

# Impacts and cratering on Titan: a pre-Cassini view

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**Abstract.** Titan is an outstanding object in terms of its usefulness for comparative planetology. While essentially an icy giant satellite like Ganymede and Callisto, its thick atmosphere makes it a very different place. Also, while Titan's thermal history may have similarities with Ganymede and Callisto, Titan's much larger distance from its primary and from the sun profoundly affects impactor speed distributions and gravitational focusing. The likely characteristics of the impact crater population on Titan are reviewed under two broad categories: number and spatial distribution, and morphology, with particular emphasis on impact-atmosphere interactions. Additionally, measurement prospects for Cassini are considered. © 1997 Elsevier Science Ltd

## 1. Introduction

Impact cratering is perhaps the most pervasive surface geological process in the solar system: impacts are certainly the most apparent features on planetary satellites. For discussion on cratering in general, good overviews are given by Chapman and McKinnon (1986) and the book by Melosh (1989). In this paper, I attempt to make broad predictions of the likely nature of cratering on Titan, whose surface has not yet been revealed at sufficient resolution to investigate cratering. A key feature of Titan, and indeed the reason we know so little about its surface, is its thick (1.5 bar) haze-laden nitrogen atmosphere. The interactions of an atmosphere with an incoming bolide, and its influences on the subsequent impact processes such as melt formation and ejecta dispersal, are among the poorest-understood aspects of impact science. In this paper, I aim to survey the various aspects of impact cratering about which Titan may teach us to put Titan in context

in the solar system and to provide a starting point for future, more detailed, modelling efforts.

I investigate Titan cratering under two broad headings: distribution and morphology. Under the former, I consider the likely number, and size and spatial distributions of impact features on Titan. The latter comprises the predicted nature of impact features; both craters themselves.

Additionally, I review recent work on Titan's atmospheric chemistry in so far as it concerns impacts. Despite the reducing nature of Titan's atmosphere, the abundance of CO<sub>2</sub> (albeit a trace) argues for the injection of oxygen-bearing material (probably water) into the stratosphere, due to meteoric ablation.

Finally, I consider the capabilities of Cassini instrumentation for investigating cratering on Titan, to document not only the craters themselves, but the effects of impactors on atmospheric chemistry and the sediment budget on the surface.

## 2. Distribution

### 2.1. Crater size distribution

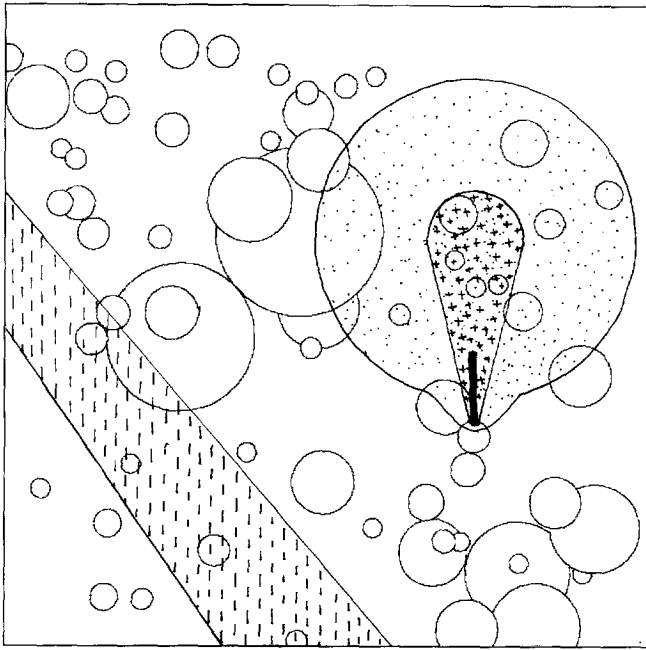
Lorenz (1993) suggested that if Titan's surface is old, then by interpolating between the impact populations (Plescia and Boyce, 1985) of Hyperion and Rhea (the satellites adjacent to Titan) there should be around 200 craters of greater than 20 km diameter every 10<sup>6</sup> km<sup>2</sup>. Engel *et al.* (1995) suggested, extrapolating from Plescia and Boyce's (1985) Iapetus figure, the same value. Engel *et al.* (1995) point out that the saturation crater density may be only 210 per 10<sup>6</sup> km<sup>2</sup>.

Griffith and Zahnle (1995) obtained least-squares fits to the impact crater populations on Rhea and Iapetus, determined by Lissauer (1988).

$$\log_{10} N(>D) = -2.73 \log_{10} D - 0.064 \quad (1a)$$

for Rhea and

$$\log_{10} N(>D) = -2.70 \log_{10} D + 0.109 \quad (1b)$$



**Fig. 1.** A synthetic Titan surface, with a distribution of craters from equation (1a). The box has 600 km sides. The dotted area shows the footprint of the Huygens probe descent imager DISR low-resolution imager, the area marked with crosses indicates the medium-resolution imager. The solid area the footprint of the probe altimeter, which operates from a lower altitude than DISR, hence the line is shorter than that swept by the imager footprints. The dashed region a segment of a radar SAR swath, the edges of which are not parallel—during a Titan flyby the spacecraft distance, and hence the beam footprint, changes. The relative orientations and locations of the SAR and Huygens footprints are arbitrary—the Cassini tour is still being designed

for Iapetus, where  $D$  is in km and  $N$  in craters per  $\text{km}^2$ . These relations give crater numbers of 242 and 395 per  $10^6 \text{ km}^2$  respectively for a minimum diameter of 20 km (see below). Titan's total surface area is  $\approx 83 \times 10^6 \text{ km}^2$ , so the total number of craters is likely to be of the order of 20 000. Figure 1 shows a section of Titan's surface with an impact crater population generated by the Rhea production function above, with the footprints of several Cassini instruments shown. The influence of different gravities of Rhea, Iapetus and Titan is small compared with the range of estimates above. An atmosphere may also affect the cratering process (and hence the size distribution), for example by influencing the deposition of energy into the target material (Brackett and McKinnon, 1992), but a much more important effect is the screening of small impactors (see below).

Since Titan's surface is completely unknown in age, and may indeed be quite young (since the potential for cryovolcanism exists—see, e.g. Lorenz (1996a,b), the crater distributions above may be far too high. On the other hand, the breakup of Hyperion (Farinella *et al.*, 1983, 1990) probably significantly augmented the number of craters on Titan. Since Hyperion is in a resonant orbit with Titan, most of the fragments from the proto-Hyperion breakup would have rained down on Titan within a relatively short time and would be sufficient to completely saturate the surface with impacts.

