

Hiding Titan's ocean: densification and hydrocarbon storage in an icy regolith

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Abstract. The problem of regolith densification is examined with application to Titan, with the objective of establishing the pore volume in the regolith. A large pore volume could store the large volume of hydrocarbons required to exist on Titan's surface by photochemical models, forming a "buried ocean", a surface model more compatible with observational constraints than the global ocean model favored in the last decade. A sophisticated coupled model of methane diffusion and regolith densification is presented. It is found that several hundred meters equivalent "ocean" depth can be concealed in the regolith. Further, methane diffusion through the regolith is fast enough to maintain the atmospheric methane concentration against photolysis. "Hydrothermal" methane circulation may play a significant role in surface chemistry and heat transport. Copyright © 1996 Elsevier Science Ltd

Introduction

Titan's surface has presented an enigma to planetary scientists; see, e.g. reviews by Lorenz (1993) and Lunine (1993). Photochemical models predict that a layer of hydrocarbon liquids should exist near Titan's surface. However, observations by radar and in the near-infrared—which can penetrate the optically-opaque haze-laden atmosphere of Titan—suggest that a large part of the surface is solid. Stevenson (1992) reconciled this contradiction by suggesting that the liquid could fill space in a porous crust. Here we examine, by means of a model of regolith densification, how much pore space may be available for "concealing" liquid hydrocarbons. We consider not only the traditional case of an ice regolith, con-

sidered previously by Eluszkiewicz and Stevenson (1990), where heat transport by (porosity-dependent) conduction establishes a temperature gradient which in turn influences subsequent densification, but also mixed rock-ice regoliths and a "soaked" regolith where flow of methane also plays a heat transport role.

Hydrocarbon inventory

The amount of liquids that must be accounted for is suggested only by photochemical models. The present atmospheric inventory of methane should have been depleted photochemically in $\sim 10^7$ years: its existence at present suggests a possible surface or near-surface reservoir of methane. Although this idea ran foul of Voyager measurements of the near-surface methane humidity, it was noted by Lunine *et al.* (1983) that the dominant photochemical product, ethane, is a liquid at Titan surface temperatures, and a mixed ethane-methane ocean would have a low humidity, compatible with the Voyager measurements. A photochemical model by Yung *et al.* (1984) suggested methane photolysis over the age of the solar system should lead to an ethane layer 600 m thick.

Dubouloz *et al.* (1989) argued, on the basis of Voyager 1 radio-occultation data on the near-surface atmospheric temperature and composition, and the ocean-atmosphere equilibrium, that the ocean could be methane-rich (up to 10 km deep) or ethane-dominated (~ 700 m). This range depended explicitly on the ethane "partial depth" of 600 m derived by Yung *et al.*

A more recent photochemical model suggests a rather lower ethane production (Lara *et al.*, 1994), giving a partial depth of the order of 250 m or less: the range of allowable ocean depths is correspondingly smaller.

To avoid the appearance of a "global ocean", either topography must account for the bright regions seen at radar (Muhleman *et al.*, 1990) and near-IR wavelengths

(Lemmon *et al.*, 1993, 1995; Smith *et al.*, 1996), or there must be a porous regolith able to hide several hundred meters equivalent of liquid. For discussion see Stevenson (1992), Lara *et al.* (1994), Lorenz (1993), and Lunine (1993).

Regolith depth

A starting point for considering how much regolith may be present today is how much regolith may have been produced by impacts. Estimates of mean regolith depths based on Monte-Carlo simulations (Veverka *et al.*, 1986) are Callisto (870 m), Ganymede (320 m), Rhea (1900 m), Tethys (1600 m) and Dione (740 m). Analogy with Titan might suggest thicknesses of ~ 500 – 1500 m. Much of this would have been generated in Titan's early history, during the late heavy bombardment. Thompson and Sagan (1992) interpolating crater densities elsewhere in the Saturnian system, suggest about 300 m of impact-derived regolith should have been generated over the last 3.5 Gyr. An additional factor, difficult to quantify, is the possibility that debris from the disrupted body, of which Hyperion is but a remnant, fell on Titan. Over 90% of the fragments of proto-Hyperion would have fallen on Titan within ~ 100 years (Farinella *et al.*, 1990). This volume of material corresponds to a thickness of impacting substance (not considering excavation during impact) of 300–8000 m depth (Lorenz, 1993). Thus we investigate regolith with an original thickness of 10,000 m. Although this initial regolith is at the upper limit of what might be expected, we find that the porosity at depth decreases sufficiently quickly that most of the residual pore volume is in the upper kilometer or so, so the results are not very sensitive to the initial regolith depth, as long as this figure is > 1 km, which the above considerations suggest is likely.

