A comparative study on the volcanism and tectonics of the terrestrial planets and satellites

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Abstract

Origin of diversity of the volcanic activities of the terrestrial planets and icy satellites around the outer gaseous planets is discussed in view of energy sources and heat removal processes. Although available energy increases with planetary radius, dependence of thermal activities on the body size is not clear for the Saturnian and Uranian satellites. Tidal heating mechanisms should have been effective for some of the satellites, such as Enceladus, Ariel, and Triton.

The amounts of tectonic stresses induced by mantle convection are estimated for the terrestrial planets and some icy satellites, by applying the thermal boundary layer theory. Absence of compressive deformation features on the icy satellites may be due to the low yield strength and viscosity of water ice or due to the predominant extensional stress fields associated with internal structural evolution. The relations between planetary dimension and shapes and the maximum topographic reliefs are discussed in view of their thermal activities.

Key words : tidal heating, tectonic stress, mantle convection

Introduction

Images of planetary surfaces accumulated by a number of space probes have revealed that there is a wide variation in the volcanic activities and tectonic styles among the terrestrial planets (e.g., Carr et al., 1976; Carr, 1981; Head et al., 1991, Solomon et al., 1991).

On the earth, there are a number of active volcanoes situated along the plate boundaries as well as inside the plates exhibiting distinct styles of activity. These features are highly reflected their tectonic setting and peculiar magma composition. They can be regarded to be a consequence of plate tectonic processes operated in the outer portion of the Earth.

If one turns to observe surface features of the nearby planets, it can be easily found that plate tectonic processes are unique to the Earth and that the volcanic activities and tectonic styles of the other bodies are quite different from those of the Earth (e.g., Head and Solomon, 1981). Differences in volcanic activity and tectonic style are significant even for those regarded to be twin bodies, namely the Earth and Venus (e.g., Kaula, 1990; Kawakami et al., 1990). What caused such differences is one of the significant problems in the planetary evolution studies.

On the other hand, the Voyager encounters with

Jupiter, Saturn, Uranus, and Neptune have revealed surface features of many icy satellites (Smith et al., 1979 a, b, 1981, 1982, 1986, 1989). The Galilean satellites should be regarded to be terrestrial planets, not only because of their physical dimensions, but of their wide variety of geological features comparable to the terrestrial planets (Soderblom, 1980). The satellites of Saturn, Uranus, and Neptune also show quite interesting surface features resulting from volcanism and tectonics quite different from those of the Earth (e.g., Soderblom and Johnson, 1982; Croft and Soderblom, 1991, Smith et al., 1989). Since there exists active volcanism on Io and Triton as well as volcanic landforms on rather small icy satellites, it is quite important to study energetics and tectonic stress states of these bodies in view of comparative planetology. In order to do this, simple models are useful to characterize volcanic and tectonic activity of objects having a wide variation of physical, chemical, and environmental conditions.

In the present paper, in order to discuss the origin of the diversity of volcanic activities and tectonic styles, energy sources and heat removal processes in the planetary evolution is considered. Vigor of mantle convection and tectonic stresses induced by convective motion are estimated using the thermal boundary layer model described by Turcotte and Schubert

Table 1. Energy sources for planetary thermal activities.

planetary body	Accretional energy per unit mass (MJ/kg)	Total accretional energy (10^{30} J)	Temperature increase (K)	Radioactive heating (10^{11} W)	Tidal heating (10^{11} W)
Earth	37.5	224	31,000	370	-
Venus	32.2	157	27,000	300	-
Mars	7.57	4.88	6,300	40	-
Mercury	5.42	1.79	4,500	15	-
Moon	1.69	0.12	1,400	6.8	-
Io	1.97	0.18	1,600	4.9	17
Ganymede	2.27	0.34	1,300	5	-
Callisto	1.79	0.19	1,000	3	-
Enceladus	0.01	8×10^{-7}	10	10^3	?
Triton	0.18	4×10^{-3}	100	0.8	0.01

(1982). Some remarks will be given for the differences of volcanic activities and tectonic styles among the solid bodies in the solar system. Detailed comparison between the Earth and Venus will be given by Fujii (this issue).

Energetics of planetary thermal evolution

The energy sources which drive volcanism on planetary bodies are gravitational energy of accretion, energy released by radioactive elements such as U, Th, K, and possibly Al, and energy released by tidal dissipation.

