(ii) by fluorescent scattering of solar radiation by preexisting CO_2^+ ions

$$CO_2^+(\tilde{X}^2\Pi_g) + h\nu \longrightarrow CO_2^+(\tilde{A}^2\Pi_u, \tilde{B}^2\Sigma^+_u) \quad (2)$$

and (iii) by simultaneous excitation and ionization of CO_2 by photoelectron impact

$$e + CO_2 \longrightarrow e + CO_2^+(\tilde{A}^2\Pi_u, \tilde{B}^2\Sigma^+_u) + e \quad (3)$$

If we assume that the neutral component of the atmosphere of Mars is effectively pure CO_2, the total rates of population of the electronic levels of CO_2^+ by photo-ionization can be estimated directly from the cross-section data at 584 Å (*2*) and the incident solar flux. The resulting rates of production of the $\tilde{A}^2\Pi_u$ and $\tilde{B}^2\Sigma^+_u$ states of CO_2 in a vertical column are, respectively, 3.4×10^9 cm^{-2} sec^{-1} and 3.3×10^9 cm^{-2} sec^{-1} for the solar zenith angle of 27°, appropriate to the Mariner 6 observations. If no deactivation or cascading occurred, the rates of population would also be the emission intensities. Deactivation is negligible, radiative lifetimes being about 10^{-7} second (*3*), but the $\tilde{C}^2\Sigma^+_g$ state of CO_2^+ decays to both $\tilde{A}^2\Pi_u$ and $\tilde{B}^2\Sigma^+_u$ states. We assume arbitrarily that the branching ratio is 4 : 1. The predicted intensities of the two emission systems in kilorayleighs are given in Table 1.

As described by Chamberlain (*4*), the fluorescent scattering efficiences g for the two systems may be calculated from the laboratory lifetime data (*3*). We obtain for a zenith angle of 0° $g = 1.1 \times 10^{-2}$ for the $\tilde{A}^2\Pi_u$ state and $g = 1.2 \times 10^{-3}$ for the $\tilde{B}^2\Sigma^+_u$ state. If we assume that the ionic component of Mars is pure CO_2^+, the fluorescent intensities can be derived from data on the total electron content. The total electron content derived from Mariner 6 occultation data (*5*) at a solar zenith angle of 56° was 1.1×10^{12} cm^{-2}. A linear scaling procedure based upon an ionization balance between photo-ionization and dissociative recombination leads to a total electron content of 1.4×10^{12} cm^{-2} at 27°. The corresponding fluorescent intensities are given in Table 1.

Table 1. Predicted intensities (in kilorayleighs) of the $\tilde{X}^2\Pi_g$-$\tilde{A}^2\Pi_u$ and $\tilde{X}^2\Pi_g$-$\tilde{B}^2\Sigma^+_u$ band systems of CO_2^+ on Mars and Venus at a solar zenith angle of 27°.

System	Mars		Venus	
	$\tilde{A}^2\Pi_u$	$\tilde{B}^2\Sigma^+_u$	$\tilde{A}^2\Pi_u$	$\tilde{B}^2\Sigma^+_u$
Photo-ionization	4.1	3.5	17.5	15.0
Fluorescent scattering	15.3	1.6	82.6	8.4
Photoelectron impact	0.5	0.1	2.2	0.6
Total	19.9	5.2	102.3	24.0

The energy flux of photoelectrons with sufficient energy to simultaneously excite and ionize CO_2 is about 3×10^{11} ev cm^{-2} sec^{-1} on Mars. Detailed calculations of the energy degradation of photoelectrons in a pure CO_2 atmosphere, which are similar to calculations carried out for photoelectrons in air (*6*), yield the intensities given in Table 1.

Table 1 compares the contributions from the three sources. Photo-ionization is the strongest source of the ultraviolet bands at 2890 Å, and fluorescent scattering is the strongest source of the Fox-Duffendack-Barker band system. Photoelectron impact contributes not more than 5 percent for either transition.

Our conclusion that electron impact excitations are unimportant is unlikely to be altered by modifications in the adopted neutral and ionic composition or in the laboratory cross-section data. Data on the actual intensities observed from Mariner 6 would be of great interest. Similar predictions for Venus at a solar zenith angle of 27°, based upon Mariner 5 data, are also included in Table 1.

