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Atmospheres as a Clue to the Early Solar System

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Atmospheres as a clue to the early Solar System

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Conditions that could have applied in the environments of the major planets when they were forming make it possible that the present icy mantles of the larger satellites were then oceans and vapour atmospheres encasing silicate–ferrous cores. The major constituents are explored by comparison with the present atmospheres of the terrestrial planets. It is further suggested that the primary condensations during the formation of the Solar System were the Sun and the major planets, and that the terrestrial planets and satellites were a secondary formation. Some observational data are offered in support of the arguments and future tests are suggested.

1. Introduction

The members of the Solar System have reached an overall equilibrium, which has in large measure obliterated the features peculiar to its origin. The appearance is of a single dynamical system with many common basic features among its members, even though each is also unique in some special way. There is ample evidence from the terrestrial planets and satellites for an early period of collisional encounters over a wide range of energies, including several catastrophies. There is, however, little coherent evidence of the conditions leading to this phase. The only firm information available is that based on current observations, which owe much to recent remote measurements from space vehicles.

The *Voyager* missions have provided important new data for the physical properties of the outer regions of the Solar System, and especially of the icy satellites. As a result, the inner and outer regions of the Solar System can now be compared in a quantitative way not possible before, and the whole viewed as a more complete entity. The relations between the various planets and satellites are reconsidered on this basis in an attempt to gain a deeper understanding of conditions in the early Solar System.

2. Members of the system

The larger members of the Solar System (see, for example, Beatty *et al.* 1981; Cole, 1984*a*) are eight planets in stable orbits about the Sun together with a small outer planet (Pluto) and some 40 satellites orbiting the planets. All but three of the satellites (this includes the two orbiting Mars, which are really large boulders) are associated with the four major planets.

The characteristic feature of these assorted bodies is the form of their internal equilibrium (Cole 1984*a*, 1987), which is the balance between the force of self-gravity and the material resistance to excessive compression. For the Sun the resistance arises from non-degenerate gas forces of the ionized interior (see, for example, Shu 1982). For planetary bodies, where the material is not ionized by pressure and thermal forces are irrelevant, the resistance arises from the macroscopic material incompressibility.

This equilibrium puts general upper and lower mass limits defining a planetary body of given material composition. At the upper end the internal pressure forces must nowhere exceed the strength of the constituent atoms: at the lower end the mass must be sufficient to provide a self-gravitational force at least comparable to the interatomic forces and so ensure an overall figure of revolution for the body. For a planet composed of pure hydrogen the upper mass limit is about 2.1×10^{27} kg with the associated maximum radius of about 1.1×10^{5} km, values only slightly larger than the observed mass and radius of Jupiter. Planets composed of heavier materials will have a higher mass limit but smaller radius. The lower mass limit for a silicate–ferrous body is a few times 10^{19} kg, or about the mass of Mimas.

The gap between planet and star is filled theoretically by intermediate bodies called brown dwarfs (for a hydrogen composition they span the mass range of *ca.* 2×10^{27}–*ca.* 2×10^{29} kg). They remain rather hypothetical bodies and none are yet known for detailed study. Because the mass range from planetary body to star is continuous from the theoretical point of view there is probably a need to understand the main properties of brown dwarfs before an overall understanding of a planetary system orbiting a star can be properly achieved.

3. Material compositions

The abundance of the chemical elements (see Beatty *et al.* 1981) in the Sun (the solar abundance, which includes all the elements of the periodic table) is special in that it is similar to that for main sequence stars and to the cosmic-ray spectrum. It is accepted as the standard against which the compositions of the different members of the Solar System are to be judged.

The two most abundant chemically active elements are hydrogen (overwhelmingly so) and oxygen, helium being essentially chemically inert. The most abundant compounds will then be hydrogen-based materials such as water, ammonia, methane, and the more complicated molecules involving especially H,O,C and N. (These elements, with other traces, are incidentally the basis of living tissue.) Oxygen compounds, of which the silicates provide a rich example, would be expected to be relatively rare.

The planetary and satellite compositions broadly reflect this overall composition. The major planets are fluid bodies composed predominantly of hydrogen and hydrogen-based molecules, but with a proportion of silicates and ferrous materials. They have masses in excess of 8×10^{25} kg. For the Sun, Jupiter and Saturn the fractional abundances (see Beatty *et al.* 1981) by number of molecules in the outer atmospheres are similar for hydrogen, CH_4 and NH_3, although helium is deficient in the outer regions of Saturn and water in both Jupiter and Saturn. This gives no clue about conditions inside.

The compositions of Jupiter and Saturn are not far removed from the solar abundance, although models (see, for example, Cole 1984*a*) of these planets suggest a rather lower proportion of the lighter elements than would be expected if the solar abundance were followed closely. Uranus and Neptune have a still greater proportion of heavier materials, probably including water. Each planet would, however, become a star not unlike the Sun by the addition of more material of its present composition.

