

CARBON MONOXIDE IN JUPITER'S UPPER ATMOSPHERE: AN EXTRAPLANETARY SOURCE

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ABSTRACT

Ablation of meteoroidal material in Jupiter's atmosphere may provide substantial quantities of H₂O. Subsequent photochemistry can convert H₂O and CH₄ to CO and H₂. The associated source of CO could account for the observations by Beer, Larson, Fink, and Treffers, and Beer and Taylor, and would explain the relatively low rotational temperatures inferred by Beer and Taylor. Meteoritic debris might also provide spectroscopically detectable concentrations of SiO.

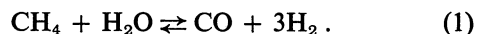
Subject headings: planets: abundances — planets: atmospheres — planets: Jupiter

I. INTRODUCTION

Beer's (1975) discovery of CO in the highly reducing environment of Jupiter's atmosphere poses a number of interesting problems. Resolution of these problems may be expected to provide important new information on the role of dynamical transport in the lower atmosphere and may offer us, in addition, an opportunity to study the influence of meteoritic debris on the composition and chemistry of the stratosphere.

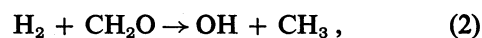
Beer's (1975) initial paper reports identification of a number of lines from *P*(17) to *R*(9) in the fundamental vibration-rotation band near 4.7 μm, and attribution of these lines to CO in Jupiter's atmosphere is confirmed by the more recent observations of Larson, Fink, and Treffers (1978) and Beer and Taylor (1978). The column density of CO appears to lie between 4 × 10¹⁷ molecules cm⁻² and 8 × 10¹⁷ molecules cm⁻². Beer and Taylor (1978) argue that the rotational temperature is relatively low, 125 ± 25 K, and conclude on this basis that CO is most probably concentrated in the stratosphere. They believe that their data would not allow mixing ratios for tropospheric CO much greater than 10⁻¹⁰. This conclusion would impose powerful constraints on the range of possible sources for Jovian CO. It would appear to exclude, for example, a dominant contribution to the observed CO due to upward transport of the gas from deep regions of the troposphere as suggested by Prinn and Barshay (1977). Further data are obviously desirable in order to clarify this issue. Larson, Fink, and Treffers (1978) regard the matter as still subject to question.

Thermochemical considerations (Barshay and Lewis 1978; Prinn and Barshay 1977) suggest that CH₄ and CO should be the most abundant carbon compounds at temperatures above 920 K, with an equilibrium described by



Exchange between various forms of oxidized carbon,

notably CO, CO₂, CH₂O, and CHO, is rapid. Conversion of oxidized carbon to reduced carbon, which proceeds mainly through



is somewhat slower (Prinn and Barshay 1977). The concentration of CO in the upper atmosphere is vanishingly small in equilibrium. The rate for reaction (2) is sufficiently slow so that vertical mixing can supply a significant disequilibrium concentration of CO, as pointed out by Prinn and Barshay (1977). A uniform mixing ratio of 10⁻⁹, which could account for the column densities observed in the visible atmosphere, would require an effective vertical diffusion coefficient of about 2 × 10⁸ cm² s⁻¹ in regions of the atmosphere where the temperature is about 1000 K. The diffusion coefficient should not exceed about 2 × 10⁶ cm² s⁻¹ if CO is concentrated in the stratosphere, as suggested by Beer and Taylor (1978).

We shall describe here another possible source for the observed CO. Our approach is in some sense complementary, yet antithetical, to that given by Prinn and Barshay (1977). We consider a possible source of oxygen at high altitudes associated with the ablation of meteoroidal material. There are reasons to believe that the concentration of micrometeoroids may be reasonably high in the vicinity of Jupiter (Humes 1976). Analogy either with meteorites observed in the upper atmosphere of the Earth (Brownlee *et al.* 1975; Brownlee, Tomandl, and Olszewski 1977) or with comets (Donn *et al.* 1976) suggests that these objects should contain appreciable amounts of oxygen. Most of this oxygen should be transformed rapidly to H₂O on entering the reducing environment of Jupiter. Subsequent photochemistry can provide a significant source of stratospheric CO which will be transported downward and converted ultimately to CH₄ and H₂O in the deep atmosphere.

The model is described in § II. Results are presented and discussed in § III. Some implications of a meteoroidal source for Jovian CO are considered in § IV.

