Lecture 32

Planetary Accretion –
Growth and Differentiation of Planet Earth

Reading this week: White Ch 11 (sections 11.1 -11.4)
Today – Guest Lecturer, Greg Ravizza

1. Earth Accretion
2. Core Formation – to be continued in Friday's lecture

Next time

The core continued, plus, the origin of the moon

Growth and Differentiation of Planet Earth

Last lecture we looked at *Boundary Conditions* for earth's early history as a prelude to deducing a likely sequence of events for planetary accretion.

Today we consider three scenarios for actual accretion:

❖ homogeneous condensation/accretion, followed by differentiation – *Earth accretes from materials of the same composition AFTER condensation, followed by differentiation*

❖ heterogeneous condensation/accretion (partial fractionation of materials from each other) before and during planetary buildup – *Earth accretes DURING condensation, forming a differentiated planet as it grows.*

❖ something intermediate between these two end-members

Growth and Differentiation of Planet Earth

Both homogeneous and heterogeneous accretion models require that the **core segregates** from the primitive solid Earth at some point by melting of accreted Fe.

The molten Fe sinks to the Earth's center because it is denser than the surrounding silicate rock and can flow through it in the form of droplets.

However, the original distribution of the Fe and the size and character of the iron segregation event differ between these two models.

Heterogeneous accretion

Earth builds up first from cooling materials (initially hot and then cool). This requires a very rapid build up of earth, perhaps within 10,000 yrs of the initiation of condensation.

- **The early core of the earth contains the refractory elements/early condensates.**
- **Silicate and metal phases condense next, with denser metal particles being concentrated near the core.**
- **Subsequent heating due to impacts and radioactive decays causes the lower Fe-rich mantle and refractory element core to swap places (within the first 10-100 million years of condensation).**

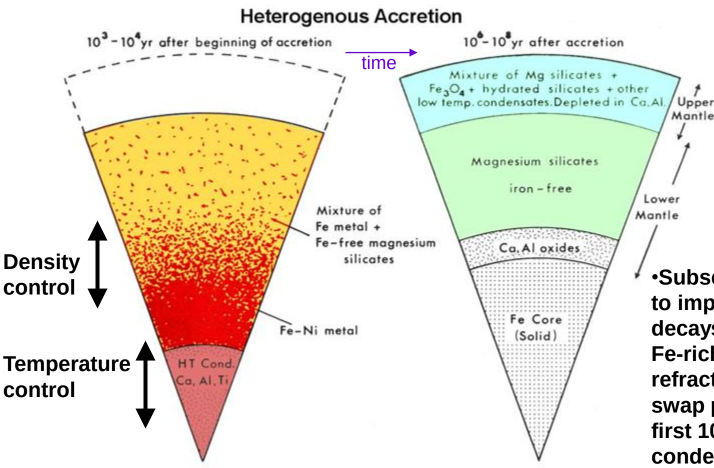


Figure 7.2 Zonal structure of Earth suggested by heterogeneous accumulation hypotheses of Turekian and Clark (1969) et al. (1972).

Heterogeneous accretion

The pluses of this model:

1. Qualitatively explains the **density differences among the stoney planets**.

   Because of the approximately radial temperature gradient in the solar nebula, each planet accretes with a different proportion of higher-T and lower-T condensates.

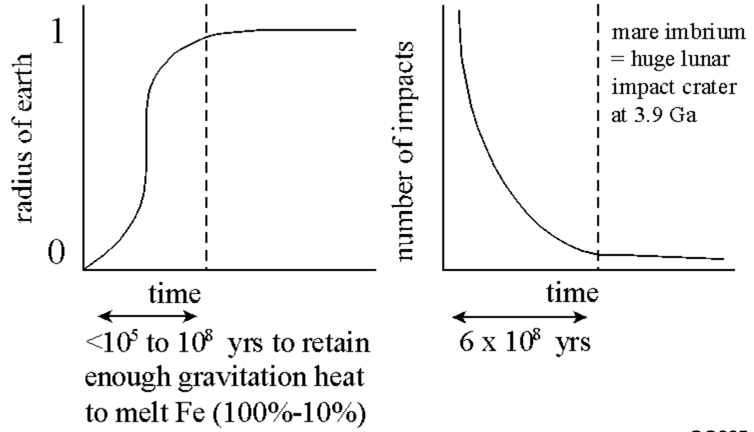
2. Helps explain the fact that **some siderophile elements are more abundant in the mantle than what we expect for complete equilibrium between molten Fe and silicate minerals**.

