Understanding Cratonic Flood Basalts

Paul G. Silver\textsuperscript{1}, Mark Behn\textsuperscript{1,2}, Katherine Kelley\textsuperscript{1}, Mark Schmitz\textsuperscript{1,3}, and Brian Savage\textsuperscript{1}
\textsuperscript{1}Carnegie Institution of Washington, DTM
\textsuperscript{2}Now at Woods Hole Oceanographic Institution
\textsuperscript{3}Now at Department of Geosciences, Boise State University

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Abstract
The origin of continental flood basalts remains a mystery. These eruptions often produce millions of cubic kilometers of basalt on timescales of only a million years. Although flood basalts are found in a variety of settings, no locale is more puzzling than cratonic areas such as southern Africa, where strong, thick lithosphere is breached by these large basaltic outpourings. Conventionally, flood basalts have been interpreted as melting events produced by one of two processes: 1) elevated temperatures associated with mantle plumes and/or 2) adiabatic-decompression melting associated with lithospheric thinning. In southern Africa, however, there are severe problems with both of these mechanisms. First, the rifting circumstances of several well-known basaltic outpourings clearly reflect lithospheric control rather than the influence of a deep-seated plume. Specifically, rift orientation is controlled by preexisting lithospheric mantle fabric revealed by seismic anisotropy, and rift timing is correlated with stress perturbations to the lithosphere associated with nearby collisional orogens (collisional rifts). Second, the substantial lithospheric thinning required for adiabatic decompression melting is inconsistent with xenolith evidence for the continued survival of thick lithosphere beneath flood basalt domains. As an alternative to these models, we propose a new two-stage model that interprets cratonic flood basalts not as melting events, but as rapid drainage events that tap previously-created sublithospheric reservoirs of molten basalt formed over a longer time scale. Reservoir creation/existence (Stage I) requires long-term (e.g. >>1 Ma) supersolidus conditions in the sublithospheric mantle, which could be maintained by an elevated equilibrium geotherm (appropriate for the Archean), a slow thermal perturbation (e.g. thermal blanketing or large-scale mantle upwelling), or a subduction-related increase in volatile content. The drainage event (Stage II) occurs in response to an abrupt stress-induced increase in the lithospheric fracture permeability to melt. Such a model accounts for the rapidity of flood basalts, the evidence for lithospheric control of rifting, and the continued survival of cratonic lithosphere.

Introduction
From time to time, the Earth’s surface is pierced by tremendous outpourings of basalt, the primary melt product of the Earth’s mantle. These so-called flood basalts are unlike the basalts from arc volcanoes that decorate convergent plate boundaries, nor are they like basalts that are erupted along the globe-encircling length of mid-ocean ridges. They are indeed more akin to “floods”, characterized by eruptions of millions of cubic kilometers of basalt that can spread over a large fraction of a continent with great rapidity – often in less than one million years. Flood basalts occur in both continental and oceanic settings, where they are called continental flood basalts and oceanic plateaus, respectively. They are often associated with continental breakup, as in the case of Parana, Karoo, and Deccan flood basalts accompanying the breakup of Gondwanaland. Others are associated with rifting in stable cratons that does not lead to breakup, such as the Mid-Continent Rift in North America, or the Ventersdorp Rift of southern Africa. Their occurrence is not predicted by plate tectonics and consequently their origin remains enigmatic. What causes them to occur? How do they erupt so quickly? Where does the basalt come from?
Much of the thinking on the creation of flood basalts is based on the classic paper of White and McKenzie [1], which interprets flood basalts as melting events that are caused by one of two processes: (i) a high-temperature plume that serves to replace subsolidus sublithospheric mantle with supersolidus mantle, or (ii) the thinning of the lithosphere, thereby reducing the pressure of mantle material that is sitting just below the base of the lithosphere (Fig. 1a,b). The first of these is inspired by the plume hypothesis [2,3], that such outpourings are the result of deep, high-temperature plumes, while the second is an adaptation of the adiabatic-decompression melting model that so successfully accounts for the formation of mid-ocean-ridge basalts [4]. In cratonic environments, the necessary lithospheric thinning could be caused by either stretching of the lithosphere [1], or through some form of convective instability and lithospheric delamination, such as that advocated for Tibet (e.g. [5, 6]). The latter process would account for regions where there is little surface indication of the required lithospheric stretching needed for adiabatic decompression, although such models predict uplift following delamination.