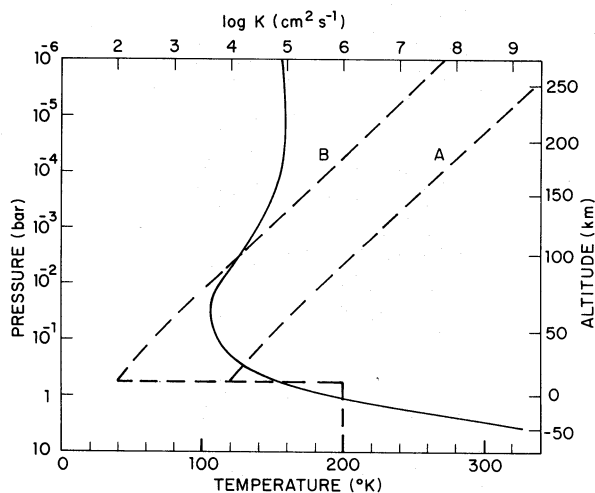


FIG. 1a

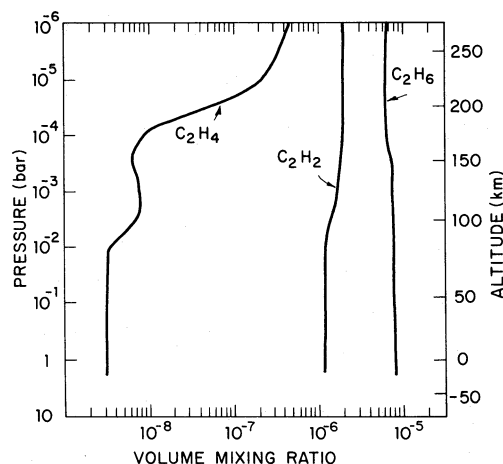


FIG. 1b

FIG. 1a.—Jovian temperature profile. The assumed temperature profile (*solid curve*) for the visible atmosphere of Jupiter is plotted as a function of both altitude and pressure. The vertical diffusion coefficients K (*dashed lines*) are shown for models A and B (see text).

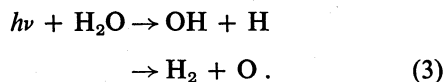
FIG. 1b.—Volume mixing ratios of Jovian hydrocarbons. The mixing ratios of C_2H_6 , C_2H_4 , and C_2H_2 are taken from the work of Strobel (1974). The mixing ratio of CH_4 is assumed to be 10^{-3} by volume.

II. PHOTOCHEMICAL MODEL

We adopt a model for the upper atmosphere taken from Wallace, Prather, and Belton (1974), with extension to the lower atmosphere following Divine (1970) and Trafton and Stone (1974). The model is consistent with recent data (Wallace 1976) and is shown in Figure 1. Zero on the altitude scale is located arbitrarily at a pressure of 1 bar. Mixing ratios of CH_4 and He are assumed constant with altitude, with values set equal to 10^{-3} (Prinn and Owen 1976) and 10^{-1} (Carlson and Judge 1976), respectively. Abundances of C_2H_6 , C_2H_4 , and C_2H_2 were taken from Strobel (1974) and are also reproduced in Figure 1.

We consider a downward flux of H_2O of magnitude $\phi_{H_2O^\infty}$ (molecules $cm^{-2} s^{-1}$) at a level in the atmosphere corresponding to a pressure of 10^{-6} bar. The data reported by Humes (1976) suggest an influx of material to Jupiter of magnitude $3.4 \times 10^{-13} g cm^{-2} s^{-1}$. If the composition of Jovian meteoroids is similar to that of carbonaceous chondrites, as appears to hold for meteorites observed in the terrestrial atmosphere (Brownlee *et al.* 1975), we might anticipate an influx of H_2O to Jupiter as large as $5 \times 10^9 cm^{-2} s^{-1}$. Given the obvious uncertainty attached to this result, we propose to investigate a range of values for $\phi_{H_2O^\infty}$.

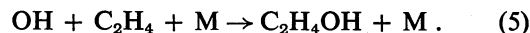
Water in Jupiter's stratosphere is removed efficiently by photolysis,



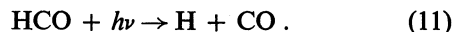
Production of CO is initiated by reactions such as



and



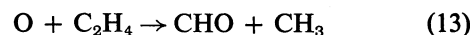
The primary products in reactions (4) and (5) have been observed mass spectrometrically (Kanofsky *et al.* 1974; Morris, Stedman, and Niki 1971), but details of their structure and subsequent chemistry are unknown. We assume that they isomerize rapidly to their most stable forms, CH_2CO and CH_3CHOH , and that they are removed by the reaction sequence



Reactions of atomic oxygen with hydrocarbons

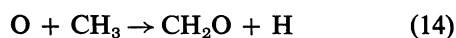


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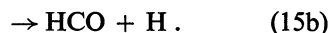
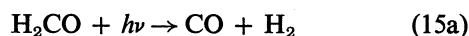


provide a somewhat smaller source of CO. Experimental studies of reactions (12) suggest that both paths are important (Kanofsky *et al.* 1974; Williamson and Bayes 1969; Jones and Bayes 1973). We assume

rapid collisional decomposition of the C_2HO fragment to CO and CH. Carbon monoxide is formed also by the fast radical-radical reaction

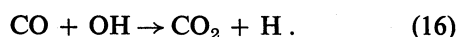


followed by photolysis of formaldehyde

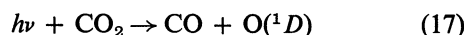


The detailed path for production of CO is somewhat uncertain. The initial reactions in which the C—O bond is formed, however, are relatively well defined, and it would seem plausible that the production rate for CO should be approximately equal to the combined rates for reactions (4), (5), (12), (13), and (14).

Carbon monoxide is removed chemically by reaction with OH:



The loss is transitory, since CO_2 is rapidly photolyzed to CO



at wavelengths below 1700 Å. Transport to the lower atmosphere represents the ultimate sinks for both CO and CO_2 . A summary of the chemical model is given in Table 1.

