Chapter 5 Cryovolcanism

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Abstract Cryovolcanism has been observed in several bodies in the solar system, most notably Saturn's moon Enceladus, where jets of water vapor and other constituents are spewed into space. Here we review cryomagmatism and cryovolcanism, which are the subsurface and surface processes resulting from the mobilization and migration of fluids generated in the interiors of icy bodies. While these have no counterparts on Earth, they are important processes in the Solar System, particularly in the icy moons of the outer Solar System. We discuss mechanisms of cryomagmatism and cryovolcanism, the possible compositions of cryomagmas, and the observational evidence found so far in extraterrestrial bodies, ranging from plumes to surface features interpreted as cryovolcanic in origin.

Keywords: cryovolcanism, cryomagma, cryolava, cryoclastic, cryovolcano, ice, water, brine, ammonia, explosive, effusive, plume, dome

5.1 Introduction

In the frigid outer Solar System, beyond the orbit of Mars, ice-rich bodies are found as satellites of the giant planets or as individual bodies orbiting the sun (dwarf planets or asteroids) (Figure 5.1). These icy bodies formed in locations of the proto-planetary nebula that harbored water and other compounds that were too volatile to remain in the warmer, inner portions of the accretionary disk. Whereas such

objects typically possess a rocky interior, they also have substantial outer layers of H₂O and possibly other ices, depending on their location. The surfaces of icy satellites exhibit a wide range of ages, from ancient, impact-scarred terrains, to youthful, actively resurfacing terrains. Endogenic geologic activity may be manifested at the surface as tectonism and cryovolcanism (i.e., icy cold volcanism). On Titan, the only icy body to host a substantial atmosphere, these endogenic processes are overprinted by atmosphere-driven processes such as eolian and fluvial activity. This chapter is concerned with cryomagmatism and cryovolcanism: the subsurface and surface processes resulting from the mobilization and migration of fluids generated in the interiors of icy bodies.

Cryovolcanism can be defined as *"The eruption of liquid or vapor phases (with or without entrained solids) of water or other volatiles that would be frozen solid at the normal temperature of the icy satellite's surface"* (Geissler, 2015). This means that ice and water (± other compounds in solution or suspension) at the conditions prevailing on the icy bodies can be considered to be the geologic equivalent of rock and magma on the terrestrial planets, and the geological features of icy bodies can be interpreted in light of this analogy. Cryovolcanism might be manifested at the surface as *cryolava flows* of aqueous or nonpolar liquids, possibly containing suspensions of solid grains and gas bubbles (Kargel, 1995). Alternatively, *cryoclastic* eruptions might produce gas-rich sprays of droplets and particles. The consequences of interior cryomagmatic processes such as diapirism or intrusions might also be visible in the surface geology (e.g., Schenk and Jackson, 1993; Head et al., 1999; Nimmo and Manga, 2002; Pappalardo and Barr, 2004; Michaut and Manga, 2014; Manga and Michaut 2017).

There are good reasons to believe that water is the dominant component of cryomagmas on most icy bodies. Although there is much insight to be gained by using our understanding of silicate volcanic processes to interpret the origins of features on icy satellites, there are some key differences in material properties, interior structures, and environments that complicate the application of models of eruptive processes and interpretation of surface features. Water has unusual thermal and physical properties, causing it to behave in ways dissimilar to silicate magmas. For example, it expands on freezing and so is denser in liquid form than in its solid state. Unlike silicate magma therefore, there is no buoyancy force to initiate ascent and eruption of cryomagmas, but yet we observe landforms, terrains, and activity that therefore require consideration of possible alternative mechanisms for delivering cryomagmas from depth to the surface. Furthermore, unlike silicates, the melting temperature of water ice increases with decreasing pressure, and so decompression melting within an ice shell will not occur. In addition, the viscosity of water is orders of magnitude less than even low-viscosity silicate magmas, meaning that the

morphologies of features produced by water cryomagmas might be difficult to interpret.

Water is, however, an excellent solvent, and so aqueous cryomagmatic liquids on icy bodies are unlikely to be pure. A range of possible additives exists, including magnesium- and sodium-based salts, ammonia, and volatile species including N₂, CO, CO₂, SO₂ or H₂ (Kargel, 1991). The presence of any such impurities is likely related to location within the Solar System: more volatile compounds would have migrated out to cooler portions of the accretionary disk during planet formation, such that cryomagmatic fluids in the far reaches of the Solar System might be more exotic, and possess a wider range of properties, than the water-based fluids expected to dominate at Jupiter and Saturn. Indeed, the surfaces of icy satellites reveal a wide range of features of possible cryovolcanic origin, and much work lies in understanding their compositions and mechanisms of formation via processes that have an imperfect analog in silicate volcanism.

Cryovolcanism might also be considered an imperfect analog to silicate volcanism because the latter is entirely the result of endogenic activity – the manifestation of internal processes that drive magmatic materials to the surface – whereas in some cases, apparent cryovolcanic activity on icy satellites appears to be a response to dominantly exogenic processes, such as solar heating of volatile materials, eruption of impact-induced melt bodies, or the tidal opening of fissures to release sprays of ocean water.

Figures 5.2 and 5.3 provide a visual comparison of the bodies of the Solar System that possess features interpreted to be cryovolcanic in nature, and **Table 5.1** lists the key characteristics of these bodies. These bodies range widely in size and origin. Ceres and Pluto are both dwarf planets, but Ceres resides in the asteroid belt just beyond the orbit of Mars, and Pluto orbits in the far outer Solar System beyond Neptune. The other bodies are satellites of the giant planets: Jupiter, Saturn, Neptune and Uranus. These bodies exhibit a range of surface compositions (Table 5.1), but we know from their bulk densities, spectroscopy, and geophysical observations, that they all contain a substantial amount of H₂O ice that may exist as an outer shell of relatively pure ice (e.g., Europa), or as a mixture of ice and rock (e.g., Ceres).

Throughout the history of outer Solar System exploration, planetary scientists have been repeatedly surprised as spacecraft data have revealed a wealth of alien landforms and features indicative of endogenic activity. The outstanding example is tiny Enceladus, a 510 km diameter Saturnian satellite that spews a constant plume of vapor and particles to feed Saturn's diffuse E-ring. Of the other bodies, Europa, Titan, Triton, and Pluto exhibit youthful surface units containing enigmatic geologic features.

The surfaces of other bodies shown in Figure 5.2 are substantially older, but nevertheless display intriguing features suggestive of endogenic activity.

Body	Diameter (km)	Bulk density (kg m ⁻³)	Gravity (m s ⁻²)	Eccen- tricity	Mean Surface Temp. (K)	Atm. Press. (Pa)/ Composition	Surface Composition	Subsurface Ocean [*]
Ceres	950	2100	0.26	0.076	235	_	H ₂ O, salts, hydrated minerals	Maybe
Europa	3122	3010	1.31	0.009	103	0 ₂ , 10 ⁻⁷	H ₂ O (salts)	Yes
Ganymede	5262	1940	1.43	0.0013	113	0 ₂ , 0.2–1.2 x 10 ⁻⁷	H ₂ O (other?)	Multiple?
Enceladus	504	1610	0.11	0.0047	80	_	H ₂ O (organics, CO ₂)	Yes
Tethys	1061	985	0.15	0.0001	86	_	H_2O	Unlikely
Dione	1123	1480	0.23	0.0022	87	_	H_2O	Maybe
Titan	5150	1881	1.35	0.0288	93	94% N ₂ <6% CH ₄ 1.5 x 10 ⁵	H ₂ O (hydrocarbons, organics)	Yes
Miranda	471	1200	0.079	0.0013	86	-	H ₂ O	?
Ariel	1158	1590	0.26	0.0012	60	_	H ₂ O (NH ₃ ?)	?
Triton	2705	2050	0.78	0.000016	58	N ₂ , 1.4	N_2	Maybe
Pluto	2374	1860	0.62	0.2488	40	N ₂ , 1.0	H ₂ O (N ₂ , CO ₂ , CH ₄ , NH ₃)	Maybe
Charon	606	1680	0.28	0.0002	53	_	H ₂ O	?

Table 5.1. Characteristics of icy bodies for which cryovolcanism has been identified or inferred.

* Some bodies are established ocean worlds based on multiple independent observations; others may have an ocean based on a single type of observation and/or theoretical models

Although understanding cryovolcanism as a possible mechanism of formation of surface features is an end in itself, arguably the greater significance of cryovolcanism is that it provides a means of delivering materials from the interiors of icy bodies to their surfaces. In recent years, thanks to groundbreaking missions that started with the twin Voyager spacecraft in the 1970s and continued through the Galileo (1989–2003), Cassini (1997–2017), Dawn (2007–2018), and New Horizons (launched 2006) missions, we have come to understand that a surprising number of icy satellites harbor global oceans beneath their ice shells (Hendrix et al., 2019). There is good evidence to suggest that global subsurface oceans of liquid water exist on Europa, Ganymede, Enceladus, and Titan, and liquid layers or oceans may even still exist within Ceres, Triton, and Pluto. In some cases, these subsurface oceans might supply the requisite radiation protection, key elements, and energy sources to provide habitable conditions for the development of simple lifeforms. Communication between the ocean and the surface in the form of cryovolcanism might therefore provide an essential pathway for biosignatures originating in the ocean to be delivered to the surface where they could possibly be detected by spacecraft instruments searching for signs of extraterrestrial life. This chapter discusses how some icy bodies harbor interior liquid layers or reservoirs, what the compositions and properties of those liquids might be, and how those fluids might be propelled from the interior of the bodies to the surfaces. Furthermore, we take a tour of the outer Solar System and examine the range of features that have been observed on icy bodies and provide some explanations for their origins.

5.2 Cryomagma Production

Any discussion of the generation of cryomagmatic fluids needs to start with consideration of how largescale bodies of fluid are formed on icy bodies. On a silicate body, melt can be generated in a number of ways, including by decompression during upwelling of warm mantle, and by addition of solidusdepressing fluids to the overlying mantle wedge by dewatering from subducting lithosphere. However, in a lithosphere composed of water ice, these processes have no analog. There are two main ways in which liquids might form within ice shells: by large-scale melting and formation of a global subsurface ocean, or by formation of discrete fluid bodies within the ice shell. We first examine the ways in which subsurface oceans are formed on icy bodies, with the notion that cryomagmatic fluids might derive ultimately from these oceans.

5.2.1 Generation of interior oceans

Over the past couple of decades, our understanding of the existence of ocean worlds in our Solar System has advanced dramatically. Spacecraft observations of icy bodies has shown that many worlds have surfaces indicative of a highly active geology, in the present or in the past. This is often our first hint that such bodies might have subsurface oceans. However, in other cases, such as Jupiter's moon Callisto, seemingly ancient cratered surfaces might also conceal subsurface oceans. Theoretical modeling of the thermal evolution and interior structures of icy bodies, using their size and bulk density as a starting point, can provide insights into when liquid water may have been present in an icy body, and whether it might persist today (e.g., Vance et al., 2018). However, it is only with more focused study by spacecraft instruments that the suite of data necessary to confirm liquid water can be assembled. For example, at the Jupiter system, the Galileo magnetometer detected an induced magnetic field at Europa, which arises as a result of the electrical response of a conducting layer to Jupiter's time-varying magnetic field (Khurana et al., 1998; Kivelson et al., 1999, 2000). The most plausible explanation is a global, near-surface salty ocean. The same technique later revealed subsurface oceans at Ganymede and Callisto (Zimmer et al., 2000; Kivelson et al., 2002).

Other methods for identifying subsurface oceans include examination of the gravity signature, variations in the shape of the satellite, satellite libration (variations in rotation rate), and obliquity (angle between the spin axis with respect to the orbital plane). These techniques provide information about the distribution of mass within the body (via the moment of inertia), and the presence of a layer (i.e., an ocean) that decouples the ice shell from the satellite interior (Nimmo, 2018). Figure 5.3 shows the inferred internal structures of the icy bodies shown in Figure 5.2.

How do subsurface oceans form within icy bodies, and what maintains them in the present day, given that these are relatively small bodies orbiting in the colder reaches of the Solar System? During the formation of the Solar System, planets and satellites accrete from smaller bodies that themselves form from the circum-solar planetary disk. As these bodies collide and grow, the impacting debris liberates its kinetic energy in the form of heat upon impact with the growing body. This heat of accretion can be sufficient to melt larger bodies, which in turn can lead to differentiation of the body, in which denser, heavy components sink towards the center of the body, leaving lighter components to rise up to form the outer layers. This reorganization of mass within the body itself generates heat as potential energy transforms to thermal energy via frictional processes. However, both of these processes take place early in Solar System history and modeling of the subsequent cooling shows that heat generated then should not persist to the present day. Furthermore, small bodies or bodies forming in less dense portions of the accretionary disk may not have experienced sufficient accretional heating to fully differentiate.

On planetary bodies that are dominantly silicate in composition, decay of radioactive elements takes over as the dominant heat-generation mechanism, initially through short-lived radioisotopes with half-lives on the order of 0.1– 10 Ma (dominantly ²⁶Al, but with contributions from ⁴¹Ca, ⁵³Mn, ⁶⁰Fe, and others), and then through long-lived radioisotopes with half-lives on the order of 1 Ga or greater (e.g., ⁴⁰K, ²³²Th, ²³⁵U, ²³⁸U). On Earth, this is the continuing source of internal heat. In the outer Solar System, however, the icy bodies have much smaller amounts of radioactive elements in part because of their smaller sizes, but also because they contain lower proportions of silicates overall. The bulk densities given in Table 5.1 provide a measure of the proportion of silicate materials present within the body; densities less than ~3000–4000 kg m⁻³ indicate a substantial proportion of H₂O, and given that radioactive elements are lithophiles (i.e., they concentrate in silicates), bodies with low densities will have proportionally less radioactive material available for internal heating.

We are left to conclude that, except for larger bodies, radiogenic heating may not be sufficient to

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generate and maintain an internal ocean in low-density icy satellites. However, orbital dynamics reveal that tides raised within the bodies by gravitational interactions with their parent planet can be a significant source of internal heat. This requires special circumstances and explains why relatively few icy bodies exhibit evidence of ongoing internal melting; the Jupiter System is a good example (Wiesel, 1980). Io, Europa, and Ganymede experience a 4:2:1 orbital resonance, so that, for every one orbit that Ganymede completes around Jupiter, Europa completes two, and Io completes four (Figure 5.4a). Thus, the satellites find themselves in periodic alignments, such that the repeated gravitational interactions maintain elliptical orbits with a finite eccentricity. This eccentricity means that during its orbit, the satellite will find itself at a variable distance from Jupiter, ranging from apojove (farthest point) to perijove (closest point). Therefore, Jupiter exerts a varying gravitational force on the satellite, which responds by deforming to a variable extent (Figure 5.4b). Heat is generated in the interior in response to the mechanical deformation caused by the continuously changing shape of the satellite (tidal flexing).

Nowhere is tidal heating more clearly expressed than at Io, the closest of the Galilean satellites to Jupiter, and the most volcanically active body in the Solar System (see Chapters 2 and 4). In Io's case, the volcanism is silicate-based: heating is so extreme that water cannot exist there in any significant amount. However, moving out from Io to Europa's orbit, the gravitational interactions are less extreme and a ~100 km thick water/ice layer exists over a differentiated interior (Figure 5.3). Because Ganymede and Callisto are farther from Jupiter, they experience successively less tidal heating. Their bulk densities are substantially less than 2000 kg m⁻³, indicating a far smaller proportion of silicates than Europa. Callisto may only be partly differentiated, which might have resulted from slower accretion (and therefore less primordial heating) that far out in the planetary nebula, such that ²⁶Al was exhausted before accretion was completed. Nevertheless, Ganymede and Callisto are understood to have liquid oceans (Khurana et al., 1998; Kivelson et al., 1999; 2002; Zimmer et al., 2000). However, the substantial ice shell thicknesses (150 km and 80–150 km, respectively; Kivelson et al., 2002) act as an impediment to possible cryovolcanism, with Callisto exhibiting an ancient cratered surface indicative of minimal to no endogenic activity being expressed at the surface.