Although Titan's near-surface crust may have significant silicate and organic components, since it is heated enough during accretion to melt, we assume it differentiated sufficiently that the dominant regolith material is water ice.

Regolith compaction

A regolith, by virtue of its porosity, has a lower thermal conductivity than the bulk material from which it is made. Thus the temperature gradient induced in it by geothermal heat flow becomes high. This in turn leads to high temperatures at depth, which encourage the grains to sinter together and the porosity to decrease: thus a regolith's thickness is self-limiting.

The process of densification has been investigated by many workers: see, e.g. Smoluchowski and McWilliam (1984), Eluszkiewicz (1990), and Kossacki and Leliwa-Kopystyński (1993). We use essentially the methodology of the latter, with appropriate modifications to take into account the presence of methane (see later).

Table 1. Abundances, decay constants and heat generation of radioactive isotopes (see Prialnik *et al.*, 1987)

Isotope	λ (year ⁻¹)	E (J kg ⁻¹)	X (initial)
⁴⁰ K	5.50×10^{-10}	1.72×10^{-12}	1.1×10^{-6}
²³² Th	4.99×10^{-11}	1.65×10^{-13}	5.5×10^{-8}
²³⁸ U	1.54×10^{-10}	1.92×10^{-13}	2.20×10^{-8}
²³⁵ U	9.71×10^{-10}	1.86×10^{-13}	6.30×10^{-9}

Model description

In an attempt to determine the evolution of the pore space within Titan's regolith we consider two kinds of model: a "dry" regolith which does not contain liquid in the pores and a regolith partially "soaked" with liquid hydrocarbons.

In each case we assume that the regolith is initially homogeneously porous and undergoes compaction by closing of pores. Ice and rock are incompressible compared with porous ice or porous ice-rock mixtures. The bulk modulus of compressibility of a porous ice-rock mixture is of the order of 10^6 – 10^8 Pa for a wide range of porosity and of rock abundance (for example 0.35×10^7 in the case of porosity of about 0.35 and rock abundance of 0.57)—private communication by Leliwa-Kopystyński (according to experiments by Leliwa-Kopystyński and Maeno). On the other hand, for nonporous ice and for rock the bulk modulus of compressibility is 10^{10} and 10^{12} Pa, respectively. To calculate the rate of compression we have used the formula proposed by Leliwa-Kopystyński and Maeno (1993) and applied for icy satellites by Kossacki and Leliwa-Kopystyński (1993) because of its applicability to porous mixtures of ice and rock, as well as for pure ice. We compare the results with those for "power law creep", although this, like most existing formulas, applies only to porous ice without rocky admixtures (see, e.g. Ashby, 1988).

The heat flux at the bottom of the regolith depends on the current heat production within Titan's interior. It is assumed that the decay of radioactive isotopes is the only heat source within Titan's interior (we neglect accretional and tidal (Sohl *et al.*, 1995) heating. The composition of chondritic meteorites (see Table 1) was taken as the starting point to estimate the radiogenic heat production of Titan's rocky core.

Self-compaction of dry regolith

The self-compaction of porous ice (or ice/rock) regolith is described by means of equations given hereafter. Due to the compressibility of the considered medium we chose the Lagrangian form of equations. For the same reason, as an independent spatial variable we use mass instead of distance from the center of Titan or depth.

The equation for the internal energy change that, when written in a way suitable to calculate the evolution of the temperature, has the form

$$C \frac{DT}{Dt} = \frac{p}{\rho^2} \frac{D\rho}{Dt} - 4\pi r^2 \frac{\partial}{\partial m} \left(4\kappa\pi\rho r^2 \frac{\partial T}{\partial m} \right) + Q \quad (1)$$

where Q is the radioactive heat source per unit mass and is given by

$$Q = \sum_i \lambda_i E_i X_i c_m. \quad (2)$$

The quantity λ is the decay constant, E is the total energy released during the decay of a unit of mass of an isotope, X is the mass concentration of isotope within the rocky component of regolith and c_m is the rock concentration within regolith. The sum is over the total number of radioactive isotopes taken into account. The quantity κ denotes the thermal conductivity and is calculated as the volume average

$$\kappa = \rho \left(\kappa_i \frac{1-c_m}{\rho_i} + \kappa_r \frac{c_m}{\rho_r} \right) \quad (3)$$

and C is the effective specific heat

$$C = c_m C_r + (1-c_m) C_i. \quad (4)$$

The symbols κ_i , κ_r , C_r and C_i mean heat conductivity of rock, of ice, specific heat of rock and specific heat of ice, respectively. In this work $C_r = 1200 \text{ J kg}^{-1} \text{ K}^{-1}$ and $\kappa_r = 4 \text{ W m}^{-1} \text{ K}^{-1}$. For C_i and κ_i we use the simple relations

$$\kappa_i = \frac{488.19}{T} + 0.4685 \text{ W m}^{-1} \text{ K}^{-1} \quad (5)$$

(according to Hobbs (1974)) and

$$C_i = 7.037T + 185 \text{ J kg}^{-1} \text{ K}^{-1} \quad (6)$$

(according to Ellsworth and Schubert (1983)).

