Geologie en Mijnbouw 65: 91–101 (1986)

Archaean global dynamics

N.J. Vlaar
Department of Theoretical Geophysics, Institute of Earth Sciences, University of Utrecht, P.O. Box 80.021, 3508 TA Utrecht, The Netherlands

Received 7 March 1986; accepted in revised form 1 June 1986

Abstract

Archaean global tectonics and continental growth is founded on the existence of a differentiated upper mantle of average lherzolite composition. A thick segregated basalt layer constitutes the upper part of the upper mantle. The geodynamic state is determined by a high deeper mantle temperature and a mobile cool shallow basalt layer and this can explain the growth of the Archaean craton and its geothermal state. Of particular importance for continental tectonics are isostatic conditions which differ from the present in allowing for large relative and absolute vertical displacements.

Introduction

There is no direct evidence in the Archaean geological record that oceanic crust and lithosphere, if existing, were involved in continental evolution. In order to understand this evolution, insight must be gained in the global tectonic regime in general. Most studies regarding Precambrian global tectonics and its role in the evolution of the continent tend to conform to an extrapolation of present day plate tectonics, sometimes taking into account higher mantle temperatures. The latter is then assumed to affect lithospheric thickness, the vigour of plate motions, and local geotherms. Most of these studies incorporate aspects of present day plate tectonics without questioning their validity in the context of a different thermal regime.

Plate tectonics as it manifests itself is an intricate process of interaction between temperature dependent parameters. The process of differentiation and segregation of the oceanic lithospheric layering is strongly dependent on mantle temperature, and this is particularly so for the thickness of the segregated layers. The stability of the layering when created at a spreading ridge depends on this thickness, and hence on mantle temperature (Vlaar 1985). Plate tectonics requires a stable lithosphere layering lasting long enough for the plates to acquire sufficient strength to be able to act like rigid plates. It has also been demonstrated that the driving forces of plate tectonics are strongly temperature dependent (Vlaar & Wortel 1976), which may influence the tectonic interaction at a convergent plate boundary (Wortel & Cloetingh 1985). Regarding Precambrian, and in particular Archaean global tectonics, the question should be posed as to what extent present day plate tectonic conditions are fulfilled (Vlaar 1985). High mantle temperatures as indicated by Archaean komatiites may have influenced global dynamics in such a way that the plate tectonic concept loses its validity altogether. Extrapolation of present to Archaean conditions therefore may not lead to a proper understanding of geology.

Apart from those who advocate the plate tectonic mechanism also for the Archaean, there are others who invoke a different global tectonic style in order to comply with geological data (e.g. Kroener 1985), though the subject has not become less ambiguous. One of the outstanding problems of the Archaean is the discrepancy between high mantle temperatures as indicated by komatiite extrusion temperature and the relatively low temperatures determined from mineral paragenesis in high grade terrains. A further problem concerns the excessive relative and absolute vertical displacements in Archaean continental crust, which, on the one hand enabled supracrustals to be buried to depths of 40–50 km, and on the other hand caused an uplift of the same order of magnitude to bring the high grade terrains to the present day surface. The petrogenesis and extrusion of dense komatiite liquids also is not well understood.

The present paper addresses these and related problems and places them in the framework of a model for Archaean global tectonics. I depart from the assumption that the earth's mantle had acquired its basic stratification as we know it today, as early as the Archaean. In particular is it assumed that the upper mantle had been segregated and could act as a reservoir for the differentiation of the continental crust, and that global tectonic processes were determined by the dynamics of the upper mantle. A subtle role is played by the interplay between the geothermal and geodynamic regime on the one hand, and the melting of basalt on the other hand.

