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Abstract: In the comment the model offered by J. A. Percival, and R. N. Pysklywec ("Are Archean lithospheric keels inverted?" Earth and Planetary Science Letters, 2007. Vol. 254, pp. 393-403) is discussed. The discussion is focused on the analysis of the values of parameters and constants accepted for the model, as well as the comparison of P-T-conditions within the crust and upper mantle proposed in the model with P-T-conditions of numerous geologic, petrological and geothermal features known in the Late Archean. It is shown that the P-T-conditions of the model cannot support the formation of a huge eclogite layer at the bottom of the crust in the Late Archean. Based on known facts of the maximum of komatiite magmatism, worldwide magmatic activity, and the second worldwide pulse of granulite metamorphism, it is shown that the thermal boundary conditions accepted for the model are too low for the Late Archean. It is also shown that incorrect values were used for some parameters and constants in the model.

Cover Letter

Dear Mr. Carlson,

I would like to thank you for your remarks which helped me significantly improve my "Comment on “Are Archean lithospheric keels inverted?” by J. A. Percival, and R. N. Pysklywec, Earth and Planetary Science Letters, 2007.Vol. 254, pp. 393-403".

I would like to ask you publish my corrected version of my manuscript in your journal.

Best regards,

Arkady Pilchin

Comment on "Are Archean lithospheric keels inverted?" by J. A. Percival, and R. N. Pysklywec, Earth and Planetary Science Letters, 2007.Vol. 254, pp. 393-403

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ABSTRACT

In the comment the model offered by J. A. Percival, and R. N. Pysklywec ("Are Archean lithospheric keels inverted?" Earth and Planetary Science Letters, 2007. Vol. 254, pp. 393-403) is discussed. The discussion is focused on the analysis of the values of parameters and constants accepted for the model, as well as the comparison of P-T-conditions within the crust and upper mantle proposed in the model with P-T-conditions of numerous geologic, petrological and geothermal features known in the Late Archean. It is shown that the P-T-conditions of the model cannot support the formation of a huge eclogite layer at the bottom of the crust in the Late Archean. Based on known facts of the maximum of komatiite magmatism, worldwide magmatic activity, and the second worldwide pulse of granulite metamorphism, it is shown that the thermal boundary conditions accepted for the model are too low for the Late Archean. It is also shown that incorrect values were used for some parameters and constants in the model.

Key words: Archean, heat capacity, heat conductivity, pressure, temperature

1. Physical and petrological analysis of parameters, petrologic composition and boundary conditions of the model

Percival and Pysklywec (2007) report results of their numerical modelling in an attempt to explain the origin of the post-tectonic granite "blooms" and widespread crustal melting in many Archean cratons. They offer an inversion model, which they call a "Thermo-mechanical model of lithospheric inversion." The main part of the model is related to heat transfer; however, there are no explanations regarding the kinds of methods and equations used for calculating the amounts and rate of heat transfer. Oddly, there is also no equation, reference, or even mention regarding the final representation of the solution to the problem. Values of physical parameters accepted for the model are not supported by references, and in most cases have wrong or even unrealistic values. The petrologic composition of the lithosphere accepted for the model is in conflict with the main kinds of Archean rocks and their composition. Boundary temperature conditions are in obvious conflict with known petrological and geothermal facts for the Archean. These discrepancies of the model are analyzed in this comment, and the corresponding values and boundary temperatures are compared with their real values.

1.1.Analysis of physical parameters and processes accepted for the model

It is unclear from the (Percival and Pysklywec 2007) what method (methods) was used for the calculation of heat transfer for their inversion model, but it is obvious that for solid lithosphere heat transfer would be controlled by heat conduction prior to and after the inversion; while it would be controlled by some other dynamic process during the inversion, which could play a more important role than heat conduction in the parts of the lithosphere involved in the inversion. Unfortunately, Percival and Pysklywec (2007) did not provide any information or references regarding these problems, nor the methods of their solution or their resulting conclusions. At the same time, the use of the main thermal parameters (heat conductivity λ, specific heat capacity c_p, heat production A_0), as well as the boundary thermal conditions, shows that some kind of geothermal method (or methods) including the heat conduction equation was used for the model.

It is also unclear from (Percival and Pysklywec 2007) what kind of geothermal model, whether layered or single layer, one- two- or three-dimensional, was used for the model of inversion.

Nevertheless, let us analyse the values accepted for the model, since the values of these thermal parameters govern the rates and amounts of heat transfer, as well as the distribution of temperature and heat flow within the lithosphere.

