COMPUTATION BOOK

Name SAGAN Number of Book _____

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First known protozoa (foraminifera - hard parts) - Ordovician

porifera (similar to colonial protozoa; some w. calcareous or silicon skeleton) - Cambrian

coelenterates (e.g., corals) - Cambrian

bryozoa (colonial animals w. calcareous skeleton) - Ordovician

brachiopoda (brachs) - Late cambrian

mollusca (clam shells) - Cambrian

annelida (tracks + burrows, occasionally, a fossil) - Cambrian

arthropoda (e.g., trilobites) - Cambrian

echinodermata (calcareous exterior) - Cambrian

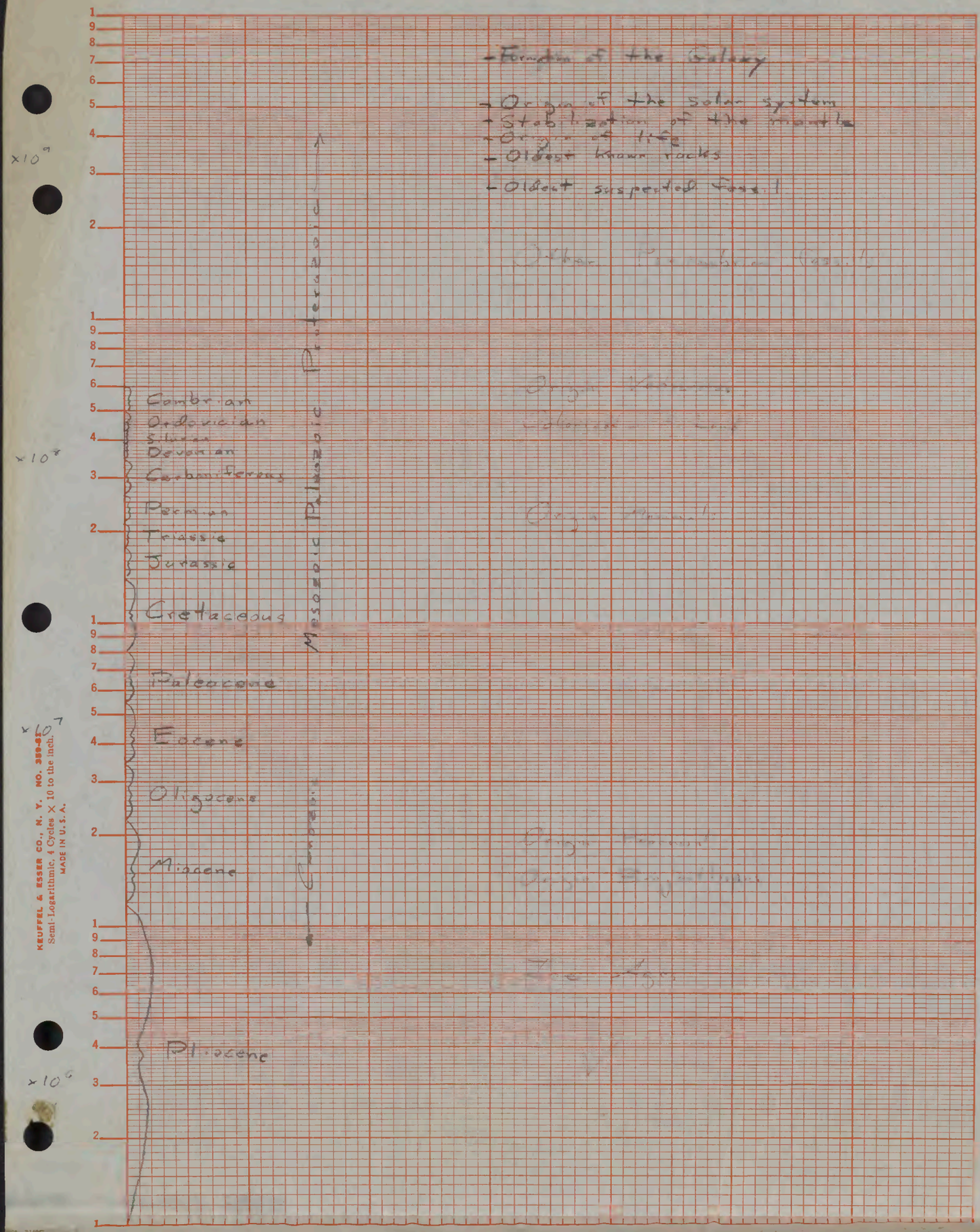
chordata (bony armour of fishes) - Middle ordovician

thallophyta (e.g., calcareous algae) - Very precambrian

bryophyta (mosses + liverworts) - Carboniferous

trachoophyta (vascular land plants) - Ordovician

And, e.g., bacteria should be much earlier.
Suspected Pre-cambrian fossils include coelenterates, brachiopods, thallophytes, porifera, + annelids.



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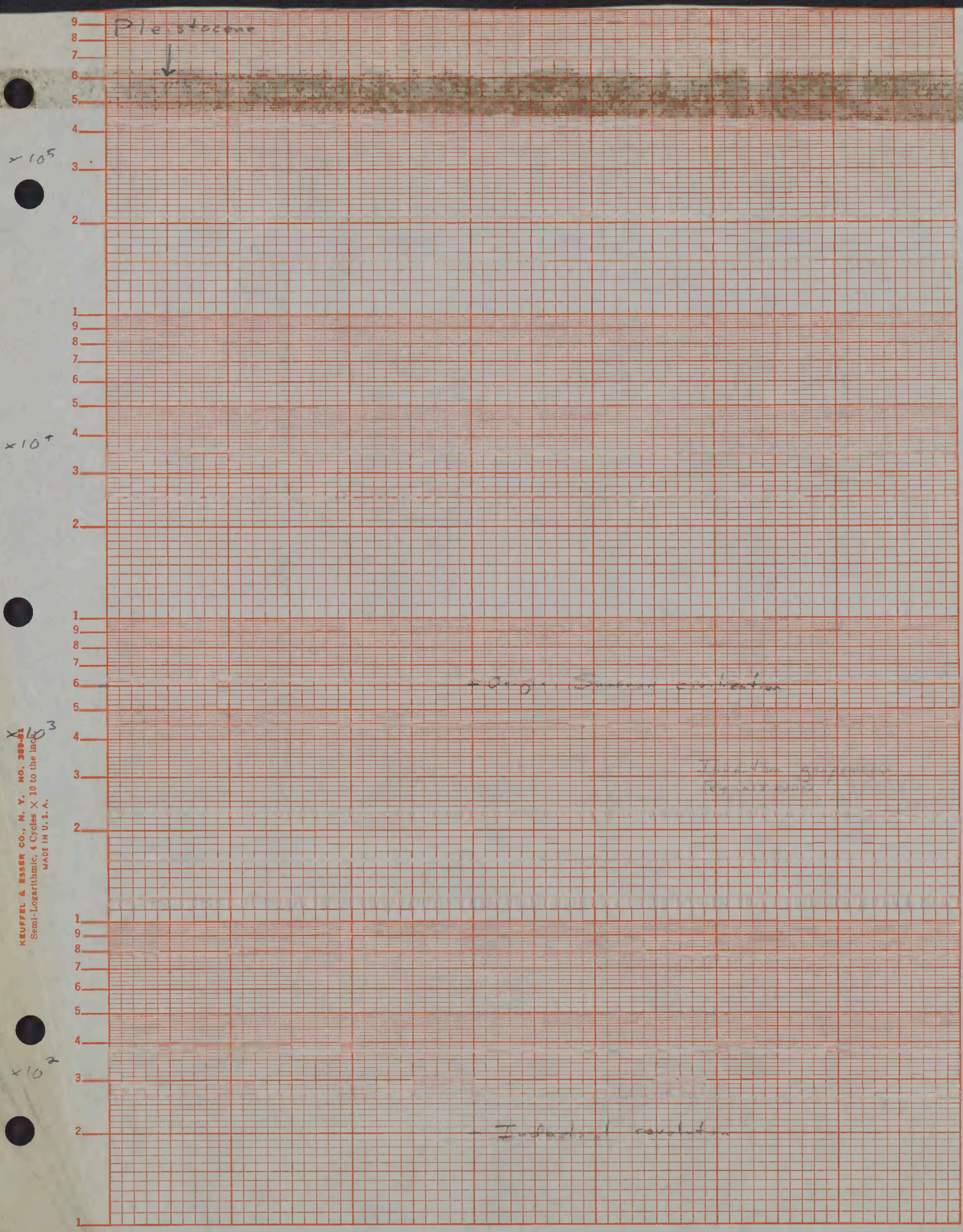
Pleistocene
×10^5
×10^4
×10^3
×10^2
Tropic Summer insolation
Tropic Summer insolation
Tropic Summer insolation
Tropical insolation
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Some Points on the Origin of Life

Natural selection of primitive polynucleotides can occur even before polynucleotide – polypeptide association. E.g., a selection pressure towards poly-A-T to avoid UV radiation damage; + away from poly-A-T to increase info. capacity. For the main evolutionary path, some orig. association of poly-n + poly-p is required. Auto-complex concentration provides one concentration mechanism.

Invoke van der Waals – London resonance interactions between poly-n + poly-p, based on similarity in bond lengths. Then a given poly-n has in its vicinity a certain specific poly-p for some period of time (e.g. a few %) > average. ∴ some simple enzymatic catalyses occur more frequently in the vicinity of one poly-n than of another. The Horowitz hypothesis on the origin of biochemical reaction chains can then be introduced at a very early stage.

Are the poly-ps preferentially produced in simulated primitive environments those which can catalyze any useful primordial reactions?

No. bits in a hg. animal:

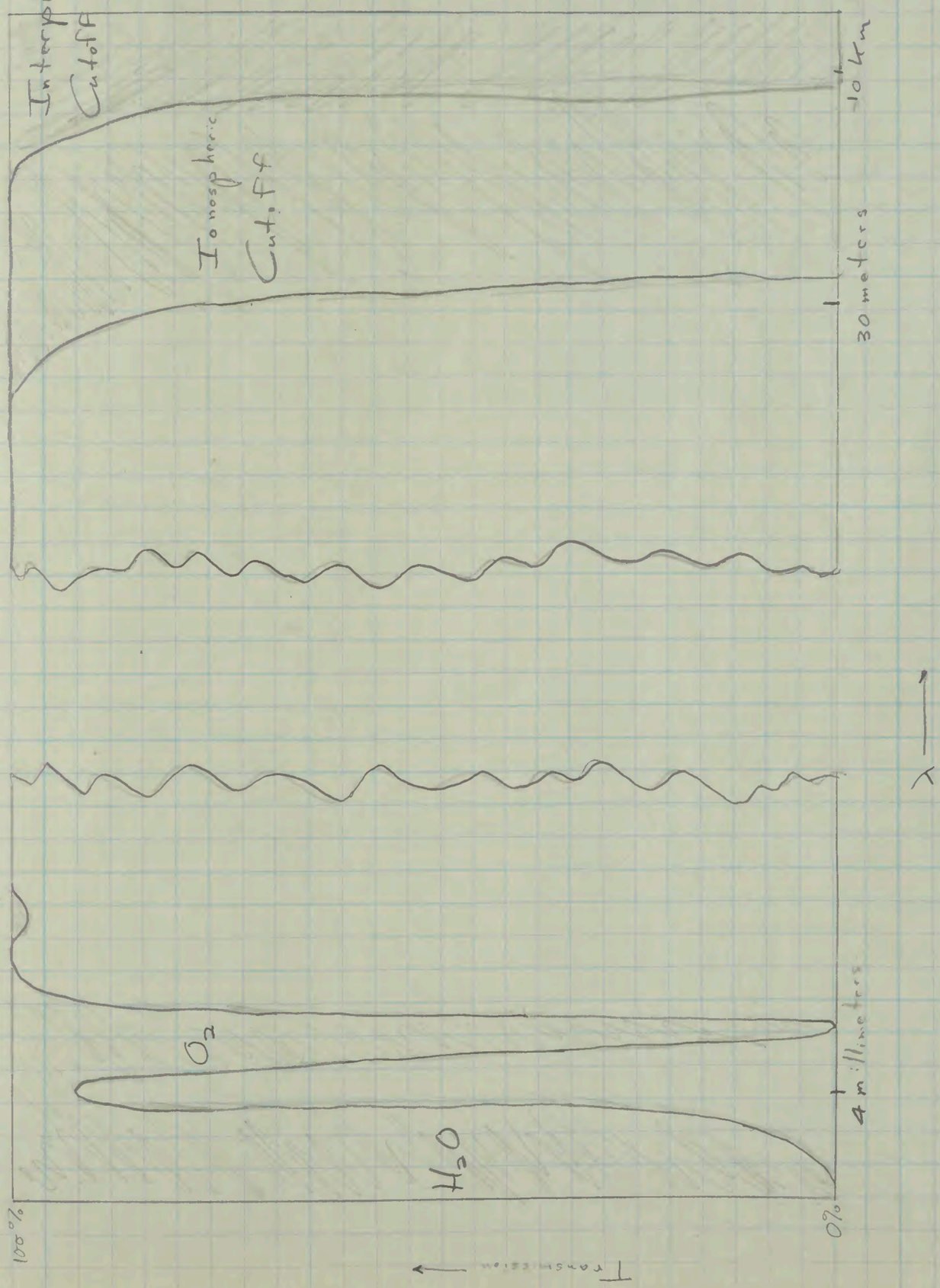
$\sim 10^3$ nucleotide pairs/gene × $\sim 10^4$ genes/chromosome

× ~ 10 chromosomes/organism ≃ 10^8 bits/organism

For a virus ~ 10^3 bits/organism.

The diffs. in size are >> the diffs. in info.

Radio Transmission Spectrum of the Earth's Environment
(Lilley, Space Age Astron., Fig. 1)



The interplanetary + ionospheric cutoffs are determined from the plasma frequency,

$$\omega_p = \left(\frac{4\pi n_e e^2}{m_e}\right)^{1/2}$$

or, numerically,

$$f_p = \frac{\omega_p}{2\pi} = 8.97\times10^3\, n_e^{1/2} \qquad \text{[in cgs units]}$$

(Spitzer, p. 50).

For ⊕ ionosphere,

$n_e \sim 10^6$, $f_p \sim 9\times10^6$ cps, +

$$\lambda = \frac{3\times10^{10}}{9\times10^6} = \tfrac{1}{3}\times10^4 = 3000\text{ cm} = 30\text{ meters}$$

For i.p. environment,

$n_e \sim 10\text{ cm}^{-3}$, $f_p \sim 3\times10^4$ cps, +

$$\lambda = \frac{3\times10^{10}}{3\times10^4} = 10^6\text{ cm} = 10\text{ km}.$$

Lilley chooses $n_e \sim 100\text{ cm}^{-3}$ + gets $\lambda = 3$ km.

