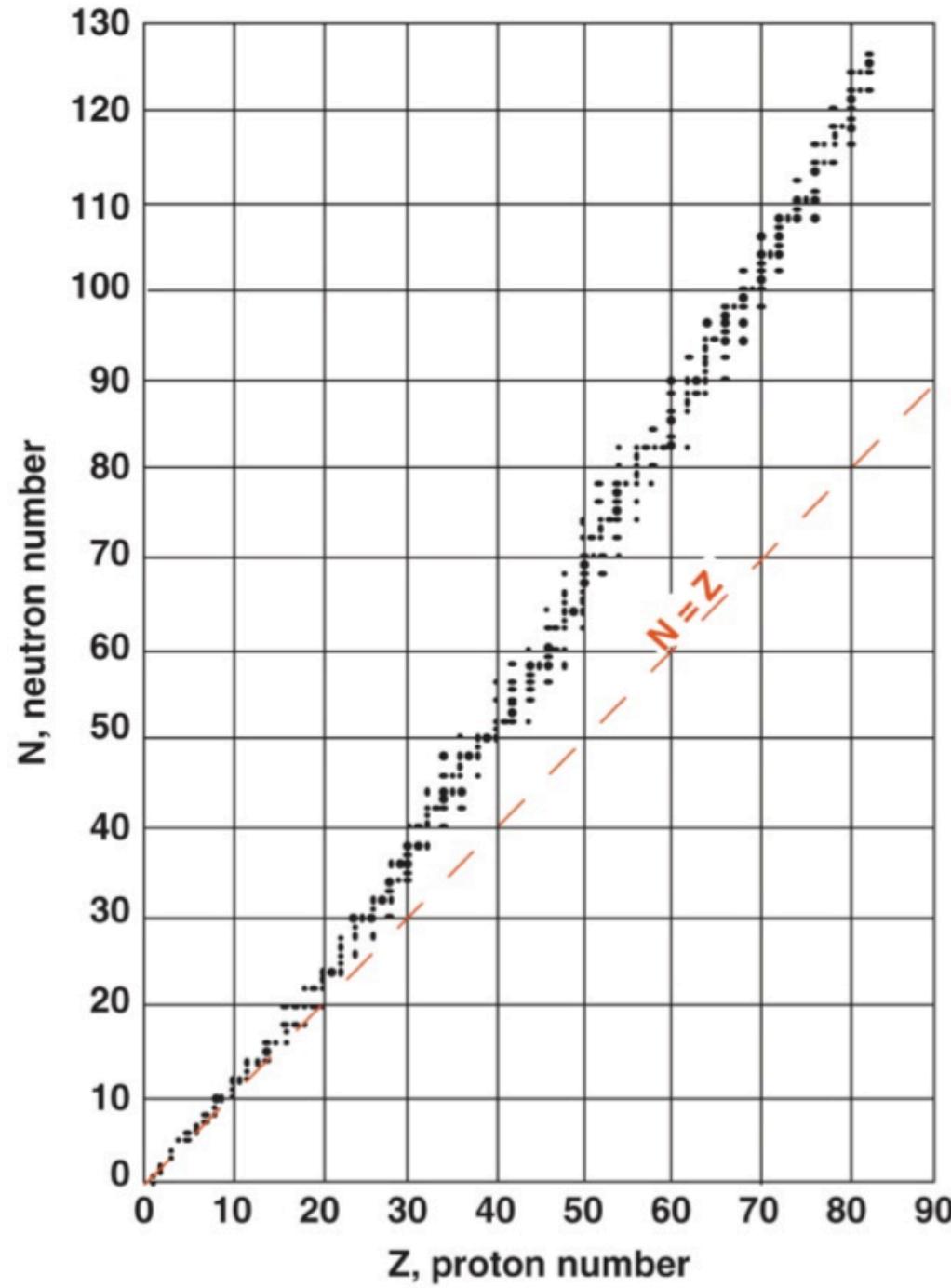
Next Lecture



Methods of Numerical Age Dating

Isotope Geochemistry





I-b Properties of nuclides

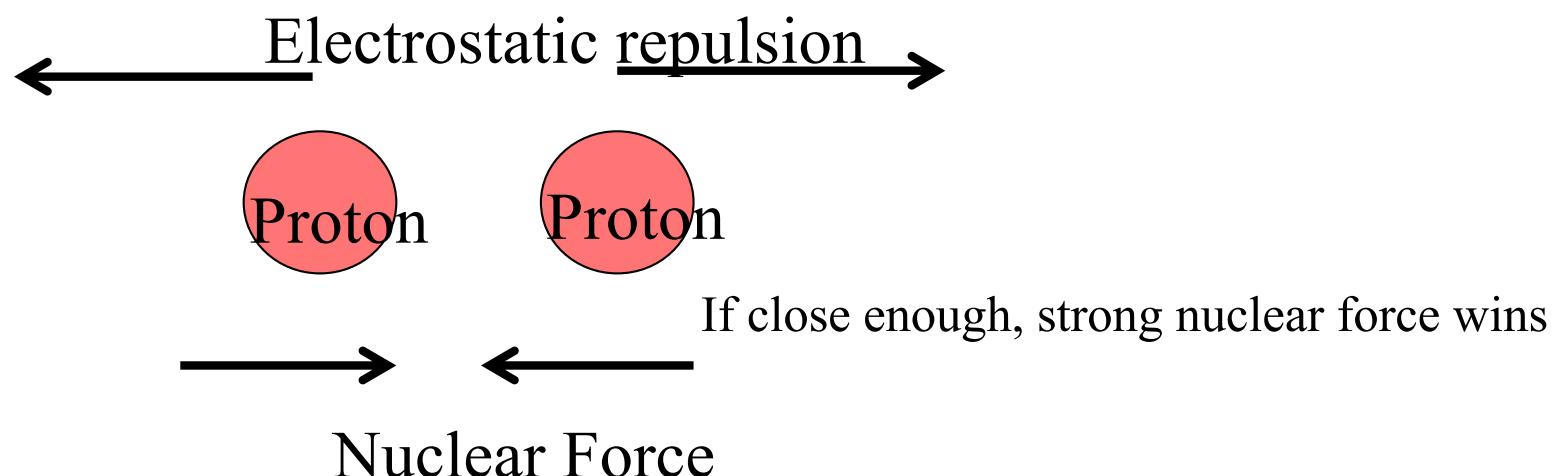
- Mass:

The mass of a nuclide is given in amu (atomic mass unit)

Definition: $M(^{12}\text{C})=12 \text{ amu}$

$m_{\text{proton}} \approx m_{\text{neutron}} \approx 1 \text{ amu}$

- The mass of one atom of carbon 12 (^{12}C) is determined by mass spectrometry and is equal to $1.9922 \times 10^{-23} \text{ g}$
- Electrostatic repulsion vs. nuclear attraction



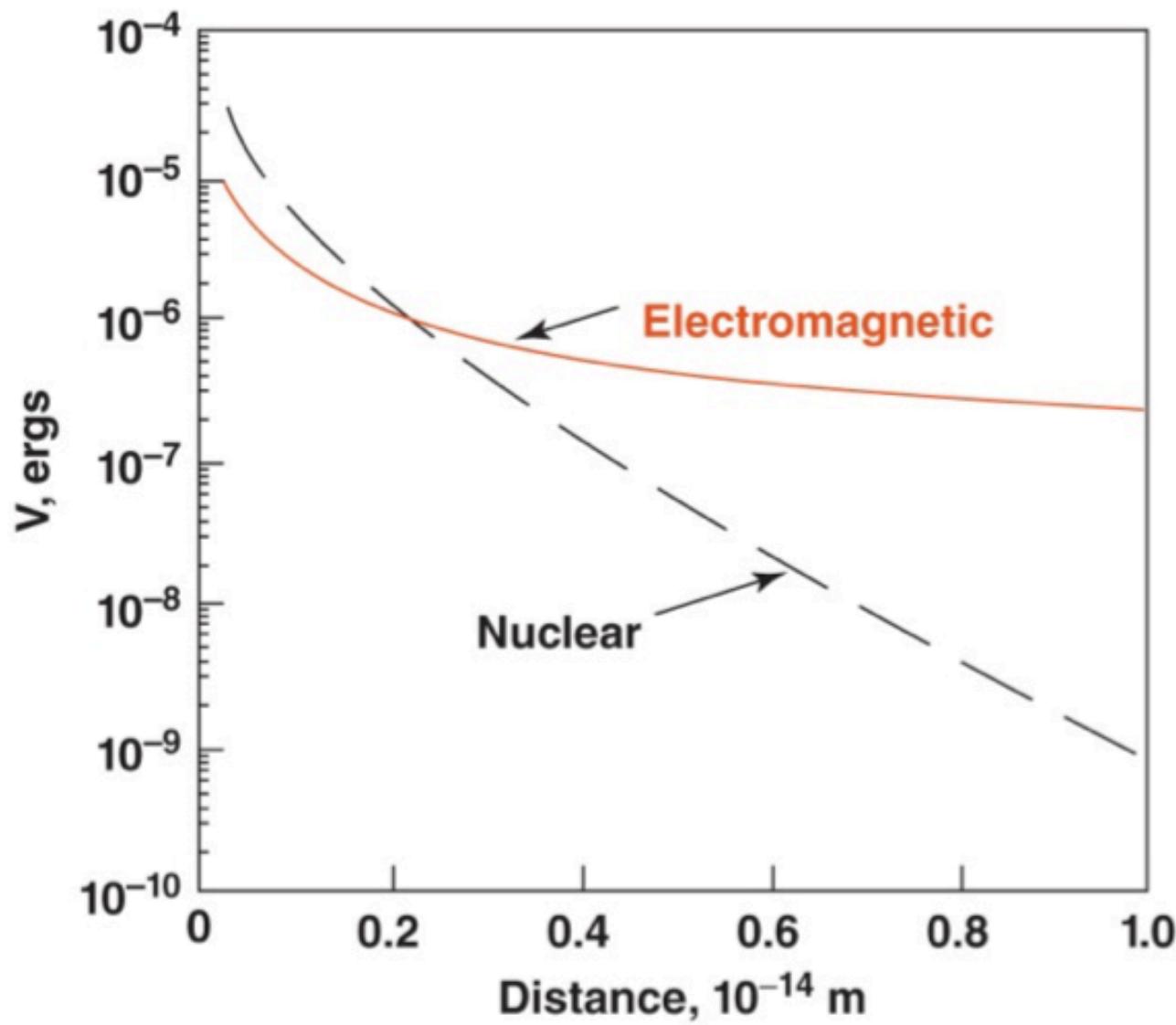
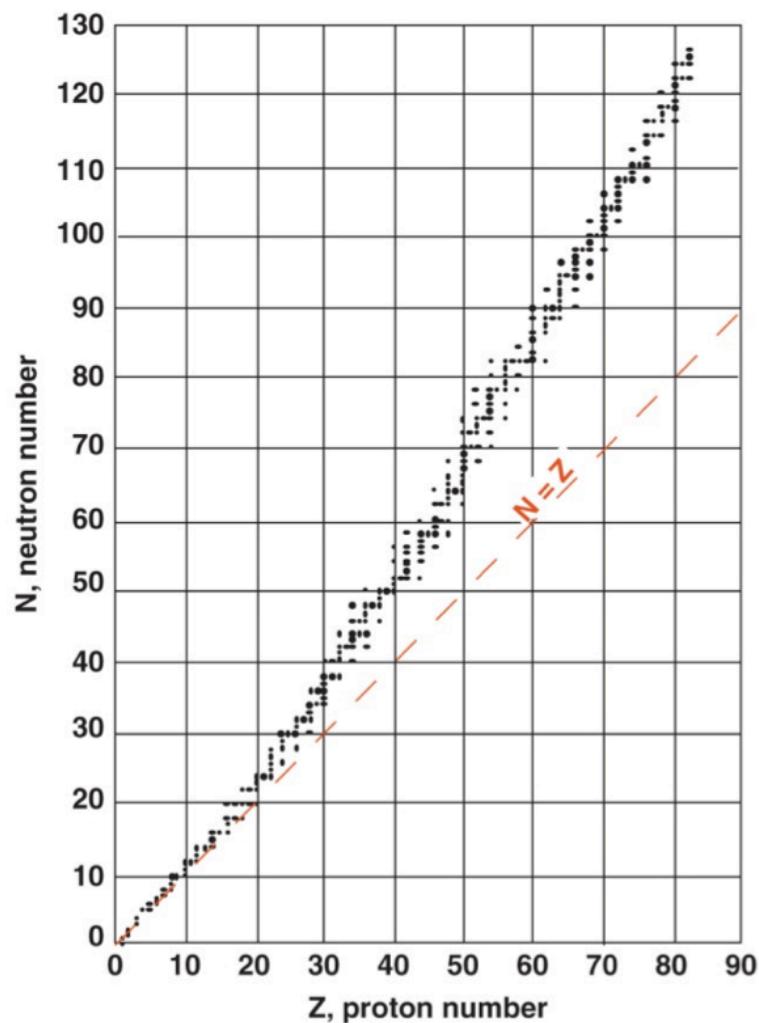


Figure 1.2 The nuclear and electromagnetic potential of a proton as a function of distance from the proton.



The elements deviate from the 1:1 line.

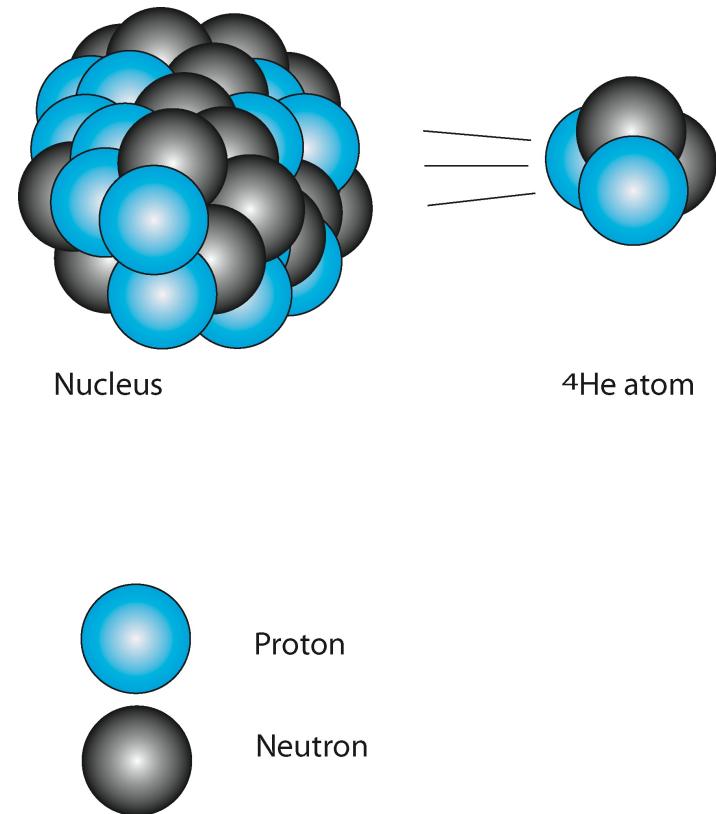
As the number of protons increases in a nucleus, the electromagnetic force increases faster than the strong nuclear force. Therefore larger atoms needs more and more neutrons to remain stable.

II. Natural Radioactivity

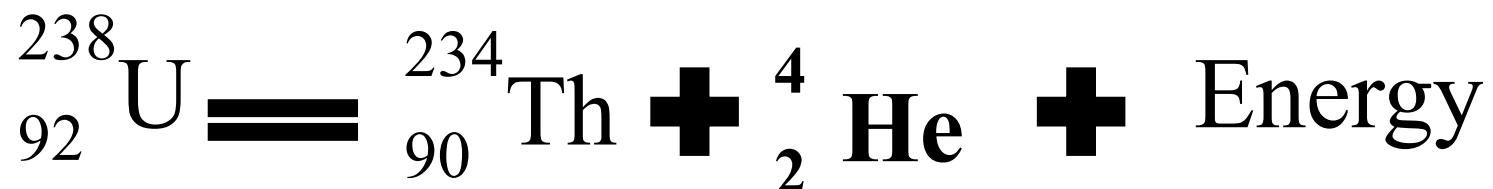
- Radioactivity is the emission of particles from unstable nuclei. Radioactivity allows a nucleus to evolve from an unstable state to a more stable configuration.
- The emitted particles by radioactivity are of three types: α , β , γ .
 $\alpha = {}^4\text{He}$ nucleus;
 $\beta^- = \text{electron}$ or $\beta^+ = \text{positron}$;
 $\gamma = \text{photons}$

α - decay

- Occurs for unstable nuclides with $A \geq 56$ (except ${}^5\text{He}$, ${}^5\text{Li}$ and ${}^6\text{Be}$)
- Consists of 2 proton and 2 neutron



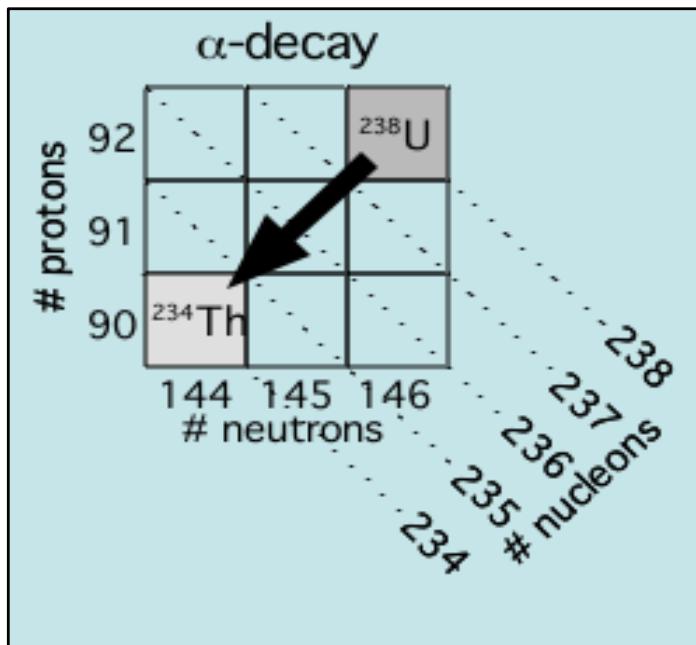
	<i>Atomic No.</i>	<i>Neutron No.</i>	<i>Mass No.</i>
<i>Parent</i>	Z	N	Z+N = A
<i>Daughter</i>	Z-2	N-2	Z-2+N-2 = A-4



α - decay

Emission of an α particle =⁴He nucleus (2 neutrons, 2 protons)

The number of nucleons A decreases by 4 and its Z by 2.



Example: $^{238}\text{U} \rightarrow ^{234}\text{Th} + ^4\text{He}$

Mass balance:

^{238}U	238.0508 amu
^{234}Th	-234.0436 amu
^4He	-4.00260 amu

Mass defect 0.0046 amu

$$= 6.86 \times 10^{-10} \text{ J/decay}$$

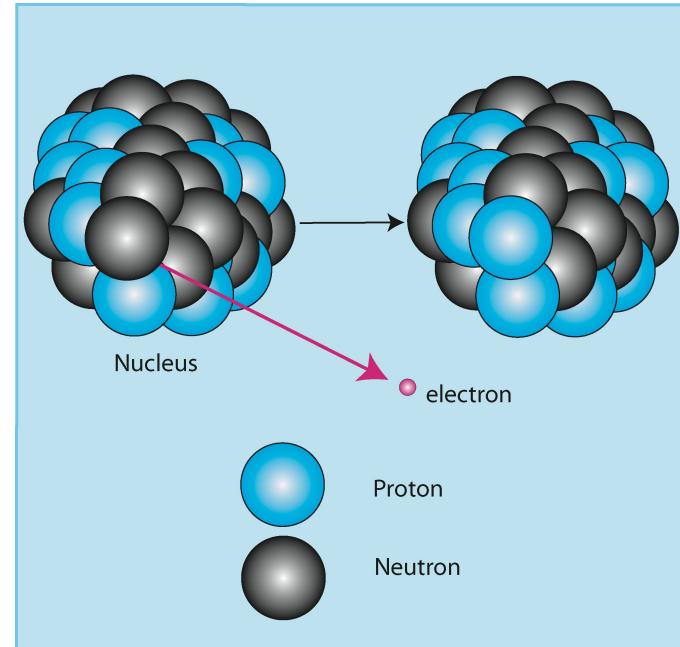
$$= 1.74 \times 10^{12} \text{ J/g } ^{238}\text{U}$$

$$=\sim 480,000 \text{ KW-H (1.6 Billion BTUs)}$$

This is the principal mode of decay of heavy isotopes such as U, Th, ²⁰⁹Bi

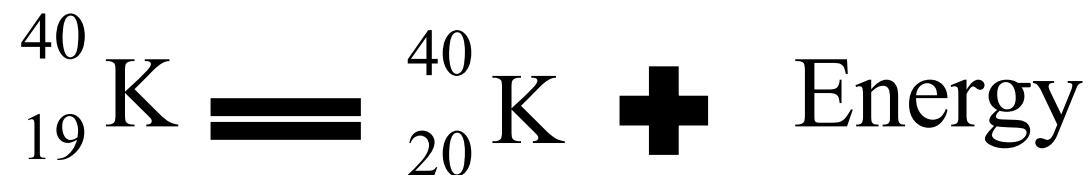
β^- decay

Decay by emitting a negatively charged beta particle



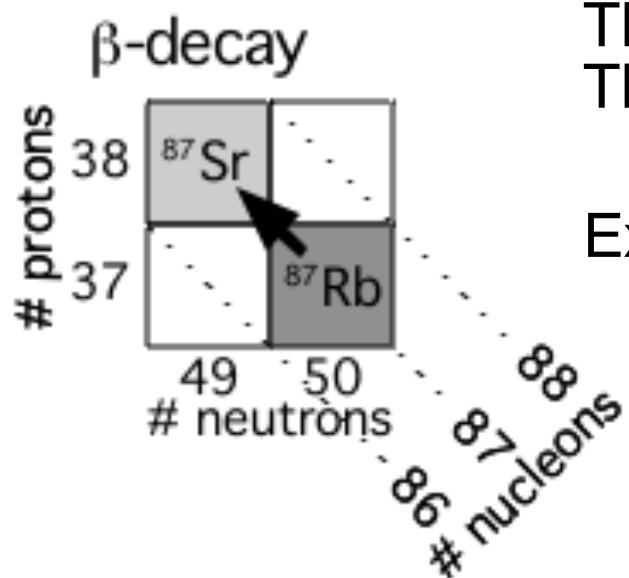
Neutron = Electron and Proton

	<i>Atomic No.</i>	<i>Neutron No.</i>	<i>Mass No.</i>
<i>Parent</i>	Z	N	Z+N =A
<i>Daughter</i>	Z+1	N-1	Z+1+N-1 = A

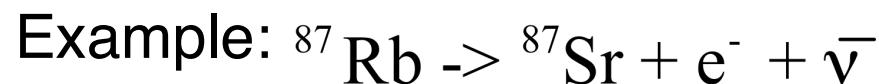


β^- decay

Emission of an electron (and an antineutrino) during the conversion of a neutron to a proton



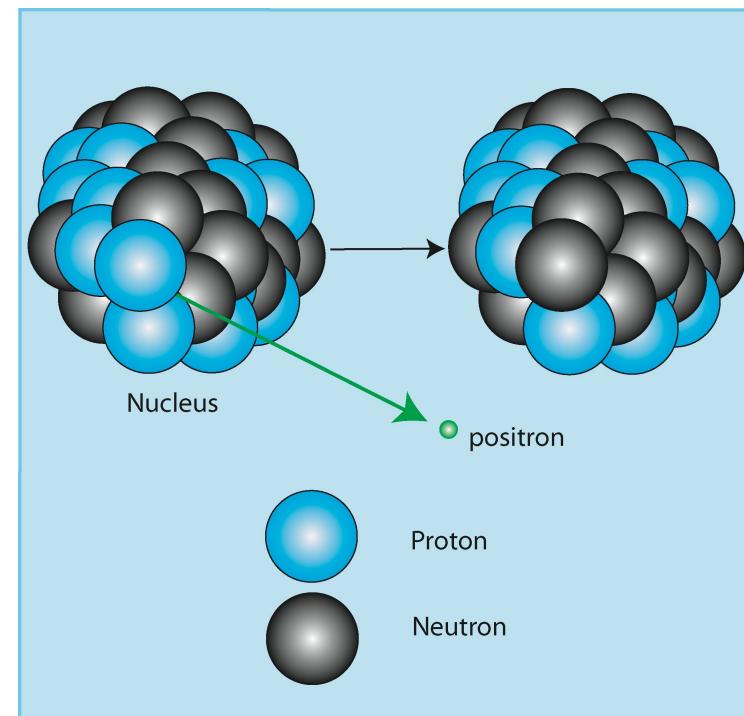
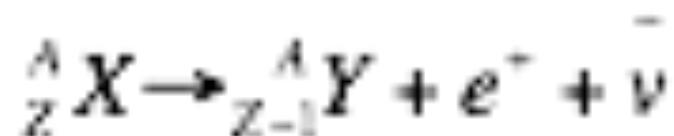
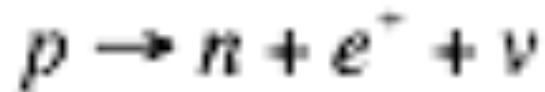
The number of nucleons does not change.
The atomic number increase by one:



This is the preferred mode of decay for nuclei with an excess number of protons relative to the stability valley.

