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Thermal evolution of Indian cratonic lithosphere

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ABSTRACT
Thermal evolution of the Indian cratonic lithosphere can be deciphered by using the heat conduction equation with the thermal properties and other related information, such as heat flow values and pressures, temperatures and ages derived from rocks of various terrains. Some thermal causes reflect changes in the surface and basal boundary conditions and some others reflect input of heat inside the lithosphere as a result of magmatic or tectonic events. The later processes can be posed as initial value problem of the heat conduction equation. Surface heat flow and heat generation data are used to infer the present thermal structure of the Indian cratonic lithosphere. Metamorphic pressure and temperature data have been used to construct the paleogeotherms. Formation of the Cudappah basin is ascribed to thermal perturbation of the lithosphere. Formation of charnockites has been ascribed to thermal perturbation by flux of CO_2 from mantle depths, or uplift and erosion of thrust sheets emplaced on the lithosphere or a combination of both. The few problems of the Indian cratonic lithosphere, which have been addressed as initial/boundary problems or changes in the heat sources or addition of advection term in the heat conduction equation are described in this review.

INTRODUCTION

Earth has been losing heat from its interior to its environment since its origin. In the outer core and mantle, transport of heat towards earth's surface is by thermal convection. Near the surface, the heat is transported by conduction. This top layer in which heat is transported mainly by heat conduction is termed as the lithosphere. The layer underlying the lithosphere is rheologically weak and is called the asthenosphere. Lithosphere is subjected to cooling from top, which is enhanced by erosion and heating by heat flow from asthenosphere and by decay of radioactive elements concentrated in its upper regions. All geological processes recorded on its surface are linked to earth's thermal evolution. In principle it should be possible to extrapolate the present day observations backwards to find the thermal evolution of the lithosphere. However such a backward extrapolation is not a well-posed problem. A small change in the present condition will blow up the solution when extrapolating to the past. Prediction of future state from the present state using heat conduction equation however is a well-posed problem. Thus to get history, we need to know something of the past and then match with the observations of today to have confidence in the history. We thus need to study the geological products of the past. Exposed rocks do provide clues to this. Present day heat flow and heat generation measurements can constrain the present thermal state of the Indian cratonic lithosphere. Pressure, temperature and time provided by the studies of metamorphic rocks and xenoliths are very helpful in constraining the snapshot of paleothermal state of the lithosphere. Similarly uplift and subsidence recorded in sedimentary formations can also yield information about the thermal states at the time of their formations.

Globally this has been very active area of research since formulation of the heat conduction equation by Fourier (1822). Fourier used it for propagation of oscillation of surface temperatures into subsurface. Kelvin (1863) used a transient solution of one dimensional heat equation for estimating the age of the earth, fitting with an observed value of temperature gradient at the earth's surface. A large class of solutions of the heat conduction equation as summarized in Carslaw and Jaeger (1958) has found applications in understanding geophysical processes underlying geological phenomena (Jaeger, 1965, Turcotte and Schubert, 2002). In this paper we focus on those works that use geological and geophysical data with heat conduction models to address the issue of evolution of Indian cratonic lithosphere.

Inference From Present Heat flow And Heat Generation Data

For inferring geotherm we need the governing equation for the temperature distribution denoted by T(z), z- depth coordinate increasing with depth. In the simplest case of the vertical heat transport with homogeneous thermal conductivity, the governing equation is

$$K\frac{d^2T}{dz^2} = -A(z) \tag{1}$$

Here thermal conductivity and heat production are denoted by K and A(z) respectively. The surface temperature can be taken as constant, usually taken as zero for geothermal studies of lithosphere. The radiogenic heat distribution is generally taken as

$$A(z) = A_0 \exp(-z/d) \tag{2}$$

This form of distribution has been advanced as it satisfies, in presence of differential uplift and erosion, the linear surface heat flow and heat generation relationship, called Birch-Lachenbruch relationship (Roy et al, 1968, Lachenbruch, 1970 and Negi and Singh, 1979). Rao et al (1976) presented such a relationship for the Indian region. For the given values of the surface heat flow, q_s, and temperature, T_s, the solution of eqn (1) and (2) is given by

$$T = T_s + \frac{(q_s - A_0 d)z}{K} + A_0 d^2\left(1 - \exp\left(-z/d\right)\right) \quad (3)$$