The regular three-dimensional heat conduction equation (Carslaw and Jaeger, 1959; Kappelmeyer and Haenel, 1974; Eppelbaum and Pilchin, 2006) is:

$$\rho c \frac{\partial T}{\partial t} = \lambda \left(\frac{\partial^2 T}{\partial^2 x} + \frac{\partial^2 T}{\partial^2 y} + \frac{\partial^2 T}{\partial^2 z} \right) + A_0 \qquad (1)$$

Here T –temperature, t- time, x and y – horizontal dimensions, z – vertical dimension, λ – thermal conductivity, A_o – heat production, ρ – density, and c_p – specific heat capacity. Heat conductivity λ regulates the value of heat flow Q, which is in accordance with equation:

$$Q = \lambda G , \qquad (2)$$

where G is the geothermal gradient.

For some reason unexplained, Percival and Pysklywec (2007) accepted constant values for heat conductivity (λ = 3 W/m/K), specific heat capacity (c_p = 1.25×10^3 J/kg/K), and absolute absence of heat production (A_o = 0). A comparison of values accepted by Percival and Pysklywec (2007) for heat conductivity and specific heat capacity with their published values shows that the values accepted for the model are too high. Published values of a thermal conductivity of 2.5-2.6 W/m/K for granite, 5.0 W/m/K for quartzite, 4.57 W/m/K for olivine, and 2.3 for diabase W/m/K (Côté and Konrad, 2005) are in conflict with values accepted by

Percival and Pysklywec (2007). Some other publications point to even lower values of thermal conductivity, which for granite and basalt are as low as 2.2 w/m/K and 2 w/m/K, respectively (Lide 2005). Furthermore, the value of thermal conductivity greatly decreases with the increase of temperature (Clark Jr., 1966; Vosteen and Schellschmidt, 2003). At a temperature of 200°C, the values of thermal conductivity for granite, tonalite, and syenite could drop to as low as 2.14, 2.31, and 2.09 1.4 $Wm^{-1}K^{-1}$, respectively (Clark Jr., 1966); and at a temperature of 500 °C, its value is about 1.5 $Wm^{-1}K^{-1}$ for the main kinds of magmatic rocks and about 1.4 $Wm^{-1}K^{-1}$ for metamorphic rocks (Vosteen and Schellschmidt, 2003), and these values would continue to drop with increase of temperature. This means that according to equation (2) the values of heat flow for the parameters of the model would be approximately 50-100% or more greater than their real values.

Similarly, the value of specific heat c_p, which is 752 J/(kg*K) for granodiorite, 775 J/(kg*K) for diorite, 720 J/(kg*K) for granite, and 827±855 J/(kg*K) for peridotite (Schärli and Rybach, 2001), is much lower than that which was accepted for the model. Interestingly, all these referenced values of specific heat were measured for rocks from eclogite-bearing regions. This means that according to the model, during the inversion hot olivine replacing sinking eclogite would deliver more heat (about 46% or more greater) than its real value to the lower crust.

In the model by Percival and Pysklywec (2007), the main transfer of heat is taking place through the transfer of mass during inversion within the lithosphere, and the density change accepted for this numerical experiment is exclusively dependent on the thermal effect taking place on the density; unfortunately, the formula was also published with an obvious misprint.

This raises two problems related to the inversion model: 1) which part of the olivine-composed lithosphere mantle would replace a sinking eclogite plate?; and 2) how is it possible for any rock/volume to expand under high-pressure conditions in the lower crust and upper mantle? The first question raises serious problem, because the temperature at the top of the olivine layer (depth 60 km) is 730°C and at the bottom of the layer (depth 200 km) it is 1350°C. This means that depending on which part of the olivine layer would replace a sinking eclogite layer, a different amount of heat energy would be delivered to the lower part of the crust or its replacement. It is easier to accept from a mechanical point of view that an eclogite layer would be replaced by olivine present in the uppermost part of the olivine layer right below the eclogite plate. However, in such a case the amount of heat energy transferred to the original depths of the eclogite layer would be very small, taking into account close values of temperature of both layers at about 60 km depth, and the limited size of the eclogite layer. From Figure 2, 6 and 7 of (Percival and Pysklywec 2007), one could assume that the highest temperature of olivine replacing sinking eclogite is 1260°C and lower. This means that the difference in temperature between that of the eclogite plate and that of the replacing olivine would be about 530°C. Taking that and using the value of specific heat capacity accepted by Percival and Pysklywec (2007), brings the excess of heat energy to 662.5 J/g of olivine. Averaging this amount of heat for a crust layer between 0-60 km and a width about 3 times that of the eclogite plate, one could receive 73.61 J/g for this crust layer. A simple comparison of this amount of heat with results presented in Table 1 shows that even the amount of heat released by radioactive sources in ultramafic rocks during 20 M.y. is significantly higher, and the amount of heat released by radioactive sources in the crust of granitic composition during 20 M.y. would be about 7-12 times greater than the amount of heat delivered by the inversion proposed in the model by Percival and Pysklywec (2007).