## 2.2. Geographical distribution and spin state

A synchronously rotating satellite will experience an asymmetry in impactor flux. Using the method outlined in Shoemaker and Wolfe (1982), the asymmetry may be determined for long-period comets of low inclination for Titan as about 5. Figure 2 in English and McBride (1995) show the predicted mass flux on Titan of interplanetary meteoroids (which may well have a different velocity and angular distribution from macroscopic impactors, due to their sensitivity to electrodynamic and radiation pressure forces) as a function of angle from the direction of motion, varying from  $10^{-16} \text{ kg m}^{-2} \text{ s}^{-1}$  on the trailing side ("anti-apex") to  $1.5 \times 10^{-16} \text{ kg m}^{-2} \text{ s}^{-1}$  on the apex. Averaging by eye over the angle range suggests a leading hemisphere/trailing hemisphere ratio of about 3.

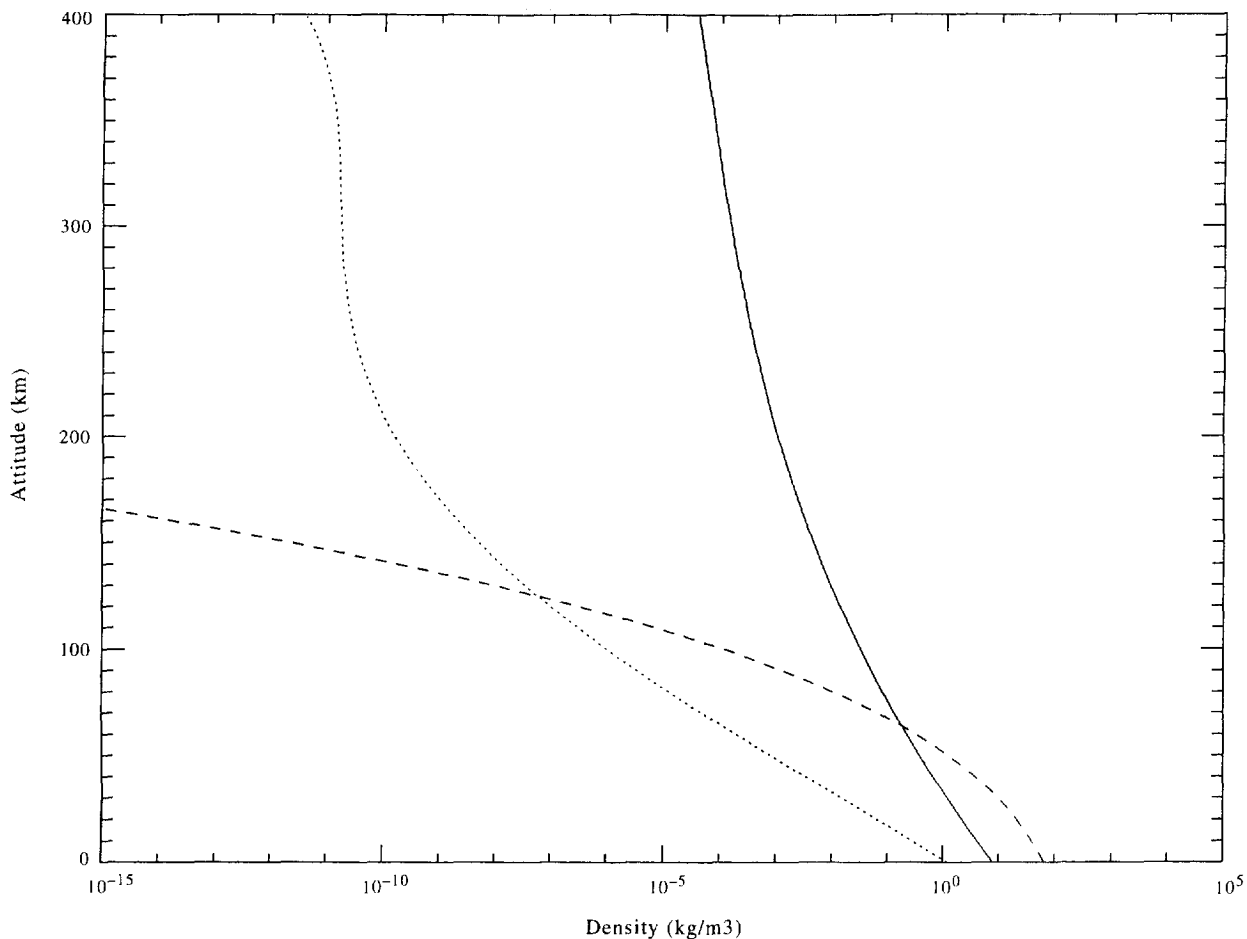
There is a factor of  $\approx 5$  between the crater densities on Rhea's leading and trailing hemispheres (Farinella *et al.*, 1990). This asymmetry is interpreted as due to a population of Saturnocentric impactors—Rhea, orbiting closer to Saturn than Titan does, would be expected to have a smaller asymmetry. If, as they contend, there is a connection with the disruption of proto-Hyperion and Titan has not changed its spin state since that event (somewhat unlikely, as a large off-normal impact is likely to break synchronous lock) then the asymmetry for Titan could be rather larger.

The fact that Titan's brightest regions (Smith *et al.*, 1996) are on the leading face of Titan, where the impactor flux would be highest, may be relevant here, although their preferred interpretation is that the bright region corresponds to mountains.

## 2.3. Atmospheric interaction—limits on size

Unsurprisingly, the massive atmosphere of Titan is likely to have a profound effect on the observed crater population, as on Venus. Figure 2 shows the density profiles of Venus, Earth, and Titan. The profiles shown are simple models of the form  $\log_{10}[\rho(z)] = c_0 + c_1z + c_2z^2 + c_3z^3 + c_4z^4$ , with  $\rho$  in  $\text{kg m}^{-3}$  and  $z$  in km. For the convenience of other workers, I present the coefficients here. For Earth ( $c_0 = 0.09$ ,  $c_1 = -0.063$ ,  $c_2 = -5.2 \times 10^{-5}$ ,  $c_3 = 8.56 \times 10^{-7}$  and  $c_4 = -1.28 \times 10^{-9}$ ), give good agreement (r.m.s. error 6% for the altitudes 0, 10, 20, 30, 40, 50, 60, 70, 80, 90, 100, 150, 200, 300, 400 km) with the CIRA 76 atmosphere up to 400 km; Venus ( $c_0 = 1.8$ ,  $c_1 = -0.02$ ,  $c_2 = -0.0002$ ,  $c_3 = -1.7 \times 10^{-6}$  and  $c_4 = 0$ ) gives an error of 2% up to 100 km altitude (comparing with the profile of Seiff *et al.* (1985)) and for Titan ( $c_0 = 0.9$ ,  $c_1 = -0.029$ ,  $c_2 = 5.2 \times 10^{-5}$ ,  $c_3 = -8.0 \times 10^{-9}$  and  $c_4 = -6 \times 10^{-11}$ ) gives an r.m.s. error of 4% between 0 and 400 km, compared with the atmosphere profile derived from the Voyager radio-occultation by Lellouch and Hunten (1987)—see also Lellouch *et al.* (1983). Note that the  $c_0$  Titan term does not equal the logarithm of the surface atmospheric density ( $5.3 \text{ kg m}^{-3}$ ) as the set of coefficients is optimized to fit the set of 15 altitudes above, of which 0 km is only one.