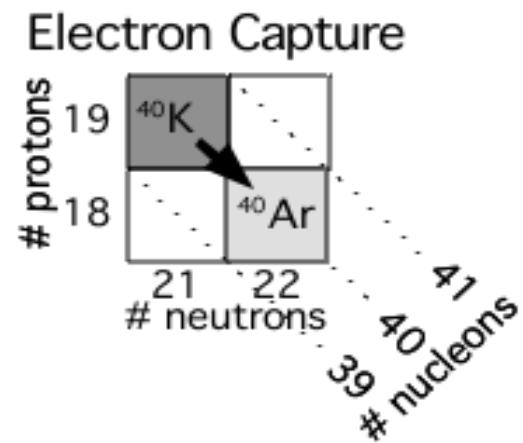
β^+ decay and electron capture

- A proton transforms into a neutron with emission of a positron and a neutrino:



β^+ decay and electron capture

Emission of a positron (and a neutrino) *or* capture of an electron from an internal layer. This is the conversion of a proton into a neutron



The number of nucleons does not change. The atomic number decreases by one :

Examples: ${}^{40}_{19}\text{K} \rightarrow {}^{40}_{18}\text{Ar} + e^- + \nu$

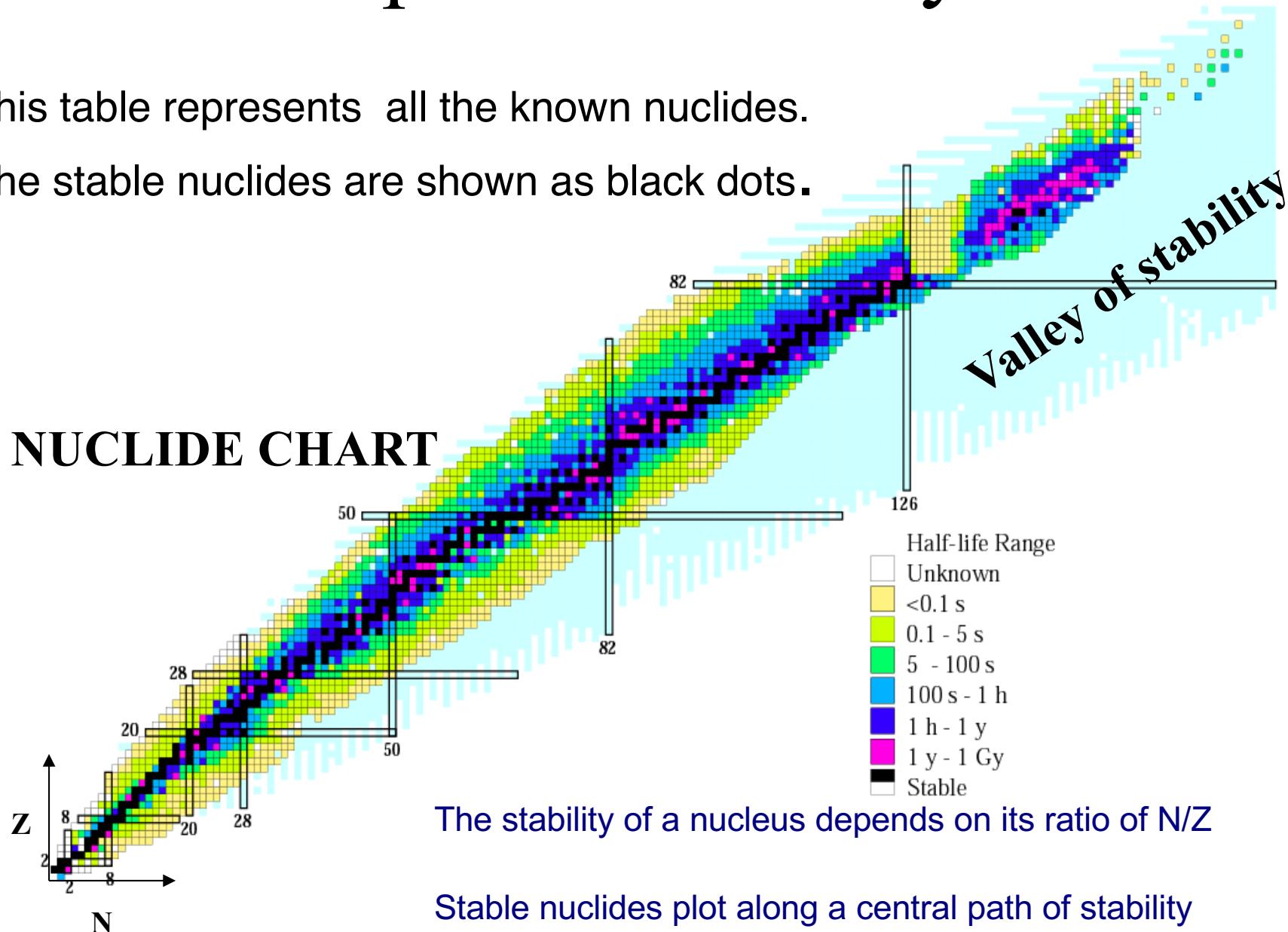
These are the decay mode for nuclei with an excess of protons relative to the stability valley.

Central path of stability

This table represents all the known nuclides.

The stable nuclides are shown as black dots.

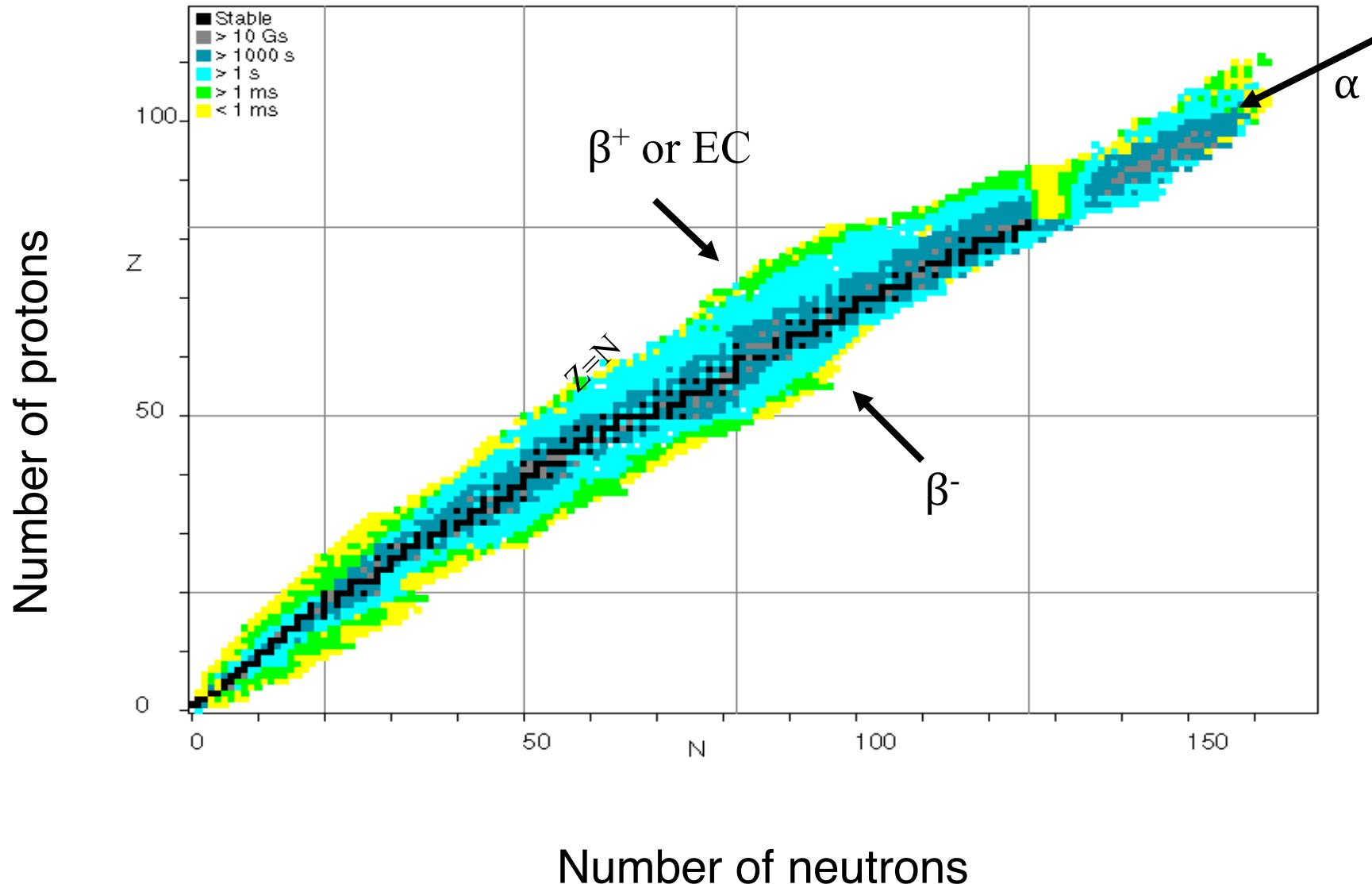
NUCLIDE CHART



The stability of a nucleus depends on its ratio of N/Z

Stable nuclides plot along a central path of stability
At low masses, $N/Z \approx 1$ ($Z=N$)
At high masses, $N/Z \approx 3$

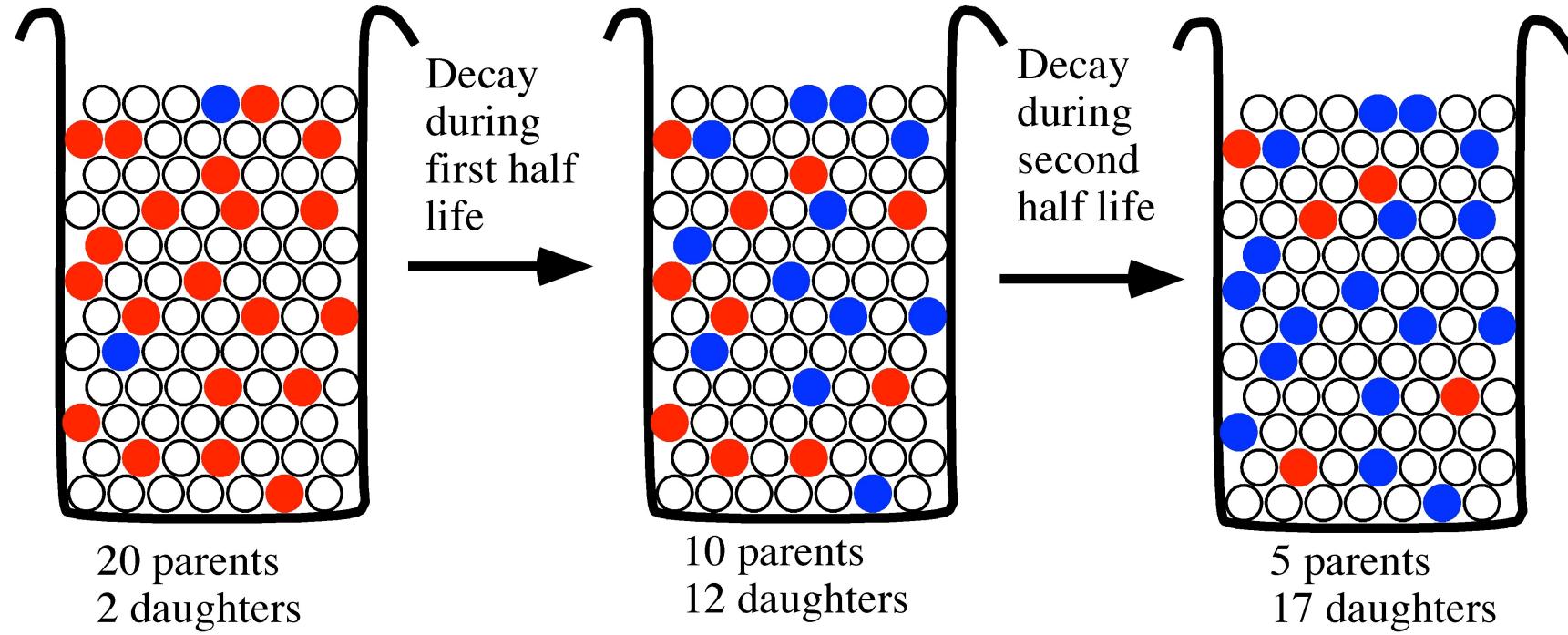
NUCLIDE CHART



Isotopes



Dating with Radioactivity



*The actual number of atoms that decay (**radioactive parent**) continually decreases and the number of stable daughter atoms (**radiogenic daughter**) increases.*

Simple Radioactive Decay

Radioactive decay is a stochastic process linked to the stability of nuclei. The rate of change in the number of radioactive nuclei is a function of the total number of nuclei present and the decay constant λ .

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

The sign on the left hand is negative because the number of nuclei is decreasing. Rearranging this equation yields

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

and integrating yields

$$-\ln N = \lambda t + C$$

C is the integration constant. We solve for C by setting $N = N_0$ and $t = t_0$. Then

$$C = -\ln N_0$$

Substituting for C gives

$$-\ln N = \lambda t - \ln N_0$$

We rearrange

$$\ln N - \ln N_0 = -\lambda t$$

Rearrange again

$$\ln N/N_0 = -\lambda t$$

Eliminate the natural log

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

And rearrange

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

...continue...

Unfortunately, we don't know N_0 a priori, but decayed N have produced radiogenic daughters D^* .

Therefore

$$D^* = N_0 - N$$

Replacing N_0 with $N e^{\lambda t}$ yields

$$D^* = N e^{\lambda t} - N$$

Rearranged

$$D^* = N (e^{\lambda t} - 1) \quad \text{or, for small } \lambda t, \quad D^* = N \lambda t,$$

The number of daughter isotopes is the sum of those initially present plus those radiogenically produced.

$$D = D_0 + D^*$$

Therefore,

$$D = D_0 + N (e^{\lambda t} - 1) \quad \text{or, for small } \lambda t, \quad D = D_0 + N \lambda t,$$

This is the basic radioactive decay equation used for determining ages of rocks, minerals and the isotopes themselves. D and N can be measured and λ has been experimentally determined for nearly all known unstable nuclides. The value D_0 can be either assumed or determined by the **isochron method**.

How to obtain a “half-life”?

The rate of decay of a radioactive parent nuclide to a stable daughter product is proportional to the number of nuclides present at any time (Rutherford and Soddy, 1902)

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

(1)

whereby N is number of nuclides , t is time and λ is the decay constant

Integration of 1

$$\int_{N_0}^N \frac{dN}{dt} = -\lambda \int_0^t dt$$

results in

$$\ln \frac{N}{N_0} = -\lambda t$$

(2)

which can be rewritten as

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

(3)

Half life: $N=N_0/2$ and $t=t_{1/2}$

$$t_{1/2} = \frac{\ln 2}{\lambda} = \frac{0.693}{\lambda}$$

Points to Remember

$$-dN/dt = \lambda N$$

N = # atoms of the radioactive nuclide, t = time,
 λ = decay constant = probability that an atom will decay in a unit time

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

$$N_0 = N \text{ at } t=0$$

$$t_{1/2} = (\ln 2) / \lambda = .693/\lambda$$

Half-life = time it takes for half a sample to decay (t when $N = \frac{1}{2}N_0$)

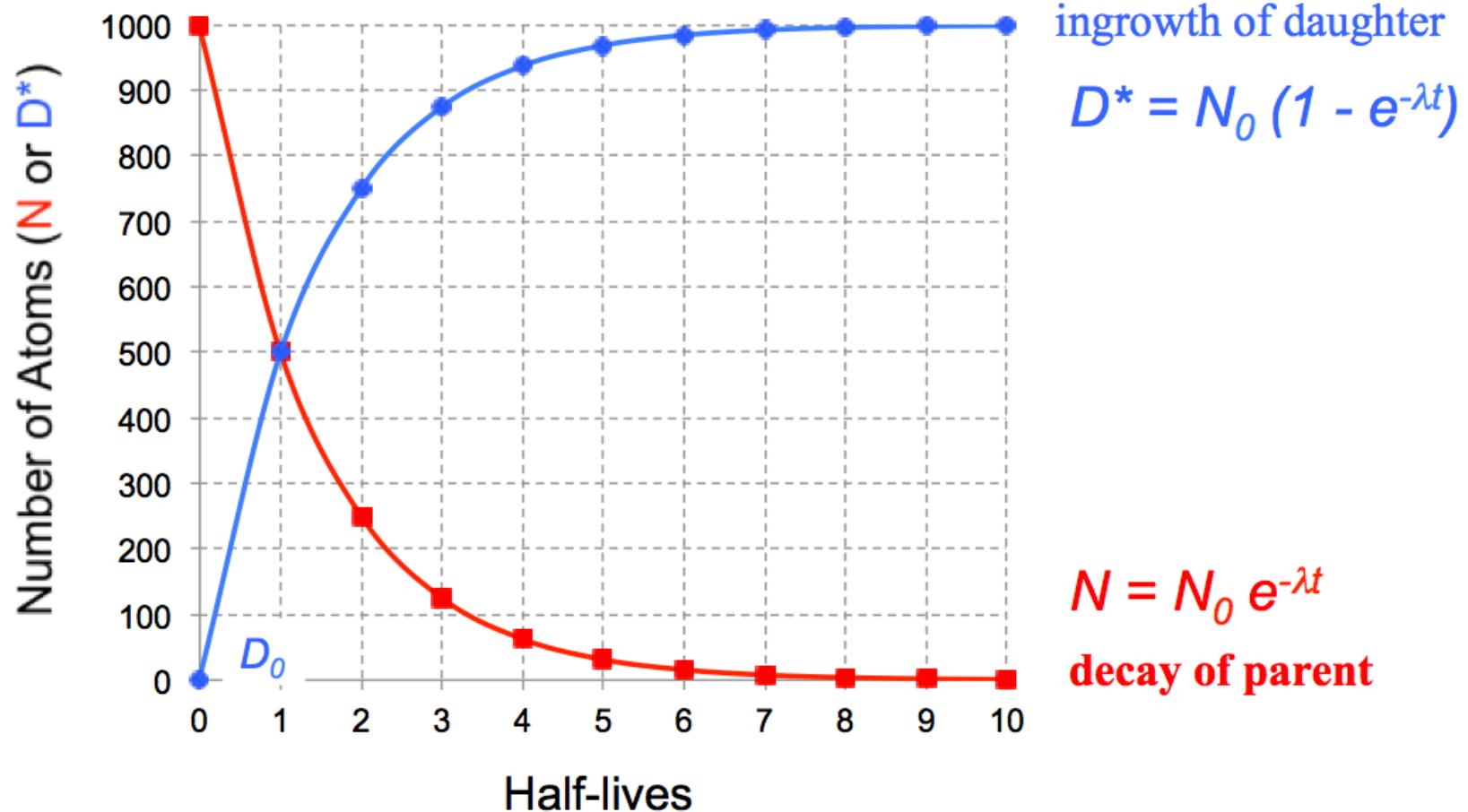
$$D = D_0 + N(e^{\lambda t} - 1)$$

D^* = radiogenic daughter = # daughter atoms produced by radioactive decay of a parent, P ($D = D_0 + D^*$)

Divide by a stable, non-radiogenic isotope of the daughter element to get ratios e.g. for $^{87}\text{Rb} \rightarrow ^{87}\text{Sr} + \beta^-$:

$$^{87}\text{Sr}/^{86}\text{Sr} = (^{87}\text{Sr}/^{86}\text{Sr})_0 + ^{87}\text{Rb}/^{86}\text{Sr} (e^{\lambda t} - 1) \text{ [Example]}$$

Simple Decay: Radioactive Parent \Rightarrow Stable Daughter



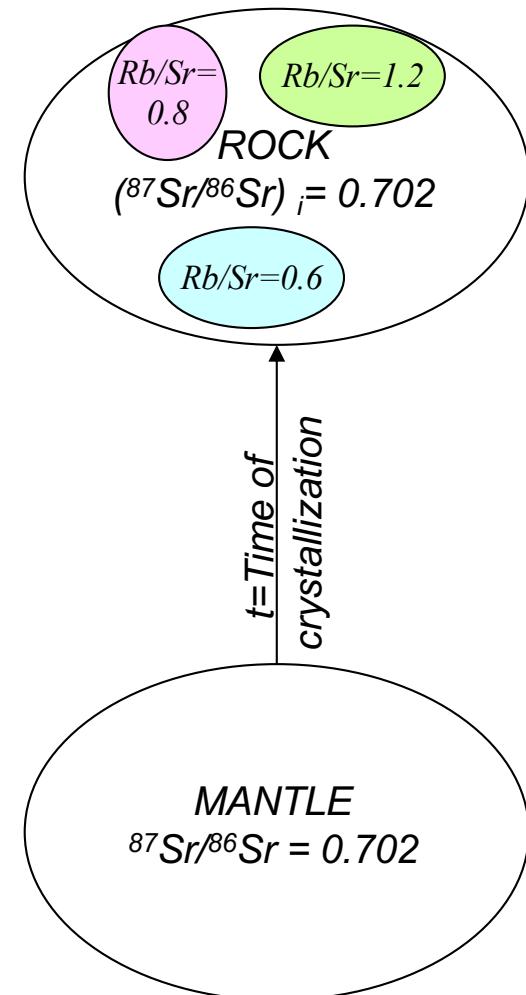
^{87}Rb - ^{87}Sr decay equation (widely used in river geochemistry)

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

measured

measured

*when you crystallize a rock,
you will always have some Sr
present*



Uses of Radiogenic Isotopes

- Age Determination
- Source Identification
- Mixing Calculation

Dating

We can use the radioactive decay equation to calculate the age of a sample. We can measure the present day ratios and λ , but we still have 2 unknowns: R_0 and t . What can we do?