   In this model, siderophile elements in some of the later condensates are never equilibrated with molten iron and therefore remain in the silicate mantle.

Heterogeneous accretion   The minuses of this model:

1. requires very fast accretion (10^3-10^4 yrs to complete). Inconsistencies:

a. Most models of accretion rate (see last lecture)
b. the occurrence of large impact craters at 3.9 Ga on the moon



Early Earth History Summarized

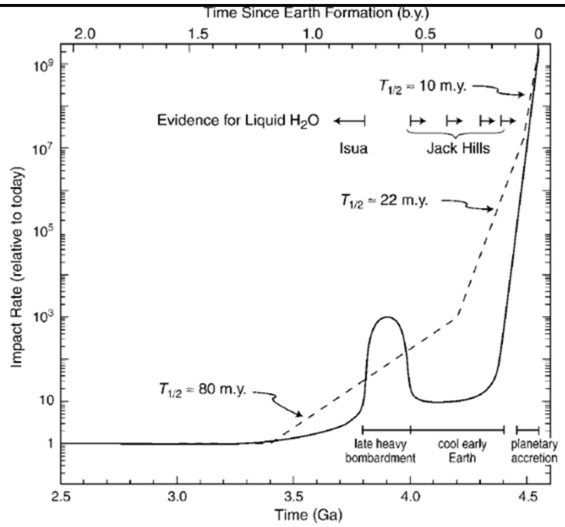


Figure 3. Estimates of meteorite-impact rate for first 2 b.y. of Earth history. Two hypotheses are shown: exponential decay of impact rate (dashes, Hartmann et al., 2000) and cool early Earth–late heavy bombardment (solid curve, this study). Approximate half-life is given in million years for periods of exponential decline in flux. In either model, spikes occurred owing to isolated large impacts. Evidence for liquid water comes from high-$\delta^{18}O$ zircons (>4.4 to >4.0 Ga) and sedimentary rocks (Isua, 3.8–3.6 Ga). The cool early Earth hypothesis (solid curve) suggests that impact rates had dropped precipitously by 4.4 Ga, consistent with relatively cool conditions and liquid water.

Heterogeneous accretion More **minuses** of this model:

2. Requires an Fe-free lower mantle with Fe/(Mg+Fe) ~0

 This is because Fe is segregated as metal while still hot and Fe doesn't enter silicates until cooler temperatures (~450° C at atmospheric pressure).

 But…seismic elasticity of the lower mantle requires Fe/(Mg+Fe) ~0.1

3. Requires a refractory element enriched (Ca, Al, Ti) rich lower mantle and an upper mantle depleted in these elements, which is also not observed.

Heterogeneous accretion More **minuses** of this model:

4. It is difficult to get ~10% of one or more of the light elements into the core as required by the seismic data.

5. It creates a volatile depleted lower mantle (not observed).

To summarize the minuses, it creates a compositionally layered mantle with respect to Fe, Ca, Al and volatile element abundances that are not observed.

Some of this could have subsequently been erased by mantle convection and mixing after accretion, but the Fe-free silicate requirement is a fatal flaw for this model.

Homogeneous accretion

Earth builds from cool materials first and then becomes hotter.

The more commonly accepted notion: Earth accreted mostly homogeneously after condensation was complete.

