Comparison of the Tectonic Conditions on Venus with Tectonic Conditions of Early Archean Earth.

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Introduction

Analysis of energy released by the processes of accretion, formation of the core, differentiation of Earth's layers, heavy bombardment of the planet's surface, and energy released by the decay of short-living and long-living radioactive elements shows that the amount of energy produced was sufficient to melt the entire planet (Pollack, 1997; Schubert et al., 2001). Additionally, the energy from a possible collision of Earth with a Mars-size body (e.g., Cameron, 1997) (the so-called Moon-forming event) at about 4.48 Ga (Halliday, 2008) would have been sufficient to melt the entire planet in and of itself as well. These facts lead most researchers to accept that after accretion Earth was covered by a magma-ocean up to 1000 km deep (Ohtani, 1985, Labrosse et al., 2007) or even deeper (Righter and Drake, 1997, Rubie et al., 2004). Some evidence derived from lunar rocks also indicates an intense "Late Heavy Bombardment" at about 3.85-3.9 Ga (Koeberl, 2006; Marty and Meibom, 2007), which could also create local magma oceans. All this means that at least during the Hadean the surface temperature of Earth was very high and the surface was represented by either a planet-wide or local magma-oceans. The emergence of magma-oceans after accretion was likewise proposed for Mars (Ward and Brownlee, 2003; Anderson, 2007; Elkins-Tanton, 2008; Erkaev et al., 2014) and Venus (Ward and Brownlee, 2003; Anderson, 2007).

It was shown earlier that Earth's magma-ocean was stratified by density (Anderson, 2007; Pilchin and Eppelbaum, 2009, 2012; Eppelbaum et al., 2014) and iron content, with its increase with depth (Pilchin and Eppelbaum, 2009, 2012; Eppelbaum et al., 2014).

Even though some researchers believe that plate-tectonics on Earth operated from the Archean (e.g., Kröner, A., 1981; Sleep, 1992; Condie and O'Neill, 2010) most researchers consider plate tectonics to have started no earlier than the Proterozoic (e.g., Hamilton, 1998, 2003; Eriksson et al., 2001; Sharkov and Bogatikov, 2001; Deming, 2002; Bjørnerud and Austrheim, 2004; Lenardic et al., 2004; Anderson, 2007; and many others). At the same time, most researchers believe that plate tectonics is unique to Earth (e.g., Stern, 2007) and did not take place on other planets like Venus (e.g., Ward and Brownlee, 2003; Anderson, 2007) and Mars (e.g., Ward and Brownlee, 2003; Anderson, 2007).

It was shown in (Pilchin and Eppelbaum, 2002; Pilchin, 2005) that subduction, obduction and uplifting are anti-isostatic tectonic processes violating the law of gravity and thus requiring the presence of gigantic external forces. Obduction is of course the most graphic process of plate tectonics, as obductions of oceanic plates onto continental margins are found in many regions throughout the world. The best example is the obduction and formation of the Oman ophiolite, where an oceanic plate up to 20 km thick obducted onto the Arabian Peninsula in the Cretaceous (e.g., Eppelbaum et al., 2014). Another example shows that during an early obduction in the pre-Appalachians region, the continental margin was detached and thrust onto the Archean Superior province of Canada for ~300-500 km, forming the entire Grenville province (Pilchin and Pilchin, 2014). Dramatic obductions are also known in such regions as, Fennoscandian Caledonides, the Urals, the Alps, and many other regions.

Subduction of a dense rigid plate into the mantle is one of the key features of plate tectonics. It is part of a process of re-working of the oceanic plate to compensate for the formation of new lithosphere in spreading centers (mid-ocean ridges). It is a unique process that is supposed to be put in motion by the negative buoyancy of a subducting slab, slab pull force, ridge push force, and basal drag force. For subduction to take place, the net of all these forces must overcome the immense forces of friction (Buiter et al., 2001; Pilchin and Eppelbaum, 2002; Eppelbaum et al., 2014) between the submerging slab and its surrounding rocks. The slab pull force, when the negatively buoyant head of the slab pulls the entire plate, is generally considered as one of the dominant acting forces (e.g., Wilson, 1993). The main friction forces applied to an obducting or subducting plate may be explained by Byerlee's law (e.g., Eppelbaum et al., 2014) and defined by a friction coefficient (μ) within the range of 0.6–0.85 (Byerlee, 1978).