Vertical transport of CO, CO_2 , and H_2O is described using the one-dimensional diffusion approximation with flux (molecules $cm^{-2} s^{-1}$), ϕ_i , given by

$$\phi_i(z) = -K(z)N(z) \frac{\partial f_i}{\partial z}(z), \quad (18)$$

where K is an effective diffusion coefficient ($cm^2 s^{-1}$), $N(z)$ is the total number density (cm^{-3}), and f_i is the volume mixing ratio of either CO, CO_2 , or H_2O . The number density, n_i , of species i satisfies a continuity equation of the form

$$\frac{\partial \phi_i}{\partial z} = P_i - L_i, \quad (19)$$

where P_i and L_i denote volume production and loss rates, respectively. Production and loss rates depend parametrically on the density of other short-lived species, j , including H, CH, 1CH_2 , CH_3 , C_2H_3 , C_2H_5 , O, $O(^1D)$, OH, HCO, H_2CO , CH_2CO , CH_3CHO , and CH_3CHOH . Concentrations of these species are obtained by solving a set of chemical equilibrium equations,

$$P_j = L_j, \quad (20)$$

with the reaction scheme and reaction rates given in Table 1.

The partial pressure of H_2O is constrained not to exceed the equilibrium vapor pressure. In practice, concentrations of H_2O were obtained by solving

equations (18) and (19) with specified downward flux $\phi_{H_2O}^\infty$, at a background pressure of 10^{-6} bars, and with the number density set equal to 1.5×10^2 molecules cm^{-3} at a background pressure of 30 mbars. Results are relatively insensitive to the precise location of the lower boundary. Maximum conversion of H_2O and CH_4 to CO and H_2 occurs between pressure levels of 0.5 and 10 mbars, with details of the conversion profile influenced by the choice of $K(z)$.

The mixing ratio of CO_2 is assumed to be vanishingly small in the troposphere, at pressure levels above 0.5 bars. The lower boundary condition on CO allows a range of values for the appropriate mixing ratio, from 10^{-10} to 10^{-9} , as suggested by the analysis of Prinn and Barshay (1977). The fluxes of CO and CO_2 at the upper boundary, 10^{-6} bars, are set equal to the net, height-integrated sources above this level.

III. RESULTS

It is usually assumed that minimum values of the eddy diffusion coefficient should occur near the tropopause. Estimates for $K(z)$ near the tropopause range from $10^2 cm^2 s^{-1}$ (Sagan and Salpeter 1976) to about $2 \times 10^4 cm^2 s^{-1}$ (Strobel 1973). Observations of $L\alpha$ by *Pioneer 10* (Carlson and Judge 1974) and *Copernicus* (Jenkins, Wallace, and Drake 1974) indicate values for K of order $10^8 cm^2 s^{-1}$ near the tropopause (Carlson and Judge 1974; Strobel 1974). Estimates for K in the lower troposphere range as high as $10^7 cm^2 s^{-1}$ (Gierasch 1976; Sagan and Salpeter 1976; Prinn and Barshay 1977).

We assume here that the variation of K with altitude above the tropopause may be represented by

$$K(z) = K_0[N_0/N(z)]^\alpha, \quad (21)$$

where N_0 and $N(z)$ are total number densities at the tropopause and altitude z , respectively. We consider a range of values for K_0 with α equal to either 1 or 0.5 (Lindzen 1971; Strobel 1974). Results are insensitive to the choice of $K(z)$ in the troposphere except insofar as deep tropospheric mixing may influence the supply of CO from the high-temperature region below.

Results for two models, characterized by values of $K_0 = 10^4$, $\alpha = 1$, $\phi_{H_2O}^\infty = 2 \times 10^9$ (model A) and $K_0 = 10^2$, $\alpha = 1$, $\phi_{H_2O}^\infty = 1 \times 10^7$ (model B), respectively, are given in Figures 2 and 3. The column densities of CO are similar for both models, about $4 \times 10^{17} cm^{-2}$ above a level where the pressure is approximately 2 bars. Model B allows for a longer residence time of oxygen compounds above the tropopause. It therefore permits relatively complete conversion of H_2O to CO, in contrast to model A in which significant concentrations of H_2O are removed by condensation. This point is illustrated by the flux profiles given in Figures 2b and 3b. Condensation of H_2O is important for values of K_0 greater than $3 \times 10^3 cm^2 s^{-1}$.

Concentrations of photochemical intermediates are given, for models A and B, in Figures 4 and 5. The most abundant oxygen compounds, in addition to