In the Saturn system, Titan, with a substantial eccentricity deriving from its origin as reaccreted debris from the impact-induced destruction of several satellites (Asphaug and Reufer, 2013), but having a large orbital distance from Saturn, is thought to be mildly tidally heated. However, Titan is large enough to have sufficient silicate component that allows radiogenic heating to be significant. As a consequence, increasing pressure with depth in the thick ice shell, which reduces the ice melting point,

allows formation of an ocean. Furthermore, at Saturn and beyond, the presence of ammonia on icy satellites, which reduces the melting point of ice mixtures, may play an important role in the generation and maintenance of oceans (e.g., Kargel, 1995). For other Saturnian satellites, tidal heating is critical. Diminutive Enceladus has a 2:1 resonance with Dione, and had a past 2:1 resonance with Janus. Miranda-Ariel-Umbriel are close to 4:2:1 resonance, and Tethys-Mimas experience a 2:4 resonance (Dougherty and Spilker, 2018). Resonances arise due to the outward drift of satellites away from their parent planet over time, which means that satellites can evolve into and out of resonances. This in turn means that the extent of tidal heating and endogenic geologic activity on a given body can vary over time, and in the present day we might observe only older terrains and landforms representing the surface manifestation of past oceans that have since refrozen.

Triton is notable among the icy satellites for being in a retrograde orbit around Neptune, i.e., it is orbiting in the opposite direction to Neptune's rotation. This fact, along with its size and bulk composition (density) similarity to Pluto (Table 5.1), suggest that Triton did not form as a satellite within the Neptune system, but rather originated as an object within the Kuiper Belt (the belt of small, icy bodies left over from Solar System formation orbiting beyond Neptune; Figure 5.1), and was captured into orbit around Neptune. The capture and the subsequent highly eccentric orbit would have led to a major heating episode of the body, keeping Triton in a molten state for a billion years (Ross and Schubert, 1990). However, the capture is thought to have occurred early in Solar System history, such that the early heating could not be responsible for the youthful terrain observed today. Triton's present eccentricity is low (0.000016; Table 5.1) and so eccentricity-driven tidal heating is negligible today. Radiogenic heating, while greater than tidal heating, is not sufficient to maintain an ocean on Triton. However, a relatively slow decay of higher past eccentricity, combined with radiogenic heating, may allow a subsurface ocean to exist today, especially if ammonia is a significant component (Gaeman et al., 2012). Furthermore, Nimmo and Spencer (2015) note that Triton's high orbital inclination (angle between the satellite's orbit and Neptune's equatorial plane) drives its obliquity, leading to an additional source of diurnal cycles of deformation. This obliquity-driven tidal heating could therefore contribute energy for endogenic activity.

Closer to home, Ceres is the largest body in the asteroid belt and the only dwarf planet in the inner Solar System. Its mean density of 2162 kg m⁻³ (Park et al., 2016; Table 5.1) suggests a water:rock ratio of 25:75 by mass (Castillo-Rogez et al., 2019), indicating that it is the most water-rich body in the inner Solar System (after Earth). Early heating by short-lived radioisotope decay likely triggered global melting of Ceres' volatiles, and allowed for the formation of an ice-rich crust (30-40 % ice, 60-70 % silicates) and a rocky mantle (Castillo-Rogez and McCord, 2010). A variety of evolutionary scenarios support the preservation of fluids within Ceres until the present day. The "mudball model" suggests that a global, muddy ocean existed for the first 3 Ga of Ceres' evolution and that regional, muddy seas consisting of hundreds of kilometers (depth) of liquid still persist today (Bland and Travis, 2017; Travis et al., 2018). Conversely, a liquid brine layer that is tens of kilometers thick may currently exist at the base of Ceres' crust (Neveu and Desch, 2015; Castillo-Rogez et al., 2019; Quick et al., 2019). This brine layer could represent the remnants of a frozen ocean (Neveu and Desch, 2015; Ammannito et al., 2016; Castillo-Rogez et al., 2018; 2019; Quick et al., 2019), the preservation of which was made possible by the abundance of carbonate and chloride salts in the liquid, and gas clathrate hydrates in the crust (Neveu and Desch, 2015; Castillo-Rogez et al., 2018, 2019). Clathrate hydrates are a form of ice in which gas molecules are trapped within the water ice lattice cages. These materials are highly insulating and could have inhibited freezing of residual fluids in the interior of the dwarf planet. Recent endogenic activity on Ceres could be driven by: 1) convection in Ceres' heterogenous crust which is replete with salts, clathrates, ice, and phyllosilicates (De Sanctis et al., 2016; Bland et al., 2019); 2) convection in a muddy mantle (Neveu and Desch, 2015; Bland and Travis, 2017; Travis et al., 2018; Ruesch et al., 2019); 3) the continued freezing of a remnant ocean (Neveu and Desch, 2015; Quick et al., 2019); 4) hydrothermal circulation triggered by heat supplied from large impacts (Bowling et al., 2019; Hesse and Castillo-Rogez, 2019); or 5) some combination of these mechanisms (Raymond et al., 2020).

Radiogenic heating in Pluto's core was likely high enough to produce a subsurface ocean that persists today (Robuchon and Nimmo, 2011; Bierson et al., 2018). The presence of this ocean is consistent with observational evidence from the New Horizons mission that shows widespread extension on Pluto (Moore et al., 2016), and with modeling results that suggest that Pluto should not be in a contractional regime (Hammond et al., 2016). Additionally, the current location of, and the positive gravity anomaly associated with, Sputnik Planitia, a large N₂-ice filled impact basin, necessitates the presence of a thick subsurface ocean beneath an ice shell that has been locally thinned (Johnson et al., 2016; Nimmo et al., 2016). Tidal heating would have diminished early in Pluto's history (Cheng et al., 2014), and the amount of radiogenic heating within Pluto would not prevent this ocean from freezing over billion-year timescales. However, maintaining this ocean to the present day is possible if it contains substantial amounts of salts, ammonia, and/or methanol (Johnson et al., 2016; Nimmo et al., 2016). An insulating "cap" of clathrates at the base of Pluto's ice shell may also prevent freezing, allowing for a thick subsurface ocean to be maintained within Pluto today (Kamata et al., 2019). The pattern of widespread

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normal faulting (Moore et al., 2016) suggests that Pluto is undergoing global expansion (Keane et al., 2016), indicative of gradual freezing of a subsurface ocean.

The > 100 kg m⁻³ density contrast between Pluto and its satellite Charon (Table 5.1) suggests that they are compositionally distinct bodies with different rock/ice ratios (Bierson et al., 2018). An ice-rich Charon could have been produced if it formed as a result of a giant impact into Pluto. A grazing impact would have caused a portion of the impactor to be captured into orbit around Pluto, and formation of an icy disk of material that subsequently reaccreted onto this impactor could have formed Charon with the ice-rich composition that it has today (Canup, 2011). The presence of pervasive extensional features on Charon's surface (Moore et al., 2016) indicate that it once harbored a subsurface ocean, the gradual freezing of which caused global expansion (Stern et al., 2015; Beyer et al., 2017). While energy from the Charon-forming impact would have been insufficient to induce ocean formation (Canup 2005; 2011), Charon likely contained a high enough silicate fraction and a high enough ammonia content (Grundy et al., 2016) for an ocean to form from melting of some fraction of an early ice shell. Additional energy imparted by tidal heating (Barr and Collins, 2015) could have also triggered the formation of an early ocean. This ocean may have undergone multiple episodes of freezing and melting (Beyer et al., 2017; Desch and Neveu, 2017). Such an ocean is unlikely to persist today (Hussmann et al., 2006), but cryovolcanic activity at Charon's surface may have been triggered during the course of its gradual freezing (Manga and Wang, 2007; Desch and Neveu, 2017; Beyer et al., 2017; 2019).

In summary, there are a variety of heating mechanisms that can produce internal oceans from which cryomagmas might be derived, some satellites retain these oceans in the present day, and others may formerly have had oceans that produced recognizable surface features prior to and during the refreezing of the subsurface water.

5.2.2 Formation of liquid bodies within the ice shell

Rather than originating from a subsurface ocean, it is possible that some cryovolcanic eruptions might be sourced from discrete liquid reservoirs within the ice shell. As we will see in Section 5.4, there are significant difficulties to be overcome in delivering water from a subsurface ocean all the way to the surface of an icy body, especially for those bodies having thicker ice shells. Other bodies may never have experienced sufficient heating to generate a global liquid water layer. We therefore consider the possibility that isolated fluid reservoirs might form within the ice shell, which would leave shorter distances for ascending fluids to traverse to the surface. Two main mechanisms of forming subsurface reservoirs considered here are formation in situ due to melting induced by, for example, an ascending

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plume of warm ice; and the emplacement of liquid intrusions in the ice shell sourced from the ocean below.

To understand how these mechanisms would work, we first need to consider the physical, mechanical, and thermal state of the ice layer. With depth into the interior of an ice lithosphere, the temperature rises, and there may be a transition in the way that ice response to applied forces. The cold, outermost portion of the ice shell typically responds in a brittle manner, revealed by the presence of tectonic features, fractures, and the maintenance of topography. At depth, however, the response to applied stresses might become ductile, and if certain threshold conditions are met, warmer ice may thermally convect: warm ice originating at the base of the ice shell wells up because it is less dense than the colder surrounding ice. As it rises and encounters the overlying cold, brittle "lid", the warm ice spreads, cools, increases in density, and then proceeds to sink back down into the warmer interior. In this way a convective circulation is set up beneath the overlying stagnant, conductive lid (**Figure 5.5**).

Numerous studies have investigated the conditions for convection within the ice shells of icy bodies (e.g., Sohl et al., 2003; Tobie et al., 2005; Han and Showman, 2005; Mitri and Showman, 2005, 2008; Barr and Showman, 2009; Nimmo and Bills, 2010; Lefevre et al., 2014). The onset of convection is defined by the Rayleigh number, $Ra = \alpha \rho g \Delta T b^3/(\kappa \eta)$, where α is the coefficient of thermal expansion, ρ is the ice density, g is acceleration due to gravity, ΔT is the temperature difference between the bottom and top of the ice shell, b is the ice shell thickness, κ is the ice thermal diffusivity, and μ is ice viscosity, which depends on temperature, ice grain size, and strain rate in the ice shell. The critical Rayleigh number for convection is on the order of $10^3 - 10^4$ (Barr and Pappalardo, 2005), and convection is facilitated in thick ice shells with large temperature contrasts and low ice viscosities. If some portion of the ice shell is convecting, this has important implications for heat transfer from the warm ice–ocean interface to colder, shallower levels (and hence for the thermal gradient), the mechanical response to applied stresses (which is temperature-dependent), and the formation of liquid bodies by melting or intrusion. Figure 5.5b shows the consequences for thermal gradients of convective vs. purely conductive heat transfer through an ice shell.

Sotin et al. (2002) developed a numerical model to investigate the consequences of thermal convection on Europa. By accounting for the strong temperature-dependence of ice viscosity, they found that, for an ice shell undergoing convection and experiencing diurnal tides, tidal energy becomes focused in the rising warm, lower-viscosity plumes, such that melting can be achieved at shallow depths as the plume spreads laterally. Up to 10% of the upwelling ice could form a melt lens at the top of the

plume (Figure 5.6), which could increase in size with continued convective input of warm ice from below. Any impurities (such as salts or ammonia ices; Kargel et al., 2000; see Section 5.3) in the nearsurface ice would lower the melting temperature and facilitate formation of eutectic mixtures above the plume (Schmidt et al., 2011).

An alternative mechanism for forming melt reservoirs is through the emplacement of cryointrusions such as sills or laccoliths, derived from the ocean, into the near-surface layers (Collins et al., 2000; Dombard et al., 2013; Michaut and Manga, 2014; Craft et al., 2016; Manga and Michaut, 2017). A transition from upward propagation of a vertical fluid-filled fracture to horizontal sill emplacement might be induced on encountering a rheologic interface (e.g., the brittle-ductile transition), a stress regime that causes the lateral injection of fluid rather than continued ascent (e.g., compressional nearsurface stresses), or a density change due to porosity of near-surface ice. However, detailed models of fracture propagation on Europa suggest this the transition from vertical to lateral emplacement is difficult to achieve (Craft et al., 2016). Assuming that this transition could take place, Michaut and Manga (2014) and Manga and Michaut (2017) modeled the process of sill formation on Europa fed by a dike-like fluid-filled fracture originating at the ocean-ice interface (Figure 5.7). They considered that the stresses induced in the base of the ice shell by a pressurized ocean (Manga and Wang, 2007) were sufficient to initiate a vertical fracture to feed the sill, and assumed the sill formed at the brittle-ductile transition. Addressing the processes of sill intrusion, cooling and freezing, flexural support of the water weight, and topographic relaxation, Michaut and Manga (2014) found that deep sills are favored over shallow sills by the driving pressures attainable in a pressurized ocean $(10^4 - 10^6 \text{ Pa}; \text{Manga and Wang},$ 2007), and that deep sills, by causing downward flexure of the basal ice as opposed to uplift of the surface, would help to maintain ocean pressure and hence continue to feed the sill.

Another mechanism for producing localized melts involves the tidal working and concentration of stresses at fractures penetrating the ice shell from the surface (e.g., Nimmo and Gaidos, 2002; Kalousová et al., 2016; **Figure 5.8**). In this instance, existing fractures experience strike-slip motion resulting from the time-varying diurnal stress field. The resulting heating of the ice in proximity to the fracture is predicted under some circumstances to cause melting, and potentially form a fluid body (Nimmo and Gaidos, 2002; Kalousová et al., 2016). Finally, large impacts are expected to produce substantial amounts of melt, and may in some cases penetrate to an underlying ocean. Cooling and refreezing of the melt can produce an isolated reservoir in the subsurface, as has been proposed to

explain features within impact craters on Europa (Steinbrügge et al., 2021) and Ceres (Hesse and Castillo-Rogez, 2019; Bowling et al., 2019).

Formation of melt reservoirs within the ice shell, whether by in situ melting or sill emplacement, may have a range of surface expressions resulting from the topographic, tectonic, and thermal responses to the presence of the body of water (Sotin et al., 2002; Schmidt et al., 2011; Dombard et al., 2013; Michaut and Manga, 2014; Manga and Michaut, 2017; Figures 5.6 and 5.7). Regardless of how the presence of such reservoirs is manifested at the surface, subsurface reservoirs provide a potential source of cryomagmas if conditions prevail to deliver such fluids to the surface (see Section 5.4.1). The longevity of reservoirs filled with negatively-buoyant fluid reflects a balance between cooling/freezing of the water and the tendency to subside wholesale or percolate downwards under gravity (Michaut and Manga, 2014; Kalousová et al., 2014; Lesage et al., 2020). Large reservoirs would have more pronounced effects on the surface than small reservoirs, and be a more voluminous source for cryovolcanic eruptions. Large reservoirs would also heat the surrounding ice to a greater extent, facilitating the downward motion of the dense fluid as a negatively buoyant diapir, and potentially removing the reservoir as a source of cryomagma. The extent to which these competing processes influence Europa's ability to erupt cryomagmas is still under debate (Michaut and Manga, 2014; Kalousová et al., 2020).

