Geochemical Journal, Vol. 50, pp. 27 to 42, 2016 doi:10.2343/geochemj.2.0398

INVITED REVIEW

Origin of Earth's oceans: An assessment of the total amount, history and supply of water

HIDENORI GENDA*

Earth-Life Science Institute, Tokyo Institute of Technology, Ookayama, Meguro-ku, Tokyo 152-8550, Japan

(*Received April 9, 2015; Accepted October 15, 2015*)

The presence of water on Earth has played important roles in shaping the solid regions of the planet as well as in the origin and evolution of life. This paper addresses three fundamental aspects of Earth's water; (1) the quantity of water on the surface and in the interior that Earth possesses, (2) the length of time that surface oceans have been present, and (3) the mechanism(s) by which this water was supplied or generated. From geochemical and geophysical analysis, and high-pressure experimental works, the water content in the Earth's mantle can be estimated to be from one to ten times the present ocean mass. Although it is difficult to estimate the water content in the Earth's core, recent high-pressure experimental work indicates copious amounts of hydrogen in the core. From geological and geochemical evidence, the Earth's surface oceans appear to have existed since very early in the Earth's history, perhaps even since the Earth's formation. However, changes in the ocean volume throughout the Earth's history have not been well determined. Several possible water sources and supply mechanisms have been proposed, in association with theories regarding planet formation in our solar system. Since there are several uncertainties concerning the process of planet formation, the origin of the Earth's water is still in question.

Keywords: origin of water, age of oceans, water in Earth, planet formation, snow line

INTRODUCTION

Our Earth is often referred to as an "aqua planet," due to the large quantity of water on its surface (71% of the Earth's surface is covered by oceans) compared to the other terrestrial planets in our solar system, such as Mercury, Venus and Mars. However, the mass of the Earth's oceans ($M_{\mathrm{oce}} = 1.4 \times 10^{21}$ kg) is only 0.023 wt% of the planet's total mass ($M_{\mathrm{E}} = 6.0 \times 10^{24}$ kg) (see Fig. 1). Even if the water in the Earth's interior (the mantle and core) is taken into account, the mass fraction of water does not exceed 2 wt% of the total planetary mass. In contrast, Uranus and Neptune, the outermost planets in the solar system, are composed primarily of H_2O (60–70 wt%) (e.g., Guillot, 2005). Therefore, Earth is not truly an aqua planet from the viewpoint of its H_2O content, and strictly speaking Uranus and Neptune are not aqua planets either, because their abundant H_2O is not in the form of liquid, but ice.

Water molecules (H_2O) are expected to be abundant in our solar system, since hydrogen is the most abundant element in the universe and oxygen is the third most abundant element in the solar system (Anders and Ebihara, 1982). The second most abundant element, helium, is chemically inert and does not normally form molecules. In the protoplanetary disk from which our solar system planets formed, there were numerous H_2O molecules, likely amounting to two to three times the mass of the rock and iron that were the main building blocks of the terrestrial planets. Since H_2O molecules condense into ice beyond the Jovian planet region, it is a natural consequence that Neptune and Uranus were formed mainly from these ices. If the C/O ratio in an extrasolar system is more than unity, oxygen atoms are likely to combine with carbon atoms to form CO, and so H_2O is not produced. However, since the C/O ratios in most extrasolar systems are less than unity, as confirmed by observations of various stars (Petigura and Marcy, 2011), H_2O could be as abundant in extrasolar systems as in our solar system.

The distance from the central star (or from the sun in our solar system) is the most important factor determining the stability of liquid water on a planetary surface (e.g., Kasting *et al.*, 1993). If a planet orbits close to a star, all liquid water will be vaporized due to the runaway greenhouse effect of water vapor. A planet with oceans has an upper limit of planetary radiation (Kasting, 1988; Nakajima *et al.*, 1992; Kodama *et al.*, 2015). If a planet receives a solar energy flux above this upper limit,

---

*E-mail address: genda@elsi.jp

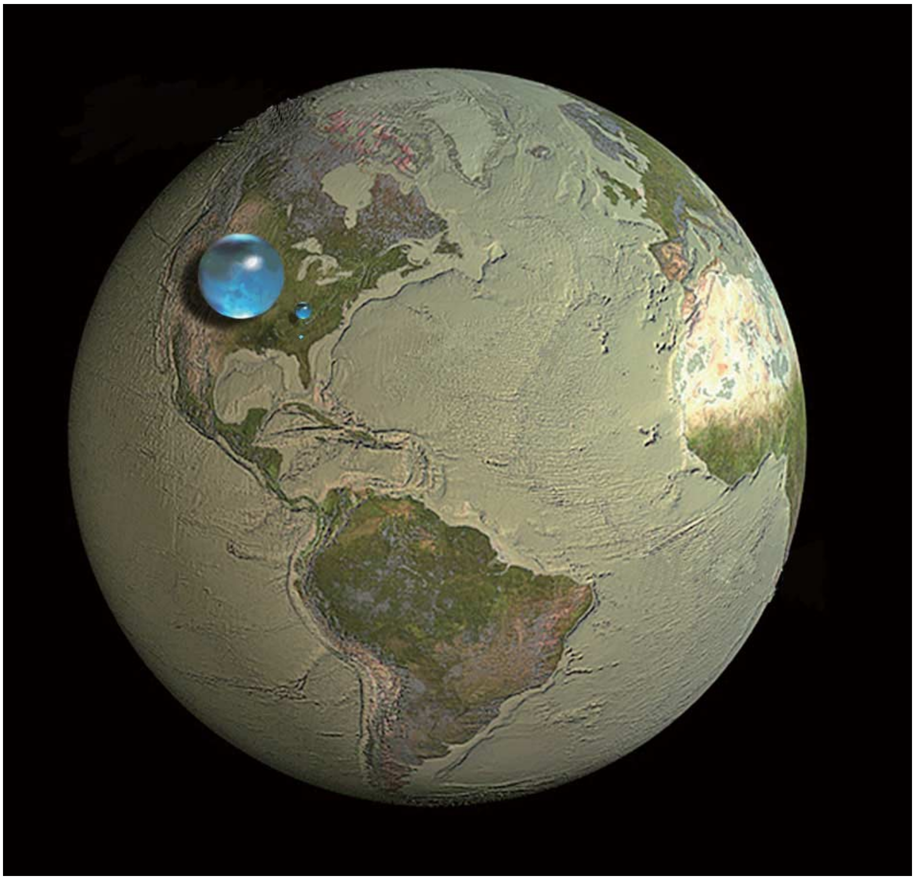

Fig. 1. The amount of water on Earth compared with the overall globe. (Note: two of the three droplets are quite small and may be difficult to see.) The largest droplet represents the volume of the Earth's oceans, the second largest droplet represents the total volume of water other than that in the oceans (such as in rivers, lakes, ground water, water vapor and the water in living organisms), and the smallest droplet represents the volume of water that humans can readily utilize. From Howard Perlman, USGS; globe illustration by Jack Cook, Woods Hole Oceanographic Institution; © Adam Nieman.

all water is vaporized. Conversely, a planet far from its star will hold only frozen water. The amount of atmospheric greenhouse gases is also important in determining the stability of liquid water. As an example, if Earth had no greenhouse gases (that is, no CO_2 or water vapor), its surface temperature could be as low as –15°C, and therefore it would be fully covered with ice. The adequate distance of Earth from the sun and a suitable amount of greenhouse gases in the Earth's atmosphere have made this planet habitable.

The water on Earth also plays many important roles in the solid regions of the planet. In plate tectonics, water acts a lubricant (e.g., Regenauer-Lieb *et al.*, 2001), and water also promotes long-term climate stability through the carbon cycle between the atmosphere and the planet (Walker, 1982; Tajika and Matsui, 1992). If, for example, Earth had ten times its present volume of oceanic water, there would be no dry continents and the Earth's environment would be quite different (Maruyama *et al.*, 2013). Therefore, an adequate amount of liquid water on the Earth's surface is also important in terms of forming and preserving the present Earth's surface environment.

Liquid water is one of the most important compounds required to support life (at least terrestrial life as we now know it), and many believe that liquid water played a significant role in the generation of living organisms on the planet. Indeed, H_2O molecule has many unique characteristics. Hydrogen bonding between molecules gives water its relatively high melting and boiling temperatures compared to other compounds with comparable molecular weights (Fig. 2). For this reason, liquid water is stable up to a reasonably high temperature, and so chemical reactions in aqueous solutions can potentially proceed rapidly, which has been advantageous with regard to the birth and evolution of life. The unique properties and high abundance of H_2O molecules in the universe suggest that extraterrestrial life will also be water-based.

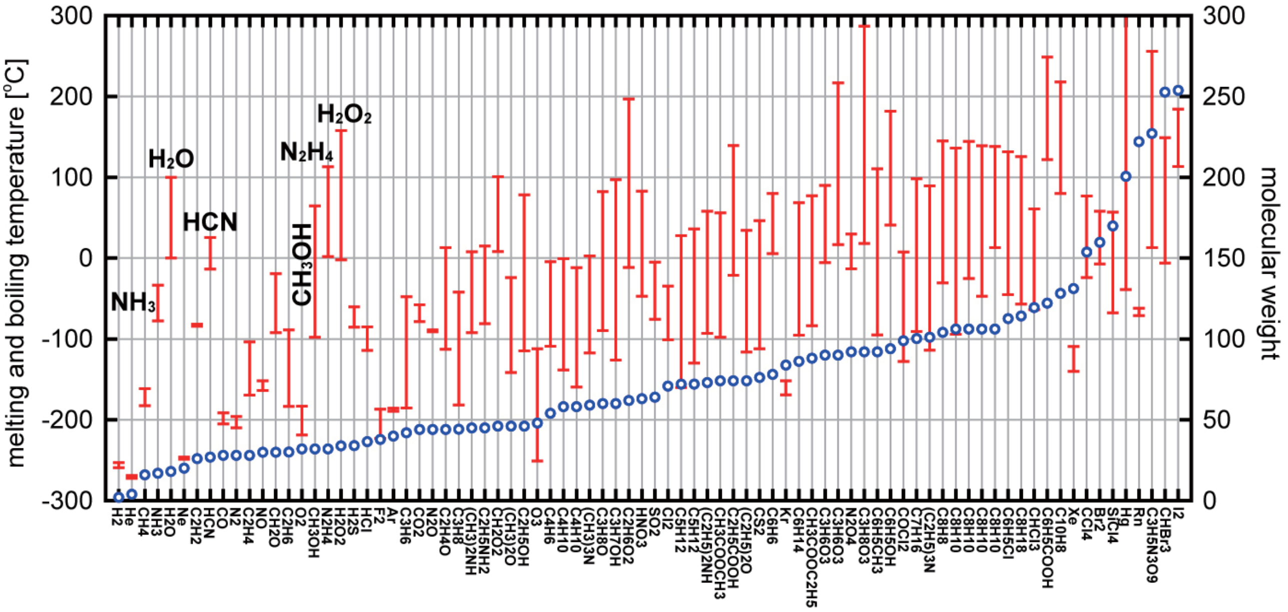

Fig. 2. Melting and boiling temperatures at 1 atm (lower and upper ends of red bars) and molecular weights (blue circles) of various compounds. Data are taken from Perry's Chemical Engineers' Handbook (Green and Perry, 2008), and Abe (2009).