Table 1

It is also clear from Table 1 that this difference between the heat energy held by the inversion of olivine and that produced by radioactive elements during 40 M.y. would be more than doubled; that would make energy transfered by the inversion in the model account for only about 4-7% of energy released by heat production within the crust during 40 M.y. This means that ignoring heat produced by radioactive sources, as it did Percival and Pysklywec (2007), is a huge mistake that leaves one of the main sources of heat energy out of model. On the top of that, results of previous researches (Kappelmeyer and Haenel, 1974) show that the portion of terrestrial heat production at depths between 0-100 km is about 50%- and that of between 100-200 km is about 25%- of all heat energy released within these layers. Ignoring such huge amounts of heat energy is irresponsible and would obviously lead to mistakes in the calculations of the temperature regime by up to 50-100%. For example, the total crustal contribution to the surface heat flow density ($37mWm^{-2}$) from radiogenic heat sources calculated for Fennoscandian shield as $26mWm^{-2}$ (Kukkonen and Lahtinen, 2001); that is 70.3% of total heat flow. The kind of solution one could get by neglecting over 70% of energy released cannot be acceptable for any model. It should also be taken into account that the estimations of heat released by heat production in Table 1 were made for its present values, which should be almost doubled for the Archean (Rybach and Wormald, 1999; Mareschal and Jaupart, 2006). Some published values of heat production (in J/(g*M.y.)) show that it reaches 28.34 for the central Fennoscandian shield of Finland (Kukkonen and Lahtinen, 2001), 48.42-64.96 for Late Archaean Bundelkhand granite of India (Menon et al. 2003), 43.7 for the Archean basement of Penokean Orogen in N. Michigan (Attoh 2000), 94.48-177.15 for granite intrusions of Lac de Gras in Slave Province (Mareschal et al. 2004), 118.1 for Gaborone Granite Complex of SE Botswana (Rybach and Wormald 1999), etc.. If so significant an amount of heat released within the crust of Archean cratons is not sufficient from then until now for melting rocks, it is impossible for such a small amount of heat transferred by inversion to be capable of generating numerous magma chambers and the formation of numerous batholiths within a cold crust as accepted by Percival and Pysklywec (2007). Moreover, since in such an inversion some amount of hot olivine would be above layers of cold olivine, it would definitely create conditions for part of the heat to be transferred downwards. This means that even the relatively small amount of heat transferred by inversion would not be totally used for melting crustal rocks.

It is unacceptable to ignore the laws of compressibility in modeling density values of rocks at great depths. In reality, how is it possible for any rock volume to expand under a pressure of about 1.68 GPa at the depth of 60 km for the model by Percival and Pysklywec (2007)? The effect of compressibility on the density, volume, and seismic velocity of rocks and minerals is well known (Birch, 1966; Anderson, 1989; Pilchin and Eppelbaum, 1997, 2002 and 2005; Cattin et al., 2001; Pilchin, 2005). The thermodynamic equation for any matter in the most general form is presented in Pilchin and Eppelbaum (2002) and as equation (4) in Pilchin (2005):

$$P = P_o + \frac{\alpha}{\beta}\left(T - T_o\right) - \frac{1}{\beta}\left(\frac{\Delta V}{V_o}\right). \qquad (3)$$

Here the variables are P- real pressure, P_0- normal pressure (usually lithostatic pressure for solid rocks), α – coefficient of thermal expansion, β – compressibility, T – real temperature, T_0 – normal temperature defined from condition of equilibrium, ΔV – change of volume, V_0 – initial volume prior to its change. The condition of equilibrium when the effect of thermal expansion would be totally compensated by the effect of compressibility is defined as $\Delta V=0$ (correspondingly it also means no change of density), and it would be

$$p = p_o + \frac{\alpha}{\beta}\left(T - T_o\right). \qquad (4)$$

It is obvious that at low temperatures the effect of pressure (compressibility) could be greater than the effect of thermal expansion, and this would lead to an increase of density. The presence of an external tectonic force or pressure (for example during continent-continent collision, obduction, or subduction) would compress the volume, so its change would be negative $\Delta V<0$. In such a case, it would lead to an increase of density. All this and much more is described in detail in (Pilchin and Eppelbaum 2002, 2005). In contrast, using only the effect of thermal expansion, Percival and Pysklywec (2007) left their model with the only possible outcome of a decrease of density as a result of increase of volume. Moreover, when temperature increases at great depths, thermal expansion would not actually take place until it would generate a pressure greater than the lithostatic pressure (see Pilchin and Eppelbaum, 2002, 2005). Only then would real expansion be possible, since for matter to expand the effect of thermal expansion must overcome the effect of pressure (or compressibility), as is clear from the condition of equilibrium. A comparison of the density change calculated for data of the model by Percival and Pysklywec (2007) and the density calculated using published data with the composite effect of temperature and pressure is presented in Table 2.

It is obvious from Table 2 that the density of olivine actually increases under P-T-conditions when counting the effect of compressibility.

Table 2

The most graphic example of the effect of compressibility is expressed in the statement "At 10 kilobars (equivalent to a depth of about 35 kilometres) the melt would be 4-5% denser than at the surface" (Hall, 1995; page 69). Of course, this effect would be smaller for solids, but it would not be zero.