Archaean global tectonics

Crucial data concerning the state of the earth in its early stages are the rate of heating by radioactive elements, and the melting temperature of ultramafic peridotitic komatiite magma. Radioactive heating in the Archaean is estimated to be two to three times its present day value. The komatiite melting temperature is indicative for the temperature temperature of the Archaean upper mantle, that is in excess of 1650° C (Bickle et al. 1977). The corresponding temperature of the upper mantle, at a depth of some 100 km, taking into account possible latent heat of fusion and adiabatic selfheating, should thus be put in excess of 1750° C. This temperature can be taken to be representative for the late Archaean, the last occurence of high MgO komatiite lavas. Earlier Archaean then should be characterised by even higher mantle temperatures. A consequence of an increase of mantle temperature is the decrease by an order of magnitude of the apparent viscosity with each 100° C. (Weertman 1970), implying a very low viscosity in the subsolidus part of the Archaean upper mantle, and a strongly enhanced convective regime. Convective and advective heat transport must have been a dominant feature of Archaean global tectonics. In view of the vigour of convection in the early stages of the earth's evolution, the gross chemical layering of the earth should have been virtually completed in the beginnings of the Archaean era. Mass exchange between separate mantle layers hereafter could only take place at a substantially reduced rate. Seismological and high pressure research suggests that the upper mantle transition layer between 400 and 700 km depth must be explained by heterogeneous chemical layering (Bass & Anderson 1984), whereas the shallower upper mantle satisfies a grossly homogeneous peridotite petrology. One can therefore apply the model of a grossly homogeneous shallow upper mantle of undepleted peridotite (lherzolite) composition as the main reservoir for crustal accretion since early Archaean. High temperature and vigour of convection must have resulted in dehydration and degassing, leaving us with a dry upper mantle reservoir. Mass and heat transport by means of diapirism and magmatism is of a penetrative character. A lherzolite diapir, rising along an adiabatic ascent path, traverses the basalt melting range. I follow Sleep & Windley (1982), who devised a schematic model for the differentiation of a rising lherzolite diapir (figure 1). This model provides for an almost complete melting of the basalt fraction between the 5 and 30% melting curves for dry lherzolite. The melting curves are assumed to be linear in the temperature – depth plane. According to Arndt (1977), segregation of the melt from its matrix is virtually total, even at very small melting percentages. Therefore, a lherzolite diapir which has traversed the melting range, leaves it with a basalt liquid fraction, which segregates, leaving behind a solid depleted harzburgite layer. The thickness of the basalt layer on top depends on the temperature at which the lherzolite diapir has entered the melting range. It is assumed that within the melting range, due to the latent heat of fusion, the temperature drops 100° C along the adiabatic ascent path. Basalt magma, reaching the earth's surface at 1650°C would be generated by a lherzolite diapir entering the melting range from below at 1800° C and 140 km depth. Whether also komatiite follows the same adiabatic trajectory with a 100° C drop in temperature due to latent heat depends primarily on the way it is generated. The above process combined with vigorous convection, segregates the upper mantle into a layering of basalt- harzburgite- lherzolite, the thickness of the layers depending on mantle temperature. As shown in Fig. 1, in case of a 1650° C extrusion temperature the basalt layer is about 40 km and the harzburgite layer about 100 km thick. The peridotite layer still residing in the melting range has not been depleted completely of its basalt fraction. At higher mantle temperatures earlier in the Archaean, the segregated layering should have been correspondingly thicker. If cooling of the top layers would not take place, the basalt layer would be in a completely liquid state, whereas the harzburgite layer should be solid, apart from a possible minor amount of partial melt.

The above scheme results in a gravitationally stable stratification of the upper mantle layers with respect to each other, though the separate layers, being grossly homogeneous, could acquire internal instability. The process of partial melting and segregation appears to be irreversible at a fixed mantle temperature. Reversibility can occur when the temperature has lowered sufficiently to enable the basalt- eclogite phase transition to take place. This can be the case if the upper basalt layer remains solidified over longer periods, as the reaction time for the phase change has been demonstrated to be of the order of tens of millions of years for dry basalt at moderate temperatures (Ahrens & Schubert 1975). If the phase transition depth is 50 km, an upper basaltic layer of less than 50 km depth could survive without becoming negatively buoyant. If, at some stage, eclogite forms, it may sink to greater depth because of greater density. Although segregation of liquid basalt from lherzolite appears to be effective, it is unclear whether the reaction of eclogite with residual harzburgite may easily take place. It is thus not possible to decide whether eclogite reacts with harzburgite to form 'pristine' mantle, again or whether it sinks to a deeper level in the upper or lower mantle.

It seems clear that the Archaean has been characterised by a thermal regime which promoted the existence of a thick segregated basalt layer of 40 km or more. Although the gross upper mantle layering was gravitationally stable, the separate layers may have suffered internal instability because of being largely homogeneous. The strongest instability should have resulted from cooling and solidifying at the top and thus would cause sinking of cooled solid basaltic crust which subsequently could remelt again. In the early Archaean this situation should have resulted in a predominantly 'average' liquid basaltic layer, whereas towards the late Archaean, at lower mantle temperatures, this layer should have been largely solid. Strong convective instability, the layer being episodically partly above the basalt liquidus, is an effective mechanism for the earth's heat loss. The rate of cooling could only be limited by the slower rate of subsolidus convection in the layers below. Cooling of the upper basalt layer results also in conductive and convective cooling from the top of the lower harzburgite layer downward. When convective instability sets in, reheating starts. This should lead to episodic convective mobility with secularly decreasing vigour during the progress of the Archaean.

A conclusion to be drawn from the above is that Archaean global tectonics were determined by the fact that the shallow upper mantle, through heating from below, was episodically above its liquidus in its lower parts and evolved towards a more solid state. Archaean geodynamical conditions must therefore have been strongly different from those at present.