Herein, three fundamental aspects regarding water on Earth are discussed. These are the actual quantity of water on the planet, including that in the mantle and core, the time length that surface oceans have been present, and the mechanism by which the planet acquired its present water.

EARTH'S WATER BUDGET

Surface water

The amount of water in the oceans on the Earth's surface (M_{oce}) has been determined with reasonable precision, and is reported to be 1.37×10^{21} kg. There are other water reservoirs on or around Earth, such as ice, groundwater, lakes, rivers, soils, sedimentary rocks, crust and the atmosphere, and the total mass of water in these reservoirs is estimated to approximately 5×10^{20} kg (Mottl *et al.*, 2007), a value that is about one half of M_{oce}.

Water in the mantle

The Earth's mantle could potentially represent a huge reservoir of water because of its large mass (4.0×10^{24} kg or 67 wt% of the Earth's mass). If the entire mantle contains on average 0.1 wt% water, this equates to 4.0×10^{21} kg of water, corresponding to 3 M_{oce}. Therefore, determining the water content in the mantle is vital in terms of assessing the overall water budget of the planet. However, precise and direct measurement of the water content in the mantle is quite difficult because we cannot access the mantle directly. Even the deepest hole drilled to date (the Kola Superdeep Borehole in Russia) reached a depth of only 12 km (Kremenetsky and Ovchinnikov, 1986), and thus did not exit the Earth's crust. Nevertheless, we can infer the water content in the mantle by means of the extensive analysis of mantle-derived rock samples, geophysical probing methods such as electrical conductivity assessments of the mantle, and high-pressure experiments of hydrous minerals. Table 1 summarizes the estimated water content in the Earth's mantle determined by these methods.

Studies of water and trace elements in mid-ocean ridge basalts (MORBs), which are formed by partial melting of the upper mantle, indicate that the upper mantle contains between 50 and 200 ppm H_2O (Michael, 1988; Dixon *et al*., 2002; Hirschmann, 2006). High-pressure experiments of olivine, the major component of the upper mantle, have shown that the water capacity of olivine increases with pressure, approaching 0.4 wt% (=4000 ppm by weight) at the lowermost regions of the upper mantle (410 km) (Bell *et al*., 2003; Koga *et al*., 2003). Since this measured value is not the actual water content in the upper mantle, but rather represents the holding capacity, the water content in the upper mantle estimated from MORB samples shows that the upper mantle is not fully saturated with water.

The electrical conductivity of the Earth's upper mantle also shows that the upper mantle is relatively dry. It is known that the electrical conductivity of mantle minerals is strongly correlated with their water content. Therefore, measuring the electrical conductivity of the Earth's

Table 1. Estimated water contents in the Earth's mantle

Mantle	Mass (10^{21} kg)	Natural samples	Electrical conductivity	High-pressure experiments	Water mass (M_{oce})
Upper mantle	615	50–200 ppm[1]	<0.09 wt%[4]	<0.4 wt%[6]	0.02–2
Transition zone	415	~1 wt%[2]	~0.1 wt%[5]	~2 wt%[7]	0.3–5
Lower mantle	2955	300–1000 ppm?[3]	—	~0.2 wt%[8]	0.6–4

Data from [1] Hirschmann (2006), [2] Pearson et al. (2014), [3] Dixon et al. (1997, 2002), Hauri (2002), [4] Wang et al. (2008), [5] Yoshino et al. (2008), Huang et al. (2005), [6] Bell et al. (2003), Koga et al. (2003), [7] wadsleyite: Inoue et al. (1995, 1998), ringwoodite: Kohlstedt et al. (1996), [8] Murakami et al. (2002), Litasov et al. (2003).

mantle can provide information concerning the water concentration. The electrical conductivity has been ascertained based on electromagnetic induction theory by measuring magnetic and electric fields both on the ground and using artificial satellites (Olsen, 1999; Kuvshinov and Olsen, 2006; Püthe and Kuvshinov, 2013). Combining the observed data with the experimentally determined electrical conductivity of high-pressure hydrous minerals, the water contents have been estimated at 10–900 ppm in the upper mantle (Wang *et al.*, 2008), and at <0.1 wt% (Yoshino *et al.*, 2008) and at 0.1–0.2 wt% (Huang *et al.*, 2005) in the transition zone. There is no reliable data concerning the lower mantle.

The transition zone, ranging from 410 to 660 km, is thought to be a potentially huge water reservoir. Wadsleyite and ringwoodite are high-pressure polymorphs of olivine, and dominate the transition zone. High-pressure experiments on the solubility of H_2O in these minerals show that wadsleyite and ringwoodite can contain 3.3 wt% (Inoue *et al.*, 1995, 1998) and 2.2 wt% H_2O (Kohlstedt *et al.*, 1996), respectively. Since about 60 wt% of the transition zone is composed of these minerals, the transition zone could potentially contain a mass of water equal to 5 M_{oce}. However, the water content estimated by electrical conductivity measurements of the transition zone (approximately 0.1 wt%) does not support a water saturated transition zone (Yoshino *et al.*, 2008; Huang *et al.*, 2005).

Recently, Pearson *et al.* (2014) found a hydrous ringwoodite inclusion in a diamond sourced from Juína, Brazil. This was the first discovery of ringwoodite in natural samples on Earth, although this mineral has been identified in meteorites and synthesized in laboratories. The water-rich nature of this inclusion was direct evidence for a wet transition zone, at least locally. Pearson *et al.* (2014) estimated that the transition zone contains approximately 1 wt% water.

Geochemical studies of oceanic island basalts (OIBs) suggest that the water content of the OIB source ranges from 300 to 1000 ppm (Dixon *et al.*, 1997, 2002; Hauri, 2002). If OIBs come from the lower mantle, this region could also represent a significant water reservoir. High-pressure experiments of lower mantle minerals support the prediction of a large amount of water in the lower mantle. Murakami *et al.* (2002) reported that the major minerals of the lower mantle could hold water up to 0.2 wt% for Mg-perovskite, 0.4 wt% for Ca-perovskite, and 0.2 wt% for magnesiowüstite.

Thus, although an exact determination of the water content in the entire Earth's mantle is difficult, the value evidently is in the range of 1–10 M_{oce} (see Table 1).

Hydrogen in the core

Experimental studies of hydrogen partitioning between the hydrous silicate melt and molten iron show that more than 95% of H_2O reacts with Fe to form FeH_X at 7.5 GPa (Okuchi, 1997). Therefore if Earth had held water during the magma ocean stage, a significant fraction of the hydrogen in the water would have been incorporated into the Earth's core.

Seismic velocity data shows that the density of the Earth's liquid outer core is 5–10 wt% less than that of pure iron (Birch, 1952; Dubrovinsky *et al.*, 2000; Dewaele *et al.*, 2006). To account for this density deficit, the core must contain a considerable amount of light elements, such as Si, O, S, C and/or H (e.g., Poirier, 1994; Hirose *et al.*, 2013). In the extreme case in which H is considered as the only light element in the core, the density deficit could be entirely reconciled by assuming 0.5–1.0 wt% H in the form of FeH_X ($X = 0.28–0.56$) (Narygina *et al.*, 2011). Since the mass of the outer core is 1.9×10^{24} kg, this amount of H corresponds to H in the water of 60–120 M_{oce}.

Recently, Nomura *et al.* (2014) showed experimentally that the solidus temperature of pyrolite (a theoretical rock representing the mantle) is 3570 K at the core-mantle boundary (CMB), a value that is about 400 K lower than previously thought. Since a lower temperature at the CMB also implies a lower temperature at the core, the melting temperature of the outer core must be greatly depressed in order for the inner part of the core to solidify. Since such a large depression is impossible without the presence of hydrogen in the core (Alfè *et al.*, 2007; Sakamaki *et al.*, 2009), Nomura *et al.* (2014) concluded that the outer core must contain high concentration of hydrogen corresponding to an amount of water equal to

Fig. 3. Pillow lava basalt in Isua, Greenland, which erupted under water at 3.8 Ga. The grey-green portion is the core of the pillow and is mantled by a dark green portion, which in turn is rimmed by pale-colored chilled margins, together with the matrix. The scale is given by a hammer at the top of the photo. Courtesy of Museum of Evolving Earth, Tokyo Institute of Technology.

80 M_{oce}. Therefore, the core is potentially a huge reservoir of water (up to approximately 100 M_{oce}). However, since we have no direct samples from the Earth's core, we cannot confirm the hydrogen content in the core.

HOW LONG HAVE THE EARTH'S OCEANS EXISTED?

Geological and geochemical constraints

At present, it is thought that the oceans have existed since the very early stage of the Earth's history. Direct evidence for the existence of an ancient ocean is provided by pillow lava formed throughout the Earth's history and located in various locations around the planet. This lava is characterized by pillow-shaped masses, as shown in Fig. 3, and is formed when hot lava flows into water and cools rapidly. The oldest pillow lavas were found in the Isua Supracrustal Belt, southwestern Greenland, and were determined to have formed at 3.7–3.8 Ga (Appel *et al.*, 1998; Maruyama and Komiya, 2011). Therefore, the Earth's oceans must have existed at 3.8 Ga.

The oldest rock sampled to date comes from Acasta Geneiss, northwestern Canada, and is dated to 4.0 Ga (Bowring *et al.*, 1989; Bowring and Williams, 1999), thus there are no geological rock records on Earth before 4.0 Ga, and the first approximately 500 Ma of the Earth's history remains poorly understood. However, it is possible to locate zircon ($ZrSiO_4$) that can be dated before 4.0 Ga. Zircon is a mineral that is highly resistant to erosion, weathering and metamorphism. During the Hadean era which ranged from 4.5 Ga (the Earth's formation) to 4.0 Ga, all Earth's rock records prior to 4.0 Ga has been missing and/or has been destroyed likely by early intense meteor bombardment, but zircon grains are thought to have survived this era (e.g., Marchi *et al.*, 2014). Hadean zircon grains have high oxygen isotope (δ^{18}O) values (Wilde *et al.*, 2001; Mojzsis *et al.*, 2001; Valley *et al.*, 2014), indicating that the Earth's oceans existed during the Hadean era. High δ^{18}O values of zircon grains compared with the mantle are produced by low-temperature interactions between rocks and liquid water. Zircon grains with the ages of 3.91 to 4.28 Ga have high δ^{18}O values ranging from 5.4 ± 0.6 to 15.0 ± 0.4‰, suggesting interactions between the continental crust and oceans (Mojzsis *et al.*, 2001). Again, this presents evidence for the existence of Earth's oceans during the Hadean era.

The early formation of the Earth's atmosphere deduced by noble gas geochemistry (Hamano and Ozima, 1978; Staudacher and Allègre, 1982) is also consistent with the presence of oceans on early Earth. As an example, Hamano and Ozima (1978) estimated the timing and degassing rate of Ar from the Earth's interior, and showed that more than 80% of the internal Ar must have degassed to the surface during the Hadean era to explain the present low ^{40}Ar/^{36}Ar ratio in the air (=295.5) compared to the extremely high $^{40}Ar/^{36}Ar$ ratios in MORB samples (>30,000). The degassing of Ar likely also involves the degassing of other volatiles, such as CO_2 and H_2O.