The terrestrial planets and the satellites are quite different, being almost totally devoid of free hydrogen. To achieve approximately the solar abundance for the Earth, some 3×10^{26} kg of hydrogen and/or helium would need to be added to it. Such a quantity of hydrogen and helium would hardly be lost through the Jeans mechanism (Jeans 1908; Lewis & Prinn 1984) over the present lifetime of the Earth although a substantial part of it could perhaps have been swept away (see Shu 1982) by emissions from the Sun in an early period of development. Whether all the hydrogen and helium could have been lost in this way is problematic. It would seem more likely that the compositions of the terrestrial planets were always fundamentally different from those of the major planets. The same is true of the satellites.

There is, then, a present dichotomy of objects in the Solar System both in relation to mass and composition: those with masses greater than 8×10^{25} kg which are especially rich in free hydrogen (atomic or molecular), and separately those with masses below about 7×10^{24} kg with no significant free hydrogen. This distinction is not one simply of hydrogen loss but (disregarding special hypotheses, see Goody & Walker (1985); Chamberlain (1978)) more reasonably reflects a difference of material composition at the time of formation. Each of the hydrogen-rich major bodies is surrounded by bodies that are now deficient in free hydrogen: the Sun is surrounded by the terrestrial planets, and the major planets by the icy satellites.

If the Solar System were approached unknown from outside it would be most logical to view it as five hydrogen-dominated primary bodies each surrounded by silicate–ferrous secondary bodies. The total mass of the secondary bodies in each case increases with the mass of the primary body (see § 7.1). The primary bodies are in orbits about the centre of mass of the system which happens to be about one solar radius from the centre of the biggest of them.

4. The satellites

Of the non-hydrogen bodies the satellites (being the smallest) are the most likely still to show signs of early conditions and the satellites of the major planets are the most numerous of these bodies. The recent *Voyager* missions have provided considerable data about the physical characteristics of the icy satellites and these are listed in table 1. It is seen that, with the exceptions of Io and Europa, the densities are generally below 2000 kg m^{-3}. Spectroscopic evidence, consistent with the low temperatures of the satellite environments (*ca.* 70–80 K), shows the surfaces to be ices, and particularly water. The cosmic abundance indicates these to be the most abundant hydrogen compounds. The remaining material is accepted as being silicate with a ferrous component.

4.1. *Proportions of silicates in the icy satellites*

The ice is supposed to be water and the observed satellite densities are then the result of a simple mixture of silicate–ferrous material and water ice. Unfortunately, other ice mixtures could have a density similar to water ice and the model is not unique. Io has a density of 3550 kg m^{-3} and has no water ice on its surface. This density will be taken as typical (in the absence of other evidence) of the silicate–ferrous materials in the outer region. The masses of silicate–ferrous components calculated on this basis are included in table 1.

It is seen that the larger satellites have proportions of silicates of about 70% but that the smaller ones have larger proportions of ice. Tethys is essentially an ice ball. It is noticeable that the satellites of Uranus have a consistently higher density than those of Saturn. This could imply either a higher proportion of silicate–ferrous material (not really-supported by figure 2) or a higher proportion of the ferrous component giving a heavier silicate–ferrous component.

Table 1. Mass fraction of silicate material, radius (assumed silicates collected together), mass of mantle, and ratio of mass of silicates to mass of satellite for the satellites of the major planets

(The silicate density is assumed to be 3550 kg m^{-3} and the density of ice as 940 kg m^{-3}. Data for Charon are also included although these are less certain than the others: data for Miranda are excluded.)

body	mean density kg m^{-3}	radius 10^6 m	M_{sil} 10^{20}kg	R_{sil} 10^6 m	$M_{\text{sil}}/M_{\text{P}}$	e
Io	3550	1.816	892	1.82	—	2.7
Europa	3047	1.563	457	1.45	0.94	1.5
Ganymede	1938	2.638	1038	1.91	0.70	3.96
Callisto	1815	2.41	693	1.67	0.65	3.0
Titan	1934	2.56	944	1.85	0.70	3.9
Triton	1994	1.90	410	1.40	0.71	2.3
Pluto	1983	1.21	105	0.89	0.71	1.5
Mimas	1400	0.195	0.19	0.11	0.44	0.02
Encaledus	1200	0.25	0.22	0.11	0.29	0.03
Tethys	1031	0.525	0.67	0.17	0.11	0.05
Dione	1430	0.56	4.82	0.32	0.46	0.33
Rhea	1300	0.765	8.96	0.39	0.37	0.34
Iapetus	1165	0.72	4.59	0.31	0.25	0.33
Ariel	1650	0.58	7.81	0.37	0.58	0.43
Umbriel	1440	0.595	5.90	0.34	0.47	0.46
Titania	1590	0.805	19.09	0.51	0.55	0.48
Oberon	1500	0.775	14.64	0.46	0.50	0.49
[Charon	1890	0.59	11.04	0.42	0.68	—]

There is currently no way of distinguishing between these three possibilities. It is, however, true that the mean density of Io is some 6% higher than that of the Moon and this could reflect a higher ferrous component, in this case. Unfortunately Io is not associated with Uranus.

