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Meteoric Material – an Important Component of Planetary Atmospheres

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Interplanetary dust particles (IDPs) interact with all planetary atmospheres and leave their imprint as perturbations of the background atmospheric chemistry and structure. They lead to layers of metal ions that can become the dominant positively charged species in lower ionospheric regions. Theoretical models and radio occultation measurements provide compelling evidence that such layers exist in all planetary atmospheres. In addition IDP ablation products can affect neutral atmospheric chemistry, particularly at the outer planets where the IDPs supply oxygen compounds like water and carbon dioxide to the upper atmospheres. Aerosol or smoke particles from incomplete ablation or recondensation of ablated IDP vapors may also have a significant impact on atmospheric properties.

1. INTRODUCTION

Atmospheric ablation of infalling meteoroids releases foreign atoms, molecules, and ions into the surrounding atmosphere. The deposited material becomes part of a planet's ionosphere and can play key roles in minor species aeronomy. In addition, the cosmic particles contain a large percentage of volatile elements like oxygen and carbon, which, when deposited at Earth, mostly blend unnoticed into the atmosphere. However, for the outer-solar-system planets they contribute to the deposition of water and/or carbon dioxide molecules that may not otherwise exist in the upper atmospheres of these planets. Each planet's interaction with the interplanetary particle population is different due to its unique atmosphere, size and position in the solar system. The biggest contrast in the effects of the ablated meteoric material is between the outer planets (Jupiter, Saturn, Uranus and Neptune) with reducing atmospheres, and the inner terrestrial planets (Venus, Earth and Mars) with oxidizing atmospheres. This paper describes some of these differences in the atmospheric and ionospheric consequences of the constant rain of interplanetary dust particles (IDPs). In this paper IDP refers to any cosmic particle with radius less than ~ 1 cm regardless of its origin.

2. BACKGROUND

2.1. Properties of Dust in Solar System

Accurate quantitative predictions of the effects of IDPs on an atmosphere require a knowledge of the IDP mass and velocity distributions at the planet. Extensive radar and visual meteor observations and spacecraft measurements provide information on the velocities, total impact flux and mass (size) distribution at the Earth [*Grün et al.*, 1985; *Love and Brownlee*, 1993; *Taylor*, 1995]. Knowledge of the IDP spatial density and velocity distribution decreases the further one goes from Earth. Spacecraft-borne dust detector measurements from around the solar system combined with zodiacal light observations partially constrain the particle flux as a function of heliocentric distance [e.g., *Leinert and Grün*, 1990; *Grün*, 1994]. However, the spacecraft data contain only a small number of events and sample a range of particle sizes that may miss a sizeable portion of the population. Because of uncertainties in the mass flux and other properties of IDPs at planets other than the Earth, measurements of atmospheric and ionospheric phenomena related to IDPs provide valuable benchmarks for testing the realism of the assumed IDP distributions.

Comets and asteroids are believed to be the main source of the inner solar system IDPs. The asteroidal component is not as important in the outer solar system, where ice-rich particles should be more numerous. An IDP will impact an atmosphere with a velocity between the escape velocity of the planet and a maximum value [*Öpik*, 1958; see Table 1]. The maximum speed is that of a particle on a retrograde parabolic orbit around the Sun in a head-on collision with the planet (combined with gravitational focusing by the planet). Interstellar material with greater velocities has been observed but is not likely to be a significant contributor to the deposited mass [*Grün*, 1994]. Note from Table 1 that the range of impact speeds is much broader for the terrestrial planets. For particles encountering the planet from random directions, the most probable impact angle is 45°.

Although IDPs with a wide range of masses impact an atmosphere, most of the $\sim 10^8$ kg of yearly deposited mass at the Earth arises from small grains with radii in the size range ~0.01 to 1 mm [e.g., see the reviews in *McDonnell*, 1978]. Very small IDPs with radii less than ~10 μm radiate the frictional heat and do not evaporate for typical Earth entry velocities, while impacts of very large bodies are relatively rare on time scales relevant to the lifetime of meteoric material in the atmosphere. Free-molecular flow conditions generally apply to IDP interactions with an atmosphere. In ablation models it is commonly assumed that the particles are spheres with uniform composition, and fragmentation is ignored.

More refined measurements of the IDP distribution functions and their physical characteristics throughout the solar system are needed. At current levels of understanding, models of atmospheric consequences should be considered as order-of-magnitude estimates rather than as precise, quantitative predictions.

2.2. Ablation Process

A fast moving meteoroid heats as the kinetic energy of colliding atmospheric molecules is converted into internal energy. The same collisions can also sputter material off the surface [*Johnson*, 2002]. This energy transfer causes melting and evaporation of surface material off the particles [*Öpik*, 1958]. The amount of material released and the altitude profile of the mass deposition depends on the particle's mass, velocity, size, physical structure, thermodynamic properties and chemical composition, and on the mass density profile of the atmosphere. Three equations describe the physics of meteoroid ablation: (1) the momentum equation, relating the reduction in speed to atmospheric drag; (2) the ablation equation, describing mass loss from the meteoroid by evaporation and sputtering; and (3) the heat equation, relating the bulk heating of the meteoroid to frictional heating, radiative energy loss and evaporative cooling. A fourth equation is used to track a particle's position as a function of time.