Modern global tectonics is determined by the

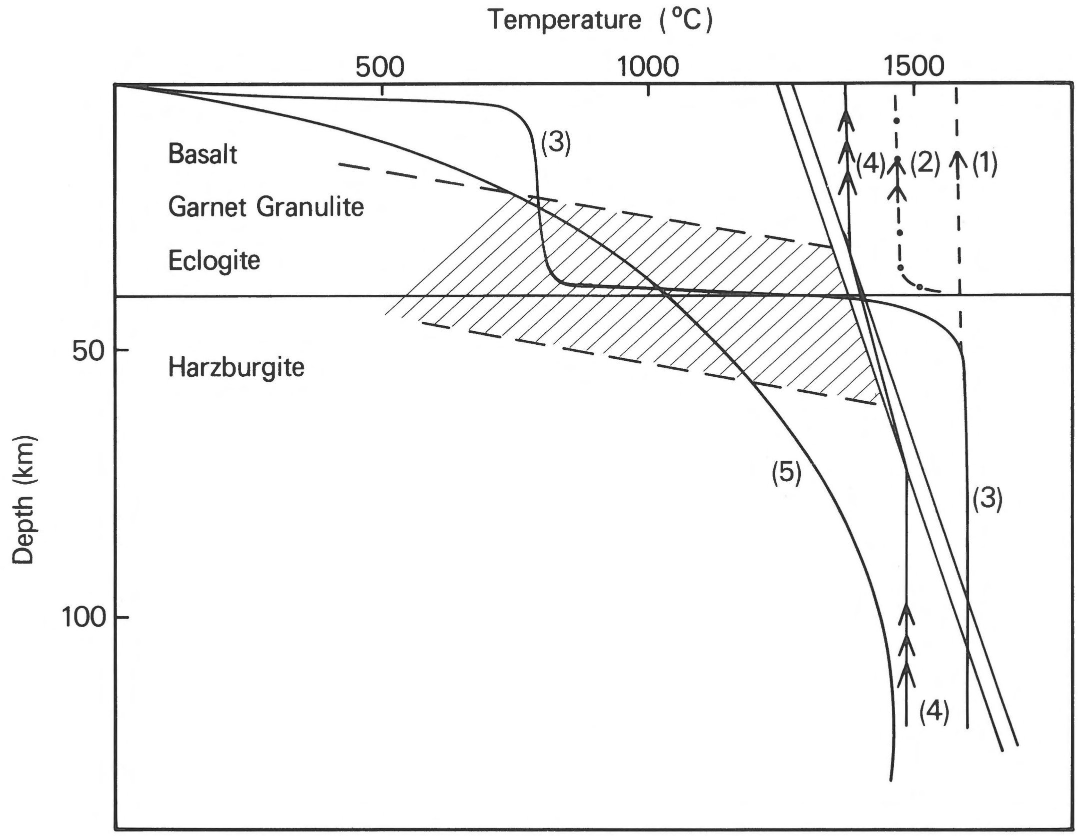

Fig. 1. Archaean geodynamic regime: (1) Ascent path of partially melted harzburgite (komatiite). (2) Associated path of liquid basalt, at lower temperature because of latent heat of fusion. (3) 'Average' geotherm in a convective regime in the basalt layer, with thermal boundary layers at the top and the bottom. (4) Adiabatic ascent path of segregating peridotite at a mid-ocean ridge, extruding at 1375° C. (5) Associated conductive geotherm in a stagnant and cooling lithosphere.

Shaded area: schematic basalt-garnet granulite-eclogite transition. Double bars: 5–30% melting range of lherzolite (after Sleep & Windley 1982).

circumstance that a thin segregated basaltic layer is cooled below its solidus virtually immediately after its creation at an oceanic spreading ridge, and finds itself subsequently in a gravitationally stable position with respect to the underlying upper mantle layers. Stability persists to the point where thermal contraction and consequent increase in density offsets the compositionally stable layering. For modern oceanic lithosphere this occurs at a lithospheric age of about 30 Ma (Vlaar & Wortel 1976, Oxburgh & Parmentier 1977). By this age the lithosphere has acquired sufficient strength to prevent spontaneous downbuckling of the upper layers and lithospheric plates are strong enough to transmit the large stresses required for plate tectonics. This situation, however, should be ruled out for the Archaean, as stability of the surface layers, and thus sufficient strength, could not be reached in a regime which was episodically above its liquidus. Archaean global tectonics can thus be envisaged to have been chaotic and of a rather random style. However, following Condie (1984), a tonalitic low density protocontinent in the early Archaean, as a result of partial melting of near surface solidified basalt, appears to be plausible.