4.2. *Could the silicates and ices be differentiated?*

It is not known from the appearance of the surfaces whether the silicates and ices are separated (differentiated) or form a homogeneous mixture. From the appearance (V. R. Baker personal communication) of the craters it would seem that either the silicate material is in the form of fine grains mixed with the ice or the materials are largely differentiated. As one indication, photographs of the surface of Miranda show the silicates and ices there to occupy separate regions: this satellite could well have been disrupted early on and subsequently reformed (see also Marcialis & Greenberg 1987).

The self-gravitational energy E_{h} for a homogeneous body of ice and silicates is greater than the energy E_{d} for a body where the silicates form a core with an ice mantle (Cole 1984*b*; Cole & Herniman 1985). The energy difference $(E_{\text{d}}-E_{\text{h}})$ can be compared with the binding energy E_{b} between the atoms of the material: we express the ratio $(E_{\text{d}}-E_{\text{h}})/E_{\text{b}} = e$. Plastic compression conditions inside would allow the silicates and ice to separate, the interiors becoming differentiated even if they were not formed in that way. The flow conditions would be complicated but slow Stokes flow can be assumed to give a reliable estimate of the time needed for separation. Over a period of some 3 Ga, for $e > 1$ a fairly complete separation of silicates, ferrous materials and ices is possible under conditions where plastic flow can occur, whereas for $e < 1$ separation is not expected if it were not there initially; the intermediate case $e \sim 1$ could refer to partial separation. The values of e for the satellites are included in table 1.

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It is seen that, on this basis, differentiation of materials could have occurred for the larger satellites, with masses in excess of about 10^{22} kg, even if the materials were homogeneous when the satellites were formed, the silicate–ferrous materials now forming cores. This would not have occurred for the smaller bodies with masses below about 10^{20} kg: partial differentiation is possible for the bodies of intermediate masses, the larger silicate–ferrous particle masses having separated out from the remainder.

The present equatorial location of the large crater Herschel of Mimas has been suggested (Cole 1986) as possible evidence for the interior differentiation of this small body (where separation would not otherwise have been expected). The situation is confused, however, because alternatively the present figure of Mimas could be responsible for this effect; but that could well have followed the present orientation rather than having caused it. The structure of Miranda is known from *Voyager* photographs to be far from a homogeneous mix of silicates and ice but the geological structure of this body is highly complex. There remains ambiguity in the interpretation of simple observations.

Apart from any heat of formation, the interior of the satellites would have been heated by radioactive material decay although nothing is yet known of the radioactive content of the materials. Information about the interior state of differentiation would give indications of the possible radioactive content. There is an urgent need to determine, presumably by using *in situ* seismic measurements, whether the larger *and* the smaller satellites are differentiated bodies or not.

4.3. *The quantity of silicates*

A convenient way of comparing the silicate–ferrous content of the different satellites (assumed of the same density for each body) is to suppose it collected into spheres of appropriate radii. This would represent the actual interior core radius if the materials are differentiated. These 'radii' are compared in figures 1 and 2 by using the silicate–ferrous masses listed in

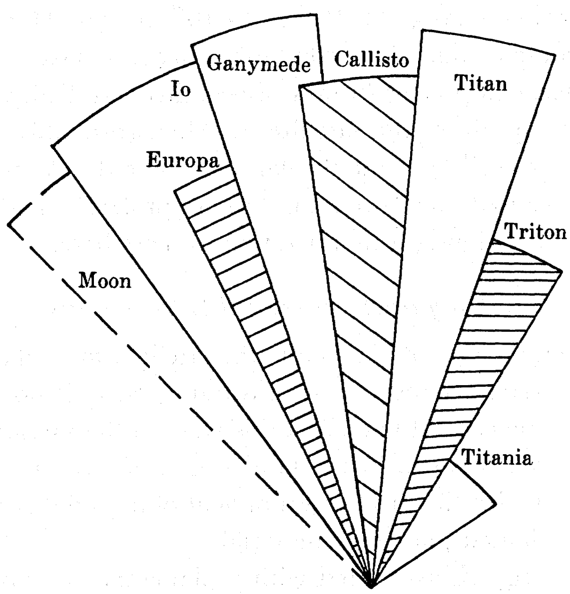

FIGURE 1. The relative sizes of the silicate–ferrous contents of the larger satellites, under the assumption that the material is aggregated together.