Rao et al (1976) also obtained Moho temperatures and showed that Indian shield is hotter than other shields. More complex model using temperature dependent thermal conductivity has been constructed by Singh and Negi (1982) for the crust, Singh (1981) for the lithosphere, and Manglik and Singh (1991) for the continental crust and underlying mantle lithosphere together. In this case crustal temperatures are higher than the case with constant thermal conductivity. Now with more measurement, there is no well-defined linear relationship between surface heat flow and heat generation data. Recently a radiogenic heat model has been constructed using the heat generation values of rocks at depth by their exposures at the surface due to tectonic processes (Roy and Rao, 2000). This model puts more radiogenic heat in the crust with lesser values for Moho heat flows. Geotherms are constructed by dividing the crust/lithosphere into number of homogeneous layers having constant thermal conductivity and heat generation and using the following relationships between temperatures, T_t and heat flow, q_t on the top of a layer of thickness, Δz with the temperatures, T_b and heat low, q_b at the bottom of the layer as (Singh, 1968; Manglik, 2006; Furlong and Chapman, 2013)

$$T_b = T_t + \frac{q_t \Delta z}{K} - \frac{A \Delta z^2}{2K} \quad (4i)$$

$$q_b = q_t - A\Delta z \quad (4ii)$$

This new empirical heat production model makes Indian shield as cold as other shields. Steady state geotherms for different regions and with different sets of assumptions have been constructed (Rao et al, 1976, Singh, 1981, Singh and Negi, 1982, Gupta et al, 1991, Thiagarajan et al, 2001, Ramana et al, 1999 and 2003, Manglik and Singh, 2002, Roy and Rao, 2003, Manglik, 2006, Singh, 2007, Kumar et al, 2007a,b, Manglik and Singh, 2009).

Thermal structure has been used to find the thickness of the Indian lithosphere. Singh (1980) estimated the thickness of the Indian lithosphere and found, for the first time, it to be less than 100 km in the western part of the northern shield and more than 100 km in the eastern part of the northern shield. Later studies by Negi et al (1986) also found such estimation of about 100 km for the entire Indian region. However, those studies wherein the radiogenic heat contribution are higher (Roy and Rao, 2003; Kumar etal 2007a, b), lead to thicker Indian lithosphere. Receiver function analyses of the seismological data have been used to argue for thin Indian lithosphere (Kumar et al, 2007). Later studies showed that there are mid lithospheric discontinuities, which are taken as lithosphere - asthenosphere interface, making Indian lithosphere thin (Bodin et al, 2014).

Thermal structure has also been used to constrain the rheology of the lithosphere and used to estimate the thickness of the elastic lithosphere. For the northern part of the shield the elastic lithosphere was found as 30-35 km (Singh 1981) and for the southern part of the shield it was found as 50 km (Bhattacharji and Singh, 1984). Manglik and Singh (1991, 1992,1999) used more realistic rheological stratified constitutive relationships for the crust and lithosphere and obtained the thicknesses of brittle crust and also brittle mantle for normal, strike slip and thrust faulting. They also estimated the thickness of the rheological lithosphere and the basal shear stress at the top of the asthenosphere. Manglik (2015) has recently discussed the comparative estimates of Indian elastic lithosphere obtained from rheological laws and from admittance analysis of the gravity and topography data.

Inference From Metamorphic Pressure-Temperature-Time Data

Given mineral assemblages in cratonic rocks of various times, it is possible to estimate paleogeotherms and infer the cooling of the cratonic lithosphere. An estimate of cooling of the Indian cratonic lithosphere was made by Singh (1984) by using the pressure and temperature data [age of metamorphism: -3Ga; pressure: 6-7kbar; temperature: 750-850ºC] for Sargur schists of Karnataka (eastern block), a part of the Western Dharwar craton (Rollinson et al, 1981).

For constructing the paleogeotherm, we need to estimate the distribution of the radiogenic heat in the crust from the given present distribution and also the value of the surface temperature. For extrapolation of the present radiogenic heat we need to consider that erosion has taken place, radioactive heat has decayed and also the nature of possible overlying rock column changed with time. Taking the overburden pressure as 6.5 kbar, the total erosion is about 24 km. There has been a suggestion in the literature that the preexisting crust also contained a thick volcanic pile. West and Mareschal (1979) have assumed 5-10 km volcanic piles emplaced during short time interval of 5-10 Ma. Singh (1980) assumed a thickness of 20 km in case the eruptions took over long period of time. This layer has zero radiogenic heat. For the Kolar location in south India, Rao et al (1976) have proposed the following form of heat production in crust

$$A(z) = A_0 \exp(-z/7500) \qquad (5)$$

Correcting for the radioactive decays and 4 km of the erosion of the radiogenic heat producing crust, the form of the heat production arrived by Singh (1984) is

$$A(z) = 0 \quad \text{for} \quad z \quad 0 \le z \le l(= 20\ \text{Km})$$

$$A(z) = A_t \exp\left(-\frac{(z-20000)}{7500}\right) \quad \text{for} \quad z \ge l$$

$$(A_t = 7.3.10^{-6}\text{Wm}^{-3}) \qquad (6)$$

The solution of eqn (1) is to be obtained for radiogenic model given by eqn (6) with following boundary conditions:

$$T(z=0) = 0 \qquad (7i)$$

$$K\frac{dT}{dz} = q^* \quad \text{for} \quad z \to \infty \qquad (7ii)$$

The heat flow at great depth is denoted by q*. The solution in the zero radiogenic layer is

$$T_u = Mz \qquad (8)$$

The solution in the bottom half space is

$$T_b = \frac{q^* z}{K} - (A_t d^2/K)\exp\left(-\frac{z-l}{d}\right) + N \qquad (9)$$

