

Chapter 2

Accretional History of Earth

2.1 The Accreting Material

Before describing the possible scenarios for planetary accretion and their relation to degassing of a primordial atmosphere it is essential to have some idea of just *what* was being accreted. There is general agreement that known meteorites are the best candidate materials based on their ages and compositions. The similarity between the compositions of planets, meteorites, and the refractory element components of the sun argues persuasively for a common origin. Presumably, Earth's volatile elements came from the same or a very similar source. Although extant meteorites may not very precisely match terrestrial compositions, either in terms of elemental abundances or isotopically, the complexities involved in accretional history may be a sufficient explanation. Furthermore, comparisons between meteorites and upper mantle compositions are confounded by the equilibration between metal and silicates during core formation as well as the effect of a possible late veneer. This is probably even more true when considering volatiles, which were undoubtedly affected by various earlier processes in the solar nebula. The idea that most of Earth's volatiles came from late comet impacts has been suggested (Frank et al. 1986; Hartogh et al. 2011), but considerations of elemental and isotopic chemistry apparently limit such contributions to less than 50% (Drake 2005; Altwegg et al. 2015). This could still be a significant amount, but there is little evidence *in favor* of a major cometary component.

Considering the complex processes that would accompany just the latest stages of accretion, it is questionable whether any simple comparison of meteorite types with primitive upper mantle compositions (including considerations of the late veneer) should be expected to produce a close match with particular meteorites or meteorite classes. Earth's upper mantle and Earth's volatile reservoirs must express the results of mixing of a wide variety of late-arriving objects with the pre-existing accreted material, and with complex processing involving impact, magmatic, hydrothermal and atmospheric chemistry. It seems unlikely that any simple model comparisons would be definitive. Aspects of some of these accretional processes relevant to Earth's early atmosphere will be discussed in more detail in the next

section. Even if volatile components of extant meteorites do not precisely match the chemistry of Earth's atmosphere, the volatiles in that atmosphere must have come from some collection of volatile-bearing accreted materials, whether early or late arriving. Exactly what combination of the various groups of meteorite might apply to any particular (terrestrial) planet is to some extent constrained by the known chemistry of each planet; obviously we require a mix that results in a suitable final average composition. This leaves aside, for now, the question of the order of accretion, or some possible degree of inhomogeneity in the accretion process.

In the case of Earth, for which voluminous compositional data has been available for decades, the so-called “chondritic” model for the bulk silicate composition has long been considered a reference case. When a suitable iron meteorite mix is added we can get a reasonable overall composition for the Earth. The “chondritic” refers, of course, to chondritic meteorites, by far the most abundant of all meteorite classes. Chondrules are spheroidal particles of silicate minerals (mostly olivine and orthopyroxene, which are iron-magnesium silicates) that comprise a large proportion of the meteorites collectively known as stony meteorites. It is thought that chondrites may have been formed by freezing of liquid silicate droplets in the hot solar nebula as it underwent cooling. They are accompanied by a variety of additional silicates, some of which were refractory enough to have remained solid in the nebula, silicates altered by processes of hydration or other reactions in the cooling nebula, and various amounts of iron-nickel metal. They are often put into separate classes depending upon the amount of metal: H for high-iron, L for low-iron and LL for very low-iron. In addition there is a separate class, the carbonaceous chondrites, which have unusually large amounts of volatile components, in particular carbon and carbon compounds. Some rare stony meteorites are chondrite-free and have mineral compositions very different from the chondrites. They may be, in part, the result of early magmatic processes on meteorite parent bodies. Table 2.1 provides an indication of the compositional ranges, with the emphasis on iron content. Metallic meteorites come in a range of types as well, but in general they are largely iron-nickel with a few minor additional components. A final class of meteorite is composed of a roughly 50–50 mix of iron-magnesium silicates and metal.

Table 2.1 Antarctic meteorite summary

Type	Mass in grams	Mass fraction	Metal fraction in class	Overall metal fraction
CC	35254	0.010	0.000	0.000
H	914898	0.255	0.143	0.037
L	1667145	0.466	0.027	0.013
LL	284338	0.079	0.000	0.000
Other stony	89114	0.025	0.000	0.000
Stony-iron	202348	0.057	0.500	0.028
Iron	387799	0.108	1.000	0.108
Total	3580896			0.186

H, L, and LL are the three main categories of chondrites

CC Carbonaceous chondrites of all sub-types

Table 2.2 Volatile components in Meteorites. (after Anders and Ebihara 1982; Huss 1987; Palme and O'Neill 2003; Bradley 2004; Schaefer and Fegley 2007). Orgueil is taken as a typical carbonaceous chondrite for these calculations