The possible mechanisms for the formation of isolated fluid bodies within an ice shell have received most attention in the context of Europa in an effort to understand the formation of a range of surface features observed there. However, the same considerations could be used to explore conditions leading to melt reservoir formation on other icy bodies. For any icy body that does not currently possess a global water layer (as far as we know), one might consider the formation of liquids at the silicate–ice interface due to localized thermal activity.

The mechanism of formation of a fluid reservoir in an ice shell has implications for the composition of the resulting cryomagmas. For in situ melting by warm ice plume, the composition of the resulting liquid in the reservoir would depend partly on the composition of the ascending warm ice but also on the composition of the ambient near-surface ice, both of which may contain non-ice components (similar to solicate mantle plumes on Earth). The composition of fluid in reservoirs fed by fractures originating at the ocean–ice interface would reflect ocean composition. In both cases, fractionation within the intrusive body during cooling would change the composition of the residual liquid as the water froze, leaving a liquid phase successively more concentrated in dissolved non-ice compounds. In the next

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section, we consider the range of plausible compositions of fluids and ices across the outer Solar System.

5.3 Cryomagma Compositions and Properties

There are a number of ways by which the compositions of cryomagmas on icy bodies can be constrained, including multispectral measurements of surface compositions, theoretical modeling based on the inferred stability of different materials at a given location in the Solar System, inferences from geomorphology, and in rare cases, direct sampling of erupting plumes. Many icy satellite surfaces (e.g., Europa and Enceladus) have a dominantly water ice spectral signature (Table 5.1), and only on close examination are the subtle signatures of non-ice compounds revealed. Other bodies, such as Ceres, Titan, Triton, Pluto, and Charon, show more complex surface compositions (Table 5.1). In seeking to link surface compositions to the compositions of cryomagmas, we have to ascertain whether the compositional signature is spatially coincident with a particular type of geologic feature, and furthermore, whether that feature is plausibly derived from the interior.

As discussed in Section 5.1, water is expected to have condensed in abundance beyond the orbit of Mars. Candidate cryomagmas are therefore likely to be water-based, but some amounts of non-H₂O constituents are likely to be found in aqueous fluids. For bodies such as Europa and Enceladus, whose oceans lie in direct contact with the rocky interior (Figure 5.3), exchange between the silicates and ocean water via aqueous alteration or hydrothermal processes would introduce a variety of compounds into the ocean. Multispectral observations of Europa's surface, for example, indicate a range of compositions associated with dark-toned geologic features that imply provenance from the interior, and thus provide evidence of the ocean composition. These compounds include magnesium and sodium sulfates, sodium carbonate, sodium chloride (McCord et al., 1998, 1999) and radiolytically-altered sulfurrich species including sulfuric acid hydrate, SO₂, H₂S, and sulfur polymers; Carlson et al., 1999, 2002), suggesting deposition at the surface by evaporation of water derived from briny or acidic ocean. The Galileo magnetometer results at Europa also require a salty ocean (Khurana, 1998; Kivelson 1999). Saltrich particles have also been detected in Enceladus' plume, implying a briny ocean beneath the ice crust (Postberg et al., 2009).

At Titan, gravity data imply the presence of a dense (briny) ocean (Mitri et al., 2014), with candidate salt species including sodium chloride, magnesium sulfate, and ammonium sulfate (Fortes et al., 2007; Mitri et al., 2014; Vance et al., 2018). Most, but not all, models predict the existence of a high-pressure ice layer sandwiched between the ocean and the silicate core (Sohl et al., 2003; Tobie et al., 2005;

Fortes, 2012; Mitri et al., 2014; Vance et al., 2018), which would inhibit exchange of materials between the core and ocean. However, more recent work suggest that the high-pressure ice layer itself could have undergone melting and convection to allow communication with the ocean and hence influence ocean composition (Kalousová and Sotin, 2020).

Ammonia has also been considered to be a plausible component of icy bodies, based on theoretical modeling of its condensation during Solar System formation (Lewis, 1972; Prinn and Fegley, 1989; Kargel, 1995). Being more volatile than water, it is unlikely to have condensed in any quantity at Jupiter's relatively warm location in the protoplanetary disk but, together with methane, it is found in the Saturn system and beyond. Ammonia has been directly detected in Enceladus' plume, for example (Waite et al., 2009). In the cold, far reaches of the outer Solar System, CO, CO₂, and N₂ are likely to be found. Compounds such as methanol and formaldehyde are observed in comets (representing primitive Solar System species) and might be minor constituents of the more distant bodies (Kargel, 1995; Geissler, 2015).

Table 5.2 and **Figure 5.9** summarize the properties of some proposed cryomagmas (based on Kargel et al., 1991; Kargel, 1995). The addition of salts to pure water has the effect of slightly lowering the melting temperature of the solid phase, which might facilitate the production of the liquid phase (Kargel, 1991). However, adding salts also increases the density of the melt, which serves to exacerbate the problem of negative buoyancy with respect to the ice shell. The addition of certain salts can result in development of a melting interval, in which solid and liquid phases can coexist, which might facilitate the development and eruption of icy slurries (Quick and Marsh, 2016), with more complex rheologies than the liquid phase alone.

Ammonia should make up to 15% of the mixture of ices in a fully equilibrated primordial nebulae (Geissler, 2015), and ammonia is indeed seen in abundance in comets and extrasolar nebulae (Greenberg, 1998; Rizzo et al., 2014). However, even if ammonia is present in significant quantities within a given body, it is highly reactive and would be consumed by many reactions that would occur in an ocean, producing ammonium salts (e.g., Kargel, 1992; Fortes et al., 2007). Therefore, there may only be a limited range of conditions under which free ammonia–water solutions would exist. However, in recent years, ammonia has been inferred at Titan (Niemann et al., 2010), detected at Enceladus (Waite et al., 2009), and possibly at Miranda (Bauer et al., 2002), Pluto (Dalle Ore et al., 2019), and Charon (Brown and Calvin, 2000), but with the exception of Pluto and Charon, only in minor quantities. The residence time of ammonia on the surface of an icy body is limited because it is readily photolyzed to N₂

or dissociated in the space environment to form other compounds; any detections of ammonia on the surfaces or in the atmospheres of icy bodies would suggest that it is being replenished somehow. Ammonium chlorides and other ammonium compounds have been detected on Ceres (Rivkin et al., 2006), so ammonia as a precursor seems plausible there. The question remains as to what extent free ammonia is a factor in cryomagmatism.

Liquid	Melting point (K)	Liquid density (kg m ⁻³)	Liquid viscosity (Pa s)	Solid density (kg m ⁻³)
Water 100% H ₂ O	273	1.000	0.0017	0.917
Brine 81.2% H ₂ O, 16% MgSO ₄ , 2.8% Na ₂ SO ₄	268	1.19	0.007	1.13
Ammonia–water 67.4% H ₂ O, 32.6% NH ₃	176	0.946	4	0.962
Ammonia–water–methanol 47% H ₂ O, 23% NH ₃ , 30%C ₃ OH	153	0.978	4000	-
Data from Kargel (1005)				

Table 5.2 Properties of candidate cryomagmas (eutectoid composition).

Data from Kargel (1995)

Accepting that ammonia might be present in oceans or water reservoirs on icy bodies, we can explore its properties (Table 5.2). Ammonia substantially depresses the melting temperature of aqueous fluids, thereby lowering the thermal barrier to melting and the formation of oceans or intra-ice liquid reservoirs. Methanol, which is highly soluble in water even at low temperatures, has a similar effect. If these compounds are present within a cold ice shell, production of ammonia–water (or ammonia–methanol–water) cryomagmas is energetically more favorable than production of brines. Even modestly sized bodies might be able to achieve a melting temperature of 176 K if substantial concentrations of ammonia are present (Croft et al., 1988; Kargel 1995). The ice shell overlying an ammonia–water ocean would be colder and more rigid than for brine or pure water oceans, and it may therefore be stable against thermal convection (e.g., Nimmo and Bills, 2010), which has implications for transport processes through the ice shell. Although methane does not substantially reduce the melting point of ice, it marginally reduces the melting point of N₂ to 62 K and so might be a useful constituent of cryomagmas on say, Triton and Pluto (Kargel, 1995).

Ammonia in water reduces the density contrast between the fluid and ice phases (Croft et al., 1988), to the point where the densities of ammonia–water mixtures get close to ice densities, especially if ammonia is also incorporated in the solid phase. Although the fluid phase is only ever positively buoyant with respect to pure ice under a restricted set of conditions (Croft et al., 1988), the presence of

ammonia at least reduces the barrier to ascent (see Section 5.4.1). Another key property of ammonia– water mixtures is the existence of a melting interval and the ability to form partial melts. This means that a range of liquid concentrations might be produced and extracted from the ice phase, and/or that crystal-rich slurries might be produced. Kargel et al. (1991) showed ammonia–water slurries develop substantially more complex rheologies and higher viscosities, in some cases approaching those of silicate magmas.

Kargel (1995) summarized the properties of a range of possible cryomagma compositions, and made predictions for the types of cryovolcanism that would result. Figure 5.9b shows the "mobility index" for various cryomagmas which represents the viscosity of a given fluid scaled by the gravity of a given icy body (Kargel et al., 1991). Water or brines, with their low viscosities, would tend to produce smooth, low-lying ponds or plains features, rather than construct observable topography, assuming that a mechanism was capable of delivering the fluids to the surface. Near-peritectic ammonia–water mixtures have higher viscosities than water or brines, similar to that of corn syrup at room temperature. The higher viscosities of two-phase slurries of ice crystals suspended in the liquid phase of ammonia–water mixtures could lead to the construction of features with discernible relief, such as flow fronts, domes, and ridges. With the addition of methanol, which is more likely to be present on satellites of Uranus and Neptune, viscosities increase by orders of magnitude, approaching those of andesite and dacite (Kargel et al., 1991). Such viscosities would be capable of producing features with substantial relief. High-relief features could also be produced by effusions occurring dominantly in a solid state (e.g., warm or impure water ice), with perhaps small amounts of melt lubricating crystal boundaries (Jankowksi and Squyres, 1988).

Looking far out in the Solar System to Triton and Pluto, where N_2 and methane are common, fluid phases of these substances would have extremely low viscosities (lower than water; Table 5.2) and any features produced would be transient because of their propensity to flow in the solid state or to sublime, even at the cold surface temperatures there (38 K at Triton, 44 K at Pluto).

In summary, compositions of most cryomagmas are likely to be dominantly water, with the addition of non-H₂O components such as salts (e.g., chlorides or sulfates of magnesium, sodium, and ammonium) or perhaps more exotic species (e.g., free ammonia or methanol). The specific composition, together with the ambient conditions on bodies ranging from Ceres to Pluto, can produce a wide range of cryomagma thermal and rheological properties, which might explain the diverse geomorphologies observed on icy bodies.

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5.4 Cryomagma Eruption Mechanisms

We have little in the way of direct evidence of cryovolcanism on icy satellites. Only in the case of active plumes on Enceladus, and perhaps Europa and Triton, do we have any direct data with which to constrain models of cryovolcanic eruption mechanisms. In the majority of other cases, we have observations of surface features that have been interpreted as cryovolcanic, but which may also plausibly be explained by other processes. The difficulty of understanding the origins of such features is exacerbated by that fact many of the characteristics of the ice shell and ocean of a given body are not known with certainty, and these characteristics are vital for developing well-constrained models of formation mechanisms. Among these poorly known characteristics are the thickness of the ice shell and composition of the ocean (H_2O + impurities), which control the temperature of the ocean and the temperature gradient within the ice shell, and by extension the mechanical and rheological properties of the ice shell. These characteristics are key to understanding whether thermal convection operates within the ice, whether and how far fractures can propagate, whether fluid reservoirs form within the ice shell, and whether conditions are favorable for ascent from depth to the surface.

There are two key conditions that must be satisfied for liquid cryogmagmas to reach the surface: the negative buoyancy of the fluid phase must be overcome, and some form of conduit between the liquid body and the surface must be established. Alternatively, the dissociation of any gas clathrate hydrates or the movement of ice in the solid state may also provide mechanisms of delivery of subsurface materials to the surface, but these are not considered cryovolcanic processes per se. **Figure 5.10** illustrates the wide range of mechanisms that have been proposed to promote delivery of materials (solid or fluid) from the interior to the surface (or near-surface) of icy bodies. Note that not all of these mechanisms (and potentially none of them) operate on a given body. Even so, they serve as a starting point to discuss issues surrounding cryomagmatism and cryovolcanism, from which we might assess their validity.

5.4.1 Overcoming negative buoyancy

A major impediment to the ascent and eruption of cryomagmas is the fact that water is denser than ice, which means that, on differentiated bodies, an ocean underlying an ice shell is in a stable configuration and, barring additional factors, will tend to stay where it is. This is not the case for silicate volcanism, where the molten phase is less dense than the solid phase and the resulting positive buoyancy will encourage ascent through the country rock. Only in the shallow crust, where rock densities are generally lower (due to porosity and/or composition) will buoyancy-driven ascent stall, requiring pressurization of the magma or exsolution of volatiles to allow continued ascent. Similarly, the negative buoyancy of water with respect to ice therefore requires some specific conditions to be met if ascent and eruption is to be possible. One option is to somehow make the liquid buoyant, which might be achieved by adding low-density substances or exsolved volatiles to the water; or by advocating a densifying agent such as a silicate grains (or, less effectively, salts) to the ice shell. Another option is to pressurize the fluid phase to overcome the negative buoyancy and drive the fluid to the surface.

In considering a subsurface ocean, Manga and Wang (2007) proposed that pressurization of an entire ocean might be achieved as it progressively freezes and the ice shell thickens. The volume change from liquid to solid state would not only pressurize the ocean but also lead to tensile stresses that might fracture the ice shell to provide pathway to the surface for the ocean water. Indeed, extensional tectonic features on bodies such as Tethys, Dione, Miranda, and Ariel are attributed to global expansion caused by freezing of their oceans. By modeling ocean pressurization due to ice shell thickening, Manga and Wang (2007) found that freezing of a 1–10 km ice thickness globally would induce ocean pressures exceeding the tensile strength of ice (~10⁴–10⁶ Pa; Kehle, 1964; Dempsey et al., 1999; Lee et al., 2005; Litwin et al., 2012), although the pressures would not be sufficient to drive ocean water to the surface. Instead, water-filled intrusions within the ice shell might be formed. For bodies smaller than Europa, freezing can generate higher overpressures, which might be adequate for ascent and eruption of ocean water. For small bodies whose ocean has frozen completely, it is therefore plausible that they experienced large-scale surface flooding by such processes.

For liquid water or brine reservoirs contained within the ice (see Section 5.2.2) partial freezing of those reservoirs would induce pressurization, provided the walls of the reservoir did not deform in response to the increase in volume resulting from the transition from denser liquid to less dense solid phase (Fagents, 2003; Lesage et al., 2020). If a fracture opens by some means between the reservoir and the surface (see Section 5.4.2 for discussion of fracture formation mechanisms), the excess pressure may drive the fluid to the surface, the eruption lasting as long as the pressure overcomes the negative buoyancy and the weight of the fluid in the column. Lesage et al. (2020) demonstrated that small (0.1 to 2.6 km diameter), shallow liquid reservoirs on Europa would freeze on a shorter timescale than the surrounding ice could deform by viscous relaxation, meaning that pressurization, wall fracturing, and eruption of reservoir fluid would be the likely consequence. Similar mechanisms of pressure-driven

ascent have been proposed for Ceres (Neveu and Desch, 2015; Quick et al., 2019) and Pluto (Martin and Binzel, 2021). It has also been speculated that pressure gradients associated with, for example, diurnal tidal stresses or surface topography could induce ascent of melt bodies contained within the ice shell, effectively pumping fluid to the surface (Wilson et al., 1997; Collins et al., 2000; Showman et al., 2004; Mitri et al., 2008). On the other hand, Kalousová et al. (2014, 2016) found that water bodies produced above warm ice plumes or at frictionally heated strike-slip faults on Europa would be gravitationally unstable and would migrate down to the ocean on geologically short time scales.

