FORMATION OF THE TERRESTRIAL PLANETS

WILLIAM M. KAULA
University of California, Los Angeles
Department of Earth and Space Sciences
Box 951567
Los Angeles, CA 90095, U.S.A.

Abstract

The early phases of formation in the inner solar system were dominated by collisions and short-range dynamical interactions among planetesimals. But the later phases, which account for most of the differences among planets, are unsure because the dynamics are more subtle. Jupiter's influence became more important, leading to drastic clearing out of the asteroid belt and the stunting of Mars's growth. Further in, the effect of Jupiter-- both directly and indirectly, through ejection of mass in the outer solar system-- was probably to speed up the process without greatly affecting the outcome. The great variety in bulk properties of the terrestrial bodies indicate a terminal phase of great collisions, so that the outcome is the result of small-N statistics. Mercury, 65 percent iron, appears to be a residual core from a high-velocity collision. All planets appear to require a late phase of high energy impacts to erode their atmospheres: including the Earth, to remove CO_2 so that its ocean could form by condensation of water.

Consistent with this model is that the largest collision, about 0.2 Earth masses, was into the proto-Earth, although the only property that appears to require it is the great lack of iron in the Moon. The other large differences between the Earth and Venus, angular momentum (spin plus satellite) and inert gas abundances, must arise from origin circumstances, but neither require nor forbid the giant impact. Venus's higher ratio of light to heavy inert gases argues for it receiving a large icy impactor, about 10^{-6} Earth masses from far out, requiring some improbable dynamics to get a low enough approach velocity. Core formation in both planets probably started rather early during accretion.

Some geochemical evidences argue for the Moon coming from the Earth's mantle, but are inconclusive. Large scale melting of the mantle by the giant impact would plausibly have led to stratification. But the "lock-up" at the end of turbulent mantle convection is a trade-off between rates: crystallization of constituents of small density difference versus overall freezing. Also, factors such as differences in melting temperatures and densities, melt compressibilities, and phase transitions may have had homogenizing effects in the subsequent mantle convection.

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1. Introduction

The terrestrial planets are distinguished from the major planets in being (1) closer to the Sun, (2) smaller, (3) made mainly of rock, and (4) separated by a marked gap, the asteroid belt. Property (1) explains properties (2) and (3): so close to the Sun that high velocities inhibited merger and fostered fragmentation, and it was hot enough that ices could not condense. But property (4) depends mainly on the innermost major planet being the largest, Jupiter, and thus having a disrupting effect by scattering smaller bodies through the next zone interior to it. The depopulation of this zone would have occurred even if a sizable planet had formed there; such a body could have been fragmented by Jupiter scattering (Wetherill, 1992). The smallness of Mars also arises from this Jupiter effect. However, between Mars and the Earth there is a qualitative difference: the "Weidenschilling Limit" (Weidenschilling, 1975). The minimum perturbation for Jupiter to eject a body from the solar system is equal to perturbation required to send a body from Jupiter to about 1.3 AU. Hence bodies going closer to the Sun than this limit are much more likely to be ejected from the solar system upon returning to the vicinity of Jupiter than are bodies not perturbed so far inward. The effect on Mercury's zone of bodies perturbed inward by the Earth and Venus is similar, in principle: there was intense fragmentation in this zone as well.

TABLE 1: Properties of the Planets

Planet	Solar Dist. AU	Mean Motion n_J	Mass/ Zone M_E/AU^2	Mass M_E	Mean Density* Mg/m^3	Metal Iron %	Atmo-sphere bars	Rota-tion Ω_E
Mercury	0.4	49.	.10	.06	5.6	65.	0.	.008
Venus	0.7	19.	.61	.85	3.95	30.	90.	.004
Earth	1.0	12.	.40	1.00	4.02	30.	1.	1.0
Moon	1.0	12.		.01	3.34	0.	0.	.037
Mars	1.5	6.3	.011	.10	3.65	15.	.01	1.0
Asteroids	2.8	2.5	.000	.00	3.4	0.	0.	—

*Reduced to a pressure of 1.0 GPa. n_J: mean motion of Jupiter: M_E: mass of Earth