Using the equations and assumptions described above, models of the IDP ablation rate profiles and atmospheric effects at the Earth have been made [e.g., *Lebedinets et al.* 1973; *Hunten et al.*, 1980; *McNeil et al.* 1996; *Grebowsky and Pesnell*, 1999]. The results are consistent with available measurements. The investigations at Earth show that, with our current understanding, there are adequate modeling tools for exploring IDP effects at the other planets.

2.3. Ionospheric Processes

Ions formed by the impact of fast-moving ablated neutral atoms with atmospheric molecules are an immediate consequence of the ablation of a high-speed meteoroid (see reviews in *McDonnell* [1978]). Ion densities are initially very high in the train of a particle but are quickly reduced by diffusion. The number of ions deposited by one typical grain is not very large. For example, at Earth, for a nominal 20 km/s entrance speed an IDP with mass 10^{-6} g, (i.e., a diameter of a few 10's of microns) directly deposits 10^{12}-10^{13} ions [*Lebedinets et al.*, 1973]. However, the lifetimes of atomic ions are long. Hence the continual influx of IDPs can seed persistent meteor ion layers. The impact ionization rate of the freshly vaporized neutrals, moving at the meteoroid speed v in the Earth's atmosphere, varies as v^{α}, where α is ~3-4 for chondritic materials [*Bronshten*, 1983].

This source of ionization is particularly significant for planets with large escape velocities (see Table 1) and for high-speed meteor streams. It is not important for particles in the sporadic background entering the atmospheres of the low mass terrestrial planets.

Photoionization will also ionize the ablated neutral atoms. The wavelengths of the ionization thresholds of the metals extend from ~1500 Å for Si to ~2700 Å for K [*Swider*, 1969]. Depending upon the atmospheric composition and pressure profile, the wavelengths of the solar spectrum that can ionize specific neutral meteoric species may or may not be absorbed above the ablation region. The importance of this mechanism varies from species to species because of the differing ionization potentials, and the photoionization rate depends on the distance from the Sun. Also, because the ionization potentials of metal atoms are low compared to the ionization potentials of ambient ionospheric species (see Table 2), charge exchange with ionospheric ions is another mechanism for ionizing the neutral metal atoms [*Swider*, 1969]. The importance of this ionization source depends on the magnitude of the ambient ionospheric ion concentrations in the ablation region.

The chemical loss processes for the meteoric ions and the dynamical rearrangement of the ions will differ from planet to planet. The loss of the meteoric atomic ions at high altitudes is by radiative recombination with electrons (for Mg^+ this recombination rate is $\sim 4\times10^{-12}$ cm^3 s^{-1}) resulting in very long ion lifetimes (up to many weeks on Earth). With decreasing altitude, the atomic ions interact more with atmospheric neutral molecules (through two- or three-body interactions) to produce molecular metal ions. For example, Mg^+ reacting with O_3 (with a rate of $\sim 8\times10^{-10}$ cm^{3}s^{-1}) yields MgO^+, whereas the three body reactions occur at rates $\sim 1\times10^{-30}$ cm^6 s^{-1}, yielding molecular ions such as MgN_2^+, $Mg^+{\bullet}CO_2$ and MgO_2^+ [see *McNeil et al., 1996* and *Pesnell and Grebowsky*, 2000a, for a discussion of these rates]. The resulting metal ion molecules can undergo rapid dissociative recombination, providing another loss channel for atomic meteoric species. These reactions are characteristic of the inner planets. For the outer planets analogous processes can exist in which hydrogen and hydrocarbon species are the reactants for the meteoric ions. Aerosol layers (such as mesospheric cloud particles on Earth) can reduce the metal ion concentrations through clustering, as can charge capture onto dust particles.

A simple balance between these loss processes and ion production, including the effects of diffusion, would produce a single ion layer in the ablation region. Complexities are introduced by ionosphere dynamics whose importance is established for Earth; converging ion drifts due to combining neutral wind drag or electric fields with a magnetic field produce very narrow ion layers (see review by *Kelley*, [1989], and the modeling of *Carter and Forbes*, [1999]). One can assume with confidence that the terrestrial dynamic processes also apply, but to different degrees, to the other planets with magnetic fields. The very structured low altitude layers observed at Jupiter (see Figure 1a) and similar complexities observed at Saturn [*Kliore et al.*, 1980] and Neptune [*Lyons*, 1995] have been attributed to such processes.