Cratonic southern Africa represents a particularly stringent test of these two mechanisms. There have been several massive outpourings or intrusions of continental basalts in the Precambrian of southern Africa, and thanks to the results from the recent Kaapvaal Project [7], it is now possible to examine the characteristics of these magmatic events in unprecedented detail. This information is in the form of new analyses of crust and mantle xenolith suites (chronology, geothermobarometry) from southern Africa’s numerous kimberlite pipes, as well as new seismic constraints on crust and upper mantle provided by the associated Southern African Seismic Experiment.

Analysis of the rifting characteristics of four magmatic events (Table 1, Fig. 2) spanning the time period 1.8-2.7 Ga – Ventersdorp (2.71 Ga), Great Dyke (2.57 Ga), Bushveld (2.06 Ga), and Soutpansberg Trough (1.88 Ga) – shows that they were closely related to collisional orogens in two important ways. First, the actual orientation of the rifts suggests that they have exploited pre-existing mechanical anisotropy in the lithospheric mantle that was produced by strain-induced lattice preferred orientation of olivine [8,9]. The alignment of olivine, hypothesized to be the source of the mechanical anisotropy and manifested in the fast polarization direction of split shear waves, primarily reflects the collisional deformation associated with the ca 2.9 Ga Kimberly and Pietersburg orogens along the western and northern margins of the Kaapvaal craton, respectively [10,11]. For all four magmatic events, the rift orientations are parallel to this mantle fabric. Indeed, there is increasing evidence that a close relationship exists between mechanical anisotropy and seismic anisotropy in other cratonic environments such as Australia [12] and the Canadian Shield [13], as well as in oceanic environments [14].

Second and particularly striking, the timing of these rifting events are virtually synchronous with subsequent major orogens (Table 1, Fig. 2b). The Neo-Archean Ventersdorp rift is temporally linked with compressional deformation and crustal thickening in the Central and Southern Marginal Zones of the Limpopo Orogen [15-17]. Similarly, the Great Dyke was emplaced during the exhumation of the Northern Marginal Zone of the Limpopo Orogen over the Zimbabwe Craton [18-20], and the Paleoproterozoic emplacement of the Bushveld-Molopo Farms Complex was nearly
synchronous with the Magondi Orogeny along the NW margin of the Zimbabwe-Kaapvaal Cratons [21-23]. Significant dextral transpressional reactivation of the Triangle and Palala Shear Zones of the Limpopo Belt are also recorded at this time [24-26]. Subsequent extensional reactivation of the Palala Shear Zone and formation of the Soutpansberg Trough coincided with compressional deformation in the Kheis Belt along the western Kaapvaal Craton [27-28]. We interpret such synchronicity as evidence that these magmatic outpourings were produced by collisional rifts, i.e. rifts that formed in the stress field produced by an adjacent collisional orogen, through extension perpendicular (or at a high angle) to the principle compressive stress axis. Other more recent examples of collisional rifts include the Rhine Graben, and the Baikal Rift [29-31].

This close temporal and spatial relationship between magmatic events and collisional orogens renders the plume explanation highly unlikely, as it would require that plume events just happen to occur with each major orogen. Rather, it appears that these magmatic events are caused by stress perturbations in a lithosphere weakened by mechanical anisotropy, not melting due to deep-seated, high-temperature mantle plumes.