- 1) **Assume zero initial daughter.** This approach can be valid if you know something about mineralogy. For example, zircons are often used for U-Pb dating because the mineral incorporates U, but not lead. Therefore any Pb measured is radiogenic. Similarly, micas accept Rb but not Sr. Another example is K-Ar dating of volcanic rocks. Ar, the daughter, is lost upon eruption, and the only Ar present is radiogenic Ar.
- 2) **Use 2 different isotope systems.** For example, if you know what the age should be from a system where you think you know the initial daughter ratio (e.g. U-Pb), you can calculate the initial daughter ratio.
- 3) **Assume one and calculate the other.** If we assume the initial ratio and calculate an age, it is called a “Model Age”
- 4) **Use an isochron diagram**

The Isochron

The radioactive decay equation is in the form of a line:

$$D = D_0 + P(e^{\lambda t} - 1) \dots y = b + xm$$

Plot D ratio vs. P/D for several comagmatic or cogenetic samples and draw a best fit line through the data

y-intercept = initial D ratio, slope is related to t

This line is called an “**Isochron**”

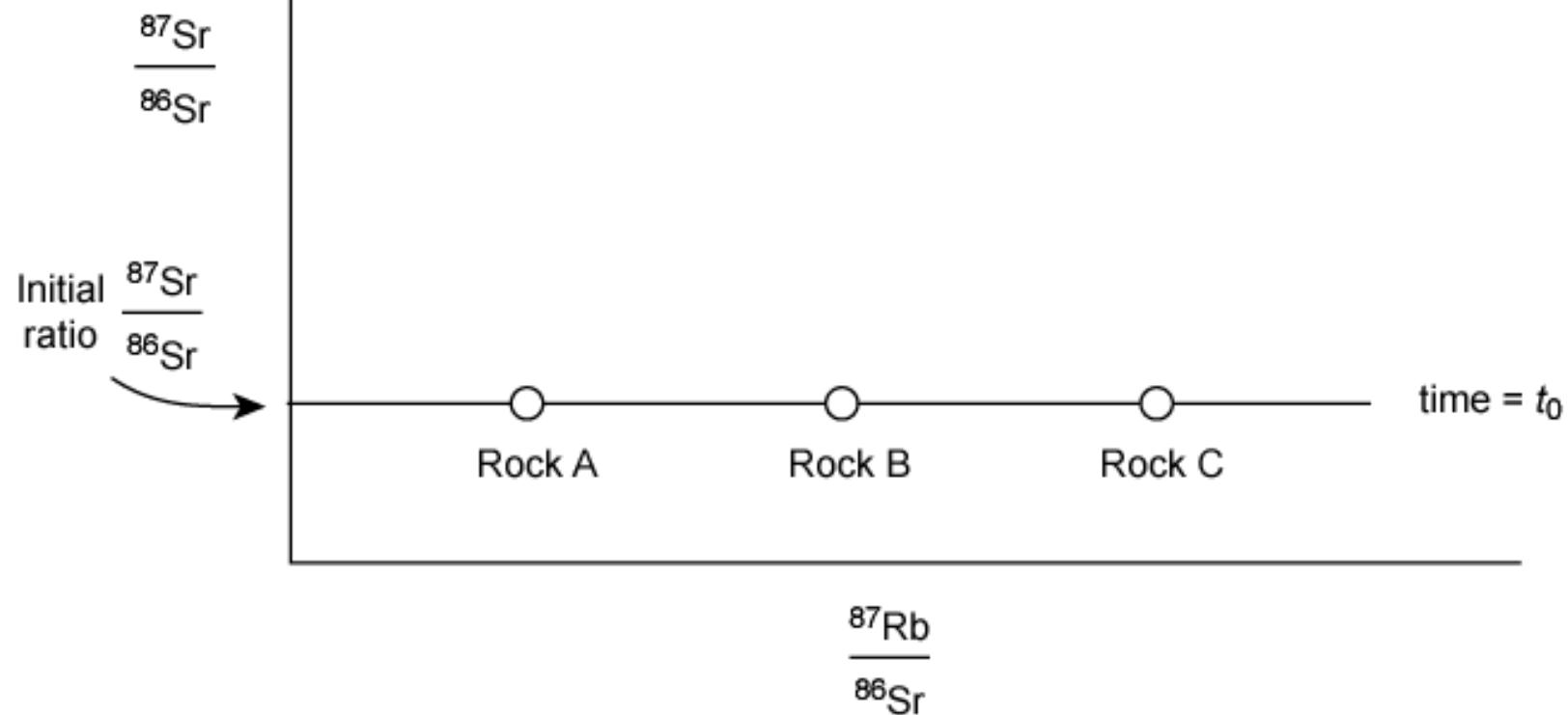
Represents true age if:

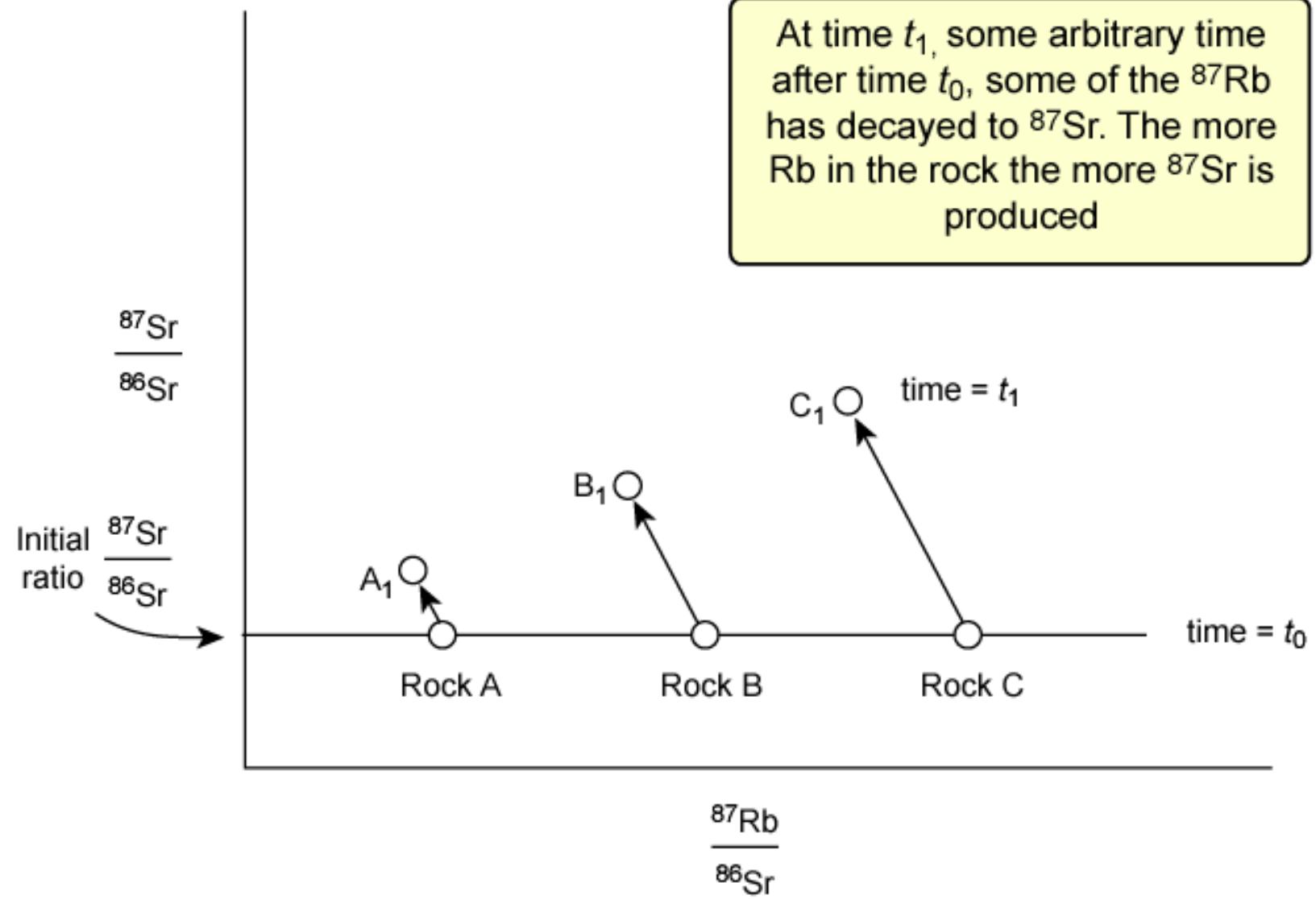
- (1) The system was at isotopic equilibrium at time $t = 0$.
i.e. all the samples formed with the same initial daughter isotope ratio
- (2) Closed system since formation

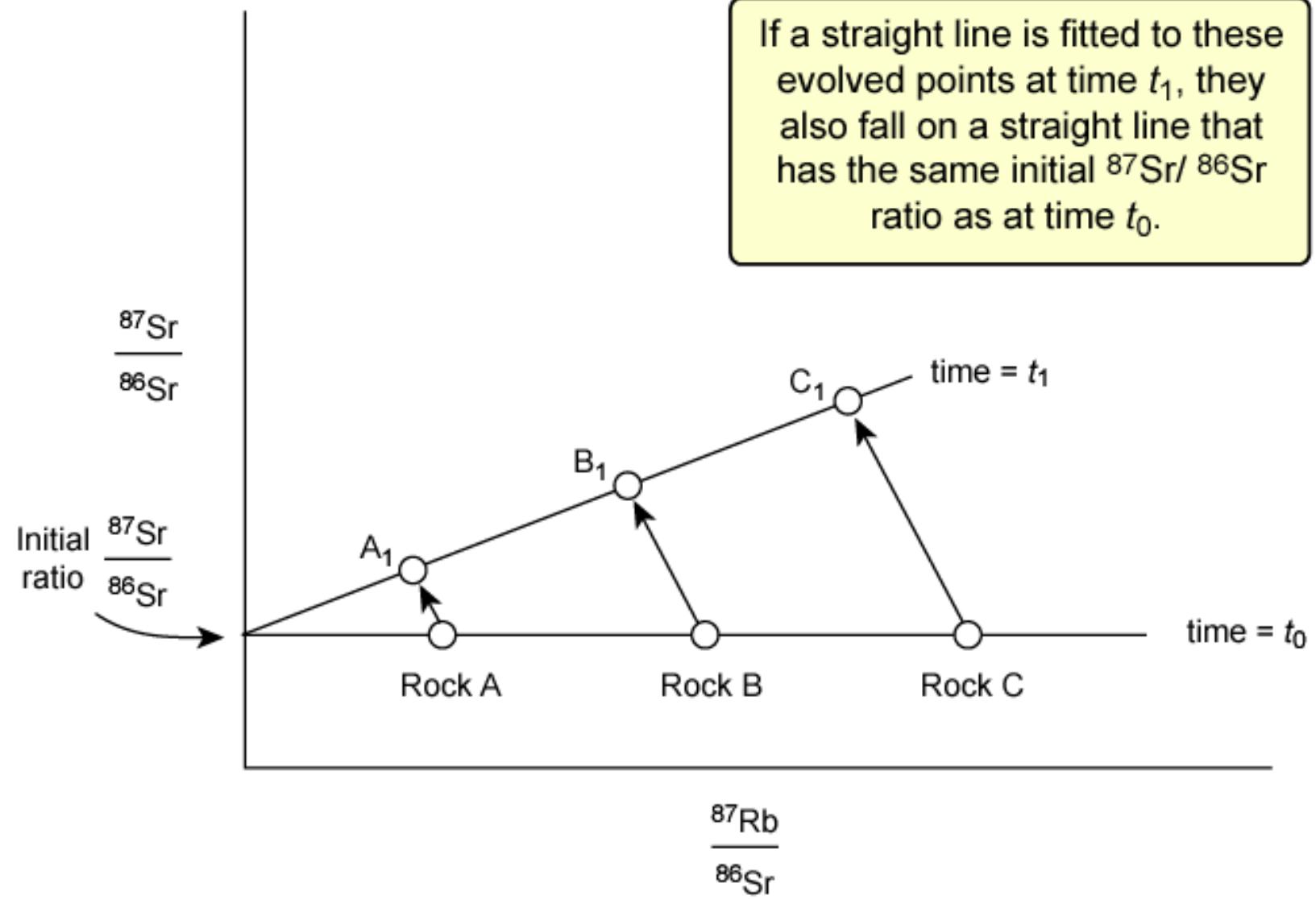
Whole-rock isochron represents age of formation

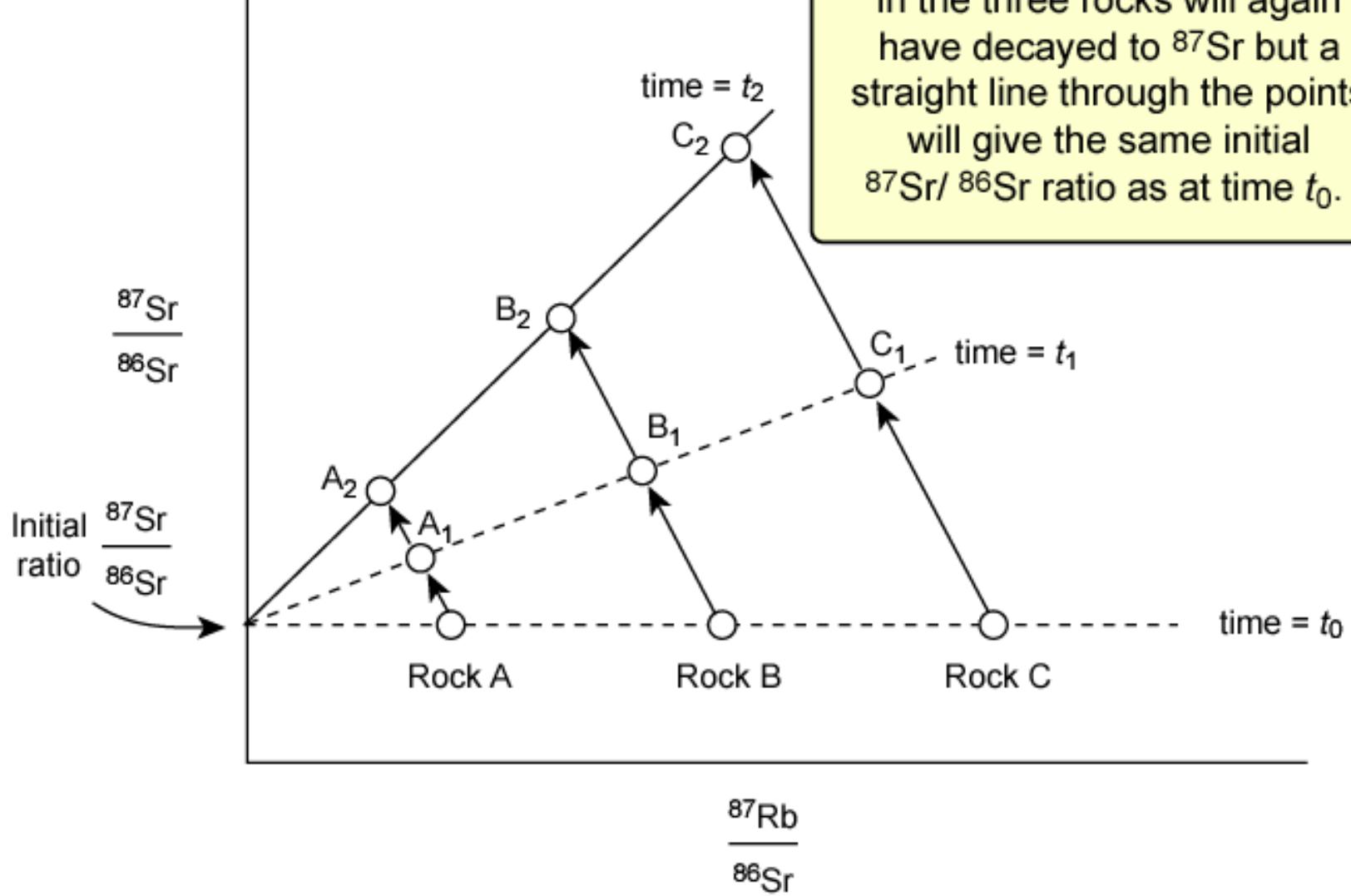
Mineral isochron represents age of last metamorphosis

Start by plotting three rocks with different Rb contents at time t_0



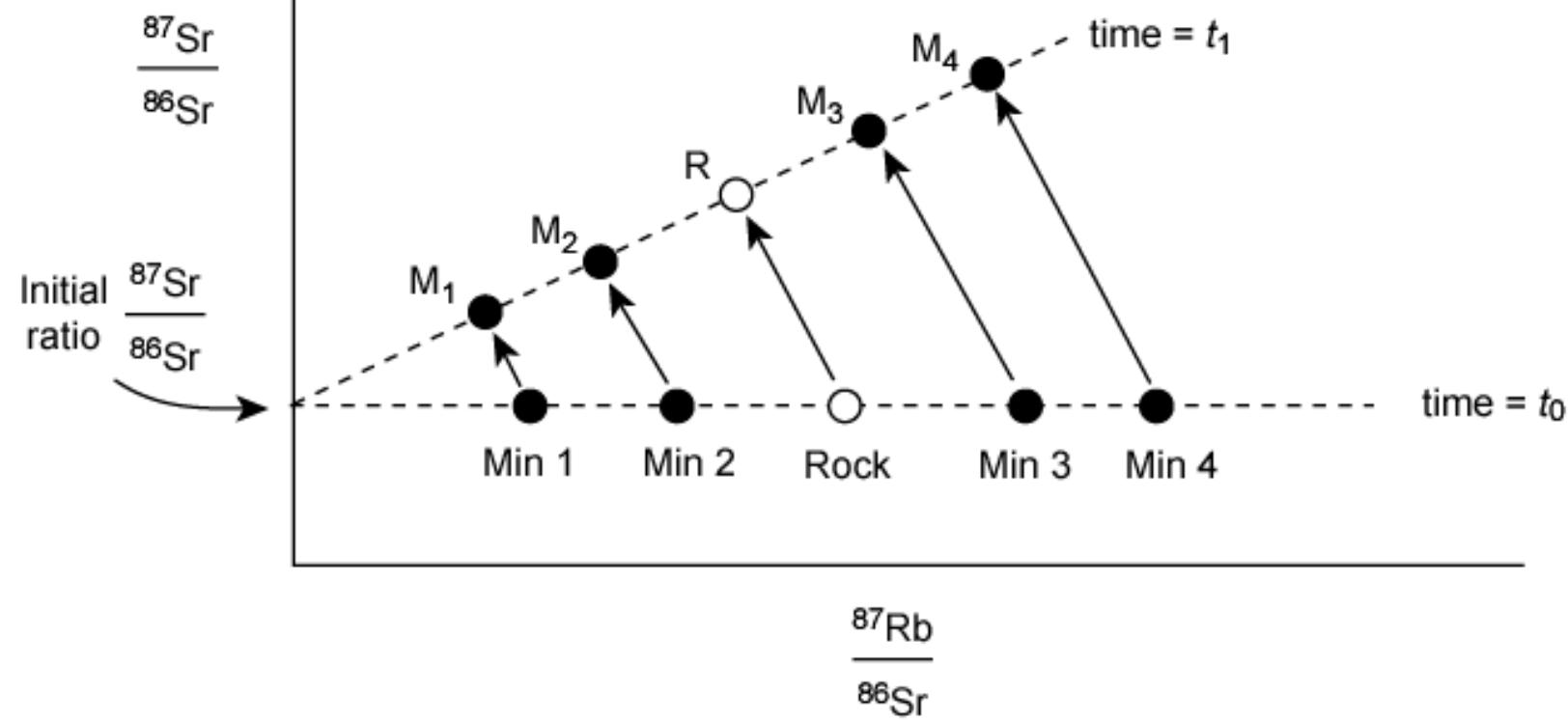






Similarly after time t_2 , the ^{87}Rb in the three rocks will again have decayed to ^{87}Sr but a straight line through the points will give the same initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio as at time t_0 .