Rather than pressurizing the fluid phase, another mechanism of inducing ascent might be the presence of impurities (i.e., non-H₂O compounds) dissolved within the ocean water that reverses the negative buoyancy. Unfortunately, in the Jupiter system, the most abundant non-H₂O compounds in the ocean are thought to be salts such as magnesium and sodium sulfate hydrates, sodium carbonate, or sodium chloride (McCord et al., 1998, 1999, 2002), which only serve to increase the liquid density (Table 5.2; Figure 5.9a; Section 5.3). The same is true in the Saturn system and beyond. However, the possible presence of ammonia in ocean water would decrease the density contrast and diminish the buoyancy barrier to ascent, but is unlikely to succeed in making the fluid phase positively buoyant.

Alternatively, if the additional species is a dissolved volatile, exsolution and the formation of bubbles can have a dramatic effect. Likely candidates throughout the Solar System are CO₂, SO₂, N₂, H₂, and CH₄, based on cosmic abundances (Crawford and Stevenson, 1988; Kargel 1995) and detections made, for example, in the plume of Enceladus (Waite et al., 2009, 2017; Perry et al., 2015). Although dissolved when under pressure (i.e., at depth beneath an ice shell), if the fluid is decompressed the volatiles can exsolve, form bubbles, and the bulk fluid becomes less dense than the surrounding ice and therefore tends to rise. There are at least two scenarios in which volatile-driven eruptions might be possible. First, if a fracture penetrates to an ocean or a subsurface fluid reservoir, water would rise to its level of hydrostatic equilibrium, at about nine-tenths of the way to the surface, in the same way that an iceberg floats at an equilibrium with nine-tenths of its volume below the water. Under the action of diurnal tides, which induce cycles of tension and compression in the ice shell, water or water-ice slurries may be pumped farther up (Greenberg et al., 1998). However, a more dramatic effect will be the consequences of exposure of the water column to the near-vacuum ambient conditions of most icy bodies. The surface of the water column will immediately start to boil because the vapor pressure of water exceeds the ambient pressure (~0 Pa). In addition, any volatiles dissolved within the water (e.g., CO_2 , SO_2 , CH_4) would exsolve to produce bubbles. The combination of H_2O vapor boiling at the surface of the water column, and other volatiles exsolving and rising through the water, would release of spray of droplets and vapor, which may then rise up through the remaining length of the fracture and evolve into a jet or plume. Some combination of these processes is likely operating to drive Enceladus' plume (see Section 5.5.1; Porco et al., 2014; Matson et al., 2012; Goldstein et al., 2018; Postberg et al., 2018).

Second, Crawford and Stevenson (1988) proposed that fractures propagating upward into Europa's ice shell from the ocean—ice interface would possess a low-pressure zone at the fracture tip that would allow any volatiles (e.g., CO₂, CO, SO₂) dissolved within the water to exsolve into a discrete gas phase. The elastic properties of the ice would likely prevent penetration of the fracture all the way to the surface, and the cracks would instead pinch off, enclosing some proportions of gas and fluid. The gas phase would impart buoyancy to the mixture and potentially drive the crack all the way to the surface where it might erupt a vapor-rich spray and/or produce a limited-volume liquid effusion.

As an alternative to lowering the density of the liquid phase, conditions might prevail in which the density of the ice shell is greater than that of the liquid. Early in the evolution of a body, before differentiation has advanced substantially, heating might lead to the formation of fluids in the body's outer portion. If this region is still rich in rocky material, even dense brines would be positively buoyant. For a body in which differentiation has progressed substantially, the influx of dense material (e.g., micrometeorites) onto the surface of an icy body might increase the density of the ice shell. Fagents (2003) found that a uniformly mixed fraction of a dense component required < 4 % by volume for the density to be sufficiently high for pure water to be buoyant on Europa. However, the availability of these volumes and the extent to which these dense materials could mix into the ice is unclear (Showman et al., 2004), as is their longevity once in the ice.

Finally, in considering the negative buoyancy of cryomagmas, we might take a lesson from Earth's Moon. The lunar basalts were substantially denser than the anorthosite flotation crust through which they ascended to erupt at the surface (Wieczorek et al., 2001). Head and Wilson (1992) recognized that even so, lunar basalt magma would rise a substantial fraction of the way to the surface in an open conduit to a level of hydrostatic equilibrium. If the top few kilometers of crust were missing, say as a result of a large impact, then the pressure gradients associated with the topographic variation would permit the magma to erupt within the resulting impact basin. Examining the stresses related to filling of a large impact basin, McGovern and Litherland (2011) showed that the combination of flexural and membrane stresses set up by the basin loading creates zones of enhanced ascent for magmas around the margins of the basins, where the stress gradient favors upward motion of negatively buoyant

magmas. Although most icy bodies generally do not retain substantial relief, this mechanism might be a factor, at least transiently, in some cases, and could be could be important for bodies such as Pluto where possible cryovolcanic features are observed adjacent to the impact basin Sputnik Planitia (McGovern et al., 2021; Section 5.5.3).

5.4.2 Formation of eruptive conduits

Many proposed eruption mechanisms require the opening of a fracture to connect the ocean or fluid reservoir with the surface. In practice, it might be difficult to achieve this connection, especially for bodies with thick ice shells, because with increasing depth in the ice shell, the overburden pressure acts to close any fractures. In general, fractures that might be exploited as eruptive conduits could be the result of either global stress patterns arising from a change in the shape or volume of the body, or more localized stress fields associated with, for example topographic loading (McGovern and Litherland, 2011) or the freezing of fluid reservoirs contained within the ice shell (Lesage et al., 2020). One might also consider the role of impacts in creating a fractured, weakened ice mega-regolith that could facilitate movement of fluids in the near-surface (Buczkowski et al., 2018; Raymond et al., 2019; Steinbrügge et al., 2020). Table 5.3 summarizes the sources of global stress fields that act on planetary bodies. Some of these sources of stress (e.g., differentiation, despinning) would mainly have acted early in the history of a body, and may leave no discernible trace today, while others (e.g., diurnal stress fields, non-synchronous rotation) are ongoing, and may be contributing to the surface geology presently observed.

In investigating the ways in which fractures might be produced, we consider the following scenarios (Figure 5.10): (i) fractures penetrating downwards from the surface; (ii) fluid-filled fractures propagating upward from the base of the ice shell; (iii) fractures initiating within the ice shell, and then propagating both upwards and downwards; (iv) fractures resulting from refreezing liquid reservoirs within the ice shell.

Fractures propagating downwards from the surface

There is ample evidence that many icy bodies are heavily tectonized, as evidenced by fractures and lineaments, which suggest that stress fields exist that are able to overcome the tensile or shear strength of ice to produce these features at the surface (Table 5.3). In some cases (e.g., Europa, Enceladus), these stress fields might be some combination of stresses due to diurnal tides, non-synchronous rotation, and perhaps true polar wander (Hurford et al., 2007a,b; Rhoden et al., 2010). In other cases

Table 5.3. Sources of global stress fields in planetary bodies.

Source of Stress	Description				
Diurnal tides	 Dynamic stress field migrates across the globe in patterns of tension and compression on the timeframe of the orbital period: Eccentricity-driven: magnitude of tidal bulge varies with orbital position (Fig. 5.4); Obliquity-driven: tidal bulge migrates in latitude. 				
Non-synchronous rotation	Ice shell decoupled from core rotates faster than interior, inducing stress field in shell as it reorients with respect to tidal bulge.				
Polar wander	Mass heterogeneity in ice shell causes it to reorient with respect to interior; outer shell migrating over tidally deformed interior induces changing stress field in shell.				
Orbital migration	Magnitude of tidal deformation changes as a satellite moves closer to (or farther from) their parent planet; this increases (or decreases) stresses as the satellite orbit migrates.				
Internal differentiation	Change in mass distribution within satellite leads to change in shape and consequent change in stress field; reduces tidal response and induces compression in sub-/anti-planet regions; tension in leading/trailing hemispheres. In most cases differentiation occurs early in a body's history.				
Rotational stress	Change in rotation rate changes shape of body and changes stress field as a result: increase in rotation rate (spin-up) increases oblateness, inducing tensional stress at low latitudes; spin-down increases sphericity, inducing tensional stresses at high latitudes.				
Volume changes	Expansion or contraction due to cooling or heating of body can induce large isotropic tensional or compressional stress fields. Whether a body experiences expansion or contraction on cooling depends in detail on the ice:rock ratio, internal structure, and size of the body. Smaller ice-rich bodies likely expand on cooling.				

(e.g., Tethys, Dione, Miranda, and Ariel), tectonism might have been dominated by extensional stresses due to a change in the volume of the body on freezing of a subsurface ocean. However, the ability of a fracture to penetrate to an ocean or cryomagma reservoir is limited by the lithostatic pressure and by the inability of the deeper, warmer portions of the ice shell to respond to imposed stresses in a brittle manner by fracturing. Thus, depending on the depth to the ocean or water body, there may be only very limited conditions under which a fracture might penetrate to liquid water.

The applied stresses must be sufficient to overcome the tensile strength of the ice. Although pure laboratory ice has a tensile strength of order 10^6-10^7 Pa (Schulson, 201, 2006), analyses of in situ sea ice suggest lower tensile strengths, ~ 10^4-10^5 Pa (Kehle, 1964; Dempsey et al., 1999). Fracturing and porosity in a satellite's ice shell may lead to significantly lower tensile strengths in the upper portion (Lee et al., 2005). Tensile stresses on Europa or Enceladus due to diurnal tides reach a maximum of 0.1 MPa (Hurford et al., 2007a,b), whereas stresses of 1–10 MPa are achievable through non-synchronous rotation or true polar wander on Europa (Leith and McKinnon, 1996; Schenk et al., 2008), or by diapirinduced reorientation of Enceladus' ice shell (Nimmo and Pappalardo, 2006). As noted above, ice shell thickening due to freezing can induce stresses of 1–3 MPa on Europa, and 1–10 MPa on Enceladus (Manga and Wang, 2007). It therefore seems quite plausible that the threshold stresses for fracturing the ice can readily be attained, at least on these two bodies. The question is, then, how far do the fractures penetrate?

Rudolph and Manga (2009) developed a numerical model to investigate the feasibility of fracturing an ice shell from the surface downwards, treating the lower portion of the shell as capable of undergoing viscoelastic relaxation of applied stresses. Adopting plausible mechanical properties for the ice, they explored a range of ice shell thicknesses and applied stresses, and concluded that it would be difficult to propagate a fracture in Europa's ice shell all the way to the ocean unless the ice shell was less than 2.5–5 km thick, which is at the low end of the range of plausible estimates (1–32 km; Billings and Kattenhorn, 2005). In contrast, the lower gravity on Enceladus allows much greater fracture penetration, readily reaching the ocean for ice shell thicknesses <25 km, which is consistent with observations of active plumes and inferences of ice shell thicknesses < 20 km (Čadek et al., 2016).

In general, the results of Rudolph and Manga (2009) for Europa concur with the earlier analysis of Crawford and Stevenson (1988), which concluded that propagation from the surface to the ocean is impossible except for very high extensional stresses (of order 10 MPa) applied over timescales that are short compared to the relaxation time of the ice ($< 10^4-10^5$ s); otherwise the deeper, warmer ice will respond to imposed stress by creep rather than brittle failure. Given that Europa and Enceladus have the thinnest ice shells of the bodies considered in this chapter, the problem is only exacerbated for bodies with thicker shells. However, even if penetration of surface fractures all the way to the ocean might not be plausible for a given body, there remains the possibility that they might intersect near-surface fluid reservoirs.

Fractures propagating upwards from the ice-ocean interface

Crawford and Stevenson (1988) also investigated the upward propagation of liquid-filled cracks from the base of Europa's ice shell (Figure 5.10). The presence of water in the fractures allows for greater propagation distances for a given applied stress because the pressure of the water on the fracture walls opposes the hydrostatic pressure in the ice that acts to close the fracture. However, the issue remains as to whether warm basal ice in contact with the ocean is capable of responding in a brittle manner to applied stresses. For Europa, where salts are likely to be the main non-H₂O component of the water, this might be a tough proposition. However, if ammonia is present in any quantity in the oceans of other bodies, this would cause a substantial lowering of the ocean temperature and a commensurately cooler and more brittle, although stronger ice shell. Crawford and Stevenson (1988) also suggested that periodic stresses (such as diurnal stresses) could cause failure due to stress corrosion cracking (analogous to 'metal fatigue') at much lower stresses.

Accepting that this is a viable mechanism, Crawford and Stevenson (1988) used linear elastic fracture mechanics (LEFM) to track the fate of upward-propagating cracks. As noted in section 5.4.1, volatiles exsolving in the low-pressure crack tip could permit the tip to pinch off and continue to rise under its buoyancy. In this way, the crack might make it to the surface to erupt a gas-rich spray of entrained water droplets, even if conditions were not possible for an open fracture to span the whole thickness of the ice shell.

Similarly, Mitri et al. (2008) proposed that, by analogy with terrestrial ice shelves, small defects in the base of the ice shell can propagate upward as larger fractures given an appropriate stress field. They used LEFM to investigate propagation from an ammonia–water ocean on Titan to investigate the conditions under which ascent to the surface might occur. Without invoking volatile exsolution, they found that progressive freezing of water in the pocket would increase the concentration of ammonia, thereby decreasing the negative buoyancy, and that large-scale pressure gradients associated with tides, non-synchronous rotation, solid state convection or long-wave topography might overcome the small remaining density contrast and drive upward motion and eruption at the surface.

Fractures initiating within the ice shell

As discussed in Section 5.4.1, progressive freezing of a subsurface ocean and thickening of an ice shell can pressurize the subsurface ocean and increase tangential stresses in the shell. For Europa, the tensile stress was predicted to peak within a few km of the surface (Nimmo, 2014), at which point a fracture could initiate and propagate both upwards toward the surface and downwards toward the ocean (Manga and Wang, 2007); tidal stresses might also help fracture propagation. If the fracture penetrated to the ocean, it would fill with water and assist with upward propagation, although it would be unlikely to reach all the way to the surface (Manga and Wang, 2007). In that case, transition from vertical to horizontal fracture propagation at a rheologic boundary, and the consequent development of sills or laccoliths, were considered possible outcomes (Manga and Wang, 2007; Craft et al., 2016; Manga and Michaut, 2017).

In the event that a liquid reservoir forms within the ice shell (Section 5.2.2), the volume change on refreezing of some proportion of the liquid could lead to pressurization within the reservoir that may be sufficient to cause failure of the reservoir walls and propagation of a fluid-filled fracture to the surface (Fagents, 2003; Section 5.4.1). If the surrounding ice behaves rigidly and does not deform to

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accommodate the volume change in the reservoir, it is possible to calculate the amount of reservoir liquid that must freeze for reservoir pressure to fracture the ice and drive liquid to the surface. For Europa, Lesage et al. (2020) calculated that, for water reservoirs located at depths of 2 - 10 km, approximately 4 - 13% of the reservoir volume would have to freeze to induce an eruption. Continued cooling and freezing of the reservoir could lead to multiple eruptions before the reservoir fully solidified (Lesage et al., 2021).

