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The life, death and afterlife of a raindrop on Titan

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Abstract. A model is presented which describes the descent rate and evaporation rate of methane raindrops on Titan. The model, using conventional aerodynamics, with raindrop distortion parameterized by the Weber number, gives excellent agreement with terrestrial raindrop data. Terminal descent velocities for drops of various sizes at different altitudes are presented, and it is found that the largest raindrops may be larger than those on Earth (9.5 mm diameter vs 6.5 mm diameter) yet fall much more slowly (1.6 m s^{-1} vs 9.2 m s^{-1}. Under standard conditions on Titan, raindrops evaporate before they reach the ground: profiles showing the shrinking of drops due to evaporation during their descent are shown for various values of relative humidity. A 500 m increase in elevation can lead to a tenfold increase in rain mass flux, leading to increased "washing" of highland terrain. It is pointed out that evaporating raindrops will leave behind their condensation nuclei: fall times for these are presented, and it is noted that they may significantly affect visibility in the troposphere. The effects of additional factors on raindrop behaviour, such as the nonideal solubility of nitrogen in methane, are briefly considered.

Nomenclature

a	Semi-major axis (half-"width") of flattened droplet
a_0	Equivalent spherical radius of droplet
b	Semi-minor axis (half-"height") of flattened droplet
B	Bond number ($=\rho a_0^2 g/\gamma$)
C_d	Drag coefficient
C_{d0}	Drag coefficient of undistorted sphere
D	Coefficient of diffusivity
D_0	Diffusivity coefficient at standard conditions
Eo	Eotvos number (= Bond number)
f_v	Ventilation coefficient
g	Acceleration due to gravity
k	Flattening ratio
L	Specific heat of evaporation
M	Mass of droplet
m_a	Relative molecular mass of atmosphere
m_d	Relative molecular mass of volatile droplet component
m_i	Relative molecular mass of involatile droplet component
N_{Re}	Reynolds number ($=2a_0\rho V/\mu$)
N_{Sc}	Schmidt number ($=\mu/\rho_a D$)
N_{We}	Weber number ($=\rho_a V^2 a/\gamma$)
P	Atmospheric pressure
P_0	Standard pressure (101,325 Pa)
q	Fraction of surface area where convective updraughts occur
R_0	Universal gas constant
RH	Relative humidity
S	Drag area
S_0	Drag area of undistorted sphere
T	Temperature
T_0	Standard temperature (273K)
U	Updraught velocity
V	Terminal velocity of raindrop
X	Parameter
Y_a	Mole fraction of volatile raindrop component in atmosphere
Y_d	Mole fraction of volatile raindrop component in droplet
γ	Surface tension
μ	Coefficient of dynamic viscosity
ρ_a	Density of atmosphere
ρ_l	Density of liquid

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Introduction

Titan is perhaps one of the most fascinating bodies of the solar system and has attracted particular interest recently in connection with the preparations for the joint NASA/ESA *Cassini/Huygens* mission (Lebreton, 1992; Lebreton and Matson, 1992). One of the most exciting aspects is the possibility of seas of liquid hydrocarbons on the surface (Lunine *et al.*, 1983; Flasar, 1983) and of methane rainfall (Eshleman *et al.*, 1983; Flasar, 1983; Thompson and Sagan, 1984; Toon *et al.*, 1988; Lorenz, 1993).

In this paper, I present a model of the behaviour of individual raindrops on Titan, and consider the fate of raindrops as they descend towards the ground. Previous treatments of rain on Titan (e.g. Toon *et al.*, 1988) have suffered from poor estimates of fall velocities of raindrops: an accurate model and results are presented here.

The scope of this paper is limited to near-surface conditions and large methane drops and the evolution and behaviour of such drops as they approach or reach the surface of Titan (the "life and death" of the raindrops). The processes associated with the "birth" of raindrops, namely cloud microphysics and convection, are not treated here, nor is the high-altitude formation of the photochemical aerosols (Scattergood *et al.*, 1992; Cabane *et al.*, 1992; Toon *et al.*, 1992) that may act as the condensation nuclei for such raindrops.

Nature of rain on Titan

The *Voyager* 1 radio-occultation experiment indicates that the lower atmosphere of Titan may have appreciable methane "humidity", and therefore clouds and rain have long been suspected. Attempts to constrain the extent and type of clouds (Thompson and Sagan, 1984; Toon *et al.*, 1988; Griffith *et al.*, 1991) from ground-based and *Voyager* measurements have not yet been conclusive, although the best indications are (Toon *et al.*, 1988; Griffith *et al.*, 1991) that if clouds exist, they are either patchy, occupying only a fraction of Titan's disc, or are optically thin.

Vertical atmospheric motions associated with rainfall could be triggered by the interaction of Titan's zonal winds (Flasar *et al.*, 1981) with topography (i.e. orographic rain), or by convection: in the lower atmosphere about 1% of the solar flux may be transported upwards by convection (McKay *et al.*, 1991).