	Mass Fraction in each type			Mass Fraction in "Average mix"			Mass fraction in mix as compounds		
	H	C	N	H	C	N	H ₂ O	CO ₂	N ₂
CI (Orgueil)	0.020	0.032	0.0032	0.0031	0.0016	0.00006			
H	0.0032	0.0012	0.000034						
L	0.0034	0.0016	0.000034				.028	.0057	0.00006
LL	0.006	0.0024	0.00005						

Even though there are several types of chondritic meteorites they are sufficiently similar to one another and to the known and presumed composition of Earth's mantle to justify the chondritic model of mantle composition, as a first approximation. The basic conclusion is that the Earth was formed by accretion of about 2/3 mass fraction of some kind of average chondrite with about 1/3 iron meteorite (to account for Earth's metallic core), without necessarily specifying a particular accretion sequence or history of accretion rate.

Because we are mostly interested in the volatiles that came to form the atmosphere and oceans it seems reasonable to focus on the components of meteorites which provided those volatiles. The Antarctic Meteorite Program has given us what appears to be a representative sample of meteoritic material arriving on Earth at the present time. Table 2.1 is a summary of those samples, from a compilation made in June 2011 (http://curator.jsc.nasa.gov/antmet/us_clctn.cfm for the most recent compilation).

The mass fraction of iron (actually an iron-rich iron-nickel alloy) in the Earth is close to 0.33. Although this is a factor of two higher than that found in the Antarctic collection summary, the difference is small enough to suggest that the material arriving now is a reasonable approximation to bulk Earth, and suggests homogeneous accretion over time. This does not, of course, prove homogeneous accretion, but it is difficult to imagine how significant inhomogeneity of early accretion could be followed 4.5 BY later by an infalling meteorite mix so similar to bulk Earth.

These values must be considered reasonable (at best) approximations given the rarity of samples and the possibility that even averages of several samples may not be representative. In addition the most volatile-rich carbonaceous chondrites (e. g., Orgueil in Table 2.2) comprise only about half of the mass of Antarctic carbonaceous chondrite samples, the other carbonaceous chondrite classes being typically lower in volatiles. This suggests somewhat smaller values of available volatiles (for degassing) should be taken if using the mass of all carbonaceous chondrites in calculating the percentage of accreting volatiles attributed to them. With that caveat it is possible to calculate the volatiles available from a given mass of "average meteorite" mix. In this analysis no volatile contribution is attributed to the iron meteorite fraction, and thus represents a kind of minimum since iron meteorites actually contain some carbon and nitrogen. For this calculation the following volatile contents (mainly using the average values from Schaefer and Fegley 2007) are combined

Table 2.3 Volatiles on Earth

H ₂ O in hydrosphere = 1.4×10^{24} g	(Walker 1977)
H ₂ O in sediments = 1.5×10^{23} g	(Walker 1977)
Total C in crust = 1.0×10^{23} g	(Walker 1977, Wedepohl 1995, Hayes and Waldbauer 2006)
Carbonate rocks in crust (as C) = $8\text{--}8.5 \times 10^{22}$ g	(Walker 1977, Hayes and Waldbauer 2006)
CO ₂ in atmosphere = 4.5×10^{18} g	(Walker 1977)
CO ₂ production from volcanoes = 1×10^{14} g/yr	(Walker 1977; Hayes and Waldbauer 2006)
Oxygen in atmosphere = 1.2×10^{21} g	(Walker 1977)
Nitrogen in atmosphere = 4×10^{21} g	(Walker 1977)
Nitrogen in crust = 1.4×10^{21} g	(Walker 1977) also see Wedepohl 1995
Total iron resources (mostly BIF) = 8×10^{17} g	(Brobst and Pratt 1973)
C/N in cyanobacteria = 6 (mass), 7 (moles)	(Chorus and Mur 1999)
C/P in cyanobacteria = 42 (mass), 109 (moles)	(Chorus and Mur 1999)
P in Cont. crust = 1.7×10^{22} g ca. 780 ppm	(Wedepohl 1995)
Rate of erosion of continents = $3\text{--}5 \times 10^{16}$ g/year	(Wilkinson 2005; Walker 1977)

with the meteorite percentages from Table 2.1, adjusting to a total chondrite fraction of 81 %.

Each metric ton of meteorite mix provides volatiles equivalent to about 28 kg of water, 6 kg of carbon dioxide and 60 g of nitrogen. Most of the water and carbon dioxide actually come from the ordinary chondrite component, along with about half of the nitrogen. Table 2.2 includes data relevant to volatile budgets for the near-surface parts of the Earth, including the crust.

