Ancient continental lithosphere dislocated beneath ocean basins along the mid-lithosphere discontinuity: a hypothesis

Zhensheng Wang[1], Timothy M. Kusky[1, 2, *], Fabio A. Capitanio[1, 3]

1. *State Key Lab of Geological Processes and Mineral Resources, Center for Global Tectonics, School of Earth Sciences, China University of Geosciences, Wuhan, China*
2. *Three Gorges Research Center for Geohazards, China University of Geosciences, Wuhan, China*
3. *School of Earth, Atmosphere and Environment, Monash University, Clayton, 3800 VIC, Australia*

* *Corresponding author: tkusky@gmail.com*

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Abstract

The documented occurrence of ancient continental cratonic roots beneath several oceanic basins remains poorly explained by the plate tectonic paradigm. These roots are found beneath some ocean-continent boundaries, on the trailing sides of some continents, extending for hundreds of km or farther into oceanic basins. We postulate that these cratonic roots were left behind during plate motion, by differential shearing along the seismically imaged mid-lithosphere discontinuity (MLD), and then emplaced beneath the ocean-continent boundary. Here, we use numerical models of cratons with realistic crustal rheologies drifting at observed plate velocities to support the idea that the mid-lithosphere weak layer fostered the decoupling and offset of the African continent's buoyant cratonic root, which was left behind during Meso-Cenozoic continental drift and emplaced beneath the Atlantic Ocean. We show that in some cratonic areas, the MLD plays a similar role as the lithosphere-asthenosphere boundary for accommodating lateral plate tectonic displacements.

Keywords:

Cratonic root; Mid-lithosphere discontinuity; Oceanic basins; Plate tectonics; Continental drift; Oceanic plateau

Key points

- Cratonic roots can be left behind beneath oceanic basins during plate motion, by differential shearing along the seismically imaged MLD.
- This process might have taken place in the SW Africa-Atlantic margin, and the Ontong Java Plateau.
- A weak MLD likely plays a similar role as the LAB for accommodating lateral plate tectonic displacements.

1. Introduction

Modern oceanic basins are typically floored by lithosphere formed at spreading ridges or areas near mantle plume heads within oceanic basins. The age of the oceanic lithosphere increases away from spreading centers to continental margins, where modern oceanic lithosphere is no older than 300 Ma [Müller et al., 2008]. The thickness of most of the oceanic lithosphere increases with the square root of its age, until ~ 100 km [Doin and Fleitout, 1996], however, high seismic-velocity anomalies are found beneath some oceanic basins, indicating that the lithosphere extends in these places to greater depths [e.g., King and Ritsema, 2000; Deen et al., 2006; Begg et al., 2009]. An example is the presence of thick lithosphere beneath the southern Atlantic margin adjacent to Africa where several studies reveal high shear wave velocity anomalies (2-7 %) at depths between 100-175 km or deeper [e.g., King and Ritsema, 2000; Deen et al., 2006; Begg et al., 2009] (Figure 1). This anomaly is laterally continuous with high-velocity anomalies in the cratonic area adjacent to it (where the anomaly is ~8%), yet it is found 300-1300 km away from the passive margin. Some of these areas have been proposed to be continental lithospheric roots, possibly ancient cratonic lithosphere of Proterozoic or Archean ages (>1000 Ma) [Bonatti, 1990; Chazot et al., 2005; Deen et al., 2006; Begg et al., 2009; Coltorti et al., 2010], which were dislocated or embedded beneath oceanic basins for about 300-1300 km away from passive margin (Figure 1) [Bonatti, 1990; Chazot et al., 2005; Deen et al., 2006; Begg et al., 2009; Coltorti et al., 2010; Huismans and Beaumont, 2011].

This geophysically detectable, atypically thick and old lithosphere is distinct from regions yielding ancient magmatic components derived from heterogeneous mantle sources contaminated by ancient lithospheric fragments [Harvey et al., 2006; Liu et al., 2008; Hamelin et al., 2013], and also distinct from hyper-extended continental margins [Brune et al., 2014], where continental crust can be trapped within the forming oceanic lithosphere during stretching. These observations perhaps are consistent with suggestions that some parts of cratonic mantle lithosphere are not permanently attached to the drifting continental lithosphere, as previously thought, thus potentially highlighting novel aspects of continent stability [Artemieva and Mooney, 2002; Begg et al., 2009; Huismans and Beaumont, 2011; Beaumont and Ings, 2012; Huismans and Beaumont, 2014; Kaban et al., 2015].

However, why such volumes of what appears to be thick continental cratonic lithosphere are found beneath oceans remains controversial. Some of these phenomena are likely contributed by small-scale convection related down-welling [e.g., King and Ritsema, 2000]. Some think that the shift of buoyant cratonic roots in continental interiors can be caused by strong basal drag with over thick (>300 km) cratonic roots [Artemieva and Mooney, 2002; Kaban et al., 2015], some propose that buoyant cratonic roots are dislocated with relatively small offsets (typically <100 km) by forces that arise from lateral pressure gradients during the continental breakup process [Huismans and Beaumont, 2011; Beaumont and Ings, 2012; Huismans and Beaumont, 2014], whereas others support fragmentation and lateral disruption of the cratonic lithosphere along weak zones during rifting [Begg et al., 2009; O'Reilly et al., 2009].

