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Dynamics of the Arctic and adjacent petroleum basins: a record of plume and rifting activity

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Abstract

The Late Paleozoic and Mesozoic history of high-latitude petroleum and coal basins is investigated and compared with the history of plume magmatism in the same areas. The sedimentation rates in all discussed cases are proven to be the fastest (more than 100 m per 1 Myr) during rifting events. Other peaks of rapid deposition may be associated with collisional mountain growth and/or climate change.

Keywords: basin; subsidence history; sedimentation; rifting; plume; Siberian Arctic

Introduction

There are three types of rift basins in the Siberian Arctic distinguished according to their age and subsidence history. (1) Late Paleozoic–Mesozoic basins with highest sedimentation rates in the Late Devonian (about 380–360 Ma) and angular unconformities at the base of C_2 and T_{1-2}. They are the Voronin trough (Fig. 1 in Vernikovsky et al., 2013) and, possibly, also a basin along the margin of the Novaya Zemlya plate, as well as rift basins in the eastern (Vilyui–Verkhoyansk) and southern (Minusa and Kuznetsk) periphery of the Siberian craton. Their evolution was discussed in detail with the example of the Kuznetsk basin, which is referred to in many publications (Belyaev et al., 2008; Geology..., 1959; Polyansky et al., 2004). (2) Permian–Triassic–Jurassic rifts with most rapid subsidence and deposition in the latest Permian and Early–Middle Triassic (260–230 Ma). They are the Koltogory–Urengoi rift at the base of the West Siberian petroleum basin, a rift recorded in the lowermost sediments of the Yenisei–Khatanga basin, the Ust'-Lena rift, the Saint Anna trough, and the Chukchi basin. The basins of types 1 and 2 experienced faster sedimentation during later phases of rifting or mountain growth. (3) Cretaceous basins and rifts that subsided and accumulated sediments most rapidly in the Hauterivian–Aptian (137–110 Ma) or Cenomanian–Turonian (100–90 Ma). These are the rifts in the eastern Russian Arctic (North and South Chukchi basins) (Artyushkov, 2010; Khain et al., 2011).

The highest sedimentation rates in all three types of basins correlate with peaks of plume magmatism, of which some are considered in our another paper from this issue (Dobretsov et al., 2013). The plume activity culminated in the Devonian (370–400 Ma), as appears in the Kuznetsk and Vilyui basins, in the Permian (two events at 290–270 and 260–240 Ma, the more prominent later one known to produce the Siberian Trap Province), in the Jurassic (210–180 and 170–150 Ma events), and in the Cretaceous (130–110 and 100–90 Ma).

The relationship among plume magmatism, rifting, and sedimentation was modeled in a simplified way by McKenzie (Jarvis and McKenzie, 1980; McKenzie, 1978) and specified in later studies (Dobretsov, 2010; Dobretsov et al., 2010; Khain, 2010; Polyansky et al., 2000, 2004; Sheplev and Reverdatto, 1994). The model implies upwelling of hot asthenosphere (in early versions) or a lens-shaped head of a mantle plume (in a later version) and the ensuing thinning and extension of the lithosphere during rifting followed by postrift cooling and isostatic rebound. In different modifications of the models, the lithosphere is assumed to stretch and attenuate rapidly (McKenzie, 1978), with exponential acceleration until cessation (Jarvis and McKenzie, 1980), or at a constant rate (Sheplev and Reverdatto, 1994).

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More sophisticated models of pull-apart rifting were discussed in (Polyansky, 2002; Polyansky and Dobretsov, 2001).

Modeling method

The actual sediment thickness in a basin (basin depth) is predicted in the following way. The sedimentation process is modeled in 1D, as a function of depth, as producing several layers, for which properties of different lithologies are specified by the user (density, thermal conductivity, compaction, and initial surface porosity). The method implies successive reconstruction of the paleotemperatures of sediments and the paleodepths of the basin assuming a variable number of layers with known thermal-mechanical properties of porous rocks. The algorithm consists in reconstructing subsidence depths for each layer back into time since the presumed end of the basin evolution. The weight of the sedimentary column decreases progressively and each successive layer, from the youngest one, is stripped back to the surface (Steckler and Watts, 1978), while its thickness is corrected for compaction.

We estimated the possible amount of stretching for a two-layer model developed after the one-layer model of McKenzie (1978). Rifting parameters were calculated for a two-layer lithosphere (crust plus lithospheric mantle) stretched nonuniformly with depth. The stretching percentage presumably correlated with the volume of vertical basaltic dikes. The crust was assumed to be isostatically balanced during deposition.

