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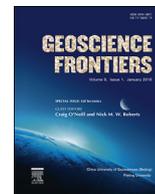


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Research Paper

# Stagnant lid tectonics: Perspectives from silicate planets, dwarf planets, large moons, and large asteroids

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## ARTICLE INFO

## Article history:

Received 7 November 2016

Received in revised form

22 May 2017

Accepted 16 June 2017

Available online 1 July 2017

## Keywords:

Stagnant lid

Solar system

Plate tectonics

Planets

Moons

## ABSTRACT

To better understand Earth's present tectonic style—plate tectonics—and how it may have evolved from single plate (stagnant lid) tectonics, it is instructive to consider how common it is among similar bodies in the Solar System. Plate tectonics is a style of convection for an active planetoid where lid fragment (plate) motions reflect sinking of dense lithosphere in subduction zones, causing upwelling of asthenosphere at divergent plate boundaries and accompanied by focused upwellings, or mantle plumes; any other tectonic style is usefully called “stagnant lid” or “fragmented lid”. In 2015 humanity completed a 50+ year effort to survey the 30 largest planets, asteroids, satellites, and inner Kuiper Belt objects, which we informally call “planetoids” and use especially images of these bodies to infer their tectonic activity. The four largest planetoids are enveloped in gas and ice (Jupiter, Saturn, Uranus, and Neptune) and are not considered. The other 26 planetoids range in mass over 5 orders of magnitude and in diameter over 2 orders of magnitude, from massive Earth down to tiny Proteus; these bodies also range widely in density, from 1000 to 5500 kg/m<sup>3</sup>. A gap separates 8 silicate planetoids with  $\rho = 3000$  kg/m<sup>3</sup> or greater from 20 icy planetoids (including the gaseous and icy giant planets) with  $\rho = 2200$  kg/m<sup>3</sup> or less. We define the “Tectonic Activity Index” (TAI), scoring each body from 0 to 3 based on evidence for recent volcanism, deformation, and resurfacing (inferred from impact crater density). Nine planetoids with TAI = 2 or greater are interpreted to be tectonically and convectively active whereas 17 with TAI < 2 are inferred to be tectonically dead. We further infer that active planetoids have lithospheres or icy shells overlying asthenosphere or water/weak ice. TAI of silicate (rocky) planetoids positively correlates with their inferred Rayleigh number. We conclude that some type of stagnant lid tectonics is the dominant mode of heat loss and that plate tectonics is unusual. To make progress understanding Earth's tectonic history and the tectonic style of active exoplanets, we need to better understand the range and controls of active stagnant lid tectonics.

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## 1. Introduction

These are wonderful times to be a geoscientist. One of the most exciting opportunities that have recently appeared is to compare Earth tectonics to that of other bodies in the Solar System. This is now possible because all of the larger solid objects – planets, moons, and asteroids – have been imaged and studied to various extents over the past half-century. With 2015 visits to 1 Ceres by the Dawn spacecraft and the Pluto-Charon system by the New

Horizons spacecraft, we now have looked in some detail at all the large solar system bodies. These have been imaged using visible light and reflected as well as thermal infrared for 24 planetoids and synthetic aperture radar (SAR) for Venus and Saturn's largest moon, Titan. For some of these bodies, only a small portion has been imaged in any detail, but this is enough to let us begin considering what these bodies tell us about the range of tectonic behaviors that exist in the Solar System.

Terminology for the larger bodies going around the Sun has long been established by astronomers, who refer to planets, moons, and asteroids. This terminology is being modified as we learn more about the smaller bodies, especially those in the asteroid belt and the Kuiper Belt (only known to exist since 1992; Jewitt and Luu, 1993). For example, The International Astronomical Union in

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Peer-review under responsibility of China University of Geosciences (Beijing).

2006 redefined a planet as (1) a body that orbits the Sun; (2) is massive enough for its gravitational field to make it approximately spherical; and (3) cleared its orbital neighborhood of smaller objects around its orbit. In contrast, a “dwarf planet” is a body that: (1) is in orbit around the Sun, (2) has sufficient mass for its self-gravity to overcome rigid body forces so that it assumes a hydrostatic equilibrium (nearly round) shape, (3) has not cleared the neighborhood around its orbit, and (4) is not a satellite. Under this new definition, Pluto and 1 Ceres are dwarf planets in the Kuiper Belt and asteroid belt, respectively.

The terminology established by astronomers is less useful for geoscientists, who are less concerned with orbital mechanics and more interested in the composition and tectonic style of the bodies themselves, especially the extent to which these are Earth-like. There is no accepted term encompassing all large objects in our solar system. Because this paper is written by geoscientists to compare broadly similar solid bodies for the purpose of better understanding the tectonic evolution of Earthlike bodies, we informally and for brevity refer to all of the larger solar system bodies orbiting the Sun as “planetoids”. The term planetoid is often used as a synonym for asteroid but that is not how we use the term here. For comparison with Earth, we focus on planetoids with the following characteristics: (1) they have solid surfaces that have been imaged (removing Jupiter, Saturn, Uranus, and Neptune from consideration); and (2) they have approximately spherical shapes. The first characteristic allows us to examine the body and determine whether it is tectonically active or not. The second characteristic is explained in terms of rock (or ice) strength (yield stress) relative to stresses created by topography. With this consideration in mind, we take 400 km as the minimum diameter for the bodies

we consider, because this is close to the limit between spherical and irregular-shaped bodies for both silicate and icy bodies ( $\sim 525 \text{ km}^2$  Vesta and  $\sim 420 \text{ km}$  Proteus are the largest irregular silicate and icy bodies, respectively). The 30 largest bodies in our solar system are listed in [Table 1](#), along with their approximate diameters, densities, masses, whether or not spherical, likely composition, and the names of the missions that studied these bodies. These include the 8 planets (Mercury, Venus, Earth, Mars, Jupiter, Saturn, Uranus, and Neptune), two asteroids (Ceres and Vesta), 19 moons (Earth’s Moon; 4 moons of Jupiter: Io, Europa, Ganymede, and Callisto; 6 moons of Saturn: Enceladus, Tethys, Dione, Rhea, Titan, Iapetus; 5 moons of Uranus: Miranda, Ariel, Umriel, Titania, and Oberon; 2 moons of Neptune: Proteus and Triton; and two Kuiper Belt objects (Pluto and Pluto’s moon Charon). Several additional Kuiper belt dwarf planets that are in the appropriate size range have been discovered (e.g. Eris, Orcus, Sedna, Quaoar) but we do not include them as insufficient observations exist to determine whether or not these are tectonically active.

The wealth of information about planetoids invites geoscientists to contribute their perspectives about how these bodies evolved. Our focus is on tectonics, for which we modify the definition of [Burke \(2011\)](#): “The large scale evolution of planetoid lithospheres” (In section 3 below we define lithosphere). Our survey of the 26 large planetoids reveals that plate tectonics is an unusual tectonic style in the Solar System today. Most planetoids are either tectonically inactive (dead) and most active planetoids are characterized by stagnant lid tectonics.

Here we consider the other large solid objects in the Solar System to see if these can provide new insights into Earth’s tectonic history, especially the possibility that there were times when Earth

**Table 1**  
Summary of 30 largest Planetoids (Planets in Boldface).

Body	Diameter (km)	Density (g/cm <sup>3</sup> )	Mass (kg)	Spherical?	Likely composition	TAI <sup>c</sup>	Pertinent Missions (All NASA except noted)
<b>Mercury</b>	4900	5.4	$3.3 \times 10^{23}$	Yes	Silicate	1	Mariner 10 (1974); MESSENGER (2011–2015)
<b>Venus</b>	12,100	5.2	$4.9 \times 10^{24}$	Yes	Silicate	3	Venera (USSR), Vega (USSR), Magellan
<b>Earth</b>	12,700	5.5	$6 \times 10^{24}$	Yes	Silicate	3	Many
Moon	3500	3.3	$7.3 \times 10^{22}$	Yes	Silicate	0	Apollo 11–16
<b>Mars</b>	4220	3.9	$6.4 \times 10^{23}$	Yes	Silicate	2	Many <sup>b</sup>
<i>Ceres</i>	940	2.2	$9 \times 10^{20}$	Yes	Ice (MSIB <sup>a</sup> )	1	Dawn (2015–present)
<i>Vesta</i>	525	3.5	$2.6 \times 10^{20}$	No	Silicate	0	Dawn (2011–2012)
<b>Jupiter</b>	1,34,000	1.3	$1.9 \times 10^{27}$	Yes	?		Voyager 1, 2 (1979); Galileo (1995–2003)
Io	3600	3.5	$8.9 \times 10^{22}$	Yes	Silicate	2	Voyager (1979); Galileo (1995–2003); New Horizons 2007
Europa	3100	3	$4.8 \times 10^{22}$	Yes	Silicate	2	Voyager (1979); Galileo (1995–2003)
Ganymede	5300	1.9	$1.5 \times 10^{23}$	Yes	Ice	1.5	Voyager (1979); Galileo (1995–2003)
Callisto	4800	1.8	$1.1 \times 10^{23}$	Yes	Ice	0	Voyager (1979); Galileo (1995–2003)
<b>Saturn</b>	1,16,000	0.7	$5.7 \times 10^{26}$	Yes	?		Pioneer 11 (1979); Voyager 1, 2 (1980); Cassini (2004–present)
Enceladus	504	1.6	$1.1 \times 10^{20}$	Yes	Ice (MSIB <sup>a</sup> )	2	Voyager 1 (1980); Cassini (2004–present)
Tethys	1100	1	$6.2 \times 10^{20}$	Yes	Ice (MSIB <sup>a</sup> )	1.5	Cassini (2004–present)
Dione	1100	1.5	$1.1 \times 10^{21}$	Yes	Ice (MSIB <sup>a</sup> )	1.5	Voyager 1 (1980); Cassini (2004–present)
Rhea	1500	1.2	$2.3 \times 10^{21}$	Yes	Ice (MSIB <sup>a</sup> )	1	Voyager 1 (1980); Cassini (2004–present)
Titan	5200	1.9	$1.3 \times 10^{23}$	yes	Ice & tholin	2	Voyager, Cassini (2004–present)
Iapetus	1500	1.1	$1.8 \times 10^{21}$	Yes	Ice (MSIB <sup>a</sup> )	0	Cassini (2004–present)
<b>Uranus</b>	51000	1.3	$8.7 \times 10^{25}$	Yes	?		Voyager 2 (1986)
Miranda	470	1.2	$6.6 \times 10^{19}$	Yes	Ice (MSIB <sup>a</sup> )	1	Voyager 2 (1986)
Ariel	1200	1.6	$1.4 \times 10^{21}$	Yes	Ice (MSIB <sup>a</sup> )	1	Voyager 2 (1986)
Umbriel	1200	1.4	$1.2 \times 10^{21}$	Yes	Ice (MSIB <sup>a</sup> )	1	Voyager 2 (1986)
Titania	1600	1.7	$3.5 \times 10^{21}$	Yes	Ice (MSIB <sup>a</sup> )	1	Voyager 2 (1986)
Oberon	1500	1.6	$3.0 \times 10^{21}$	Yes	Ice (MSIB <sup>a</sup> )	1	Voyager 2 (1986)
<b>Neptune</b>	49000	1.6	$1.0 \times 10^{26}$	Yes	?		Voyager 2 (1989)
Proteus	420	1.3	$4.4 \times 10^{19}$	No	Ice (MSIB <sup>a</sup> )	1	Voyager 2 (1989)
Triton	2700	2.1	$2.1 \times 10^{22}$	Yes	Ice (MSIB <sup>a</sup> )	3	Voyager 2 (1989)
<i>Pluto</i>	2400	1.9	$1.3 \times 10^{22}$	Yes	Ice (MSIB <sup>a</sup> )	2	New Horizons (2015)
Charon	1200	1.7	$1.5 \times 10^{21}$	Yes	Ice (MSIB <sup>a</sup> )	1.5	New Horizons (2015)

<sup>a</sup> MSIB: Medium-sized icy body ([Czechowski and Leliwa-Kopystynski, 2002](#)).

