Subsidence of the West Siberian Basin: Effects of a mantle plume impact

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ABSTRACT

Comparison of modelling results with observed subsidence patterns from the West Siberian Basin provides new insight into the origin of the Siberian Traps, and constrains the temperature, size, and depth of an impacting mantle plume head during and after the eruption of the Siberian Traps at the Permian-Triassic boundary (250 Ma). We compare synthetic subsidence patterns from 1-D conductive heat flow models to observed subsidence from backstripping studies on sedimentary sections from wells in the basin interior. This results in a best-fit scenario with a 50-km thick initial plume head with a temperature of 1500 °C situated 50 km below the surface, and an initial regional crustal thickness of 34 km, in agreement with published values. Backstripping and modeling results agree very well, including a 60-90 Myr delay between the rifting phase and the first regional subsidence below sea level. Regional subsidence patterns indicate that the plume head was present across a minimum area of ~2.5 million km^2. These results re-emphasize the viability of a mantle plume scenario for the Siberian Traps, provide

important constraints on the dynamics of mantle plume heads and suggest a thermal control for the subsidence of the West Siberian Basin.

Keywords: plume head, Siberia, basin, subsidence.

INTRODUCTION

The West Siberian Basin (WSB) is one of the largest intra-continental rift basins in the world, with an area of roughly 3.5 million km^2, including prolongations in to the Yenisey-Khatanga Trough and the Kara Sea (Fig. 1). The basin is associated with the Siberian Traps, which form the largest Phanerozoic continental flood basalt province. These events are commonly interpreted to be the result of the impact of a mantle plume head at the base of the Siberian lithosphere (e.g. Richards et al., 1989).

An important feature of the WSB is the regional delay, on the order of 60-90 Myrs, in the onset of sedimentation after the initial rifting (Saunders et al., 2005). The cause of this delay has been inferred to be decay of uplift generated by the thermal effect of a mantle plume (Campbell and Griffiths, 1990; Saunders et al., 2005), but so far this hypothesis has not been quantified. The subsidence patterns within the WSB also present an ideal opportunity to examine the generic spatial extent of spreading plume material beneath lithosphere. Previously this has only been estimated from the areal extent of volcanism (d'Acremont et al., 2003) and numerical models of plume head spreading (Campbell, 2007). By comparing subsidence patterns predicted by 1-D conductive heat flow models with those from backstripping, this study provides 1) a quantitative explanation for the cause of the subsidence delay after initial rifting, 2) independent estimates of the lateral extent of the Siberian mantle plume head beneath the lithosphere, and 3) new constraints on mantle plume head dynamics.

GEOLOGICAL BACKGROUND

At the end of the Permian (~250 Ma) the WSB underwent extension associated with eruption of the Siberian flood basalts (Reichow et al., 2009). Figure 1 shows the locations of the main rifts and the extent of the flood basalts. The flood basalts reach a maximum thickness of 3 km in the northwest of the Siberian Craton. An average thickness of 2 km is deduced by Braitenberg and Ebbing (2009) from gravity modelling. Most authors agree that the flood basalts in the WSB were erupted sub-aerially (Westphal et al., 1998; Saunders et al., 2005; Vyssotski et al., 2006; Saunders et al., 2007) and therefore it is likely that the basement hadn't started to subside at the time.

During the Triassic, sediment deposition continued within the rifts at a reduced rate, but there was no deposition (or at least preservation) outside the rifts. In areas adjacent to the major rifts, sedimentation began in the Early Jurassic (~200 Ma). The first basin wide transgression did not occur until the Callovian (~165 Ma) (Peterson and Clark, 1991; Vyssotski et al., 2006). This is visible in the backstripped subsidence curves from Saunders et al. (2005) (Fig. 2). Sedimentation throughout the rest of the Mesozoic and early Cenozoic was strongly influenced by changes in global sea level. There was little further tectonic activity within the basin after the rifting.

