Tectonophysics xxx (2008) xxx-xxx

Contents lists available at ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

Styles of lithospheric extension controlled by underplated mafic bodies

Tadashi Yamasaki [a,b,*], Laurent Gernigon [c]

[a] Dublin Institute for Advanced Studies, 5 Merrion Square, Dublin 2, Ireland
[b] Department of Tectonics, Faculty of Earth and Life Sciences, VU University, Amsterdam, De Boelelaan 1085, 1081 HV Amsterdam, Netherlands
[c] Norges Geologiske Undersøkelse (NGU/Geological Survey of Norway), leiv, Eirikssons vei 39, N-7491 Trondheim, Norway

A R T I C L E I N F O

Article history:
Received 30 August 2007
Received in revised form 2 April 2008
Accepted 22 April 2008
Available online xxxx

Keywords:
Localization of deformation
Underplated mafic bodies
Lithospheric rheology
Continental rift
Passive margin
Continental break-up

A B S T R A C T

The role of underplated mafic bodies (UPMB) in the localization of deformation is examined using a two-dimensional thermo-mechanical finite element model. Rheological heterogeneity brought about by the UPMB is linked with the following two main physical effects: the UPMB material is assumed to have (1) an anomalous high temperature and (2) a mafic crustal rock composition that is intrinsically weaker than the mantle. The thermal effect will disappear rather quickly, but the rock composition can have an effect at any stage of extension. We show that the UPMB has a strong influence on the style of lithospheric extension, which depends on the thermal condition of the uppermost mantle at the time when the UPMB is emplaced. Since the strength contrast between the weakened and non-weakened regions is the most important factor, the UPMB works more efficiently on the localization of deformation for a colder uppermost mantle. Dependence of the localization of deformation on the temperature and thickness of the UPMB (i.e. more significant localization for the UPMB with higher temperature and greater thickness) also depends on the thermal condition of the uppermost mantle.
However, the width of the UPMB has a strong influence on the localization for any thermal condition of the uppermost mantle; a greater amount of thinning is distributed into a narrower weakened region. Such a model behaviour implies that various styles of lithospheric extension, including inward or outward migration of deformation and asymmetric extension, can be simply obtained by considering the emplacement of the UPMB, which also plays an important role in controlling the onset of continental break-up.

1. Introduction

Many continental rifts and passive margins display a high level of volcanic activity (see Fig. 1), before, during or shortly after extension (e.g., Ruppel, 1995; Neumann et al., 1995; White and McKenzie, 1989; Symonds et al., 1998; Gladczenko et al., 1997; Eldholm et al., 2000; Berndt et al., 2001; Menzies et al., 2002; Geoffroy, 2005; Mackenzie et al., 2005). The presence of high seismic velocity lower crust ($V_p \sim 7.1 \sim 7.8$ km/s) often observed along volcanic margins (Planke et al., 1991; Eldholm et al., 2000; Bauer et al., 2000; Mjelde et al., 2007) and magmatic rifts (e.g., Ro and Faleide, 1992; Mohr, 1992) is referred to as mantle-derived mafic intrusions, or so-called magmatic underplating (Cox, 1980; Mutter et al., 1984) or magmatic inflation (Thompson and McCarthy, 1990). Such underplated mafic bodies (UPMB) have been discussed in relation to (i) the high grade metamorphism of lower crustal material (e.g., Rey, 1993), (ii) thickening of the crust (Gans, 1987; Mohr, 1992), (iii) permanently higher post-rift topography (Lachenbrunch and Sass, 1978, 1990; Lister et al., 1991) and (iv) significant positive gravity anomalies (Hutchinson et al., 1990).

However, in most kinematic and dynamic models of rift formation and evolution, the effects of magmatism have been neglected (e.g. McKenzie 1978; Royden and Keen 1980; Wernicke, 1985;Braun and Beaumont 1987; Buck 1991; Davis and Kusznir, 2004). The generation of melt associated with rifting has been investigated for a given kinematic model of extension, but without considering effect of melts on the dynamics (e.g., McKenzie and Bickle, 1988; White and McKenzie, 1989; Latin and White, 1990; Harry et al., 1993; Harry and Leeman, 1995). As many prior studies have emphasized that most of the rifted margins are affected by magmatism (e.g., Gernigon et al., 2004; Geoffroy, 2005; Ebbing et al., 2006; Gernigon et al., 2006; Lizarralde et al., 2007), one of the most important issues to address is how magmatic activity influences the rifting dynamics.

The aim of the present study is principally to evaluate, using a two-dimensional thermo-mechanical finite element model, the possible influence of UPMBs on localizing extensional deformation. The understanding of the development of continental rifts and passive margins requires knowledge of a process to explain how the deformation is localized. Continental rifts formed by large-scale extensional tectonics are characterized by locally thinned crust and high surface heat flows (e.g., Artemjev and Artyushkov, 1971; McKenzie, 1978). Continental

* Corresponding author. Department of Tectonics, Faculty of Earth and Life Sciences, VU University, Amsterdam, De Boelelaan 1085, 1081 HV Amsterdam, Netherlands. Tel.: +31 20 5983921; fax: +31 20 5989943.
E-mail address: tadashi.yamasaki@falw.vu.nl (T. Yamasaki).

0040-1951/$ – see front matter
doi:10.1016/j.tecto.2008.04.024

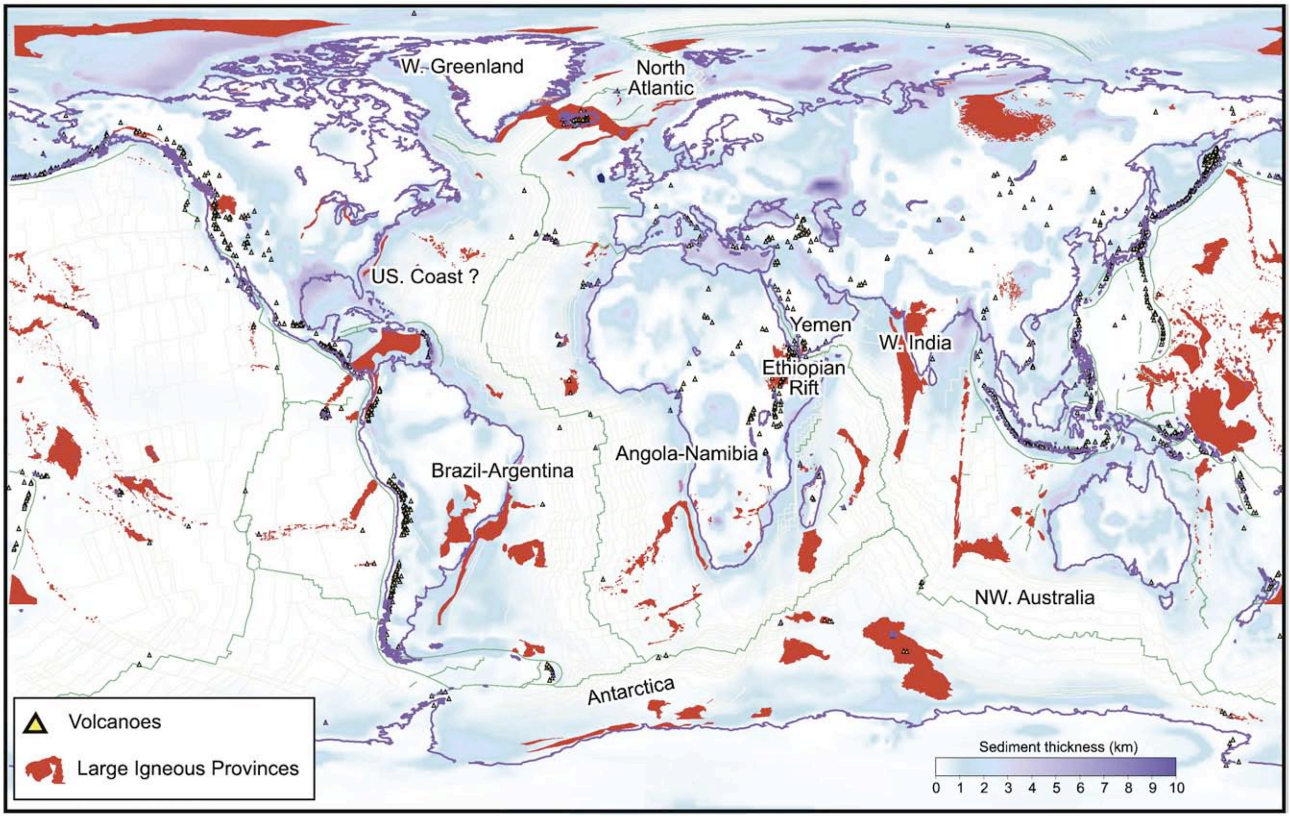

Fig. 1. Distribution of sedimentary basins, large Igneous Provinces, the main volcanic rifts and volcanic rifted margins in the world. This map illustrates the importance of the magmatism associated with sedimentary basins, continental margin and rift systems. Data after Coffin and Eldholm (1994) and Laske and Masters (1997). Isochrons are after Müller et al. (1997).

break-up is an end product of continental extension, and the break-up point usually focuses and concentrates within a local area (e.g., Ebinger and Casey, 2001). Thus, the problem which we also have to consider is what controls the localization of deformation. In this study, we focus on the UPMB as a possible origin of rheological heterogeneity, and attempt to investigate specifically a correlation between magmatism and rifting dynamics from the point of view of the localization of deformation. As well as the rheological heterogeneities brought about by a locally thickened crust (e.g., Braun and Beaumont, 1987; Bassi, 1991; Bassi et al., 1993; Govers and Wortel, 1993; Bassi, 1995; Govers and Wortel, 1995), temperature variation in the lithosphere (e.g., Govers and Wortel, 1999; Pascal et al., 2002) or pre-existing faults (e.g., Braun and Beaumont, 1989; Dunbar and Sawyer, 1988; 1989a,b; Melosh and Williams, 1989; Williams and Richardson, 1991), the presence of UPMBs in the continental lithosphere is likely to modify the thermal structure and mechanical properties (e.g., Frey et al., 1998; Buck, 2006), which could then control the evolution of the rift system. Analogue modelling studies by Callot et al. (2001, 2002) have previously demonstrated that localization can be brought about by low viscosity bodies in the lower crust and/or the uppermost mantle, interpreted as plume-related partial melts. Corti et al. (2003a) and Corti et al. (2007) also carried out analogue modelling experiments to investigate effects of melts on the progression of continental rifting into localized break-up. However, even though one of the most important physical mechanisms in such soft point models depends on temperature, it is difficult to include a temperature-dependent rheology in analogue modelling. Therefore, it is important to investigate the role of the UPMB in relation to its thermal aspects using numerical modelling (e.g., Gac and Geoffroy, submitted for publication).

This study particularly considers in detail the likely rheological heterogeneity brought about by the UPMB. We first discuss the role of the UPMB as the initial rheological heterogeneity, while respecting the initial crustal and thermal structure of the lithosphere. Since magmatic activity possibly precedes rifting (e.g., Burke and Dewey, 1973; Sengör and Burke, 1978; Griffiths and Campbell, 1991; Menzies et al., 2002; Courtillot and Renne, 2003), the UPMB could work as the initial rheological heterogeneity. In fact, Buck (2006) emphasized the importance of magmatic intrusion as a possible means of reducing the entire lithospheric strength to initiate the rifting with a reasonable tectonic force.

The basic characteristics of the UPMB, i.e. its width, thickness and temperature are specifically examined in this study, as well as the relative importance of the UPMB to a locally thickened crust for the localization of deformation. Whether the lateral variation in the thickness of the UPMB has the potential to bring about the asymmetric structures found in rift and passive margins is also examined in this study. We lastly demonstrate the possible localization of deformation that can be brought about by the UPMB during the extensional process.

2. Model description

The two-dimensional finite element model is schematically shown in Fig. 2. We examine the response of the lithosphere to an applied constant boundary velocity at the edge of the model. The UPMB is introduced as a material with an anomalous high temperature and mafic lower crustal composition. The mechanical problem is coupled with the thermal problem through temperature-dependent viscosity. The plane strain condition, which seems appropriate for the geological structure of a rift, is assumed in this study. The lithosphere has three compositional layers: wet quartzite upper crust, anorthite lower crust, and wet dunite mantle.