Both the geological and geochemical evidence discussed above show that the Earth's oceans have existed since the very early stages of Earth's history. Moreover, water in lunar samples, such as volcanic glasses (Saal *et al.*, 2008; Hauri *et al.*, 2011), apatite crystals in lunar volcanic rocks (McCubbin *et al.*, 2010; Boyce *et al.*, 2010) and plagioclase crystals in lunar highland rocks (Hui *et al.*, 2013), has been discovered recently. Especially, the presence of water in plagioclase strongly supports the supply of water on Earth and Moon prior to their formation (i.e., at 4.5 Ga), because lunar plagioclase was crystallized from lunar magma ocean related to lunar formation just after the giant impact. Therefore, Hui *et al.* (2013) concluded that the Moon must have contained a significant amount of water at the time at which it was formed. They also estimated that the water content in the lunar magma ocean could have been as high as 320 ppm, a value similar to the concentration of water in the upper mantle of present Earth, although this high value of water concentration might be overestimated if the recent calibration of the FTIR work is applied (Losq *et al.*, 2015). Geophysical data concerning the moon's electrical conductivity and tidal quality factor also support the existence of a wet lunar interior (Karato, 2013) containing as high water concentration (approximately 100 ppm) as the Earth's upper mantle. Therefore, water was presumably supplied to Earth as well as the Moon prior to their formation.

Ocean volume changes throughout the Earth's history

There is almost no doubt that the Earth's oceans have existed throughout the Earth's history. However, estimating the ocean volume in the past is much more difficult than simply confirming their presence. The volume of the ocean would have varied by means of the loss of hydrogen into outer space and/or the exchange of water between the Earth's interior and the oceans.

The escape rate of hydrogen into space from present-day Earth is estimated at approximately 3 kg/s (Hunten, 1982). If this rate was constant as far back as 4.5 Ga, the total loss of hydrogen would be 4×10^{17} kg, corresponding to 3.9×10^{18} kg of water or 0.2% of the present Earth's ocean mass. Therefore, the current escape rate of hydrogen has little effect on the ocean volume. This rate is controlled by the diffusion of the hydrogen carrier (H_2O in the present Earth's atmosphere) in the atmosphere. In the case of present-day Earth, the cold trap of water vapor (i.e., condensation of water vapor) results in very low concentrations of water vapor ($\sim 10^{-7}$) in the upper atmosphere (e.g., Goody, 1995). Therefore, the current escape rate of hydrogen is extremely low. In the case of very high surface temperatures, such as on Venus, the cold trap vanishes and extreme hydrogen loss takes place. This phenomenon was responsible for water loss on the ancient Venus (Kasting *et al.*, 1993; Hamano *et al.*, 2013).

If the concentration of CH_4 in the ancient atmosphere was more than 10 times higher than that in present Earth, CH_4 would have replaced H_2O as the main hydrogen carrier, leading to a high hydrogen escape rate. Pope *et al.* (2012) reported that Archean oceans were depleted in deuterium ($\delta D = -25 \pm 5$‰) based on the analysis of serpentine from the approximately 3.8 Ga Isua Supracrustal Belt in western Greenland. Tomiyasu *et al.* (2013) also reported that Archean oceans depleted in deuterium ($\delta D = -24 \pm 5$‰) judging from the analysis of pillow basalts from the approximately 3.5 Ga Barberton Greenstone Belt in South Africa. Hydrogen escape into space increases the value of δD over time, since hydrogen is lighter and thus escapes more readily than deuterium (Hunten, 1973; Genda and Ikoma, 2008). Pope *et al.* (2012) concluded that the early Earth oceans were about 26% more voluminous than at present. High concentrations of CH_4 in the early Earth atmosphere would have been responsible for this significant loss of hydrogen. The D/H ratio of ancient oceans would be the key to tracing the history of changes to the ocean volume.

The exchange of water between the Earth's interior and the oceans would also change the ocean volume. Water is outgassed from the mantle via volcanism at mid-ocean ridges, ocean islands, arcs and back-arc basins. In addition, water is lost from the ocean at subduction zones via the subduction of slabs with water-bearing rocks such as sediments, altered oceanic crust and serpentinized lithospheric mantle. In subduction zones, the present ingassing water flux is estimated to be 9×10^{11}–1.9×10^{12} kg/yr (Rea and Ruff, 1996; Javoy, 1998). Present outgassing water fluxes at mid-ocean ridges and arcs are estimated to be 2×10^{11}–5×10^{11} kg and 1.9×10^{11}–1.5×10^{12} kg/yr (Bounama *et al.*, 2001; Hilton *et al.*, 2002), respectively. Therefore, present ingassing and outgassing water fluxes are almost comparable. However, if this balance were to change, the ocean volume would also change significantly. As an example, if either outgassing or ingassing of water were to completely terminate, about 5 times the volume of the present oceans would be transferred during 4.5 Ga. Moreover, the past outgassing and ingassing water fluxes are more difficult to estimate than the present values. Therefore, the exchange of water between the Earth's interior and oceans must have played a significant role in varying the volume of the Earth's oceans over time.

SUPPLY OF WATER ON EARTH

How did the Earth's water originate? Where did the

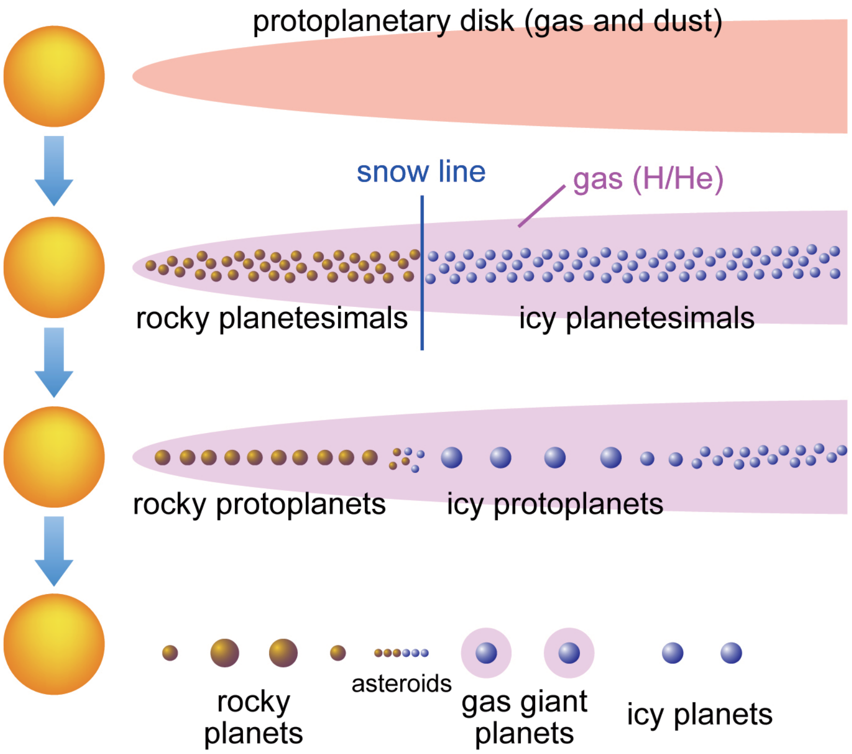

Fig. 4. Summary of planet formation in our solar system. A protoplanetary disk composed of gas and dust forms around the sun during star formation and planets are generated from this disk.

Earth's water come from? Here, three possible sources of water and supply mechanisms are reviewed. The process of water supply to Earth is closely related to the dynamics and chemistry of planet formation. First, the standard scenario of planet formation in our solar system is discussed, after which possible water sources and supply mechanisms are detailed.

Planet formation in our solar system

An outline of planet formation is provided in Fig. 4. Planets are formed in a protoplanetary disk around a star (otherwise known as a nebular disk, or simply nebula), composed of gas and dust. This disk is typically the by-product of star formation (e.g., Armitage, 2011). Terrestrial planets are made primarily from the dust component of the disk, and cores of Jovian planets are made from dust component (rock and ice), following which these cores gather the surrounding gas component (H_2 and He).

There are several stages involved in planet formation. The first stage is the formation of a large number of kilometer-sized bodies known as planetesimals by accretion of dust particles (Johansen *et al.*, 2007; Cuzzi *et al.*, 2008). Rocky planetesimals are formed inside the snow line (the boundary between condensation and vaporization of H_2O), and icy planetesimals are formed outside the snow line.

In the second stage, these planetesimals collide to produce protoplanets (e.g., Genda *et al.*, 2015a). Within the terrestrial planet region in our solar system (inside the snow line), a few tens of rocky Mars-sized protoplanets were formed (Wetherill, 1985; Kokubo and Ida, 1998, 2000; Kokubo *et al.*, 2006). Since solid materials were plentiful in the Jovian planet region (beyond the snow line) due to the presence of ice as well as rocky materials at large distances from the sun, a few huge icy protoplanets (often called cores) with masses several times those of Earth were formed (Tanaka and Ida, 1997; Kokubo and Ida, 2002; Kobayashi *et al.*, 2010). These huge protoplanets began to gravitationally attract the surrounding nebular gas in a runaway manner, eventually forming gas giant planets such as Jupiter and Saturn (Mizuno, 1980; Bodenheimer and Pollack, 1986; Pollack *et al.*, 1996; Ikoma *et al.*, 2000). Since the formation timescales of protoplanets beyond Saturn were much longer than the dissipation timescale of the nebular gas, these protoplanets did not capture a large amount of nebular gas. These icy protoplanets remain in the outer solar system forming Neptune and Uranus (Kokubo and Ida, 2002).

After the formation of Jupiter and Saturn, the final stage of terrestrial planet formation was characterized by collisions among rocky protoplanets (Chambers and

Wetherill, 1998; Agnor *et al.*, 1999; Kokubo and Genda, 2010; Genda *et al.*, 2012). These collisions are referred to as giant impacts. A giant impact (likely the last one) would be responsible for the formation of the Moon (Genda and Abe, 2003a; Canup, 2004; Ćuk and Stewart, 2012). In the terrestrial planet region, rocky planets such as Earth and Venus were formed during this stage. Because of the significant effects of the gravitational pulls of Jupiter and Saturn, terrestrial planets were not formed in the region between Mars and Jupiter, corresponding to the asteroid belt.

Planetesimals around Earth's orbit

If the building blocks of Earth (that is, planetesimals at approximately 1 AU) had contained small amounts of water, Earth would have originated as a wet planet. When a planet grows to lunar size (approximately 10^{23} kg), water will begin to degas from the planetesimals, and a significant steam atmosphere will form, followed by the formation of oceans (Matsui and Abe, 1986; Zahnle *et al.*, 1988). Therefore, Mars-sized protoplanets might have had surface oceans during terrestrial planet formation. Although giant impacts between these protoplanets were highly energetic events, significant amounts of water would have been able to survive these impacts (Genda and Abe, 2003b, 2005; Schlichting *et al.*, 2015).

The main question regarding this scenario is whether or not the planetesimals at approximately 1 AU contained water. According to the classical model of the protoplanetary disk, the temperature distribution in the disk is determined by the intensity of the solar radiation. In the case of our sun, the temperature of a body around 1 AU is estimated to have been 270 K (Hayashi, 1981). Since H_2O condenses to ice below 160–170 K at the gas pressure in a protoplanetary disk (approximately 1 Pa), the estimated temperature would have been too high to allow H_2O condensation. Therefore, planetesimals in the 1 AU region would have been dry. This disk model predicts that the snow line would have been located in the vicinity of the main asteroid belt, which is consistent with the observed water distribution of the main asteroid belt, in which C-type asteroids are distributed outside, and S-type asteroids inside (Gradie and Tedesco, 1982; Mothé-Diniz *et al.*, 2003).

