Composition and the Early History of the Earth

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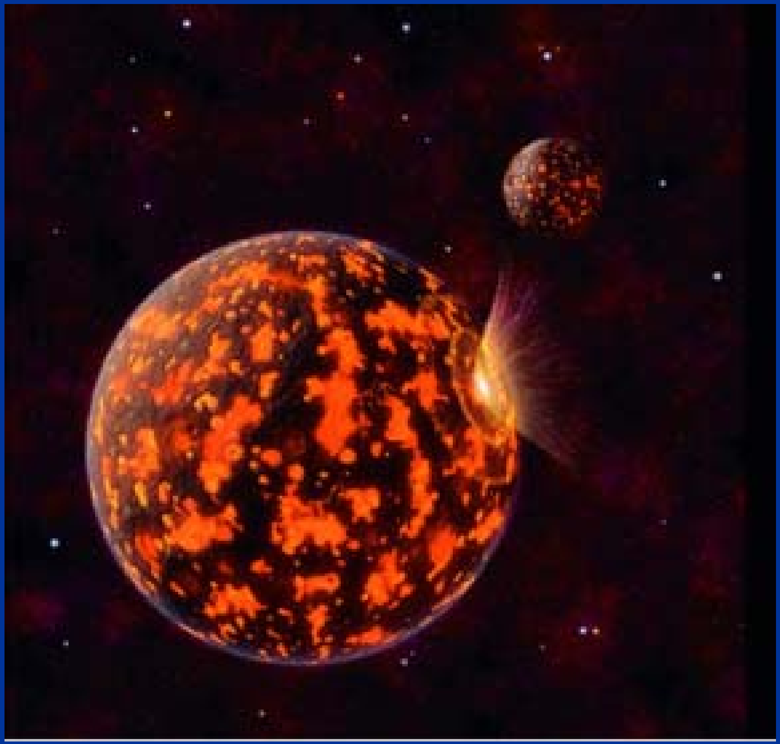
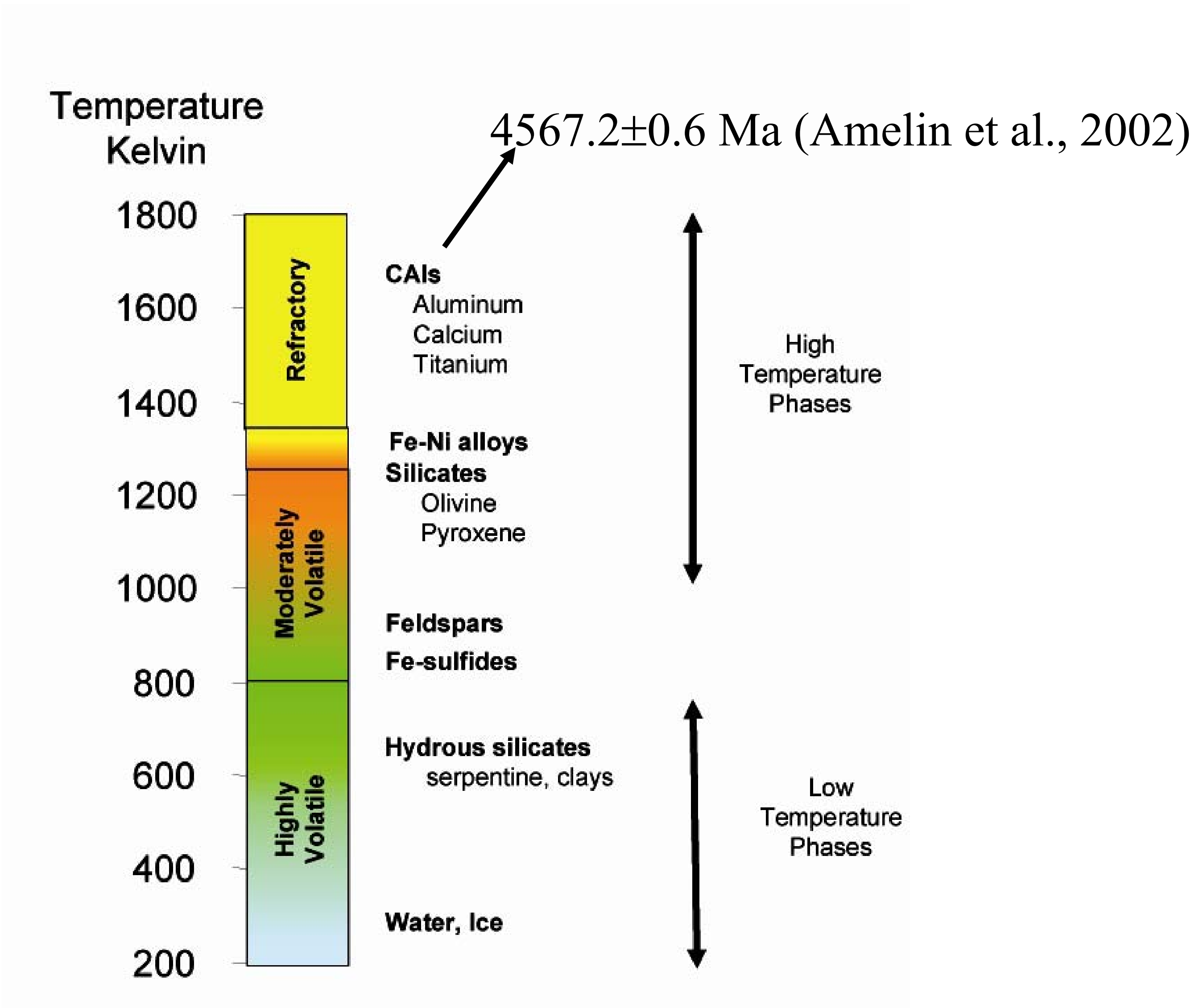
CIDER 2006



What we will cover in this lecture

- Composition of Earth
- Short lived nuclides and differentiation of the Earth
- Atmosphere and the initial volatile inventory



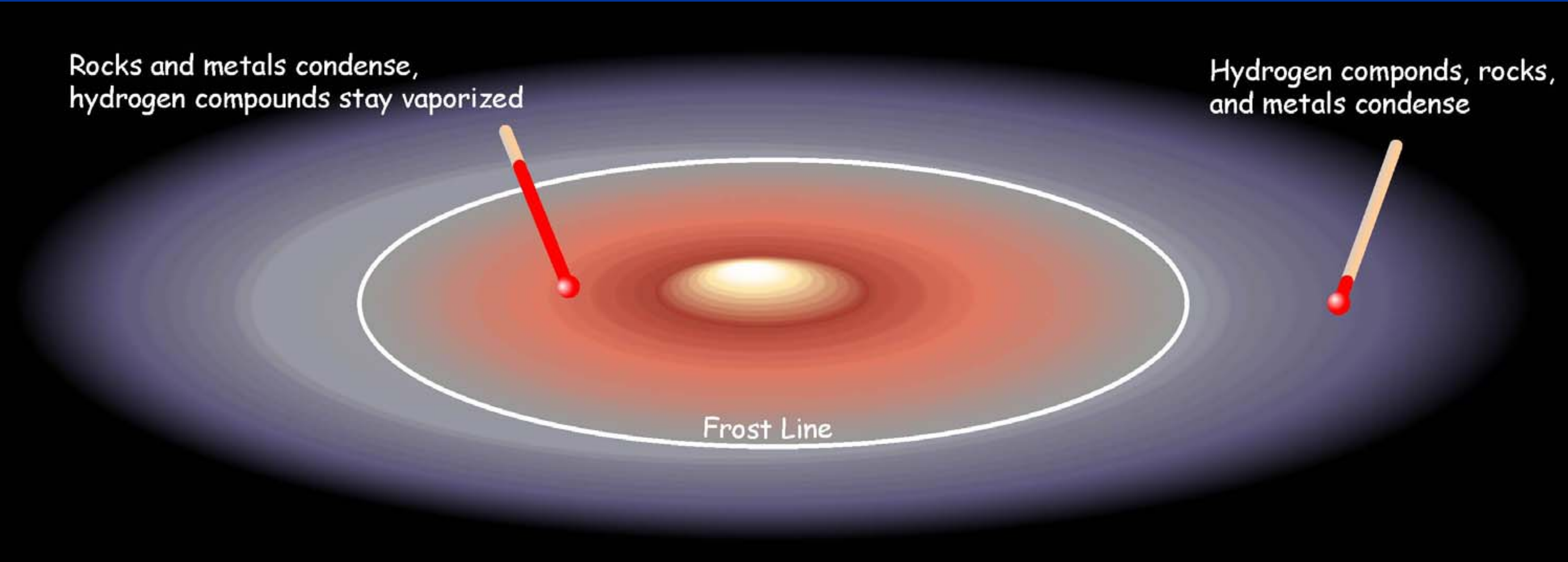
Temperature
Kelvin
4567.2±0.6 Ma (Amelin et al., 2002)
1800
1600
1400
1200
1000
800
600
400
200
Refractory
Moderately Volatile
Highly Volatile
CAIs
Aluminum
Calcium
Titanium
Fe-Ni alloys
Silicates
Olivine
Pyroxene
Feldspars
Fe-sulfides
Hydrous silicates
serpentine, clays
Water, Ice
High Temperature Phases
Low Temperature Phases

Planet Formation in the Solar Nebula

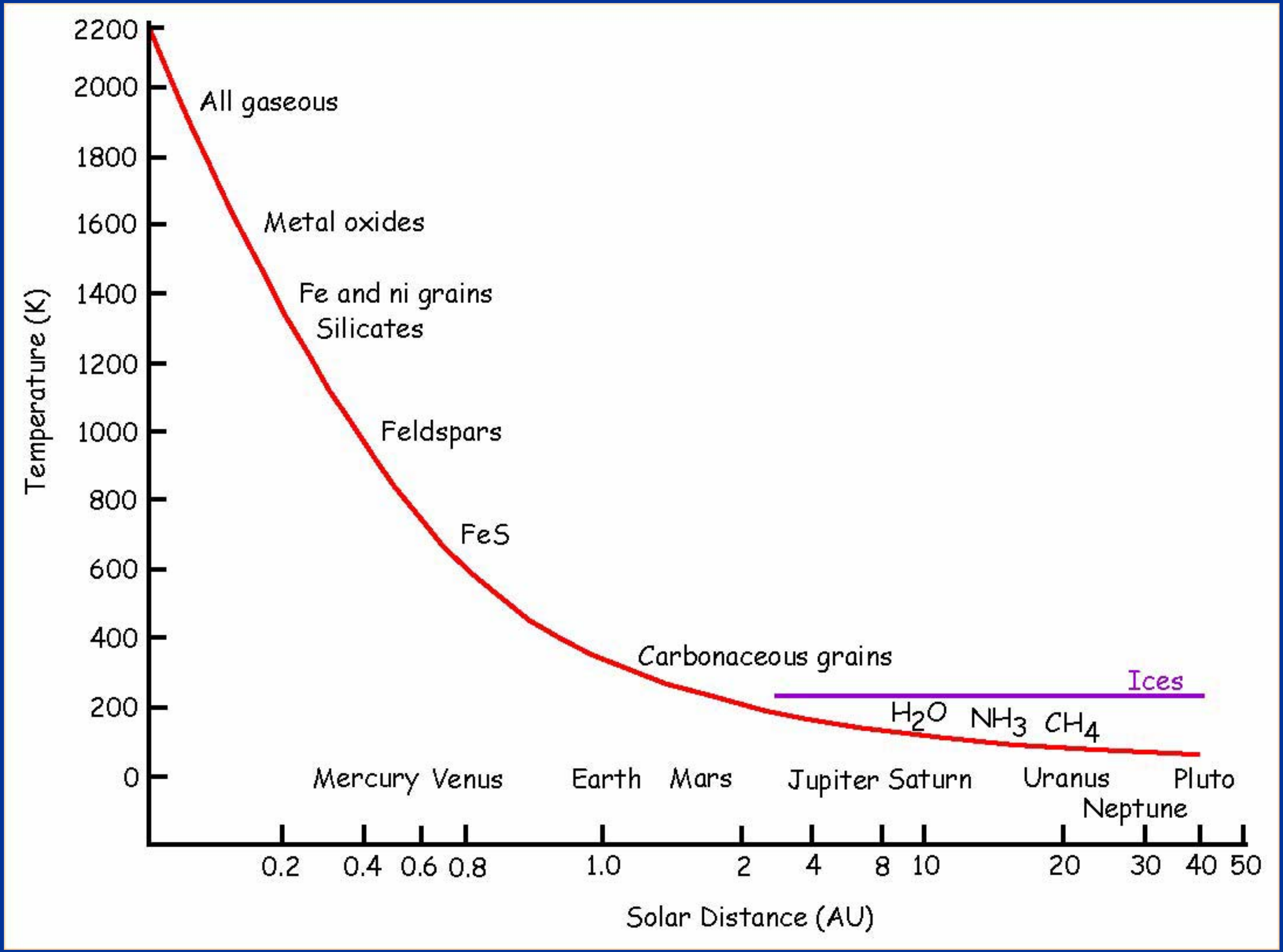
The great temperature differences between the hot inner regions and the cool outer regions of the nebula determined what kinds of condensates were available to form planets.

