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ABSTRACT

Mariner IV data on the Martian ionosphere have been variously interpreted in terms of E, F_1, and F_2 regions. Problems exist with each of the interpretations that will not likely be resolved without additional observations. The data that can be expected to be obtained from the Mariner 1969 mission may resolve the question of the appropriate interpretation, but detailed questions will certainly remain - for example, relating to latitudinal variations. Therefore, orbiters should be emphasized over fly-bys in future missions, as their data return should far exceed that of fly-bys and permit far more accurate description of the Martian atmosphere.

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The Mariner IV occultation experiment provided valuable data on the atmosphere of Mars, but it has also left room for a wide range of speculation, particularly relative to the ionosphere. The main conclusions from Mariner IV concerning the Martian ionosphere are that the altitude of the ionization peak is much lower than expected, the rate of decrease of ionization above the peak is surprisingly rapid, and the maximum ion concentration is less than was anticipated (Kliore et al., 1965). The ionosphere has been interpreted variously as analogous to the earth's E, F_1, and F_2 regions, and in fact two or more different models have been proposed for each of these analogues. These models involve neutral atmospheric particle concentrations at the ionization peak of approximately 5×10^{12}, 3×10^{10}, and 2×10^{9} cm^{-3}, respectively. The range of speculation is indicated in Figure 1, where temperature profiles appropriate to several of the models are indicated. The corresponding atmospheric density profiles are indicated in Figure 2. The E model is that of Chamberlain and McElroy (1966) for which the assumed atmospheric composition is 44% CO_2 and 56% N_2. Two F_2 models are shown, the solid line indicating that of Johnson (1965) and the dashed line that of Fjeldbo, Fjeldbo, and Eshleman (1966); both of these models are for nearly pure CO_2 atmospheres with atomic oxygen the principal constituent in the upper atmosphere. The F_1 model is also for a nearly pure CO_2 atmosphere.

The F_2 models offer little in the way of justification for the low temperature thermosphere which characterizes them; the low temperature is a conclusion reached on the basis of a particular assumption as to the physics that controls the ion peak. The strongest argument against F_2 interpretations in general is that calculations of the temperature distribution on the basis of radiation exchange, such as those made by Chamberlain and McElroy (1966) and Gross, McGovern, and Rasool (1966), indicate that the upper thermosphere should reach temperatures in the vicinity of 400°K or higher.

The F_2 model of Johnson (1965) suffers the additional limitation that laboratory data (Fehsenfeld et al., 1966) indicate that the reaction,

$$O^+ + CO_2 \rightarrow O_2^+ + CO \quad ,$$

proceeds so rapidly that the development of an F_2 region would be completely suppressed with the relative composition assumed in the model. This difficulty is avoided in the model of Fjeldbo et al. (1966) by the assumption of a very cold region between 60 and 80 km (see Figure 1), which reduces the CO_2 concentrations at higher levels to a sufficient degree to permit the existence of an F_2 region.

The E models tend to fall at the other extreme from the F_2 models. The E distribution in Figure 1 shows the temperatures calculated for the region above 150 km by Chamberlain and McElroy (1966), which they joined to a temperature model calculated by Prabhakara and Hogan (1965) for the region below 150 km; as mentioned above, this particular model assumed a composition of 44% CO_2 and 56% N_2. A feature of this particular E region model was the requirement that the atmosphere be well mixed up to 190 km in order to maintain a composition distribution that would suppress the formation of an F region; this would require a rate of eddy mixing of about 2×10^8 cm^2/sec in order to counteract the effects of molecular diffusion which tend to allow the various atmospheric constituents to distribute themselves according to their own molecular weights. The heat transport that would be associated with such a rapid rate of eddy mixing completely invalidates the calculated temperature profile, which did not take such eddy transport into account. Recently, McElroy (private communication) has suggested that radiation effects associated with the eddy motion might in large degree cancel the effect of eddy heat transfer over a large altitude region, so the importance of eddy transport in calculating temperature profiles from heat balance considerations might be small. It seems likely that such radiation effects can be important only when the optical depth is small, so that there can be radiation exchange with the environment external to the planet rather than with the immediately surrounding environment. Further, the time constant for equilibration by radiation exchange would have to be short compared to the lifetimes of turbulent eddies. These latter are not known but are probably less than 10^3 seconds; this would correspond to eddies of scale size less than 5 km and an eddy mixing coefficient of 2×10^8 cm^2/sec. Since scale sizes are not apt to be larger than 5 km (the scale height is about 10 km), the eddy time constants are probably shorter than 10^3 sec, perhaps less than 10^2 sec, and the radiative time constants would have to be shorter than this if radiation is to cancel the eddy heat transport. Since the radiative time constants may be quite short in an atmosphere that is not diluted with an infrared inert gas, the restriction indicated by the eddy transport consideration may not be valid.