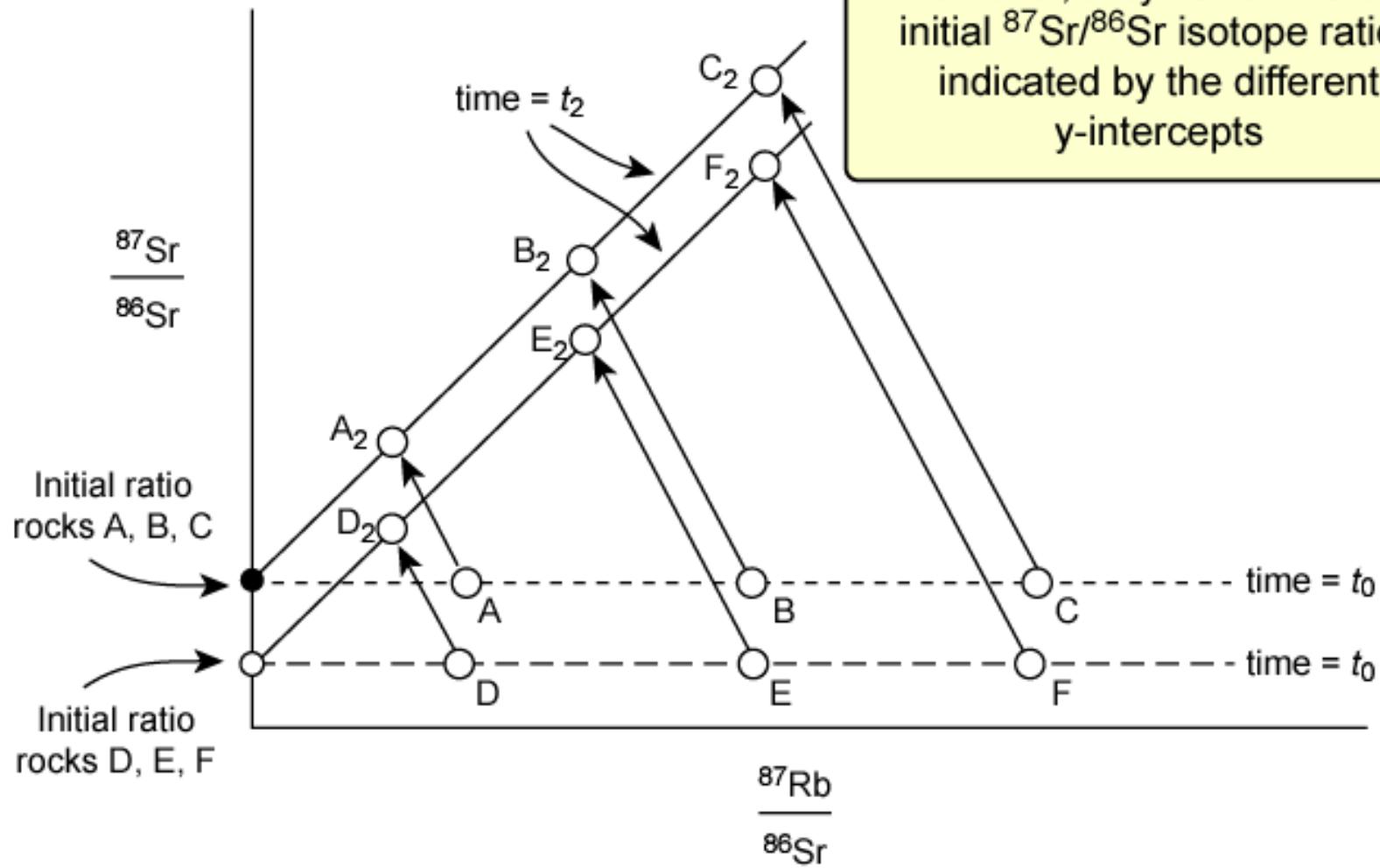
One rock with four different minerals each with different Rb/Sr ratios. Also evolve so that they lie along a steeper but still linear line going through the same initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio

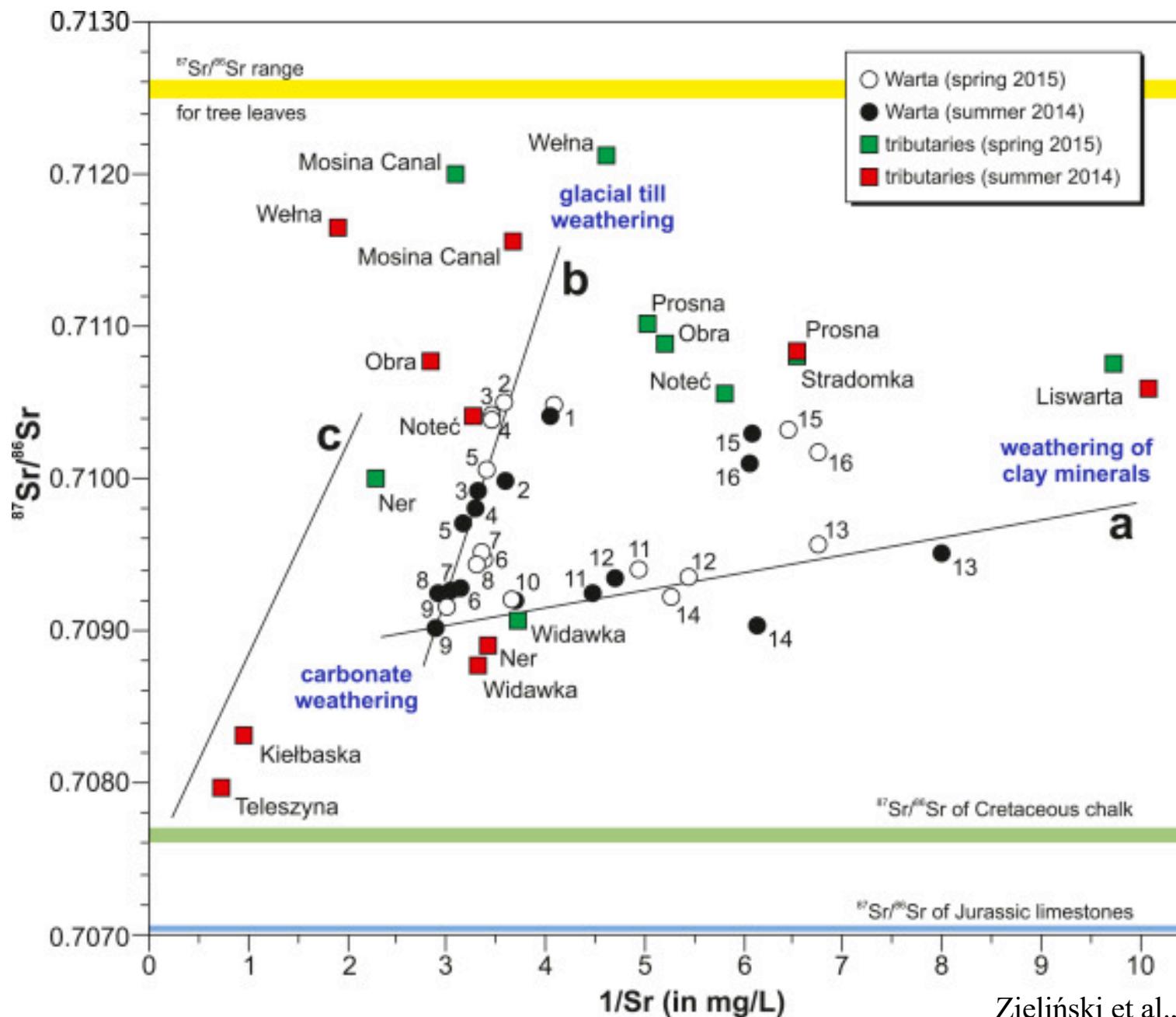


The initial ratio

- How do we know if a series of sediments are co-genetic?
- For sediments to be co-genetic, implies that they are derived from the same parent material.
- This parent material would have had a single $^{87}\text{Sr}/^{86}\text{Sr}$ isotope value, ie the initial isotope ratio
- Therefore, all samples derived from the same parent magma should all have the same $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratio
- If they don't, it implies that they are derived from a different parent source.

Two suites of samples that are not co-genetic





Mixing Theory

- Geochemistry often tries to model variations in measured composition as the result of mixing of a small number of *components* [N.B. a different use of component from the thermodynamic usage] or *end members*
 - This reduces highly multivariate data to a few manageable dimensions
 - It allows identification of the end members with particular source or fluxes, hence a meaningful interpretation of data
 - Many geochemical processes are easily understood in terms of mixing or unmixing:
 - river water + ocean water = mixing
 - primary liquid – fractionated crystals = unmixing
- We will work out the mixing relations for several spaces:
 - Element-element plots
 - Element-ratio plots (including elemental and isotope ratios)
 - Ratio-ratio plots

Mixing Theory

- Mixing is simplest to see and understand when there are only two end members: *Binary mixing*
- For concreteness, instead of a bunch of general symbols, let's do all this with two end members, a and basalt b , with the following compositions:

	a	b
[Sr]	100 ppm	400 ppm
[Nd]	2 ppm	20 ppm
$^{87}\text{Sr}/^{86}\text{Sr}$	0.712	0.704
$^{143}\text{Nd}/^{144}\text{Nd}$	0.511	0.513

- The same relationships will apply for mixing of major elements as for trace elements.
- The same relationships will apply for ratios of major elements, ratios of trace element concentrations, and ratios of isotopes.

Binary Mixing I: element-element

- This is the simplest case. Binary mixing in concentration-concentration space always generates *lines*.
- Let mixtures be generated with mass fraction f_a of end member a and f_b of end member b , such that $f_a + f_b = 1$.
- Then for two species, say Sr and Nd, we have conservation of atoms and mass in the form

$$[\text{Sr}]_{\text{mix}} = f_b[\text{Sr}]_b + (1 - f_b)[\text{Sr}]_a$$

$$[\text{Nd}]_{\text{mix}} = f_b[\text{Nd}]_b + (1 - f_b)[\text{Nd}]_a$$

- This can be written

$$[\text{Sr}]_{\text{mix}} - [\text{Sr}]_a = f_b([\text{Sr}]_b - [\text{Sr}]_a)$$

$$[\text{Nd}]_{\text{mix}} - [\text{Nd}]_a = f_b([\text{Nd}]_b - [\text{Nd}]_a)$$

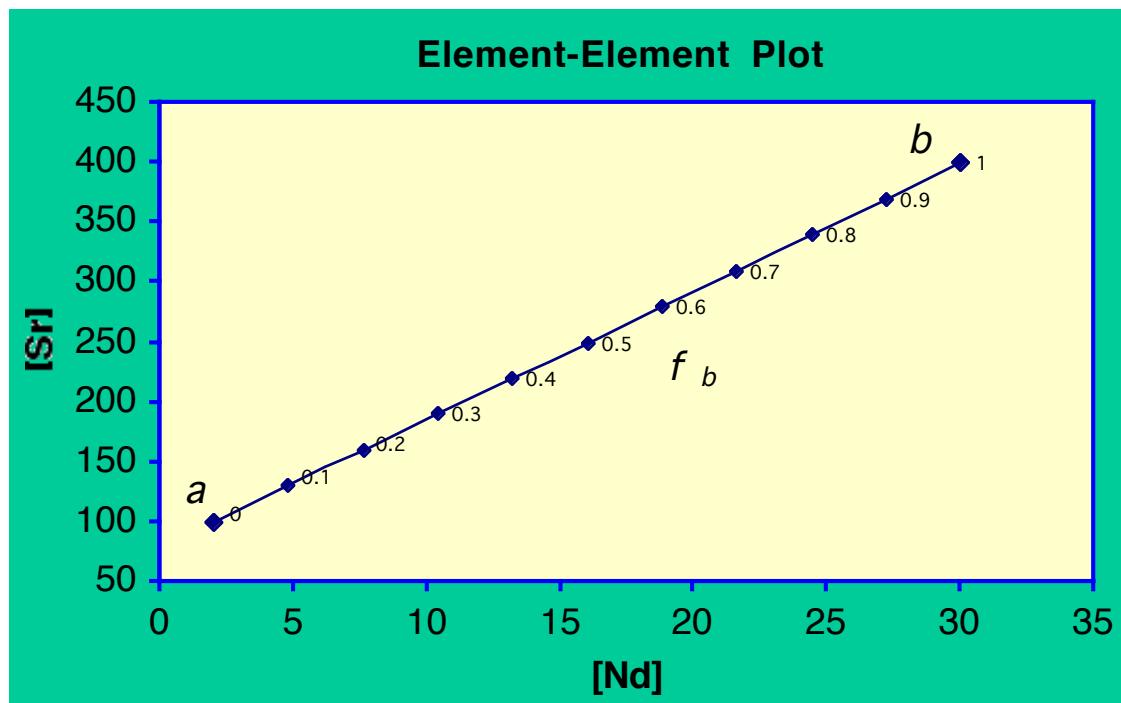
- Dividing these two equations gives the equation of the mixing relationship in $([\text{Sr}], [\text{Nd}])$ space:

$$[\text{Sr}]_{\text{mix}} - [\text{Sr}]_a = \frac{[\text{Sr}]_b - [\text{Sr}]_a}{[\text{Nd}]_b - [\text{Nd}]_a} ([\text{Nd}]_{\text{mix}} - [\text{Nd}]_a)$$

Binary Mixing I: element-element

$$[Sr]_{mix} - [Sr]_a = \frac{[Sr]_b - [Sr]_a}{[Nd]_b - [Nd]_a} ([Nd]_{mix} - [Nd]_a)$$

- This is the equation of a *line* with slope $\frac{[Sr]_b - [Sr]_a}{[Nd]_b - [Nd]_a}$
- Passing through points $([Nd]_a, [Sr]_a)$ and $([Nd]_b, [Sr]_b)$



Binary Mixing I: element-element

- If you *know* the compositions of the end members, you can solve for f_b using the lever rule:

$$f_b = \frac{[Sr]_{mix} - [Sr]_a}{[Sr]_b - [Sr]_a} = \frac{[Nd]_{mix} - [Nd]_a}{[Nd]_b - [Nd]_a}$$

- If you *don't know* the end members, but only the data, what can you learn from a graph showing a linear correlation?
 - You can infer that if generated by mixing there are only two end members, otherwise the data would fill a triangle
 - You can infer that both end members lie on the mixing line, outside the extreme range of the data on both ends if they must have positive amounts (*additive mixing*)

Binary Mixing II: Ratio

- In geochemistry we very frequently work with ratios, either isotope ratios or ratios of concentrations.
 - Sometimes a ratio is all you can measure accurately
 - Sometimes ratios have significance where concentrations are more or less arbitrary (example: during fractionation of olivine from a basalt, [Sr] will change because it is incompatible and the amount of liquid is decreasing, but [Sr]/[Nd] will not)
- You might think that the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of mixtures could be obtained in the same way as for [Sr]

$$(^{87}\text{Sr}/^{86}\text{Sr})_{\text{mix}} = f_b (^{87}\text{Sr}/^{86}\text{Sr})_b + (1 - f_b) (^{87}\text{Sr}/^{86}\text{Sr})_a$$

- You would be wrong!
- The isotope ratio of the mixture is going to be weighted by the concentration of Sr in each end member
 - More generally, the weighting of ratios in the mixture is controlled by the denominator of the ratio, in this case ^{86}Sr .

Binary Mixing II: Ratio

- Let's do it right: we have

$$[{}^{87}\text{Sr}]_{mix} = f_b [{}^{87}\text{Sr}]_b + (1 - f_b) [{}^{87}\text{Sr}]_a$$

$$[{}^{86}\text{Sr}]_{mix} = f_b [{}^{86}\text{Sr}]_b + (1 - f_b) [{}^{86}\text{Sr}]_a$$

- Taking the ratio of these, we have

$$\left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}}\right)_{mix} = \frac{f_b \left[\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}}\right]_b + (1 - f_b) \left[\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}}\right]_a}{f_b \left[\frac{{}^{86}\text{Sr}}{{}^{86}\text{Sr}}\right]_b + (1 - f_b) \left[\frac{{}^{86}\text{Sr}}{{}^{86}\text{Sr}}\right]_a}$$

- And substituting $[{}^{87}\text{Sr}] = ({}^{87}\text{Sr}/{}^{86}\text{Sr})[{}^{86}\text{Sr}]$ for a and b :

$$\left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}}\right)_{mix} = \frac{f_b \left[{}^{86}\text{Sr}\right]_b \left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}}\right)_b + (1 - f_b) \left[{}^{86}\text{Sr}\right]_a \left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}}\right)_a}{f_b \left[{}^{86}\text{Sr}\right]_b + (1 - f_b) \left[{}^{86}\text{Sr}\right]_a}$$

- We can approximate this using $[\text{Sr}]$ instead of $[{}^{86}\text{Sr}]$ as the weighting factors.

Binary Mixing II: Ratio

- Now let's consider plotting the isotope ratio of the mixture against an elemental concentration, perhaps $[Nd]$.
- If we eliminate f_b between the mixing equation for $(^{87}\text{Sr}/^{86}\text{Sr})$ and $[Nd]$, we obtain the following equation (using the approximation $\text{Sr} \sim ^{86}\text{Sr}$):

$$A\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{mix} + B[Nd]_{mix}\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{mix} + C[Nd]_{mix} + D = 0$$

$$A = [Nd]_b[\text{Sr}]_a - [Nd]_a[\text{Sr}]_b$$

$$B = [\text{Sr}]_b - [\text{Sr}]_a$$

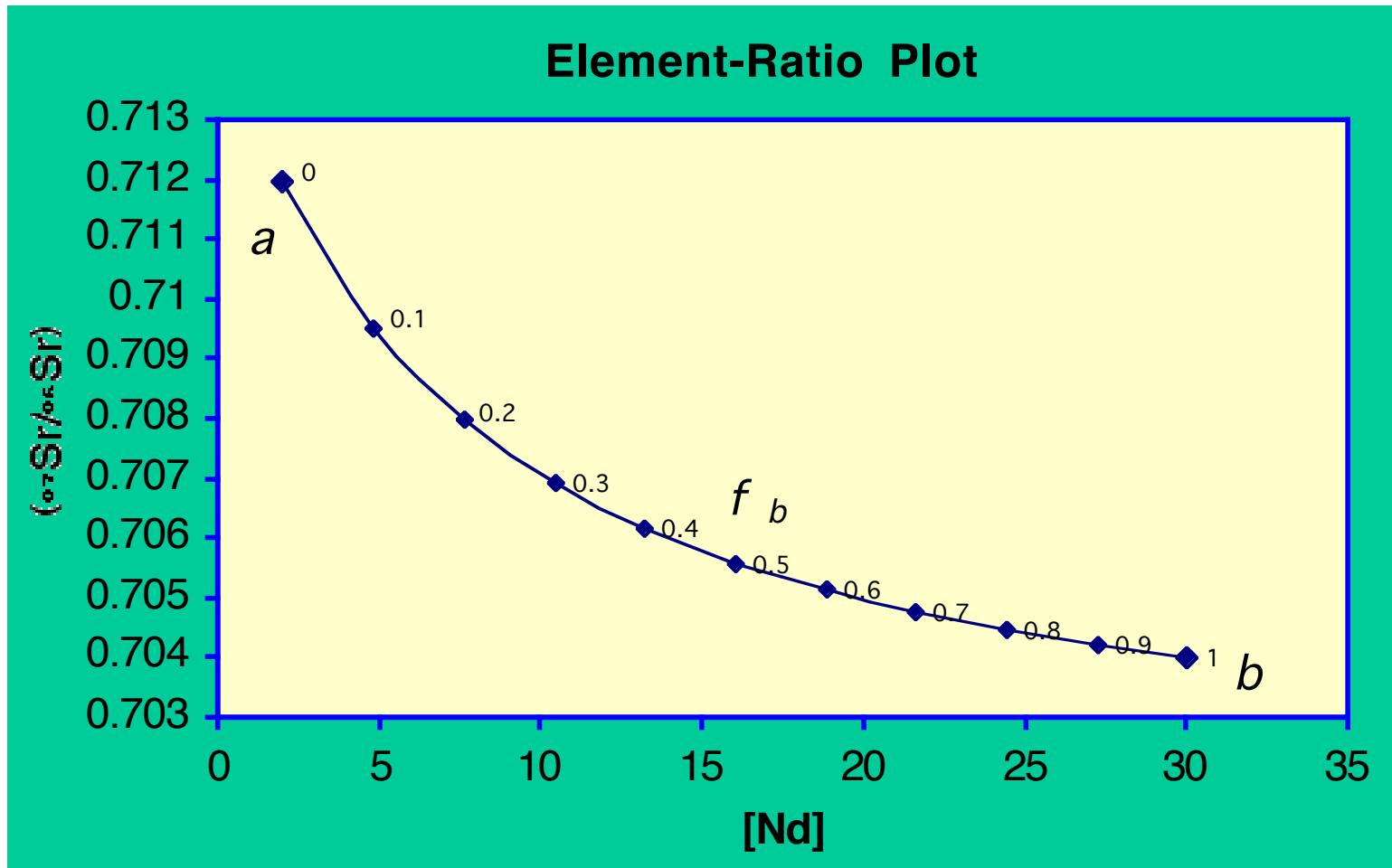
$$C = [\text{Sr}]_a\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_a - [\text{Sr}]_b\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_b$$

$$D = [Nd]_a[\text{Sr}]_b\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_b - [Nd]_b[\text{Sr}]_a\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_a$$

- Equation $Ax + Bxy + Cy + D = 0$?

Binary Mixing II: element-ratio

- In general, element-ratio mixing generates a *hyperbola*.
 - The only case in which it is linear is $B = [\text{Sr}]_b - [\text{Sr}]_a = 0$



- The index of curvature $r = [\text{Sr}]_b / [\text{Sr}]_a$ tells you “how hyperbolic” the hyperbola is going to be.

Binary Mixing III: Inverse Element-Ratio

- Although there is nothing special about the element-ratio case A/B vs. B (thus $(^{87}\text{Sr}/^{86}\text{Sr})$ vs. [Sr] is still hyperbolic), there is an especially useful test for mixing if you plot A/B vs. 1/B (in this case, $(^{87}\text{Sr}/^{86}\text{Sr})$ vs. 1/[Sr]).
 – Going back to our hyperbolic equation, if we replace $[\text{Nd}]_{mix}$, $[\text{Nd}]_a$ and $[\text{Nd}]_b$ with $[\text{Sr}]_{mix}$, $[\text{Sr}]_a$ and $[\text{Sr}]_b$ we have:

$$A\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{mix} + B[\text{Sr}]_{mix}\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{mix} + C[\text{Sr}]_{mix} + D = 0$$

$$A = [\text{Sr}]_b[\text{Sr}]_a - [\text{Sr}]_a[\text{Sr}]_b = 0, B = [\text{Sr}]_b - [\text{Sr}]_a$$

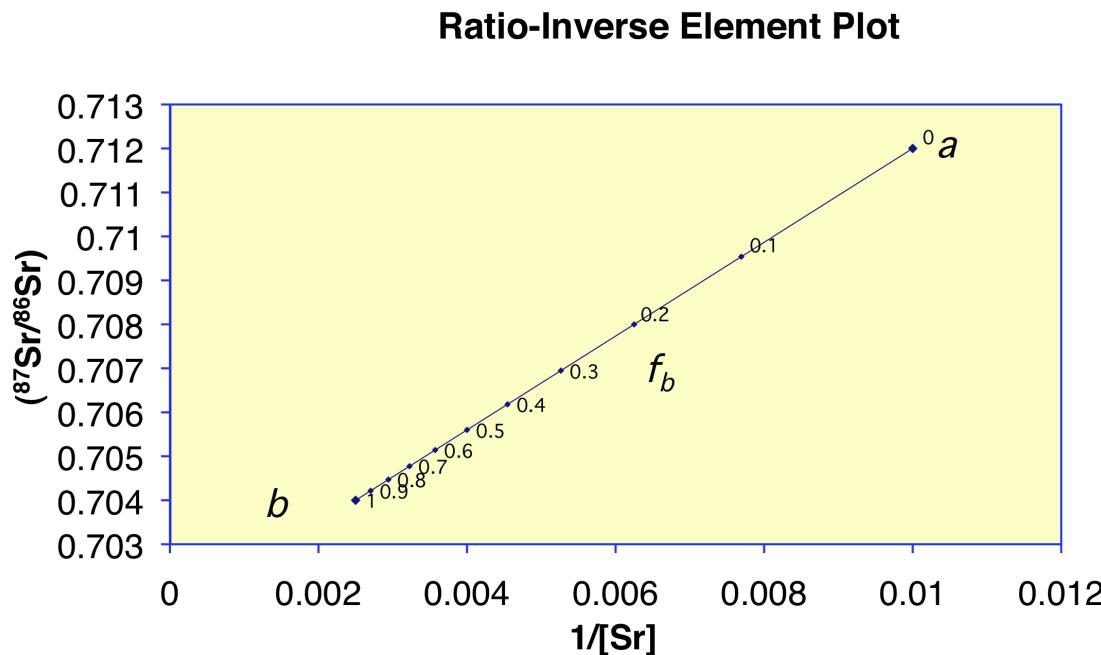
$$C = [\text{Sr}]_a\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_a - [\text{Sr}]_b\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_b$$

$$D = [\text{Sr}]_a[\text{Sr}]_b\left(\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_b - \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_a\right)$$

- Or, $B\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{mix} + C + D\frac{1}{[\text{Sr}]_{mix}} = 0$, which is a *line*.

Binary Mixing III: inverse element-ratio

- Mixing in A/B vs. 1/B space always generates a line.



- The value of $r = [\text{Sr}]_b/[\text{Sr}]_a$ now controls how hyperbolic the *spacing* of equal increments of mixing fraction are along the line. Since linear correlation is easy to calculate, it is much easier to test whether data are consistent with binary mixing in this space than in ratio-element space.

Binary Mixing IV: Ratio-Ratio

- Our final case is plots of ratios against ratios, whether isotope ratios, trace element ratios, or major element ratios.
- For example, let's do $(^{87}\text{Sr}/^{86}\text{Sr})$ vs. $(^{143}\text{Nd}/^{144}\text{Nd})$.
- We now have two equations of the same form:

$$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{mix} = \frac{f_b[\text{Sr}]_b \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_b + (1-f_b)[\text{Sr}]_a \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_a}{f_b[\text{Sr}]_b + (1-f_b)[\text{Sr}]_a}$$

$$\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_{mix} = \frac{f_b[\text{Nd}]_b \left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_b + (1-f_b)[\text{Nd}]_a \left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_a}{f_b[\text{Nd}]_b + (1-f_b)[\text{Nd}]_a}$$

Where once again for the particular case of small variations in isotope ratios I have weighted by concentration rather than by the stable denominator isotope.