Near Mercury's orbit, metal started to condense. Moving outwards to Venus and Earth, more rock condensed

Only beyond the frost line, which lay between the present-day orbits of Mars and Jupiter, were temperatures low enough for hydrogen compounds to condense into ices.



So, the outer solar system contained condensates of all kinds, and since ice was nearly three times more abundant, ice dominated the mixture.



1 AU ~= 149,597,870 km

Estimating Earth composition

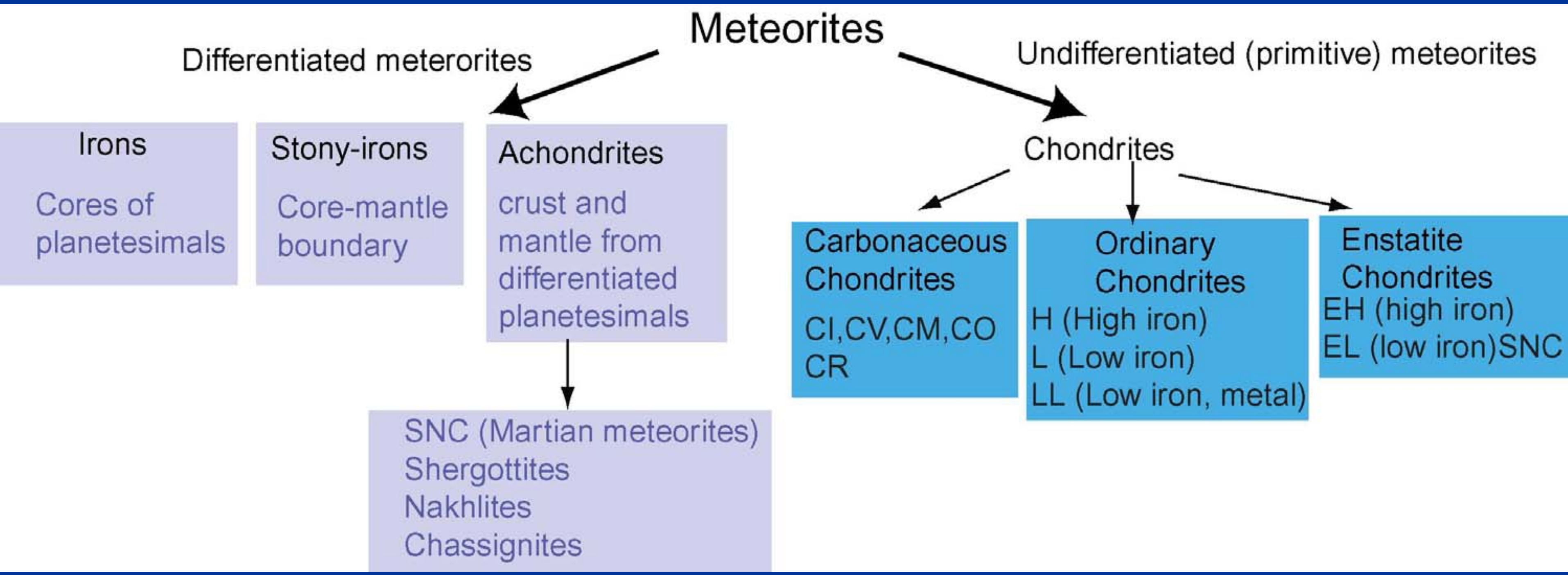
Assume earth has a bulk major element and refractory element composition similar to that of CI chondrites

-- does not predict abundances of moderately volatile and volatile species

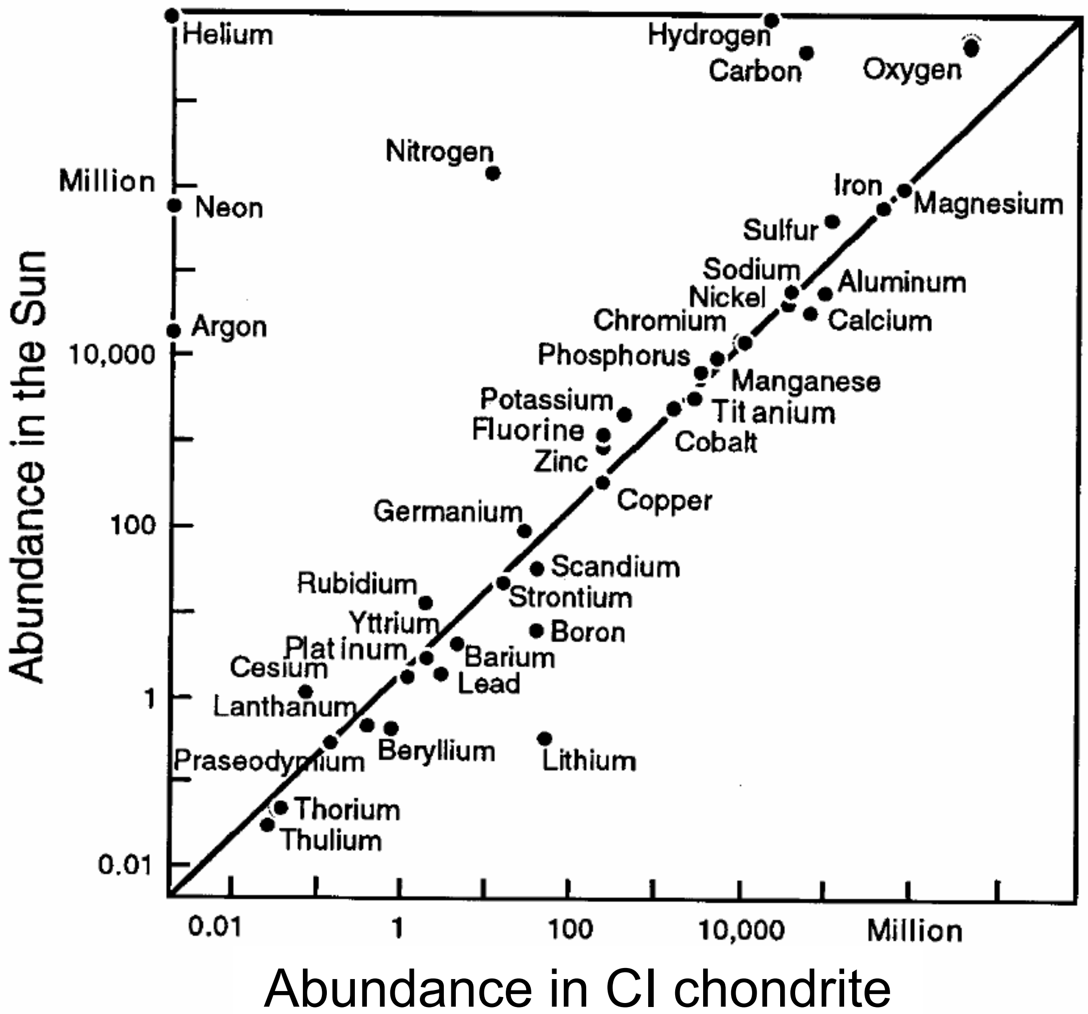
Simply assuming that all elements are present in chondritic proportions will lead to erroneous results

Earth composition
Meteorites – the building blocks

Meteorite classification



Meteorites – the building blocks



Primitive meteorites have elemental abundances that are identical to abundances in the solar photoshpere (except for the volatile species)

So CI chondrites are good starting point for inferring the composition of the Earth (at least for the non-volatile species)