It was shown earlier (Pilchin, 1986; Pilchin and Eppelbaum, 2002; Pilchin, 2005; Eppelbaum et al., 2014) that the serpentinization process takes part in every known obduction and every formation of ophiolite/serpentinite thrusts. Moreover, it was demonstrated (Pilchin, 1986; Pilchin and Eppelbaum, 2002; Pilchin, 2005) that the serpentinization process, resulting in a vast increase of volume (up to 40%; Hess, 1955), may generate immense pressure capable of facilitating the obduction processes. The increase of volume of peridotite during its serpentinization, at only 1%, could cause a composite unloading of pressure of about 1000 MPa (Pilchin, 1986; Pilchin and Eppelbaum, 2002). Since both the obduction and subduction processes are intrinsically related to the oceanic lithosphere containing a peridotite layer, which is the main rock type required for conducting serpentinization under certain/specific temperature and water presence conditions, it is clear that the serpentinization process could effectively play a role in the generation of great external pressures. It was shown (Pilchin, 1986; Pilchin and Eppelbaum, 2002; Pilchin, 2005) that the serpentinization process taking place in one part of a plate generates gigantic forces (pressures), which would be applied to the other parts of the plate.

For any plate to be stable and capable of transferring a pushing force in a specific direction it must either be elastic or a significant portion of it must contain elastic lithosphere. The elastic lithosphere is characterized by rock temperatures in the range between 573-673 K and 873 K (e.g., Pilchin and Eppelbaum, 2009, 2012; Eppelbaum et al., 2014). Below this temperature range rocks within the plate may be too brittle, and above this range rocks may be in plastic condition, which is not conducive for the plate to move as a single unit. Problems related to the elastic lithosphere (and its portions) are discussed in greater detail in (Pilchin and Eppelbaum, 2012).

Tectonic conditions on Earth in the Early Archean

Since plate tectonics, and the subduction process in particular requires dense and cold rigid plates (e.g., Anderson, 2007), it is clear that one of the most important characteristics to investigate is the temperature regime of the lithosphere or its parts forming a plate.

It was shown in (Pilchin, 2011) that from the beginning of Earth's accretion to the time when the temperature of the surface dropped below ~843-729 K (the upper temperature limit of the transformation of ferrous iron into ferric iron (TFFI)), magnetite could not have existed. This fact allows us to accept the emergence of the first banded iron formations (BIFs) as the time when Earth's surface temperature dropped below ~843 K (Pilchin, 2011). The first BIFs (Algoma type formations) were deposited as early as 3.85 Ga in Greenland and 3.5 Ga in Australia (Klein, 2005). These facts allowed (Pilchin, 2011; Pilchin and Eppelbaum, 2012; Eppelbaum et al., 2014) to conclude that the surface temperatures dropped below 843 K sometime within the range of 3.85-3.5 Ga. This means that throughout most of the Archean, even the surface rocks on Earth were not in elastic condition, and could not have formed a rigid and elastic plate.

Another important point in Earth's evolution is the time the water-ocean began to form at about 3.42-3.26 Ga, when surface temperatures dropped to ~600-500 K (Pilchin and Eppelbaum, 2012; Eppelbaum et al., 2014). Until that point, all water on Earth existed within the early Earth atmosphere (Pilchin and Eppelbaum, 2006, 2009, 2012; Pilchin, 2011; Eppelbaum et al., 2014). This means that the start of formation of the oceanic crust/lithosphere required for plate tectonics was unlikely before the Middle Archean. Moreover, when the water-ocean began forming under temperatures of ~600-500 K, the elastic part of the lithosphere ran from the surface, having no contact with the convecting mantle, and was too thin to operate as a long single plate.