The foregoing plausibilities about the main properties of the terrestrial planets in Table 1 are consistent with a scenario of hierarchical growth from an assumed initial population of small planetesimals. The earlier stages of this scenario are now well confirmed by models based on short-range dynamical interactions. However, other effects must be invoked to arrive at bodies similar to the actual planets (Lissauer, 1993). This review, as appropriate to a volume on comparative planetology, will concentrate on problems associated with the later phases of the formation of the terrestrial planets, since they account for virtually all their differences in properties. This is not to say that there are no difficulties with the early phases. In particular, simple gravitational instability of dust (Safronov, 1972; Goldreich and Ward, 1973) is probably an insufficient mechanism for initial formation of planetesimals in a nebula turbulent from infalls and convection (Tscharnuter and Boss, 1993); other mechanisms of coagulation must be invoked (Cuzzi et al.., 1993; Weidenschilling and Ruzmaikina, 1994). Also, both gas drag and gravitational interaction with the nebula plausibly affected planetesimal orbits, and thence growth of embryos (Ward, 1993b). I divide the topics into (1) terminal accretion; (2) volatile gain and loss; (3) core formation; and (4) the giant impact.

2. Terminal Accretion

Models starting with a large swarm of planetesimals a few kilometers in size, such as that of Wetherill and Stewart (1993), evolve within about 10^5 years to several embryos of Moon to Mercury size in nearly circular, co-planar orbits, plus a swarm of smaller planetesimals that cannot grow because their relative velocities are enhanced to more than their escape velocities from perturbations by the embryos. This result depends on the correct equipartition of energy at encounters (Stewart and Wetherill, 1988), and has been obtained by both analytic and numerical techniques using the algorithm of Opik (1976). Fragmentation and drag are also taken into account.

To carry forward models allowing for close interactions only, larger eccentricities must be assumed. With correspondingly increased inclinations, these models take well over 10^8 years to accumulate bodies similar to the terrestrial planets (Kaula, 1990; Wetherill, 1991). Clearly, other effects need to be incorporated: at a minimum, perturbations by Jupiter. Wetherill (1992) implemented an Opik algorithm in the asteroid belt, adding the accelerations arising from sixteen resonances with Jupiter, as well as ejection by Jupiter, and shows that velocities are developed sufficient to fragment embryos of 0.02 Earth mass. However, within the Weidenschilling limit further effects must operate, whose exploration will entail considerable computational effort. Integrations of the present system obtain oscillations of the planetary eccentricities of about 0.06, almost all associated with periodicities of about 10^5 years. Hence *if* the number of terrestrial planets were doubled, there would be close encounters, leading to further enhancement of eccentricities, and thence collisions. If interactions of these embryos with a significant declining mass of nebula and planetesimals in the asteroid belt and further out are taken into account, the eventual problem may be how to limit and time this loss, in order not to enhance terrestrial planet eccentricities excessively (Ward, 1988, 1993a).

In any case, taking enhanced eccentricities as given, the consequences for the planetary embryos are severe: e.g.,with eccentricities of 0.10, approach velocities of

3 km/sec are obtained in the Earth's vicinity. Kaula (1990) started with the planets already 81 percent accumulated, plus a population of planetesimals ranging in mass up to 0.02 Earth masses. Eccentricities of both planet and planetesimal orbits were taken into account in calculating probabilities of close encounters and approach velocities in a modified Opik algorithm. The results were that proto-Mercury always received sufficient energy to cause fragmentation; proto-Mars received less, but still appreciably more than proto-Earth or proto-Venus. The range of summed colllisional energy inputs for the two larger bodies was great: from 10^6 to 2×10^7 J/kg. Angular momentum absorption was positively correlated with energy absorption, leading to spins ranging from 1 cycle/4 hrs to 1 cycle/40 hours, and obliquities up to 160°: retrograde in 30% of cases. Twelve percent of cases coupled a fast prograde rotation of one body with a slow retrograde rotation of the other.

The outcome of planetesimal interaction models such as Wetherill and Stewart (1993) appears inexorable: an inner solar system with a few dozen embryos of 0.01 to 0.06 Earth mass. Mechanisms to enhance the eccentricities of these embryos are evident, but are yet to be worked out in detail. The inevitable consequence for the terminal phases of accretion is large impacts, leading to a wide range of possible outcomes for fragmentation, loss of volatiles, spin rates, obliquities, and satellites of the proto-planets. Another important effect is the high degree of mixing among zones of the inner solar system, leading to a blurring of any compositional gradients from condensation in the nebula. The main substantive question (as distinguished from technical difficulty) is the relative phasing of terrestrial planet growth with loss of nebula, growth of Jupiter, and loss of planetesimals. At present, it appears that the losses of nebula and planetesimals must occur rather early, and perhaps the growth of Jupiter rather quickly, in order that relative velocities in the inner solar system not be enhanced too much.

3. Volatile Gain and Loss

Experimental evidences relevant to the acquisition of volatiles by the terrestrial planets include:

(1) The H_2O content of ordinary chondrites is generally 0.2-0.3% by weight; it is almost never less than 0.1%, and can be over 10% in carbonaceous chondrites, which are probably the most abundant (Mason, 1962).