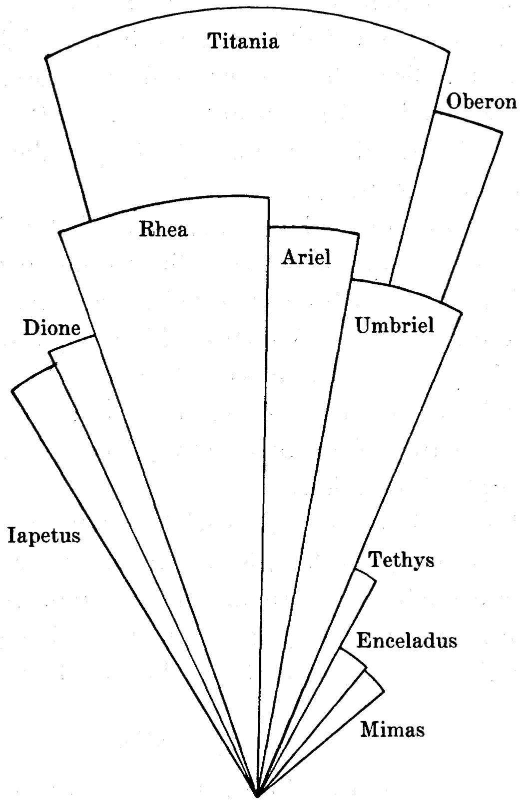

FIGURE 2. The relative sizes of the silicate–ferrous contents of the smaller satellites, under the assumption that the material is aggregated together.

table 1, assuming a common silicate–ferrous density of 3550 kg m^{-3}. Three groupings of object are discernable: the first includes the largest satellites (Galilean, Titan, Triton), and incidentally Pluto, with masses in the range 1×10^{22} to 2×10^{23} kg. The Moon (a comparable body) and Io are included for comparison. The second grouping has masses in the range 4×10^{20} to 2×10^{21} kg. The remainder, comprising Mimas, Enceladus and Tethys, lie in the mass range 10^{19} to 7×10^{19} kg. For comparison, the satellites of Mars (probably captured bodies) have masses in the range 10^{15} to 10^{16} kg, and the largest asteroid (Ceres) has a mass of order 10^{18} kg. The object that caused the crater on Mimas probably had a mass in the range 10^{15}–10^{17} kg (Cole 1985). One senses a quantization of mass here.

4.4. *Early conditions external to the satellites*

It is usually supposed that the parent planet and the satellites formed at about the same time. The free-fall times of the materials and effects of rotation make it possible that the satellites formed more quickly than the central planet. If this were so the satellites could have been in orbit while the central planet was still in the later stages of forming. The formation of any planetary body is associated with the production of heat of material compression which raises the temperature of the condensed planetary material.

The self-gravitational energy E_g associated with a planetary body of mass M_p and equilibrium radius R_p is GM_p^2/R_p where G (6.67×10^{-11} m^3 kg^{-1} s^{-2}) is the constant of gravitation. For a body of Jupiter mass (1.99×10^{27} kg) and radius (7×10^7 m) the energy is $E_g = 3.8 \times 10^{36}$ J

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(Cole *et al.* 1986). Part of this energy will be retained by the condensed body and part will be radiated away: suppose half is retained. The radiated energy must pass outwards through the surface, we presume for the present purposes as a black body. If this process takes the time t_f, the surface radiation flux E_s will be $E_s = [GM_p^2/8\pi R_p^3 t_f]$. The rate of energy loss will depend partly on t_f (that is the amount of time necessary for the formation) and also partly on the conditions external to the forming planet. High dust concentrations outside the surface would retard the loss of heat and could make the Newton linear temperature law more appropriate than the T^4 law that we have used.

For a black body radiating into free space we would use the Stefan–Boltzmann T^4 law. Because black-body radiation would provide the lowest surface temperatures the values quoted are a minimum for any formation mechanism over a period of 10^6–10^7 years. The surface fluxes and temperatures calculated this way for Jupiter and Saturn are listed in table 2 together with the consequent temperatures at two satellite distances. Over the formation period each major planet imitates the Sun, though on a restricted scale, and heats its environment. Like the present Sun it will have been surrounded by silicate–ferrous satellite bodies. Thermal conditions in the neighbourhood of the major planets during this period could have been not dissimilar to those near the Sun now.

TABLE 2. THE MEAN SURFACE FLUXES AND THE SURFACE TEMPERATURES FOR JUPITER AND SATURN DURING THE FORMATION PERIOD; THE FLUXES AND TEMPERATURES AT TWO SATELLITE ORBITS ARE INCLUDED

time of contraction s	body	mean surface flux $\mathrm{J\,m^{-2}\,s^{-1}}$	surface temperature K	orbit temperature K	
3.16×10^6	Jupiter	8.70×10^5	1980	Europa	650
				Callisto	387
	Saturn	1.26×10^5	1220	Mimas	690
				Titan	270
3.16×10^7	Jupiter	8.70×10^4	1115	Europa	365
				Callisto	220
	Saturn	1.26×10^4	630	Mimas	390
				Titan	151

The surface temperatures of the satellites would be above the boiling points of the non-silicate materials concerned. In that case the ices would be melted and so form atmospheres and 'oceans'. As the central planet cooled the atmospheres would freeze, from the most distant members inwards. On this basis the present ice mantles are to be interpreted as frozen atmospheres.