The equation for the pressure within regolith, which is assumed to be hydrostatic

$$p(m) = \frac{G}{4\pi} \int_m^{M_{\text{sat}} + M_{\text{reg}}} \frac{m'}{[r(m')]^4} dm' \quad (7)$$

where

$$r(m) = \left[\frac{3}{4\pi} \int_{M_{\text{sat}}}^m \frac{1}{\rho(m')} dm' + R_{\text{sat}}^3 \right]^{1/3}. \quad (8)$$

The term R_{sat} denotes the radius of Titan without regolith (the distance of the regolith's bottom from the center of Titan) while the symbols M_{sat} and M_{reg} mean the mass of Titan without regolith and the mass of regolith, respectively.

For the initial thickness of the regolith, for the reasons explained earlier, we have attributed the value 10 km. We calculate the initial thickness of regolith R_{sat} so as to obtain for the fully compacted regolith the sum $R_{\text{reg}} = R_{\text{sat}} + D_{\text{reg}}$ is approximately equal to the present total radius of Titan. Therefore, we set $R_{\text{sat}} = 2570 \text{ km}$ to get the present 2575 km surface radius.

The term ρ_{ref} is the density of the bulk (nonporous) material from which the regolith is made and is given by

$$\rho_{\text{ref}} = \frac{\rho_i \rho_r}{(\rho_i c_m + \rho_r (1-c_m))}. \quad (9)$$

The symbols ρ_i , ρ_r mean the density of ice and the density of rock, respectively. The quantity c_r is the volume concentration of rock within the porous regolith and is related to the mass concentration of rock via the formula

$$c_v = \frac{c_m \rho_i}{(\rho_i c_m + \rho_r (1-c_m))}. \quad (10)$$

The actual density of the regolith is related to the porosity via the equation

$$\psi = 1 - \frac{\rho}{\rho_{\text{ref}}}. \quad (11)$$

An empirical formula for the rate of densification (modified from Leliwa-Kopystyński and Maeno (1993))

$$\frac{d\psi}{dt} = -\psi \left(\frac{p}{10^6} \right)^{k_1 \psi} \times \exp \left(k_2 + k_3 \psi + k_4 \psi T + \frac{k_5 + k_6 \ln(1-c_v)}{T} \right) \quad (12)$$

where p is the compaction pressure in Pa and the values of the coefficients are: $k_1 = 9.8$, $k_2 = 8.727$, $k_3 = 54.11$, $k_4 = -0.2625 \text{ K}^{-1}$, $k_5 = -4663 \text{ K}$, $k_6 = 6000 \text{ K}$.

For pure water ice, the compaction rate may be expressed by power law creep

$$\frac{d\psi}{dt} = \frac{-3.1}{B^{1/2}} D_{\text{crp}} (1-\psi) \left[\frac{Bp}{3\sigma_0(1-\psi^2)} \right]^n \text{ for } \psi > 0.1 \quad (13)$$

$$= -1.5 D_{\text{crp}} (1-\psi) \psi \left[\frac{1.5 p}{n \sigma_0} \frac{1}{1-\psi^{1/n}} \right]^n \text{ for } \psi < 0.1. \quad (14)$$

Here ψ_0 has the sense of the initial (uncompressed) porosity and should be about 0.41 to avoid a discontinuity for $\psi = 0.1$,

$$B \equiv \psi_0 / (\psi_0 - \psi),$$

R is the universal gas constant $8.314 \text{ J K}^{-1} \text{ mol}^{-1}$,

$$D_{\text{crp}} = 10^{-6} \exp[-Q_{\text{crp}}((T_m/T) - 2)/(RT_m)].$$

Q_{crp} is the activation energy in the general flow law

$$\dot{\epsilon} = A \sigma^n \exp \left(-\frac{Q_{\text{crp}}}{RT} \right). \quad (15)$$

The values of the parameters are the following: $\sigma_0 = 18 \times 10^7 \text{ Pa}$, $Q_{\text{crp}} = 36,000 \text{ J mol}^{-1}$ and $n = 4.7$.