For ionospheric model of ♀,

$n_e \sim 10^9$ cps, $f_p \sim 3\times10^8$ cps, +

$$\lambda = \frac{3\times10^{10}}{3\times10^8} = 100\text{ cm} = 1\text{ meter}.$$

$n_e \sim 10^{10}$ cps $\Rightarrow$ $\lambda = 30$ cm. ~~The resulting Gaussian window is the very narrow, being bounded at short wavelength by free-~~
I.e., everything longward of 30 cm does not pass through the ionosphere. Because of free-free emission, everything longward of ~3 cm (+ arriving from the surface) does not pass through the ionosphere.

On the Milieu Intérieur
(Baldwin, Comparative Biochemistry, chap. I)

"La fixité du milieu intérieur est la condition de la vie libre"
— Claude Bernard

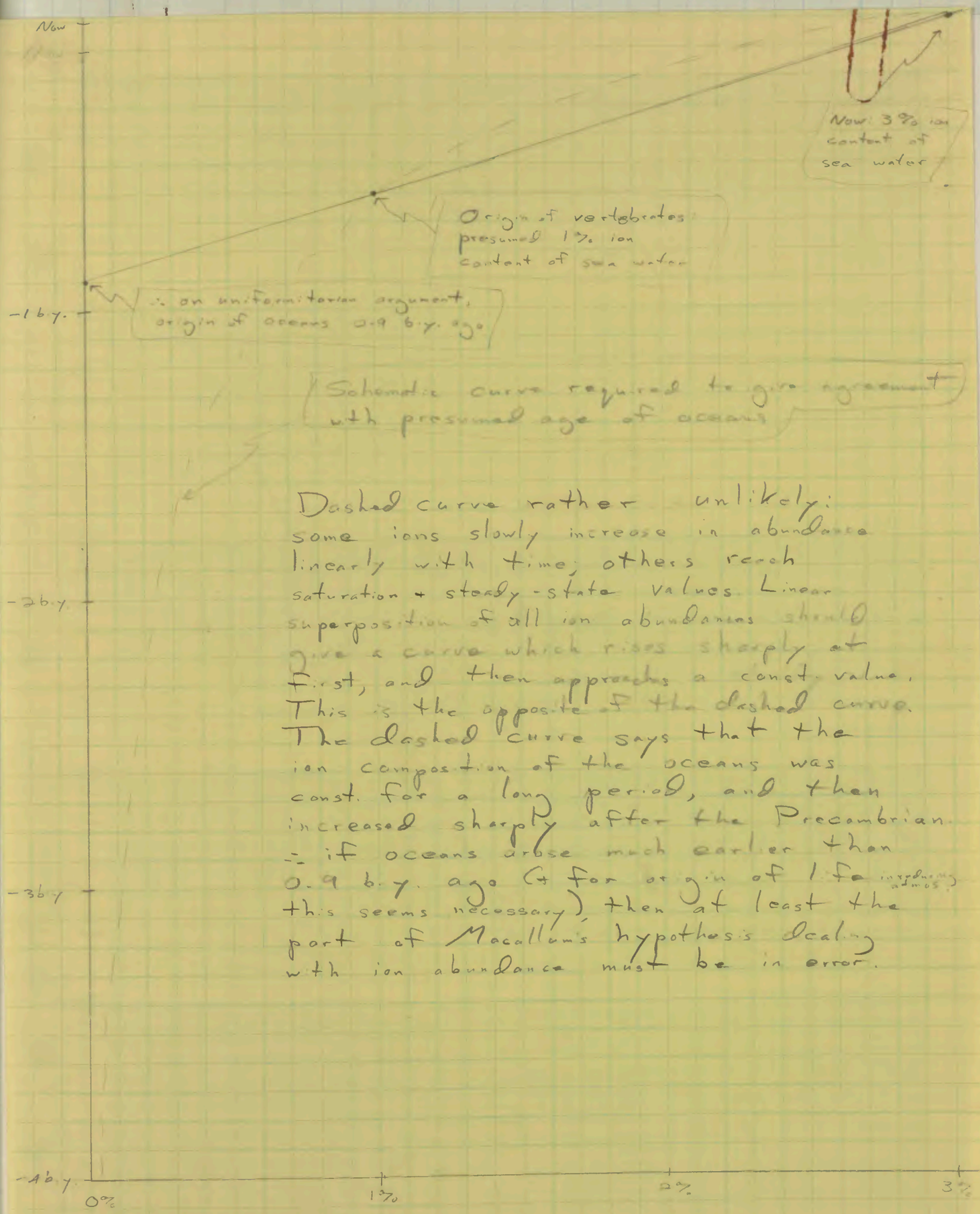
Among the virtues of marine environment for the origin of life are its high specific heat, so that ΔT is sm., + high viscosity so no violent mechanical disturbances. [Also ice lighter than water; otherwise continual deposition of ice until no liquid left. Liquid ammonia is less dense than solid ammonia, + so there it works in the wrong direction.] On other hand, much less O_2 in water than in air [∴ if life arose in ocean, difficult to use O_2 even if atmosphere were oxidizing!] Contemporary organisms are very sensitive to changes in such internal properties as ionic constituents, pH, T, etc. With no regulating physiology the most primitive organisms must have lived in a very stable environment ⇒ the sea. But colonization of dry land ~~would~~ or fresh water required major regulating developments.

The relative composition of the ions in blood plasma of widely differing ~~all~~ animals are remarkably alike, + strikingly similar to that of contemporary seawater (Baldwin, Table III, p. 10). [What of ionic comp. of cells?] Macallum suggested that the resemblance is due to the fact that the circulatory fluids are simply substitutes for the sea water at the time of origin of metazoa, i.e., that blood is modified seawater. [∴ ionic composition seawater cond. in time. Cf. Rubey]. What discrepancies do exist, Macallum attributes to slow changes in oceanic ionic comp.; e.g., more K + less Mg in Archean Ocean. Crabs more nearly resemble modern sea water. Macallum concludes crabs shut off from seawater only recently. "... life began in the ocean millions of years ago, ... the blood systems of diff. groups of animals closed off at diff. times, ... the composition of the bloods of their descendants has remained practically unmodified ever since owing to the influence of heredity."

Potts has objected to Macallum's assumption that once the animal is shut off from sea water, its ionic composition remains fixed. The body surface of many organisms is permeable to ions, + a change in ionic composition of seawater will result in a change in the ionic composition of such marine organisms. Differential permeability → internal ionic distribution must change. ∴ the ionic diff between present seawater + some present organisms' blood must be attributed to differential permeability, + not to Δ ionic comp of seawater during geological time. In fact diff. perm. works in the right direction: K moves faster than Na, Mg slower.

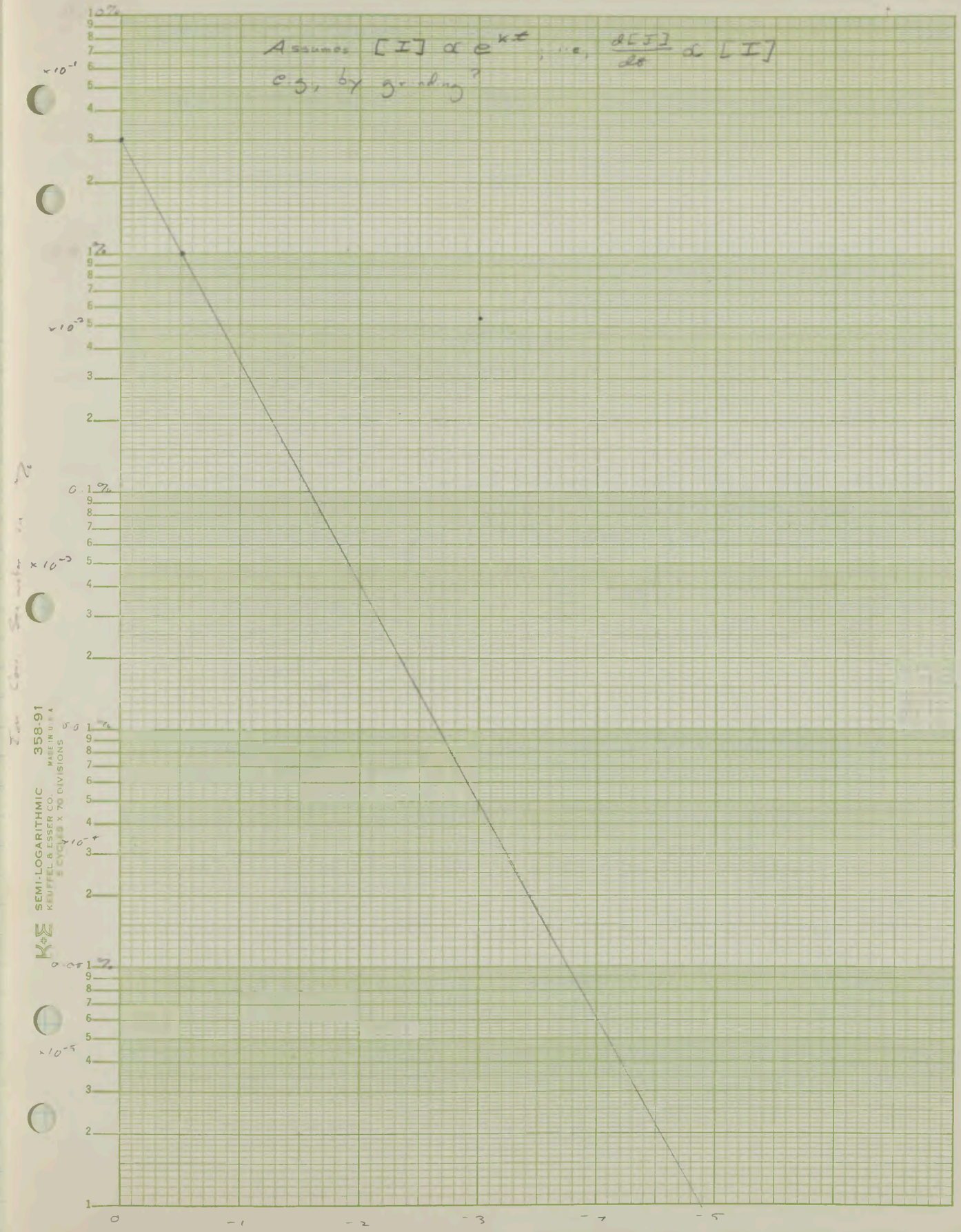
On this model we expect that as the excreting apparatus becomes more efficient, more exchange occurs, the effect of differ. permeability is enhanced, and the ionic comp. differs more + more from seawater [For sea animals. The argument becomes invalid once an animal arrives on dry land. There no exchange w. the ocean exists, + Macallum's original idea on the fixity of the shut-off circulatory fluid remains valid. ∴ to some extent ionic comp. seawater has remained fixed in time, at least from the Paleozoic.] [This is a potential tool for evolutionary analysis. E.g., comparison of ionic comp. of dolphins + sharks.]

Fresh water eels spend their adult lives in low salt conc. environments; yet they swim to marine localities to spawn their young (e.g., migration to the Sargasso Sea). This is merely to provide an adequate ionic environment for the young eels and their still underdeveloped osmoregulatory apparatus (also cf. recapitulation)



Dashed curve rather unlikely: some ions slowly increase in abundance linearly with time; others reach saturation + steady-state values. Linear superposition of all ion abundances should give a curve which rises sharply at first, and then approaches a const. value. This is the opposite of the dashed curve. The dashed curve says that the ion composition of the oceans was const. for a long period, and then increased sharply after the Precambrian. ∴ if oceans arose much earlier than 0.9 b.y. ago (+ for origin of 1 fa involving atmos.) this seems necessary, then at least the part of Macallum's hypothesis dealing with ion abundance must be in error.

= 120 K°/3



5.9×10^{-3}

$= 120\ K^{0}/3$

Intermission:

Nocturnal Cooling on Mars + Elsewhere

Brunt:

$$\tau = \frac{3}{4}\pi \rho^2 c_p^2 \kappa \left[\frac{\Delta T}{S(0)}\right]^2$$

$$\therefore \frac{\Delta T}{S(0)} = \left[\frac{4\tau}{3\pi \rho^2 c_p^2 \kappa}\right]^{1/2}$$

This is especially okay since it appears under $\sqrt{\ }$

What value to choose for the net radiation flux, $S(0)$, leaving the ground at night (which, by hypothesis, is just balanced by the heat conducted to the surface from underground)? Since $S(0) = \varepsilon \sigma \overline{T_s^{\bullet 4}}$, the value depends critically on the choice of $\overline{T_s^{\bullet}}$. But $\exists$ compensating b.e.'s. For $S(0)$ lg., ΔT is lg. But ΔT cannot be so lg. that the $\overline{T_s^{\bullet}}$ from ΔT is $\ll$ the $\overline{T_s^{\bullet}}$ from $S(0)$.

(i) $\overline{T_s^{\bullet}} \simeq 240°K$.

For medium fine, dry quartz sand,

$$\rho c_p \kappa^{\frac{1}{2}} = 0.0140 \text{ cal } (K°)^{-1} \text{ cm}^{-2} \text{ sec}^{-1/2}.$$

1 cal. = 4.2 joules = 4.2×10^7 ergs.

