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WET MODEL OF SATURN'S IONO-SPHERE: WATER FROM THE RINGS

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Current theoretical models of Saturn's ionosphere are similar to those of Jupiter's because of the gross similarity of their upper atmospheres. In the case of Jupiter, the theoretical models can be fit reasonably well to ionospheric electron density profiles[1,2] obtained from the Pioneer and Voyager Radio Science occultation experiments.[3,4,5] In contrast, the theoretical models of Saturn's ionosphere are inconsistent with both the ionospheric electron density profiles[6,7] obtained from the Pioneer and Voyager occultation observations[8,9,10] and the large diurnal variation of maximum ionospheric electron density deduced from studies of Saturn lightning discharges.[11,12] We propose a radically different model of Saturn's ionosphere in which water plays a major role as a minor constituent present by virtue of downward diffusion from an external source. Our model Saturn ionosphere is a classical "F2" type layer resulting from the photodissociative production of H^+ from H_2 and rapid chemical loss due to a series of charge exchange reactions with water. A planetwide influx of $\sim 4 \times 10^7$ molecules cm^{-2} s^{-1} of water from the rings is consistent with the observed ionospheric electron densities. An enhanced influx of water occurs at latitudes ($-38°$, $+44°$) magnetically connected to the inner edge of Saturn's B ring[12] which results from an electromagnetic erosion process[13] contributing substantially to the (local) upper atmosphere water content. Present-day influx at these latitudes may be as large as $\sim 2 \times 10^9$ molecules cm^{-2} s^{-1}.

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The essential features of previous theoretical models of Saturn's
ionosphere are summarized below. The major topside ion, H^+, is produced via
photoioniziation of atomic H or photodissociative ionization of H_2. Both H
and H^+ are extremely long-lived above the homopause ($\sim$1100 km altitude; all
altitudes referenced to the 1 bar pressure level) and are removed by downward
diffusion to the vicinity of the homopause where appreciable concentrations of
heavier gases (most importantly methane) are encountered. H^+ ion loss occurs
essentially at the homopause via charge exchange and electron-ion
recombinations, resulting in a peak ion concentration of $\sim$2x10^5 cm^{-3} at $\sim$1200
km altitude. This very simplified picture emerges from a consideration of
many possible chemical reactions and detailed calculations taking into account
the production, diffusion and chemical losses of many atmospheric
constituents.[7]

Figure 1 is a representative summary of the dawn and dusk electron density
(as a function of altitude) profiles obtained from analyses of the Pioneer and
Voyager radio occultation data. In Figure 1, mid-latitude Voyager observations
are compared with a typical model ionospheric electron density profile.
(Figure 2 shows all of the radio science observations.) The measured electron
densities are all approximately an order of magnitude less than that calculated
on the basis of current models. Maximum observed ionospheric densities are
$\sim$1x10^4 cm^{-3}, compared with a calculated maximum density of 2x10^5 cm^{-3} in all
previous models. The peak electron density is observed at between 2000 and
3000 km altitude instead of the calculated 1200 km altitude. The discrepancy
in the altitude of the peak is particularly significant in light of the thermal
structure of the atmosphere. Above $\sim$1600 km, the atmosphere is isothermal with
a temperature of $\sim$800 °K and an atmospheric scale height of

∿400 km. Below that the temperature decreases linearly to ∿125°K at 1000 km altitude where the atmospheric scale height is ∿60 km.[7] The difference in the observed and modeled altitudes of peak ion density is thus many atmospheric scale heights. The large (∿2 order of magnitude) diurnal variation of maximum ionospheric electron density (Figure 3) deduced from analyses of Saturn lightning discharges[11,12] is even more difficult to understand on the basis of current models of Saturn's ionosphere. Without an additional loss mechanism for rapid removal of the long-lived H^+ ions, no diurnal variation is expected since the time constant for loss of H^+ above the homopause is ∿400 hrs, many planetary rotations. This is due to the slow electron recombination reaction $H^+ + e \rightarrow H + h\nu$ @ 2×10^{-12} cm^{-3} s^{-1} and the large diffusion time constant below the 2500 km altitude of the ionospheric peak.