Recently, a Mid-Lithospheric Discontinuity (MLD, at depths between ~60 and ~160 km, and generally between ~80 and 100 km) has been detected in most cratonic areas, showing low seismic velocities and a distinct seismic anisotropy signature [Thybo and Perchuć, 1997; Yuan and Romanowicz, 2010; Wölbern et al., 2012; Sodoudi et al., 2013; Selway et al., 2015; Calò et al., 2016; Aulbach et al., 2017]. Seismic velocity reductions across this layer are very large (3–10%) and greater than those measured across the lithosphere-asthenosphere boundary (LAB) in continental areas (~1% or slightly larger) [Selway et al., 2015 and references therein; Aulbach et al., 2017]. The fast-axis direction of azimuthal anisotropy in some cratonic areas (e.g., North America) changes sharply at MLD depths [Yuan and Romanowicz, 2010]. Proposed models for the nature of the MLD include enhanced partial melting or grain boundary sliding, enrichment in infiltrated frozen melts, pyroxenes, phlogopite, amphibole or carbonates, and layering of minerals with distinct orientation or changes in deformation creep mechanisms (boundary between diffusion- and dislocation-creep) [Selway et al., 2015; Fei et al., 2016; Aulbach et al., 2017]. The general view is that some MLD's are sufficiently weak layers in the mid-lithosphere [e.g., Thybo and Perchuć, 1997; Liao *et al.*, 2013; Liao and Gerya, 2014 and references therein], where decoupling and shearing between the upper and lower part of the cratonic lithosphere will most likely occur [Liao *et al.*, 2013; Liao and Gerya, 2014].

The occurrence of the MLD is also well-documented in the Tanzania and Kalahari Cratons of Africa. Here the cratonic roots are inferred to be buoyant, as constrained by deep-sourced xenolith compositions [Poudjom Djomani et al., 2001]. The MLD is possibly related to melt infiltration and metasomatism [Wölbern et al., 2012; Sodoudi et al., 2013], which can lead to localized MLD weakness and shearing. Thus, we postulate that the MLD in the African cratons is likely a weak layer that locally underwent strong shearing, decoupling the cratonic lithospheric root and allowing its lateral offset underneath the Atlantic oceanic basin, during motion of the African plate (Figure 1).

We test the viability of this mechanism by means of numerical modeling, and find that the large lithospheric offset beneath the continent-ocean transition zone can be generated by shearing along a weak MLD during plate motion. Our results are consistent with geological examples and have broad implications for understanding the origin of some other poorly understood thick, ancient lithosphere beneath oceanic basins.

2. Numerical modeling of lateral lithospheric root offset

The process described here relies on the mechanical strength of the MLD layer and lateral shearing during plate motions. The mechanical competence of the MLD is determined by its rheological properties, that is its averaged viscosity (η_{MLD}) and thickness (h_{MLD}). These values can be inferred from laboratory-constrained flow laws of mantle-forming minerals, yielding averaged viscosities of 10^{19}-10^{23} Pa s [Hirth and Kohlstedt, 2003] and geophysically-constrained MLD thicknesses, ranging from 10 to 50 km [Selway et al., 2015]. The shearing during plate motions is mainly driven by the plate velocities (v_{plate}) along the model's upper boundary, which is here constrained by present-day plate velocities of 0 to 20 cm/yr [Müller et al., 2008; Seton et al., 2012; Zahirovic et al., 2015] with respect to a fixed reference frame, here placed at the model's mantle lower boundary. Thus, we test the role of these parameters by numerical modeling (see the details in the supporting information), where a vertically-layered cratonic lithosphere embeds a layer with different predefined thicknesses and effective viscosities in between the upper and lower lithosphere, representing the MLD, and undergoes a range of shearing velocities acting as plate velocity on the top of the plate. Figure 2 shows two representative examples out of 225 models we tested, with a weak and a strong MLD in between the upper and lower lithosphere. The related methods and details for modeling are described in the supporting information [Ranalli, 1995; Moresi et al., 2003; Gerya and Stöckhert, 2006; Capitanio et al., 2010; Rodríguez González et al., 2012; Lee et al., 2011; Li et al., 2013; Wang et al., 2015; Wang et al., 2016].

The results from all models are displayed on a MLD thickness-effective viscosity-plate velocity (h_{MLD}-η_{MLD}-v_{plate}) space through data interpolation, where we measure, as diagnostic indicators, the decoupling velocity ($v_{\mathrm{decoupling}}$) and the second invariant of the deviatoric strain rate tensor ($\dot{\varepsilon}_{II}$). The decoupling velocity is the differential velocity between the lower and the upper sections of the lithosphere and provides an estimate of the lateral offset rates of the lithospheric root, while the strain rates in the MLD represent the instantaneous shearing within the MLD. We have also tested the length of the cratonic block along the plate motion direction, although we found the outcomes are not sensitive to this parameter (Figure S2).

We find that shearing, quantified by the strain rate $\dot{\varepsilon}_{II}$, increases with a reduction in MLD strength which results from either increasing the MLD thickness or lowering its viscosity. Instead, the lateral offset rate, quantified by $v_{\mathrm{decoupling}}$, increases with plate velocity. Our results show that in models with vanishing MLD thickness, or large MLD viscosity, only minor cratonic root offset and shearing are measured, with $v_{\mathrm{decoupling}}$ of 0.01 to 0.1 cm/yr and $\dot{\varepsilon}_{II}$ $10^{-17.5}$ to $10^{-16.5}$s^{-1}, in spite of the plate velocity being as large as 20 cm/yr (Figures 3a and 3b). In this case, the lateral offset of the lower lithosphere is 10-100 km within 100 Ma for tested plate velocities of 1 to 20 cm/yr. The models with large MLD thicknesses and low MLD effective viscosity produce the largest offset of the lower lithosphere (Figures 3a and 3b), with decoupling velocities of 1 to 10 cm/yr, which can lead to lower lithosphere lateral offsets as large as 1000-10000 km within 100 Ma, and strong shearing between the upper and lower lithosphere along the weak MLD with strain rates of $10^{-15.5}$ to $10^{-14.5}$ s^{-1}.

3. Applications to geological cases

The SW Africa region underwent tectonic translation at plate velocities of 0.8 to 6 cm/yr, with an average value of ~2.5 cm/yr, since the breakup between Africa and South America 130 Ma ago (Figure 1) [Müller et al., 2008; Seton et al., 2012; Zahirovic et al., 2015]. The largest ~1300 km wide lateral offset of continental lithosphere beneath the ocean basin, initiated by 130 Ma, could have been achieved at an average decoupling velocity of ~1 cm/yr. The lower bounds imposed by averaged plate and decoupling velocities constrain the condition of the average MLD rheology near SW Africa (thickness and effective viscosity) based on the modeling results (Figure 3) imply an average MLD viscosity of 10^{19} to $5*10^{20}$ Pa s and thickness of 10 to 50 km (Figures 3c and 3d), to allow decoupling velocities of ~0.5-1.5 cm/yr.