The continuity of temperature and heat flux across the interface z=l, gives the values of the constants M and N as

$$M = (q^* + A_t d)/K \qquad (10)$$

$$N = \frac{A_t d}{K}(l + d) \qquad (11)$$

Fitting eqn (9) with the metamorphic data gives the value of q* as 45 mWm^{-2}. This value is to be compared with the present value of Moho heat flow, which is 33 mWm^{-2}. The Moho temperature for a crust having thickness of 40 km is 1118^{0}C compared to present Moho temperature of 500^{0}C. Thus, the Moho temperature has reduced to less than half at present. Singh (1984) inferred cooling of the Indian shield mantle by using a parameterized model of mantle convection. Davies (1980) has given the following relationship between average mantle temperature and mantle heat flow, for a particular form of viscosity model of the mantle, as

$$\frac{T_t}{T_0} = (q_t^*/q_0^*)^{1/11} \qquad (12)$$

Using values of Moho heat flow, we get

$$T_t = 1.03T_0 \qquad (13)$$

Schubert et al (1980) gave the value of present mean temperature of mantle as 2550 K. Using this value, Singh (1984) got the mantle cooling rate as 25^{0}C/Ga.

Now better measurements of heat flow and heat generation at many locations within Indian craton, since this study, have been made. These recent studies show that the present Moho heat values are 12-20 mWm^{-2}. This then implies more cooling of the Indian crust than estimated in the above study (Singh, 1984).

Inference From Xenolith Data

Ganguly et al (1995) have constructed paleogeotherm for the Indian shield using the pressure and temperature data of Proterozoic mantle xenoliths. These xenoliths have ages of about 1 Ga. They used the following solution of eqn (1) with eqn (2) relating temperatures at depths z_1 and z_2 as

$$T(z_2) = T(z_1) + \frac{q^*}{K_z}(z_2 - z_1) - \frac{A_0 d^2}{K_z}\left(e^{-z_2/d} - e^{z_1/d}\right) \qquad (14)$$

Ganguly et al (1995) used the following form of thermal conductivity function varying with depth:

$$K(z) = 2.5\frac{W}{mK} \quad \text{for} \quad z \le 30\ km \qquad (15i)$$

$$K(z) = 3.10 - 1.40.10^{-2}z + 2.18.10^{-4}z^2 - 5.34.10^{-7}z^3 \quad \text{for } 30 \le z \le 200\ km \qquad (15ii)$$

This form is obtained by fitting the data of Schatz and Simmons (1972) on thermal conductivity for continental shield. Ganguly et al (1995) used the heat flow and heat generation data as given in Gupta et al (1991). According to Gupta et al (1991), the surface heat generation and depth scale of exponential heat production model are 1.75.10^{-6}Wm^{-3} and 11.5 km, respectively. Surface heat production at 1*Ga* can be obtained by correcting for radioactive decays. Using average decay rate as 2.38 µW/yr, the surface heat generation at $A(1Ga)$=2.22 µWm^{-3}. Taking the surface temperature as 298 *K*, geotherm passing through the P-T arrays requires reduced heat flow as 20.92 mWm^{-2}.

Roy and Mareschal (2012) also fitted the geotherm to the xenoliths PT- arrays in Dharwar craton. They used more xenoliths data than used by Ganguly et al (1995). They constructed a large number of geotherms varying the values of Moho temperature, mantle heat flow and mantle radiogenic heat. The best fitting models had Moho heat flow lying within 18 – 20 mWm^{-2}, close to the finding of Ganguly et al (1995).

Perturbations In The Thermal Structure

Normal geotherms are perturbed due to changes in the basal boundary conditions due to additional heat flow from deeper mantle such as plumes or accretion of subducting slabs or due to influx of percolating fluids such as CO_2 or overthrusting of slabs or erosion. Growth of cratons takes place by action of plumes or addition by subducted slabs. How will thermal structure evolve in these cases? We refer to two studies which have been done to explain formation of Cuddapah basin. Bhattacharjee and Singh (1984) used vertical subsidence by eclogitization in lower crust and Anand et al (2003) used lithospheric stretching processes. Heating due to carbon dioxide flux through lithosphere and emplacement of overthrusted layers are used to explain the generation of charonockites by Ganguly et al (1995). Manglik et al (2005) and Manglik and Singh (2005) studied thermal evolution after thermal perturbation induced by magma underplating and convective thinning of lithosphere, respectively.