Important aspects:

a. Heat builds up as the planet accretes.

b. Sometime afterwards, the core formed by Fe melting, accompanied by other chemical transformations (see next slide)

c. Requires later mantle overturn during core formation.

Homogeneous Accretion

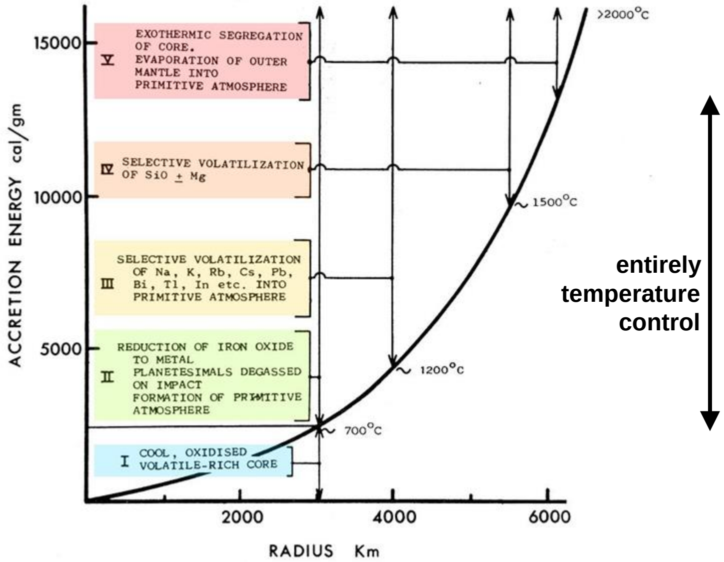


Figure 7.1 Relationship between energy of accretion and radius of a growing Earth-sized terrestrial planet. The principal stages of accretion are also shown in relation to the energy of accretion and approximate surface temperatures. (After Ringwood, 1970; with permission.)

Homogeneous accretion

This model begins with **relatively oxidized solids** that have been physically segregated from much of the nebular H_2 (i.e., like C1-C3 chondrites).

However, there is enough H_2 present in the early stages for Fe to be in the reduced form (i.e., as Fe metal rather than FeO).

Later, as the young Earth heats up, some of that iron combines with silicate material. This silicate becomes more enriched in oxidized Fe^{+2} through the following reaction:

$MgSiO_3 + Fe + H_2O \leftrightarrow (Mg,Fe)_2SiO_4 + H_2$ (favored at higher T)

although some SiO_2 may also be lost at even higher T by:

$MgSiO_3 \leftrightarrow Mg_2SiO_4 + SiO_2$ (g) (also favored at higher T)

Homogeneous accretion

Built into the model is the notion that about 10% of the silicate earth retained volatiles during accretion and the other 90% was degassed during accretion.

The “primitive mantle” composition mentioned last lecture is essentially this 90-10% mix.

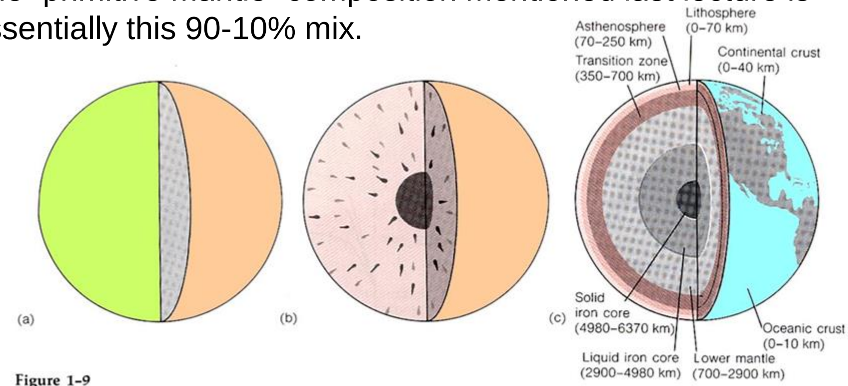


Figure 1-9
The early Earth (a) was probably a homogeneous mixture with no continents or oceans. In the process of differentiation, iron sank to the center and light material floated upward to form a crust (b). As a result, the Earth is a zoned planet (c) with a dense iron core, a surficial crust of light rock, and, between them, a residual mantle.