Binary Mixing IV: ratio-ratio

- This time eliminating f_b between the mixing equations gives

$$A \left(\frac{87\text{Sr}}{86\text{Sr}} \right)_{mix} + B \left(\frac{87\text{Sr}}{86\text{Sr}} \right)_{mix} \left(\frac{143\text{Nd}}{144\text{Nd}} \right)_{mix} + C \left(\frac{143\text{Nd}}{144\text{Nd}} \right)_{mix} + D = 0$$

$$A = [\text{Nd}]_b[\text{Sr}]_a \left(\frac{143\text{Nd}}{144\text{Nd}} \right)_b - [\text{Nd}]_a[\text{Sr}]_b \left(\frac{143\text{Nd}}{144\text{Nd}} \right)_a$$

$$B = [\text{Nd}]_a[\text{Sr}]_b - [\text{Nd}]_b[\text{Sr}]_a$$

$$C = [\text{Nd}]_b[\text{Sr}]_a \left(\frac{87\text{Sr}}{86\text{Sr}} \right)_a - [\text{Nd}]_a[\text{Sr}]_b \left(\frac{87\text{Sr}}{86\text{Sr}} \right)_b$$

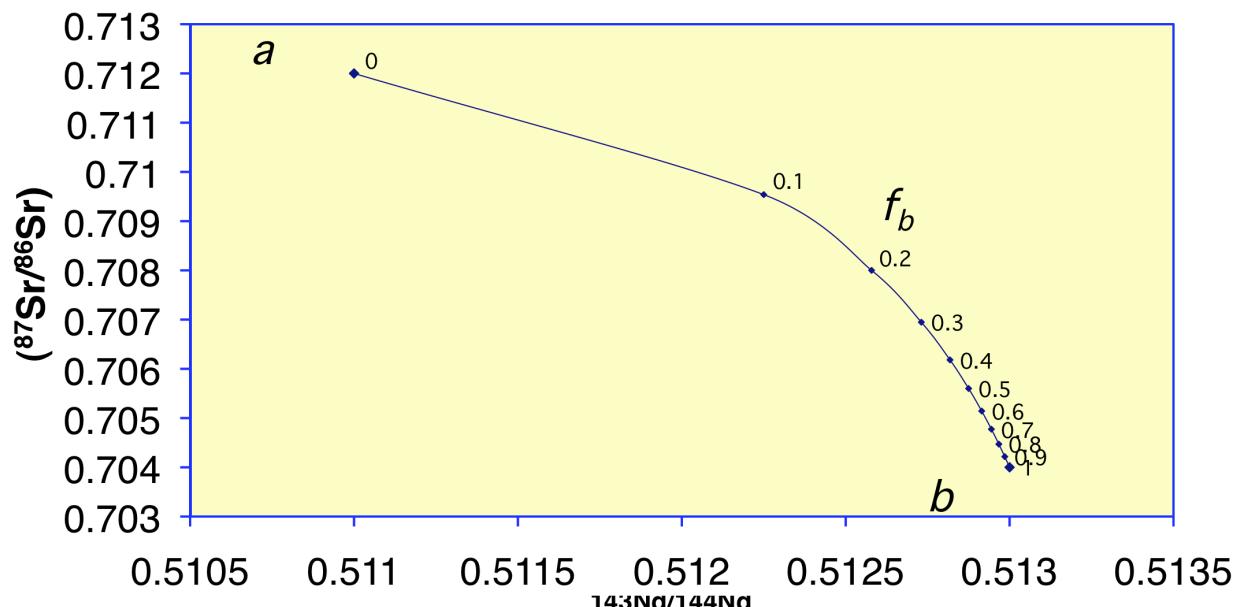
$$D = [\text{Nd}]_a[\text{Sr}]_b \left(\frac{143\text{Nd}}{144\text{Nd}} \right)_a \left(\frac{87\text{Sr}}{86\text{Sr}} \right)_b - [\text{Nd}]_b[\text{Sr}]_a \left(\frac{143\text{Nd}}{144\text{Nd}} \right)_b \left(\frac{87\text{Sr}}{86\text{Sr}} \right)_a$$

- Still a hyperbola. Now term B gives the curvature index
- $r = ([\text{Sr}]_a / [\text{Sr}]_b) / ([\text{Nd}]_a / [\text{Nd}]_b)$

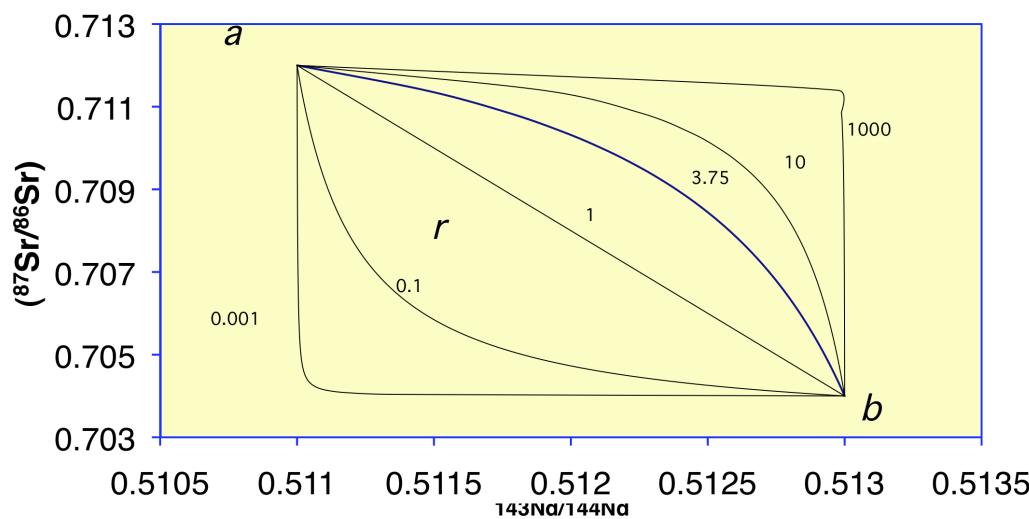
Binary Mixing IV: ratio-ratio

This looks different from the element-ratio hyperbola because now the *spacing* of equal increments of mixing fraction is no longer regular.

Ratio-Ratio Plot

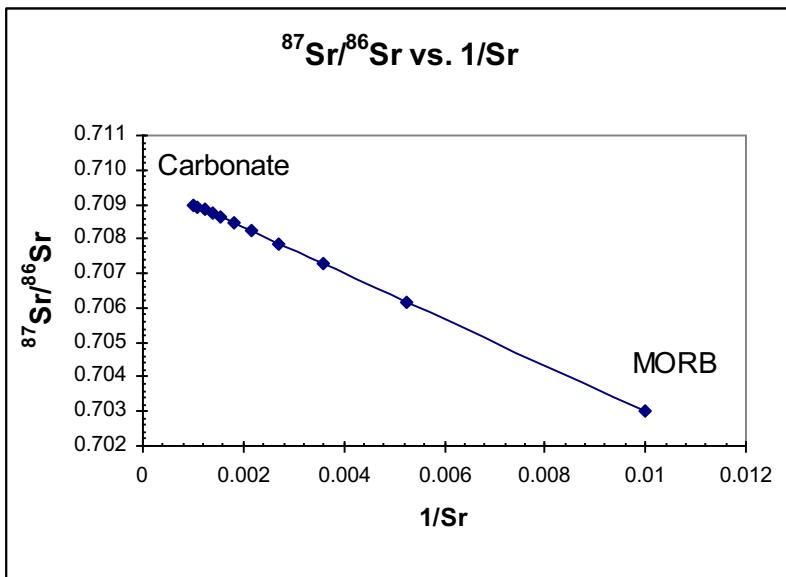
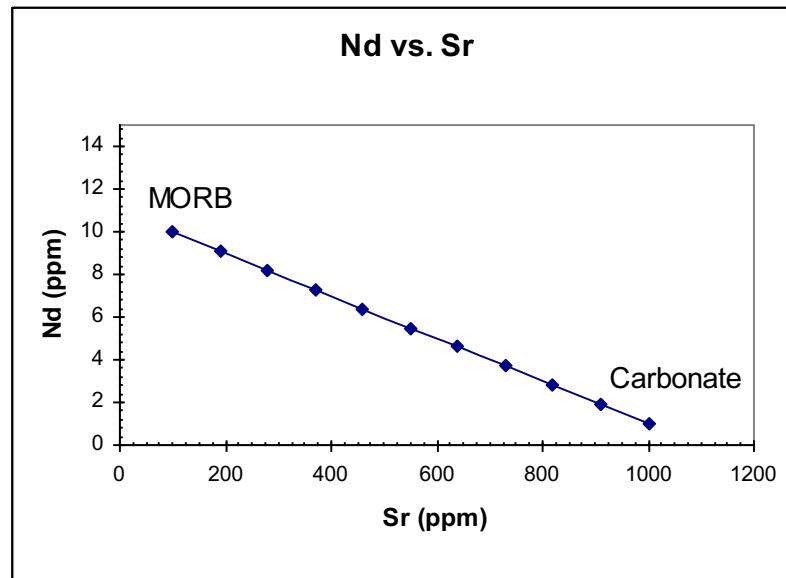
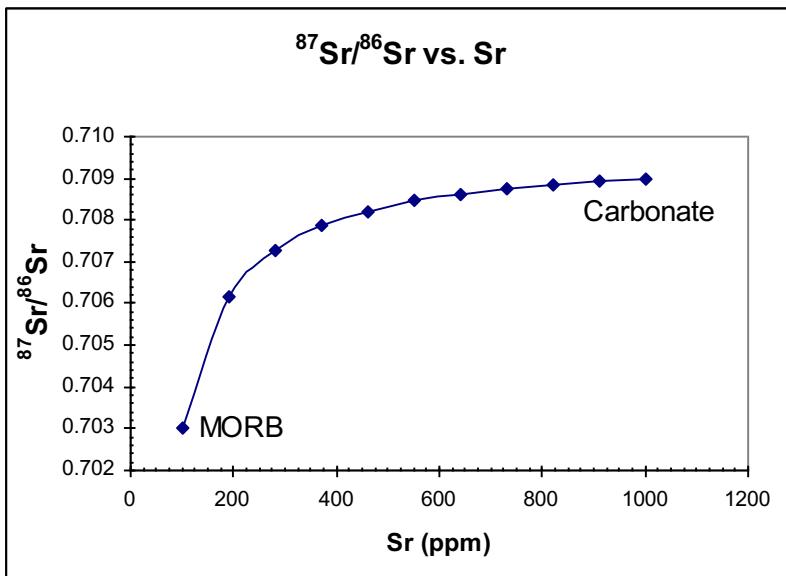


Ratio-Ratio Plot



Given an array of ratio-ratio data, you can constrain the curvature parameter as well as the isotope ratios of the end members.

Some mixing plots



RADIOACTIVE DECAY

Radioactive Isotopes

Sm Radioactive (Parent)

Os Radiogenic (Daughter)

Rd Radiogenic and Radioactive

H																			He
Li	Be																		
Na	Mg																		
K	Ca	Sc	Ti	V	Cr	Mn	Fe	Co	Ni	Cu	Zn	Ga	Ge	As	Se	Br		Kr	
Rb	Sr	Y	Zr	Nb	Mo	Tc	Ru	Rh	Pd	Ag	Cd	In	Sn	Sb	Te	I	Xe		
Cs	Ba	La	Hf	Ta	W	Re	Os	Ir	Pt	Au	Hg	Tl	Pb	Bi	Po	At	Rd		
Fr	Ra	Ac																	
		La	Ce	Pr	Nd	Pm	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu			
		Ac	Th	Pa	U														

- two extremely interesting and important aspects of radioactive decay makes it so useful as a chronometer.
 - Unstable nuclei decay to stable ones at **rates independent of all environmental influences**.
 - Each nucleus has a **fixed probability of decaying** per unit time. Nothing affects this probability (e.g., temperature, pressure, bonding environment, etc.)

Radiogenic Isotopic variations is a function of:

1. Time (of decay since the system is closed)
2. Different parent/Daughters ratio

Radioactive Decay

- Basic equation of radioactive decay: first-order rate law: **Curie-Rutherford-Soddy Law:**

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

The minus sign simply indicates N (present-day parent concentration) decreases.

λ is the decay constant- probability of decay per unit time . Unit: time⁻¹.

- Integrating the decay equation, we get:

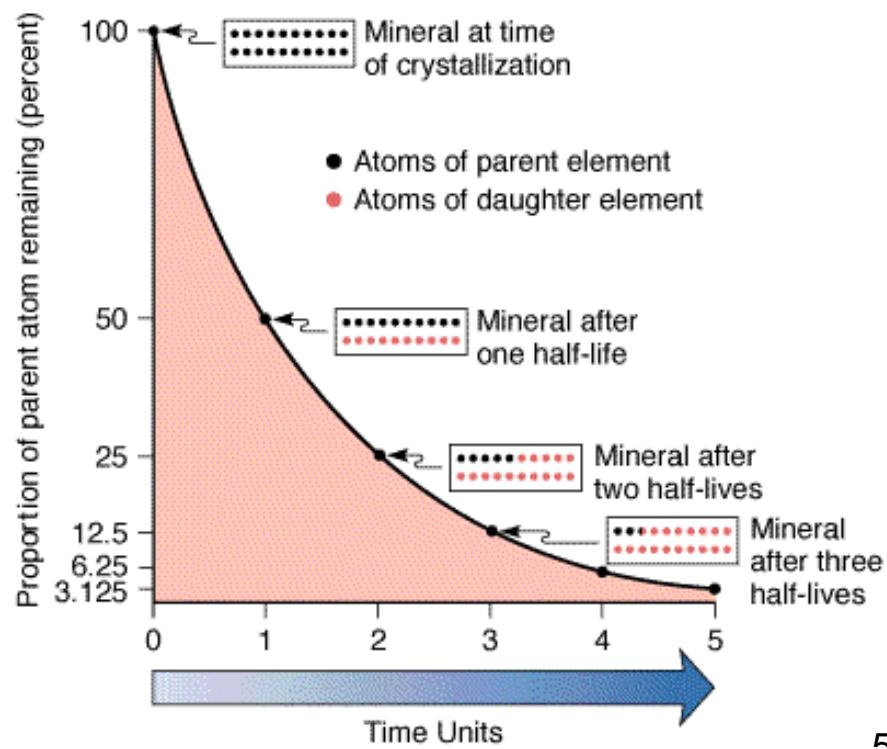
$$N = N_0 e^{-\lambda t}$$

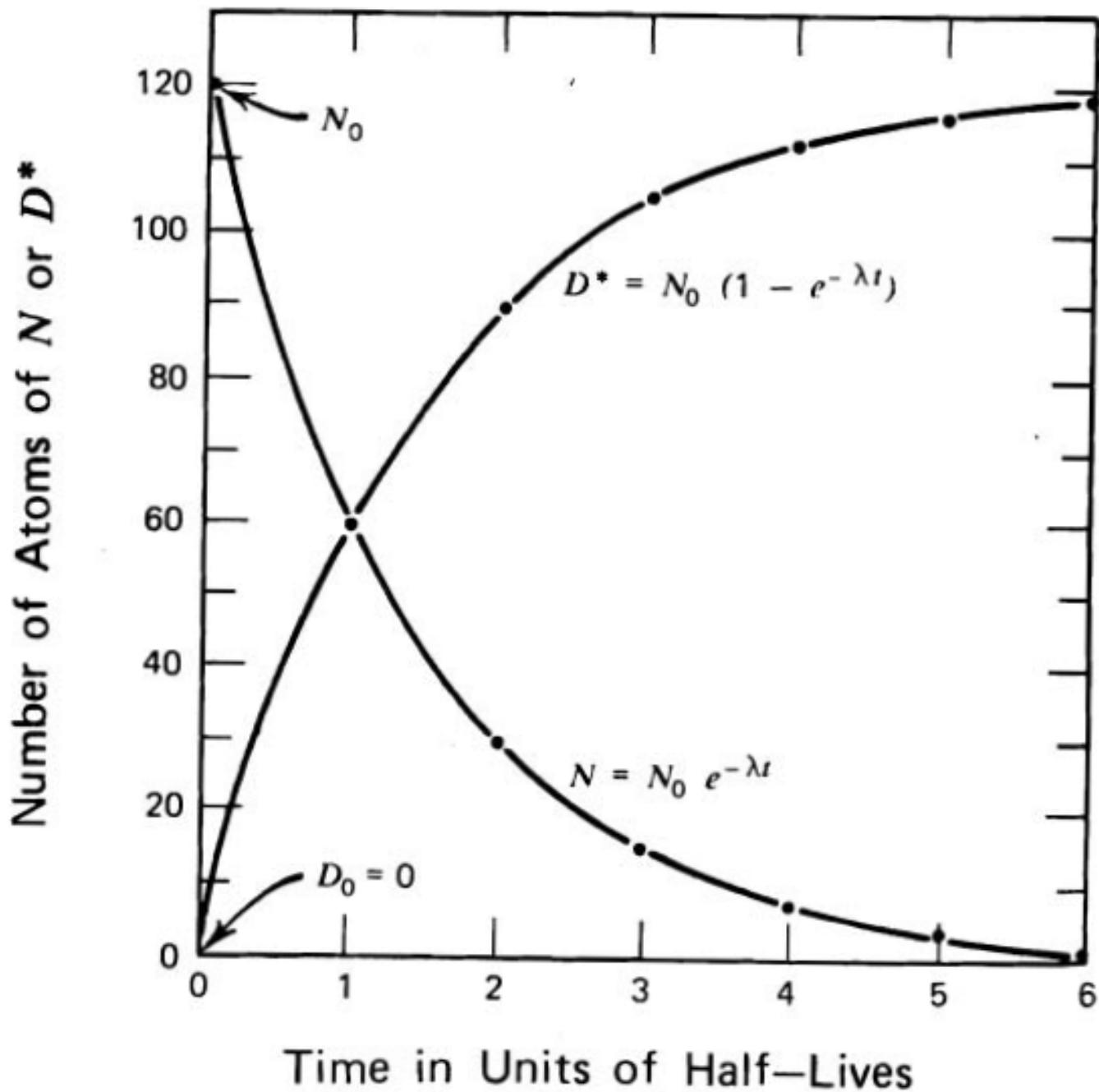
Half-life: $t_{1/2} = \frac{\ln 2}{\lambda} = \frac{0.693}{\lambda}$

The **mean life** τ of a parent nuclide:
(average life expectancy of a radioactive atom)

$$t\tau = -\frac{1}{N_0} \int_{t=0}^{t=\infty} t \cdot dN$$

$$\tau = \frac{N}{\lambda N} = \frac{1}{\lambda}$$





The decay of the parent produces a daughter (D^*), or **radiogenic nuclide**.

$$D^* = N_o - N$$

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

- However, there might be some daughter isotope present in the system (initial value) to begin with.

Therefore: $D = D_o + D^*$

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

D = number of daughter atoms

N = number of existing parent atoms

D_o = number of initial daughter atoms

λ = decay constant

t = time elapsed

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

Note that this equation is **independent of N_0** .

solve the above equation for age of the system (t):

$$t = \frac{1}{\lambda} \ln \left[\frac{D - D_0}{N} + 1 \right]$$

Practical limitations on age range:

Very young rocks: cannot measure tiny amount of daughter accurately

Very old rocks: cannot measure tiny amounts of parent left accurately

Applicability of an Isotopic system depends on λ .

Geologically Useful Long-Lived Radioactive Decay Schemes

TABLE 2.1: Geologically Useful Long-Lived Radioactive Decay Schemes

Parent	Decay Mode	λ	Half-life	Daughter	Ratio
^{40}K	β^- , e.c., β^+	$5.5492 \times 10^{-10} \text{y}^{-1}$ *	$1.28 \times 10^9 \text{yr}$	^{40}Ar , ^{40}Ca	$^{40}\text{Ar}/^{36}\text{Ar}$
^{87}Rb	β^-	$1.42 \times 10^{-11} \text{y}^{-1}$ ^f	$48.8 \times 10^9 \text{yr}$	^{87}Sr	$^{87}\text{Sr}/^{86}\text{Sr}$
^{138}La	β^-	$2.67 \times 10^{-12} \text{y}^{-1}$	$2.59 \times 10^{11} \text{yr}$	^{138}Ce , ^{138}Ba	$^{138}\text{Ce}/^{142}\text{Ce}$, $^{138}\text{Ce}/^{136}\text{Ce}$
^{147}Sm	α	$6.54 \times 10^{-12} \text{y}^{-1}$	$1.06 \times 10^{11} \text{yr}$	^{143}Nd	$^{143}\text{Nd}/^{144}\text{Nd}$
^{176}Lu	β^-	$1.867^t \times 10^{-11} \text{y}^{-1}$	$3.6 \times 10^{10} \text{yr}$	^{176}Hf	$^{176}\text{Hf}/^{177}\text{Hf}$
^{187}Re	β^-	$1.64 \times 10^{-11} \text{y}^{-1}$	$4.23 \times 10^{10} \text{yr}$	^{187}Os	$^{187}\text{Os}/^{188}\text{Os}$, ($^{187}\text{Os}/^{186}\text{Os}$)
^{190}Pt	α	$1.54 \times 10^{-12} \text{y}^{-1}$	$4.50 \times 10^{11} \text{yr}$	^{186}Os	$^{186}\text{Os}/^{188}\text{Os}$
^{232}Th	α	$4.948 \times 10^{-11} \text{y}^{-1}$	$1.4 \times 10^{10} \text{yr}$	^{208}Pb , ^4He	$^{208}\text{Pb}/^{204}\text{Pb}$, $^3\text{He}/^4\text{He}$
^{235}U	α	$9.8571 \times 10^{-10} \text{y}^{-1}$ [‡]	$7.07 \times 10^8 \text{yr}$	^{207}Pb , ^4He	$^{207}\text{Pb}/^{204}\text{Pb}$, $^3\text{He}/^4\text{He}$
^{238}U	α	$1.55125 \times 10^{-10} \text{y}^{-1}$	$4.47 \times 10^9 \text{yr}$	^{206}Pb , ^4He	$^{206}\text{Pb}/^{204}\text{Pb}$, $^3\text{He}/^4\text{He}$

Note: the branching ratio, i.e. ratios of decays to ^{40}Ar to total decays of ^{40}K is 0.117. ^{147}Sm and ^{190}Pt also produce ^4He , but a trivial amount compared to U and Th.

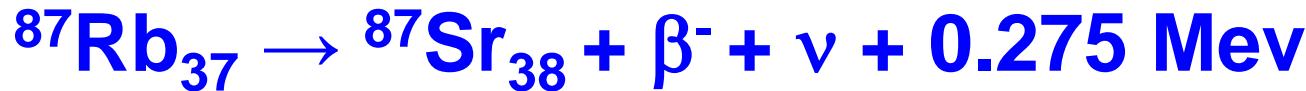
*This is the value recently suggested by Renne et al. (2010). The conventional value is $5.543 \times 10^{-10} \text{y}^{-1}$

^fThe officially accepted decay constant for ^{87}Rb is that shown here. However, recent determinations of this constant range from $1.421 \times 10^{-11} \text{y}^{-1}$ by Rotenberg (2005) to $1.399 \times 10^{-11} \text{y}^{-1}$ by Nebel et al. (2006).