The thickness of the elastic part of the lithosphere reached values no greater than 9-12 km only in the Late Archean, with a total thickness of the lithosphere less than 100 km, including a forsterite layer and possible melt pockets (Pilchin and Eppelbaum, 2012; Eppelbaum et al., 2014). This makes it very problematic for plate tectonics to start in the Late Archean, because about 90% of the lithosphere was either in plastic or molten condition. Further evidence of the temperature conditions of the Earth crust and lithosphere in the Archean can be obtained from conditions of the formation of granulites. It was determined (Pilchin, 2011; Pilchin and Eppelbaum, 2012; Eppelbaum et al., 2014) using the formation conditions of granulites that temperatures in the lower crust were very high in the Archean (see. Table 1).

Table 1. Thermodynamic conditions of the metamorphic processes of granulites of different origin (adapted from Pilchin and Eppelbaum, 2009, 2012)

Rock (*n*)	Average *T*, K	Average *P*, MPa	Average *P*/*T*, MPa/°C	Average depth of lithostatic pressure, km	Average geothermal gradient (K/km)
Granulites average* (*n* = 601)	1075	822	1.02	~28	28.6
Granulites of:					
Early and Middle Archean	1059	880	1.14	30.3	25.3
Late Archean	1127	871	0.87	30.0	27.8
Early Palaeo-Proterozoic	1040	667	0.86	23.0	32.6
End of Palaeo-Proterozoic	1087	825	1.01	28.4	28.0

*only granulites of the Archean and Early Proterozoic age

The data in Table 1 can be used to characterize the temperature conditions within the lower crust of the forming lithosphere during the Archean and Early Proterozoic. It is clear that during these periods, temperatures at depths of ~23-30 km were high and the thickness of the elastic part of the lithosphere was about 9-12 km.

It is known (e.g., Pilchin and Eppelbaum, 2002; Pilchin, 2005) that the obduction process and formation of ophiolite/serpentinite thrusts involve remnants of the oceanic peridotite layer. Any signs of obduction processes or ophiolites older than the Early Proterozoic are unknown (e.g., Pilchin and Eppelbaum, 2002; Pilchin, 2005).

Researches show that the subcontinental lithospheric mantle (SCLM) is buoyant for all lithospheric ages, with greatest buoyancy for the Archean (e.g., Poudjom Djomani et al., 2001; Griffin et al., 2003).

It is also generally accepted that eclogite is the only rock which could be denser than rocks of the asthenosphere (e.g., Anderson, 2007). This means that for a plate to be negatively buoyant, it must be composed of significant amounts of eclogite. It is also known (Pilchin and Eppelbaum, 2009, 2012) that high density rocks have a higher content of iron. This means that eclogites with the greatest density would be those containing greater amounts of iron (higher content of almandine and aegirine).

Current tectonic conditions on Venus

There is significant similarity between the present composition of the Venus atmosphere and that of early Earth from the Hadean until the beginning of the water-ocean formation at the end of the Early

Archean. The main difference being that the Venus atmosphere contains only negligible amounts of water. Most researchers (e.g., Lewis, 1995) believe that Venus lost water through the escape of hydrogen. The average surface temperature of Venus is ~740 K; ~650 K at the top of Maxwell Montes and ~755 K in its deep depressions (e.g., Basilevsky and Head, 2003; Fegley, 2004). Such surface temperatures were found on Earth at least before 3.5 Ga. The average pressure on the Venus surface is ~9.20-9.50 MPa (Seiff, 1983); ~4.5 MPa at the top of Maxwell Montes and ~11.0 MPa at depressions about 2 km deep (Basilevsky and Head, 2003). To put things in perspective, the pressure of the early Earth atmosphere before the water-ocean began to form was at least ~35 MPa (Pilchin, 2011; Pilchin and Eppelbaum, 2012; Eppelbaum et al., 2014).