TABLE 1
PHOTOCHEMICAL RATE DATA

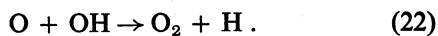
No.	Reaction	Rate Expression†	References and Notes
1.....	${}^1\text{CH}_2 + \text{H}_2 \rightarrow \text{CH}_4^* \rightarrow \text{CH}_3 + \text{H}$	7.0-12	Braun, Bass, and Pilling 1970
2.....	${}^1\text{CH}_2 + \text{CH}_4 \rightarrow \text{C}_2\text{H}_6^* \rightarrow \text{CH}_3 + \text{CH}_3$	1.9-12	Braun, Bass, and Pilling 1970
3.....	$\text{CH} + \text{CH}_4 \rightarrow \text{C}_2\text{H}_4 + \text{H}$	2.5-12	Braun, McNesby, and Bass 1967
4.....	$\text{CH} + \text{H}_2 \xrightarrow{\text{M}} \text{CH}_3$	1.0-12	Braun, McNesby, and Bass 1967
5.....	$\text{H} + \text{CH}_3 + \text{M} \rightarrow \text{CH}_4 + \text{M}$	$k_2 = 3.32-10$ $k_3 = 7.09-29$	Cheng and Yeh 1977
6.....	$\text{CH}_3 + \text{CH}_3 + \text{M} \rightarrow \text{C}_2\text{H}_6 + \text{M}$	$k_2 = 4.04-11$ $k_3 = 4.76-27$ $k^{-1} = k_2^{-1} + (k_3\text{M})^{-1}$	Fitted to the data of van den Berg <i>et al.</i> 1968, and the RRKM calculations of Waage and Rabinovitch 1971
7.....	$\text{H} + \text{H} + \text{M} \rightarrow \text{H}_2 + \text{M}$	$8.0-33 \left(\frac{T}{300}\right)^{-0.65}$	Fit to the data of Trainor, Ham, and Kaufman 1973
8.....	$\text{H} + \text{C}_2\text{H}_2 + \text{M} \rightarrow \text{C}_2\text{H}_3 + \text{M}$	$k_2 = 9.2-12 \exp\left(\frac{-1213}{T}\right)$ $k_3 = 1.1-30 \exp\left(\frac{-770}{T}\right)$ $k^{-1} = k_2^{-1} + (k_3\text{M})^{-1}$	Fit to the data of Payne and Stief 1976
9.....	$\text{H} + \text{C}_2\text{H}_3 \rightarrow \text{C}_2\text{H}_2 + \text{H}_2$	7.0-12	Benson and Haugen 1967; Volpi and Zocchi 1966
10.....	$\text{H}_2 + \text{C}_2\text{H}_3 \rightarrow \text{C}_2\text{H}_4 + \text{H}$	$5.0-12 \exp\left(\frac{-3000}{T}\right)$	Benson and Haugen 1967
11.....	$\text{H} + \text{C}_2\text{H}_4 + \text{M} \rightarrow \text{C}_2\text{H}_5 + \text{M}$	$k_a = 1.71-11 \exp\left(\frac{-755}{T}\right)$ $k_b = 1.02 + 21$ $k^{-1} = k_a^{-1} + k_b\text{M}^{-1/2}$	Kurylo <i>et al.</i> 1970 Penzhorn and Darwent 1971
12.....	$\text{H} + \text{C}_2\text{H}_5 \rightarrow \text{C}_2\text{H}_6^* \rightarrow \text{CH}_3 + \text{CH}_3$	6.0-11	Kurylo <i>et al.</i> 1970
13.....	$\text{OH} + \text{CH}_4 \rightarrow \text{CH}_3 + \text{H}_2\text{O}$	$2.36-12 \exp\left(\frac{-1710}{T}\right)$	Davis <i>et al.</i> 1974
14.....	$\text{OH} + \text{C}_2\text{H}_6 \rightarrow \text{C}_2\text{H}_5 + \text{H}_2\text{O}$	$1.86-11 \exp\left(\frac{-1232}{T}\right)$	Greiner 1970
15.....	$\text{OH} + \text{C}_2\text{H}_4 + \text{M} \rightarrow \text{C}_2\text{H}_4\text{OH} + \text{M}$	$k_2 = 1.37-12 \exp\left(\frac{388}{T}\right)$ $k_3 = 5.81-31 \exp\left(\frac{1258}{T}\right)$ $k^{-1} = k_2^{-1} + (k_3\text{M})^{-1}$	Fit to the data of Davis <i>et al.</i> 1975, and Atkinson <i>et al.</i> 1977
16.....	$\text{OH} + \text{C}_2\text{H}_2 \rightarrow \text{C}_2\text{H}_2\text{O} + \text{H}$	1.65-13	Davis <i>et al.</i> 1975
17.....	$\text{O} + \text{C}_2\text{H}_4 \rightarrow \text{CHO} + \text{CH}_3$ $\rightarrow \text{C}_2\text{H}_2\text{O} + \text{H}_2$	$5.15-12 \exp\left(\frac{-569}{T}\right)$ $2.71-13 \exp\left(\frac{-569}{T}\right)$	Davis <i>et al.</i> 1972; Kurylo and Huie, 1973; Pruss <i>et al.</i> 1974
18.....	$\text{O} + \text{C}_2\text{H}_2 \rightarrow \text{C}_2\text{HO} + \text{H}$ $\rightarrow \text{CH}_2 + \text{CO}$	$6.0-12 \exp\left(\frac{-1370}{T}\right)$ $6.0-12 \exp\left(\frac{-1370}{T}\right)$	Fitted to the data of Westenberg and deHaas 1969, below 300 K.
19.....	$\text{O} + \text{CH}_3 \rightarrow \text{CH}_2\text{O} + \text{H}$	1.0-10	Morris and Niki 1972; Washida and Bayes 1976
20.....	$\text{OH} + \text{H}_2 \rightarrow \text{H}_2\text{O} + \text{H}$	$1.8-11 \exp\left(\frac{-2334}{T}\right)$	Smith and Zellner 1974
21.....	$\text{OH} + \text{CO} \rightarrow \text{CO}_2 + \text{H}$	$2.1-13 \exp\left(\frac{-115}{T}\right)$	Greiner 1969
22.....	$\text{OH} + \text{OH} \rightarrow \text{H}_2\text{O} + \text{O}$	1.7-12	Clyne and Down 1974
23.....	$\text{O} + \text{CO} + \text{M} \rightarrow \text{CO}_2 + \text{M}$	2.2-36	Stuhl and Niki 1971; Slinger <i>et al.</i> 1972
24.....	$\text{O}({}^1D) + \text{H}_2 \rightarrow \text{OH} + \text{H}$	1.0-10	Davidson <i>et al.</i> 1977
25.....	$\text{O}({}^1D) + \text{CH}_4 \rightarrow \text{OH} + \text{CH}_3$ $\rightarrow \text{H}_2 + \text{H}_2\text{CO}$	1.3-10 1.3-11	Davidson <i>et al.</i> 1977 Hampson and Garvin 1975
26.....	$\text{O}({}^1D) + \text{H}_2\text{O} \rightarrow \text{OH} + \text{OH}$	2.3-10	Streit <i>et al.</i> 1976
27.....	$\text{OH} + \text{H}_2\text{CO} \rightarrow \text{H}_2\text{O} + \text{HCO}$	1.4-11	Morris and Niki 1971
28.....	$\text{H} + \text{H}_2\text{CO} \rightarrow \text{H}_2 + \text{HCO}$	$2.2-11 \exp\left(\frac{-1890}{T}\right)$	Westenberg and de Haas 1972
29.....	$\text{H} + \text{HCO} \rightarrow \text{H}_2 + \text{CO}$	5.0-11	Estimate
30.....	$\text{H} + \text{CH}_3\text{CHOH} \rightarrow \text{H}_2 + \text{CH}_3\text{CHO}$	1.0-13	Estimate
31.....	$\text{CH}_3 + \text{CH}_3\text{CHOH} \rightarrow \text{CH}_4 + \text{CH}_3\text{CHO}$	1.0-13	Estimate