In general, Archaean petrology should find its foundation in basalt as source material. The Archaean thermal and tectonic regime can be cast between the extremes of a completely melted basalt layer at a temperature which is compatible with upper mantle temperature, and a completely solid basalt layer which has been cooled effectively from top to bottom. The thermal relaxation time associated with conductive heating of a layer 50 km thick, is well known to be about 20 Ma (e.g. England & Bickle 1984). It therefore takes about this time to arrive at an equilibrium geotherm which is compatible with the temperature below. However, before this temperature is reached, melting of the basalt layer starts from the bottom and initiates diapirism and magma penetration to a shallower level, thereby reducing the reheating time considerably. At the other extreme of a completely melted basalt layer on a harzburgite substratum, cooling at the surface and solidification results in the sinking of basalt and rapid convective cooling of the entire basalt layer within a much shorter time than for conductive cooling. Although convective and conductive cooling from the top and reheating from the bottom appears to be both spacially and temporally a random process, the average thermal state could be represented by an average boundary layer model (figure 1). Boundary layers and large thermal gradients are involved on top and bottom of an adiabatic gradient core. The lower boundary layer involves cooling of the top of the harzburgite solid substratum. The 'average' convective geotherm being below the basalt solidus, implies an average solid state of the basalt layer. Reheating from below gives rise to melting of the latter's base and thus to episodic magma generation and convective overturning. On a secular time scale the above mechanism will come to a halt when the mantle temperature has dropped to the point that melting of the base of the basalt layer becomes exceptional. Apart from incidental periods of magma generation, the average geothermal state then is characterised by a basalt layer which is predominantly below its solidus and a geotherm which is mostly of a conductive type. This situation should dominate the early Proterozoic.

Archaean high grade terrains

Episodic and random weakening of the upper basaltic layer by penetration of basaltic lavas from below, and the consequent convective instability, together with the high mantle temperatures prevailing during the Archaean, must have prevented modern type plate tectonics and subduction. It is not ruled out, however, that locally and episodically weak compressional and tensional stress regimes may have occured and mobility of a differentiated protocrust may have resulted. An early protocontinent could have been formed by the sweeping together of 'tonalite islands' (Condie 1984), floating on a basaltic substratum, and formed by the partial melting of basalt which had solidified at the earth's surface. From early Archaean onwards, continental fragments have existed, and moreover, crustal thicknesses may have exceeded modern values. From the foregoing it ensues that Archaean continental growth should have been caused mainly by the addition to the crust by magmas derived from a basaltic parent.

Archaean high grade terrains which at present are exposed at the earth's surface, contain metamorphic mineral assemblages which reflect burial depths at intermediate crustal levels of 30–40 km at temperatures of 750–850° C (Tarney & Windley 1977, Perkins & Newton 1981). At present, Archaean high grade terrains are underlain by a normal continental crust of 30–40 km thickness. If the present crust has not been thickening by later underplating, it should have reached thicknesses of 70 km at the late Archaean. Pressure-temperature relations suggest that geotherms in Archaean continental crust did not differ essentially from modern shield geotherms. Experimental data from kimberlite nodules estimated by Mercier & Carter (1975) give 500° C at 40 km depth. This contrasts paradoxically with the 2–3 times higher heat production in the Archaean and the high mantle temperatures inferred from komatiites. An equilibrium geotherm satisfying these conditions does not appear to be feasible as can be concluded from the following arguments.

A one- dimensional stationary temperature field can be obtained by integrating the heat conduction equation, giving:

$$q(z) = k\,dT/dz = -\int_0^z Q(z')\,dz' + q(0)$$

$$T(z) = \int_0^z dz' \, [q(0) - \int_0^{z'} Q(z'') \, dz'']/k(z')$$

where z is the depth, T(z) the temperature, k(z) the coefficient of thermal conduction, q(z) the heat-flow, q(0) the heatflow through the earth's surface, and where T(0) = 0. Assuming that both the depth distribution of radioactive elements and the thermal conductivity are equal to modern values, the threefold heat generation in the Archaean and hence threefold heat flow, causes the temperature T(z) to be three times the present continental temperatures. Following Mercier & Carter (1975) this results in 1500° C at 40 km depth in Archaean high grade terrains. This is about double the value found by pyroxene geothermometry. If all heat producing elements were concentrated shallower than 40 km, the required low temperature of the high grade terrains could be reached. In this case however, heat flow from below 40 km should vanish, which is conflicting strongly with high upper mantle temperature. If all radioactive heat production were concentrated below 40 km, a temperature close to 2000° C at this level would follow, which is also conflicting with petrological findings. The petrological data can only be explained by assuming an extremely high thermal conductivity in the upper 40 km, and an extremely low one below this depth, combined with low crustal heat production. England (1979) indeed demonstrated that under these extreme circumstances a stationary geotherm can be found to match petrological constraints. As there is no reason to assume that thermal conductivity in the Archaean differed drastically from present day values, a stationary geotherm complying with high mantle temperatures as indicated by komatiite melting, should be ruled out. If an equilibrium geotherm existed during longer periods in the Archaean, the conclusion should be that the temperature at deeper crustal and subcrustal levels did not differ substantially from present day values. This indeed is consistent with the model of a basaltic substratum, which by rapid convective cooling is most of the time at a low temperature not exceeding 1000° C, and which, episodically is perturbed by magmatic events.

Wells (1980) considered the thermal history of episodic over- and underaccretion of magma and inferred that the high grade metamorphic conditions are compatible with intrusion and extrusion temperatures in excess of a stationary geothermal state. However, he assumed that these temperatures are in a range of 850–1000° C, and hence, considerably below the extrusion temperature of komatiite. England & Bickle (1984) ascribed the relatively low temperature in high grade terrains to tectonic crustal thickening and subsequent rapid erosion. Nonetheless, these authors concluded that the Archaean continental thermal regime is not appreciably different from the present one. In view of the foregoing, a cool shallow upper mantle of basaltic composition, as proposed by the present global tectonic model, appears to be justified by the petrological data of Archaean high grade terrain.