<sup>b</sup> Mariner 4–9 (1965–1971); Viking (1976); Global Surveyor (1997–2006); Pathfinder (1997); Phoenix (2008); Exploration Rovers (2004–present); Mars Science Lab./Curiosity (2012–present).

<sup>c</sup> Tectonic Activity Index.

did not have plate tectonics and what Earth's pre-plate tectonic convective style might have been. Plate tectonics is a special planetary convection style that is characterized by independent motion of lithospheric fragments, with new lithosphere forming at mid-ocean ridges and consumed in subduction zones and continental collision zones. Earth is the only planetoid known to now have plate tectonics. An implicit assumption is that all these bodies have similar origins as condensates of the solar nebula and for this reason are appropriate to compare. Our expectation is that an inventory of Solar System planetoids with solid surfaces will reveal which are tectonically active and which are tectonically dead, and that examining the active ones will further illuminate some of the tectonic styles that the Earth might have experienced before it developed plate tectonics.

We approach this by first characterizing the composition and then the tectonic behavior of the 26 large planetoids with solid surfaces that have been imaged, the “Big 30” minus the large gaseous planets of Jupiter, Saturn, Uranus, and Neptune. We first summarize the terminology used for characterizing the strong outer layers of rocky and icy planetoids. Then we explain how we characterize a planetoid as tectonically active or dead. We follow this by summarizing the interior weaknesses that allow tectonically activity for both icy and rocky planets. We conclude that we have much to learn from comparative “planetoidology” but that the significant differences between the icy and the rocky bodies limit what insights from the former can usefully be applied to understanding the latter.

## 2. Planetoid compositions and structure

Planetoids have a wide range of densities, indicating a wide range of compositions. We have some information about the composition of these bodies, principally from density estimates determined from their orbits. The 30 largest bodies of our Solar System (excluding the Sun) are readily separated into three groups: (1) Four gaseous/icy giants with densities of 700–1700 kg/m<sup>3</sup> (0.7–1.7 g/cm<sup>3</sup>); (2) Eight bodies with densities between 3000 and 5500 kg/m<sup>3</sup> (3.0–5.5 g/cm<sup>3</sup>); and (3) 18 small bodies with solid surfaces and densities between 1000 and 2200 kg/m<sup>3</sup> (1–2.2 g/cm<sup>3</sup>). Note that the densities of the 4 gaseous/icy giant planets are enormously increased by self-compression, the densities of Earth and Venus are significantly increased by self-compression (the uncompressed densities would be about 4.1 and 4.0 g/cm<sup>3</sup>, respectively), but the densities of Mars and smaller bodies are much less affected. Fifteen of the small planetoids have been termed “medium-sized icy bodies” (Czechowski and Leliwa-Kopystynski, 2002). The density vs. diameter variations of these 30 solar system objects are shown on Fig. 1 and the compositional implications of the density spectrum is explored in Fig. 2, which also shows a clear density gap between 2200 and 3000 kg/m<sup>3</sup> (2.2 and 3.0 g/cm<sup>3</sup>) for these planetoids. The four gaseous/icy giants do not figure much in this discussion; focus is on the 26 smaller, denser planetoids. The densest eight bodies must be composed of silicates and iron whereas the lightest 18 bodies contain much more ice.

The big 30 planetoids are sufficiently large that they mostly differentiated into compositional layers according to densities: Fe-Ni core at the center overlain by Mg-rich silicate mantle overlain by Ca-Al-rich silicate crust, overlain by fluids and ices, all enveloped by a gaseous atmosphere. This is the situation for Earth, which has a bulk density of 5500 kg/m<sup>3</sup> (5.5 g/cm<sup>3</sup>), which is the weighted mean for Earth's compositionally distinct shells: inner solid and outer liquid core (9900 to 13,000 kg/m<sup>3</sup>), mantle (3300 to 5700 kg/m<sup>3</sup>), crust (2700–3300 kg/m<sup>3</sup>), water and ice (~1000 kg/m<sup>3</sup>) and atmosphere (<1000 kg/m<sup>3</sup>). We assume that most of the big 30 planetoids are similarly differentiated into compositionally-distinct

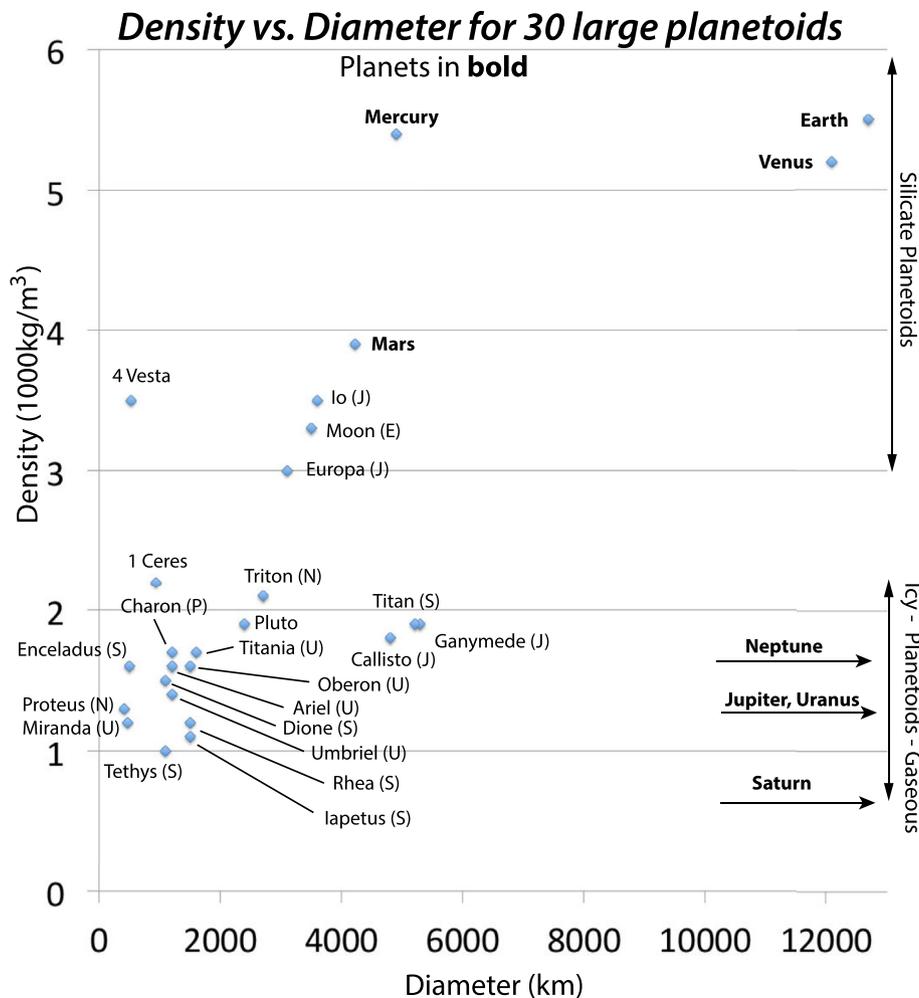
layers although we can only constrain the compositions and proportions of each layer for a few of them, for example Earth and Earth's moon. To do this will require experiments designed to resolve interior planetary structure, such as the Juno mission, now in orbit around Jupiter, which will constrain Jupiter's interior structure by measuring gravity, or the planned NASA InSight (Interior Exploration using Seismic Investigations, Geodesy and Heat Transport) mission that will place a single geophysical lander on Mars to study its deep interior using a seismometer.

Some of the smallest, lightest planetoids may be undifferentiated and not have rocky and Fe-Ni cores, such as Tethys (1000 kg/m<sup>3</sup>, 1100 km in diameter), Iapetus (1100 kg/m<sup>3</sup>, 1500 km in diameter), Miranda (1200 kg/m<sup>3</sup>, 470 km in diameter) and Proteus (1300 kg/m<sup>3</sup>, 420 km in diameter).

## 3. Lid nomenclature and behavior

The term lithosphere was introduced by Barrell (1914) to describe the strong outer shell or lid of the solid Earth, and the concept has since evolved. Earth's lithosphere has three useful definitions: thermal, chemical, and rheological/mechanical (Anderson, 1995). Thermal lithosphere ( $L_T$ ) has a conductive geothermal gradient; chemical lithosphere ( $L_C$ ) has compositional and isotopic characteristics that have been isolated from well-mixed asthenosphere for some time; and rheological lithosphere ( $L_R$ ) is defined by strength. So defined, these different lithospheres have different thicknesses; generally  $T_T \sim T_C > T_R$ . Lithosphere is underlain by asthenosphere, which is weak enough to flow. On Earth, the asthenosphere lies a few hundred km below the surface, and below the asthenosphere the viscosity of Earth's mantle increases with depth. Convection encompasses the entire mantle, resulting in an adiabatic temperature gradient except where phase transitions inhibit vertical flow. On smaller bodies, pressures are much lower and the entire mantle may be “asthenosphere”, so in this review the term is used in a general sense to mean material that has a low enough viscosity to flow and convect. Earth's asthenosphere may be weak partly because of concentrated volatiles such as H<sub>2</sub>O and CO<sub>2</sub>. The concept of asthenosphere was developed for silicate Earth, but we also use it here to describe regions of liquid water which may underlie icy “lithosphere” on some planetoids (e.g., Europa, Enceladus and Dione). Thermal lithosphere is the outer solid part of a planetoid dominated by conductive heatflow whereas underlying asthenosphere is characterized by convection. The multifaceted definition of lithosphere developed for Earth is especially useful where the full range of geoscientific tools – geophysical, geodetic, geochemical, and isotopic – can be brought to bear. This is not yet the case for the planetoids considered here; instead we focus on the rugged, rocky (silicate-dominated) or icy surfaces that mark the upper surface of the lithosphere.

Lithosphere can also be called “lid”, in reference to its behavior as the external thermal boundary layer characterized by conductive temperature profile. This thick planetary blanket is sometimes called thermal lid. Any planetoid with a hard, rocky or icy surface must have a significant outer thermal boundary layer and thus a lid. We can distinguish planetoids with a single lid from those with a fragmented lid. Plate tectonics is a variety of fragmented lid but not the only variant. Dead planetoids necessarily have a single lid, for example Mercury or Earth's moon. Tectonically active planetoids are characterized by vigorous convection, and convection requires a layer of lower viscosity beneath the thermal lid, either asthenosphere for rocky planets or subterranean ocean for icy planetoids. We know that a planetoid has such a weak layer because it is tectonically active, which can be inferred from the nature of its solid surface.



**Figure 1.** Plot of density vs. diameter for 30 planetoids listed in Table 1. Note that the four giant planets plot far to the right. Letters in parentheses show which planet a moon orbits.

It is important to distinguish lithosphere and asthenosphere from compositionally-distinct layers crust and mantle. For Earth, lithosphere includes crust and uppermost mantle.

Earth is a tectonically active planetoid with a style of fragmented lid tectonics called plate tectonics. Plate tectonics is a style of convection for active planetoids where lid fragment (plate) motions reflect sinking of dense lithosphere in subduction zones, causing upwelling of asthenosphere at divergent plate boundaries and accompanied by focused upwellings, or mantle plumes. This definition implies that active planetoids with fragmented lids dominated by water ice cannot have plate tectonics, because water ice I (low pressure water ice) is buoyant relative to the underlying weak layer (often liquid water) and so cannot drive plate motions by sinking in subduction zones.