An alternative scenario is that individual aerosol particles may simply accrete methane and rain out when they pass through supersaturated regions of the troposphere. This scenario ["rain without clouds", Toon *et al.* (1988)] is plausible due to the relatively low number density of aerosols able to act as condensation nuclei.

However raindrops may form, the purpose of this paper is to consider their behaviour and fate near the surface.

Raindrop model

A problem with rain studies is that, to date, they have been almost exclusively devoted (perhaps not unnaturally) to terrestrial rain, so many results and expressions are empirical in nature, and inapplicable to rain in other planetary environments.

The model used here to predict sizes and descent rates of methane drops in the Titan atmosphere was first developed in studies (Lorenz and Zarnecki, 1992) prompted by the concerns (now largely allayed) that supercooled methane drops could freeze on the *Huygens* probe during its descent. The model includes the variation of drag coefficient with Reynolds number and models the deformation of drops due to aerodynamic forces.

Descent velocity for spheres

The steady-state descent velocity V of a sphere moving under the action of aerodynamic drag and gravity (i.e. falling at its terminal velocity) is straightforward to compute:

$$Mg = (\tfrac{1}{2})\rho_a S C_d V^2, \tag{1}$$

where the mass of the drop is simply:

$$M = (\tfrac{4}{3})\pi\rho_l a_0^3 \tag{2}$$

and for a spherical droplet the drag reference area is:

$$S = S_0 = \pi a_0^2. \tag{3}$$

The most awkward element in the equation is the drag coefficient of a sphere. There are various empirical relations for this quantity as a function of Reynolds number [e.g. see Davies (1945): for a detailed review, see also Clift *et al.* (1978)]. Here, the relation:

$$C_d = (24/N_{Re})(1+0.197N_{Re}^{0.63}+2.6\times10^{-4}N_{Re}^{1.38}) \tag{4}$$

is used for convenience. The Reynolds number (the ratio of inertial to viscous forces) is given as:

$$N_{Re} = 2a_0\rho_a V/\mu. \tag{5}$$

For low Reynolds numbers ($N_{Re} < 1$) the viscous effects dominate and the expression above reduces to Stokes' law ($C_d = 24/N_{Re}$). For higher Reynolds number, the drag coefficient tends to lie between 0.6 and 1.0. The relation does not predict the drop in drag coefficient associated with the transition from laminar to turbulent flow, but this never happens in raindrops, since drops break up before reaching such high Reynolds numbers.

The above method (and, for that matter, the unmodified Stokes' law) breaks down for very rarified atmospheres and very small droplets, where the frequency of collision of gas molecules with the droplet is so low that a correction factor for the slip flow regime must be applied. However, this occurs only for droplets so small (typically submicron) that we would consider them aerosol particles, rather than drops, and they are therefore beyond the scope of this paper.

Deformation of drops

It is found that the above method, while effective in predicting fall velocities of rigid spheres and small droplets, breaks down for large drops, which are deformed by aerodynamic forces.

To model this deformation, we modify the drag area and the drag coefficient by factors related to the flattening ratio ($k = b/a$) of the drop.

$$S = Sk_0^{-2/3}, \tag{6}$$

$$C_d = C_{d0}/k, \tag{7}$$

where we assume that the drop is deformed into an oblate spheroid (typically in free air the drop will be oscillating—jelly-like—about a mean, oblate shape, possibly becoming momentarily prolate: here it is assumed that it suffices simply to model the average shape).

Pruppacher and Beard (1970) found that drops smaller than 0.5 mm in radius have their deformations given as:

$$k = [1-(9/16)a_0\rho_a V/\gamma]^{0.5}, \quad (8)$$

as suggested by Imai (1950), while for the size range 0.5 mm $< a_0 <$ 4.5 mm the deformation is linearly related to drop radius:

$$k = 1.030-0.124a_0 \; (a_0 \text{ in mm}). \quad (9)$$

Note that strictly speaking for large droplets the shape is not an oblate spheroid, but has a flattened bottom—for a fuller discussion see Pruppacher and Beard (1970) and Clift *et al.* (1978). Green (1975) suggests that an oblate spheroid is an adequate approximation for most purposes.

The linear relation with diameter, though, is of little use in this application. The linearity may be fortuitous for the terrestrial parameter regime, and in any case, the constant of proportionality will be different for other planets. Since the deformation is due to a balance between aerodynamic and surface tension, the ratio between these forces, the Weber number:

$$N_{We} = a_0 V^2 \rho_a/\gamma, \quad (10)$$

seems a more appropriate parameter on which to derive a general relation for deformation. It can be seen (Fig. 1) that, for terrestrial raindrops at least, deformation is a reasonably linear function of Weber number. Thus for Weber numbers $N_{We} > 0.1$:

$$k = 0.97-0.072N_{We}, \quad (11)$$

while for the smaller drops with $N_{We} < 0.1$, the relation derived by Imai (1950) is used, which can be expressed as:

$$k = [1-(9N_{We}/16)]^{1/2} \quad (12)$$

[Green (1975) gives an alternative fit for flattening ratio as a function of Bond number $B = \rho a_0^2 g/\gamma$ (the ratio of gravity to surface tension forces, also known as the Eotvos number *Eo*)]. His relation:

$$k = [(4/17)(17B/4+1)^{1/2}+13/17]^{-3/2}, \quad (13)$$

also fits terrestrial drop data quite well, and since for a drop falling at terminal velocity weight and aerodynamic forces are balanced, the Bond number and Weber number are related. For the results presented here, however, the Weber number formulae above have been used: using the Weber number formulation is more closely related to physical reality, and allows the model to be applied to unsteady cases, e.g. for accelerating or unsteady drops.