Even if the snow line was located beyond 1 AU, there is the possibility that dry dust grains in the surrounding nebular gas could have absorbed water vapor on their surfaces (Stimpfl *et al.*, 2006), and so planetesimals made from these dust grains would contain small amounts of water. Izidoro *et al.* (2013) performed numerical simulations of terrestrial planet formation taking absorbed water into account, and showed that the contribution of absorbed water at approximately 1 AU was comparable to the water supplied by water-carrying asteroids beyond the snow line.

Moreover, recent protoplanetary disk models suggest that the snow line moves with time, and so may have been temporarily located inside 1 AU (Chiang and Goldreich, 1997; Sasselov and Lecar, 2000; Oka *et al.*, 2011). Here, it is assumed that the numerous dust particles blocked the transmission of light from the sun, such that the temperature distribution was lower than predicted by the classical disk model. During planetesimal formation, there were presumably numerous dust particles in the protoplanetary disk. If this is correct, icy planetesimals such as comets were inevitably formed in the vicinity of 1 AU (Machida and Abe, 2010). However, this is inconsistent with the very small amount of water currently on Earth. Additionally, if dusts and/or pebbles that originated beyond the snow line migrate inward in the gaseous disk (e.g., Ciesla, 2009; Simon *et al.*, 2011; Guillot *et al.*, 2014), these water-carrying particles could contribute to the Earth's building blocks, and could affect Earth's water budget in some degree. Therefore, the absorption of water onto the surface of dust grains, the timing of planetesimal formation, the evolution of the snow line, and radial mixing of small particles in a gaseous disk are the important factors in understanding the water content of Earth's building blocks, and further detailed research is required.

External sources of water (asteroids and comets)

As noted the amount of water in Earth's oceans is tiny compared with the Earth's mass. Therefore, even if the main building blocks of the planet were completely dry, the small addition of water-carrying objects onto Earth is sufficient to explain the present Earth's oceans. As an example, since carbonaceous chondrites and comets have approximately 5 wt% and 80 wt% water, respectively, the accretion of these objects to provide only 1% of Earth's mass is sufficient to have contributed the present ocean mass.

In the above scenario, Earth receives a tiny amount of volatile-carrying material after it is fully formed. This is referred to as the late-veneer hypothesis. Originally, this hypothesis was proposed to explain the excessive concentration of highly siderophile elements (HSEs) such as Ru, Rh, Pd, Re, Os, Ir, Pt and Au in the Earth's mantle (e.g., O'Neill and Palme, 1998). Additionally, the relative abundance of HSEs in the Earth's mantle is similar to that in CI chondrites, which belong to the carbonaceous chondrites group and are highly oxidized and thus free of metallic iron (e.g., O'Neill and Palme, 1998). Therefore, if a tiny amount of metallic iron-free objects such as CI chondrites impinged on the planet following the formation of the Earth's core, the excessive abundance of HSEs in the mantle would be explained. Because CI chondrites are rich in volatiles such as water and organic compounds,

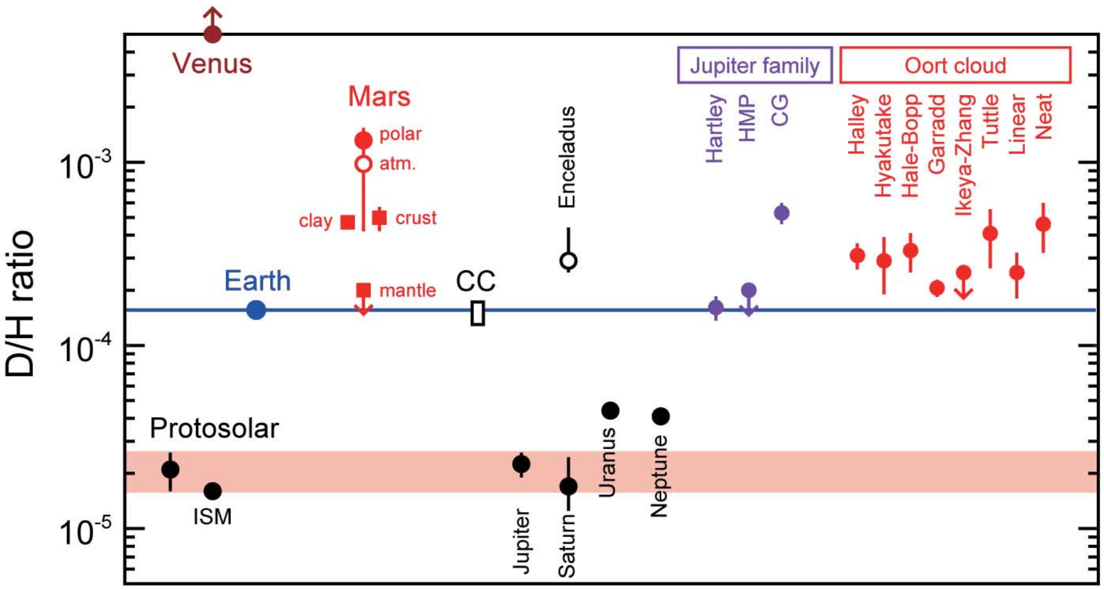

Fig. 5. D/H ratios in various objects in the solar system. All data are from Table 2. Legend: ISM = interstellar medium, CC = carbonaceous chondrites, HMP = Honda-Mrkos-Pajdušáková, CG = Churyumov-Gerasimenko.

the late veneer has also been considered a possible source of both the Earth's water and atmosphere.

Several mechanisms by which water-carrying objects can land on Earth from outside the terrestrial planet region have been proposed to date. Due to the gravitational disturbance of the asteroid belt region by Jupiter, it has been proposed that a large fraction of asteroids is ejected, some of which collide with Earth (Morbidelli *et al*., 2000; Raymond *et al*., 2009), which is consistent with the extreme mass depletion of the present asteroid belt. The migrations of Uranus and Neptune also result in the scattering of icy planetesimals (that is, comets), and a number of these bodies have entered the terrestrial planet region and are considered to be responsible for the late heavy bombardment around 4 Ga (Gomes *et al*., 2005) that would be recorded on lunar samples. It should be noted that the amount of water supplied to Earth by comets has likely been minor (Gomes *et al*., 2005), although comets have significantly affected the noble gas budget on Earth due to their high noble gas concentration (Dauphas, 2003).

Recently, in order to explain the reason why Mars is so small compared to Earth and Venus, Walsh *et al*. (2011) proposed the new "Grand Tack" scenario. Here, Jupiter migrates inward to the present orbital location of Mars and then returns to its present position during the planet formation stage, resulting in significant scattering of asteroids and icy planetesimals. In this scenario, water equal to 5–10 M_{oce} could have been supplied on Earth (O'Brien *et al*., 2014).

Another possible explanation for the low mass of Mars is that the planetesimal disk was originally truncated in the vicinity of Mars' orbit, such that a small Mars was naturally formed (Hansen, 2009). A truncated disk would be expected if the protoplanetary disk was a dynamic environment experiencing turbulence (Papaloizou and Nelson, 2003; Jin *et al*., 2008). Izidoro *et al*. (2014) carried out numerical simulations concerning terrestrial planets originating from a truncated disk and succeeded in making small terrestrial planets around 1.5 AU such as Mars. They also showed that Earth-sized planets in the vicinity of 1 AU would have received a sizeable quantity of water comparable to M_{oce}.

Nebular gas

If a planet is formed in a nebular gas, it will gravitationally attract the surrounding gas, generating a hydrogen-rich atmosphere (Hayashi *et al*., 1979; Mizuno *et al*., 1980; Pollack *et al*., 1996). When a protoplanet grows to lunar size (approximately 10^{23} kg) through the successive accretion of planetesimals, it begins to attract the surrounding nebular gas (Ikoma and Genda, 2006). If oxygen had been supplied from oxides in the Earth's interior to the hydrogen atmosphere, water molecules would have been formed on the planetary surface (Sasaki, 1990). The amount of water resulting from this process would depend on the oxides that were available in Earth. For example, iron oxides such as FeO and Fe_3O_4 can react with atmospheric hydrogen to produce a mass of water comparable to the mass of hydrogen that the planet attracts (Ikoma and Genda, 2006). Therefore, if a planet attracts a large quantity of hydrogen and a magma ocean is formed that can effectively react with the atmospheric

Table 2. D/H ratios in various objects in the solar system

Object	D/H ($\times10^{-4}$)	Reference
Protosolar	0.21 ± 0.05	Geiss and Gloeckler (1998)
Interstellar medium	0.16 ± 0.01	Linsky *et al.* (2006)
Earth (SMOW)	1.56	De Wit *et al.* (1980)
Venus	160 ± 20	Donahue *et al.* (1982)
Mars		
atmosphere	4.2–15.4	Fisher (2007)
estimated polar ice	~12.5	Villanueva *et al.* (2015)
clay mineral	4.7 ± 0.3	Mahaffy *et al.* (2015)
crust	4–6	Boctor *et al.* (2003), Greenwood *et al.* (2008)
mantle	<2	Usui *et al.* (2012)
Carbonaceous chondrites	1.3–1.7	Robert (2003)
Jupiter	0.225 ± 0.035	Lellouch *et al.* (2001)
Saturn	0.170 (+0.075/–0.045)	Lellouch *et al.* (2001)
Enceladus	2.9 (+1.5/–0.4)	Waite *et al.* (2009)
Uranus	0.44 ± 0.04	Feuchtgruber *et al.* (2013)
Neptune	0.41 ± 0.04	Feuchtgruber *et al.* (2013)
Jupiter family		
103P/Hartley 2	1.61 ± 0.24	Hartogh *et al.* (2011)
45P/Honda-Mrkos-Pajdušáková	<2.0	Lis *et al.* (2013)
67P/Churyumov-Gerasimenko	5.3 ± 0.7	Altwegg *et al.* (2015)
Oort cloud		
1P/Halley	3.1 ± 0.5	Balsiger *et al.* (1995), Eberhardt *et al.* (1995)
C/1996 B2 Hyakutake	2.9 ± 1.0	Bockelée-Morvan *et al.* (1998)
C/1995 O1 Hale-Bopp	3.3 ± 0.8	Meier *et al.* (1998)
C/2009 P1 Garradd	2.06 ± 0.22	Bockelée-Morvan *et al.* (2012)
153P/Ikeya-Zhang	<2.5	Biver *et al.* (2006)
8P/Tuttle	4.09 ± 1.45	Villanueva *et al.* (2009)
C/2002 T7 Linear	2.5 ± 0.7	Hutsemékers *et al.* (2008)
C/2001 Q4 Neat	4.6 ± 1.4	Weaver *et al.* (2008)

hydrogen, a mass of water comparable to that of the present Earth's oceans would result. However, the generation of water via gravitationally attracted nebular atmosphere is inconsistent with some geochemical constraints observed in Earth (Genda and Ikoma, 2008), as discussed below.