Vyssotski et al. (2006) reports the crustal thickness of the WSB varying from a minumum of 34 km within the rifts to ~44 km at its margins. The average thickness across the basin is ~38 km, which is signifacantly thinner than the Siberian craton to the east or the Ural mountains to the west. The thickness of the lithosphere beneath the WSB is not certain. Artemieva and Mooney (2001) used heat flow data to model the geothermal gradient, and calculated a lithosphere thickness of ~ 125 km across the WSB. In comparison, Priestley and McKenzie (2006) used shear wave velocity gradients to calculate a thickness of ~180 km.

It is important to assess the viable range of possible temperatures, thicknesses and depths of a plume head impacting on the base of the WSB lithosphere. The temperature can be estimated from the composition of the erupted volcanics. Saunders et al. (2005) used trace element ratios to argue for a decrease in the depth of melting from > 100 km to 100-50 km over the evolution of the flood basalts. We interpret this to show that the plume effectively thins the overlying lithosphere to ~50 km.

The thickness of a plume head spreading out beneath continental lithosphere was investigated by Nyblade and Sleep (2003) who calculated thickness in the order of 10s of km. They also calculate that a 40 km thick plume head with an excess temperature of 200 ˚C will thin the lithosphere by 50 km. In contrast, Campbell's (2007) model of plumes originating at the core mantle boundary show that the plume head may have had a thickness of 175 ± 25 km. Therefore this parameter is not well-constrained for any possible plume head, including the Siberian example.

METHODOLOGY

A 1-D forward thermal model was used to examine whether the cooling of a thermal anomaly generated by a plume head could cause the subsidence in the WSB. Conductive heat flow through the lithosphere and upper mantle and the associated subsidence of the column are calculated numerically. An in-depth description of the model is found in Holt et al. (2010). Temperature T for each depth z is integrated over time t by combining Fourier's law for conductive heat flow with conservation of energy and a radiogenic heat production A:

$$\frac{\partial T}{\partial t} = \frac{1}{C_p \rho}\{\frac{\partial}{\partial z}(k\frac{\partial T}{\partial z})\} + \frac{A}{C_p \rho} (1).$$

Specific heat C_p is related to the rock type, while density ρ is dependent on both temperature and rock type (Holt et al., 2010). The resulting density profile is used to calculate the isostatic height of the column relative to sea level calibrated to a column of mid-ocean ridge material with 2.7 km of water overlying a 7 km basaltic-gabbroic crust above a peridotitic upper mantle. For negative surface heights we extend the top of the column to sea level with water. This allows comparison with water-loaded tectonic subsidence calculations from backstripping. Similarly to water loading, sediment loading is also calculated to compare the model to the observed Moho depths. The sedimentation rate is assumed to keep pace with the subsidence. Since the Moho depths are those observed today and not the initial conditions, the model was run to produce a final crustal thickness, including sediments deposited in the basin that matches the present-day crustal thicknesses.

The temperature at the top (0 °C) and bottom (1381 °C) of the model are fixed using a potential temperature at the surface of 1330 °C, an adiabatic temperature increase of 0.3 °C/km and a model depth of 170 km. The model is benchmarked against sea-floor spreading models (Parsons and Sclater, 1977; Stein and Stein, 1992) for heat flow and bathymetry.

The initial conditions of the model are set up to match our current knowledge of the WSB as closely as possible. The crust is composed of a layer of flood basalt at the surface above a granitic upper crust and a lower crust with a density corresponding to granulite and mafic intrusions related to the volcanics at the surface. The thickness of the crust is varied, as are the flood basalt thickness and the upper – lower crust ratio to model the changes across the WSB. The crust is underlain by mantle lithosphere down to the top of the plume head. The initial temperature profile is a linear gradient from the surface, and intersects the mantle adiabat at the top of the plume head. The (initially constant) temperature, thickness and position of the underlying plume head are varied. Below the plume head, the temperature follows the mantle adiabat. This results in the base of the lithosphere settling at ~ 150 km (using the 1200 °C definition of Stein and Stein (1992)) after 250 Myrs, which is an intermediate value of the various estimates of present-day lithospheric thickness.