It should be also stated that other mechanical properties were also ignored in the modelling by Percival and Pysklywec (2007). If an eclogite plate or layer is within the solid lithosphere, it is impossible to disregard friction between solid rocks, which is characterised by a very high friction coefficient of ≥ 0.6, known as Byerlee's Law (see for example Sammis and Biegel, 2004). Such strong friction would obviously overpower a small difference in the density of an eclogite layer and the underlying lithosphere and asthenosphere. At temperatures below 730°C for the crust and eclogite layer, it is difficult to expect that friction would not play a role in the sinking of solid eclogite through solid olivine.

The rapid formation of an eclogite layer with density 3500 kg/m^3 from rocks of the lower crust with density 2800 kg/m^3, requires a reduction of volume of about 25%. This means that such a transformation would cause the formation of a depression above the area where this transition took place. For the formation of an eclogite layer with thickness of about 20 km, such a depression would reach a depth of about 5 km. This fact was also not addressed in the model.

The statement of the authors saying that "… the eclogitic mass became gravitationally unstable once oriented vertically …" (Percival and Pysklywec, 2007; page 400) completely contradicts the proposed model, not to mention the laws of Mechanics, because it is not possible that a 20 km-thick and 150 km or 400 km wide plate would be able to rotate from a horizontal to vertical position without even disturbing the layers of the middle and upper crust (see Fig. 2 and Fig. 5 in Percival and Pysklywec, 2007).

1.2.Analysis of petrologic composition of lithosphere accepted for the model

The model offered by Percival and Pysklywec (2007) also has problems with petrologic composition. Firstly, there is no known massive eclogites of the Archean age (see for example Pilchin and Eppelbaum, 2006), and certainly there is no known facts of the existence of such a huge eclogite plate/layer of this age as proposed. Schulze (1989) concluded that the amount of eclogite in the upper 200 km of the subcontinental upper mantle is perhaps <1% by volume. Carlson et al. (2005) show that "In general, high-velocity bodies consistent with large masses of eclogite are not observed in the continental mantle" and that "… at present, eclogite appears to be a relatively minor component of the continental lithospheric mantle." (paragraph 6 in Carlson et al., 2005).

Some other authors also point at the rarity of blueschists and eclogites older than 2 Ga (Möller et al., 1995). However, some researches show that Archean age signatures were found in eclogites from kimberlites (Pearson et al. 2003). Research show (Baldwin et al. 2004) that eclogites from the East Athabasca mylonite triangle in northern Saskatchewan with peak eclogite facies metamorphism 1,904.0±0.3 Ma, contain zircon from one eclogite sample of 2.54 Ga, which was considered as a minimum age for the emplacement or earliest metamorphism of the gabbroic protolith. This means that the presence of Archean signatures in eclogites xenoliths from kimberlites does not necessarily mean that these signatures are pointing at the age of eclogites themselves and not their protoliths.

In another case, inclusions of rare garnet and clinopyroxene trapped within diamonds occurring in eclogite xenoliths from the Siberian craton were more depleted in incompatible trace elements and have lower Mg# than the eclogite host minerals (Ireland et al., 1994). This could mean that the eclogite inclusions within diamonds have possibly nothing to do with eclogites in the kimberlites themselves. It would then be incorrect to use the age of eclogite inclusions in diamonds for estimations of the age of eclogite xenoliths in kimberlites.

Is it possible that in the lower crust P-T-conditions were favorable for the formation of an eclogite layer?

Taking into account that in the model presented the Archean crust was homogeneous prior to the formation of an eclogite layer, the P-T-conditions right before the formation of the eclogite layer were: T=730°C at the base of the 60 km thick crust, and the pressure P was 1.68 GPa at the bottom of the crust and 1.12 GPa at the depth of 40 km near the top of the forming eclogite layer, and the ratio P/T = 2.3 MPa/°C. The value of the average geothermal gradient within such a 60 km thick crust with surface temperature of 20°C would be 11.83 °C/km. It was shown earlier that for the formation of eclogites the minimum geothermal gradient should be about 6.3 ºC/km (Christensen, 1979). A comparison of these values with the average values of these parameters for known eclogites (Table 3) shows that the pressure in the model is too low, and both the temperature and geothermal gradient are high for the formation of eclogite.