The observational results thus require an additional and effective loss mechanism for H^+ ions, since ion production via photodissociative ionization of H_2 and photoionization of atomic H is well understood. Several possible loss processes have been quantitatively evaluated but none have proven adequate. These include losses due to an increased methane abundance which requires an unreasonably high eddy diffusion coefficient[14], removal of H^+ by vibrationally excited H_2[6], and/or the presence of large vertical drifts.[7,8] The possibility of an interaction between the ionosphere and ring system was recognized as early as 1975[15] and a suggestion involving H^+ ion loss due to the presence of OH radicals from the rings[16,17] was made to explain the low ionospheric densities inferred from the Pioneer observations.

We propose a simple model ionosphere dominated by the removal of H^+ ions via charge exchange with water and subsequent electron recombination. Table 1 summarizes the important chemical reactions and reaction rates; the reactions involving water have been largely drawn from work relating to comet models.[18,19] The important reactions are H^+ ion production via dissociative photoionization of H_2 (A3) and loss via the charge exchange with H_2O (B1). The H_2O^+ ion is very short lived, reacting with H_2 (B2) to produce H_3O^+, which is then readily removed by electron recombination (B3–B5). As in previous models, the H_2^+ ions produced by photoionization of H_2 (A2) are rapidly removed via reaction with H_2 (B6) and charge exchange (B7) and are not a major topside ion. The direct photoionization of atomic hydrogen (A4) does not significantly contribute to the H^+ ion density due to the limited availability of atomic H.

The OH produced in B4 and B5 has a short lifetime ($\tau < 1s$) at the altitude of the peak ion density due to a reaction (B9) with H_2, producing H_2O and H. Thus water acts as a catalyst in the rapid removal of H^+; each H^+ ion removed returns, on average, 3 or 4 H atoms with no associated chemical loss of H_2O. Photodissociation of H_2O (A5) yields OH which in the presence of appreciable H_2 (B9) rapidly reforms H_2O (as would OH from any other source). There is no net loss of water due to charge exchange or photoionization. Thus the density of water above the homopause is given by[20]

$$N_W = N_0 \, EXP(-9Z/H) + FH_0/8D_0 \, [EXP(-Z/H) - EXP(-9Z/H)]$$

where N_0, H_0, and D_0 are the H_2O density, the atmospheric scale height, and molecular diffusion coefficient at a reference (Z=0) altitude, F is the downward flux of H_2O and H is the atmospheric scale height. Assuming that the

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water is readily removed by freezing out just above the homopause or by mixing and chemical loss at the homopause, the density of water throughout the (isothermal) region of interest above 1600 km altitude is just $N_w = (FH/8D_0)EXP(-Z/H)$. Since the diffusion coefficient $D=D_0/N$, where N is the atmospheric density, the downward flux of water results in a water density proportional to the density of the H_2 atmosphere through which it flows under the condition of constant flux, i.e., $N_w/N = cnst = FH/8N_0D_0$.

Under these conditions a classical "Bradbury Layer", analogous to the earth's F2 layer can form at an altitude where the chemical τ_c and diffusive τ_d time constants are comparable.[20,21] For such a layer, the ionospheric peak forms above the level of maximum ion production due to the upward diffusion of ions. Below that altitude the increasing chemical loss (charge transfer with water) effectively removes the major ion, so that most of the ionization anticipated by previous models is lost. The loss coefficient is $\beta = k[H_2O]$ where k is the rate coefficient of reaction B1 in Table 1. The required water concentration at the altitude of the ionospheric peak can be estimated by $N_m \tilde{=} q(h_m)/\beta(h_m)$ where N_m is the maximum electron density at the altitudes (h_m) of the ionospheric peak, and the ion production function q and loss β are evaluated at the altitude of the peak. Taking $h_m \sim 2500$ km, $q \sim 1$ cm^{-3} s^{-1} and $N_m \sim 10^5$ cm^{-3} appropriate to the equatorial ionosphere at noon yields a water concentration of ~ 1250 cm^{-3} at 2500 km altitude. This requires a 4×10^7 cm^{-2} s^{-1} downward flux of water molecules. The maximum densities observed at the dawn and dusk terminators where the EUV production rate is ~ 0.1 of the maximum are ~ 1 order of magnitude lower than the noon density, as expected. Equating the chemical and diffusive time constants ($1/\beta \sim H^2/D$) yields an expected peak altitude between 2,000 and 2500 km.