D/H ratios in our solar system and other geochemical constraints

The deuterium/hydrogen (D/H) ratios in various planetary materials can give exceptional insights into the origin of Earth's water. Figure 5 and Table 2 summarize the D/H ratios for the Earth's oceans (SMOW or standard mean ocean water), and of various planetary materials, including the water sources discussed above, such as carbonaceous chondrites, comets and protosolar nebula. Unfortunately, little is known about the D/H ratios for the Earth-forming planetesimals around 1 AU that could potentially have held water.

The D/H ratio for the Earth's oceans (SMOW) is 1.56×10^{-4} (de Wit *et al.*, 1980), a value that is similar to the D/H ratio for carbonaceous chondrites, which ranges from 1.3×10^{-4} to 1.7×10^{-4} in bulk samples (Robert, 2003). In contrast, the D/H ratios for comets are generally higher than that of SMOW. The long-period comets, which originate from the Oort cloud, all have values approximately twice that of SMOW (e.g., Drake and Righter, 2002). Recently, Hartogh *et al.* (2011) reported that the Jupiter family comet 103P/Hartley2 has a D/H value similar to that of Earth's oceans ($(1.61 \pm 0.24) \times 10^{-4}$) based on studies using the Herschel Space Observatory. Lis *et al.* (2013) reported that the upper limit of D/H for another Jupiter family comet, 45P/Honda-Mrkos-Pajdušáková, was 2.0×10^{-4}, suggesting that it also has an Earth's ocean-like D/H. However, Altwegg *et al.* (2015) reported that the D/H ratio for the Jupiter family comet 67P/Churyumov-Gerasimenko ($(5.3 \pm 0.7) \times 10^{-4}$) is more than twice the SMOW value, based on observations made by the Rosetta spacecraft. Therefore, there seems to be a wide range of D/H values within the Jupiter family comets.

The atmosphere of Venus has an extremely high D/H ratio ($(1.6 \pm 0.2) \times 10^{-2}$) (Donahue *et al.*, 1982), more than 100 times as high as the SMOW value, as the result of the extreme loss of water from this planet (e.g., Chassefière *et al.*, 2012). The D/H ratio for the Martian atmosphere changes significantly with the season, ranging from 4.2×10^{-4} to 15.4×10^{-4} (Fisher, 2007), while the D/H ratio for polar ice (the main reservoir of H_2O on Mars) is estimated to be approximately 1.2×10^{-3} (Villanueva *et al.*, 2015), which is about 8 times the SMOW value. The D/H ratio for hydrated crust and/or ground ice is estimated to be 2–3 times the SMOW value from analysis of martian meteorites (Usui *et al.*, 2015). Although the D/H ratios at the Martian surface (the atmosphere, polar ice, clay minerals and crust) are higher than the SMOW value, Usui *et al.* (2012) reported that the D/H ratio for the Martian mantle is comparable to the SMOW value. Assuming that the D/H ratio for the Martian mantle did not change during 4.5 Ga, the loss of water from Mars would increase the D/H ratio at the Martian surface (e.g., Kurokawa *et al.*, 2014), which is consistent with the recent in situ D/H analysis of ancient martian clays (~3 Ga) (Mahaffy *et al.*, 2015).

The D/H ratio for hydrogen in the protosolar nebula estimated from the solar composition is about 5 to 7 times lower than the SMOW value (Geiss and Gloeckler, 1998). Therefore, considering only D/H ratios, the Earth's oceans were likely supplied by carbonaceous chondrites or a combination of comets and protosolar nebula.

However, the origin of the Earth's water is not so simple, if other geochemical constraints are considered. For example, using the Os isotopic compositions of the mantle and of meteorites, Meisel *et al.* (1996, 2001) argued that the material of the late veneer is more similar to that of enstatite chondrites or ordinary chondrites than to that of carbonaceous chondrites, meaning that the main source of the late veneer was not carbonaceous chondrites.

Another example is the Xe/Kr ratio. The Xe/Kr ratios on Earth and also on Mars ($^{130}Xe/^{84}Kr = 8.4 \times 10^{-3}$ for Earth) are much lower than that in any of the chondrites (2.0×10^{-1} for carbonaceous chondrites). This is often termed the missing xenon problem (e.g., Ozima and Podosek, 2002). In order to solve this problem, a source with low Xe/Kr ratio is required. Since low-temperature experiments have shown that amorphous water ice can trap significant amounts of noble gases, and that the associated Xe/Kr ratio is very low (Owen *et al.*, 1992; Bar-Nun and Owen, 1998), comets likely played an important role in supplying noble gases to Earth. Dauphas (2003) has pointed out that the Earth's volatiles have the dual origins, coming from both carbonaceous chondrites and comets, based on noble gas abundances. Considering both D/H ratios and the other geochemical constraints, comets alone cannot account for the entire water budget of Earth, nor can meteorites and water from the primordial nebula. The most likely means by which water was delivered to Earth, which also agrees with dynamical models for the formation of terrestrial planets, involves asteroids from the outer part of the asteroid belt (e.g., Raymond *et al.*, 2005, 2009; Izidoro *et al.*, 2013, 2014; Walsh *et al.*, 2011).

SUMMARY

The presence of oceans distinguishes Earth from the other planets in our solar system. Water is also essential for the origin and evolution of life, the stability of the surface environment and the evolution of the planetary interior. Although the mass of the Earth's oceans is very small (0.023 wt%) compared to the mass of entire Earth, our planet potentially has water in its interior equal to many times the ocean mass. The estimated amount of water in the Earth's mantle ranges from 1 to 10 M_{oce}, and the Earth's core could contain up to the equivalent of 100 M_{oce} as hydrogen.

The Earth's oceans have existed throughout the planet's history. Planetary formation theory predicts that water supply on Earth occurred during its formation with much less after its formation. The origin of the Earth's water is still much in question, because of several uncertainties concerning the planet formation process, including the position of the snow line in the protoplanetary disk, the lifetime of the nebula gases, the timing of the formation of Jupiter and migration of planets. Fortunately, recent observations of the birth place of planets in extrasolar systems by large telescopes such as ALMA (Atacama Large Milimeter Array) have begun to reduce some of these uncertainties concerning the planet formation process (e.g., Wyatt, 2008; Dutrey *et al.*, 2014; Genda *et al.*, 2015b).

The number of detected extrasolar planets, which are planets around other stars, has increased as observation technologies have advanced, and now exceeds 1000 (Fabrycky *et al.*, 2014). Extrasolar planets with masses comparable to that of Earth have also been recently detected (Weiss and Marcy, 2014), and some are thought to be terrestrial or rocky planets based on the relationship between their mass and radius. In the near future, planets with oceans (and even life!) may be detected. By comparing Earth with these extrasolar planets, we will be able to understand more about the origin of Earth's water.

Acknowledgments—I thank anonymous reviewers for valuable comments and suggestions on the manuscript. I also thank guest editors Dr. J. I. Simon and Dr. T. Usui for having the opportunity to write this review paper. This work was supported by Research Grant 2015 of Kurita Water and Environment Foundation.

REFERENCES

Abe, Y. (2009) Habitable planets: Their formations and evolutions. *Planetary People* **18**, 194–215 (in Japanese).

Agnor, C. B., Canup, R. M. and Levison, H. F. (1999) On the character and consequences of large impacts in the late stage of terrestrial planet formation. *Icarus* **142**, 219–237.

Alfè, D., Gillan, M. J. and Price, G. D. (2007) Temperature and composition of the Earth's core. *Contemp. Phys.* **48**, 63–80.

Altwegg, K., Balsiger, H., Bar-Nun, A., Berthelier, J. J., Bieler, A., Bochsler, P., Briois, C., Calmonte, U., Combi, M., De Keyser, J., Eberhardt, P., Fiethe, B., Fuselier, S., Gasc, S., Gombosi, T. I., Hansen, K. C., Hässig, M., Jäckel, A., Kopp, E., Korth, A., LeRoy, L., Mall, U., Marty, B., Mousis, O., Neefs, E., Owen, T., Rème, H., Rubin, M., Sémon, T., Tzou, C.-Y., Waite, H. and Wurz, P. (2015) 67P/Churyumov-Gerasimenko, a Jupiter family comet with a high D/H ratio. *Science* **347**, 1261952.

Anders, E. and Ebihara, M. (1982) Solar-system abundances of the elements. *Geochim. Cosmochim. Acta* **46**, 2363–2380.

Appel, P. W. U., Fedo, C. M., Moorbath, S. and Myers, J. S. (1998) Recognizable primary volcanic and sedimentary features in a low-strain domain of the highly deformed, oldest known (≈3.7–3.8 Gyr) Greenstone Belt, Isua, West Greenland. *Terra Nova* **10**, 57–62.

Armitage, P. J. (2011) Dynamics of protoplanetary disks. *Annu. Rev. Astron. Astrophys.* **49**, 195–236.

Balsiger, H., Altwegg, K. and Geiss, J. (1995) D/H and ^{18}O/^{16}O ratio in the hydronium ion and in neutral water from in situ ion measurements in comet Halley. *J. Geophys. Res.* **100**, 5827–5834.

Bar-Nun, A. and Owen, T. (1998) Trapping of gases in water ice and consequences to comets and the atmospheres of the inner planets. *Solar System Ices* (Schmitt, B., De Bergh, C. and Festou, M., eds.), 353–366, Kluwer, Dordrecht.

Bell, D. R., Rossman, G. R., Maldener, J., Endisch, D. and Rauch, F. (2003) Hydroxide in olivine: A quantitative determination of the absolute amount and calibration of the IR spectrum. *J. Geophys. Res.* **108**, 2105.

Birch, F. (1952) Elasticity and constitution of the Earth's interior. *J. Geophys. Res.* **57**, 227–286.

Biver, N., Bockelée-Morvan, D., Crovisier, J., Lis, D. C., Moreno, R., Colom, P., Henry, F., Herpin, F., Paubert, G. and Womack, M. (2006) Radio wavelength molecular observations of comets C/1999 T1 (McNaught-Hartley), C/2001 A2 (LINEAR), C/2000 WM1 (LINEAR) and 153P/Ikeya-Zhang. *Astron. Astrophys.* **449**, 1255–1270.

Bockelée-Morvan, D., Gautier, D., Lis, D. C., Young, K., Keene, J., Philllips, T., Owen, T., Crovisier, J., Goldsmith, P. F., Bergin, E. A., Despois, D. and Wootten, A. (1998) Deuterated water in comet C/1996 B2 (Hyakutake) and its implications for the origin of comets. *Icarus* **133**, 147–162.

Bockelée-Morvan, D., Biver, N., Swinyard, B., de Val-Borro, M., Crovisier, J., Hartogh, P., Lis, D. C., Moreno, R., Szutowicz, S., Lellouch, E., Emprechtinger, M., Blake, G. A., Courtin, R., Jarchow, C., Kidger, M., Küppers, M., Rengel, M., Davis, G. R., Fulton, T., Naylor, D., Sidher, S. and Walker, H. (2012) Herschel measurements of the D/H and ^{16}O/^{18}O ratios in water in the Oort-cloud comet C/2009 P1 (Garradd). *Astron. Astrophys.* **544**, L15 (6 pp.).

Boctor, N. Z., Alexander, C. M. O'D., Wang, J. and Hauri, E. (2003) The sources of water in Martian meteorites: Clues from hydrogen isotopes. *Geochim. Cosmochim. Acta* **67**, 3971–3989.