The values of the parameters for water ice are also listed in Eluszkiewicz (1990) and for nitrogen and methane ice in Eluszkiewicz (1991) (those two papers also contain references to the original literature). However, it should be borne in mind that these parameters are not "constants" they vary with temperature (for example, it is often found that $n \rightarrow n+2$ at low temperatures), and depend on stress conditions and microstructure. Further, power law

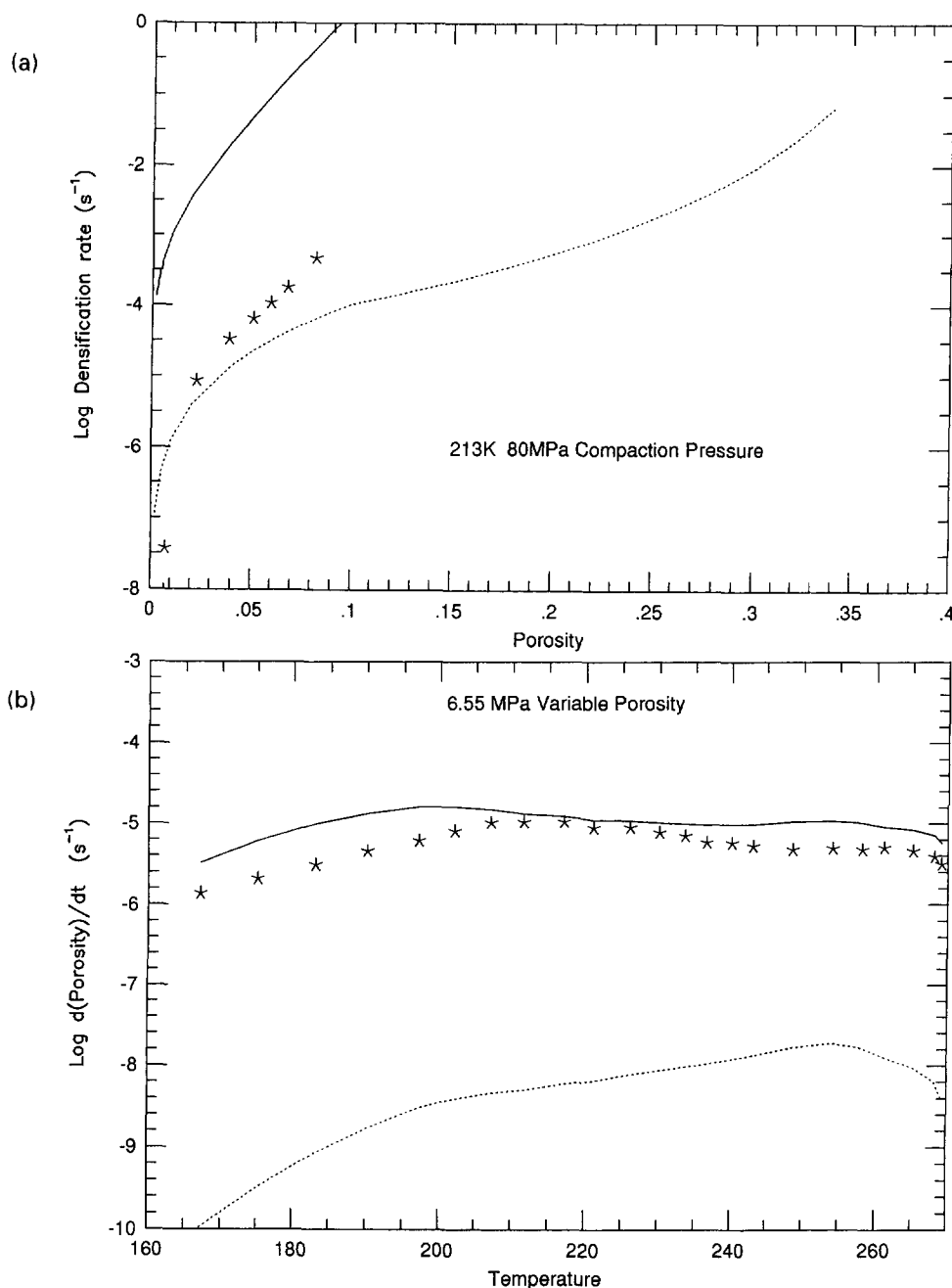


Fig. 1. Compaction rates. Experimental data (stars) from Leliwa-Kopystyński *et al.* (1994), and Leliwa-Kopystyński and Kossacki (1995), with the empirical rheology (solid line) and power-law creep (dotted line) shown for comparison. At high pressures (a) power-law creep is closer to the experimental data, but at pressures more appropriate to Titan's regolith (b), the empirical relation holds up well. In both cases, experimental data lies between the empirical and power law curves. Note in (b) the porosity at which the measurements were taken was not held constant, but decreases for higher temperatures

creep is strictly applicable only for ice regoliths, without a rocky component.

Although in some situations (see Fig. 1a) the power law creep appears to better describe the compaction rate, at the temperature and pressure in the regolith on Titan, at depths of a few km (100–180 K and a few MPa pressure) the empirical formula is a better representation (Fig. 1b). In any case, it seems to be a robust upper limit to the compaction rate, and hence gives a lower limit to pore ocean depth.