Fingerprinting the building blocks with oxygen isotopes

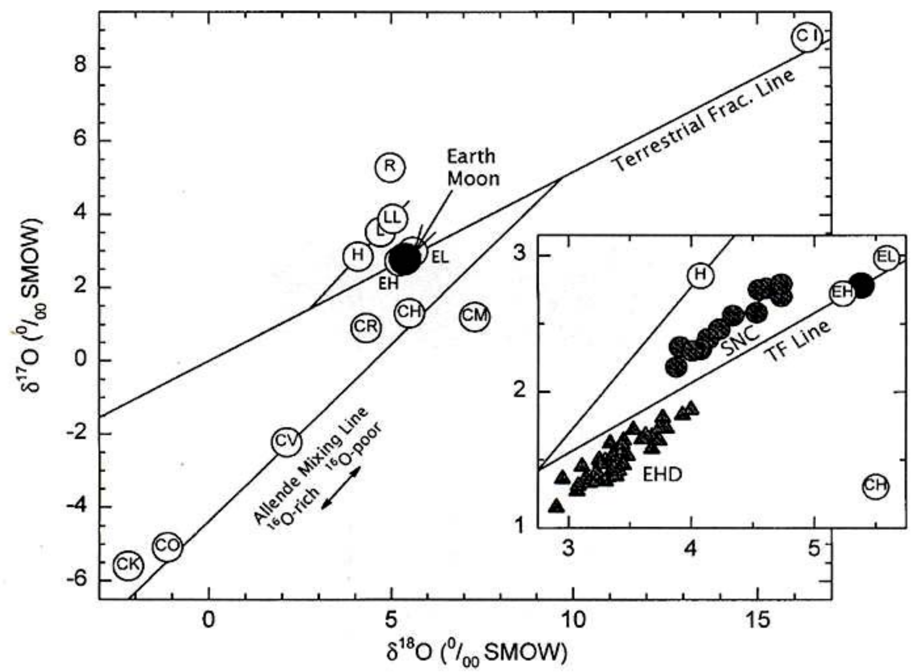


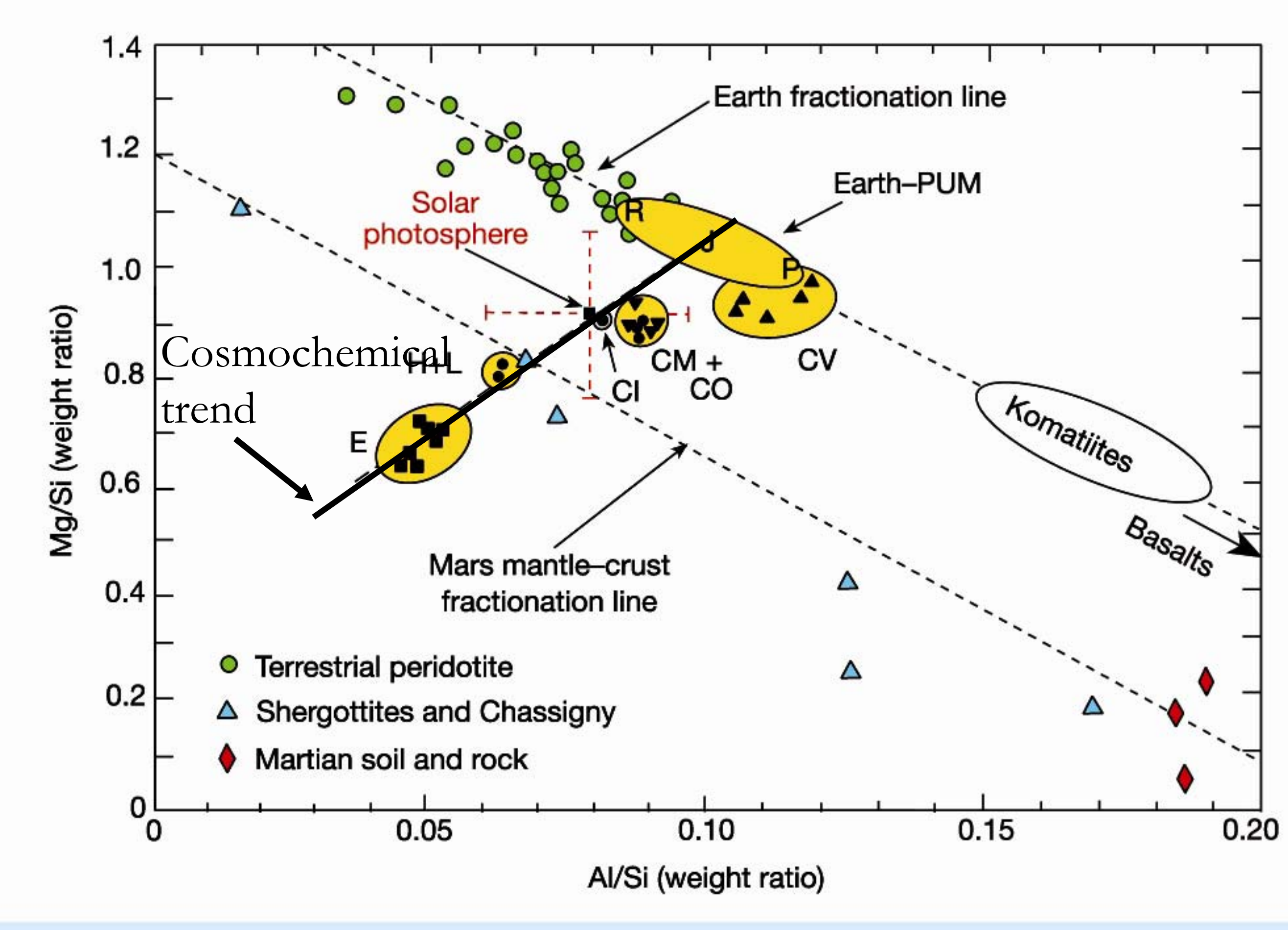
FIG. 1. Oxygen isotope systematics for chondrites and the Earth–Moon system. The terrestrial fractionation line and the mixing line of Allende components are also shown. The insert shows an enlarged region for differentiated meteorites such as EHD (eucrites, howardites, diogenites) and SNC (shergottites, nakhlites, chassignites), which plot parallel to the terrestrial fractionation line. All data are from Clayton and co-workers (e.g., Clayton 1993, Clayton *et al.* 1991, Clayton and Mayeda 1983, 1984, 1996 and references therein).

Only the enstatite chondrites have oxygen isotope composition that falls on the terrestrial fractionation line

Or make the Earth by mixing chondrites that plot above and below the fractionantion line → not much lie above the line though…..

Or the Earth was made of material that does not exist anymore

Bulk composition of the Earth



Drake and Righter, 2002

Bulk composition of the Earth

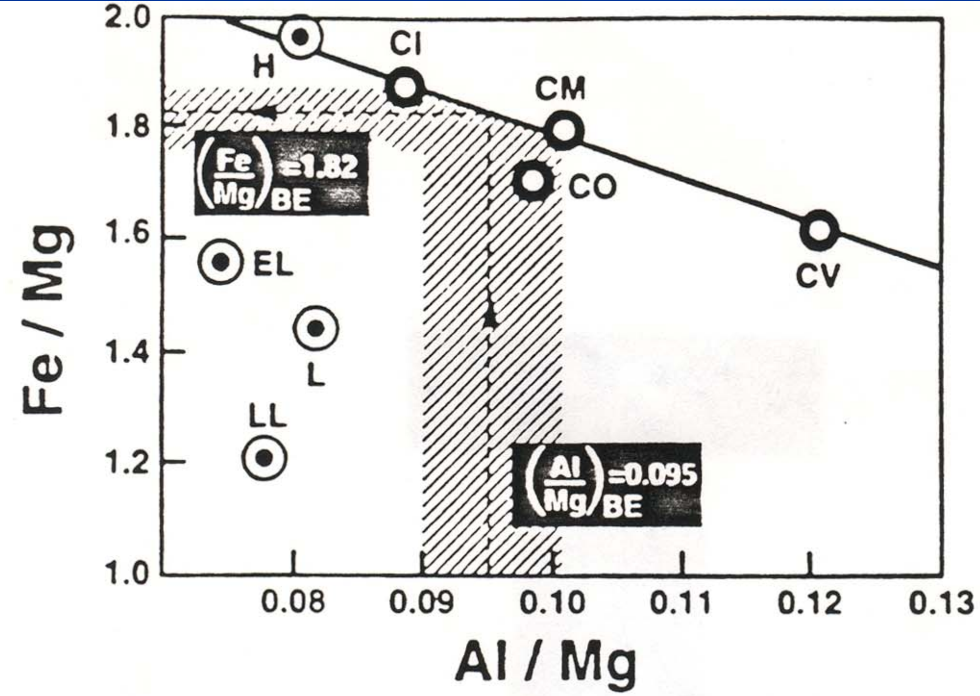
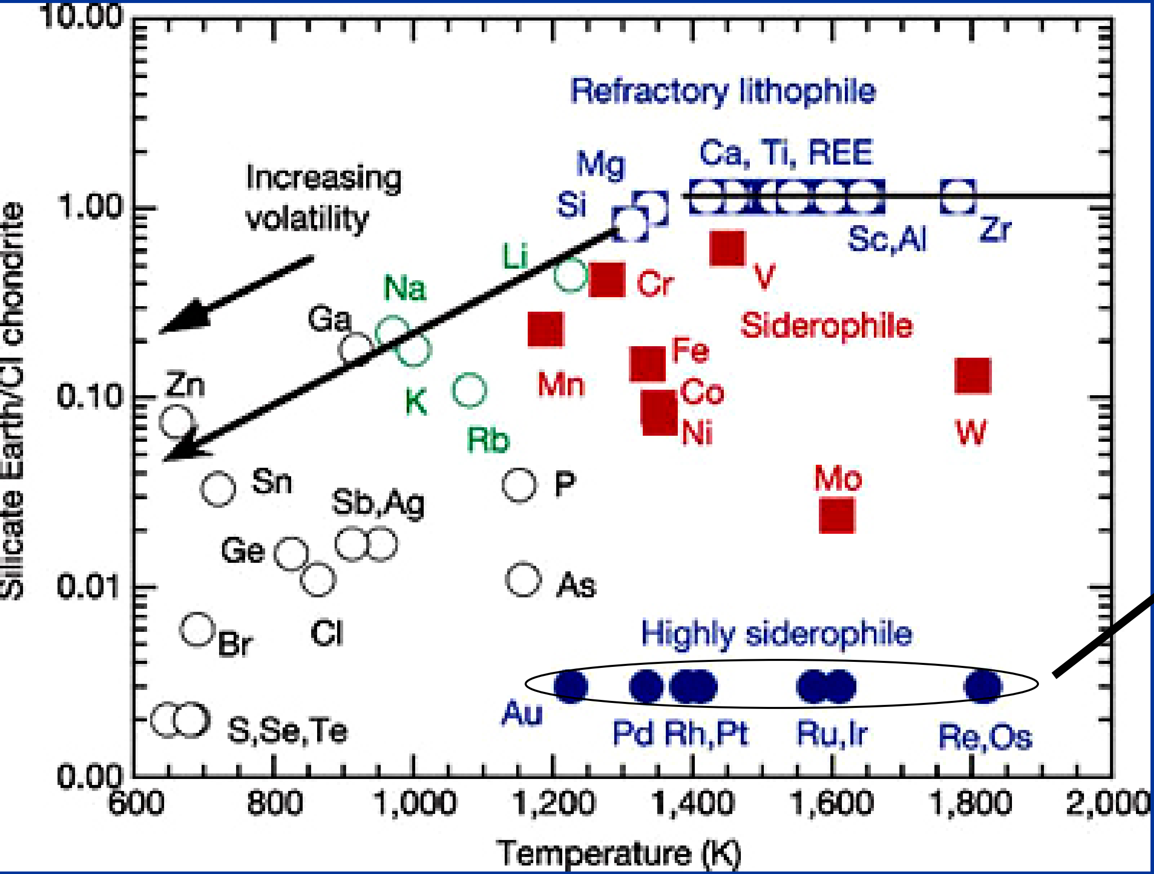


Fig. 3. (Al/Mg) vs. (Fe/Mg) ratios for a suite of chondrites (references and symbols are the same as in Fig. 2). We deduced the $(Fe/Mg)_{BE}$ value from the linear array based on $(Al/Mg)_{BE}$ = 0.095.

Estimating Earth Composition



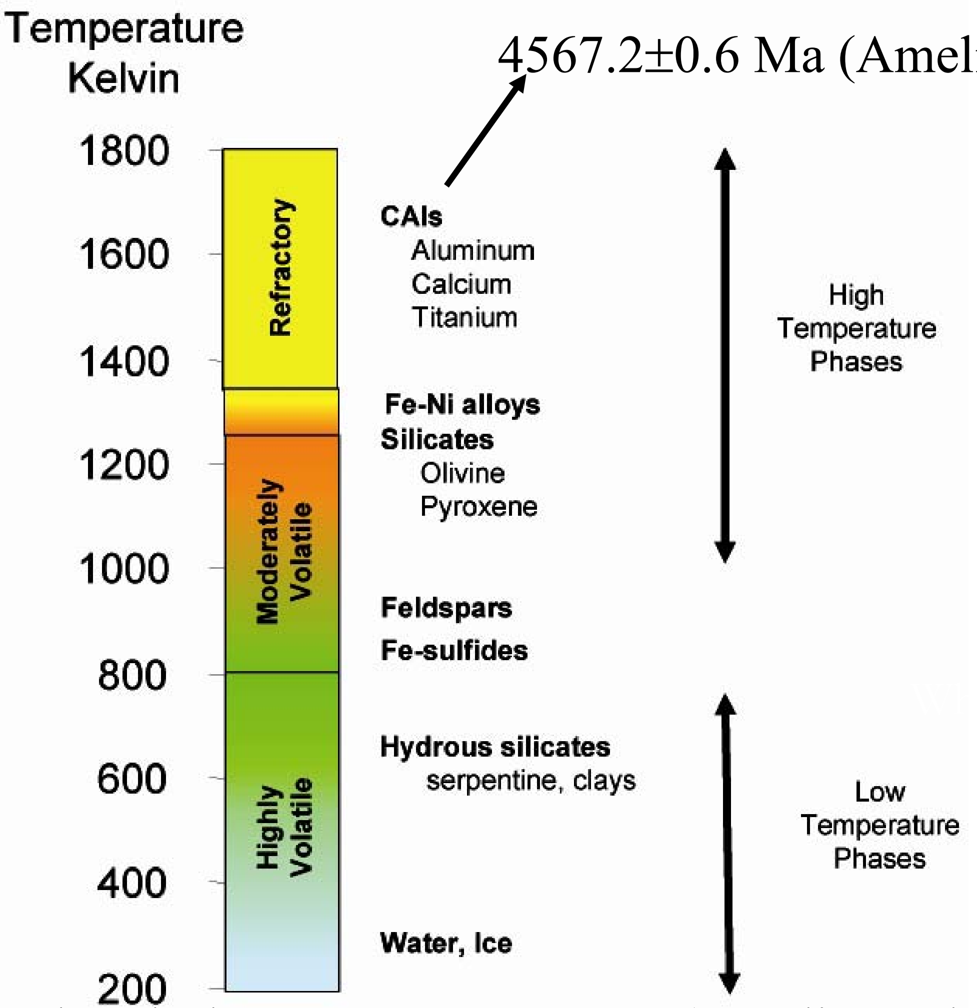
Lithophile – silicate loving
Siderophile – iron loving
Chalcophile – sulfur loving
Atmophile – species prefer gas/ fluid phase