Table 3

The value of the ratio P/T for the model is roughly that of the lowest value of the ratio within the eclogite range of stability. This means that the P-T-conditions of the model are not favourable for the formation of eclogite, and even if some amount of eclogite were formed under such conditions, it would not be a significant amount under any circumstances, and would certainly not be an eclogite layer. To analyse this matter in greater detail, available data on eclogites was separated into three data sets: eclogites formed under low temperatures with T≤570°C, eclogites formed under moderate temperatures with 570°C<T≤720°C, and eclogites formed under high temperatures with T>720°C. The value 570°C was accepted as the upper temperature limit for low temperature eclogites formation, because this is the temperature of the transformation from ferrous to ferric iron oxides (see for example Pilchin and Eppelbaum, 1997, 2006, 2007 and references there). The reason for using this value as the upper limit of low temperature eclogites is related to fact that aegirine/acmite, one of the main components of omphacite, which in turn is one of the main components of eclogite, contains both ferric and ferrous iron oxides; however, ferric iron oxide is unstable at temperatures above 570°C

(Pilchin and Eppelbaum, 1997, 2006, 2007 and references there). This means that there are problems for the formation and stability of aegirine, and correspondingly eclogite, at a greater temperature than this. Though it should also be taken into account that the temperature of the range of stability of ferric iron oxide could increase above 570°C within layers rich in oxygen, it cannot be applied to the Archean known for anoxic conditions in both the lithosphere and atmosphere (Pilchin and Eppelbaum, 2006 and references there). In fact, the temperature of the transformation of ferrous to ferric iron oxides reduces to ~450°C with even a small increase of pressure (Pilchin and Eppelbaum, 1997, 2006, 2007 and references there).

Analysis of data on eclogites using different groupings of data in Table 3 show that the eclogites in the model should be regarded as eclogites of moderate temperature formation. However, the pressure, the ratio P/T, and depth of eclogites formation in the model are too low, and the geothermal gradient is too high compared with the average values of these parameters for eclogites formed under moderate temperatures (Table 3). Furthermore, the differences between these parameters are even greater than those between the eclogites of the model and the average parameters for all eclogites. It is also clear from Table 3 that the parameters of the model do not fit the corresponding parameters for eclogites of orogenic belts. Moreover, it was shown earlier (Pilchin and Eppelbaum, 2005) that lithostatic pressure alone is not enough for the formation of eclogites, and that P-T-conditions involving overpressure are required. All this makes the formation of a thick eclogite plate in the Archean in the lower crust under the P-T-conditions presented in the model impossible. Another problem in the model is the value of density 3500 kg/m^3 accepted; in most publications values of the density of eclogite presented are smaller than 3500 kg/m^3. For example, Hall (1995) presents an average density of 3390 kg/m^3 with a density range of 3340-3450 kg/m^3, and Kappelmeyer and Haenel (1974) presented a value of the density of eclogite of about 3200 kg/m^3. Data presented in Clark Jr. (1966) shows that only for eclogites of California and Norway Caledonides is the density greater than 3376 kg/m^3. This means that only Phanerazoic eclogites have high densities. This is related to the high content of dense almandine, which is rich in iron. Taking into account that dense eclogites contain high amounts of both almandine and aegerine, while Archean rocks are known for the lowest content of iron (Griffin et al. 2003; Pilchin and Eppelbaum, 2006, 2007) and during the Archean significant amounts of iron were removed from rocks and deposited as banded iron formations (BIFs), it is clear that there was not enough iron in the Archean rocks for the formation of iron-rich and dense eclogites. This means that even if there were conditions favourable for the formation of some eclogites, those would be Mg-rich and they would not be dense. Therefore, the density of eclogite 3500 kg/m^3 accepted by Percival and Pysklywec (2007) for their model is too great.

There are also no known massive peridotites or peridotites of any significant amount of the Archean age, and olivine is not even among the essential minerals of Archean rocks (see for example Pilchin, 2005).

Some researches demonstrate that at some point of its early evolution Earth was covered by a magma-ocean (Pollack, 1997; Solomatov, 2000; Walter and Trønnes, 2004), which was a few hundred kilometers deep (Walter and Trønnes, 2004). Some scientists believe that the magma-ocean could have remained for 100–200 My (Pollack, 1997), and even longer (Solomatov, 2000). Taking into account that the density of basalt melts is within the range of about 2600-2680 kg/m^3 (see Table 4) and the density of komatiite melt is about 2745 kg/m^3 (see Table 4), it is clear that solid forsterite (density of about 3100-3300 kg/m^3 and melting point of about 1890°C Ganschow and Klimm (2005)), as well as solid and melted (melting temperature of about 1203°C) fayalite (see Table 4), would sink very quickly in the magma-ocean.

Table 4

Moreover, the fact that both forsterite and fayalite are denser than komatiites means that most olivine would sink to depths of komatiite magma formation and possibly deeper. Similarly, most ultramafic and some mafic rocks, and possibly their melts, especially those rich in iron would also sink in the magma-ocean.