TABLE 1—Continued

No.	PHOTOLYTIC PROCESS	PROBABILITY		REFERENCES AND NOTES
		At $\lambda\alpha$	At all other λ	
32.....	$\text{CH}_4 + h\nu \rightarrow \text{}^1\text{CH}_2 + \text{H}_2$ $\rightarrow \text{CH} + \text{H} + \text{H}_2$	0.92 0.08	1.0 0.0	Sun and Weissler 1954; Mount <i>et al.</i> 1977; Rebert and Ausloos 1972. See discussion by Strobel 1973.
33.....	$\text{C}_2\text{H}_6 + h\nu \rightarrow \text{C}_2\text{H}_2 + 2\text{H}_2$ $\rightarrow \text{C}_2\text{H}_4 + \text{H}_2$ $\rightarrow \text{C}_2\text{H}_4 + 2\text{H}$ $\rightarrow \text{CH}_4 + \text{}^1\text{CH}_2$ $\rightarrow \text{CH}_3 + \text{CH}_3$	0.26 0.14 0.30 0.25 0.05	0.38 0.47 0.13 0.02 ...	Schoen 1962; Okabe and Becker 1963; Hampson and McNesby 1965; Akimoto <i>et al.</i> 1965; Lias <i>et al.</i> 1870
34.....	$\text{C}_2\text{H}_4 + h\nu \rightarrow \text{C}_2\text{H}_2 + 2\text{H}$ $\rightarrow \text{C}_2\text{H}_2 + \text{H}_2$		0.5 0.5	Zelikoff and Watanabe 1953; Sauer and Dorfman 1961; Back and Griffiths 1967
35.....	$\text{C}_2\text{H}_2 + h\nu \rightarrow \text{C}_2\text{H}_2^*$		1.0	Nakayama and Watanabe 1964; Takita <i>et al.</i> 1968, 1969; Okabe 1975. See discussion by Strobel 1973.
36.....	$\text{CO}_2 + h\nu \rightarrow \text{CO} + \text{O}(\text{}^1D)$ $\rightarrow \text{CO} + \text{O}(\text{}^3P)$		1.0 1.0	$\lambda < 1700 \text{ \AA}$ Nakata <i>et al.</i> 1965 $\lambda > 1700 \text{ \AA}$ Shemansky 1972; Slanger and Black 1978
37.....	$\text{H}_2\text{O} + h\nu \rightarrow \text{H} + \text{OH}$ $\rightarrow \text{H}_2 + \text{O}$		0.75 0.25	$\lambda < 1100 \text{ \AA}$ Watanabe and Jursa 1964 $\lambda > 1100 \text{ \AA}$ Watanabe and Zelikoff 1953; McNesby <i>et al.</i> 1962
38.....	$\text{H}_2\text{CO} + h\nu \rightarrow \text{H}_2 + \text{CO}$ $\rightarrow \text{H} + \text{HCO}$		0.63 0.37	McQuigg and Calvert 1969 Calvert <i>et al.</i> 1972
39.....	$\text{HCO} + h\nu \rightarrow \text{H} + \text{CO}$	$1.0 \times J(\text{H}_2\text{CO})$		Estimated
40.....	$\text{CH}_2\text{CO} + h\nu \rightarrow \text{CH}_2 + \text{CO}$		1.0	Knox <i>et al.</i> 1952; Braun <i>et al.</i> 1970
41.....	$\text{CH}_3\text{CHO} + h\nu \rightarrow \text{CH}_4 + \text{CO}$ $\rightarrow \text{CH}_3 + \text{HCO}$		0.3 0.7	Calvert and Pitts 1966 Cundall and Davies 1967
Solar flux.....		$900 \leq \lambda \leq 1350 \text{ \AA}$ $1400 \leq \lambda \leq 1850 \text{ \AA}$ $1900 \leq \lambda \leq 8000 \text{ \AA}$		Hinteregger 1977 Heroux and Swirbalus 1976 Vernazza <i>et al.</i> 1976

† Units are $\text{cm}^3 \text{ molecule}^{-1} \text{ s}^{-1}$ for bimolecular reactions, $\text{cm}^6 \text{ molecule}^{-2} \text{ s}^{-1}$ for termolecular reactions. The notation 1.0–11 is intended to be read as 1.0×10^{-11} .

