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Article

Why supercontinents became shorter lived as the Earth evolved

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A R T I C L E I N F O

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A B S T R A C T

Periodic assembly and break-up of supercontinents since at least two billion years ago (Ga), dubbed the supercontinent cycle, provides the first-order tectonic control on the evolution of the Earth System including episodic orogenic events, mineralization, the formation and closure of oceans and superoceans, and even the evolution of life. However, the lifespan of the supercontinents appears to decrease with time, from ~300 million years (Myr) for Nuna/Columbia, to 200–250 Myr for Rodinia and ~150 Myr for the youngest supercontinent Pangaea. To understand what caused such a secular decrease in supercontinental lifespan, we conduct 3-D geodynamic modeling using realistic tectonic settings. The results show that the yield stress of newly formed orogens during the assembly of a supercontinent provides the dominant control on the lifespan of the supercontinent, implying that the yield stress of young orogens becomes lower with time. We hypothesize that the decreasing mantle temperature due to Earth's secular cooling might have caused new orogens to become weaker.

1. Introduction

Geological and paleomagnetic evidence suggests the recurrence of supercontinental events over the past 2 billion years at an interval of every ~600 Myr [1–5]. Each such event involves intense and widespread global orogenesis during the continental assembly stage, including continental collisions [6], followed by widespread magmatic and rifting events, commonly involving plume activities, during the break-up of the supercontinent [7,8]. Such episodic global geodynamic events are accompanied by sea level changes [9–12], cooler climatic episodes [13], biosphere upheavals [9,14], and episodes of mineralization. Excluding megacontinents like Gondwana that commonly assemble first as part of the supercontinent assembly [15], the three widely accepted supercontinents show an intriguing decreasing trend in their lifespans when the birth of a supercontinent is taken at the time when over 75% of all known continental areas at the time were together [3,16], and the break-up time of a supercontinent is taken as the time when the largest remaining continental mass during the break-up phase of a supercontinent becomes <75% of global continental area: 300–400 Myr for the Mesoproterozoic supercontinent Nuna/Columbia [17–20], 200–250 Myr for the early Neoproterozoic supercontinent Rodinia [21–23], and ~150 Myr for Pangaea [3,4,24]. Such a systematic change has direct implications to related global tectonic, climatic and biological events and evolution, and likely reflects fundamental secular changes in the Earth System. However, so far little is known as to what factor(s) controlled such a systematic change.

2. Model setup

In this study, we build 3-D spherical geodynamic models to simulate the mantle-plate coupling process during the tenure of a supercontinent until its break-up, using realistic plate tectonic settings. We then use these models to examine factors that might influence the lifespan of a supercontinent. Each model starts from a global temperature field featuring a degree-1 mantle structure from pure thermal convection calculations [25], consisting of a super-downwelling and an antipodal super-upwelling [25,26], with continents randomly placed on the surface of the super-downwelling hemisphere (Fig. S1 and Methods in Supplementary materials online). From such a common initial mantle structure, the models then execute self-consistently with a planted thermochemical layer at the bottom of the lower mantle. During the model evolution, the originally scattered continents are able to first assemble to form a supercontinent, driven by the degree-1 mantle convection. Oceanic subduction surrounding the supercontinental margin (i.e., the subduction girdle [27]) then starts to progressively transform the super-downwelling under the newly assembled supercontinent into a super-upwelling, leading to the formation

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of a degree-2 mantle structure [25]. Two large thermo-chemical piles (i.e., the seismic tomography-observed large low shear velocity provinces (LLSVPs) [28]), either beneath or antipodal to the supercontinent, form during this mantle transition process. The sub-supercontinent super-upwelling [29–31], along with retreating subduction slabs [32–34] (Movie S1 online), eventually drives the break-up of the supercontinent. Such a transient mantle structure and dynamic evolution, which encompasses the assembly and break-up of a supercontinent as previously hypothesized from 2-D geodynamic modeling [35] or geological/geophysical observation-based syntheses [1,29], is for the first time realized in global self-consistent 3-D modeling here.