Taking into consideration that: the region of Venus's accretion within the solar nebula was relatively hot (see Table 2) and was most likely able to collect more heat energy during accretion than Earth or Mars; Venus, as any other planet, was under heavy bombardment during its early evolution; and that it receives much more heat energy from the Sun than Earth or Mars; it is likely that Venus never had a surface temperature lower than it currently does. Moreover, the upper crustal rocks on Venus are mostly basaltic (e.g., Grotzinger et al., 2013), which requires a minimum surface temperature for the Venus magma-ocean to have been 1273-1473 K, with the actual temperature likely being even higher. This means that up to this point, Venus never had conditions favorable for the liquefaction of water and formation of a water-ocean. It also means that it never had an oceanic crust or oceanic lithosphere. The high temperature of the Venus surface indicates that it has an extremely thin thickness of the elastic portion of its lithosphere (starting right from the surface), and it is not possible for it to form into a rigid plate even now. Moreover, it is also conceivable that the Venusian lithosphere has not yet finished its formation, and molten rocks could be present at very shallow depths. Such a lithosphere would not be capable of carrying or transmitting significant forces. At the same time, the possible presence of melt at shallow depths within Venus will not allow any thin plate to descend.

In contrast, Mars had a much lower temperature in the region of its accretion within the solar nebula (see Table 2), and even though it was able to collect enough heat energy to form a magma-ocean, it had no chance to collect enough to keep it hot or above the freezing point of water after the Late Hesperian.

Table 2. Temperature in the solar nebula at the distance of the planets, and the present surface (or cloud-top) temperatures of the planets

Planet	Temperature in Solar Nebula at the Distance of the Planets, K			Present Average Surface Temperature, K
	After Kaufmann III, 1994	After Kaufmann III and Freedman, 1999	After Kaler, 1994	
Venus	~750	~750 K	800 – 850	740
Earth	~650 – 700	~650-700	600	290
Mars	~450 – 460	~500	500	240

Plate tectonics on the inner planets

In many cases, when talking about the negative buoyancy of lithospheric plates, researchers are comparing the density of the asthenosphere and/or mantle with that of the SCLM (e.g., Poudjom Djomani et al., 2001; Griffin et al., 2003). In order for a plate to have negative buoyancy, its density (SCLM + crust) must be greater than that of mantle rocks. This means that a significant portion of the plate must be composed of eclogite (e.g., Anderson, 2002), since eclogite has greater density (up to 3500 kg/m^3; e.g., Percival and Skulski, 2000) than peridotite (~3300-3340 kg/m^3), which is considered to be the main rock within the uppermost mantle (subcontinental lithosphere). However, as it was shown in (Pilchin and

Eppelbaum, 2009, 2012), a density of ~3500 kg/m^3 can only be achieved for eclogite with a maximum iron content (garnet as almandine and clinopyroxene as aegirine), and in case of Mg- and Ca-rich garnets the density of eclogite would be much lower. For example, Hall (1995) presents an average density of eclogite of 3390 kg/m^3 with a density range of 3340-3450 kg/m^3, and Kappelmeyer and Haenel (1974) presented a value of density of eclogite of about 3200 kg/m^3. Data presented in (Clark, 1966) illustrate that only for eclogites of the California and Norway Caledonides is the density greater than 3376 kg/m^3. Anderson (2007) shows that the density of eclogite increases from 3.24 g/cm3 at depth ~60 km, to 3.37 g/cm3 at >90 km, 3.43 at > 130 km, and 3.49 g/cm3 at >220 km. Though this data suggests that a density of 3500 kg/m^3 is a little high for eclogite, nevertheless this density was accepted for estimations to show the maximum hypothetically possible density difference between eclogite and the peridotitic upper mantle. The results of calculations of the required amount of eclogite necessary within a plate to make it denser on average than mantle rocks by only 0.1 g/cm^3 (10 kg/m^3) are presented in Table 3.

Table 3. Thickness of eclogite layer with density 3500 kg/m^3 required to make the average density of a subducting slab exceed that of the subcrustal lithospheric mantle (3300 kg/m^3) and asthenosphere (3340 kg/m^3)

Kind of lithosphere	Average thickness of crust, km	Average density of crust, kg/m3	Minimum necessary thickness for an eclogite layer with a density of ~3500 kg/m^3 to create density excess
			10 kg/m^3
Continental	30-40	2850	72-97 km
Oceanic	7-10	2900	15-22 km
Transitional	10-30	2880	23-68 km

