Evolution of the earth's mantle and core: Convective cycle within the mantle and different plate tectonic environments related to magma generation

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The earth is considered to have been formed in several stages in a solar nebula by accretion of carbonaceous chondrites. The moon might have been formed toward the end of such a process by collapse of a sedimentary ring around the proto-earth or due to collision of the proto-earth with a large comet. The thermal energy related to the impact resulted in the formation of a magma ocean. The process of density stratification of Fe–Ni alloy at the core and other silicates (magnesiowustite, pervoskite, garnet, spinel, etc.) is the subject of modern experimental studies. The large difference in temperature between the core and the mantle–crust interface resulted in convection, either involving whole mantle right from the outer core or initiating two-layered convection (650 km seismic discontinuity separating the two convective cycle systems). Earthquake and volcanism in the mid oceanic ridges, oceanic plate boundaries and subduction zones are surface manifestations of energy flow through mantle convection.

THERE is a general consensus among earth scientists that the earth was formed by accretion of planetesimals in a solar nebula, constituted mainly of gasses (H, He), ice particles and rock materials. Urey[1] suggested that a particular type of meteorite called the type I chondrite may represent the primitive dust particles in the solar nebula[2]. Ringwood[2] visualized the following five stages of formation of the planet earth:

In the early stage of accretion the energy involved was small and the process slow. The surface temperature was low and mainly controlled by the latent heat of vaporization of the volatile components in the accreting material, and the key reaction was $CH_4 + H_2O \leftrightarrow CO + 3H_2$. During this stage primordial sulphur was trapped to form (Fe,Ni)S, which coexisted with magnetite, olivine and hydrated Fe–Mg silicates. The accretion energy was 2000 cal/g, and the surface temperature < 500°C.

In the second stage the mass of the accreting earth increased and the energy imparted by careering planetesimals became high enough to cause degassing and sufficient rise in its surface temperature. Carbonaceous materials then reduced iron oxides to metallic iron and a reducing atmosphere began to grow. The surface temperature of the earth probably rose between 700 and 1200°C, and the accretion energy was around 5000 cal/g.

In the third stage of development, the earth's mass increased to about one fifth of its present mass. The surface temperature possibly increased from 1200 to 1500°C, and a number of the more volatile components (viz. Na, K, Rb, Cd, Hg, Bi, etc) evolved into the primitive atmosphere. The accretion energy at this stage was around 8000 cal/g.

In the fourth stage, with further increase of the nucleus and the rate of accretion, the surface temperature exceeded 1500°C causing volatilization of even the major silicate phases. At this stage volatilization of silica and Mg as SiO or MgO might have taken place before they evaporated into the primitive atmosphere. The accretion energy was probably higher than 10,000 cal/g.

According to Ringwood[2] the nucleus of the earth developed further in the final stage. Prior to the formation of the core, the earth had accreted as a homogeneous aggregate of metal and silicate phases with a more oxidized nucleus surrounded by metal-rich reduced materials. Also, large amount of heat was evolved (600 cal/g) due to conversion of a major part of the gravitational energy into heat[3], which raised the temperature of the earth by 2000°C. Urey[4–6] suggested that segregation of the iron-rich core took place over a long geological time, setting the scene for mantle convection. Runcorn[7,8], Munk and Davis[9] further nurtured this model. Ringwood[10] considered the core-formation to be rapid, facilitated by rising temperature that would have lowered the viscosity. For, if the sinking were slow as suggested by Urey, the giant metal liquid droplet would have frozen below 500 to 700 km. Sinking of the metallic core toward the centre would have buoyed up the oxidized volatile-rich

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nucleus forcing convection in the mantle, and thoroughly mixing this oxidized zone with magnesian and iron-poor silicates. Outside the solid earth, the atmosphere containing silica-rich volatiles would be subjected to turbulent motion, and spun into a disk which would cool to form a sediment ring of planetisimals. Later, this silica-rich sediment ring would collapse to form its only satellite, the moon.

According to Sasaki[11] the heat produced by the impact of falling planetesimals was retained by the blanketing effect of the primary solar atmosphere. When the mass of the planet exceeded $0.2\,M_E$, where M_E is the present mass of the earth equal to 5.97×10^{24} kg, the rain of planetesimals on the earth was sufficiently intense to produce large amount of heat. Not only did this cause melting of the silicate materials but also reduction of iron oxide and some of the silicates to metallic iron and a reduced form of silica. The oxygen so liberated in a progressively reducing atmosphere combined with the already present hydrogen in the primitive atmosphere to produce large amounts of H_2O. The high surface temperature of 4000 K (at Venus' mass: $0.8\,M_E$) and 5000 K (at earth's mass) would also force intense atmospheric convection, while a deep ocean of molten magma formed with a life time as long as that of the solar nebula (10^6–10^4 yr). Metal silicate fractionation and core formation could then have been rapid and simultaneous with accretions.

Despite uncertainties about the manner in which terrestrial planets accreted, Presnal[12] believes that a broad consensus has emerged over the past 25 years about the formation of terrestrial planets which did not occur exclusively by accretion of a large number of small objects (ordered accretion) but by stochastic accretion of planetesimals that increased in size and decreased in number with time. Modelling by Wetherill[13] suggested that only about 10^5 years would be needed to sweep most planetesimals less than about 10^{26}–10^{27} g in mass, leaving only objects of the size of the Moon, Mercury and Mars. The final growth of the earth to its present mass of 6×10^{27} g would thus take about 2×10^8 years to complete. Wetherill[13,14] argued that the terrestrial planets suffered several large and giant impacts in the later stages of their growth. Support for this model is provided by the spins of the Earth and Mars. According to Presnall[12], calculation of Dones and Tremaine[15–16] shows that the rapid spins of these planets could not have been produced by ordered accretion, but only by several giant impacts.

Weatherill[13,14] argued that the energy released by giant impacts is sufficient to produce partial, and perhaps even complete melting of bodies at least as small as Mercury. Such melting, he suggested would initiate core formation simultaneously with the growth of the protoplanets. The moon is now generally considered to have been formed toward the end of the earths' accretionary history by collision with a very large planetesimal[17–20] which is believed to be about the size of Mars. Wetherill[13] and Dones and Tremain[15,16] have even suggested that the largest impactor may have had a mass about three times as much. Presnall further suggests that a collision of such a body with the proto-Earth would have an energy release several times that needed for complete melting (see also Weatherill[13,14]). Because of the high probability of very large impacts prior to the Moon-forming event, both the proto Earth and the giant impactor would be expected to be already differentiated into a mantle and one or both of the colliding bodies could have been at least partly molten at the time of impact. Stevenson[19], for example, estimated that the proto-Earth could have been approximately 50% molten prior to the Moon-forming impact. Thus the suggestion of Wetherill about the formation of the moon is quite different from that of Ringwood[2].

It is well known that the Earth is depleted in volatile elements relative to those in C1 chondrites. If the accretion history of the Earth includes one or more giant impacts, as favoured here, a necessary consequence is extensive volatilization[18,21]. Both Mg and Si are slightly volatile relative to the strictly refractory lithophile elements, and Si is slightly more volatile[22] than Mg. Thus, one or more giant impacts would be expected to cause substantial depletion of both Si and Mg, Si more than Mg, exactly as observed. Following Hart and Zindlar[23], Presnall[12] favours volatilization by a giant impact to explain the upper mantle abundance of Mg and Si. According to this model, the Earth accreted from chondritic materials, which produced the observed abundance ratio in the Earth, of refractory lithophile elements and the volatile elements (including Mg and SiO) which show varying degrees of depletion. Thus, the abundance of both volatile and refractory elements in the Earth is readily explained as a natural consequence of a well-known process.