Press & Siever, Earth

Homogeneous accretion

Heat is required to melt the Fe that will ultimately form the core, the three main sources being:

a. Accretion b. Radioactivity c. Gravity

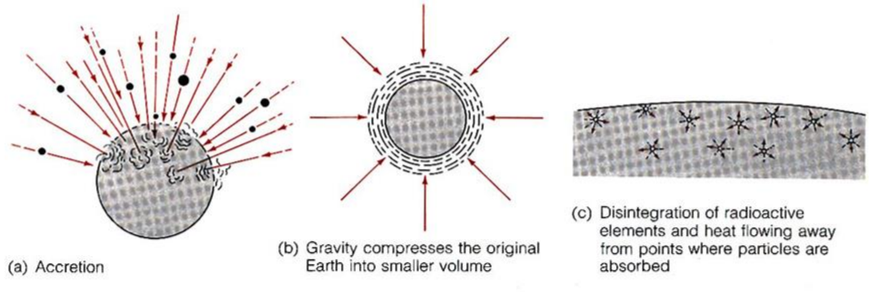


Figure 1-6
Three mechanisms that would cause the early Earth to heat up. (a) In accretion, impacting bodies bombard the Earth and their energy of motion is converted to heat. (b) Gravitational compression of the Earth into a smaller volume causes its interior to heat up. (c) Disintegration of radioactive elements releases particles and radiation, which are absorbed by the surrounding rock, heating it.

Press & Siever, Earth

Homogeneous accretion   The **pluses** of this model:

1. provides a mechanism to get volatile elements to core.

2. provides a better mechanism to get most siderophile elements to core …

and makes siderophile element abundances in the upper mantle low.

3. provides a heat source for early mantle melting/ formation of proto continents.

Homogeneous accretion   The **minuses** of this model:

1. The degassing of all but 10% of the volatile elements:

Matches the moderately volatile concentrations well (1300°-600° category from last week)

Does not match the lower T volatiles, which are too abundant.

2. The siderophile element abundances of the upper mantle are too high for equilibration of molten Fe-Ni and silicate.

3. Suggests a cool accretion, with heat for melting Fe coming later.

Is there enough heat to melt Fe?

Core Formation

What is clear from this discussion is that the role of the core in the accretion process is crucial (as one might expect given that the core is 32.5% of the Earth's mass…

Core formation requires:

- Immiscible components (iron metal and silicate).
- Macro-segregation of components: at least one of which was molten or mostly molten.
- Substantial difference in density of components.
- Gravitational settling

gravitational energy from Fe sinking to the core from a melt annulus near the surface, provides additional gravitational heating to induce more melting.

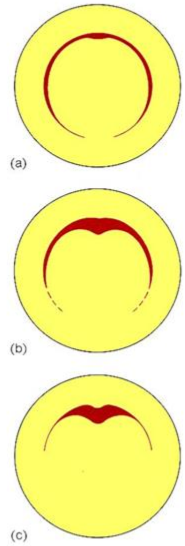


Figure 1-8 *Press & Siever, Earth*
(a) The melting of iron leads to the formation of a heavy liquid layer. Drops begin to develop in later stages (b, c) and sink toward the center.

Core Formation

Models of the thermal profile for post-accretionary earth suggest that it is difficult to accumulate enough **accretional heat** within the early earth to melt Fe in the interior if Earth starts out cool, because the Fe melting temperature increases with pressure.

Melting seems possible near the surface if accretion is rapid, but this won't form a core of the size needed.