5. Atmospheres as outgassed materials

An early atmosphere can arise from accreted or deposited materials, or by the emission of gas by the forming silicate–ferrous materials (generally called outgassing). Outgassing could have been important if the silicate material were at sufficient temperature during the formation period, which is likely. The accreted materials would presumably be similar to those forming the body.

Experience of larger silicate bodies can provide clues to the outgassing of the smaller bodies, and it is natural to turn to the terrestrial planets for this purpose. If all the atmospheres had outgassed from essentially similar silicate–ferrous centres and contiguous materials at their

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formation, they would all be expected initially to have been similar in the major components. The present atmospheres have, of course, been affected by evolutionary changes (see Holland *et al.* 1986).

5.1. *Aspects of the composition of the terrestrial atmospheres*

The atmospheres observed now will have lost the less-massive lighter components over the lifetime of the planets, the lightest being lost the more readily. Mercury would have lost virtually all its atmosphere even if it had initially been as massive as 10^{19} kg and indeed it now has no atmosphere except a tenuous collection of solar-wind particles (mainly hydrogen). The present atmosphere composition (see Taylor 1986) for the five most abundant components of Venus, Earth and Mars are collected in table 3. These are very different. The proportions of components are closely similar for Venus and Mars, but the mass of the martian atmosphere is some four orders of magnitude smaller: the composition of the Earth's atmosphere is different and its mass is intermediate between those of Venus and Mars.

Table 3. Data for the four principal components of the atmospheres of the planets Venus, Earth and Mars

	Venus		Earth		Mars	
component	mass/kg	%	mass/kg	%	mass/kg	%
carbon dioxide	4.7×10^{20}	96.0	3×10^{15}	0.033	3×10^{16}	95.0
nitrogen	10^{19}	3.5	4×10^{18}	77.0	5×10^{14}	3.5
oxygen	—	—	10^{18}	21.0	3×10^{13}	0.13
water	2×10^{16}	0.01	3×10^{16}	1.0	3×10^{12}	0.03
argon 40	3×10^{16}	0.007	6×10^{16}	0.9	4×10^{14}	1.6
total	4.8×10^{20}	99.62	5.1×10^{18}	99.99	3.1×10^{16}	100.3
mean molecular mass	43.4		29.0 (dry)		43.5	
mean surface pressure/bar	92		1		7×10^{-3}	
temperature/K	730		288		220	

5.2. *The Earth*

The Earth's atmosphere loses CO_2 through precipitation of water, and through the action of living material. Living material now also puts oxygen into the atmosphere. It is possible (Carmichael 1982) that as much as 1% of the crustal mass (that is 1% of 2.4×10^{22} kg) is due to absorbed CO_2 and if this is true the early atmosphere would have contained about 2.4×10^{20}kg of CO_2 which has since been removed. This quantity cannot be stated precisely because conditions are rather more complicated (Houghton 1977), but we can take this value for our present arguments. The composition of the Earth's atmosphere then would be as in table 4. This proportional composition is astonishingly similar to the present composition of Mars, except for the enhanced oxygen content for the Earth. The excess of atmospheric oxygen is conventionally associated with the presence of living material, and so will have increased from an initial value. The inferred excess can, in fact, be acounted for (N. Brindle, personal communication) by a build-up over a period of about 10^7 years followed by the equilibrium balance which applies now. The ocean water (of mass about 10^{21} kg) must be regarded as part of the initial material budget: these data are also included in table 4.

TABLE 4. HYPOTHETICAL COMPOSITIONS OF THE EARTH'S ATMOSPHERE WITH CRUSTAL CARBON DIOXIDE INCLOUDED, AND WITH BOTH THE CRUSTAL CO_2 AND THE OCEAN WATER

component	with crustal CO_2 mass/kg	%	with crustal CO_2 and oceans mass/kg	%
carbon dioxide	2.4×10^{20}	98.0	2.4×10^{20}	19.3
nitrogen	4.1×10^{18}	1.7	4.1×10^{18}	0.33
oxygen	1.1×10^{18}	0.45	1.1×10^{18}	0.09
water	3×10^{16}	0.02	1×10^{21}	80.7
total	2.45×10^{20}	100.0	1.24×10^{21}	100.0

5.3. *Mars*

Although relatively thin now, the martian atmosphere cannot have been so in the past. This conclusion follows from the observed channels (Baker 1982) in the surface that have the appearance of having been caused by flowing liquid. This must have been water: CO_2 will not flow under martian atmospheric temperatures and pressures, and the temperature would not have been low enough for either nitrogen or oxygen to liquify.