Ratio of these elements are chondritic → used to argue for a late veneer after core formation

Elemental abundance versus 50% condensation temperature (Wood et al., 2006)

Relative to volatility trend, some elements are grossly depleted in silicate portion of the earth

4567.2±0.6 Ma (Amelin et al., 2002)



Chondrules -> 4564.7±0.6 Ma (Amelin et al., 2002)
Eucrites (achondrites) ~4560 Ma
So the Earth must also have formed within a few (ten) million years after the start of the solar system

Extinct radionuclides and early Earth history ($^{182}Hf \rightarrow {}^{182}W$, $^{146}Sm \rightarrow {}^{142}Nd$, $^{129}I \rightarrow {}^{129}Xe$)

Age equation for a long-lived radioactive element

$$\left(\frac{D_r}{D_s}\right)_{today} = \left(\frac{D_r}{D_s}\right)_{initial} + \left(\frac{P_r}{D_s}\right)_{today} (e^{\lambda t} - 1)$$

Measurable quantities

For an extinct radionuclide we have

$$\left(\frac{D_r}{D_s}\right)_{today} = \left(\frac{D_r}{D_s}\right)_{initial} + \left(\frac{P_r}{D_s}\right)_{initial}$$

But P_r does not exist anymore…..

$$\left(\frac{D_r}{D_s}\right)_{today} = \left(\frac{D_r}{D_s}\right)_{initial} + \left(\frac{P_r}{P_s}\right)_{initial} \left(\frac{P_s}{D_s}\right)$$

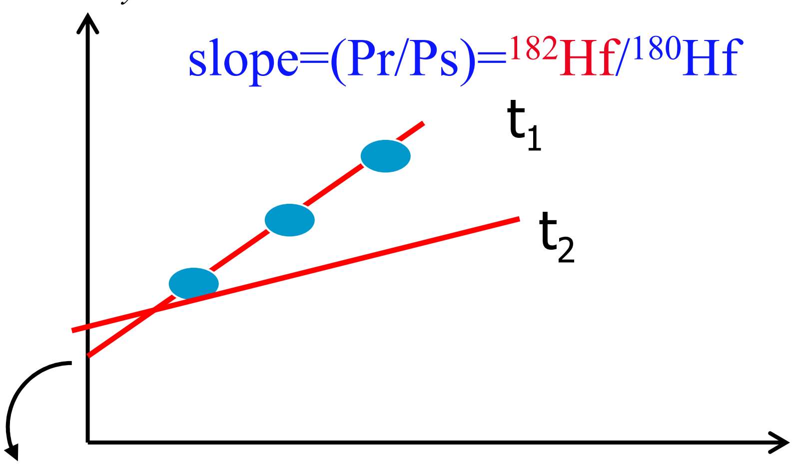
P_r is radioactive parent
D_r is daughter produced from radioactive decay of P_r
D_s is stable isotope of D_r and not produced by radioactive decay
P_s stable isotope of P_r
λ is the decay constant

Extinct radionuclides: $^{182}Hf \rightarrow {}^{182}W$

$$\left(\frac{D_r}{D_s}\right)_{today} = \left(\frac{D_r}{D_s}\right)_{initial} + \left(\frac{P_r}{P_s}\right)_{initial}\left(\frac{P_s}{D_s}\right)$$

$$\left(\frac{^{182}W}{^{183}W}\right)_{today} = \left(\frac{^{182}W}{^{183}W}\right)_{initial} + \left(\frac{^{182}Hf}{^{180}Hf}\right)_{initial}\left(\frac{^{180}Hf}{^{183}W}\right)$$

$$\left(\frac{^{182}W}{^{183}W}\right)_{today}$$



$$\left(\frac{^{182}W}{^{183}W}\right)_{initial}$$

$$\left(\frac{^{180}Hf}{^{183}W}\right)$$

$$\left(\frac{^{182}Hf}{^{180}Hf}\right)_2 = \left(\frac{^{182}Hf}{^{180}Hf}\right)_1 \exp[-\lambda\Delta t]$$

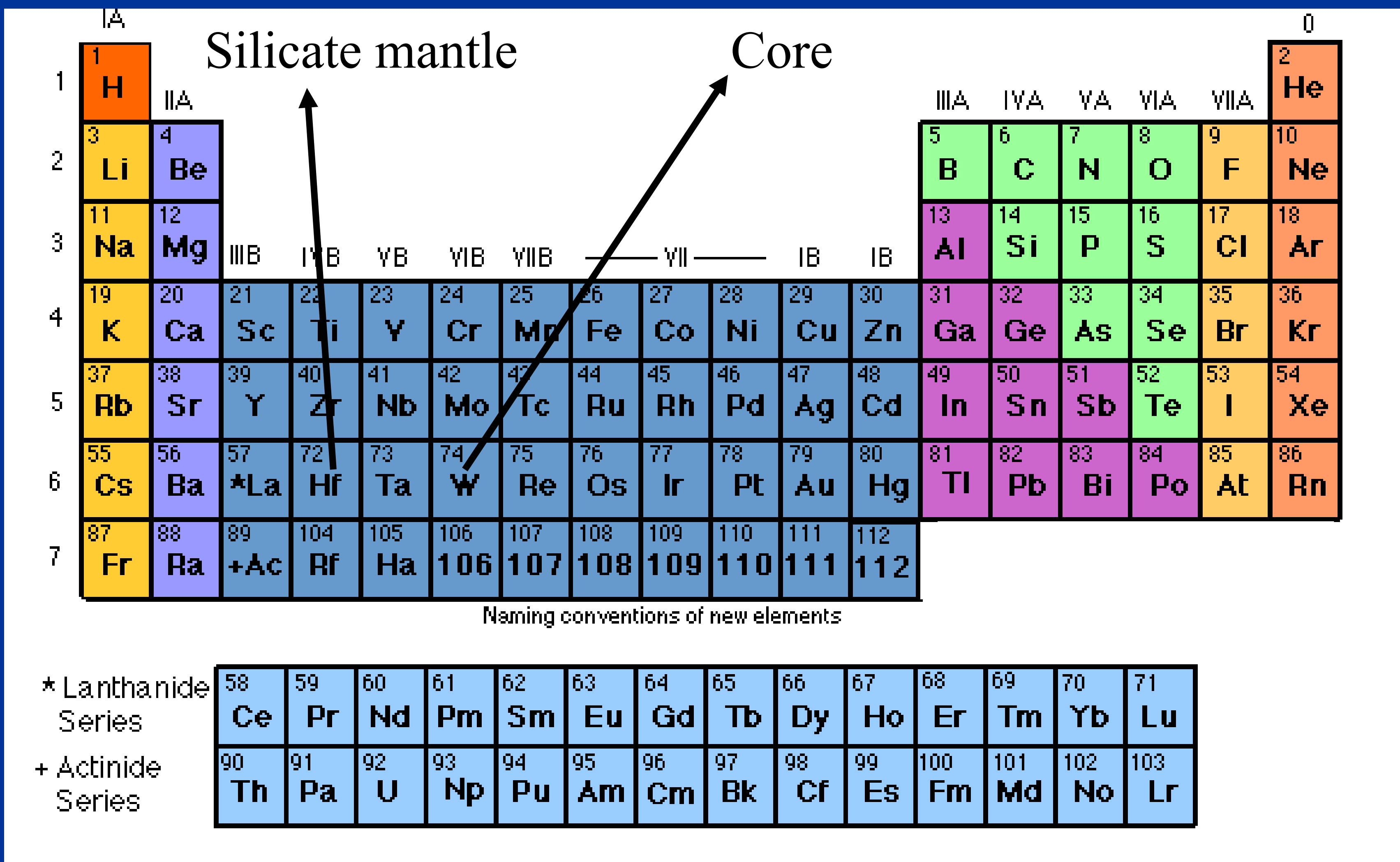
$$\Delta t = t_2 - t_1$$

Hafnium –Tungsten (^{182}Hf-^{182}W) systematics; $t_{1/2}$ = 9 M.y

Hf is highly *lithophile*, retains in the silicate portion of the Earth.

W is *moderately siderophile*, some enters core, some retains in the silicate mantle.