(2) The C/H atomic ratio of chondrites varies from 0.1 to 1.0; Anders and Ebihara (1982) take 1/7 as typical

(3) CI chondritic meteorites have not suffered as much loss of inert gases relative to cosmic abundances as the terrestrial planets (Fig. 1) (Owen et al., 1992).

(4) Argon is trapped in amorphous ice at temperatures below 30 K, and Neon at temperatures below 20 K (Bar-Nun et al., 1988).

(5) The ratio of H_2O solubility in magma to that of CO_2 decreases with temperature, but is always at least an order-of-magnitude greater (Mysen, 1977).

(6) Incipient devolatilization of a CM chondrite occurred at a shock pressure of about 11 GPa, and complete devolatilization at about 30 GPa (Tyburczy et al., 1986).

Taking 3% as the H_2O of impactors (1), by the time a terrestrial protoplanet is large enough to start devolatilizing impactors (6), it will contain 1.5 Earth oceans of water. It will fully devolatilize impactors when it has six oceans (Ahrens, 1990). It has long been recognized that a terrestrial planet of Mars's mass or larger will heat up to silicate melting, if formed mainly from planetesimals of 100 km or more (Safronov, 1972). A further heating factor is that H_2O outgassed from impactors will create a steam atmosphere, which will increase until pressure sufficient to dissolve it in magma is achieved. Such an atmosphere has an infra-red opacity sufficient to maintain silicate melting temperatures at the surface, so long as appreciable infalls continue

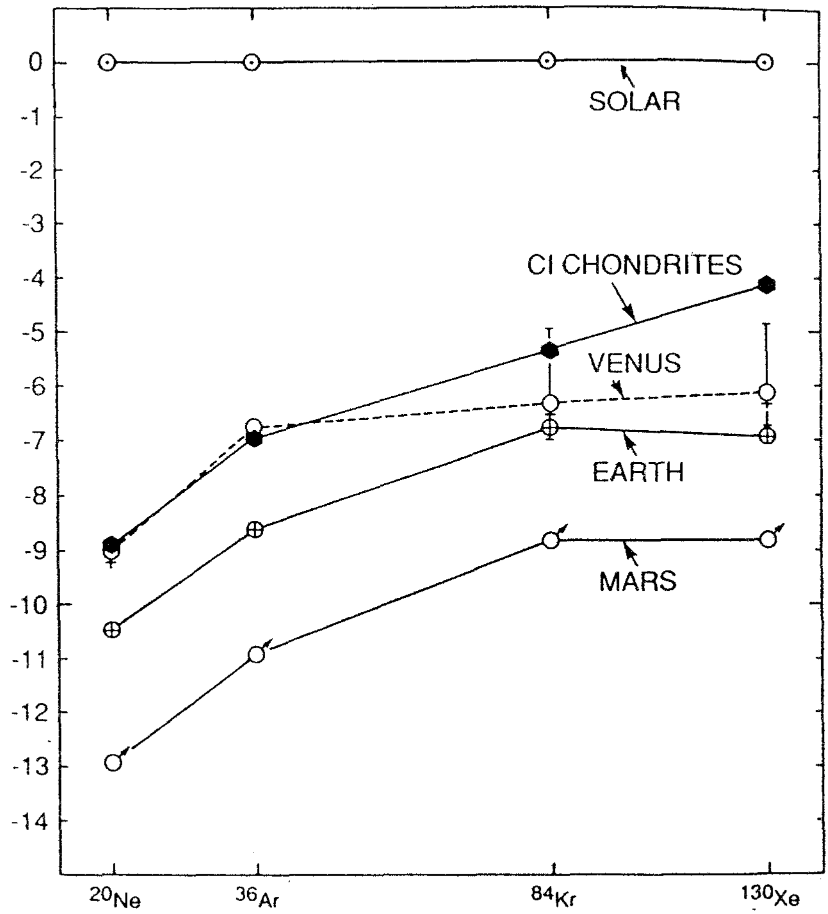

Figure 1. Inert gas abundances relative to solar abundances (normalized to silicon). The ordinate is log(abundance); the abscissa is in atomic mass units (AMU).