More realistically, the velocity of the African plate was faster during the interval from 130 - 90 Ma following break-up (Figure 1), where absolute plate velocities ranged between 2.5 and 6 cm/yr, after which this has been constantly below 2.5 cm/yr since 90 Ma. Likely, the root's offset occurred in this stage, when faster plate velocity maximized the lithospheric shearing. The largest offset of ~1300 km between 130-90 Ma could have occurred at an average rate of 3.25 cm/yr, in this case. The numerical models with plate motions at such velocities reproduce compatible decoupling rates, that is >3.25 cm/yr, for MLD weaker than ~$10^{19.5}$ Pa s and thicknesses of ~40 km.

The estimated thickness and averaged viscosities in these models are compatible with the available constraints on lithospheric discontinuities. The minimum MLD thickness required according to the models is ~25 km, consistent with the well-documented 30-40 km MLD thickness in many cratons [Yuan and Romanowicz, 2010; Wölbern et al., 2012; Sodoudi et al., 2013; Selway et al., 2015; Aulbach et al., 2017]. The model results suggest that the MLD viscosity constraints for the SW African case (10^{19} - $5*10^{20}$ Pa s) are somewhat outside the formal viscosity range of mantle lithosphere by laboratory constraints (10^{21}-$5*10^{24}$ Pa s) [Selway, 2015], factors such as water, partial melt, and crystal alignment and shear weakening, all reduce viscosity. The difference of 1-2 orders of magnitude in viscosity can be explained reasonably well by wet mantle rocks with ~400-2800 ppm hydrogen or partially melted mantle rocks with 5-10 wt% melt [Hirth and Kohlstedt, 2003; Wang et al., 2016] and assisted by the crystallographic preferred orientations of mantle minerals and the consequent viscous anisotropy decrease (by 1 order of magnitude) [Hansen et al., 2012], or favored by recrystallization of melts with basaltic composition and smaller grain size [Hirth and Kohlstedt, 2003].

4. Discussion

The process explained here is compatible with examples of cratonic lithosphere delamination, in which lithospheric plates are dismembered vertically. In this case, shearing and decoupling also occurs along the weak MLD, while sinking is the consequence of negatively-buoyant roots. This mechanism has been proposed to explain the observed lithospheric root loss in the North China Craton [Chen et al., 2014]. Similar to other kinds of metasomatism-related lithospheric weakening [Wenker and Beaumont, 2017], the weakness of the MLD has also been demonstrated to be important for craton extension and thinning processes [e.g., Liao et al., 2013; Liao and Gerya, 2014].

The numerical models here on the lateral offset of cratonic roots are comparable with previous works focusing on buoyant cratonic roots dislocated by forces that arise from lateral pressure gradients during breakup process [e.g., Huismans and Beaumont, 2011; Beaumont and Ings, 2012; Huismans and Beaumont, 2014] or by strong basal drag with over thick (>300 km) cratonic roots in continental interiors [e.g., Artemieva and Mooney, 2002; Kaban et al., 2015]. The forces that arise from lateral pressure gradients and basal drag can make the continental lithospheric mantle unroofed or shallowly roofed in the oceanic basins during breakup, with lateral offsets typically <100 km for models without sufficiently weak MLD's [e.g., Huismans and Beaumont, 2011; Beaumont and Ings, 2012; Huismans and Beaumont, 2014] and >100 km for models with sufficiently weak MLD's [e.g., Liao et al., 2013; Liao and Gerya, 2014]. Besides, small-scale convection-related down-welling and deformation of the lithospheric step at craton-ocean margins can also contribute to the displacement [King and Ritsema, 2000]. Thus, small-scale convection, forces that arise from lateral pressure gradients, basal drag, shearing along a sufficiently weak MLD, as well as lateral disruption of the cratonic lithosphere during rifting [Begg et al., 2009; O'Reilly et al., 2009] likely accompany each other to accomplish such a long distance offset (300-1300km) of cratonic roots.

Our model results indicate large lithospheric offset along the MLD can be achieved during the post-rift stage after the oceanic lithosphere has thermally stabilized, whereas previous model results [e.g., Liao et al., 2013; Liao and Gerya, 2014] have shown that the models at the time of rifting would develop exhuming channel flows which would modify/inhibit the decoupling of the lower lithosphere. All of these studies [Liao et al., 2013; Liao and Gerya, 2014 and this study] use a 'dry' olivine flow law for the asthenosphere, which would induce a stronger basal drag than the one using 'wet' olivine. In a 'wet' olivine case with approximately a factor of 10 lower viscosity for the asthenosphere, the basal drag can be reduced by a factor of 10, thus the decoupling velocity and displacement of the lower lithosphere would be also reduced.