For water to exist freely on the surface its partial atmospheric pressure (near the freezing point) must have been about 7×10^2 N m^{-2} (or 10^{-2} bar): this would correspond to a mass of water about 3×10^{16} kg in the atmosphere. The present observed mass of water vapour (see table 1) is about 8×10^{12} kg. Over its lifetime, Mars would be expected to lose naturally only about 10^8 kg of water so the present low water-vapour pressure must be due either to absorption of water in the surface, or to catastrophic loss, or to both causes. There is ample evidence (Greeley 1985) from the *Mariner* data to suggest substantial 'sludge' regions (ejecta blankets) around existing features (for instance, the crater Yuty and the crater Arandas) and this is most likely to be water. Crustal freezing must then involve about 3×10^{16} kg of water from the atmosphere lying in the surface crustal region, being a volume of water of about 3×10^{13} m^3. Spread evenly over the surface (although surface topology would intervene) it would form a layer about 20 cm thick. It is likely (Cole 1986; Carr 1987) that the water content is much greater than this and that a remanent ocean is also included underground. If ocean water were outgassed on the Earth's scale, though proportionately to the martian mass, this would be some 10^{20} kg, or a surface layer uniform over the surface about 700 m thick. Estimates of martian crustal water by other authors have fallen within these two limits. The elucidation of the actual water content of the martian crust is an important future requirement.

CO_2 is known to be deposited in the polar regions, and the amount will be quite large, although the thickness of the polar caps is not known. Judging from the quantity of water required to form the observed surface channels, the corresponding mass of CO_2 on the basis of the present relative proportions of constituents would need to be about 2×10^{20} kg. This much could be contained in extensive polar caps some 1.5 km thick. The reduction of the nitrogen content presents a more difficult problem, though it is possible that some proportion has entered the crustal material.

The overall structure of the Tharsis region can probably be properly understood only if it were linked with an encounter with a substantial body (Connell 1986; see also Connell & Woolfson 1983; Wilhelms & Squyres 1984). This could cause atmosphere to be lost or gained but the colliding body would itself have been of silicate–ferrous composition.

5.4. *Venus*

This planet provides considerable problems in relation to water. Conditions now are not appropriate to its free existence on the surface. If the initial outgassing had included water on the Earth's scale it should now either be in the atmosphere or dissociated into hydrogen and oxygen. Account must then be made for some 10^{20} kg of hydrogen initially. The Jeans mechanism would probably account for no more than the loss of 10^{19} kg (that is some 10%) so the atmosphere should be richer in hydrogen now than is observed. Again a comparable quantity of oxygen would be expected, much of which would not have been lost by conventional mechanisms. Unfortunately, the observed atmospheric oxygen content is infinitesimal (Chamberlain 1986).

A catastrophic exchange of atmosphere cannot be ruled out. The rotational and orientational characteristics of Venus suggest a past collision event and this could have affected the atmosphere in ways that cannot be ascertained at the present time. There is an urgent need to have information about the dynamics of the Venus crust and mantle, and the role of volcanoes in the atmospheric compositional budget.

5.5. *A standard atmosphere for a silicate body?*

The present atmospheric compositions of the terrestrial planets are not inconsistent with the hypothesis that all started with atmospheres of similar composition and quantity proportional to the total mass of the body, but these evolved differently (particularly for Venus) or all constituents were lost (for Mercury over a period of about 10^9 years).

Whether a common outgassing is assumed or not, it is probably necessary to suppose some earlier catastrophe for both Venus and Mars to account for the full details of the present physical conditions for these two planets. If the initial outgassing were supplemented by material from outside, the extraneous material would need (not unexpectedly) to have been of essentially the same composition as the outgassed material.

There is, then, the possibility of 'common' initial atmospheres for the terrestrial planets (derived from outgassing and perhaps accretion) even though each has subsequently evolved differently. To construct this we account for the common CO_2 proportions for Mars and Venus, and the oxygen and water content of Mars, together with the ocean content of Earth. The general proportions of the four principal components of such an hypothetical atmosphere for a silicate–ferrous body are essentially as for Earth in table 4.

This atmosphere and ocean structure, if realistic, would be the standard against which atmospheres of other silicate–ferrous bodies can be judged. If the icy satellites were once siliciate–ferrous cores with atmospheres and oceans it would be natural to suppose this composition to be the basic form in these cases as well, modified by local circumstances.