So ideal for tracking core formation



$^{182}W^*$ Evolution due to decay of ^{182}Hf

$t_{1/2}$ = 9 m.y

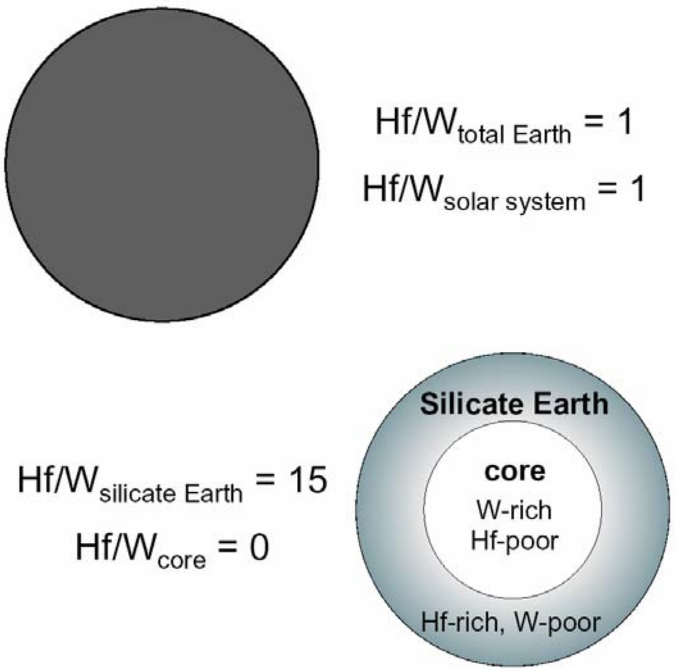
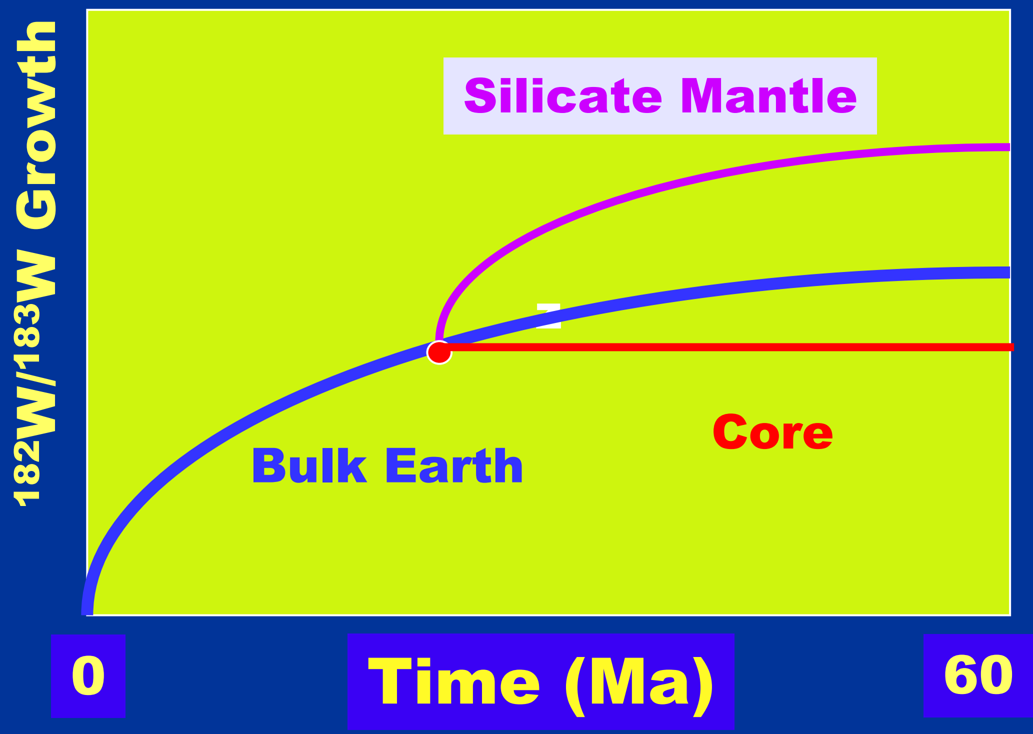
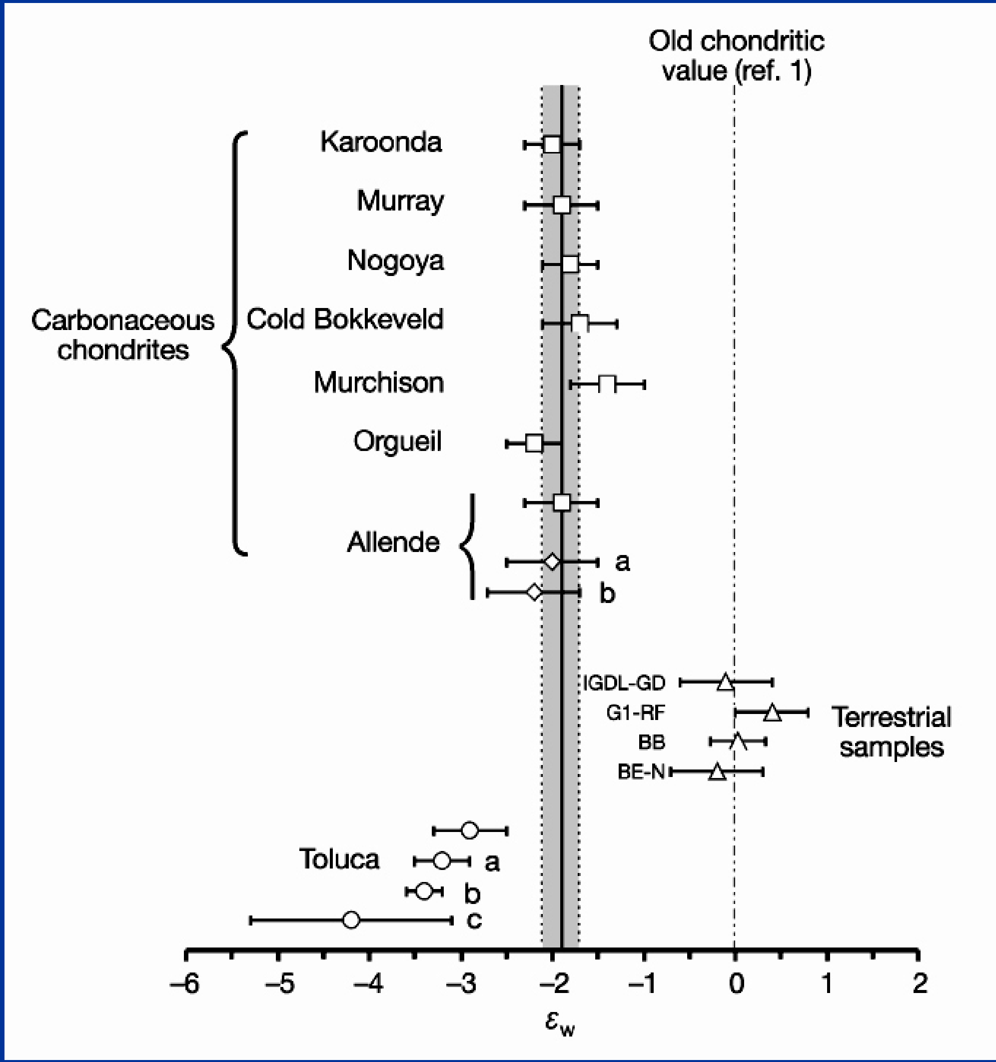


Fig. 2. Schematic showing the fractionation of Hf from W in the early solar system.



So what do the data look like?



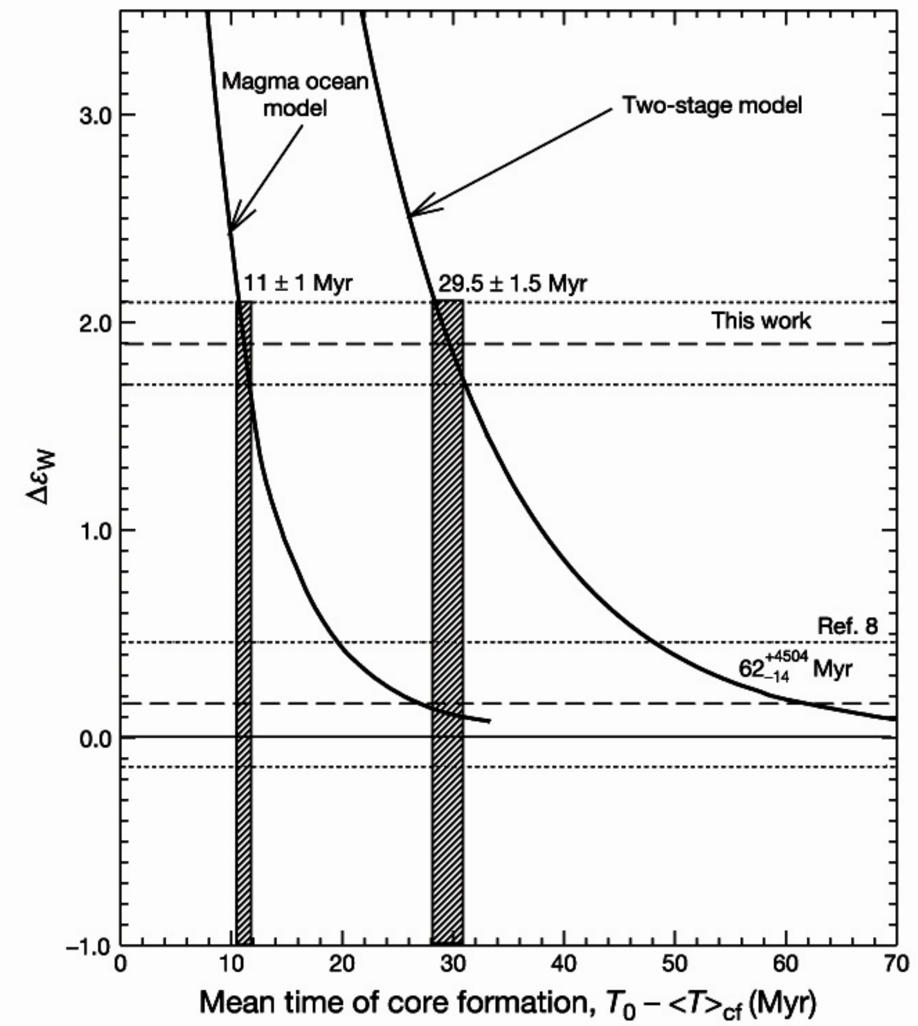
Kleine et al. 2002, Yin et al. 2002

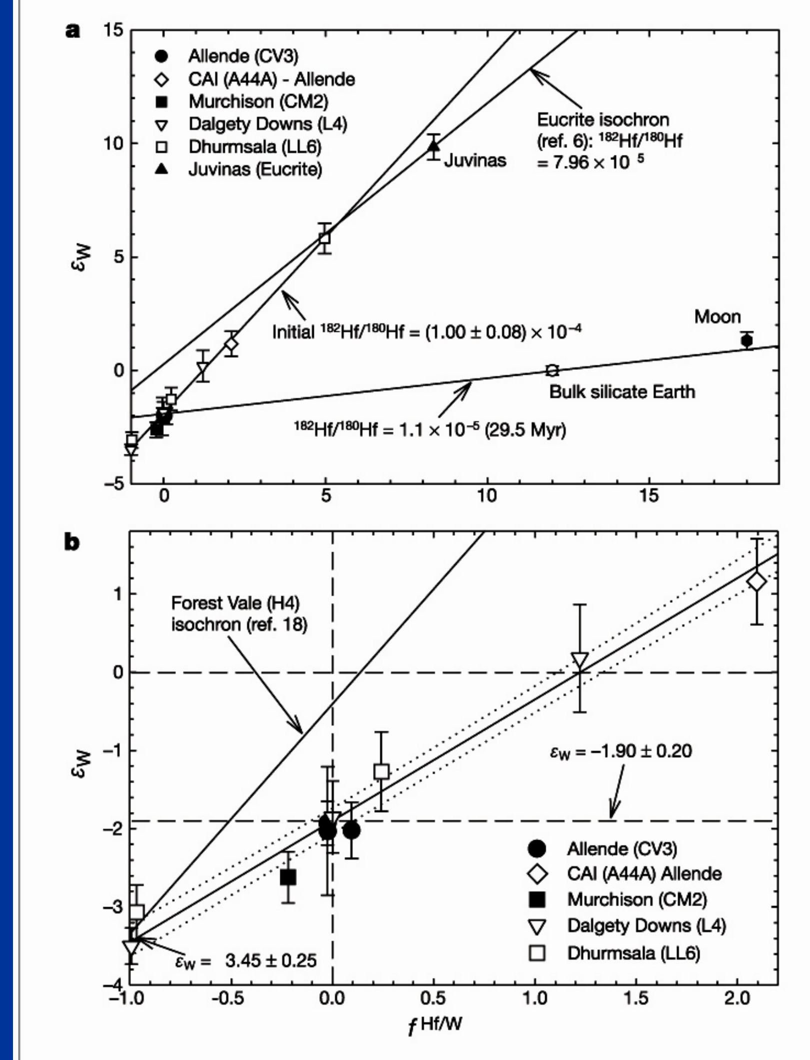
Compared to primitive meteorites, Earth's mantle has an excess of ^{182}W. Core formation must have happened when ^{182}Hf was still alive.

Mean time of core formation was 11 Myr and core-formation was completed within 30 Myr

Moon forming impact at ~30 Myr (Yin et al 2002; Jacobsen 2005)