Cudappah Basin Evolution

In local isostasy, it is hypothesized that at the depth of compensation, all the crustal sections have the same weight. Taking these crustal columns as having uniform density, in one of the models, it is inferred that higher topography has deeper roots. Thus lowering topography will require changing the density of the subsurface, making it larger than its surrounding. Density of the material depends upon its constitution and the thermodynamic state, such as pressure and temperature. Thus, to induce subsidence changes in the composition and thermodynamic states are needed. This is possible by bringing high density material from greater depths to shallow regions or by changing the thermal or stress regimes. Subsidence is also possible if the low density material is brought to the upper levels of the crust. Subsidence can also be induced in the context of regional isostasy by changing the load patterns in the lithosphere. These loads can be both internal and on the earth's surface.

Cudappah basin is an intracratonic basin. Bhattacharjee and Singh(1984) proposed cooling of heat source at depth and resulting contraction as the mechanism for subsidence involved in formation of Cudappah basin following the work of Haxby et al (1976) for intracratonic Michigan basin. A heat source in the form of an excess temperature at depth d, extending to a depth of $2d$ is hypothesized. This forms initial condition for the time dependent heat conduction equation. The surface temperature is fixed to zero and so is the bottom temperature at depth of $3d$. We need solution of the following initial boundary value problem, ignoring the radiogenic heating in the lithosphere:

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \tag{16}$$

The boundary conditions are

$$T(0) = 0 \tag{17i}$$

$$T(d) = 0 \tag{17ii}$$

The following initial condition is more general than in Haxby et al (1976):

$$T(z,0) = 0 \;\; for \; 0 \le z \le a \tag{18i}$$

$$T(z,0) = \Delta T fora \le z \le b \tag{18ii}$$

$$T(z,0) = 0 \;\; forb \le z \le l \tag{18iii}$$

The general solution for any initial condition distribution for the above boundary condition is (Osizik, 1993):

$$T(z,t) = \sum_{m=1}^{\infty} e^{-\kappa\beta_m^2} \frac{2}{l} X(\beta_m, z) \int_0^l X(\beta_m, z')T(z',0)dz' \tag{19}$$

$$\beta_m = (n\pi/l) \tag{20}$$

$$X(\beta_m, z) = sin(\beta_m z) \tag{21}$$

The integral in eqn (19) is evaluated as

$$\int_0^d X(\beta_m, z')T(z'0)dz' = \Delta T \int_a^b sin\,(\beta_m z')dz'$$

$$= \Delta T(cos\beta_m a - cos\beta_m b)/\beta_m \tag{22}$$

Thus the solution of the problem is as given in Jaupart and Mareschal (2011):

$$T(z,t) = \left(\frac{2\Delta T}{l}\right) \sum_{m=1}^{\infty} e^{-\kappa\beta_m^2} sin(\beta_m z)(cos\beta_m a - cos\beta_m b)/\beta_m \tag{23}$$

The heat flux from the top surface is obtained from eqn(23) as

$$q_s = \left(\frac{2K\Delta T}{l}\right) \sum_{m=1}^{\infty} e^{-\kappa\beta_m^2 t}(cos\beta_m a - cos\beta_m b) \tag{24}$$

The total heat loss over the time duration t from the top surface is

$$Q_s - \int_0^t q_s(\tau)d\tau = \left(\frac{2K\Delta T}{l\kappa}\right) \sum_{m=1}^{\infty} (1 - e^{-\kappa\beta_m^2 t})(cos\beta_m a - cos\beta_m b)\,/\beta_m^2 \quad 25)$$

The total heat loss as time tends to infinity from top surface is

$$Q_a(t \to \infty) = \left(\frac{2K\Delta T}{l\kappa}\right) \sum_{m=1}^{\infty} (cos\beta_m a - cos\beta_m b) / \beta_m^2 \tag{26}$$

Using the following formula

$$\sum_{m=1}^{\infty} \frac{cosnx}{n^2} = \frac{\pi^2}{6} - \frac{\pi x}{2} + \frac{x^2}{4} \tag{27}$$

The total heat loss from both top and bottom surface of the heated body

$$Q_a = \frac{\rho_m C\Delta T}{l}(b-a)(2l-a-b) \tag{28}$$

For $a = b - a = l - b = l/3$ we get as in Haxby et al (1976)

$$Q_a = \rho_m Cl\nabla T/3 \tag{29}$$

The total load exerted by contraction is

$$P_0 = \rho_m \alpha \Delta T gl/3 = \frac{\alpha g Q_a}{C_p} \tag{30}$$

Load and the subsidence are obtained by calculating the flexure of the plate. For a disc load of radius a, the subsidence at the centre of the disc, $w(0)$ is given by (Haxby et al 1976)

$$w(0) = \frac{P_0}{(\rho_m - \rho_s)g}\left(1 + \frac{a}{B} ker' \frac{a}{B}\right) \tag{31}$$