5. Hypothesis and implications for thick, ancient lithosphere beneath oceanic basins

The process proposed and modeled here can be divided into 3 stages. In the initial stage (Stage 0, Figure 4a), cratonic lithosphere with a weak MLD is still stable because of insufficient weakness of the MLD, a large distance from a continental margin, or protection by surrounding orogenic belts. The lithosphere at this stage is likely similar to the lithospheres of the present Superior, Slave, Yilgarn and western North China Cratons [Chen et al., 2014; Selway et al., 2015; Aulbach et al., 2017], where Mid-Lithosphere Discontinuities are imaged within the cratons, but are probably not weak enough and protected by surrounding orogenic belts [e.g., Lenardic et al., 2000] so do not show offset of roots beneath oceans. For the cratons with sufficiently weak MLD's (low viscosity and large thickness), the cratonic roots would start to be displaced laterally during on-craton rifting, making cratonic roots dislocated beneath the newly formed oceanic basins [Liao *et al.*, 2013; Liao and Gerya, 2014]. In the following stage (Stage 1), the upper lithosphere moves fast during plate motion, leaving the buoyant cratonic root behind beneath the oceanic-continent transition along the weakly coupled MLD (Figure 4b). The best example for this stage is likely the lithosphere near the SW Africa margin of the Atlantic. This process can be sustained by large plate velocities, driving progressive shear, although progressive thinning of the MLD results in decreased decoupling (Figure 4c), so that the displaced buoyant cratonic root is anchored beneath, and consequently coupled to, the oceanic plate, moving with it (Stage 2). This results in a new, hybrid oceanic/continental lithosphere, with the upper section formed at an ocean ridge, and with a cratonic root trapped beneath it. Upon emplacement in the oceanic realm, where geotherms are hotter, the MLD undergoes heating and potentially melting. Melting in the lithospheric layers rich in amphibole, phlogopite, frozen melts, or volatile matter [Griffin et al., 2003; Thybo and Perchuć, 1997; Foley, 2008; Wölbern et al., 2012; Selway, 2015], can occur followed by upward migration of melt and intrusion into a new, or reworked, MLD in the oceanic area (Figure S5) above the former MLD. We speculate that a likely example of this stage is the layered lithosphere in the Ontong Java oceanic plateau. Here, lower lithosphere with ancient continental features (~95-120 km or ~280 km, 0.9-1.7 Ga) [Ishikawa et al., 2011; Tharimena et al., 2016] underlies the young oceanic lithosphere (<85 km and <160 Ma) with a clearly detected 'MLD' (40-80 km) in between [Ishikawa et al., 2011; Tharimena et al., 2016]. Although a proposed cause is plume impingement [Ishikawa et al., 2011; Tharimena et al., 2016], this configuration is also compatible with the Stage 2 in our model. Additionally, other discrete segments of ancient continental roots beneath oceanic basins in the Atlantic [Bonatti, 1990; Coltorti et al., 2010] can also be examples of Stage 2.

The model proposed here is further corroborated by similar processes occurring on the western side of the Atlantic rift-to-drift evolution. Thick lithospheric roots are also imaged by seismic studies along the American side of the Atlantic, beneath the North and South American continental margins [e.g., King and Ritsema, 2000; O'Reilly et al., 2009] (Figure S6). Following the Atlantic opening, similar shearing and offset of continental roots might have occurred on both sides of the oceanic spreading, so that some North and South American continental lithosphere underwent the same processes as those constrained in the SW African margin. This supports the idea that the mechanism proposed here is of global relevance, especially for the craton-ocean margins with relatively dry asthenosphere or strong basal drag. It reveals that in some cratonic areas, the MLD plays a similar role as the lithosphere-asthenosphere boundary (LAB) for plate tectonics. Both the LAB and the MLD can serve as the boundary for accommodating lateral plate tectonic displacements.

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References

Amante, C. and B.W. Eakins (2009), ETOPO1 1 Arc-minute global relief model: procedures, data sources and analysis, NOAA Technical Memorandum NESDIS NGDC-24, National Geophysical Data Center, NOAA. doi:10.7289/V5C8276M

Artemieva, I. M., and W. D. Mooney (2002), On the relations between cratonic lithosphere thickness, plate motions, and basal drag, Tectonophysics, 358, 211-231.

Aulbach, S., M. Massuyeau, and F. Gaillard (2017), Origins of cratonic mantle discontinuities: A view from petrology, geochemistry and thermodynamic models, Lithos, 268, 364-382.

Beaumont, C., and S. J. Ings (2012), Effect of depleted continental lithosphere counterflow and inherited crustal weakness on rifting of the continental lithosphere: general results, J. Geophys. Res., 117, 10-1029.

Begg, G. C., W. L. Griffin, L. M. Natapov, S. Y. O'Reilly, S. P. Grand, C. J. O'Neill, J. Hronsky, Y. P. Djomani, C. J. Swain, and T. Deen (2009), The lithospheric architecture of Africa: Seismic tomography, mantle petrology, and tectonic evolution, Geosphere, 5, 23-50.

Bonatti, E. (1990), Subcontinental mantle exposed in the Atlantic Ocean on St Peter-Paul islets, Nature, 345, 800-802.

Brune, S., C. Heine, M. Pérezgussinyé, and S. V. Sobolev (2014), Rift migration explains continental margin asymmetry and crustal hyper-extension, Nat. Commun., 5, doi:10.1038/ncomms5014

Calò, M., T. Bodin, and B. Romanowicz (2016), Layered structure in the upper mantle across North America from joint inversion of long and short period seismic data, Earth Planet. Sci. Lett., 449, 164-175.

Capitanio, F. A., D. R. Stegman, L. Moresi, and W. Sharples (2010), Upper plate controls on deep subduction, trench migrations and deformations at convergent margins, Tectonophysics, 483, 80-92.

Chazot, G., S. Charpentier, J. Kornprobst, R. Vannucci, and B. Luais (2005), Lithospheric Mantle Evolution during Continental Break-Up: The West Iberia Non-Volcanic Passive Margin, J Petrol., 46, 2527-2568.

Chen, L., M. Jiang, J. Yang, Z. Wei, C. Liu, and Y. Ling (2014), Presence of an intralithospheric discontinuity in the central and western North China Craton: Implications for destruction of the craton, Geology, 42, 223-226.

Coltorti, M., C. Bonadiman, S. Y. O'Reilly, W. L. Griffin, and N. J. Pearson (2010), Buoyant ancient continental mantle embedded in oceanic lithosphere (Sal Island, Cape Verde Archipelago), Lithos, 120, 223-233.

Deen, T. J., W. L. Griffin, G. Begg, S. Y. O'Reilly, L. M. Natapov, and J. Hronsky (2006), Thermal and compositional structure of the subcontinental lithospheric mantle: Derivation from shear wave seismic tomography, Geochem. Geophys. Geosyst., 7, 329-349.

Doin, M. P., and L. Fleitout (1996), Thermal evolution of the oceanic lithosphere: an alternative view, Earth Planet. Sci. Lett., 142, 121-136.