ε is the deviation of a sample from a reference or standards in parts per 10,000





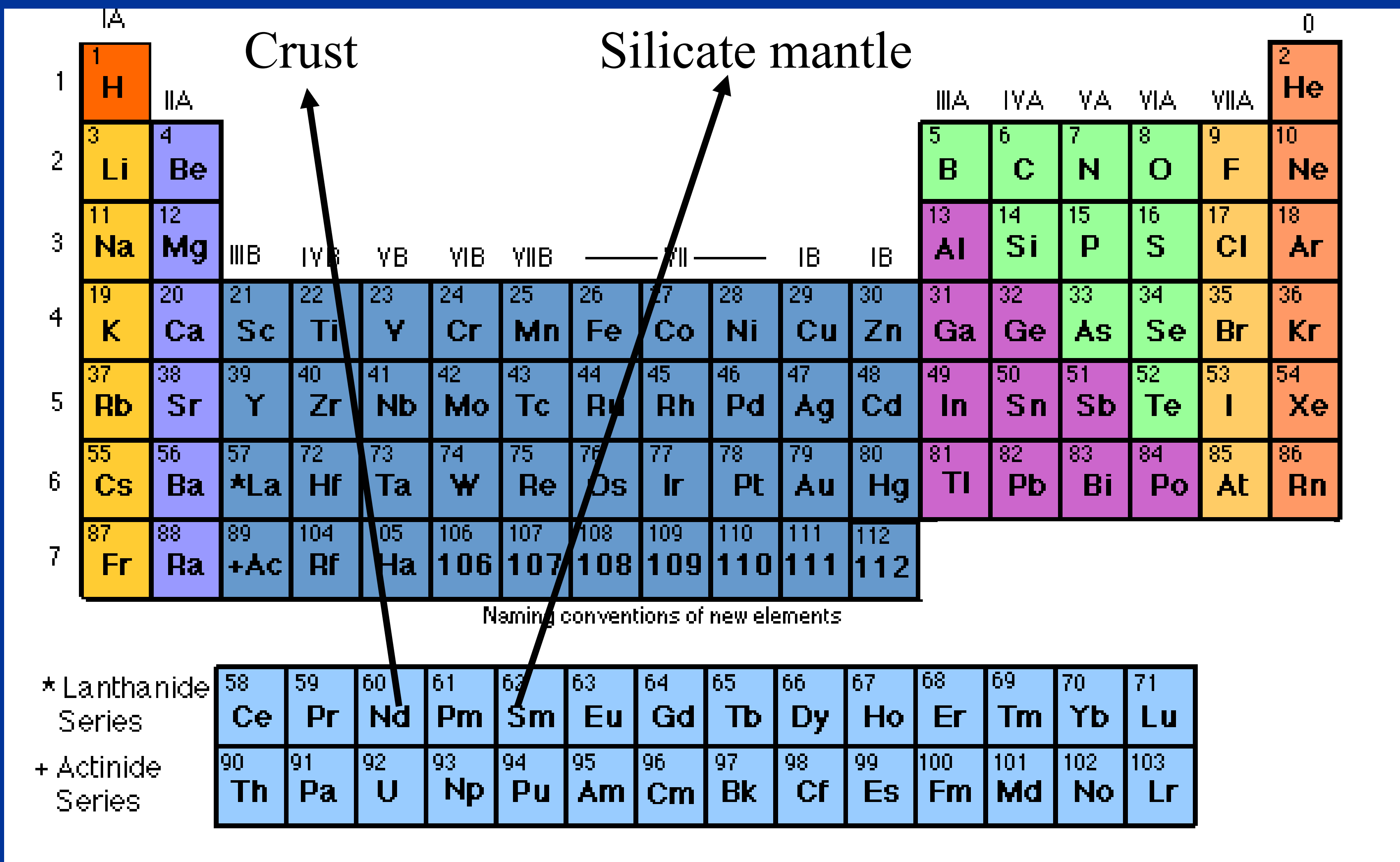
Yin et al., 2002

Mean time of core formation was 11 Myr and core-formation was completed within 30 Myr

Moon forming impact at ~30 Myr (Yin et al 2002; Jacobsen 2005)

Samarium –Neodymium (^{146}Sm-^{142}Nd) systematics; $t_{1/2}$ = 103 M.y

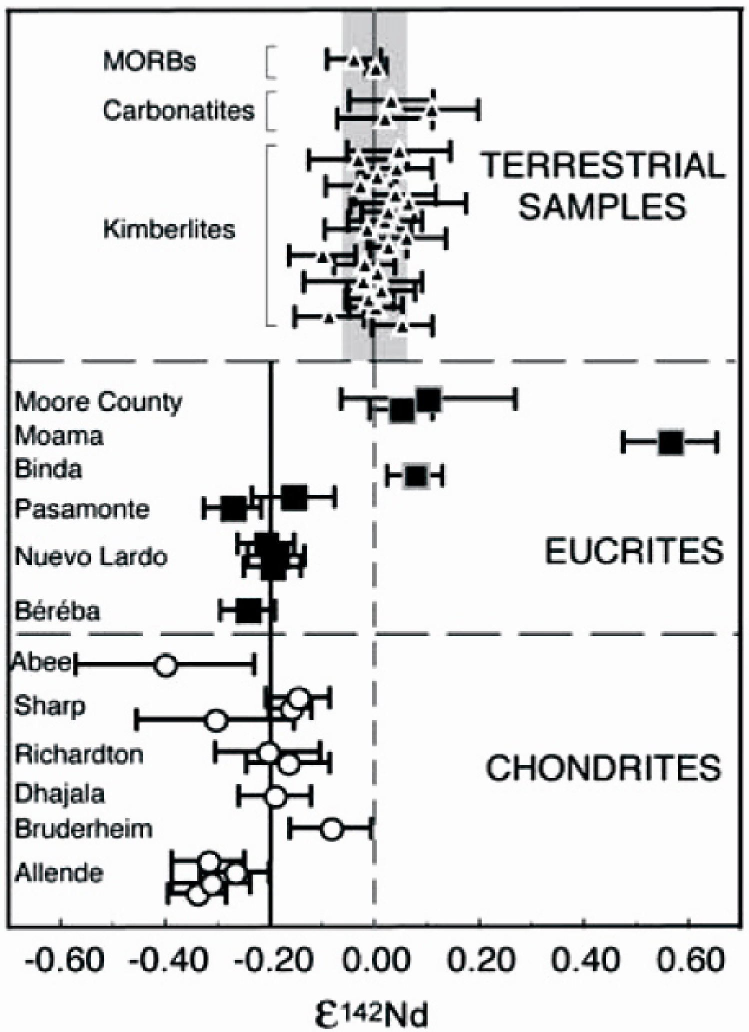
Both Sm and Nd are *lithophile*. Both are incompatible. However Nd is more incompatible than Sm. So ideal for tracking crust-mantle differentiation.



The early differentiation of the Earth's mantle: evidence from ^{142}Nd

$^{146}Sm \rightarrow {}^{142}Nd$; $t_{1/2}$= 103 My.

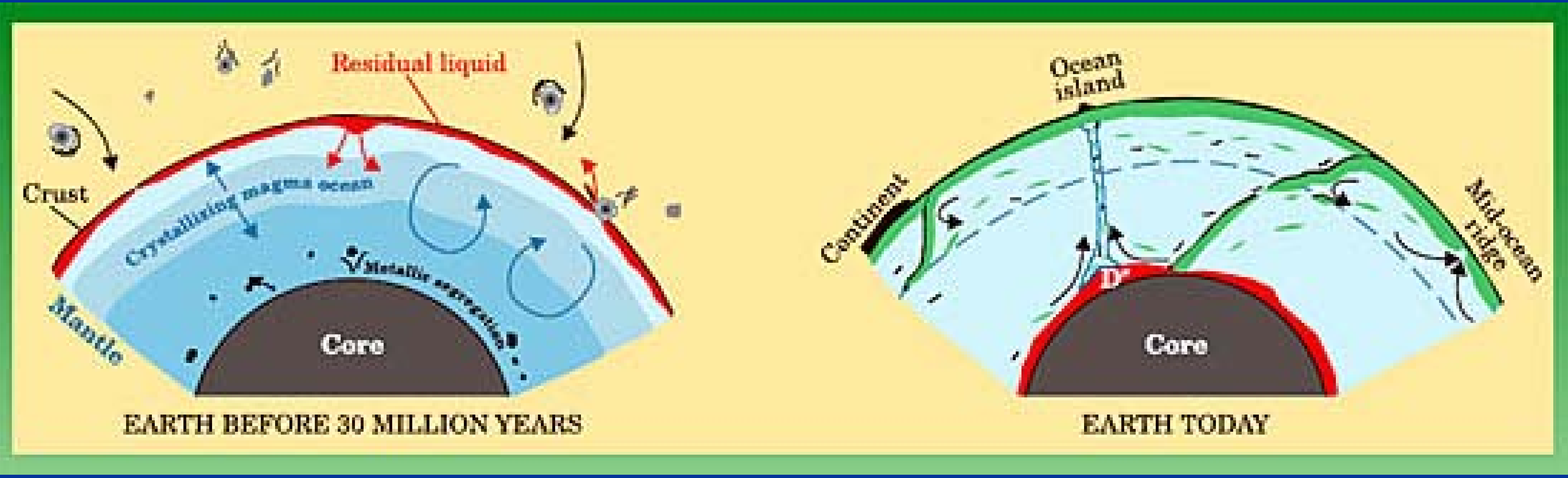
$^{147}Sm \rightarrow {}^{143}Nd$; $t_{1/2}$ = 109 Gy



Early crust-mantle differentiation; ~30 Myr after start of the solar system.

Boyet and Carlson (2005)

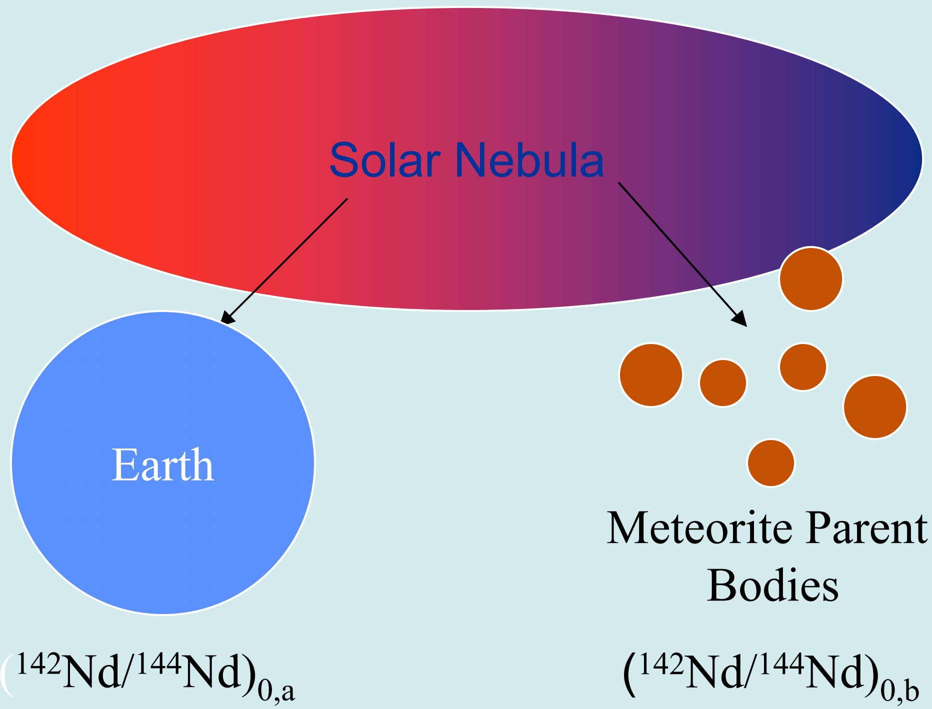
Boyet and Carlson (2005) attributed the ^{142}Nd difference due to formation of an early enriched layer (at ~ 30 Myr) that subsequently sank back into the mantle; this hidden layer is not sampled today at either mid ocean ridge volcanism or ocean island volcanism.