Here density of mantle (sediment) is denoted by ρ_m (ρ_s), and g is the acceleration due to gravity. Flexural parameter B is given by

$$B = \left[\frac{D}{(\rho_m - \rho_s)g}\right]^{1/4} \tag{32}$$

The flexural rigidity D depends upon Young's modulus, E, plate thickness, h, and Poisson's ration, ν as

$$D = \frac{E h^3}{12(1-\nu^2)} \tag{33}$$

The plate thickness is determined from joint analysis of the gravity and topography data. The plate thickness is also derived by using rheology of the lithosphere. Bhattacharjee and Singh (1984) used olivine rheology to plot the depth variation of shear stress with depth using the thermal structure obtained from heat flow and heat generation values (Singh, 1981). Upper part would fail in fracture rather than flow as per Coulomb law relating shear stress with lithostatic pressure. The value of plate thickness is determined as 50 km. The value of B is estimated as 96 km. The radius of basin after correcting for shortening comes as 200 km. Thus the value of a/B is taken as 2. Substituting these values in eqn (31), the relation between load and central subsidence is

$$P_0 = 1.02.10^3 w(0) \tag{34}$$

Thus the relation between subsidence and excess temperature, ΔT, with value of coefficient of thermal expansion $\alpha = 4.10^{-5}$/°C, is

$$d\Delta T = 7.64.10^{-3} w(0) \tag{35}$$

The maximum thickness of the basin is taken ~10 km based on Deep Seismic Sounding studies (DSS) (Kaila et al, 1979). If excess temperature is 800°C, the vertical extent of heat source would be about 100 km. In case the density of sediments is taken as 2900 Kgm^{-3}, instead of 2500 Kgm^{-3} as earlier, the vertical extent of heat source is reduced to 48 km. There has been now reinterpretation of DSS data (Reddy et al, 2010) and the thickness is found to be much less than that taken in Bhattacharji and Singh (1984). Thus the both excess temperature and the thickness of the heat sources will be substantially less than derived above.

Anand et al (2003) have proposed lithospheric stretching model (McKenzie, 1978), for the formation of Cuddapah basin. Mathematically this model is described as solving the heat conduction equation with initial and boundary condition as:

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2}, \qquad 0 < z < l \tag{36}$$

$$T(z,0) = \frac{T_L z}{l/\beta}, \quad 0 \le z \le \frac{l}{\beta}; \quad = T_L, \quad \frac{l}{\beta} \le z \le l \tag{37}$$

$$T(0,t) = 0 \tag{38}$$

$$T(l,t) = T_L \tag{39}$$

This equation describes the process of reducing the thickness of lithosphere from l to l/β with earth surface fixed and lithosphere-asthenosphere boundary having temperature as T_L to rise to depth of l/β. β is termed as the stretching factor. The temperature remains linear within the lithosphere and uniform in the asthenosphere at T_L. Then the whole lithosphere cools and subsides. In case this stretching has resulted into lithosphere thickness reducing to zero, this would have been model for thermal cooling for lithosphere from mid-oceanic ridge, as it ages. In the stretching model, there would be no unconformity at the base of basin as there will be no uplift and no erosion. So if there is unconformity at the base of basin this model as it stands cannot be used. It would need modification if there is evidence of stretching.

This mathematical problem can be transformed to earlier case by substituting in eqn (36) the following:

$$T = \theta + T_L z/l \tag{40}$$

We then have

$$\frac{\partial\theta}{\partial t} = \kappa\frac{\partial^2\theta}{\partial z^2}, \qquad 0 < z < l \tag{41}$$

The boundary and initial conditions are transformed as

$$\theta(0,t) = 0 \tag{42}$$

$$\theta(L,t) = 0 \tag{43}$$

$$\theta(z,0) = \frac{T_L z}{l}(\beta - 1)\ 0 \le z \le l/\beta \tag{44i}$$

$$= T_L(1 - z/l) \quad l/\beta \le z \le l \tag{44ii}$$

Thus the solution of this initial boundary value problem is written as

$$\theta(z,t) = \sum_{m=1}^{\infty} e^{-\kappa\beta_m^2 t}\frac{2}{l}X(\beta_m,z)\int_0^l X(\beta_m,z')\theta(z',0)dz' \tag{45}$$

The integral in eqn (45) is evaluated as

$$\int_0^l X(\beta_m,z')T(z'0)dz'$$

$$= \frac{T_L}{l}(\beta-1)\int_0^{l/\beta} z'\sin(m\pi z'/l)dz' + T_L\int_{l/\beta}^{l}(1-z'/l)\sin(m\pi z'/l)dz'$$

$$= T_L\beta l\frac{\sin(m\pi/\beta)}{(m\pi/\beta)^2} \tag{46}$$

Thus

$$\theta(z,t) = T_L\frac{2\beta}{\pi^2}\sum_{m=1}^{\infty}\frac{\sin(m\pi z/l)\sin(m\pi\beta)}{m^2}exp(\kappa m\pi^2 t/l^2) \tag{47}$$