The komatiite problem

The petrology and magmatism of komatiites have yet evaded an unambiguous understanding. High MgO (32%) ultramafic komatiite lavas are restricted to the Archaean, and in particular to the granite- greenstone belts. Though their origin may be problematic, their melting temperature of 1650° C is indicative of the temperature of their mantle source and the Archaean earth's interior. Some current ideas concerning komatiite petrogenesis center on the question whether komatiite has been generated by melting of a fertile lherzolite mantle source or whether it is the product of multistage melting. Deriving komatiite from fertile lherzolite, in view of the high MgO content, requires a very high degree of partial melting of the mantle source rock near the peridotite liquidus (Green 1975). Mantle temperatures in excess of 1800° C at a depth of 100 km are required. However, at this temperature, the low density basalt fraction should have left the source and have accumulated at a shallower level leaving behind a depleted peridotite mantle. An alternative is the suggestion that komatiite is the initial partial melting product of fertile lherzolite, generated at still greater depths in excess of 150 km (O'Hara et al.

1975). Nisbett & Walker (1982) rejected this alternative on the grounds that it can not provide for the voluminous extrusion of komatiite lavas. Moreover, also in this case, basalt would be the first fraction to be segregated. From the foregoing it can be concluded that Archaean basalts and komatiites cannot be derived simultaneously from the same mantle source. Before komatiite would be generated, the basalt should have been extruded, leaving a depleted source behind. This suggests that basalts and komatiites in greenstone belts are derived from spatially and compositionally different sources. In the model, discussed in this paper, these sources can be identified as the upper basalt layer and its harzburgite substratum, respectively.

A difficulty concerning the emplacement of lavas in greenstone belts is their stratigraphic position. Invariably, komatiites are extruded first in the early stages of greenstone belt formation, and are overlain stratigraphically by basaltic komatiites and basalts. This requires a komatiite liquid to be formed before basalt is melted. In view of melting temperatures, at first sight, this seems to be impossible unless their separate sources have strongly differing melting temperatures, the komatiite at or near its liquidus, and the basalt below its solidus. This can be achieved by the global tectonic model presented in this paper. Komatiite lava is supposed to be generated by partial melting of depleted peridotite. The harzburgite layer in the present model is assumed to be the komatiite source. The requirement for separate sources for basalt and komatiite is hereby met. Komatiite appears to possess low trace element and incompatable element abundances, and light depleted R.E.E. patterns (Weaver & Tarney 1979) indicating a depleted mantle source.

The model must thus be completed by allowing for a komatiite layer, derived from harzburgite, at the base of the shallow basalt layer, possibly grading upward into basaltic komatiite and basalt. This layering is completely stable, unless komatiite is in the liquid and basalt is in the solid state. Komatiite liquid (at 1 atm.) and at liquidus temperature has a density of 2.8 g cm^{-3} (Nisbett & Walker 1982) which is definitely lower than the density of a basaltic solid (3 g cm^{-3}). The situation of a solid basaltic layer resting on melted komatiite, in our model, could be realised if after a cooling episode, reheating starts from below. Komatiite, basaltic komatiite, and basalt should then melt in succession. After the basalt has started melting, komatiite and basaltic komatiite cannot rise to the surface any longer. The extrusive succession with komatiite at the base of the lower volcano- sedimentary sequences is a general feature of Archaean greenstone belts.

Tectonics and magmatism

Continental crust had differentiated from the mantle already in early Archaean and was floating on denser mantle. The Archaean continental gradient was not essentially different from the present day one, indicating an average cool shallow upper mantle. The latter, as has already been made evident, had to consist of a thick layer of a grossly basaltic composition. Episodic heating from below resulted in convective overturn of this layer and thus in effective heat loss of the earth and accompanying magma generation, predominantly of a basaltic source. Accretion of the Archaean continent was mainly due to magmatic additions of basalt or its fractionates, i.e. tonalites and granodiorites. Thermal perturbations caused by intrusions gave rise to large scale anatexis of the continental crust which resulted in crustal remobilisation and generation of more felsic fractionates and emplacement of granitic batholiths to a shallower level. Anatexis and intrusion should have caused gravity tectonic features like folding and apparent crustal shortening. Greenstone belt vulcano-sedimentary sequences laid down in subsiding intracontinental basins can be ascribed to stretching and thinning of continental crust and lithosphere (Hawkesworth et al. 1975, Bickle & Erickson 1982, Nisbet 1984). A typical feature is that greenstone belts rest unconformably on continental gneissic basement. Following initial subsidence typified by shallow water sediments, the first extrusive magmas are dominantly komatiites.