2.4. Neutral Processes

Not all of the ablated material is ionized. Although both atoms and molecules evaporate from the IDP, subsequent atmospheric collisions will tend to break the molecules into component atoms [*Öpik*, 1958] of which only a fraction is ionized. Ablated neutral atoms have decreasing lifetimes with decreasing altitude (either because of chemical reactions transforming them into molecular species or due to condensation of the atoms about any existing aerosols.) Layers of meteoric neutral atoms form and peak at an altitude below the corresponding metal ion layers [e.g., *McNeil et al.*, 1996]. The process of recondensation of refractive species in the dense vapor could also yield small dust (or smoke) particles [e.g., *Hunten et al.*, 1980]. Hence, in addition to the small particles that do not ablate, a broad range of particle sizes are left in the atmosphere that will slowly diffuse or fall downward to leave traces far below the ablation zone.

Unlike the metal ions, which can be dominant ionospheric species, the neutral meteoric species are always minor constituents of the neutral atmosphere. Nevertheless, by either acting as catalysts or introducing reactive species that would otherwise be absent, they can have a persistent impact on the chemistry of the major atmospheric constituents. The introduction of oxygen, carbon, nitrogen, and sulfur atoms is particularly important for the giant planets. Furthermore, the spectral lines of the metals stand out against the background and can be used as tracers for chemical and dynamical studies with lidar [*Plane et al.*, 1999] and all-sky imagers [*Taylor et al.*, 1995].

3. TERRESTRIAL PLANETS

Meteoric ion and neutral particle measurements at Earth have revealed the major chemical and dynamic processes that control the metal distributions in the lower ionosphere. Prominently observed is a persistent metal ion layer between 85 and 100 km, the region of maximum meteor activity. Most of the metal ion species have maximum concentrations at the same altitude. Fe and/or Mg are typically the dominant metal ions (Figure 1b), consistent with a chondritic meteoroid source. The altitude of the peak, the peak ion concentration, and the shape of the layer vary with atmospheric temperature and composition and are affected by atmospheric winds and electric fields (produced in the magnetosphere or by the atmosphere-generated dynamo).

The dynamical processes can produce multiple, and very thin, ion layers. The metal ions at Earth are important components of the ionosphere. Metal neutral atoms charge exchange with ambient ions, thereby reducing their concentrations. Also there is a loss of ambient ions through enhanced dissociative recombination when the electron concentration increases due to the presence of the metal ions. Regions of the lower terrestrial ionosphere are often composed predominantly of metal ions, particularly at night. Specific references and discussion of all these features can be found in the reviews in *Murad and Williams* [2001].

For the other planets, there are no low altitude observations of the ion composition, but radio occultation measurements provide electron density profiles. If metal ions prevail, then, as at Earth, one would expect to see prominent electron density layers below the main ionosphere peak. Figure 1c shows evidence for such low altitude layers at the terrestrial planet Venus. *Butler and Chamberlain* [1976] were the first to note that the Venusian lower layer might be attributable to metal ions. No published radio occultation data for Mars shows the presence of such layering, but the published occultation studies do not show detailed electron density profiles at the low altitudes where IDP effects could occur.

For Mars and Venus, the ionospheric dynamics that lead to the complex, terrestrial metal ion distributions are not anticipated to play a significant role. The motions of metal ions at Earth are strongly constrained by the Earth's magnetic field. Venus has no intrinsic magnetic field, so ionospheric dynamics could be simpler. Mars is more complex in this regard because of the existence of localized remnant magnetic fields in its crust. Still, it is reasonable to assume that such processes will not affect the metal ion distributions outside of these magnetized regions.

Model studies for Mars [*Pesnell and Grebowsky*, 2000a] and Venus [*Pesnell and Grebowsky*, 2000b] have only considered the Mg profile. It is anticipated that Fe^+, the other major metal ion, would behave similarly. The calculations were steady state in one dimension. The Mars and Venus models employ the same ablation and chemical schemes used for Earth, but with background atmospheric and ionosphere models appropriate to the other planets. The assumption is made that the IDPs have the same mass distribution as at Earth. *Pesnell and Grebowsky* [2000a,b] showed that the CO_2 atmospheres of Mars and Venus do not strongly absorb the Mg-ionizing photons until well below the ambient ionosphere, so that photoionization is the main ionization process. Also, although the atmospheric composition of the three terrestrial planets differ, the chemistry pathways controlling Mg and Mg^+ involve the same reactions, albeit at different rates.

The Mars and Venus models are shown in Figure 2. Modeled Mg profiles for Earth follow similar patterns

[*McNeil et al.*, 1996; *Grebowsky and Pesnell*, 1999]. The Mg^+ layers are formed by the ionization of neutral Mg atoms diffusing up from the lower-altitude, peak-ablation region. On the topside of the Mg layer, almost all of the neutrals become ionized. The Mg^+ concentration drops with decreasing altitude because it is transformed, via 3-body reactions with atmospheric neutral molecules, into molecular ions. A prominent layer with peak concentration similar to values at Earth is predicted near 80 km at Mars. However, for Venus only a very weak metal ion layer in the vicinity of 110 km is modeled. The Venus and Mars predictions appear counter to the radio occultation measurements for the two planets — i.e., no such layer has been reported for Mars while a prominent layer (due to a non-meteoric origin?) is sometimes seen at Venus (Figure 1c). More intensive investigations of the occultation measurements would help to resolve these differences.