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1 December 1969; revised 12 January 1970 ■

Viscosity of Lunar Lavas

Abstract. *The viscosity of a synthetic silicate liquid with the composition of a lunar rock has been determined experimentally and found to be lower than that of any previously studied volcanic rock on earth. This fact suggests that lunar lava flows will be very thin and extensive unless they are ponded, and that lava tubes should be common and of larger dimensions than those on earth. Coarse crystallinity can be a feature of rapidly cooled surface lavas.*

Perhaps the most startling characteristics of the lunar samples returned from the Apollo 11 mission are their high contents of iron and titania, which exceed those of any known terrestrial volcanic rocks. The coarsely crystalline nature of many of the rocks suggests that the unusual composition was responsible for the low viscosity and for the ability of the rocks to flow and vesiculate but still to grow crystals of large dimension at the surface or at shallow depths where cooling is relatively rapid.

In order to investigate this possibility, we have determined the viscosity of a synthetic silicate liquid of lunar composition over a range of temperature above its liquidus. A 200-g sample was prepared with the same composition (in percentage by weight), except for the last two components listed, as that given for specimen 22 in the report of the preliminary examination team of the Apollo 11 samples (*1*): SiO_2, 43; TiO_2, 11; Al_2O_3, 7.7; FeO, 21; MnO, 0.26; MgO, 6.5; CaO, 9.0; Na_2O, 0.40; K_2O, 0.21; Cr_2O_3, 0.41; ZrO_2, 0.14; and NiO, 0.04. A solution was prepared of nitrates of Al, Fe, Mg, Ca, Na, K, Mn, and Cr in a mixture of ethanol and water. To this a weighed quantity of tetraethyl orthosilicate was added, and a gel was precipitated by the addition of ammonia. Water and NO_2 were then driven off by slowly heating the sample at successively higher temperatures. Titanium dioxide was added and ground together with the sintered material, and then the combined sample was heated to fusion at a maximum temperature of 1500°C.

On the basis of observations of the crystallization of samples cooled from temperatures between 1250° and 1500°C we believe that the liquidus temperature is probably between 1300° and 1380°C. Magnetite, ilmenite, clinopyroxene, and plagioclase were the only minerals identified in samples heated in air; in argon magnetite was not crystallized. The rapid growth of crystals during quenching makes it difficult to estimate the liquidus temperature, but the fact that a sharp change of viscosity was observed near 1370°C, both in air and in argon, indicates that below this temperature the liquid was not homogeneous.

Viscosity was measured by a counterbalanced sphere method similar to that described by Shartsis and Spinner (*2*). The apparatus consists essentially of a platinum bob suspended from one arm of an analytical balance placed on a rigid stand over a vertical SiC resistance furnace. In our runs the platinum sphere had a diameter of 0.5 cm and was suspended by a platinum wire 0.59 mm in diameter. The molten specimen was placed in an iron crucible 5 cm in diameter and 5.5 cm high, and this in turn was placed inside an iron tube from which air was evacuated by a current of argon. The furnace has vertical heating elements 35 cm long and can be controlled to ± 5°C at 1000°C near the center. The apparatus was calibrated with National Bureau of Standards soda-lime-silica glass No. 710 from 800° to 1500°C over a range of viscosity of 10^2 to 10^6 poises. The vertical movement of the platinum sphere in the molten rock was 0.702 cm. When counterweights greater than the weight of the sphere and wire in the sample are used, the motion is up, and when lighter weights are used, the motion is down; thus two load-velocity plots can be made for each temperature. These define straight lines with a slope proportional to viscosity. The error in the measurements is less than 0.05 on the logarithmic scale of poises. Another series of five measurements was made in air with a platinum crucible, in order to provide a basis for evaluating the effects of the oxidation of iron. Because of the low content of volatile components in the liquid, it is believed that the viscosity measurements made in argon must be very close to those that would be obtained if the same measurements could be made in a vacuum.

Viscosity-temperature relations for the sample in air and in argon are shown in Fig. 1, together with data for other common lavas. The viscosity of the synthetic lunar sample ranges from 4.5 poises at 1495°C to 10 poises at 1395°C in air and 5.0 poises at 1450°C to 7.1 poises at 1395°C in argon. These values are comparable to the viscosity of heavy motor oil at room temperature and much lower than any previously measured viscosity for volcanic rocks on earth. The measurements in air and argon are very close, and part of the difference may result from a surface layer of oxidized material on the sample measured in air. Surface effects of this kind have been described for slags (*3*). The temperature range of the measurements was limited on the high side by the capacity of the equipment. At temperatures below 1395°C the behavior of the platinum sphere was erratic, thus suggesting that the liquid was no longer homogeneous.