Thus the solution for this model is (McKenzie, 1978)

$$T(z,t) = T_L\left\{\frac{z}{l} + \frac{2\beta}{\pi^2}\sum_{m=1}^{\infty}\frac{\sin(m\pi/\beta)\sin(m\pi z/L)}{m^2}exp(-m^2\pi^2\kappa t/l^2)\right\} \tag{48}$$

As time tends to infinity, the temperature will relax to

$$T(z,\infty) = T_L z/l \tag{49}$$

Thus the elevation at a time *t* over the depth at time tends to infinity is (McKenzie, 1978)

$$e(t) = \alpha\int_0^l\big(T(z',\infty) - T(z',t)\big)dz' \tag{50}$$

Using eqns (48) a and (49), the expression for e(*t*) is

$$e(t) = \alpha T_L l\left[\frac{4\beta}{\pi^3}\sum_{m=1}^{\infty}\frac{\sin((2m-1)\pi/\beta)}{(2m-1)^3}exp\left(-(2m-1)^2\pi^2\kappa t/L^2\right)\right] \tag{51}$$

Approximate form of Eqn (51) is used by taking only first terms in the summation for small values of β, as shown by McKenzie (1978) that second term is very small in comparison with the first term. Thus

$$e(t) = \alpha T_L l\frac{4\beta}{\pi^3}\sin(\pi/\beta)exp(-\pi^2\kappa t/l^2) \tag{52}$$

Subsidence is given by

$$h(t) = e(0) - e(t)$$

$$= \alpha T_L l\frac{4\beta}{\pi^3}\sin\left(\frac{\pi}{\beta}\right)\left(1 - exp(-\pi^2\kappa t/l^2)\right) \tag{53}$$

Subsidence has two parts, initial tectonic subsidence, obtained by applying Airy isostatic hypothesis and later on thermal subsidence as given by above equation. Matching the subsidence data with the theoretical curve using eqn (53), the unknown value of stretching factor, β is obtained. For Cuddapah basin a stretching factor is arrived by Anand et al (2000, 2003), based on the subsidence and geochemical data. Geochemical signatures of both volcanic flows and sills show that degree of partial melting is 10-15%, the depth interval of partial melting ranges from120 km to 70-80 km, and the potential mantle temperature is taken as 1500ºC. Thus total thickness of melt is 4-5 km. For producing such a thick melt layer, the required stretching of the lithosphere is 1.6-1.8 following work of McKenzie and Bickle (1988).

Charnockitization By Advection Of Carbon Dioxide From Mantle

Formation of charnockites requires heating of the crustal rocks to more than 800ºC at depths of 20 km to metamorphose rocks without their partial melting. As rocks always contain some form of water, it is difficult to avoid partial melting. However, if the percolating fluid through lithosphere is carbon dioxide, it is possible to raise the temperature without melting, as carbon dioxide reduces the activity of water. To quantify such processes, the governing equation has advection terms:

$$\overline{\rho C}\frac{\partial T}{\partial t} - (\rho C\phi)_f v\frac{\partial T}{\partial z} = \frac{\partial}{\partial z}\left(K_e\frac{\partial T}{\partial z}\right) + A(z) \tag{54}$$

The velocity of upward moving percolating carbon dioxide is denoted as ν. $\overline{\rho C}$ is average values for solid and liquids. Subscript *f* refers to fluid. In absence of any heat source and in steady state, eqn (54) reduces to

$$\frac{\partial}{\partial z}\left(K_e\frac{\partial T}{\partial z}\right) + (\rho C\phi)_f v\frac{\partial T}{\partial z} = 0 \tag{55}$$

The boundary conditions for this equation are

$$T(0) = T_s \tag{56}$$

$$T(L) = T_L \tag{57}$$

The solution of this equation satisfying the boundary conditions is given in Brady (1988):

$$T(z) = T_s + (T_L - T_s)\frac{1-exp\,(-bz/L)}{1-exp\,(-b)} \tag{58}$$

$$b = \frac{(\rho C\phi)_f vL}{K_g} \tag{59}$$

Eqn (58) is fitted to the pressure and temperature data for charnockites /granulites. As done in Ganguly et al (1995), the resulting value of *b* is 7 for a value of *L* as 90km. The values of density, heat capacity and porosity for CO_2 are taken as 930 – 1530 kgm^{-3}, 1550 *J/KgK* and 1% respectively. The thermal conductivity for the lithosphere is taken as 2.5 *W/mK*. The Darcy velocity $v\phi$ is $0.3.10^{-2}$ *m/yr*. Ganguly et al (1995) also addressed the issue of how much time it will take to achieve the target temperature at various depths. The transient solution of the eqn (54) is obtained with the initial condition obtained by solving eqn (54) with its left hand side as zero. The method of eigenvalue - eigenfunction has been used to build transient solution. Computations are presented in Ganguly et al (1995). The evolution of temperature at 10 and 20 km depths for various values of Darcy velocity has been calculated and it is seen that required temperatures are attained in about 30*Ma* with Darcy velocity as ~ $0.5.10^{-2}$*m/yr*.

