clearly biological character of the bacterium-like organism, they are significant inasmuch as they appear to show a greater degree of structural complexity. Their evident physical association with mineralogical structures of the matrix (Figs. 11 and 12) and their similarity in morphology and distribution to filiform structures seen optically in thin sections establish that these elongate threadlike forms are indigenous to the chert. Their origin, however, is difficult to determine.

The fibrillar (Fig. 11), branching (Figs. 11 and 12) morphology of these complex structures is suggestive of the high degree of molecular and polymeric order characteristic of living systems; their general appearance is not dissimilar from degraded plant material. However, the fact that they appear to lack such structures as cell walls or transverse septae, and are generally quite irregular in form, suggests that although their molecular components are probably biogenic the branching morphology may be an artifact of inorganic processes operating during crystallization of the matrix. Additionally, these forms might represent ordered remnants of organic material produced abiotically in the early stages of organic evolution (*20*). Although we regard these threadlike forms as almost certainly biogenic, additional investigation is necessary to clarify their mode of origin.

The Middle Precambrian Gunflint chert from southern Ontario contains the oldest known structurally preserved evidence of multicellular plant life (*2, 7*). The diversity and complexity of this approximately 2-billion-year-old microfossil assemblage, and the occurrence of possible biogenic remnants in sediments thought to be older than 2.5 billion years (*15, 21*), have constituted putative, yet somewhat equivocal, evidence suggesting that biological systems originated early in Precambrian time. The occurrence of bacterium-like microfossils, presumably biogenic organic filaments, and complex biologically important hydrocarbons in the Early Precambrian Fig Tree chert establishes that organisms were in existence at least 3.1 billion years ago, and indicates that life on earth must have originated during the preceding 30 percent of the earth's history.

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Upper Atmosphere and Ionosphere of Mars

Abstract. *It is argued that the single-layer ionosphere at 125 kilometers discovered in the Mariner IV occultation experiment is an F1 region coinciding with the ultraviolet photoionization peak. The CO_2 density there must be of the order of 10^{11} molecules per cubic centimeter. Such a density is consistent with the properties of the lower atmosphere by Mariner IV and the temperature model of Chamberlain and McElroy if the atmosphere is mainly CO_2 below 70 kilometers. The absence of an F2 region can be explained even if the density ratio of O to CO_2 is 100 at 230 kilometers on the basis of the rapid conversion of O^+ to O_2 by CO_2. Thus a model with an exospheric temperature of 400°K, a modest degree of CO_2 dissociation, and diffusive separation above 70 kilometers is possible.*

Chamberlain and McElroy (*1*) have shown that radiation from CO_2 and CO in the Martian upper atmosphere is not rapid enough to inhibit the development of a thermosphere in the region where the solar ionizing ultraviolet is absorbed. According to their calculations the temperature must rise from the neighborhood of 160°K at 100 km to about 400°K at 300 km. Thus the low densities required at 125 km that would permit the single ionospheric layer there (*2*) to be an F2 region (*3*) are not attainable. It is also difficult to understand why, if the ionosphere is an F2 region, it disappears in the Martian night (*2*).

If the temperature profile above 50 km calculated by Chamberlain and McElroy (*1*) is correct and if the Martian lower atmosphere consists mainly of CO_2 (or CO_2 and another heavy gas) with density and scale height similar to those measured by Mariner IV, then the

Table 1. Martian atmospheric model. Surface pressure, 7 mbar.

Altitude (km)	Temp. (°K)	Density (cm^{-3})		
		CO_2	O	CO
0	220	2×10^{17}		
50	160	1×10^{15}		
100	160	2×10^{12}	3×10^{10}	3×10^{10}
150	180	4.6×10^{9}	2×10^{9}	6×10^{8}
200	320	2.5×10^{7}	2×10^{8}	2.2×10^{7}
250	400	2.4×10^{5}	7×10^{7}	1.6×10^{6}

gas density near 125 km would be close to 10^{11} molecules per cubic centimeter. In such a case the maximum in the photoionization rate would occur there when the sun is 20 degrees above the horizon. Hence there is a good possibility that the ionosphere is an F1 region. Since the ionization rate would be about 1.2×10^{3} cm^{-3} sec^{-1} at the maximum, the electron density would be 9×10^{4} cm^{-3} (as observed) if the effective recombination coefficient were 1.5×10^{-7} cm^{3} sec^{-1}. This would correspond to the expected value for O_2^+ and could mean that this ion is the one predominantly formed as a result of ion-molecule reactions.