THERMAL SUBSIDENCE CURVES

Backstripped wells from Saunders et al. (2005) were used to test the models as they cover a broad range of locations within the WSB, with different crustal and sedimentary thicknesses (Fig. 1). The backstripping analysis showed earlier onset of sedimentation in the rifts, around 250 – 240 Ma, and a greater amount of subsidence than the wells from the wider basin, where the onset of sedimentation was between 200 and 160 Ma (Fig. 2).

The backstripped wells are shown alongside the modeled subsidence in Fig.2. Best fits are obtained with a plume head from 50 km to 100 km depth, with a temperature of 1500 °C. An initial crustal thickness of 34 km produces a subsidence curve that fits the wells outside the rifts (N, Sa, Sl, Su), giving a final crustal thickness including the sediments of 37 km. This is equivalent to present crustal thicknesses across the WSB, outside of the main rifts (Vyssotski et al., 2006). Subsidence curves for wells within the rifts are successfully modeled by an initial crustal thickness of 30 km (Fig. 2). Changes in only the initial crustal thickness are sufficient to fit subsidence patterns in the rifts and in the wider basin. Well data from outside the rifts is representative of more regional subsidence patterns. The 3 km contour of sediment thickness from Vyssotski et al. (2006) is a proxy for regional subsidence on the scale described above, and implies that the plume head lay under this area of over 2.5 million km^2, with little variation in depth to the top of the plume head.

Sensitivities of model parameters are shown in Fig. 3. In each case one parameter is varied compared to a standard set up (Fig 3a). Results are very sensitive to crustal thicknesses (Fig. 3b); a 46 km thick crust never subsides below sea level so is unlikely to form a basin, whereas a 38 km thick crust starts with an elevation of 1400 m, which then subsides, dropping below sea level after 88 Myrs. For a crustal thickness of 34 km, the initial elevation is 360 m and the model drops below sea level after 14 Myrs. The likely variation in the flood basalt thickness in the WSB has a much smaller effect on the subsidence of the basin than the crustal thickness.

Plume temperatures were varied between 1400 – 1600 °C to cover the range of plausible temperatures following melting and thinning of the lithosphere (Fig.3c). Results reveal that a hotter plume results in more initial uplift and increases the length of time before a basin begins to form. However, plume temperature has only a relatively minor influence. Furthermore, for all model runs the final temperature profile, and therefore the final subsidence, is the same.

To test the effect of the depth of the plume the model was run for a 50 km thick plume head situated various depths. The effect of plume head thickness was also tested. Fig. 3d shows that where the plume reaches a depth of 50 km the model starts significantly above sea-level regardless of its thickness. For a 120 km thick plume the initial elevation is 1700 m whereas a 50 km thick plume has an initial elevation of 1400 m and the onset of sedimentation is hastened. There is little difference when compared to varying the depth of the plume. If a 50 km thick plume begins at 100 km depth then its initial elevation is 350 m above sea level, whereas if the plume is 20 km deeper then the initial elevation of the plume is 250 m below sea level. The results illustrate that the subsidence curve is much more sensitive to the depth of the plume head than the thickness of the plume head. This is because the temperature contrast of the plume with the normal geotherm is greater at shallower depth, which will cause a greater reduction in the density of the material and therefore more initial uplift. Variations in radioactive heat production were previously shown to have little effect on the either the total subsidence or the shape of the subsidence curve (Holt et al., 2010), and therefore are not discussed in this study.