Reasonableness of such a value of Darcy velocity is shown by applying the Darcy law for flow of carbon dioxide. The Darcy velocity is related to pressure (P) gradient with depth as

$$v\phi = \frac{\kappa}{\rho_f\,\eta}\frac{dP}{dz} \tag{60}$$

Here κ, ρ_f and η are respectively permeability, fluid density and fluid viscosity. The pressure gradient is given by

$$\frac{dP}{dz} = (\rho_s - \rho_f)g \tag{61}$$

Combining eqn (60) and (61), the Darcy velocity is

$$v\phi = \frac{\kappa}{\eta}(\rho_s - \rho_f)g \tag{62}$$

With κ,η,ρ_s and ρ_f respectively as $10^{-18}m^2$, $1.5.10^{-4}$Pa.s, 2930 kgm^{-3} and 1200 kgm^{-3}, the value of $v\phi$ is $(4\text{-}5)10^4$m/yr. This is similar to earlier estimates for fitting the pressure- temperature arrays and time required for raising the temperatures at the target levels.

Ganguly et al (1995) estimated total carbon dioxide required for the charnockitization in south India. The total area of charnockitic rocks in south India and Srilanka is $1.5.10^5km^2$. Mass flux of CO_2 is $\rho v\phi$ which is $4kg/m^2yr$ or 90 mol/m^2yr. This fluxing took place over 30Ma. Thus, the total amount of carbon dioxide required is c.4.10^{20} mol. The inventory of carbon dioxide in the upper mantle is estimated to lie within $2.1.10^{22}$ – $4.2.10^{23}$mol (Des Marais, 1985). CO_2 has been out gassing in the past and during the Archean 0.09 to 15 times of the above amount would have been out gassed. Ganguly et al (1995) estimated that c.0.01 – 2% of the total Archean CO_2 outflow would have occurred through the south India, in case whole charnockitic terrain is formed by this processes. Ganguly et al (1995) termed these values as excessive.

Charnockitization With Thrusting And Crustal Thickening

Charnockitization can also be achieved with emplacement of thrust sheet increasing the radiogenic heat in the crust and further raising the temperature with uplift and erosion. With rapid thrusting a slab of crust is emplaced over the existing crust. Thus, lithosphere had initially saw-tooth temperature-depth profile. The whole lithosphere then underwent uplift and erosion. The rocks while rising to the surface are subjected to increasing temperature reaching a maximum and then decreasing temperatures. The plot maximum temperature and the depth at which this occurs is called metamorphic gradient (Singh and Ganguly, 2014). This is what is recorded by charnockitic rocks. The thickness of the slab and rate of uplift and erosion are changed to fit metamorphic data. For example initial condition is in simplest case of no radiogenic heat in the lithosphere and emplacement of a slab of thickness, *d* can be written as

$$T(z,0) = Gz \quad 0 \le z \le d \tag{63i}$$

$$= G(z-d) \quad d \le z \le l \tag{63ii}$$

The constant G is the uniform temperature gradient in lithosphere. In presence of radiogenic heat, the profile departs from linearity and can be obtained by using the solution given by eqn (3) for exponential model of radiogenic heat distribution. The governing equation for time dependent solution in this case is obtained by solving the following equation:

$$\rho C\left(\frac{\partial T}{\partial t} - w\frac{\partial T}{\partial z}\right) = K\frac{\partial^2 T}{\partial z^2} + A(z,t) \tag{64}$$

The uplift and erosion rate is denoted by w. The boundary conditions are

$$T(0,t) = 0 \tag{65}$$

$$K\frac{\partial T}{\partial z}\bigg|_{z=l} = q^* \tag{66}$$

The solution of this equation can be obtained using method of Laplace transforms (Carslaw and Jaeger, 1959) or numerical methods. Ganguly et al (1995) used a numerical algorithm by Haugerud (1986) to simulate this process for two thicknesses of slab as 25 and 35 km and uplift/erosion rate of 0.1 and 0.05 mm/yr. The basal boundary condition is assumed as constant heat flux 21.3 mW/m^2. The initial saw tooth profile is built from the solution of steady geotherms as given in the above analysis of metamorphism due to CO_2 flux. It is seen that a 30 – 35 km thrust slab with uplift/erosion rate as 0.1 mm/yr yields metamorphic arrays as observed in the charnockitic rocks. Presence of such overthrust sheets are seen in seismic images of south India, which have been interpreted in terms of compressive tectonics and crustal thickening associated with subduction processes (Reddy et al, 2003; Rajendra Prasad et al, 2007, Vijay Rao and Reddy, 2010).