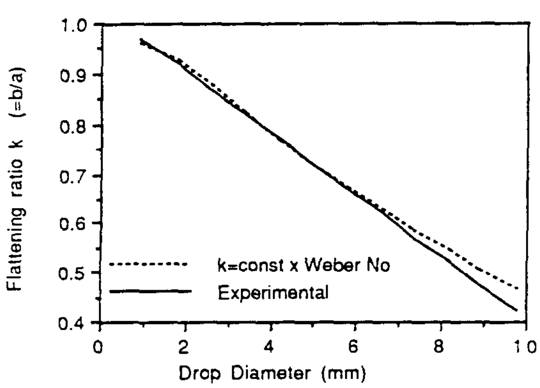

Fig. 1. Deformation ratio of terrestrial k ($=b/a$) as a function of the diameter of the (undeformed) drop. The k = constant $\times N_{We}$ agrees well with the linear relation of k with diameter found by Pruppacher and Beard (1967)

Figure 2a and b compares experimental measurements of terminal velocities with those predicted by the method described above, by approximating drops as spheres, and using Stokes' law. It is seen that Stokes' law breaks down for terrestrial raindrops of 0.1 mm diameter or larger. The rigid sphere approximation works well up to diameters of about 3 mm, above which the deformation of the drops becomes significant. Indeed, for drops larger than 4 mm the terminal velocity is virtually constant, with the increase in flattening of the droplet almost completely compensating for the increase in mass.

Note that the equations for descent rate, Reynolds number and deformation are all coupled, so to solve them a "first guess" terminal velocity must be input to derive N_{Re} and k, which must be folded back into the equations. For large drops, several iterations are usually required before convergence can be achieved.

Figure 3 gives a plot of descent rates for raindrops of various sizes at two altitudes on Titan, and on Earth for comparison. It is seen, as might be expected from the dense atmosphere and low gravity, Titan raindrops fall considerably slower than their terrestrial counterparts, with the maximum velocity for the largest drops (see next section) of 1.6 m s^{-1}. Toon *et al.* (1988) state that a 1 mm drop on Titan would fall 10 km in 1 hr, an estimate, based presumably on Stokes' law, that overestimates the fall rate by a factor of 2–3.

Clift *et al.* (1978, p. 179) give the relation $V_{max} = 2(\rho_l g\gamma/\rho_a^2)^{0.25}$ to predict the maximum velocity attainable by a large free-falling drop—independent of size or atmosphere viscosity—which yields 1.65 m s^{-1}, in excellent agreement with the result of the present model.

For the convenience of use by other workers, the descent rates of drops for various sizes and various altitudes are tabulated in Table 1. Note, however, that above 14 km, the equilibrium state (Kouvaris and Flasar, 1991; Thompson *et al.*, 1992) of the methane–nitrogen mix is solid: metastable supercooled drops will have descent velocities as in the table, but particles of ice (i.e. hailstones) will fall rather faster than drops of the same size.

Maximum drop size

An important parameter for work on raindrops is the maximum size a drop can attain. This is dictated (as is the drop shape) by the balance between surface tension holding the drop together and the aerodynamic forces which threaten to disrupt the drop. Photographic work by Matthews and Mason (1964) and references therein show that the disruption mechanism is as follows: for increasing drop sizes and descent rates, the drop becomes more and more flattened, until ultimately the centre of the