DISCUSSION

Uplift has been suggested in conjunction with the eruption of a number of other large igneous provinces (LIP's) including the Emeishan traps, the Deccan traps, the North Atlantic LIP and Yellowstone (Saunders et al., 2007). Our modeling shows that if the lithosphere is significantly thinned or heated, then isostasy alone will be enough to cause the uplift seen. This is in agreement with the recently proposed isostatic cause for the abnormal elevation seen in the American Cordillera (Hyndman and Currie, 2011) although the reason for the temperature anomaly is markedly different in each case. We have not included dynamic uplift in our model as it is a transient effect only acting while the plume was beneath the basin. Whether a plume can thin lithosphere, as our results indicate, is matter of debate both in active plume such as Hawaii and in numerical models. Ribe and Christensen (1999) calculated that the majority of the Hawaiian swell could be accounted for by dynamic uplift with only minor thinning of the lithosphere to 89 km. Likewise in the model of Nyblade and Sleep (2003) there is some lithospheric thinning, but it limited to the rheological boundary layer at the base of the lithosphere. In contrast Li et al., (2004) showed that seismic data indicates that the lithosphere beneath Hawaii is only 50-60 km thick. Similarly the model of d'Acremont et al., (2003) that focus on implementing a realistic crust and mantle lithosphere rheology shows that strain rate and stress weakening of the lithosphere enhances its erosion by a plume head. Our study provides another line of independent evidence supporting thinning of the lithosphere. Geochemical evidence from the North Atlantic LIP (Kerr, 1994) shows such thinning is associated with other flood basalt provinces. The areal extent that the lithosphere is thinned over will likely vary somewhat between LIP's as the volumes of flood basalts do. d'Acremont et al (2003) use the extent of flood basalts visible at the surface to estimate that the plume head thins the lithosphere over an area ~3 million km^2. Our method, using the subsidence patterns to give an indication of this, is preferable as flood basalts can flow large distances from where they are erupted whereas the subsidence is more directly linked to the state of the lithosphere beneath it.

CONCLUSIONS

Comparing thermal subsidence curves from numerical heat flow models with backstripping results illustrates that the West-Siberian Basin subsidence is well explained by a 1500 °C, 50-km-thick plume head, impinging on a overlying lithosphere that is ~50 km thick after plume head emplacement with a 30 km thick crust in the rifts and a 34 km thick crust in the wider basin. The modeling predicts that the subsidence in the WSB is most sensitive to the crustal thickness within the basin and depth to the top of the plume head, and shows less sensitivity to (reasonable) variation in the temperature and thickness of the plume head and the thickness of the flood basalts. Thermal modeling results show an excellent fit to both the magnitude and timing of subsidence, within the rifts and the wider basin. We conclude that the flood basalts, rifting and subsequent basin formation associated with the West Siberian Traps are best explained by a plume head which spreads out beneath the entire basin, while eroding the lithosphere to ~50 km thickness, and covering an area of over 2.5 million km^2. We propose that quantitative comparison between modeled and observed subsidence of continental flood basalt provinces is a fruitful way to provide tighter constraints on the volume, lateral extent, and thermal erosion effects of mantle plume heads.

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Figure Captions

Figure 1. Location of the West Siberian Basin (dashed line), the 3 km sedimentary thickness contour (dotted line) and the neighboring Siberian Craton (dot dash line). Red dots mark backstripped wells from Saunders et al. (2005). N=Novoporto-130, S=SG-6, Sa=Samotlar-39, Sl=Salym-184, Su=Surgot-51, U=Urengoy-414. Modified from Allen et al. (2006).

Figure 2. Backstripped water-loaded subsidence from wells across the basin (Saunders et al., 2005) is compared to the subsidence produced by the forward model. The wells within the rifts are fitted best by a model with an initial 30 km thick crust. This does not fit the rift phase of SG-6 well because rifting is not included in the model, however it is a close fit to the thermal subsidence phase. Delayed onset in sedimentation seen from the wells outside the rifts is matched by a model with an initial crustal thickness of 34 km. In each of the above models the plume lies at 50-100 km depth and has an initial temperature of 1500 ˚C.

Figure 3. The sensitivity of the model to b) the thickness of the crust, and the variation in the flood basalt thickness, c) the temperature of the plume head and d) effect of the depth and thickness of the plume head. In each case only one parameter is varied while the rest are kept as a standard model, shown in a) and described in the text. It represents initial set up of the model and does not represent the best fitting model. The gray area on each graph encapsulates the backstripped subsidence curves shown in Fig. 2 from Saunders et al., (2005).