H_2O and CO , are CO_2 , CH_2CO , CH_3CHOH , H_2CO , O , OH , and HCO . The large concentrations of O and OH in model A will form significant quantities of molecular oxygen by the reaction



We expect maximum concentrations of O_2 of order

$10^5 \text{ molecules cm}^{-3}$ and $10^2 \text{ molecules cm}^{-3}$ to occur about 0.5 mbar, for models A and B, respectively. Column densities of these species are too small, however, to permit detection with present spectroscopic techniques.

We may note that fluxes of H_2O required to account for observed levels of CO are relatively modest. The

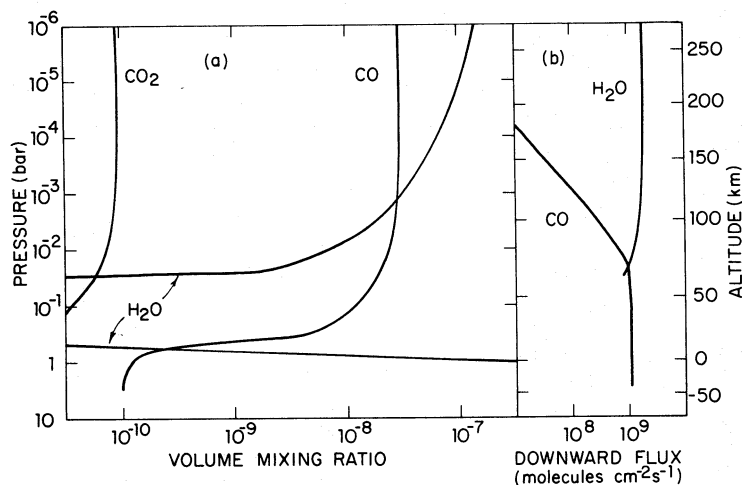


FIG. 2a.—Profiles of H_2O , CO , and CO_2 for model A. The mixing ratios of the primary oxygen-containing species are shown for model A where the downward flux of H_2O is $2 \times 10^9 \text{ cm}^{-2} \text{ s}^{-1}$, and the vertical diffusion coefficient is given in Fig. 1a. (b) Downward flux of H_2O and CO for model A. The abrupt cutoff of the H_2O flux near 60 km is due to the formation of H_2O ice. Water vapor is assumed to be saturated below this level.

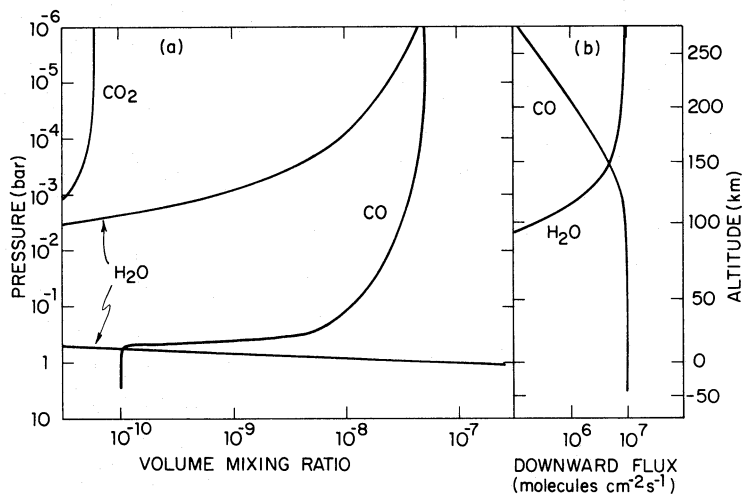


FIG. 3.—(a) Profiles of H_2O , CO , and CO_2 for model B. In this case model B has the vertical diffusion coefficient given in Fig. 1a and a downward H_2O flux of $1 \times 10^7 \text{ cm}^{-2} \text{ s}^{-1}$. (b) Downward flux of H_2O and CO for model B.