Using the numbers in Table 2.2 and comparing with Table 2.3 we can see that it takes a rather small amount of meteorite mix (compared to the mass of Earth's mantle) to provide all of the excess volatiles on Earth as shown in Table 2.4. Note that the relative amounts of volatiles in the meteorite mix and Earth are quite similar, which suggests an initial meteorite mix may well have provided Earth's volatiles. Even if there remain residual volatiles in the mantle, this result strongly indicates the necessity of major volatile loss from Earth, or a high degree of early (inhomogeneous) accretion of volatile-poor material. The suggestion that massive volatile loss accompanied the presumed moon-forming impact (Lock et al. 2014) implies relatively late timing for such loss. In any case, whether Earth experienced “dry” inhomogeneous accretion or large-scale volatile loss at a late accretional stage, most of Earth's extant volatiles could have been added late in accretional history, prob-

Table 2.4 Volatiles available from Meteorite mix equal to 1 % of Earth's mass

	Potential from degassing (see Table 2.2)	Earth's near-surface inventory (see Table 2.3)
H ₂ O	1.7×10^{24} g	1.55×10^{24} g
CO ₂	3.4×10^{23} g	3.7×10^{23} g
N ₂	3.6×10^{21} g	5.4×10^{21} g
H ₂ O/CO ₂	4.8	4.2
N ₂ /CO ₂	0.010	0.015

ably with other late veneer material (Ballhaus et al. 2013). It is certainly possible that the earlier high temperature conditions of Earth's interior may have left some volatiles in the interior. The fact that they survived that hot phase suggests that they would probably contribute little to subsequent degassing.

2.2 Accretional History

Little detail of accretional history was available until the advent of high speed computing and the formulation of accretion models depending on massive calculations. In fact, earlier ideas about the actual timing and rate of accretion left open the possibility that accretion could have been slow enough to avoid a hot early planet. Such a possibility would have allowed for the formation of a (relatively) cold Earth followed by continuous radiogenic heating and gradual release of gases over geologic time, as originally proposed by Rubey (1951, 1955). Modeling of accretional scenarios in the 1970s indicated fairly rapid formation of terrestrial planets. These results were supported by geochemical studies indicating timing of core formation on Earth at about 50 million years after solar system (meteorite) origin (Wood et al. 2006; Halliday 2008). Such results are also consistent with the suggested moon-forming impact at about the same time. Some recent models (Halliday and Wood 2009) suggest that accretion *ended* with the moon-forming impact (MFI) at about 4.5 BYBP), which seems rather unlikely. Although the Late Heavy Bombardment (LHB) (Koeberl 2003, 2006a, b; Marchi et al. 2012) added only a small amount of additional material, its occurrence suggests a significant amount of material remaining to accrete following the MFI, such accretion taking place between the MFI and LHB, a 600+ million year interval.

Although precise timing for these events is difficult, if not impossible, to determine, the energetics of the moon-forming impact would surely have resulted in core formation if it had not already taken place. The models also indicate continued accretion for tens of millions of years and at least as late as the LHB, as the unaccreted residue impacted. Accretional models in Fig. 2.1 show various results for different average accretion rates. Also shown are a few cases incorporating a moon-forming impact at 50 million years of a Mars-sized object. All except the highest accretion rate (which is probably so high that it would result in core formation earlier than geochemistry suggests) have accretional tails extending out to 100 million years or so, and involving post MFI accretion of a few percent of Earth's total mass (Fig. 2.2).

An interesting aspect of this late accretional era is the Late Heavy Bombardment seen in the cratering history of the Moon. Although this surge in accretion rate produced significant effects on the lunar surface, the quantity of new material added to Earth (or the Moon) was relatively small, in particular in terms of additional volatiles.

These models indicate that the thermal state of Earth's surface was largely established well before the end of accretion. The MFI was a highly energetic event,