The working model proposed by Presnall involves (i) accretion from chondritic material, (ii) a giant moon-forming impact to melt and partially devolatilize the earth, (iii) gravitational separation of immiscible iron and silicate liquids to form the core and mantle, and (iv) equilibrium separation of immiscible iron and silicate liquids to form the core and mantle, and (v) equilibrium crystallization of the mantle from the bottom up without significant factionation. The nearly total exclusion of fractional crystallization from the model is driven by the powerful constraint of a strictly chondritic Ca/Al ratio for the primitive upper mantle. This model is simpler than other proposals that employ various scenarios for fractionation of the magma ocean to produce a compositionally layered mantle. In his view there is no known data for the present mantle that demands heterogeneity and warrants a more complex model.

Evolution of the lower and upper mantle

If the primitive meteoritic debris have a composition similar to that of carbonaceous chondrites, how is it that it got fractionated to form a layered body with a crust, mantle and core? Agee[24] made high pressure melting experiments on the Allende CV3 carbonaceous chondrites at 24, 26 and 26.5 GPa. He found that a FeO-rich magnesiowustite is an abundant crystallizing phase at the liquidus temperature (Figure 1). According to Agee[24], when the chondritic earth experienced a high temperature molten stage, the FeO-rich magnesiowustite could have been segregated to the deepest levels of the earth's interior, resulting in the depletion of the initial FeO content of the primitive chondritic mantle (Figure 1).

Phase equilibria studies on Allende meteorite (Figure 2) show that magnesiowustite, majorite (SiO_2 – 50.93, FeO – 11.60, MgO – 26.11), and ferromagnesian silicate (with perovskite-like structure) are the crystallizing phases at temperatures near the silicate liquidus and 26 GPa pressure (Figure 2). Above this, only magnesiowustite and perovskite are expected to be stable in the presence of a carbonaceous chondrite silicate melt, which at this pressure is similar to a molar mixture of 81% komatiite and 19% Fe_2SiO_4.

Because of its high density relative to coexisting chondritic silicate liquid, magnesiowustite may segregate downward in a deep magma ocean, and may be added to an earlier formed sulphide melt protocore. If this process is efficient, the core could be composed primarily of Fe–Ni–S–O. The amount of magnesiowustite that would thus be added to the core would depend on the amount that may be fractionated from the magma ocean, which is unknown if it is assumed that the magnesiowustite fractionation is an important mechanism in removing excess Fe^{2+} from the mantle, then mass balance considerations suggest that removal of 32 wt% magnesiowustite is close to the correct value. This amount would decrease the FeO/(FeO + MgO) ratio of the chondritic silicate magma ocean to the level of peridotite xenoliths.

With reference to Figure 3, Agee[24] suggested that cooling and crystallization of the chondritic earth was associated with (i) sulphide liquid and silicate liquid immiscibility, leading to the formation of the protocore (Figure 3 *a*); Figure 3 *b* suggests magnesiowustite fractionation at the base of the early mantle, Figure 3 *c* shows flotation of silicates of perovskite structure and fractionation in the mid-mantle. Figure 3 *d* shows olivine flotation in the shallow mantle and garnet flotation in the transition zone. He concludes that compositional layering produced by the five experimentally predicted differentiation stages obey mass balance conditions and is consistent with a shallow mantle source for peridotite xenoliths, a transition zone at 540 km rich in garnet a lower mantle with a super chondritic Si/Mg and a core of Fe–Ni–S–O.

Takahashi[25] studied a peridotite (KLB-1) of composition similar to that expected to occur in the upper mantle. He observed that olivine occurred as the liquidus phase at all pressures (Figure 4). The second mineral to crystallize with increasing pressure was a Ca-poor orthopyroxene (up to 3 GPa), pigeonitic clinopyroxene (up to 7 GPa) and pyroxene-rich garnet (> 7 GPa). The melting temperature interval of the peridotite was more than 600°C wide at 1 atm but narrowed to about 150°C at 14 GPa. The partial melts along the peridotite solidus became increasingly more MgO-rich as pressure increased throughout the pressure range studied[25]. At 5–7 GPa the partial melts formed within 50°C of the solidus contained more than 30 wt% MgO, and were similar to Al-depleted peridotitic komatiite, which are common in Archaean volcanic terrains. Due to increase of the enstatite com-