It can be seen that, although the Venus surface density is higher, the fall-off of density on Titan is much slower.



**Fig. 2.** Comparison of the atmospheric densities of Venus (dashed curve), Titan (solid line) and Earth (dotted line). While Venus's surface density is higher, the sharp drop in temperature with altitude, and the higher gravity, make it much more compact, while Titan's has significant density to very high altitudes

due in part to the heating of the stratosphere by aerosols on Titan and the low gravity which increase the scale height. Thus an incoming bolide will “feel” the atmosphere earlier on Titan (e.g. the density at 90 km, where meteors burn up on Earth, occurs at above 500 km altitude on Titan) than on Earth or Venus.

Ignoring ablation effects, an impactor will penetrate an atmosphere with a significant portion of its incident velocity if  $d > 0.2P_0/\rho_p g$  (Melosh, 1989), where  $P_0$  is the surface pressure (1.5 bar),  $g$  the gravitational acceleration ( $1.35 \text{ m s}^{-2}$ ),  $\rho_p$  the projectile density and  $d$  the projectile diameter. For Titan, this yields  $d > 120 \text{ m}$  for ice impactors. By comparison, for rock impactors, the values are 3 m,  $\approx 1 \text{ cm}$  and 200 m for Earth, Mars and Venus, respectively.

Thompson and Sagan (1992) give the expression  $D = 38r^{0.88}$  (although note that following their derivation, the exponent should in fact be 0.81, not 0.88) for crater diameter  $D$  as a function of (ice) impactor radius  $r$  (both in km), with an assumed impact speed of  $18 \text{ km s}^{-1}$ : the velocity dependence is  $D \propto v^{0.54}$  (the expression is derived from Shoemaker and Wolfe (1982)). A 120 m diameter ice projectile would cause a crater on Titan of diameter 4 km for an impact velocity of  $18 \text{ km s}^{-1}$ . This velocity is

probably an over-estimate of the typical impact speed—English and McBride (1995) (Fig. 1) suggest  $10 \text{ km s}^{-1}$  is more typical for meteoroids; Griffith and Zahnle (1995) (Fig. 2) suggest  $10 \text{ km s}^{-1}$  for Uranus–Neptune planetesimals and  $\approx 20 \text{ km s}^{-1}$  for Oort cloud comets; Jones and Lewis (1987) use  $6 \text{ km s}^{-1}$  for the “population I” Saturnian impactors and  $3 \text{ km s}^{-1}$  for impactors from the Hyperion breakup (most of this is due to Titan's escape velocity of  $2.6 \text{ km s}^{-1}$ ).

Melosh's (1989) expression describes the aerodynamic braking of an impactor (i.e. loss of velocity). An impactor may also be prevented from making a crater by breakup when the aerodynamic pressure on it exceeds its strength. Engel *et al.* (1995) investigated this with a simple model, and derived a cutoff in the crater distribution at around 20 km diameter. More recently, Ivanov *et al.* (1997) have investigated breakup using a more detailed numerical model of stresses on the bolide. Their work indicates the cut-off at rather smaller sizes than Engel *et al.* (1995), e.g. at 10 km diameter, the expected crater number density of Ivanov *et al.* (1997) is more than half that of an airless Titan. At 2 km, the population is one tenth. By comparison, Herrick and Phillips (1994) find these thresholds for Venus at around 10 and 20 km, respectively.

These studies only determine whether an impactor suffers mechanical failure during entry—fragments may themselves survive with sufficient velocity to form a crater (and indeed will be a principal source of the few small craters that will be found). Passey and Melosh (1980) developed a theory for the atmospheric breakup of impactors, which leads to the “strewn fields” of meteorites on Earth—see also Chap. 9 of Melosh (1989). They predicted that separation of disrupted impactors would be rather larger on Venus than on Earth, a prediction verified by the Magellan mission. Their model assumes the dispersion occurs due to the impulse imparted by the stagnation pressure between two fragments before they disperse.

The separation developed is given by

$$\Delta Y \approx (\rho_g / \rho_p)^{0.5} Z_b / \sin \theta. \quad (2)$$

Since  $\rho_g \approx \rho_s \exp(-Z_b/H)$ , this separation reaches a maximum at  $Z_b = 2H$ . For Earth, this expression gives, with  $\theta = 45^\circ$ ,  $\Delta Y \approx 0.5$  km, while for Venus  $\Delta Y \approx 4$  km. Clearly, since this is a single number from an idealized model, it cannot describe all cases, let alone extreme ones (shallow impacts, weak impactors). Although Herrick and Phillips (1994) in a detailed survey found one multiple impact with a separation of 36 km (which may well have been due to a binary asteroid in any case), most are separated by less than 8 km (Herrick and Phillips, Fig. 8). Thus, Passey and Melosh’s formula seems a reasonable guide. On Titan, for  $\theta = 45^\circ$ , we find  $\Delta Y \approx 2$  km. As this value is smaller than the diameter associated with impactors able to penetrate the atmosphere, most disrupted impactors will form a large, single, shallow crater than a cluster of smaller ones. Note that (see below) craters are likely to be shallow in any case.

#### 2.4. Crater chains

Interest in multiple impactors on satellites has been stimulated by the breakup of comet Shoemaker–Levy 9 by tidal forces during its close approach to Jupiter: following breakup near perijove, the comet’s fragments each followed separate, slightly different, orbits, making them form a “string of pearls” near apojove. Melosh and Schenk (1994) proposed that the crater chains seen on Ganymede and Callisto (3 and 13, respectively) were caused by similarly disrupted comets. The considerable excess of crater chains on the Jupiter-facing side of Callisto favors this origin, as does the absence of suitable candidates for the large impact basins that should be associated with secondary crater chains.