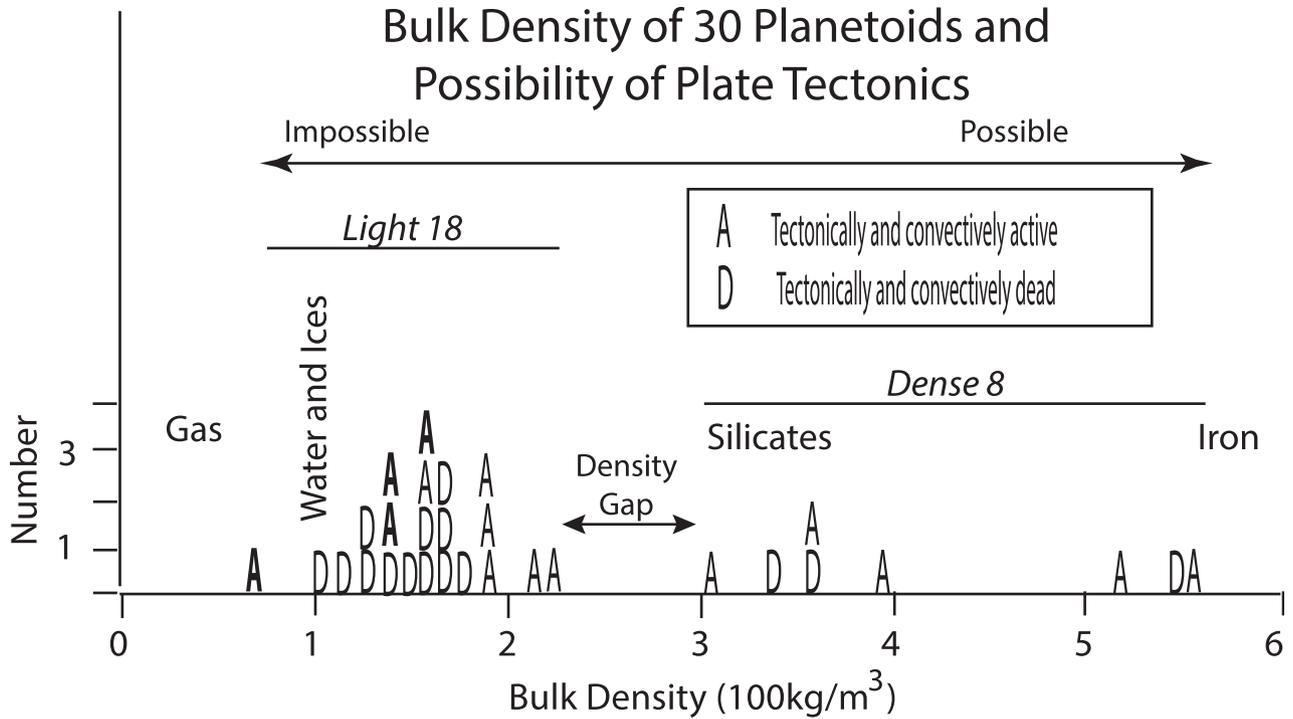
For tectonically active planetoids, a useful distinction between lid behavior of plate tectonics and “stagnant lid” tectonics is that in the first case, all of the lithosphere is involved in convection, whereas in the second case, only the warmer, weaker part of the lid is involved, by dripping off or delaminating (Stevenson, 2003).

#### 4. Planetoids: dead or alive?

We want to know whether a planetoid is tectonically dead or alive. This reflects whether or not the body convects internally, which is exceedingly difficult to determine directly, even on Earth. However, convective activity can be inferred from three independent considerations: (1) evidence that the lithosphere

has been recently deformed (inferred from imaged faults and folds); (2) evidence of a vigorous asthenosphere, (inferred from imaged volcanoes or eruptions); and (3) evidence of recent planetary resurfacing (inferred from impact crater abundance). We consider that clear evidence for the first two criteria indicates tectonic activity whereas crater density is a negative indication and dense cratering indicates a planetoid that has been tectonically dead for some time. These observations are summarized in Table 2 and used to assess whether a planetoid is tectonically active or dead. All of these assessments result from a combination of visible and near infrared (VNIR) and synthetic aperture radar imaging (SAR) of planetoid surfaces, supplemented by identification of geysers and eruption plumes. Note that “maybe” in especially the “Recent Volcanism” category is scored as 0.5.

It may be that an active planetoid doesn’t show all three evidences. Because we do not want to mischaracterize active planetoids, we assume that a planetoid with two or more of the three criteria is tectonically active (2 or more on “Tectonic Activity Index” in Table 2); otherwise we consider the planetoid in question to be tectonically dead. This list is tentative and there is likely to be some disagreement, particularly for the 3 planetoids with Tectonic Activity Index of 1.5, but future space missions will resolve uncertainties (for example ESA JUICE mission to Jupiter’s moons Ganymede, Callisto, and Europa and NASA’s new Europa mission). Our assessments of each of the planetoids using the three categories are explained below.



**Figure 2.** Spectrum of densities for 30 planetoids listed in Table 2. These bodies can be composed of 4 types of matter: gas, water and other ices, silicates, and iron. Gas is only an important component in the 4 giant planets: Jupiter, Saturn, Uranus, and Neptune, identified with bold letters. Note the “density gap” between the 8 silicate-dominated planetoids with density >3 g/cm<sup>3</sup> and the 18 “icy” planetoids with density between 1 and 2.2 g/cm<sup>3</sup>. Note which planetoids even have possibility of being active silicate bodies and even have the potential to have plate tectonics.

4.1. Faults and folds

Lithospheric deformation is revealed by its breaking (faulting), shearing, or flexing (folding). Identification of faults and related linear features like shear zones, grabens or mountain ranges on a planetoid is evidence that tectonic activity happened, but when? It is not always possible to easily distinguish recently active from long dead faults. For example, surface of the tectonically dead planet Mercury shows reverse faulting as a result of as much as 7 km of radial contraction due to cooling and solidification of its core (Byrne et al., 2014); high-angle reverse faults such as Beagle Rupes are predominantly oriented N–S and may be billions of years old (Watters et al., 2004).

The surfaces of tectonically active planets with sufficiently strong lithosphere like Earth are affected by three types of faults: normal (extensional), reverse (compressional) and strike-slip. Faulting is often associated with shearing and folding. Deformation on Earth is concentrated at plate boundaries but on stagnant lid planets is likely to be more broadly distributed. The surface of Venus is affected by all three types of faults, with a preponderance of rift-related normal faults called chasmata that extend for several thousand kilometers (Platz et al., 2015). Belts of wrinkle ridges interpreted to mark regions of contractional deformation are up to 20 km wide and can be thousands of km long (Platz et al., 2015). Strike-slip faults thought to mark several tens of kilometers of offset are also documented for Venus (Koenig and Aydin, 1998; Harris and Bédard, 2014). In addition, coronae are large circular features on Venus, with complex structures marked by circular faults and volcanic rocks. Coronae mark sites of plume-like mantle upwelling (Grindrod and Hoogenboom, 2006). Martian lithosphere is also cut by normal, reverse and strike-slip faults (Platz et al., 2015). Deformation centers on the Tharsis volcanic province, from which extensional faults radiate outwards. Here is the most

spectacular rift in the Solar System, the Valles Marineris, 4000 km long, 200 km wide and up to 7 km deep.

Faulting is also documented for many of the planetoids in the outer solar system. Faults have not been mapped on Io, probably because it is so volcanically active, but its great mountains (50–400 km long, 5–15 km tall) are thought to be great fault-bound tilted lithospheric blocks (Schenk and Bulmer, 1998) or thrust blocks (Bland and McKinnon, 2016). The great mountains on Io appear uplifted and tilted which implies a tectonic origin. These are likely associated with faulting, although the brittle-ductile transition may lie close to the surface (Kirchoff et al., 2011). Europa is an example of a planetoid with a fragmented lid (Patterson et al., 2006). It is tectonically riven and is one of the youngest (crater density age of 40–90 Ma) surfaces of the planetoids, implying rapid recycling of its icy lid. The fact that Europa’s strike-slip faults are dextral in the southern hemisphere and sinistral in the northern hemisphere suggests that tidal forces or differential rotation of the lid may be responsible (Rhoden et al., 2012), and that Europa’s plates are not driven by lithospheric

**Table 2**  
Features of active planetoids (Planets in boldface).

Body	TAI <sup>a</sup>	Outer lithosphere	Tectonic style	Plumes?
<b>Venus</b>	3	Silicate	Stagnant lid	Yes
<b>Earth</b>	3	Silicate	Fragmented lid	Yes
<b>Mars</b>	2	Silicate	Stagnant lid	Yes
Io	2	Silicate	Stagnant lid	Yes
Europa	2	Silicate	Fragmented (ice) lid	Yes
Enceladus	2	Ice (MSIB <sup>b</sup> )	Fragmented (ice) lid	Yes?
Titan	2	Ice & tholin	Stagnant lid	Yes?
Triton	3	Ice (MSIB <sup>b</sup> )	Stagnant lid	Yes?
Pluto	3	Ice (MSIB <sup>b</sup> )	Stagnant lid	Yes?

<sup>a</sup> Tectonic Activity Index.  
<sup>b</sup> Medium-sized Icy Body.

density as found for Earth. Regardless, [Kattenhorn and Prockter \(2014\)](#) inferred a brittle, mobile, plate-like fragmented icy outer layer that rides over a weak, watery zone deeper zone.

Icy Ganymede is another fragmented lid planetoid, its surface riven by a global fault network. Polygons separated by linear grooved lanes, which seem to be sites of focused extension. Grooved lanes are typically 5 to 10 km across, often comprised of parallel blocks of crustal material separated by faults. Faults in the groove lanes truncate and modify features within the polygons, indicating that groove lanes are younger than polygons ([Collins et al., 1998](#)).

Several of Saturn's six major moons show evidence of faulting. Icy Enceladus reveals a protracted history of faulting: extensional, compressional, and strike-slip. Its ridged plains mark a zone of convergence and folding 500 km wide, with a crater-based age of 10 to 100 Ma ([Kargel and Pozio, 1996](#)). Geologic activity in the southern polar region is documented in the form of water vapor and ice particles emanating from fractures known informally as "tiger stripes" ([Nahm and Kattenhorn, 2015](#)). Based on gravity and shape data, [Beuthe et al. \(2016\)](#) concluded that a  $38 \pm 4$  km thick ocean lies beneath a  $23 \pm 4$  km thick icy shell.

[Collins et al. \(2009\)](#) grouped together Dione, Tethys, and Rhea (satellites of Saturn) and Titania (satellite of Uranus) because their surfaces indicate modest lithospheric deformation. The surface of Tethys is cut by normal faults. The main tectonic feature is a graben complex called Ithaca Chasma, which is 2–3 km deep surrounded by a rim up to 6 km high ([Giese et al., 2007](#)). The densely cratered surface of Dione and Rhea are also cut by normal faults ([Collins et al., 2009](#); [Byrne et al., 2015](#)). Gravity and shape data for Dione indicate that a  $99 \pm 23$  km thick icy shell overlies a  $65 \pm 30$  km thick global ocean ([Beuthe et al., 2016](#)).

Titan is the largest satellite and the largest Tholin-rich body in the Solar System, with far more hydrocarbons than found on Earth ([Lorenz et al., 2008](#)). Its soft icy hydrocarbon-rich surface does not preserve faults well but the SAR instrument onboard Cassini returned images that reveal morphotectonic features such as mountains, ridges, and faults. [Solomonidou et al. \(2013\)](#) identified six regions on Titan where they identified faults and fault-like features.

Iapetus exhibits a giant equatorial ridge that may reflect contraction of lithosphere that is thinner at the equator ([Beuthe, 2010](#)). Most striking are several large troughs on the bright, trailing hemisphere. These troughs appear to be extensional but they are clearly ancient ([Singer and McKinnon, 2011](#)).

The five major satellites of Uranus are all icy bodies and show a range of surface features due to deformation of their lids. Miranda's surface is faulted in places and has at least 3 coronae, each of which are over 200 km across and are ovoid or polygonal in shape. Arden corona, the largest, has ridges and troughs with up to 2 km of relief. Explanations for what caused these coronae vary widely, from major impact to tidal flexing to convection in the outer icy shell ([Hammond and Barr, 2014a](#)). On Ariel, several 75–200 km wide, 3–5 km deep linear chasmata, interpreted as grabens and are evidence for faulting ([Schenk, 1991](#)). The surface of Umbriel is not well-imaged but seems to preserve evidence of an ancient tectonic system and thus faults ([Helfenstein et al., 1989](#)). The surface of Titania reveals a branching network of faults and graben, 20–50 km wide and 2–5 km deep ([Collins et al., 2009](#)). The surface of Oberon is not well imaged but shows a graben system ([Croft, 1989](#)).