It is also difficult to appeal to lithospheric thinning, either by lithospheric stretching or delamination, to produce adiabatic decompression melting beneath the southern African craton. The first issue is timing. Under the assumption that the million-year flood basalt time scale is a melting-event time scale (i.e. residence time of melt at depth is negligible), there are few processes that can realistically occur over such a short period of time. For example, pressure-release melting by lithospheric stretching requires stretching factors (or β factor in the terminology of [1]) of order two. Thus, to produce large volumes of melt in 1 My would require a strain rate of $10^6$/yr, which is an order of magnitude higher than the strain rates of present-day rifts ($~10^7$/yr or less). This includes the most prominent extensional environments active today, such as the Baikal Rift, East African Rift, and Basin and Range. (Baikal Rift: $0.3 \times 10^7$/yr, [32]; East African Rift: $0.5 \times 10^7$/yr [33], Basin and Range: $<1 \times 10^7$/yr, [34]). On the other hand, it is more difficult to reject delamination on the basis of timing. While [1] argued that this process is too slow, others suggest that it can occur as rapidly as a few million years [5,6].

Even if lithospheric delamination could conceivably occur in a sufficiently short period of time, however, it still must be determined if delamination indeed has occurred in the geological setting of interest. The most direct means of testing for delamination in cratonic environments is to study the geothermobarometry and chronology of mantle xenoliths from kimberlite pipes to determine whether the mantle portion of the lithosphere was removed at the time of magmatism. These xenoliths provide an age – either of basalt depletion, or in the case of diamond growth, the age of diamond formation, and an estimate of the P, T conditions. This represents the state of the mantle as it was at the time of kimberlite eruption, which, in our case, was well after the magmatic events. If the mantle lithosphere was indeed removed, then there should be no mantle xenoliths with dates prior to the magmatic event in question. In the case of diamond formation ages from silicate or sulfide diamond inclusions, diamond formation should only occur 100-300 My after the hypothesized delamination event, which is the
time necessary to regrow the lithosphere by simple convective cooling, and thus return the lithospheric mantle to the diamond stability field.

Data collected in Southern Africa during the Kaapvaal project permits such a test. For two of the magmatic events, the Bushveld (2.06 Ga) and Ventersdorp (2.71 Ga), there are kimberlite pipes that are near or in the source zone of the basaltic intrusions. For the Bushveld, there are two studied kimberlites on the northern and southern edge of the intrusion, Klipspringer and Premier, respectively, and one, Palmietgat, which is within the Bushveld. The most extensive data set is from the Premier mantle xenolith suite. A study of Re-Os minimum depletion ages, $T_{RD}$, [35, Fig. 2] reveals a range of ages, most of which in fact predate the Bushveld. The available xenoliths with well-constrained pressures do not reveal any younging with depth as might be expected from delamination. Indeed the deepest xenolith (7 GPa) is the oldest, with a $T_{RD}$ of 3.5 Ga. For the Klipspringer pipe, sulfide diamond inclusions yield Re-Os model ages of 2.55 Ga [36], significantly earlier than the Bushveld intrusion. Finally, the Palmietgat pipe has thus far yielded 4 sulfide inclusion diamond ages, one of which is 3.0 Ga. [37]. Regarding the Ventersdorp, there are numerous 2.9-3.2 Ga diamond formation ages for the nearby mines at Kimberley [38,39], which, as in the case of the Bushveld, predate the Ventersdorp event.

From the evidence available from this extensive xenolith data set, we conclude that the occurrence of delamination is highly unlikely. Indeed, it is very difficult to find any evidence for delamination from the xenolith record of any cratonic environment worldwide. The only clear evidence for delamination is for the North China Craton [40]. Delamination has also been invoked for the Wyoming craton [41] although more recent data in fact argues for lithospheric preservation [42]. We thus conclude that the presently accepted mechanisms for flood basalts – mantle plumes and lithospheric thinning - are incapable of accounting for the southern African Precambrian magmatic events discussed above.