Wichman and Wood (1995) have considered tidal disruption of impactors more generally, noting one possible example on the moon (the Davy Crater Chain). While crater chains due to secondary impacts or volcanism are common on the moon, a disrupted impactor origin is suggested for Davy because of the circularity of the individual craters (secondaries are more typically elongate or irregular) and their raised rims (volcanic craters tend not to have such rims). They go on to examine the likelihood of crater chains on other solar system satellites and conclude that Titan is the best possibility for finding other examples.

Wichman and Wood (1995) calculate impact probabilities for crater chain (by scaling the size of the Roche limit of the primary planet to that of Jupiter and multiplying by the ratio of the cross-sectional area of the satellite to the spherical surface defined by its orbital radius) and the probable crater chain length. The crater chain length is computed for a close approach of 1.5 planetary radii, a velocity of  $15 \text{ km s}^{-1}$  and a projectile diameter of 1 km (in fact it is possible to back-calculate the probable size of the impactor prior to breakup from the crater-chain length—see McKinnon and Schenk (1995)).

The expected length for Titan computed by Wichman and Wood (1995) of  $\approx 20$  km is similar to that of Ganymede. However, this is based on an approach velocity of  $15 \text{ km s}^{-1}$ , which is rather high for the Saturnian system: Saturn’s greater distance from the sun reduces heliocentric encounter velocity, and acceleration in its shallow gravity well is also smaller than for Jupiter. A lower encounter velocity would give a somewhat larger dispersion (i.e. chain length), although the sensitivity is modest (Melosh and Schenk, 1994). The chains discovered to date on Ganymede and Callisto have been typically somewhat ( $\approx 5 \times$ ) larger than the predicted values of Wichman and Wood’s simple model. Schenk *et al.* (1996) make a detailed study of the chains on Ganymede and Callisto, and for periapses of 1.5 Saturn radii, predict chain lengths for Titan between 60 and 120 km (depending on the velocity). However, they point out calculations by Asphaug and Benz (1996) that comet disruption of this type may be impossible in the Saturnian system, as to produce enough stress to actually pull the comet apart would require a periape radius smaller than one Saturn radius. This is somewhat consistent with the paucity of confident crater chain candidates on other Saturnian satellites.

If crater chains are found, there are complicating factors. In Titan’s thick atmosphere, the aerodynamic wakes of the impactors may interact. Also, since a tidally disrupted comet breaks into small impactor fragments, the possibility exists that Titan’s atmosphere will screen out a significant fraction of these fragments. If the larger fragments are distributed randomly, we may expect to find on Titan interrupted crater chains, where smaller craters that would have appeared in a chain have not been created as the corresponding impactor was disrupted by the atmosphere. Schenk *et al.* (1996) observe, however, that the larger craters in a chain tend to be near the center of it, so truncated, rather than interrupted, chains are more likely.

### 3. Surface morphology

#### 3.1. Airburst scars

The Tunguska event in Siberia in 1908 levelled huge tracts of forest by the explosive breakup of an impactor in the atmosphere: although no crater was formed, the shock wave stripped or knocked over trees for tens of kilometers. Attention on such phenomena was focused recently on the SL-9 impacts with Jupiter. On Venus, such events have been attributed (Zahnle, 1992) as the likely cause of radar-

dark “shadows”, often associated with impact craters. These are interpreted as pulverized surface material, often with a margin of bright blocky rubble.

By crude analogy (Lorenz, 1993) it is natural to expect the same type of features on Titan. However, more detailed consideration suggests that such features are more difficult to form on Titan than on Venus for two reasons. First is the lower shocked gas density—Zahnle notes that the “wall” of gas from the explosion has a density of  $\rho_s(1+\gamma)/(1-\gamma)$  where  $\gamma$  is the ratio of specific heats. For Venus, with  $\rho_s = 66 \text{ kg m}^{-3}$  and carbon dioxide with  $\gamma = 1.3$ , this is  $\approx 500 \text{ kg m}^{-3}$ . Titan has  $\rho_s = 5.4 \text{ kg m}^{-3}$  and  $\gamma = 1.4$  the shocked gas density is only  $31 \text{ kg m}^{-3}$ —a shockwave therefore has an order of magnitude less momentum or rubble-generating capability than on Venus. Secondly, the shockwave spreads, so that a lower breakup creates higher shock pressures—on Venus the scale height is much lower than on Titan (see Fig. 2), so if breakup occurs on Titan it is likely to take place at a higher altitude, so the shockwave hitting the ground is much weaker. Unless Titan’s surface material breaks up much more easily than rock on Venus, we would not therefore expect there to be many “shadow” features. It is possible, however, that regolith could be scoured away by the atmosphere’s response to an impactor breakup.

### 3.2. Crater morphology

Aside from the general cratering literature mentioned in the introduction, some investigations of cratering on icy satellites in particular were performed in the wake of the Voyager encounter with the Galilean satellites: see Greeley *et al.* (1982) and Passey and Shoemaker (1982).

On large icy satellites, many large craters have upward-doming bottoms, presumably due to viscous relaxation of the crustal material—see Section 3.3. Schenk (1993) found that the depth  $h$  of craters on Ganymede could be related to their diameter  $D$  by  $h = 0.22D^{0.44}$ , for craters below 40 km diameter width (see also Schenk (1989)). Above 40 km diameter, craters had a near-constant depth (about 1 km) and feature central pits. Figure 15 of Schenk (1993) shows that the transition diameter from central peak to central pit craters scales approximately with the reciprocal of gravity: thus a similar value for Titan might be expected as for Ganymede and Callisto (30–40 km).

It may be noted that these depths are considerably shallower than the lunar craters of the same diameter. Simple craters there have a depth/diameter ratio of about 0.2, whereas 0.1 is more common for the icy satellites. Also, the transition diameter from simple craters to those with central peaks is smaller—10 km on the icy satellites, versus 20 km on the moon.

Note that these depth/diameter ratios apply to craters caused by single, intact impactors. Atmospherically disrupted impactors whose fragments have not sufficiently separated to create distinct craters will form a single large crater which has a shallower depth/diameter ratio than the original impactor would (Melosh, 1989).

Multi-ring impact basins are formed by the crustal response to large impacts, especially if the lithosphere is

thin. McKinnon and Melosh (1980) investigate multi-ring basins on Ganymede and Callisto. As Titan may have had a similar thermal history, it too may have a number of multi-ring basins.