Such facts as the absence of olivine/peridotite xenoliths in the late Archean komatiites; that the peridotite component was less than 1% of the Archean greenstone belts (Nalivkina, 1978);

that in the Archean magmatic rocks of the Superior Province of Canada the content of olivine is less than 0.4-0.9% (Goodwin, 1977); and that among Archean rocks olivine is absent or extremely rare, and is not even one of the main mafic minerals (Pilchin, 2005) point to a lack of olivine/peridotite in the uppermost mantle in this period. This is in agreement with data presented in Figure 6 of (Pearson et al., 2003) showing that for most Archean regions, garnet peridotite xenoliths were delivered from depths usually greater than about 70 km and that the quantity of xenoliths significantly increases with depth, with most xenoliths originating deeper than 100 km. Since the delivery of xenoliths to the surface is much easier from shallower depths, this also points to a lack of peridotites in the uppermost mantle. Researches on P-T-conditions of equilibration of peridotite xenoliths show that peridotites were equilibrated under pressures 2-6 GPa (Pearson et al., 2003), 2.5–5.0 GPa and 3-7 GPa (Lee et al., 2003); for densities accepted for the model by Percival and Pysklywec (2007), that corresponds to minimum depths of about 70 km, 85 km and 100 km. This means that olivine could not be present in significant amounts at shallow depths in the mantle under forming Archean cratons, and especially at depths less than 100 km. In fact, it is possible that there was an extreme lack of olivine content at depths between 60 and 70 km.

1.3.Analysis of boundary temperature conditions accepted for the model

The model by Percival and Pysklywec (2007) has two main boundary temperatures, 1350°C at the depth 200 km and 20°C at the surface, and one intermediate boundary at the depth 60 km with a temperature of 730°C. Also, conditions with no internal heat sources within the lithosphere were accepted for the model. It was shown earlier (Pilchin and Eppelbaum, 2005; Eppelbaum and Pilchin, 2006) that for homogeneous layers heat production demands a decline of the geothermal gradient with depth. Of course, without heat production there should be a constant value for the geothermal gradient within a homogeneous layer. Since the model uses the same incorrect values of thermal parameters for the entire lithosphere with no heat production within the lithosphere, after a long enough time of maintaining boundary conditions the values of the geothermal gradient within the lithosphere at the surface and the depth of 200 km should be constant or at least close to constant. In reality, the geothermal gradient for the model is 11.83 °C/km for depths 0-60 km, 4.43 °C/km for depths of 60-200 km, and 6.65 °C/km for 0-200 km. This means that the temperature of 730 °C selected for the depth of 60 km is not correct for the accepted model, and using the average geothermal gradient for the lithosphere proposed the temperature should actually be about 399°C. However, with such low temperatures within the crust there is no chance to attain any melting conditions for the inversion model.

Since one of the main purposes of the modeling is to explain events related to the evolution of the Superior province of Canada during 2.72-2.58 Ga, the Yilgarn craton of Australia during 2.68-2.61 Ga, and the Slave craton of Canada during 2.64-2.50 Ga (see Fig. 1 in Percival and Pysklywec, 2007), let us analyse available petrologic and geothermal data that could be used for the estimation of temperature within the crust and the mantle parts of the lithosphere at about that period of the Late Archean.

Among the most important features and processes of the evolution of Earth in the Late Archean there are: 1) the maximum temperatures in the mantle corresponding to the widespread formation of komatiites in the Middle and Late Archean, with the peak of komatiites formation around 2.8-2.7 Ga (Arndt *et al*., 1979; Arndt and Nisbet, 1982; Barley *et al*. 2000; Svetov and Smolkin, 2003); 2) the worldwide magmatic event during the Late Archean (2.75-2.65 Ga) (Condie, 2001; Barley *et al*., 2005); and 3) the second worldwide pulse of granulite metamorphism beginning from about 2.7 Ga (Percival, 1994). These events were followed by the superplume events of 2.51-2.45 Ga and around 2.25 Ga (Heaman, 1997; Bekker et al., 2003, Barley et al., 2005; Pilchin and Eppelbaum, 2006 and references there); and the formation of Early Proterozoic dyke swarms started near the Archean – Early Proterozoic border (Aspler and Chiarenzelli, 1998; Buchan *et al*., 1998, Mertanen *et al*., 1999, Buchan and Ernst, 2004).

Takahashi (1990) has shown that the latest Archean komatiites may have been produced at ≥4 GPa and ≥1650°C. Such P-T-conditions are present at depths of about 120-130 km, being