Continents in our models are considered as chemically-distinct materials which are stronger (i.e., with higher yield stress) and more buoyant than the oceanic lithosphere (Table S1 online). Five continents with a total area covering ~30% of Earth's surface are used. Weaker continental orogens (Table S1 online) are automatically generated when two converging continents become close enough (i.e., within 300 km). Once this happens, we extend the orogen along strike to cover the adjacent regions where the distance between the opposing continental margins is <600 km (Fig. S2a and Methods in Supplementary materials online). Minor isolated gaps of <1500 km in dimension between the fully assembled continents are also converted to orogens once the supercontinent assembly is completed (Fig. S2b online), mimicking the accretion of minor terranes along orogens during supercontinent assembly [6,36]. The relatively weak orogens will act as a stress guide, promoting the future supercontinent break-up [37]. The time-averaged viscosity profile of orogens from the formation of a supercontinent to its break-up is given in Fig. S3a (online), along with that of continental and oceanic lithosphere for comparison.

Pseudo-plastic yielding of the modeled oceanic lithosphere (see Methods in Supplementary materials online), which simplifies the brittle and plastic deformation into instant yielding (i.e., time-independent), allows for dynamically induced Earth-like ocean-ocean or ocean-continent subduction to occur in our models [38]. In order to enable density instability-induced ocean-continent subduction, we automatically create low viscosity (weak) zones along continental margins [39,40] when the adjacent oceanic lithosphere along such margins becomes older than 200 Ma (the time when spontaneous subduction can occur due to negative buoyancy of the cold oceanic lithosphere).

3. Results

Fig. 1 shows four critical evolutionary stages for the reference case (Case 1), where the yield stress of orogens (σ_{orogen}) is 100 MPa. The initial, semi-randomly distributed, cratonic continents in this model (Fig. 1a) move toward the center of the super-downwelling (longitude/latitude at around 0°/0°; Fig. 1e) after the model starts. At ~100 Myr, the average velocity of the continents' motion toward the center of the upcoming supercontinent ($v_{\mathrm{toward\_center}} = \delta d_{\mathrm{cratons}}/\delta t$, where δd is the finite change of the mean distance between the geometric centers of the continents, and δt is the finite time interval), which is used for distinguishing continental assembly (i.e., $v_{\mathrm{toward\_center}} < 0$) from dispersal (i.e., $v_{\mathrm{toward\_center}} > 0$), changes from negative to zero (Fig. S3b online). It marks the completion of the supercontinent assembly. At that time the remaining minor oceans inside the supercontinent are converted to orogens (beige in Fig. 1c). After that, the $v_{\mathrm{toward\_center}}$ changes from zero to positive (Fig. S3b online).

Because the orogens are weaker than the continental cratons (Table S1 online), continental break-up preferentially occurs along such weak orogens [37]. At ~275 Myr, the continental blocks are visibly dispersed, with ~15% of the orogens replaced by upwellings at depths below 20 km (Fig. 1d). We take this 15% crustal conversion ratio as the threshold marking the break-up of the supercontinent. Deep subduction is well developed around the supercontinent by that time (Fig. 1h; Movie S1 online) which is coupled with the formation of two partially connected antipodal LLSVP-like thermo-chemical piles, one beneath the supercontinent and the other beneath the superocean (Movie S1 online; note that the evolution of plume-like features can be found in (a)–(c) of Movie S1 online). From the supercontinent formation at ~100 Myr till its break-up at ~275 Myr, the supercontinent survived for ~175 Myr in the reference model (Case 1).

3.1. Strength of orogen vs. supercontinental lifespan

We first test the influence of changing yield stress for orogens (σ_{orogen}) on the lifespan of a supercontinent. In Cases 2 and 3, we changed σ_{orogen} from 100 MPa for Case 1 to 50 and 150 MPa, respectively, but kept other parameters the same (Table 1). The resulting supercontinent survival time in the two cases are 102 and 194 Myr (Fig. S4 online), respectively, which are substantially different (i.e., deviations >10%) from that of the reference case (175 Myr). These two tests, combined with the reference case, suggest a positive correlation between supercontinental lifespan and the strength of young orogens that join the cratons to form the supercontinents. Model outcomes with σ_{orogen} ranging from 50 to 300 MPa at 50 MPa steps (Cases 1–6 in Table 1) are all plotted in Fig. 2, illustrating a positive, near linear trend between supercontinental lifespan and the strength of orogens. The longest modeled supercontinental lifespan is 272 Myr with σ_{orogen} = 300 MPa, which is comparable with the life span of Nuna as recently defined by an updated paleomagnetic analysis [20] (~300 Myr). In comparison, the shortest lifespan is ~100 Myr when σ_{orogen} = 50 MPa (Fig. 2).