**A New Model for Cratonic Flood Basalts**

We propose an alternative model for cratonic flood basalts that is compatible with a collisional rifting origin, leaves the cratonic lithosphere intact, and satisfies the short 1 My time scale for these events. The basic point of departure is that a flood basalt is not a melting event, but instead constitutes the *drainage* of a pre-formed sublithospheric molten basaltic reservoir. The million-year time scale is then a reservoir-exhaustion time scale, rather than a melt-production time scale. The magmatic event is the result of an abrupt, stress-induced increase in the melt permeability of the lithosphere through the formation of narrow lithospheric fractures that subsequently become dikes. That such lithospheric fractures can form near the base of the lithosphere is dramatically demonstrated by kimberlite eruptions that can originate from depths in excess of 200km. The initial stress perturbation can have many causes, although collisional rifting is probably the dominant mechanism for rifts that do not lead to continental breakup. We thus propose a two-stage model that decouples the process of magma formation from magma transport to the Earth’s surface. This decoupling has several desirable features. First, the formation and maintenance of a magma reservoir can occur over periods of time
much longer than 1My, which significantly broadens the class of allowable formation mechanisms. Second, this model leaves the lithosphere intact since narrow lithospheric dikes are unlikely to disrupt the lithosphere.

Stage I: Formation/Maintenance. We hypothesize that a reservoir of molten basalt is formed at the base of the lithosphere and is maintained over extended periods of time (long compared to 1 My). The presence of such a reservoir requires two conditions: that the sublithospheric mantle is often above the solidus, and equally important, that the lithospheric mantle and underlying molten reservoir are in equilibrium. This second consequence, in turn, requires a vertical discontinuity in both solidus temperature and in mechanical strength (Fig. 4). There is indeed good evidence for such a discontinuity beneath ancient cratons, since they appear to be both melt and volatile depleted from previous basalt-extraction events [43,44]. As an illustration of this discontinuity, we have used the MELTS algorithm [45,46] to calculate the melting behavior of both a fertile peridotite (MM3) and a depleted peridotite (DMM) representing the asthenosphere and lithosphere respectively. For a base-of-lithosphere depth of 180 km (6 GPa) and for a potential temperature of 1750ºC the fertile asthenosphere yields a >10% melt fraction while the deleted lithosphere does not melt. If the lithospheric base is at 240 km (8 GPa), then a potential temperature of 1850ºC leads to the same contrast in melt behavior. Regarding lithospheric strength, Sleep [47] has estimated that the lithosphere must be at least 20 times more viscous than the underlying mantle in order for cratonic lithosphere to avoid entrainment over billions of years. In fact, the actual viscosity contrast of cratonic lithosphere is likely much higher. One study of the effect of devolatilization [48] suggests that this contrast can be as high as 500, apparently more than enough to stabilize the lithosphere.

Probably the most significant consequence of this molten-reservoir model is that the subcratonic mantle often remains above the solidus. This raises two questions: (i) Is there any evidence for such high base-of-lithosphere temperatures in the present or the past, and (ii) what is the possible source of this heat? Some information on recent cratonic geotherms is provided by the geothermobarometry applied to mantle xenoliths from the large number of Cretaceous kimberlites that erupted in southern Africa during the breakup of Gondwanaland. A recent compilation by James et al. [49] based on low-temperature on-craton nodules reveals a roughly linear gradient from 0ºC at the surface to about 1300ºC at a depth of about 200 km (Fig. 5). A major source of uncertainty, however, is the depth to the base of the chemically distinct lithosphere. If it extends to 250 km, a number that is compatible with recent seismic tomography beneath southern Africa [50], this linear gradient implies a base-of-lithosphere temperature of about 1550ºC. The high-temperature xenoliths, however, suggest an inflected, rather than linear geotherm, with a steeper thermal gradient near the base of the craton. There have been a variety of interpretations of this inflected geotherm that are permitted by the data. One interpretation is that high temperature melts have infiltrated the lithosphere, effectively advecting heat into the base of the lithosphere. Since these melts had to have come from greater depth, it implies that the sublithospheric mantle is at least partially molten despite the apparently subsolidus temperature for fertile dry mantle at the top of the asthenosphere. A second interpretation is that this inflection represents a
disequilibrium conductive geotherm resulting from a thermal perturbation applied to the base of the lithosphere. With this interpretation, the linear gradient implied by these high-temperature xenoliths can be extrapolated to 250 km depth, giving a base-of-lithosphere temperature of 1850ºC (Fig. 5). We note that this extrapolated temperature is near the fertile-mantle dry solidus at base-of-lithosphere depth. A third interpretation of this inflected geotherm is that it reflects the apparently strong temperature dependence of thermal conductivity [51], which would predict an equilibrium inflected geotherm. The nearly factor-of-two increase in thermal gradient suggested by the high-temperature xenoliths is broadly consistent with the predicted temperature dependence of this parameter. We note that all three interpretations would suggest the presence of partial melt at sublithospheric depths. In the Archean, it is possible to have had an equilibrium geotherm with a semi-permanent supersolidus asthenosphere, given the evidence for higher sublithospheric Archean mantle temperatures. Estimates of this increase, based on the chemistry of komatiites, suggest a value between 100º and 500ºC, depending on the assumed komatiite source region ([52] and references therein). Finally, it is well established that water addition can dramatically reduce the peridotite solidus at high pressures by more than 500ºC [53,54], so that a volatile-rich asthenosphere could actually be supersolidus at a significantly lower temperature. Indeed, the collision that created the collisional rift presumably signifies the termination of an extended phase of subduction, which might be responsible for increasing the volatile content of the asthenosphere in adjacent regions.