The origin of the force driving the extension is not considered. Even if the lithosphere has a uniform rheological structure localization of deformation can be obtained by a non-uniform driving force. The rising plume may control a location where deformation could be concentrated (e.g., Sengör and Burke, 1978; Burov and Guillou-Frottier, 2005; Burov et al., 2007). Additionally, in the laboratory analogue model of Tirel et al. (2006), localization of deformation is obtained by a

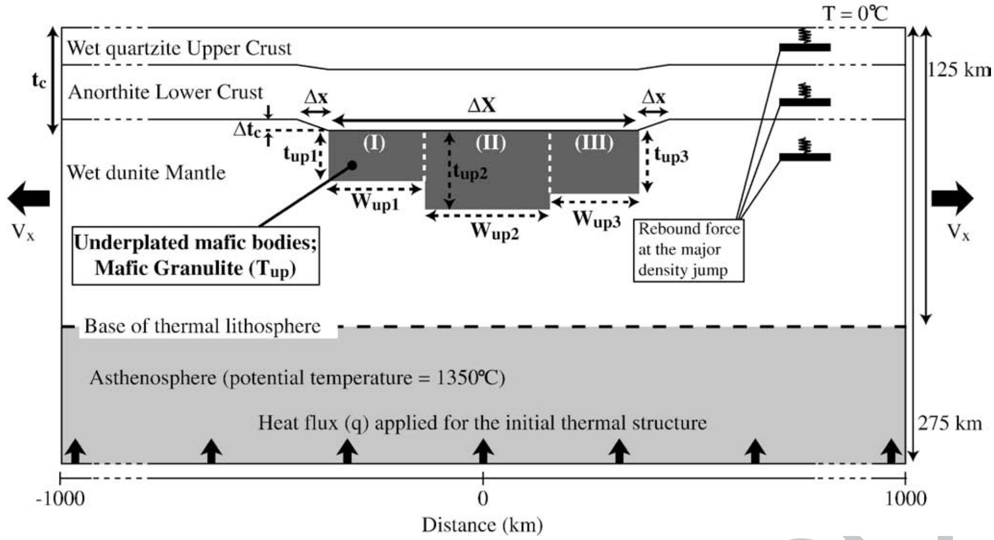

Fig. 2. Schematic figure of the numerical model adopted in this study. The continental lithosphere has three compositional layers; wet quartzite upper crust, anorthite lower crust and wet dunite mantle. The initial thickness of the crust is t_c, and the thickness of the lower crust is $t_c/2$. A constant horizontal boundary velocity (V_x=1 cm/year) is applied at the right and left side ends of the model. The Winkler restoring force is applied at the following major density interfaces so as to calculate the vertical surface movements: at the surface, the boundary between upper and lower crust and at the Moho. The thermal boundary condition of the model is as follows. The horizontal heat flow is zero at the right and left side boundaries of the model. The temperature of the surface is fixed to be 0 °C. In order to obtain the steady state solution of Eq. (5) the reduced heat flow is applied at the bottom of the model, satisfying the condition that the temperature at the base of a given thermal lithosphere (at a depth of 125 km) is equal to the potential temperature of the asthenosphere (1350 °C). At depths below the thermal lithosphere the temperature is set to be the potential temperature of the asthenosphere. In the transient thermal problem, the temperature below the depth of 125 km is imposed to be 1350 °C as a constant temperature boundary condition. Three domains of the UPMB (I~III) with abnormally high temperature and mafic lower crustal composition are characterized by W_{up1} and t_{up1}, W_{up2} and t_{up2}, and W_{up3} and t_{up3}, respectively. All domains of the UPMB are assumed to have uniform temperature Tup. For the model with a locally thickened crust, the crust is thickened by ⊗t_c over a distance ⊗X in the centre of the model. The transition from thickened to normal crust takes place over a distance of ⊗x, and the Moho depth linearly decreases from t_c to t_c−⊗t_c (km). The initial topographic height resulting from the isostatic response to the thickened crust is not taken into account in the setup of the model.

velocity discontinuity between the surrounding rigid plates and the plate in which deformation is taking place. In the present study, however, localization of deformation is particularly discussed in relation to the rheological heterogeneity brought about by the UPMB, so that we just apply a constant velocity (V_x=1 cm/yr) at the left and right side boundaries of the model.

Since the formation process of UPMBs is still poorly understood, we simply assume that the UPMB is emplaced instantaneously, as is adopted in Frey et al. (1998), without considering how the melt is generated and how it migrates upwards. Additionally, even though both the timing of the emplacement and volume of the UPMB are important factors in controlling magma related heterogeneities (e.g., Chapman and Furlong, 1992), it is still difficult to exactly estimate the timing of the magmatism and the quantity of generated melt prior to and/or during extension. For example, as shown in Hirth and Kohlstedt (1996), the process of decompressional melting cannot be investigated without information of the initial water content in the mantle peridotite. Therefore we make simple assumptions about the timing and distribution of the UPMB.

The UPMB is described by three basic parameters, its width (W_{up}), thickness (t_{up}) and temperature (T_{up}). Each of three domains of the UPMB (labelled I, II and III in Fig. 2) are characterized by W_{up1} and t_{up1}, W_{up2} and t_{up2}, and W_{up3} and t_{up3}, respectively. All the domains are assumed to have a uniform temperature T_{up}.

In this study, the UPMB is regarded as the primal origin of rheological heterogeneity for localization of deformation. However, the rheological heterogeneity brought about by a locally thickened crust is also a plausible origin of the heterogeneity, as adopted in many prior studies (e.g., Braun and Beaumont, 1987; Bassi, 1991; Bassi et al., 1993; Govers and Wortel, 1993; Bassi, 1995; Govers and Wortel, 1995). Therefore, we perform several models with a locally thickened crust in the Section 3.4 in order to investigate the condition where the UPMB could play a dominant role in localizing the deformation. For the model with a locally thickened crust, the continental lithosphere has a crustal thickness t_c locally thickened by ⊗t_c over a distance ⊗X in the centre of the model (see Fig. 2). The transition from thickened to normal crust takes place over a distance of ⊗x, over which the Moho depth linearly decreases from t_c to t_c−⊗t_c. The initial topographic height resulting from the isostatic response to the thickened crust is not taken into account in the setup of the model. A finite element code, tekton ver. 2.1, developed by Melosh and Raefsky (1980, 1981) and Melosh and Williams (1989), is used to solve the mechanical equilibrium equation

$$\nabla \cdot \sigma + X{-}0 \qquad (1)$$

where $\boldsymbol{\sigma}$ is the stress tensor and X is the body force. Large strain and displacement effects are incorporated by using an updated Lagrangian method. The vertical movements of the major density interfaces, i.e. the surface, the boundary between upper and lower crust, and the Moho, are calculated by using the Winkler restoring force (e.g., Williams and Richardson, 1991).The constitutive relationships that relate the stress to the strain are required to solve the mechanical equilibrium equation. The viscoelastic–perfect plastic rheology model with temperature-dependent viscosity has been incorporated into the code (Yamasaki et al., submitted for publication). Under low deviatoric stress conditions, rock deformation is elastic, in which the strain is linearly related to the stress by the generalized Hooke's law (e.g., Ranalli, 1995). However, under larger deviatoric stress conditions, different deformation mechanisms become more important, depending on the temperature and pressure conditions.

Under low temperature and pressure conditions, deformation of rocks takes place in a brittle manner. The brittle stress σ_b is given by Byerlee's law (Byerlee, 1967, 1978):

$$\sigma_b = \varsigma\left(1-^{*}\right)z \qquad (2)$$

where z is the depth in km, ζ is a constant (24 MPa/km) and υ^* (0.4) is the density ratio of pore water to rock matrix (e.g., Brace and Kohlstedt, 1980; Ranalli, 1995). The elastic–perfect plastic rheology is adopted for the brittle deformation (e.g., Flecher and Hallet, 1983;

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

Braun and Beaumont, 1987; Bassi, 1995). The state of the stress is assumed to be controlled by the Von Mises yield criterion F,

$$F = J_{2D} - \frac{1}{3}\sigma_y^2 = 0 \tag{3}$$

where J_{2D} is the second invariant of deviatoric stress and σ_y is the yield stress that is equivalent to σ_b in Eq. (2).

On the other hand, under high temperature and pressure conditions, ductile deformation is the dominant mechanism. A non-linear Maxwell viscoelastic rheology is adopted for the ductile behaviour of rocks (e.g., Melosh and Raefsky, 1980; Braun and Beaumont, 1987; Bassi, 1995; Govers and Wortel, 1995). The viscous flow is controlled by power law creep (e.g., Kirby, 1983; Carter and Tsenn, 1987), where the effective viscosity η can be written in the form

$$\eta = \frac{1}{2A^*} J_{2D}^{\frac{1-n}{2}} \exp\left(\frac{Q}{R\Theta}\right) \tag{4}$$

where A^* is a material constant, Q is the activation energy, n is the stress exponent, R is the universal gas constant and Θ is the absolute temperature. Flow law parameters of each rock composition are shown in Table 1.

Since power law creep is a function of temperature, the distribution of temperature in the lithosphere must be calculated at all stages of deformation. The time-dependent heat conduction equation is given by

$$\rho C_p \frac{\partial T}{\partial t} = \nabla \cdot (K \nabla T) + H \tag{5}$$

where ρ is the density, C_p is the specific heat, T is the temperature, t is the time, K is the thermal conductivity and H is the radiogenic heat

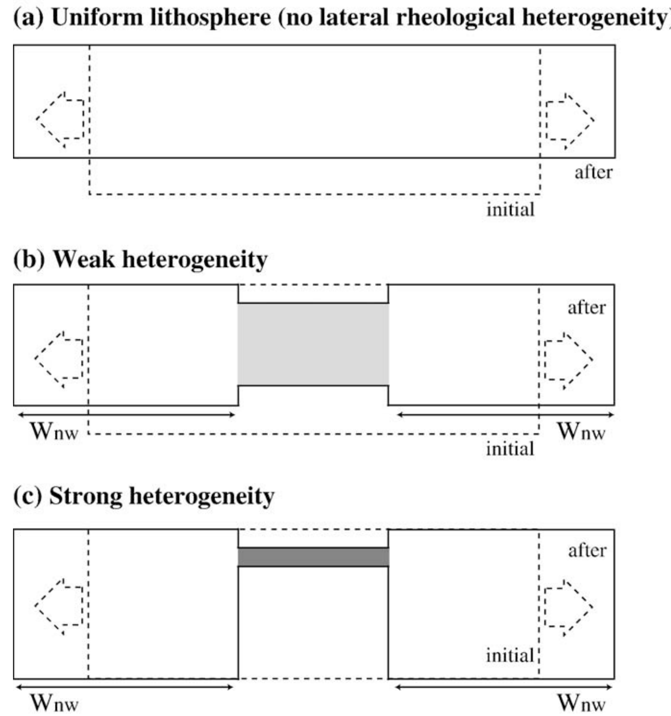

Fig. 3. Localization of extensional deformation; (a) extension of the uniform lithosphere with no lateral rheological heterogeneity, (b) extension of the lithosphere with weak rheological heterogeneity, and (c) extension of the lithosphere with strong rheological heterogeneity.

Table 1
Model parameter values used in this study

Symbol	Meaning	Value
E	Young's modulus	1.5×10^{11} Pa
ν	Poisson's ratio	0.25
ζ	Depth dependence of brittle failure	24.0 MPa/km
υ^*	Density ratio of pore water to rock	0.4
C_p	Specific heat	1050 $\mathrm{Jkg^{-1}K^{-1}}$
R	Universal gas constant	8.314 $\mathrm{Jmol^{-1}K^{-1}}$
T_a	Potential temperature of the asthenospher	1350 °C
T_s	Temperature at the surface	0 °C
ρ_{uc}	Mass density of the upper crust	2800 $\mathrm{kgm^{-3}}$
ρ_{lc}	Mass density of the lower crust	2900 $\mathrm{kgm^{-3}}$
ρ_m	Mass density of the mantle	3300 $\mathrm{kgm^{-3}}$
H_{uc}	Heat production in the upper crust	1.37 $\mathrm{\mu Wm^{-3}}$
H_{lc}	Heat production in the lower crust	0.45 $\mathrm{\mu Wm^{-3}}$
H_m	Heat production in the mantle	0.02 $\mathrm{\mu Wm^{-3}}$
K_{uc}	Thermal conductivity of the upper crust	2.56 $\mathrm{Wm^{-1}\,K^{-1}}$
K_{lc}	Thermal conductivity of the lower crust	2.60 $\mathrm{Wm^{-1}\,K^{-1}}$
K_m	Thermal conductivity of the mantle	3.20 $\mathrm{Wm^{-1}\,K^{-1}}$
Flow law parameters of power law creep		
Wet quartzite: Koch et al. (1989)		
A^*_{uc}	Pre-exponent	1.10000×10^{-21} $\mathrm{Pa\text{-}ns^{-1}}$
n_{uc}	Power	2.61
Q_{uc}	Activation energy	145 $\mathrm{kJ\ mol^{-1}}$
Anorthite: Shelton and Tullis (1981)		
A^*_{lc}	Pre-exponent	5.60000×10^{-23} $\mathrm{Pa\text{-}ns^{-1}}$
n_{lc}	Power	3.20
Q_{lc}	Activation energy	238 $\mathrm{kJ\ mol^{-1}}$
Wet Aheim dunite: Chopra and Paterson (1984)		
A^*_m	Preexponent	5.4954×10^{-25} $\mathrm{Pa\text{-}ns^{-1}}$
n_m	Power	4.48
Q_m	Activation energy	498 $\mathrm{kJ\ mol^{-1}}$
Mafic granulite: Ranalli (1995)		
A^*_{upmb}	Preexponent	8.8334×10^{-22} $\mathrm{Pa\text{-}ns^{-1}}$
n_{upmb}	Power	4.2
Q_{upmb}	Activation energy	445 $\mathrm{kJ\ mol^{-1}}$

source. We ignore the effect of the latent heat associated with the solidification of melts. Parameter values used in this study are shown in Table 1. We have written an additional finite element code to solve the transient heat conduction equation (Eq. (5)). Overlapping thermal and mechanical finite element grids are used to solve the heat and mechanical equilibrium equations sequentially. Advection of heat is incorporated in a way that the finite element geometry is updated at every time step with the displacements computed by the mechanical algorithm. The thermal boundary conditions are as follows. At the left and right side boundaries of the model the horizontal heat flow is zero. The temperature at the surface is held at 0 °C. The potential temperature of the asthenosphere is assumed to be 1350 °C. The initial thermal structure is calculated from the steady-state solution of Eq. (5). The reduced heat flow is applied to satisfy the condition that the temperature at the base of the thermal lithosphere (at a depth of 125 km) is equal to the potential temperature of the asthenosphere (i.e. the thickness of the thermal lithosphere a is defined by the 1350 °C isotherm). Below a depth of 125 km, the temperature is assumed to be identical to the asthenospheric potential temperature. In the transient thermal problem, the temperature at a depth greater than 125 km is imposed to be 1350 °C as a constant temperature boundary condition.

3. Results and discussion

Extensional deformation of the lithosphere results in its thinning following the conservation of mass. If a lithosphere that has no lateral rheological heterogeneities is extended by an applied boundary velocity, thinning of the lithosphere will occur uniformly (see Fig. 3(a)). However, if lateral rheological heterogeneity is present in the lithosphere, the thinning will be distributed, particularly into a weak region (e.g., Fernàndez and Ranalli, 1997). In this case, the strength contrast of the heterogeneity is an important factor in controlling how much of the thinning is distributed into the weak region. If the strength contrast is

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

weak, there will be a small amount of thinning in the weak region because extension is also significantly accommodated by thinning in the non-weakened region (Fig. 3(b)). On the other hand, for a stronger strength contrast, extensional deformation is mostly accommodated by thinning the weak region (see Fig. 3(c)). It should be noted that in the numerical model the strength contrast is, in practice, recognized by the difference in the strain rate. Higher strain rate is obtained in the weakened region, which results in the localization of deformation.