Fei, H., S. Koizumi, N. Sakamoto, M. Hashiguchi, H. Yurimoto, K. Marquardt, N. Miyajima, D. Yamazaki, and T. Katsura (2016), New constraints on upper mantle creep mechanism inferred from silicon grain-boundary diffusion rates, Earth Planet. Sci. Lett., 433, 350-359.

Foley, S. F. (2008), Rejuvenation and erosion of the cratonic lithosphere, Nat. Geosci., 1, 503-510.

Gerya, T., and B. Stöckhert (2006), Two-dimensional numerical modeling of tectonic and metamorphic histories at active continental margins, Int. J. Earth Sci., 95, 250-274.

Griffin, W. L., S. Y. O'Reilly, L. M. Natapov, and C. G. Ryan (2003), The evolution of lithospheric mantle beneath the Kalahari Craton and its margins, Lithos, 71, 215-241.

Hamelin, C., A. Bezos, L. Dosso, J. Escartin, M. Cannat, and C. Mevel (2013), Atypically depleted upper mantle component revealed by Hf isotopes at Lucky Strike segment, Chem. Geol., 341, 128-139.

Hansen, L. N., M. E. Zimmerman, and D. L. Kohlstedt (2012), Laboratory measurements of the viscous anisotropy of olivine aggregates, Nature, 492, 415-418.

Harvey, J., A. Gannoun, K. W. Burton, N. W. Rogers, O. Alard, and I. J. Parkinson (2006), Ancient melt extraction from the oceanic upper mantle revealed by Re–Os isotopes in abyssal peridotites from the Mid-Atlantic ridge, Earth Planet. Sci. Lett., 244, 606-621.

Hirth, G., and D. Kohlstedt (2003), Rheology of the upper mantle and the mantle wedge: A view from the experimentalists, in Inside the Subduction Factory, AGU Monogr., vol. 138, edited by J. Eiler, pp. 83–105, AGU, Washington, D. C.

Huismans, R., and C. Beaumont (2011), Depth-dependent extension, two-stage breakup and cratonic underplating at rifted margins, Nature, 473, 74-78.

Huismans, R. S., and C. Beaumont (2014), Rifted continental margins: The case for depth-dependent extension, Earth Planet. Sci. Lett., 407, 148-162.

Ishikawa, A., D. G. Pearson, and C. W. Dale (2011), Ancient Os isotope signatures from the Ontong Java Plateau lithosphere: Tracing lithospheric accretion history, Earth Planet. Sci. Lett., 301, 159-170.

Kaban, M. K., W. D. Mooney, and A. G. Petrunin (2015), Cratonic root beneath North America shifted by basal drag from the convecting mantle, Nat. Geosci., 8, 797-800.

King, S. D., and J. Ritsema (2000), African hot spot volcanism: small-scale convection in the upper mantle beneath cratons, Science, 290, 1137-1140.

Lee, C. A., P. Luffi, and E. J. Chin (2011), Building and destroying continental mantle, Annu. Rev. Earth Planet. Sci., 39,59–90.

Lenardic, A., L. Moresi, and H. Mühlhaus (2000), The role of mobile belts for the longevity of deep cratonic lithosphere: the crumple zone model, Geophys. Res. Lett., 27, 1235-1238.

Li, Z., Z. Xu, T. Gerya, and J. Burg (2013), Collision of continental corner from 3-D numerical modeling, Earth Planet. Sci. Lett., 380, 98-111.

Liao, J., T. Gerya, and Q. Wang (2013), Layered structure of the lithospheric mantle changes dynamics of craton extension, Geophys. Res. Lett., 40, 5861-5866.

Liao, J., and T. Gerya (2014), Influence of lithospheric mantle stratification on craton extension: Insight from two-dimensional thermo-mechanical modeling, Tectonophysics, 631, 50-64.

Liu, C. Z., J. E. Snow, E. Hellebrand, G. Brügmann, A. V. D. Handt, A. Büchl, and A. W. Hofmann (2008), Ancient, highly heterogeneous mantle beneath Gakkel ridge, Arctic Ocean, Nature, 452, 311-316.

Moresi, L., F. Dufour, and H. Mühlhaus (2003), A Lagrangian integration point finite element method for large deformation modeling of viscoelastic geomaterials, J Comput. Phys., 184, 476-497.

Müller, R. D., M. Sdrolias, C. Gaina, and W. R. Roest (2008), Age, spreading rates, and spreading asymmetry of the world's ocean crust, Geochem. Geophys. Geosyst., 9, Q4006, 10.1029/2007GC001743

O'Reilly, S. Y., M. Zhang, W. L. Griffin, G. Begg, and J. Hronsky (2009), Ultradeep continental roots and their oceanic remnants: A solution to the geochemical “mantle reservoir” problem? Lithos, 112, 1043-1054.

Poudjom Djomani, Y. H., S. Y. O Reilly, W. L. Griffin, and P. Morgan (2001), The density structure of subcontinental lithosphere through time, Earth Planet. Sci. Lett., 184, 605-621.

Ranalli, G. (1995), Rheology of the Earth, Deformation and Flow Processes in Geophysics and Geodynamics, 2nd ed., Chapman and Hall, London.

Rodríguez González, J., A. M. Negredo, and M. I. Billen (2012), The role of the overriding plate thermal state on slab dip variability and on the occurrence of flat subduction, Geochem. Geophys. Geosyst., 13(Q01002), 1-21.

Selway, K., H. Ford, and P. Kelemen (2015), The seismic mid-lithosphere discontinuity, Earth Planet. Sci. Lett., 414, 45-57.

Selway, K. (2015), Negligible effect of hydrogen content on plate strength in East Africa, Nat. Geosci., 8, 543-546.

Seton, M., R. D. Müller, S. Zahirovic, C. Gaina, T. Torsvik, G. Shephard, A. Talsma, M. Gurnis, M. Turner, and S. Maus (2012), Global continental and ocean basin reconstructions since 200 Ma, Earth-Sci. Rev., 113, 212-270.

Sodoudi, F., X. Yuan, R. Kind, S. Lebedev, J. M. C. Adam, E. Kästle, and F. Tilmann (2013), Seismic evidence for stratification in composition and anisotropic fabric within the thick lithosphere of Kalahari Craton, Geochem. Geophys. Geosyst., 14, 5393-5412.