The variation of electron density with altitude can also be understood using simple approximations to "Bradbury layer" behavior that have been widely used in the study of the earth's F2 layer.[20,21] The ion distribution well above the peak is controlled by ambipolar diffusion, so $N \sim \exp(-Z/H_p)$ where H_p is the plasma scale height. This distribution is indicated in Figure 2 assuming $800°K$ electron and ion temperatures, for which $H_p = 4H$. The H^+ ion density distribution below the peak is controlled by chemical loss, with $N(h) \sim q/\beta(h)$. The ion production, q, varies slowly, so to first approximation we expect $N(h)$ below the peak to increase with height as $\exp(h/H)$, as is illustrated in Figure 2. This reflects the increased loss at greater depths due to greater water concentration. The peak has a width of a few scale heights and occurs at an altitude (2500 km) where the diffusion time constant is a few hours, as does the earth's F2 layer. However, the rapid nighttime decay of the equatorial ionospheric density evidenced in Figure 3 requires a more rapid loss, perhaps a downward drift related to the ionosphere/protonosphere exchange described below.

Figure 3 shows an ionospheric density minimum just after dusk and a steady increase in near-equatorial ionospheric density throughout the night. We attribute this phenomenon to a diurnal ionosphere – protonosphere exchange similar to that proposed for the earth's nighttime F2 layer.[20] During daylight hours, photoproduction in Saturn's ionosphere pumps protons outward along field lines to great heights. At night, a relaxation back towards the ionosphere occurs to replenish the nighttime ionospheric density. If we assume a uniform density (n) column of protons at rest at dusk, under constant downward acceleration (a) the proton influx Φ_p = nat increases linearly with time. Supposing the nighttime maximum ionospheric density occurs at high

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altitudes where loss is negligible, the increase in the maximum density with time is related to the influx by $N_m(t) = \int \Phi_p/H \, dt = (na/2H)t^2$. This increase is indicated on figure 2 by the dashed line, which fits the nighttime observations well with $na/2H \sim 2 \times 10^{-5}$ cm^{-3} s^2, corresponding to a proton influx increasing to a maximum of $\sim 2 \times 10^7$ cm^{-2} s^{-1} just before dawn. The equivalent ion production rate at dawn due to this exchange ($\sim .2$ cm^{-3} s^{-1}) is comparable to or greater than the EUV production. Thus, we envision a very dynamic ionosphere, characterized by ion production and transport into a vast protonosphere reservoir by day coupled with relaxation and return by night.

The increased ion losses at latitudes magnetically connected to the inner edge of Saturn's B-ring[12] can be explained by an increased local water influx of as much as 2×10^9 molecules cm^{-2} s^{-1}. The erosional process described by Northrop and Hill[13] provides a mechanism whereby sub-micron charged particles (collection of water molecules), unstable in Saturn's ring plane, precipitate along magnetic field lines into the ionosphere. If these particles are deposited directly at the level of the ionospheric peak (~ 2500 km) the required flux is less than 2×10^9 cm^{-2} s^{-1}. At this maximum flux, the ring mass loss is $\sim 6 \times 10^{-14}$ g cm^{-2} s^{-1}. The large dayside extinction was observed at latitudes of -37 to -39°;[12] the maximum southerly extent of the Voyager 1 spacecraft. Averaged over $\sim 4^\circ$ latitude extent of the presumed source region in the rings ($1.52 - 1.62$ R_S), the total ring mass loss is $\sim 1.2 \times 10^6$ g s^{-1} or 1.6×10^{12} g yr^{-1}, comparable to the estimated total mass influx from interplanetary meteoroids.[22] At this rate, 60 g cm^{-2} of material can be removed from the source region in $\sim 30 \times 10^6$ years, suggesting that this process may be responsible for the large decrease in optical depth of the rings inward of 1.62 R_S.