Given the range of permissible time scales for reservoir formation and maintenance, there are several phenomena that could significantly elevate the temperatures at the base of the lithosphere. For example, at the long-term end, thermal blanketing [55] can increase base of lithosphere temperature by hundreds of degrees in about 300 My. Given the large thermal inertia of thick lithosphere, this could still lead to a perturbed geotherm similar to the ‘kink’ in Figure 5, based on a simple analysis of one dimensional heat flow. Similarly, the process of advective thickening of the lithosphere, thought to be the way in which thick cratons are initially formed [11,43], should produce a decrease in heat loss through the lithosphere, and a corresponding increase in the temperature of the sublithospheric mantle, assuming the heat entering the sublithospheric mantle from below is unchanged. There may be large-scale thermal anomalies in the deep mantle that ultimately reach the base of the lithosphere, such as those inferred to be associated with lower mantle low velocity anomalies beneath southern Africa [56]. Finally, there might be a continual rain of ‘failed’ plumes that deposit their heat at the base of the lithosphere because they are unable to penetrate through cratonic lithosphere.

Stage II: Drainage. The actual eruption of basalt onto the surface is interpreted as the drainage of the basaltic reservoir established in Stage I, through the creation of localized lithospheric fractures. Under normal circumstances the overlying lithosphere is effectively impermeable to this reservoir of magma sitting below it (Fig. 1). While it is true that island arc basaltic magmas have very short residence times, as determined by uranium series observations (e.g. [57]), these regions are always tectonically active and have very thin lithosphere directly beneath the location of the arc. This environment differs markedly from a stable cratonic environment, characterized by very thick
lithosphere and the virtual absence of tectonic activity except for participation in occasional collisions. We assume that lithospheric fractures only form in response to perturbations in lithospheric stress due to one of these collisions. The million-year eruption time for flood basalts is then interpreted as the time necessary to drain this sublithospheric reservoir. Given the $10^6$ km$^3$ volume of flood basalts, and assuming this represents a 10% melt of a peridotite source, then the partial melt region (melt plus residue) is $10^7$ km$^3$. Assuming the reservoir is in the shape of a disk with thickness ranging from 100 to 10 km, corresponding to a range of disk radii of 178-564 km, the required melt migration velocity to completely drain the reservoir would be $18-56$ cm/yr. This number is consistent with estimates for melt migration near spreading ridges, which are at least 100 cm/yr [58], assuming that the flow through the dike is the rate-limiting step. At this rate, a system of dikes a kilometer wide and total length of 1,000 km would drain such a reservoir in $10^6$ year. Finally, these fractures are unlikely to disrupt the lithosphere because they are highly localized. As discussed by Rubin [59], through-going lithospheric dikes have to reach a minimum critical width of only a few meters for sublithospheric magma to travel to the surface. Below this width, the walls of the dike freeze out the magma, thus blocking the propagation of magma through the fracture. This critical thickness is sufficiently small to ensure that the thermal perturbation to the lithosphere from the passage of basaltic magma is indeed sufficiently localized to have a negligible effect on the overall strength and stability of the lithospheric mantle. Even if the dike were continually active over 1 My, the thermal perturbation would conduct only a few kilometers into the host rock, assuming a thermal diffusivity of $10^6$ m$^2$/s, so that a larger-scale perturbation to the lithosphere is unlikely. On the other hand, thermal perturbations of the crust are certainly possible, given the much lower melting temperature of crustal rocks. Indeed Schmitz and Bowring [60] noted the remarkable contrast between a thermally perturbed, extended crust overlying an apparently physically undisturbed mantle lithosphere in response to the Ventersdorp flood basalt event.