6. Consequences for the icy satellites

We can now return to the icy satellites. If the previous arguments were to hold true throughout the Solar System the present mantle compositions should be similar in their bulk composition to the hypothetical Earth atmosphere including ocean water listed in table 4. The frozen atmosphere–ocean combination would have a mean density of about 966 kg m^{-3}, whereas the frozen Earth's atmosphere without the oceans would have a density of about

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880 kg m^{-3}. For comparison, a frozen Venus atmosphere would have a density of about 1894 kg m^{-3}. Pure water ice would have a density at atmospheric pressure of 960 kg m^{-3} although polymorphic forms (Hogg 1975; Poirier 1982) associated with higher pressures (appropriate to the larger satellites) could have densities as high as 1500 kg m^{-3}. Reference to table 1 shows that the mean densities preclude a density as high as 1894 kg m^{-3} appropriate to a major concentration of CO_2. They do, however, allow for mixtures of water and CO_2; methane ice could also be a significant constituent because its density is relatively low.

The composition of the atmosphere of Titan (the only atmosphere for an icy satellite) is 85% nitrogen, 15% argon and 1% methane. The mass of the present atmosphere is about 10^{19} kg, which is (see table 1) about 0.02% of the total mass of the mantle. This is, within the present accuracy, comparable to the total nitrogen content to be expected on the basis of the arguments of § 5.5, and is understandable if the remaining constituents have frozen. The observations are, therefore, not inconsistent with the arguments for an outgassed frozen atmosphere. The reason why the other satellites have not got atmospheres is unclear although it is presumably associated with lower surface temperatures.

The data of tables 1 and 4 show the proportions of water in the terrestrial planets to be low in comparison with the icy satellites.

7. Primary and secondary bodies

We now make the hypothesis that the terrestrial planets be regarded as secondary condensations like the satellites but unlike the major planets which were separate primary condensations comparable to the Sun. Only the Sun became massive enough to form a star but the major planets shared a common formation process with it. The Sun and terrestrial planets must, then, share with the major planets and their satellites, common relation between the masses, self-gravitational energies and angular momenta of the primary and secondary bodies.

The data for the mass of the primary M_p (Sun, Jupiter, Saturn, Uranus and Neptune) and the total masses M_s of the secondaries (the terrestrial planets, and the icy satellites) as a fraction of M_p are collected in table 5. Also collected in the table are the self-gravitational energies $E = GM^2/R$ for a body of mass M and radius R: E_p there is this energy for a primary body and E_s the sum of the energies of the corresponding secondaries. The ratios of the rotational angular momenta of the primaries ω_p and the *orbital* angular momenta ω_0 of the secondaries as fractions of ω_p are also collected in table 5.

In any test of our hypothesis care must be taken with data for Uranus and Neptune. Uranus

Table 5. Data for the Sun, the major planets, the terrestrial planets regarded as secondary to the Sun, and the satellites of the major planets.

(The properties of the primary body are mass M_p, gravitational energy E_p and rotational angular momentum ω_p; the properties of the secondary bodies are mass M_s, gravitational energy E_s and orbital angular momentum ω_0.)

primary	M_p	(M_s/M_p)	E_p/E_s	ω_p/ω_0
Sun	1.99×10^{30}	5.939×10^{-6}	5.84×10^{10}	2.5×10^{-1}
Jupiter	1.899×10^{27}	2.071×10^{-4}	2.65×10^{6}	7.2×10^{-3}
Saturn	5.689×10^{26}	2.479×10^{-4}	7.44×10^{5}	1.2×10^{-2}
Neptune	1.03×10^{26}	5.53×10^{-4}	2.51×10^{5}	4.5×10^{-2}
Uranus	8.66×10^{25}	2.825×10^{-4}	9.05×10^{6}	1.2×10^{-2}

has a rotation axis making an angle of 98° with the normal to the ecliptic and the satellite system is unlike those for the other major planets in not having a large member. The satellite system for Neptune is rudimentary, there being no satellites of intermediate mass. Numerical analysis now will therefore involve only data for the Sun, Jupiter and Saturn systems.

7.1. *Properties of primary and secondary bodies*

Empirical relations are readily found to link the mass of the primary to collective features of the corresponding secondaries. These are constructed using only the data for the Sun, Jupiter and Saturn systems. First, the sum of the masses M_s of the secondaries is found to be related to the mass M_p of the primary by

$$M_s = 7.54 \times 10^8 \times M_p^{0.535}. \tag{1}$$

The total satellite mass increases with that of the primary, roughly as the square root. Second, the common logarithm of the energy ratio $y = \lg [E_p/E_s]$ is related to the common logarithm of the mass of the primary $m = \lg M_p$ by the quadratic formula

$$y = 56.95 - 4.82m + 0.109m^2. \tag{2}$$

The data for satellite masses and energies are shown in figures 3 and 4.

