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Cryolava flow destabilization of crustal methane clathrate hydrate on Titan

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ABSTRACT

To date, there has been no conclusive observation of ongoing endogenous volcanic activity on Saturn's moon Titan. However, with time, Titan's atmospheric methane is lost and must be replenished. We have modeled one possible mechanism for the replenishment of Titan's methane loss. Cryolavas can supply enough heat to release large amounts of methane from methane clathrate hydrates (MCH). The volume of methane released is controlled by the flow thickness and its areal extent. The depth of the destabilisation layer is typically ≈30% of the thickness of the lava flow (≈3 m for a 10-m thick flow). For this flow example, a maximum of 372 kg of methane is released per m^2 of flow area. Such an event would release methane for nearly a year. One or two events per year covering ~20 km^2 would be sufficient to resupply atmospheric methane. A much larger effusive event covering an area of ≈9000 km^2 with flows 200 m thick would release enough methane to sustain current methane concentrations for 10,000 years. The minimum size of "cryo-flows" sufficient to maintain the current atmospheric methane is small enough that their detection with current instruments (e.g., *Cassini*) could be challenging. We do not suggest that Titan's original atmosphere was generated by this mechanism. It is unlikely that small-scale surface MCH destabilisation is solely responsible for long-term (> a few Myr) sustenance of Titan's atmospheric methane, but rather we present it as a possible contributor to Titan's past and current atmospheric methane.

1. Introduction

It has been pointed out (Tobie et al., 2006; Lunine et al., 1989, 2009; Choukroun and Sotin, 2012), that Saturn's moon Titan may have an upper crust rich in methane clathrate hydrates (MCH) which formed early in Titan's history. With an estimated mass of $\sim 2 \times 10^{17}$ kg (Choukroun et al., 2010; Choukroun and Sotin, 2012), methane is an important component of Titan's present atmosphere. This amount of methane, which photo-dissociates under the influence of solar UV would normally disappear within 10-100 My (Yung et al., 1984). There is therefore a requirement for the replenishment of this atmospheric component over at least this time period, and longer, over geologic timescales (e.g., Atreya et al., 2006; Lunine et al., 2009; Niemann et al., 2010) if Titan's atmospheric methane component is indeed ancient. One possibility is that volcanic processes transport and release these gases from Titan's interior, although so far there has been no direct observation of on-going volcanic activity. Still, some process has recently supplied a considerable amount of methane to Titan's atmosphere.

Additionally, the relative geological youthfulness of Titan's surface (e.g., Lorenz et al., 2004) argues for internal processes that may have delivered gases to the atmosphere. We have investigated one such mechanism for the release of methane, namely the thermal destabilization of MCH by volcanic heat. We have modeled the thermal effects of the emplacement of likely "cryolavas" of ammonia-water composition. Firstly, we examine how such a volcanic process behaves thermally in order to determine event detectability via remote sensing, and, secondly, we model the penetration of the thermal wave into a methane-rich substrate. The destabilization of these clathrates would release methane into the atmosphere. We also note that meteorite impacts would destabilize some portion of a MCH crust, with the volumes of methane subsequently released being commensurate with the size and frequency of the impacts. In this paper we confine ourselves to considering endogenic processes that can contribute to the replenishment of atmospheric methane.

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The scale of volcanic activity required to maintain Titan's current methane atmosphere is quantified. Once the scale of activity is understood, the observations required for detecting it can be designed and data obtained by spacecraft. Titan can then be examined in the light of a better understanding as to what to look for both geomorphologically and thermally to determine whether Titan can truly join Io and Enceladus as volcanically active satellites. The techniques thus developed are applicable to the interpretation of data for other satellites, such as Ganymede and Europa, to seek evidence of effusive cryovolcanic activity.

2. Cryovolcanism on Titan

Volcanism has had a major role in shaping the surfaces of the terrestrial planets and moons across the Solar System. On an increasing number of bodies (Enceladus: e.g., Spencer et al., [2013]; Europa: e.g., Fagents et al., [2003] and Roth et al., [2013]; Triton: e.g., Kargel and Strom [1990] and Kargel [1994]), such activity continues to play an important role in planetary evolution, resurfacing the planet and yielding products that elucidate the internal chemistry and processes that are taking place. We use the terms "cryovolcanism" and "cryolava" to refer to processes and morphologies that resemble volcanism on the Earth, but where some much colder material plays the role of silicate magma and lava. Typically, cold water-ice mixtures with ammonia, methane, and other materials are referred to in this context. In a similar vein, activity on Enceladus is technically geysering, but as this process (as on Earth) is delivering heat from within Enceladus to the surface, we include it as a volcanic process.

The *Cassini* spacecraft has re-written what is known or suspected about the role of cryovolcanism in the evolution of the outer Solar System. The discovery on Enceladus of water-rich plumes (Porco et al., 2006) venting from cracks in the surface accompanied with relatively high heat flow from the vent regions (Spencer et al., 2006, 2013) led to the inevitable conclusion that internal heat exchange processes drive these activities.