The boundary conditions are the surface porosity ψ_s

constant versus time and the surface temperature T_s . In all the models we ran, we held the surface temperature constant at 91 K, near its present level. We fixed the initial porosity throughout the regolith, at ψ_s , either 0.4 or 0.5 and examined the effect of rock fraction c_m of 0.4. See Table 2 for a summary of the models we ran.

Self-compaction of partially "soaked" regolith

To account for the presence of liquid hydrocarbons in some part of the pore space it is necessary to introduce

Table 2. Model characteristics

Model	Rock fraction	Initial porosity	Methane present?	Rheology
1	0.0	0.5	No	Empirical
2	0.0	0.5	Yes	Empirical
3	0.0	0.4	Yes	Empirical
4	0.4	0.5	Yes	Empirical
5	0.4	0.4	No	Empirical
6	0.0	0.4	No	Power law
7	0.0	0.4	Yes	Power law

modifications in the model. We analyze the regolith as two-layered. The upper layer is “dry”, with pores filled with saturated vapor and the bottom one has pores filled with liquid hydrocarbons.

Heat transport within the liquid and the heat outflow related to the evaporation of methane (which is significantly more volatile than ethane) could affect the temperature distribution and thus also compression rate of the regolith. Assuming that the heat transport through the soaked bottom layer is determined by the convective motion of the liquid methane and therefore is very efficient, this part of the regolith is approximated as isothermal.

Evaporating methane diffuses through the upper part of the regolith with a rate dependent on the temperature, the size and geometry of pores as well as on the pressure of methane vapor and of nitrogen contained within pores. It should be noticed, however, that if the nitrogen pressure within the pores is the same as in the atmosphere and if the average pore radius is of the order of a millimeter, the influence of pore size on the diffusion rate can be safely neglected because the mean free path is in such a case much shorter than the pore size. Thus, for simplicity we assume a unimodal pore size distribution rather than a more physically-plausible, but numerically more complicated, power-law distribution. We assume the number of pores per unit volume of regolith and the lengths of individual pores remain constant. This means that the value of pore radius is proportional to the square root of the actual porosity. To account for the possibility of existence of very thin pores, we define an effective diffusion coefficient the similar way to Clifford (1991). He suggested, considering the diffusion of water vapor in the Martian regolith, the reciprocal of the effective diffusion coefficient as the sum of the reciprocals of two diffusion coefficients for extreme cases: the bulk diffusion coefficient dependent only on collisions between water and carbon dioxide particles (big pores—in our case this corresponds to methane and nitrogen) and the Knudsen diffusion coefficient (small pores). In our calculations, for the bulk diffusion we adopted a formula for the diffusion of water vapor in nitrogen atmosphere. On the other hand, instead of the Knudsen diffusion coefficient alone, we have combined it with the coefficient for Hagen–Poiseuille diffusion to avoid restricting the applicability of the equations to a narrow range of vapor pressure (see Steiner and Kömle (1991)).

Since the “soaked” layer of regolith is approximated as isothermal, the current value of its temperature is cal-

culated by a heat balance equation. This includes the heat flux through the bottom (bedrock) and upper (boundary between layers) surfaces as well as on the radioactive heat production within the layer (where the regolith includes a rocky component). The heat flux through the upper surface depends on the heat transport by evaporating methane as well as on the thermal conductivity of the solid ice/rock matrix.

The pressure of methane vapor at the boundary between layers of regolith is taken to be equal to the saturation pressure at the relevant temperature. We also investigated the case where we held the methane pressure close to the regolith surface at zero, but this gave essentially the same results.

The process of methane condensation within the upper layer of regolith is not included in the model. The assumption of phase equilibrium at the interface between layers of regolith probably leads to an overestimation of the pressure of methane vapor at that level (and in consequence within the whole “dry” layer of regolith). However, it is difficult to obtain a much more reasonable estimation of pressure.

The following equations are added to the model to take the methane flow into account.

The equation for the temperature of the bottom layer of regolith

$$\frac{dT_{\text{bot}}}{dt} = \frac{S4\pi R_{\text{sat}}^2 + M_{\text{br}}c_m Q - FH_{\text{met}}4\pi R_{\text{bound}}^2 + \kappa \frac{\partial T}{\partial r} 4\pi R_{\text{bound}}^2}{CM_{\text{br}} + C_{\text{met}}M_{\text{met}}} \quad (16)$$

where R_{bound} is the central distance (the distance from Titan's center) of the boundary between regolith layers, M_{br} the mass of the bottom layer of regolith (without liquid methane), S the heat flux per surface unit area of bedrock, $H_{\text{met}} = 5.1 \times 10^5 \text{ J kg}^{-1}$ is the latent heat of methane evaporation, $C_{\text{met}} = 4000 \text{ J kg}^{-1} \text{ K}^{-1}$ and F is the flux of methane vapor through the boundary between layers (“soaked” and “dry”). The density of liquid methane is taken as $\rho_{\text{met}} = 612 - 1.8T$ (in kg m^{-3}). All methane properties used in this paper are taken from data in the *Encyclopedie des Gaz*, Elsevier, 1976.