(Abe and Matsui, 1985). The pressure from this equilibrium atmosphere is estimated to be 30 MPa (300 bars) (Abe, 1993). If the planet is far enough from the sun and its atmosphere is dominantly H_2O, when energy inputs from infalls and the magma ocean decline the H_2O will condense and rain out. This appears to be the case for the Earth, but possibly not Venus. The problem is the carbon content of the atmosphere. If the 1/7 C/H atomic ratio applies (2), then a model which has about 3×10^{21} kg of H_2O, most of it dissolved in the magma ocean (e.g., Zahnle et al.,1988) will also have more than 10^{21} kg of CO or CO_2 , which stays in the atmosphere, since it is not soluble in magma (5). This is enough CO_2 for 15 Mpa (150 bars) pressure. While CO_2 is not as opaque as H_2O, this is more than enough to prevent condensation of water: see, e.g. figure 9 of Kasting (1988), which shows that 100 bars of CO_2 will produce a surface temperature of 480 K with solar radiation 0.8 times the present radiation at the Earth.

Hence mechanisms to remove atmospheres are needed. A compositionally insensitive mechanism is impact erosion (Ahrens, 1993). Effective erosion requires impactors appreciably larger than the atmospheric scale height. The planetesimal growth models produce bodies amply large, even though fragmentation is expected to grind them down in the late phases. The apparent loss of silicates by proto-Mercury argues for a strong impact regime. In the late phases, comet influx from the outer solar system will be at high enough velocities to be erosional for rather small impactors (Weissman, 1989). The parallelism of the Earth and Mars inert gas abundances (Figure 1) argues for a compositionally insensitive mechanism. Other characteristics of Figure 1 argue for compositionally-sensitive effects. The slope of the CI chondrite curve reflects different adsorption characteristics operating early in the formation process. But the differences therefrom of the planets requires processes peculiar to atmospheres, such as mass fractionation in hydrodynamic escape arising from the early high solar EUV flux (Hunten et al., 1987). Fractionation at an earlier stage, such as hyrdodynamic escape from a meteorite parent body, is indicated by marked trends in the isotoes of krypton (Pepin, 1991). But escape seems insufficient to explain the differences between Venus and the Earth in Ne/Kr and Ar/Kr (Pepin, 1991; Zahnle, 1993), so comets are invoked. However, many small comets would mitigate against differences between Venus and the Earth. Owen et al. (1992) suggest a single impact of an extra large body from the Uranus: Neptune zone to explain the argon excess on Venus (4), but nobody's model of the outer solar system has temperatures low enough to explain the neon. A body carrying Venus's 10^{16} kg of excess argon and having a 2.0 Mg/m^3 density and solar Ar/Si would have a mass less than 10^{-6} Earth masses-- a diameter less than 100 km, of which there must have been millions. The problem is the dynamics of delivery. It cannot be direct: the *minimum* approach velocity to Venus would be 12.9 km/sec, leading to an impact velocity of 16.4 km/sec-- very erosive, instead of accretive. Instead, perturbations by Uranus and/or Neptune must drop its perihelion to the vicinity of Jupiter, which in turn must remove enough energy to flip its perihelion into an apohelion. Then, most difficult, the Earth must quickly remove enough energy to bring its apohelion well within Jupiter's orbit. Assuming 10^8 bodies, 3×10^6 might be "dropped" to Jupiter; then 10^4

of these "flipped", and one of these "deboosted" to eventually ease into Venus. This last step seems forbiddingly improbable (Kaula, 1994).

4. Core Formation

The apparent unavoidability of magma oceans on the Earth and Venus, due to large impacts and steam atmospheres, facilitates the collection of large enough pockets of metallic iron to penetrate any solid silicate mantle below it. Thus core formation probably occurred continuously from the time planets were a few times 0.01 Earth mass, (Stevenson, 1990). But the chemical disequilibrium between the metallic core and the upper mantle suggests there was an evolution toward more oxidizing conditions as the impact rate decreased and magma oceans solidified (Newsom and Sims, 1991). These conditions apply as well to Mercury, which had the greatest energy inputs from impacts, and evidences a history of cooling since origin. Mars is more problematical; its high mantle density and high moment-of-inertia indicate more iron was retained in the mantle, while its surface topography suggests a warming history, as would occur with delayed core formation.

5. The Giant Impact

The *only* property requiring a great impact in the late phases of the formation of the Earth is the extraordinary depletion of the Moon in iron: not only must any metallic core be negligible, but also the silicate Fe/Mg ratio of the bulk Moon must be as low as that in the upper mantle, appreciably less than in chondritic meteorites. The alternative of a screening mechanism in a swarm around proto-Earth founders on the time scale of accretion of such a swarm into one body being much shorter than the time scale of infall of additional bodies from heliocentric orbits, so that most of the time the proto-lunar material is clumped in a body big enough to be insensitive to differences among infalling bodies. The high angular momentum and big satellite of the Earth can be explained by a circum-terrestrial swarm, provided that it starts early in planetary growth and the Earth's tidal dissipation factor (1/Q) is big enough (Harris and Kaula, 1975).