The most intriguing possibilities for other current cryovolcanic activity in the Saturnian system are found on Titan. Driven by the need to explain the presence of atmospheric methane, speculation that cryovolcanism was taking place (e.g., Kargel, 1992; Lorenz, 1996, 2002) has resulted in searches for surface features possibly produced by cryovolcanic processes. *Cassini* observations have identified several candidate cryovolcanic features. Recent reviews of candidate features of cryovolcanic origin are found in Solomonidou et al. (2013a) and Lopes et al. (2013).

Daytime reflectance data obtained by the *Cassini* Visible and Infrared Imaging Spectrometer (VIMS) identified possible cryovolcanic features at Tortola Facula (Sotin et al., 2005), and Tui and Hotei Regios (Barnes et al., 2005, 2006; Nelson et al., 2009; Soderblom et al., 2009), although brightness variations in VIMS data of Hotei proposed by Nelson et al. (2009) were not found in reanalysis of the data (Soderblom et al., 2009; Solomonidou et al., 2013b). *Cassini* Radar data yielded a number of possibilities: for example, at Ganesa Macula and flow-like features in a number of other locations (Elachi et al., 2005; Lopes et al., 2007, 2013). However, subsequent examination has cast doubt on an exclusively volcanic formation model (Moore and Pappalardo, 2011) at most of these locations. Additionally, subsequent radar data of Ganesha Macula showed that this was a flat feature, and did not have the volcanic dome or pancake-like topography proposed by Lopes et al. (2007).

Perhaps the most reasonable candidates so far are at and around Sotra Facula and flows of unknown origin at Hotei Regio (e.g., Lopes et al., 2013). The region around Sotra Facula has been interpreted as a complex of multiple cryovolcanic features dominated by two mountains, Doom Mons and Erebor Mons (Lopes et al., 2013). "Finger-like" flows of perhaps 100 m thick have been proposed to explain some features apparent in *Cassini* radar data. A volcanic origin for Sotra Facula would support the idea that ammonia-ice mixtures might form analogous features to terrestrial basalts due to their similar dynamic viscosities (e.g., Kargel et al., 1991; Kargel, 1992; Lorenz and Shandera, 2001).

In other work unrelated to Sotra Facula, ammonium sulphate-driven explosive volcanism was proposed by Fortes et al. (2007).

The features proposed as possible volcanoes on Titan are somewhat- to highly-eroded. There is little or no reliable evidence in *Cassini* data suggesting large-scale, ongoing volcanic activity; for example, eruptions covering thousands of square kilometers with new lava flows. This raises the questions of where and when activity required to maintain the current levels of atmospheric methane is taking place, and the styles and areal extents of activity associated with resupply processes. Rather than concentrate on the eroded stubs of features that might once have been volcanoes, our focus is instead on smaller-scale occurrences of processes that can supply atmospheric methane and then determining whether these events are potentially detectable in *Cassini* data.

To date, no anomalous thermal emission that would be unambiguous evidence of endogenic volcanic activity (the "smoking gun" described by Davies et al., 2009) has been detected by any *Cassini* instrument. Some possible reasons for this non-detection are discussed below. What is certain is that Titan's atmospheric methane needs to be replenished to explain current levels. If cryovolcanism is indeed present on Titan and the crust of Titan is composed of MCH, then thermal interaction is inevitable. The release of methane by thermally destabilizing these clathrates is examined in this paper.

3. Methane clathrate hydrates (MCH)

As noted above, the upper ~5 km of Titan's crust may be rich in MCH. This shell would have formed early in Titan's history (Lunine et al., 1989, 2009; Tobie et al., 2006; Choukroun and Sotin, 2012). We propose that thermal destabilization of MCH may be a contributing mechanism for atmospheric methane replenishment.

MCH are stable at 94 K and 1.5×10^5 Pa (1.5 bars), the temperature and pressure at the surface of Titan (e.g., Sloan and Koh, 2008; Choukroun et al., 2013, and references therein). Destabilization occurs on raising the temperature of the MCH. We will show that destabilisation can be achieved through thermal interaction with lava flows which, if comprised of a 16 wt% ammonia-water mixture as modeled here, have liquidus temperatures of ≈250 K (e.g., Kargel, 1992). On the surface of Titan, it is the partial pressure (about 5% of atmospheric pressure) of methane that determines the temperature at which MCH destabilize. That temperature is ≈150 K (Fig. 1). However, if the clathrates are underneath a lava flow, then the lithostatic pressure (to which has to be added the Titan atmospheric pressure) controls the destabilisation temperature. The destabilisation temperature T_d (K) as a function of lithostatic pressure P (in bars) is given empirically (e.g., Sloan and Koh, 2008) by

$$T_\mathrm{d} = -\frac{948.17}{\log_{10}(P) - 4.90185} \tag{1}$$

The destabilisation temperature just below the surface is therefore 201 K. Beneath 1-m-, 10-m- and 100-m-thick cryolava flows of density 1000 kg/m^3, the destabilization temperatures are 201 K, 202 K and 213 K, respectively. Within the flow substrate, the destabilization temperatures are 213 K (at a depth of 1 m), 214 K (at a depth of 10 m) and 257 K (at a depth of 100 m). At a depth of 1 km, the destabilisation temperatures are 255 K for 1-m- and 10-m-thick flows, and 257 K for a 100-m-thick flow. As depth increases, the effect of flow thickness diminishes and the increasing destabilisation