The amounts of eclogite required to achieve negative buoyancy for a lithospheric slab received (see Table 3) are not realistic even for present Earth conditions, and they would have been inconceivable for conditions in the Archean. It should be noted that if we accept higher densities for the mantle (for example Poudjom Djomani et al. (2001) estimated the density of the Primitive Mantle at 3390 kg/m^3), the required thickness for the eclogite part of a slab would be much greater than those shown in Table 3. Artemieva (2003) likewise indicates a density of ~3390 kg/m^3 in the asthenosphere under the Fennoscandian Shield. Moreover, it is clear that if we use the real densities of eclogites found in different regions rather than the highest possible value used in our calculations, the necessary thickness for it to increase the density of a slab beyond that of the uppermost mantle would be much greater. At the same time, even negative buoyancy for a plate does not guarantee the start of subduction, since the force generated by negative buoyancy must also overcome friction forces between the plate and its surrounding rocks (Byerlee's law). It should also be pointed out that no significant amounts of eclogites are known for the Archean (e.g., Pilchin and Eppelbaum, 2002; Eppelbaum et al., 2014), and the only Archean eclogites so far reported are ones containing 2.87 Ga zircon from the Uzkaya Salma area and 2.70 Ga eclogite from Shirokaya Laba area of Kola Peninsula, Fennoscandian Shield, Russia (Mints et al., 2010). There is still controversy about the origin of eclogites (usually related to diamonds and diamond inclusions) in the Archean lithosphere, many of which are highly magnesian (Griffin et al., 2003) and therefore less dense. Griffin et al. (2003) also state that eclogites and eclogite-paragenesis diamonds are much more abundant in Proterozoic settings than in the Archean. All the above would suggest that the presence of significant amounts of eclogites in Archean plates is unlikely, and those plates would not have had negative buoyancy. It was also shown that the highest densities of eclogites (greater than 3376 kg/m^3; Clark, 1966) are found only in the Phanerozoic (Pilchin and Eppelbaum, 2002; Eppelbaum et al., 2014).

Another problem with eclogite is its mineral and chemical composition. Eclogite mostly consists of garnet (almandine, pyrope, and grossular) and clinopyroxene (usually omphacite, jadeite, and aegirine).

Garnets are extremely rich in aluminum as Al_2O_3, having a content of 20.48 % for almandine, 25.29% for pyrope and 22.64 % for grossular. Jadeite and omphacite also have high aluminum content as Al_2O_3: 22.38 % for jadeite and 7.24 % for omphacite. Aegirine contains no aluminum, but it contains ferric iron (Fe^{3+}). On the other hand, aluminum content as Al_2O_3 is quite low in mantle rocks. According to a pyrolite model by Ringwood (1966) and other models by Mason (1966), Hofmann (1988), and Hart and Zindler (1986), the aluminum content in mantle rocks is 3.5%, 3.1%, 4.06%, and 4.1%, respectively. It is therefore evident that mantle rocks with such a low content of Al could not be considered source rocks for the formation of large amounts of eclogites. However, K-feldspars (18.32 % Al_2O_3) and plagioclases (albite with 20.35 % Al_2O_3 and anorthite with 35.84 % Al_2O_3) may well be excellent sources of aluminum for the formation of garnets and clinopyroxenes, and consequently eclogites. Moreover, the direct transformation of albite to jadeite at specific thermodynamic conditions is well known. However, K-feldspars and plagioclases are located in the upper crust of continents, while eclogites are formed on average under a pressure corresponding to a depth of ~64 km (Pilchin, 2011; Eppelbaum et al., 2014; see also Table 4). This means that in order for K-feldspars and plagioclases to form eclogites, they must first somehow be delivered to depths of ~ 64 km. This would only be possible by applying a strong force pushing the upper crust layer to that depth. Also, it is known that in the Archean the temperatures were in general greater than in later periods, because of the overall cooling of Earth. This means that eclogites formed in the Archean would most likely be high temperature ($T > 993$ K) or medium temperature ($843 < T < 993$) ones, and that rocks of the upper crust necessary to form the eclogite must have been pushed to depths averaging ~70-90 km. Additionally, it is clear from Table 1 that the temperature at a depth of ~64 km during the Archean was much greater than the ~856 K required for the formation of eclogite (see Table 4).