Bodenheimer, P. and Pollack, J. B. (1986) Calculations of the accretion and evolution of giant planets: The effects of solid cores. *Icarus* **67**, 391–408.

Bounama, C., Franck, S. and von Bloh, W. (2001) The fate of Earth's ocean. *Hydrol. Earth Syst. Sci.* **5**, 569–575.

Bowring, S. A. and Williams, I. S. (1999) Priscoan (4.00–4.03 Ga) orthogneisses from northwestern Canada. *Contrib. Mineral. Petrol.* **134**, 3–16.

Bowring, S. A., Williams, I. S. and Compston, W. (1989) 3.96 Ga gneisses from the Slave province, Northwest Territories, Canada. *Geology* **17**, 971–975.

Boyce, J. W., Liu, Y., Rossman, G. R., Guan, Y., Eiler, J. M., Stolper, E. M. and Taylor, L. A. (2010) Lunar apatite with terrestrial volatile abundances. *Nature* **466**, 466–469.

Canup, R. M. (2004). Dynamics of lunar formation. *Annu. Rev. Astron. Astrophys.* **42**, 441–475.

Chambers, J. E. and Wetherill, G. W. (1998) Making the terrestrial planets: N-body integrations of planetary embryos in three dimensions. *Icarus* **136**, 304–327.

Chassefière, E., Wieler, R., Marty, B. and Leblanc, F. (2012) The evolution of Venus: Present state of knowledge and future exploration. *Planet. Space Sci.* **63–64**, 15–23.

Chiang, E. I. and Goldreich, P. (1997) Spectral energy distributions of T Tauri stars with passive circumstellar disks. *Astrophys. J.* **490**, 368–376.

Ciesla, F. J. (2009) Two-dimensional transport of solids in viscous protoplanetary disks. *Icarus* **200**, 655–671.

Ćuk, M. and Stewart, S. T. (2012) Making the Moon from a fast-spinning Earth: A giant impact followed by resonant despinning. *Science* **338**, 1047–1052.

Cuzzi, J. N., Hogan, R. C. and Shariff, K. (2008) Toward planetesimals: Dense chondrule clumps in the protoplanetary nebula. *Astrophy. J.* **687**, 1432–1447.

Dauphas, N. (2003) The dual origin of the terrestrial atmosphere. *Icarus* **165**, 326–339.

De Wit, J. C., Van der Straaten, C. M. and Mook, W. G. (1980) Determination of the absolute hydrogen isotopic ratio of V-SMOW and SLAP. *Geostand. Newsl.* **4**, 33–36.

Dewaele, A., Loubeyre, P., Occelli, F., Mezouar, M., Dorogokupets, P. I. and Torrent, M. (2006) Quasihydrostatic equation of state of iron above 2 Mbar. *Phys. Rev. Lett.* **97**, 215504.

Dixon, J. E, Clague, D. A., Wallace, P. and Poreda, R. (1997) Volatiles in alkalic basalts from the North Arch Volcanic Field, Hawaii: Extensive degassing of deep submarine-erupted alkalic series lavas. *J. Petrol.* **38**, 911–939.

Dixon, J. E., Leist, L., Langmuir, C. and Schilling, J.-G. (2002) Recycled dehydrated lithosphere observed in plume-influenced mid-ocean-ridge basalt. *Nature* **420**, 385–389.

Donahue, T. M., Hoffman, J. H., Hodges, R. R. and Watson, A. J. (1982) Venus was wet: A measurement of the ratio of deuterium to hydrogen. *Science* **216**, 630–633.

Drake, M. J. and Righter, K. (2002) Determining the composition of the Earth. *Nature* **416**, 39–44.

Dubrovinsky, L. S., Saxena, S. K., Tutti, F. and Rekhi, S. (2000) *In situ* X-ray study of thermal expansion and phase transition of iron at multimegabar pressure. *Phys. Rev. Lett.* **84**, 1720–1723.

Dutrey, A., Di Folco, E., Guilloteau, S., Boehler, Y., Bary, J., Beck, T., Beust, H., Chapillon, E., Gueth, F., Huré, J.-M., Pierens, A., Piétu, V., Simon, M. and Tang, Y.-W. (2014) Possible planet formation in the young, low-mass, multiple stellar system GG Tau A. *Nature* **514**, 600–602.

Eberhardt, P., Reber, M., Krankowsky, D. and Hodges, R. R. (1995) The D/H and $^{18}O/^{16}O$ ratios in water from comet P/Halley. *Astron. Astrophys.* **302**, 301–316.

Fabrycky, D. C., Lissauer, J. J., Ragozzine, D., Rowe, J. F., Steffen, J. H., Agol, E., Barclay, T., Batalha, N., Borucki, W., Ciardi, D. R., Ford, E. B., Gautier, T. N., Geary, J. C., Holman, M. J., Jenkins, J. M., Li, J., Morehead, R. C., Morris, R. L., Shporer, A., Smith, J. C., Still, M. and Van Cleve, J. V. (2014) Architecture of *Kepler*'s multi-transiting systems. II. New investigations with twice as many candidates. *Astrophy. J.* **790**, 146 (12 pp.).

Feuchtgruber, H., Lellouch, E., Orton, G., de Graauw, T., Vandenbussche, B., Swinyard, B., Moreno, R., Jarchow, C., Billebaud, F., Cavalié, T., Sidher, S. and Hartogh, P. (2013) The D/H ratio in the atmospheres of Uranus and Neptune from *Herschel*-PACS observations. *Astron. Astrophys.* **551**, A126.

Fisher, D. A. (2007) Mars' water isotope (D/H) history in the strata of the North Polar Cap: Inferences about the water cycle. *Icarus* **187**, 430–441.

Geiss, J. and Gloeckler, G. (1998) Abundances of deuterium and helium-3 in the protosolar cloud. *Space Sci. Rev.* **84**, 239–250.

Genda, H. and Abe, Y. (2003a) Modification of a proto-lunar disk by hydrodynamic escape of silicate vapor. *Earth Planets Space* **55**, 53–57.

Genda, H. and Abe, Y. (2003b) Survival of a proto-atmosphere through the stage of giant impacts: the mechanical aspects. *Icarus* **164**, 149–162.

Genda, H. and Abe, Y. (2005) Enhanced atmospheric loss on protoplanets at the giant impact phase in the presence of oceans. *Nature* **433**, 842–844.

Genda, H. and Ikoma, M. (2008) Origin of the ocean on the Earth: Early evolution of water D/H in a hydrogen-rich atmosphere. *Icarus* **194**, 42–52.

Genda, H., Kokubo, E. and Ida, S. (2012) Merging criteria for giant impacts of protoplanets. *Astrophy. J.* **744**, 137 (8 pp.).

Genda, H., Fujita, T., Kobayashi, H., Tanaka, H. and Abe, Y. (2015a) Resolution dependence of disruptive collisions between planetesimals in the gravity regime. *Icarus* **262**, 58–66.

Genda, H., Kobayashi, H. and Kokubo, E. (2015b) Warm debris disks produced by giant impacts during terrestrial planet formation. *Astrophys. J.* **810**, 136 (8 pp.).

Gomes, R., Levison, H. F., Tsiganis, K. and Morbidelli, A. (2005) Origin of the cataclysmic Late Heavy Bombardment period of the terrestrial planets. *Nature* **435**, 466–469.

Goody, R. (1995) *Principles of Atmospheric Physics and Chemistry*. Oxford Univ. Press, New York.

Gradie, J. and Tedesco, E. (1982) Compositional structure of the asteroid belt. *Science* **216**, 1405–1407.

Green, D. W. and Perry, R. H. (2008) *Perry's Chemical Engineers' Handbook*. McGraw-Hill Professional, 2400 pp.

Greenwood, J. P., Itoh, S., Sakamoto, N., Vicenzi, E. P. and Yurimoto, H. (2008) Hydrogen isotope evidence for loss of water from Mars through time. *Geophys. Res. Lett.* **35**, L05203.

Guillot, T. (2005) The interiors of giant planets: Models and outstanding questions. *Annu. Rev. Earth Planet. Sci.* **33**, 493–530.

Guillot, T., Ida, S. and Ormel, C. W. (2014) On the filtering and processing of dust by planetesimals. I. Derivation of collision probabilities for non-drifting planetesimals. *Astron. Astrophys.* **572**, A72.

Hamano, K., Abe, Y. and Genda, H. (2013) Emergence of two types of terrestrial planet on solidification of magma ocean. *Nature* **497**, 607–610.

Hamano, Y. and Ozima, M. (1978) Earth-atmosphere evolution model based on Ar isotopic data. *Terrestrial Rare Gases* (Alexander, E. C., Jr. and Ozima, M., eds.), 155–171, Cent. Acad. Publ., Tokyo.

Hansen, B. M. S. (2009) Formation of the terrestrial planets from a narrow annulus. *Astrophys. J.* **703**, 1131–1140.

Hartogh, P., Lis, D. C., Bockelée-Morvan, D., de Val-Borro, M., Biver, N., Küppers, M., Emprechtinger, M., Bergin, E. A., Crovisier, J., Rengel, M., Moreno, R., Szutowicz, S. and Blake, G. A. (2011) Ocean-like water in the Jupiter-family comet 103P/Hartley 2. *Nature* **478**, 218–220.

Hauri, E. H. (2002) SIMS analysis of volatiles in silicate glasses, 2: isotopes and abundances in Hawaiian melt inclusions. *Chem. Geol.* **183**, 115–141.

Hauri, E. H., Weinreich, T., Saal, A. E., Rutherford, M. C. and Van Orman, J. A. (2011) High pre-eruptive water contents preserved in lunar melt inclusions. *Science* **333**, 213–215.

Hayashi, C. (1981) Structure of the solar nebula, growth and decay of magnetic fields and effects of magnetic and turbulent viscosities on the nebula. *Prog. Theor. Phys. Suppl.* **70**, 35–53.

Hayashi, C., Nakazawa, K. and Mizuno, H. (1979) Earth's melting due to the blanketing effect of the primordial dense atmosphere. *Earth Planet. Sci. Lett.* **43**, 22–28.

Hilton, D. R., Fischer, T. P. and Marty, B. (2002) Noble gases and volatile recycling at subduction zones. *Rev. Min. Geochem.* **47**, 319–370.

Hirose, K., Labrosse, S. and Hernlund, J. (2013) Composition and state of the core. *Annu. Rev. Earth Planet. Sci.* **41**, 657–691.

Hirschmann, M. M. (2006) Water, melting, and the deep Earth H_2O cycle. *Annu. Rev. Earth Planet. Sci.* **34**, 629–653.

Huang, X., Xu, Y. and Karato, S. (2005) Water content in the transition zone from electrical conductivity of wadsleyite and ringwoodite. *Nature* **434**, 746–749.

Hui, H., Peslier, A. H., Zhang, Y. and Neal, C. R. (2013) Water in lunar anorthosites and evidence for a wet early Moon. *Nat. Geosci*. **6**, 177–180.

Hunten, D. M. (1973). The escape of light gases from planetary atmospheres. *J. Atmos. Sci*. **30**, 1481–1494.

Hunten, D. M. (1982) Thermal and nonthermal escape mechanisms for terrestrial bodies. *Planet. Space Sci*. **30**, 773–783.