Tharimena, S., C. A. Rychert, and N. Harmon (2016), Seismic imaging of a mid-lithospheric discontinuity beneath Ontong Java Plateau, Earth Planet. Sci. Lett., 450, 62-70.

Thybo, H., and E. Perchuć (1997), The seismic 8 discontinuity and partial melting in continental mantle, Science, 275, 1626-1629.

Wang, H., J. van Hunen, and D. Graham Pearson (2015), The thinning of subcontinental lithosphere: The roles of plume impact and metasomatic weakening, Geochem. Geophys. Geosyst., 16, 1156-1171.

Wang, Z., T. Kusky, and F. Capitanio (2016), Lithosphere thinning induced by slab penetration into a hydrous mantle transition zone, Geophys. Res. Lett., 43, 11567-11577.

Wenker, S., and C. Beaumont (2017), Can metasomatic weakening result in the rifting of cratons? Tectonophysics, in press, http://dx.doi.org/10.1016/j.tecto.2017.06.013.

Wölbern, I., G. Rümpker, K. Link, and F. Sodoudi (2012), Melt infiltration of the lower lithosphere beneath the Tanzania craton and the Albertine rift inferred from S receiver functions, Geochem. Geophys. Geosyst., 13, 1-20.

Yuan, H., and B. Romanowicz (2010), Lithospheric layering in the North American craton, Nature, 466, 1063-1068.

Zahirovic, S., R. D. Müller, M. Seton, and N. Flament (2015), Tectonic speed limits from plate kinematic reconstructions, Earth Planet. Sci. Lett., 418, 40-52.

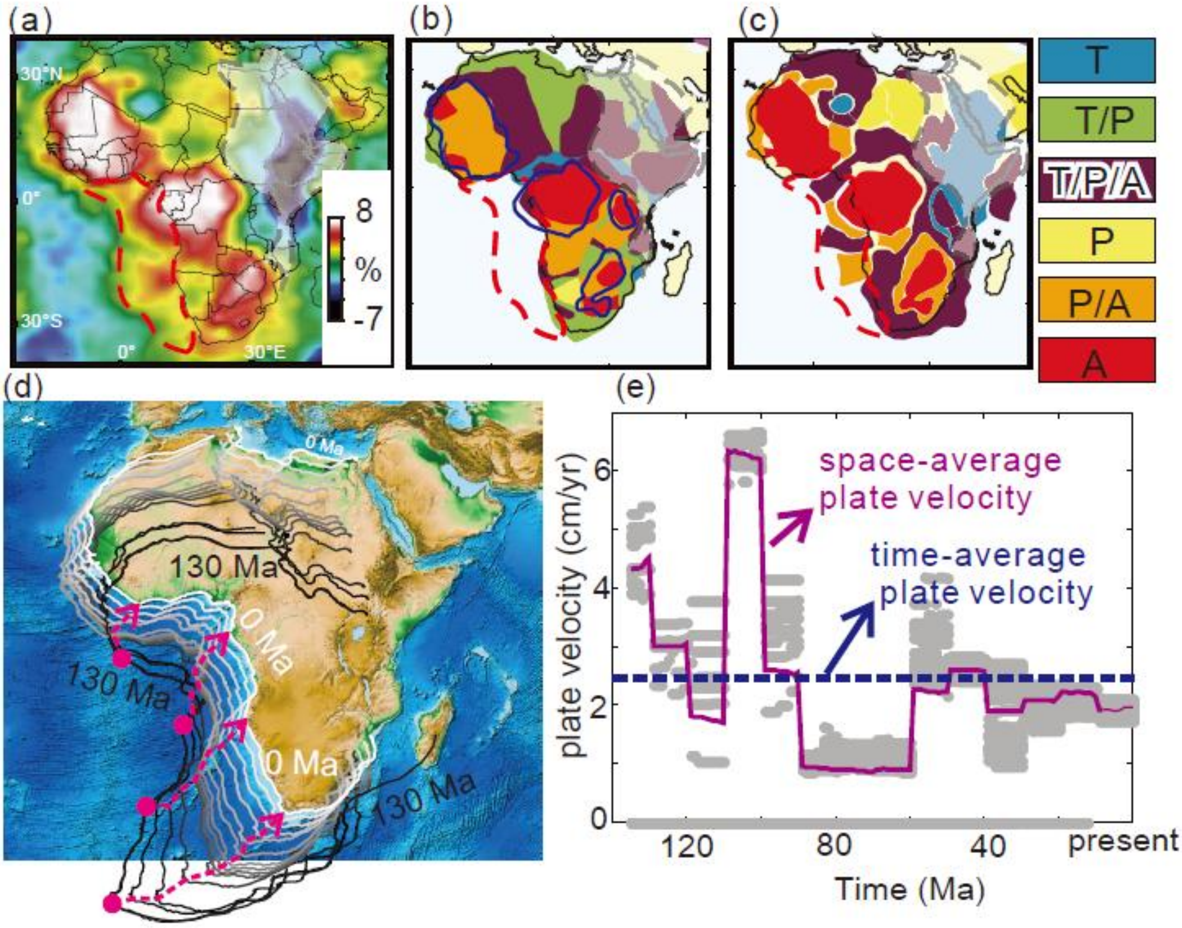

Figure 1. Lateral lithosphere offset near SW Africa during plate motion. (a) 100-175 km depth Vs difference section near Africa from the 4.5 km/s average velocity of this depth (after Begg et al., 2009). (b) Age units of upper lithosphere (~ 0-100 km depth) (after Begg et al., 2009). (c) Age units of lower lithosphere (~ 100-175 km depth) (after Begg et al., 2009). "A" means the regions that experienced their last major tectonothermal events >2.5 Ga ago; "P" is formed or modified between 2.5-1.0 Ga; "T" means formed or modified <1.0 Ga. Red dashed line outlines the distinct region with oceanic upper lithosphere and ancient cratonic roots, extending 300-1300 km into the Atlantic Ocean; black dashed line outlines the region with ancient / reworked upper lithosphere but without ancient lower lithosphere. Blue lines in (b)

outlines the ancient lower lithosphere with last major tectonothermal events >2.5 Ga (red "A" unit in c) to show the lateral offset between the upper and lower lithosphere. (d) Paleo-position (130 Ma - present, black-gray-white lines) and plate motion path line of the African plate through the past 130 million years, topographic map is from ETOPO1 (Amante and Eakins, 2009). (e) Average plate velocity through time relative to a fixed mantle reference frame based on the velocity of control points (gray circles, resolution is 3×3 degree within the whole African plate, see supporting information) on the African plate.