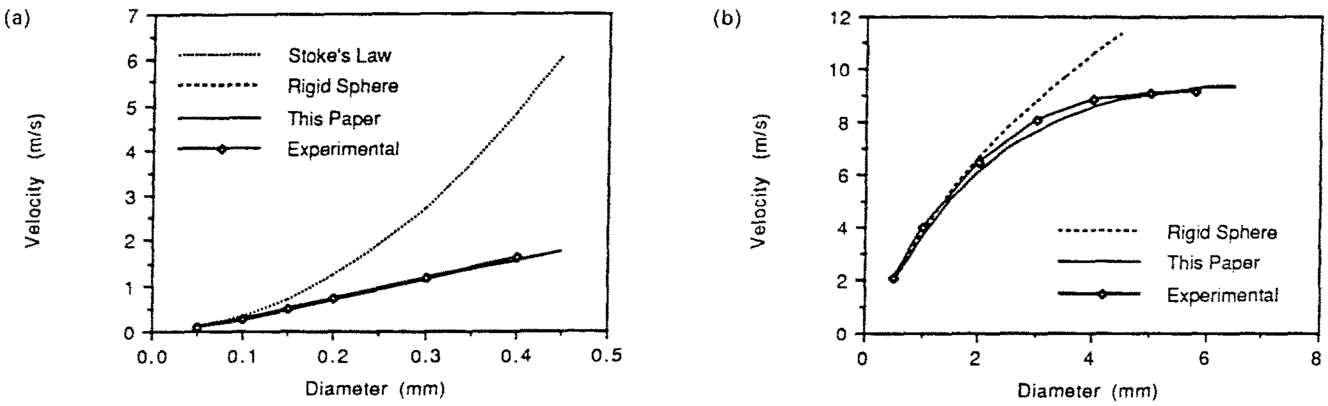

Fig. 2. Terminal descent velocities for terrestrial drops, showing results for Stokes' law, rigid sphere [equation (4)], the model used in this paper (with k = constant $\times N_{\mathrm{We}}$) and experimental data from Mason (1971). Panel (a) shows small drops: since drop deformation is small, the rigid sphere curve is identical to, and hidden by, the "this paper" curve. Panel (b) shows large drops—Stokes' law is so inaccurate that it is not shown

drop becomes so thin that it is blown upwards, forming a sac, which quickly bursts.

Naturally, there is strictly no such thing as a hard limit on drop size, since disruption is not instantaneous, and depends on ambient conditions, such as the turbulence of the atmosphere. Thus the maximum drop size is something of a statistical concept. However, it is known that drops much larger than about 6 mm diameter are rarely found in terrestrial rain. Further, studies by Komabayasi *et al.* (1964) showed that while 50% of drops of diameter 7 mm broke up within 20 s, the same fraction of drops of diameter 6 mm took 100 s to break up.

Matthews and Mason (1964) give the relation:

$$V^2 d = 8\gamma / n\rho C_{\mathrm{d}}, \tag{14}$$

as defining the upper limit of drop size, with n a factor between 1 and 2, and $d = 2a$. This can be rewritten using the Weber number definition above, and assuming the nC_{d} product equal to 1, as:

$$N_{\mathrm{We}} = 4. \tag{15}$$

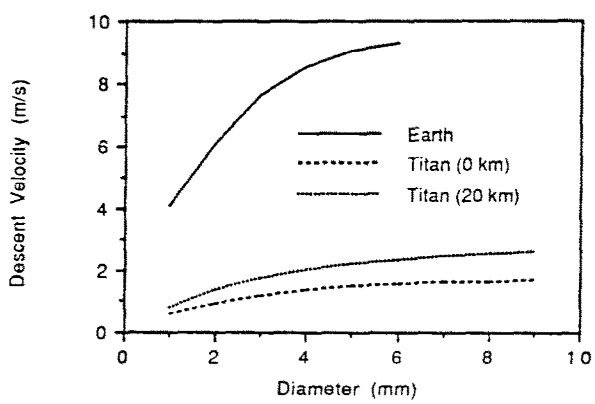

Fig. 3. Terminal descent velocities for Earth and Titan. The terminal velocities for raindrops on Earth and Titan are shown: clearly drops on Titan fall rather slower. In the thinner air at 20 km altitude on Titan, drops fall 50% or so faster than at ground level

Applying the parameters for water gives a limit size for terrestrial drops of 6.5 mm, which compares well with the experimental data cited above. For Titan conditions, the result is 9.5 mm.

[Note: Clift *et al.* (1978) give a formula, $d_{\max} = 4(\gamma/g\rho_{\mathrm{l}})^{0.5}$ which suggests that the maximum size of a drop is given by the relation $Eo = 16$, suggesting the largest drop has a diameter of 10 mm on Earth, or 18 mm on Titan: this relation may be appropriate for controlled laboratory conditions—Clift *et al.* (1978) is primarily a chemical engineering book—but in the turbulent "real world", such large drops do not occur, and so this formula is inappropriate to describe natural raindrops.]

Evaporation of raindrops

The usual method of computing the evaporation rate is to assume first that the drop is at rest, and compute the mass flux diffusing from it, then to modify this mass flux by an empirical coefficient f_{v} to take into account the ventilation of the drop by the airflow around it (which for large terrestrial drops can increase the flux by a factor of 15).

Thus:

$$\frac{\mathrm{d}M}{\mathrm{d}t} = \frac{f_{\mathrm{v}} 4\pi a_0 D m_{\mathrm{d}}}{R_0 T}(P_{\mathrm{v}} - P_{\mathrm{d}}), \tag{16}$$

where D is the molecular diffusivity of the raindrop material, P_{v} is the ambient vapour pressure of the raindrop material [equals the saturation vapour pressure P_{sat} of the drop fluid multiplied by the ambient relative humidity (RH), or the ambient atmospheric pressure P multiplied by the ambient mole fraction Y_{a} of the vapour]. Thus:

$$P_{\mathrm{v}} = (\mathrm{RH})P_{\mathrm{sat}} = PY_{\mathrm{a}}. \tag{17}$$

P_{d} is the vapour pressure of the raindrop fluid. Since (as on Earth) there may be involatile solutes in the raindrop (in Titan's case, the most likely candidate is ethane), these will depress the vapour pressure of the volatile component from the saturation level. Applying Raoult's law, this is