Hutsemékers, D., Manfroid, J., Jehin, E., Zucconi, J.-M. and Arpigny, C. (2008) The $^{16}OH/^{18}OH$ and OD/OH isotope ratios in comet C/2002 T7 (LINEAR). *Astron. Astrophys*. **490**, L31–L43.

Ikoma, M. and Genda, H. (2006) Constraints on the mass of a habitable planet with water of nebular origin. *Astrophy. J.* **648**, 696–706.

Ikoma, M., Nakazawa, K. and Emori, H. (2000) Formation of giant planets: Dependences on core accretion rate and grain opacity. *Astrophys. J.* **537**, 1013–1025.

Inoue, T., Yurimoto, H. and Kudoh, Y. (1995) Hydrous modified spinel, $Mg_{1.75}SiH_{0.5}O_4$: A new water reservoir in the mantle transition region, *Geophys. Res. Lett*. **22**, 117–120.

Inoue, T., Weidner, D. J., Northrup, P. A. and Parise, J. B. (1998) Elastic properties of hydrous ringwoodite (γ-phase) in Mg_2SiO_4. *Earth Planet. Sci. Lett*. **160**, 107–113.

Izidoro, A., de Souza Torres, K., Winter, O. C. and Haghighipour, N. (2013) A compound model for the origin of Earth's water. *Astrophys. J.* **767**, 54 (20 pp.).

Izidoro, A., Haghighipour, N., Winter, O. C. and Tsuchida, M. (2014) Terrestrial planet formation in a protoplanetary disk with a local mass depletion: A successful scenario for the formation of Mars. *Astrophys. J.* **782**, 31 (20 pp.).

Javoy, M. (1998) The birth of the Earth's atmosphere: The behaviour and fate of its major elements. *Chem. Geol*. **147**, 11–25.

Jin, L., Arnett, W. D., Sui, N. and Wang, X. (2008) An interpretation of the anomalously low mass of Mars. *Astrophys. J.* **674**, L105–L108.

Johansen, A., Oishi, J. S., Mac Low, M.-M., Klahr, H., Henning, T. and Youdin, A. (2007) Rapid planetesimal formation in turbulent circumstellar disks. *Nature* **448**, 1022–1025.

Karato, S. (2013) Geophysical constraints on the water content of the lunar mantle and its implications for the origin of the Moon. *Earth Planet. Sci. Lett*. **384**, 144–153.

Kasting, J. F. (1988) Runaway and moist greenhouse atmospheres and the evolution of Earth and Venus. *Icarus* **74**, 472–494.

Kasting, J. F., Whitmire, D. P. and Reynolds, R. T. (1993) Habitable zones around main sequence stars. *Icarus* **101**, 108–128.

Kobayashi, H., Tanaka, H., Krivov, A. V. and Inaba, S. (2010) Planetary growth with collisional fragmentation and gas drag. *Icarus* **209**, 836–847.

Kodama, T., Genda, H., Abe, Y. and Zahnle, K. J. (2015) Rapid water loss can extend the lifetime of the planetary habitability. *Astrophys. J.* **812**, 165 (11 pp.).

Koga, K., Hauri, E., Hirschmann, M. M. and Bell, D. (2003) Hydrogen concentration analyses using SIMS and FTIR: comparison and calibration for nominally anhydrous minerals. *Geochem. Geophys. Geosyst*. **4**, 1019.

Kohlstedt, D. L., Keppler, H. and Rubie, D. C. (1996) Solubility of water in the α, β and γ phases of $(Mg,Fe)_2SiO_4$. *Contrib. Mineral. Petrol*. **123**, 345–357.

Kokubo, E. and Genda, H. (2010) Formation of terrestrial planets from protoplanets under a realistic accretion condition. *Astrophy. J. Lett*. **714**, L21–L25.

Kokubo, E. and Ida, S. (1998) Oligarchic growth of protoplanets. *Icarus* **131**, 171–178.

Kokubo, E. and Ida, S. (2000) Formation of protoplanets from planetesimals in the solar nebula. *Icarus* **143**, 15–27.

Kokubo, E. and Ida, S. (2002) Formation of protoplanet systems and diversity of planetary systems. *Astrophys. J.* **581**, 666–680.

Kokubo, E., Kominami, J. and Ida, S. (2006) Formation of terrestrial planets from protoplanets. I. Statistics of basic dynamical properties. *Astrophys. J.* **642**, 1131–1139.

Kremenetsky, A. A. and Ovchinnikov, L. N. (1986) The Precambrian continental crust: Its structure, composition and evolution as revealed by deep drilling in the U.S.S.R. *Precambrian Res*. **33**, 11–43.

Kurokawa, H., Sato, M., Ushioda, M., Matsuyama, T., Moriwaki, R., Dohm, J. M. and Usui, T. (2014) Evolution of water reservoirs on Mars: Constraints from hydrogen isotopes in martian meteorites. *Earth Planet. Sci. Lett*. **394**, 179–185.

Kuvshinov, A. and Olsen, N. (2006) A global model of mantle conductivity derived from 5 years of CHAMP, Ørsted, and SAC-C magnetic data. *Geophys. Res. Lett*. **33**, L18301.

Lellouch, E., Bézard, B., Fouchet, T., Feuchtgruber, H., Encrenaz, T. and de Graauw, T. (2001) The deuterium abundance in Jupiter and Saturn from ISO-SWS observations. *Astron. Astrophys*. **670**, 610–622.

Linsky, J. L., Draine, B. T., Moos, H. W., Jenkins, E. B., Wood, B. E., Oliveira, C., Blair, W. P., Friedman, S. D., Gry, C., Knauth, D., Kruk, J. W., Lacour, S., Lehner, N., Redfield, S., Shull, J. M., Sonneborn, G. and Williger, G. M. (2006) What is the total deuterium abundance in the local galactic disk? *Astrophys. J.* **647**, 1106–1124.

Lis, D. C., Biver, N., Bockelée-Morvan, D., Hartogh, P., Bergin, E. A., Blake, G. A., Crovisier, J., de Val-Borro, M., Jehin, E., Küppers, M., Manfroid, J., Moreno, R., Rengel, M. and Szutowicz, S. (2013) A *Herschel* study of D/H in water in the Jupiter-family comet 45P/Honda-Mrkos-Pajdušáková and prospects for D/H measurements with CCAT. *Astrophy. J. Lett*. **774**, L3 (5 pp.).

Litasov, K., Ohtani, E., Langenhorst, F., Yurimoto, H., Kubo, T. and Kondo, T. (2003) Water solubility in Mg-perovskites and water storage capacity in the lower mantle. *Earth Planet. Sci. Lett*. **211**, 189–203.

Losq, E. L., Cody, G. D. and Mysen, B. O. (2015) Alkali influence on the water speciation and the environment of protons in silicate glasses revealed by ^1H MAS NMR spectroscopy. *Am. Mineral*. **100**, 466–473.

Machida, R. and Abe, Y. (2010) Terrestrial planet formation through accretion of sublimating icy planetesimals in a cold nebula. *Astrophy. J.* **716**, 1252–1262.

Mahaffy, P. R., Webster, C. R., Stern, J. C., Brunner, A. E., Atreya, S. K., Conrad, P. G., Domagal-Goldman, S., Eigenbrode, J. L., Flesch, G. J., Christensen, L. E., Franz,

H. B., Freissinet, C., Glavin, D. P., Grotzinger, J. P., Jones, J. H., Leshin, L. A., Malespin, C., McAdam, A. C., Ming, D. W., Navarro-Gonzalez, R., Niles, P. B., Owen, T., Pavlov, A. A., Steele, A., Trainer, M. G., Williford, K. H., Wray, J. J. and the MSL Science Team (2015) The imprint of atmospheric evolution in the D/H of Hesperian clay minerals on Mars. *Science* **347**, 412–414.

Marchi, S., Bottke, W. F., Elkins-Tanton, L. T., Bierhaus, M., Wuennemann, K., Morbidelli, A. and Kring, D. A. (2014) Widespread mixing and burial of Earth's Hadean crust by asteroid impacts. *Nature* **511**, 578–582.

Maruyama, S. and Komiya, T. (2011) The oldest pillow lavas, 3.8–3.7 Ga from the Isua supracrustal belt, SW Greenland: Plate tectonics had already begun by 3.8 Ga. *J. Geography* **120**, 869–876.

Maruyama, S., Ikoma, M., Genda, H., Hirose, K., Yokoyama, T. and Santosh, M. (2013) The naked planet Earth: Most essential pre-requisite for the origin and evolution of life. *Geoscience Frontiers* **4**, 141–165.

Matsui, T. and Abe, Y. (1986) Evolution of an impact-induced atmosphere and magma ocean on the accreting Earth. *Nature* **319**, 303–305.

McCubbin, F. M., Steele, A., Hauri, E. H., Nekvasil, H., Yamashita, S. and Hemley, R. J. (2010) Nominally hydrous magmatism on the Moon. *Proc. Natl. Acad. Sci. USA* **107**, 11223–11228.

Meier, R., Owen, T. C., Matthews, H. E., Jewitt, D. C., Bockelée-Morvan, D., Biver, N., Crovisier, J. and Gautier, D. (1998) A determination of the DHO/H_2O ratio in Comet C/1995 O1 (Hale-Bopp). *Science* **279**, 842–844.

Meisel, T., Walker, R. J. and Morgan, J. W. (1996) The osmium isotopic composition of the Earth's primitive upper mantle. *Nature* **383**, 517–520.

Meisel, T., Walker, R. J., Irving, A. J. and Lorand, J.-P. (2001) Osmium isotopic compositions of mantle xenoliths: A global perspective. *Geochim. Cosmochim. Acta* **65**, 1311–1323.

Michael, P. J. (1988) The concentration, behavior and storage of H_2O in the suboceanic upper mantle: Implications for mantle metasomatism. *Geochim. Cosmochim. Acta* **52**, 555–566.

Mizuno, H. (1980) Formation of the giant planets. *Prog. Theor. Phys*. **64**, 544–557.

Mizuno, H., Nakazawa, K. and Hayashi, C. (1980) Dissolution of the primordial rare gases into the molten Earth's material. *Earth Planet. Sci. Lett*. **50**, 202–210.

Mojzsis, S. J., Harrison, T. M. and Pidgeon, R. T. (2001) Oxygen-isotope evidence from ancient zircons for liquid water at the Earth's surface 4,300 Myr ago. *Nature* **409**, 178–181.

Morbidelli, A., Chambers, J., Lunine, J. I., Petit, J. M., Robert, F., Valsecchi, G. B. and Cyr, K. E. (2000) Source regions and timescales for the delivery of water to the Earth. *Meteorit. Planet. Sci*. **35**, 1309–1320.

Mothé-Diniz, T., Carvano, J. M. and Lazzaro, D. (2003) Distribution of taxonomic classes in the main belt of asteroids. *Icarus* **162**, 10–21.

Mottl, M. J., Glazer, B. T., Kaiser, R. I. and Meech, K. J. (2007) Water and astrobiology. *Chemie der Erde* **67**, 253–282.

Murakami, M., Hirose, K., Yurimoto, H., Nakashima, S. and Takafuji, N. (2002) Water in Earth's lower mantle. *Science* **295**, 1885–1887.