Neptune's moon Triton has a surface with abundant ridges and graben like those of Europa that are thought to be tectonic in origin ([Prockter et al., 2005](#)).

Pluto and Charon have only recently been imaged. Although faults have not been imaged, Pluto has young, rugged mountains and Charon has rift valleys up to 10 km deep, so lithospheric

deformation is indicated on both. Some of the processes operating on Pluto appear to have operated geologically recently ([Stern et al., 2015](#)). Pluto also shows recent surface polygons that resemble Triton's cantaloupe terrain. Pluto and Charon show different levels of tectonic and magmatic activity. Pluto shows ongoing geological activity centered on a vast basin of thick, active ices (resurfaced  $< \sim 10$  Ma; [Moore et al., 2016](#)). Charon does not appear to be active, but experienced major extensional deformation and cryovolcanic resurfacing  $\sim 4$  billion years ago ([Moore et al., 2016](#)).

#### 4.2. Volcanism

Volcanism reflects asthenospheric melting, including basalt volcanism on silicate planetoids and cryomagma eruptions from deep watery zones (including geysers) on icy bodies. In most cases melting indicates the existence of a silicate asthenosphere or water-rich zone beneath the icy shell. Volcanic eruptions advect internal heat (from the asthenosphere) through weaknesses in the lid. In the case of silicate-dominated planetoids such as the Earth, the advected liquid is eutectic melt of silicate mantle asthenosphere called magma.

Four of the seven silicate bodies either have active volcanism (Earth, Io, Venus) or show signs of recent volcanism (Mars). Thermal infrared imaging of Venus volcanoes by Venus Express has revealed several that appear to be recently active ([Shalygin et al., 2015](#)). In contrast, the ancient surface of Vesta shows that basaltic volcanism happened early in the history of the Solar System, as do the ancient lava flows of Mercury and Earth's Moon. In the case of the ice-dominated planetoids the erupted liquid is mostly water, and such eruptions are called cryovolcanism. For either the case of planetoids with silicate or ice lids, recent volcanism indicates that asthenosphere or watery weak zone underlies the overlying solid lid shallow at shallow-enough depth that melts reach the surface. This is strong evidence of a tectonically active planetoid.

Cryovolcanism has recently been inferred for Ceres ([Ruesch et al., 2016](#)). The ice-covered surfaces of Jupiter's moons Ganymede and Europa show fractures through some of which liquid water may have erupted, although we cannot be sure when ([Fagents, 2003](#); [Bland et al., 2009](#)). Active cryovolcanism is documented for Saturn's moon Enceladus ([Porco et al., 2006](#); [Hansen et al., 2008](#)) and for Jupiter's moon Europa ([Quick et al., 2015](#)). Icy plumes rising as high as 200 km are documented for Enceladus ([Garisto, 2016](#)). Smooth plains on Dione and Tethys may reflect erupted water; evidence of plasma sourced from these bodies suggests active volcanism ([Burch et al., 2007](#)). Uranus' satellites Miranda and Ariel show evidence for ancient cryovolcanism ([Schenk, 1991](#)). Recent studies of Titan's surface combining synthetic aperture radar (SAR) as well as visible and thermal infrared indicate that there are a few active cryovolcanoes ([Lopes et al., 2013](#); [Solomonidou et al., 2013](#)). Triton, Neptune's only large satellite, shows evidence for cryovolcanism including nitrogen geysers ([Duxbury and Brown, 1997](#)). A unique morphological feature on Triton is its "cantaloupe terrain", containing quasicircular shallow depressions typically 25–35 km in diameter, with slightly raised rims, reminiscent of cantaloupe melon skin. Triton's cantaloupe terrain is suggested to represent cryovolcanic explosion craters, such as terrestrial maars, on the basis of their similar morphologies. An alternative mode of formation is by diapirism: the organized cellular pattern of the cantaloupe terrain resembles the surface above some terrestrial salt diapirs, and the terrain may have formed due to density inversion in a layered crust composed partly of ice phases more complex than only water ice ([Schenk and Jackson, 1993](#)). Maybe this is an expression of a mantle plume on an icy planet? The surface of Pluto shows continuing geological activity centered on a vast basin filled with volatile ices with a crater

retention age no greater than  $\sim 10$  million years. Possible cryovolcanic features include mounds with central depressions that appear to be constructional. One such feature is  $\sim 6$  km tall and 225 km across (Moore et al., 2016).

#### 4.3. Impact craters

The density of impact craters provides an indirect assessment of planetary tectonic activity: a tectonically active planetoid will be resurfaced frequently and so will have a surface with few impact craters whereas a dead planetoid will not have been resurfaced for a long time and so will have an ancient surface with many impact craters. It is possible for a planetoid with a dense atmosphere to be tectonically dead but to be resurfaced by liquids and aeolian processes, but that concern only applies to the 2 objects with dense atmospheres, Titan and Venus. Venus lacks surface liquids and its craters do not seem to be eroded by aeolian processes, although Titan shows evidence that its surface is modified by liquid and atmospheric flow. Calculating an age for when a planetoid was last resurfaced requires many assumptions, but for our purposes it is enough to know qualitatively that a planetoid surface has a high or low abundance of impact craters. It should be acknowledged that several planetoids show regions with distinctly different crater densities, for example Mars and most of the moons of Uranus, but our simple classification does not capture these variations. If any part of a planetoid's surface is densely cratered, we list this as "high" Impact Crater Density in calculating TAI.

Mercury has a densely cratered surface (Strom et al., 2008). Venus has  $\sim 900$  identified impact craters, suggesting that it was resurfaced in the past 300–700 Ma, either catastrophically (Schaber et al., 1992; Herrick, 1994; Strom et al., 1994) or by continuous processes (Hansen and Young, 2007; O'Rourke et al., 2014). Earth has 188 documented impact craters < <http://www.passc.net/EarthImpactDatabase/>>. The surface of Earth's Moon is saturated with impact craters. Robbins and Hynek (2012) reported 384,343 impact craters on the surface of Mars with diameters  $> 1$  km. Vesta's surface is characterized by abundant impact craters (Jaumann et al., 2012). The surface of Ceres also appears to be dominated by impact craters.

The Galilean satellites show a range of crater densities, fewer on the inner satellites, more on the outer ones. Io shows no visible impact craters. Europa's surface is lightly cratered (Moore et al., 2001). Ganymede and Callisto are heavily cratered.

Five of the six Saturnian satellites are heavily cratered. Enceladus is densely cratered with different regions showing threefold variations in impact densities (Kirchoff and Schenk, 2009). Tethys, Dione, Rhea, and Iapetus are all heavily cratered (Kirchoff and Schenk, 2010). Titan is more heavily cratered than Earth, but much less than Mars or Ganymede: the area fraction covered by craters is comparable with that of Venus (Neish and Lorenz, 2012).

For the five satellites of Uranus, Miranda is variably but densely cratered (Plescia, 1988). Ariel, Umbriel, Titania, and Oberon also have high impact crater densities (Zahnle et al., 2003). For Neptune's two moons of interest, Proteus is densely cratered (Croft, 1992) whereas Triton's surface has a low density of impact craters (Schenk and Zahnle, 2007).

Some regions of Pluto and most of Charon are heavily cratered, corresponding to ages  $> 4$  Ga. Other places on Pluto, most notably the  $N_2$  ice field Sputnik Planum, have few craters, and therefore appear to be young (Moore et al. 2016).

#### 4.4. Summary of planetoid tectonic activity

To summarize our threefold assessment of tectonic activity on the 26 planetoids with solid surfaces: there are 5 planetoids with

"Tectonic Activity Indexes" of 3: Venus, Earth, Io, Titan and Triton. These are clearly active bodies. There are another 4 planetoids with scores of 2: Mars, Europa, Enceladus, and Pluto. These are probably active bodies. Three bodies (Ganymede, Tethys, and Dione) have scores of 1.5 because we are unsure whether they have active volcanism; we do not think they are tectonically active. Nine of the planetoids have scores of 1; these are probably dead. Four planetoids score zero; these are Earth's Moon, Vesta, Callisto, and Iapetus, and are almost certainly dead. Active planetoids commonly show evidence for extensional deformation, some exhibit strike-slip faulting, and several show evidence for contractional tectonism.

Table 2 summarizes the distinctive features of the nine active planetoids. Five of them have the densities indicating that they are silicate bodies, four are icy bodies. Six are single lid planetoids and three have fragmented lids; Earth has a particular style of fragmented lid known as plate tectonics. Europa has something like plate tectonics, but fundamentally differs because plate motions are not driven by sinking of dense lithosphere in subduction zones. Hoppa et al. (1999) identified 121 strike-slip faults up to 800 km long with up to 42 km of offset on Europa. Europa's youthful surface is produced at extensional zones that at least superficially resemble terrestrial mid-ocean spreading ridges. However there is little evidence of large-scale contraction (destructive plate margins) that is required in plate tectonics to balance the observed extension or to recycle lid fragments. Our interpretation of Europa's tectonics differs from that of Kattenhorn and Prockter (2014), who identified five periods of discrete plate motions. They argued that subduction (followed by melting of the subducted ice, which they term "subsumption") may recycle surface material into Europa's interior; they further suggest that Europa is the only planetoid other than Earth that has something akin to plate tectonics.

### 5. Comparative planetoidology caveats

The fact that we have imaged the 30 largest planetoids encourages comparative geoscientific assessment to elicit insights about planetoid tectonics. On the other hand, the variability in size, composition, and tidal interactions encourages caution in this effort. The four largest of the 30 (Jupiter, Saturn, Uranus, and Neptune) have dense atmospheres and nothing is known about the nature of their interiors, which are nevertheless likely to be rocky and active due to their size. The solid surfaces of the other 26 planetoids have all been imaged and otherwise studied by space missions (Table 1) and these studies allow tectonic assessments. It is unclear how useful comparisons are of the 9 active bodies, which range in mass by more than 4 orders of magnitude. Below we outline several differences between the planetoids and Earth, each of which presents challenges to using especially the low density planetoids as analogues for Earth tectonic styles.

#### 5.1. Likely differences between tectonics of rocky and icy planetoids

There are several reasons that all of the 26 planetoids with imaged solid surfaces are not strictly comparable. First, these planetoids are readily separated into two groups of 8 denser bodies and 18 less-dense bodies (Table 1; Fig. 2). These two density groups correspond to rocky vs. icy compositions that are likely to convect differently. The dense eight must be dominated by silicates and iron whereas the light 18 are icy bodies that may (density  $\sim 2000$  kg/m<sup>3</sup>) or may not (density  $\sim 1000$  kg/m<sup>3</sup>) have rocky and iron-rich interiors. Table 3 summarizes some of ways that these two great types of planetoid material are similar and differ. Earth is one of the 8 silicate-dominated bodies, can we learn anything about its possible past tectonic history by considering the 18 planetoids with ice-dominated surfaces? We undertake this exploration with the

understanding that icy and rocky lithospheres should behave differently. Below we explore some of the likely differences between the 26 solid planetoids.

