Granitoids localization in collisional overthrust structures subject to thermal conditions-numerical modeling

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High temperature metamorphism and widespread granite magmatism are common peculiarities of many ancient continental collision structures. Deeply eroded areas of the Archean and Proterozoic continental shields formed in the process of tectonic evolution including horizontal shortening and collision expose at the surface middle to the lower crust rocks uplifted by overthrusting from the depths of 20—40 km. This phenomenon seems to be valid also for modern collision zones. The massive layer of granite melt approximately of 10 km thickness is observed at 10—15 km depth by seismic methods in the resent orogens such as the Himalayas and Caucasus [Rosen, Fedorovsky, 2001]. The Palaeozoic Variscan orogen in Europe is characterized by large volumes of felsic granites intruded during HT-LP metamorphism and exposed in deeply eroded parts of the Variscan belt (South Bohemian Batholit) [Gerdes et al., 2000].

Fundamental features of collision zones reflect the effect of the main tectonic event — horizontal shortening in compression setting, collision of two continental plates accompanied by the thickening of the crust and the surface uplift. Extensive development of horizontal and oblique motions of crustal plates and blocks leads to the disturbances in the thermal regime, heat flow, the surface and Moho topography. The main petrologic mark of such collision is granite melt generated at different depth's levels and exposed at the surface as a result of the denudation and uplift.

Thermal-kinematic model of continental collision calculates pressure, velocity and temperature fields and includes horizontal shortening, brittle overthrusting in the upper crust compensated by the lower crust viscous flow and erosion of the thickened crust. Finite-element 2D modeling is used to examine the thermal and kinematic conditions for high-temperature metamorphism and the depth and timing of crustal melting in the case of rheologically layered lithosphere [Parphenuk et al., 1994; Parphenuk, 2005]. The thermal effects during collision process and postorogenic stage are studied.

Calculation of lithospheric temperature field is based on crustal thickening by overthrusting with dip angle of faulting of 15—30° and shortening rate in the range of 0.5—2 cm/y. Total amount of horizontal shortening is 100 km. We assumed that erosion and concurrent sedimentation begins after an additional crustal portion of a substantial thickness is exposed. The equation of energy conservation is solved for the case of Lagrange coordinates with material time derivative [Turcotte, Shubert, 1985]:

$$c_i \rho_i \frac{DT}{Dt} = \lambda_i \nabla^2 T + H_i ,$$

where c is specific heat, ρ is density, λ is the thermal conductivity, and H is the heat generation rate. The model consists of three layers: 20 km brittle upper crust (i=1), 20 km ductile lower crust (i=2) and 80 km lithospheric mantle (i=3) with different thermal, kinematic and rheological parameters.

The calculations show the possibility of the partial melting and granite melt generation subjected to the main thermal parameters of the model — the initial temperature distribution (heat flow higher than 60 mW/m^2) and radiogenic heat production. The temperature increase can be fairly significant (up to 250 °C) at depths level of 10—30 km confirming the idea of crustal origin of the continental collisional granites. The partial melting (wet granite solidus) starts in the thickened crust in the vicinity of thrust fault and widens in the direction of thrusting and to the lesser depths. Maximum temperature increase along the fault zone is 300 °C.

2D collision model confirms the observation of a wide range of *PT*-conditions over short distances in metamorphic belts [Chamberlain, Karabinos, 1987]. As a result of upward motion along the fault, the additional loading redistribution in the course of erosion, and the viscous compensation at the level of the lower crust, *PT*-histories will be completely different for the points along the thrust fault. For example, for the convergence and erosion rates of 0.5 and

0.025 cm/y, respectively, maximum erosion level of ~9 km of uplifted crust is reached at 45 Ma after the onset of collision (at postcollisional stage). The material will be transported from a depth of 20 km to a depth of about 5 km while the rocks that were initially located at a depth of 3.5 km will experience a rather complicated *PT*-evolution. At the postcollisional stage (after 20 Ma of shortening) the compressional regime is changed by extension with very low velocities.

Local effect may result from the frictional heating along the slip zone during the overthrusting. The frictional heat production is proportional to the product of the shear stress across the slip plane, the velocity of obduction and coefficient of friction. The additional heating can rise the temperature by 50—150 °C at 10—20 Ma in the vicinity of slip zone in the case of horizontal shortening rate of up to 4 cm/y and thrust sheet thickness up to 20 km [Brewer, 1981]. This local and moderate additional heating (in comparison with crust thickening effect) is much less in the case of slower thickening.

The following set of parameters is critical to initiate crustal melting and granite formation: initial temperature distribution with heat flow density value higher than 60 mW/m^2, relatively high radiogenic heat production, slow crustal thickening (0.5—1 cm/y for 10—20 Ma) and slow exhumation. The results of the modeling confirm other model estimates [England, Thompson, 1984; Gerdes et al., 2000].

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