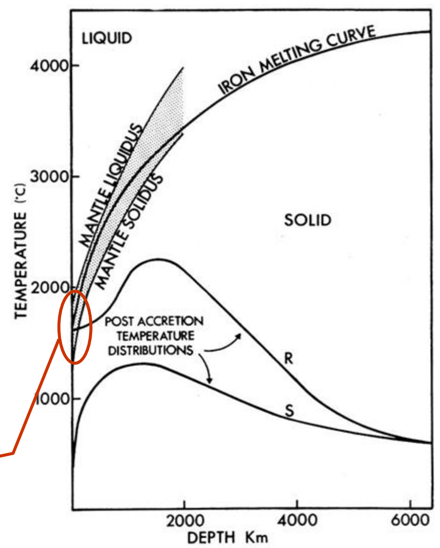


Figure 3.5 Estimates of melting temperatures of iron and melting interval of mantle (Higgins and Kennedy, 1971; Kennedy and Higgins, 1973) together with estimates of post-accretional temperature distributions within the Earth and prior to core formation. S = Safronov (1972a), R = Ringwood (1975a).

Core Formation

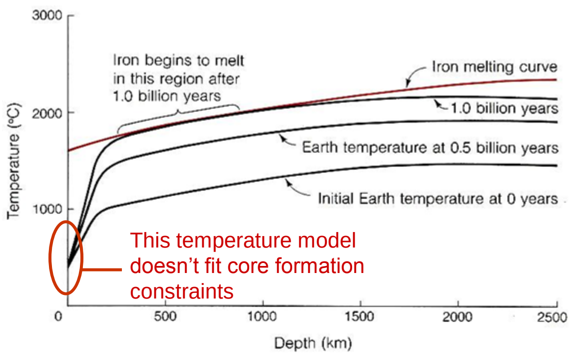


Figure 1-7 *Press & Siever, Earth*
Temperature in the Earth's interior at different times in its history, according to a calculation by T. C. Hanks and D. L. Anderson. The lowest curve shows the initial temperature due to accretion and compression at 0 years. After 500 million years radioactivity warmed the Earth to the temperature shown by the next curve. After one billion years the interior heated to the melting point of iron at depths between 400 and 800 km, and iron began to melt in this region.

Adding heat from **radioactive decay** helps, but thermal models for heating the Earth from an initially cool temperature at the end of condensation imply that the interior doesn't heat up enough to melt much Fe until at least 1 Gyr after accretion,

This is much too slow to fit the bounds on the time of core formation from isotopic evidence (we'll get to this evidence soon).

Core Formation

One possibility is that the **melting temperature of iron was depressed** by the addition of some FeO. It's not clear how likely this is or how it may have happened.

A magma ocean:
With melting of iron throughout much of the mantle, it's difficult to avoid melting at least the upper several hundred kilometers of the mantle as well.

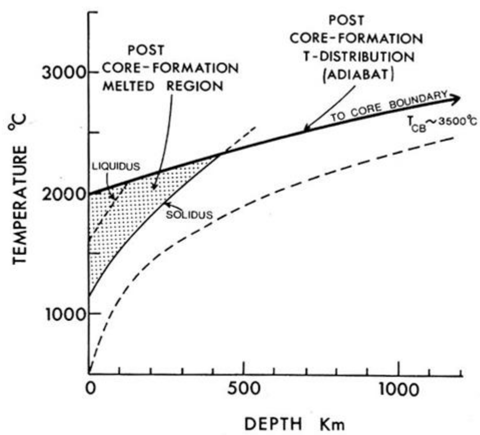


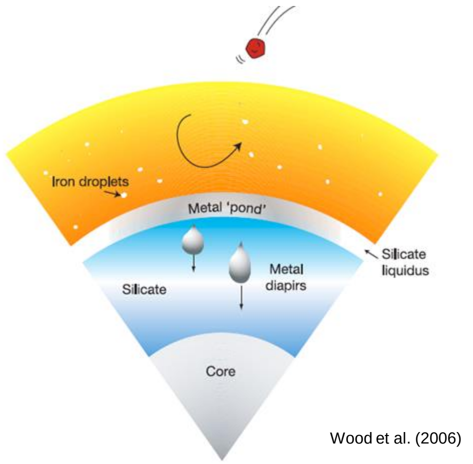
Figure 3.6 Maximum temperature distribution in mantle following catastrophic core formation, in relation to pyrolite solidus and liquidus. The mantle temperature distribution followes an adiabat from 2000°C, based on data of Birch (1952) and Verhoogen (1956). Note extensive molten and partly molten zone extending to a depth of 400 km. Also shown is an estimate of present mantle temperature distribution (broken line). (From Ringwood, 1975a, with permission.)