Figure 1
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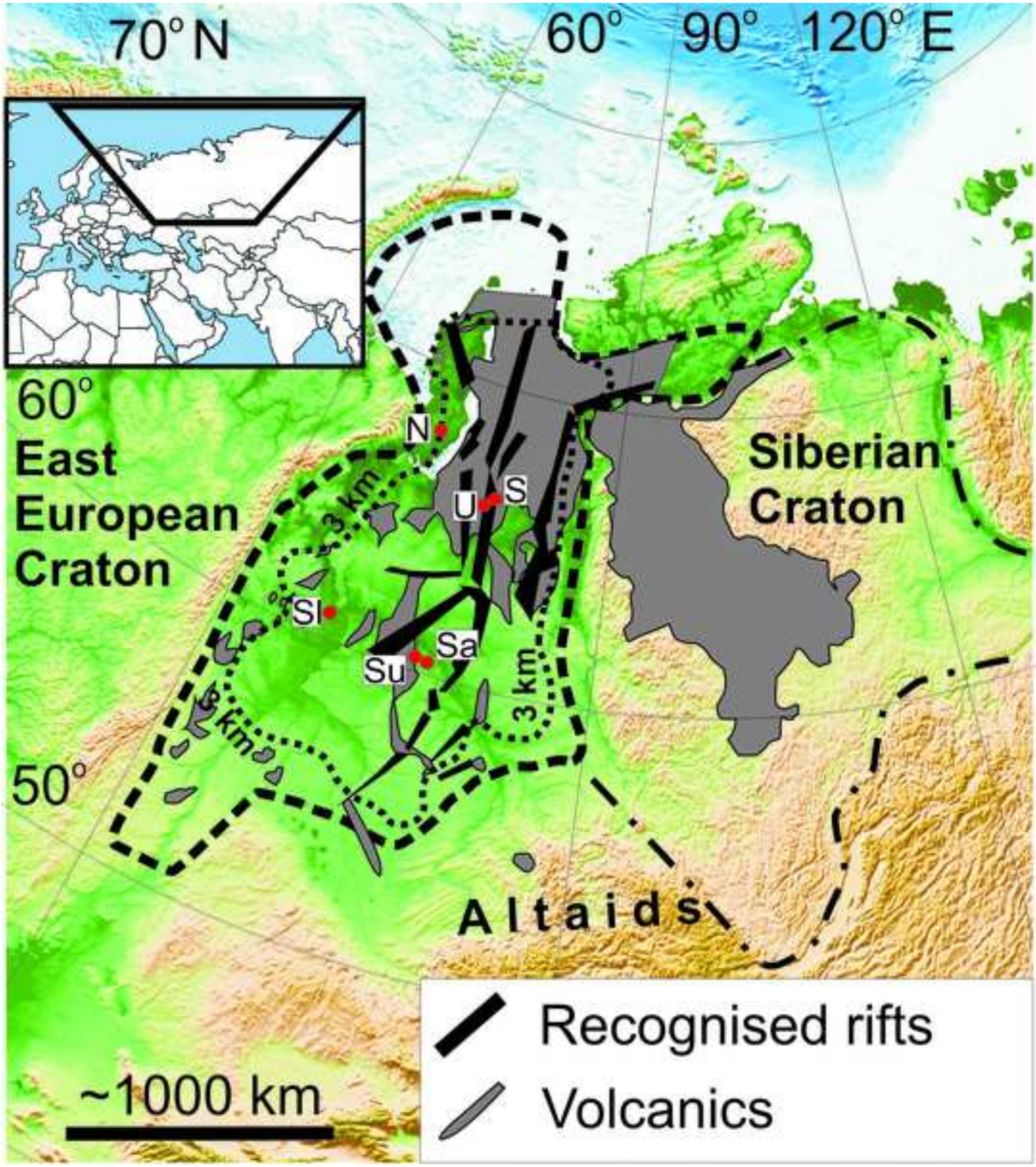