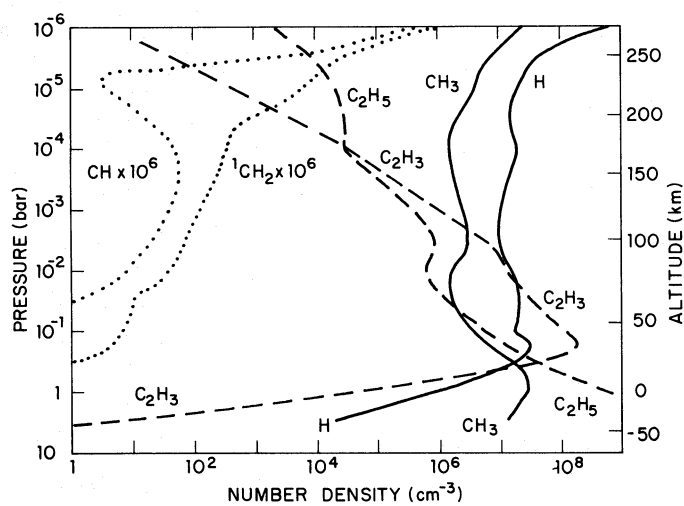


FIG. 4a

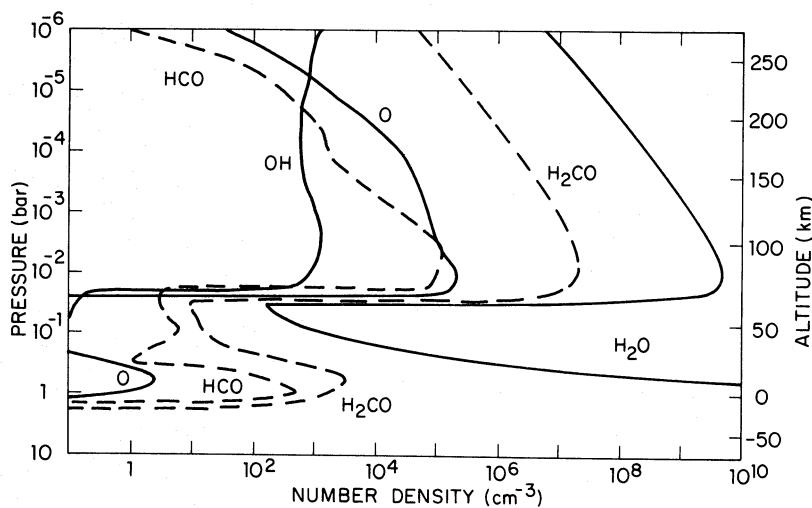


FIG. 4b

FIG. 4.—Profiles of chemical constituents for model A. Fig. 4a shows those radicals which do not contain oxygen. Figs. 4b and 4c give concentrations for most of the oxygen-bearing species in the chemical model.