The flux of methane vapor is expressed by

$$F = \frac{p_{\text{sat}}(T(R_{\text{bound}})) - p_{\text{met}}(R_{\text{reg}})}{\int_{R_{\text{bound}}}^{R_{\text{reg}}} \frac{dr'}{D_{\text{eff}}}} \quad (17)$$

where p_{met} is partial pressure of methane and p_{sat} the pressure of equilibrium between gas and liquid phases of methane. For p_{sat} we used the relation

$$\log_{10}(p_{\text{sat}}) = 4.05 - \frac{455}{T}. \quad (18)$$

The equation for the effective diffusion coefficient (see Clifford, 1991) is

$$D_{\text{eff}} = \frac{D_n D_r}{D_n + D_r} \quad (19)$$

where

$$D_n = \frac{1.38 \times 10^{-23} T^{2.072} 1.37 \times 10^{18}}{p_{\text{atm}}} \quad (20)$$

is the coefficient of water vapor diffusion in nitrogen (since the molecular mass of methane is close to that of water, we assume the same value will be representative for methane in nitrogen), and p_{atm} , the atmospheric pressure close to the surface is taken as 60 kPa at the beginning of Titan's history and 160 kPa at present, and

$$D_r = D_c \frac{1}{1 + K_n} + D_k \frac{K_n}{1 + K_n} \quad (21)$$

which describes the influence of pore size on the rate of gas diffusion through a porous medium (Steiner and Kömle, 1991). The term D_c is the coefficient for Hagen–Poisuille flow

$$D_c = \frac{\psi r_p^2 d_m^2}{2\tau^2} \left(\frac{\pi^3}{mkT} \right)^{1/2} p_{\text{met}} \quad (22)$$

D_k is the coefficient for Knudsen flow

$$D_k = \frac{8\psi r_p}{2\tau^2} \left(\frac{kT}{2\pi m} \right)^{1/2} \quad (23)$$

and K_n is defined as

$$K_n = \frac{128kT}{18\pi^2 r_p d_m^2 p_{\text{met}}} \quad (24)$$

is the modified Knudsen number (see Steiner and Kömle, 1991).

The expression for the pressure of methane within the upper layer of regolith is

$$p_{\text{met}}(r) = p_{\text{sat}}(T(R_{\text{bound}})) - F \int_{R_{\text{bound}}}^r \frac{dr'}{D_{\text{eff}}} \quad (25)$$

The equations listed in this subsection contain two parameters not present in the formulas for “dry” regolith: the initial average pore radius r_p and the methane pressure in the atmosphere, just above the surface of regolith, $p_{\text{met}}(R_{\text{reg}})$. In the presented calculations we used $r_p = 1 \times 10^{-5}$ m and $p_{\text{met}}(R_{\text{reg}}) = 0$.

We note also (following a suggestion by Stevenson) that the compaction pressure p is the difference between the pore pressure and the mean stress on the regolith. Thus, equation (7) earlier should have the term $(1 - \psi \rho_{\text{met}}/\rho)$ inserted inside the integral, for that region where liquid methane is present. This has, however, only a modest effect.

The methane inventory in our models has been assumed to be equivalent to a 200 m global layer of liquid ($D_{\text{oc}} = 200$ m).

It was beyond the scope of this investigation to consider a methane-soaked regolith with this primordial surface temperature, since methane would freeze. Therefore, most of the calculations fix the surface temperature at 91 K, a little above the freezing temperature of methane.

The heat flow through the crust is assumed to be due solely to radiogenic production. Tidal heating, for Titan's present orbit at least, is significantly lower than the radiogenic production (Sohl *et al.*, 1995).

None of the above assumptions is critical, as the important epoch is that following the episode of most regolith formation, which could have been very brief if the proto-Hyperion breakup played a significant role. The seven models we present are adequate to establish the variability of the results.

Results and discussion

Temperature and porosity profiles for different models were generated after the models had settled down to a near-equilibrium, where porosity is sufficiently low that the temperature gradient is low enough to make further densification extremely slow. The porosity evolution for five depths (0–4 km) is shown in Fig. 2.