Vlaar (1982) ascribed crustal stretching, initial subsidence, and subsequent volcanism, precursive to rift formation, to crustal underaccretion of hot material that is lighter than surrounding mantle rocks. For the formation of greenstone belts, this light material should originate from the base of the shallow upper mantle layer and consist sequentially of liquid komatiite, basaltic komatiite, and basalt with decreasing density. Located below the lighter continental crust, it should cause anatexis, granite diapirism, and crustal stretching. Greenstone belts could only survive if deposited on a lighter continental crust. In an oceanic environment, the basement would be loaded to the point of negative buoyancy. In the oceanic realm, following Condie (1984), partial melting of solidified basalt near the surface, generated tonalite. Therefore, tonalite may be a main constituent of buoyant primitive continental crust. This crust could thicken by subsequent addition of underaccreted magmas. As large horizontal stresses in a plate tectonic setting do not appear to be possible, compression is not considered to be a cause of Archaean crustal thickening.

A gradually thickening continent and a secularly cooling substratum may have changed the style of magmatisme during the Archaean into a more felsic character. Particularly, the more common occurrence of andesite in later Archaean can be ascribed to such a changing environment.

An important feature of Archaean tectonics is that the isostatic conditions differ in an essential fashion from now. Whereas modern continental and oceanic crust rests on a dense peridotite substratum, Archaean continental crust, according to the foregoing, was underlain by low density basalt. Isostasy is controlled by the density contrast between crust and substratum. Taking an Archaean continental crust of d km of density ϱ_c floating in a basaltic substratum of density ϱ_b, a simple model of isostatic equilibrium yields:

$$d(\varrho_b - \varrho_c) = h\varrho_b - h_w\varrho_w$$

where h is the height of the continental surface above the ocean bottom and h_w and ϱ_w are ocean depth and density respectively. We have assumed that h is larger or equal to h_w, i.e. that the continent is elevated above sea level. The case $h = h_w$ also represents $h < h_w$, i.e. that the continental crust is overlain by the ocean.

Archaean high grade terrains contain a.o. supracrustals of shallow water origin which have been buried subsequently to depths of up to 40 km. The sediments presumably have been derived from volcanic complexes rising above sea level. Taking $h = h_w$, continental surface being at or below sea level, $\varrho_w = 1$, $\varrho_c = 2.8$, and $\varrho_b = 3\,g\,cm^{-3}$, we obtain $d = 10\,h$. Taking for the average ocean depth $h = h_w = 4$ km, we arrive at a crustal thickness of 40 km in this case. This implies that a crust of less than 40 km would be submerged in an ocean with a depth of 4 km. Taking $h_w = 4$ km and $d = 30$ km, the continental surface is 1 km below the ocean surface. For $d = 50$ km, $h = 4.75$ km, the continental surface being 0.75 km above sea level. These simple isostatic considerations have considerable implications. Due to the small density difference $\varrho_b - \varrho_c = 0.2\,g\,cm^{-3}$, a thick crust can exist, thickness having only a small influence on the height of the continental surface relative to sea level. Loading of the crust from above under isostatic conditions, adding d' km on top, only results in (1/15) d' km increase of height. On a crust which has been weakened by intrusions such that isostasy can be easily accomplished, the deposition of large piles of sediments is possible without influencing the depth of deposition in any considerable way. By maintaining isostatic conditions, large relative and absolute vertical displacements of columns of continental crust are possible. In view of this, large volcanic complexes rising above sea level and nearby depositional basins are required to invoke large relative vertical displacements. Intrusion of batholiths and consequent ductile regime, combined with relative vertical movements, suggests gravity tectonics to be the main agent for supra- and intracrustal movements. The characteristics of Archaean continental crust are therefore a large thickness combined with a moderate surface relief, and in order to preserve isostasy, large relative and absolute vertical displacements. Archaean tectonics, in view of the foregoing, supports the existence of a light basaltic substratum.

Thinning and stretching of the continental crust caused by underaccretion of komatiitic to basaltic lavas, leading to the formation of ensialic greenstone belt basins, is also accompanied by considerable subsidence. In order to maintain shallow water sedimentation in the initial phases of basin formation (Bickle & Eriksson, 1982), and due to the smaller density difference between the continental basement and its (partly) fluid substratum, a much thinner crust is required. Taking again $h_w = 4$ km, and $h = 3$ km, which is approximately valid for a shallow water environment, and for the density difference $\varrho_b - \varrho_c = 0.1$ g cm^{-3}, we find a crustal thickness $d = 30$ km. For $h = 2$ km, that is for sediment deposition at 2 km below sea level, $d = 20$ km. These values for the crustal thickness including the greenstone pile, indicate considerable crustal thinning due to the intrusive uprise of the (partly) fluid substratum. Subsequent solidification and conductive cooling provides for protracted deepening of the basin. Due to the specific isostatic conditions, extremely deep ensialic basins can be formed. Oceanic crust would not be able to maintain a gravitationally stable stratification, as has been argued in the foregoing, and could therefore not serve as a sedimentary basin.