Models based on an E region interpretation may require average temperatures below 100 km that are greater than those calculated on the basis of radiative equilibrium. The molecular weight assumed for the E model shown in Figure 1 is 35, whereas the occultation data for the lower atmosphere indicate a heavier molecular weight, a value of 44 characteristic of a nearly pure CO_2 atmosphere being favored by the occultation results. Spectroscopic data also indicate a nearly pure CO_2 atmosphere. If the molecular weight is 44, the temperatures below 120 km must be increased by a factor 1.25 over those associated with molecular weight 35 in order to preserve the densities required at 120 km for any particular model. Temperatures 1.25 times greater than those indicated in Figure 1 for the E model seem to be too high to be easily acceptable from an energy balance viewpoint. Other E models may of course avoid the particular difficulties of the E region model indicated in Figure 1, but E models in general are apt to require unattractively high temperatures below 100 km, or a substantial proportion of a light constituent to reduce the molecular weight, in order to maintain sufficiently high atmospheric density at the altitude of the ionization peak.

The intermediate possibility for interpretation of the Martian ionosphere is that it is analogous to an F_1 region. The ionization scale height above the peak for an F_1 region is related to the ionizable constituent, and not necessarily to the predominant ion, since chemical reactions may quickly change the ion produced by the photoionization to another species. Therefore to satisfy the requirements of an F_1 model with a warm thermosphere, it is necessary that the ionizable constituent have a large molecular weight, preferably as high as CO_2; atomic oxygen in an isothermal F_1 region would lead to the same temperature requirement as an F_2 interpretation. Further Gross, McGovern, and Rasool (1966) have drawn attention to the fact that a rise in temperature with height through the ionospheric region would lead to a reduction of the ionization scale height above the peak, and Donahue (private communication) has noted that the temperature distribution calculated by Chamberlain and McElroy for the thermosphere is consistent with an F_1 interpretation. The model indicated F_1 in Figure 1 illustrates this concept. It is drawn for a pure CO_2 atmosphere and uses the temperatures calculated by Chamberlain and McElroy for the 44% CO_2 atmosphere, adjusted so that the same temperatures occur at the same CO_2 concentrations. The temperature at the ionization peak is 275°K, higher than the 235° that would be required for an isothermal F_1 region in a CO_2 atmosphere; the scale height gradient is sufficient to bring the ionization scale height down near the value observed in Mariner IV. In this model, the temperature below 100 km has been arbitrarily lowered so as to produce the required concentration of about $3 \times 10^{10}/cm^3$ at the ionization peak; the only justification to be offered for this reduction in temperature below the calculated values is that eddy transfer might act to cool much of this atmospheric region by transferring heat downward. If McElroy's conclusion that radiation effects associated with eddy transfer can cancel the eddy transport of heat applies through this region, then it must be admitted that the F_1 model requires temperatures that are too cold from a heat balance viewpoint. This possible discrepancy appears at its worst to be no more severe than that associated with the E region interpretation, partly because, at the high latitude winter location at which the occultation took place, the temperature should be lower than indicated by the energy balance calculations of Prabhakara and Hogan, and partly because any distortion of the temperature distribution, calculated on the basis of radiative heat balance, due to the effects of eddy transport should be in the direction of producing a cooler upper atmosphere.

The question of water vapor on Mars is also relevant to the ionospheric problem. If the atmospheric water vapor content is 14 microns, as indicated by Kaplan, Munch and Spinrad (1966) and by Shorn et al. (1967), and if this is uniformly mixed through an approximately 10-km thick layer in the troposphere, the concentration would be 5 x 10^{13} molecules/cm^3, and the frost point would be near 200°K. Temperatures much lower than 200°K in the troposphere would then produce clouds. The mixing ratio would be near 2 x 10^{-4}. In general, of the various models that have been considered, only the E region models have involved temperature distributions that are consistent with the claimed water content of the atmosphere and the absence of dense white clouds. The presence in the photochemical region of water vapor or its decomposition products in relative concentrations near 2 x 10^{-4} is probably a significant factor to take into account in connection with the photochemistry and the radiative heat balance, but so far this has not been done.