### 3.3. Crater obliteration and relaxation

Craters may “disappear” by a number of processes—burial under lava flows or aeolian sediments, subduction (to the impact science community’s frustration, a significant process on Earth), and erosion/deposition by fluvial action. Lorenz *et al.* (1995) examined aeolian transport rates and found them very small. Further, the budget of sediments available for aeolian transportation may be limited by how much fine sediment is available. On Mars, high winds in the thin atmosphere enable aeolian erosion to create more particulates. On Venus and Titan, the thick atmosphere forces winds near the surface to be slower, such that aeolian erosion is small. The only sediments available may be those formed by impact ejecta (see, e.g. Garvin, 1990). Lorenz *et al.* (1995) estimated the equivalent depth of sub-cm particles on Titan due to impacts as less than half a meter global equivalent depth—too thin a veneer to cover more than a handful of kilometer-deep craters. Lorenz and Lunine (1996) examined erosion rates and concluded that 20 km craters would be preserved for at least 2 Gyr, in the absence of extreme climate fluctuations.

Crater relief may be degraded with time by viscous relaxation—see Scott (1967), Parmentier and Head (1981) and Solomon *et al.* (1982). On the Galilean satellites this leads to “palimpsests”—relics of craters that have little or no topography, but a pronounced circular region of different albedo—the crater has been flattened out. Albedo-only features may be more difficult to identify on Titan, since a veneer of photochemical material may cover low, flat regions.

Viscous relaxation acts as a probe of the variation of viscosity with depth. If Titan’s surface is water ice and rock, it will be very stiff at the low surface temperature, and relaxation of small craters will be very slow. The warmer interior, however, will be more fluid, so the degradation of large craters versus smaller ones can indicate the thermal gradient (large craters degrade quicker anyway due to scale effects, but the departure of the observed degradation behavior from this known effect can be determined). The influence of ammonia on the viscosity of water ice is not well known, but may be crucial.

### 3.4. Lake formation

Lorenz (1994) noted that craters on Titan would slowly fill with photochemical products (mostly ethane) deposited from the atmosphere, forming circular, annular or multi-ring (“bullseye”) lakes. The Manicougan structure in Canada is a prime terrestrial example of a ring lake (somewhat deepened by glacial erosion), and the Elgygytyn basin in Russia and Bosumtwi in Ghana are lakes formed in craters. There are about a couple of dozen examples in total, mostly in Canada and Finland (see, e.g. Hodge,

1994). Additional examples relevant to predicting crater lake morphology on Titan, although volcanic in origin, are Crater Lake in Washington state, U.S.A., and many volcanic island atolls.

Since most craters will have central peaks or updomed floors, the ring-lake type should be most common. By analogy with other icy satellites, most craters with updomed floors may have a central pit (which could also fill with liquid), “bullseye” lakes should not be uncommon.

Note that since liquid ethane and methane are rather less dense ( $\approx 500\text{--}600\text{ kg m}^{-3}$ ) than the expected crustal material (water ice,  $\rho = 900\text{ kg m}^{-3}$ ), there will be a restoring tendency to raise the center of the crater by viscous relaxation, even if the crater is completely filled with liquid. Thus, as the crust bows upwards, liquid may be displaced, so crater lakes may be the *sources* of fluvial systems, rather than only their sinks.

If a multi-ring impact basin (which typically have inward-facing scarps) were partially filled with liquid, it would be a spectacular structure to observe.

### 3.5. Atmospheric interaction with ejecta: impact melt

Rays, so striking on the moon, are not expected on bodies with significant atmospheres: the atmosphere brakes ejecta whose subsequent motion (unless the ejecta escapes in the tunnel carved in the atmosphere by the impactor) is dominated by atmospheric motions.

Thompson and Sagan (1992) investigated the expansion of an impact vapor plume into Titan’s 1.5 bar atmosphere, and noted that craters on Titan might have substantial liquid ejecta blankets, up to four times the radius of the crater itself (the liquid here is melted crustal ice, and should not be confused with liquid hydrocarbons that might subsequently fill the crater). They compare these ejecta blankets with those of the so-called pedestal craters on Mars (Carr *et al.*, 1977). These are believed to indicate the presence of subsurface volatiles (Barlow and Bradley, 1990); Schultz and Gault (1979, 1984) believe the lobate shape of ejecta flow is an artifact of an atmosphere, but that volatiles can enhance it (Schultz, 1988).

The atmospheric interpretation is not universally accepted, and must be called into question given the work of Horner and Greeley (1982) who find pedestal craters on Ganymede. They suggest this may reflect the easier creation of impact melt on an icy surface than on a rocky one. Since both an icy surface and surface volatiles are expected on Titan, ejecta blankets seem likely.

Craters on Venus have significant (and often very long) flows of ejecta—discussed in some detail by Chadwick and Schaber (1993). Schultz (1992) contends that these are due largely to the atmosphere’s motions immediately after the impact. If this is the case, then since the atmospheric density is an order of magnitude lower on Titan than Venus, one might expect the flows to be somewhat shorter on Titan than Venus. More particularly, environmental cooling of a melt flow on Venus will be slow, leading to long flows (especially since flow length scales correlate with gravity) compared with any on Titan. The Venusian surface is only  $\approx 30\%$  cooler than the melting

temperature, while on Titan (at 94 K, compared with water–ammonia melting temperatures of 176–273 K) a melt flow will be rapidly quenched by convection and entrainment of cold surface material. Thus, while ejecta blankets seem likely, the long Venusian-style flows do not.

### 3.6. Parabolas

Piercing of the atmosphere, and ballistic ejecta dispersal, has been noted in recent work on understanding the kill mechanisms associated with the terrestrial K/T impact and the demise of the dinosaurs. Additionally, some atmospheric nuclear weapons testing, notably Teak (3.8 MT at  $\approx 100\text{ km}$  altitude) generated effects that could only be understood by ballistic transport of the fireball.

Purely ballistic transport of ejecta leads to the formation of rays and secondary craters. In an atmosphere, expansion of the ejecta cloud is constrained, preventing the throwing of ejecta over planetary distance scales. A small impact will create a fireball that turns into a mushroom cloud, and fine ejecta is carried upwards and may be dispersed into a streak downwind. However, if the impact is sufficiently energetic, the explosion will puncture the atmosphere and ejecta can be dispersed ballistically and re-enter the atmosphere globally. The puncture can only occur if the expanding fireball caused by the impact has sufficient energy that its radius (were it to expand to ambient conditions) exceeds a scale height, such that the fireball can “sense” the top of the atmosphere. Vervack and Melosh (1992) express the minimum projectile radius  $a_p$  that can cause such an explosion as

$$a_p = H \left( \frac{P_0}{S \rho_p v_i^2} \right)^{1/3\gamma} \quad (3)$$

with  $\gamma$  the ratio of specific heats,  $v_i$  the impact velocity,  $H$  the scale height of the atmosphere, and  $P_0$  and  $\rho_p$  defined as before.  $S$  is a factor related to the target material (Melosh, 1989) and is  $\approx 1.3$  for rock, and 1.6 for ice.