Core Formation

A recent proposal is that early collisional heating melted produced a deep magma ocean, which persisted as accretion continued.

Temperatures below this may have remained relatively cool, meaning that the lower mantle would have retained many of it's siderophile elements …

because the ***lower*** mantle didn't equilibrate with the diapirs of molten Fe that subsequently sank through it.



Wood et al. (2006)

Figure 3 | The deep magma ocean model. Impacting planetesimals disaggregate and their metallic cores break up into small droplets in the liquid silicate owing to Rayleigh–Taylor instabilities. These droplets descend slowly, re-equilibrating with the silicate until they reach a region of high viscosity (solid), where they pond in a layer. The growing dense metal layer eventually becomes unstable and breaks into large blobs (diapirs), which descend rapidly to the core without further interaction with the silicate. Note that the liquidus temperature of the silicate mantle should correspond to pressure and temperature conditions at a depth above the lower solid layer and plausibly within the metal layer as indicated.

Core Formation Timing

Timing constraints on core formation are critical for understanding Earth's time-temperature path during accretion:

1. Physical models of accretion suggest that the core formed nearly simultaneously with accretion.

2. Pb isotopes provided one of the first important bounds on the timing of core formation, which now know is too high.

A comparison of $^{206}Pb/^{204}Pb$ in chondrites and values estimated for the early mantle (from Pb ores in dated Archean rocks) imply that *core formation must have occurred **within about 500 Myr** after the chondrites formed*.

Note: ^{206}Pb is produced by ^{238}U decay, whereas ^{204}Pb is an s-process nuclide (non-radiogenic).

This limit would rule out purely homogeneous accretion, but is now believed to be much too long to be correct.

Core Formation Timing

3. Early crust? Temperatures are most favorable for melting Fe in the upper part of the Earth, however, impact heating would be impeded if a substantial crust existed early on because much of the impact energy is spent removing the crust, not melting Fe-rich mantle).

The oldest dated (by U-Pb dating) mineral grains on Earth are zircons; zircon is a crustal mineral. They indicate that we had some sort of crust by 4.2-4.3 Ga.

This sets a likely *upper bound on core formation of* ***250 to 350 Myr*** *after chondrite formation*, *although this also now appears too long for other reasons.*

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Core Formation Timing – Extinct radionuclides

Several short-lived isotope systems provide important evidence on the rate of core formation, as well as other aspects of early solar-system differentiation (see Lect. 30).

Some examples of the most recently exploited short-lived radionuclide systems are in the table below. Four of these are particularly good at telling us when Fe melted during accretion.

Table 1 Short-lived radioactive nuclides once existing in solar system objects.[a]