It is clear from the foregoing that there is a wide diversity of opinion concerning the probable nature of the Martian ionosphere. This matter is not apt to be resolved until more data, and different types of data, are obtained for the upper atmosphere. Especially valuable would be the altitude distribution of several airglow emissions, especially resonance radiation from atomic oxygen. This can provide independent information on the temperature of the upper thermosphere, and it will assist in identifying the controlling physics and chemistry of the ionosphere.

Especially important in regard to the collection of new data is the relative roles of fly-bys and orbiters. So far as many measurements are concerned - occultation measurements included - a fly-by provides roughly the equivalent of a rocket sounding in the earth's atmosphere. An orbiter, on the other hand, is better compared with an earth satellite, where many measurements can be obtained from a single flight. One would not seriously entertain the thought of mapping the earth's atmosphere above 200 km by means of a series of hundreds of rocket flights; instead, a satellite is much more efficient for this purpose. On Mars, we are faced by a similar need for many measurements and mapping, this clearly indicates the need to emphasize orbiters over fly-bys. The differential cost between a fly-by and an orbiter is not large in a relative sense - unlike the comparison between rockets and satellites on earth where the cost ratio is about 100 or more - and this makes it all the more essential that orbiters be used rather than fly-bys. Radar mapping or repeated occultation experiments could give some indication of surface topography as well as surface pressure - at a time when further surface pressure measurements are not apt to be of any great significance unless coupled with topographic information (i.e. geocentric distance). Repeated occultation and airglow measurements, and probably other physical measurements in addition, will be required to provide the required ionospheric data that will be much more definitive than the presently available data. Even if the Mariner fly-by in 1969 is highly successful, important questions, such as those of geographic variation, will remain that will be answered only by an orbiter, or by an unrealistically large number of additional fly-bys.

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FIGURE CAPTIONS

Figure 1. Temperature distributions suggested for the Martian atmosphere on the basis of E, F_1, and F_2 interpretations of the Martian ionosphere. The E model is that of Chamberlain and McElroy (1966). The F_2 model, dashed line, is that of Fjeldbo et al. (1966). The F_2 model, solid line, is that of Johnson (1965).

Figure 2. Particle concentration distributions suggested for the Martian atmosphere for the temperature distributions indicated in Figure 1.

REFERENCES

Chamberlain, J. W. and M. B. McElroy, 1966, "Martian Atmosphere: The Mariner Occultation Experiment," Science, 152, 21-25

Fehsenfeld, F. C., E. E. Ferguson, and A. L. Schmeltlekopf, 1966, "Thermal Energy Ion-Neutral Reaction Rates III. The Measured Rate Constant for the Reaction $O^+(^4S) + CO_2(^1\Sigma) \rightarrow O_2^+(^2\pi) + CO(^1\Sigma)$," J. Chem. Phys., 44, 3022-3024

Fjeldbo, G., W. C. Fjeldbo and V. R. Eshleman, 1966, "Models for the Atmosphere of Mars Based on the Mariner 4 Occultation Experiment," J. Geophys. Research, 71, 2307-2316

Gross, S. H., W. E. McGovern and S. I. Rasool, 1966, "Mars: Upper Atmosphere," Science, 151, 1216-1221

Johnson, F. S., 1965, "Atmosphere of Mars," Science, 150, 1445-1448

Kaplan, L. D., G. Munch and H. Spinrad, 1964, "An Analysis of the Spectrum of Mars," Astrophys. J., 139, 1-15

Kliore, A., D. L. Cain, G. S. Levy, V. R. Eshleman, G. Fjeldbo and F. D. Drake, 1965, "Occultation Experiment: Results of the First Direct Measurement of Mars's Atmosphere and Ionosphere," Science, 149, 1243-1248

Prabhakara, C. and J. S. Hogan, 1965, "Ozone and Carbon Dioxide Heating in the Martian Atmosphere," J. Atmos. Sciences, 22, 97-109