The exceptionally low values of viscosity are consistent with those predicted on the basis of bridge-density relationships described by Murase (*4*) and are similar to viscosities measured in certain iron-rich blast furnace and welding slags (*5*). They can be directly related to the high-iron and high-titania contents and the low-silica content.

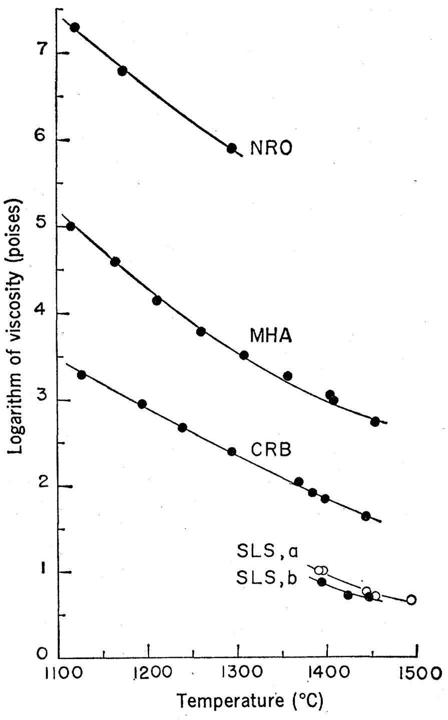

Fig. 1. Logarithm of viscosity as a function of temperature for a synthetic lunar sample (*SLS*) in air (*a*) and in argon (*b*). For comparison, data are shown for three terrestrial lavas, a rhyolite obsidian from Newberry caldera (*NRO*), an andesite from Mount Hood (*MHA*), and a Columbia River basalt (*CRB*). The viscosity of the NRO sample was measured by a bending method; viscosities for the other samples were measured by the counterbalanced-sphere method.

Recognition of the extremely low viscosity of lunar lavas leads to several significant conclusions about geologic features of the moon. Most obvious, of course, is the great fluidity of such lavas and the long distances they may flow as long as their temperatures remain near or above their liquidus. The distances lavas flow are a function primarily of their velocity and cooling rates. Velocity can be estimated from the Jeffreys formula for laminar flow

$$V = \frac{g \sin \alpha \, d^2 \rho}{3 \eta}$$

where V is the mean velocity, g is the acceleration of gravity, α is the slope angle, d is the thickness of the flow, ρ is its density, and η is its viscosity. Although lunar gravity is only one-sixth that of the earth, the fact that the viscosity of lunar lavas is at least an order of magnitude less than that of terrestrial basalts and the density is higher [about 2.95 as opposed to 2.60 at liquidus temperatures (*6*)] indicates that, for the same slope angle and thickness, the velocity of lavas on the moon will be about twice as great as that of a basalt on earth. The high velocities that would be attained in flows, even on very gentle slopes, would indicate Reynolds numbers of the order of 10^5 to 10^6. Under such conditions, flow would almost certainly be turbulent, and the velocity would become less dependent on viscosity and more a function of the roughness, shape, and slope of the channel.

Lavas of Icelandic volcanoes commonly flow more than 25 km on slopes as low as 1 degree and with thicknesses of less than a meter. Individual flows of a Columbia River basalt have been traced as far as 200 km on slopes that could not have been greater than a degree or two, but many of these lavas ponded in broad depressions and attained thicknesses of many meters.

Cooling rates of lavas are primarily a function of radiant heat losses at the surface, the losses from conduction to the ground and air being only enough to cool a flow 1° or 2°C per hour. Radiant heat losses can be evaluated from the Stefan-Boltzmann equation for blackbody radiation

$$Q/t = \sigma T^4$$

where Q is the total energy radiated per square centimeter, t is time in seconds, T is absolute temperature, and σ is a constant with the value 5.67×10^{-5} erg sec^{-1} cm^{-2} deg^{-4}. Above temperatures of about 1200°C the radiant heat loss is very high, and, because of the fourth-power relation of the temperature, it falls off rapidly at lower temperatures.