Fractionation[b]	*Parent nuclide*	*Half-life* (Myr)	*Daughter nuclide*	*Estimated initial solar system abundance*	*Objects found in*	*Reference*
	^{41}Ca	0.1	^{41}K	$10^{-8} \times {}^{40}Ca$	CAIs	(1)
	^{26}Al	0.7	^{26}Mg	$(4.5 \times 10^{-5}) \times {}^{27}Al$	CAIs, chondrules, achondrite	(2)
Nebular	^{10}Be	1.5	^{10}B	$(\sim 6 \times 10^{-4}) \times {}^{9}Be$	CAIs	(3)
	^{53}Mn	3.7	^{53}Cr	$(\sim 2\text{-}4 \times 10^{-5}) \times {}^{55}Mn$	CAIs, chondrules, carbonates, achondrites	(4)
	^{60}Fe	1.5	^{60}Ni	$(\sim 3 \times 10^{-7}) \times {}^{56}Fe$	achondrites, chondrites	(5)
	^{107}Pd	6.5	^{107}Ag	$(\sim 5 \times 10^{-5}) \times {}^{108}Pd$	iron meteorites, pallasites	(6)
	^{182}Hf	9	^{182}W	$10^{-4} \times {}^{180}Hf$	planetary differentiates	(7)
Planetary	^{129}I	15.7	^{129}Xe	$10^{-4} \times {}^{127}I$	chondrules, secondary minerals	(8)
	^{92}Nb	36	^{92}Zr	$10^{-4} \times {}^{93}Nb$	chondrites, mesosiderites	(9)
	^{244}Pu	82	Fission products	$(7 \times 10^{-3}) \times {}^{238}U$	CAIs, chondrites	(10)
	^{146}Sm	103	^{142}Nd	$(9 \times 10^{-4}) \times {}^{147}Sm$	chondrites	(11)

References: (1) Srinivasan *et al.* (1994, 1996), (2) Lee *et al.* (1977), MacPherson *et al.* (1995); (3) McKeegan *et al.* (2000); (4) Birck and Allègre (1985), Lugmair and Shukolyukov (1998); (5) Shukolyukov and Lugmair (1993a), Tachibana and Huss (2003); (6) Chen and Wasserburg (1990); (7) Kleine *et al.* (2002a), Yin *et al.* (2002); (8) Jeffery and Reynolds (1961); (9) Schönbachler *et al.* (2002); (10) Hudson *et al.* (1988); and (11) Lugmair *et al.* (1983)
[a] Some experimental evidence exists suggesting the presence of the following additional isotopes, but confirming evidence is needed (half-lives are given after each isotope): ^{7}Be—53 d (Chaussidon *et al.*, 2002); ^{99}Tc—0.2 Myr (Yin *et al.*, 2000); ^{36}Cl—0.3 Myr (Marty *et al.*, 1997); ^{205}Pb—15 Myr (Chen and Wasserburg, 1987). [b] Environment in which most significant parent–daughter fractionation processes occur.

(McKeegan & Davis 2003)

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Core Formation Timing

Isotope anomalies of short-lived, now extinct radionuclides, for which parent and daughter isotopes have different affinities for molten iron and silicate:

Collectively, 4 systems indicate that Earth had molten Fe early on during accretion.

1. Ni isotopes

Recall from lecture 30 that high $^{60}Ni/^{58}Ni$ in Ca-poor achondrites ("eucrites") relative to terrestrial and lunar rock values implies that differentiation of small meteorite parent bodies to segregated Fe metal from rocky mantle in first **1-2 Ma** of solar system history, very soon after the last nearby r-process event. (Shukolyukov & Lugmair, *Science*, 1993).

($^{60}Fe \rightarrow {}^{60}Co \rightarrow {}^{60}Ni$, $t_{½} = 1.5$ Ma)

Core Formation Timing

2. Ag isotopes:
Isotopic anomalies of silver (Ag) resulting from short lived ^{107}Pd ($T_{½} = 6.5 \times 10^6$ yrs) decay

(Ag is preferentially taken into the Fe phase relative to Pd)

3. Cr isotopes
Isotopic anomalies of chromium (Cr) resulting from short lived ^{53}Mn ($T_{½} = 3.7 \times 10^6$ yrs) decay.

(Cr is preferentially taken into the Fe phase relative to Mn)

Both of these systems show that core formation in meteorite parent bodies began within about **15 Ma** of the last r-process nucleosynthetic event (supernova) (White, 1997). *See the McKeegan & Davis 2003 table 2 slides back for the phases involved.*

Core Formation Timing

4. W isotopes
Tungsten (W) isotope anomalies are caused by decay of ^{182}Hf to ^{182}W ($T_{1/2}$ ~ 9 Myr).

(W is preferentially taken into the Fe phase relative Hf).