Meteoric material can also play roles in the neutral atmospheric chemistry and aerosol distributions at the terrestrial planets. Neutral meteoric species seldom exceed part-per-billion mixing ratios in the terrestrial atmosphere. Nevertheless, meteoric metals are possible agents for the chlorine-catalyzed destruction of ozone on the Earth [*Rodriguez et al.*, 1986; *Aikin and McPeters*, 1986, *Prather and Rodriguez*, 1989]. On Venus, the only direct evidence of meteor chemistry is ultraviolet nightglow in which NO emission was seen along with what may have been an extended meteor trail [*Huestis and Slanger*, 1993].

Solid meteoritic dust in the terrestrial-planet atmospheres may affect neutral atmospheric chemistry, cloud and haze formation, radiative balance of the atmosphere, and global climate change. Tiny smoke particles formed in the mesosphere and thermosphere from the recondensation of refractory ablated vapor settle throughout the atmosphere, and larger (radius greater than ~1 μm) unablated or partially ablated micrometeorites could dominate the large-particle component of the stratospheric aerosol layer [*Hunten et al.*, 1980]. These particles could provide condensation nuclei for noctilucent clouds in the Earth's mesosphere [e.g., *Thomas*, 1991]. An increasing body of evidence also suggests that the sulfate aerosols in the stratosphere contain meteoric debris, perhaps from heterogeneous condensation of sulfuric acid about meteoric smoke particles [*Cziczo et al.*, 2001]. The solid meteoric debris can also provide surfaces upon which heterogeneous chemical reactions occur [e.g., *Summers and Siskind*, 1999]. If meteoric dust does play an important role as condensation nuclei for stratospheric aerosols or for mesospheric noctilucent clouds, the effects on the radiative balance and global energy budget of the atmosphere could be profound. In the atmospheres of Mars and Venus, solid meteoric debris could have the same effect as on Earth. *Turco et al.* [1983] and *Michelangeli et al.* [1993] demonstrated that recondensed meteoric particles can act as nucleation sites for aerosol formation in the upper atmospheres of these planets.

4. OUTER PLANETS

Occultations of radio signals from spacecraft provide evidence for very structured and time varying ionospheres on the outer planets (see review by *Kar* [1996]). A Jovian profile is shown in Figure 1a. Neutral atmosphere measurements, in combination with IDP-atmosphere interaction models, provide independent evidence for the importance of the meteoroid inputs.

As a needed input to the models, spacecraft measurements provide a constraint on the IDP number density as a function of heliocentric distance [e.g., *Humes* 1980; *Grün,* 1994]. The window of possible IDP impact velocities into each of the outer planetary atmospheres is quite small (see Table 1). Hence ablation rate effects that are dependent upon the particle speeds are fairly well defined. However, knowledge of the velocity distribution outside of a planet's gravitational sphere is necessary to estimate the flux enhancements due to gravitational focusing. The enhancement is less than a factor of ~ 2 for the inner planets; but, for the outer planets the enhancement can range from a factor of ~6 to 200, depending on the assumed IDP orbits [*Moses et al.*, 2000]. The relative contributions from different particle sizes are unknown for the outer solar system. All ablation calculations to date have used the terrestrial mass distribution. The approach taken in the IDP-atmosphere studies has been to make reasonable estimates of the total impact flux and velocity distribution for plausible sources and orbits of the particles. Comparisons of the model results with observed electron density profiles and neutral atmospheric constituents then allow the validity of the assumptions to be checked.

In the outer solar system, the IDPs are probably comet-like mixtures of ices, silicates, and complex organic materials. During ablation, water and other volatiles will be released at higher altitudes than metal and silicate vapor because of the lower vaporization temperature of the ices. Also, since outer planetary atmospheres consist predominantly of H_2, He and hydrocarbons (Table 1), meteoric oxygen must be considered, unlike at the inner planets, where meteoric oxygen becomes a trivial component of the overall atmospheric oxygen budget. Ablated material is ionized by charge exchange and impact ionization; photoionization is not important. Detailed model studies have been made of the meteoric ionization for Jupiter [*Kim et al.*, 2001], Saturn [*Moses and Bass*, 2000], Neptune [*Lyons*, 1995], as well as for Saturn's moon Titan [*Molina-Cuberos et al.*, 2001].