Table 4. Thermodynamic conditions necessary for the formation of eclogites of different origins (after Pilchin, 2011)

Eclogites Formed at Temperatures, K, (*n*)	Average *T*, K	Average *P*, GPa	Average *P*/*T* *, MPa/C	Average Depth of Lithostatic Pressure, km	Average Geothermal Gradient (K/km)
Average, (*n*=556)	856	1.90	3.23	~64	9.1
$T < 843$, ($n = 279$)	762	1.51	3.10	50	9.8
$843 < T < 993$ (216)	910	2.13	3.34	70	9.1
$T > 993$ (61)	1075.3	2.77	3.46	90	8.7

*Average value *P*/*T* was calculated as the average of the *P*/*T* ratios

This means that during the Archean there would have been problems with initiating a slab pull force, and correspondingly the subduction process without strong external forces. Another problem with a slab pull force is related to the fact that it is an extensional force between the head and the rest of a slab, while the ultimate tensile strength is very low for rocks (e.g., Clark, 1966) and the value of their stress limit declines with an increase in plate size. This would certainly result in detachment of the slab's head from the rest of the plate very early in the subduction process. Moreover, it was shown earlier (Solomatov, 2004 and references therein) that the presence of water has the potential of reducing rock strength to almost zero. In other words, even if the slab's head would not get detached from the rest of a plate, the start of its dehydration would immediately cause it to break up. On the other hand, the highest among the strength limits for rocks is their compressive strength (e.g., Clark, 1966), which means that strong compressive forces in both vertical (down) and horizontal directions would be required for subduction to start.

An obducting plate would generally (except in cases of obduction of an oceanic plate onto another oceanic plate) face strong friction against solid rocks of the continental margin that must obey

Byerlee's law (e.g., Eppelbaum et al., 2014) and have a friction coefficient within the range of 0.6–0.85 (Byerlee, 1978). Therefore, the presence of a gigantic horizontal force would be required to overcome this enormous friction and support the main thrusting against the force of gravity. In contrast, a subducting slab would face friction forces from all sides, which would increase the magnitude of the friction net force. At the same time, a subducting slab must force its way through the layers of mantle rocks, which are under vast pressure, and this would also require immense additional net forces.

Estimations of pressure generated by the main forces of plate tectonics gave values of: ~45-50 MPa for slab pull force (Bott, 1993; Price, 2001); ~20-40 MPa for ridge push force (Price, 2001); and ~20 MPa for basal drag force (Bird, 1998). In contrast, the maximum pressure involved in creating the metamorphic margin during obduction of the Oman ophiolite was 1.5-2.5 GPa (e.g., Eppelbaum et al., 2014; Duretz et al., 2015). All pressures involved in other obduction processes measure in the hundreds of MPas, up to single GPas (Pilchin, 2005; Eppelbaum et al., 2014). It is clear that such pressures cannot be generated by slab pull, ridge push, and basal drag forces. This means that for obduction or subduction processes to take place, giant external forces (external to the obducting or subducting part of a plate) are required.

It was shown earlier that the immense forces required for initiating the subduction process could be provided by the serpentinization of rocks and minerals (Pilchin, 1986, 2005; Pilchin and Eppelbaum, 2002; Eppelbaum et al., 2014). These researches discuss in detail problems related to the formation of obduction, as well as demonstrating that under certain conditions it may trigger the subduction process. A simple explanation of this is presented in Figure 1.