In comparing these bodies, we should consider both the thermal and rheological definitions of lithosphere. The thermal definition has rheological implications as well because material behaviors reflect homologous temperature ( $T_H$ ), defined as the ratio of a material's absolute temperature to its melting temperature. The behavior of solids changes from brittle to ductile at  $T_H \sim 0.50$  (Frost and Ashby, 1982), thus at much lower temperatures for water ice (melting temperature  $\sim 273$  K) than for silicate rock (melting temperature  $\sim 1500$  K). Although ice is brittle at the cold temperatures found in the outer Solar System, its low  $T_H$  means that the tectonics of icy planetoids may be influenced by a ductile and/or liquid layer (asthenosphere) beneath the brittle thermal boundary layer (lithosphere). This inference is supported by evidence that several of icy planetoids are likely to have subsurface oceans of liquid water. The detection of induced magnetic fields near Jupiter's satellites Europa, Ganymede, and Callisto is one of the most interesting findings of the Galileo mission to Jupiter (Khurana et al., 1998; Zimmer et al., 2000). These magnetic fields cannot be generated in solid ice or in silicate rock. Instead, the presence of these magnetic fields suggest the existence of electrically conducting subsurface "oceans" of liquid water beneath the satellites' outermost icy shells that may contain more water than all terrestrial oceans combined (Hussman et al., 2006). Ceres may also host subsurface pockets or a layer of brine (DeSanctis et al., 2016). These subsurface bodies of water are a kind of asthenosphere beneath the planetoid's icy lithosphere. The presence of watery asthenosphere beneath icy lithosphere allows for surface deformation like that seen on Europa and Enceladus (see Fig. 4).

### 5.2. Double lithosphere and asthenosphere

Some interesting possibilities for tectonic activity are provided by some of the icy planetoids and these possibilities are explored in Fig. 5. Icy planetoids with higher densities are likely to have an icy outer layer surrounding a rocky core (Fig. 5, row 3). In this case it is possible to have an outer icy lithosphere underlain by a watery asthenosphere (Fig. 5, row 3A), both of which could be underlain either by a tectonically alive silicate lithosphere and asthenosphere (Fig. 5, row 3C), or by a tectonically dead silicate body (Fig. 5, row 3B). We cannot tell from the icy surface what is the nature of the underlying silicate body, even whether or not it is tectonically active. It is unclear what useful information can we extract about Earth's likely tectonic styles from such situations.

### 5.3. Gravity fields

The range in mass of the active planetoids varies by more than 4 orders of magnitude and thus so do the gravitational fields. Because all tectonics ultimately reflects buoyancy and thus density,

variations in gravitational fields affect planetoid tectonics. One effect is that topographic relief is easier to maintain on a small, low-gravity planetoid than on a large, high-gravity planetoid; examples of this are Olympus Mons and the great faults scarps of Valles Marineris on Mars, the equatorial ridge on Iapetus, and Verona Rupes on Miranda. Active planetoid tectonics reflects a balance between driving forces such as mantle convection, as expressed by the Rayleigh number (see section E below) and the resistive strength of the lithosphere. Mantle convection simulations show that increasing planetoid gravity increases resisting stresses (O'Neill and Lenardic, 2007), but simulations and scaling arguments indicate that buoyancy forces probably increase faster (Valencia et al., 2007; van Heck and Tackley, 2011), as also indicated by the observation that only the largest planetoid – Earth – has plate tectonics.

### 5.4. Tidal vs. residual vs. radioactive heating

A key way in which the tectonics of smaller planetoids orbiting large bodies (Jupiter, Saturn, Uranus, Neptune) differ from those of larger planetoids reflects the influence of tides. Inner satellites around the 4 largest planets are affected by tides that not only influence their orbital and rotational evolution, but are also a major source of stress and heat. For these bodies, tides provide energy for deformation and maintain asthenospheric weaknesses, leading to tectonic and cryovolcanic activity. On several satellites, the observed tectonic features may reflect tidal flexing and heating or changes in their tidal figures. Io's hyperactivity is due it being the innermost satellite of giant Jupiter combined with its significant orbital eccentricity forced by a Laplace resonance with the satellites Europa and Ganymede, which also creates much tectonic activity on Europa. Enceladus is another good example of a small active planetoid that owes its activity to tidal deformation. It is spectacularly active; its south polar terrain is a site of young tectonism, with several tall icy plumes venting into space. In the case of Neptune's moon Triton, its capture and evolution from highly eccentric to circular orbit is thought to have stressed the lithosphere (Prockter et al., 2005).

Other than tidal activity for inner satellites of large planets, the major difference in planetoid energetics reflects the size, reflecting the slower cooling rate of larger planetoids, so that much of the heat flow from the larger solid bodies is original heat of accretion (Stevenson, 2003). Larger bodies are more likely to be active than smaller planets. Planetoid energy reflects initial heat of accretion and radioactivity for large silicate planets plus tidal energy for the inner satellites of the giant planets.

### 5.5. Rayleigh numbers

In order to further investigate differences between rocky and icy planetoids we evaluated approximate Rayleigh number ( $Ra$ ) for their outermost shell (i.e. for their silicate and icy mantle layer, respectively; Barr and McKinnon, 2007).  $Ra$  is a dimensionless number that reflects the likelihood that a fluid will convect (e.g. Turcotte and Schubert, 2014). When the Rayleigh number is below a critical value for that fluid, heat transfer is primarily conductive; when it exceeds the critical value, heat transfer is primarily in convective. The Rayleigh number is computed as

$$Ra = g x^3 \Delta \rho / (\eta \kappa)$$

where  $g$  is acceleration of gravity at the surface ( $m/s^2$ ),  $x$  is thickness of the mantle layer (m);  $\Delta \rho$  is characteristic density difference driving mantle convection (30 and 10  $kg/m^3$  for rocky and icy mantle respectively (Tackley, 2000; Barr and McKinnon, 2007));

**Table 3**  
Comparison between silicate and ice rheologies.

Properties	Water ice	Silicate	Similar?
Melting T (K)	273	950–1500	No
Density ( $kg/m^3$ )	1000	3000	No
Young's modulus (GPa)	$\sim 10$	$\sim 100$	No
Low strain deformation	Elastic	Elastic	Yes
High strain, low T def.*	Brittle	Brittle	Yes
Low strain, high T def*	Ductile	Ductile	Yes

\*def. = deformation.

Modified after Solomonidou et al. (2013).

$\eta$  is viscosity of the mantle ( $10^{22}$  and  $10^{16}$  Pa s for rocky and icy mantle respectively (Parmentier and Head, 1981; Kaufmann and Lambeck, 2000);  $\kappa = 10^{-6}$  m<sup>2</sup>/s is thermal diffusivity of the mantle. Ra for the 26 solid planetoids are listed in Table 4 and graphically compared in Fig. 6, which shows that the TAI of rocky planetoids positively correlates with their Ra. This correlation could be interpreted in term of the mantle convection vigor: the largest bodies with strong gravity and thick silicate mantle (Earth, Venus) convect more vigorously and therefore have the strongest tectonomagmatic activity at the surface. The temperature dependence of viscosity is important but we cannot address this variable for most of the planetoids. In contrast, icy planetoids do not show a similar trend (Fig. 6) and their surface activity is thus likely regulated by more complex dependencies related to their internal structure and accretion history, and/or by tidal dissipation. Assuming that all bodies of the same type have the same interior viscosity and driving density contrast is clearly a major simplification and leads to the calculated Rayleigh numbers only being rough estimates (e.g. the Moon and Mercury are calculated to be super-critical whereas convection has probably died out; Earth's calculated Ra is lower than normally estimated) but to treat each body separately would require detailed modeling of each's structure and thermal evolution, which is beyond the scope of the present survey.

### 5.6. Other tectonic drivers

Forces other than internal convection can drive planetoid tectonics. Contraction due to cooling (for example Mercury), expansion reflecting interior volume changes due to ice or water phases, and despinning have been common in the evolution of smaller planetoids and satellites. Except for Saturn's moon Hyperion, all

the large satellites of the outer Solar System have despun and are in (or very close to) synchronous rotation, showing the same face toward the planet they orbit. Their characteristic tectonic patterns have thus been sought for on all planets and large satellites with an ancient surface. There is observational evidence on several bodies for the global faulting pattern expected for contraction or expansion, though the pattern is seldom isotropic as predicted (Beuthe, 2010). Other driving mechanisms for tectonics include deformation of the surface above rising diapirs of warm silicates or ice and motion of subsurface material toward large impact basins as they relax. In the case of Venus, radial compressive stress generated outwards from uplifted areas above mantle plumes is also proposed as the origin of thrusts and foldbelts (Harris and Bédard, 2014).

## 6. Discussion

In the following discussion, we use the results from our survey of 26 solid planetoids in our Solar System survey to briefly explore what this tells us about the likely tectonic evolution of Earth and other planetoids: (1) How common is plate tectonics in space, and what are the minimum requirements for it to occur?; (2) How important are mantle plumes and other focused upwellings for active planetoids?; and (3) Recognizing that stagnant lid tectonics is the dominant planetoid style, what is the range of stagnant lid tectonic behaviors?

### 6.1. How common is plate tectonics?

Given the broad range of planetoid size, composition, and heat sources of the 26 large solid planetoids in our Solar System, what if

**Table 4**  
Summary of 30 largest planetoids.

Body	Grav. Accel.	Density (kg/m <sup>3</sup> )	Silicate fraction	Viscosity (Pa s)	Mantle (km)	Ra	tau/t_solar	TAI
Mercury	3.67	5357		$1 \times 10^{22}$	650.0	3022	42.27	1
Venus	8.93	5283		$1 \times 10^{22}$	3050.0	760260	257.75	3
Earth	9.93	5594		$1 \times 10^{22}$	2885.0	715185	283.94	3
Moon	1.59	3252		$1 \times 10^{22}$	1420.0	13661	21.57	0
Mars	3.72	3924		$1 \times 10^{22}$	1689.5	53773	80.90	2
Ceres	0.27	2069	0.46	$1 \times 10^{16}$	105.9	322652	1.56	1
Vesta	0.25	3432		$1 \times 10^{22}$	262.5	14	0.49	0
Jupiter	28.24	1508		$1 \times 10^{16}$	67000.0			
Io	1.83	3643		$1 \times 10^{22}$	1200.0	9501	22.82	3
Europa	1.33	3077	0.90	$1 \times 10^{16}$	100.0	1333011	16.92	2
Ganymede	1.43	1924	0.40	$1 \times 10^{16}$	800.0	729668352	49.45	1.5
Callisto	1.27	1900	0.39	$1 \times 10^{16}$	275.0	26498685	40.56	0
Saturn	11.31	697	0.00	$1 \times 10^{16}$	58000.0			
Enceladus	0.12	1641	0.28	$1 \times 10^{16}$	50.0	14446	0.45	2
Tethys	0.14	890	0.00	$1 \times 10^{16}$	550.0	22751520	2.13	1.5
Dione	0.24	1578	0.25	$1 \times 10^{16}$	202.8	2024837	2.13	1.5
Rhea	0.27	1302	0.13	$1 \times 10^{16}$	369.0	13706385	3.96	1
Titan	1.28	1766	0.33	$1 \times 10^{16}$	798.0	651927451	47.60	3
Iapetus	0.21	1019	0.01	$1 \times 10^{16}$	599.5	45997305	3.96	0
Uranus	8.93	1253	0.11	$1 \times 10^{16}$	13288.4			
Miranda	0.08	1214	0.09	$1 \times 10^{16}$	128.5	169180	0.39	1
Ariel	0.26	1547	0.24	$1 \times 10^{16}$	228.2	3082757	2.54	1
Umbriel	0.22	1326	0.14	$1 \times 10^{16}$	287.1	5261499	2.54	1
Titania	0.36	1632	0.27	$1 \times 10^{16}$	279.9	8002148	4.51	1
Oberon	0.36	1698	0.30	$1 \times 10^{16}$	246.1	5302276	3.96	1
Neptune	11.12	1623	0.27	$1 \times 10^{16}$	8644.9			
Proteus	0.07	1134	0.06	$1 \times 10^{16}$	128.5	141381	0.31	1
Triton	0.77	2038	0.45	$1 \times 10^{16}$	314.6	23937881	12.83	3
Pluto	0.60	1796	0.35	$1 \times 10^{16}$	357.5	27516605	10.14	2
Charon	0.28	1658	0.29	$1 \times 10^{16}$	204.7	2383640	2.54	1.5