(1) Gravitational energy of accretion

Since the accretional theory of planet formation (e.g., Safronov, 1972) became popular in the 1960s, the gravitational energy of accretion has been regarded to be one of the efficient energy of planetary heating (e.g., Hanks and Anderson, 1969). Although it has been regarded that the Moon is too small to be molten, the lunar samples returned to the Earth revealed that the interior of the Moon was extensively molten early in its history. The concept of hot origin of the Moon resulted in the magma ocean hypothesis (see, Taylor, 1982). Mizutani et al. (1972) suggested that the gravitational energy of accretion is enough for heating the lunar upper mantle if the time scale of lunar formation is short.

The accretional energy deposited in the planetary body averaged over the total mass is given by

$$E = h \frac{5}{2} GM / R = h \frac{20}{3} \pi \rho R^2 \quad (1)$$

where h is the fraction of energy deposited in the planetary interior, G is the gravitational constant, R is planetary radius, and ρ is the density. The accretional energy is proportional to the square of the radius of planetary body. Assuming that all the accretional energy is deposited to the interior ($h=1.0$), the value of E and associated temperature increase for each planetary body are given in Table 1.

Since the accretional energy is released by the

impact of planetesimals to the planetary embryo, the energy is roughly proportional to the impact velocity, which is rms of random velocity of planetesimals and escape velocity of the planetary embryo. This means that impact velocity increases with the size of planetary embryo. According to the energy balance consideration (e.g., Kaula, 1980), post-accretional temperature profile ($T(r)$) can be expressed by

$$T(r) = h GM(r)/C_p r [1 + ru^2/2 Gm(r)] + T_e \quad (2)$$

$$\approx 4 \pi \rho h r^2 / C_p$$

where r is the radial distance from the planet center, $m(r)$ is the mass within the radius r , u is the random velocity of planetesimals, C_p is the heat capacity, and T_e is the ambient temperature. Equation (2) suggests that the internal temperature approximately increases with the square of radial distance from the center. Detailed numerical modeling of post-accretional temperature profiles (e.g., Squyres et al., 1988) confirmed that accretional heating for icy Saturnian and Uranian satellites is insufficient to melt the interiors, unless significant amounts of volatile phases having low melting temperature (such as NH_3 or CH_4) were present.

(2) Radioactive elements

The energy released by radioactive elements has been regarded as a major energy source of planetary heating since the discovery of radioactivity. The amount of energy released by the radioactive elements is proportional to their abundance and given by

$$Q = C_U (q_U + \alpha q_{Th} + \beta \gamma q_K) \quad (3)$$

where C_U is the abundance of U, α is the abundance ratio of Th to U, C_{Th}/C_U , β is the C_K/C_U ratio, γ is 1.19×10^{-4} , and q_U , q_{Th} , q_K are the energy release rates from the unit mass. The chondritic abundance of $C_U = 12$ ppb, $\alpha = 3.3$, and $\beta = 6.8 \times 10^4$ gives $Q = 1.59 \times 10^{-7}$ J / (g · year) (Mizutani and Watanabe, 1977). In the case of icy satellites, the amount of radioactive elements in the constituent ice is not known. If the amount is

Fig. 1. A model of Io's lithosphere (Kawakami and Mizutani, 1987; Kawakami, 1989b). Molten layer beneath the lithosphere is maintained by tidal heating. The same amount of magma flowing over the surface is supplied at the top of molten layer by remelting of crustal materials.

negligible, the energy is less significant for those with smaller silicate core. The energy released by the decay of radioactive elements for the specific body is given in Table 1.

(3) Tidal heating

Just before the Voyager 1 encounter with the Jovian system, tidal dissipation has been postulated to be important energy source which heats up the inner most Galilean satellite Io (Peale et al., 1979). The previous estimations of tidal dissipation energy for the Earth and Moon are too small to cause significant effects for their thermal evolution (e.g., Kaula, 1964; Peale and Cassen, 1978). However, satellites orbiting quite near the heavy primary has high probability of heating effects by tidal heating.

Since the tidal heating is caused by mechanical deformation cycles induced by the changing rotational and revolutional configurations, there are two major mechanisms of tidal dissipation. One is the result of the forced eccentricity of the satellite orbit, the other is the result of the non-zero inclination and obliquity of satellite rotation vectors. The former mechanism is effective for Io and the latter is effective for Triton.