W anomalies indicate that all known iron meteorite parent bodies segregated their metal phase within **5 Ma** of each other,

regardless of their composition or size and that core formation on Earth probably happened within **30 Ma** of accretion (various authors).

Core Formation Chemistry

All of the model-based predictions about how and when the core formed depend on *WHAT else* the core is made of besides Fe.

- Iron meteorites contain 5-10% Ni. Removing about 6% of the Ni in a model chondritic Earth to the core fits primitive-mantle abundance estimates pretty well, and is compatible with seismological data.

- As we've discussed, the core must *also* contain about 10% of some light element(s) in order to fit the seismological data.

So this gets us back to: what is the *light element or elements* in the core?

Core Formation Chemistry

The light element(s) required to match the core's density is (are) critical for understanding how other elements partitioned into the core as it is formed.

Candidate light elements include: **O, S, Si, C, P, Mg** and **H.**

For various reasons, **S and/or O** are the most likely light elements.

FeS is miscible with Fe liquid at low and high temperatures

FeO miscibility requires high pressures and temperatures.

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Core Formation Chemistry

Evidence for FeS in the core:

Sulfur is severely depleted in the silicate Earth (more so than O).

S is more depleted than elements with similar volatility, like Zn, so this depletion is probably not just because S is a volatile element.

Iron meteorites contain considerable amounts of FeS (the mineral troilite), demonstrating that S was extracted into the cores of the meteorite parent bodies.

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Core Formation Chemistry

However, none of those three rationale are definitive about FeS vs FeO in the core.

The proportion of S and O relative to Fe affects the solubility of the siderophile and chalcophile elements in the core, so we can turn to these other elements for clues.

Two things are apparent from the tables on the next slide:

a) Sideophile elements are not as low in the mantle as would be expected from pure metal-silicate equilibration

b) Chalcophile elements are depleted in the silicate Earth relative to chondrites, but not as depleted as many of the siderophiles are. This *could* argue against much S in the core.

Core Formation Chemistry

a) Sideophile elements are not as low in the mantle as would be expected from pure metal-silicate equilibration

b) Chalcophile elements are depleted in the silicate Earth relative to chondrites, but not as depleted as many of the siderophiles are. This *could* argue against much S in the core.

Table 2.2 Distribution of some siderophile elements between mantle and core, as compared to experimental Fe metal–silicate coefficients.

Element	Observed K(Mantle/Core)	Experimental K(Silicate/Fe metal)
Involatile	0.033 avg	<0.001
Co	0.1	0.005
Ni	0.08	0.0006
Re	0.001	5×10^{-4}–10^{-6}
Os	0.004	Probably similar to Au and Re
Ir	0.002	
Pt	0.01	
Volatile	0.11 avg	0.005
Au	0.01	$< 3 \times 10^{-5}$
Cu	0.14	0.02–0.003
Ge	0.013	0.001
As	0.4	< 0.01

siderophile elements are 5-350 times more enriched than expected for complete equilibrium. Volatile siderophiles appear to be even more enriched than non-volatile ones.

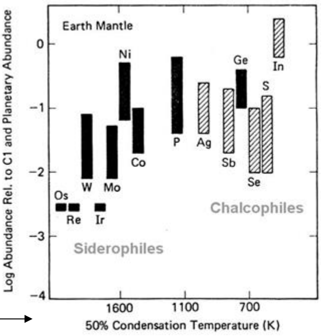

Figure 13.15 Abundances of siderophile elements (black bars) and chalcophile elements (hatched bars) in the mantles of the earth relative to C1 chondrite and bulk planet abundances. The depletions from 0 are due to fractionation of metal and sulfide during core formation. The horizontal axis is the temperature at which 50% of the element would be condensed in the nebula.

We return to this discussion of inferring core chemistry next lecture.

Besides light elements, we also need to consider what other transition metals are in the core, such as Ni.