5.4.3 Presence of gas clathrate hydrates

The possible presence of gas clathrate hydrates in icy satellites provides an additional mechanism by which eruptive activity might take place. Gas clathrate hydrates are compounds in which gas molecules become incorporated in cages of frozen hydrogen-bonded water molecules. They form under conditions of low temperature and/or elevated pressure, provided an appropriate gas species is present. Across the icy bodies of the Solar System, CH₄ (and other hydrocarbons), CO₂, SO₄, H₂S, N₂, CO, and O₂ have been considered as possible clathrate-forming species (Prieto-Ballesteros et al., 2005; Mousis et al., 2015; Safi et al., 2017). For example, at Europa, clathrates of CO₂, SO₂, H₂S, and CH₄ are stable in most of the crust and interior, although their presence depends on the provision of free gas from the Europan interior (Prieto-Ballesteros et al., 2005), and the formation of CO₂ clathrate would be inhibited by the presence of sulfates (e.g., $MgSO_4$) in the ocean (Safi et al., 2017). Similarly, CH_4 , N_2 and CO_2 clathrates have been considered for Enceladus (Kieffer et al., 2006; Safi et al., 2017; Mousis et al., 2015). At Titan, methane, ethane, and other hydrocarbons are likely gas species. Methane clathrates are stable from the surface to the deep interior, and could have formed a substantial crustal layer early in Titan history (Tobie et al., 2006). Methane and ethane clathrates might even continue to form today owing to the Titan's meteorological methane cycle (Mousis et al., 2015) and the photochemical production of ethane from atmospheric methane (Mousis et al., 2016).

Clathrates readily decompose under low pressures and upon warming. Stevenson (1982) noted that the presence of clathrates in the subsurface of icy bodies might lead to explosive cryoclastic eruptions if they become destabilized by contact with warm materials (e.g., cryomagmas) rising from below. The production of a pressurized gas might then explode through the overlying ice to produce a gas-rich explosion of ice particles. For Titan, the decomposition of clathrates might help to explain the replenishment of atmospheric methane, which would otherwise be destroyed by photochemical processing over ~30 Ma timescales (Wilson and Atreya, 2004; Horst, 2017). A cubic meter of methane clathrate can liberate up to ~120 kg of gaseous methane which, depending on ambient temperature and

pressure conditions, could provide a substantial driver for eruption of cryomagmas. Notably, the presence of ammonia (NH₃) in a cryomagma substantially reduces the stability of methane clathrate, so an ammonia–water cryomagma in contact with a clathrate-rich crust could generate substantial gas to drive explosive or effusive cryovolcanic eruptions (Choukroun et al., 2010).

Another consequence of the presence of clathrates in ice shells is that their low thermal conductivities means that they are effective insulators, which would lead to warmer temperatures within the ice shell and would help to promote solid state convection (Kalousová and Sotin, 2020) or other internal activity. Finally, the densities of some clathrates are greater than that of water ice (Prieto-Ballesteros et al., 2005). If clathrates are present in substantial quantities, therefore, this might provide a driving mechanism for aqueous cryomagmas or ice diapirs to ascend buoyantly towards the surface.

5.4.4 Solid state convection and diapirism

An alternate mechanism of bringing materials from depth to the near-surface involves movement of warm ice still in the solid state, perhaps lubricated by minor amounts of melt. In this case, convection within the ductile portion of the ice shell could be initiated by thermal means, such that warm ice rises by virtue of its buoyancy with respect to the surrounding cooler ice. As discussed in Section 5.2.2, the plausibility of thermal convection in an ice shell depends on a range of parameters that are typically not well known: the ice shell temperature profile (which in turn depends on interior heat flow and the ocean temperature/composition), the thickness of the ice shell, and the rheology of the ice (which depends on grain size, strain rate, and temperature). Hence different models do not always agree that convection is possible on a given body (e.g., Tobie et al., 2005; Mitri and Showman, 2008; Nimmo and Bills, 2010; Hemingway et al., 2013). Alternatively, compositional heterogeneity in the ice shell might induce upwelling of ice as a result of the density contrast between ice containing varying amounts of impurities (e.g., Pappalardo and Barr, 2004; Han and Showman, 2005). Upwelling warm ice might also be manifested in the near-surface layers as diapirs; i.e., discrete bodies of warm ice intruding the cooler, brittle lithosphere (Quick and Marsh, 2016). Warm ice impinging on impure ice having a lower melting point (e.g., salt-rich ice), might then lead to melting and effusion of surface flows. For example, Head and Pappalardo (1999) suggested that thermal perturbation by diapirs of salty ice could induce mobilization and release of brine at the surface. The diapirs themselves might lead to a range of surface expressions (Rathbun et al., 1998; Pappalardo and Barr, 2004).

5.5 Manifestation of Cryovolcanism in the Outer Solar System

5.5.1 Explosive Cryovolcanism: Active Plumes and Cryoclastic Deposits

The identification of plumes of material erupting from the interiors of icy bodies can be interpreted in the context of explosive silicate volcanism, with which we are familiar on Earth, Jupiter's moon Io, and which we understand to have occurred on Mars, the Moon, Mercury, and (maybe) Venus (see chapter 4). Whereas there are some similarities to silicate volcanism, i.e., particles are ejected into the environment by expanding gases, the details of the subsurface mechanisms may differ markedly on an icy body.

Enceladus

The most dramatic, unequivocal example of eruptive behavior on an icy body is in the form of a plume of vapor and particles emanating from the prominent "Tiger Stripe" fracture system (sulci) in the south polar region of Enceladus (Figure 5.11). Although Voyager 2 revealed the diverse surface terrains in Enceladus in 1981, it did not image the south polar region, so it was not until the Cassini spacecraft toured the Saturn system between 2004 and 2017 that the spectacular eruptions were observed. In early 2005, the magnetometer instrument provided the first hints that something anomalous was occurring at Enceladus: unusually intense oscillations in the Saturn's magnetic field were observed in the vicinity of Enceladus. In the months that followed, the magnetometer measured deflections in the magnetic field at Enceladus, which suggested that something electrically conductive was perturbing the field lines (Dougherty et al., 2006). Subsequently, observations of stellar occultations at Enceladus by the Ultraviolet Imaging Spectrometer (UVIS) revealed the presence of an asymmetric atmosphere composed of water vapor centered on the south polar region (Hansen et al., 2006; Waite et al., 2006). The Ion Neutral Mass Spectrometer (INMS) confirmed water vapor as the dominant constituent of the plume (Waite et al., 2006), and the Cosmic Dust Analyzer (CDA) detected a concentration of micron-sized particles at the south pole (Spahn et al., 2006). Furthermore, on close approaches to Enceladus, the Composite Infrared Spectrometer (CIRS) resolved thermal anomalies (exceeding 140 K) associated with the sulci fractures (Spencer et al., 2006). In sum, these observations all suggested that the sulci were themselves the source of a plume of vapor and ice particles erupting into space. Visual confirmation of the active plume came with Imaging Science Subsystem ISS images (Porco et al., 2006), which also revealed that some fraction of the plume materials escapes Enceladus' low gravity (0.113 m s^{-2}) to feed Saturn's E-ring. A suite of follow-on observations has shown that the plume consists dominantly of water vapor, with traces of CO₂, N₂, ammonia, methane and other hydrocarbons (Waite et al., 2009,

2017; Perry et al., 2015), as well as ice particles, salts (Postberg et al., 2009), SiO₂ nanoparticles (Hsu et al., 2015), molecular hydrogen (H₂; Waite et al., 2017), and complex organic molecules (Postberg et al., 2018). The silica nanoparticles (Hsu et al., 2015) are particularly important because they indicate a process of active hydrothermal venting from the core–ocean interface, and the methane is evidence of serpentinization that could produce the H₂. These observations have important implications for habitability for Enceladus' interior.

We now know that the plume is fed by ~ 100 discrete jets of ice and salt-rich particles and vapor emanating from vents up to 9 m wide arranged along fissures up to 130 km long (Goguen et al., 2013; Porco et al., 2014). The jets of material then coalesce to form the bulk plume. The jets are rich in ice particles, with large (salt-rich) particles falling out to produce deposits with distinct coloration (Schenk et al., 2011), and smaller (ice) particles entrained by the vapor to produce the bulk plume and escape Enceladus' gravity to feed the E-ring (Kempf et al., 2010). Vapor velocities range up to 1 km s⁻¹, with mass fluxes of 300 kg s⁻¹, and remain roughly constant (Hansen et al., 2006, 2017). However, the plume brightness varies diurnally, and is thought to reflect a varying column density of particles (with typical velocities and mass fluxes of < 230 m s⁻¹ and 50 kg s⁻¹ respectively; Ingersoll and Ewald, 2011). This diurnal variation is a response to the varying tidal stress with position along Enceladus' eccentric orbit around Saturn, such that plumes appear brightest near apoapse, when tensional stresses holding the fissures open are at their greatest (Hurford et al., 2007; Hedman et al., 2013).

At first, a variety of models were postulated for the source of the jets, including vapor derived from melting and vaporizing or subliming ice in fractures undergoing shear heating (Nimmo et al., 2007), evaporation of the surface of a water column in fractures (Kite and Rubin, 2016), and volatiles exsolving from liquid-filled fractures (Matson et al., 2012). However, the INMS measurements of salts in the plume (Postberg et al., 2011), was seen as good evidence for the existence of a large subsurface body of water in contact with (and reacting with) the silicate core, from which the salts were derived. Furthermore, the presence of molecular hydrogen, identified by INMS in 2015, has been interpreted to result from ongoing hydrothermal reactions at the ocean–core interface (Figure 5.12; Waite et al., 2017). Enceladus' gravity signature, shape, and libration are consistent with a global internal body of liquid (rather than an isolated subsurface reservoir), lying beneath an ice crust approximately 20 km thick, but expected to be thinner at the south pole where tidal heating is greater (less et al., 2014; Čadek et al., 2016). Enceladus' small size and very low gravity allows for tensional stresses to open fractures to greater depths than would be possible on a larger body such as Europa, which helps explain the continuous jetting of material from the subsurface ocean.

Europa

Europa's extensively fractured, youthful surface was first revealed by the visits to the Jupiter System by the two Voyager spacecraft in 1979, and was documented extensively by the Galileo mission between 1995 and 2003. The paucity of impact craters implies a surface age of 20–180 Ma (Schenk et al., 2004)—a geological blink of an eye—which in turn suggests that the body has been actively resurfacing throughout its history. Analyses of tidal stresses over Europa's surface indicate that any given location will experience cycles of compressional, shear and tensional stresses, as the satellite travels around Jupiter in an eccentric orbit (Greenberg et al., 1998). It might be expected, therefore, that Europan fractures undergoing tension may open to expose water in a shallow ocean, and provide a pathway to the surface for vapor, sprays or water from the subsurface. Indeed, explosive cryovolcanism is one of a number of theories proposed for forming ridge systems flanked by dark margins (e.g., Rhadamanthys Linea and other "triple bands"; Fagents et al., 2000; Quick and Hedman, 2020).

The young surface age and extensive fracturing prompted a number of searches for active plumes over the years, with mixed results. An early analysis of a single, low-resolution Voyager 2 image indicated the possible presence of an active plume above Europa's bright limb (Cook et al., 1983), but later reexamination of the data concluded that the feature identified was an artifact of the vidicon image (Pappalardo et al., 1999; Phillips at al., 2000). Later, a careful search for surface changes over the 25-year timeline between Voyager and Galileo missions led to no positive identifications of changes in surface features that might be attributable to explosive venting or some other cryovolcanic process (Phillips et al., 2000). De La Fuente Marcos and Nissar (2002) reported Earth-based telescope observations of a transient brightening of Europa in 1999, and interpreted it as possible evidence of eruptive activity.

Similarly, a plume search campaign during the Galileo mission, adopting favorable illumination and phase angle conditions, also failed to identify active plumes. On orbit E17 of the mission, a series of images at 72 m/pxl resolution were targeted along the limb of Europa from 2°S to 15°N at 215° longitude; 15 images were acquired along the surface parallel to the limb and 15 images in the dark sky above the limb to search for diffuse glows above the surface or anomalously bright regions on the surface. None were identified, although it was noted that the search targeted areas of the Europan surface that were expected to be under compression at the time of the observations (Phillips et al.,

2000). During orbits G7 and C20, images of Europa in eclipse were acquired to search for glowing "hotspots" of warm plume material above the disk of Europa. While this technique worked well for identifying hot (>1000 K) silicate volcanism on Io, the lack of detections at Europa again failed to identify any plumes.

It was not until recently that more convincing evidence of active plumes at Europa was found using data acquired by the Hubble Space Telescope. Using the Space Telescope Imaging Spectrometer (STIS) Roth et al. (2014) observed anomalous H and O emissions over a period of 7 hours in December 2012, producing a feature that extended above the surface of the south pole regions (Figure 5.13a). They interpreted these observations as a plume of water vapor 200 km high, erupting at 500–700 m s⁻¹. However, there were no other detections of plumes in similar images from 1999, or in 19 other observations between 2012 and 2015 (Rhoden et al., 2015). Using an independent technique, Sparks et al. (2016) examined images of Europa in transit across the face of Jupiter, processing the data to remove the contribution to the scene of Jupiter's disk, and revealing a feature interpreted to be a plume > 100 km high above Europa's limb in four of 12 transit images (Figure 5.13b). Repeat observations revealed a 50 km tall plume in one of the same locations as previous observations (Sparks et al., 2017). A recent reanalysis of Galileo magnetometer and Plasma Wave spectrometer data acquired during its closest encounter with Europa revealed an anomaly in magnetic field and plasma wave density that is consistent with the presence of plume erupting from a location close to the equator (Jia et al., 2018). However, analysis of Galileo Photopolarimeter Radiometer measurements of surface thermal properties showed no evidence of hotspot sources of potential plumes during the Galileo flybys (Rathbun and Spencer, 2020).

We can conclude from these investigations that Europa's plumes appear to be infrequent, transient features, quite different from Enceladus' continuous output. Indeed, Paganini et al. (2019) have suggested that the large-scale plumes initially imaged by HST are outliers and that Europa's plume activity may be more small-scale and sporadic in nature, and hence difficult to identify using Earth-based techniques. Calculations of column density and total mass in the observed Europan plumes are greater than for Enceladus plume (~10¹⁷ cm⁻² vs. 10¹⁶ cm⁻², and ~10⁶ kg vs. 10⁴ kg; Hansen et al., 2006, 2020; Roth et al., 2014; Sparks et al., 2016), and higher eruption velocities are required for a plume to reach a given height on Europa because of its higher gravity (1.31 m s⁻² vs. 0.113 m s⁻²). One might speculate that these differences are caused by differences in manifestation of tidal energy and response to imposed

stresses, differences in subsurface plumbing, and possibly differences in ocean composition, volatile content, and temperature.

Although it is not possible to construct detailed models of Europa plume eruptions without further observations to confirm their existence and collection of more detailed data, we can speculate that exposure of water at the surface of Europa would produce a spray of droplets and vapor as the water boiled into the near-vacuum conditions, perhaps aided by exsolution of dissolved gases. This could be the case both for effusion of fluid on to the surface, and for exposure of a water column in a fracture. Although Roth et al. (2014) and Sparks et al. (2016) identified the approximate area on Europa's surface that served as the source for the plumes, it was not possible to identify specific vents or cryovolcanic features. The youth and complexity of Europa's terrain leaves open the possibility that any number of surface features might provide a cryovolcanic source. For example, **Figure 5.13c–e** shows prominent lineaments with dark, diffuse margins, which are widespread on Europa. Deposition of cryoclastic material erupted from a medial fracture is one origin suggested for these deposits, although thermally produced lag deposits of dark, non-ice materials is an equally plausible origin (Fagents, 2003). Liquids exposed at chaos and lenticulae might also be sources for plumes.