The outer planet ionosphere models include hydrogen and hydrocarbon ions as well as one or more meteoric elements. Examples of the modeled ion profiles for Jupiter

[*Kim et al.*, 2001] and Neptune [*Lyons*, 1995] are shown in Figure 3a and 3b. Both calculations assume a cometary composition for the ablated IDP atoms. The incident meteoroid mass flux for the Jovian calculation in Figure 3a is the terrestrial influx value of 1.5×10^{-16} g cm^{-2} s^{-1}. This flux magnitude yields a metal ion peak density comparable to the occultation measurements of a few 10^4 cm^{-3} [*Hinson et al.* 1998]. The incident flux used for the Neptunian model of Lyons was 7×10^{-16} g cm^{-2} s^{-1}, a value that lies between the *Moses* [1992] predictions for randomly inclined, highly eccentric particles and low inclination, low eccentricity particles. Saturn's ionosphere was modeled by *Moses and Bass* [2000], with Mg^+ as sole representative of the major metal ions. An IDP influx of 3×10^{-16} g cm^{-2} s^{-1} was employed, based on models of the Infrared Space Observatory (ISO) observations of H_2O and CO_2 in Saturn's upper atmosphere [*de Graauw et al.*, 1997; *Feuchtgruber et al.*, 1997; *Moses et al.*, 2000]. The altitudes and maximum densities of the metal ion layers in all three of these model studies are in agreement with the magnitude and altitude zone of the low-altitude ionospheric layers probed by radio occultation. However, these simple models do not match the complexities of the occultation-measured layers (e.g., time varying and sharp multiple layers). *Lyons* [1995] and *Moses and Bass* [2000] showed that such structures can be produced by the inclusion of ion transport associated with vertical ion drift shears, as occurs in the Earth's ionosphere. Waves in these motions will yield multiple layers. Ionospheric structures for atmosphere-bearing satellites of the outer planets may also have a meteoric origin. *Molina-Cuberos et al.* [2001] predict meteoric ion concentrations (shown in Figure 3c) on Titan comparable to those of the background ionosphere produced by photochemistry.

The 1-D average model ionospheres in Figure 3 and the model for Saturn have Mg^+ and/or Fe^+ as the major low-altitude ion. *Kim et al.* [2001] and *Moses and Bass* [2000] also modeled local time variations for Jupiter and Saturn, respectively. They found double-layer ionosphere structures (as in Figure 3b) that were most prominent just before dawn. The upper layer was due to H^+, whereas metal ions dominated the lower layer. The molecular ion concentrations decay through the night. All the models use charge exchange as a source of metal ions, but the Jupiter calculation also includes impact ionization of the ablated neutral atoms. Because of the high incident velocity of the IDPs at Jupiter, this ion production source is dominant. For the Saturn and Neptune studies, recondensation onto dust particles is the primary low-altitude loss channel for the ablated metal atoms. For Jupiter *Kim et al.* [2001] include the loss of meteoric ions by three body reactions with H_2 and hydrocarbons. Such rates have yet to be measured for Mg^+ and Fe^+. The *Kim et al.* [2001] model depicted in Figure 3a used the laboratory-measured rates of the three body reaction of sodium with H_2 scaled for Mg^+, but considered a similar Fe^+ process unlikely. The modeled dominance of Mg^+ over Fe^+ simply reflects the relative composition of their parent neutral species in the incoming meteoroids. Laboratory studies are lacking for many of the reactions of metal atoms with hydrocarbons. The species H_3O^+ in Figure 3a is the result of the meteoric oxygen.

Because of the complex structures produced by ionospheric dynamics, it is difficult to use the occultation measurements to precisely constrain the IDP input fluxes. Neutral atmosphere measurements provide another avenue for exploring the IDP characteristics. Recognition that the oxygen-bearing vapor from extraplanetary sources could participate in stratospheric photochemistry on the outer planets first arose when CO was discovered in Jupiter's atmosphere [*Beer*, 1975]. Because carbon monoxide is a disequilibrium species in Jupiter's atmosphere, *Prather et al.* [1978] suggested meteoroid ablation as a source of water vapor in Jupiter's upper atmosphere — subsequent photochemistry would then convert the H_2O to CO. However CO could also be produced in the deep atmosphere and transported to the upper troposphere where it is observed [e.g., *Prinn and Barshay*, 1977; *Fegley and Lodders*, 1994]. The same ambiguities apply to the interpretation of the observed CO at Saturn and Neptune [e.g., *Noll et al.*, 1986; *Marten et al.*, 1993].

The definitive evidence for influxes of atmosphere-modifying material on planets other than Earth was the discovery of H_2O in the stratospheres of all the giant planets and CO_2 on Jupiter, Saturn, and Neptune [*de Graauw et al.*, 1997; *Feuchtgruber et al.*, 1997; *Bergin et al.*, 2000]. Intrinsic water vapor on the outer planets will condense deep in the tropospheres and should not observed in the stratospheres. Also although CO_2, like CO, may be transported upwards from the deeper regions in which it is thermodynamically stable, the predicted CO_2 mixing ratios in such a scenario are much smaller than what is observed [*e.g., Lellouch et al.,* 1998]. Therefore an external source is needed [e.g., *Feuchtgruber et al.,* 1997] and the most likely source is meteoroid ablation [*Moses et al.,* 2000]. Modeled IDP effects in the Saturnian atmosphere from the latter study are shown in Figure 4. The oxygen species could affect the abundance of some of the less abundant atmospheric hydrocarbon molecules. IDP ablation will also introduce nitrogen- or sulfur-containing species or other volatile vapors that are not normally present in the upper atmospheres of the giant planets. Many aspects of the interaction of meteoric vapor with outer planetary atmospheres remain to be explored.