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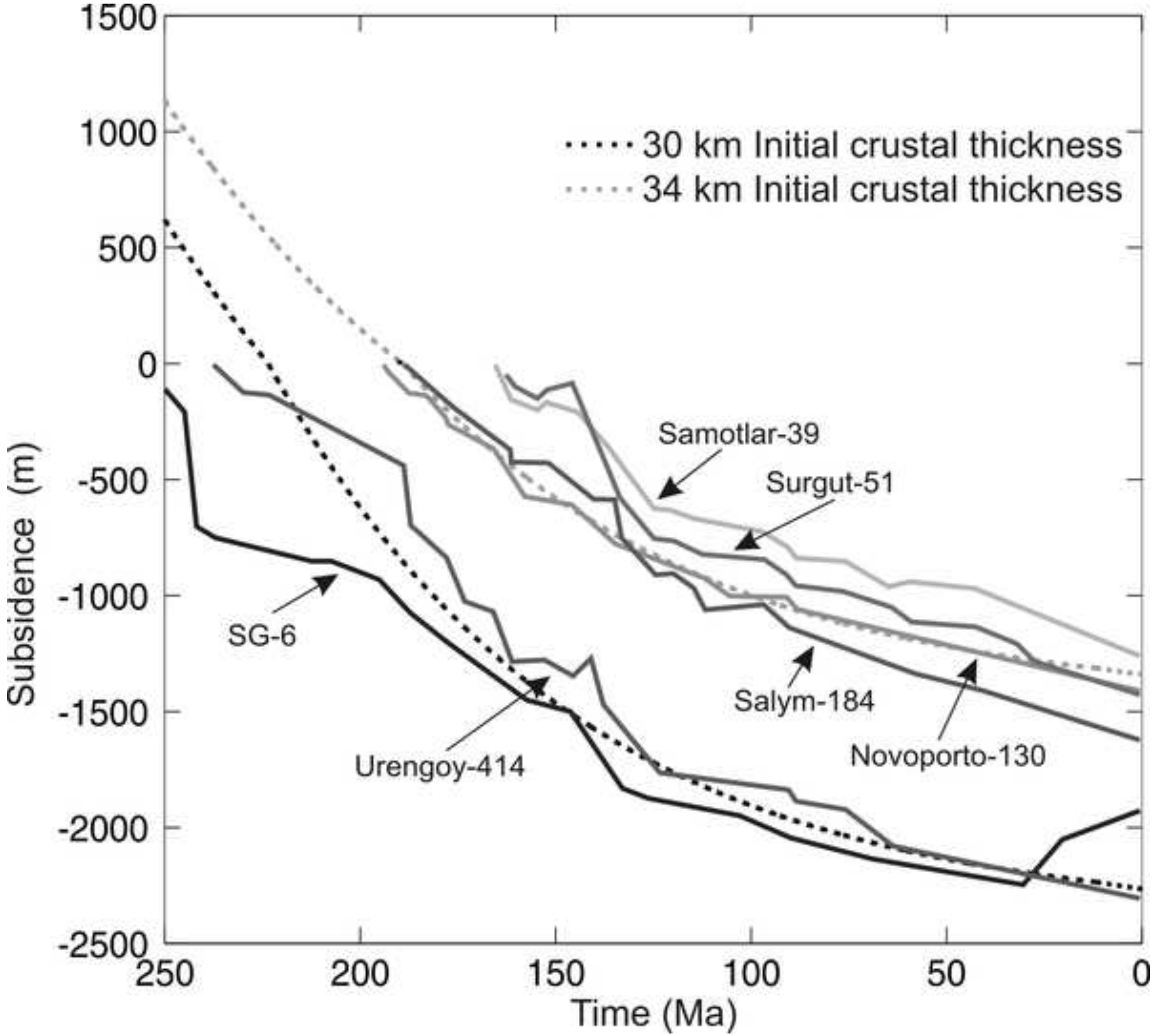

Figure 3

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