The radio science observations of Saturn's ionosphere and the inferences drawn from studies of Saturn lightning can be largely understood on the basis of a new model of Saturn's ionosphere in which water plays a major role. The planetwide influx of water (4×10^7 molecules cm^{-2} s^{-1}) required is approximately an order of magnitude greater than that expected from photosputtering of the rings.[23] Morfill et al.[22] estimate that micrometeoride impacts on Saturn's icy rings produce between .75 to 6×10^6 g/sec of vapor, ~10% of which is deposited on Saturn's atmosphere. This results in an estimated influx of between 0.6 and 4.5×10^7 molecules cm^{-2} s^{-1}. In spite of the uncertainties in this estimate,[22] micrometeoride bombardment of Saturn's rings is an attractive source mechanism because it provides an explanation of the vast differences between Jupiter's and Saturn's ionospheres. An atmospheric influx of water has been previously suggested to explain features of ultraviolet spectra of Saturn[24] and Voyager infrared spectra of Titan.[25] The increased influx of water at latitudes magnetically connected to the inner edge of the B-ring is not inconsistent with expected erosion rates. Finally, we infer a diurnal ionosphere – protonosphere exchange in which protons pumped into the protonosphere by day return by night.

TABLE 1

CHEMICAL REACTIONS IN SATURN'S IONOSPHERE: H_2, H_2O, H

ION PRODUCTION		RATE*	REF.
A1	$H_2 + h\nu \rightarrow H + H$	4.5×10^{-10} s^{-1}	[7]
A2	$H_2^+ + e$	5.4×10^{-10}	[7]
A3	$H + H^+ + e$	9.5×10^{-11}	[7]
A4	$H + h\nu \rightarrow H^+ + e$	7.3×10^{-10}	[7]
A5	$H_2O + h\nu \rightarrow OH + H$	1.0×10^{-7}	[19]
	$OH^+ + H + e$	5.5×10^{-10}	[19]
	$H_2 + O(^1D)$	1.4×10^{-8}	[19]
	$H_2O^+ + e$	3.3×10^{-9}	[19]
	$H_2 + O^+ + e$	5.8×10^{-11}	[19]
	$H^+ + OH + e$	1.3×10^{-10}	[19]

CHARGE EXCHANGE/RECOMBINATION			
B1	$H_2O + H^+ \rightarrow H_2O^+ + H$	8.2×10^{-9} cm^3 s^{-1}	[18]
B2	$H_2O^+ + H_2 \rightarrow H_3O^+ + H$	6.1×10^{-10}	[18]
B3	$H_3O^+ + e \rightarrow H_2O + H$	4.3×10^{-7}	[18]
B4	$H_2 + OH$	4.3×10^{-7}	[18]
B5	$H + H + OH$	4.3×10^{-7}	[18]
B6	$H_2^+ + H_2 \rightarrow H_3^+ + H$	2.0×10^{-9}	[7]
B7	$H_3^+ + e \rightarrow H_2 + H$	1.0×10^{-7}	[7]
B8	$H^+ + e \rightarrow H + \nu$	2.0×10^{-12}	[7]
B9	$OH + H_2 \rightarrow H_2O + H$	3.6×10^{-11}	[18]

* Photoionization rates normalized to 10 A.U. by $1/R^2$ scaling of reference values

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Figure Captions

Figure 1 Model calculations of mid-latitude ionospheric electron density as a
function of altitude (EUV) and Voyager observations of Saturn's
ionsphere at mid-latitudes, adapted from Atreya et al.[7]

Figure 2 Ionospheric electron density profiles deduced from Voyager 1 and 2
and Pioneer radio science data, adapted from Atreya et al.[7]
Altitude is referenced to the 1 bar pressure level.