^tThis is the value recommended by Söderlund et al. (2004).

[‡]Value suggested by Mattinson (2010). The conventional value is $9.8485 \times 10^{-10} \text{y}^{-1}$.

Isochron Dating



$$(\lambda = 1.42 \times 10^{-11} \text{ yr}^{-1}; T_{1/2} = 48.8 \text{ Ga})$$

$$^{87}\text{Sr} = ^{87}\text{Sr}_0 + ^{87}\text{Rb}(e^{\lambda t} - 1)$$

As it turns out, it is generally much easier, and usually more meaningful, to measure ratio of two isotopes precisely than the absolute abundance of one. We, therefore, measure the ratio of ^{87}Sr to a non-radioactive isotope, which by convention is ^{86}Sr . We can recast the above equation as:

$$\frac{^{87}\text{Sr}}{^{86}\text{Sr}} = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_0 + \frac{^{87}\text{Rb}}{^{86}\text{Sr}} (e^{\lambda t} - 1)$$

Measured for today

Estimated from isochrons

Measured for today

Assumptions in age determination

- 1) System was **closed** between $t = 0$ and time t (usually the present time)
 - no transfer of the parent or the daughter element into or out of the system
- 2) At $t = 0$, Concentration of parent must be different in different phases in the system, but concentration of initial daughter must be the same.
- 3) we must also know λ accurately

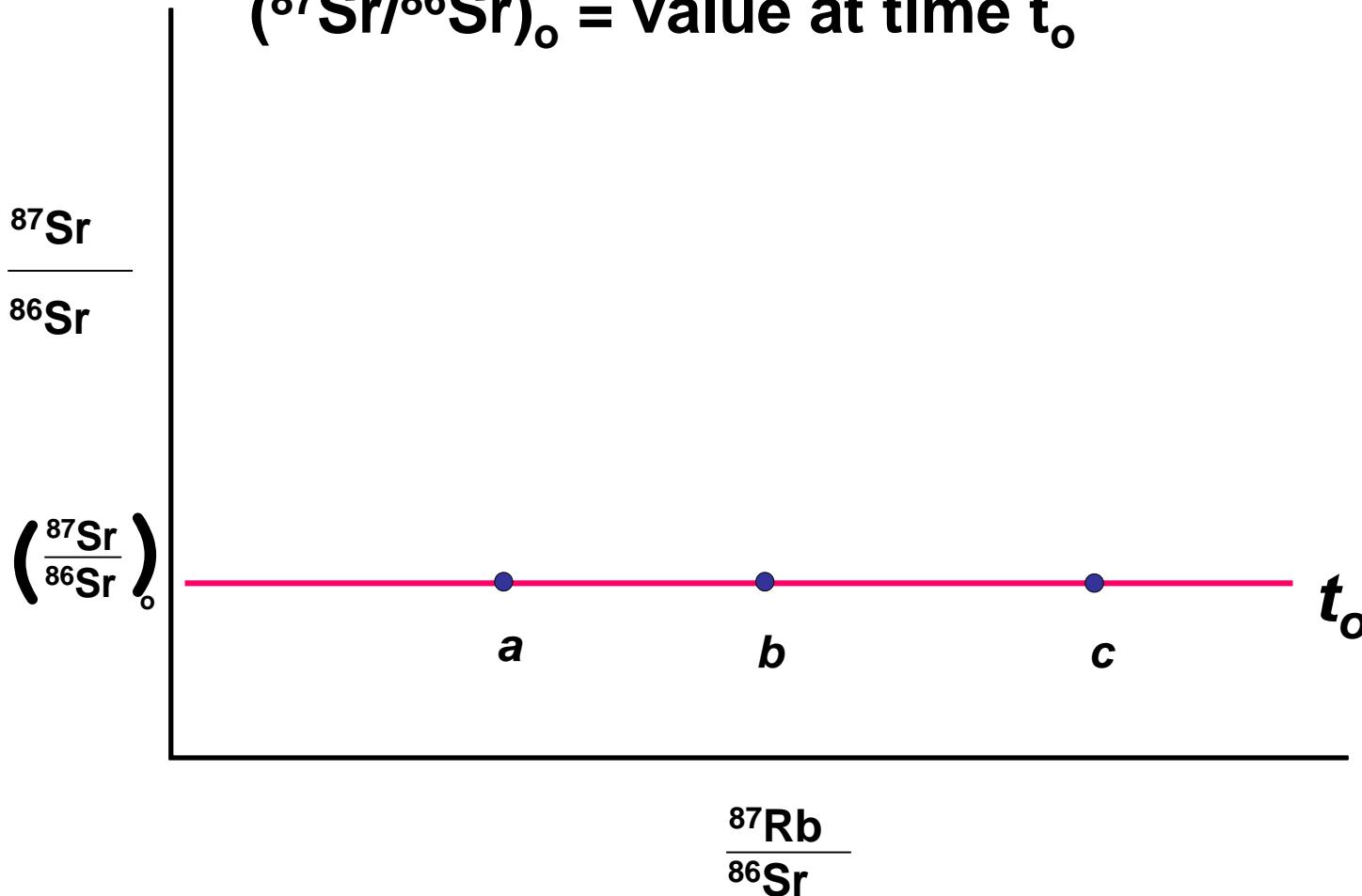
Violation of these conditions is the principal source of error in geochronology. Other errors arise from errors or uncertainties associated with the analysis.

Must satisfy: System is closed

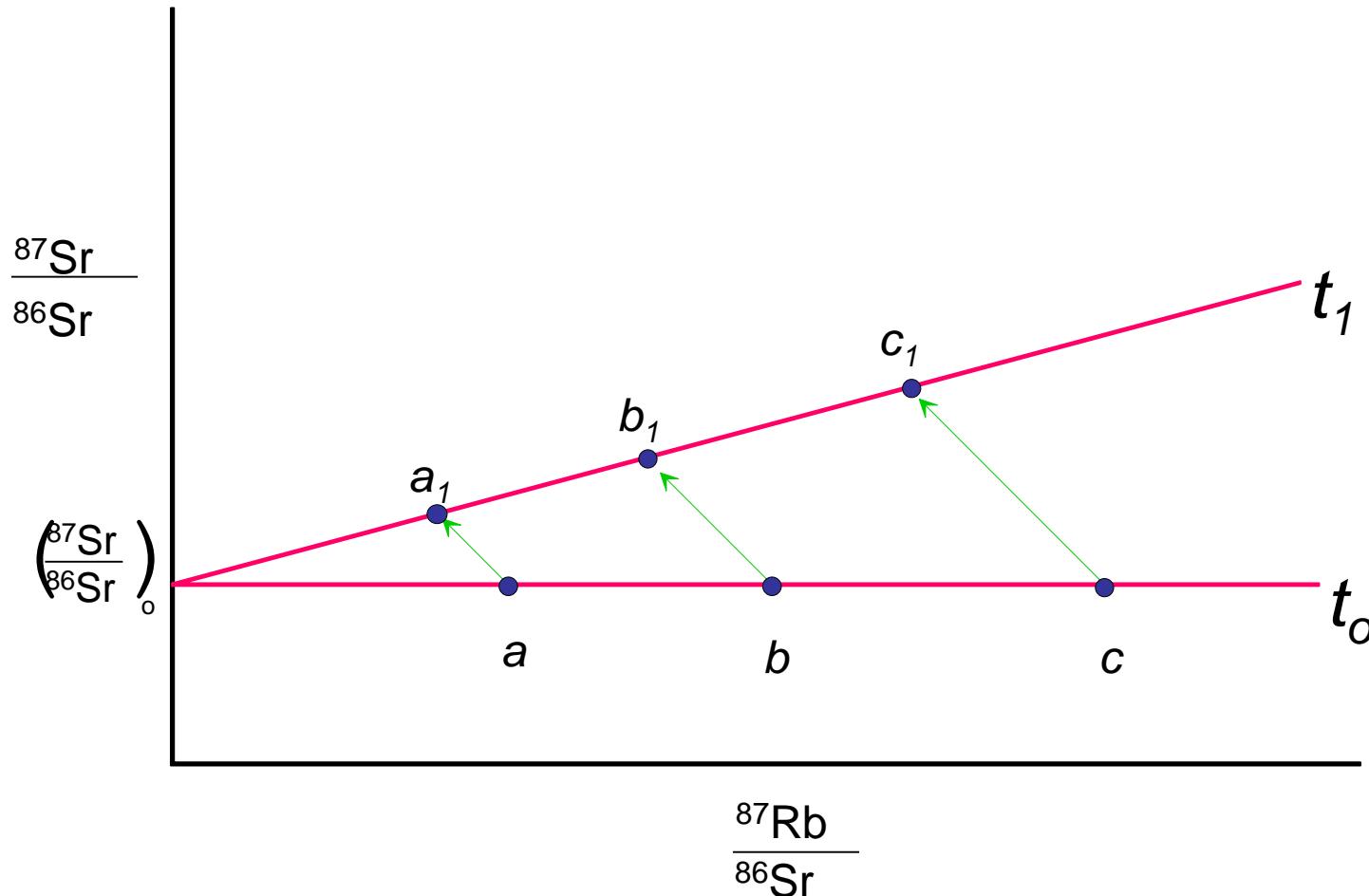
Begin with 3 rocks plotting at **a, b, c** at time **t_o**

They must have same value of initial $^{87}\text{Sr}/^{86}\text{Sr}$

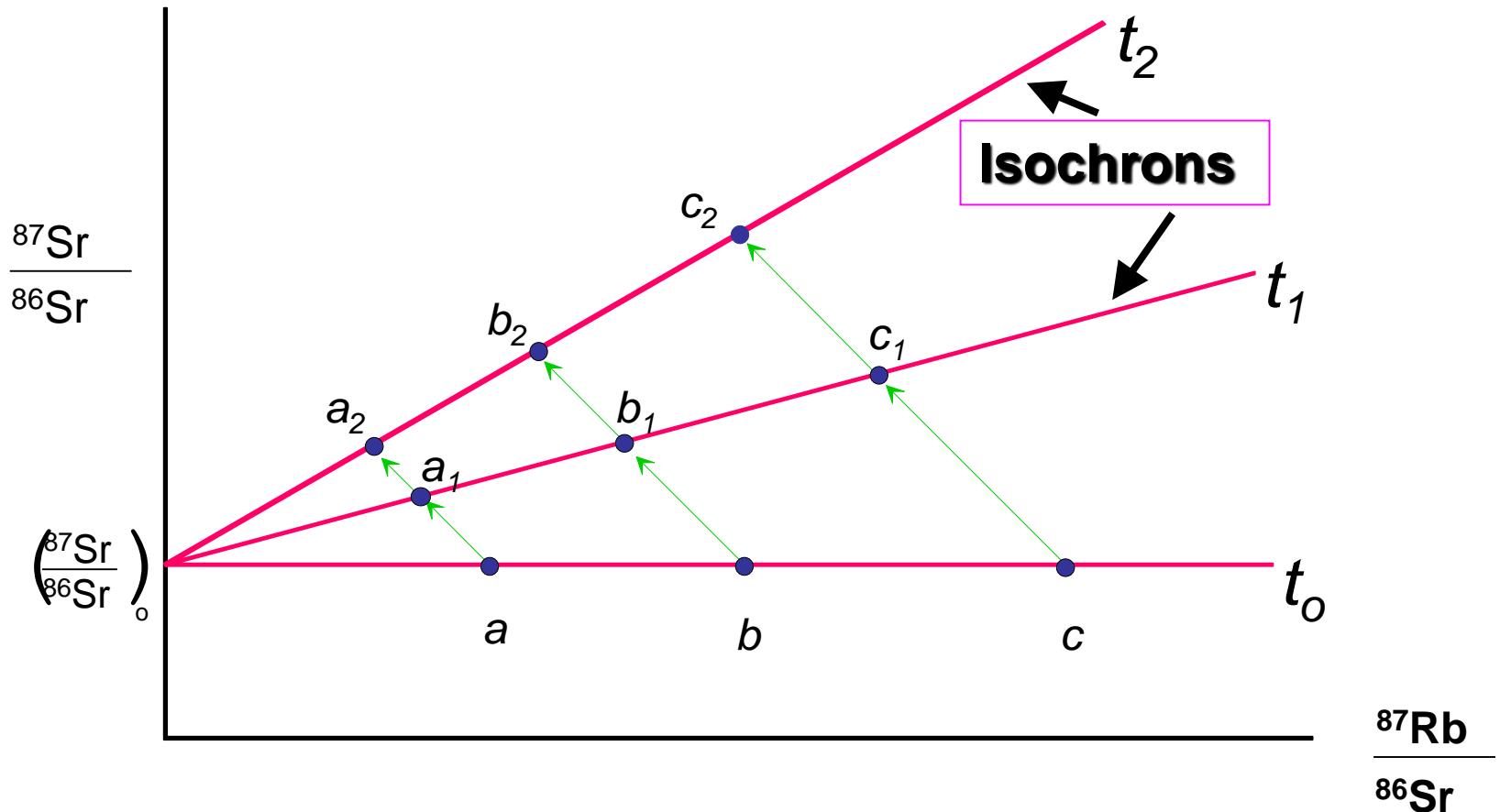
$$(^{87}\text{Sr}/^{86}\text{Sr})_o = \text{value at time } t_o$$



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



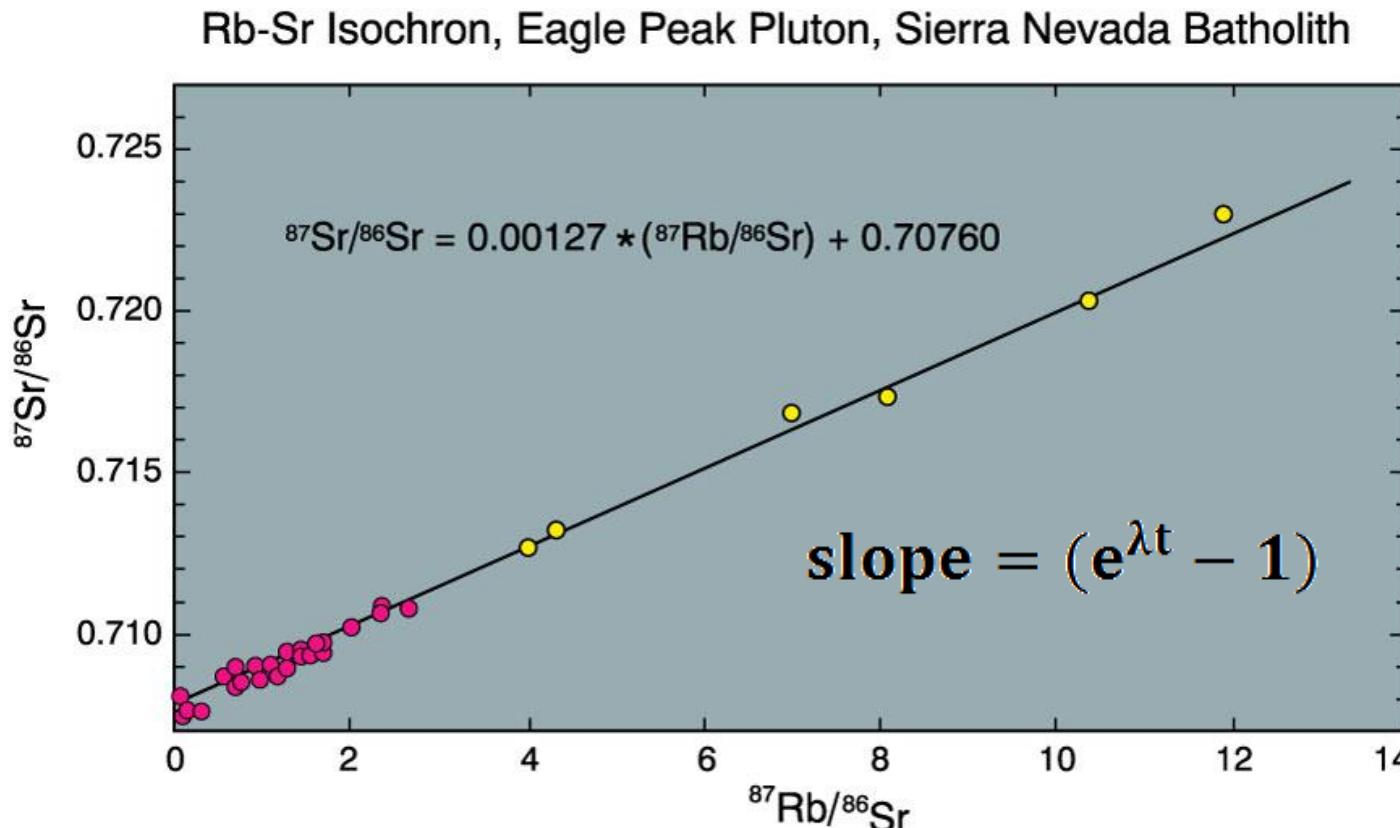
At time t_2 each rock system has evolved → new line. Again still linear and steeper line



All points on a single isochron line have same age.

Isochron technique produces 2 valuable things:

1. The age of the rocks (from the slope)
2. $(^{87}\text{Sr}/^{86}\text{Sr})_0$ = the initial value of $^{87}\text{Sr}/^{86}\text{Sr}$, which we did not know beforehand



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.



Indian Institute of Technology, Kanpur

Department of Earth Sciences

ESO213A: Fundamentals of Earth Sciences

Lecture 29. Layered Earth

Santanu Misra

Department of Earth Sciences
Indian Institute of Technology, Kanpur
smisra@iitk.ac.in • <http://home.iitk.ac.in/~smisra/>



Aims of this lecture

- Earth's interior
- Different layers and their properties

Reading:

- Marshak's Book (Chapter 2)
- Grottinger & Jordan's book (Chapter 14)
[for the entire week]

Surface of the Earth



- Our experience with Earth is limited to its surface.
- Dramatic elevation changes—mountains, canyons—are tiny “scratches” on this surface.
- Our Earth is much more vast and complex than the surface suggests.

The Layered Earth



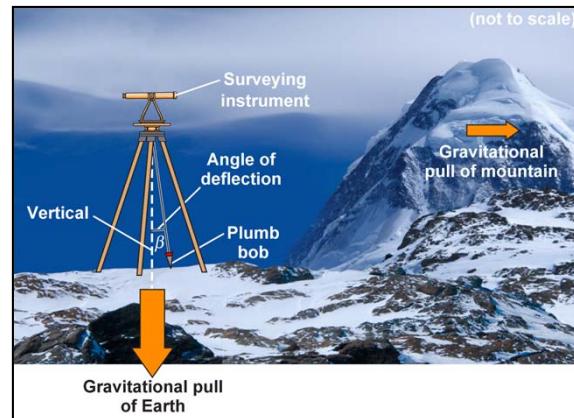
- How do we know that the Earth has a layered interior?
- Early speculations sought to explain:
 - The source of lava.
 - Gem and mineral enrichment.
 - Earthquakes.
- Early guesses were wrong.
 - Open caverns to the interior
 - Flowing lava, air, water



The Layered Earth



- The first key to understanding Earth's interior: **density**.
 - A plumb bob is deflected by a nearby mountain mass.
 - Degree of deflection can be used to calculate Earth's mass.
 - The density from this method (4.5 g/cm^3) is much higher than the density of the thin outer crust (2.5 g/cm^3).
 - This suggests that density must increase with depth.



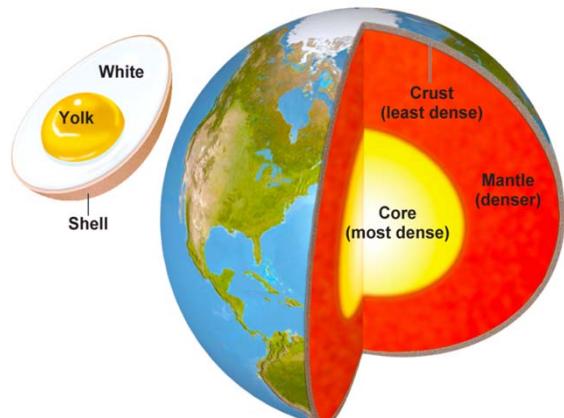
Watch "Schiehallion experiment" here: <https://vimeo.com/16031284>

The Layered Earth



- The first key to understanding Earth's interior: **density**.

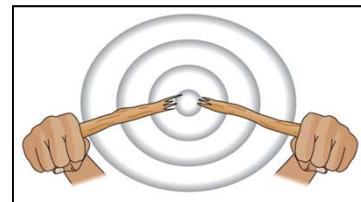
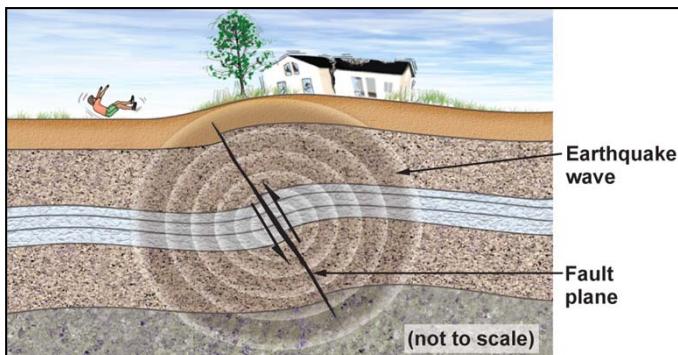
- In 1896, Emil Wiechert made important contributions.
 - He determined that metal must be present in Earth's center
- We now know he was correct.
- His ideas led to a model:
 - Earth is like an egg.
 - Thin, light crust (eggshell)
 - Thicker, more dense mantle (eggwhite)
 - Innermost, very dense core (yolk)
- Other density observations:
 - The land doesn't have large tides, hence Earth's center must be solid.



The Layered Earth



- Earthquakes: seismic energy from fault motion.
 - Seismic waves provide insight into Earth's interior.
 - Seismic wave velocities change with density.
 - We can determine the depth of seismic velocity changes.
 - Hence, we can tell where densities change in Earth's interior.



The Layered Earth



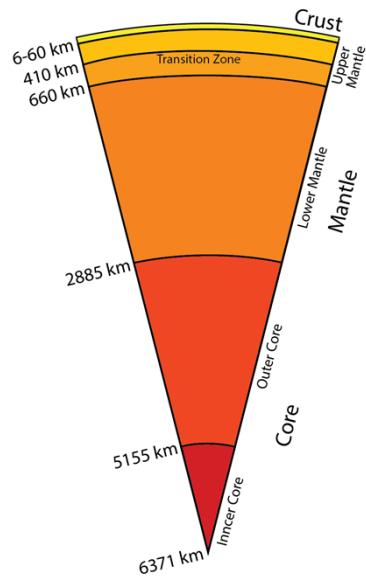
- Geologists strived to understand the nature of the layers.
 - Studied meteorites as analogues for core and mantle.
 - Conducted laboratory experiments.
 - Density measurements of rocks from the interior
 - Characteristics of mantle-derived rocks and minerals
 - Determined high P and T stability field of rocks and minerals



The Layered Earth



- End result of a century of investigation?
 - We know much about the nature of Earth's interior.
 - This knowledge continues to evolve.
- Earth's layers consist of the crust, upper, transitional, and lower mantles, and liquid outer and solid inner cores.
- Much complexity characterizes even within these layers.