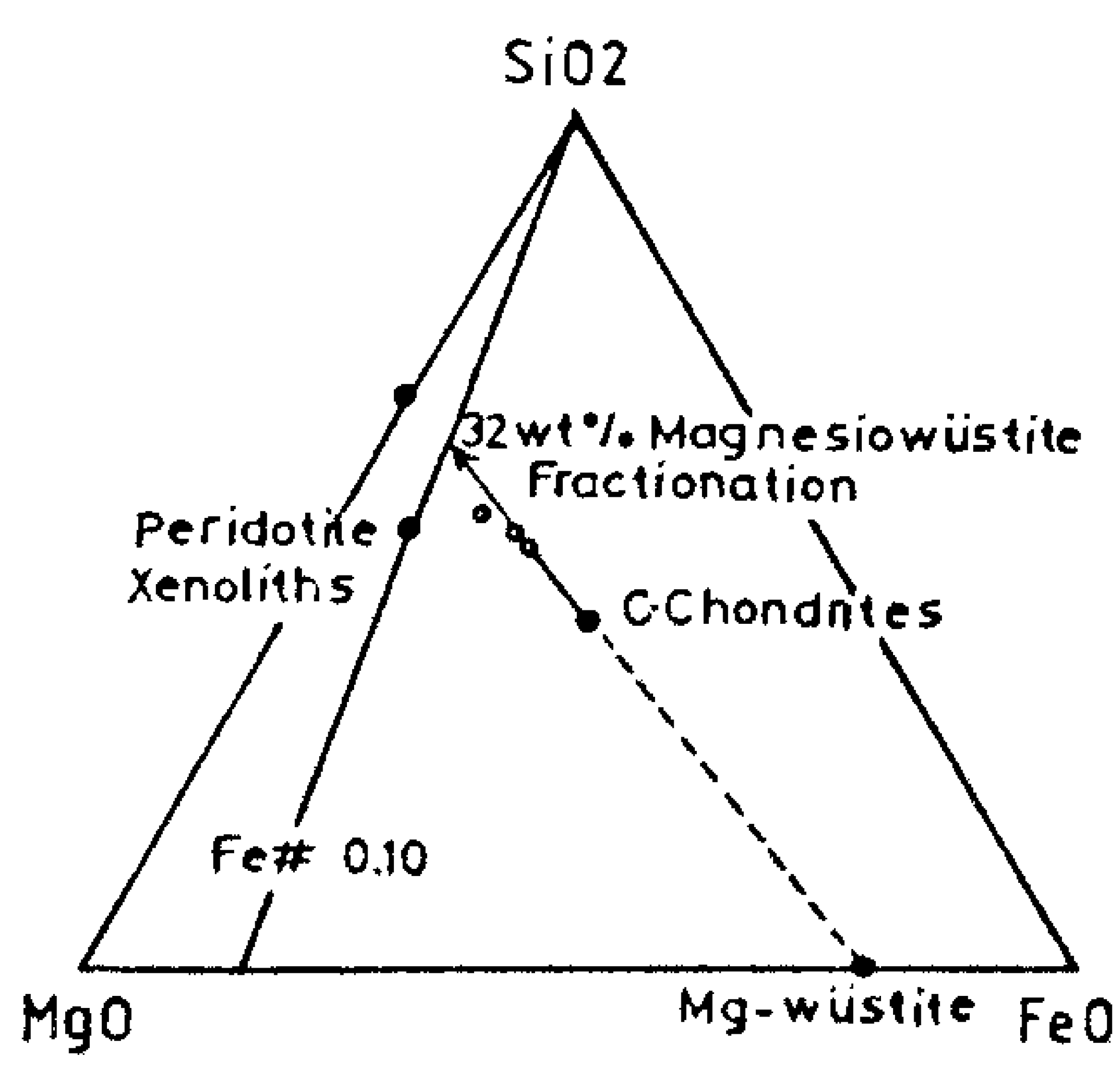

Figure 1. Composition diagram of SiO_2–FeO–MgO in wt% showing the intersection of the chondritic 'trend' and the FeO/FeO + MgO = 0.10 tie line. The intersection is labelled chondritic mantle. The intersection composition is unlike any meteorite type and resembles a pyroxenite rather than a representative lherzolite (after Agee[39]).

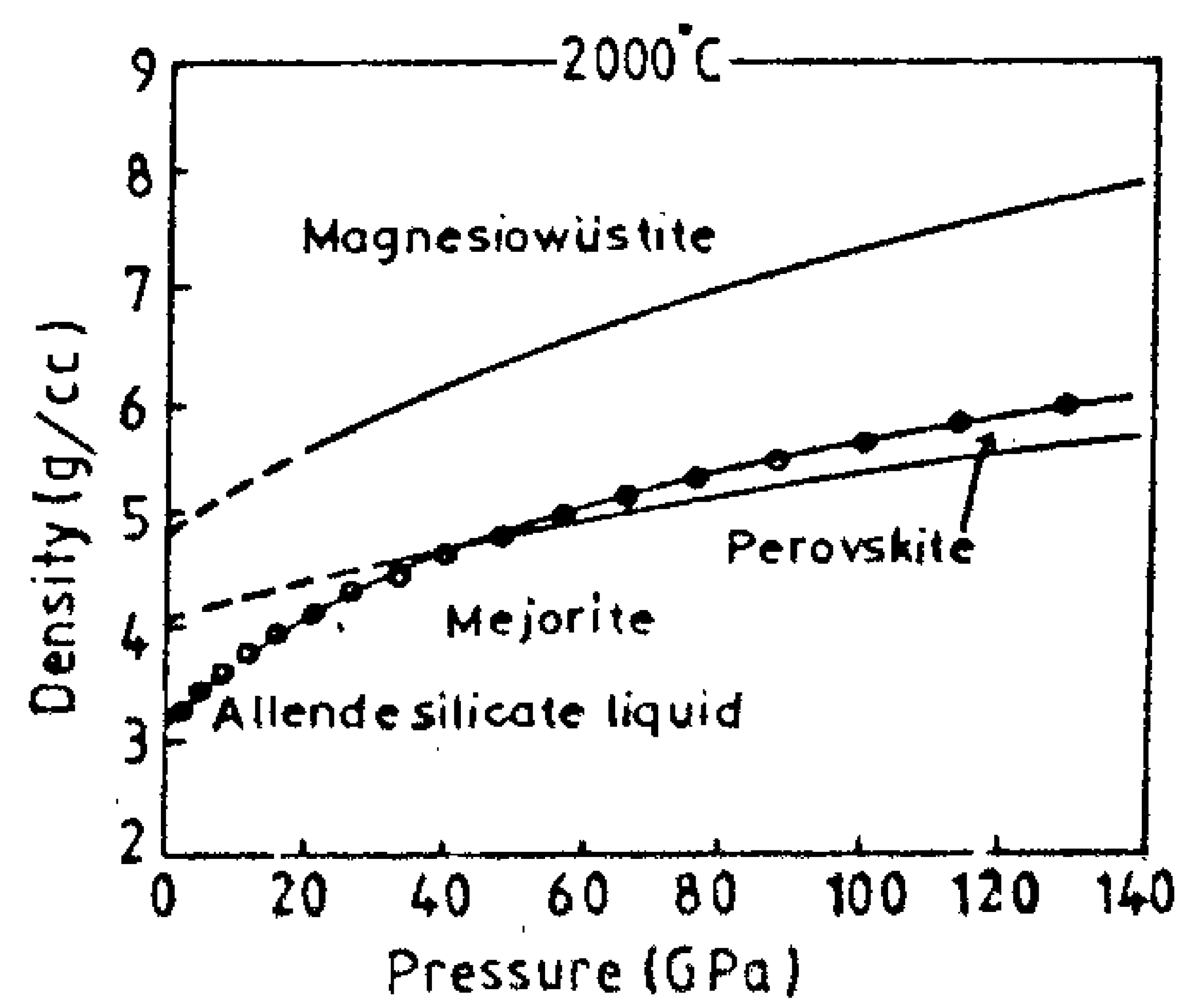

Figure 2. Density versus pressure diagram showing third-order Birch–Munaghan isothermal compression curves at 2000°C for phases that are in equilibrium during partial melting of Allende meteorite at 26 GPa (after Agee[39]).