The amount of tidal dissipation for homogeneous bodies having non-zero eccentricity is given by (Peale et al., 1979; Kawakami and Mizutani, 1987)

$$E_T = 42 \pi \rho^2 R^7 n^5 e^2 / 19 \mu Q \quad (4)$$

where n is the mean motion, e is the eccentricity of the satellite orbit, μ is rigidity, Q is the quality factor which characterizes anelasticity of material constituting the bodies. Here we assumed that $\rho g R > \mu$, which is valid for large bodies. It should be noted here that the tidal dissipation is dependent on the 7th power of planetary radius, square of the mean motion and the square of the eccentricity. This implies that only large body orbiting quite near the heavy primary should have been affected by tidal heating. Since the eccentricity of satellite orbit decays in a rather short

time scale, orbital resonance between satellite orbits should be efficient mechanism for maintaining a long-term tidal heating (e.g., Yoder and Peale, 1981; Greenberg, 1982).

For the inclination-obliquity mechanism, the energy is given by

$$E_T = 4 \pi \rho^2 R^7 n^5 i^2 / 19 \mu Q \quad (5)$$

where i is the obliquity of the rotational axis (Kawakami, 1989a). If an estimation of the amount of tidal dissipation in Triton under current dynamical configuration is conducted using eq. (5), the tidal heating would be too small to cause any active volcanism on Triton.

It has been suggested that tidal heating was effective heat source for Europa (Cassen et al., 1980), Enceladus (Peale et al., 1980; Lissauer et al., 1984), the Uranian moons (Peale, 1988; Dermott et al., 1988; Tittlemore, 1990), and Triton (Goldreich et al., 1989; Ross and Schubert, 1990). Surface geology of these satellites shows evidences of a variety of resurfacing processes as shown in the Kawakami et al., (this issue). However, estimation of temporal variation of tidal heating remains a quite difficult problem, because calculation of orbital evolution requires information of orbital resonances among mutual moons or planetary rings, which exhibits chaotic behaviors during the passages of the resonant states (e.g., Tittlemore, 1990).

(4) Heat removal mechanisms

While the amount of energy heating up planetary interior is proportional to its radius with powers of 5, 3, and 7 for the accretional, radioactive, and tidal heating energies respectively, the energy escaped from its surface is proportional to its surface area, πR^2 . The thermal activity of planets is expected to increase drastically with its radius. There exist three major mechanisms in the lithospheric heat transfer (Solomon and Head, 1982). They are heat conduction, plate tectonic processes, and heat pipe mechanism.

Fig. 2. Axial ratios of satellite shapes as a function of their maximum axis. Data are from Smith et al. (1981, 1986), Croft (1992), Thomas (1988), and Dermott and Thomas (1988). En: Enceladus, Mr: Miranda, Pr: Proteus, Mi: Mimus, Hy: Hyperion, Am: Amalthea, Ja: Janus, Pm: Prometheus, Pn: Pandora, Ep: Epimetheus, He: Helene.

The creation, spreading, and subduction of oceanic plate carry significant part of the terrestrial heat flow (Sclater et al., 1980). Since plate tectonics is operated only on the Earth, the heat transfer mechanism associated with plate creation and consumption should not be applicable to the other solid bodies in the solar system.

The heat pipe mechanism which carry hot molten material from the deep interior to the surface is quite effective means for heat transfer (O'Reilly and Davis, 1980; Turcotte, 1989). However, operation of this mechanism require existence of hot molten materials and suitable tectonic setting to provide conduits to the surface. The ubiquitous volcanic landforms and lava fields on Io and Venus suggests that a large amount of heat flow has been transferred by this mechanism (Solomon and Head, 1982). Fig. 1 shows a model of Io's lithosphere which allows thick lithosphere with high heat flow (Kawakami and Mizutani, 1987). The thick lithosphere is based on the consideration that the rugged mountains on Io are supported by the strength of thick lithosphere. In the contrary, if the observed heat flow is transferred by the lithospheric conduction, the thickness of lithosphere must be as thin as a few km, which contradicts to the existence of rugged mountains. The only way to resolve these observations is to introduce heat pipe mechanism as shown in Fig. 1. The fact that a large amount of heat from Io is released from a restricted areas also suggests the operation of heat pipe mechanism.