Triton

Farther out in the Solar System, the Neptune system has not been visited by a spacecraft since Voyager 2 flew by in the summer of 1989. It made a series of remarkable observations of Neptune's large moon Triton (2707 km diameter), which revealed a very young surface age (<55 Ma; Schenk and Zahnle, 2007) with diverse terrain types, and a surface composition of 55% N₂, 15–35% H₂O and 10–20% CO₂ ices (Thompson and Sagan, 1990; Cruikshank et al., 1993; Stern 1994). It has a nitrogen atmosphere with a pressure of 1.4–1.9 Pa (McKinnon and Kirk, 2014). As noted in Section 5.2.1, Triton's orbital evolution could provide a source of internal heat to maintain a subsurface ocean and drive endogenic activity (Nimmo and Spencer, 2015).

During the Voyager 2 flyby, four active nitrogen-driven plumes were observed in Triton's southern hemisphere, rising to heights of 8 km, with mass fluxes of solids and gas of 10 kg s⁻¹ and < 400 kg s⁻¹, respectively (Soderblom et al., 1990). The plumes were associated with dark, fan-shaped deposits on the surface (Figure 5.14a), and 120 other such deposits (4–117 km long) were observed on the surface, suggesting that plumes were a common occurrence.

Two models were proposed for these plumes (Figure 5.14b). The solar-driven model (Smith et al., 1990; Soderblom et al., 1990; Kirk et al., 1990; Brown et al., 1990), proposes solid state greenhouse, in

which a few-meter thick seasonal translucent layer of N_2 ice overlying a permanent N_2 substrate promotes sublimation of the subsurface nitrogen. The vapor pressure builds until the surface ice ruptures and allows the gas to vent to the surface, entraining darker particulates. A mere 4 K temperature difference is sufficient to drive the observed plumes with an eruption velocity of 180 m s⁻¹. However, a substantial subsurface storage zone is needed to accommodate the accumulating N_2 gas.

The second model (Kirk et al., 1995) proposes a cryovolcanic origin for the plume, in which water plus ammonia hydrate or decomposing clathrate hydrates erupt explosively through vents 15–20 m wide at velocities of 600–950 m s⁻¹. This model bears reexamination because it is now recognized that obliquity tides can provide heat and possibly drive endogenic activity on Triton (Nimmo and Spencer, 2015).

Although the solar model was consistent with the seasonal concentration of solar insolation being in the southern hemisphere at the time of the Voyager 2 flyby, the north was not imaged well so we are unable to conclude whether it is a solar process alone. If Triton were to be visited by another spacecraft in the next 20 years (during the northern summer) the persistence of plume activity in the south would imply an endogenic rather than solar-driven origin.

Elsewhere on Triton, smooth high plains representing overlapping flat to undulating quasi-circular patches (60–120 km diameter) with central pits have been interpreted as cryoclastic deposits driven by volatiles (N₂, CO₂, H₂, CH₂, H₂S) exsolving from an aqueous (ammonia–water) cryomagma (Helfenstein et al., 1992; Croft et al., 1995). Nitrogen- or methane-driven cryoclastic eruptions could result in eruption speeds of 300–500 m s⁻¹ and produce deposits up to 300 km in diameter, which would account for the smooth highlands deposits and quasicircular dark patches (Croft et al., 1995). Other features on Triton interpreted to result from explosive cryovolcanism include cones 7–15 km in diameter and 100–200 m high, with summit pits and flanks at the angle of repose, indicative of particulate grains consistent with cryoclastic ejecta (Croft et al., 1995).

Ceres

Anomalous detections of water vapor outgassing at Ceres by ESA's Hershel Space Observatory were attributed to active explosive cryovolcanism on the dwarf planet (Küppers et al., 2014). This outgassing was observed to be occurring at a rate of 6 kg s⁻¹ (Küppers et al., 2014), and if an endogenic source was confirmed, would have represented the first instance of active cryovolcanic plumes in the inner Solar System. However, observations by the Dawn spacecraft failed to confirm the presence of active plumes on Ceres. It is likely that, rather than being a result of explosive cryovolcanism, these anomalous outbursts were due to other mechanisms such as the local sublimation of surface ice (Landis et al.,

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2017), sputtering of surface ice triggered by solar energetic protons (Villarreal et al., 2017), or the seasonal migration of a polar cap (A'Hearn and Feldman, 1992).

Although no active plumes have been observed on Ceres, it is has been postulated that explosive venting was responsible for the formation of the faculae, or "bright spots" in Ceres' Occator crater (Figure 5.15) (De Sanctis et al., 2016; Krohn et al., 2016; Nathues et al., 2017; Zolotov, 2017; Quick et al., 2019; Ruesch et al., 2019a). Zolotov (2017) proposed that the faculae were emplaced on Occator's floor as a result of post-impact boiling. However, Nathues et al. (2015, 2017) suggested that Occator is 30 Ma older than the brightest faculae within it. The activity that emplaced the faculae is thought to have ceased relatively recently, within the past 2 Ma. Owing to the unknown mechanical properties of the faculae, and because crater-based chronology on young surfaces may be imprecise, these ages have been disputed (Neesemann et al., 2019). Notwithstanding, based on crater counting and morphological grounds (Nathues et al., 2017; Stein et al., 2019), the proposed age differences between the faculae and Occator crater argue for independent formation mechanisms for these features.

The faculae in Occator crater are mostly composed of sodium carbonates, which may represent deposits left behind by brines that erupted from Ceres' interior onto Occator's floor (DeSanctis et al., 2016). It has been proposed that the faculae are the result of eruptions of brines at the surface (Quick et al., 2019; Ruesch et al. 2019a). Upon exposure to Ceres' zero-pressure surface environment, exsolving volatiles and eruptive boiling would produce a spray of droplets and salt particles launched on ballistic trajectories into an eruption plume. Quick et al. (2019) suggested that the deposition of salts from the eruption plume could have formed two prominent faculae, Cerealia and Vinalia (Figure 5.15), and found that less than 2 wt.% of a driving volatile such as such as water vapor, CO₂, CH₄ or NH₃ could have created deposits of the sizes observed. Any ice within the deposit would readily sublime away, leaving a bright, salt-rich residual deposit (Ruesch et al., 2019a). The proximity of the faculae to fractures on Occator's floor suggests that these fractures are linked to subsurface conduits that could have facilitated both effusive and explosive eruptions (De Sanctis et al., 2016; Buczkowski et al. 2018).

5.5.2 Effusive Cryovolcanism

It is important to note that, unlike plume activity, there have been no observations of active effusive eruptions on icy bodies. Instead, we see landforms that are suggestive of having been formed by flowing fluids at the surface. In some cases, the landforms have very little relief, as might be expected of a low-viscosity fluid (i.e., a water cryomagma) flooding the surface. In other cases, the landforms have considerable relief, implying substantial viscosity and/or strength, which might be explained if the

effusions are predominantly in the solid state (as ice-rich slurries, or warm ductile ice, with perhaps a small fraction of lubricating fluid), or if the cryomagmas have a more exotic composition (e.g., ammonia–water slurries; Kargel et al., 1991; Kargel, 1995; Section 5.3; Figure 5.9b).

In studying these landforms, we endeavor to deduce their mechanisms of formation from their morphological, topographic, stratigraphic, and compositional attributes, and to develop conceptual and mathematical models of their emplacement, in an attempt to quantify and constrain their volumes, eruption rates, material properties, among other characteristics. However, in some cases, a cryovolcanic origin is not the only reasonable explanation for a given feature type, and it is important consider other (endogenic or exogenic) processes of formation. It is frequently the case that it is not possible to rule out a given mechanism of formation without additional remote sensing data. We must also remain aware that a feature that is suggestive of a particular origin in images of a given resolution, may look quite different (and suggest other origins) at another resolution (e.g., Fagents, 2003; Moore and Pappalardo, 2011).

Upon exposure at the surface of an icy satellite, water will experience simultaneous boiling and freezing in the cold, vacuum environment, generating a spray of vapor, droplets, ice crystals (Allison and Clifford, 1987) and, potentially, particles of non-H₂O compositions. As the liquid continues to cool, accumulating ice will eventually form a crust that suppresses further vaporization and insulates the liquid, but would be subject to repeated disruption due to the turbulent motions of the flowing fluid. The flow front might be a chaotic mix of ice blocks, vaporizing water, and water breakouts. However, the formation of ice crystals and crust will modify the bulk rheology of the flow, leading to a higher viscosity. However, without composition-modifying contaminants, the resulting flow deposits are likely to lack substantial relief, and would readily pond in topographic lows.

Smooth plains

Relatively smooth, bright, sparsely-cratered surfaces exist on several icy bodies, and have been interpreted as the product of major effusions of water cryomagmas at the surface. The Voyager imaging data of Europa, for example, showed a young surface crioss-crossed by lineaments, mottled by dark regions, with intervening bright plains that were smooth at Voyager resolutions (no better than 4 km; Lucchitta and Soderblom, 1982; **Figure 5.16a**). With the arrival of Galileo data in the 1990s, however, it became apparent that these bright plains are in fact heavily tectonized, with ridges and fractures observed at a range of scales down to the limits of resolution (Figure 5.16b). Similarly, Enceladus has surfaces that appeared smooth at Voyager resolutions (Smith et al., 1982) but were revealed by the

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Cassini mission to comprise low-relief ridges and fractures suggestive of compressional, extensional, and shear deformation (Figure 5.16c). For both Europa and Enceladus, it is conceivable that these surfaces were originally emplaced as cryovolcanic effusions, but have since been overprinted by tectonic processes so as to remove evidence of their emplacement.

Saturn's moons Tethys (1062 km diameter) and Dione (1128 km diameter), which occupy orbits beyond that of Enceladus, both possess cratered surfaces that have a variety of ages, based on crater counts. Both have relatively youthful plains that lack large impact craters, and in which pre-existing relief has become subdued (Figures 5.16d,e). Whereas Tethys' lightly cratered plains are centered on the trailing hemisphere (the side that faces away from orbital motion), Dione's are centered on the leading hemisphere, contrary to what is expected for a tidally locked satellite. It is possible that these terrains resulted from a period of endogenic activity (e.g., surface flooding) perhaps related to a past 2:3 resonance between the satellites (Matson et al., 2009). However, both satellites lack clear morphological indicators of cryovolcanism.

Neptune's moon Triton exhibits sparsely-cratered, low relief terrains that appear to have been resurfaced (Smith et al., 1989; Strom et al., 1990; Stern and McKinnon, 2000). In addition to the southern hemisphere plume deposits (Section 5.5.1), Triton has diverse surface types strongly indicative of endogenic processes (Croft et al., 1995; **Figure 5.17a**). Roughly 40% of Triton's surface was imaged during the Voyager 2 flyby in 1989, so our understanding of its geology is limited. Crater counts on the surfaces observed reveal ages as young as 6–55 Ma (Schenk and Zahnle, 2007), meaning that extensive resurfacing has continued to the recent past, perhaps as a result of a slow decay of its past high eccentricity in combination with radiogenic heating (Section 5.2.1). The observed surface is composed of ~55% nitrogen ice, 15–35% water ice, and 10–20% CO₂ ice, with minor amounts of methane and SO₂ (Thompson and Sagan, 1990; Cruikshank et al., 1991, 1993; Stern and Trafton, 1994). In terms of plains units, smooth materials occupy four quasi-circular walled depressions 80–390 km in diameter, as well as terraced areas adjacent to some of these depressions, and smooth high plains farther to the east (Fig. 5.17a). The fact that smooth plains are bounded by the walls of the depression, and the smooth plains embay all irregularities of the walls, suggests that this unit was emplaced within the depression as a fluid cryolava, possibly an ammonia–water mixture (Croft et al., 1995).

Pluto has extensive bright, smooth plains in the Sputnik Planitia region of Tombaugh Regio (Figure 5.17b). Sputnik Planitia may have originated as an impact basin into an ice shell overlying an ocean, and subsequently infilled with the plains material (Nimmo et al., 2016). The plains are devoid of impact

craters and are therefore extremely young (possibly < 10 Ma; Trilling, 2016). However, the plains do *not* represent cryovolcanic resurfacing. They consist predominantly of nitrogen ice (with minor methane and carbon monoxide; Stern et al., 2015), which accumulates as frost on the bordering mountains and then appears to flow, glacier-like, back into the planitia (Figure 5.17c). The surface of the plains exhibits polygonal cells with marginal troughs, which are thought to represent convection of the nitrogen ice (McKinnon et al., 2016; Trowbridge et al., 2016). Although we cannot point to Pluto's plains as representing cryovolcanism, there are clearly dynamic processes at play in shaping their surface. Furthermore, the loading of the basin by the nitrogen ice would have set up an extensional stress regime surrounding the basin, which could have facilitated cryomagma ascent to the surface (McGovern et al., 2021). This might explain the presence of Wright and Piccard Montes to the south of Sputnik Planitia (see section 5.5.3), and possible cryovolcanic deposits in the Virgil Fossae area to the west.

Pluto's satellite Charon (1212 km diameter) hosts Vulcan Planitia, rolling plains covering roughly half of the mappable surface (Figure 5.16f). Although this unit consists of a variety of textures, it is believed to have been emplaced by the flow of icy lavas (Stern et al., 2015; Moore et al., 2016; Beyer et al., 2019). Convex margins near the southernmost part of Vulcan Planitia may represent converging flow fronts, and a bounding moat separating Vulcan Planitia from Oz Terra to the north implies substantial flow viscosities (Schenk et al., 2018; Beyer et al., 2019). Based on the likely incorporation of NH₃ into Charon (Grundy et al., 2016) and its probable sequestration into an ancient ocean, the cryolavas that formed Vulcan Planitia were likely ammonia-rich (Borrelli and Collins, 2018; Beyer et al., 2017; 2019). Initial modeling suggested that these cryolavas may have been extremely viscous with rheologies similar to glacial ice (Beyer et al., 2019). However, cryolavas with much lower viscosities and rheologies similar to terrestrial basalt are possible (Quick, 2019). The favored emplacement style for Vulcan Planitia flows is similar to that of the emplacement of the lunar maria in which there were many localized sources for multiple individual flows (Beyer et al., 2019). These flows may have effused onto Charon's surface as a result of the upwelling of an NH₃-H₂O ocean during episodes of lithospheric expansion and moderate disruption. Conversely, the pressurization of a subsurface ocean as it gradually froze may have driven NH₃-rich fluids to the surface (Manga and Wang, 2007). However, in the case that lithospheric expansion led to severe disruption, NH₃-rich material may have been delivered to Charon's surface after lithospheric blocks foundered into the viscous mantle below. The Vulcan Planitia flows appear to have embayed large massifs (Butler and Kubrick Montes) and a cluster of mountains (Clarke Montes) (Stern et al., 2015; Moore et al., 2016). Although these mountains were originally proposed to be cryovolcanoes (Desch and Neveu, 2017), New Horizons imagery suggests that their morphologies are

not consistent with a cryovolcanic origin (Robbins et al., 2019; Beyer et al., 2019).

Lobate or flow-like features

Prior to the Galileo mission arriving at Europa, there had been much speculation based on Voyager data that Europa would reveal a wealth of effusive cryovolcanic features. For example, Lucchitta and Soderblom (1982) suggested that the dark, mottled regions of Europa's surface (that we now term chaos terrain) might be the result of eruptions of muddy slurries. Similarly, Wilson et al. (1997) modeled the emplacement of Thrace and Thera Maculae (Figures 5.16a, 5.18a), two dark, apparently lobate features in Europa's southern hemisphere, as effusive eruptions of dirty water-ice slurries. The acquisition of image data by the Galileo spacecraft did much to disabuse us of the notion that Europa was covered with cryovolcanic deposits. High-resolution data for Thrace Macula revealed a complex structure in which ridges in the background terrain can be traced into the interior of the macula where they become progressively degraded. The interior of Thrace appears to lie at or slightly above the elevation of the surrounding terrain (Prockter and Schenk, 2016). Only at the margins is there good evidence of lowviscosity fluids leaking out of the raised interior of the feature and being diverted and confined by the pre-existing topography of the background ridged plains. Thrace Macula has been interpreted as disruption and darkening of the existing surface by a thermal source (diapir or freezing liquid body) impinging from below that may have mobilized near-surface brines to produce the dark, low-lying deposits (Fagents, 2003). Thera Macula, on the other hand, was shown to have a very different morphology and topography, and is in fact topographically low with respect to the surrounding terrain. This has been explained as the result of subsidence and disaggregation of existing terrain due to the presence of a liquid reservoir in the subsurface (Schmidt et al., 2011; Figure 5.6b).