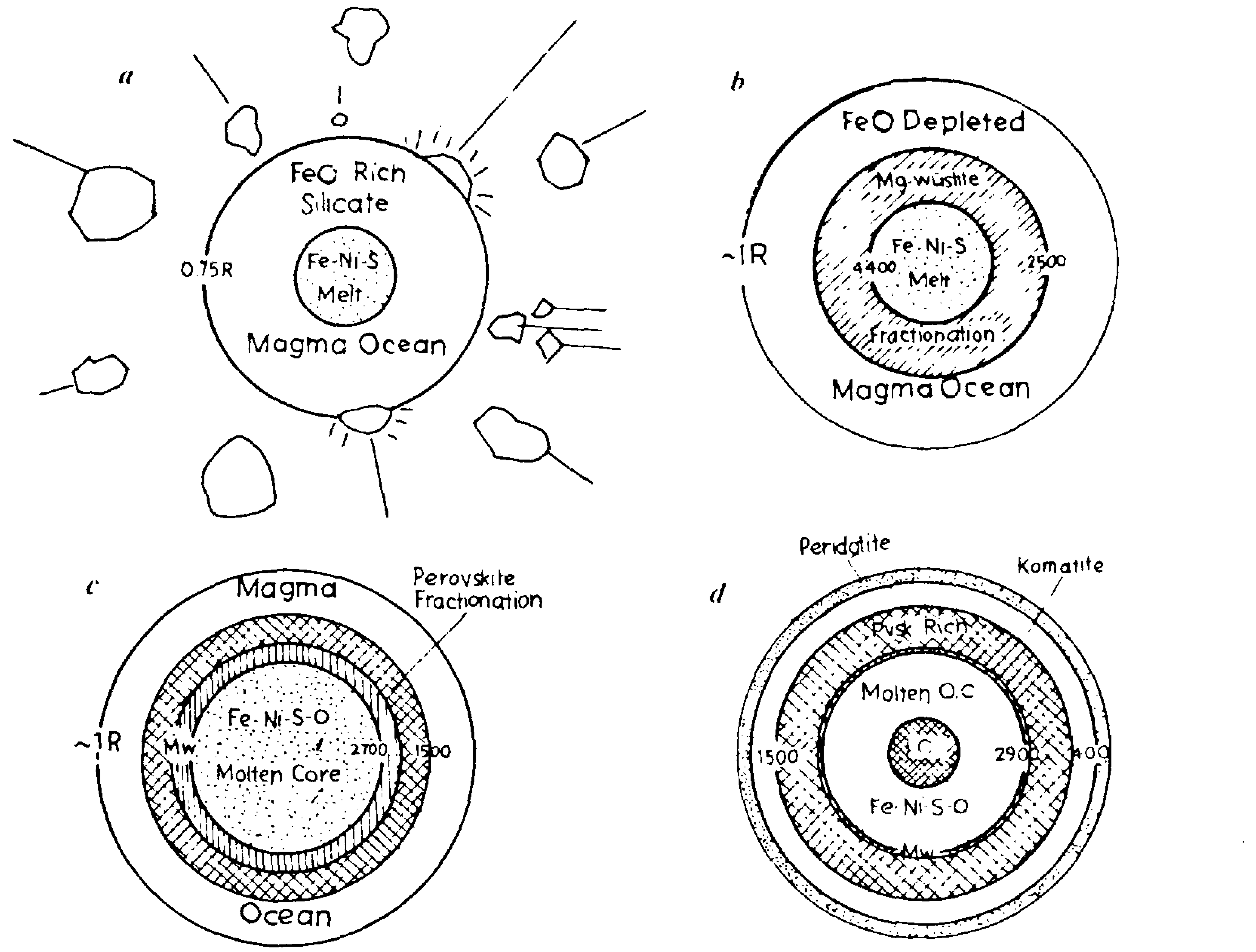

Figure 3 *a–d*. ***a***, Segregation of Fe–Ni–S protocore. Agee used elemental abundances of Alende meteorite to estimate the volume proportion of the silicate and Fe–Ni–S melts. Continuous impact of new chondritic materials possibly increases the size of the earth and caused more heating with a Mars size object resulting in the formation of a moon, it might have been another source of chondritic meteorite and heat for the proto-earth (after Agee[39]). ***b***, Formation of Fe–Ni–S proto-core accompanied by 30% magnesiowustite fractionation. If all the fractionated magnesiowustite were incorporated in the core, the radius will be ~ 400 km larger than the present core. According to Agee, in the molten silicate mantle FeO/FeO + MgO was 0.10. Although fractionation of magnesiowustite and FeO solution in the core is shown separately in the figure, this might have happened simultaneously (after Agee[39]). ***c***, Later stages of differentiation in which most of the FeO from magnesiowustite is dissolved in the molten core. Agee considers that the predominant crystalline phase at this time is Mg-perovskite, which results in the enrichment of mantle below 700 km in SiO_2. According to Agee the depth of perovskite accumulation occurs in a layer that contains its level of natural buoyance (~ 1680 km) in the silicate melt (after Agee[39]). ***d***, Mostly solidified mantle after magmatic differentiation. Agee thinks that formation of a liquid with bulk composition similar to komatiitic liquid results due to mass balance. Layer of komatiitic melts as well as that of crystalline phases (olivine, clinopyroxene, majorite, β-phase, Mg and Ca-perovskite) occupy position at different levels depending on pressure temperature condition of fractionation from the magma ocean. According to Agee with the exception of peridotite the composition of layer may not be the same as that at present (after Takahashi[25]).

ponent in the clinopyroxene solid solution at high pressure and temperature, the orthopyroxene liquidus field narrowed as pressure increased and disappeared at 3–5 GPa. Harzburgites, which are common in the peridotites from ophiolite suites, may represent relicts after separation of the melts formed at depths lower than 100 km. He considers that lherzolites showing protogranular texture, might have been produced as a single pyroxene peridotite residue formed by partial melting at relatively greater depths (> 100 km). Takahasi[25] also suggests a diapiric model consistent with the genesis of a komatiitic magma formed by partial melting of a mantle peridotite at a depth of 150–200 km. He made the following observations: (i) convergence of the liquidus and solidus of the peridotites at pressure > 14 GPa, (ii) the near-solidus partial melt composition very close to the bulk rocks at 14 GPa, and (iii) change in the liquidus mineral from olivine to majorite garnet at pressures between 16 and 20 GPa. Based on the above conclusions he proposed that the upper mantle peridotite was generated originally as a magma (or magmas) by partial melting of the primitive earth at 400–500 km depth.

There is a sharp discontinuity at 670 km, which has been accepted as the boundary between the upper and lower mantle[26]. To clarify the nature of the 670 km discontinuity and the state of the lower mantle, Ito and