Discussion
There are geochemical consequences of this two-stage model that should be reflected in basaltic melts. Melt chemistry is a function of both temperature and pressure, and can thus be used, in principle, to constrain the depth range of mantle melting. Indeed, this sensitivity has been exploited to invert for the melt column (minimum and maximum depth of melting) using both trace and major elements. For example [1] used the relative abundances of the heavy rare-earth elements (HREE) in erupted lavas. Garnet, which is stable in the mantle above about 3 GPa, has a strong preference for the HREE, and thus melts created in the garnet stability field record a HREE signature indicating some fraction of the melt originated below about 90 km depth. Beyond this constraint, however, HREE modeling is less reliable, because we presently lack comprehensive partitioning and melting constraints over the range of temperatures, pressures, and compositions appropriate for the asthenospheric and lithospheric conditions. This limitation is illustrated by the estimated minimum melting depth for the Ventersdorp flood basalts of ~30 km depth by [1]. This estimate, effectively requiring complete removal of the lithospheric mantle, is in direct conflict with the xenolith record, as discussed earlier. Major element inversions [61,62], based on FeO and Na$_2$O
concentrations, appear to hold greater promise. FeO constrains the maximum depth of melting by exploiting the pressure dependence of FeO-partitioning in olivine. Na$_2$O, on the other hand, is simply incompatible and consequently its concentration is inversely proportional to the total melt fraction. Combining these two provides estimates of melt-column depth and potential temperature. Application of this methodology to Basin and Range basalts [62] yields minimum depth estimates that, for this area, are consistent with independent estimates of the depth to the base of the lithosphere. They also find inferred temperatures of 1425ºC in the west beneath thin lithosphere, and much higher temperatures in excess of 1700ºC in the east beneath thicker lithosphere. This is a promising methodology that could be used in cratonic environments as well. At present, however, this approach has yet to be calibrated above 4GPa, which is a necessary step in evaluating the hypothesis of melts originating at pressures that are nearly double this value. We nevertheless note that the Ventersdorp basalts have higher concentrations of FeO than average mid-ocean ridge basalts (at 8 wt.% MgO: ave. FeO$_{ven}$=~11 wt.% [63], ave. FeO$_{MORB}$=~9.1 wt.% [64], which is consistent with a deeper origin.

This two-stage model for flood basalts implies that in the past much of the sublithospheric cratonic mantle may have been partially molten. Indeed, if in the Archean the geotherm was high enough to be above the fertile-peridotite solidus, then there may have even been a global partially molten layer. An important question to ask is whether such layers exist today in isolated pockets or ‘puddles’ beneath the lithosphere. Clearly, cratonic flood basalts have continued to occur since the mid-Proterozoic. The best known flood basalts unrelated to continental breakup in cratonic areas are the basaltic eruptions associated with the Neo-Proterozoic (1.1 Ga) Mid-Continent Rift in North America, and the Permian (250 Ma) Siberian flood basalts. This latter event is arguably the largest of all known flood basalts, and its occurrence in the recent geologic past suggests that such melt zones may indeed persist to the present. If so, they should be detected seismically as thin low velocity zones beneath the lithosphere. The detection of low velocity zones is a notoriously difficult problem in seismology, since downgoing waves would not bottom in the zone. Nevertheless, we note that in the last decade, there have been a variety of studies that provide evidence of reduced seismic velocity between the base of the lithosphere and the 410-km discontinuity, that suggest the presence of melt in various parts of the world: China, (e.g. [65]), Siberia, Southern Africa [66], the Arabian plate [67], western North America [68,69], and even globally [70]. We are actively developing methodologies that are optimized for the detection of such sub-lithospheric reservoir zones.