Lithosphere-Asthenosphere (LAB) Boundary Condition

Several boundary conditions have been used at the LAB. As the heat transfer is described by the heat conduction equation, there are three possibilities of the boundary conditions, each representing different nature of underlying processes in the asthenosphere. These possibilities are: Dirichlet, Neumann or Robbin type boundary conditions. In the Dirichlet case, the temperature is prescribed. In the Neumann case the heat flux is prescribed. And in the Robbin case a combination of temperature and heat flux is prescribed. In the plate model of lithosphere (McKenzie, 1967), the basal condition is defined as given temperature condition. It is known that the mantle is in convection and asthenosphere undergoes small-scale convection (Parsons and McKenzie, 1978, Jaupart and Mareschal, 2011). The given temperature at this interface thus demands this asthenosphere to be in vigorous thermal convection. It also demands that heat flows vertically with no lateral heat transport. If this convection is not vigorous, this boundary condition is not appropriate. Another boundary condition used is by prescribing the heat flow at the interface, which deepens due to cooling, as a Stefan moving boundary problem. This is used in well known CHBILIS model (Cough 1975, Doin and Fleitout, 1996, Gliko and Mareschal, 1989, Manglik, 1994, Manglik et al, 1992, 1995). In this case lateral heat transport would be taking place in the asthenosphere. Robbin type boundary condition is used by Richter (1984, 1985), Ramana (1993) and Ramana and Singh (1994). Thus, these boundary conditions entirely depend upon the nature of the small-scale convection at the base of the lithosphere. As knowledge about small-scale convection evolves so would be the knowledge about implications to thermal state of basal part of the cratonic lithosphere.

Isopycnicity Hypothesis

A major point of discussion in craton evolution has been the isopycnicity hypothesis of Jordan (1978), which proposes larger thickness under cratons, as cratonic keel. It exploits dependence of density both on temperature and composition. Thermal cooling induces an increase in the density, whereas partial melting could lead to decrease in the density due to removal of high density incompatible elements and their transport with melts. Samples of both partial melts emplaced in crust and residues in the mantle are available for cratons in the form of volcanic rocks and xenoliths, former brought to the surface at the time of partial melting and later brought up later on with volcanic activities. Geochemical studies of these samples point to two end-member processes in cratonic keel formation. In the first, a plume head impinges the base of cratonic lithosphere partially melting the basal part of lithosphere. Melts are emplaced in or on crust and residues remain in cratonic mantle lithosphere. In other, the subducted oceanic lithosphere/arcs join the craton at depth, building craton by accreted slab stacking. In this process too partial melting takes place in cratonic mantle, but at lower pressures and in presence of water compared to plume model. Geochemical and petrological data favor later more than former.

Eaton and Perry (2013) have considered evolution of a cylindrical body of cratonic keel, having maximum thickness as 260 km and diameter as 3000 km. In case of heating by a plume, the initial thermal state is described by its basal boundary having a fixed temperature varying with depth of the base of the lithosphere and extrapolated to the past to the age of 3Ga. The upper surface has constant temperature as zero. On this initial state, a plume adds a transient decaying temperature anomaly with time constant of 50 Myr and lateral width of 400 km. In the other case, the initial temperature state is formed by vertically stacking two mantle wedges/ slabs. The thermal state of slabs is that of thermally mature lithosphere. The evolutions of both hot and cold initial states shows that after one billion years, the thermal states of the cratonic keel was similar for both the cases. Thus, it is not possible to infer initial thermal condition from today's conditions to choose from two alternative schemes of cratonic stability. However, it would be interesting to apply both end member models and also their combinations to the Indian cratonic conditions and derive their geochemical implications towards explaining the observations.

CONCLUDING REMARKS

Heat conduction theory has been used to construct the knowledge about the earth, ever since it was formulated by

Fourier. Kelvin applied this theory to estimate the age of the earth by studying its cooling. Now the mathematical derivations of Kelvin are used to constrain the age of the oceanic lithosphere. Several geological processes are shown to be consequences of transmission of energy from the mantle represented as enhanced temperature or enhanced heat flow or a combination of both. Heat is also added in form of magmatic intrusion. Lithosphere also cools faster by erosion. All these effects can be modeled using heat conduction equation with changing initial condition or boundary condition or source functions or thermal properties. The mathematical solutions of such an initial-boundary value problem are compared with observations of surface heat flow or metamorphism of rocks or characteristics of emplacement of igneous bodies along with accompanying xenoliths, subsidence and uplift of lithosphere and associated formation of sedimentary basins. Only a limited number of such studies have been carried out for understanding Indian geological terrains and these have been summarized in this paper. It is pointed out that such an equation based thinking for understanding the Indian geology should also be tried along with data based thinking, which has more attraction being closer to the geological reality.

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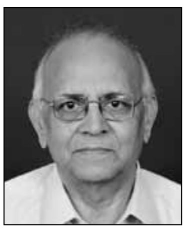

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