$\therefore$ in c.g.s. units, $\rho c_p \kappa^{\frac{1}{2}} = 5.9 \times 10^5$ erg $(K°)^{-1}$ cm^{-2} sec$^{-1/2}$

τ = 24 hrs 37 min = 24.62 hrs = 8.87×10^4 secs.

$$\therefore \Delta T = 0.9 \times 5.67 \times 10^{-5} \times (2.4)^4 \times 10^8 \times 2 \times \left[\frac{8.87 \times 10^4}{3\pi \times 5.9^2 \times 10^{10}}\right]^{1/2}$$

$$\cancel{= 3.37 \times 10^3 \times [1.59 \times 10^{-2}]^{1/2}}$$

$$\cancel{= 3.37 \times 10^5}$$

$$= \frac{3.37 \times 10^3}{5.9 \times 10^5 \times 3} \times [8.87 \times 10^4]^{1/2} = \frac{3.37 \times 10^5}{5.9 \times 10^5 \times 3} \times 2.98 \times 10^2$$

$$= 120 \text{ K}^{\circ}$$

In gen., on ☾ at night, vapor pressure H_2O is effectively 0 so far as IR opacity goes. Effect of CO_2 on net outgoing radiation should be very sm.; i.e., negligible nighttime radiation from atmos CO_2 to ground, because of steep diff. in temp. between ground + air (Mintz). ∴ $\varepsilon\sigma\overline{T_s}^4$ O.K. ∴, in gen.,

$$\Delta T = 0.9 \times 5.67\times10^{-5} \times 2\times10^{8} \times 2.98\times10^{2} \times \left[\frac{T}{100}\right]^4$$

$$\times \frac{1}{\sqrt{3\pi}} \times \frac{1}{\rho\, c_p \sqrt{\kappa}}$$

$$= 9.6\times10^{5} \times \frac{(T/100)^4}{\rho\, c_p \sqrt{\kappa}}$$

∴, e.g., $T = 240$, $\rho c_p \sqrt{\kappa} = 5.9\times10^5$ gives

$$\Delta T = \frac{9.6\times10^5}{5.9\times10^5} \times 3.3\times10 = 53^\circ K.$$ Doesn't check. It's okay; w. penal correction overleaf.

$$\therefore \overline{T_s}^\circ = 200^\circ K \Rightarrow \Delta T = \frac{9.6\times10^5}{5.9\times10^5}\cdot 16 = 26^\circ K.$$

This is a contradiction, because we know $\overline{T_s}^\circ$ sometimes is ~ 266°K. ∴ $\Delta T = 26^\circ K \Rightarrow \overline{T_s}^\circ > 240^\circ K$, contrary to assumption. $\overline{T_s}^\circ = 240^\circ K$ is okay. It gives $\Delta T = 53^\circ K$, so $\overline{T_s}^\circ = 293^\circ K = 20^\circ C$, which checks.

Now sandy clay w. 15% moisture gives $\rho c_p \sqrt{\kappa}$ which is $0.0358/0.0140 = 2.56$ larger. ∴ ΔT would be 2.56 smaller. E.g., $\overline{T_s}^\circ = 240^\circ K$ would give $\Delta T = 21^\circ K$ + $\overline{T_s}^\circ = 261^\circ K$. This is contrary to observation.

If we know $\overline{T_s}^\circ$ and ΔT, we can derive $\rho c_p \sqrt{\kappa}$. From Dobson's data, we know, e.g., approximate T at 7 A.M. local time, + max. daytime temp. This gives all we need, if we assume T at 7 A.M. = avg. nighttime temp. (should be approx. correct).

∴ from Abbot, $\overline{T_s}^\circ \simeq 225^\circ K$, $\Delta T \simeq 50\,K^\circ$

$$\therefore \rho c_p \sqrt{\kappa} = \frac{9.6\times10^5 \times (2.25)^4}{50}$$

$$= 4.9\times10^5 \text{ ergs } (K^\circ)^{-1}\,cm^{-2}\,sec^{-1/2}$$

$$= \frac{4.9\times10^5}{4.2\times10^7} = 0.012 \text{ cal } (K^\circ)^{-1}\,cm^{-2}\,sec^{-1/2}$$

∴ somewhat larger than ordinary granite, and, very likely actually T_s° is not slightly less, + $\rho c_p \sqrt{\kappa}$ is therefore less, further accentuating the degree.

Now note that the T_s° used in computing $S(0)$ is not here the avg. nighttime temp. as for ξ; it is the minimum temp. But specifying $\rho c_p \sqrt{\kappa}$, we can then derive T_s° (v. Johnson, p. 158).

Sutton derives the product $\sqrt{\rho c_p \kappa}$ from his analysis. ∴ we can find both κ and ρc_p. Also applicable to the moon.

Sutton (Ap. J. 131: 459, 1960, p. 466) calculates diurnal temp. variation on ☾ + compares w. observations. Values fall between computation for $(\kappa\rho c_p)^{1/2} = 0.004$ + 0.01.

For $(\kappa\rho c_p)^{1/2} = 0.004$, $\Delta T = 140\,K^\circ$, $T_s^\circ = -110^\circ C = 163^\circ K$. Then,

$$\rho c_p \sqrt{\kappa} = \frac{9.6\times10^5}{1.4\times10^2} \times (1.63)^4$$

$$= 4.8\times10^4 \text{ ergs } (K^\circ)^{-1}\,cm^{-2}\,sec^{-1/2}$$

$$= 0.0011 \text{ cal } (K^\circ)^{-1}\,cm^{-2}\,sec^{-1/2}$$

Approximately of range

For $(\kappa\rho c_p)^{1/2} = 0.01$, $\Delta T = 90\,K^\circ$, $T_s^\circ = -70^\circ C = 203^\circ K$. Then,

$$\rho c_p \sqrt{\kappa} = \frac{9.6\times10^5}{9.0\times10} \times (2.03)^4$$

$$= 1.81\times10^5 \text{ ergs } (K^\circ)^{-1}\,cm^{-2}\,sec^{-1/2}$$

$$= 4.3\times10^{-3} \text{ cal } (K^\circ)^{-1}\,cm^{-2}\,sec^{-1/2}$$

$$\therefore \frac{\rho c_p \sqrt{\kappa}}{(\rho c_p \kappa)^{1/2}} = \sqrt{\rho c_p} = \frac{4.3 \times 10^{-3}}{1.0 \times 10^{-2}} = 0.43$$

$\therefore \rho c_p = 0.185$, intermediate between snow + pumice and

Also, $\sqrt{\kappa} = \dfrac{\rho c_p \sqrt{\kappa}}{\rho c_p} = \dfrac{4.3 \times 10^{-3}}{1.85 \times 10^{-1}} = 2.3 \times 10^{-2}$

$\therefore \kappa = 5.3 \times 10^{-4}\ cm^2\ sec^{-1}$.

Corresponds to no known material; ~~approximately~~ within factor 2 of values for clay and pumice + still water. But ρc_p far off for these.

Back to $(\kappa \rho c_p)^{1/2} = 0.004$: $\rho c_p \sqrt{\kappa} = 0.0011$.

$$\therefore \sqrt{\rho c_p} = \frac{1.1 \times 10^{-3}}{4 \times 10^{-3}} = 2.75 \times 10^{-1}$$

$\rho c_p = 7.6 \times 10^{-2}$. Range of snow.

$$\sqrt{\kappa} = \frac{1.1 \times 10^{-3}}{7.6 \times 10^{-2}} = 1.4 \times 10^{-2}$$

$\kappa = 2 \times 10^{-4}\ cm^2\ sec^{-1}$, even worse.

Conclusion of this exercise:

Brunt's eq. gives perfectly reasonable results by itself (for, e.g., dry pumice and) if we assume the value of ΔT ~~[illegible]~~ is that of Gibbard. If we take it to be that of Sinton, the method gives absurd results for familiar terrestrial materials. On the other hand Sinton's computation based on Woolich + Jaeger show agreement. What diff. in assumption is made in the two theories?

Order of my treatment of Sutton's calc.

Heat conduction eq.:

$$\frac{\partial T}{\partial t} = \frac{\kappa}{\rho c_p} \frac{\partial^2 T}{\partial x^2}.$$

Approx.:

$$\frac{\Delta T}{\Delta t} = \frac{\kappa}{\rho c_p} \frac{(\Delta T)^2}{(\Delta x)^2}.$$

B.c.s: at any level flux is given by

$$F = \kappa \frac{\partial T}{\partial x}.$$

Approx.:

$$F = \kappa \frac{\Delta T}{\Delta x}.$$

At surface, flux is

$$F = \sigma T^4 - I,$$

where I is solar heating.

During the day, assume

$$I = \frac{1-A}{r^2} G \cos \frac{2\pi t}{P}$$

where G is solar const. $\therefore$

$$\frac{\Delta T}{\Delta t} = \frac{\kappa}{\rho c_p} \left(\frac{F}{\kappa} \right)^2 = \frac{F^2}{\rho c_p \kappa}.$$

$$\therefore \frac{\Delta T}{\Delta t} = \frac{1}{\rho c_p \kappa} \left[\sigma T^4 - \frac{1-A}{r^2} G \cos \frac{2\pi t}{P} \right]^2.$$

∴ at night,

$$\frac{\Delta T}{\Delta t} = \frac{S(0)^2}{\rho c_p K},$$

or completely converting to previous notation,

$$\Delta T = \frac{S(0)^2}{\rho c_p K} \tau.$$ ∴ S(0) is avg nighttime flux

Brunt's treatment:

$$\frac{\partial T}{\partial t} = K_s \frac{\partial^2 T}{\partial z^2}$$

Approx:

$$\frac{\Delta T}{\Delta t} = K_s \left(\frac{\Delta T}{\Delta z}\right)^2$$

Newton's law of cooling:

$$F = -\rho c_p K_s \frac{\partial T}{\partial z}$$

Approx:

$$F = -\rho c_p K_s \frac{\Delta T}{\Delta z}.$$

∴ Sutton's & Brunt's treatments are the same with the understanding that Sutton's

$$K = \rho c_p K_s.$$

∴ Sutton's $(K \rho c_p)^{1/2} = \rho^{1/2} c_p^{1/2} K_s^{1/2}$

∴ Brunt's sol. is

$$\frac{\Delta T}{\Delta t} = K_s \left(\frac{F}{-\rho c_p K_s}\right)^2$$

or $$\boxed{\Delta T = \frac{S(0)^2}{\rho^2 c_p^2 K_s}\,\tau}$$

or within a factor of $2/\sqrt{\pi}$ of the rigorous solution.

∴ the use of Sutton's nos. was misleading. We must first convert. ∴ Sutton's values lie between

$$\rho c_p K_s^{1/2} = 0.01 \text{ and } 0.004.$$

This is between the values for snow + sand. Let us try to derive these values:

Sutton's data show for $\frac{1}{3}\tau_{sp}$ $\Delta T = +25°C - (-65) = 90\,K°$. $\overline{T_s}° \approx T_s(7\,A.M.) = 210°K$.

$$\therefore \rho c_p \sqrt{K_s} = \frac{S(0)}{\sqrt{\Delta T}}\sqrt{\frac{\tau}{3}}$$

$$= \frac{0.9 \times 5.67 \times 10^{-5} \times (2.1)^4 \times 10^8 \times 2.98 \times 10^2}{1.73 \times 9.5}$$

$$= 1.63 \times 10 \times 10^5 = 1.63 \times 10^6 \text{ erg } (K°)^{-1} cm^{-2} sec^{-1/2}$$

$$= 3.88 \times 10^{-2} \text{ cal } (K°)^{-1} cm^{-2} sec^{-1/2}.$$

This is just in the range of sandy clay w. 15% moisture + is higher than Sutton's values. The whole question of the exact numerical value turns on the extrapolation to sunrise temp. (~6 A.M. local time). Sutton's obs range from −70°C to −110°C for this quantity. In T, this covers a very wide range. Gifford has higher values of $\overline{T_s}°$, + ∴ the $\rho c_p \sqrt{K_s}$ derived from his data is also higher. Gifford's $\overline{T_s}° \simeq 210°K$ as we have calc., + $\Delta T \simeq 70K°$.

∴ to determine moisture content of Martian soil need T obs. at sunrise. ∴ Mariner obs.

Exactly the same procedure can be used for the moon. E.g. Sinton in Kopal (p. 411) gives some pts. The curve drawn through them gives

$$\overline{T_s^\circ} \simeq 120^\circ K, \quad \Delta T \simeq 380 - 120 = 260^\circ K.$$

For the moon $\tau = 27^d.3 = 27.3 \times 8.64 \times 10^4 = 2.35 \times 10^6$ secs

$$\therefore \rho c_p \sqrt{K_s} = \frac{0.9 \times 5.67 \times 10^{-5} \times (1.2)^4 \times 10^8 \times \sqrt{2.35} \times 10^3}{1.73 \times \sqrt{2.6} \times 10 \times 4.2 \times 10^7}$$

$$= 0.9 \times 10^{-2} \text{ cal } (K^\circ)^{-1} \text{ cm}^{-2} \text{ sec}^{-1/2}$$

Compares very favorably w. Sinton's 0.23×10^{-2} cal $(K^\circ)^{-1}$ cm^{-2} sec$^{-1/2}$. These values of the thermal inertia are approx. those of snow (prob. freshly fallen — good analogy!)

Back to gen. princ.: the eq. is

$$\Delta T = \frac{S(0)^2}{\rho^2 c_p^2 K_s} \Delta t.$$

$\therefore$ can proceed by iteration to compute entire temp profile: Start w. max T. Compute ΔT for sm. t using T_{max}. This gives new T in $S(0)$. Etc., each operation giving the T for the next sm. Δt.

∴ below a few tens of cms on ♂ + a few metres on the ♀ ΔT is very small indeed, + T, though low, is const.

Note the two b.c.'s give the skin depth ~~at~~ from which the rad. is effectively coming:

$$S(0) = \rho c_p K_s \frac{\Delta T}{\Delta z}$$

∃ diff. between ΔT in time & ΔT in space + this eq. can be solved for the gradient directly. Alternatively, can specify a value of ΔT + derive skin depth Δz.

$$\therefore \boxed{\Delta z = \rho c_p K_s \frac{\Delta T}{\varepsilon \sigma T_s^4}}.$$

Must know material to put in $\rho c_p K_s$.

First try ♀: Take values for snow.

$$\begin{array}{r} -1 \\ -1 \\ -3 \\ -8 \\ \hline -13 \end{array} \qquad \begin{array}{r} +2 \\ +5 \\ \hline +7 \end{array}$$

$$\therefore \Delta z = 1.0\times10^{-1} \times 2.7\times10^{-3} \times \frac{2.6\times10^{2}\times4.2\times10^{7}}{0.9\times5.67\times10^{-5}\times(1.2)^4\times10^8}$$

$$= 3.4\times10^{-6}\times4.2\times10^{7}$$

$$= 1.4\times10^{2} \text{ cm or } 1 \text{ meter}.$$

∴ the topmost meter of lunar surface cools during a lunation. Below that, regions of approx. const. temp. should be found, as radio data indicate.

For ♂: Take values for gravity sand.

$$\begin{array}{r} -1 \\ -3 \\ -5 \\ \hline -12 \end{array} \qquad \begin{array}{r} +1 \\ +7 \\ +5 \\ \hline +13 \end{array}$$

$$\Delta z = 1.9\times1.65\times10^{-1}\times2\times10^{-3}\times\frac{7.0\times10\times4.2\times10^{7}}{0.9\times5.67\times10^{-5}\times(2.1)^4\times10^8}$$

$$= 1.9\times10 = 19 \text{ cm}.$$

For ♀, ΔT is about same as for ♂, but T is 3× more.
∴ Δz is $3^4 = 81$ × less.

$$\therefore \Delta z \sim 2 \text{ mm},$$

+ we see an extremely thin surface region. But the radio waves come from much further down. ∴ how can there be a phase effect on ♀, unless $\rho c_p K_s$ is a factor

of $> \frac{200\,cm}{0.2\,cm} = 10^2$ less?