Table 1. Terminal descent velocities for methane drops on Titan at various altitudes. Terminal velocities of water drops on Earth are shown for comparison. N.B. terrestrial drops are not larger than about 6 mm

	Descent velocity of drops (m s^{-1})					
Diameter (mm)	Titan (0 km)	Titan (10 km)	Titan (20 km)	Titan (30 km)	Titan (40 km)	Earth (0 km)
0.1	0.04	0.044	0.048	0.052	0.055	0.236
0.2	0.107	0.121	0.138	0.156	0.174	0.668
0.3	0.173	0.198	0.231	0.269	0.311	1.11
0.4	0.236	0.271	0.32	0.38	0.448	1.538
0.5	0.295	0.341	0.406	0.487	0.583	1.942
0.75	0.431	0.499	0.602	0.735	0.9	2.854
1	0.547	0.638	0.776	0.958	1.188	3.606
1.25	0.65	0.761	0.931	1.159	1.451	4.339
1.5	0.737	0.865	1.062	1.331	1.689	4.993
2	0.904	1.065	1.317	1.661	2.113	6.098
2.5	1.042	1.232	1.533	1.948	2.5	6.965
3	1.157	1.372	1.715	2.194	2.836	7.636
4	1.33	1.587	1.999	2.581	3.375	8.53
5	1.454	1.737	2.198	2.859	3.77	9.02
6	1.538	1.84	2.338	3.055	4.054	9.2
7	1.594	1.91	2.43	3.191	4.254	—
8	1.632	1.957	2.498	3.284	4.393	—
9	1.655	1.987	2.539	3.346	4.487	—

equal to the saturation vapour pressure of the raindrop fluid, multiplied by its mole fraction in the droplet:

$$P_d = P_{sat} Y_d. \qquad (18)$$

As the volatile component evaporates, this mole fraction, Y_d, and hence the vapour pressure and evaporation rate will decrease. The saturation vapour pressure is a function of temperature (see the Appendix).

Pruppacher and Rasmussen (1978) found that the ventilation coefficient could be calculated as a function of Reynolds number and Schmidt number (the ratio of viscosity to diffusivity of the atmosphere) as follows:

$$f_v = 1+0.108X^2 \quad \text{for} \quad 0 < X < 1.4 \qquad (19)$$

and:

$$f_v = 0.78+0.308X \quad \text{for} \quad 1.4 < X < 51.4, \qquad (20)$$

with:

$$X = N_{Re}^{0.5} N_{Sc}^{0.33}, \qquad (21)$$

where N_{Sc} is the Schmidt number ($\mu/\rho_a D$).

The temperature T_a is assumed to be the temperature of both the droplet and the atmosphere. Strictly, the droplet should be cooled slightly with respect to the atmosphere by virtue of the loss of energy due to the evaporation of the volatile component of the droplet. A more refined model could include this factor, and suitable expressions are given in Pruppacher and Rasmussen (1978).

Evaporation (or growth) of droplets can be computed for a given altitude range Δh (100 m, for example), using the method above to obtain the rate of change of mass, noting that the time interval Δt to descend through Δh is given by:

$$\Delta t = \Delta h/(V-U), \qquad (22)$$

where V is the terminal velocity of the drop relative to the atmosphere, and U the velocity (positive upwards) of the ambient air. In an updraught, a drop will take longer to traverse a given altitude range, allowing more time for evaporation or condensation.

Collision and growth of raindrops

The growth of drops by direct condensation onto an existing drop or a nucleation centre can be computed directly using the evaporation relation described above, with a relative humidity greater than 100%. This is the dominant process where nucleation centres are limited (which may be the case on Titan, which may even have rainfall without clouds). However, a significant process in terrestrial clouds is the growth of drops by collision. In a cloud, droplets of different sizes will have different terminal velocities and so move relative to one another. When droplets collide, they may coalesce to form one larger droplet. Very small droplets are inefficient at colliding, since, having low inertia, they follow the streamlines around a larger drop, rather than impacting upon it. Thus, there is an efficiency parameter associated with a given pair of drops, defining the likelihood of coalescence, although for most drops (above a size of a few microns) the efficiency is virtually unity.

By following the trajectories of individual drops in a cloud with an "ambient" drop population, the evolution of the population may be investigated. Typically, "medium-sized" drops will be scavenged rapidly by collisions, leaving behind small droplets which are inefficient at colliding, and large drops which have terminal velocities high enough that their residence time in the cloud is short. However, such a rigorous and computationally intensive simulation is not considered further in this paper.

Results of evaporation model

The results of the evaporation/descent model described in the earlier sections indicate that, for the 70% maximum relative humidity indicated by the *Voyager* data (Flasar, 1983) and a 3 km cloudbase, rain (assuming pure methane rain, and ignoring the effect of nitrogen solubility) evaporates before it reaches the ground (see Fig. 4).