3.2. Model sensitivity to other parameters

Since the thermo-chemical layer at the bottom of the mantle may have evolved with time in both its density and thickness, we test here (Cases 7–10 in Table 1; Fig. S5 online) if changing these two parameters would cause changes in the lifespan of the supercontinents. Fig. 2 inset shows that increasing or decreasing the density ($\Delta\rho_{\mathrm{tcl}}$), or the thickness ($D_{\mathrm{tcl}}$), of the thermo-chemical layer produces negligible impact on the model results (within ± 10 Myr of the reference case).

In Case 11 (Table 1), we examine if varying the mantle's radioactive heating rate would impact the lifespan of a supercontinent through more vigorous mantle convection. By increasing the internal heating rate by ~1/3 (e.g., a hotter mantle as in Nuna time) from the reference case (Pangea time) [41] but ignoring the potential impact of a hotter mantle on the strength of young orogens, the model only reduced the lifespan of the supercontinent by 5 Myr, demonstrating the insensitivity of supercontinent survival time to the mantle's heat-controlled convection vigor (Fig. S6 online).

We further examine, in Case 12, model sensitivity to the strength of the oceanic lithosphere (σ_{ocean}), which determines the overall effective viscosity of oceanic lithosphere. With different strengths, the size of subducted oceanic lithosphere changes [42], which may result in a varying driving force acting on the dispersal continents through subduction retreat. However, by increasing σ_{ocean} from 125 MPa in the reference case by 40% to 175 MPa, the lifespan of the supercontinent reduces by only ~6 Myr from the ~175 Myr for the reference case (Table 1; Fig. S6 online). This implies that the break-up of a supercontinent is mostly controlled by the push force of the sub-supercontinent plumes other than the pull force of subduction retreat [31,39].

Overall, our test cases show that the density and thickness of the lower mantle thermos-chemical layer, the internal heating rate