Figure 3 Diurnal variation of maximum ionospheric electron density inferred
from Voyager observations of lightning-generated radio emissions in
Saturn's atmosphere, adapted from Kaiser et al.[12] The dashed line
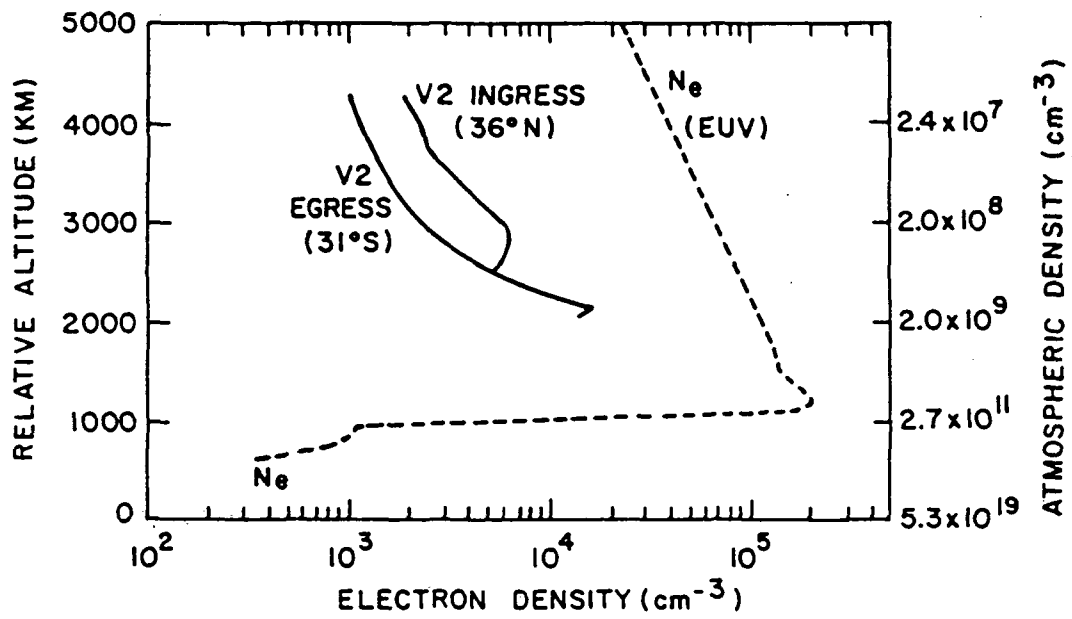indicates a model fit to the nighttime increase of maximum
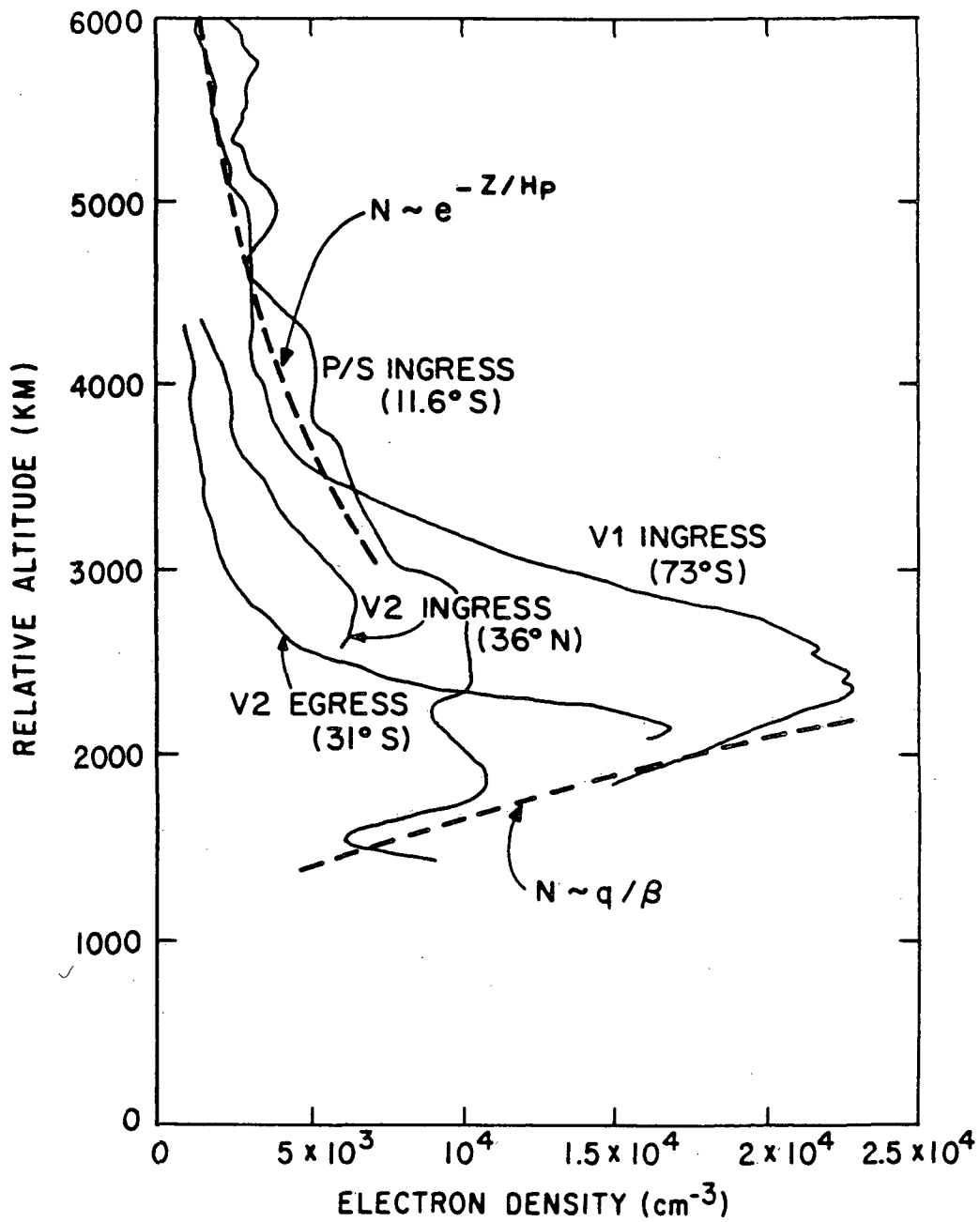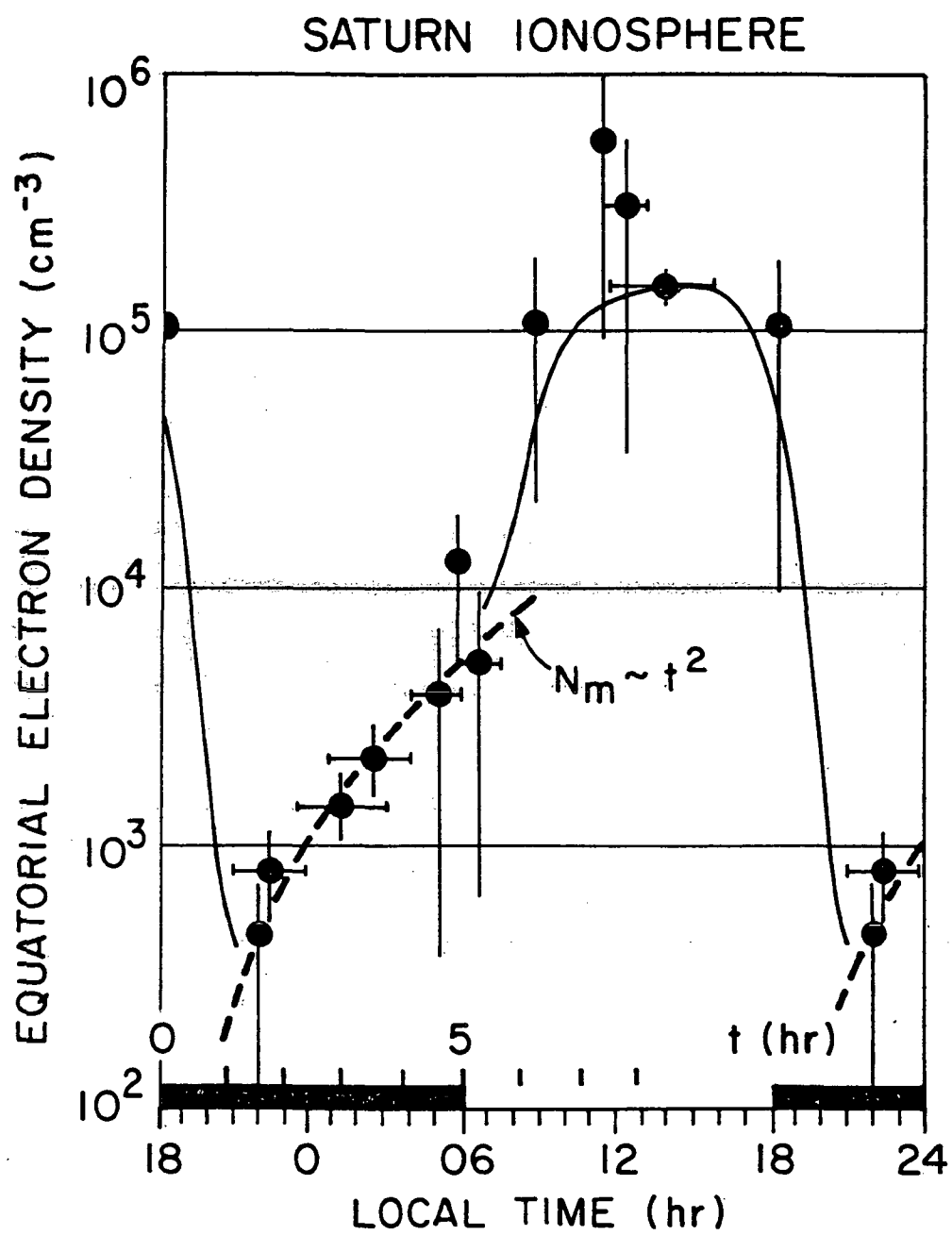ionospheric electron density.

Figure 1

Figure 2

SATURN IONOSPHERE

Figure 3

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16. Abstract

    Current theoretical models of Saturn's ionosphere are difficult to reconcile with the ionospheric electron density profiles obtained from the Pioneer and Voyager radio occultation observations and the large diurnal variation of maximum ionospheric electron density deduced from studies of Saturn lightning discharges. We propose a new model of Saturn's ionosphere in which water plays a major role as a minor consitutent present by virtue of downward diffusion from an external source. Our model Saturn ionosphere is a classical "F2" type layer resulting from the photodissociative production of H^+ from H_2 and rapid chemical loss due to a series of charge exchange reactions with water. A planetwide influx of about 4×10^7 molecules $cm^{-2} s^{-1}$ of water from the rings is consistent withthe observed ionospheric electron densities and estimates of influx due to micrometeoride bombardment of the rings. An enhanced influx of water occurs at latitudes $(-38°, +44°)$ magnetically connected to the inner edge of Saturn's B ring which results from an electromagnetic erosion process contributing substantially to the (local) upper atmosphere water content. Present-day influx at these latitudes may be as large as close to 2×10^9 molecules $cm^{-2} s^{-1}$.

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