Schorn, R. A., H. Spinrad, R. C. Moore, H. J. Smith and L. P. Giver, 1967, "High Depression Spectroscopic Observations of Mars II. The Water-Vapor Variations," Astrophys. J., 147, 743-752

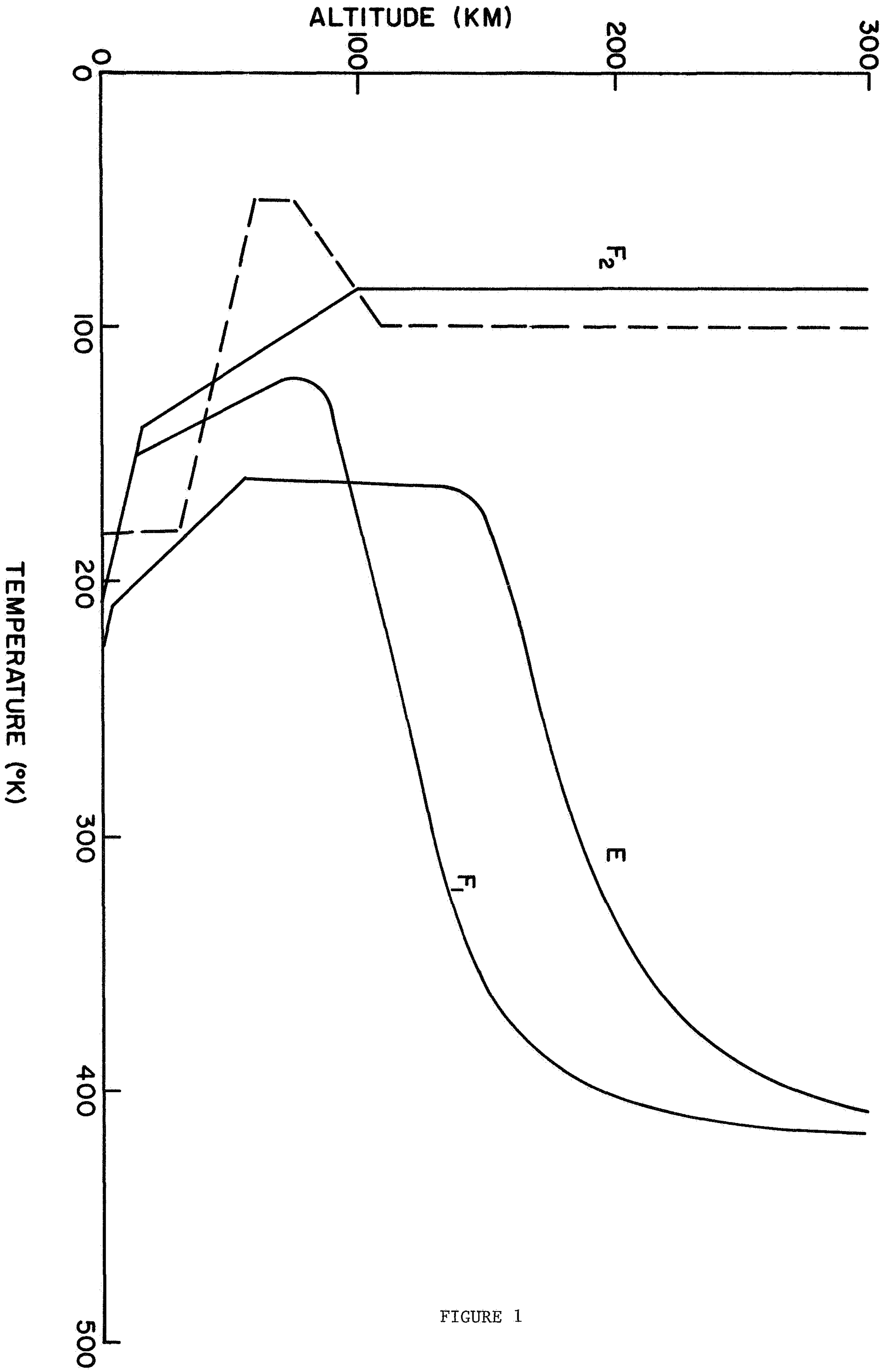
ALTITUDE (KM)
0
100
200
300
TEMPERATURE (°K)
0
100
200
300
400
500
F_2
E
F_1