70-80 km higher than the depth of 200 km for which Percival and Pysklywec (2007) accepted a temperature value of 1350°C. Some authors show that komatiites could possibly have been formed at even greater temperatures of about 1700 – 1850°C (Svetov and Smolkin, 2003) at depths of about 160–180 km. Since komatiites have the highest melting temperatures among all known magmatic rocks, their formation could be viewed as an indicator of the maximum temperature in the upper mantle. Taking this data into account, along with results of geothermal research (Pollack, 1997) showing that throughout the entire Archean the temperature in the mantle dropped by no more than 200°C, there is little evidence of the temperature 1350°C at a depth of 200 km accepted by Percival and Pysklywec (2007) for modelling of thermal conditions in the mantle during the Late Archean. The fact of a worldwide magmatic event taking place during the Late Archean (2.75-2.65 Ga) with its peak at about 2.7 Ga (Condie, 2001, 2005) also points to high mantle temperatures. Approximately 35% of Late Archean greenstones have plume affinities (Condie, 2001); and also, interestingly, the oldest well-documented continental flood basalts are the Ventersdorp (2.7 Ga; Kaapvaal craton, South Africa) and Fortescue (2.77 Ga; Pilbara craton, Australia) (Condie, 2001). The fact that the second worldwide pulse of granulite metamorphism took place in different cratons from before 2.7 Ga to about 2.5 Ga (a small fraction of published data on granulites formed during the second worldwide pulse of granulite metamorphism is presented in Table 5) also contradicts the geothermal conditions accepted for the model by Percival and Pysklywec (2007).

Table 5

It is evident from Table 5 that the second worldwide pulse of granulite metamorphism took place in all cratons (Superior province, Slave craton, Yilgarn craton, and Caapvaal craton) mentioned in the paper by Percival and Pysklywec (2007).

In reality, it is clear from Table 3, that the average depths and temperatures of the formation of granulites of about 28 km and 800°C are in complete disagreement with the value of temperature 730°C at depth 60 km accepted for the model by Percival and Pysklywec (2007). Interestingly, calculations using geothermal methods for the crust part of the lithosphere in the Slave province during the Archean show (Mareschal and Jaupart, 2006) that temperature at the depth of 50 km was 880°C for stratified distribution of heat sources, and at the depth of 40 km it would be about 875°C if the distribution of radioactive sources was uniform for the crust. Calculations of temperature at the depth 150 km of the same region (Mareschal and Jaupart, 2006) using different geothermal models gave 1250-1400 °C for a crust with thickness of 40 km and 1350 °C for a crust with thickness of 50 km. These values also completely contradict to the model by Percival and Pysklywec (2007).

Analysis of the P-T-conditions for granulites formed during different periods of the Archean and Early Proterozoic are also presented in Table 3. It is clear from Table 3 that the P-T-conditions of granulite metamorphism were on average the highest during the second, Late Archean (at ~ 2.7 Ga), pulse of granulite metamorphism. This fact also supports a high, and possibly maximal, thermal regime during this period of the crust and mantle evolution. This data, along with the maximums of komatiite magmatism and worldwide magmatism, supports the fact that at this point in the Archean there was a possible temperature maximum for the entire Archean period. It is known that the second worldwide pulse of granulite metamorphism is related to the formation of mostly mafic granulites; such as, for example, in the Kapuskasing Structural zone, Superior province (Percival and West 1994). How is it then possible to form such huge amounts of mafic granulites at depths of about 30 km within felsic crust with a thickness of about 60 km, as is accepted by the model of Percival and Pysklywec (2007)? It is obvious that the model would not work with mafic rocks, which have much higher melting points than those of felsic rocks.

The facts of the superplume event and the formation of giant dyke swarms during the time period on the border of the Archean and Early Proterozoic and the earliest Early Proterozoic, as well as the presence of komatiites and picrites in the Early Proterozoic formed under temperatures of up to 1750°C (Svetov and Smolkin, 2003), and the third, Early Proterozoic, worldwide pulse of granulite metamorphism at about 2.0-1.8 Ga show that the temperature in the mantle was very high at least until the end of the Early Proterozoic. It is for certain that both geothermal and petrologic data support a high temperature regime in the mantle during the Archean. That is the reason why Pollack (1997) called the presence of signs of Archean signatures in some diamonds enigmatic.

Another fact contradicting such a model of inversion is the known presence of extremely dense ultra-high metamorphic rock layers existing at and near the surface for hundreds of millions of years without such an inversion as described in the model having ever taken place. For example, a giant eclogite-peridotite layer is present on the surface in the Norwegian Caledonides for as long as 400 Ma (see Pilchin, 2005 and references there), but no inversion ever took place there, even though the difference between the average density of this layer and the underlying rocks is much greater than the value accepted for the model.

2. Conclusions

From all of the above, it is clear that: 1) the model (models) offered by Percival and Pysklywec (2007) uses incorrect values of the thermal characteristics of rocks that led to about a 50-100% increase of heat flow and about a 50% increase of the amount of heat energy transferred during inversion; 2) the ignoring of heat production leads to significant mistakes in the determination of thermal parameters of the crust; 3) incorrect P-T-conditions for the Archean accepted for the model led to significant discrepancies between temperatures accepted as boundary conditions for the model and the real temperature conditions supported by numerous petrologic facts and geothermal calculations; 4) the petrologic composition of the model is in conflict with known petrologic and geologic facts, as well as with P-T-conditions of the stability of eclogite.