It is assumed that silicate melt exists beneath the lithosphere and hot molten material from this layer is transferred to the surface. When the amount of material outpouring from hotspots is averaged over entire surface of Io, the corresponding resurfacing

rate is of the order of a few mm/year (Kawakami, 1989 b). The same amount of magma should be supplied by remelting of the bottom of lithosphere by tidal heating. When this advective process operates steadily, the material composing lithosphere must be recycled within a time scale of 1–10 Ma. The recycling of material through the lithosphere accompanying melting and solidification processes may lead to the accumulation of sulfur rich layer on Io.

The mode of material transfer through the Venusian mantle is also quite interesting, as will be discussed by Fujii (this issue). In contrast to the predominant vertical migration of material in Io's lithosphere, both the lateral deformation of lithosphere and vertical hot plume/cold plume activities in Venusian mantle make the mode of material transfer rather complicated.

The surfaces of other solid objects, such as the Moon, Mercury and Mars are covered with one thick lithosphere (see Pictorial, this issue, p. 1–5). The majority of heat flow from these objects were removed by the lithospheric conduction. Let us consider a spherical body with initially homogeneous temperature without internal heat sources for simplicity. The time scale of conductive cooling of this body is given by

$$\tau = R^2 / \kappa \quad (6)$$

where κ is the thermal diffusivity. The cooling time is proportional to the square of radius, indicating that duration of thermal activity is longer for larger bodies. The longer duration of volcanic activity for larger bodies would reflects their thermal inertia. Although the conductive heat transfer is predominant for the icy satellites around the Saturnian and Uranian satellites, thermal activities inferred from surface morphology do not depend solely on their sizes.

Thermal relaxation of shapes and topographic reliefs

Planetary shapes and topographic reliefs produced by impact crating or tectonic processes provide important information for thermal history of planetary lithosphere. Existence of topographically relaxed craters has been used to estimate viscosity and temperature of planetary lithospheres (e.g., Passey and Shoemaker, 1982). Similarly, transition of planetary shapes from irregular to sphere should provide an important constraint for planetary thermal evolution.

Fig. 2 shows the ratio of the maximum axis (A-axis) to the minimum axis (C-axis) of satellite's shape as a function of the maximum radius. Bodies with radius larger than 250 km can be represented by sphere in shape. The condition for shape evolution toward sphere is dependent on whether the differential stress caused by its irregular shape exceeds the yield strength of constituting material, and on whether the time scale of creep deformation is shorter than the age

of the solar system.

If the constituent material is expressed by the Maxwell body, the relaxation time is given by

$$\tau = \eta / \mu \quad (7)$$

where η is the viscosity. The viscosity of material is generally exponential function of temperature. When $\eta = 10^{21}$ Pa.s and $\mu = 4$ Gpa is adopted as a nominal values, τ is of the order of 10^5 years. Thus, it is expected that bodies which experienced a high temperature enough for viscous relaxation would have spherical shapes. On the other hand, the differential stress, $\rho g \delta R$ is generally smaller than the yield stress of material in low temperature. The yield strength of ice just below the melting temperature is 4 MPa (Arakawa and Maeno, 1992), which is comparable to the differential stress of the object with $R = 250$ km with maximum relief, $\delta R = 25$ km. Thus, the transition between irregular to spherical shapes observed in the icy satellites implies that interior of large icy satellite was heated to a high temperature enough for creep deformation.

Although deep interior of planet is heated to a high temperature, it is expected that the surface layer remains cold to maintain short-wavelength topographic undulations. Thus topographic relief is also considered to reflect thermal history of planetary lithosphere. Fig. 3 shows the maximum relief of topography as a function of planetary radius (Croft, 1992). The maximum relief for irregular satellites with radius less than about 200 km increases with its size suggesting a self-similar nature in their shapes. On the other hand, the ice-covered Galilean satellites are almost completely relaxed suggesting that their internal temperature increased even in the superficial layer. The maximum reliefs of the Saturnian and Uranian satellites with higher volcanic or resurfacing activities show smaller maximum reliefs. However, the differences of volcanic activity among twin bodies such as Ganymede and Callisto, Titania and Oberon, Ariel and Umbriel can not be explained by a simple energy-size considerations. The differences for the pairs of Uranian moons were possibly due to the tidal heating (Tittemore, 1990). The low volcanic activities of Saturnian satellites, Rhea and Iapetus, compared to the smaller Saturnian moons are enigmatic.