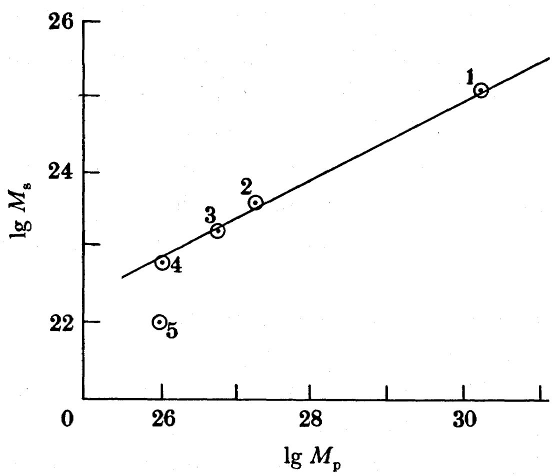

FIGURE 3. Showing the relation – expressed logarithmically – between the mass M_p of the primary body and the total masses M_s of the satellites. The solid line is derived from equation (1) of the text: the circled points refer to Sun (1), Jupiter (2), Saturn (3), Neptune (4), and Uranus (5).

7.2. *The cases of Uranus and Neptune*

The data for Neptune fit the formulae (1) and (2) to within 10% which is probably better than our present knowledge of the mass and radius of Triton. The Neptune system could have lost a mass up to some 2×10^{22} kg with self-gravitational energy up to 10^{28} J. The mass and self-energy of Pluto/Charon (Tholen 1985; Tedesco *et al.* 1987) are respectively 1.36×10^{22} kg and 9×10^{27} J, values close to our upper limits. A past association between Neptune and Pluto/Charon has been proposed by several authors and this would not be inconsistent with (1) and (2).

The values of the mass and self-gravitational energy of Uranus differ from (1) and (2) by a

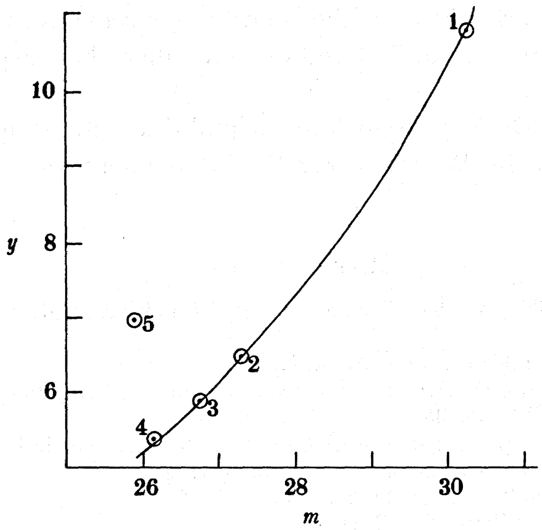

FIGURE 4. The relation between the logarithm of the ratio of self-gravitational energies $y = \lg [E_p/E_s]$ and the logarithm of the mass M_p of the primary body $m = \ln M_p$. The solid line is derived from equation (2) of the text: the circled points refer to Sun (1), Jupiter (2), Saturn (3), Neptune (4), and Uranus (5).

factor 2 or 3. This corresponds to deficits of mass and energy of respectively 5×10^{22} kg and 10^{29} J, which each easily contain the corresponding values for Pluto/Charon. This amount of matter, if a single body, would have a radius about 1.7×10^{6} m and a density about 2500 kg m^{-3}. Numerical rational functional interpolation gives an associated angular momentum deficit of about 10^{35} kg m^{2} s^{-1} and could imply an hypothetical orbit of some 8.5 Uranus radii, which is between the present orbits of Ariel and Umbriel. The energy in orbiting Uranus would be 6×10^{29} J: that for Pluto/Charon about the Sun is about 1.5×10^{29} J. Pluto/Charon have rotation axes closely parallel to that of Uranus and a large orbital inclination (17.2°) to the ecliptic. A past link between Pluto/Charon and Uranus remains a possibility.

It is suggested that there has been a significant past rearrangement of the initial satellite systems of Uranus and Neptune and the unambiguous association of Pluto/Charon with one or other of these planets would help the further refinement of the present arguments.

8. Final comments

The arguments presented here raise questions to be answered in detail in the future. For instance, there should be a substantial quantity of martian water underground. Is it there and in what quantity? And what is the mean thickness of the polar caps? What are the crustal dynamics of Venus? Are the icy satellites (and particularly the smaller ones) separated in their silicate/ferrous and ice components or not? What is the composition of the satellite silicate materials and the ices? Are the compositions of the silicates in the outer Solar System the same as those in the inner system: if not, how do they differ? How does Pluto/Charon fit into the pattern? There is one additional conjecture. The motion of Jupiter is the major perturbation of the solar motion about the centre of total mass. The Jupiter orbital period is 11.86 years and this is closely the semiperiod for solar magnetic activity: is this coincidence? This does, incidentally, introduce magnetism which has been ignored so far.

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Our appreciation of the early history of the Solar System remains rudimentary. It would be an advantage if astronomers were to find and observe other planetary systems.

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