Residual micrometeorites and smoke particles formed from ablated-IDP vapors can also affect photochemistry through the attenuation of solar ultraviolet radiation. They might cause localized heating in the upper atmosphere [e.g., *Rizk and Hunten*, 1990], which could affect atmospheric dynamics. They can act as condensation nuclei to facilitate stratospheric haze formation and provide surfaces upon which heterogeneous reactions could occur.

Moons of the outer planets sample the same IDP population as their parent planet (although in different proportions). The presence of CO [*Lutz et al.*, 1983], CO_2 [*Samuelson et al.*, 1983], and H_2O [*Coustenis et al.*, 1998] in Titan's atmosphere suggests an extraplanetary supply of oxygen. *Ip* [1990] and *English et al.* [1996] show the credibility of this interpretation by modeling the meteoroid-ablation deposition of oxygen-bearing materials into the moon's atmosphere. Triton, the largest moon of Neptune, may also provide a laboratory to study the IDP population at Neptune. The surface pressure of Triton's atmosphere of roughly 10 μbar to abate many IDPs may be sufficient [*Strobel and Summers*, 1995].

5. SUMMARY

Sufficient theoretical understanding exists to determine the ramifications that IDPs may have on all planetary atmospheres. Details of these interactions are measured at Earth, where meteoric ions often dominate the lower ionosphere and where it is suggested that IDPs could lead to the development of aerosol layers and possibly affect the stratospheric ozone population. Radio occultation measurements on the other atmosphere-bearing planets find ionospheric layers similar to the metal layers on Earth. Model studies support a metal ion interpretation. For the inner planets of Mars and Venus there is ambiguity between the modeled and measured layers. For the outer planets and Titan, IDPs can be a source for the water and carbon dioxide observed in these atmospheres. What is missing most in our understanding are measurements of ionospheric and atmospheric structures in other atmospheres and IDP properties throughout the solar system.

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Figure Captions

Figure 1. Planetary ionosphere measurements indicative of metal ion layers (indicated by arrows). (a) A Jupiter occultation [from *Hinson et al.*, 1998] with prominent narrow low altitude layers. (b) A sounding rocket measurement of terrestrial metal ion layers [from *Kopp*, 1997]. (c) A Venus radio occultation measurement of a low altitude layer. (Excerpted with permission from *Kliore et al.*, *Science*, *205*, page 99, 1979. Copyright 1979, American Association for the Advancement of Science.)

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Figure 2. Model calculations of the Mg ion and neutral layers for (a) Mars and (b) Venus [*Pesnell and Grebowsky*, 2000a,b].

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Figure 3. Meteoric ion models. (a) Model of Jovian ionosphere [from *Kim et al.* 2001] — altitude is referenced to 1 bar atmospheric pressure level. (b) Neptune model. (Reprinted with permission from *Lyons, Science*, *267*, page 649, 1995. Copyright 1995, American Association for the Advancement of Science.) (c) Metal ion model for Saturn's moon Titan [*Molina-Cuberos et al.*, 2001, reprinted with permission from Elsevier Science].

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Figure 4. Modeled dominant molecular neutral species resulting from IDPs in Saturn's atmosphere [*Moses et al.*, 2000].

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METEORIC COMPONENT OF PLANETARY ATMOSPHERES

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Table 1: Properties of Solar System Bodies with Significant Atmospheres

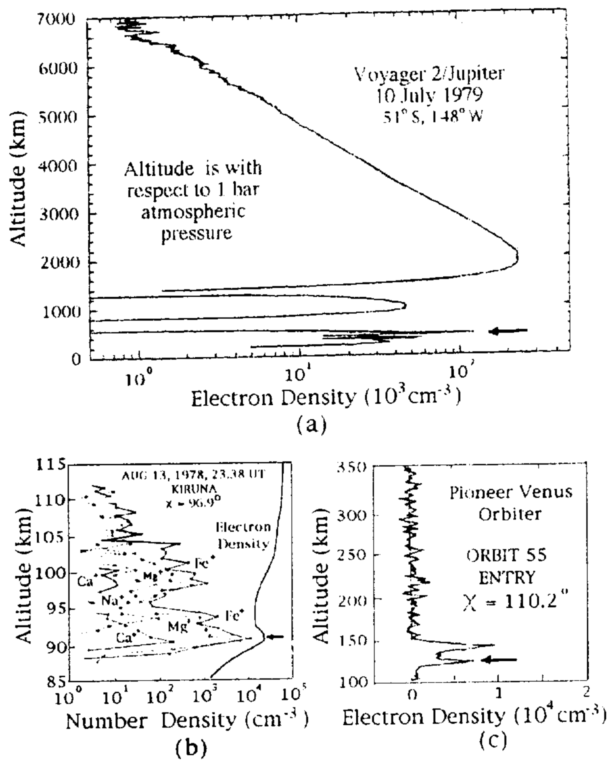
Object	Escape Speed[1] (km/s)	Maximum Speed[2] (km/s)	Important Atmospheric Constituents[3]
Venus	10.4	86	CO_2, CO, N_2, O
Earth	11.2	73	N_2, O_2, O, H_2O, O_3
Mars	5.0	59	CO_2, N_2, CO, O_2, O, O_3
Jupiter	60.0	69	H_2, He, CH_4, C_2H_6, C_2H_2, H
Saturn	36.0	44	H_2, He, CH_4, C_2H_6, C_2H_2, H
Titan	2.6	24	N_2, CH_4, H_2, H, C_2H_6
Uranus	21.0	27	H_2, He, CH_4, C_2H_6, H
Neptune	24.0	28	H_2, He, CH_4, C_2H_6, CO, H
Triton	1.5	18[4]	N_2, CO, Ar, CH_4