It has been observed that there exists a large misfit between the mass center and shape center of the Moon (Kaula, 1975). The difference has been regarded to be the difference of crustal structure between the Earth-facing side and opposite side (Toksöz et al., 1974). The thickness of the lunar crust facing the Earth is thinner and covered with young basaltic lavas. It has also been found that large gravitational anomalies (mascon) exist over the basins covered with basaltic lavas (Kaula, 1975). These geophysical observations have suggested that the lunar upper mantle has been cooled to low temperature within a billion

Fig. 3. Maximum topographic relief as a function of their sizes. Data are from Thomas (1988) and Croft (1992). Ga : Ganymede, Ca : Callisto, Eu : Europa, Ob : Oberon, Um : Umbriel, Rh : Rhea, Te : Tethys, Ta : Titania, Ar : Ariel, Di : Dione, En : Enceladus, Mr : Miranda, Mi : Mimus, Pr : Proteus, Hy : Hyperion, Ja : Janus, Ep : Epimetheus, Pm : Prometheus, Pn : Pandora, He : Helene, Ca : Calypso.

years after solidification of the magma ocean. These observations were used to constrain thermal structure of the Moon (Pullan and Lambeck, 1980). The thermal evolution model proposed by Toksöz et al. (1978) shows a temperature evolution apparently consistent with the above observations. Future thermal evolution studies of icy satellites should be explored with shape and heterogeneity information as constraints.

Mantle convection and tectonics

Orogenic belts produced by compressive deformation of the lithosphere are ubiquitous on the Earth and Venus. These features would be related to collision of continental plates or accumulation of crustal materials over the mantle return flows. Formation of these tectonic features can be ultimately attributed to the stresses induced by mantle convection. Although some compressive deformation features interpreted to be thrust faults were identified on the Mercury, their formation would be unrelated to mantle convection. Their origin was postulated to be tidal despinning (Melosh, 1977) or the Carolis-forming impact (Fleitout and Thomas, 1982). For Mars, many tectonic features were interpreted on the basis of regional tectonic setting (e.g., Banerdt et al., 1992).

On the other hand, quite few compressive features can be observed on the icy satellites. While extensional features such as large chasmata, rifts or grooves are quite commonly observed. A number of mechanisms were proposed which resulted in extensional tectonics on the icy satellites. A significant volume expansion associated with phase transition of ices was proposed for the origin of the grooves on

Fig. 4. (a) A model of structural evolution of Ganymede. Ganymede was assumed to be accreted homogeneously. Differentiation of silicates and ice begins from the hotter outer layer. (b) Phase diagram of ice as a function of depth for the homogeneous Ganymede model. Differentiation of silicates and ice causes upward migration of ice phase, resulting in phase transition from phase VII to V or VI.

Ganymede (Kawakami and Mizutani, 1980). If Ganymede and Callisto were homogeneous mixtures of ice and rocky materials just after the accretion, subsequent gravitational differentiation to form silicate core and icy mantle should have resulted in a significant volume expansion, mainly due to the phase changes of ice VII to ice V or VI (Fig. 4). Dehydration reactions of silicates are considered for Europa (Finnerty et al., 1981), and expansion or contraction associated with internal thermal evolution would be effective for small bodies (Ellsworth and Schubert, 1983; Hillier and Squyres, 1991). Although all or some of these processes should be responsible for surface cracking or groove formation, it is not possible to determine which processes should have been involved for the specific extensional features.

An important problem to be explained is the absence of compressive deformation features on the icy satellites. If mantle convective flows in the large icy satellites such as Ganymede and Callisto are strong enough, drag forces exerted to the bottom of lithosphere should have been produced orogenic belts. In order to estimate such possibilities, compressive stress induced by convective flow is estimated using the thermal boundary layer model described by Turcotte and Schubert (1982).