Addition of the effects of nitrogen solubility (Kouvaris and Flasar, 1991; Thompson *et al.*, 1992) is a significant complication to the model, and has not yet been attempted. It is believed that the current approach of assuming pure methane drops is qualitatively accurate (the maximum droplet nitrogen concentration below 3 km altitude is only 20%, and due to the slightly nonideal behaviour of the nitrogen–methane mix, the vapour pressure of methane is only reduced by 10% (Kouvaris and Flasar, 1991, Fig. 7). Thus, since the evaporation of the drop is in any case controlled by the methane in the drop [atmospheric nitrogen will quickly (Hibbard and Evans, 1968) be dissolved into or exsolved from the drop to attain equilibrium with whatever methane is in the drop] the present model should overestimate evaporation by only 10% or so [as a very crude estimate, using the curves on Fig. 4, the effect of nonideality may be seen by following a curve for a higher ambient humidity (e.g. 80% instead of 70%)].

Whether or not nitrogen is included, evaporation will be slower (and therefore the probability of reaching the surface higher) for colder or more humid regions, such as the perhaps cooler polar regions. Stevenson and Potter (1986) have argued that the observed temperature distribution on Titan suggests polar temperatures may be "pinned" by liquid polar caps of methane. The growth of such caps can be imagined if methane rain occurs globally and is unable to reach the ground in the warmer equatorial regions, yet may do so at the poles. However, the temperature estimates used by Stevenson and Potter were based on observations of the 530 cm^{-1} emission temperature, which may have a significant stratospheric contribution (Toon *et al.*, 1988), so the inferred surface temperature difference may not be real, although this point remains contentious (Flasar and Conrath, 1992).

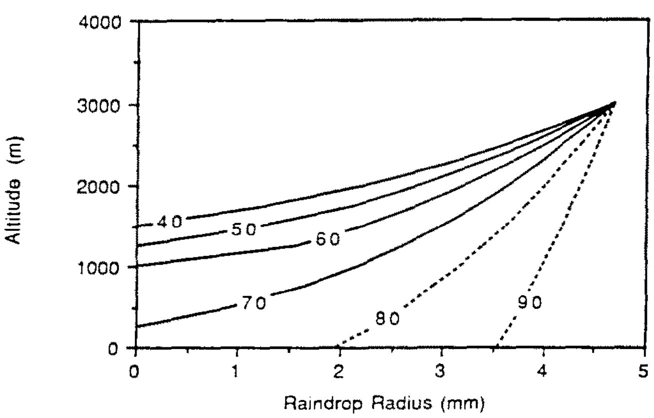

Fig. 4. Evaporation of drops on Titan. A 4.5 mm radius drop (approximately the largest possible) is released from the estimated cloudbase of 3 km and allowed to fall and evaporate in air of given relative humidity. It is seen that for humidities of 70% and lower (70% is the limit on surface humidity imposed by *Voyager* 1 data), the drop completely evaporates

Perhaps the most likely locations for rainfall to actually reach the surface are on mountains, where the surface is elevated with respect to the reference sphere of radius 2575 km on which the atmosphere profile is based: here the raindrops have a shorter distance to travel to reach the surface, and the atmosphere is slightly cooler.

As seen from Fig. 4, assuming drops start with a radius of 4.5 mm, the radius of the residual drops reaching topography elevated by 1000 m above the reference surface is about double those reaching topography at 500 m. The mass flux is therefore about an order of magnitude higher: this has intriguing implications (see later) for the appearance of mountains on Titan.

Ethane mists: "rain-ghosts"

The raindrops may have originally condensed on aerosol particles or tiny ethane droplets: indeed aerosol modellers (Toon *et al.*, 1992) assume "rainout" as the sink process for the flux of photochemical products. It is conventionally assumed that the aerosol number density in the lower troposphere is small as, sooner or later [Toon *et al.* (1991) assume within 1 year] drops are washed out and deposited on the surface. However, if, as is argued here, the rain cannot reach the ground, the aerosols will be redeposited in the atmosphere above the surface.

The aerosol particles are so small that they have very low sedimentation velocities, and hence long residence times in the atmosphere. While they are trapped inside a raindrop, however, they will be transported by it at the raindrop's much higher fall speed.

If there is convective activity in the lower levels of the atmosphere, then the aerosols will be unable to simply fall through, but may accumulate until rainout occurs. If raindrops grow by collision, the aerosols that formed the nucleation centres for the original smaller drops will be collected together. If the raindrop subsequently evaporates during its descent to the ground, the aerosol material (tholin or ethane or both) will be released (although it will be slightly diluted by some residual methane which is allowed to stay unevaporated because its vapour pressure is lowered by dissolution in the ethane). The aerosol will slowly descend—see Fig. 5 which shows (long) fall times for droplets released at 1 km altitude.