Vervack and Melosh (1992) investigated whether an atmosphere-piercing ejecta plume could account for the dark parabolic features associated with some impacts on Venus. They found that the parabolic features could indeed be reproduced by a model of re-entrant ejecta, with a mean size falling off with distance from the impact, being winnowed by zonal winds. Since small ejecta falls slower, its horizontal displacement by wind is larger.

Lorenz and Melosh (1996) investigated this possibility for Titan, since the thick atmosphere and zonal wind field (Flasar, 1997) prompt many comparisons between Titan and Venus. For Venus, the criterion in equation (3) gives an impactor radius of  $\approx 680\text{ m}$ , whereas for Titan the figure is  $\approx 730\text{ m}$ .

One peculiarity of Titan is the sharp increase of temperature with altitude, owing to sunlight absorption by the photochemical haze. This increase in temperature causes an increase in scale height (from about 20 km to about 40 km) above the first two scale heights. (Thus, Titan’s atmosphere stretches deep into space-orbital trajectories are difficult to sustain below 1000 km altitude; on Venus, the temperature (and scale height) fall rapidly with altitude, so the atmosphere becomes significant only

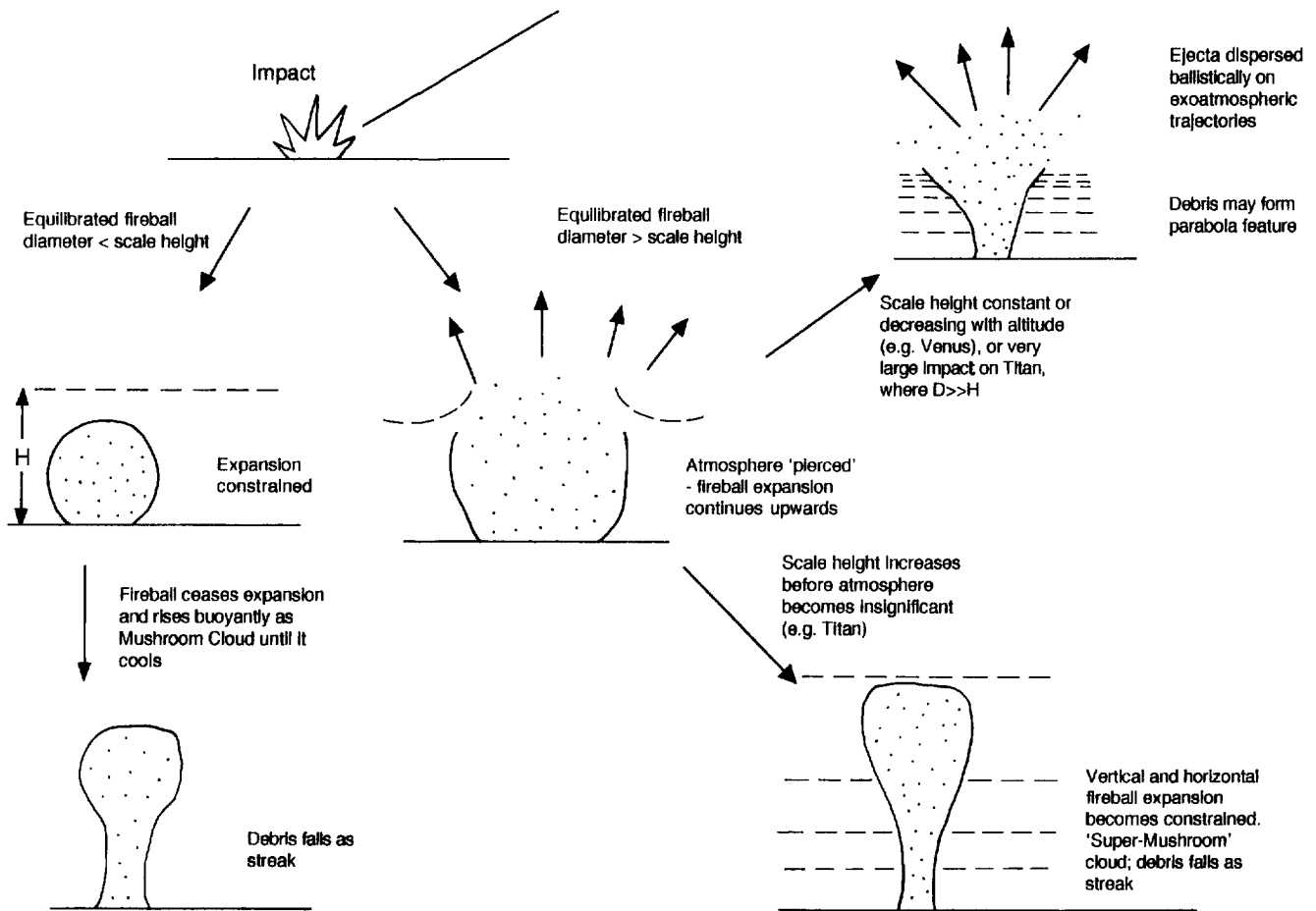


Fig. 3. Schematic diagram of fireball evolution. Constrained fireballs and mushroom clouds are common for small impactors in all atmospheres. "Blowout" occurs only for larger impactors. The reconstrained blowout is peculiar to Titan

below 200 km altitude or so, see Fig. 2. When the sharply different radii of Titan and Venus are considered, 2575 km versus 6310 km, the difficulty of impact ejecta attaining ballistic, but non-escaping, trajectories on Titan may be appreciated.)

Thus, while an impactor may create an ejecta cloud with an equilibrated diameter greater than a scale height at the surface, its subsequent expansion may be constrained—see Fig. 3. Lorenz and Melosh (1996) concluded that streaks of wind-smearred ejecta would be possible, and even Titan-girdling bands (assuming the wind direction to be exclusively zonal), but parabolae are relatively unlikely.