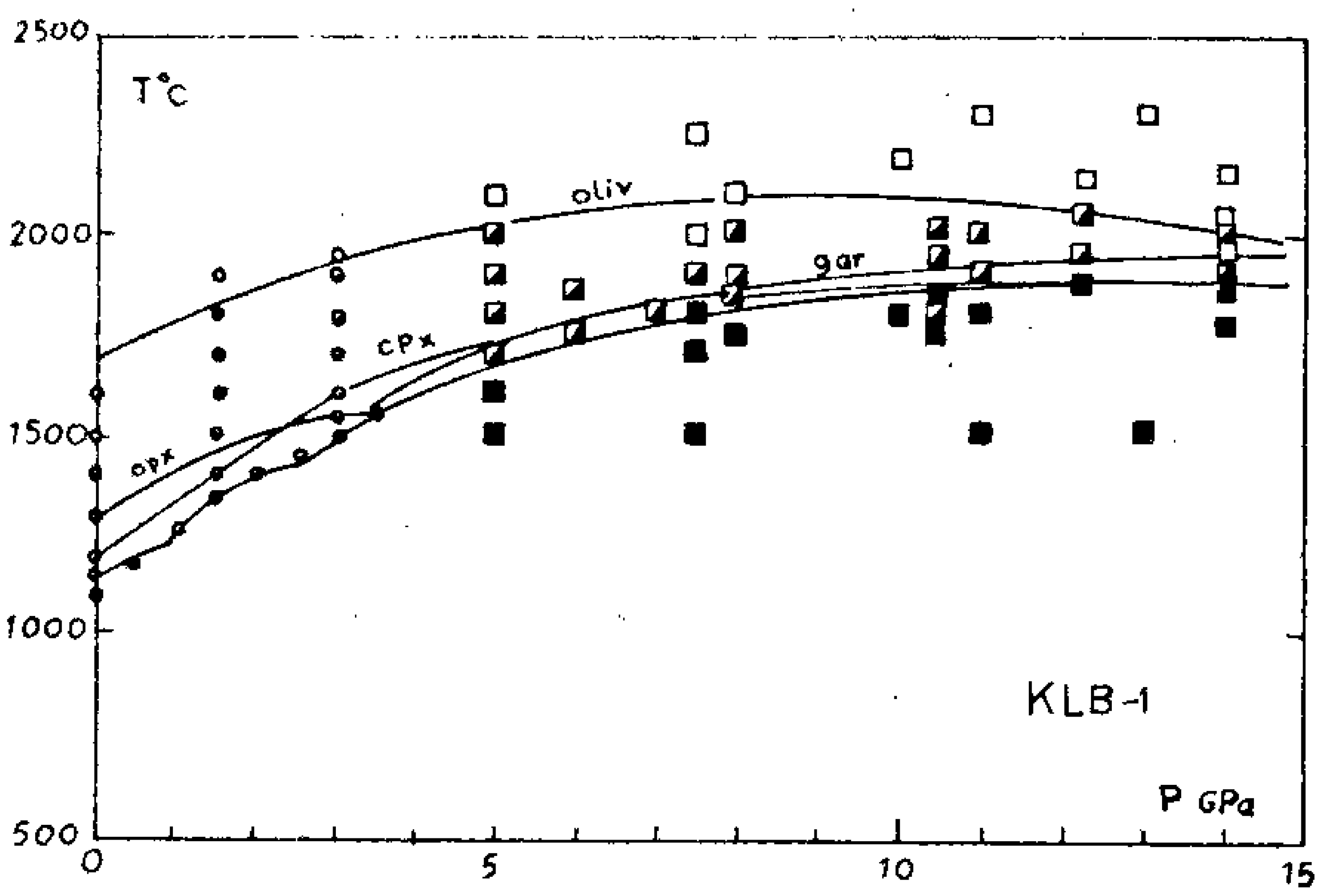

Figure 4. Melting phase relations of a fertile peridotite KLB-1 up to 14 GPa under dry conditions[25]. Melting temperature interval decreases drastically from about 600°C at 1 atm to about 150°C at 14 GPa. Squares denotes experimental data points. Only selected data points are shown for the experiments below 3.5 GPa (small circles).

Takahashi[27] studied the system MgO–FeO–SiO_2 and $CaSiO_3$–$MgSiO_3$–Al_2O_3 at pressures between 10 and 27 GPa and at temperatures up to 1600°C. They found that the dissociation of the spinel phase into an assemblage of perovskite and magnesiowustite could be responsible for the 670 km discontinuity. They found the dissociation to be complete within the pressure interval of only less than 1 GPa, thereby making the discontinuity very sharp. By considering detailed phase relations and estimated physical properties of the lower mantle, they determined the possible composition of the lower mantle in the system MgO–FeO–SiO_2. In the case of the $CaSiO_3$–$MgSiO_3$–Al_2O_3 system, the stability field of majorite was observed to expand rapidly toward the $CaSiO_3$–$MgSiO_3$ join in the pressure range of 10–18 GPa, and retrograde toward the grossular-pyrope join at pressures higher than 23 GPa, dissociating to an assemblage comprising $MgSiO_3$-rich perovskite and an unquenchable 'CaO-rich phase' with diopside-like composition. Majorite is therefore, expected to be an important constituent in the transition zone, and the dissociation of majorite could contribute to the 670 km discontinuity. The complete dissociation of majorite however, requires a fairly large pressure interval (~ 39 GPa), which produces $MgSiO_3$-rich perovskite, a 'CaO-rich phase' with small amounts of stishovite, and an 'Al_2O_3-rich phase' as the lower mantle constituents.

Ohtani *et al.*[28] studied a model chondritic mantle composition having 50.18 wt%, SiO_2, 3.62% Al_2O_3, 7.16% FeO, 36.18% MgO and 2.5% CaO. They prepared the chondritic starting material at 1150°C under controlled partial pressure of oxygen. The starting material contained olivine, pyroxene and a small amount of anorthite. Their study on the melting relation of the model chondritic mantle at pressures up to 20 GPa and temperatures in excess of 2000°C, is summarized in Figure 5. They noted that at present between 12 and 15 GPa the liquidus phase changes from olivine to majorite. The liquid coexisting with 20 GPa has peridotitic composition. They further noted that the CaO/Al_2O_3 ratio of the liquidus of majorite at 20 GPa is lower that of the subsolidus majorite: partial melting and majorite fractionation at the base of the upper mantle could produce a peridotitic liquid with a CaO/Al_2O_3 ratio greater than that of the chondritic starting material, which agrees well with mantle geochemistry.

Convective cycles in the upper mantle

Because of the transfer of heat from the core to the surface of the earth the mantle is in a slow convective motion. Different models of convective cycles have been summarized by Wyllie[29], who divided the various interpretations into two categories: (i) involving whole mantle convection and (ii) involving a layered mantle convection (Figure 6 *a* and 6 *b*). The models, which involve two-layered convection with occasional transfer of materials between the two layers is shown in Figure 6. The discussion about primitive mantle of peridotite composition is given in Figure 6 *a*. According to Depaolo and Wasserburg[30] this layer has been depleted by chemical differentiation and growth of the crust. Figure 6 *b* depicts a mantle of peridotite composition with addition of materials from the subducted lithosphere which has been transported to the core-mantle boundary by the convection of the whole mantle. Figure 6 *c* is a schematic diagram as visualized by Anderson[31]. The diagram shows the presence of a depleted mantle with subduction of the lithosphere which enriches the eclogite composition of the upper mantle. This is denoted as piclogite. Figure 5 *d* shows the two layer convection with subducted lithospheric slabs accumulated as megaliths at the depth[2] of 670 km.

According to Fyfe[32] the earth can lose energy from the interior by two major processes of conduction and convection. Active convection requires an appropriate Raleigh number for the layers of the interior, and modern analysis of the problem shows that the Ra's for the interior in general greatly exceed those required for the initiation of convection. While in general, pressure increases viscosity, new data from high-pressure experiments and seismic modelling has shown that the expected increase in viscosity is offset by phase changes and changes in the coordination numbers, as simple crystal structures dominate materials at depth. In fact, viscosity changes little through the outer 100 km of the mantle.

Magmatism associated with different tectonic regimes

Volcanic activities associated with oceanic ridges, rift valleys, different collision zones or regions of hot-spot activities are surface manifestations of convective cycles within the earth's mantle. The regions of intense volcanism along the plate boundaries are the zones of earthquakes. Volcanic processes in these regions are briefly described below.

Oceanic ridges

In all major ocean basins there are ridges (1000–3000 m above the ocean floor) with a total length of more than 60,000 km. The mid-oceanic ridges are the regions, where new oceanic lithosphere is generated. The piles of lava thus formed, are pushed away from the ridge axis as the ocean floor spreads at the rate of 1 to 10 cm per year. Hot mantle materials rise along an adiabate below the spreading centre. The change in temperature with depth in this region[33] is only about 0.5°K/km. At a relatively shallow depth (50–100 km) the ascending mantle cuts across the solidus and partial melting takes place. Geochemical and petrological studies thus constrain the depth at which, more basaltic melt is generated. At depths greater than 65–80 km a small amount of melting takes place in garnet lherzolite. The maximum amount of melting occurs at a depth of 20–65 km in the spinel peridotite field, and little melting at depths above 20 km in the plagioclase peridotite stability field. Petrological and geochemical data suggest that mid-oceanic ridges are produced by 10–15% partial melting of the sub-ridge mantle. Turcotte and Morgan[33] suggest that in order to form a 6 km oceanic crust from a mean amount of peridotite fusion (10%), a 60 km thick column of melted mantle should exist below this crust as it moves away from the spreading centre. They further believe that the melt is generated along grain boundaries but accumulates and rises to produce the oceanic crust through the residual solid matrix because of its differential buoyancy. There is sufficient evidence to suggest that magma is collected in a magma chamber. Field studies show that there are sheeted dyke complexes, which are an essential feature of the formation of oceanic crust at spreading centres. A schematic representation of this process as visualized by Turcotte and Morgan[33] is shown in Figure 7. They believe that the magma is stored in a magma chamber at the base of the sheeted