Other features that appeared in low resolution Galileo images to have lobate morphologies and be associated with ridges or fractures that could have provided a pathway to the surface for an ascending cryomagma, were shown in higher resolution data to have characteristics that do not support origins as a cryolava flow. For example, Figures 5.18b and 5.18c shows apparently lobate features at resolutions of 550 m/pixel and 30 m/pixel, respectively. Although such features can look flow-like at lower resolutions, higher resolutions often reveal a different story – in this case, complex, disrupted preexisting ridged terrain is seen to lie within the margins of the feature. Instead, these features might be the surface manifestation of upwelling warm ice or diapirs (e.g., Pappalardo et al., 1988).

Intriguing candidate effusions were also identified by Kattenhorn and Prockter (2014), associated with possible subduction zones on Europa. There is ample geological evidence for crustal extension and

creation of new surface area at dilational bands, such that an enduring mystery has been where this extension is compensated by loss of surface area. Detailed mapping of the northern trailing hemisphere allowed Kattenhorn and Prockter (2014) to identify areas of substantial crustal loss (20,000 km²). They proposed a model whereby cold brittle plates become subducted into the warm convecting interior, and produce cryolavas as a result of heating of salt-rich ice being entrained from the surface, perhaps assisted by frictional heating between the downgoing plate and the overlying ice (Figure 5.18d). These melts may then be forced to the surface by the pressure gradients associated with the subducting plate and surroundings; a number of lobate features (possible cryolava flows) were identified in the 228 m/pixel data (Figure 5.18e). Although lessons learned previously at Europa suggest that caution should be exercised in the search for cryolava flows, these are indeed intriguing features that will have to await data from future missions to decipher.

Several fields of lobate features have been identified on Titan (Lopes et al., 2007; 2013) as possibly due to effusive cryovolcanism, although the low spatial resolutions of Cassini's Synthetic Aperture Radar (SAR), varying from ~350 m to 1.5 km, and limited data from the Visible and Infrared Spectrometer (VIMS) data, make it hard to distinguish cryovolcanic from fluvial origins (Moore and Pappalardo, 2011). A review of the flow-like features on Titan that may be cryovolcanic is given in Lopes et al. (2013) who conclude that Mohini Fluctus, in the Sotra Patera complex (Figure 5.18f; Section 5.6.3) is the most plausible cryovolcanic flow candidate on Titan. They also concluded that flow-like features in the Hotei Regio area are viable cryovolcanic candidates, though a fluvial origin has also been proposed (Moore and Pappalardo, 2011).

Several potential cryovolcanic flows have been identified in New Horizons images of Pluto. One example extends from a depression on the southwest flank of Pluto's Wright Mons (Singer et al., 2016; see Section 5.5.3), and recent work suggests that cryovolcanic activity has occurred in the geologically recent past in the Virgil Fossae complex (Cruikshank et al., 2019), including a possible cryovolcanic flow and cryoclastic deposits (Figure 5.18g). This flow may represent effusion from multiple sources of an NH₃-water fluid along a fault that defines the south wall of the Virgil Fossae trough. Ponding in impact craters and potential breakouts from this flow are also observed. These breakouts are perpendicular to the flow and extend up to 3 km from the main trough. This enigmatic flow-like feature may have also been accompanied by explosive eruptions that emplaced cryoclastic deposits 100 km from the main trough (Dalle Ore et al., 2019). The association of ammonia compounds with the Virgil Fossae features suggests that they are relatively recent (<10⁹ years) and that Pluto is an active cryovolcanic body.

Dome features

During the Galileo mission, numerous examples of dome-like features were identified on Europa, and various endogenic origins were proposed (Pappalardo et al., 1999). While some positive-relief features appear to be simple uplifts of the surface with or without disrupting the existing surface, others have the appearance of being superposed on existing terrain and/or lying within annular moats (Figure 5.19a-d); Fagents, 2003; Pappalardo and Barr, 2004; Quick et al., 2017). Typical sizes of such features range from a few km to >10 km in diameter (Fagents et al., 2000; Fagents, 2003; Quick et al., 2017), although resolution limitations preclude a robust assessment of the full size range (Greenberg et al., 1999; Riley et al., 2000). The striking similarity of positive-relief, dome-like features to terrestrial lava domes led to the consideration that these were emplaced as cryovolcanic effusions of viscous fluids (Figure 5.10; Fagents, 2003; Quick et al., 2017). Bulk viscosities calculated through modeling of the domes as radiallyspreading viscous gravity currents lie in the range 10⁶ to 10⁹ Pa s. However, these bulk viscosity values include the brittle insulating crust (Allison and Clifford, 1987) that prevents fluid from vaporizing in Europa's zero-pressure surface environment. The viscosities of dome-forming cryomagmas at the time of eruption may be 4 – 5 orders of magnitude lower than these values (Schenk et al., 1991; Lorenz, 1996). Hence Europa's extrusive domes may form from fluids that are $10 - 10^5$ Pa s at the time of eruption, implying a rheology far more substantial than that of water cryomagma, possibly indicative of an ice-rich slurry (Fagents, 2003; Quick et al., 2017). Quick and Marsh (2016) modeled the heat transfer from warm ice diapirs rising through the ice shell to form extrusive features. They found that, depending upon their ascent speed and the degree of softening of their preferred ascent pathway, diapirs may move fast enough to transport cryomagmatic melt to shallow levels in Europa's crust.

Other plausible mechanisms of dome formation exist. The common spatial association with similarly-sized depressions and low-albedo spots, as well as uplifted but undisrupted domes, led Pappalardo et al. (1998) to infer that these domes, pits and spots had a common origin; they proposed impingement of ice diapirs on a brittle surface 'lid', could lead to the range of morphologies observed. Furthermore, warming and release of near-surface brine pockets might explain the negative relief and low-albedo moats (Head and Pappalardo, 1999). Rathbun et al. (1998) modeled the dynamics of diapiric ascent and found that diapirs likely originated at depths less than a few tens of kilometers, suggesting that liquid water may have recently existed at those depths in Europa's ice shell. Figueredo et al. (2002) proposed that the unique morphology of Murias Chaos (Figure 5.19d) could be explained breaching of the surface by a warm ice diapir and subsequent lateral flow of the warm ice over the surface (Figure

5.20a).

The similarity of the surface textures of dome-like features to those of larger disrupted chaos areas, led to the adoption of the term 'microchaos' to describe the former. The larger chaos terrains were explained as the coalescence of large or multiple ice diapirs at the surface, such that the existing crust became broken into blocks within a disaggregated matrix (Figure 5.20b). Another school of thought suggested that chaos areas, large and small, were the result of a melting-though of the ice shell by focusing of tidal heating such that surface ice became disrupted into rafts or icebergs floating in a liquid matrix (Carr et al., 1998; Greenberg et al., 1999; Riley et al., 2000). Collins et al. (2000) argued that the energy required to melt entirely through the ice shell would be extremely difficult to achieve, and instead proposed that chaos areas form over discrete bodies of water contained within the ice shell (Figure 5.20b). This scenario has been shown to satisfy observations of the range of morphologies and scales, from simple pit-like depressions to negative-relief chaos, to positive relief, dome-like features, resulting from volume changes and flexural/brittle response during various stages of the formation and refreezing of these liquid bodies (Figures 5.6, 5.7; Sotin et al., 2002; Schmidt et al., 2011; Walker and Schmidt, 2015; Michaut and Manga, 2014; Manga and Michaut, 2017). Although the domes, pits, and spots may not represent true cryovolcanic effusions, these features argue for endogenic formation processes, and the presence of subsurface fluid (cryomagmatic) bodies in the past (which have refrozen to produce uplifted areas) and existing today (manifested as pits or depressions; Schmidt et al., 2011), are of substantial astrobiological interest (Ruiz et al., 2007).

On Ceres, Sizemore et al. (2019) catalogued multiple large domes (diameters ≥ 30 km) and small mounds (diameters < 10 km) (Figure 5.19e-h), several of which have morphologies and geological settings that may be indicative of a cryovolcanic origin (Buczkowski et al., 2016; Sori et al., 2017). Of these large domes, Sori et al. (2018) identified 22 with aspect ratios that are consistent with an extrusive origin, similar to Ahuna Mons (see Section 5.5.3). These authors speculated that these domes were emplaced within the last 100 Ma, and that they formed from material that included >40 vol% of ice. Recently, Bland et al. (2019) suggested that many of these domes may have formed as a result of solid-state diapirism in Ceres' ice-rich crust. Quick et al. (2019) suggested that, like Ahuna Mons, Cerealia Tholus (Fig. 5.19g) may have a cryovolcanic origin, although emplacement as an intrusive dome is also possible (Scully et al., 2020). Most formation hypotheses for domes on Ceres involve the upward migration and, in many cases, the extrusion of briny fluids onto Ceres' surface. Ceres' dome population may be a result of both intrusive and extrusive processes.

Linear features

A final category of possible effusive cryovolcanic feature involves predominantly linear features forming ridges or bands. Some features, such as the bright grooved terrain of Ganymede, might indicate erupting or upwelling material within tectonic features such as graben. Other features exhibit substantial relief and, if they represent extrusive features, would have had a substantial viscosity and strength, requiring exotic cryomagma compositions (Figure 5.9) or ice in mostly solid state. Although these features may have origins that are not conclusively cryovolcanic, in part because of a lack of data, we review here these unusual features for which an origin remains to be definitively assigned.

Voyagers 1 and 2, together with the Galileo spacecraft, imaged essentially all of Jupiter's largest moon, Ganymede. The surface of Ganymede is broadly divided into two terrains: dark, ancient terrain covering \sim 35% of the planet, and younger, bright terrain that is commonly grooved (Figure 5.2). The bright terrain shows morphologic characteristics that are consistent with the emplacement of water-icebased lavas (liquid water, warm ice, or icy slush) (Schenk, 1991; Schenk and Moore, 1995; Schenk et al., 2001; Pappalardo et al., 2007). The topography and morphology of the bright, grooved terrain is consistent with it having been formed by extension, and it most likely consists of domino-style tilt blocks (Pappalardo and Greeley, 1995). Within these extensional zones are bright, smooth bands that fill topographically low regions. Although no lobate flow fronts, sinuous ridges, or vent-like features are observed, the embayment relations, in addition to the relatively high albedo and low crater-counts on these smooth bands suggest that they were formed by extrusion of cryolavas. Schenk et al. (2001) propose that initial tectonic extension produced a graben, which was then flooded by low-viscosity cryolavas, and that continued extension then produced tilt-blocks. Sippar Sulcus is a good example of this terrain (Figure 5.21a). Superficially similar bands are observed on Europa (Figure 5.21b). These have been interpreted as spreading centers that initiate along zones of weakness (fractures or the medial troughs of double ridges) in the brittle ice layer, which then separated and spread to accommodate upwelling of warm, convecting ice (Prockter et al., 2002). This process is somewhat analogous to terrestrial mid-ocean spreading centers, but as such does not represent eruptive activity.

The surfaces of Uranus' moons have only been imaged by the Voyager 2 flyby in 1986, so that only portions of those icy bodies were imaged. Even so, the geomorphology exposed on the surfaces of these tiny moons reveals some astonishing endogenic and possibly cryovolcanic processes. Miranda is Uranus' innermost moon, and is only 472 km in diameter (Smith et al., 1986; Pappalardo et al., 1997). During the Voyager 2 flyby, Miranda's southern hemisphere was imaged; the highest resolution obtained was just

under a kilometer per pixel. Three ovoid to rectangular coronae were observed. These coronae are characterized by concentric ridges and smooth bands that show a lower impact crater density than the surrounding plains. One of the earliest interpretations of this bizarre morphology was that Miranda was broken apart by a massive impact event and then gravitationally reassembled (Smith et al., 1986). An alternative interpretation is that the coronae represent volcano-tectonic activity. Beddingfield et al. (2015) use impact crater size-frequency distributions to conclude that, of the three coronae, Elsinore corona is the oldest (perhaps as old as 3.5 Ga) while Arden and Inverness coronae are substantially younger (~1.6 Ga). This age range is more consistent with a volcano-tectonic origin for the coronae than a giant impact origin. In this model, the coronae are the surface expression of diapiric upwelling, causing extension (normal faulting) and cryovolcanism (Pappalardo and Greeley, 1995). Schenk (1991) identified two distinct morphologic types of "flow-like ridges" in Elsinore Corona: 1) long (20-50 km), linear ridges that are 300–600 m tall and 3–5 km wide, characterized by a shallow groove along the crest of the ridge; and 2) a traceable flow band that extends >250 km long, 8–10 km wide and 100–500 m tall (Figure 5.21c). These ridges superpose the surrounding terrain, and their morphologies are consistent with extrusion and emplacement of a high-viscosity ($^{1}0^{7}-10^{16}$ Pa s) material—possibly a mixture of ammonia and water ice (Jankowski and Squyres, 1988; Schenk, 1991; Kargel et al., 1995).

Uranus' moon Ariel (diameter = 1158 km) also displays evidence for cryovolcanic ridges (Figure 5.21d). Like Miranda, Ariel displays morphologic evidence for extrusions of viscous fluid (plausibly a mixture of water ice and ammonia, with other additives possible; Jankowski and Squyres, 1988; Kargel et al., 1995). Voyager 2 imaged ~35% of Ariel. Plescia (1987) identified three main geologic terrains on Ariel: cratered terrain, ridged terrain and plains. Some of the plains are located in the floors of east–west oriented troughs that are interpreted to be graben (Schenk, 1991); and locally, the margins of the plains are lobate, morphologically similar to lava flow margins. Plains within the graben are commonly brighter and smoother than plains elsewhere, and also commonly contain a linear to sinuous medial trough (Plescia, 1987). Schenk (1991) interprets these intra-graben plains to be cryovolcanic, with the medial troughs acting as possible vents (see also Kargel, 1995; Peterson et al., 2015). These observations are consistent with a period of global resurfacing: the cratered terrain formed first, then the graben, and finally volcanic extrusion (Schenk, 1991; Peterson et al., 2015).

At Neptune, Triton displays multiple long, cross-cutting troughs that are interpreted to be graben (Figure 5.21e). Some graben span the entire imaged area of the moon, so maximum lengths cannot be measured. Medial ridges are observed within some graben floors and appear to be superposed on the

graben, suggesting that the ridges are younger. These ridges have been interpreted to be extrusions of viscous cryolava (possibly a mixture of water ice and ammonia) that erupted through dikes that intersect the graben floors (Smith et al., 1989). In must be remembered, however, that data quality and quantity is limited for these far distant bodies, and so we await additional information from future missions to draw firmer conclusions about the role of cryovolcanism in producing these features.