Nakajima, S., Hayashi, Y. and Abe, Y. (1992) A study on the "runaway greenhouse effect" with a one-dimensional radiate-convective equilibrium model. *J. Atmos. Sci*. **49**, 2256–2266.

Narygina, O., Dubrovinsky, L. S., McCammon, C. A., Kurnosov, A., Kantor, I. Y., Prakapenka, V. B. and Dubrovinskaia, N. A. (2011) X-ray diffraction and Mössbauer spectroscopy study of *fcc* iron hydride FeH at high pressures and implications for the composition of the Earth's core. *Earth Planet. Sci. Lett*. **307**, 409–414.

Nomura, R., Hirose, K., Uesugi, K., Ohishi, Y., Tsuchiyama, A., Miyake, A. and Ueno, Y. (2014) Low core-mantle boundary temperature inferred from the solidus of pyrolite. *Science* **343**, 522–525.

O'Brien, D. P., Walsh, K. J., Morbidelli, A., Raymond, S. N. and Mandell, A. M. (2014) Water delivery and giant impacts in the 'Grand Tack' scenario. *Icarus* **239**, 74–84.

Oka, A., Nakamoto, T. and Ida, S. (2011) Evolution of snow line in optically thick protoplanetary disks: Effects of water ice opacity and dust grain size. *Astrophy. J*. **738**, 141 (11 pp.).

Okuchi, T. (1997) Hydrogen partitioning into molten iron at high pressure: Implications for Earth's core. *Science* **278**, 1781–1784.

Olsen, N. (1999) Long-period (30days–1year) electromagnetic sounding and the electrical conductivity of the lower mantle beneath Europe. *Geophys. J. Int*. **138**, 179–187.

O'Neill, H. St. C. and Palme, H. (1998) Composition of the silicate Earth: Implications for accretion and core formation. *The Earth's Mantle: Composition, Structure, and Evolution* (Jackson, I., ed.), 3–126, Cambridge Univ. Press, Cambridge.

Owen, T., Bar-Nun, A. and Kleinfeld, I. (1992) Possible cometary origin of heavy noble gases in the atmospheres of Venus, Earth and Mars. *Nature* **358**, 43–46.

Ozima, M. and Podosek, F. A. (2002) *Noble Gas Geochemistry*. 2nd ed., Cambridge Univ. Press, Cambridge, 286 pp.

Papaloizou, J. C. B. and Nelson, R. P. (2003) The interaction of a giant planet with a disc with MHD turbulence—I. The initial turbulent disc models. *Mon. Not. Roy. Astron. Soc*. **339**, 983–992.

Pearson, D. G., Brenker, F. E., Nestola, F., McNeill, J., Nasdala, L., Hutchison, M. T., Matveev, S., Mather, K., Silversmit, G., Schmitz, S., Vekemans, B. and Vincze, L. (2014) Hydrous mantle transition zone indicated by ringwoodite included within diamond. *Nature* **507**, 221–224.

Petigura, E. A. and Marcy, G. W. (2011) Carbon and oxygen in nearby stars: Keys to protoplanetary disk chemistry. *Astrophys. J*. **735**, 41 (10 pp.).

Poirier, J.-P. (1994) Light elements in the Earth's core: A critical review. *Phys. Earth Planet. Inter*. **85**, 319–337.

Pollack, J. B., Hubickyj, O., Bodenheimer, P., Lissauer, J. J., Podolak, M. and Greenzweig, Y. (1996) Formation of the giant planets by concurrent accretion of solids and gas. *Icarus* **124**, 62–85.

Pope, E. C., Bird, D. K. and Rosing, M. T. (2012) Isotope composition and volume of Earth's early oceans. *Proc. Natl. Acad. Sci. USA* **109**, 4371–4376.

Püthe, C. and Kuvshinov, A. (2013) Determination of the 1-D distribution of electrical conductivity in Earth's mantle from *Swarm* satellite data. *Earth Planets Space* **65**, 1233–1237.

Raymond, S. N., Quinn, T. and Lunine, J. I. (2005) Terrestrial planet formation in disks with varying surface density profiles. *Astrophys. J*. **632**, 670–676.

Raymond, S. N., O'Brien, D. P., Morbidelli, A. and Kaib, N. A. (2009) Building the terrestrial planets: Constrained accretion in the inner Solar System. *Icarus* **203**, 644–662.

Rea, D. K. and Ruff, L. J. (1996) Composition and mass flux of sediment entering the world's subduction zones: Implications for global sediment budgets, great earthquakes, and volcanism. *Earth Planet. Sci. Lett*. **140**, 1–12.

Regenauer-Lieb, K., Yuen, D. A. and Branlund, J. (2001) The initiation of subduction: Criticality by addition of water? *Science* **294**, 578–580.

Robert, F. (2003) The D/H ratio in chondrites. *Space Sci. Rev*. **106**, 87–101.

Saal, A. E., Hauri, E. H., Cascio, M. L., Van Orman, J. A., Rutherford, M. C. and Cooper, R. F. (2008) Volatile content of lunar volcanic glasses and the presence of water in the Moon's interior. *Nature* **454**, 192–195.

Sakamaki, K., Takahashi, E., Nakajima, Y., Nishihara, Y., Funakoshi, K., Suzuki, T. and Fukai, Y. (2009) Melting phase relation of FeH_x up to 20 GPa: Implication for the temperature of the Earth's core. *Phys. Earth Planet. Inter*. **174**, 192–201.

Sasaki, S. (1990) The primary solar-type atmosphere surrounding the accreting Earth: H_2O-induced high surface temperature. *Origin of the Earth* (Newsom, H. E. and Jones, J. H., eds.), 195–209, Oxford Univ. Press, New York.

Sasselov, D. D. and Lecar, M. (2000) On the snow line in dusty protoplanetary disks. *Astrophys. J*. **528**, 995–998.

Schlichting, H., Sari, R. and Yalinewich, A. (2015) Atmospheric mass loss during planet formation: The importance of planetesimal impacts. *Icarus* **247**, 81–94.

Simon, J. I., Hutcheon, I. D., Simon, S. B., Matzel, J. E. P., Ramon, E. C., Weber, P. K., Grossman, L. and DePaolo, D. J. (2011) Oxygen isotope variations at the margin of a CAI records circulation within the solar nebula. *Science* **331**, 1175–1178.

Staudacher, T. and Allègre, C. J. (1982) Terrestrial xenology. *Earth Planet. Sci. Lett*. **60**, 389–406.

Stimpfl, M., Walker, A. M., Drake, M. J., de Leeuw, N. H. and Deymier, P. (2006) An ångström-sized window on the origin of water in the inner solar system: Atomistic simulation of adsorption of water on olivine. *J. Cryst. Growth* **294**, 83–95.

Tajika, E. and Matsui, T. (1992) Evolution of terrestrial proto-CO_2 atmosphere coupled with thermal history of the earth. *Earth Planet. Sci. Lett*. **113**, 251–266.

Tanaka, H. and Ida, S. (1997) Distribution of planetesimals around a protoplanet in the nebula gas. II. Numerical simulations. *Icarus* **125**, 302–316.

Tomiyasu, F., Ueno, Y. and Dewit, M. (2013) Hydrogen isotopic composition of Earth's early ocean estimated from archean MORB in Barberton Greenstone Belt. *Goldschmidt 2013 Conference Abstract*, 2340.

Usui, T., Alexander, C. M. O'D., Wang, J., Simon, J. I. and Jones, J. H. (2012) Origin of water and mantle-crust interactions on Mars inferred from hydrogen isotopes and volatile element abundances of olivine-hosted melt inclusions of primitive shergottites. *Earth Planet. Sci. Lett*. **357–358**, 119–129.

Usui, T., Alexander, C. M. O'D., Wang, J., Simon, J. I. and Jones, J. H. (2015) Meteoritic evidence for a previously unrecognized hydrogen reservoir on Mars. *Earth Planet. Sci. Lett*. **410**, 140–151.

Valley, J. W., Cavosie, A. J., Ushikubo, T., Reinhard, D. A., Lawrence, D. F., Larson, D. J., Clifton, P. H., Kelly, T. F., Wilde, S. A., Moser, D. E. and Spicuzza, M. J. (2014) Hadean age for a post-magma-ocean zircon confirmed by atom-probe tomography. *Nat. Geosci*. **7**, 219–223.

Villanueva, G. L., Mumma, M. J., Bonev, B. P., Di Santi, M. A., Gibb, E. L., Böhnhardt, H. and Lippi, M. (2009) A sensitive search for deuterated water in comet 8P/Tuttle. *Astrophys. J*. **690**, L5–L9.

Villanueva, G. L., Mumma, M. J., Novak, R. E., Käufl, H. U., Hartogh, P., Encrenaz, T., Tokunaga, A., Khayat, A. and Smith, M. D. (2015) Strong water isotopic anomalies in the martian atmosphere: Probing current and ancient reservoirs. *Science Express*, doi:10.1126/science.aaa3630.

Waite, J. H., Jr., Lewis, W. S., Magee, B. A., Lunine, J. I., McKinnon, W. B., Glein, C. R., Mousis, O., Young, D. T., Brockwell, T., Westlake, J., Nguyen, M.-J., Teolis, B. D., Niemann, H. B., McNutt, R. L., Perry, M. and Ip, W.-H. (2009) Liquid water on Enceladus from observations of ammonia and ^{40}Ar in the plume. *Nature* **460**, 487–490.

Walker, J. C. G. (1982) Climatic factors on the Archean Earth. *Palaeogeogr., Palaeoclimatol., Palaeoecol*. **40**, 1–11.

Walsh, K. J., Morbidelli, A., Raymond, S. N., O'Brien, D. P. and Mandell, A. M. (2011) A low mass for Mars from Jupiter's early gas-driven migration. *Nature* **475**, 206–209.

Wang, D., Li, H., Yi, L. and Shi, B. (2008) The electrical conductivity of upper-mantle rocks: Water content in the upper mantle. *Phys. Chem. Minerals* **35**, 157–162.

Weaver, H. A., A'Hearn, M. F., Arpigny, C., Combi, M. R., Feldman, P. D., Tozzi, G.-P., Dello Russo, N. and Festou, M. C. (2008) Atomic deuterium emission and the D/H ratio in comets. Asteroids, Comets, Meteors 2008, LPI Contribution No. 1405, paper id. 8216.

Weiss, L. M. and Marcy, G. W. (2014) The mass-radius relation for 65 exoplanets smaller than 4 Earth radii. *Astrophy. J. Lett*. **783**, L6 (7 pp.).

Wetherill, G. W. (1985) Occurrence of giant impacts during the growth of the terrestrial planets. *Science* **228**, 877–879.

Wilde, S. A., Valley, J. W., Peck, W. H. and Graham, C. M. (2001) Evidence from detrital zircons for the existence of continental crust and oceans on the Earth 4.4 Gyr ago. *Nature* **409**, 175–178.

Wyatt, M. C. (2008) Evolution of debris disks. *Annu. Rev. Astron. Astrophys*. **46**, 339–383.

Yoshino, T., Manthilake, G., Matsuzaki, T. and Katsura, T. (2008) Dry mantle transition zone inferred from electrical conductivity of wadsleyite and ringwoodite. *Nature* **451**, 326–329.

Zahnle, K. J., Kasting, J. F. and Pollack, J. B. (1988) Evolution of a steam atmosphere during Earth's accretion. *Icarus* **74**, 62–97.