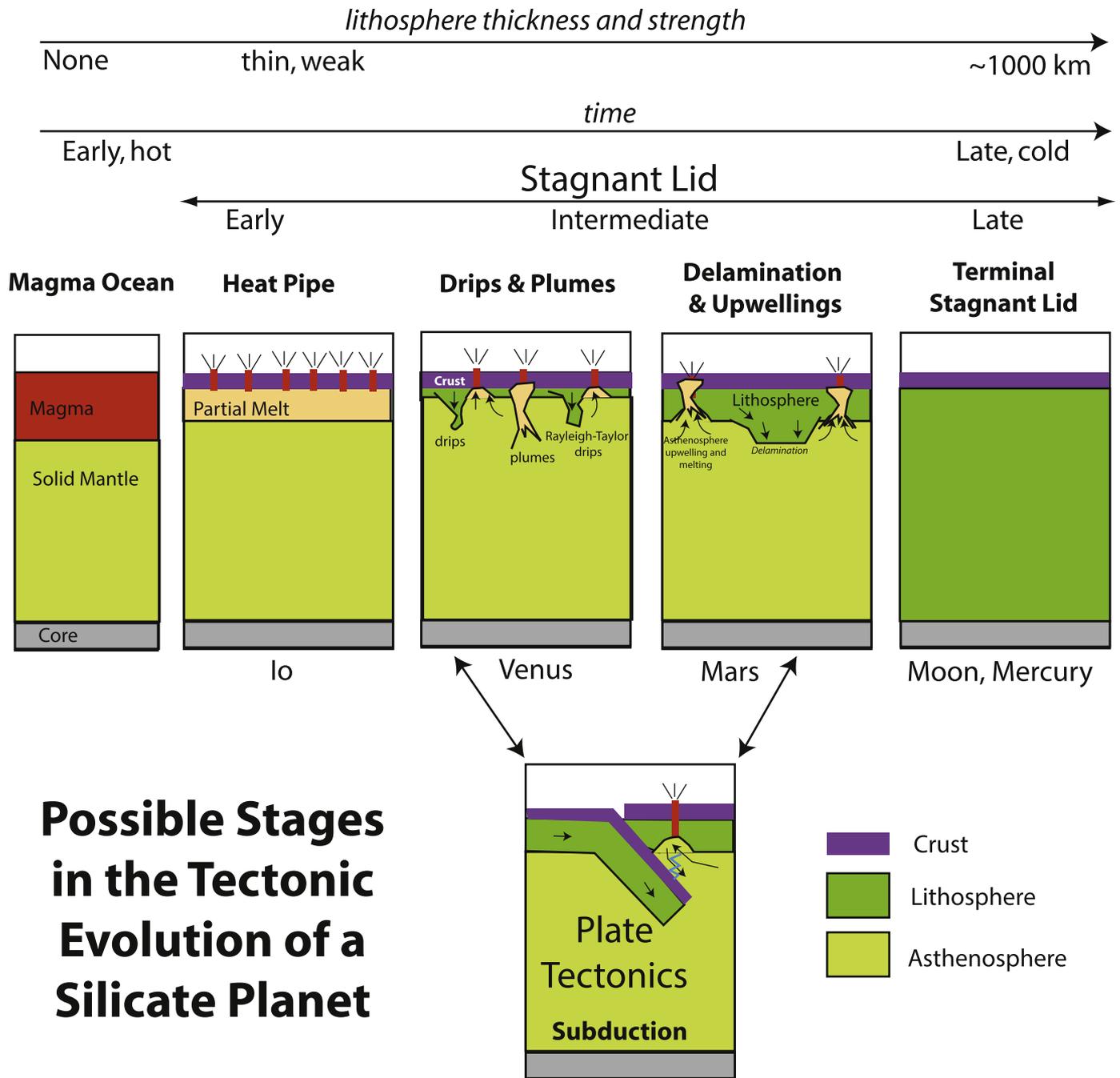
Grav. Accel. = gravitational acceleration (m/sec<sup>2</sup>)

Silicate fraction for icy planets only based on assumption of constant density of silicate (3300 kg/m<sup>3</sup>) and ice (1000 kg/m<sup>3</sup>).

Mantle thickness from Wikipedia or computed on the basis of the silicate fraction.

Ra = Rayleigh number.

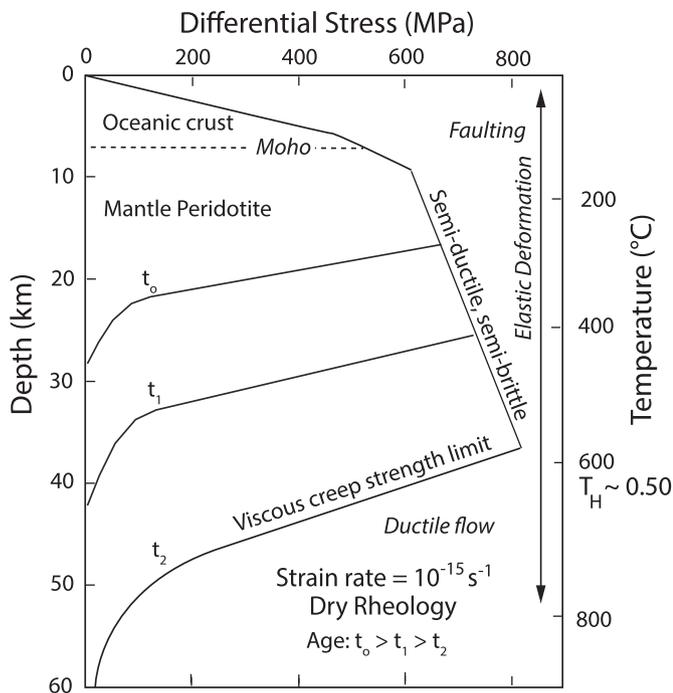
tau/t\_solar = ratio of characteristic conductive cooling time scale of the planet and solar system age (4.56 Ga); note convection is ignored.



**Figure 3.** Possible evolution of magmatotectonic styles for a large silicate body, like the Earth. Plate tectonics requires certain conditions of lithospheric density and strength to evolve and is likely to be presaged and followed by stagnant lid tectonics.

anything can we deduce about the likelihood of plate tectonics vs. stagnant lid tectonics occurring on a solid body? Plate tectonics is a style of convection for active planetoids with a fragmented silicate lid, where lithospheric fragments move independently on underlying weak asthenosphere, and where lid fragment (plate) motions reflect sinking of dense lithosphere in subduction zones and upwelling of asthenosphere beneath spreading ridges (Fig. 7A and B). Plate tectonics can only occur on an active body. For a plate tectonic regime to exist, stresses associated with mantle convection must exceed the strength of the lithosphere sufficiently to allow plate rupture but not so much stronger that the plate breaks up into drips or so that the bottom of the lithosphere separates from the top of

the lithosphere (delaminates; Gerya et al., 2015). These conditions are sufficiently restrictive that plate tectonics currently operates only on Earth, and mantle convection in most terrestrial planets and moons is probably in a stagnant lid regime, where the entire lithosphere of the planet is a single, globe-encircling plate. Stagnant lid behavior can occur on planetoids that are tectonically active or dead. We reaffirm from our survey that plate tectonics is unusual and that stagnant lid tectonics is the dominant mode; only one in 26 bodies (4%) show plate tectonic behavior, with 96% of the solid bodies showing either dead or active stagnant lid behavior. If we restrict consideration to active planetoids, then the frequency of plate tectonics increases slightly to one in 9 (11%). If we further



**Figure 4.** Evolution of lid strength over the life of a cooling silicate planetoid, beginning ( $t_0$ ) using 60 Ma oceanic lithosphere of Earth as an example (modified after Kohlstedt et al., 1995; Tackley, 2000). Any differential stress to the left of the strength envelope results in recoverable (elastic) deformation. The solid line on the upper part of the strength envelope shows where strength is limited by frictional sliding following Byerlee's Law (Byerlee, 1968) whereby strength increases with lithostatic pressure (depth). Differential stress greater than this results in brittle failure (faulting). The solid line in the bottom part of the figure shows where strength is limited by temperature-dependent viscous creep, and the dotted line connecting these represents semiductile, semibrittle flow. These different behaviors reflect the homologous temperature ( $T_H$ )  $\sim 0.50$ , which for mantle peridotite occurs  $\sim 600$  °C. Ice has a strength profile with a similar shape but because it melts at much lower  $T$ , the brittle-ductile transition happens at much lower  $T$ . Cooling silicate body would experience overall deepening of 600 °C.

restrict consideration to active planetoids with density of  $3 \text{ g/cm}^3$  or greater, we see that 1 in 5 (20%) show plate tectonics and 80% show some variant of active stagnant lid behavior, from the vigorous heat-pipe tectonics of Io to the sluggish plume-and-rift tectonics of Mars. The inescapable conclusion is that stagnant lid tectonics is the dominant mode of heat loss from planetoids and that plate tectonics is very unusual, restricted to a minority of active silicate bodies.

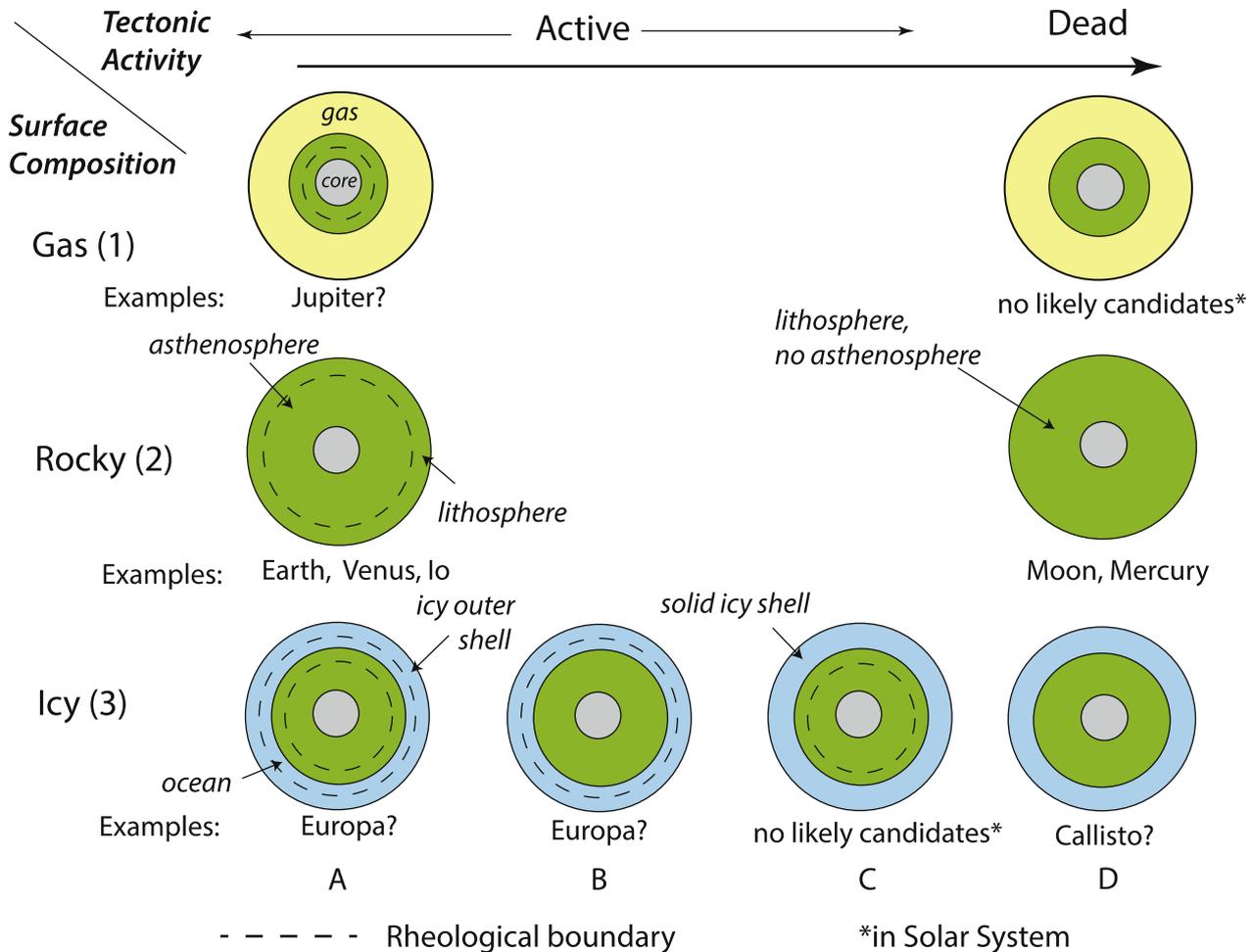
These observations beg the question: what are the minimum requirements for plate tectonics to occur? Clearly, the planetoid must be active, it must have tectonic and convective vigor. Any planetoid can be active; although size is important in retaining original heat, small inner satellites of large planets – for example, Jupiter's innermost satellite Io – can be heated by tidal flexing. Although Earth is an internally heated body, reflecting a combination of primordial and radiogenic heat – there is no obvious reason why a tidally heated silicate satellite couldn't also have plate tectonics. Beyond the activity requirement, there are three additional conditions required for plate tectonics to occur: (1) it must have a thick outer skin, or lithosphere, above a weak underlayer, or asthenosphere; (2) the lithosphere must be strong enough to remain coherent during subduction and plate motion (e.g., van Hunen and van den Berg, 2008; Crameri and Tackley, 2015), but weak enough to be broken into multiple plates; and (3) lithosphere and asthenosphere must be composed of materials that are denser when colder so that lithosphere can evolve from