FIGURE 1

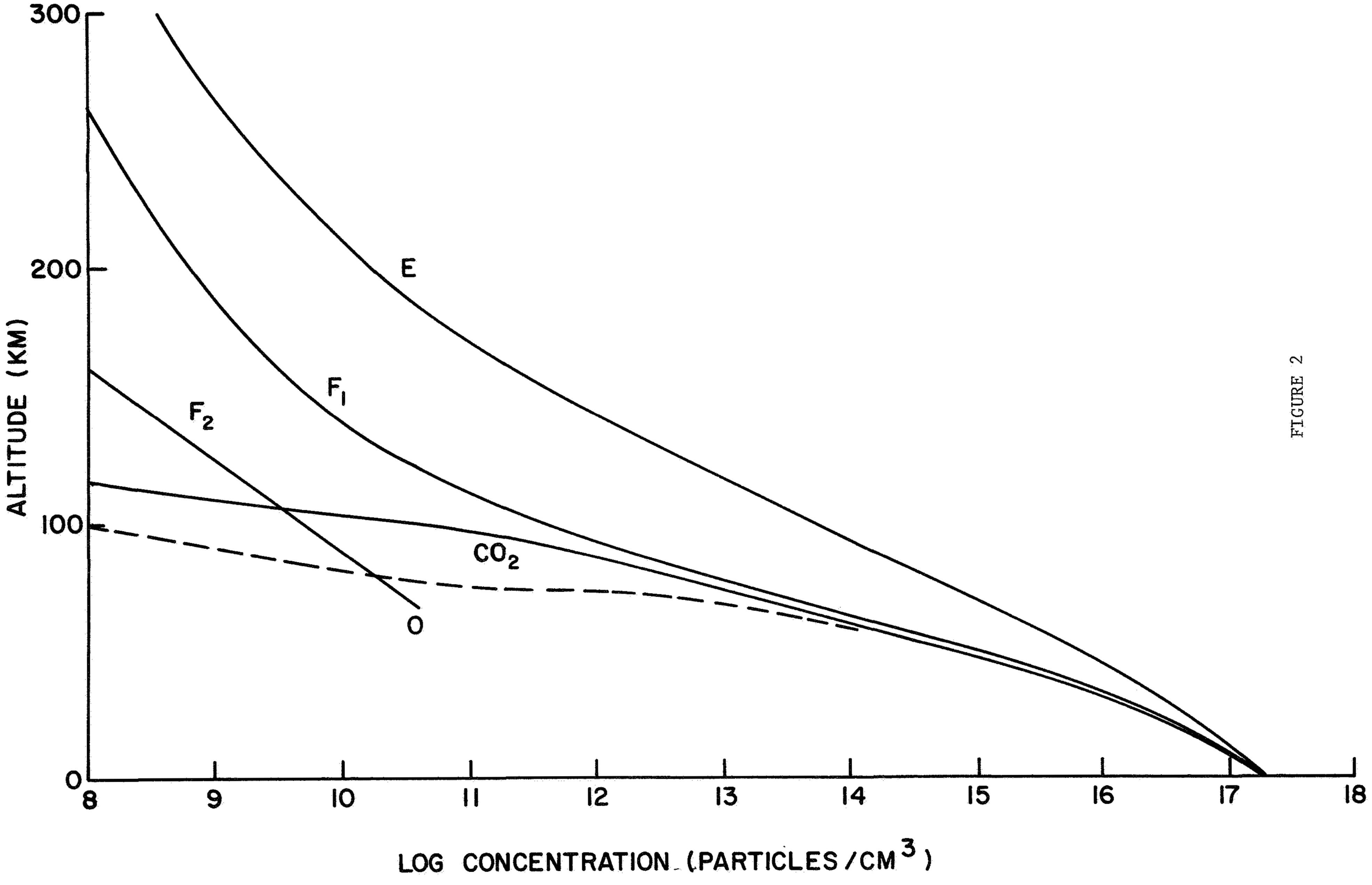
ALTITUDE (KM)
300
200
100
0
E
F_1
F_2
CO_2
O
8
9
10
11
12
13
14
15
16
17
18
LOG CONCENTRATION (PARTICLES/CM^3)

FIGURE 2