According to the two-dimensional thermal boundary layer model of convection heated from below, horizontal velocity (u) of flow beneath the thermal boundary layer is given by

$$u = 0.271 \kappa / b \text{ Ra}^{2/3} \quad (8)$$

where Ra is the Rayleigh number,

$$\text{Ra} = \rho g \alpha \delta T b^3 / \nu \kappa \quad (9)$$

κ is thermal diffusivity, α is thermal expansion coefficient, g is gravity, δT is the temperature difference between top and bottom of the convection cell, ν is viscosity, and b is the thickness of convection cell. The shear stress (σ) at the bottom of the thermal boundary layer is

$$\sigma = 2\mu u / b. \quad (10)$$

The thickness of boundary layer (δ) is given by

$$\delta / b = 1.7 \text{ Ra}^{-1/3} \quad (11)$$

If one specifies nominal parameters in these equations, thickness of the boundary layer, shear stress and horizontal velocity at the bottom of the boundary layer can be estimated. The results for the terrestrial planets and some icy moons are shown in Table 2. In this calculation, most critical parameter is the viscosity. When the value $\nu = 10^{22}$ Pa.s is accepted as a nominal value for silicates, the calculated thermal boundary layer thickness, plate velocity, and stress of the Earth is 28 km, 9.7 cm/year, and 21 MPa. These values are within a factor of two of the estimated values (e.g., Parson and McKenzie, 1978; Kanamori and Anderson, 1975; Minster and Jordan, 1978), although relation between the state of stress in the lithosphere and drag force due to the mantle convection has not been clarified (Richardson et al., 1980).

For the rheology of ice, the value $\nu = 10^{17}$ Pa.s has been adopted. The low viscosity of ice yields the Rayleigh number in the order of magnitude higher than those of silicate bodies, which results in high

Table 2. Physical properties and strength of mantle convection estimated from the thermal boundary layer theory

physical properties	Earth	Venus	Mars	Mercury	Moon	Ganymede	Enceladus
R (km)	6370	6051	3393	2439	1740	2631	251
g (m/s ²)	9.81	8.60	3.72	3.78	1.62	1.44	0.08
b (km)	2900	2800	1800	600	1200	800	200
α (10 ⁻⁵ /K)	3	3	3	3	3	10	10
κ (m ² /s)	10 ⁻⁶	10 ⁻⁶	10 ⁻⁶	10 ⁻⁶	10 ⁻⁶	10 ⁻⁶	10 ⁻⁶
δT (K)	2000	2000	2000	2000	2000	300	300
ν (Pa.s)	10 ²²	10 ²²	10 ²²	10 ²²	10 ²²	10 ¹⁷	10 ¹⁷
Ra	6x10 ⁶	5x10 ⁶	4x10 ⁵	1x10 ⁴	5x10 ⁴	3x10 ⁸	2x10 ⁵
δ (km)	28	29	42	42	55	3	11
u (cm/y)	9.7	8.9	2.6	0.7	1.0	480	14.6
σ (MPa)	21	20	9.1	7.0	5.1	0.04	0.5
L (km)	20-40	15-40	18-230	-	> 50	-	-

horizontal velocity. However, low viscosity of ice could not sustain high tectonic stress in the lithosphere. Thus, the difference of rheological properties between silicates and ice would be a likely reason for the absence of compressive features on the icy bodies. The mode of coupling of mantle convection and tectonic deformation of lithosphere for the Earth/Venus and Ganymede/Callisto are schematically illustrated in Fig. 5.

Although the thickness of thermal boundary layer and state of stress in the planetary lithosphere is difficult to estimate, the elastic thickness of the lithosphere can be inferred from the topography or gravity data. The thickness of the elastic lithosphere of the Earth can be derived from the bending of subducting oceanic lithosphere or topographic features resulting from mass loading by the seamounts (e.g., Bondie et al., 1981). According to these studies, the thickness of elastic lithosphere is in the range from 20 to 40 km, roughly comparable to that of thermal lithosphere calculated in Table 2.

The values of lithosphere thickness (L) estimated for the Moon, Mars and Venus are also listed in Table 2. Solomon and Head (1980) compiled tectonic features of lunar maria and estimated elastic thickness of lithosphere comparing stresses predicted from the mascon loading and observed tectonic features. The results suggest the thickness of elastic lithosphere increased with time and that significant lateral variation in lithospheric thickness ranging from <25 km to >75 km prevailed at the time of basin formation. Similar procedure was applied to Martian volcanic constructs, yielding that thickness of elastic lithosphere is quite heterogeneous ranging from 18 to 230 km (Solomon and Head, 1990). On the other hand, Sandwell and Schubert (1992) interpreted the annular moats and outer rises around large Venus coronae comparable to the trench-outer rise system and ob-

tained the thickness of elastic lithosphere in the range from 15 to 40 km.