Notes:
[1]From *Tholen et al.* [2000].
[2]Calculated as described in text using data from *Tholen et al.* [2000].
[3]From *Yung and DeMore* [1999]. Due to changes with altitude, the actual mixing ratios are not listed.
[4]Includes orbital velocity of Triton around Neptune.

Table 2: Characteristic Ionization Potentials[1]

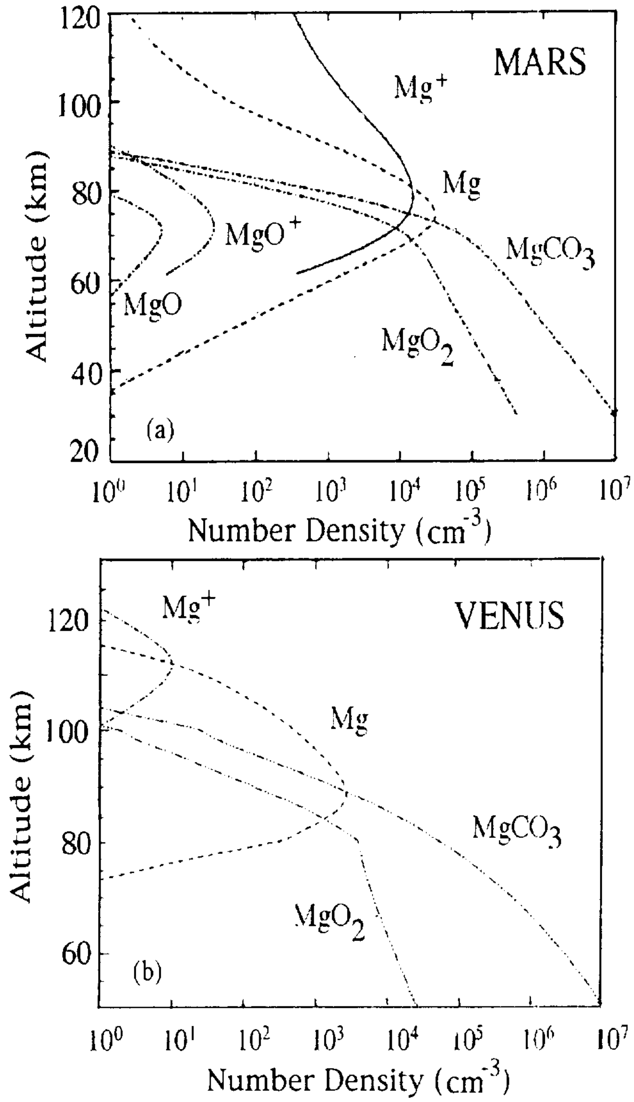
Meteoric Species	I. P. (eV)	Atmospheric Species	I. P. (eV)
Si	8.15	H_2	15.43
K	4.34	He	24.59
Na	5.14	O	13.62
Fe	7.90	O_2	12.07
Mg	7.65	N_2	15.58
Li	5.39	CO_2	13.77
Ni	7.64	CH_4	13.00
Ca	6.11	Ar	15.76

Notes:
[1]Data from *Däppen* [2000].