Thus, beneath an active raincloud there will be a zone where there are some aerosols and raindrops, then below this a region where all the raindrops have evaporated and there are only aerosols. Thus, after the rain ceases, there will be a cloud of aerosol particles left—a "ghost" of the raincloud. If rainfall is a rare occurrence, then this "ghost" cloud may contain a large number of aerosol droplets, accumulated in the convective cloud layers over weeks or months. Such a ghost cloud may have sufficient optical depth (see below) to be detectable with instrumentation on the *Huygens* probe.

If we take the ethane production rate of 1.5×10^9 cm^{-2} s^{-1} [Lara *et al.* (1992), although note that Yung *et al.*

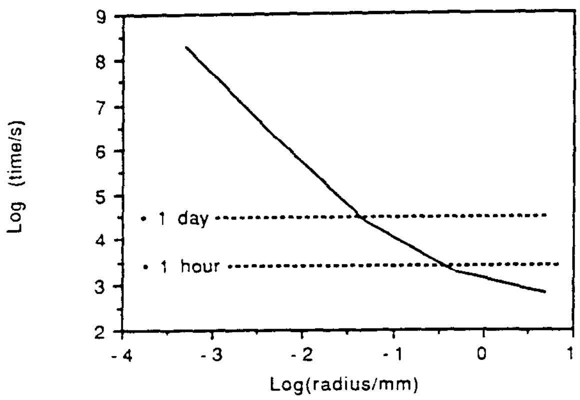

Fig. 5. Fall times for droplets on Titan from 1 km. While raindrops take about $\frac{1}{2}$ h to reach the surface, small aerosol particles may take days

(1984) give a rather higher value of ethane production rate of 5.8×10^9 cm^{-2} s^{-1}) and assume it is deposited in the troposphere near the surface as 1 μm radius drops, these will fall at about 0.02 mm s^{-1} and will have a number density of about 1.6×10^8 m^{-3}, giving an extinction of about 4.8×10^{-4} m^{-1} or 0.5 km^{-1} (compare this with about 20 km^{-1} for thick terrestrial fog). This type of haze should be easily detectable with the optical instrumentation on the *Huygens* probe (e.g. a 100 m thick layer of such a haze above a surface of albedo 0.2, seen from the probe 500 m above it and about 1 km away, would cause a 50% change in brightness: such contrasts will be easily observed). On the other hand, if aerosols are accumulated in the cloud layer, they may be more likely to be gathered into larger particles: if the ethane flux above is deposited as 10 μm radius particles instead of 1 μm, the extinction is a factor of 10^4 less (as well as having a lower number density for a given ethane deposition rate, such drops also fall much faster).

The existence of a near-surface ethane mist was proposed by Lunine *et al.* (1983) and has been investigated using near-i.r. observations (inconclusively) by Griffith *et al.* (1991). It will be an interesting challenge to determine whether any mists detected by *Huygens* are due to the condensation of ethane vapour near the surface, aerosol generation above an ocean (bubble-bursting and spindrift), or are the ghosts of past rainstorms.

The size (and sedimentation time) of the aerosols will depend first on the size of the particles that are delivered to the troposphere from above, and then on how much convective activity there is (how long the aerosol is "detained" in the troposphere) before rainout occurs. As to whether (or to what extent) aerosol particles are accumulated together by collision-induced droplet growth, this will depend on many parameters, such as cloud droplet number density, updraught strength and so on.

The fate of raindrops that reach the surface

Some rain may reach the surface, whether due to elevated ground, a low cloudbase, or reduced evaporation during descent.

If the "ground" is liquid, for example a methane–ethane lake, the drop will be absorbed. Such lakes or seas will be in equilibrium with the atmosphere—implications of this equilibrium on the allowable compositions are discussed by Dubouloz *et al.* (1989). If a large amount of rain is deposited in the sea, the sea composition will become methane-rich and will attempt to eventually lose methane by evaporation. However, the rate at which it can lose methane to re-attain equilibrium depends on the efficiency of ocean/atmosphere exchange processes (which are poorly understood, even on Earth).

If a drop lands on porous ground (as might be expected from impact-tilled icy regolith) the drop may be absorbed, percolating downwards into the soil. The area of liquid exposed to the atmosphere will be small, and vapour transport in the intergranular spaces will be relatively slow, so evaporation may be substantially reduced. If the regolith contains nonvolatile materials soluble in methane (such as photochemical aerosol) these materials may be dissolved, and will reduce evaporation further by lowering the methane vapour pressure. As the methane trickles down through the regolith, it may carry the material with it: rain could wash the surface clear of dark photochemical material. This scenario has been proposed by Griffith *et al.* (1991) to account for the "dirty ice" near-i.r. albedo of Titan: some regions have the bright icy bedrock or regolith exposed by rain, while other areas are dark, either (dirty) seas/lakes, or surface regions which have not been washed clean.