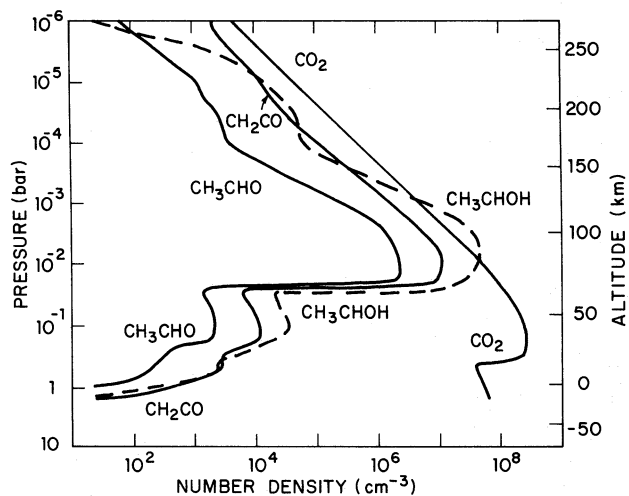


FIG. 4c

FIG. 4.—Continued

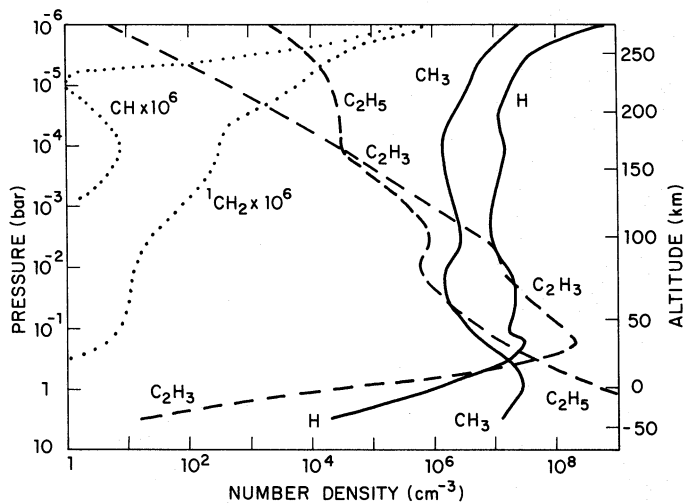


FIG. 5a

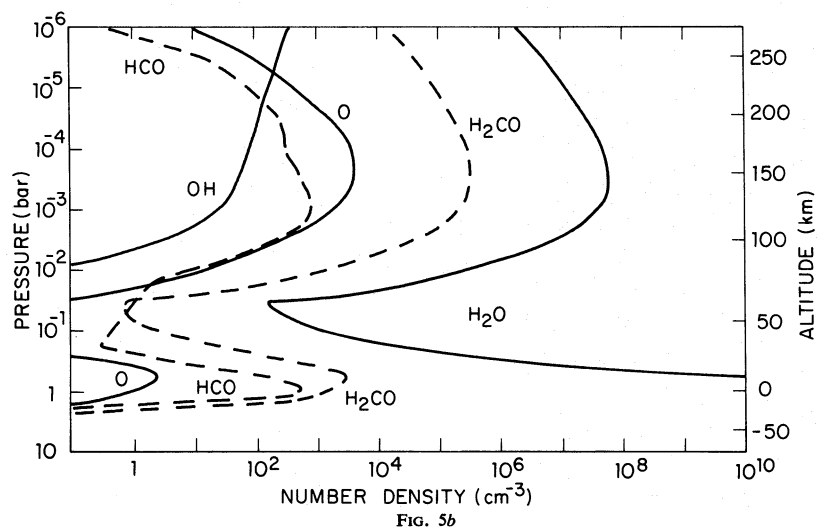


FIG. 5b

FIG. 5.—Profiles of chemical constituents for model B, as in Fig. 4

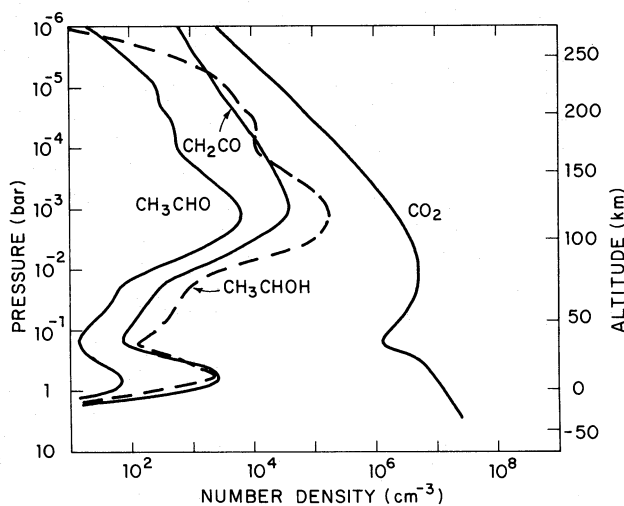


FIG. 5c

FIG. 5.—Continued

flux in model A represents approximately 40% of the flux deduced earlier based on meteoritic inputs given by Humes (1976). The more stable stratosphere envisaged in model B would allow H_2O fluxes smaller than this value by a further factor of 2×10^2 . A summary of models yielding acceptable column abundances of CO is given in Figure 6.

IV. CONCLUDING REMARKS

Reasonable fluxes of H_2O associated with ablation of meteoroidal debris, in combination with reasonable estimates for the eddy diffusion coefficient in Jupiter's lower stratosphere, imply column densities of CO consistent with observation (Beer 1975; Larson, Fink, and Treffers 1978; Beer and Taylor 1978). The models indicate a significant stratospheric concentration of CO, as suggested by Beer and Taylor (1978).

A preferential concentration of CO in the troposphere would require either an appreciable reduction in the range of values considered above for the high-altitude source of H_2O or a large increase in the rate assumed for mixing in the lower stratosphere, or both. Conversely, a preferential concentration of CO in the stratosphere, if confirmed by further observations, may be used to place useful constraints on the rate of dynamical mixing in the deep troposphere (Prinn and Barshay 1977).

Meteoritic debris might provide detectable concentrations of disequilibrium compounds in addition to CO. Using relative abundances observed for carbonaceous chondrites (Brownlee *et al.* 1975; Mason 1971) and constraining meteoroidal fluxes to produce a CO column of $4 \times 10^{17} \text{ cm}^{-2}$, we estimate column abundances of SiO , PH_3 , and GeH_4 of magnitude $1.4 \times 10^{17} \text{ cm}^{-2}$, $1 \times 10^{15} \text{ cm}^{-2}$, and $8 \times 10^{12} \text{ cm}^{-2}$, respectively, for model A. The corresponding values for model B are smaller by a factor of 2. These estimates assume maximum conversion of Si, P, and Ge into these compounds with

the further assumption that the troposphere should represent the only significant sink. The column densities inferred here for PH_3 and GeH_4 are too small to account for the abundances of these compounds reported by Ridgway, Wallace, and Smith

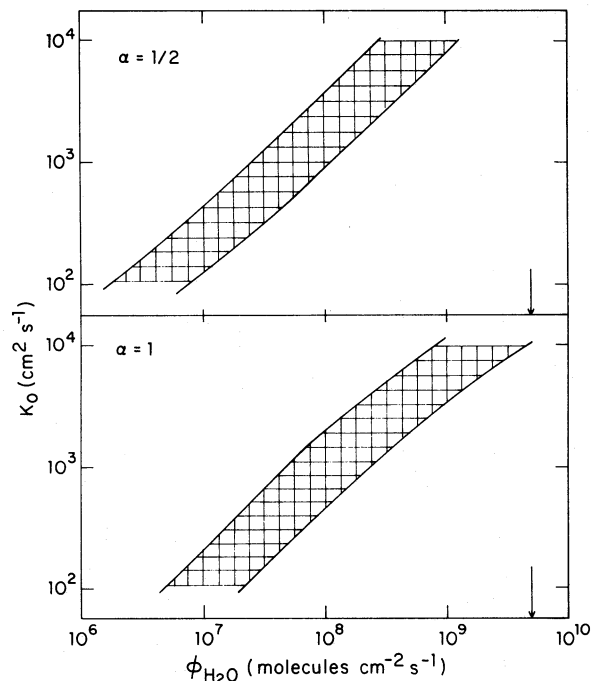


FIG. 6.—Domain of permissible values for the meteoroidal influx of H_2O and the vertical diffusion coefficient in the lower stratosphere. The cross-hatched area includes all values of these parameters which result in CO column abundances consistent with observations, 2×10^{17} to $8 \times 10^{17} \text{ cm}^{-2}$. The downward flux of H_2O is designated by $\phi_{\text{H}_2\text{O}}$. The vertical diffusion coefficient is assumed to have the form $K = K_0(N_0/N)^\alpha$. The arrow points to the flux of H_2O expected from the meteoroidal influx reported by Humes (1976).

(1976), by Larson, Treffers, and Fink (1977) and by Treffers *et al.* (1978). We expect that ablation of silicon compounds in the atmosphere of Jupiter should provide significant sources of SiO, with much smaller yields of SiH₄ (Kukielka *et al.* 1971; Gardner 1974). Silicon monoxide should be photochemically stable in the atmosphere of Jupiter and might be spectroscopically detectable with careful observations in either the X¹Σ⁻-A¹Π transition between 2100 and 2300 Å, or in the fundamental vibration-rotation band at 8 μm (Herzberg 1950). The albedo of Jupiter is unexpectedly low between 2100 and 2300 Å (Caldwell 1977), and preliminary calculations suggest that absorption by

SiO could be dominant in this spectral interval. The observed spectrum of Jupiter near 8 μm (Ridgeway, Larson, and Fink 1976) is not inconsistent with this possibility. Positive identification of SiO would provide independent support for the meteoritic model developed here.

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