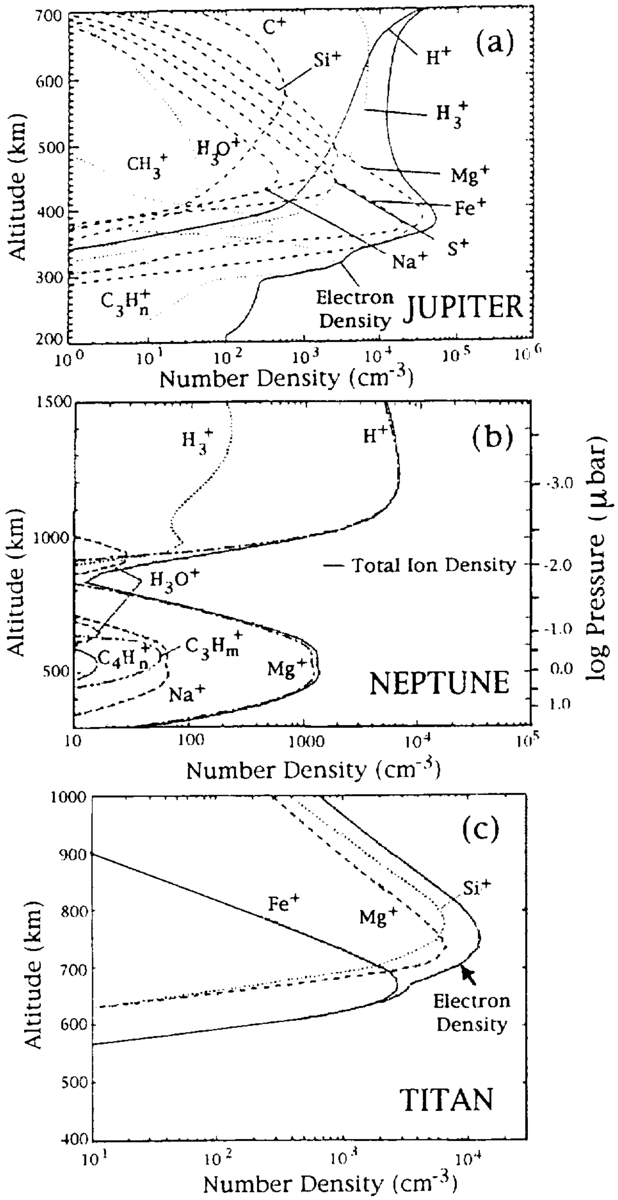


Voyager 2/Jupiter
10 July 1979
51° S, 148° W
Altitude is with
respect to 1 bar
atmospheric
pressure
Altitude (km)
Electron Density ($10^3 cm^{-3}$)
(a)
AUG 13, 1978, 23.38 UT
KIRUNA
Electron
Density
Altitude (km)
Number Density (cm^{-3})
(b)
Pioneer Venus
Orbiter
ORBIT 55
ENTRY
χ = 110.2°
Altitude (km)
Electron Density ($10^4 cm^{-3}$)
(c)

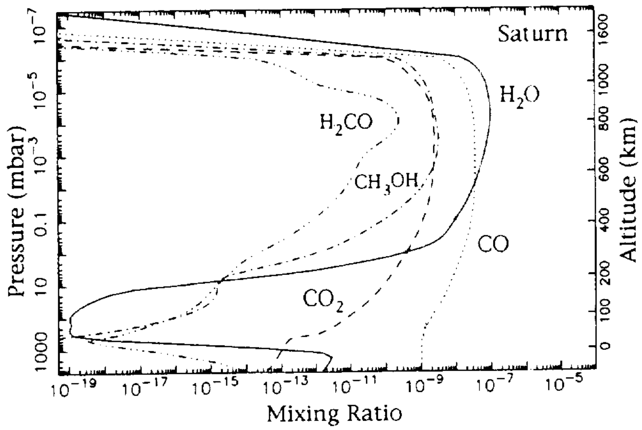
(FIGURE 1)


MARS
Mg+
Mg
MgCO3
MgO+
MgO
MgO2
(a)
Altitude (km)
Number Density (cm-3)
VENUS
Mg+
Mg
MgCO3
MgO2
(b)
Altitude (km)
Number Density (cm-3)

(FIGURE 2)


(a)
JUPITER
Altitude (km)
Number Density (cm^{-3})
C$^+$
Si$^+$
H$^+$
H$_3^+$
H$_3$O$^+$
CH$_3^+$
Mg$^+$
Fe$^+$
S$^+$
Na$^+$
C$_3$H$_n^+$
Electron Density
(b)
NEPTUNE
Altitude (km)
log Pressure (μbar)
Number Density (cm^{-3})
H$_3^+$
H$^+$
— Total Ion Density
H$_3$O$^+$
C$_4$H$_n^+$
C$_3$H$_m^+$
Mg$^+$
Na$^+$
(c)
TITAN
Altitude (km)
Number Density (cm^{-3})
Fe$^+$
Mg$^+$
Si$^+$
Electron Density

(FIGURE 3)



Saturn
H_2O
H_2CO
CH_3OH
CO
CO_2
Pressure (mbar)
Altitude (km)
Mixing Ratio

(FIGURE 4)

Date: Tue, 13 Nov 2001 10:20:09 -0500
To: Joseph.M.Grebowsky.1@gsfc.nasa.gov
From: Molly McDonough <mmcdonou@pop300.gsfc.nasa.gov>
Subject: AGU Paper

Mr. Grebowsky:

Could you please forward 2 copies of your AGU paper entitled "Meteroric Metaerial- An important component of planetary atmospheres" to Molly McDonough, Code 293. We have received a copy of the DAA but, not a copy of the paper. We need copies of both for archiving.

Thank you,
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