The Archaean-Proterozoic transition

The transition from Archaean to Proterozoic is marked by a change in tectonic style. The permobile Archaean regime gives way to more stable lateral growth of the continent by the formation of continental platforms. Greenstone belts can be traced into the Proterozoic, though MgO komatiites appear to be restricted to the Archaean proper. The latter indicates that upper mantle temperature had dropped to below the komatiite melting point. The model of this paper requires an almost stable solidified basaltic substratum, which only incidentally could be heated from below to above the basalt liquidus and give rise to early Proterozoic ensialic thermal events. The internal stability of the basaltic subcrustal layer, with a dominantly subsolidus geotherm, as would be prevailing, should induce the basalt to undergo the high pressure phase transition to garnet- granulite below a certain depth. This phase transition in dry basalt has an extremely long reaction time in the order of tens of millions of years at intermediate crustal depths (Ahrens & Schubert 1975). By the limited depth of some 50 km for the basalt upper mantle layer, the eclogite phase change cannot, or can only marginally, be reached. The garnet- granulite transition can take place at the shallower level of some 15 km. The density of garnet- granulite (3.2 g cm^{-3}) thus allows a stable position with respect to underlying harzburgite (3.3 g cm^{-3}). This transition should have far reaching consequences for already stabilised Archaean cratons. The high grade terrains with their deep roots, probably in excess of 40 km, due to isostatic requirements, should be pushed upward due to the densification having taken place. Taking an average ocean depth of 4 km, densities as above, and densification below 15 km from 3 g cm^{-3} to 3.2 g cm^{-3}, an average surface uplift of a 40 km thick crustal pile is seen to be 1.3 km above sea level. For a 50 km thick crust this figure becomes 2.57 km. The densification event leads to a considerable uplift of the thicker parts of the Archaean cratonic crust and thus generates a source for sediments to be deposited on the now stabilised ocean bottom. Erosion in turn generates continuing uplift. When the erosion base has returned to sea level, and erosion ceases, the thickness of the continental crust has been reduced to 29.5 km.

A crust of 40 km before uplift thus lost 10.5 km on top by erosion, and a crust of 50 km lost 20.5 km. These figures are too low to explain the palaeopressures of over 10 kbar in the high grade terrains. This discrepancy can be overcome by having topography above sea level. Taking the several parameters as before, and taking the continental surface at 1 km above sea level, the equilibrium crustal thickness after erosion is 17.5 km. Crustal piles of 40 and 50 km respectively are reduced by erosion to 23.5 and 33.5 km respectively. These amounts are of the required magnitude.

The proposed mechanism accounts for the extremely large uplift of Archaean high grade terrains. Erosion and subsequent deposition on stable ocean floor accounts for the creation of large Proterozoic shield platforms, and hence, lateral accretion of the continent at the expense of Archaean cratonic crust. Considering that greenstone belts have been generated as basins in a stretched continental crust, and assuming that the total depth of the crust included these belts and did not exceed 17.5 km, the greenstone belts would not be subject to uplift like the adjacent high grade terrains. The basalt-garnet granulite phase change is most probably the future Moho. In the above model, it is situated at a depth of 20 km below the present day average continental surface. A larger and more correct value could be obtained by adjusting the depth of the phase transition and simultaneously increasing original crustal thickness by the same amount.

Nothwithstanding uncertainties in the model parameters, the foregoing model of Archaean isostasy yields a good approximation to observed geology.

Concluding remarks

The upper mantle temperature is the controlling parameter for the evolution of global tectonics. However, contrary to what is usually assumed, the relation is rather intricate. Dynamical quantities which are strongly influenced by mantle temperature, are, apart from viscosity, also the extent of segregation of mantle peridotite, the thickness and stability of the segregated layering, and especially the geothermal state relative to the melting curves of peridotite and basalt. The latter plays a subtle role in determining the thermal and dynamical state of the upper mantle layering.

The Archaean in particular is characterised by a thick (50 km) shallow upper mantle of basaltic composition resting on top of a depleted harzburgite layer. The Archaean regime is determined by convective cooling through the shallow basalt layer from the surface downward, due to internal instability of this layer. A main effect is that the latter is predominantly in the solid state, and is therefore cooler than is compatible with a conductive regime complying with large heat production and high upper mantle temperature. Episodic reheating from the bottom upwards to above the melting temperature can mobilise komatiitic and basaltic lavas. Hereby the paradox can be explained that Archaean mantle can generate high temperature komatiitic lavas and at the same time can maintain the relatively low temperature as is indicated by mineral paragenesis in high grade terrains.

A major effect of the upper mantle stratification is the deviating response of Archaean crust to loading and unloading. The small density difference between crust and shallow basaltic upper mantle favours large vertical movements and generation of thick crustal piles by sedimentation and volcanism. The mechanically weakened crust is particularly suited for vertical and gravity tectonics caused by the uprise of plutons. The continental crust cannot support a large relief above sea level for protracted periods, and large horizontal stresses as in modern plate tectonics must be ruled out.