Variation of lithospheric thickness with time is potentially one of the important constraints for planetary thermal history calculations (Schubert et al., 1979). According to the detailed thermal evolution models studied by Spohn (1991), the predicted thickness variation is dependent on the vigor of mantle convection, the mode of heat transport in the lithosphere, and degree of mantle differentiation. The observed lateral heterogeneities in lithospheric thickness mentioned above could not tightly constrain the thermal evolution models at present, because the degree of lateral heterogeneities appears to be too large to indicate global average value. Rather, it would be valuable to explore the causes and consequences of lateral heterogeneities in view of mantle dynamics or regional tectonics to constrain tectonic history of the terrestrial planets.

Summary

The tectonic style of the terrestrial planets and icy satellites is controlled by heat sources and the mode of heat transfer through the lithosphere. Since the accretional energy and radiogenic heat increases drastically with body size, duration and strength of thermal and tectonic activities are generally greater for the objects with larger dimensions. Tidal heating is a likely energy source for some of the small icy satellites with exceptionally high resurfacing activities, such as Enceladus, Ariel, and Triton.

Heat flowing out from planetary interior is carried by conduction, outpouring of hot material (magma) through heat pipes, or creation and subduction of oceanic lithosphere. The fraction of heat flow carried by these processes is strongly dependent on planetary size, chemistry, and on the mechanical properties of lithosphere. While heat transfer associated with

plate tectonics is unique to the Earth, the heat pipe mechanism is predominant on Io and Venus. Ridge network and corona array on Venus would be surface manifestation of transient tectonic style from plume tectonics to plate tectonics as discussed by Fujii (this issue).

Tectonic features produced by compressive stresses are ubiquitous only for the Earth and Venus. Although vigor of convective flow in the ice mantle might have been strong for Ganymede and Callisto, the tectonic stresses caused by convective flow was insufficient to form orogenic belts, because the rheology of ice is too weak to sustain large compressional stresses in the lithosphere.

The evolution of thickness and mechanical properties of lithosphere is a consequence of thermal evolution of lithosphere. Variations of these parameters with time have been analyzed from topographical informations to constrain thermal and mechanical history of lithosphere. However, it has been obvious that there exists strong lateral heterogeneities in the planetary lithospheres. Regionality of lithospheric or crustal properties will provide important information for evolutionary histories of planets. The direction of studies in structure and dynamics of the Earth's interior in the past may suggest that studies on the causes and consequences of the lateral heterogeneities and on the mechanisms maintaining such heterogeneities over long time should be a next step for planetary dynamics studies. Efforts to understand internal structure of planets by means of geophysical and geological observations should be required for future progress.

On the other hand, more sophisticated and integrated models involving tectonics, thermal history, and global differentiation should be valuable for organizing and interpreting the geological knowledge and for making future exploration plans. However, importance of the observational data has greatly increased since it has been realized that rare and probabilistic events, such as the catastrophic giant impact event to form the Earth's Moon (Hartmann et al., 1986), affected the solar system history more seriously than previously considered. The strange surface markings observed on Miranda is the other example (Greenberg et al., 1991). If such stochastic events affected evolution of planets significantly, models of planetary evolution will be helpful to make clear the uniqueness of each object. Understanding the uniqueness of each object by means of both the observational and theoretical studies would be a major challenge in planetary sciences.

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Kawakami, S., 1994, A comparative study on the volcanism and tectonics of the terrestrial planets and satellites. *Jour. Geol. Soc. Japan*, **100, 71-80.** (川上紳一, 1994, 地球型惑星, 氷衛星の火山活動とテクトニクスの比較惑星学. 地質雑, **100**, 71-80)

地球型惑星と氷衛星の火山活動度に認められる多様性の原因を, 内部熱源量と熱輸送形態に基づいて考察した. 惑星内部の熱源量は, 惑星の大きさとともに増大するが, 熱的活動度と天体サイズとの対応関係は, 土星や天王星の衛星では成立しないものがある. エンセラダス, アリエル, トリトンでは潮汐摩擦による加熱が有効であった可能性が高い.

マンツルの対流運動によって生じる応力の大きさを熱境界層モデルを用いて推定した. 氷衛星に圧縮性の応力でできた地形がない理由として, 氷の降伏応力が小さいこと, または氷衛星の内部分化に伴う伸張性応力が卓越したことが挙げられる. 氷衛星の大きさと形状, および地形の最大起伏量は, それらの天体の熱的活動度をよく反映している.