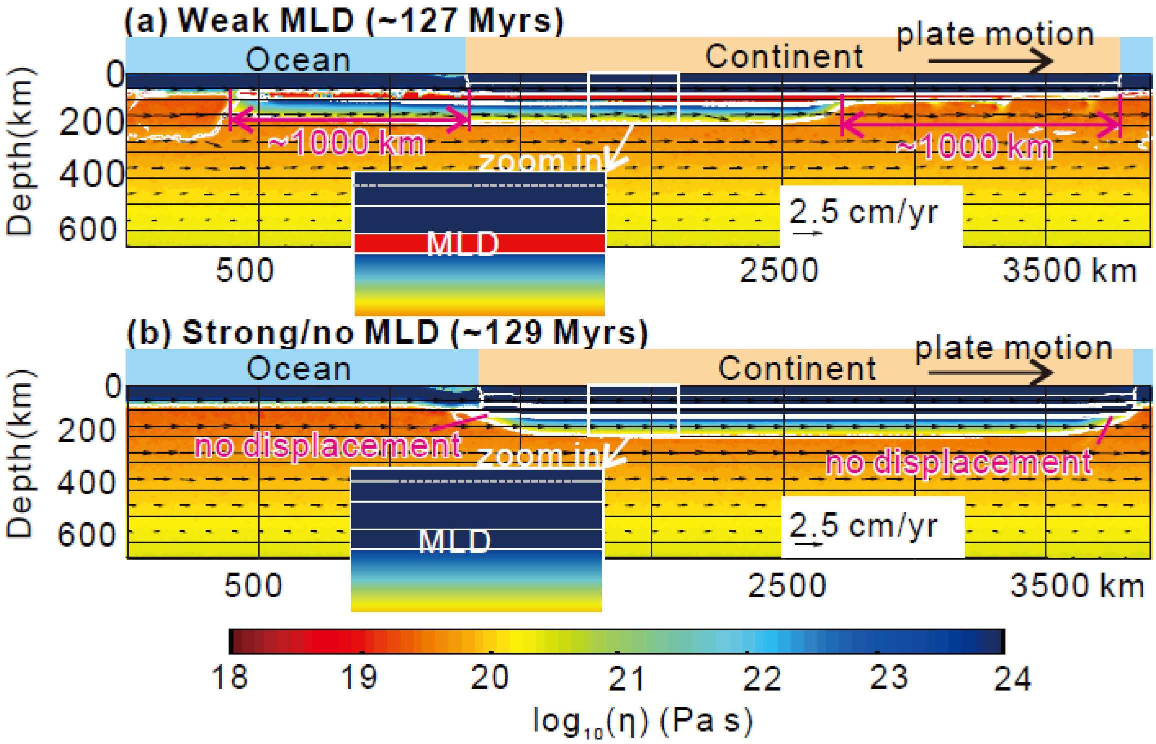

Figure 2. Model example of a ~2500 km length cratonic lithosphere under plate velocity v_{plate}= 2.5 cm / yr. (a) Example with 30 km thick MLD owning a viscosity of 10^{19} Pa s, significant offset (~1000 km) is calculated after ~127 Myrs of plate motion (v_{plate}= 2.5 cm / yr). (b) Example without MLD, no significant offset is observed after similar period (~129 Myrs) of plate motion (v_{plate}= 2.5 cm / yr). For all models in this paper, oceanic crust and oceanic mantle lithosphere are 0-10 km deep and 10-80 km deep, respectively. The upper crust (0-15 km deep), lower crust (15-40 km deep), and cratonic mantle lithosphere (40-200 km deep) are assigned to cratonic regions.