The pore space in the various models is computed, as an equivalent depth (that of the “hidden ocean”), and shown in Fig. 3. Since it was seen in Fig. 2 that although most porosity is lost in the first 2 Myr of densification and the end-state is reached only asymptotically, so Fig. 2 has logarithmic time axis, allowing easy extrapolation to the present.

For ice only, it is clear that the empirical expression (models 1–3—lines on the figure) gives much lower pore volumes than power-law creep (crosses and asterisks). Since the empirical expression is optimized for ice-rock mixtures, it may overestimate the compaction rate for ice only. For these cases, the effects of initial porosity and the presence of methane are small. Even in this case, however 200–400 m of ocean could be hidden in the regolith.

The less conservative power law creep rheology (models 6 and 7) gives much more pore volume—a couple of kilometers' worth. The presence of a rock component (models 4 and 5) and the empirical rheology considerably enhance the pore volume, although not by much compared with ice-only power-law creep.

In summary, it appears entirely plausible that the equivalent ocean depth of the regolith pore volume could be several hundred meters or more. This is in agreement with the earlier order-of-magnitude estimate by Eluszkiewicz and Stevenson (1990).

Methane transport

We find that the rate of diffusion of methane through the crust (assuming no condensation occurs) is at most $\sim 2 \times 10^5$ kg s⁻¹. This value was not very sensitive to the parameter variations we tried.

This rate is $\sim 100 \times$ the rate at which methane is lost from the present atmosphere by photolysis. The heat flow corresponding to the latent heat of vaporization for this flux of 2×10^5 kg s⁻¹ is $\sim 10^{11}$ W: about 25% of the total (present) geothermal heat flow.

At such a rate, the likely methane reservoir would be depleted from the regolith and deposited on the surface. It would then trickle back down (mixed, perhaps, with ethane and other photochemical products) into the regolith before evaporating again—a hydrothermal cycle would result.

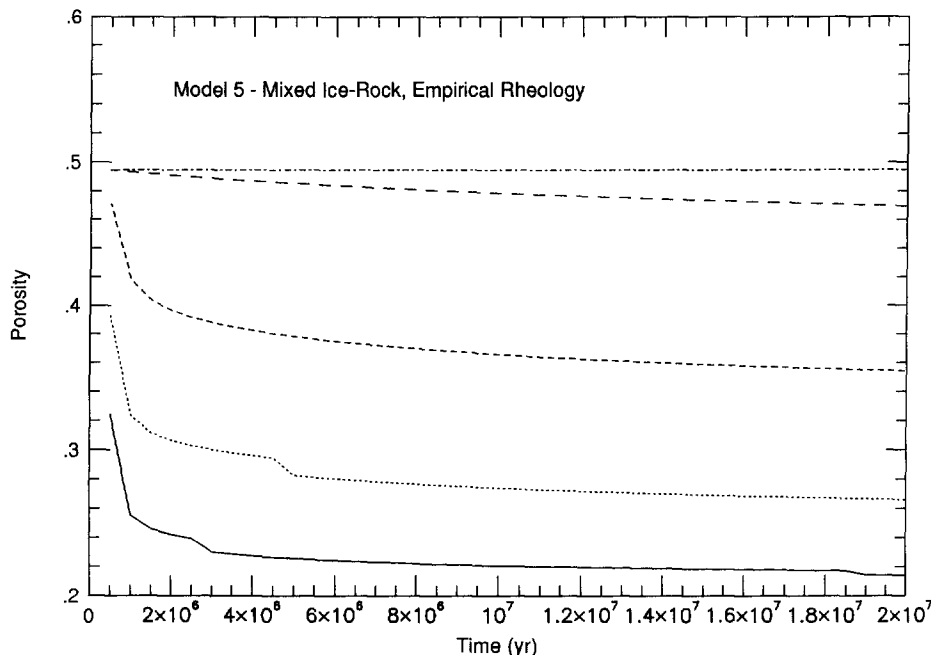


Fig. 2. Porosity evolution as a function of time and depth for model 4 (empirical rheology and 40seem that porosity quickly falls, and settles to a near-constant value after ~ 10 Myr. The kink at ~ 3 Myr is a numerical artifact caused by the rebinning of data as the regolith settles. Curves are for the following depths (0 km : dash-dot, 1 km : long dash, 2 km : short dash, 3 km : dots, 4 km : solid line)