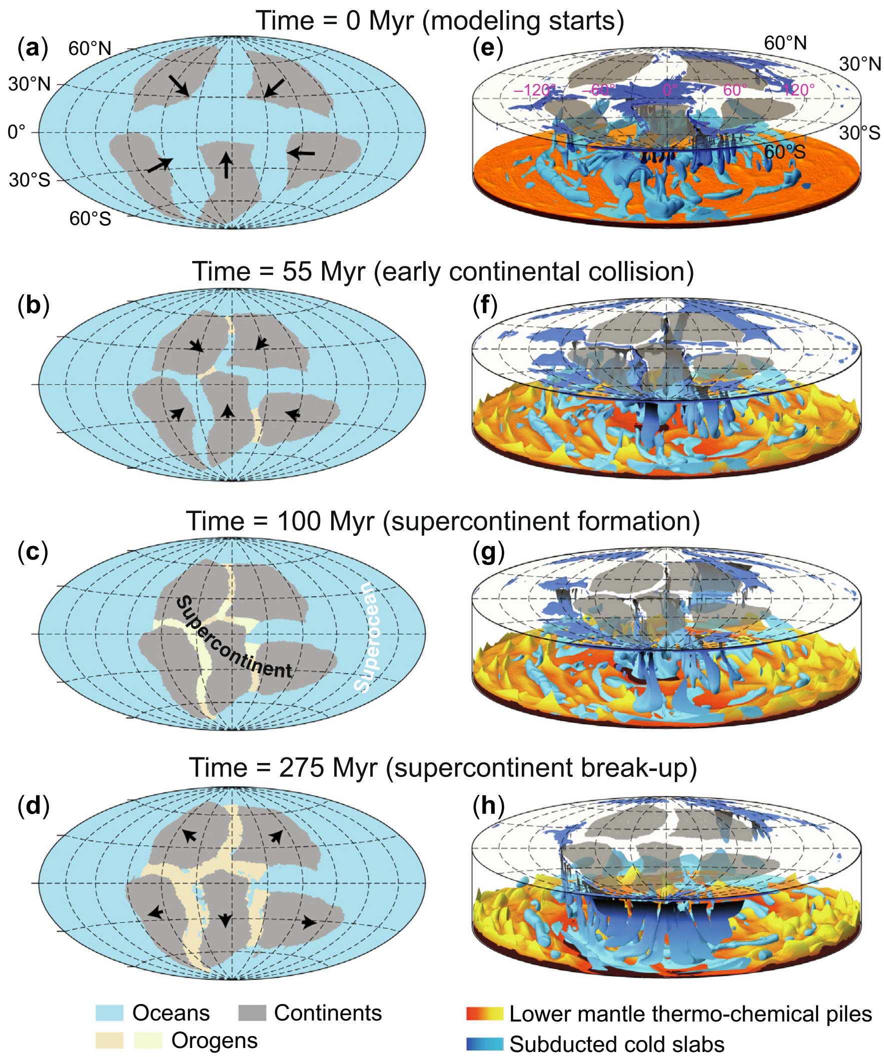

Fig. 1. Evolutionary results for Case 1 at four critical time points. (a)–(d) Continental drift and orogenesis during the supercontinent assembly and its eventual break-up. The orogens are converted to new oceans when materials at 20-km depth are replaced by upwelling mantle during continental break-up. Black arrows in (a), (b), and (d) denote the motion directions of the continents. (e)–(h) The evolving mantle structure, including cold downwellings (dark blue to blue) and the undulating lower mantle thermo-chemical layer (red to yellow).

Table 1
Results of test cases.

Case No.	Thermo-chemical layer thickness (D_{tcl}, km)	Thermo-chemical layer extra density ($\Delta\rho_{tcl}$, kg m^{-3})	Yield stress of orogens (σ_{orogen}, MPa)	Yield stress of oceanic lithosphere (σ_{ocean}, MPa)	Non-dimensional internal heating H	Assembly/ break-up time (Myr)	Lifespan ($t_{lifespan}$, Myr)
1	250	30	100	125	85	100/275	175
2	250	30	50	125	85	106/208	102
3	250	30	150	125	85	113/307	194
4	250	30	200	125	85	112/335	223
5	250	30	250	125	85	112/353	241
6	250	30	300	125	85	106/378	272
7	250	60	100	125	85	104/277	173
8	250	15	100	125	85	107/275	168
9	550	30	100	125	85	114/295	181
10	100	30	100	125	85	108/284	176
11	250	30	100	125	114	178/348	170
12	250	30	100	175	85	178/347	169

of the mantle, and the strength of the oceanic lithosphere, all show minor influences on the model outcomes (±10 Myr). The supercontinental lifespan is most sensitive to the yield stress of young orogens with a clear near-linear positive correlation, suggesting that the lifespans of supercontinents are predominantly controlled by the strength of the supercontinent-forming orogens.

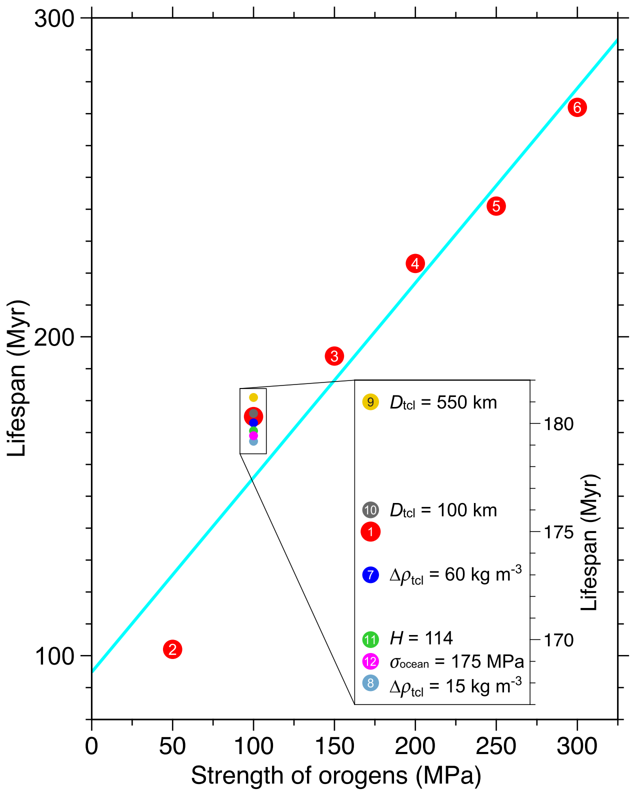

Fig. 2. Relationship between the yield stress of supercontinent-forming young orogens and the lifespan of the supercontinents. Numbered red dots are tested cases (Table 1) with different σ_{orogen}, with the cyan line being a linear fit for the red dots using least squares regression. Dots with colors other than red are the results of cases with variable parameters as annotated in the zoom-in panel, with numbers on the dots showing respective case numbers as in Table 1.