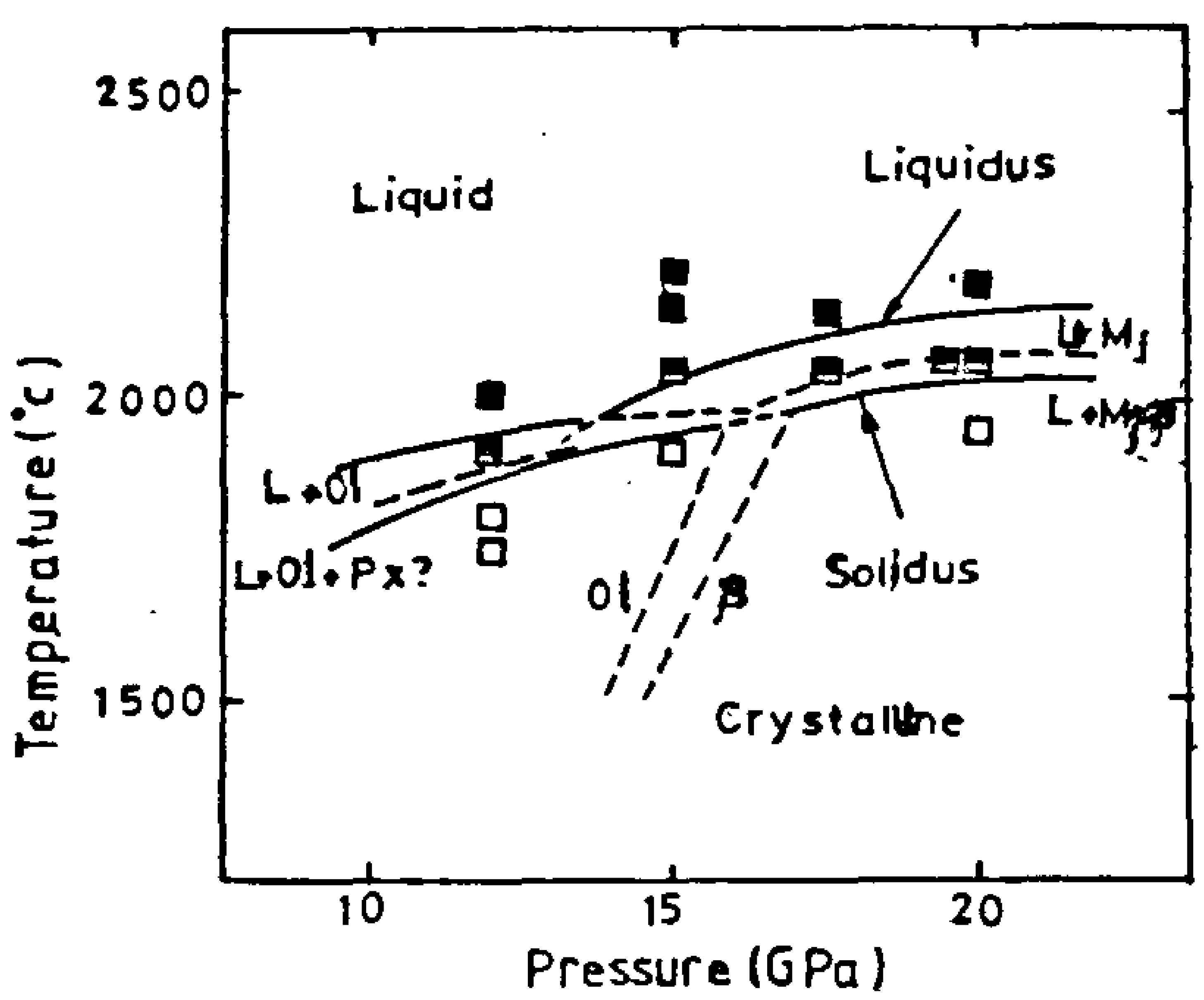

Figure 5. Melting relations of a model chondritic mantle to 20 GPa; solid squares, super-liquidus; semisolid squares, crystal + liquids; blank squares, subsolidus; L, liquid; Ol, olivine; Px, clinopyroxene; β, modified spinels; Mj, Majorite (after Ohtani *et al.*[28]).

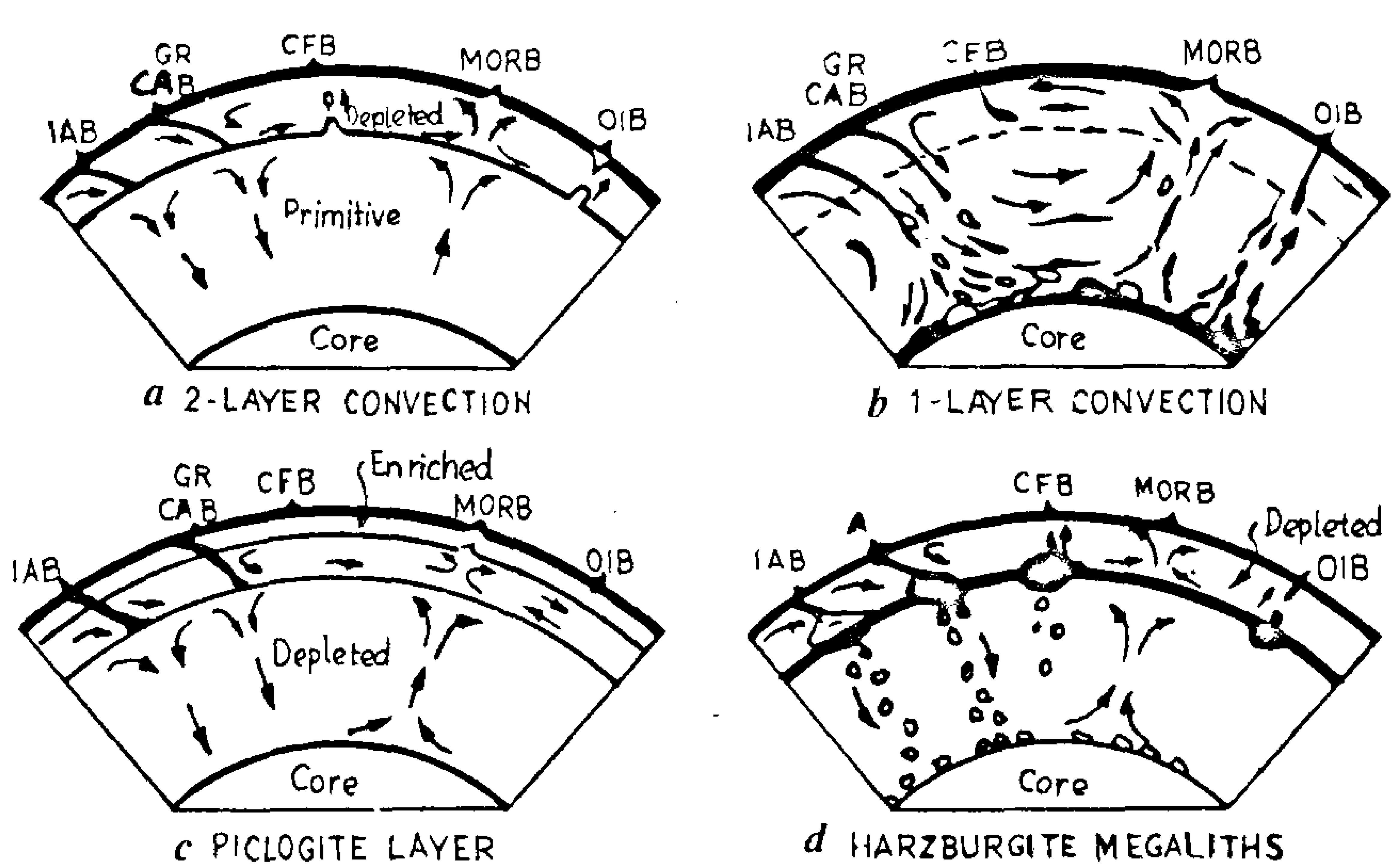

Figure 6. Convective cycles in the mantle as proposed by different workers (after Wyllie[31]) (for explanation, see text). IAB – Island arc basalt; GR/CAB – Granite/calc-alkalic basalt; CFB – Continental flood basalt; MORB – Mid oceanic ridge basalt; OIB – Ocean island basalt.