We have confirmed that the least amount of a thinning of the lithosphere cannot be avoided in the non-weakened region even for the model with significant strength contrast. Extension applied at the edge of the model is also accommodated by a thinning of the non-weakened region, and the total amount of extension accommodated in the non-weakened region increases as W_{nw} (width of non-weakened region) increases. That is, the amount of thinning distributed into the weakened region becomes smaller for the model with larger W_{nw}. Therefore, the absolute amount of thinning in a significantly deformed region can be evaluated only for a specific size of model. The most important point to be investigated is the sensitivity to an assumed heterogeneity.

3.1. Dependence on the initial lithospheric structure

Rheological heterogeneity brought about by the UPMB is linked with the following two physical effects: (1) the higher temperature of the UPMB results in weakening through the temperature-dependence of the viscosity and (2) the mafic crustal composition of the UPMB is significantly weaker than the mantle, even without any thermal effects.

Fig. 4(a) and (b) shows the initial temperature distributions for the models with and without the UPMB, with t_c=30 and 40 km, respectively. The temperature at the Moho is higher for the model with t_c=40 km than for the model with t_c=30 km. Therefore, the difference in temperature between the UPMB and normal mantle is larger for the model with t_c=30 km than for the model with t_c=40 km.

This implies that the UPMB could be more effective in localizing the deformation for the model with t_c=30 km. When t_c is 30 km, the

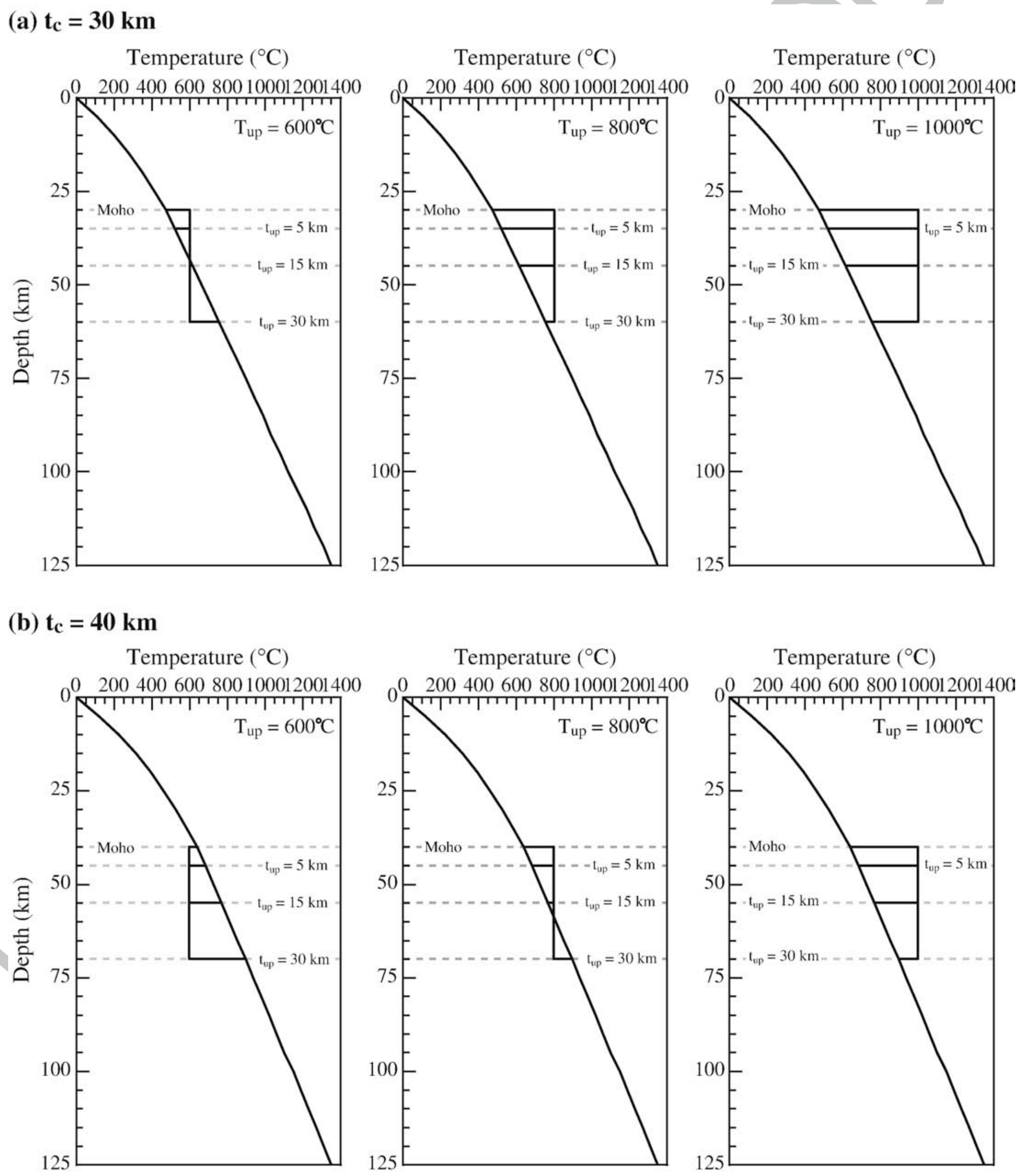

Fig. 4. Temperature distributions for the models with and without an UPMB that has an abnormal temperature T_{up} (T_{up}=600, 800 and 1000 °C) and a thickness of t_{up} (t_{up}=5, 15 and 30 km); (a) t_c=30 km and (b) t_c=40 km.

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

temperature of the UPMB for the model with T_{up}=800 and 1000 °C is greater than that of the normal mantle from the Moho to the base of the UPMB. For the model with T_{up}=600 °C, the temperature of the UPMB is higher than that of normal mantle to a depth of about 45 km, but lower at a greater depth. When t_c is 40 km, with T_{up}=600 °C the temperature of the UPMB is never higher than that of the uppermost

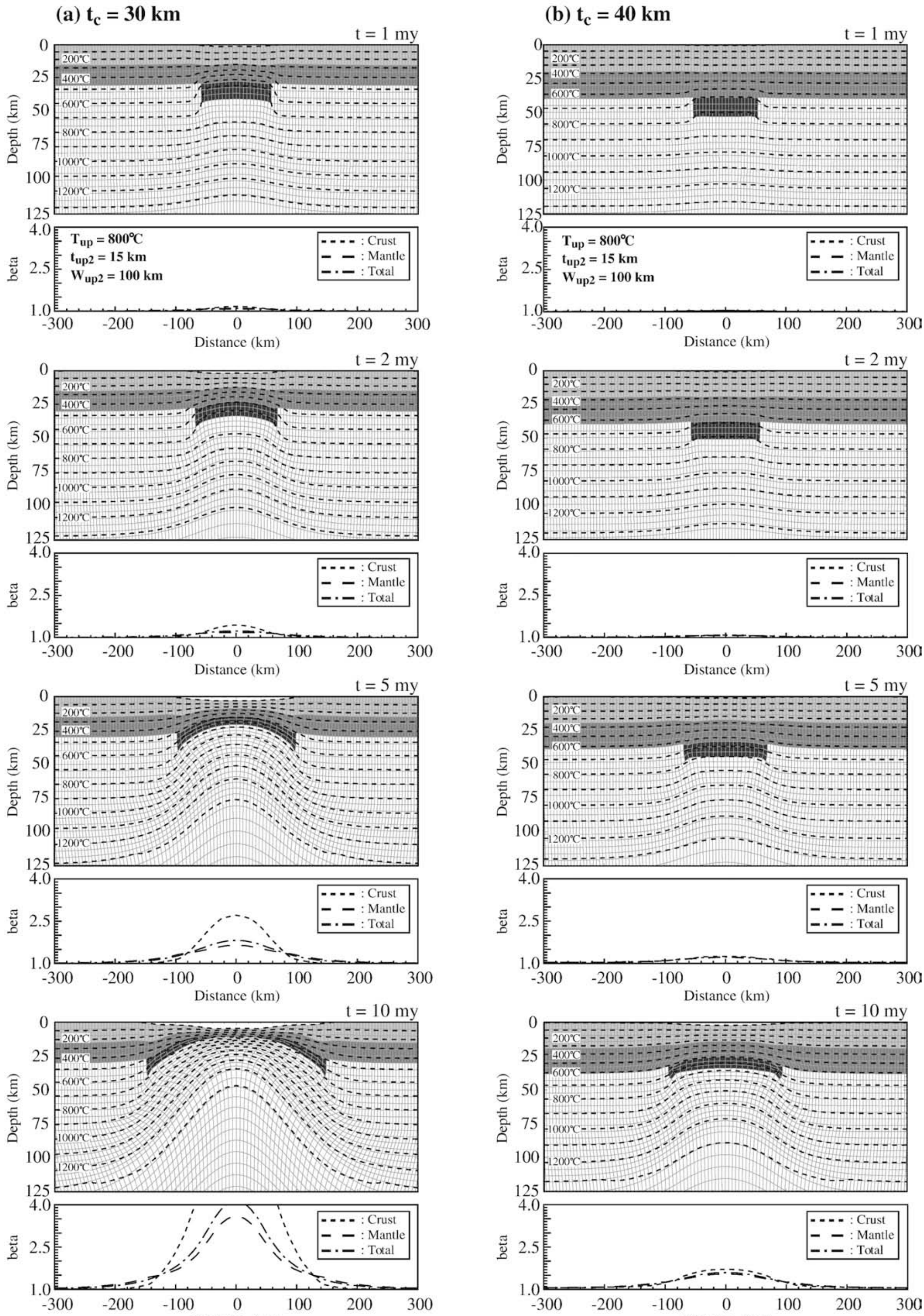

Fig. 5. Time-dependent deformed grids with superimposed temperature contours for the UPMB models with ⊗t_c=0 km, W_{up1}=W_{up3}=0 km, W_{up2}=100 km, t_{up2}=15 km and T_{up2}=800 °C; (a) t_c=30 km and (b) t_c=40 km. Contours of temperature are shown at 100 °C intervals from 0 to 1300 °C. Stretching factors of the crust (dotted curve), mantle (dashed curve) and total (dotted and dashed curve) are depicted for each time. The stretching factor is defined by the ratio of the thickness before extension to the thickness after extension.

mantle for the whole range of t_{up}. Even for the model with T_{up} = 800 °C, the temperature is higher only to a depth of about 55 km. However, for the model with T_{up} = 1000 °C, the UPMB results in a higher temperature anomaly over the whole depth range where the UPMB is situated.

Fig. 5(a) and (b) shows the extensional deformation of the lithosphere with increasing time for the models with t_c = 30 and 40 km, respectively, in which only the centre domain (II) is used to describe the characteristic of the UPMB (i.e. $W_{up1}=W_{up3}=0$). A locally thickened crust prior to extension is not taken into account ($\otimes t_c=0$ km). It is assumed that T_{up} is 800 °C, W_{up2} is 100 km and t_{up2} = 15 km. It can be clearly seen that a greater amount of thinning is distributed into a weakened region for the model with t_c = 30 km than for the model with t_c = 40 km. The total stretching factor at time t = 10 my reaches up to about 4.0 for the model with t_c = 30 km, but to only about 1.6 for the model with t_c = 40 km. A greater strength reduction through the increase in temperature is obtained by the UPMB for the model with t_c = 30 km because of the smaller initial geothermal gradient in the normal mantle (see Fig. 4).

It should be also noted that the temperature anomaly caused by the UPMB disappears within a few million years. The period that the UPMB can keep a temperature higher than the normal mantle may practically depend on the size of the UPMB and the time-scale of its formation process (e.g., Chapman and Furlong, 1992): the period could be shorter for a smaller UPMB and/or for a slower formation of the UPMB. On the other hand, another effect relating to the difference in composition between the UPMB and mantle could work on localizing deformation at any stage of extension, because this effect has no relation to temperature.

Although only the results of the model with a = 125 km are shown, the UPMB model is applicable for any thicknesses of the thermal lithosphere. Since the localization of the deformation brought about by the UPMB is dependent on the degree of strength reduction, it is clear that the UPMB works more effectively for the colder lithosphere than for the hotter lithosphere.

3.2. The basic model parameters describing the characteristics of the UPMB

Fig. 6 shows how the stretching factor varies as a function of distance for models with t_c = 30 km and 40 km, in which only the domain II is used to describe the characteristic of the UPMB. A locally thickened crust prior to extension is not taken into account ($\otimes t_c=0$ km). The times are 6 and 10 my for the models with t_c = 30 and 40 km, respectively. Fig. 6(a) shows the dependence of extensional deformation on T_{up}, with W_{up2} = 100 km and t_{up2} = 15 km. Three models are shown for T_{up} = 600, 800 and 1000 °C. The dependence on T_{up} is not seen in the model with t_c = 30 km. Even when the temperature of the UPMB is 600 °C, the temperature anomaly necessary to produce a strength contrast strong enough to distribute the thinning mostly into a weak zone is obtained below the Moho (see Fig. 4(a)). On the other hand, for the model with t_c = 40 km, the localization of the deformation is dependent on T_{up}; a higher UPMB temperature leads to a higher degree of thinning distributed into a weak region, because the strength contrast between weakened and non-weakened regions is larger for higher UPMB temperature. It is also important to note that extensional deformation can be localized even for the model with t_c = 40 km and T_{up} = 600 °C, i.e. the temperature of the UPMB is lower than that of the normal mantle (Fig. 4(b)). This is because the mafic crustal material is intrinsically weaker than the mantle without any thermal effects.