buoyant (relative to asthenosphere) when young to dense and gravitationally unstable when it is old. This buoyancy reversal is required to allow lithosphere to sink in subduction zones. Details of an active planetoid's lithosphere-asthenosphere boundary and will vary depending on composition and gravitation field, which control phase relationships at depth. Water also plays a key role in weakening silicate asthenosphere and lubricating the plate interface in subduction zones (Regenauer-Lieb, 2006; Gerya et al., 2008; Crameri et al., 2012; Dymkova and Gerya, 2013). The absence of surface water (for example on Mars) does not preclude water being present in minerals in the lithosphere or asthenosphere.

Icy planetoids cannot experience plate tectonics (at least in their outer icy shells because requirements 2 and 3 above are not satisfied, although plate tectonics may occur in their rocky interior (Fig. 5). Although icy shells are likely to be underlain by weaker material and icy shells readily fragment, plates of water ice are less dense than the underlying water and cannot sink. Other ices ( $\text{CO}_2$ ,  $\text{CH}_4$ ,  $\text{NH}_3$ , tholins) perhaps may be denser than liquids of the underlying weak zone but they are not strong enough to remain coherent during sinking to pull a surface plate. For example, Jupiter's moon Europa has dilational bands that resemble terrestrial mid-ocean spreading zones, evidence for strike-slip faulting, as well as evidence for the removal of  $\sim 20,000 \text{ km}^2$  of the surface. Kattenhorn and Prockter (2014) interpreted some zones as subduction-like convergent plate boundaries that abruptly truncate older geological features and is flanked by potential cryolavas on the overriding ice. Kattenhorn and Prockter (2014) proposed that Europa's ice shell has a brittle, mobile, plate-like system above convecting warmer ice. Although this is grossly similar to plate tectonics, this is a form of ice convection, not plate tectonics, because it involves materials that are not strong and/or dense enough to pull the plates. This is "ice floe tectonics", driven by underlying convection currents (Fig. 7C–F), not by strong dense oceanic lithosphere as required by plate tectonics. Alternatively, Europa has a high enough density ( $3.0 \text{ g/cm}^3$ ) that it appears to be a mostly silicate body with an icy lid (Fig. 5), and similar surface deformation features described above could manifest true plate tectonics of the underlying silicate core. We emphasize that only silicate bodies can have plate tectonics.

Earth is presently in a "tectonic Goldilocks" situation favoring plate tectonics, where oceanic lithosphere is dense enough to sink in subduction zones but strong enough to remain coherent during subduction and plate motion; at the same time, it is weak enough to form new plates (especially during continental rifting) and new subduction zones. How long can conditions conducive to the plate tectonics exist? We do not know, but this question lies at the heart of the controversy about when plate tectonics began on Earth (Korenaga, 2013) and the related question, when will plate tectonics here stop and return to stagnant lid mode (Sleep, 2000)? Lithospheric density and strength are certain to change over the life of a tectonically active planet, from thin, weak, buoyant lithosphere early to increasingly thick, strong, dense lithosphere later (Fig. 3); changes in convective style are also likely. As a result of cooling and progressive lithospheric thickening, a silicate planet will experience several magmatotectonic styles (Fig. 3), among which plate tectonics is only one of several possibilities.

We emphasize that the scenario depicted in Fig. 3 is oversimplified and largely speculative, although it is based on observations of other silicate bodies in our Solar System, including five tectonically active ones (Venus, Earth, Mars, Io, and Europa) and two that are tectonically dead (Mercury and Earth's moon). We emphasize that there is no scientific consensus on Earth's tectonic regimes through time, and that Earth (and by implication, other tectonically active silicate bodies) may have experienced episodes



**Figure 5.** Simplified compositional and tectonic summary for the 30 large planetoids of the solar system discussed here. Three compositional types are identified, based on the nature of the innermost visible layer: gas, rocky material, or ice. Tectonically active planetoids require one or more strong outer layers (lithosphere or ice shell; above dashed line) underlain by an equal number of weak zones (asthenosphere or ocean; below dashed lines).

when plate tectonic and stagnant lid episodes alternated (Sleep, 2000; Van Hunen and Moyen, 2012).

### 6.2. How important are mantle plumes and other focused upwellings (FUs) for active planetoids?

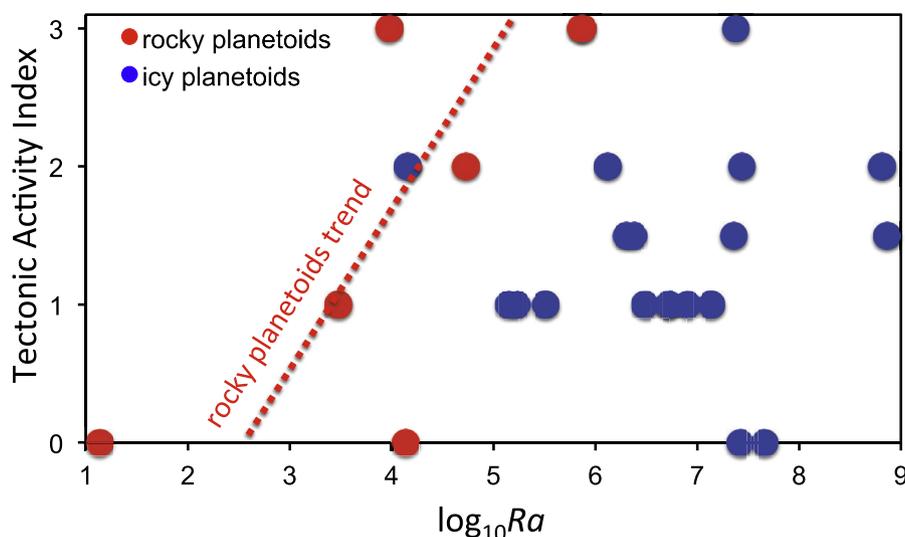
Mantle plumes or other kinds of focused upwellings (FUs) are common features of tectonically active planetoids. Downwellings must also exist, in order to conserve mass. On Earth, FUs have a wide range of scales, from hemispheric (the African and Pacific Superswells; French and Romanowicz, 2015) to 'petit spots' of a few square kilometers (Hirano et al., 2001; Table 2). Even for our relatively well-studied home planet, it is controversial which if any of these come mantle upwellings originate at the core-mantle boundary and which have shallower upper mantle sources (Foulger, 2011), although improved seismic imaging is yielding increasing evidence that many plumes are rooted at the base of Earth's mantle (Montelli et al., 2004; French and Romanowicz, 2015). Our discussion is consistent with either interpretation.

Mantle plumes are easiest to recognize on the rocky planetoids (Earth, Venus and Mars), where they leave prominent topographic, tectonic and magmatic imprints on the surface (e.g., White and McKenzie, 1995; Grindrod and Hoogenboom, 2006; Keller and Tackley, 2009 and references therein; Zhong, 2009; Golabek et al., 2011; Burov and Gerya, 2014; Gerya, 2014; Gerya et al., 2015).

Among the silicate planetoids, FUs range from closely-spaced Ionian heat pipe mode to Venusian coronae and arachnoids (Gerya, 2014) to Earth multiscale intraplate volcanism to the Martian supervolcanoes like Olympus Mons, which are themselves embedded into the much larger plume-related structure of the Tharsis rise. Except for eruptions from rifts, which never evolve to true seafloor spreading on SL bodies, FUs are perhaps the only way to transport molten material – magma or water – to the surface on bodies characterized by active stagnant lid tectonics.

It is not surprising that FUs are found on all active silicate planetoids. Plumes are likely to form on any planetoid with vigorous mantle convection that includes a thermal boundary layer at its base and/or at an internal interface, because this creates significant lateral temperature and buoyancy variations and large stresses at various depths and locations beneath the lid. From this prospective, heat-pipe regime of the surface of Io and vapor jets on Enceladus may directly reflect vigorous convection and plume activity in the planetoids' interiors heated by radioactive decay and tidal dissipation (e.g., Běhounková et al., 2010 and references therein; Moore and Webb, 2013).

Understanding the causes and recognizing the surface manifestations of FUs and mantle plumes is relatively straightforward for the active silicate planetoids, but not for the icy planetoids. First of all, it is more difficult to understand how convection in an active icy planetoid might generate plumes, and how to recognize their



**Figure 6.** Relationship between the planetary Tectonic Activity Index and Rayleigh number (Ra) for rocky (silicate) and icy planetoids.

products on the surface. Conditions on icy planetoids should be favorable for the onset of convection and plumes activity since viscosity of ice is much lower than that of silicate rocks and consequently ice shells of these planetoids should be potentially in the supercritical state ( $Ra > \sim 1000$ , Turcotte and Schubert, 2014).

Many icy moons have a subsurface liquid water layer (Fig. 5; “Icy” A, B). As water has a very low viscosity compared to that of ice or rock, vigorous convection is expected, but with lateral temperature variations that are very small compared to those that occur in solid state convection. Thus, upwelling plumes in liquid water layer convection are unlikely to have a recognizable effect on an overlying solid ice layer. An exception might be if the liquid water is underlain by a silicate layer and volcanic eruptions at the top of the silicate layer create hot jets of liquid water (Fig. 5; “Icy” A).