For quartz sand, $\rho c_p K_s \sim 2\times10^{-1} \times 1 \times 2\times10^{-3} = 4\times10^{-4}$.

For still air, $\rho c_p K_s \sim 2\times10^{-1} \times 10^{-3} \times 2\times10^{-1} = 4\times10^{-4}$.

Everything else is more. ∴ something is wrong. What's wrong is that the surface of ♀ is not cooling radiatively. First, net flux has value $\sim\sigma(234)^4$ ≈ that for ☉. Second, advective heating. With this T_s^0, + quartz sand, $\Delta z \sim 20$ cm which is probably okay. But it does suggest that there should be a greater flux effect + greater surface temp. at short λ than at long. Obs. is:

Mayer, McCullough + Sloanaker (3.15 cm)

$\Delta T = 146°K$, $T^0 = 694°K$

Drake (10 cm)

$\Delta T = 78\ K°$, $T^0 = 661\ °K$.

Kuzmin + Salomonovich (8 mm)

ΔT very lg, $T^0 \sim 750°K$.

Completely consistent

Now do calc. carefully. $T_s^0 = 234°K$ lowers Δz still further, by factor $1.5\times10^{-2}/1.0\times10^{-2} = 1.5$; $19\,cm/1.5 = 12\,cm$. 21 cm measurements badly needed. But 10 cm radio waves don't come from 10 cm below surface, usually from further below ∴ not imp't except for order of mag.

Treatment assumes very little about nature of surface (pastures, sand, + air all have same $\rho c_p K_s$). It does assume surface is heated by sunlight (i.e., greenhouse effect).

What then does the lg. ΔT + on T_{min} at 8 mm mean? That the top few cm of ☿ surface cool + heat rapidly, + that the surface temp. of ☿ darkside gets well below the boiling pt. of water? If ☿ nonsynchronously rotating, ~~might~~ be condensation at night.

[N.B.: in SLAOV, we assumed $T_{ground} \approx T_{air\ near\ ground}$. Not true on ⊕. If air mud cooler, do not need to continue sublimate to as high temps., + ∴ get somewhat lower pressure. In Mintz, ~~treatment~~ the diff. in temp. was ∝ diurnal cooling of the surface.
Preceding remarks suggest surface temp. more nearly given by Kuiper + Schmidt + ∴ may be even higher than used in SLAOV. But if syn. rot., no diurnal cooling, + temp. diff. between air + surf. disappears. This is especially true if surface heated by winds! O.K.]

If ☿ syn. rotating, a straightforward interp. of 8 mm data means night surface is cooler than it is ~1 meter down. The temp. gradient required is $\Delta T/\Delta z = 300°K/100\,cm = 3\,K°/cm$. Impossible. But, of course, lower 8 mm temp. is in part due to atmos. absorption. Yet strong phase effect at 8 mm + the other data does suggest somewhat lower temps. at 8 mm. ~~[illegible]~~

On previous p. we saw rm of $T_B°(\lambda)$. What of $T_B°(\lambda)$?

Mayer et al: $T_B°(3\,cm) = 547°K$.

Drake $T_B°(10\,cm) = 540°K$.

~~Observed~~ At short λ, atmos. absorption obscures results. ∴ ~~[illegible]~~ darkside T same at 3 + 10 cm. Brightside T ~~[illegible]~~ poss. a little greater at 3 cm than at 10. ∴ no more condensation than in SLAOV.

Suppose 3 cm data comes from 3 cm + 10 cm from 10 cm.
∴ gradient is ~ 68°/7 cm = 10°/cm too much. Actually can compute gradient from the eq., + observed ΔT's.
10 cm λ gives ΔZ ~ 20 cm. 3 cm λ gives ΔZ ~ 40 cm.
∴ 68°/40 cm ~ 1.5°/cm Very steep gradient. If const., 1 meter down, temp has dropped 150°, + a few meters down it is bearable! [Sahara Conds. ~ ?]

Better treatment: Derive ΔT/ΔZ directly:

$$\frac{\Delta T}{\Delta z} = -\frac{S(0)}{\rho\, c_p K_s} = \frac{\sigma (234)^4}{4\times10^{-4}} = \frac{1.7\times10^{-2}\ \text{watt cm}^{-2}}{4\times10^{-4}\times 4.2}$$

$$= \frac{1.7\times10^{5}\ \text{erg cm}^{-2}}{4\times10^{-4}} = \frac{1.7\times10^{5}}{4.2\times10^{7}\times4\times10^{-4}}$$

$$= 0.1\times10^{2} = -10\ ^\circ/\text{cm. at surface.}$$

∴ diff λ's, coming from diff depths, will have diff temps, ~~an~~ being greater at shorter λ's on bright side. This is observed. If derived ΔT/ΔZ is good a little way down, we can find for which depth various λ's arise:

$$T(\lambda = 3\text{cm}) - T(\lambda = 10\text{cm}) = 33\,K^\circ.$$

∴ separated by ~3 cm.

Surface ~ 3 cm above λ=3 cm level ∴ T(surface) ~ 720°K, making no emissivity allowance.

Now that $\frac{\Delta T}{\Delta z}$ is known, better to compute skin depth as ff.:

We desire a relation between Δt + ΔZ:

$$\frac{\Delta T}{\Delta z} = -\frac{S(0)}{\rho c_p K_s}$$

$$\frac{\Delta T}{\Delta t} = \frac{S(0)^2}{\rho^2 c_p^2 K_s}$$

$$\therefore \frac{\Delta z}{\Delta t} = -\frac{\dfrac{S(0)^2}{\rho^2 c_p^2 K_s}}{\dfrac{S(0)}{\rho c_p K_s}} = \frac{S(0)}{\rho c_p}$$

$$\frac{\Delta T}{\Delta t} = K_s \left(\frac{\Delta T}{\Delta z}\right)^2$$

If ☿ rotates non-synchronously, then the phase effect can be interpreted as nocturnal cooling. Then, if we know the synodic period, we know $(\partial T/\partial t)_\lambda$ + $\therefore (\partial T/\partial z)_\lambda$. Alternatively, we do know $(\partial T/\partial z)_\lambda$ from the flux. $\therefore$ we can derive the period of rotation; this we have already done in SL 40V. What we need is a relation which does not involve the period of rotation directly. This is

$$\frac{\Delta T}{\Delta z} = -\frac{S(0)}{\rho c_p K_s}$$

Now for a fixed λ, what is the appropriate $S(0)$? From, say, 10 cm depth, it is certainly not $\sigma(600)^4$ because the ir won't escape from the rock. It must be $\sigma(234)^4$, the rock temp. gradient adjusting itself to the net flux from the surface, at least down to the skin depth. This gives a gradient of $-10\,K°/cm$.

Consider a sidereal period of 200^d. $\therefore \Delta t_{syn} \sim 2000^d$ $= 1.73 \times 10^8$ secs.

$$\Delta z_{3.n} = \sqrt{K_s \, \Delta t \, \Delta T}$$

$$= (2\times10^{-3} \times 1.73\times10^{8} \times 1.46\times10^{2})^{1/2}$$

$$= (5.04\times10^{7})^{\frac{1}{2}} = (50.4)^{\frac{1}{2}}\times10^{3} = 7.1\times10^{3}\ \text{cm}$$

$$= 70 \text{ meters}$$

In SLAV, we already showed this doesn't work for long periods of rot.

No, diff ΔT's are getting mixed. First use Newton's law at surface. This gives $(\Delta z/\Delta T)^{-1} \sim -10\ K^\circ/\text{cm}$ at surface. Then,

$$\frac{\Delta T}{\Delta t} = K_s \left(\frac{\Delta T}{\Delta z}\right)^2$$

The steeper the gradient the faster the cooling.

$$= 2\times10^{-3} \times 10^{2} = 0.2\ K^\circ/\text{sec.}$$

Absurdly big.

Source of error:

$$\frac{\partial^2 T}{\partial z^2} \longrightarrow \frac{\Delta T}{(\Delta z)^2}, \text{ not } \frac{(\Delta T)^2}{(\Delta z)^2}$$

∴ rederiv.:

One-dimensional eqs for slab:

$$\frac{\partial T}{\partial t} = \kappa_s \frac{\partial^2 T}{\partial z^2} \quad ; \quad S = -\rho c_p \kappa_s \frac{\partial T}{\partial z}$$

$T = 0$ at ∞

$$\frac{\Delta T}{\Delta t} = \kappa_s \frac{\Delta T}{(\Delta z)^2} \quad ; \quad S(0) = -\rho c_p \kappa_s \frac{\Delta T}{\Delta z}$$

$$\Delta z = -\rho c_p \kappa_s \frac{\Delta T}{S(0)}$$

$$\therefore \frac{\Delta T}{\Delta t} = \frac{\kappa_s \Delta T \, S^2(0)}{\rho^2 c_p^2 \kappa_s^2 \Delta T^2} = \frac{S(0)^2}{\rho^2 c_p^2 \kappa_s \Delta T}$$

$$\therefore \Delta T = \frac{S(0)}{\rho c_p \kappa_s^{1/2}} (\Delta t)^{1/2}$$

or

$$\boxed{\tau = \rho^2 c_p^2 \kappa_s \left[\Delta T / S(0)\right]^2} \qquad (7)$$

which is within a factor $2/\sqrt{\pi}$ of the exact solution. (Brunt)

Also,

$$\boxed{\Delta z = (\kappa_s \Delta t)^{1/2},}$$

from diff. eq. alone.

relating the conduction skin depth to the time of cooling. Dibson (Ap. J. 133: 1075, 1961) uses a value $2\sqrt{\pi}$ larger; Dicke (Science) a value $\sqrt{2\pi}$ smaller.

Note for ♀ that moderate phase effects and long synodic periods are consistent if the surface is a turbulent liquid. But the radar return dielectric consts. rule this out.

Reanalysis

Gibson (Ap. J., 133 : 1072, 1961)

Observed the lunar eclipse of Mar., 1960, & found that at 8.6 mm, ΔT during the eclipse was $\leq 1\,K^\circ$. Time of eclipse ~ 3 hrs. = 1.07×10^4 secs.

To analyze the results, he uses an analysis by Wesselink, based on a skin depth

$$l = 2\sqrt{\pi}\left(\frac{k\,\tau_{syn}}{\rho\, c}\right)^{1/2}$$

Using $\sqrt{k\rho c} \sim 10^{-3}$ in cal & cgs units, he must make assumptions on ρ & c to get l. He takes $\rho = 2.0\,gm\,cm^{-3}$ & $c = 0.2\,cal\,gm^{-1}$. More likely that $\rho \sim 0.2\,gm\,cm^{-3}$ ∴ k 10× larger & ρ 10× smaller. ∴ l 10× larger & all Gibson's depths should be increased by factor ~ 10.

To gain agreement between ir & these 8.6 mm results, Gibson needs 2 layers, the upper one thin & essentially transparent to radio wavelengths. He derives a value of the attenuation coeff. $\alpha \sim 0.02\,cm^{-1}$. We find $\alpha \sim 0.002\,cm^{-1}$. The electrical conductivity

$$\sigma \propto \alpha\sqrt{\text{dielectric const}}$$

Taking dielectric const ~ 1 instead of 7 as Gibson found, we find

$$\sigma \sim \frac{1.8\times10^{-1}}{\sqrt{5}\times 10} = 8\times10^{-3}\ \text{mho/meter}$$

This brings obs. back into agreement w. surface electrical conductivities. The high conductivity layer disappears (unfortunately).

To match obs., requires $C = \alpha\left(\frac{k\tau}{\rho c}\right)^{\frac{1}{2}} = 5$. The α is now smaller by 10, $(\ )^{\frac{1}{2}}$ larger by 10.

Cudlnak says his independently measured dielectric const. & polar darkening law remove necessity for Piddington & Minnett's 2-layer model & require mirror comp. down to a few meters w. $\rho \sim 0.4$ to $0.8\,gm\,cm^{-3}$ & dielectric const ~ 1.1 to 1.2.

Notes from "Development of An Instrumental System to Detect + Identify Extraterrestrial Life," Melpar Tech. Prop. P2149-047, Oct., 1962

⊕ atmosphere contains ~ 300 viable organisms m^{-3} (0.3 per liter). From absorption studies at 260 mμ, ∃ ~ 10^{-9} gm protein + 10^{-9} gm nucleic acids per liter of terrestrial air. At 1 inch depth in typical ⊕ soil no. bacteria gm^{-1} ~ 7.3×10^6. At 30 in., 1×10^5 gm^{-1} (Waxman, S. "Soil Microbiology," p. 34, 40, Wiley, 1952). This proposal assumes ∂ conc. both in atmosphere + in soil < 0.1 those for ⊕. ∴ conc. mechanism needed. Flotation is ok, but some inorganics will float up, + even trace quantities of inhibitors will block enzyme reactions. A better separation can be obtained by electro-magnetophoresis. Enzymes chosen for detection are dehydrogenases, for which a simple colorimetric assay exists through the reduction of tetrazolium salts.

Levine, S, Stevenson, H. J. B., + R. H. Bordner (Science 118: 141, 1953) report peptide absorption at 6.05 + 6.45 μ; nucleic acid absorption at 8.0 + 8.1 μ; + carbohydrate + nucleic acid absorption giving a deep bend between 8.6 + 10.0 μ.

Gas chromatography is not confused by inorganic contaminants. Can be used to distinguish families of bacteria, or diff. growth stages of the same bacteria. Some indicate that more complex organisms have greater conc. of higher molec. wt. methyl esters (above C_{20})

Comb. of approaches gives far more info. than 1 alone.

On the Interpretation of the Sinton Bands

Original frequency assignment: doublet w. separation ~0.1 μ, centered near 3.46 μ. The least squares fit (Ap. D. 126: 231, 1957; S's Fig. 3) suggests 2 bands, at 3.43 μ w. Δλ ~ 0.03 μ, + at 3.52 μ w. Δλ ~ 0.01 μ. No scan as far as 3.67 μ.