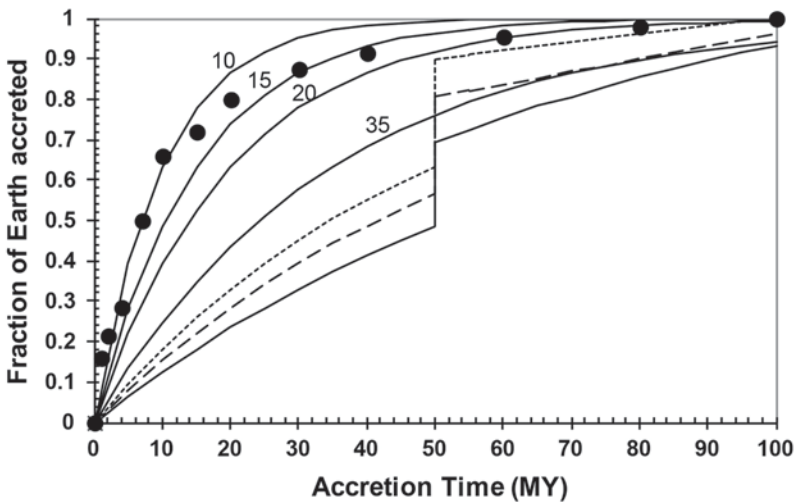
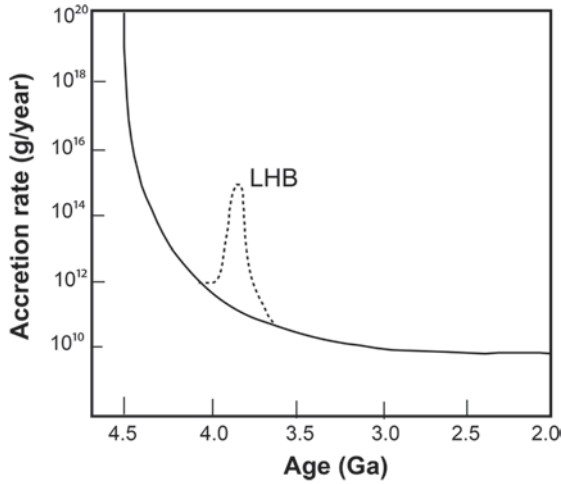


Fig. 2.1 *Accretion models for Earth.* Numbers next to curves are median accretion times in millions of years. Set of curves with “steps” at 50 million years are three cases including a Moon-forming impact at 50 MY (all after Halliday et al. 2000). Black dots are for averages of several models from Wetherill (1976). All but the most rapid of these models supports the idea that more than 1 % of Earth’s mass accreted after 50 MY, or after a Moon-forming impact at that time

Fig. 2.2 *Accretion rates for the Moon, including a “spike” for the Late Heavy Bombardment.* Accretion rates (mass/year) for Earth would have been at least 20 times higher. The very high, though rapidly falling, accretion rates in the Early Hadean could easily account for accretion of more than 1 % of Earth’s total mass following the Moon-forming impact at about 4.5 Ga, but prior to the LHB. After Koeberl 2006a



presumably resulting in a terrestrial magma ocean of unknown, but probably great, depth. The cooling and freezing of this magma ocean seemingly marks the point at which newly accreted volatiles might have accumulated on Earth’s surface. Sleep et al. (2001) calculated that, as a best guess, surface temperatures below 100C (and a solid surface, of course) would have been reached in about a million years follow-

ing the MFI, although they suggest that the continued presence of a dense carbon dioxide atmosphere may have extended this to perhaps 20 million years. Even this outer limit suggests that some liquid water was present on Earth's surface while a few percent of accretion still remained, at least for most accretion models (see Fig. 2.1). This is an important point when considering the conditions under which the latest accretion, and presumably much of the degassing of the permanent atmosphere, took place.

Although large energetic impacts may well have occurred in the 50 million years or so following the MFI, none would have approached its magnitude. These large late impacts may have been important in releasing volatiles during impact and may also have blown off some fraction of the atmosphere, but it is difficult to say what percentage of the volatile inventory was due to impacts of any particular size range, or how much might have been removed. It is important to examine the conditions that would likely have arisen from a range of sizes of impact events in order to assess the nature of the earliest atmosphere. Several aspects of surface conditions may also have changed during the production of the early atmosphere and oceans, as will become clear. An important point is that evidence from early zircons indicates that liquid water (some kind of ocean) was present as early as 4.4 billion years ago, within ca. 150 million years of the beginning of accretion, quite likely even sooner, and certainly as accretion continued (Mojzsis et al. 2001; Wilde et al. 2001; Peck et al. 2001; Valley et al. 2002; Nutman 2006).

During the last stages of accretion, when it appears most volatiles arrived as components of volatile-bearing meteorites, perhaps resembling a mixture similar to the Antarctic collection, Earth's surface would have been recognizable to us. It was probably not the volcanic nightmare so popular with illustrators, but was rather dull, with a large shallow ocean and localized volcanic activity, much of which may have been submarine. Admittedly there were a lot of rocks falling from space. To add 1% of Earth's mass over a few 100 million years gives an average of about one impact/day of a 1 m object on every square kilometer of Earth's surface, most of which would have been into a shallow global ocean. This is also equivalent to one impact each year by a 2+km diameter object. The presence of the ocean was of great importance to the processes of degassing. The next chapter discusses the nature of this early ocean, the crust beneath the ocean and the tectonic-thermal evolution of the oceans and early continents.

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