### 3.7. Submarine impact craters

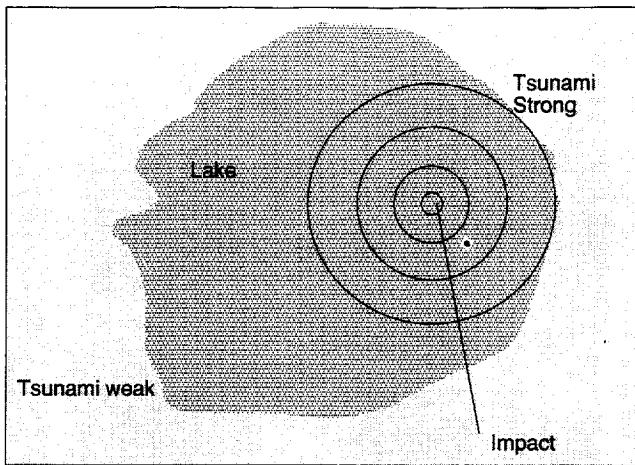
An impact into a sea with significant depth causes an impact feature in the crust with substantially modified (subdued) appearance. Such craters are termed hydroblemes. On Earth, impactors into the ocean of less than  $\approx 0.2$  km diameter will not affect the ocean floor, applying the rule of thumb that for water depth more than four times the crater depth, bottom effects are not seen in sand (Gault and Sonnett, 1982). On Titan, with a  $\approx 1$  km ocean, of lower density, it may be expected that an impact which

would produce a crater with a diameter of only a kilometer or two will generate a feature on the seabed. As this value is comparable with the diameter for impactors that are able to penetrate the atmosphere, it may be expected that hydroblemes are as common (per unit area of sea) as "normal" craters.

While no Cassini instrumentation offers the possibility to image the seabed, Titan may expose the bed to us. As a crater basin domes upwards by viscous relaxation, it will displace the liquid to the edges, forming a ring lake (Lorenz, 1993). The central region will form a circular island—perhaps with an exposed hydrobleme or two.

It may be noted also that an impact into a sea, even one that did not excavate the seabed, would give a substantial local atmospheric perturbation. Evaporating a cubic kilometer of liquid methane would inject  $\approx 5 \times 10^{11}$  kg into the atmosphere. Note that it would require a profoundly catastrophic impact to introduce an amount of methane into the atmosphere large compared with the present total abundance ( $\approx 10^{18}$  kg), although a more typical impact would produce a considerable stratospheric enhancement.

Also, tsunami formation would be expected, see for example Gault and Sonnett (1982), and Sonnett *et al.* 1987. Hills and Goda (1993) note that the strength of a tsunami is only inversely proportional from the distance



**Fig. 4.** Tsunami schematic diagram. When impact occurs, the more proximal lake edges are more strongly affected by a tsunami

of the impact, rather than proportional to the square. They give an expression for the height of a deep-water wave at a distance  $r$  from the impact of

$$h_w = 15.6 \left( \frac{10^6}{r} \right) \left[ \left( \frac{a_p}{203} \right)^3 \left( \frac{v_i}{20,000} \right)^2 \left( \frac{\rho_p}{3000} \right) \right]^{0.54}. \quad (4)$$

For a 1 km diameter ice impactor at  $10^4 \text{ m s}^{-1}$ , the tsunami at the edge of a 100 km diameter lake from a central impact (ignoring the significant wave height amplification as the wave shoals on the shallower bed near the lake rim) would be about 1 m. The formula, derived from explosive tests in water, may underestimate the wave height on Titan, as both the liquid density, and more importantly the gravity, are much lower. On the other hand, the shallowness of the lake may prevent a large tsunami from forming.

There may be some evidence of such tsunamis by, e.g. erosional or depositional features (e.g. improbably large boulders) above lake shores. It might even be possible to determine the location of a still-submerged hydrobleme by the relative heights of tsunami features (see Fig. 4). Since many lakes will be topographically constrained by crater rims, it is likely that “slosh” from tsunami may be revealed by evidence of (possibly catastrophic) liquid flow just outside the crater rim. As an aside, it is interesting to speculate (and would doubtless be interesting to model) tsunami propagation in a ring-shaped lake.

#### 4. Atmospheric chemistry

Impactors are likely to have played a significant role in the evolution of Titan’s atmosphere, both in its formation (Zahnle *et al.*, 1992; Griffith and Zahnle, 1995) and in the production of  $\text{N}_2$  and organics (Jones and Lewis, 1987). The present paper, however, will focus on the more recent and current role of impactors on Titan’s atmosphere.

Ip (1990) noted that meteor ablation on Titan would occur at rather higher altitudes (500 km versus 90 km on Earth) and that metal ions yielded by meteor ablation would be a dominant ion source in this region. Meteoric ionization may be an important factor in controlling haze

coagulation at these altitudes. Grard (1992) suggested the possibility of an enhanced ionization layer, although the initial (and a subsequent) analysis of the Voyager radio-occultation data rule out ionization levels of above  $3000 \text{ cm}^{-3}$ —see English *et al.* (1996) for discussion.

Samuelson (1983) measured an abundance of  $\text{CO}_2$  at and using a simple model, showed that this was compatible with a meteoric delivery of water. Using a terrestrial meteoric flux scaled to Titan, taking into account gravitational focusing of Saturn and Titan, and a revised photochemical model, English *et al.* (1996) showed that in fact a rather higher water flux was needed, in fact an order of magnitude more. This is due in part to treatment of condensation in the model (Lara, 1994; Lara *et al.*, 1996) and revised measurements of the abundance of CO in Titan’s atmosphere (Gurwell and Muhleman, 1995). They suggested that this could be made up by ejecta from Saturn’s outer satellites, but English and McBride (1995) showed thereafter that this contribution is very small. Additionally, they also investigated the interplanetary flux estimated by a revised interplanetary meteoroid distribution by Divine (1993) and found that gives a meteoric flux three times lower than that due to the scaled terrestrial value used in English *et al.* (1996). Thus the discrepancy remains, leaving a small number of residual possibilities:

- the models (either photochemical or meteoric or both) or the observations they are tuned to fit, are incorrect;
- if the impacting population has a comet-like composition, thereby directly injecting some  $\text{CO}_2$  into the atmosphere (as well as water), the discrepancy is much smaller (English *et al.*, 1996);
- some as-yet-unknown mechanism is able to pump water (or oxygen, at least) from the inner saturnian system outwards, in a manner analogous to the Io dust streams seen by Galileo and Ulysses, where fine particles are accelerated outwards electrostatically by Jupiter’s rotating magnetic field;
- Titan’s atmosphere is not in equilibrium, and the observed  $\text{CO}_2$  abundance reflects a perturbation by a recent large impact.

#### 5. Anticipated Cassini observations

##### 5.1. Optical remote sensing

The Cassini orbiter is expected to map Titan’s surface at resolutions of 100 m or better, with some smaller areas covered rather more closely. What exactly will be seen is open to question—HST imaging (Smith *et al.*, 1996) has demonstrated that surface contrasts exist in the 0.94 and  $1.07 \mu\text{m}$  atmospheric windows at scales of  $\approx 200 \text{ km}$ , although how visible small-scale contrasts will be is not known. In addition to imaging at these wavelengths, the ISS is also equipped with polarizers, which will reduce the contrast-reducing light contribution due to haze scattering.