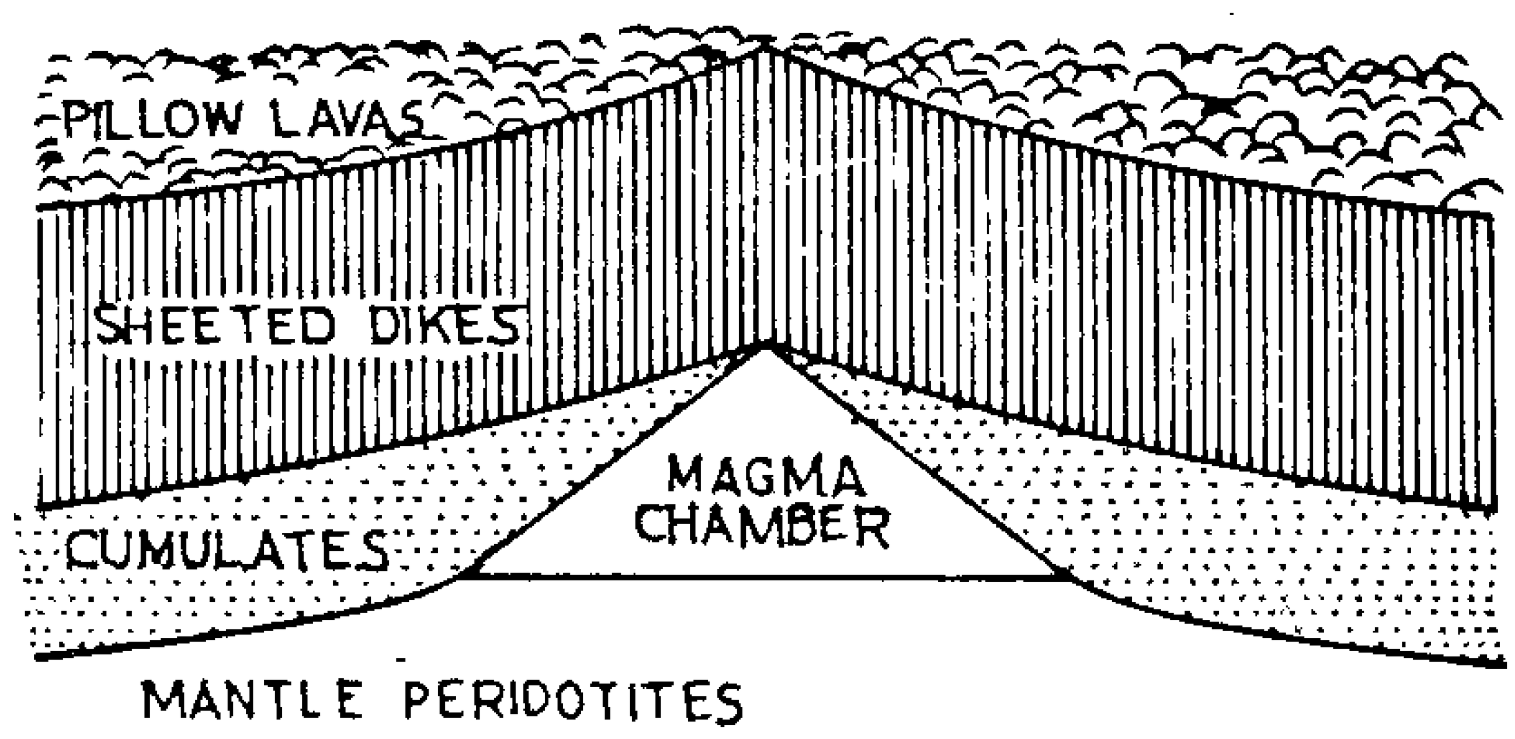

Figure 7. Illustration (not to scale) of the oceanic crust at a spreading centre (after Turcotte and Morgan[33]).

dyke complex. When the tensional stress accompanied by sea floor spreading is very large, fractures are generated within this magma chamber.

According to Cann[34,35], there are large magma chambers beneath the axes of the mid-oceanic ridges. Lavas are erupted through feeder dykes from the roof. Crystallization of the mafic magma in this region thus produces gabbro, whereas crystal fractionation produces mafic and ultramafic rocks at the floor of the dykes. This process has been termed as the 'infinite onion model' by Cann as it continuously peels off layers at the edges. Although the generation of magma may be continuous, only batches of liquids ascend upward buoyantly through the asthenosphere after reaching a certain critical size. The supply of magma to any high level reservoir is periodic and the rate of supply is proportional to the spreading rate. Prior to the addition of a new batch of magma to the chamber, the old magma materials already stored in the chamber undergo fractionation producing more evolved liquids. If the magma chamber is large, magma mixing is likely to occur, and large axial magma chambers may form in this process. This would in turn cause attenuation of earthquake shear waves. Seismic studies in the region of East Pacific Rise indeed support this model.

Nisbet and Fowler[36] however, believe that there is no attenuation of S-waves beneath the segment of the mid-Atlantic ridge below FAMOUS island, indicating that there is at present no magma pool larger than 2 km across below the ridge axis. They therefore cast serious doubt about the efficacy of this model, and consider that thermal constraints would not permit the existence of a large axial chamber beneath such a low-spreading ridge. According to Nisbet and Fowler[36], magma rises at high levels by a process of crack propagation through the brittle crust forming very small storage reservoirs at very high levels; such a model will result in the formation of a poorly layered oceanic crust.

The convective mass-energy flow from the interior is known to occur predominantly at the great narrow ocean ridge systems of the earth (the mid-Atlantic, Indian Ocean and the East Pacific Rise), which form a continuous submarine mountain range girdling nearly half the globe[32].

The submarine mountain chains are volcanic with a surface expression of basaltic lava flow. From all the mid-oceanic ridges about 15 km^3 of magma is produced annually, by rising hot mantle plumes that melt on release of pressure. About 90% of our planet's volcanism occurs at these submarine sites, and accounts for almost half the heat loss from the planet's interior. Fyfe[32] thinks that the rising convection cells have roots at depths of several hundred kilometers. Magma batches that rise into the ridges may at any time reach tens of km^3 in volume. Active magma chambers have only recently been mapped and single eruptions of 15 km^3 have been estimated. The ridge rocks are a crystalline product of essentially zero age and constitute the new crust of the Earth with a mean composition which is typical, the basalt representing the low melting fraction of the upper mantle beneath the oceanic crust.

According to Fyfe[32], early heat flow studies over and near the ridges produced a great surprise, given that the rising magmas extrude at and intrude below the surface at temperatures in excess of 1200°C. Very high heat flow values were accordingly to be observed.