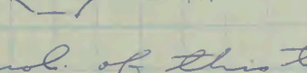
Second frequency assignment (Science, 130: 1234, 1959) Figure 1 shows weak feature at 3.43 μ; stronger features at 3.56 μ + 3.67 μ. Note immediately that here the 3.43 μ feature is weakest; in Ap. D, it was strongest. But Figure 2 shows bands at 3.46 μ, 3.59 μ, + 3.70 μ; this fig. has been divided by the solar spectrum, where Fig. 1 has not. Here 3.43 μ feature is as intense as others.

Third frequency assignment (Science, , 1961). 3.45 μ, 3.58 μ, 3.69 μ.

Now the 3.69 μ feature was not easily reproduced by Sinton in reflection spectroscopy of terrestrial vegetation. In Colthup's correlation chart (Brugel, pp. 257-3), the only org. mol. listed which has more than weak absorption near 3.69 μ are the aldehydes, both aliphatic,

$$-CH_2-CHO,$$

and aromatic,



The simplest mol. of this type is acetaldehyde,

$$CH_3CHO.$$

It may be that Sinton was misled by such a correlation chart, + took CHO to be a characteristic carbohydrate group. These same aldehydes also absorb as short as 3.45 μ; ∴ one poss. is that all three bands arise from aldehyde transitions. But there is one other category of m + s bands between 3.33 μ + 3.57 μ, CH stretch of alkanes [C-single bonds] + alkenes [C-double bonds].

Note applicability to search for carbonates in the clouds of ♀. In Sinton's ♀ spectra, ∃ no strong features between 3.33 μ + 3.60 μ. There does exist a weak dip at 3.63 μ, but Sinton is uncertain of reliability even of stronger features at 3.15 μ + 3.27 μ. Note, however, hint of a very broad continuous absorption between 3.34 μ + 3.70 μ. This might, conceivably, be due to many diff. carbonates together. But since the bands are sharp, even this seems unlikely (remember, for ♀, Sinton's resolution is better than 0.03 μ.) ∃ suspicion that variable 11.2 μ feature of Sinton + Strong is due to carbonates. If ∃ extensive carbonate absorption in 8-13 μ + at 3.75 μ, but transparency at 7820 Å, carbonates may be what we want for ♀ clouds.

But there do exist nonorganic possibilities:

C. Schaefer + F. Matossi, "Das Ultrarote Spektrum," Berlin: Springer, 1930, pp. 338 et. seq.:

Magnesite ($MgCO_3$)	strong, sharp bands	3.3 μ,	3.5 μ
Limestone ($CaCO_3$)	" " "	3.3 μ,	3.5 μ
Dolomite ($MgCO_3 + CaCO_3$)	" " "	3.36 μ,	3.5 μ
Witherite ($BaCO_3$)	" " "	3.45 μ,	3.52 μ
Cerussite ($PbCO_3$)	" " "	3.6 μ,	3.65 μ
Siderite ($FeCO_3$)	strong, broad band	3.4 μ.	

∴ it does not seem impossible, within the uncertainties of Sinton's λ-assignments that the bands result from some linear superposition of carbonate absorptions. Schaefer + Matossi's samples were in the form of thin crystals, 0.1 mm to 0.25 mm in thickness. This is just the suspected polarimetric size range of Martian surface material. But note that pulverization does alter the absorption spectrum.

I had previously thought that observation of seasonal changes in ☌ spectra would at least prove existence of life there. But hydration shifts, removes, or adds bands in this spectral region (J. M. Hunt, M. P. Wisherd, + L. C. Bonham, Anal. Chem. 22: 1478, 1950).

Note, however, that simple laboratory experience shows it is much easier to obtain spectra of organic functional groups than of inorganic functional groups in this wavelength region.

One other possibility which deserves some attention is the existence of HDO absorption between 3.52 μ + 3.85 μ. ∃ a strong HDO Q branch at ~ 3.67 μ. But Sinton's spectra quite clearly show this 3.69 μ feature in the dark areas, + not in the bright. Such a fractionation of HDO/H_2O, either biologically or abiologically, seems rather unlikely.

If the Sinton bands were as originally assigned, what checks exist? The aldehyde bands near 3.6 μ are listed by Colthup as medium. Strong bands exist between 1700 + 1900 cm⁻¹; i.e., between 5.25 + 5.90 μ. Other medium strength aldehydes lie between 800 + 1500 cm⁻¹; i.e., 6.7 + 12.5 μ. High res. spectra in 8 - 13 μ window w. good S/N might show them, but we expect m emission features to be less evident than m absorption features ∴ the failure of Sinton + Strong to see such window features is not a strong objection to the presence of aldehydes. The 5.25 to 5.90 μ region seems to be the best test.

In the same vein, the CH stretch of alkanes + alkenes is listed as m + s; alkane + alkene deformation features of similar intensity exist between 1470 + 1340 cm⁻¹; i.e., 6.8 + 7.5 μ. ∃ m alkene features near 1650 cm⁻¹ = 6.1 μ. [v. also compilation by Siebel; Beringer Fig. 102, p. 249.].

Now each feature in the 3.5 μ region has been attributed to stretch; i.e., ∃ overtones at shorter wavelengths. The overtones are weaker, sometimes by as much as factor 10; but our S/N for Stratoscope II IR is up by factor of 10 at 1.75 μ cf. 3.5 μ. ∴ we expect that if we see Sinton bands, we should see their first overtones.

W. Kaye (Spectrochimica Acta, 6 : 257, 1954) has listed these overtones for a relatively (when compared, e.g., w. Colthup's compilation) small no. of compounds. E.g., $-CH_3$, which has stretching mode at 3.4 - 3.5 μ, also absorbs at

2.2 - 2.4 μ, 1.7 - 1.8 μ, 1.1 - 1.2 μ, 0.75 μ, + 0.9 μ.

Clearly, 1.7 - 1.8 μ is first overtone of 3.4 - 3.5 μ
1.1 - 1.2 μ " " " " 2.2 - 2.4 μ

0.9 μ seems to be second overtone, + would require phy. infrared spectroscopy.

At least some aldehydes are shown by Vlage to have –CH overtones at 2.1 μ + 0.95 μ.

From the ground, ∃, on fairly dry days, windows near 6200 cm⁻¹, 7800 cm⁻¹, + 10,000 cm⁻¹, + partial transmission near 4600 cm⁻¹. ∴ window regions are:

1.0 μ 1.3 μ 1.6 μ 2.2 μ	corresponding to	0.9–1.0 μ CH second overtone 1.1–1.2 μ CH overtone 1.7–1.8 μ CH overtone 2.2–2.4 μ CH fundamental

From balloon altitudes, the 5.25 – 5.90 μ aldehyde region is completely clear, + the 6.8 – 7.5 μ hydrocarbon deformation region is almost completely clear, having only a small residual N_2O absorption [I. Strong, Pub. 414, p. 90].

From ground, note also ∃ NH + NH_2 near 1.0 μ + poss. near 1.0 μ (for = NH). All the bands at λ > 1.0 μ have characteristic widths ~0.05 to 0.1 μ. ∴ our resolution may be adequate.

Absorption By 9.4 μ + 10.4 μ CO_2 Bands in Planetary Atmospheres

Laboratory data on the 1064 + 961 cm^{-1} bands appear in "Absorption by Carbon Dioxide," by D. E. Burch, D. Gryvnak, + D. Williams in AFCRL Res. Rept. 62-698, 1962. Tracings show CO_2 abundance w + equiv. pressure P_e is running parameter. P_e arises because CO_2 is more efficient in coll. broadening CO_2 than are other molecules. For a CO_2-N_2 mixture,

$$P_e = P + 0.30\, p_{CO_2},$$

where P is total pressure, + p_{CO_2} is partial pressure CO_2.

These bands are "hot bands," + their absorption is a strong function of the temp. We will neglect this approx. for the moment, + correct it later. The lab. measurements were performed at rm. temp.

For ♀, take, for cloudtops, $P \sim 0.7$ atm. $p_{CO_2} = 0.05 \times \frac{44}{29} \times 0.7 = 5.3\times10^{-2}$ atm.

$$\therefore P_e = 0.70 + 0.3\times0.05 \simeq 0.70 \text{ atm} \simeq 530 \text{ mm Hg}.$$

Now for w. For air, 1 atm → 8 km-atm. ∴ 5.3×10^{-2} atm → $5.3\times10^{-2}\times8\times\left(\frac{29}{44}\right) = 2.8\times10^{-1}$ km atm = 280 m-atm above clouds. Plus slight g correction.

For these bands, the strong line approx. is good, so wP_e is the determining parameter. With w in cm-atm, + P_e in mm Hg, $wP_e = 2.8\times10^4 \times 5.3\times10^2 = 1.5\times10^7$. In Fig. 16 of ref. 7 a curve for $w = 5395$ cm-atm, + $P_e = 3655$ mm Hg. $5.40\times10^3 \times 3.70\times10^3 = 2.0\times10^7$. 7 another curve for $w = 4415$ cm-atm + $P_e = 2990$ mm Hg. $wP_e = 4.42\times2.99\times10^6 = 1.32\times10^7$. ∴ ♀ can fall between. The resulting curves show very substantial absorption in the 9.4 μ + 10.4 μ bands. Their absence on ♀ must then be due, either (1) to the low T's, or (2) to a high cloud opaque at 8–13 μ, but transparent at 0.8 μ.

If Aerosphere upper atmosphere right, might upper haze layer be thick enough at night to be opaque in visible (accounts for low pressures) and thin enough so lower clouds show through in visible during day (various daytime evidence for high pressures)? Also, with top of lower clouds at higher temps, we don't get quite as much as >30 atm pressure; Kuiper Feels Spinrad's data shows P ~10 atm.

The temp. effect has been studied between 23 + 77°C, + an semi-empirical interpolation was derived. Assuming we can extrapolate to low temps., we can compute the effect for low temps.

For the 9.4 μ band,

$$\begin{array}{r} 4.35 \\ 3.34 \\ \hline 1.01 \end{array}$$

$$W_e = W \exp\left[1950\left(\frac{1}{299} - \frac{1}{T}\right)\right]$$

For $T = 230°K$, $\frac{1}{299} - \frac{1}{230} = (3.34 - 4.35)\times 10^{-3} = -1.01\times 10^{-3}$.

$-1950\times 1.01\times 10^{-3} = -1.97$. ~~$e^{-19.7} = e^{-10}\times e^{-9.7}$~~

~~$= 4.5\times 10^{-5}\times 6.3\times 10^{-5} = 2.83\times 10^{-9}$, + so : unfortunately very similar relations apply to the 10.4 μ band.~~

$e^{-1.97} = 0.14$. $\therefore W_e = 0.14\times 2.8\times 10^4 = 3.9\times 10^3$ cm-atm

$W_e P_e = 3.9\times 10^3\times 5.3\times 10^2 = 2.1\times 10^6$. ∃ a curve on p. 187 of exp. ref. w. $WP = 1.42\times 10^3\times 9.6\times 10^2 = 1.4\times 10^6$. This gives abs. ~ 40%, avg. abs. ~ 20%. Sinton + Strong should have seen this. To reduce avg. abs. below 10%, we need $W_e P_e \sim 2\times 10^4$. By reducing P at cloudtops by factor 10, + reducing avg T in correction eq. above, this can almost be done. But if brightness cloudtop pressures are ~ 0.7 atm, it seems that ~~either~~ we need that high-altitude cloud. (Very similar considerations apply to the 10.4 μ band.) The cloud must be opaque in far IR, + situated at 0.1 to 0.01 the pressure level of the lower cloud. Cf. Aeolosphere model.

$$\begin{array}{r} 4.55 \\ -3.34 \\ \hline 1.21 \end{array}$$

For $T = 220°K$, $\frac{1}{299} - \frac{1}{220} = (3.34 - 4.55)\times 10^{-3} = -1.21\times 10^{-3}$

$-1950\times 1.21\times 10^{-3} = -2.36$. $e^{-2.36} = 0.095$

$\therefore W_e = 0.095\times 2.8\times 10^4 = 2.7\times 10^3$ cm-atm

$\therefore W_e P_e = 2.7\times 10^3\times 5.3\times 10^2 = 1.4\times 10^6 \Rightarrow$ 40% abs. max.
To explain Sinton + Strong, need $P_e \sim 7\times 10^{-3}$ atm.
But then CO_2 mixing ratio is very diff.; much greater.

This approach has the advantage that it only demands α + T_c w. no specification of cloud models, or relation to other P_c determinations. T_c we know directly from measurements at the same $\Delta\lambda$; α is a presumed const,

Can't we extricate ourselves by taking $P_e \ll 10^{-3}$ atm, viz., at the ~~[illegible]~~ occultation level?

Take const. CO_2 volume mixing ratio α. $\therefore \mu_{tot} = \alpha \mu_{CO_2} + (1-\alpha)\mu_{N_2} = 28.8$.

W_{CO_2} = mass CO_2 in atmosphere in gm cm^{-2} $\times [\mu_{tot} \times m_H \times L]^{-1}$

Mass CO_2 in atm = mass atm $\times \alpha \times \frac{\mu_{CO_2}}{\mu_{tot}}$

Mass atm = 0.7 atm = 0.7×1000 gm cm^{-2} = ~~[illegible]~~ $P_c \times 1000$

$$\therefore W_{CO_2} = 10^3 P_c (\text{atm})\, \alpha \frac{\mu_{CO_2}}{\mu_{tot}} \frac{1}{\mu_{tot} m_H L}$$

$$= 10^3 \times \frac{44}{(28.8)^2} \frac{P_c\, \alpha}{1.2 \times 10^{-24} \times 3 \times 10^{19}}$$

$$= 1.06 \times 10^6\, P_c\, \alpha$$

$\therefore \alpha = 0.05 \Rightarrow W_{CO_2} = 5.3 \times 10^4\, P_c$ cm-atm

$P_c = 0.7$ atm $\Rightarrow W_{CO_2} = 3.7 \times 10^4$ cm-atm ✓

$P_e \simeq P_c$ for our α.

$$\therefore \boxed{W_e P_e = 10^6 \alpha P_c^2 \exp\left[b\left(\frac{1}{299} - \frac{1}{T_c}\right)\right],} \quad P_c \text{ in atm}$$

where $b = 1950$°K for 1064 cm^{-1} + $b = 2040$°K for 961 cm^{-1}.

Roughly, exp[ ] $\simeq 0.1$ for our temps.

$\therefore W_e P_e \simeq 10^5 \alpha P_c^2$.

$\alpha = 0.05 \Rightarrow W_e P_e \simeq 5 \times 10^3 P_c^2$, with P_c in atm.