The Layered Earth



- What are the boundaries?

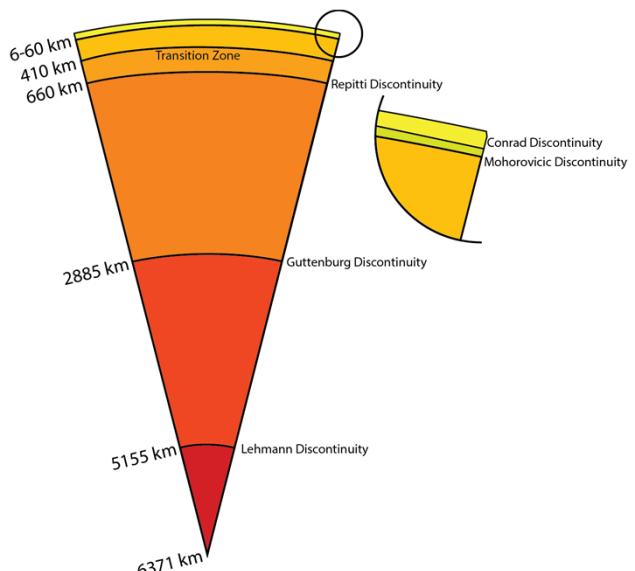
CONRAD discontinuity (15-20 km, for CC only):
between upper and lower Crust (questionable)

MOHOROVICIC discontinuity (6-60 km):
between Crust and Mantle

REPITTI discontinuity (660 km):
between upper and lower Mantle

GUTTENBURG discontinuity (2885 km):
between Mantle and Core

LEHMANN discontinuity (5155 km):
between Outer and Inner Core



The Layered Earth



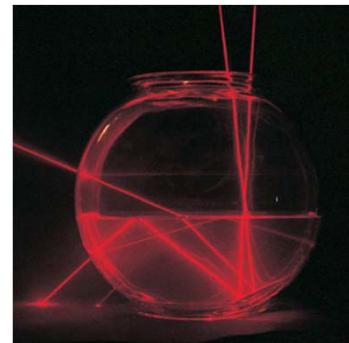
- How are the layers detected?

The velocity of waves, in general, is a function of the medium they are travelling through

The waves also reflect, and refract at the interfaces of different media

Geologists use all such wave properties, applicable to **SEISMIC** waves, together with a few more.

High Density → higher velocity
More solid like → higher velocity



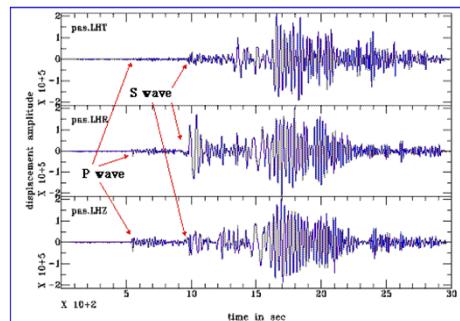
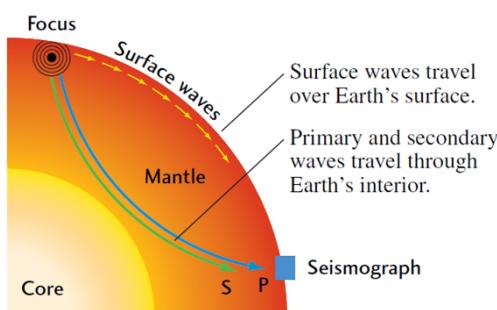
two beams of laser light enter a bowl of water from the top at slightly different angles. Both beams are reflected from a mirror on the bottom of the bowl. One is then reflected at the water-air boundary and passes through the bowl to make a bright spot on the table. Most of the light in the other beam is bent (refracted) as it passes from the water to the air, although a small amount is reflected to form a second spot on the table

The Seismic Waves



Seismic waves (*elastic waves*) are the waves of energy caused by the sudden breaking of rock within the earth or an explosion. They are the energy that travels through the earth and is recorded on seismographs.

The first waves to arrive are called primary waves, or **P waves**. The secondary waves, or **S waves**, follow. Both P waves and S waves travel through Earth's interior and known as **BODY WAVES**. Afterwards come the slower **SURFACE WAVES**, which travel around Earth's surface.



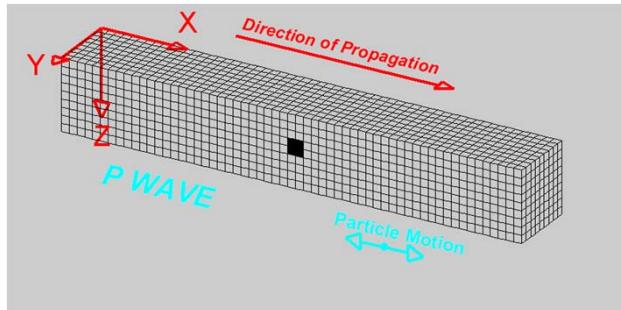
The Seismic Waves



P waves in rock are similar to sound waves in air, except that P waves travel through the solid rock of Earth's crust at about 6 km/s, which is about 20 times faster than sound waves travel through air. Like sound waves, P waves are *compressional waves*, because they travel through solid, liquid, or gaseous materials as a succession of compressions and expansions.

P waves are harmless in terms of earthquake damage.

$$V_P = \sqrt{\frac{K + \left(\frac{4}{3}\right)\mu}{\rho}}$$



<https://www.zmescience.com>

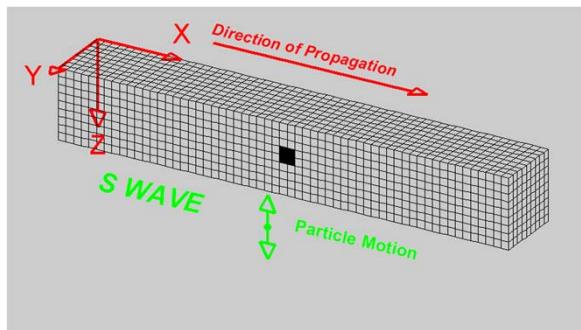
The Seismic Waves



S waves (shear waves) travel through solid rock at a little more than half the velocity of P waves. They are *shear waves* that displace material at right angles to their path of travel.

Shear waves cannot travel through liquids or gases and also rarely do any significant damage

$$V_S = \sqrt{\frac{\mu}{\rho}}$$

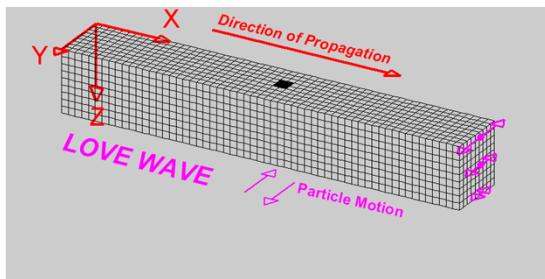


<https://www.zmescience.com>

The Seismic Waves

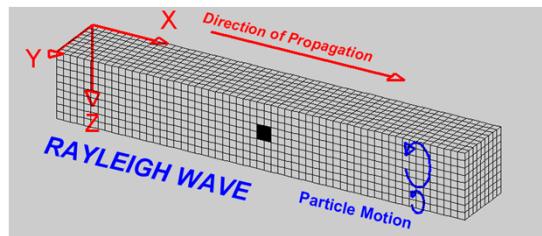


Surface waves (**Love waves**) have a transversal (perpendicular) movement and are the most destructive outside the immediate area of the epicentre. Love waves can be devastating.



<https://www.zmescience.com>

Surface waves (**Rayleigh waves**) do by far the most damage. As opposed to S and P, they propagate on the surface and carry the vast majority of the energy felt on the surface.

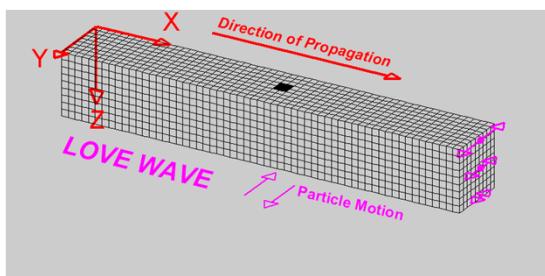


<https://www.zmescience.com>

The Seismic Waves

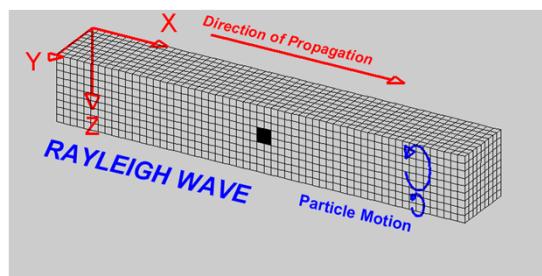


Surface waves (**Love waves**) have a transversal (perpendicular) movement and are the most destructive outside the immediate area of the epicentre. Love waves can be devastating.



<https://www.zmescience.com>

Surface waves (**Rayleigh waves**) do by far the most damage. As opposed to S and P, they propagate on the surface and carry the vast majority of the energy felt on the surface.



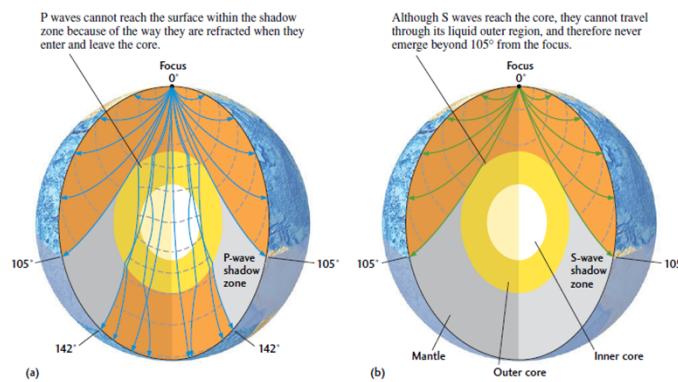
<https://www.zmescience.com>

Seismic Waves through Earth's interior



Observations of travel times and the amount of upward refraction of the ray paths, suggest that P waves travel much faster through deep Earth than at Earth's surface → **THIS IS OKAY**.

Interestingly, the P and S waves disappear beyond about 11,600 km (105°) from the source [each degree measures 111 km at the surface). Beyond about 15,800 km from the focus (142°), the P waves suddenly reappeared, although they were much delayed compared with their expected travel times. The S waves never reappeared. → **WOW**

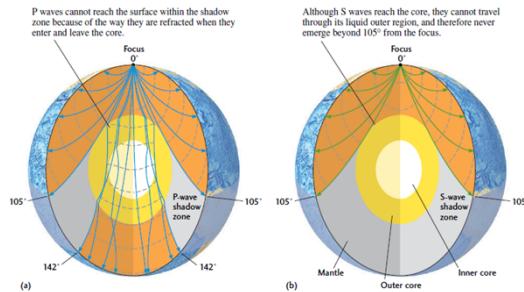


Seismic Waves through Earth's interior



P waves missing from 105 - 142° : At 105° , ray paths of P waves encounter the core-mantle boundary. At that boundary, P-wave velocity drops by almost a factor of two because of density contrast. The waves are refracted downward into the core and emerge at greater angular distances after the delay caused by their detour through the core.

S waves missing after 105° : Earth has a liquid outer core (R. D. Oldham, 1906). S waves cannot travel through the outer core, because it is liquid, and liquids have no resistance to shearing. Thus, there is an S-wave **shadow zone** beyond 105° from the earthquake focus, where S-wave ray paths encounter the core-mantle boundary.



Seismic Waves through Earth's interior

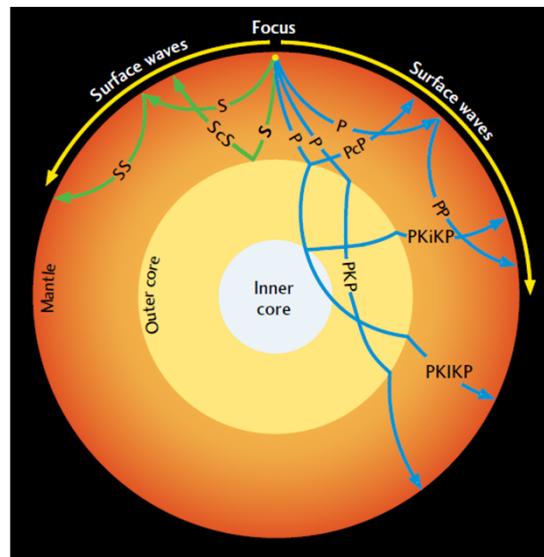


Geologists use a simple labelling scheme to describe the various ray paths taken by seismic waves.

PcP and ScS are compressional and shear waves, respectively, that are reflected by the core.

PP and SS waves are internally reflected from Earth's surface.

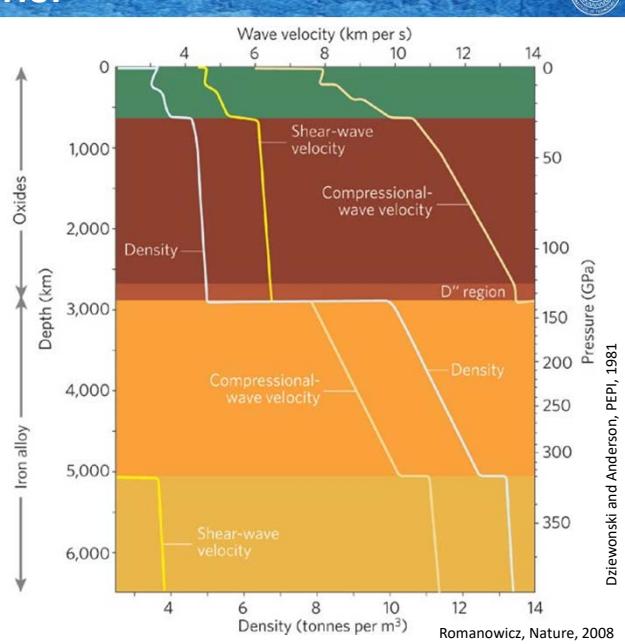
A PKP wave travels through the liquid outer core, a PKIKP wave travels through the solid inner core, and a PKIKP wave is reflected by the inner core.



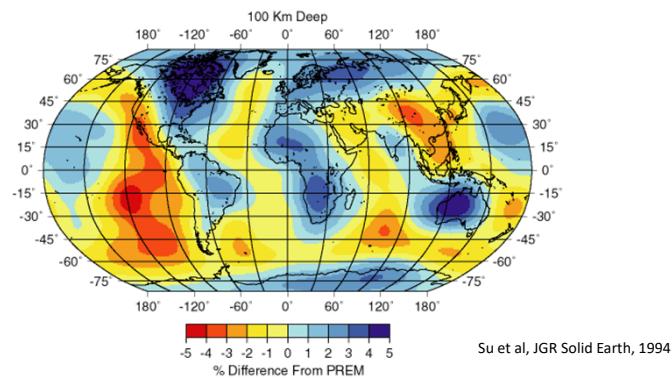
Seismic Waves through Earth's interior



The first-order structural units of Earth — its suite of concentric shells and their approximate composition — were established over the first half of the twentieth century from measurements of the travel times of seismic waves refracted and reflected inside Earth, whereas proof of the solidity of the inner core had to await the capability to record and digitize long time series and measure the frequencies of free oscillations. The '660 km' discontinuity is a phase change, and possibly a compositional change, in the silicate mantle. This illustration is of the Preliminary Reference Earth Model (PREM)

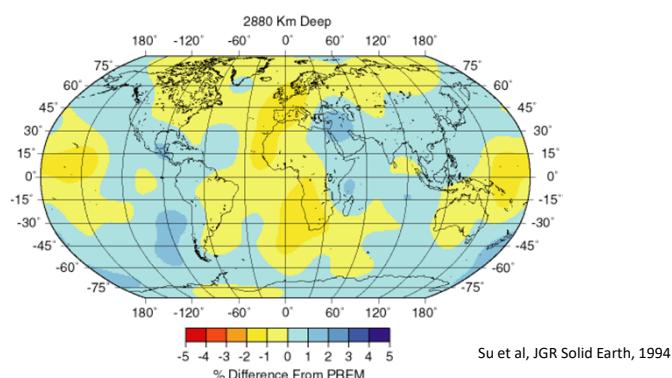


Seismic Waves through Earth's interior



In shallow (~100 km) regions where material is rising from the mantle, it should be warmer, and the velocity should be lower, in regions that are old and cold, such as beneath many of the old parts of continents, we would expect to see faster regions (assuming that temperature is the only difference). The actual variations are influenced by both temperature and composition variations, but they agree well with the ideas of plate tectonics, particularly at the divergent boundaries or oceanic spreading ridges.

Seismic Waves through Earth's interior



Shear wave variations at 2,880 km depth , just above the core-mantle boundary. See the lower-mantle velocity variations are more subdued than those in the more heterogeneous upper mantle. Also, note that the correlation with surface tectonics is gone, as you would expect for a complex convective system such as Earth's mantle.

The Layered Earth

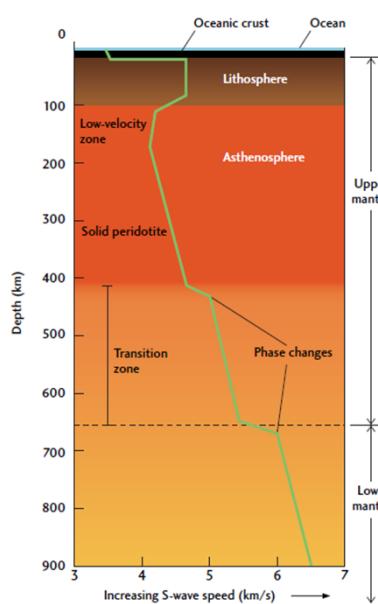
- What are the Compositions?

CRUST

- Felsic rocks typical of the upper continental crust (granite): V_p 6 km/s; 2.6 g/cc
- Mafic rocks typical of oceanic crust or the lower continental crust (gabbro): V_p 7 km/s; 2.9 g/cc
- Ultramafic rocks typical of the upper mantle (peridotite): V_p 8 km/s; 3.3 gm/cc

Mantle

- From MOHO to 410 km (peridotite: iron- and magnesium-rich silicate minerals)
- From 410-660 km (high pressure variations of olivine – the MTZ (Wadsleyite & Ringwoodite))
- From 660-2885 km (Lower Mantle – Perovskite)

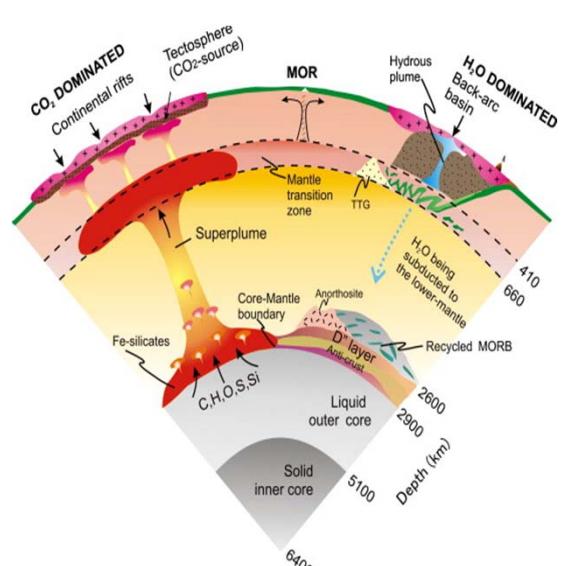


The Layered Earth

- What are the Compositions?

CORE-MANTLE BOUNDARY (CMB)

- Dramatic change in composition; from high pressure solid silicate to liquid Fe-alloy
- The last 200 km at the base of lower mantle (known as D'' layer) receive extreme heat and responsible for driving the Mantle Convection.
- Plumes also originate from the CMB.
- This zone is also the graveyard of the subducted lithosphere.



The Layered Earth



- What are the Compositions?

CORE

- Outer core Liquid and Inner Core Solid.
- Composition is Fe-Ni alloy
- We have very limited information about the core

Next Lecture



Temperature and other related properties



Indian Institute of Technology, Kanpur

Department of Earth Sciences

ESO213A: Fundamentals of Earth Sciences

Lecture 30. Mantle Dynamics

Santanu Misra

Department of Earth Sciences
Indian Institute of Technology, Kanpur
smisra@iitk.ac.in • <http://home.iitk.ac.in/~smisra/>



Aims of this lecture

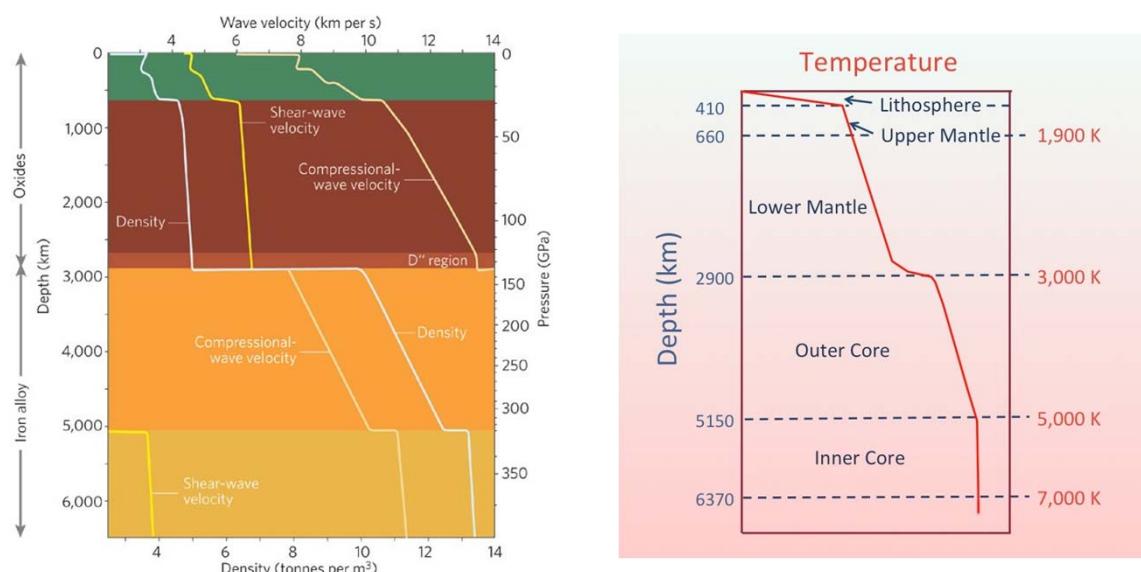
- Earth's Temperature profile
- Heat transfer in Earth's Mantle
- Mantle Convection

Reading:

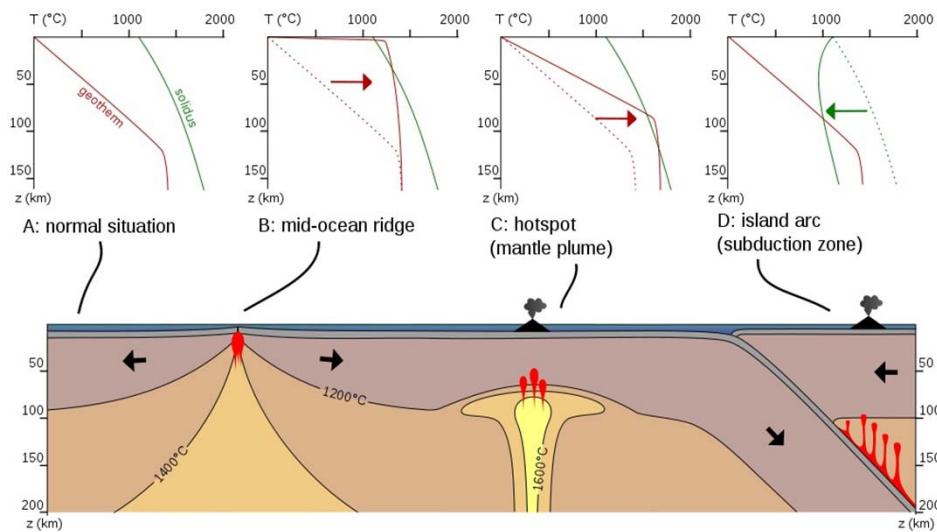
- Marshak's Book (Chapter 2)
- Grottinger & Jordan's book (Chapter 14)

Geodynamics: Turcotte and Schubert, Cambridge University Press (3rd Ed)

As a whole



Crust and upper Mantle



Heat conduction through the lithosphere



- The Earth is cooling... losing internal energy.
- Heat is being released from the Earth's interior at a rate of about 44TW/year . Averaged over the surface of the Earth, this amounts to a heat flow of about 70W/m^2 through the crust .
- Heat energy diffuses through the crust and lithosphere by conduction according to [Fourier's Law of Thermal Conduction](#).
- With magmas at volcanoes and spreading ridges, heat is being advected to the surface. Actually, this accounts for only a fraction of the heat that is brought to the surface and radiated through the atmosphere into space.