It should be emphasized that the largest impact into one protoplanet, the Earth, being considerably larger than the largest impact into another, Venus, is to be expected from the strongly stochastic nature of the terminal phase of accretion discussed above. Indeed, it is the Occamish hypothesis, and alternative hypotheses have the burden of explaining why it did not occur.

Characteristics of the giant impact indicated by modelling are: (1) widespread melting and volatilization, but very non-uniform; (2) the rapid merger of the core of any impactor with the core of the Earth; (3) the formation of a surrounding ring of gases and volatilized silicates; (4) the ondensation of some of these silicates in orbit,to form the proto-Moon mainly from impactor material; (5) almost complete loss of atmosphere due to heating sufficient to create a planetary wind, if not by hydrodynamic effects; and (6) the formation of a deep magma ocean. (Cameron and Benz, 1991; Melosh et al., 1993).

Quite consistent with a much greater rocky impact in Earth than in Venus formation, but not demanding it, are: the oxygen isotope identity of lunar and Earth ultramafic rocks; the low volatile content of The Moon (e.g., a K/U ratio of about 2000); the low oxidation level of the Moon (e.g., no Fe_2O_3); the evidences (KREEP, etc.) of a magma ocean on The Moon; the high angular momentum density of the Earth-Moon system; and the several volatile properties of the planets discussed above: lower inert gas retention in the Earth than in Venus; and ocean and atmosphere by outgassing from meteoritic composition, plus a minor veneer from the outer solar system.

The main objections to the giant impact, most articulately from Ringwood (1992 and earlier papers) were: consistency (within a factor of three) of lunar volcanic glasses (LVG's) with the Earth's mantle in abundances of siderophiles (including volatiles: e.g., germanium), implying thorough mixing of impactor and the Earth; (2) the improbability of such a big impact occurring so near the end of accretion that the Moon did not acquire metallic iron and siderophiles, and (3) the gross homogeneity of the Earth's mantle, which appears to forbid a deep magma ocean.

In regard to objection (1), the factor-of-three in the 3.5 Ga LVG's contrasts to the factors-of-hundreds for older basalts, and suggests that deep in the Moon are chunks that were only partly devolatilized. However, that the LVG's pyroclastic delivery indicates that they are peculiar in volatile content, and hence they may be as little representative of the bulk Moon as kimberlites are of the bulk Earth. Meanwhile, further compositional evidences of provenances of different sources than the Earth's mantle for lunar material, but a lack of volatility-associated fractionation, have been found by Humayan and Clayton (1994) and Norman et al. (1994). Partitioning in >2000 K melts, let alone >4000 K vapors, is not measurable and difficult to extrapolate. There could be more than one way to get scruffy within-a-factor-of-three similarities in siderophiles, so the lunar material need not have been at pressures comparable to deep in the Earth's mantle.

Regarding objection (2), the implausibility of the lateness is in the context of a lot of smaller bodies still impacting, but with high relative velocities, hence more likely to erode, than accrete, the Moon, which has an escape velocity of only 2.36 km/sec.

Objection (3), that a deep magma ocean would stratify, is more serious. Initially, magma ocean would be homogenous because it was highly turbulent (Tonks and Melosh, 1993), carrying crystals in suspension. The difficult question is how this system "locked up" without stratifying; how to get nonfractional crystallization, dependent on suspension in convective layers, pressure, rheology of partial melts, crystal size, and surface conditions (Solomatov and Stevenson, 1993). In the subsequent evolution of the mantle, factors such as rate-dependent phase transitions; lower melting temperatures of iron-rich silicates; and higher compressibility of melts making them denser than their solid matrix at pressures >8 GPa may have been homogenizing.

Ringwood's (1992) alternative hypothesis was the removal of proto-lunar material from the Earth by several smaller high-velocity impactors. The fundamental objection to this hypothesis is they just do not generate enough hydrodynamic effects to get the "second burn" necessary to place proto-lunar material in orbit, rather than escape or return to the Earth. Even if it were found that enough small high-velocity bodies could splash a lunar abundance of material off the Earth, there would remain the problem of explaining the differences between the Earth and Venus in volatiles and spin: many small events mitigate against differences, while one or two catastrophic large events favor them.

6. Conclusions

There appears to be a growing consensus on the scenario of the formation of the terrestrial planets, including the differences among them. However, there are many places where the processes connecting models and observations are quite unsure, and parts of the scenario are connected by loose plausibilities.

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