5.5.3 Major constructional landforms

Ceres' Ahuna Mons is the first constructional cryovolcanic landform to be identified in the inner Solar System (Ruesch et al., 2016). Imagery from the Dawn spacecraft revealed it to be 4 km tall and 17 km in diameter (**Figure 5.22a**), situated close to Ceres' equator at 10.5°S and 45°W. Based on the dome's morphology and the strength of the cerean lithosphere at its location, this feature is believed to have formed from multiple episodes of extrusive cryovolcanism (Ruesch et al., 2016). Recent analysis of Dawn gravity data has revealed an isostatic anomaly beneath Ahuna Mons that has been interpreted as a convective plume from which the dome-forming fluids originated (Ruesch et al., 2019b). Dynamical considerations for the ascent of cryolava to Ceres' surface in fractures suggests that the initial viscosity of the fluids that formed this feature was 10³ Pa s, indicative of a slurry-like fluid containing 30–40 vol% solid particles. Based on the composition of Ceres' crust and the spectra of the dome itself (Zambon et al., 2017), this slurry was likely a mix of ice, carbonate salts, and phyllosilicates (Ruesch et al., 2019b), suggesting that Ahuna Mons has a mud-rich composition. Ahuna Mons' upper limit age of 240 Ma suggests that it is too young to have undergone the viscous relaxation that other domes on Ceres may have experienced (Sori et al., 2017).

Other possible constructional landforms on Ceres include Cerealia Tholus in Occator crater (Figure 5.15c, 5.19g) and Yamor Mons, an ancient dome located near Ceres' north pole. Both Yamor Mons and Cerealia Tholus have been interpreted as extrusive cryovolcanic constructs (Nathues et al., 2017, 2019) that formed similarly to Ahuna Mons (Sori et al., 2018; Quick et al., 2019). Quick et al. (2019) noted that the latter may have been formed by fluids that were effused onto the surface during the gradual freezing of a subsurface brine reservoir. However, it is also possible that Cerealia Tholus is an uplift that formed in response to a laccolithic intrusion in the shallow subsurface (Schenk et al., 2019), or that the fluids that formed Cerealia Tholus originated from a shallow melt chamber that was produced as a result of the Occator forming impact (Hesse and Castillo-Rogez, 2019; Bowling et al., 2019). Recently, Raymond et al. (2020) have suggested that Cerealia Tholus may have been emplaced when brines in an endogenic reservoir were mobilized by the additional heat and substantial fracturing that occurred as a result of

the Occator forming impact. This mobilization led to the eruption of brines on the surface, and the creation of Occator Tholus and the enigmatic bright faculae.

The strongest candidates for cryovolcanic features on Titan are located in the Sotra Patera – Doom Mons – Erebor Mons region (Figure 5.22c; Lopes et al., 2013). Analysis of Cassini radar data (topography and SAR imaging) for the region showed that Doom and Erebor Montes are tall mountains (Doom is >1 km high), and Sotra Patera is a 17 km wide and >1 km deep non-circular depression interpreted to be a collapse feature adjacent to Doom Mons. Lobate flow features (collectively named Mohini Fluctus; Figure 5.18f) appear to emerge from Doom Mons, and flows also surround Erebor Mons, the mountain at the northern end of the region. Data from VIMS (Visible and Near Infrared Mapping Spectrometer) indicate that these areas are different in composition from the surrounding terrains. Of particular interest is the fact that the area is devoid of fluvial channels, at least at the available resolution, making a fluvial origin for the flows less likely. The fact that several depressions in the region, notably Sotra Patera, are not circular makes an impact origin unlikely for these features. Further support for a cryovolcanic origin for these features has come from analysis of VIMS data in combination with a radiative transfer code to account for atmospheric effects (Solomonidou et al., 2016), which showed temporal changes in the Sotra Patera region. The area became brighter up to a factor of 2 (in terms of pure surface albedo and brightness) during a period of one year (2005-2006), while the surrounding area did not present any significant changes for the same period of time, indicating a localized phenomenon. The albedo variations could be due to degassing, indicating that the region is still active, although other explanations are possible. In addition to the morphological evidence and the temporal changes, interior structure models by Sohl et al. (2014) indicate that the Doom-Sotra-Erebor region is a likely candidate for cryovolcanic activity based on the predicted spatial pattern of maximum tidal stresses. In fact, cryovolcanism on Titan can explain outgassing over time to replenish methane in the atmosphere, which is destroyed timescales of 30 Ma by photochemical dissociation (Sotin et al., 2005; Tobie et al., 2006).

Among other candidate cryovolcanic features on Titan, a radar-dark area was interpreted as a possible cryovolcanic dome from early Cassini data (Lopes et al., 2007) but additional data collected by the radar instrument later on the Cassini mission showed that the apparent construct, named Ganesa Macula, was a heavily eroded feature rather than a dome or shield. However, a dome-like feature approximately 100 km in diameter with a criss-cross fracture pattern (nicknamed the Hot Cross Bun) was interpreted from SAR data as a possible cryovolcanic bulge or laccolith (Figure 5.22c; Lopes et al., 2012;

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Malaska et al., 2016). The presence of similar uplifted, radially dissected "labyrinth" terrains on Titan has been interpreted as the surface manifestation of cryomagmatic intrusions at the brittle–ductile transition depth within the ice shell (Schurmeier, 2018).

At Pluto, Wright Mons is an annular massif that stands ~4 km tall and 150 km in diameter, and has a large central depression that is 45 km wide and 4–5 km deep (Moore et al., 2016; Singer et al., 2016; Figure 5.22d). A flow-like feature extends from a depression on its southwest flank. Piccard Mons, which shares many characteristics with Wright Mons, is ~ 6 km tall and 225 km in diameter (Moore et al., 2016; Singer et al., 2016). The lack of impact craters on both Wright and Piccard Montes is indicative of young surfaces. Although CH_4 , N_2 , and CO ices are prevalent on Pluto's surface (Stern, 1992), these ices are not strong enough to support the topography of these edifices, or to preserve impact craters (Eluszkiewicz and Stevenson, 1990; Yamashita et al., 2010). This suggests that the Montes may be composed of water-ice dominated bedrock (Stern et al., 2015).

5.5.4 Depressions, pits, and caldera-like features

In addition to the negative-relief features associated with major constructs on Titan and Pluto (Section 5.5.4; Figure 5.21b,d), there exists a range of other depressions, pits, or caldera-like depressions on the icy satellites, the characteristics and settings of which are suggestive of collapse related to eruptive or intrusive processes. In some cases, pits might represent eruptive vents or explosion craters.

On volcanoes of the rocky planets, calderas are steep-sided depressions with a wide range in diameters (on the order of 1 to >100 km). They form by collapse of the ground surface in response to the withdrawal of magma from a magma reservoir as a result of eruptions or intrusions. In some cases the eruptions that cause withdrawal are explosive, and explosive eruptions may also accompany caldera-forming events. However, in other cases, particularly for basaltic volcanoes, calderas form in response to prolonged effusive events removing volume from the reservoir. This is to say that the identification of a caldera-like feature alone does not imply a specific style of cryovolcanism; simply that volume has been removed from an underlying reservoir, leading to collapse of near surface layers forming a walled depression. In ice shells, this volume loss could simply be the result of a change from solid to liquid state within the ice, in which case, cryovolcanism is not implicated at all.

While calderas are large-scale features forming over subsurface magma storage reservoirs, the terms depression or pit might be applied to smaller scale features and/or those features for which an origin is

not clear. Some kind of volume loss or void in the subsurface is still required for formation; for example, on silicate volcanoes aligned pits might form over lava tubes.

At Europa, depressions and pits of order 5–10 km diameter (Figure 5.23a) are commonly spatially associated with domes and dark spots of similar size, and as discussed in Section 5.5.2, might be related to the presence of subsurface fluid bodies (Schmidt et al., 2011; Michaut and Manga, 2014; Manga and Michaut, 2017). Chaos terrain can also exhibit negative relief with respect to the surrounding terrain, sometimes manifesting as walled depressions, e.g., Thera Macula (Figure 5.18a), and similarly can be explained by the volume loss due to subsurface melting and/or the thermal effects of warm fluids on the near-surface ice (Figure 5.6b; Schmidt et al., 2011; Walker and Schmidt, 2015).

Castalia Macula, on the other hand, is a broad (20 x 25 km across), 350 m deep depression with gently sloping sides, distinguished by its dark, reddish coloration (Figure 5.23b; Prockter and Schenk, 2005). It lies on Europa's equator between two uplifted domes, 900 and 750 m high, which themselves exhibit evidence of lobate flows. Although the origin of the depression itself is unknown, the dark deposit has been interpreted as resulting from the flooding of Castalia Macula with a low viscosity fluid, which subsequently drained or evaporated, leaving a thin, dark, residual deposit, through which the preexisting ridged plains are visible (Prockter and Schenk, 2005).

At Ganymede, at least 18, and perhaps as many as 30, caldera-like features have been identified in Galileo imaging data (Schenk and Moore, 1995; Pappalardo et al., 2007). These features range in size from a few to several tens of kilometers, and are characterized by a bounding scarp that may or may not be closed, but have scalloped planform shapes quite distinct from impact craters, and similar to those expressed by terrestrial volcanic calderas. These depressions are filled with relatively smooth, bright material, which might represent cryovolcanic flows (Figure 5.23c).

In Titan's north polar region, a subset of lakes and depressions have been interpreted to result from explosive events analogous to maar formation on Earth. These depressions have sharp edges and raised rims and are surrounded by a raised rampart of material (Figure 5.23d). Some are circular in planform, while other have irregular, scalloped shapes, and they range from a few to tens of km in diameter (Mitri et al., 2019; Wood and Radebaugh, 2020). They exhibit morphologies inconsistent with those of impact craters. Although others have interpreted the depressions as forming from karst-like dissolution processes (Hayes et al., 2016), Mitri et al. (2019) interpreted these features as forming by explosive release of nitrogen gas during a past period of climate warming. However, the fresh appearance of these features suggest they might be more recent phenomena. Wood and Radebaugh (2020) proposed that

these features formed when a relatively warm aqueous cryomagma impinged upon and vaporized nitrogen or methane contained in porous near-surface ice (the "alkanofer"). The excavated crater then fills with liquid methane to form a lake. The depressions are therefore analogous to maar craters that form during terrestrial phreatic explosions, and the radar-bright haloes surrounding the depressions represent deposits of explosive ejecta.

Finally, Triton possesses several types of depressions at a range of scales. Some features resemble terrestrial volcanic calderas filled with fluid lava. Figure 5.23e shows one such example, Ruach Planitia. These features are characterized by irregular, steep bounding scarps ~1 km tall that enclose smooth plains, ranging from ~80 to 390 km across (Croft et al., 1995). The plains contain clusters of pits, which may be volcanic vents from which the smooth material erupted, ponded, and filled the area enclosed by the scarps. Alternatively, these pits could have been produced by explosions or collapse due to cryolava drainback (Croft et al., 1995). The bounding scarps might have formed due to collapse of the surface into an underlying, deflating cryomagma reservoir, although their convoluted nature suggests piecemeal erosional undermining of the margins. Triton's surface also contains a number of pitted features that may be cryovolcanic in origin. Pit paterae are elliptical depressions, with or without rims, and sometimes form chains (Figure 5.17; Croft et al., 1995). In some cases they are surrounded by smooth mantling deposits, which could be explosive cryoclastic deposits. Elsewhere on Triton, the so-called "cantaloupe terrain" suggests vigorous interior convective processes. In sum, Triton's diverse and youthful surface geology attests to a highly dynamic, active body that has much more to reveal next time we visit.

5.6 Summary and Future Directions

Cryovolcanism is of key importance both for understanding the geologic history and thermal evolution of ice-rich planetary bodies, and for assessing the habitability of outer Solar System bodies. The presence of subsurface liquid oceans (extant or relict) and/or discrete reservoirs within ice shells, in combination with detection of astrobiologically relevant elements (CHNOPS), complex organic molecules, and protection from the space environment, might provide conditions favorable for the development of life in the subsurface. Cryomagmatism and cryovolcanism, in various forms, might then provide for communication between the surface and subsurface. Delivery of material to the surface would provide for sampling of young ocean/subsurface materials and any biosignatures contained in those materials, which might otherwise be beyond the reach of current planetary exploration technologies. The identification of cryovolcanic features, active or fossilized, is key motivation for continued exploration of the outer Solar System, as we seek to answer the question of whether life exists, or has ever existed, beyond Earth.

Recent decades have brought numerous discoveries in the outer Solar System that have amazed planetary scientists, and acquisition of high resolution imaging data sets have revealed features of possible cryovolcanic origin from Ceres all the way out to Pluto and Charon. However, it is probably true to say that although it is by no means the dominant resurfacing process. Furthermore, cryovolcanism is not well understood from a theoretical perspective – our models rely on parameters that are commonly not well constrained (e.g., ice shell thickness, thermal profiles, cryomagma composition and properties). Interpretations of cryovolcanic features from planetary imagery must be made with caution, with consideration of the quality and resolution of the data in hand. Clearly though, the plumes erupting from Enceladus are a spectacular reminder that cryovolcanism is indeed a viable geologic process in some cases.

Europa is a well imaged body, but even so the prevalence of cryovolcanism is uncertain. Aside from the observations of possible plumes, there is good morphological evidence for some effusion of fluids at the surface, but other features are perhaps better explained by other mechanisms. Whether the variety of surface features at Europa are produced by cryovolcanism vs. interior processes (convection, diapirism, cryo-intrusions) is still a subject of active research. Even if, ultimately, Europa's diverse surface features are found to be formed by mechanisms other than cryovolcanism, they attest to a dynamic interior aided by the presence of a subsurface ocean, and processes that may deliver material close to the surface.

The difficulty in assigning a cryovolcanic origin to surface features is clearly exacerbated for bodies where data is much more limited than at the Jupiter or Saturn systems (e.g., the satellites of Uranus and Neptune). Caution is required in interpreting a cryovolcanic origins for features where image resolution is low, coverage is sparse, or there is a lack of complementary data sets. Answering lingering questions about the viability of cryovolcanism as a geologic process on icy bodies will require that future missions to collect suites of complementary data — morphology, composition, thermal emission, and topography together provide the best opportunity to interpret the geologic features of far-off icy bodies. These data sets, in combination with numerical modeling of cryovolcanic processes, and laboratory and field analog studies of icy features and materials, will allow further progress to be made.

Future planned missions scheduled to launch in the mid- to late-2020s, such as the Jupiter Icy Moons Orbiter (JUICE), Europa Clipper, and Dragonfly to Titan are expected to provide new observational data

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on cryovolcanism and its expressions, which will no doubt lead to improved modeling efforts. In addition to shedding light on the extent of surface-subsurface exchange processes on a moon with an ocean sandwiched between layers of ice, the JUICE mission will constrain the role that cryovolcanic processes, if any, play on a world with marginal tidal heating such as Ganymede. High-resolution imagery of Europa's surface by Europa Clipper may finally catch active cryovolcanism in the act on the ocean moon, possibly in the form of plumes and effusive eruptions. A follow-on Europa lander mission could provide the opportunity for us to sample cryovolcanically active regions on Europa, which would shed light on the composition and habitability of internal fluid reservoirs and/or the ocean. The DragonFly mission to Titan will produce high-resolution imagery of multiple regions on Titan's surface, allowing us to place the role that cryovolcanism has played in context with that of fluvial, aeolian, and impact processes. Future proposed missions such as Trident would execute a flyby of Triton and help elucidate the potential role of cryovolcanism in icy moon habitability at 30 AU. A possible return mission to Enceladus and missions to the Uranus and Neptune systems, which have yet to be explored by orbital spacecraft, would usher in a new era of comparative planetology focused on understanding the interior and surface expressions on extant and past ocean worlds. Knowledge of cryovolcanism and its role in habitability on these worlds may in turn be used to inform the characterization of exo-ocean worlds such as the Trappist planets. The next few decades will undoubtedly bring exciting new discoveries on the icy bodies of the Solar System.

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