The enriched layer would be enriched in heat producing radioactive elements

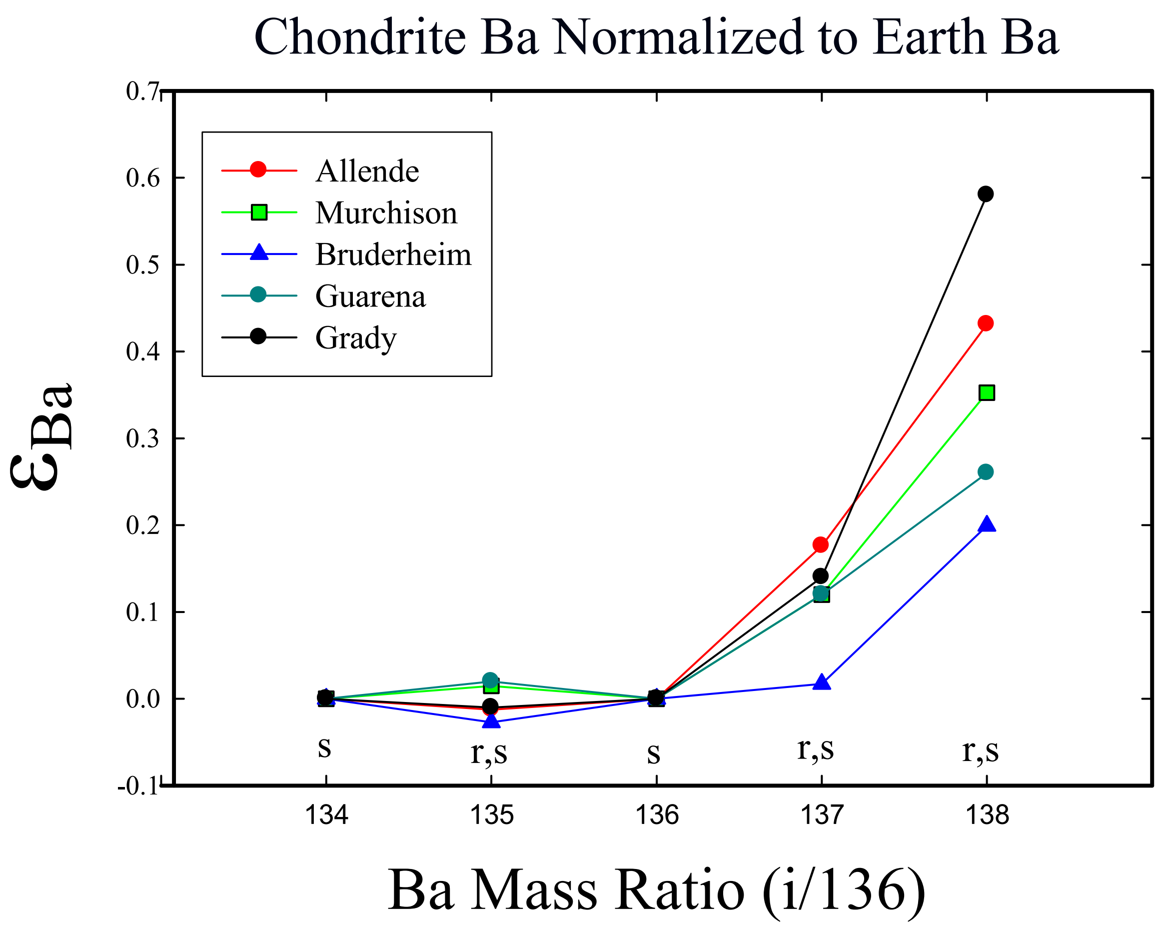
But what we one assumes that the different stellar components were not well mixed in the solar system?

Is it possible that because of incomplete mixing of r- and s- process components Earth and meteorites had different starting $^{142}Nd/^{144}Nd$ values?



To test this idea one needs to find another heavy element having isotopes produced by both the r- and s-process but not having any significant contribution from radioactive decay

Barium → isotopes are not affected by radioactive decay



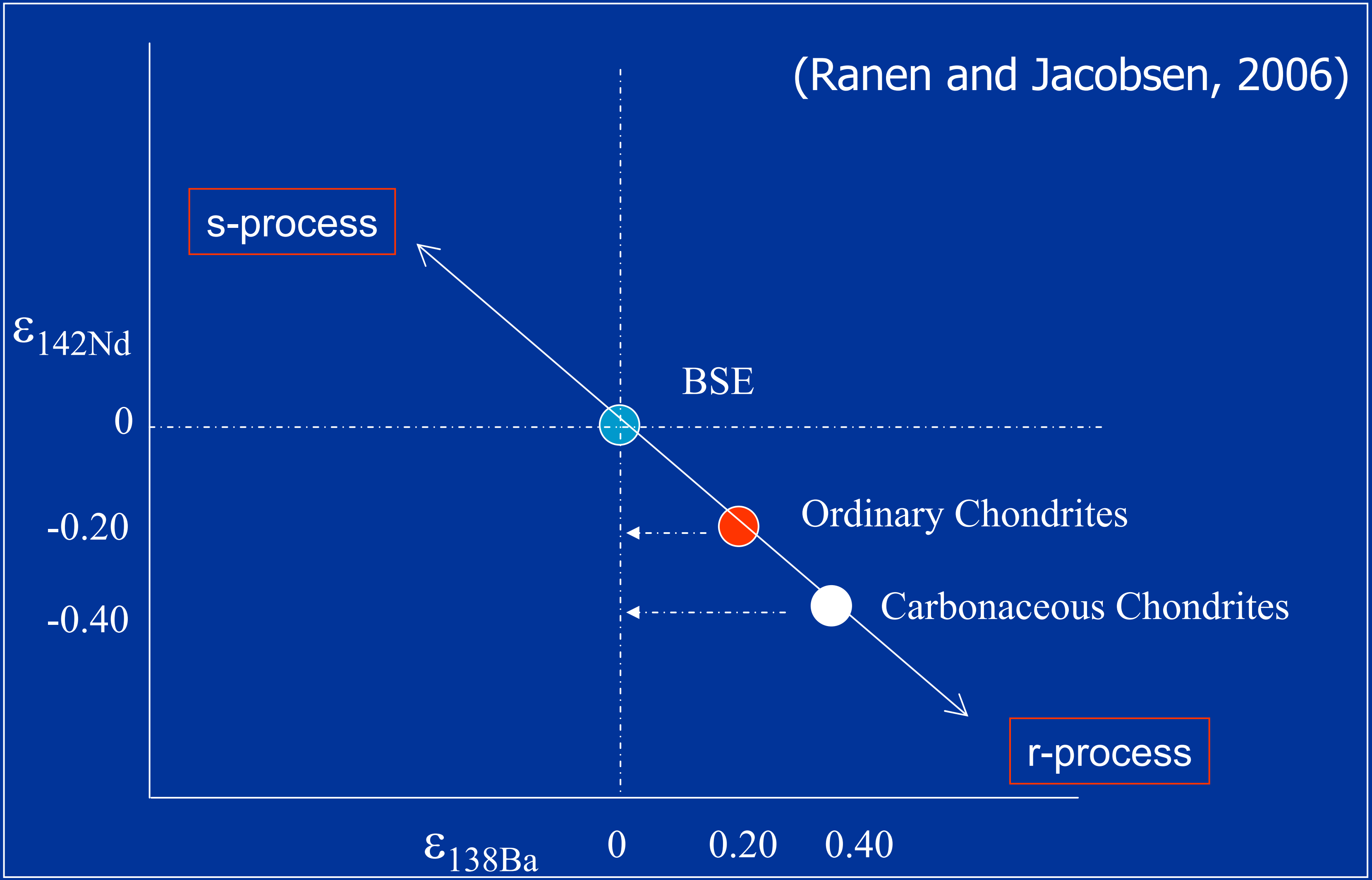
Ranen and Jacobsen, 2006

Chondrites have an excess of r-process produced Ba-isotopes

The Ba isotopic differences can only be formed by incomplete mixing of stellar components between the Earth and Meteorite parent bodies

(i.e. decay of radioisotopes cannot contribute to these variations)

Schematic results for Ba and Nd

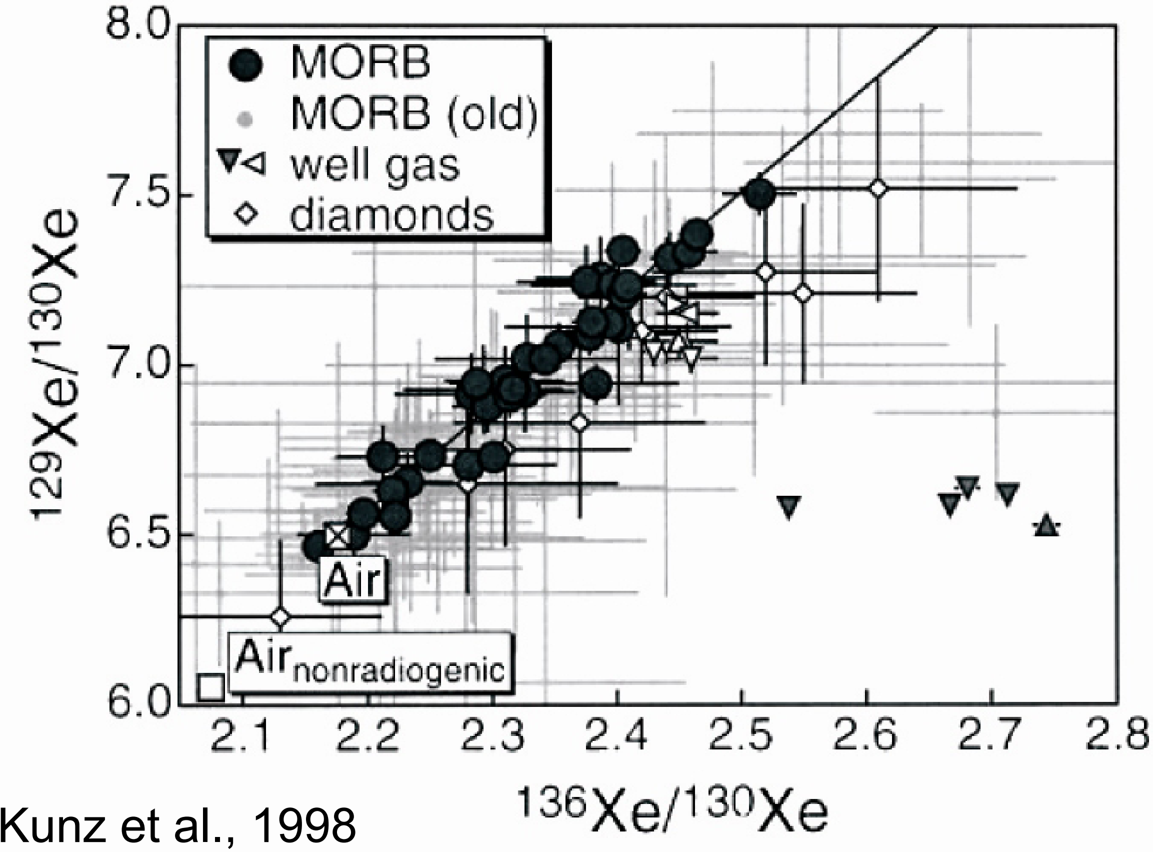


The results for Ba predict a negative ^{142}Nd effect in chondrites compared to Earth as observed → so did the Earth really undergo an early episode of crustal differentiation?

Pu-I-Xe systematics

$^{129}I \rightarrow ^{129}Xe$; $t_{1/2}$=16My

$^{244}Pu \rightarrow ^{136}Xe$; $t_{1/2}$ = 80My; 136Xe also produced from U fission



Kunz et al., 1998

Fundamental observation: MORB data differ from atmosphere, and show mixing between air and a component with excesses of both ^{129}Xe (from ^{129}I decay) and fissiogenic Xe isotopes (from Pu and/or U decay)

If Xenon in the mantle is complimentary to the atmosphere, degassing of the mantle to produce the atmosphere happens in 50-80Myr.

Alternatively, the atmosphere not formed through mantle degassing but rather came from the late veneer.

What do we know about the early atmosphere on Earth?

- Inventory established "early"
  - Uncertainties beyond this
    - (sources, mechanisms, precise timing)
- Elemental abundances and isotopic composition have been modified by atmospheric loss processes

How could volatiles be incorporated into planet Earth?

- Capture of nebular gases – atmosphere of solar composition- dissolved into magma ocean

- Impact degassing – H_2O (steam) rich atmosphere

- Delivery by cometesimals – late stage of accretion

How could volatiles be incorporated into planet Earth?

Capture of nebular gases

atmosphere of solar composition dissolved into magma ocean → dissolved hydrogen could reduce Fe-silicate to produce water (e.g., Sasaki 1990).

Noble gases would have solar composition

Elemental abundances in the mantle will be set by the atmospheric pressure and solubility in a magma

How could volatiles be incorporated into planet Earth?