Of necessity, most of the mapping imaging is at low phase angles, implying that shadows will be short. Additionally, the large scattered light contribution to surface illumination makes slope contrasts low, so photo-

clinometry will be difficult. Apart from shadow measurements and photoclinometry, the only optical topographical indicator is stereo imaging.

Surface albedo variations, over a rather wider (0.3–5.0  $\mu\text{m}$ ) wavelength range, will be measured down to resolutions of 500 m or so by the VIMS (visual and infrared mapping spectrometer) instrument, which can probe the atmosphere at window wavelengths of 1.28, 1.6 and 2 and 5  $\mu\text{m}$ , as well as the 1.07 and 0.94  $\mu\text{m}$  windows. These measurements will indicate strong albedo contrasts (e.g. lake versus shore) and probably some topography.

### 5.2. Radar mapping

The Cassini Radar experiment will conduct microwave radiometry, altimetry, scatterometry (wide-area backscatter measurements) and synthetic aperture radar (SAR) imaging. The latter mode is most relevant for crater studies, in that it provides direct two-dimensional information at the highest resolution. SAR data exists on some terrestrial impact craters from the Shuttle Imaging Radar experiments, for example, and of course all our knowledge on Venusian craters comes from radar imaging, principally from the Magellan spacecraft. The Cassini SAR will attain spatial resolutions of between 400 and 800 m, about three times poorer than Magellan.

Unlike Magellan, however, Cassini will not completely map Titan's surface. SAR imaging will generate only thin swaths of coverage near closest approach of some (perhaps 30%) of the  $\approx 40$  Titan flybys during the Cassini mission. Each swath is a few hundred km across, and over its  $\approx 3000$  km length, covers about 1% of Titan's surface.

Lorenz (1995) noted that, coverage of a surface by thin swaths is a very efficient way of detecting large craters: only a part of the crater needs to be seen to determine its area from circle-fitting techniques. Thus, although the sampling area of the swaths is limited to  $\approx 30\%$  of Titan's surface, for large craters ( $> 300$  km diameter) the effective sampling area fraction is increased by  $\approx 50\%$ . Thus, despite the incomplete coverage, Cassini should map the crater population fairly well—Lorenz (1996a) showed that of 16 random terrains, typically 10 would have more than 10% of their area covered by a radar swath pattern that covered 30% of Titan's surface.

Titan's radar appearance, and the appearance of craters in particular, is not well-constrained. Although Titan's disk-integrated radar brightness is high, and spread across this disk, how the radar scattering area is distributed at small spatial scales is not known. The attenuation of radar waves on the terrestrial surface is mostly due to liquid water: where liquid water is scarce, in dry deserts or polar caps, the radar energy may propagate some distance into the ground. Thus terrestrial radar imaging has unveiled sub-Saharan river beds and the recently discovered partially buried members of a possible crater chain at Aorounga, Chad. The Cassini radar may similarly penetrate a few meters, although it will not be able to sense the bed of lakes.

The altimeter mode may also be useful, in that vertical resolution of 80 m would allow an accurate topographical profile along the spacecraft groundtrack. The horizontal

resolution of this mode is of the order of 20 km, so it would only be useful for very large crater basins, but accurate topography measurement is vital for understanding the relaxation process that may have occurred in such large basins.

### 5.3. Other investigations

The cosmic dust analyzer (CDA) will measure the velocity and composition of dust particles, as well as their mass. The measured distributions, especially when coupled with composition measurements, will be able to constrain the impactor populations, such as those investigated by English *et al.* (1996).

The composite infrared spectrometer (CIRS) will make composition measurements with considerable spatial and altitude resolution: in particular it will measure the water and  $\text{CO}_2$  abundance profiles. The ion and neutral mass spectrometer (INMS), will directly measure elemental and molecular composition down to altitudes of 900 km. Multiple radio occultation measurements will be able to detect meteoric ionization.

Although CIRS, VIMS and the microwave radiometer could conceivably detect thermally distinct ejecta blankets on an airless Titan, the massive atmosphere has such a long radiative time constant that the surface is probably isothermal. Thus temperature contrasts between, for example, ejecta blankets and underlying material seen on Mars (Betts and Murray, 1993) will not be observed. Gross emissivity differences may be detectable, on the other hand, and for these the isothermality is in fact useful.

### 5.4. Huygens probe

The Huygens probe will descend for about 2–2.5 h by parachute in Titan's atmosphere. It carries a camera, DISR (descent imager and spectral radiometer). DISR will image a portion of Titan's surface at resolutions from  $\approx 150$  m down to 10 cm. However, comparing the area covered by DISR due to the probe's short groundtrack, suggests that the probability of "catching" a crater with DISR is only  $\approx 30\%$  for the Rhea population.

The probe radar altimeter will provide accurate topographical information, but only along a the probe's groundtrack. It is unlikely that the groundtrack will cross a crater boundary.

The probe carries a surface science package which will be able to make some basic measurements of the properties of the surface material. If the probe lands on "dry land", a penetrometer will be able to measure the surface hardness and particle size, constraining the local regolith properties. Regolith abundance may be controlled by the generation of particulates from impact ejecta and/or disrupted bolides.

## 6. Conclusions

Given the considerations above, it is certainly apparent that Titan will be both interesting and useful to study in

the context of planetary impacts. It is possible to make a number of predictions, directly testable by Cassini measurements:

- Unless Titan's atmosphere has had extended low-pressure episodes, most craters on Titan will be large (> 5 km diameter). Craters above 20 km diameter will have central peaks or domes.
- Crater lakes may be common, and may be sources, as well as sinks, of surface fluid flows.
- Contiguous ejecta blankets are likely to be present, similar to those on Mars. The long outflows seen on Venus are not expected on Titan.
- Parabolae, and shocked patches of surface, are unlikely. Streaks of wind-winnowed ejecta are possible, however. The overall sediment production by impact on Titan is probably modest.
- Erosive processes are fairly slow, so craters should have a long lifetime.
- Hydroblemes and tsunami deposits may be common—many impactors large enough to penetrate the atmosphere and subsequently form a crater are also likely to penetrate a 1 km ethane ocean.
- While Cassini will only map part of Titan's surface by radar, and optical remote sensing also has limitations, Cassini will provide good estimates of crater distributions and provide much exciting new morphology data.

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