For P_c in mm Hg,

~~$W_e P_e \simeq 5 \times 10^3 \times (76)^2 \times 10^{-4} P_c^2 = 2.9 \times 10^{-9} P_c^2$.~~

~~To reduce abs. below 10% we need $W_e P_e < 2 \times 10^{-4}$ (cm-atm)(mm Hg).~~

$\cancel{2.9\times10^{9}\,P_c^{2} < 2\times10^{4}}$

$\cancel{\therefore P_c^{2} < 6.9\times10^{-6}}$

$\cancel{\therefore P_c < 260 \text{ mm Hg}}$

$\cancel{\therefore P_c < 2.6\times10^{-3} \text{ mm Hg} = \frac{2.6\times10^{-3}}{4.6\times10^{2}} = 3.4\times10^{-6} \text{ atm}}$

~~i.e. P above the critical level.~~

we recall that 1 mm Hg = $\frac{1}{760}$ atm (as we just found).

$\therefore w_e P_e = [5\times10^{3}/(7.0)^{2}\times10^{4}]\,P_c^{2} = 8.7\times10^{-3}\,P_c^{2}$ (mm Hg)

To reduce absorption below 10% on the avg., we need $w_e P_e < 2\times10^{4}$ (cm-atm)(mm Hg).

$\therefore 8.7\times10^{-3}\,P_c^{2} < \cancel{2\times10^{4}}\ 10^{5}$ these units.

$\therefore P_c^{2} < \cancel{2.3\times10^{6}}\ 1.15\times10^{7} = 11.5\times10^{6}$

$\therefore P_c < \cancel{1.5\times10^{3} \text{ mm Hg}}\ 3\times10^{3}$ mm Hg.

~~Some numerical error still.~~

Limb Darkening as a Function of Wavelength in a Subadiabatic Atmosphere (Emission Region)

The formal solution of the equation of radiative transfer for a plane parallel atmosphere is expressed in terms of the specific intensity $I_\nu(\theta, \tau_\nu)$ at a given optical depth τ_ν, and at a angle θ, from the normal. $I_\nu(\theta, \tau_\nu)$ is given by

$$I_\nu(\theta, \tau_\nu) = I_\nu(\theta, 0)\, e^{\tau_\nu \sec\theta} + \sec\theta \int_{\tau_\nu}^{\infty} S_\nu\, e^{(\tau_\nu - \tau_\nu')\sec\theta}\, d\tau_\nu' .$$

For outgoing radiation at the top of a planetary atmosphere,

$$I_\nu(\theta, 0) = \sec\theta \int_0^\infty S_\nu(\tau_\nu')\, e^{-\tau_\nu' \sec\theta}\, d\tau_\nu' .$$

This eq. explains limb-darkening. At the center of the planetary disk, $\theta \simeq 0$, $\cos\theta \simeq 1$, $\sec\theta \simeq 1$, $e^{-\tau_\nu' \sec\theta} \simeq e^{-\tau_\nu'}$. As we look towards the limb, we are observing regions where $\theta > 0$, $\cos\theta < 1$, $\sec\theta > 1$, $e^{-\tau_\nu' \sec\theta} \ll e^{-\tau_\nu'}$, and the intensity at the limb is less.

Where local thermodynamic equilibrium applies, the Kirchhoff-Planck equation,

$$j_\nu = \kappa_\nu B_\nu(T)$$

holds. The source function is then

$$S_\nu = \frac{\kappa_\nu}{\kappa_\nu + \sigma_\nu} B_\nu(T) + \frac{\sigma_\nu}{\kappa_\nu + \sigma_\nu} \int I_\nu\, p(\cos\Theta) \frac{d\Omega'}{4\pi}$$

where κ_ν + σ_ν are the absorption + scattering coefficients, respectively, + $p(\cos\Theta)$ is the phase function for single

scattering.

For φ, we have no assurance that σ_ν does not dominate K_ν in the 8 – 13 μ window. It certainly dominates in the visible. We will consider two cases:

(1) $\sigma_\nu = 0 \quad \therefore S_\nu = B_\nu(T)$.

(2) $K_\nu = 0$, + the radiation field is isotropic.
$\therefore S_\nu = \varpi_0 I_\nu$, where ϖ_0 is the albedo for single scattering.

Case (1).

$$I_\nu(\theta, 0) = \sec\theta \int_0^\infty B_\nu[T(\tau_\nu')]\, e^{-\tau_\nu' \sec\theta}\, d\tau_\nu'$$

Now in any given wavelength interval, the Planck function can be represented as

$$B_\nu = K T^m$$

where m is given by Drayson's chart, or derived from C. D. Reed (Rev. Sc. Inst. 31: 886, 1960). We will later look into the question of replacing B_ν by B_λ.

$$\therefore I_\nu(\theta, 0) = K \sec\theta \int_0^\infty T^m(\tau_\nu')\, e^{-\tau_\nu' \sec\theta}\, d\tau_\nu'.$$

Now we wish to convert $d\tau_\nu'$ + τ_ν' to a function of T. We assume that the absorption coefficient, K_ν, is proportional to the sth power of the pressure. Usually, ∃ two cases, $s = 1$ or $s = 1/2$. ~~Further, we take~~

$$\therefore K_\nu = C' P^s.$$

Further, we take a subadiabatic atmosphere.

$$P \propto T^{\eta\gamma/\gamma-1}.$$

Also, $\rho \propto P/T$.

$$\therefore \rho \propto T^{\frac{\eta\gamma}{\gamma-1}-1} = T^{\frac{\eta\gamma-(\gamma-1)}{\gamma-1}} = T^{\frac{(\eta-1)\gamma+1}{\gamma-1}}.$$

When $\eta=1$, this reduces properly to $\rho \propto T^{\frac{1}{\gamma-1}}$.

$$\therefore d\tau_v' = -\gamma_v \rho\, dr = -C\left[T^{\frac{\eta\gamma}{\gamma-1}}\right]^s \left[T^{\frac{(\eta-1)\gamma+1}{\gamma-1}}\right] dr$$

$$= -C\, T^{\frac{\eta\gamma s}{\gamma-1}+\frac{(\eta-1)\gamma+1}{\gamma-1}}\, dr.$$

$$\therefore d\tau_v' = -C\, T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1}\, dr$$

Now the const. adiabatic lapse rate is

$$\frac{dT}{dr} = -\Gamma.$$

$\therefore dr = -dT/\Gamma$, and

$$d\tau_v' = +C\Gamma^{-1}\, T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1}\, dT$$

But from the first eq. this p.,

$$d\rho \propto T^{\frac{\eta\gamma}{\gamma-1}-2} = T^{\frac{\eta\gamma-2(\gamma-1)}{\gamma-1}} = T^{\frac{(\eta-2)\gamma+2}{\gamma-1}}\, dT.$$

$$\therefore dT \propto T^{\frac{(2-\eta)\gamma-2}{\gamma-1}}\, d\rho.$$

$$\therefore d\tau_v' \propto T^{[\eta\gamma s+(\eta-1)\gamma+1+(2-\eta)\gamma-2]/\gamma-1}\, d\rho$$

$$= T^{[\eta\gamma s+\eta\gamma-\eta\gamma-\gamma+2\gamma+1-2]/\gamma-1}\, d\rho$$

$$= T^{[\eta\gamma s+\gamma-1]/\gamma-1}\, d\rho.$$

$\therefore \tau_v' \propto$

$$d\tau_\nu' = +C\Gamma^{-1} T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1}\, dT$$

$$\therefore \tau_\nu' = \text{const.}\ T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1+1}$$

$$= \text{const.}\ T^{[\eta\gamma s+\eta\gamma-\gamma+1+\gamma-1]/\gamma-1}$$

$$= \text{const.}\ T^{[\eta\gamma(1+s)]/\gamma-1} + \text{const.}$$

$$\therefore I_\nu(\theta,0) = \text{const}\ \sec\theta$$

$$\cdot \int_0^\infty T^m e^{-\kappa T^{[\eta\gamma(1+s)]/\gamma-1}\sec\theta}\, T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1}\, dT$$

Let $x = T^{[\eta\gamma(1+s)]/\gamma-1}$.

$$\therefore dx = [\eta\gamma(1+s)]/\gamma-1\ T^{[\eta\gamma s+\eta\gamma-\gamma+1]/\gamma-1}\, dT$$

$$\therefore T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1}\, dT = \frac{\gamma-1}{\eta\gamma(1+s)}\, dx$$

$$\therefore I_\nu(\theta,0) = \text{const}\ \sec\theta \int_0^\infty T^m e^{-\kappa\sec\theta\, x}\, dx$$

Now $T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1+m}\, dT \propto$ what?

$$T = x^{\gamma-1/\eta\gamma(1+s)} \quad \therefore T^m = x^{m(\gamma-1)/\eta\gamma(1+s)}$$

$$\therefore I_\nu(\theta,0) = \text{const}\ \sec\theta \int_0^\infty e^{-\kappa\sec\theta\, x}\, x^{m(\gamma-1)/\eta\gamma(1+s)}\, dx.$$

This eq. checks properly: $m=4$, $\eta=1$, $s=1$ exponent of x becomes $2(\gamma-1)/\gamma$.

Let $\beta = m(\gamma-1)/\eta\gamma(1+s)$

This integral is listed as 568.1, p. 127 in Dwight.

$$I_\nu(\theta,0) = \text{const.}\ \sec\theta \left[e^{-\kappa \sec\theta\, x} \left\{ \frac{x^\beta}{-\kappa\sec\theta} - \frac{\beta\, x^{\beta-1}}{\kappa^2 \sec^2\theta} + - \cdots (-1)^{\beta+1} \frac{\beta!}{(-\kappa\sec\theta)^{\beta+1}} \right\} \right]_0^\infty$$

$$= \text{const.}\ \sec\theta\, (-1)^{\beta+1} \frac{\beta!}{(-\kappa\sec\theta)^{\beta+1}}$$

$$= \text{const.}\ (-\kappa\sec\theta)^{-(1+\beta)} \sec\theta = \text{const.}\ (\sec\theta)^{-\beta}$$

$$\therefore I_\nu(\theta,0) = \text{const.}\ (\cos\theta)^\beta$$

$$\therefore \boxed{\frac{I_\nu(\theta,0)}{I_\nu(0,0)} = (\cos\theta)^{\frac{m_1(\gamma-1)}{\eta\gamma(1+s)}}}$$

Now, values of these parameters:

$\gamma_{N_2}(15^\circ C) = 1.404$
$\gamma_{N_2}(-181^\circ C = 92^\circ K) = 1.47$ } adopt $\gamma_{N_2} = 1.42$.

$\gamma_{CO_2}(-75^\circ C = 198^\circ K) = 1.37$

$\alpha = 0.05 \Rightarrow 0.95 \times 1.42 + 0.05 \times 1.37 = 1.35 + 0.07 = 1.42$
$\alpha = 0.10 \Rightarrow 0.90 \times 1.42 + 0.10 \times 1.37 = 1.28 + 0.14 = 1.42$

$\therefore \gamma = 1.42$.

$\eta^\circ \sim 1$
$\eta^\bullet \sim 1.5$.

$S_{\text{[illegible]}} \sim 1$
~~[illegible]~~.

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(SUMMARY) SOVIET SCIENTISTS ARE USING RADIOASTRONOMY TO SEE THROUGH THE CLOUDS CONCEALING VENUS. THE LARGEST RADIOTELESCOPE AT THE PULKOVSKAYA OBSERVATORY HAS A REFLECTOR 2 BY 3 METERS LARGE. ITS RESOLVING POWER ON A THREE-CENTIMETER WAVELENGTH IS ADEQUATE TO DETERMINE THE TYPE OF DISTRIBUTION OF RADIO BRIGHTNESS ON THE DISC OF VENUS. IT HAS BEEN FOUND THAT THE DISTRIBUTION OF RADIO BRIGHTNESS ON VENUS DROPS AT THE EDGES OF THE DISC, AND THAT THE SURFACE OF THE PLANET IS VERY HOTTH WITH A TEMPERATURE OF ABOUT 300 TO 400 DEGREES CENTIGRADE. LEIGH-HUNT

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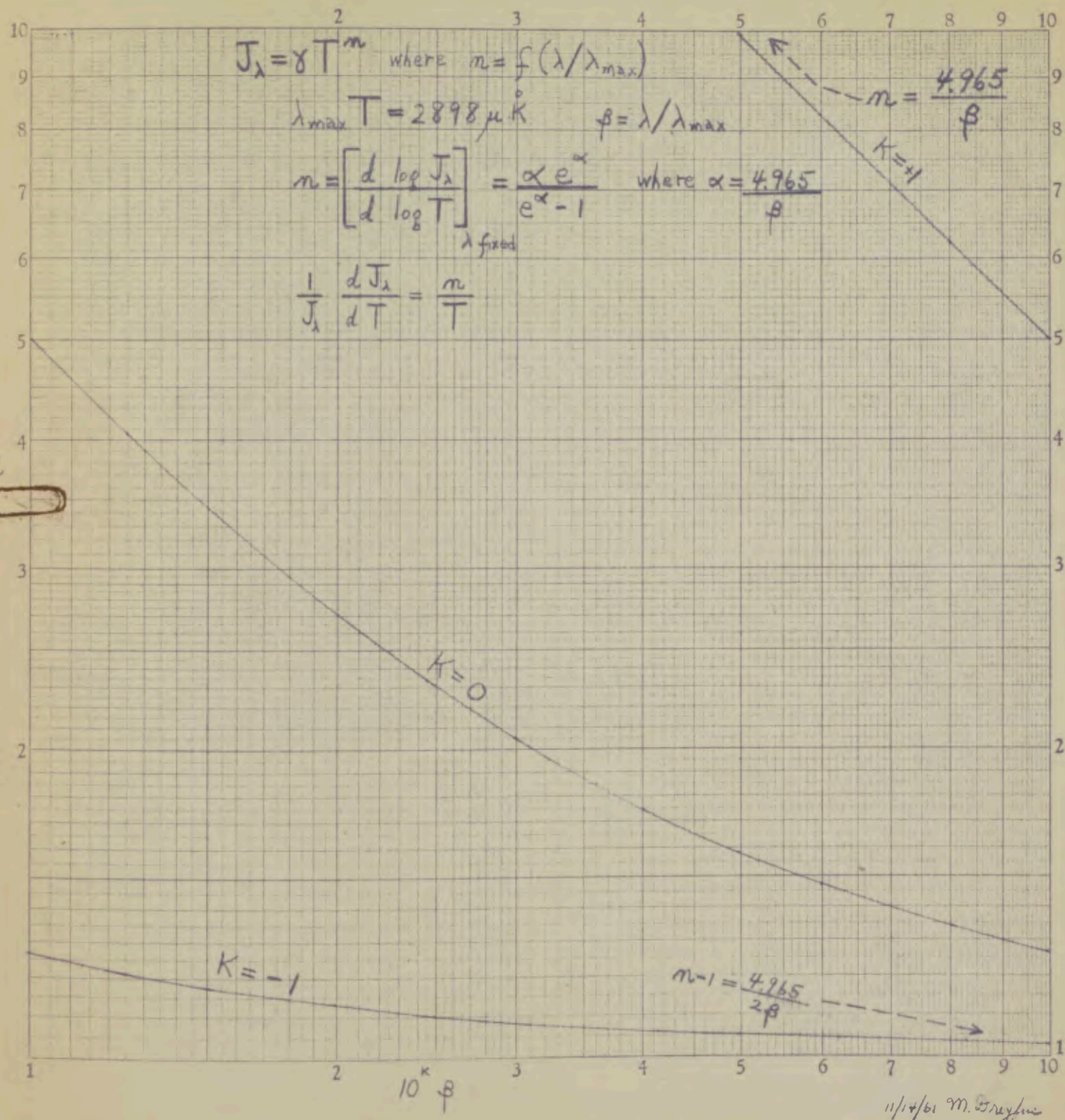
Exponential Relationship Involved In Dependence Of Blackbody Spectral Radiance J_λ On Small Changes In Absolute Temperature T (°K) At A Constant Wavelength λ