Fig. 6(b) shows the dependence on t_{up2}, with W_{up2} = 100 km and T_{up} = 800 °C, for models with t_{up2} = 5, 15 and 30 km. When t_c is 40 km, a greater amount of thinning is distributed into a weak zone for the models with greater t_{up2}. A greater reduction in the total lithospheric

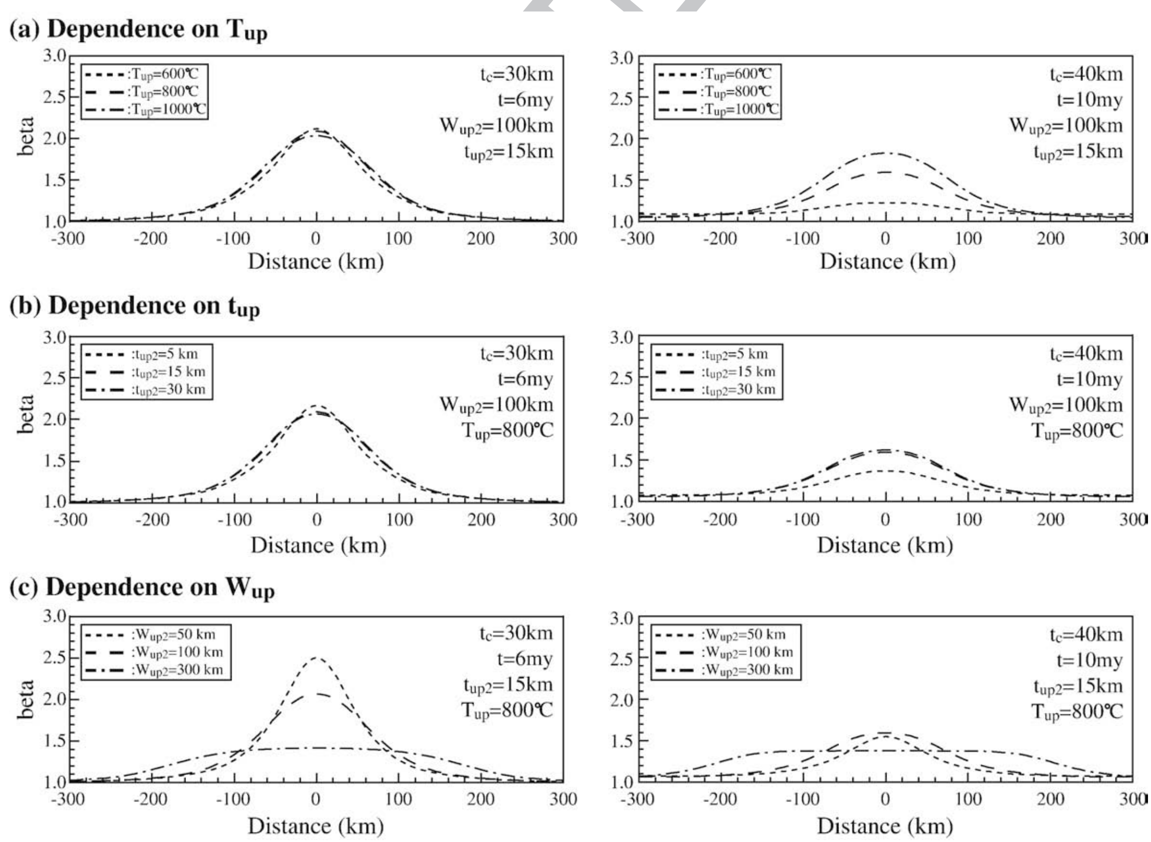

Fig. 6. Total stretching factors as a function of a distance for different (a) T_{up}, (b) t_{up2} and (c) W_{up2}, in which $\otimes t_c=0$ km and $W_{up1}=W_{up3}=0$ km. Times are 10 and 6 my for the models with t_c = 30 and 40 km, respectively.

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

strength is obtained by the model with greater t_{up2}. This is consistent with the results from analogue modelling by Corti et al. (2007).

However, the temperature of the UPMB becomes lower than the mantle temperature at a depth greater than ~57 km (see Fig. 4(b)), so that at a further depth it becomes difficult to obtain the significant strength reduction by the temperature anomaly of the UPMB. In fact, models with $t_{up2} \geq 15$ km do not further enhance the localization of the deformation, in which the stretching factor predicted by the model with $t_{up2} = 15$ km is almost the same as that with $t_{up2} = 30$ km. On the other hand, when t_c is 30 km, the model behaviour is virtually independent of t_{up2}. Even for the model with $t_{up2} = 5$ km the strength contrast is strong enough to distribute the thinning mostly into a weak region.

Fig. 6(c) shows the dependence on W_{up2}, with $t_{up2} = 15$ km and $T_{up} = 800$ °C. The results of the models with $W_{up2} = 50$, 100 and 300 km are depicted in the figures. As Corti et al. (2007) demonstrated previously using an analogue model, our numerical model also shows a strong dependence on W_{up2} for both $t_c = 30$ and 40 km. A greater amount of thinning is distributed into a weak region for the models with smaller W_{up2}. For a given amount of extension at the boundary of the model, the extensional deformation of the lithosphere is accommodated by the distribution of thinning into a narrower region for the model with smaller W_{up2}, which leads to a higher strain rate in deformed region.

3.3. Localization of deformation by slow strain rate or UPMB in the lower crust: outward migration of deformation

As described above, the distribution of thinning into a weak region is controlled by the rheological heterogeneity introduced as a result of the presence of the UPMB. Our model shows that distribution of thinning into a weak zone is larger for the models with stronger strength contrast and with a narrower UPMB. Consequently, the strain rate in the significantly deformed zone is controlled by the characteristics of the UPMB, i.e. even if the applied boundary velocity is fixed, various strain rates can be obtained.

It is important to discuss the effect of strain rate on the extensional deformation in relation to thermal diffusion. The strength of the lithosphere changes as extension progresses, as a result of the competition between the weakening caused by the increase in geothermal gradient associated with the passive upwelling of the asthenosphere and the strengthening by crustal thinning and thermal diffusion. For a smaller strain rate, the strengthening by thermal diffusion becomes the more dominant factor (e.g., England, 1983).

Fig. 7(a) shows the result of the model with an UPMB in which $W_{up1} = W_{up3} = 0$, $W_{up2} = 300$ km, $t_{up2} = 30$ km and $T_{up} = 800$ °C. A locally thickened crust prior to extension is not taken into account ($\otimes t_c = 0$ km). It can be seen that lithospheric thinning moves outward in the later rifting phase. Similar numerical model behaviour has been reported by van Wijk and Cloetingh (2002) and Corti et al. (2003b), in which outward migration of the thinning was obtained by a model with a low extensional velocity or a very wide thickened crust. Such model behaviour is attributed to the thermal diffusion in a deformed region with a slow strain rate. For the model with $W_{up2} = 300$ km, extensional deformation is accommodated by a distribution of thinning into a wide weak region, so that the obtained strain rate is low and the effect of thermal diffusion plays an important role in the style of extension. On the other hand, the strain rate in the deformed region for the models with $W_{up2} \leq 150$ km is so high (see Figs. 5 and 6)

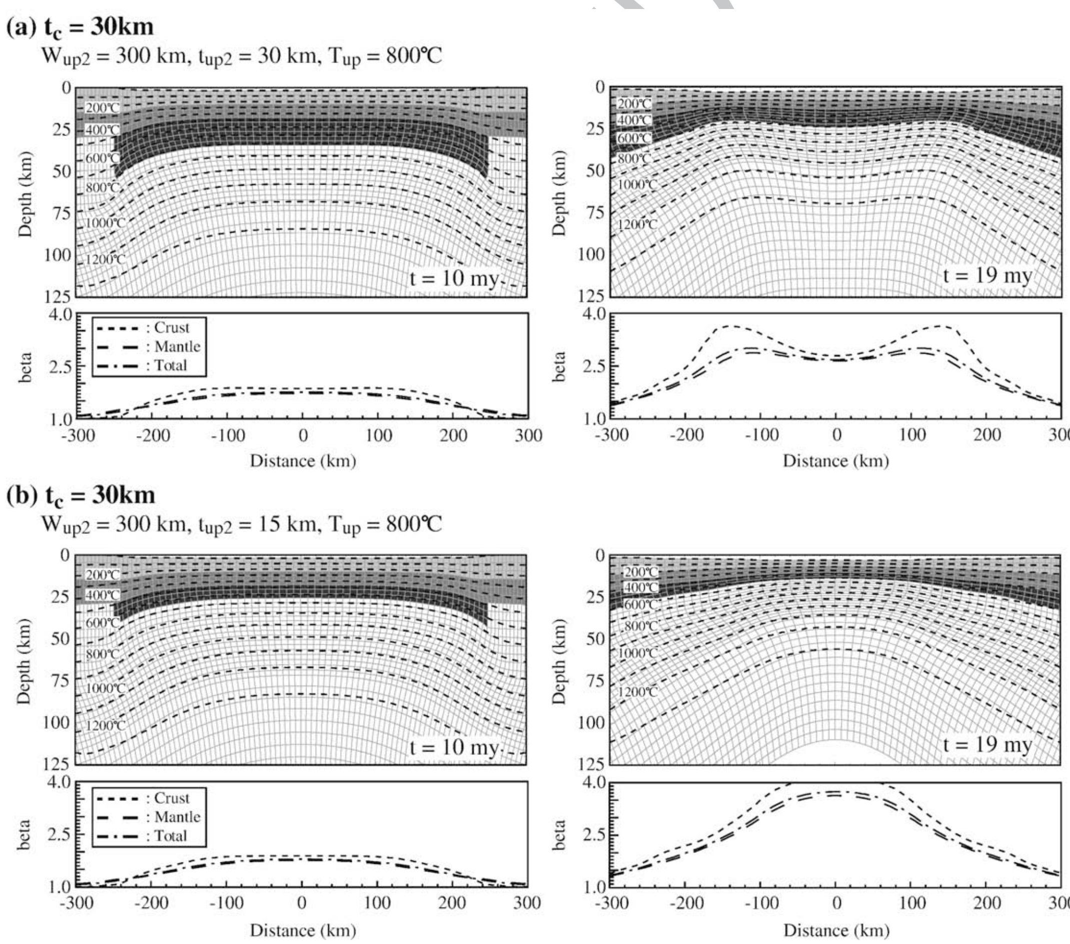

Fig. 7. Deformed grids with superimposed temperature contours for the model with $t_c = 30$ km, $\otimes t_c = 0$ km, $W_{up1} = W_{up3} = 0$ km, $W_{up2} = 300$ km and $T_{up} = 800$ °C at $t = 10$ and 19 my; (a) $t_{up2} = 30$ km and (b) $t_{up2} = 15$ km. Stretching factors of the crust (dotted curve), mantle (dashed curve) and total (dotted and dashed curve) are depicted for each model.

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

that it is difficult to obtain outward migration by thermal diffusion. Fig. 7(b) shows the result of setting t_{up2} = 15 km. Outward deformation migration is not obtained by this model. This result indicates that the replacement of the mafic lower crustal material (the UPMB) with mantle material is also an important factor in controlling the change in the total lithospheric strength. The replacement is smaller for the model with smaller t_{up2}. Therefore, even though the thermal diffusion progresses significantly, the centre of the rift does not become strong enough to obtain outward migration of deformation.

Prior numerical modelling studies have successfully explained an outward migration of deformation in terms of thermal diffusion in the deformed region (e.g., Negredo et al., 1995; van Wijk and Cloetingh, 2002; Corti et al., 2003b). However, if the UPMB is taken into account, it is generally difficult to obtain migration of the deformation to outside the UPMB region, because the UPMB has intrinsically less strength than the mantle without considering the thermal effects. Even if outward migration of deformation can be obtained by the model with the UPMB, its migration is limited to the region where the UPMB is present (see Fig. 7(a)). Therefore, in order to shift deformation to outside the UPMB region, there may be no other way than by considering a shift of magmatic activity, such as following a change in plume activity (Müller et al., 2001; Yamada and Nakada, 2006).

Only the UPMB in the uppermost mantle just below the Moho has been examined in this study. However it is likely that melts could intrude into the lower crust (e.g., Bonini et al., 2001; Corti et al., 2003a). We give an important implication for the case of the UPMB in the lower crust. As described above, the UPMB has the two physical properties important for the localization of deformation: it has an abnormally high temperature and an intrinsically weaker composition than the mantle. Because mafic granulite has a greater strength than anorthite lower crust (e.g., Ranalli, 1995; Ranalli et al., 2007), the compositional effect does not work if the UPMB is situated in the lower crust. Once the temperature anomaly caused by the UPMB in the lower crust is significantly relaxed by thermal diffusion, deformation would be localized into outside the UPMB where the felsic lower crust is intrinsically weaker than the UPMB. In the prior analogue modelling studies on extensional deformation of the lithosphere in relation to the magmatic process (Callot et al., 2001; 2002; Corti et al., 2003a; Corti et al., 2003b; Corti et al., 2007), the origin of the localization of deformation is brought about by a low viscosity body not only in the mantle but also in the lower crust. However it might be questionable to keep such a low viscosity body in the lower crust over the entire duration of rifting.

3.4. Relative importance of the UPMB to a locally thickened crust

Fig. 8 shows a relative importance of the UPMB to a locally thickened crust in which $\otimes X$ and $\otimes x$ are 100 and 25 km, respectively. Only the domain II with t_{up2} = 15 km and T_{up} = 800 °C is used to describe the characteristic of the UPMB. The time is 6 my for all models. Fig. 8(a) shows the results with t_c = 30 km and $\otimes t_c$ = 5 km. For the model without the UPMB, the total stretching factor is predicted to be about 2.2, at most. On the other hand, for a UPMB with W_{up2} = 50 km, the stretching factor at the centre of the rift reaches 2.6, which is significantly larger than that for the model without the UPMB. However, an UPMB model with W_{up2} = 100 km predicts an almost identical stretching factor to that predicted by the model without the UPMB (~2.2). In addition, the stretching factor at the centre decreases to 1.8 for the model with W_{up2} = 300 km.

Fig. 8(b) shows the results for the model with t_c = 30 km and $\otimes t_c$ = 2.5 km. The predicted stretching factor at the centre for the model without the UPMB is smaller than that even for the UPMB model with W_{up2} = 300 km. This is because a difference in t_c of 2.5 km does not result in a strength contrast strong enough to distribute thinning mostly into a region weakened by a locally thickened crust.