2. Impact degassing – H_2O (steam) rich atmosphere

- Formation of steam atmosphere starts at ~0.01 $M_{\oplus}$
- Blanketing effect of the atmosphere can raise surface temperatures high enough for melting

Noble gases would have the composition of the material that made up the planet (i.e. isotopically not solar)

But the isotopes do not add up……

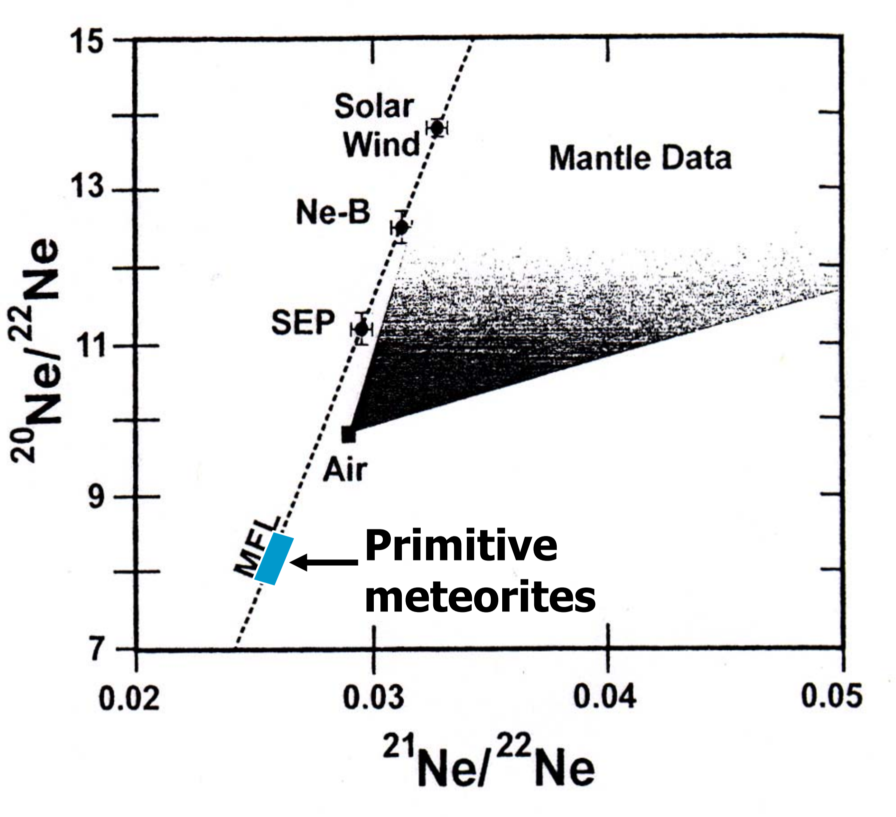


Figure from Pepin and Porcelli, 2002

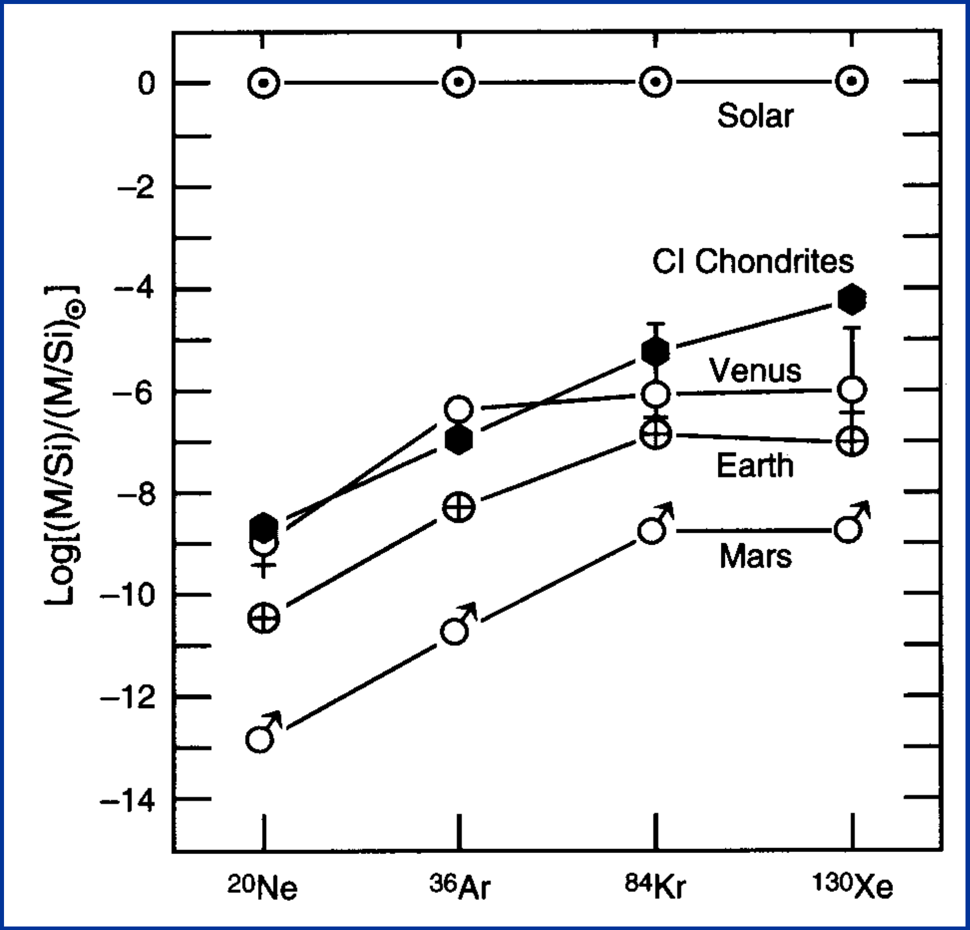
Earth started off with a solar composition

Maybe mixture of nebular and planetary gases

<u>Bottom line</u>
Need solar composition gas and need to get it into the deep mantle in high abundance

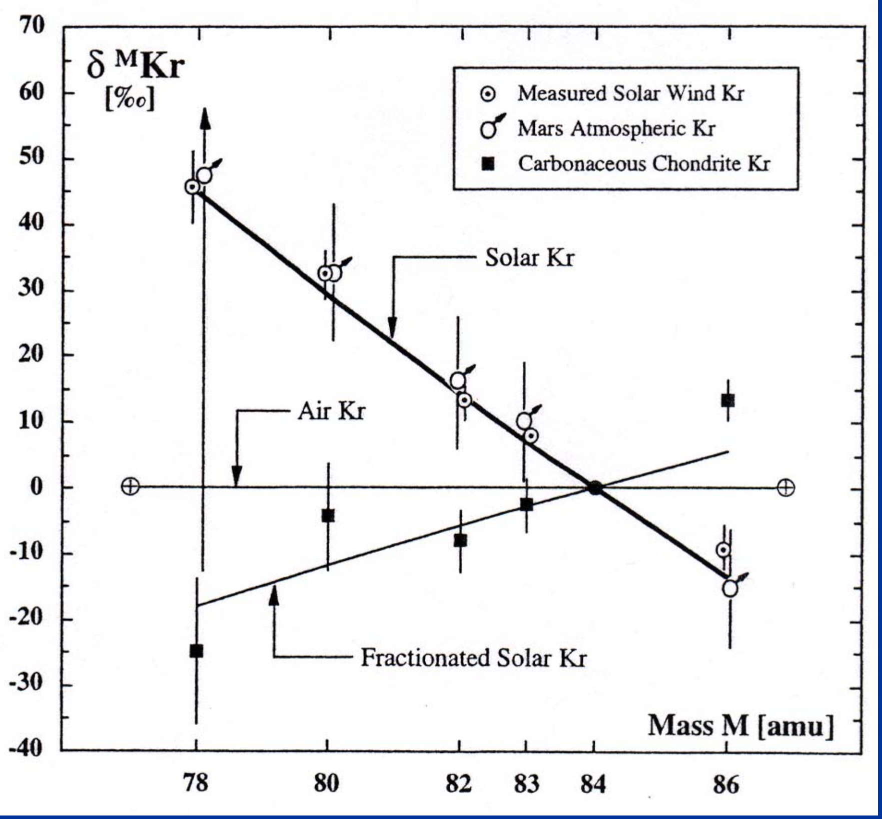
→ magma ocean is a logical way to ingas the planet (e.g., Hayashi et al., 1979; Jacobsen and Harper, 1996)

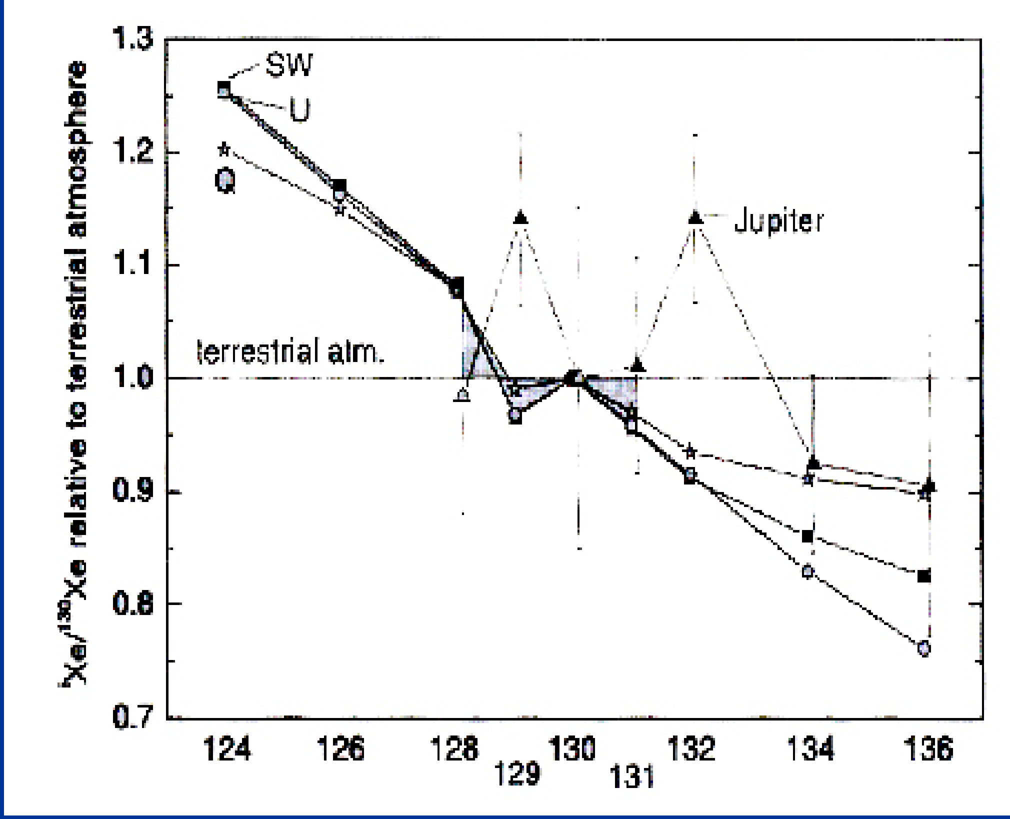
Noble gases in the atmosphere of terrestrial planets



1. Massive depletion of volatiles from Earth
2. Abundance pattern looks like carbonaceous meteorites

Krypton and Xenon in the atmosphere





>90% of the Xenon in the atmosphere was lost in the first 60-80 Myr
The Earth's atmosphere has preferentially lost the lighter isotopes → loss mechanism has to explain this

Pepin and Porcelli, 2002

What processes can strip the early atmosphere on Earth

- Hydrodynamic escape

  Extreme UV radiation from sun heats a H-rich atmosphere

  H escape fluxes can be large enough to exert upward drag forces on heavier atmospheric constituents

  Lighter isotopes will be preferentially lost

  May need unrealistically large EUV flux (500 times larger than present day and may need this for 50 Myrs)

  If significant CO_2 or CO present in the atmosphere H_2 escape flux will be limited by H_2 diffusion through CO_2

(Hunten et al., 1987; Pepin, 1991; Zahnle and Kasting, 1986)

What processes can strip the early atmosphere on Earth

Moon forming giant impact ~30 Myr after the start of the solar system (Yin et al., 2002)

leads to bulk erosion of the atmosphere

not capable of fractionating species in the atmosphere

Based on radiogenic Xenon,
30 Myr maybe too early for atmospheric loss….