[cf. C.D. Reid - Rev. Sci. Inst., 31, 886 (1960)]



11/14/61 M. Grayfine

β	.5015	.5982	.8008	.9990	1.251	1.500	2.002	2.446	2.991	4.004	4.965	5.982	8.008	9.930	12.41	15.05	19.86	24.83
n	9.900	8.302	6.213	5.005	4.046	3.435	2.707	2.337	2.050	1.745	1.582	1.472	1.342	1.271	1.213	1.174	1.130	1.103
β	31.03	41.38	49.65	62.06	82.75	99.30												
n	1.082	1.061	1.051	1.041	1.030	1.025												

Now for m:

Drayson expresses $m = m(b)$, where $b = \lambda/\lambda_{max}$. For 225°K, $\lambda_{max} \simeq 12.8\mu$. For 220°K, $\lambda_{max} = 13.1\mu$, + for 230°K, $\lambda_{max} = 12.5\mu$. Take $\lambda_{max} = 12.8\mu$.

$\therefore$ for $\lambda = 8.4\mu$, $b = 0.69$, + $m_v = \cancel{4.09}\ 7.2$

for $\lambda = 10.35\mu$, $b = 0.81$, + $m_v = 6.15$

for $\lambda = 12.5\mu$, $b = 1$, + $m_v = 5.0$.

$\therefore$ note that, on the basis of the Planck distribution alone, we expect more limb-darkening at 10.35μ than at 8.4μ (since $|\cos\theta| \leq 1$). In addition, there is the expectation that $K(10.35\mu) >> K(8.4\mu)$, so the deriv. is more valid at 10.35μ.

$\therefore$ Numerical example:

$\gamma = 1.42$, $\eta = 1.5$, $S = 1$, $m = 6.15$.

$$\therefore \beta_v = \frac{m_v(\gamma - 1)}{\eta\gamma(1+S)} = \frac{6.15(0.42)}{1.5 \times 1.42 \times 2} = 0.61.$$

For fixed $\gamma = 1.42$, $S = 1$: $\beta = 0.148\ \frac{m}{\eta}$

η / m →	5.0	6.15	7.2
1.0	0.74	0.91	1.06
1.5	0.49	0.61	0.71
2.0	0.37	0.45	0.53

The observations of Sinton + Strong give $\beta \simeq 0.5$ in the 8-13μ window as a whole.

Suppose now we have two cloud layers. The atmosphere below the upper layer will make a contribution to $I_\nu(\theta,0)$ dependent upon the transparency + cover of the upper clouds. If the cover is f, then

$$I_\nu(\theta,0) = \text{const.} \sec\theta \left\{ f \int_0^{T_1} T^m e^{-\kappa T^{[\eta\gamma(1+s)]/\gamma-1}\sec\theta} T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1} dT + (1-f)\int_0^{T_2} T^m e^{-\kappa T^{[\eta\gamma(1+s)]/\gamma-1}\sec\theta} T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1} dT \right\}$$

Consider a cloud layer + a surface of emissivity ε. Upper cloud layer has ~~transparency~~ cover f_1. Lower layer has cover f_2. $\therefore$ a fraction $f_2(1-f_1)$ of middle cloud emission escapes, when proper allowance is made for atmospheric transparency; + a fraction $\varepsilon(1-f_1)(1-f_2)$ of surface emission escapes. In this case,

$$I_\nu(\theta,0) = \text{const.} \sec\theta \cdot \left\{ f_1 \int_0^{T_1} T^m e^{-\kappa T^{[\eta\gamma(1+s)]/\gamma-1}\sec\theta} T^{[\eta\gamma s+(\eta-1)\gamma+1]/\gamma-1} dT + f_2(1-f_1)\int_0^{T_2} T^m e^{-\kappa T^{[\ ]/\gamma-1}\sec\theta} T^{[\ ]/\gamma-1} dT + \varepsilon(1-f_1)(1-f_2)\int_0^{T_s} T^m e^{-\kappa T^{[\ ]/\gamma-1}\sec\theta} T^{[\ ]/\gamma-1} dT \right\}$$

Each integral is of the form

$$\left[e^{-\kappa \sec\theta\, x} \left\{ \frac{x^{\beta}}{-\kappa \sec\theta} - \frac{\beta x^{\beta-1}}{\kappa^2 \sec^2\theta} + - \cdots (-1)^{\beta+1} \frac{\beta!}{(-\kappa \sec\theta)^{\beta+1}} \right\} \right]_0^x$$

where $x = T^{[\eta\gamma(1+s)]/\gamma-1}$

$$\therefore \int \rightarrow \left[e^{-\kappa \sec\theta\, x} \left\{ \frac{x^{\beta}}{-\kappa \sec\theta} - \frac{\beta x^{\beta-1}}{\kappa^2 \sec^2\theta} + - \cdots (-1)^{\beta+1} \frac{\beta!}{(-\kappa \sec\theta)^{\beta+1}} \right\} - (-1)^{\beta+1} \frac{\beta!}{(-\kappa \sec\theta)^{\beta+1}} \right]$$

Now $x^{\beta} = T^{\frac{\eta\gamma(1+s)}{\gamma-1} \frac{m(\gamma-1)}{\eta\gamma(1+s)}} = T^{m}$

$$\therefore \int \rightarrow \left[e^{-\kappa \sec\theta\, T^{[\eta\gamma(1+s)]/\gamma-1}} \left\{ \frac{T^{m}}{-\kappa \sec\theta} - \frac{\beta T^{m-1}}{\kappa^2 \sec^2\theta} + - \cdots (-1)^{\beta+1} \frac{\beta!}{(-\kappa \sec\theta)^{\beta+1}} \right\} - (-1)^{\beta+1} \frac{\beta!}{(-\kappa \sec\theta)^{\beta+1}} \right]$$

Note that $\beta < 1$, $T > 1$, $\sec\theta > 1$. $\therefore$, unless we have very peculiar κ, convergence is rapid.

$$\therefore \int \rightarrow \frac{T^{m}}{-\kappa \sec\theta} e^{-\kappa \sec\theta\, T^{[\eta\gamma(1+s)]/\gamma-1}} - (-1)^{\beta+1} \frac{\beta!}{(-\kappa \sec\theta)^{\beta+1}}$$

$$\therefore\ I_0(\theta,0) = \text{const.}\ \sec\theta\ \cdot$$

$$\Bigg\{ f_1 \left[\frac{T_1^{\,m}}{-\kappa\sec\theta}\, e^{-\kappa\sec\theta\, T_1^{[\eta\gamma(1+s)]/\gamma-1}} - (-1)^{\beta+1}\frac{\beta!}{(-\kappa\sec\theta)^{\beta+1}} \right]$$

$$+ f_2(1-f_1)\left[\frac{T_2^{\,m}}{-\kappa\sec\theta}\, e^{-\kappa\sec\theta\, T_2^{[\ ]/\gamma-1}} - (-1)^{\beta+1}\frac{\beta!}{(-\kappa\sec\theta)^{\beta+1}} \right]$$

$$+ \varepsilon(1-f_1)(1-f_2)\left[\frac{T_s^{\,m}}{-\kappa\sec\theta}\, e^{-\kappa\sec\theta\, T_s^{[\ ]/\gamma-1}} - (-1)^{\beta+1}\frac{\beta!}{(-\kappa\sec\theta)^{\beta+1}} \right] \Bigg\}$$

$$= \text{const.}\ \sec\theta \cdot \frac{-1}{\kappa\sec\theta}\Big[f_1 T_1^{\,m} e^{-\kappa\sec\theta\, T_1^{[\ ]/\gamma-1}}$$

$$+ f_2(1-f_1) T_2^{\,m} e^{-\kappa\sec\theta\, T_2^{[\ ]/\gamma-1}}$$

$$+ \varepsilon(1-f_1)(1-f_2) T_s^{\,m} e^{-\kappa\sec\theta\, T_s^{[\ ]/\gamma-1}} \Big]$$

$$- \text{const.}\ \sec\theta\ (-1)^{\beta+1}\frac{\beta!}{(-\kappa\sec\theta)^{\beta+1}}\Big[f_1 + f_2(1-f_1) + \varepsilon(1-f_1)(1-f_2) \Big]$$

$$= \text{const}\Big[f_1 T_1^{\,m} e^{-\kappa\sec\theta\, T_1^{[\ ]/\gamma-1}} + f_2(1-f_1) T_2^{\,m} e^{-\kappa\sec\theta\, T_2^{[\ ]/\gamma-1}}$$

$$+ \varepsilon(1-f_1)(1-f_2) T_s^{\,m} e^{-\kappa\sec\theta\, T_s^{[\ ]/\gamma-1}} \Big]$$

$$+ \text{const}'.(\cos\theta)^{\beta}\Big[f_1 + f_2(1-f_1) + \varepsilon(1-f_1)(1-f_2) \Big]$$

Now $\eta\gamma(1+s)/\gamma-1 \simeq 1.5\times 1.42 - 2/0.42 \simeq 10$. $\therefore$ all exponential term are large, + compared with the $(\cos\theta)^{\beta}$ term can be neglected.

$$\therefore I_\nu(\theta,0) = \text{const}\left[f_1 + f_2(1-f_1) + \varepsilon(1-f_1)(1-f_2)\right](\cos\theta)^{\beta},$$

and, as before,

$$\frac{I_\nu(\theta,0)}{I_\nu(0,0)} = (\cos\theta)^{\beta}$$

where $\beta = \dfrac{m(\gamma-1)}{\eta\gamma(1+s)}$

The effect of variable cloud cover is simply to change the coefficient of $(\cos\theta)^{\beta}$. In the case $\varepsilon = 1$,

$$f_1 + f_2(1-f_1) + (1-f_1)(1-f_2)$$

$$= f_1 + f_2 - f_1 f_2 + 1 - f_2 - f_1 + f_1 f_2 = 1.$$

$$\therefore I_\nu(\theta,0) = \text{const}\ \sec\theta$$

$$\left\{ -f_1(-1)^{\beta_1+1}\frac{\beta_1!}{(-x\sec\theta)^{\beta_1+1}} \right.$$

$$-f_2(1-f_1)(-1)^{\beta_2+1}\frac{\beta_2!}{(-x\sec\theta)^{\beta_2+1}}$$

$$\left. -\varepsilon(1-f_1)(1-f_2)(-1)^{\beta_3+1}\frac{\beta_3!}{(-x\sec\theta)^{\beta_3+1}} \right\}$$

$$= K_1 f_1(\cos\theta)^{\beta_1} + K_2 f_2(1-f_1)(\cos\theta)^{\beta_2}$$

$$+ K_3\,\varepsilon(1-f_1)(1-f_2)(\cos\theta)^{\beta_3}$$

Here, $|K_1| = K^{-\beta_1 - 1}$; $|K_2| = K^{-\beta_2 - 1}$; $|K_3| = K^{-\beta_3 - 1}$.

$$\beta = \frac{m(\sigma - 1)}{\eta\, \sigma\, (1+s)}$$

$$\therefore \beta \simeq \frac{0.42 - m}{1.42 \times 1.5 \times 2} = 0.0986\, m.$$

~~$T_1 = 225°K$~~

Now for 10.35μ, we derive m & β for $T = 225°K$, $285°K$, $350°K$, $400°K$, $600°K$, $700°K$.

T	λ_{max}	λ / λ_{max}	m	β
225°K	12.8μ	0.809	6.15	0.61
285	10.2μ	1.013	4.97	0.49
350	8.3	1.25	4.05	0.40
400	7.23	1.43	3.56	0.35
600	4.79	2.16	2.55	0.25
700	4.12	2.52	2.30	0.23

To first approx, then, $K_1 = K_2 = K_3$. +

$$I_\nu(\theta, 0) = \text{const}\left[f_1 (\cos\theta)^{\beta_1} + f_2 (1 - f_1)(\cos\theta)^{\beta_2} + \varepsilon (1 - f_1)(1 - f_2)(\cos\theta)^{\beta_3} \right]$$

and

$$\frac{I_\nu(\theta, 0)}{I_\nu(0, 0)} = \frac{f_1 (\cos\theta)^{\beta_1} + f_2 (1 - f_1)(\cos\theta)^{\beta_2} + \varepsilon (1 - f_1)(1 - f_2)(\cos\theta)^{\beta_3}}{f_1 + f_2 (1 - f_1) + \varepsilon (1 - f_1)(1 - f_2)}$$

For $\varepsilon \simeq 1$, as expected, denominator $\longrightarrow 1$, and

$$\boxed{\frac{I_\nu(\theta,0)}{I_\nu(0,0)} = f_1(\cos\theta)^{\beta_1} + f_2(1-f_1)(\cos\theta)^{\beta_2} + (1-f_1)(1-f_2)(\cos\theta)^{\beta_s}}$$

E.g., $\eta = 1.5$, $\lambda = 10.35\mu$, $T_1 = 225°K$, $T_2 = 400°K$, $T_s = 600°K$, $\gamma = 1.42$, $s = 1$, $f_1 = 0.99$, $f_2 = 0.5$

$$\therefore \frac{I_\nu(\theta,0)}{I_\nu(0,0)} \simeq 0.99(\cos\theta)^{0.61} + 0.005(\cos\theta)^{0.35} + 0.005(\cos\theta)^{0.23}$$

The same parameters, except $T_2 = 285°K$, $f_1 = 0.5$, $f_2 = 0.99$:

$$\frac{I_\nu(\theta,0)}{I_\nu(0,0)} \simeq 0.5(\cos\theta)^{0.61} + 0.25(\cos\theta)^{0.49} + 0.005(\cos\theta)^{0.23}$$

Note that the terms from lower layers have smaller exponents. $\therefore$ their effect becomes relatively more pronounced at smaller values of $\cos\theta$, i.e., at larger values of θ near the limb. Thus, breaks in the clouds can account for the departure from $(\cos\theta)^{1/2}$ in the limb-darkening curve of Sinton + Strong.