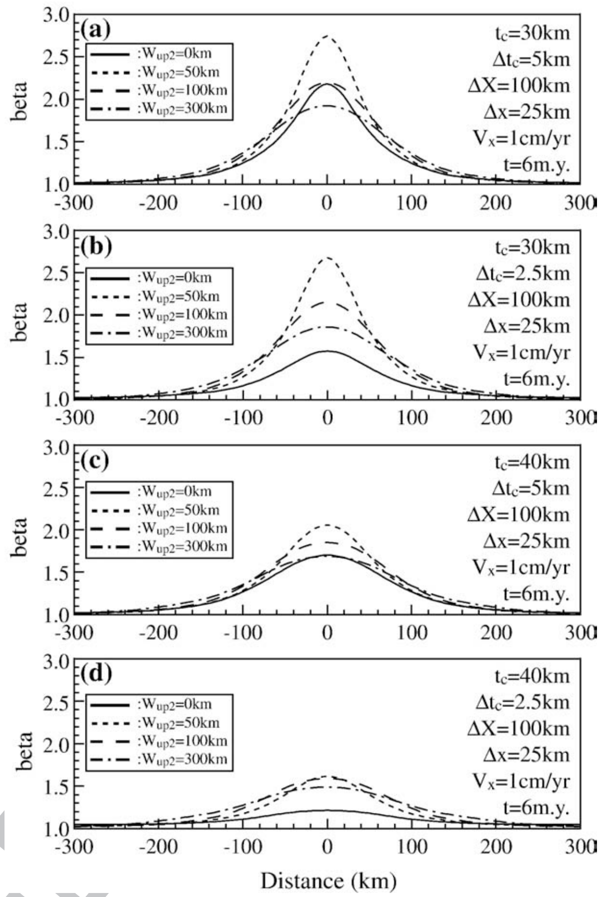

Fig. 8. Total stretching factors as a function of a distance for different widths (W_{up2}) of an UPMB, at t = 6 my, with $\otimes X$ = 100 km and $\otimes x$ = 25 km; (a) t_c = 30 km and $\otimes t_c$ = 5 km, (b) t_c = 30 km and $\otimes t_c$ = 2.5 km, (c) t_c = 40 km and $\otimes t_c$ = 5 km, and (d) t_c = 40 km and $\otimes t_c$ = 2.5 km. It is assumed that W_{up1} = W_{up3} = 0 km, t_{up2} = 15 km and T_{up} = 800 °C.

The obtainable strength contrast for the model with t_c = 40 km is weaker than for the model with t_c = 30 km, as shown in the previous subsection. The same is true for the rheological heterogeneity by a locally thickened crust. Therefore, for a given amount of extension, the predicted stretching factor (the amount of thinning distributed into a weak region) for the model with t_c = 40 km is significantly smaller than that predicted for the model with t_c = 30 km (see Fig. 8(c) and (d)). For the model with t_c = 40 km, even when $\otimes t_c$ is assumed to be 5 km (see Fig. 8(c)), the stretching factor for the model without the UPMB is predicted to be only the same as that for the UPMB model with W_{up2} = 300 km. When $\otimes t_c$ is assumed to be 2.5 km (see Fig. 8(d)), the predicted stretching factor without the UPMB is significantly smaller than that obtained with the UPMB, which is practically independent of W_{up2} (at least for W_{up2} between 50 and 300 km).

Thus, if the lateral scale of the UPMB is smaller than that of a locally thickened crust, the UPMB plays a dominant role in localization of deformation. However, if the crustal thickness variation is small, the UPMB still plays a dominant role in localization of deformation which is independent on W_{up2}.

3.5. Inhomogeneous thickness of the UPMB: implication for asymmetric structure of continental rifts and passive margins

Fig. 9 shows the results of the models where the UPMB has a variable thickness. Only two domains of the UPMB are considered (i.e., W_{up2} = 0 km). T_{up} is assumed to be 800 °C for the both domains. A locally thickened crust prior to extension is not taken into account ($\otimes t_c$ = 0 km). For the models with W_{up1} = W_{up3} = 50 km, t_{up1} = 5 km and t_{up3} = 15 km

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

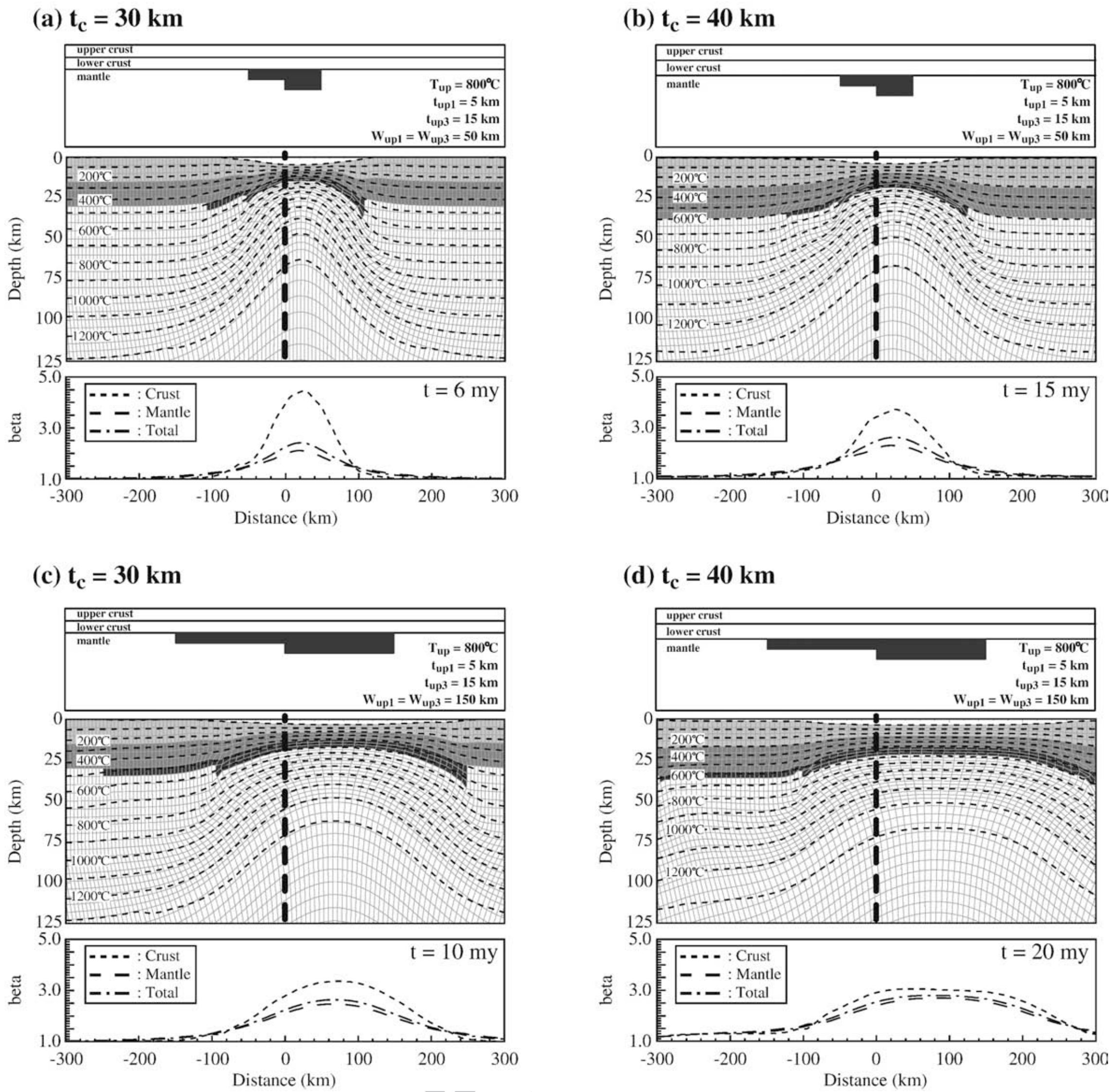

Fig. 9. Deformed grids with superimposed temperature contours for the model with T_{up}=800 °C, ⊗t_c=0 km, W_{up2}=0 km, t_{up1}=5 km and t_{up3}=15 km; (a) t_c=30 km and $W_{up1}=W_{up3}$=50 km; (b) t_c=40 km and $W_{up1}=W_{up3}$=50 km; (c) t_c=30 km and $W_{up1}=W_{up3}$=150 km, and (d) t_c=40 km and $W_{up1}=W_{up3}$=150 km. The vertical dotted line indicates the centre of the model. Stretching factors of the crust (dotted curve), mantle (dashed curve) and total (dotted and dashed curve) are shown for each model.

(see Fig. 9(a) and (b)), a greater amount of lithospheric thinning can be seen to be distributed into domain III (where there is a greater thickness of UPMB). The axis of the significantly deformed region is shifted from the centre of the model, producing an asymmetric structure about the centre of the model. However, the structure seems mostly symmetric about the axis of the significantly deformed region (although slight asymmetry is perceptible for the model with t_c=40 km).

For the models with $W_{up1}=W_{up3}$=150 km, the axis of the deformed region is shifted even more from the centre of the model (see Fig. 9(c) and (d)). For the model with t_c=30 km, the predicted structure still seems to be mostly symmetric about the axis of the significantly deformed region. The strength contrast between domains I and III is large enough to distribute a thinning into the domain III. On the other hand, for the model with t_c=40 km, a slight asymmetry about the axis of the significantly deformed region can be seen. Here the strength contrast between the two domains of the UPMB is not strong enough, so that a significant amount of thinning is distributed even into the domain I. The same physical explanation can be applied to the models with $W_{up1}=W_{up3}$=50 km. However, the lateral scale of the UPMB is too small for the asymmetric structure to be clearly perceptible (see Fig. 9(a) and (b)).

Fig. 10 shows results where three domains of the UPMB are taken into account with different thicknesses; t_{up1}=5 km, t_{up2}=30 km and t_{up3}=15 km. T_{up} is assumed to be 800 °C for all the domains. A locally thickened crust prior to extension is not taken into account (⊗t_c=0 km). Fig. 10(a) and (b) shows the results of the models with $W_{up1}=W_{up2}=W_{up3}$=50 km. The maximum amount of thinning is now distributed into domain II, and the stretching factor becomes maximum in the centre of the model. The predicted structure is asymmetric about the centre of the model and also the axis of most deformed region.

The results of the models with W_{up1}=120 km, W_{up2}=60 km and W_{up3}=120 km are shown in Fig. 10(c) and (d). The stretching factor is again maximum at the centre of the model and is clearly asymmetric even about the centre of the model. Although the greatest amount of the thinning is distributed into domain II where the UPMB has the greatest thickness (=30 km), a significant thinning also spreads into domain III where t_{up2} is 15 km. The strength contrast between domain II and III is not large enough to localize deformation only into domain II, so that there is significant deformation even in domain III (especially for the model with t_c=40 km). For the model with t_c=30 km, significant thinning is not distributed into domain I, where

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

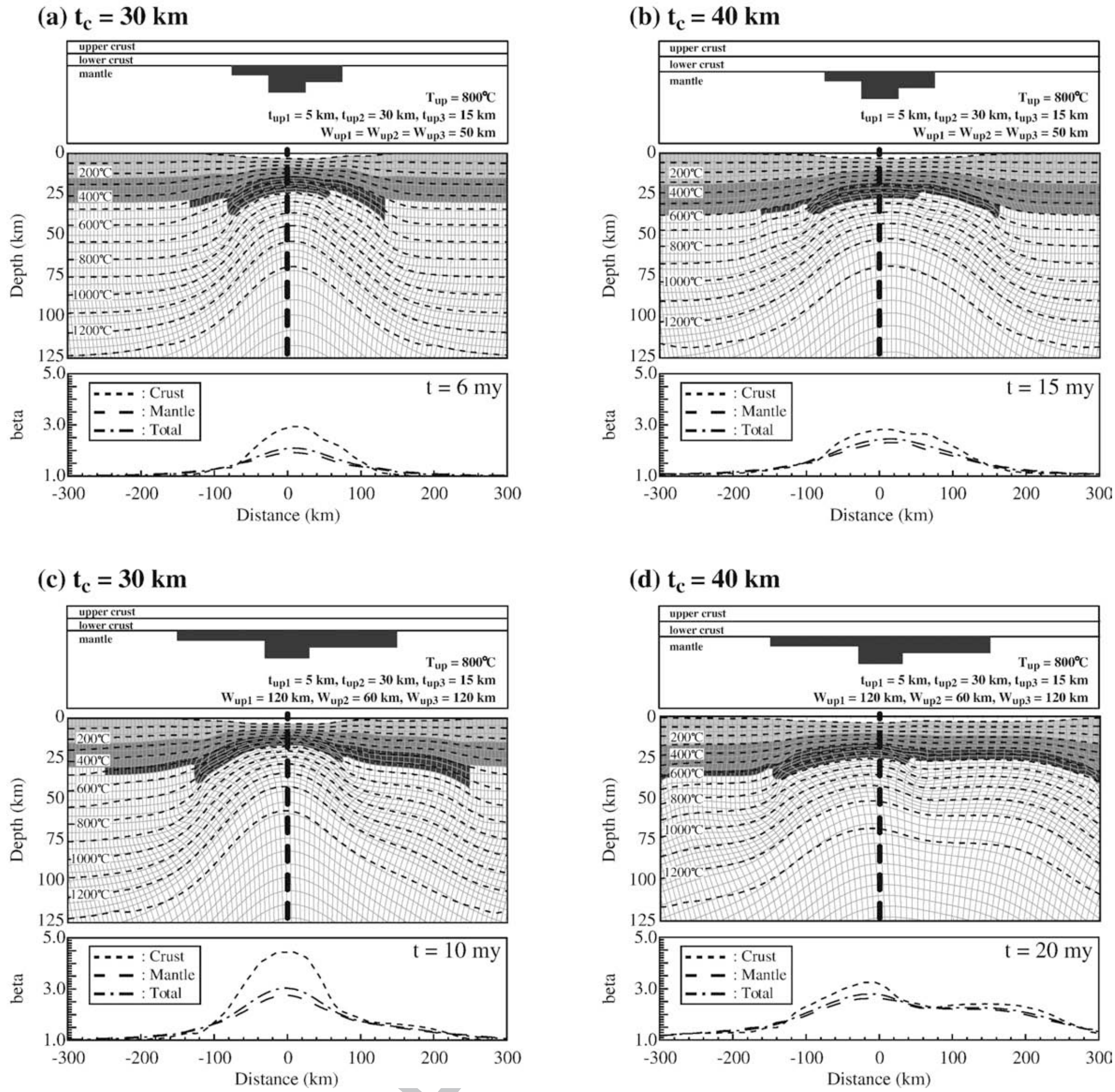

Fig. 10. Deformed grids with superimposed temperature contours for the model with T_{up}=800 °C, $\otimes t_c$=0 km, t_{up1}=5 km, t_{up2}=30 km and t_{up3}=15 km; (a) t_c=30 km and W_{up1}=W_{up2}=W_{up3}=50 km, (b) t_c=40 km and W_{up1}=W_{up2}=W_{up3}=50 km, (c) t_c=30 km, W_{up2}=60 km and W_{up1}=W_{up3}=120 km, and (d) t_c=40 km and W_{up2}=60 km and W_{up1}=W_{up3}=120 km. The vertical dotted line indicates the centre of the model. Stretching factors of the crust (dotted curve), mantle (dashed curve) and total (dotted and dashed curve) are shown for each model.

the UPMB has only a thickness of 5 km. The strength contrast between domain I and II is large enough to localize deformation into domain II. On the other hand, for the model with t_c=40 km, the strength contrast between the three domains of the UPMB is not high enough for significant thinning to be distributed, even into domain I.