4. Discussion

Our work suggests that the strength of orogens that formed the supercontinent determines the lifespan of the supercontinent. The observation that younger supercontinents have markedly shorter lifespans implies that orogens that formed during a younger supercontinent assembly are weaker than those formed during older assemblies. Although the exact causes for the weakening of supercontinent-forming young orogens with time is unclear, we speculate that the most likely cause might be Earth's secular cooling. In a secularly cooled Earth after the Archean time, a gradually lowering mantle potential temperature [43] likely leads to a lower degree of partial melting in orogenic-related processes (e.g., melting related to slab break-off, and orogenic collapse, etc.), thus thinner crust due to the reduced magmatic underplating, and weaker crust due to less stitching plutons. In fact, the entire lithosphere of young orogens would become thinner as the Earth cools, through the same mechanism as that for modeled thinning oceanic crust [44] and lithosphere [45] with time due to Earth's secular cooling. That is because a cooler mantle would produce lower-degree partial melting at the sub-orogen asthenosphere, thus producing less melts to build up the crust and leaving less depleted asthenospheric mantle to build the mantle lithosphere. A modern-day analogy would be the contrast between the lithospheric thickness of plume-induced oceanic plateaus (thicker lithosphere produced by higher degrees of mantle melting) and that of normal oceanic lithosphere produced under normal mantle temperature. In addition, less melting in the orogenic crust and lithospheric mantle during more recent time would also lead to more water being left in the system, further weakening the young orogens. Nonetheless, future observational, theoretical and modeling work is required to further examine the causes for the weakening of young orogens with time.

The model predicted weakening of young continental orogens with time might be seen as contradicting to results deduced from evolving continental geotherm [46,47] which suggest that the integrated strength in the continental lithosphere increases with time because the temperature at the base of the continental crust reduces as the mantle cools. We note that although such a conclusion is probably applicable to the old cratonic continental lithosphere, it may not be applicable to young orogens which would take time to reach thermal equilibrium by slowly rebuilding the mantle lithosphere.

The widely accepted three supercontinents discussed here does not include megacontinents like Gondwana, or highly disputed "supercontinent" like Pannotia [48–50] the existence of which has been challenged on both geological and paleomagnetic grounds (e.g., [3]). Previous 3-D geodynamic modeling already demonstrated that the mantle dynamics when there is a globally singular supercontinent are very different from the time when there are two or more megacontinents [25,51].

5. Conclusions

Our geodynamic modeling suggests that the lifespan of a supercontinent is predominantly controlled by the strength of orogens that formed the supercontinent. The secular weakening of young orogens due to Earth's secular cooling caused a stepwise reduction in the lifespan of supercontinents: ~300 Myr for Nuna, 200–250 Myr for Rodinia, and ~150 Myr for Pangea. Our modeling predicts a lifespan for the future supercontinent, Amasia [52], to be <150 Myr, possibly around 100 Myr. Future work can further examine the impact of such reducing supercontinental lifespan on other aspects of the Earth System, including tectonic cycles, climatic changes, and life evolution.

Conflict of interest

The authors declare that they have no conflict of interest.

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Author contributions

Zheng-Xiang Li and Chuan Huang conceptualized the study. Chuan Huang and Zheng-Xiang Li designed and performed the numerical model. Zheng-Xiang Li and Chuan Huang wrote the manuscript.

Appendix A. Supplementary materials

Supplementary materials to this article can be found online at https://doi.org/10.1016/j.scib.2023.01.035.

Data availability

All data and resources that are necessary for evaluating or reproducing the findings of this study (including source code, data for figures, etc.) are available at https://data.mendeley.com/datasets/7t8k3gbbn5.

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