### 6.3. Stagnant lid tectonics

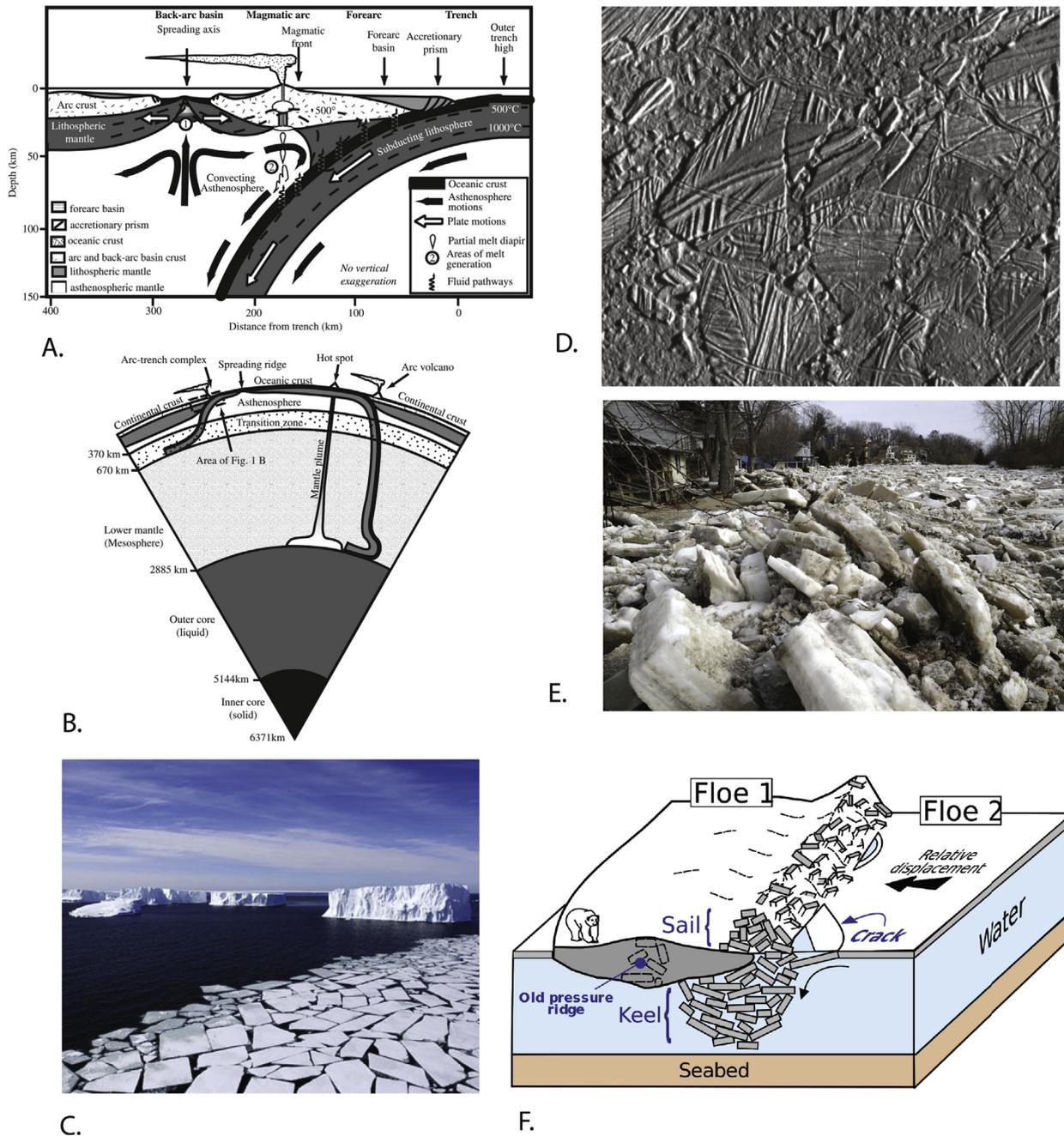
Recognizing that plate tectonics is a very unusual style of planetoid tectonics and convection and that the dominant mode of planetoid behavior is stagnant lid (SL) impels us to better understand the range of SL behaviors in the Solar System. Better understanding of SL styles is especially important because this range is poorly appreciated by the geologic community, inhibiting consideration of magmatic and deformation styles that could have characterized Earth prior to the establishment of plate tectonics. Some of this confusion results from the fact that both active and dead planetoids are characterized by SL. There is “Active” SL as shown in Figs. 3 and 5 (“Rocky”, A) which, as we discuss below, likely involves heat pipe magmatism, drips and plumes, and delamination and upwellings and there is “Dead” SL, shown as terminal SL in Fig. 3. Below we briefly examine the range of likely SL styles for silicate ( $>3 \text{ g/cm}^3$ ) planetoids.

There is a wide range of likely SL behaviors for active silicate bodies, from the present heat pipe style of Io (O’Reilly and Davies, 1981; Moore and Webb, 2013) to the corona-dominated style of upwellings that mark the Venusian surface (Reese et al., 1999; Grindrod and Hoogenboom, 2006) to the sluggish Martian style, dominated by one or a few plumes. SL tectonic style is likely to change as the body ages and cools and mantle lithosphere thickens. Dead silicate bodies also are characterized by a type of SL, for example Mercury or Earth’s moon.

Stevenson (2003) cautioned that our understanding of mantle convection is limited by three key unsolved problems: complexities

of rheology, the effects of compositional gradients, and the effects of phase transitions. This limits our ability to predict how a silicate body might evolve and so far the variety of geologic bodies in our Solar System has not led to the development of a simple evolutionary pathway. However, we can predict that lithosphere is likely to thicken with time, and this is the key first-order constraint for understanding silicate planetoid evolution. In the following we suggest that we can use what we know about the 5 large active silicate bodies (Venus, Earth, Mars, Io, and Europa) in our Solar System to develop such an evolutionary path.

Silicate planetoids start hot due to intense early heating as a result of accretion, differentiation, impacts, and radioactivity. Just-accreted silicate bodies of a large enough size likely had short-lived magma oceans, although magma oceans might exist for only a few million years before the magma ocean mostly solidifies and crust is able to form to provide a solid planetary surface (Abe, 1997; Solomatov, 2007; Elkins-Tanton, 2011). After the early magma ocean crusts over, heat pipe tectonics – the most unstable SL style – may begin (Fig. 3). Heat-pipe tectonics is magnificently expressed on Io. Heat pipes are vertical conduits through which magma traverses the lithosphere to the surface by means of buoyant ascent of magma and subsidence of cold intervening lithosphere, which is composed of older lava flows (O’Reilly and Davies, 1981). Such a scenario might have characterized Earth’s tectonomagmatic regime during Hadean time (Moore and Webb, 2013). As volcanic crust thickens, massive basalt accumulation may be thick enough or be dragged down deep enough ( $\sim 40 \text{ km}$ ) to form dense eclogite, which might limit crustal thickness (O’Rourke and Korenaga, 2012), most likely by the formation of Rayleigh–Taylor “drips” (Fig. 3) (e.g., van Thienen et al., 2004; Johnson et al., 2014). Downwellings would likely be accompanied by abundant igneous activity over complementary regions of magma upwelling. In the pure heat pipe mode, this igneous activity is purely extrusive, resulting in a relatively cold, strong crust (O’Reilly and Davies, 1981). Most likely, however, much of the magmatism is intrusive, which results in a warm, weak, deformable crust, from which delamination and dripping can more readily take place (Sizova et al., 2015; Fischer and Gerya, 2016). As youthful lithosphere cools and thickens, convective drips and delaminates broaden and become more 3-D, flanked by regions of mantle upwellings or plumes. Thicker lithosphere at this stage allows larger scale delamination. We emphasize that considerable crustal and lithospheric deformation can take place in a “stagnant lid” tectonic



**Figure 7.** Silicate planet plate tectonics (A, B) vs. ice floe tectonics (C–F). (A) Schematic section through the upper 140 km of a subduction zone, showing the principal crustal and upper mantle components and their interactions. Note that the location of the “mantle wedge” (unlabeled) is that part of the mantle beneath the overriding plate and between the trench and the most distal part of the arc where subduction-related igneous or fluid activity is found. MF stands for magmatic front (Stern, 2002). (B) Schematic section through the center of the Earth, which shows better the scale of subduction zones. Subducted lithosphere is shown both penetrating the 660 km discontinuity (right) and stagnating above the discontinuity (left). A mantle plume is shown ascending from the site of an ancient subducted slab. Dashed box shows the approximate dimensions of the shallow subduction. (C) Zone of Fig. 1 b (Stern). Weddell Sea, Antarctica <http://www.cbsnews.com/news/antarctic-ice-shelf-thinning-rapidly-in-last-decade/>. (D) Conamara Chaos, Europa. This landscape was produced by the repeated disruption of the icy crust of Europa. Conamara Chaos contains two main terrain types: fragmented and dislocated ice blocks up to 20 km across with 100–200 m high margins set in a matrix of finer ice lying at lower elevations (Carr et al., 1998). (E) Ice compression, Vermillion River, Michigan USA Jan. 2014. <http://chronicle.northcoastnow.com/2014/02/24/vermillion-river-jam-goes/>. (F) Hypothetical interaction between two floes, resulting in a pressure ridge — a linear pile-up of sea ice (Wikipedia “Pressure Ridge”).

mode, as evidenced for example by tessera terrain and transcurent shear zones on Venus (Hansen et al., 2000; Harris and Bédard, 2014); an icy body example is the grooved terrain on Ganymede (Hammond and Barr, 2014b). A similar episode of vigorously active SL may have existed on the pre-plate tectonic Earth. Ultimately a Venusian-like stagnant lid may evolve, dominated by many active mantle plumes and volcanic resurfacings that are either periodic

and catastrophic (300–700 Ma timescales; Solomatov and Moresi, 1996) or quasi-continuous (Hansen and Young, 2007; O'Rourke et al., 2014). Mantle plume interactions are likely to deform and inject magma into the lithosphere, forming coronae and arachnids (Grindrod and Hoogenboom, 2006; Gerya, 2014). If lithospheric stresses associated with delamination rupture the entire lithosphere — perhaps in association with a large mantle plume head

(Gerya et al., 2015) – a transition to plate tectonics becomes possible, if not likely. Also as the lithosphere thickened, larger mantle upwellings are required to transit the lithosphere and erupt. The present tectonic style of Mars represents an almost completely stabilized stagnant lid. Here, a single upwelling mantle plume, pouring lava out at the same sites for hundreds of millions of years, produced the immense Tharsis rise and volcanoes (Zuber, 2001). When these last few mantle plumes are extinguished, tectonic style transitions to the final SL stage – planetoid death. This stage is typified by Mercury and Earth's moon, bodies with ultra-thick, ultra-stable lithospheres, which in the Moon's case has experienced no igneous activity for 2 Ga (Hiesinger and Head, 2006), although a recent study found evidence for lunar volcanism in the last 100 Ma (Braden et al., 2014). Thus the spectrum of stagnant lid from heat pipe to ultra-stable encompasses the evolution of a planet from vigorous infancy to tectonomagmatic death.

## 7. Conclusions

There are 30 large (>300 km diameter) bodies in our Solar System (excluding our Sun), 4 of which are giant planets with presently unobservable solid surfaces. The solid surfaces of the other 26 bodies have been examined and – in tandem with other information such as density – allow comparative analysis for the purpose of understanding the range of their behaviors, encompassing icy bodies ( $\rho = 1000\text{--}2100\text{ kg/m}^3$ ) to rocky, silicate bodies ( $\rho = 3000\text{--}5500\text{ kg/m}^3$ ). It is difficult to compare ice-dominated and silica-dominated planetoids, but is required in order to consider the range of planetoid behaviors. We define the tectonic activity index, scaled for 0 to 3 to distinguish active from dead planetoids. Nine planetoids have TAI = 2 or greater and are classified as active; the other 17 have TAI <2 and are classified as dead. We define plate tectonics as a style of convection for active planetoids where lid fragment (plate) motions are mostly due to sinking of dense lithosphere in subduction zones, causing upwelling of asthenosphere at divergent plate boundaries and seafloor spreading accompanied by focused upwellings (mantle plumes). So defined, Earth is the only planetoid with plate tectonics; all others have stagnant lid (23 planetoids) or non-plate tectonic fragmented lid (2 planetoids). Dead planetoids all have stagnant lid, and excluding these reveals that the 9 active planetoids include 6 with stagnant lid, two with non-plate tectonic fragmented lid, and one with plate tectonics. Stagnant lid and non-plate tectonic fragmented lid is the dominant mode of planetoid behavior but these modes vary tremendously, from the vigorous heat pipe style of Io to the sluggish plume-and-rift style of Mars. Earth likely experienced episodes of stagnant lid and/or non-plate tectonic fragmented lid behavior before the evolution of plate tectonics and we should keep the complexity that we see in the Solar System in mind as we try to reconstruct Earth's pre-plate tectonic behavior.

## Acknowledgements

We appreciate the comments and criticisms of three anonymous reviewers. This study was made possible by two research visits of the first author to ETH, the first of which was supported by SNSF grant IZK0Z-2\_154380 and the second of which was partly supported by SNF 200021\_149252. This is UTD Geosciences contribution # 1297.

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