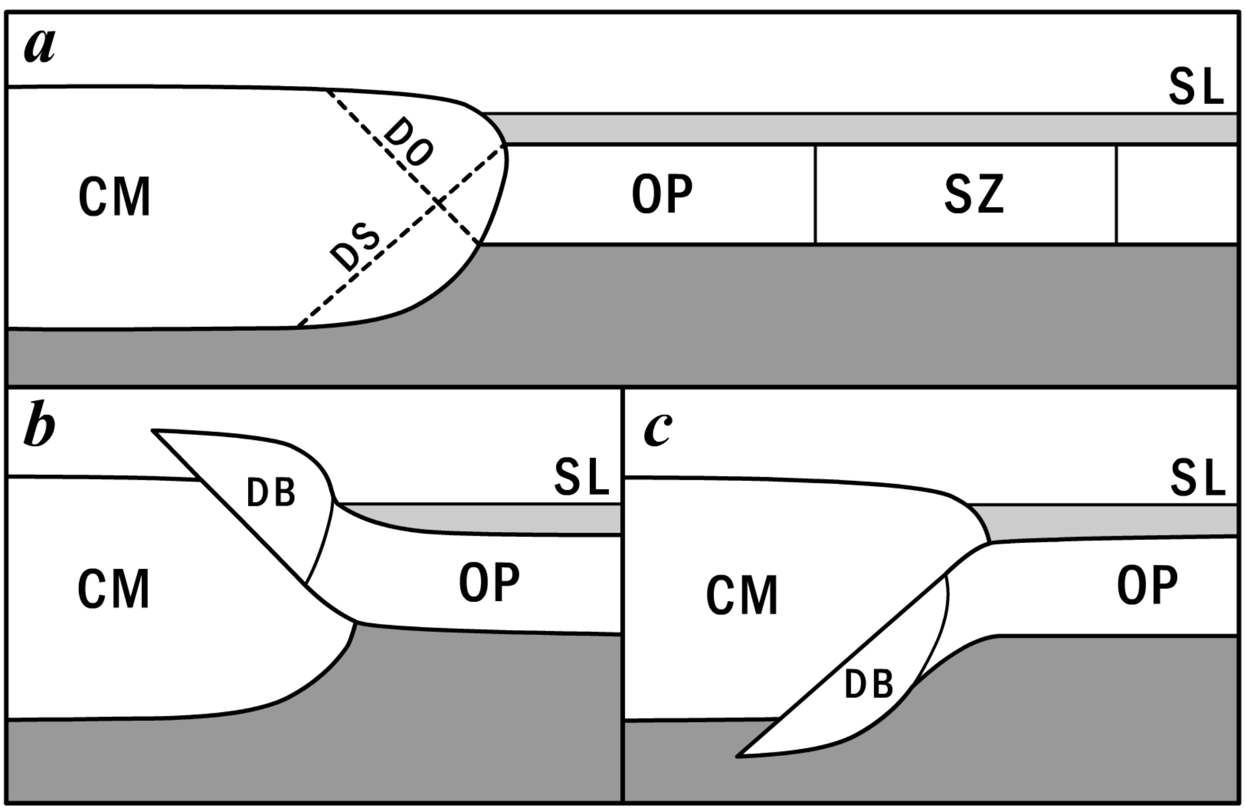

Figure 1. Simplified diagram of tectonic activity of an oceanic plate caused by serpentinization within its peridotite layer: a) initial push of activated oceanic plate against continental margin; b) formation of decollement and start of obduction process; c) formation of decollement and start of subduction process. CM – continental margin; OP – oceanic plate; SZ – zone of serpentinization of peridotite layer of oceanic plate; SL – sea level; DO – possible position of decollement for starting obduction process; DS – possible position of decollement for starting subduction process; DB – detached block of continental margin.

The serpentinization process is known to take place at temperatures below ~773-673 K and above ~473 (e.g., Hess, 1955; Pilchin and Eppelbaum, 2002; Pilchin, 2005). It is shown in Fig. 1a that the start of serpentinization of the peridotite layer in the oceanic plate engages strong forces within the entire oceanic plate, forcing it to expand and consequently begin to push against the continental margin. If the serpentinization is significant, the forces generated would be capable of destroying the continental margin. There are two possible positions for a detachment in this case: 1) upper part of the continental margin would be detached and pushed onto the continent, followed by the obducting oceanic plate (see Fig. 1b); and 2) lower part of the continental margin would be detached and pushed beneath the continent, followed by the subducting oceanic plate (see Fig. 1c).

It is important to note that only oceanic plates contain a peridotite layer, and only in this layer could strong serpentinization processes take place. If we accept that the main forces driving oceanic plates are caused by serpentinization of their peridotite layer, we can conclude that plate tectonics on Earth would stop when either the temperature of these peridotite layers (including forsterite layer; Pilchin and Eppelbaum, 2009, 2012) dropped below ~473-523 K (lower limit for the serpentinization process) or water would no longer be available to support serpentinization.