In previous dynamic models, large-scale lithospheric deformation has been usually discussed in relation to the rheological change during the extensional process without considering the importance of the initial heterogeneities. However, Dunbar and Sawyer (1988, 1989a,b) proposed the important insight that the structure of the initial heterogeneity offers the key to understanding the development of continental extension. Corti et al. (2003b) and Corti and Manetti (2006) attempted to confirm this concept by using analogue and numerical models, in which the asymmetric geometry of passive margins, wide non-volcanic margins and narrow volcanic margins, are explained by an initial asymmetric structure of a locally thickened crust. In fact, clearly the asymmetric structure cannot be obtained by the symmetric initial setup of the model.

Similarly, as shown in this study, the asymmetric structure brought about by the inhomogeneous thickness of the UPMB might explain wide non-volcanic and narrow volcanic margins, in which a greater amount of deformation focuses into the region where the UPMB has a greater thickness.

However, this kind of asymmetry is basically obtained by the shift of the significantly deformed region from the centre of the UPMB, so that it is essentially required that the UPMB have its greatest thickness at a point that is significantly distant from the centre of the UPMB (see Fig. 9). In fact, if the greatest thickness of the UPMB is located at the centre of the UPMB, it becomes difficult to explain the characteristics of the wide and narrow passive margins (see Fig. 10).

Additionally, even if the UPMB has a variation in thickness, the structure of the rift seems to be mostly symmetric about the axis of the significantly deformed region for a small lateral scale of UPMB.

Therefore, the UPMB may have to range over a region wide enough to obtain a perceptible asymmetry.

It is also important to note that the asymmetric structure brought about by the inhomogeneous thickness of the UPMB can be clearly obtained when the strength contrast between each domain of the UPMB is weak enough not to distribute deformation into a particular domain.

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

Although the asymmetric structure of continental rifts and passive margins has been usually explained in terms of the simple shear model (Wernicke, 1985) or the combined pure and simple shear models (e.g., Kusznir and Egan, 1989; Lister and Davis, 1989; Lister et al., 1991), it is not necessary for the asymmetry to be accompanied by a large-scale shear zone in the lithosphere (e.g., Corti et al., 2003b; Corti and Manetti, 2006). Our numerical models also explain the asymmetric structure of extended lithosphere without the development of a shear zone cutting the lithosphere. The presence of a shear zone might be a consequence of asymmetry, rather than its origin. However, it may still be impossible that the initial asymmetric heterogeneity is sufficient to obtain the shear zone. Other sorts of rheological weakening during the extensional process may be required to develop a large-scale shear zone in the lithosphere (e.g., Frederiksen and Braun, 2001; Huismans and Beaumont, 2002; Yamasaki, 2004; Yamasaki et al., 2006).

3.6. The UPMB during extension: from inward deformation migration to continental break-up

Fig. 11 shows localization of deformation brought about by the UPMB during the extensional process. A locally thickened crust is adopted so as to obtain an initial localization of deformation, where t_c is 40 km, $\otimes X$ is 300 km, $\otimes x$ is 25 km and $\otimes t_c$ is 5 km. Only the domain II is used to describe the characteristic of the UPMB. It is assumed that the UPMB with t_{up2} = 15 km, W_{up2} = 50 km and T_{up} = 800 °C is emplaced at time t=5 my. Extensional deformation prior to the onset of the UPMB is obviously controlled by the initial heterogeneity brought about by the locally thickened crust, and deformation takes place over a wide broad region. However, once the UPMB is emplaced, the thinning is apparently distributed into a narrower region, following rheological weakening by the UPMB. The supply of magma decreases the lithospheric strength and localizes the deformation into the magmatic zone.

In the Ethiopian rift, where the development of continental rifting into sea-floor spreading is being demonstrated, inward deformation migration is observed. Showing that border faults are inactivated prior to continental break-up and most of the strain across the rift is accommodated in the magmatic segments, Ebinger and Casey (2001) has suggested that there is migration of extensional deformation into narrower magmatic segments in the rift. Then, Keranen et al. (2004) and Maguire et al. (2006) supported the view that extensional deformation in the Ethiopian rift is strongly controlled by magmatic activity, based on a crustal seismic velocity model combined with geological studies where the distribution of faults and dike deformation zones were concentrated above high seismic velocity bodies, interpreted as intra-crustal and/or underplated intrusions. Keir et al. (2006) also pointed out a good correlation between the distributions of seismicities and magmatic activities, emphasizing an interaction between tectonic processes and magmatic activity. Similar characteristics of such an interaction between magmatism and tectonics can also be seen in Afar (e.g., Hayward and Ebinger, 1996; Manighetti et al., 1998; Doubre et al., 2007a,b), in the North Atlantic margin (e.g., Gernigon et al., 2004; Geoffroy et al., 2001; Gernigon et al., 2006), in the northwest coast of Scotland (Speight et al., 1982) and in the Ligurian Tethys (Corti et al., 2007; Ranalli et al., 2007).

Our numerical results are consistent with these observations; inward migration of deformation can be explained by rheological heterogeneity brought about by the UPMB during the extensional

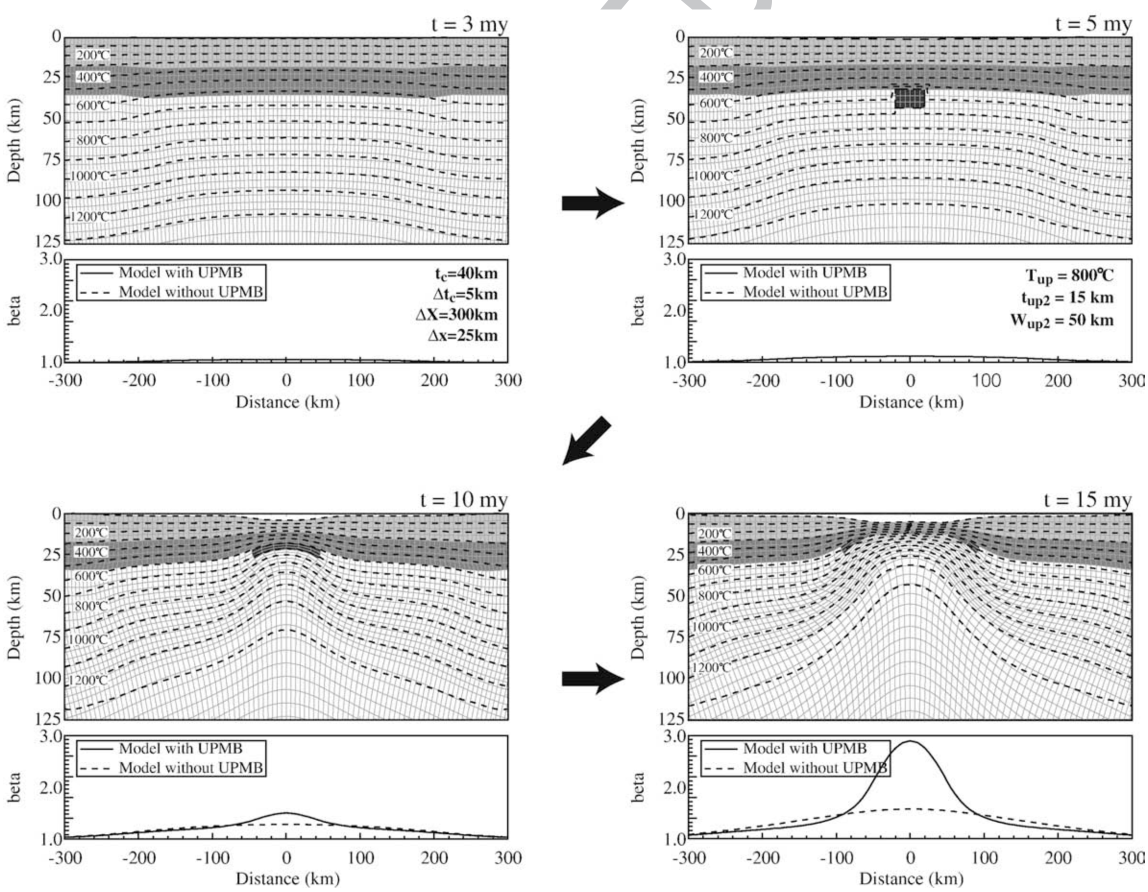

Fig. 11. Time-dependent extensional deformation of the lithosphere for the UPMB models with W_{up1} = W_{up3} = 0 km, W_{up2} = 50 km, t_{up2} = 15 km and T_{up} = 800 °C. For the initial crustal configuration, it is assumed that $\otimes X$ is 300 km, $\otimes x$ is 25 km and $\otimes t_c$ is 5 km. Emplacement of the UPMB is at t=5 my. Stretching factors of the total lithosphere for the models with and without the UPMB are depicted for each time.

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

processes. We have discussed localization of deformation brought about by the UPMB in the particular case where an UPMB is present prior to the onset of extension. However, the general model behaviour obtained in the previous sections is applicable even to localization of deformation during the extensional process. It has been shown in this study that the UPMB works more efficiently to localize a deformation for originally colder uppermost mantle. It is also well known that any material within the lithosphere suffers cooling as the lithosphere thins. Therefore, the magmatic activity in later phases of the extensional process (i.e. after a deformed region has cooled down significantly by thermal diffusion) is more appropriate to localize deformation. In this regard, the UPMB could work on significant localization of deformation even for an initially hot lithosphere if the UPMB is emplaced during the extensional process.

It is also important to note that a small amount of magmatic emplacement is more favourable to the initiation of the localization of deformation, rather than a huge amount. In this study, and also in Corti et al. (2007), it is shown that a narrower UPMB can localize deformation more effectively. Additionally, as discussed above, even a thin UPMB could work efficiently on localization of deformation if an extended lithosphere is significantly cooled by thermal diffusion. A large amount of the UPMB may blur out the localization of deformation, and the deformation may range over a wide broad region (see Figs. 6(c) and 7). The break-up triggered by a magmatic process, even that observed in the magma poor margin of the Newfoundland–Iberia rift system (Tucholke et al., 2007), may be good evidence to support the view that a small amount of the UPMB is enough to bring about significant localization of deformation. Once the additional rheological heterogeneity brought about by the UPMB occurs, the thinning of the lithosphere may be increased and local magmatic activity enhanced, which in turn successively induces subsequent localizations of deformation and magmatic activity into a more local region.

Such a localization process could explain a sudden transition from continental rifting to break-up and subsequent enhancement of major magmatic activity at the break-up point. In the Outer Vøring Basin, where inward migration of the deformation during the onset of magmatism has also been documented by Gernigon et al. (2004), Geoffroy et al. (2001) and Gernigon et al. (2006), seismic observations show that faulting migrated towards the proto-oceanic rift during the onset of major magmatism in the Paleocene, indicating that the inward migration of the deformation started before the major volcanic phase during the ultimate continental break-up during the Late Palaeocene–Early Eocene. The emplacement of a small amount of UPMB might bring about the onset of the deformation migration, and a consequent increase in lithospheric thinning at the break-up point could cause major volcanic activity there. It may be more reasonable to regard the major volcanic phase at the break-up point to be a result of the localization of the deformation, because a huge amount of magmatic emplacement prior to the onset of the deformation migration may prevent the formation of a localized break-up point.

Many studies have emphasized the importance of dike intrusions on surface extensional tectonics (e.g., Poliakov and Buck, 1998; Ebinger and Casey, 2001; Geoffroy et al., 2001; Klausen and Larsen, 2002; Keranen et al., 2004; Sigmundsson, 2006; Wright et al., 2006). However, the process of dike intrusion may have the potential only to affect the deformation of the brittle upper crust (e.g., Keranen et al., 2004). Additionally, as pointed out above, an UPMB in the crust can possibly bring about an outward migration of deformation, which from the point of view of its compositional characteristics, has more strength than the continental crust. Therefore, it may be difficult to regard the diking process in the upper crust as the main factor in controlling the break-up process. The intrusion of dikes may be, rather, a complementary and secondary product of a significantly localized deformation that is brought about by an UPMB in the uppermost mantle. Extensional tectonics controlled by the dike intrusion may indicate the latest stage of the break-up, not its initiation.