However as Fyfe[32] explains, when a hot, porous-cracked and hence permeable, material is placed in a cold liquid (deep ocean water temperatures are normally close to 0°C) cooling will occur by water circulation. The heat flow patterns near the ridge would obviously be determined by the development of large scale convective cooling cells involving the penetration of cold sea water into hot rock. At the Galapagos spreading centre and now in many ridge situations, hot water has been found discharging at temperatures of 300–400°C. In some cases, the venting is even cataclysmic, reaching temperature in excess of 400°C.

Fyfe[32] considers that the scale of this water-cooling process is impressive, and as the most simple calculation of the energetics reveals, about half of the energy of the hot basaltic liquid is transmitted to the circulating ocean water. On an annual basis, something like 10^{18} g of ocean water can thus be heated to 100°C, or about one third of the mass to 300°C. Given that the total ocean mass is 1.4×10^{24} g, the entire ocean can be processed through the ridges in a few million years, which is merely an instant in geologic time scale.

Magmatism in the rift valley

The rift valleys are large trough-like valleys formed by subsidence of the floor between subparallel marginal faults. Rift valleys are zones along which the lithosphere is ruptured because of extension, resulting in the formation of brittle cracks.

Examples of rift systems throughout the world include,

(i) Afro-Arabian rift system (Cenozoic, < 65 m.y.), (ii) The Rio Grande rift (Cenozoic), (iii) the Rhone depression, (iv) The Rhine Graben, (v) The Baikal rift of south eastern Siberia, and (vi) The graben systems at Yinchuan, Hetao, Weihe, Shanxi, North China plains (all in the Precambrian north China platform).

Rifts are principally extensional tectonic features. Common association of rifts with extensive volcanism, high heat flow, seismicity and anomalous structure of the crust and the upper mantle provide compelling evidence that rifts are not just phenomena talking place in the upper crust, but are manifestation of dynamic processes deeper in the lithosphere and asthenosphere. Rift systems comprising interconnected branches are related to lithospheric extension, and may reflect incipient stages of an evolutionary sequence in which continents are rifted apart to create a new intervening ocean. Quite often large volumes of nepheline-normative alkalic basalts, carbonatites and highly potassic mafic and ultramafic lavas are associated with rift valley systems.

Magmatism associated with collision between two plates

A very active example of such a situation is that of the Himalayas. In case of continent–continent collision, a continental crust may be subducted in spite of its larger buoyancy thickness and rigidity. In such a process a small part of a continental plate may sink at a steep angle or may remain attached to the underside of the other plate. Magmatism in the Himalayan mountain chain resulted in the emplacement of S-type granites, which have very high $^{87}Sr/^{86}Sr$ ratios (0.780). This may suggest that the acid magma might have been formed by crustal fusion during the subduction of the Indian lithosphere below the Eurasian plate. Eruption of younger K-rich andesites, dacties and rhyolites took place in the Tibetan plateau as a result of continent–continent collision. In the Himalayan example, a vast metamorphic dewatering and melting event follows the major thrust event.

Collision between an oceanic plate and a continental plate

Seismic studies of the oceanic lithosphere and earthquake generation in trench regimes show that the oceanic lithosphere, about 100 km thick, bends and sinks or is thrust into the mantle and the process is essentially initiated along the ocean–trench axes.

At subduction zones, the volatile-laden oceanic crust, with a component of sediments, which may depend on the roughness of the surface of the oceanic lithosphere, drives into the mantle, recharging it with volatiles and species like Na, K, U, and carbon and sulphur compounds. Release of the volatiles must induce convection in the overlying mantle and mobilize the most soluble and fusible components, eventually leading to the production of basaltic-andesites. Granite-rhyolite production and prograde metamorphism essentially dehydrates the basal crust and leads to a host of ore-forming processes.

Similar processes occur during continental collision and on a small scale with strike–slip boundaries, where the crust may be thickened to 60–80 km.

According to E. G. Lidiak (pers. commun.), the Benioff zone, which underlies the trench, defines narrow bands which are broken apart into different segments 100 to 500 km long and are offset from one another. The arc volcanoes occur at spatial intervals of 30 to 70 km. The basalt near the oceanic trench is quartz-normative and that away from the trench, nepheline-normative[37]. Basalts from these regions are silica-oversaturated and comparable to basalts of oceanic ridges and back-arc oceanic spreading centres. Island arc lavas are higher in Al, Rb, Sr, Ba, Pb, CO_2 and H_2O and lower in Mg, Ti, P, Zr, Nb, Ni, Cr and have lower K/Rb, Rb/Sr, Fe/Mg, CO_2/H_2O, Cl/F and Sr isotopic ratio.

According to K. Yagi (pers. commun.), in the Pacific margin of north and south American continents there has been uplift of Precambrian and Palaeozoic rocks which have been subjected to volcanic activities during late Triassic and early Jurassic. Four magmatic rock types such as low K, calc-alkalic, high K and shoshonitic lavas, which are typical of island arc environment are all observed in the South American continental margin. In contrast to island arc volcanism, the proportion of such silica-rich members as dacite and rhyolites is more abundant in regions associated with continental margin volcanism. Here, acid volcanic rocks occur as pyroclastic flow materials in association with zones of thickened continental crust.

In the case of volcanism in the Andes there are three linear segments: (i) the northern and central volcanic islands which are 140 km above the Benioff zone, that dips at 25°C to the east. The continental crust in the northern segment is 40 to 50 km thick and that in the central part is 70 km thick. The southern volcanic zone lies 90 km above the Benioff zone. The three portions are separated by two zones which are not associated with volcanism and are underlain by subduction slabs which dip to the east. Basaltic andesites and andesites (SiO_2: 53–61 wt%; $^{87}Sr/^{86}Sr$: 0.7038 to 0.7044) are found in the northern volcanic segments. Earlier, the central part erupted rhyolites and ash flow tuffs but later the region has been dominated by andesite and dacite volcanism.

Intraplate volcanism

Burke and Wilson[38] suggested that as an oceanic plate moves, it may pass over hot spots or mantle plumes of hot rock which erupting on the surface would form a submarine volcano, or even an emergent cone. As such a volcano borne by the oceanic plate is carried progressively away from the centre of the hot spot, it would, cut off from the magma supply underneath become extinct. This would ultimately result in the formation of a chain of extinct volcanoes directed away from the hot spot in the direction of the sea floor spreading (e.g. Hawaiian Islands chain). It is observed that the Pacific island volcano chains (Hawaiian–Emperor, Gambier–Tuamotu, Austral–Cook and Gilbert Mershall) swerve in direction from SE–NW to slightly west of north. Such a turning of their alignment is related to a change in the direction of the spreading of the Pacific plate 40–50 Ma. Burke and Wilson recognized a global pattern of 122 hot spots, which have been active during the past 10 Ma both within oceanic and continental plates. They identified 53 oceanic hot spots, which tend to be located close to the mid-oceanic ridges.

Kimberlites, carbonatites, feldspathoidal rocks, and A-type granite typically occur in an intraplate continental setting, although many of them may occur in other tectonic regions also. The vent areas of flood basalts however may have been initially formed within continental rifts that subsequently evolved into oceanic margins and covered by continental shelf sediments. The division of some magmatism between cratonic and rift may thus be difficult. They seem to have been derived from a subcontinental magma source, albeit flood basalts, the magma source being deep down in the mantle.

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This paper relates to mantle dynamics and its manifestation in the form of magmatism in different plate tectonic environments. Earthquake-related events produced as a consequence of mantle dynamics will be discussed elsewhere.

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