It is possible that the first serpentinization process on Earth took place at the end of the Early Archean in Barberton GSB in South Africa, when the serpentinization of komatiites of the ~ 3.286 Ga Weltevreden Formation caused it to thrust onto the ≤ 3259 Ma (Zeh et al., 2013) Belvue Road Formation and eventually itself be overlain by ≥ 3225 Ma (Zeh et al., 2013) Schoongezicht Formation. The start of this thrusting matches up with the beginning of the formation of the water-ocean. Unfortunately, komatiite deposits were represented by flows with limited horizontal dimensions, and thus their serpentinization resulted in the formation of only small local thrusts.

High surface temperatures, a mostly plastic lithosphere, the presence of magma layers at shallow depths, absence of a water-ocean and oceanic lithosphere, absence of eclogites, absence of serpentinites, and many other mentioned above reasons made it impossible for plate tectonics to operate on Earth in the Hadean - Early Archean.

In the case of Mars, many researchers believe that a water-ocean of a different size existed on the planet at some point (e.g., Carr, 1996; Clifford and Parker, 2001), and that it contained a significant amount of water, estimated at ~500-1000 m of global equivalent layer (GEL) in the Late Hesperian - Early Amazonian (Carr, 1996), and alternatively at a total of ~550–1400 m of GEL (Clifford and Parker, 2001). This means that formation of an oceanic crust/lithosphere was possible in the past. This is in agreement with recent discoveries of serpentine on the Martian surface (e.g., Ehlmann et al., 2010) in mélange terrains at the Claritas Rise and the Nili Fossae, a few southern highlands impact craters, and the regional olivine-rich stratigraphic unit near the Isidis basin. This evidence of the presence of serpentine is very important for analyzing the possibility of the existence of a Martian water-ocean, and that of plate tectonics operating on the planet in the past, because on Earth the formation of serpentine is always associated with the serpentinization process. That the presence of serpentine on Mars is uncommon (e.g., Ehlmann et al., 2010) means that serpentinization took place at some depth, and that the process was of a limited scale, preventing the evolution of thrust systems. The fact that serpentine is found in association with mélange terrains at the Claritas Rise and the Nili Fossae is solid proof that the serpentinization process in fact took place on Mars, as the formation of mélanges is typical for the serpentinization process on Earth (e.g., Pilchin, 2005). However, the presence of significant amounts of olivine on Mars (e.g., Ehlmann et al., 2010) indicates that serpentinization was not widely spread on the planet. This suggests a quick cooling of the planet's crust to below about 473-523 K, or a lack of water within the crust in contact with olivine. Ehlmann et al. (2010) came to the conclusion that serpentine-bearing materials appear to be restricted to Noachian-aged rocks in the Nili-Fossae olivine-carbonate-serpentine unit. Since it is clear that water was present on Mars in significant amounts, it is likely that the temperature at some depths within the crust dropped by the end of the Noachian below 473-523 K. This means that most of the crust was much colder than is required to form the elastic portion of the lithosphere, and it is possible plates were too thick to initiate plate tectonics. Moreover, the absence of SO_2 from the present Martian atmosphere (e.g., Catling, 2004) indicates the lack of any significant volcanic activity for a long period of time. All of the above means that Mars does not have the energy required for plate tectonics at this time, and it's ensuing cooling hinders any expectations for plate tectonic processes on the planet in the future.

It is of course too early to talk about plate tectonics on Venus, as it does not yet have thick elastic plates, nor has it yet cooled enough to form rigid plates. However, the impossibility of a water-ocean forming, as well as an oceanic lithosphere, and the impossibility of the serpentinization process taking place makes Earth-type plate tectonics on Venus unimaginable in the future.

Lastly, to briefly mention conditions on our nearest terrestrial neighbor. The presence of a 52 km thick (Taylor and McLennan, 2009) buoyant anorthosite crust on the Moon and a lack of energy sources indicate that plate tectonics was not possible there in the past, and nor will it take place there in the future.

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