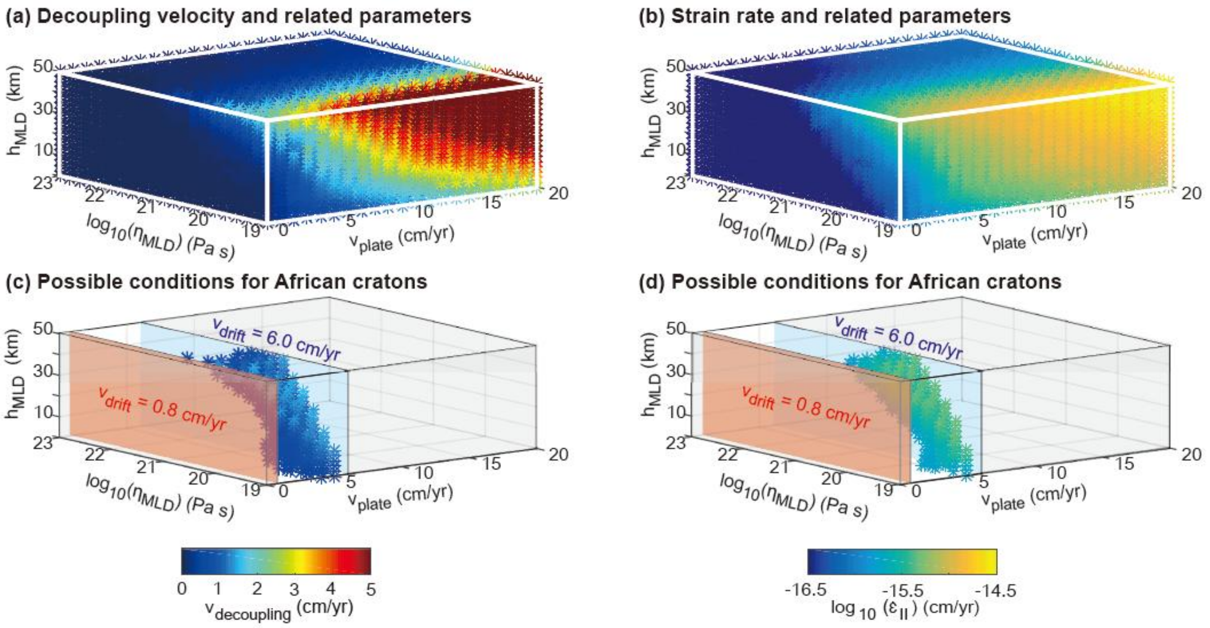

Figure 3. Dependence of decoupling velocity ($v_{decoupling}$, a and c) between the upper and lower lithosphere and the second invariant of the deviatoric strain rate tensor ($\dot{\varepsilon}_{II}$, b and d) on different parameters interpolated from more than two hundred model results. v_{plate} is the plate velocity (in cm/yr) of the cratonic upper lithosphere, η_{MLD} is the viscosity (in Pa s) of the MLD, h_{MLD} is the thickness (in km) of the MLD. (a) and (b) are the possible conditions for all cratons around the world. (c) and (d) are the possible conditions for African cratons bordered by two plate velocity slices (0.8 and 6 cm/yr for minimum and maximum plate velocities, respectively, as shown in Figure 1e), and with a decoupling velocity between 0.5-1.5 cm/yr ($v_{decoupling}$= v_{plate}- v_{root}).

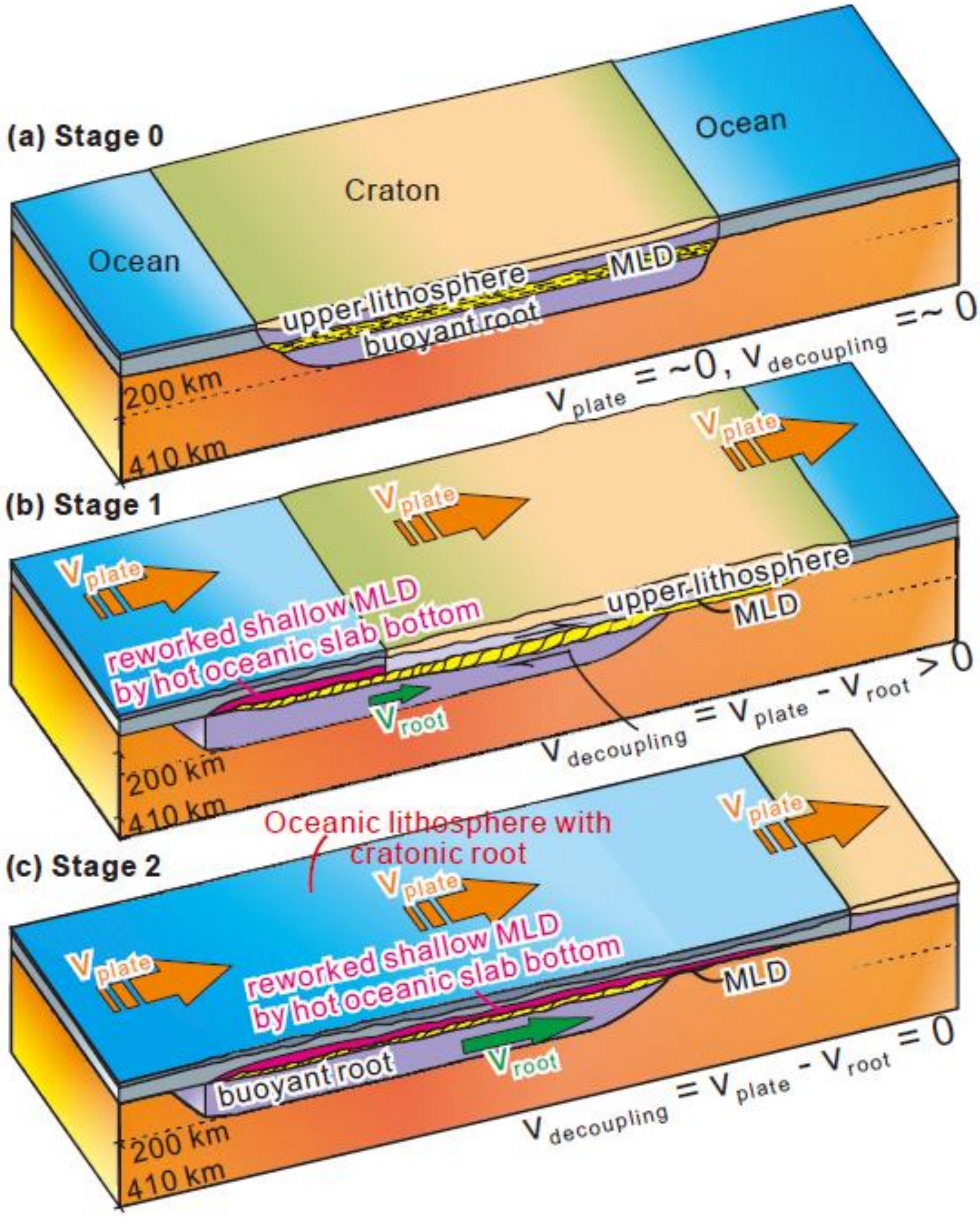

Figure 4. Simplified sketch of decoupling and lateral offset of the buoyant cratonic root beneath oceanic basins along the MLD. (a) Stage 0: Cratonic lithosphere with weak MLD's is still stable before significant plate-motion-related decoupling. (b) Stage 1: Decoupling, and sliding out of the upper lithosphere along the MLD. (c) Stage 2: MLD thinned through shearing and reworking, and displaced cratonic root being coupled with oceanic lithosphere.

Note that the buoyant root (purple) formed entirely beneath the continental lithosphere (green to orange) and has been laterally left behind during continental drift.