The suppression of an F2 peak can readily occur if the rate coefficient for

$$O^+ + CO_2 \rightarrow CO + O_2^+ \qquad (1)$$

is indeed 1.2×10^{-9} cm^{3} sec^{-1} (*4*). It is only necessary in that case that the ratio of the O to CO_2 density remain less than 10^{2} up to the altitude at which diffusion loss of O^+ begins to compete with chemical loss. The reason is that the rate of ionization of atomic oxygen in the optically thin regions of the Martian upper atmosphere is given by

$$Q(O) = 1.3 \times 10^{-7}\, n(O) \qquad (2)$$

Below the altitude at which diffusion becomes important the steady state condition for creation and loss of O^+ is

$$Q(O) = 1.2 \times 10^{-9}\, n(CO_2)\, n(O^+) \qquad (3)$$

Hence

$$n(O^+) = 10^{2}\, n(O)/n(CO_2) \qquad (4)$$

To avoid an F2 build-up of O^+ to densities greater than 10^{4} cm^{-3} it is required that the condition

$$n(O)/n(CO_2) \leqq 10^{2} \qquad (5)$$

hold up to the altitude at which

$$1.2 \times 10^{-9}\, n(CO_2) = D(O^+)/H^2(O) \qquad (6)$$

where D is the ambipolar diffusion coefficient for O^+ in O and H is the scale height of atomic oxygen (presumed to be the major constituents). If $H(O)$ is 50 km, this condition requires that at the altitude where the CO_2 density is given by

$$n(CO_2) = 3 \times 10^{14}/n(O) \qquad (7)$$

the condition (Eq. 5) be satisfied. Hence at that altitude

$$n(CO_2) = 1.7 \times 10^{6}\ cm^{-3} \qquad (8)$$

and

$$n(O) \leqq 1.7 \times 10^{8}\ cm^{-3} \qquad (9)$$

In the adopted model this CO_2 density is attained at about 230 km. From the upper limit to the oxygen density there, the maximum density at other altitudes is determined, diffusive separation being assumed. Presumably O and CO are equally abundant near 100 km. The outlines of the resultant atmospheric model are sketched in Table 1. In the model there is only modest dissociation of CO_2, but diffusive separation could occur. As Chamberlain and McElroy point out, formation of O_2 near 100 km will shield CO_2 from dissociating radiation and the low level of CO_2 dissociation is not therefore unreasonable. Presumably, then, O_2 must be present to an abundance of at least 10^{11} cm^{-3} at 100 km.

Note added in proof: Belton and Hunten (*5*) have determined with precision the partial pressure of CO_2, the surface pressure, and temperature spectroscopically. A pure CO_2 atmosphere with surface pressure 6 mbar and temperature 220°K is consistent with their measurement.

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Radiocarbon Chronology of Late Pleistocene Deposits in Northwest Washington

Abstract. *Fourteen radiocarbon dates of shells and wood from late Pleistocene sediments in northwest Washington provide evidence for correlation of the Everson interstadial with the Two Creeks interval of the midcontinent and suggest possible correlations between the Sumas and Valders stadials and between the Vashon stadial and part of the Tazewell-Cary advances.*

The latest Pleistocene deposits of the Fraser Glaciation in northwest Washington, in order of decreasing age, consist of (i) till and associated drift of the Vashon stadial, (ii) glaciomarine drift and related sediments of the Everson interstadial, and (iii) glacial drift of the Sumas stadial. This paper focuses attention on the radiocarbon chronology of the Everson interstadial, relationships of the Everson interstadial to the age of the Vashon and Sumas stadials, and possible correlations between late Pleistocene events in Washington and those in the midcontinent region and Rocky Mountains.

The presence of marine shells in late Pleistocene deposits in parts of Puget lowland of northwest Washington was noted by Reagan (*1*) and Bretz (*2*) near the early part of this century, but the origin of these deposits was not clear until Armstrong and Brown (*3*) suggested that similar fossiliferous sediments in southwestern British Columbia were deposited from floating ice in a marine environment. The origin of late Pleistocene fossiliferous marine deposits in northwest Washington has been discussed in an earlier paper (*4*), and these sediments have been included in the Everson interstadial by Armstrong, Crandell, Easterbrook, and Noble (*5*). Correlative deposits have now been traced over a large area in the Puget lowland and San Juan Islands, and 14 radiocarbon dates have been obtained from shells, wood, and peat in the deposits.

Distribution of the radiocarbon samples is shown in Fig. 1. The total area encompassed by the deposits is now known to exceed 3000 square miles (7700 km^2). Correlative deposits to the north in British Columbia span a similar area (Armstrong and others, *5*). Thus, these late Pleistocene glacial deposits constitute a unit of significant importance in the deglaciation of north-