4. Conclusions

In this study, we have investigated the influence of the UPMB on the style of lithospheric extension so as to discuss a correlation between magmatism and rifting dynamics. The numerical modelling study has demonstrated that UPMBs play an important role in controlling the localization of deformation, depending on the initial crustal and thermal structure of the lithosphere, the basic characteristics of the UPMB, the geometry of the UPMB, and the timing of the emplacement of the UPMB. This study also provides important insights into the dynamics of rifting and development of continental rifting into sea-floor spreading. It should be noted that the general model behaviour has been discussed mostly in the particular case where an UPMB is present prior to the onset of extension. However, such a model behaviour must be applicable even to the localization of deformation during extension. That is, the initial lithospheric condition assumed in this study can be regarded as the condition at the time when the UPMB is emplaced. The model with an initially located UPMB evaluates only the minimum effect of the UPMB on localization of deformation. This is because each material point in the lithosphere being extended suffers cooling, and consequently the UPMB emplaced during extension leads to a greater decrease in strength. The limitation of our model is related to the assumption in which the UPMB is a solidified mafic lower crustal material, not a melt. Rheological property of a partially molten rock should be applied in a more realistic self-consistent model, which remains as a matter to be investigated in a future.

The main conclusions of the study are as follows:

(i) The role of UPMBs in localization of deformation is linked with the following two main physical effects: they have an anomalously high temperature and a mafic crustal rock composition that has less strength than the mantle. The former effect will disappear rather shortly after the emplacement of the UPMB, but the latter one could play an important role at any stage of extension.

(ii) The most important factor in controlling localization of deformation is the strength contrast between the weakened and non-weakened regions. The strength contrast brought about by the UPMB is strongly dependent on the thermal condition of the uppermost mantle at the time when the UPMB is emplaced; colder uppermost mantle results in a larger strength contrast. Thus, the UPMB will work more efficiently on the localization of deformation for colder uppermost mantle. Dependence on the characteristics of the UPMB is also dependent on the thermal condition of the uppermost mantle. However, the width of the UPMB has a strong influence on the localization of deformation for all thermal conditions of the uppermost mantle. Such model behaviour implies that even if the extensional velocity applied at the boundary of the model is constant, various strain rates in a significantly deformed region can be obtained, depending on the characteristics of the rheological heterogeneity brought about by the UPMB.

(iii) For the model with a very wide UPMB, the effective strain rate is predicted to be very slow, and outward migration of deformation can be obtained in relation to thermal diffusion during the extensional process. However, such a migration is still limited to the region where the UPMB is situated. The outward migration of the deformation is also dependent on the thickness of the UPMB; a greater thickness of the UPMB is preferred for the outward migration.

(iv) An UPMB in the lower crust could be an origin of a weak zone only when its temperature is significantly high. Once the UPMB loses the thermal anomaly, deformation would migrate to outside the UPMB. This is because a mafic lower crustal composition generally has more strength than a normal lower crustal composition without any thermal effects.

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

(v) The relative importance of the UPMB to a locally thickened crust depends on the lateral scale of heterogeneity; if the upward migration of melts is concentrated into a region narrower than the lateral scale of the crustal heterogeneity, localization of the extensional deformation is dominantly controlled by the UPMB. When the variation in thickness of crust is small, the UPMB still plays a dominant role in localization of deformation, which is independent of the lateral scale of the UPMB.

(vi) Lateral variation in the thickness of the UPMB can lead to asymmetric rift and passive margin structures. A greater amount of deformation is concentrated into the region where a larger amount of migrated magma has accumulated below the Moho. However, the greatest thickness of the UPMB must be away from the centre of the UPMB so as to obtain the asymmetric structure found in narrow volcanic and wide non-volcanic passive margins. It is also noted that an asymmetric rift structure is more favourably obtained by the UPMB when the strength contrast between each domain of the UPMB is weak enough not to distribute deformation into a particular domain. Although our numerical model shows the potential of UPMBs to produce an asymmetric rift structure, additional weakening may be required to obtain the lithosphere-cutting shear zone that is favourable for simple shear deformation.

(vii) An UPMB that is emplaced during an extensional process results in deformation migration into a narrower region. This can be responsible for inward migration of deformation which could develop into a continental break-up as observed in continental rifts and passive margins. A small amount of UPMB emplaced in the uppermost mantle favours a localized break-up, whereas a huge amount may blur out the localization. The least amount of UPMB could result in successive localizations of deformation and magmatic activity into a more local region.

Acknowledgements

We would like to thank Peter Readman for carefully reading the manuscript and valuable suggestions. Giacomo Corti and an anonymous reviewer are acknowledged for constructive reviews of the manuscript. The calculations in this study were carried out on the computer facilities at the Dublin Institute for Advanced Studies and University College Dublin. TY thanks Kelin Wang for sharing the finite element code tekton ver.2.1 and for allowing modification of the code. Jay Melosh and Arthur Raefsky are also acknowledged as the authors of the original code. TY was supported by the CosmoGrid Project that is funded by the Program for Research in Third Level Institutions under the Irish National Development Plan and with assistance from the European Regional Development Fund.

References

Artemjev, M.E., Artyushkov, E.V., 1971. Structure and isostasy of the Baikal Rift and the mechanism of rifting. J. Geophys. Res. 76, 1197–1211.
Bassi, G., 1991. Factors controlling the style of continental rifting: insights from numerical modelling. Earth Planet. Sci. Lett. 105, 430–452.
Bassi, G., 1995. Relative importance of strain rate and rheology for the mode of continental extension. Geophys. J. Int. 122, 195–210.
Bassi, G., Keen, C.E., Potter, P., 1993. Contrasting styles of rifting: models and examples from the Eastern Canadian margin. Tectonics 12, 639–655.
Bauer, K., Neben, S., Schreckenberger, B., Emmermann, R., Hinz, K., Fechner, N., Gohl, K., Schulze, A., Trumbull, R.B., Weber, K., 2000. Deep structure of the Namibian continental margin as derived from integrated geophysical studies. J. Geophys. Res. 105, 25,829–25,853.
Berndt, C., Mjelde, R., Planke, S., Shimamura, H., Faleide, J.I., 2001. Controls on the tectono-magmatic evolution of a volcanic transform margin: the Vøring Transform Margin, NE-Atlantic. Mar. Geophys. Res. 22, 133–152.
Bonini, M., Sokoutis, D., Mulugeta, G., Boccaletti, M., Corti, G., Innocenti, F., Manetti, P., Mazzarini, F., 2001. Dynamics of magma emplacement in the centrifuge models of continental extension with implications for flank volcanism. Tectonics 20, 1053–1065.
Brace, W.F., Kohlstedt, D.L., 1980. Limits on lithospheric stress imposed by laboratory experiments. J. Geophys. Res. 85, 6248–6252.
Braun, J., Beaumont, C., 1987. Styles of continental rifting: results from dynamic models of lithospheric extension. In: Beaumont, C., Tankard, A.J. (Eds.), Sedimentary Basins and Basin-Forming Mechanisms. Can. Soc. Pet. Geol. Mem., 12, pp. 241–258.
Braun, J., Beaumont, C., 1989. Dynamic models of the role of crustal shear zones in asymmetric continental extension. Earth Planet. Sci. Lett. 93, 405–423.
Buck, W.R., 1991. Modes of continental lithospheric extension. J. Geophys. Res. 96, 20,161–20,178.
Buck, W.R., 2006. The role of magma in the development of the Afro-Arabian Rift System. In: Yirgu, G., Ebinger, C.J., Maguire, P.K.H. (Eds.), The Afar Volcanic Province Within the East African Rift System. . Spec. Pub., 259. Geological Society of London, pp. 43–54.
Burke, K., Dewey, J.F., 1973. Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks. J. Geol. 81, 406–433.
Burov, E., Guillou-Frottier, L., 2005. The plume head–continental lithosphere interaction using a tectonically realistic formulation for the lithosphere. Geophys. J. Int. 161, 469–490.
Burov, E., Guillou-Frottier, L., d'Acremont, E., Le Pourhiet, L., Cloetingh, S., 2007. Plume head–lithosphere interactions near intra-continental plate boundaries. Tectonophysics 434, 15–38.
Byerlee, J.D., 1967. Frictional characteristics of granite under high confining pressure. J. Geophys. Res. 72, 3639–3648.
Byerlee, J.D., 1978. Friction of rocks. Pure Appl. Geophys. 116, 615–626.
Callot, J.-P., Grigné, C., Geoffroy, L., Brun, J.-P., 2001. Development of volcanic passive margins: two-dimensional laboratory models. Tectonics 20, 148–159.
Callot, J.-P., Geoffroy, L., Brun, J.-P., 2002. Development of volcanic passive margins: three-dimensional laboratory models. Tectonics 21 (1052). doi:10.1029/2001TC901019.
Carter, N.L., Tsenn, M.C., 1987. Flow properties of continental lithosphere. Tectonophysics 136, 27–63.
Chapman, D.S., Furlong, K.P., 1992. Thermal state of the lower continental crust. In: Fountain, D.M., Arculus, R., Kay, R.W. (Eds.), Continental Lower Crust. Elsevier, pp. 179–199.
Chopra, P.N., Paterson, M.S., 1984. The role of water in the deformation of dunite. J. Geophys. Res. 89, 7861–7876.
Coffin, M.F., Eldholm, O., 1994. Large igneous provinces: crustal structure, dimensions, and external consequences. Rev. Geophys. 32, 1–36.
Corti, G., Manetti, P., 2006. Asymmetric rifts due to asymmetric Mohos: an experimental approach. Earth Planet. Sci. Lett. 245, 315–329.
Corti, G., Bonini, M., Conticelli, S., Innocenti, F., Manetti, P., Sokoutis, D., 2003a. Analogue modelling of continental extension: a review focused on the relations between the patterns of deformation and the presence of magma. Earth Sci. Rev. 63, 169–247.
Corti, G., van Wijk, J., Bonini, M., Sokoutis, D., Cloetingh, S., Innocenti, F., Manetti, P., 2003b. Transition from continental break-up to punctiform seafloor spreading: how fast, symmetric and magmatic. Geophys. Res. Lett. 30 (1604). doi:10.1029/2003GL017374.
Corti, G., Bonini, M., Innocenti, F., Manetti, P., Piccardo, G.B., Ranalli, G., 2007. Experimental models of extension of continental lithosphere weakened by percolation of asthenospheric melts. J. Geodyn. 43, 465–483.
Courtillot, V.E., Renne, P.R., 2003. On the ages of flood basalt events. In: Courtillot, V.E. (Ed.), The Earth's Dynamics. Comptes Rendus-Academic des Sciences. Geoscience, 335, pp. 113–140.
Cox, K.G., 1980. A model for flood basalt volcanism. J. Petrol. 21, 629–650.
Davis, M., Kusznir, N., 2004. Depth-dependent lithospheric stretching at rifted continental margins. In: Karner, G.D. (Ed.), Proceedings of NSF Rifted Margins Theoretical Institute. Columbia University Press, pp. 92–136.
Dunbar, J.A., Sawyer, D.S., 1988. Continental rifting at pre-existing lithospheric weaknesses. Nature 333, 450–452.
Dunbar, J.A., Sawyer, D.S., 1989a. How preexisting weaknesses control the style of continental breakup. J. Geophys. Res. 94, 7278–7292.
Dunbar, J.A., Sawyer, D.S., 1989b. Patterns of continental extension along the conjugate margins of the central and north Atlantic oceans and Labrador Sea. Tectonics 8, 1059–1077.
Doubre, C., Manighetti, I., Dorbath, C., Dorbath, L., Jacques, E., Delmond, J.C., 2007a. Crustal structure and magmato-tectonic processes in an active rift (Asal-Ghoubbet, Afar, East Africa): 1. Insights from a 5-month seismological experiment. J. Geophys. Res. 112 (B05405). doi:10.1029/2005JB003940.
Doubre, C., Manighetti, I., Dorbath, L., Dorbath, C., Bertil, D., Delmond, J.C., 2007b. Crustal structure and magmato-tectonic processes in an active rift (Asal-Ghoubbet, Afar, East Africa): 2. Insights from the 23-year recording of seismicity since the last rifting event. J. Geophys. Res. 112 (B05406). doi:10.1029/2006JB004333.
Ebbing, J., Lundin, E., Olesen, O., Hansen, E.K., 2006. The mid-Norwegian margin: a discussion of crustal lineaments, mafic intrusions and remnants of the Caledonian root by 3D density modelling and structural interpretation. J. Geol. Soc. (Lond.) 163, 47–59.
Ebinger, C.J., Casey, M., 2001. Continental breakup in magmatic provinces: an Ethiopian example. Geology 29, 527–530.
Eldholm, O., Gladczenko, T.P., Skogseid, J., Planke, S., 2000. Atlantic volcanic margins: a comparative study. In: Nøttvedt, A. (Ed.), Dynamics of the Norwegian Margin. . Spec. Pub. Geological Society of London, pp. 411–428.
England, P., 1983. Constraints on extension of continental lithosphere. J. Geophys. Res. 88, 1145–1152.
Fernàndez, M., Ranalli, G., 1997. The role of rheology in extensional basin formation modelling. Tectonophysics 282, 129–145.
Fletcher, R.C., Hallet, B., 1983. Unstable extension of the lithosphere: a mechanical model for basin and range structure. J. Geophys. Res. 88, 7457–7466.

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024

Frederiksen, S., Braun, J., 2001. Numerical modelling of strain localization during extension of the continental lithosphere. Earth Planet. Sci. Lett. 188, 241–251.

Frey, Ø., Planke, S., Symonds, P.A., Heeremans, M., 1998. Deep crustal structure and rheology of the Gascoyne volcanic margin, western Australia. Mar. Geophys. Res. 20, 293–312.

Gac, S., Geoffroy, L., submitted for publication. Thermo-mechanical modelling in 3D of a stretched continental lithosphere containing localized low viscosity anomalies (the soft-point theory of plate break-up).

Gans, P.B., 1987. An open-system, two-layer crustal stretching model for the eastern Great Basin. Tectonics 6, 1–12.

Geoffroy, L., 2005. Volcanic passive margin. C. R. Geosciences 337, 1395–1408.

Geoffroy, L., Callot, J.-P., Scaillet, S., Skuce, A.S., Gélard, J.P., Ravilly, M., Angelier, J., Bonin, B., Cayet, C., Perrot, K., Lepvrier, C., 2001. Southeast Baffin volcanic margin and the North American–Greenland plate separation. Tectonics 20, 566–584.

Gernigon, L., Ringenbach, J.-C., Planke, S., Le Gall, B., 2004. Deep structures and breakup along volcanic rifted margins: insights from integrated studies along the outer Vøring Basin (Norway). Mar. Pet. Geol. 21, 363–372.