The method is as follows. It stems from the isostasy principle for attenuation in the system "crust–subcrustal lithosphere" and has been modified to take into account compaction of porous sediments, thermal expansion of the lithosphere during rifting, and lithology variations. Let the crust and lithospheric mantle in the stretching lithosphere have their thicknesses and densities of h_c and ρ_c and H and ρ_m, respectively, and lie upon an asthenosphere with the density ρ_a. Their thicknesses attenuate as h/β and H/δ, respectively, and the total attenuation is

$$\varepsilon = \frac{H + h_c}{h_c/\beta + H/\delta},$$

where $\beta > 1$ and $\delta > 1$ are the stretching factors for the crust and lithospheric mantle, respectively. Extension at the account of dikes with the relative contribution γ ($0 < \gamma < 1$) to the basin volume is $\varepsilon' = \dfrac{1}{1-\gamma}$.

The sediment temperatures derived from a 1D heat transfer equation, at the depth z below the top of the nth sedimentary layer, are

$$T(z) = T_0 + Q\left[\frac{1}{\lambda_n}\left(z - \sum_{i=1}^{n-1} d_i\right) + \sum_{i=1}^{n-1}\frac{d_i}{\lambda_i}\right], \quad (2)$$

where T_0 is the surface temperature, λ_n is the average thermal conductivity of sediments, Q is the heat flow in, mW/m^2, d_i is the layer thickness, λ_i is the thermal conductivity of each layer (counted from the surface). The thermal conductivity of porous sediments is found as a geometrical mean over those of the fluid (pore water) and the solid: $\lambda = \lambda_f^{\varphi} \cdot \lambda_s^{(1-\varphi)}$, where φ is the porosity and λ_f, λ_s—thermal conductivity of fluid and solid, respectively.

Lithospheric extension produces a basin, with the maximum sediment thickness

$$S_{sed} = \frac{1}{\rho_a - \rho_{sed}}\left[(\rho_a - \rho_c)\, h\left(1 - \frac{1}{\beta}\right) + (\rho_a - \rho_m)\, H\left(1 - \frac{1}{\delta}\right)\right]. \quad (3)$$

Equation (3) relates the sediment thickness and the amount of extension for the case when deposition keeps up with subsidence. The equation does not allow for thermal expansion of the lithosphere and becomes more complicated with the latter. Namely, the initial or synrift subsidence (S_i) of a basin filled with water, at the user-specified stretching factors β and δ for the crust and the mantle, respectively, is a difference between the subsidence due to isostatic compensation E and the thermal expansion T of the lithosphere (Friedinger et al., 1991):

$$S_i = \frac{E - T}{\rho_m(1 - \alpha T_m) - \rho_w}, \quad (4)$$

while

$$E = (\rho_m - \rho_c)\, h_c\left(1 - \frac{1-\gamma}{\beta}\right)\left(1 - \frac{T_m \alpha h_c}{2a}\right), \quad (5)$$

$$T = \frac{T_m \alpha \rho_m a}{2}\left\{\left[\left(1 - \frac{1}{\delta}\right) + \left(\frac{h_c^2}{a^2} - \frac{2h_c}{a}\right)\left(\frac{1}{\beta} - \frac{1}{\delta}\right)\right](1-\gamma) + \gamma\right\}, \quad (6)$$

where $\rho_c = 2.8$, $\rho_m = 3.33$, and $\rho_w = 1.03$ (in g/cm^3) are the densities of crust, mantle, and water, respectively; $\alpha = 3.3 \times 10^{-5}$ (°C^{-1}) is the lithospheric thermal expansion coefficient, $T_m = 1350$ °C is the adiabatic temperature of the upper mantle in the model domain, $a = H + h_c$ is the lithospheric thickness, and h_c is the prerift thickness of the crust.

Inversion consists in optimization (fitting) of the stretching parameters for the crust and subcrustal lithosphere according to the depths of stratigraphic boundaries. The inversion quality is measured as the mean square deviation between the observed and modeling basin subsidence curves:

$$\sigma^2 = \frac{1}{n}\sum_{i=1}^{n}\left|h_i^{mod} - h_i^{obs}\right|^2, \quad (7)$$

where n is the number of stratigraphic units in the basin section. The minimum rms error σ^2 is found by fitting the crust and mantle stretching factors. The quality is satisfactory if the rms error is within a few percent, and the respective