The liquidus temperature of the synthetic lunar sample is somewhat higher than that of basalts, and, if the flow is extruded at or near this temperature, the radiant heat loss from the surface would be very large. A solid crust will form quickly, and the surface temperature will subsequently be governed by the rate of conduction from the molten interior. If we assume that the thermal diffusivity does not differ radically from that of other silicates, the heat loss will be greatly reduced by the insulating effect of the crust and the lava will remain fluid in its interior.

These relations favor the formation of lava tubes. On the moon the tubes will be broader than they are on the earth because of the higher stability of the roof in a low-gravity field and much longer because of the extreme fluidity of the liquid. It seems unlikely that the fluid part of a flow would be exposed except very close to the vent or for short times after the crust of a flow was ruptured.

Finally, the low viscosity implies high diffusion coefficients that would be consistent with the coarse crystallinity of the Apollo 11 samples. Even in our relatively rapid cooling of laboratory melts, crystals over a centimeter long were commonly formed, and complete crystallization occurred in samples that were cooled from 1450° to 1200°C in about 30 minutes.

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7. This work was made possible by NASA research grant No. NGR 38-003-012.

26 November 1969; revised 2 January 1970 ■

Supersaturation of Gases in Water: Absence of Cavitation on Decompression from High Pressures

Abstract. *Very high supersaturations ranging from 140 atmospheres for oxygen to more than 270 atmospheres for helium, were produced without cavitation of the water at atmospheric pressure. Thus, the lower limits for* de novo *bubble formation are increased substantially over those previously obtained.*

Bubbles normally appear when water which has been saturated with gas at a few atmospheres pressure is exposed to atmospheric pressure. Such cavitation usually is initiated by minute gas nuclei present prior to the decompression. The supersaturation at which *de novo* bubble formation occurs is not known, although it is of general interest (*1*) and has practical implications for human diving. For instance, knowing the susceptibility to cavitation for various gases would aid in developing better gas mixtures and decompression procedures (*2*).

With the present simple technique, strikingly high supersaturations without cavitation have been obtained for various gases in water. Figure 1 shows the apparatus. After trapped air was expelled from the acid-cleaned capillary observation tube through the seal *a*, gas pressure was applied and the distilled water in cup *b* was equilibrated with the gas by shaking the assembly for 45 minutes. Tests showed this time to be more than twice that needed for complete equilibration. An inherent advantage of this method is that most of the cavitation nuclei, that is, small gas pockets, dissolve when the pressure is applied initially. After equilibration, the water saturated with gas was transferred under pressure to the capillary by slowly bleeding out water at seal *a*. The assembly was then turned to the horizontal position, and the pressure in it was lowered to ambient pressure within a few seconds while the water in the capillary was observed. When no bubbles had appeared after 10 minutes, the result was recorded as "no cavitation." Bubbles rarely appeared after this initial time period. When massive cavitation occurred, it sometimes started even before the pressure was released completely.

None of the gases tested cavitated the water at supersaturations less than 130 atm. At various pressures above this value, cavitation occurred with all gases except helium, which yielded no bubbles at pressures up to 270 atm (*3*). For each of the gases, there was a gradual increase in the number of bubbles appearing as the saturation pressure was increased. Instant, massive cavitation occurred only 10 to 20 atm above the highest supersaturations without cavitation. Examination of closeup movies showed that, during light and moderate cavitation, the bubbles always appeared to form at the glass-water interface. During massive cavitation, most of the bubbles appeared at the interface; however, there were indications that some bubbles also spontaneously appeared in the water. The results (Table 1) show that maximum supersaturations without cavitation occur at different pres-

Table 1. Supersaturation of gases in water at 25°C. In column 3 the lower of the two values is the saturation pressure which usually (two out of three trials) yielded no cavitation; the higher value is the highest observed saturation without cavitation. Column 5 lists gas solubility in water at 25°C and 1 atm (*7*). Each set of pressure values is based on 12 or more determinations near the point of cavitation made with several different samples of capillaries, distilled water, and gas.

Gas	Inner diameter (mm)	Maximum supersaturation without cavitation (atm)	Saturation for beginning massive cavitation (atm)	Gas solubility (cm³/cm³ water)
Argon	1.1	135–140	160	0.03123
Oxygen	1.1	135–140	155	.02830
Nitrogen	1.1	170–185	200	.01434
Helium	1.1	> 270	> 270	.00865
Nitrogen	2.5	145–155	170	
Nitrogen	0.6	170–190	205	
Nitrogen	0.4	165–190	210	
Nitrogen	0.15	160–170	195	