Microwave limb darkening:

$\overline{T_s} \sim 640°K.$

$$\lambda_{max} = \frac{2898}{T} = 4.53\mu$$

$13.5\ mm = 1.35\times10^{0}\ cm = 1.35\times10^{4}\ \mu.$

$$\therefore b = \frac{\lambda}{\lambda_{max}} = \frac{1.35\times10^{4}}{4.53} = 2.98\times10^{3}$$

For 19 mm, $b = 4.19\times10^{3}$.

With these lg. values of b,

$$m = 1 + \frac{4.965}{2b} \simeq 1, \text{ Rayleigh-Jeans law.}$$

$\therefore$ m is much less than for IR case so limb-darkening much more moderate.

$$\frac{I_\nu(\theta,0)}{I_\nu(0,0)} = (\cos\theta)^{\beta}$$

$$\beta = \frac{m_\nu(\gamma-1)}{\gamma\eta(1+s)}$$

For $\gamma = 1.4$, $\eta = 1.5$, $s = 0$, $m = 1$,

$$\beta = \frac{0.4}{1.4\times1.5} = 0.19$$

For pressure-induced dipole transition of CO_2, $s \simeq 2$, + $\beta \simeq$ ~~0.10~~ 0.06. $\therefore$ the extreme limb-darkening found by Mariner R-2 must be a time-const. effect.

Organic Liquids on Mars

Data from E. W. Washburn, ed. Intern. Crit. Tables 3: 215 (1928). Vapor pressures:

CH_3CHO between -24.3 and $+27.5°C$:

$$\log_{10} p_{mm} = -\frac{0.05223}{T(°K)} \times 2.77 \times 10^4 + 7.820$$

At $p_{mm} = 65$ mm Hg,

$\log_{10} 65 = \cancel{4.443}\ 1.813$

$$\begin{array}{r} 7.820 \\ -1.813 \\ \hline 5.907 \end{array}$$

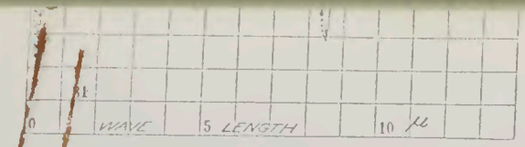
$$-5.907 = +\frac{1.445 \times 10^3}{T}$$

$$\therefore T = \frac{14.45 \times 10^2}{5.907} = 244°K.$$

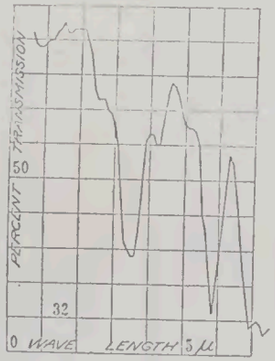
$p_{mm} = 90$ mm Hg $\rightarrow$ $\cancel{9.542}$ $\begin{array}{r} 7.820 \\ -1.954 \\ \hline 5.866 \end{array}$ $\quad T = \frac{14.45 \times 10^2}{5.866} = 247°K.$

$p_{mm} = 40$ mm Hg $\rightarrow$ $\begin{array}{r} 7.820 \\ -1.602 \\ \hline 6.218 \end{array}$ $\quad T = \frac{14.45 \times 10^2}{6.218} = 232°K.$

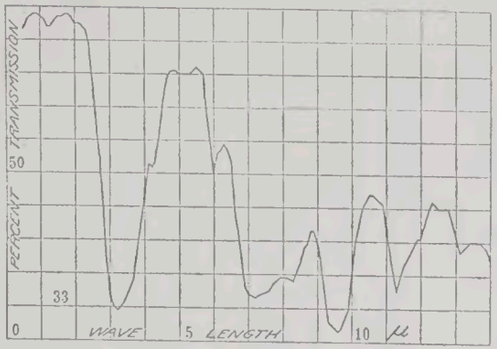
Actually, these temps. are slightly below the limits of measurement, of CH_3CHO, m.p. $\begin{array}{r} 273.1 \\ -24.3 \\ \hline 248.8 \end{array}$ but only very slightly.



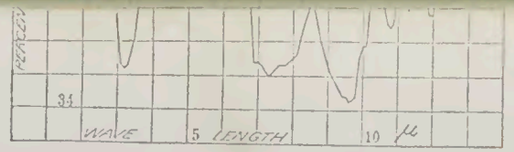
Curve 31. Isoamyl butyrate.



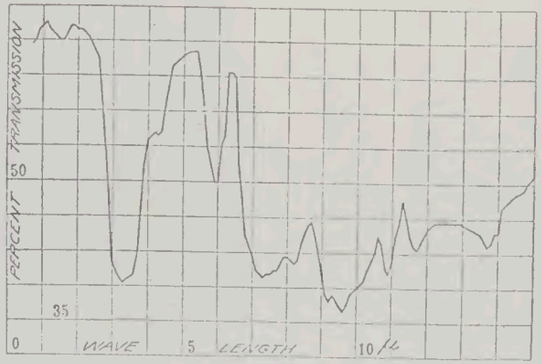
Curve 32. Isoamyl isovalerianate.



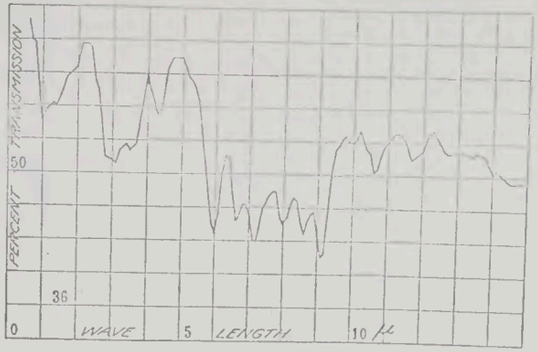
Curve 33. Ethylene glycol,



Curve 34. Propylene glycol.



Curve 35. Glycerin.



Curve 36. Acetaldehyde.

WENIGER: Infra-red Absorption Spectra. PLATE III.

For HCN, between 0 + 46 °C,

$$\log_{10} p_{mm} = -\frac{0.05223}{T} \times 27{,}875 + 7.752$$

∴ for $p_{mm} = 65$ mm Hg,

$$1.813 = -\frac{1.452 \times 10^3}{T} + 7.752$$

$$\begin{array}{r} 7.752 \\ -1.813 \\ \hline 5.939 \end{array}$$

$$\therefore T = \frac{1.452 \times 10^3}{5.939} = 245\,°K$$

Just to check,

$p_{mm} = 760$ mm Hg, $\log_{10} p_{mm} = 2.881$

$$\begin{array}{r} 7.752 \\ -2.881 \\ \hline 4.871 \end{array} \quad T = \frac{1.452 \times 10^3}{4.871} = 298\,°K = 25\,°C.$$

Observed normal boiling pt. HCN = 25.7 °C. ✓

∴ both acetaldehyde and hydrogen cyanide will be liquid or solid at night + gaseous by day over most of ⊙, certainly equatorial regions in summer + late spring (cf. Sinta in Plants + Settlers, p. 435). Also Mintz AMV, p. 1387

W. Weniger, Phys. Rev. 31: 388 (1910) "Infra-red absorption spectra" gives studied spectrum of liquid acetaldehyde. Note pronounced minima near 1.0 (v.s.), 1.4, 3.1 (v.s.), 3.6, 4.3, 6.0 (v.s.), 6.6, + 7.2 μ.

Near Infrared H_2O Absorption Bands

(Howard, Burch, + Williams, pp. 111 et seq.)

For the small quantities of water with which we are concerned, we shall almost always be — the region where the equivalent width of the band $\propto \sqrt{\ }$ water abundance. We will adopt the band $\overline{\Delta\lambda}$ at the half-power points, and derive the net $\oplus$ absorption in the band. As the abundance of water decreases, $\overline{\Delta\lambda}$ declines also; if we adopt the log. $\overline{\Delta\lambda}$, derived absorption will be minimum.

The curves of growth are of the form

$$\int A_\nu \, d\nu = a \, w^{1/2} (P+p)^n .$$

For σ, $P+p \simeq P \simeq 65$ mm Hg.

For $\oplus$, at 80,000 ft., $\simeq 15$ km, $P+p \simeq P \simeq 90$ mm Hg (cf. Dooley, p 44)

Reported stratospheric H_2O mixing ratios ~~from~~ over U.S. near 80,000 ft. range from 5×10^{-6} gm/gm (Barrett, Herndon + Carter, J. Met. 6: 367, 1949) to 4×10^{-4} gm/gm (N. Sissenwine, GRD note dated 14 Oct. 1960). The corresponding range in water abundance is

$$4\times10^{-2} \text{ gm cm}^{-2} \geq (w_{H_2O})_\oplus \geq 5\times10^{-4} \text{ gm cm}^{-2} .$$

∴ our ability to detect σ H_2O depends critically on H_2O abundance on date of flight. Spinrad at Victoria recently reported $(w_{H_2O})_\sigma < 1.4\times10^{-3}$ gm cm^{-2}. Oblique double traversal through σ atmosphere gives an increase in effective path by a factor $\simeq 4$. With σ near the zenith we can gain a factor of 2 or 3; i.e. we should be able to clearly distinguish a σ H_2O abundance equal to the vertical path water vapor abundance above Stratoscope

1.87μ H_2O band:

$$\int A_\nu \, d\nu = 152 \, w^{1/2} \, (P+p)^{0.30}$$

For ⊕, $(P+p)^{0.30} \simeq (90)^{0.30} \simeq 3.85$

For ♂, $(P+p)^{0.30} \simeq (65)^{0.30} \simeq 3.50$

For ⊕, $w = 4 \times 10^{-2}$ gm cm^{-2} $\Rightarrow \int A_\nu \, d\nu = 152 \times 2 \times 10^{-1} \times 3.85 = 117$ cm^{-1}.

With $\overline{\Delta\nu} = 400$ cm^{-1}, $\overline{A}_\nu = 0.29$

$w = 5 \times 10^{-1}$ gm cm^{-2} $\Rightarrow \int A_\nu \, d\nu = 152 \times 2.23 \times 10^{-2} \times 3.85 = 13$ cm^{-1}.

$\overline{\Delta\nu} = 300$ cm^{-1} $\Rightarrow \overline{A} = 0.044$,

For ♂, $w = 1.4 \times 10^{-3} \times 2 = 2.8 \times 10^{-3}$ gm cm^{-2} $\Rightarrow$

$\int A_\nu \, d\nu = 152 \times 5.29 \times 10^{-3} \times 3.50 = 28.2$ cm^{-1}

$\overline{\Delta\nu} = 350$ cm^{-1} $\Rightarrow \overline{A} = 0.081$.

∴ without comparison spectra, we might just barely check Spinrad, even with most favorable stratospheric water abundances. With a good comparison, we can get very much better accuracy.

♂ $w = 5 \times 10^{-4}$ gm cm^{-2} $\times 2 = 1.0 \times 10^{-3}$ gm cm^{-2} $\Rightarrow$

$\int A_\nu \, d\nu = 152 \times 3.17 \times 10^{-2} \times 3.50 = 16.8$ cm^{-1}.

$\overline{\Delta\nu} = 350$ cm^{-1} $\Rightarrow \overline{A} = 0.048$.

♂: $w = 2 \times 10^{-4}$ gm cm^{-2} $\times 2 = 4 \times 10^{-4}$ gm cm^{-2} $\Rightarrow$

$\int A_\nu \, d\nu = 152 \times 2 \times 10^{-2} \times 3.50 = 10.6$ cm^{-1}

$\overline{\Delta\nu} = 300$ cm^{-1} $\Rightarrow \overline{A} = 0.035$.

The 1.38μ band is slightly more favorable:

$$\int A_\nu d\nu = 163\ w^{1/2}\ (P+p)^{0.30}$$

The 1.1 and 0.94μ bands are much less favorable, and the 2.7 + 3.2μ bands are overlapped by CO_2. ∴ 1.38 + 1.87μ bands are the ones.

For 1.38μ,

for ⊕, $w = 4\times10^{-2}$ gm cm^{-2} ⟹ $\int A_\nu d\nu = 163\times2\times10^{-1}\times3.85$
$= 125$ cm^{-1}

$\overline{\Delta\nu} \sim 600$ cm^{-1}. ∴ $\overline{A} \simeq 0.21$

$w = 5\times10^{-4}$ gm cm^{-2} ⟹ $\int A_\nu d\nu = 163\times2.23\times10^{-2}\times3.85$
$= 14.0$ cm^{-1}

$\overline{\Delta\nu} \sim 500$ cm^{-1}. ∴ $\overline{A} \simeq 0.028$.

For ☉, $w = 1.4\times10^{-3}\times2 = 2.8\times10^{-3}$ gm cm^{-2}

$$\int A_\nu d\nu = 163\times5.29\times10^{-2}\times3.50 = 30.2 \text{ cm}^{-1}$$

$\overline{\Delta\nu} \sim 550$ cm^{-1}. ∴ $\overline{A} \simeq 0.055$.

$w = 5\times10^{-4}\times2 = 1.0\times10^{-3}$ gm cm^{-2}

$$\int A_\nu d\nu = 163\times3.17\times10^{-2}\times3.50 = 18.1 \text{ cm}^{-1}$$

$\overline{\Delta\nu} \sim 550$ cm^{-1}. ∴ $\overline{A} \simeq 0.033$

$w = 2\times10^{-4}\times2 = 4\times10^{-4}$ gm cm^{-2}

$$\int A_\nu d\nu = 163\times2\times10^{-2}\times3.50 = 11.4 \text{ cm}^{-1}$$

$\overline{\Delta\nu} \sim 500$ cm^{-1}. ∴ $\overline{A} \simeq 0.023$.

If at 1.87μ and 1.38μ, we can detect a few % change in signal cf. control, we should detect water on or. But unless the stratosphere is extremely dry on flight day, a comparison is needed. A typical situation involves a 10% stratospheric absorption superposed on a 2% Martian absorption.