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Table
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Table 1. Thermal characteristics calculated using heat production data by (Birch, 1942; Clark Jr., 1966)

Rock	**Heat production, J/(g*M.y.)**	**Heat released during 20 M.y., J/g**	**Heat released during 40 M.y., J/g**	**Maximum possible temperature effect* for 20M.y./40M.y., °C**
Early Precambrian granite Granite Gabbro Ultramafic rocks	 >41.868 25.121 8.374 4.187	 >837.36 502.42 167.48 83.74	 >1674.72 1004.82 334.96 167.48	 >670/1340 402/804 134/268 67/134

* - specific heat capacity accepted by Percival and Pysklywec (2007) was used for calculations of maximum possible temperature effect

Table 2. Comparison of density change for the model by Percival and Pysklywec (2007) with values calculated using published data and counting both effect of temperature and effectt of compressibility (In all cases for estimation of thermal effect on density the method offered by Percival and Pysklywec (2007) was used**)**

Layer, rock	**For model by Percival and Pysklywec (2007)**			**For published data**			
	Density ρ under ambient conditions, kg/m^3	**α, 10^{-5} 1/K**	**Density ρ under P-T-conditions, kg/m^3**	**Density ρ under ambient conditions, kg/m^3**	**α, 10^{-5} 1/K**	**Density ρ under P-T-conditions, kg/m^3**	**Density ρ under P-T-conditions counting effect of compressibility, kg/m^3**
Eclogite	3.5	2	3.45	3.376 [1]	3.10 [2]	3.30	3.34 [4]
Olivine	3.3	2	3.29	3.280 [1]	3.03 [3]	3.27	3.31 [5]

1 The highest density of not Phanerozoic eclogite after Clark Jr. (1966)

2 Calculated using data for thermal expansion of garnet and pyroxene from Anderson (1989).

3 Calculated for Fo92.5 using data from Singh and Simmons (1976).

4 Calculated using compressibility as average for garnets and omphacite from Pilchin and Eppelbaum (2005)

5 Calculated using bulk modulus from Chen et al. (2005).

Table 3. Thermodynamic conditions of metamorphic processes of eclogites and granulites of different origin (modified after, Pilchin and Eppelbaum, 2005, 2006; Pilchin, 2005)

Rock (n)	**Average *T*, °C**	**Average *P*, GPa**	**Average *P/T*, MPa/°C**	**Average depth of lithostatic pressure, km**	**Average geothermal gradient (°C/km)**
Eclogites, average (n=505)	585	1900	3.25	~64	9.1
Eclogites formed at T, °C					
T≤570, (n=252)	487.8	1.5	3.05	50	9.4
570 <T≤720 (n=193)	640.6	2.14	3.35	70	8.9
T> 720 (n=60)	802.3	2.77	3.46	90	8.7
Eclogites from orogenic belts (n=410)	543	1.74	3.18	58	9.1
Granulites average (n=543)	800	820	1.02	~28	28.6
Granulites of:					
Early and Middle Archean	786	880	11.42	30.3	25.3
Late Archean	854	871	8.7	30.0	27.8
Early Palaeo-Proterozoic	767	667	8.63	23.0	32.6
End of Palaeo-Proterozoic	814	825	10.1	28.4	28.0

Table 4. Densities of some rocks in their molten and solid states

Rock, mineral	**Density of molten rock, kg/m^3**	**Density of solid rock, kg/m^3**	**Reference**
Rhyolite	2140	2280	Hall, 1995
Andesite	2410	2590	Hall, 1995
Tholeiite basalt	2600	1760	Hall, 1995
Alkali olivine basalt	2680	2830	Hall, 1995
Diabase	2640	2960	Clark Jr., 1966
Basalt, Gabbro	2600	2900	Clark Jr., 1966
Rhyolite, Granite	2300-2400	2600	Clark Jr., 1966
Komatiite	2745	-	Miller et al., 1991
Fayalite	3750	4375	Chen et al., 2002

Table 5. Ages of some granulite belts formed during the second worldwide pulse of granulite metamorphism

Region, tectonic unit	Age, Ga	Reference
Africa: Limpopo Belt, Caapvaal craton	~2.72 to ~2.59	Blenkinsop et al., 2004
Canada: Pikwitonei domain	2.744 to 2.59	Percival, 1994
Pikwitonei domain (peak conditions)	2.641-2.648	Percival, 1994
Kapuskasing Structural zone	2.585-2.650	Percival, 1994
Ashuanipi Complex	2.69	Percival, 1994
Pikwitonei granulite domain	2.744 to ~2.59	Mezger et al., 1990
Slave Province	2.64–2.58	Davis et al., 2003
Orma domain, Churcill Province	2.58	James et al., 2003
Wawa gneiss domain	2.66-2.637	Moser, 1994
Minto block	2.7	Percival and Skulski, 2000
Australia: Yilgarn Craton	2.640-2.649	Nemchin et al., 1994
Wheat belt	>2.6	Newton, 1987
North Western Gneis Terrain	2.65–2.60	Solomon and Groves, 1994.