Fourier's Law of heat conduction



- A rate equation that allows determination of the conduction heat flux from knowledge of the temperature distribution in a medium

$$\vec{q}(\vec{r}) = -k\nabla T(\vec{r})$$

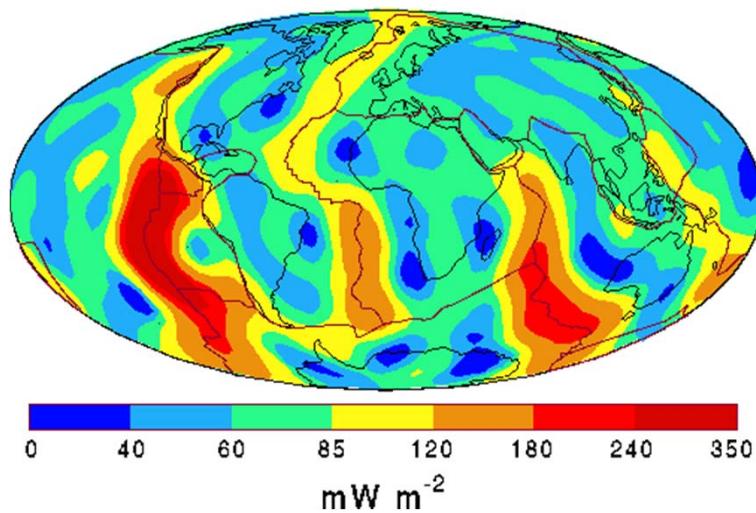
- Heat transfer is in the direction of decreasing temperature (basis for minus sign).
- Direction of heat transfer is perpendicular to lines of constant temperature (isotherms).
- Heat flux vector may be resolved into orthogonal components.
- Fourier's Law serves to define the thermal conductivity of the medium

\vec{q} : *heat flux* [$\text{W} \cdot \text{m}^{-2}$]

k : *conductivity* [$\text{W} \cdot \text{m}^{-2} \cdot \text{K}^{-1}$]

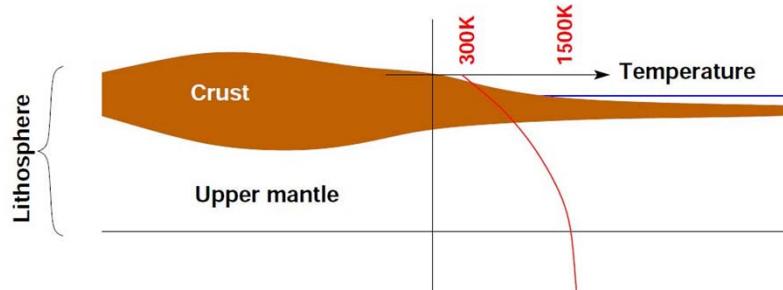
T : *temperature* [K]

Global Distribution of Heat Flow



<http://geophysics.ou.edu>

Temperature through the lithosphere and crust



- For a lithosphere with 100km depth (i.e., an average gradient of 12K/km), the average lithospheric thermal conductivity: $k \sim 4 \text{ W m}^{-2} \text{ K}^{-1}$.
- Measured conductivities of surface rocks: $k \sim 2-3 \text{ W m}^{-2} \text{ K}^{-1}$.

Thermal diffusion (conductive)



$$\nabla^2 T(\vec{r}, t) + \frac{1}{k} h_i(\vec{r}, t) = \frac{1}{D} \partial_t T(\vec{r}, t)$$

$$\nabla^2 T(\vec{r}, t) = \frac{1}{D} \partial_t T(\vec{r}, t)$$

T : temperature [K]

$D = \frac{k}{\rho C_p}$: thermal diffusivity [$m^2 \cdot s^{-1}$]

ρ : density [$kg \cdot m^{-3}$]

C_p : heat capacity at constant pressure [$J \cdot kg^{-1} \cdot K^{-1}$]

$h_i = \partial_t Q_i$: rate of heat input/volume [$W \cdot m^{-3}$]

Q_i : quantity of heat energy [J]

Thermally driven mantle convection



- The contribution of diffusive cooling of the mantle is insignificant in comparison to convective heat transport through the mantle.
- The mantle behaves like a viscous fluid on long timescales; being a fluid, it can flow and can be driven into convection by a temperature gradient.
- Heat flows out of the depths of the cooling Earth transported through the mantle between the D'' layer and the base of the lithosphere by convective fluid motions rather than conduction. This is the more effective means of moving heat through a fluid.

How does convection work?

- It's not conduction! More rapid heat transfer.
- Raise a parcel of hot rock.
- If constant entropy: Lower P => expands => larger volume => decreasing T
- This is known as adiabatic process.

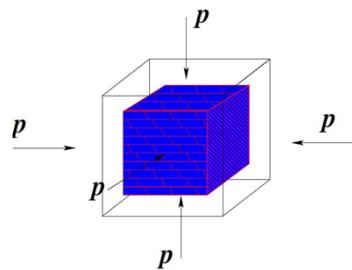
Adiabatic compression



Consider a cube of mantle material under pressure

Energy is received due to work...

$$\Delta T \propto \Delta P$$



$$\propto T$$

$$\propto \frac{1}{\rho}$$

$$\propto \frac{\alpha P}{C_p}$$

T: Temperature
P: Pressure
 ρ : Density
 α : Thermal expansion
 C_p : Heat capacity

$$\Delta T = \frac{\alpha P}{C_p} \frac{T g \rho \Delta r}{\rho}$$

Adiabatic temperature gradient



■ Need the change of temperature with pressure at constant entropy, S

$$\left(\frac{\partial T}{\partial P}\right)_S = - \left(\frac{\partial T}{\partial S}\right)_P \left(\frac{\partial S}{\partial P}\right)_T$$

Maxwell's thermodynamic relation

$$\left(\frac{\partial S}{\partial P}\right)_T = - \left(\frac{\partial V}{\partial T}\right)_P$$

Coefficient of thermal expansion

$$\alpha = \frac{1}{V} \left(\frac{\partial V}{\partial T}\right)_P$$

Specific heat

$$m C_p = T \left(\frac{\partial S}{\partial T}\right)_P$$

Re-arranging

$$\left(\frac{\partial T}{\partial P}\right)_S = \frac{\alpha T}{\rho C_p}$$

Adiabatic T gradient as a function of P
.... is this sufficient?

Adiabatic temperature gradient



$$\left(\frac{\partial T}{\partial P}\right)_S = \frac{\alpha T}{\rho C_p}$$

best to have it as function of depth...

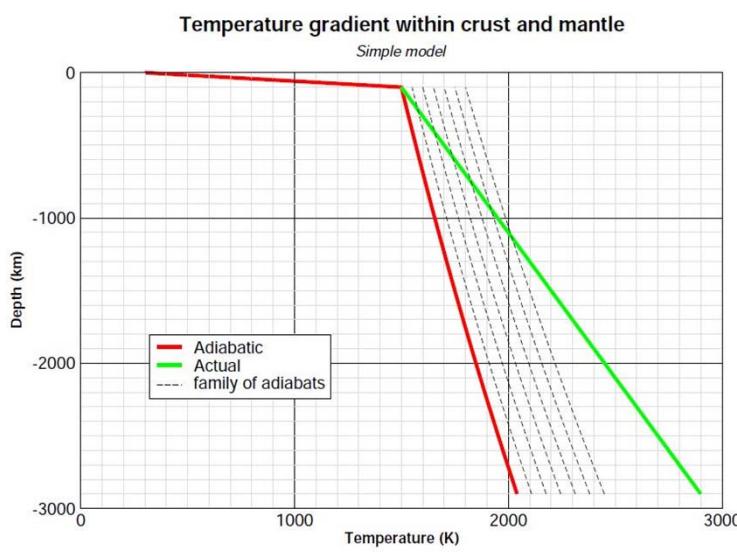
For the Earth

$$\frac{dP}{dr} = -g\rho$$

Re-arranging

$$\left(\frac{\partial T}{\partial r}\right)_S = \frac{\alpha T g}{C_p}$$

Mantle temperature?



If the actual temperature gradient exceeds the adiabatic gradient?



- If the actual temperature gradient (i.e., the increase of temperature with depth) exceeds the local adiabatic temperature gradient, then any infinitesimal displacement of a volume of mantle material will be enhanced through buoyancy, if displaced upwards or negative buoyancy if displaced downwards.
- **We have "convection"!**
- The process of convection removes heat from depth in the mantle to the base of the lithosphere where it is conducted out to the surface. The interior cools; the actual temperature gradient reduces.
- The process of convection pulls the entire mantle temperature toward the adiabatic gradient. If the temperature gradient falls to the adiabatic or below, convection ceases!
- If the temperature at the base of the lithosphere is 1500 K as corresponds to Hawaiian lava eruptions, then the adiabatic gradient to top of the D'' layer would account for a base temperature in excess of about 2100 K depending on the distributed thermal expansivity, α_p , and heat capacity, C_p , throughout the mantle.
- Heat "conducts" into the fluid mantle through the D'' boundary layer.

Mantle convection



In a fluid:

Occurs when density distribution deviates from equilibrium

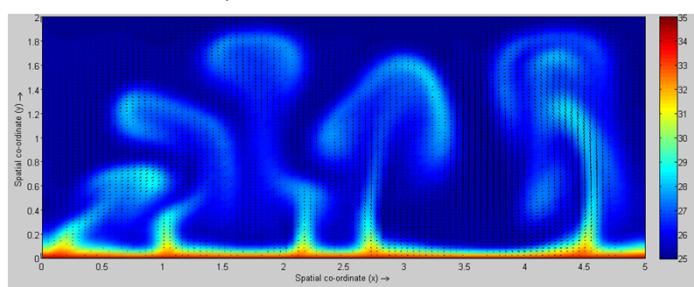
Fluid may then flow to achieve equilibrium again

In a viscous solid heated from below:

Initially heat is transported by conduction into the fluid **at the base**

Increased temperature reduces the density making the material at the base less dense than fluid above

Once the buoyancy force due to the density contrast overcomes the inertia of the fluid convection begins



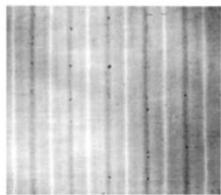
Rayleigh-Bernard convection



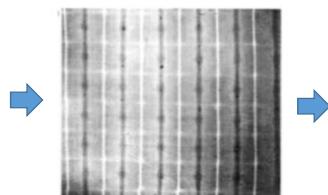
Newtonian viscous fluid:

Stress = dynamic viscosity x strain rate

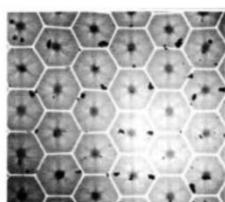
As fluid at the base heats up, initial convection is in 2D rolls



As heating proceeds, a second set of rolls forms perpendicular to the first – *bimodal*



Continued heating – *hexagonal* pattern

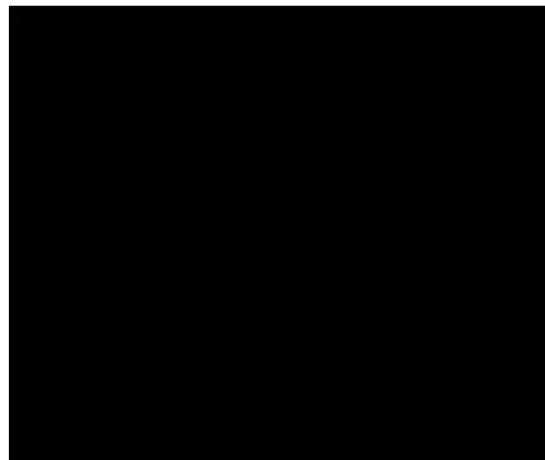


Spoke to irregular pattern

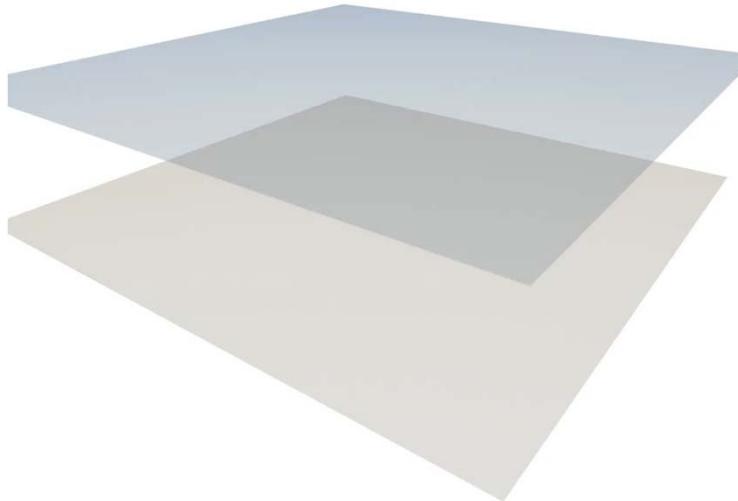
Read:

David White (1988); *The planforms and onset of convection with a temperature-dependent viscosity*; Journal of Fluid Mechanics, 191, 247-286.

Rayleigh-Bernard convection - examples



Rayleigh-Bernard convection - examples



Rayleigh number



- Convection can be *driven by internal or bottom* heating. Surely, both contribute.
- The Rayleigh number measures the ratio of the forcing-to-retardation of the convection. A nondimensional number which describes the nature of heat transfer by suggesting the relationships between buoyancy and viscosity with a fluid

■ For internal heating: $Ra = \frac{g\rho^3\alpha h_i d^5}{\eta k D}$

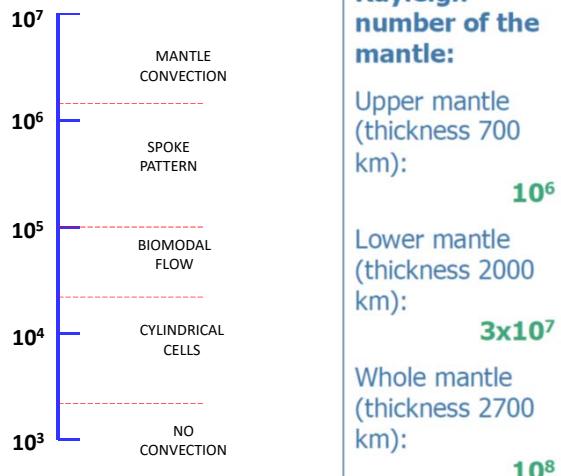
η is the local viscosity,
 $T_{sx'}$ the local adiabatic excess.
 D is mantle thickness, base of lithosphere to D".

■ For bottom heating: $Ra = \frac{g\rho^2\alpha C_p T_{sx} d^3}{\eta k}$

■ The critical Rayleigh number:

- The point at which convection initiates
- Approximately $> 10^3$ (dependent on geometry)
- By knowing the material properties and physical geometry we can determine if there will be convection and the nature of that convection

Rayleigh number and convective mode



Rayleigh number of the mantle:

Upper mantle (thickness 700 km): 10^6

Lower mantle (thickness 2000 km): 3×10^7

Whole mantle (thickness 2700 km): 10^8

Mantle convection – temperature dependence of Viscosity



Ra	10^7
T of viscosity*	$1000 \times$
η increase @ 660	0
Internal heating	0



Ra	10^7
T of viscosity*	$100 \times$
η increase @ 660	0
Internal heating	0



Ra	10^7
T of viscosity*	$10 \times$
η increase @ 660	0
Internal heating	0



Ra	10^7
T of viscosity*	$0 \times$
η increase @ 660	0
Internal heating	0



* Temperature dependence of Viscosity: --- x contrast between hottest and coldest

Mantle convection – Rayleigh number variation



Ra 10^8
 T of viscosity 1000 x
 η increase @ 660 0
 Internal heating 0



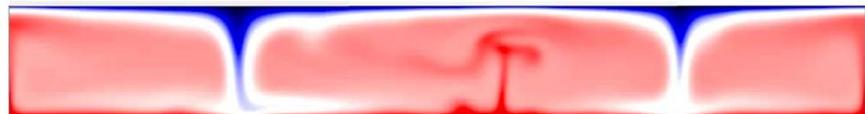
Ra 10^7
 T of viscosity 1000 x
 η increase @ 660 0
 Internal heating 0



Ra 10^6
 T of viscosity 1000 x
 η increase @ 660 0
 Internal heating 0



Ra 10^5
 T of viscosity 1000 x
 η increase @ 660 0
 Internal heating 0



* Temperature dependence of Viscosity: --- x contrast between hottest and coldest

Mantle convection – variation due to internal heating



Ra 10^7
 T of viscosity 1000 x
 η increase @ 660 0
 Internal heating 0



Ra 10^7
 T of viscosity 1000 x
 η increase @ 660 0
 Internal heating 20



Ra 10^7
 T of viscosity 1000 x
 η increase @ 660 0
 Internal heating 40



* Temperature dependence of Viscosity: --- x contrast between hottest and coldest

Mantle convection – the effect of heating - summary



Heat from below
T is fixed on upper boundary

→ **Aspect ratio of 1**
... not what we see on Earth



Heat from below
Constant heat flow across upper boundary

→ **Large aspect ratio**
... realistic?



Internal heating only

→ **No upwelling sheets**

Mantle convection – the effect of heating - summary



Isotopic ratios of oceanic basalts are very uniform but different from bulk earth values



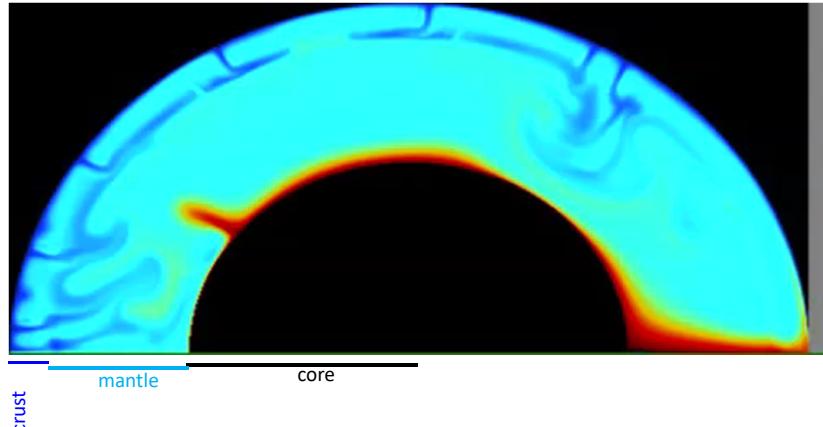
→ **The mantle is well mixed**



→ **Any body smaller than 1000 km is reduced to less than 1 cm thick in 825 Ma !**



Mantle convection – A near perfect combination

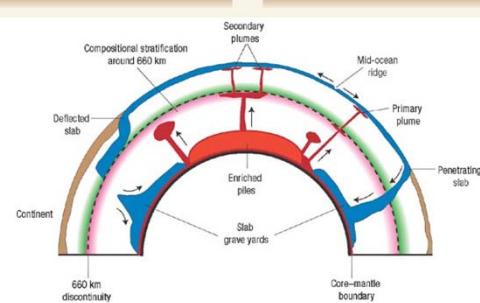
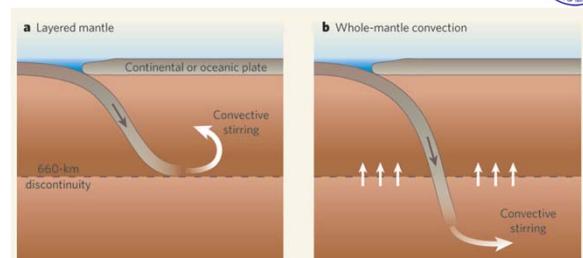


Source: Paul Tackley, ETH Zurich

Layered mantle convection?

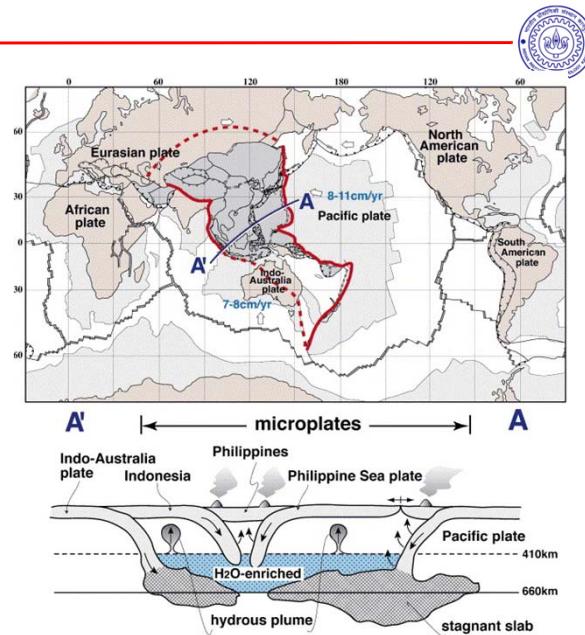


- There is/was a long-standing debate concerning the possibly layered convection in the mantle.
- There is general agreement that the 660 km phase change does retard sinking subducting plates and buoyant rising melts during convection
- There is also general agreement that plates can and do penetrate through 660 and that rising plumes and convective sheets rise through 660.
- Seismic tomography shows that we have a pooling of material around 660 as would be expected of layered convection while there remains sufficient penetration to involve the whole mantle in the convective process.



Stratified mantle

- In East Asia, the old Pacific plate subducts from the east, and the Indo-Australia plate from the south (double-sided subduction zone).
- The upper and lower mantle here are the coldest mantle regions in the world.
- This is the most active region on the Earth, indicating that the role of water is several orders of magnitude higher than that of the temperature in terms of lowering viscosity and drop of solidus. Note also the predominant occurrence of microplate in this region. Not only the fragmentation of continents but also the formation of small oceans constitutes the major reason for the dominant occurrence of microplates.
- The schematic cross-section of WPTZ is shown below to illustrate the stagnant slabs, hydrous MBL, and formation of hydrous plumes at 410 km depth by the breakdown reaction of hydrous wadsleyite enriched in incompatible elements



Some questions...

- Earth has mantle convection and plate tectonics – Why?
- When did it start? How long it will continue?
- What about other planets?