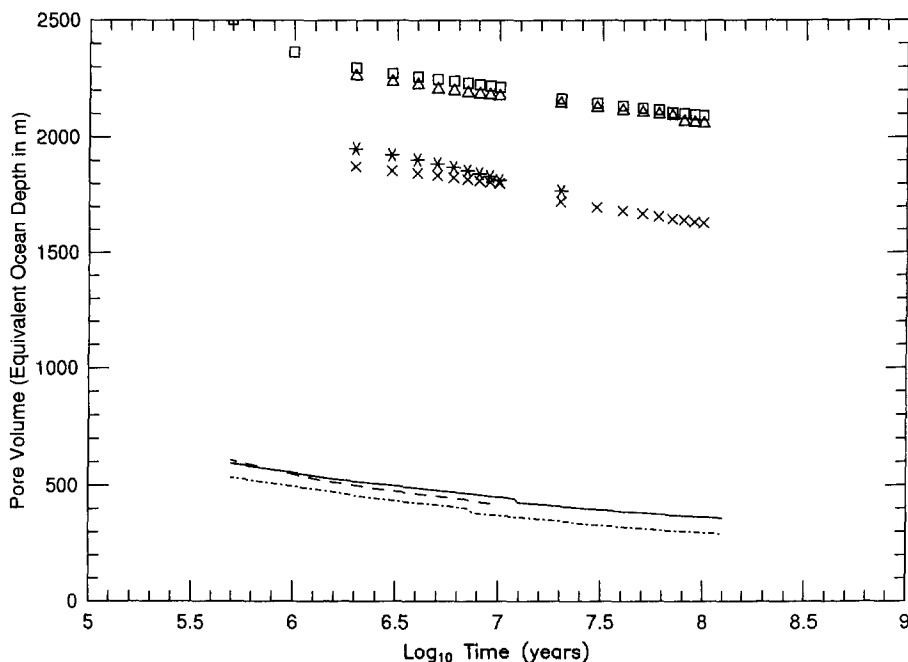


Fig. 3. Evolution of total pore volume vs time for the various models. Dashed line—1; solid line—2; dot-dash—3; squares—model 4; triangles—model 5; asterisks—model 6 and crosses—model 7. Models 1–3 form a distinct group, with the lowest pore volume. The effect of initial porosity and the presence of methane is small. 4 and 5 show that the influence of a rocky component is extremely significant in preserving pore volume, and the influence of initial porosity is small. 6 and 7 lie between the two groups, showing power law creep decreases porosity far slower than the empirical equation for pure ice. The small kinks are again due to rebinning of data as the regolith settles

In all probability, however, not all the methane in the inventory was delivered to the near-surface at an early moment in Titan's history. More likely, methane would be episodically supplied from the interior of Titan to the "buried ocean" in the regolith. However, since the transport through the regolith is sufficiently fast, the atmo-

spheric methane concentration would be buffered: the methane reservoir does not need to be exposed at the surface.

In this sense, a buried ocean behaves thermodynamically (in terms of ocean-atmosphere equilibrium) just like a surface ocean would. Note, however, that we

are not advocating that all of Titan's hydrocarbons are concealed in a porous regolith: we only suggest that most of the "deep, global ocean" can be concealed. It is likely that exposed regions of nonporous ice may exist, as well as surface reservoirs (perhaps crater lakes (Lorenz, 1994)) of hydrocarbons.

Chemical effects

As on Earth, and perhaps Mars (Clifford, 1991), hydrothermal circulation of fluids may leach out certain substances from the surface material. The same effect may occur on Titan.

Photochemical debris rains down from the atmosphere, with larger molecules in lower abundance than smaller ones. Smaller molecules are more soluble in hydrocarbon solvents: thus if hydrothermal activity has modified the surface, we might expect it to be depleted in low-C-number molecules. These aspects are discussed more fully in another paper (Lorenz and Lunine, 1995).

This effect is essentially equivalent to dissolving in methane rainfall in an atmospheric methane cycle, washing higher organics down into seas.

Eluszkiewicz and Stevenson (1990) suggested that cavernous weathering of ice by the "methanifer" could result in meter-scale caverns within 10^8 years, resulting in increased porosity at depth.

However, this suggestion was based on an incorrect experimental determination of the solubility of water ice in hydrocarbons—see Lorenz and Lunine (1995) for details.

Conclusions

Although our model will allow much more detailed investigations in the future, we limit ourselves in the present paper to the model description and the fundamental results establishing the feasibility of a "buried ocean" on Titan.

We are led to the following conclusions:

1. The regolith pore volume is sufficient to conceal most, if not all, of the hypothesized hydrocarbon ocean: even the most pessimistic estimates give several hundred meters' worth of storage volume. It follows, then, that mapping of the surface of Titan by Cassini can only establish a lower bound on the inventory of hydrocarbons.
2. Since the regolith quickly attains an equilibrium porosity profile, the regolith pore volume is not very sensitive to the initial depth, providing that the initial depth is more than the final equivalent depth.
3. Diffusion through the regolith is adequate to maintain methane levels against photolysis. The buffering of atmospheric methane is possible from a subsurface reservoir alone—there is no need for a surface reservoir.
4. The presence of methane does not affect the porosity significantly, as methane diffusion carries only a small part of the geothermal heat flux.

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