After sufficient cooling and consequent increase of stability within the dominantly subsolidus shallow upper mantle, densification due to the basalt-granulite phase change gives rise to considerable uplift of the high grade gneiss terrains. This event is assumed to have marked the Archaean- Proterozoic transition, large scale erosion of the Archaean cratons, and deposition of thick layers of sediments on stabilised oceanic crust. The advent of plate tectonics had yet to be initiated.

Important questions to be answered are related to beginnings of plate tectonics and possible remixing of the upper mantle layering by cooling, phase changes, and consequent remobilisation.

Acknowledgements

The author acknowledges financial support for his research by the 'Stichting Dr. Schuermannfonds.'

References

Ahrens, T.J. & G. Schubert 1975 Gabbro- Eclogite transition and its geophysical significance – Rev. Geoph. Space Phys. 13: 383–400

Arndt, N.T. 1977 Ultrabasic magmas and high degree melting of the mantle – Contr. Mineral. Petrol. 64: 205–221

Bass, J.D. & D.L. Anderson 1984 Composition of the upper mantle: geophysical tests of two petrological models – Geoph. Res. Lett. 11: 237–240

Bickle, M.J., C.E. Ford & E.G. Nisbet 1977 The petrogenesis of peridotitic komatiites: Evidence from high pressure melting experiments – Earth Planet. Sc. Letters 37: 97–106

Bickle, M.J. & K.A. Eriksson 1982 Evolution and subsidence of early Precambrium sedimentary basins – Phil. Trans. R. Soc. Lond. A 305: 225–247

Condie, K.C. 1984 Archaean geotherms and supracrustal assemblages – Tectonophysics 105: 29–41

England, P.C. 1979 Continental geotherms during the Archaean – Nature 277: 556–558

England, P. & M. Bickle 1984 Continental thermal and tectonic regimes during the Archaean – J. Geol. 92: 353–367

Green, D.H. 1975 Genesis of Archaean peridotitic magmas and constraints on Archaean geothermal gradients and tectonics – Geology 3: 15–18

Hawkesworth, C.J., S. Moorbath, R.K. O'Nions & J.F. Wilson 1975 Age relationships between greenstone belts and granites – Earth Planet. Sc. Lett. 25: 251–262

Kroener, A. 1985 Evolution of the Archaean continental crust – Ann. Rev. Earth. Planet. Sci. 13: 49–74

Mercier, J.C. & N.L. Carter 1975 Pyroxene geotherms – J. Geoph. Res. 80: 3349–3362

Nisbett, E.G. 1984 The continental and oceanic crust and lithosphere in the Archaean: isostatic, thermal, and tectonic models – Can. J. Earth Sci. 21:xxx–xxx

Nisbett, E.G. & D. Walker 1982 Komatiites and the structure of the Archaean mantle – Earth Planet. Sc. Lett. 60: 105–113

O'Hara, M.J., M.J. Saunders & E.L.P. Mercy 1975 Garnet-peridotite, primary ultrabasic magma and eclogite: interpretation of upper mantle processes in kimberlite, in: L.M. Ahren et al. (eds) Phys. Chem. Earth 9: 571–604

Oxburgh E.R. & E.M. Parmentier 1977 Compositional and density stratification in oceanic lithosphere – causes and consequences – Geol. Soc. Lond. 133: 343–355

Perkins, D. & R.C. Newton 1981 Charnockite geobarometers based on coexisting garnet- pyroxene- plagioclase- quartz – Nature, 292: 144–146

Sleep, N.H. & B.F. Windley 1982 Archaean plate tectonics: Constraints and inferences – J. Geol. 90: 363–379

Tarney,J. & B.F. Windley 1977 Chemistry, thermal gradients and evolution of the lower continental crust – Jl. geol. Soc. Lond., 134: 153–172

Vlaar, N.J. 1982 Lithospheric doubling as a cause of intracontinental tectonics – Proc. Kon. Ned. Akad. Wet., B 85: 469–483

Vlaar, N.J. 1985 Precambrian geodynamical constraints – A.C. Tobi & J.L.R. Touret (eds), the deep proterozoic crust in the North Atlantic provinces, 3–20. D. Reidel Publishing Company

Vlaar, N.J. & M.J.R. Wortel 1976 Lithospheric aging, instability and subduction – Tectonophysics 32: 331–351

Weaver, B.L. & J. Tarney 1979 Thermal aspects of komatiite generation and greenstone belt models – Nature 279: 689–692

Weertman, J. 1970 The creep strength of the Earth's mantle – Rev. Geophys. Space Phys. 8: 145–168

Wells, P.R.A. 1980 Thermal models for the magmatic accretion and subsequent metamorphism of continental crust – Earth Planet. Sc. Lett. 46: 253–265

Wortel, M.J.R. & S.A.P.L. Cloetingh 1985 Accretion and lateral variations in tectonic structure along the Peru-Chile trench – Tectonophysics 112: 443–462