Gernigon, L., Lucazeau, F., Brigaud, F., Ringenbach, J.-C., Planke, S., Le Gall, B., 2006. A moderate melting model for the Vøring margin (Norway) based on structural observations and a thermo-kinematic modelling: Implication for the meaning of the lower crustal bodies. Tectonophysics 412, 255–278.

Gladczenko, T.P., Hinz, K., Eldholm, O., Meyer, H., Neben, S., Skogseid, J., 1997. South Atlantic volcanic margin. J. Geol. Soc. (Lond.) 154, 465–470.

Govers, R., Wortel, M.J.R., 1993. Initiation of asymmetric extension in continental lithosphere. Tectonophysics 223, 75–96.

Govers, R., Wortel, M.J.R., 1995. Extension of stable continental lithosphere and the initiation of lithospheric scale faults. Tectonics 14, 1041–1055.

Govers, R., Wortel, M.J.R., 1999. Some remarks on the relation between vertical motions of the lithosphere during extension and the necking depth parameter inferred from kinematic modeling studies. J. Geophys. Res. 104, 23,245–23,253.

Griffiths, R., Campbell, I., 1991. Interaction of mantle plume heads with the Earth's surface and onset of small-scale convection. J. Geophys. Res. 96, 18,295–18,310.

Harry, D.L., Leeman, W.P., 1995. Partial melting of melt metasomatized subcontinental mantle and the magma source potential of the lower lithosphere. J. Geophys. Res. 100, 10,255–10269.

Harry, D.L., Sawyer, D.S., Leeman, W.P., 1993. The mechanics of continental extension in western North America: implications for the magmatic and structural evolution of the Great Basin. Earth Planet. Sci. Lett. 117, 59–71.

Hayward, N., Ebinger, C., 1996. Variations in the along-axis segmentation of the Afar rift system. Tectonics 15, 244–257.

Hirth, G., Kohlstedt, D.L., 1996. Water in the oceanic upper mantle: implications for rheology, melt extraction and the evolution of the lithosphere. Earth Planet. Sci. Lett. 144, 93–108.

Huismans, R.S., Beaumont, C., 2002. Asymmetric lithospheric extension: the role of frictional-plastic strain softening inferred from numerical experiments. Geology 30, 211–214.

Hutchinson, D.R., White, R.S., Cannon, W.F., Schulz, K.J., 1990. Keweenaw hot spot: geophysical evidence for a 1.1 Ga mantle plume beneath the Midcontinent Rift system. J. Geophys. Res. 95, 10,869–10,884.

Keranen, K., Klemperer, S.L., Gloaguen, R., EAGLE Working Group, 2004. Three-dimensional seismic imaging of a protoridge axis in the Main Ethiopian rift. Geology 32, 949–952.

Keir, D., Ebinger, C.J., Stuart, G.W., Daly, E., Ayele, A., 2006. Strain accommodation by magmatism and faulting as rifting proceeds to breakup: seismicity of the northern Ethiopian rift. J. Geophys. Res. 111 (B05314). doi:10.1029/2005JB003748.

Kirby, S.H., 1983. Rheology of the lithosphere. Rev. Geophys. Space Phys. 21, 1458–1487.

Klausen, M.B., Larsen, H.C., 2002. East Greenland coast-parallel dike swarm and its role in continental breakup In: Menzies, M.A., Klemperer, S.L., Ebinger, C.J., Baker, J. (Eds.), Volcanic Rifted Margins: Boulder, Colorado. . Spec. Pub., vol. 362. Geol. Soc. Am., pp. 137–000.

Koch, P.S., Christie, J.M., Ord, A., George Jr., R.P., 1989. Effect of water on the rheology of experimentally deformed quartzite. J. Geophys. Res. 94, 13,975–13,996.

Kusznir, J.N., Egan, S.S., 1989. Simple shear and pure shear models of extensional sedimentary basin formation: application to the Jeanne d'Arc basin, grand banks of Newfoundland. Mem. –Am. Assoc. Pet. Geol. 46, 305–322.

Lachenbrunch, A.H., Sass, J.H., 1978. Models of an extending lithosphere and heat flow in the Basin and Range Province. In: Smith, R., Eaton, G. (Eds.), Cenozoic Tectonics and Regional Geophysics of the Western Cordillera. Mem. Geol. Soc. Am., vol. 152, pp. 209–250.

Lachenbrunch, A.H., Morgan, P., 1990. Continental extension, magmatism and elevation: formal relations and rules of thumb. Tectonophysics 174, 39–62.

Laske, G., Masters, G., 1997. A global digital map of sediment thickness. EOS Trans. AGU 78 (F483).

Latin, D., White, N., 1990. Generating melt during lithospheric extension: pure shear vs. simple shear. Geology 18, 327–331.

Lister, G.S., Davis, G.A., 1989. The origin of metamorphic core complexes and detachment faults formed during Tertiary continental extension in the northern Colorado River region U.S.A. J. Struct. Geol. 11, 65–94.

Lister, G.S., Etheridge, M.A., Symonds, P.A., 1991. Detachment models for the formation of passive continental margins. Tectonics 10, 1038–1064.

Lizarralde, D., Axen, G.J., Brown, H.E., Fletcher, J.M., González-Fernández, A., Harding, A.J., Holbrook, W.S., Kent, G.M., Paramo, P., Sutherland, F., Umhoefer, P.J., 2007. Variation in styles of rifting in the Gulf of California. Nature 448. doi:10.1038/nature06035.

Maguire, P.K.H., Keller, G.R., Klemperer, S.L., Mackenzie, G.D., Keranen, K., Harder, S., O'Reilly, B., Thybo, H., Asfaw, L., Khan, M.A., Amha, M., 2006. Crustal structure of the Northern Main Ethiopian Rift from the EAGLE controlled source survey; a snapshot of incipient lithospheric break-up. In: Yirgu, G., Ebinger, C., Maguire, P.K.H. (Eds.), The Afar volcanic province within the East African Rift System. . Spec. Pub., vol. 259. Geol. Soc. Lond., pp. 269–291.

Manighetti, I., Tapponnier, P., Courtillot, V., Gruszow, S., Gillot, P.-Y., 1998. Propagation of rifting along the Arabia–Somalia plate boundary. J. Geophys. Res. 102, 2681–2710.

McKenzie, D.P., 1978. Some remarks on the development of sedimentary basins. Earth Planet. Sci. Lett. 40, 25–32.

McKenzie, D., Bickle, J.M., 1988. The volume and composition of melt generated by extension of the lithosphere. J. Petrol. 29, 625–679.

Mackenzie, G.D., Thybo, H., Maguire, P.K.H., 2005. Crustal velocity structure across the Main Ethiopian Rift: results from two-dimensional wide-angle seismic modelling. Geophys. J. Int. 162, 994–1006.

Melosh, H.J., Raefsky, A., 1980. The dynamical origin of subduction zone topography. Geophys. J. R. Astron. Soc. 60, 333–354.

Melosh, H.J., Raefsky, A., 1981. A simple and efficient method for introducing faults into finite element computations. Bull. Seismol. Soc. Am. 71, 1391–1400.

Melosh, H.J., Williams, C.A., 1989. Mechanics of graben formation in crustal rocks: a finite element analysis. J. Geophys. Res. 94, 13,961–13,973.

Menzies, M.A., Klemperer, S.L., Ebinger, C.J., Baker, J., 2002. Characteristic of volcanic rifted margins. In: Menzies, M.A., Klemperer, S.L., Ebinger, C.J., Baker, J. (Eds.), Volcanic Rifted Margins: Boulder, Colorado. Spec. Pub., 362. Geol. Soc. Am., pp. 1–14.

Mjelde, R., Raum, T., Murai, Y., Takanami, T., 2007. Continent-ocean-transitions: review, and a new tectono-magmatic model of the Vøring Plateau, NE Atlantic. J. Geodyn. 43, 374–392.

Mohr, P., 1992. Nature of the crust beneath magmatically active continental rifts. Tectonophysics 213, 269–284.

Müller, R.D., Roest, W.R., Royer, J.Y., Gahagan, L.M., Sclater, J.G., 1997. Digital isochrons of the world's ocean floor. J. Geophys. Res. 102, 3211–3214.

Müller, R.D., Gaina, C., Roest, W.R., Hansen, D.L., 2001. A recipe for microcontinent formation. Geology 29, 203–206.

Mutter, C.Z., Talwani, M., Stoffa, P., 1984. Evidence for thick oceanic crust adjacent to the Norwegian margin. J. Geophys. Res. 89, 483–502.

Negredo, A.M., Fernàndez, M., Zeyen, H., 1995. Thermo-mechanical constraints on kinematic models of lithospheric extension. Earth Planet. Sci. Lett. 134, 87–98.

Neumann, E.-R., Olsen, K.H., Baldridge, W.S., 1995. The Oslo rift. In: Olsen, K.H. (Ed.), Continental Rifts: Evolution, Structure, Tectonics. . Developments in Geotectonics, vol. 25. Elsevier, pp. 345–371.

Pascal, C., van Wijk, J.C., Cloetingh, S.A.P.L., Davies, G.R., 2002. Effect of lithosphere thickness heterogeneities in controlling rift localization: numerical modeling of the Oslo Graben. Geophys. Res. Lett. 29. doi:10.1029/2001GL014354.

Planke, S., Skogseid, J., Eldholm, O., 1991. Crustal structure off Norway, 62° to 70° north. Tectonophysics 189, 91–107.

Poliakov, A., Buck, R., 1998. Mechanics of stretching elastic-plastic-viscous layers: applications to slow-spreading mid-ocean ridges. In: Buck, R. (Ed.), Faulting and Magmatism at Mid-Ocean Ridges. . Geophys. Mono., vol. 106. American Geophysical Union, pp. 305–324.

Ranalli, G., 1995. Rheology of the Earth, 2nd ed. Chapman and Hall, London, p. 413.

Ranalli, G., Piccardo, G.B., Corona-Chávez, P., 2007. Softening of the subcontinental lithospheric mantle by asthenosphere melts and the continental extension/oceanic spreading transition. J. Geodyn. 43, 450–464.

Rey, P., 1993. Seismic and tectono-metamorphic characters of the lower continental crust in Phanerozoic areas: a consequence of postthickening extension. Tectonics 12, 580–590.

Ro, H.E., Faleide, J.I., 1992. A stretching model for the Oslo Rift. Tectonophysics 208, 19–36.

Royden, L., Keen, C.E., 1980. Rifting process and thermal evolution of the continental margin of eastern Canada determined from subsidence curves. Earth Planet. Sci. Lett. 51, 343–361.

Ruppel, C., 1995. Extensional processes in continental lithosphere. J. Geophys. Res. 100, 24,187–24,215.

Sengör, A.M.C., Burke, K., 1978. Relative timing of rifting and volcanism on earth and its tectonic implications. Geophys. Res. Lett. 5, 419–421.

Shelton, G., Tullis, J., 1981. Experimental flow laws for crustal rocks. EOS 62, 396.

Sigmundsson, F., 2006. Magma does the splits. Nature 442, 251–252.

Speight, J., Skelhorn, R., Sloan, T., Knapp, R., 1982. The dyke swarms of Scotland. In: Sutherland, D.S. (Ed.), Igneous Rocks of the British Isles. John Wiley, London, pp. 449–459.

Symonds, P.A., Planke, S., Frey, F., Skogseid, J., 1998. Volcanic evolution of the Western Australian continental margin and its implications for basin development. In: Purcell, P.G., Purcell, R.R. (Eds.), The Sedimentary Basins of Western Australia 2: Proceedings of Petroleum Exploration Society of Australia Symposium. Perth, pp. 33–34.

Thompson, G.A., McCarthy, J., 1990. A gravity constraint on the origin of highly extended terranes. Tectonophysics 174, 197–206.

Tirel, C., Brun, J.-P., Sokoutis, D., 2006. Extension of thickened and hot lithospheres: inferences from laboratory modeling. Tectonics 25 (TC1005). doi:10.1029/2005TC001804.

Tucholke, B.E., Sawyer, D.S., Sibuet, J.-C., 2007. Breakup of the Newfoundland–Iberia rift In: Karner, G.D., Manatschal, G., Pinheiro, L.M. (Eds.), Imaging, Mapping and Modelling Continental Lithosphere Extension and Breakup. Spec. Pub., vol. 282. Geol. Soc. Lon., pp. 9–46.

van Wijk, J.W., Cloetingh, S.A.P.L., 2002. Basin migration caused by slow lithospheric extension, Earth Planet. Sci. Lett. 198, 275–288.

Wernicke, B., 1985. Uniform-sense normal simple shear of the continental lithosphere. Can. J. Earth Sci. 22, 108–125.

White, R., McKenzie, D., 1989. Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. J. Geophys. Res. 94, 7685–7729.

Williams, C.A., Richardson, R.M., 1991. A rheologically layered three-dimensional model of the San Andreas fault in central and southern California. J. Geophys. Res. 96, 16,597–16,623.

Wright, T.J., Ebinger, C., Biggs, J., Ayele, A., Yirgu, G., Keir, D., Stork, A., 2006. Magma-maintained rift segmentation at continental rupture in the 2005 Afar dyking episode. Nature 442, 291–294.

Yamada, Y., Nakada, M., 2006. Stratigraphic architecture of sedimentary basin induced by mantle diapiric upwelling and eustatic event. Tectonophysics 415, 103–121.

Yamasaki, T., 2004. Localized rheological weakening by grain-size reduction during lithospheric extension. Tectonophysics 386, 117–145.

Yamasaki, T., O'Reilly, B., Readman, P., 2006. A rheological weak zone intensified by post-rift thermal relaxation as a possible origin of simple shear deformation associated with reactivation of rifting. Earth Planet. Sci. Lett. 248, 134–146.

Yamasaki, T., Miura, H., Nogi, Y., submitted for publication. Numerical modelling study on the flexural uplift of the Transantarctic Mountains. Q2

Please cite this article as: Yamasaki, T., Gernigon, L., Styles of lithospheric extension controlled by underplated mafic bodies, Tectonophysics (2008), doi:10.1016/j.tecto.2008.04.024