On Earth, mountain peaks are often covered with brilliant white caps of snow. Similarly, undersea mountains often have their upper regions covered with white deposits of calcium carbonate (from the skeletons of dead sea-creatures), while lower slopes are clear of such deposits (at the higher pressures at depth, the solubility of calcium is increased, and the deposits are dissolved away). This brings to mind the intriguing notion that mountains on Titan, since they are more likely to have their surfaces washed clean of organics by rainfall, may also have white peaks: whitecapped mountains can be found in a variety of environments in the solar system, albeit for a diverse range of reasons.

If a drop lands on a smooth, impermeable ice surface, it will probably evaporate: the low surface tension of methane (0.017 N m^{-1}, rather than 0.070 N m^{-1} for water) implies it should spread into a relatively large, flat drop, with a large surface area for evaporation. Similarly, the low surface tension suggests the drop is more likely to break into small droplets (again with a large surface area) on impact. If the drop does not evaporate immediately, it will flow along the local downhill until it either evaporates, meets porous ground, or a more substantial volume of liquid (like a lake). The high evaporation rate of drops, however, suggests that the first possibility is more likely, and that methane rivers require exceptionally (improbably?) wet conditions.

The low impact velocity (and hence low impact pressure), and the low solubility of water and other ices in cryogens (Rest *et al.*, 1990) suggests that erosion due to rainfall should be minimal (unless rain is an extremely frequent occurrence on Titan).

Conclusions and directions for further work

The question of rain on Titan remains open at present, but the properties of the atmosphere and environment on Titan suggest that, if it occurs, rainfall and its effects will be rather different from the rain we observe on the Earth. Rainfall on Titan is probably not a significant agent of erosion, but may cause variations in albedo over the surface by washing away dark photochemical material from some areas. There may also be near-surface "ghost" clouds of aerosol left behind by rainfall that has evaporated before it reaches the ground.

Directions for further work include:

(1) incorporate the effects of droplet cooling by evaporation, and the (nonideal) effects of nitrogen solubility in methane, to refine estimates of the drop evaporation rate;

(2) set up a 1-D convection model of a raincloud, using the descent and evaporation/growth model presented here for individual drops, to investigate the heat and mass transport processes within a "storm" on Titan. A possible result of such a model may be that a "rain shaft" may form—a region of air below the cloud the humidity of which has been increased by evaporating drops to the point where subsequent drops are able to reach the surface; and

(3) consider the effect on rainfall on the near-surface aerosol distribution.

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Appendix: physical properties

The viscosity of nitrogen is assumed to be well represented by the following equation

$$\mu = 1.718\times10^{-5}+[5.1\times10^{-8}(T-273)], \quad \text{(A1)}$$

where T is the atmosphere temperature in kelvin.

The diffusivity is assumed to be represented as follows:

$$D_v = D_0(T/T_0)^{1.75}(P/P_0), \quad \text{(A2)}$$

where the ambient pressure and temperature are in the same units as standard conditions (i.e. $T_0 = 273$K, $P_0 = 101.325$ Pa). Diffusivity at standard conditions D_0 for H_2O is 2.2×10^{-5} m^2 s, while for CH_4, $D_0 = 1.96\times10^{-5}$ m^2 s.

For atmospheric models which do not list all the relevant properties (e.g. the Lellouch–Hunten Titan atmosphere model lists only density and temperature), the other required properties can be obtained using the equation of state:

$$P = \rho_a R_0 T/m_a. \quad \text{(A3)}$$

For the saturation vapour pressure of methane over the temperature range of interest, the following relation has been used [although see Kouvaris and Flasar (1991) and Thompson *et al.* (1992) for the partial pressure of methane above mixtures of nitrogen and methane]:

$$\log_{10}(P_{sat}) = 0.06T-1.66 \quad \text{(A4)}$$

(T in kelvin, P_{sat} the saturation vapour pressure in pascals).

The density of liquid methane is about 450 kg m^{-3}, but to account for the presence of heavier hydrocarbons in drops (e.g. ethane, density 650 kg m^{-3}) and the dissolution of nitrogen (density 800 kg m^{-3}), a value of 600 kg m^{-3} has been used in this work.

The surface tension of liquid methane, like that of many hydrocarbons, is about 0.017 N m^{-1}.

The atmosphere model used for the studies here is the engineering model used for mission design purposes for the *Huygens* probe, due to Lellouch and Hunten (European Space Agency, Space Science Department Report ESLAB 87/199). Other models are given by Lindal *et al.* (1983) and Lellouch *et al.* (1989). A measure of the uncertainty of the model is given by the temperature uncertainty at the surface: 92.5–101K.

The nominal values from the model in the troposphere are given in Table A1.

Table A1. Titan atmosphere properties from ESLAB 87/199, used in the raindrop studies. The scale height near the surface is about 20 km, about 3 times that of Earth (i.e. Titan's atmosphere is not only dense, it is "thick")

Altitude (km)	Density (kg m^{-3})	Temperature (K)
0	5.3	97.2
1	5.14	95.9
2	4.99	94.4
3	4.83	93.0
4	4.65	92.0
5	4.45	91.4
10	3.67	85.8
20	2.24	78.6
30	1.28	74.4
40	0.7	72.7
50	0.37	73.3