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Figure Captions

Figure 1. Schematics of various models for continental flood basalts. Red zones denote areas of elevated temperature or volatile content, pink zones are molten.
(a). Plume Model.
(b). Adiabatic-decompression-melting model.
(c). Two-stage formation/drainage model.

Figure 2a. Map of southern Africa showing major rift systems with basaltic magmatism: Great Dyke, Ventersdorp (surface exposures shown), Bushveld, and Soutpansberg shown in red, along with relevant geological features. Also shown shear-wave splitting fast polarization directions (from [71]) Abbreviations: TML: Thabazimbi Murchison Lineament, CL: Colesberg Lineament, SZ, shear zone. Green filled circles denote locations of kimberlite pipes, whose mantle xenoliths sample the mantle beneath the Ventersdorp and Bushveld. Ages of nodules are denoted in blue.

Figure 2b. Schematic illustrating development of collisional rifts and basaltic magmatism in southern Africa. Arrows show compression direction of orogen. Short lines show mantle fabric inferred from seismic anisotropy, which is either being formed (yellow) or fossilized (black). Red(blue) bars denote magmatic events during(after) activity. Top-left: Emplacement of the Ventersdorp supergroup due to the early phase of the Limpopo orogen (2.71 Ga). Top-right: Emplacement of the Great Dyke due to the late phase of the Limpopo orogen (2.57 Ga). Bottom-left: Reactivation of Limpopo structures by Magondi Orogen and the creation of a collisional rift that produced the Bushveld Instrusion (2.06 Ga). Bottom-right: Creation of the collisional Soutpansberg Rift by the Kheis Orogen (1.88 Ga). After Silver et al. [9].

Figure 3. Rhenium-depletion minimum ages of xenoliths as a function of pressure (top), as well as histogram of Premier ages (bottom). Most xenoliths predate the Bushveld intrusion, and there is no evidence of younging with depth as predicted by delamination hypothesis (data from [35]). Pressures in spinel stability field (less than 2 GPa) are
nominal values. Sheared nodules have no pressure estimate but are assumed to be deep and placed arbitrarily at 6 GPa.

Figure 4. Temperature (a), melt fraction (b) and viscosity (c) as a function of pressure for a compositional model with a basalt-depleted lithosphere, DMM, from 0 to 6 GPa and a sublithospheric fertile mantle, MM3, below 6 GPa, and assuming a base-of-lithosphere potential temperature of 1750°C. Note that MM3 has melt fraction of 10% at 6 GPa while DMM does not melt at this depth. A similar situation occurs at 1850°C for a base-of-lithosphere pressure of 8 GPa (not shown). Calculations were performed using the MELTS algorithm [45]. Viscosity contrast in (c) is assumed to be controlled by volatile depletion (rather than basalt depletion), where DMM is presumed to be “dry” and MM3 “wet”, with a water concentration of $10^3 H/10^6 Si$ [72]. The jump in viscosity at the base of the lithosphere is nearly an order of magnitude. Including the effect of basalt depletion on the homologous temperature would substantially increase this viscosity contrast through the temperature dependence of viscosity.

Figure 5. P-T estimates obtained from mantle xenoliths taken from the Archean Kaapvaal craton and adjacent Proterozoic mobile belts [49]. Blue circles represent low temperature xenoliths; red circles show high temperature xenolith data. Pressures are calculated after Finnerty and Boyd [73], and temperatures are calculated using both the O'Neill and Wood [74] (open symbols) and MacGregor [75] (filled symbols) methods. Straight lines show best fit regression lines through the low and high temperature data, respectively. The regression through the high temperature xenolith data predicts a mantle potential temperature of ~1750°C at the base of the lithosphere. Curved lines represent estimates of depleted-mantle solidus (solid line) [76] and fertile-mantle (dashed) solidus assumed to be 100°C less, based on MELTS algorithm [46]. Note that extrapolated geotherm from high temperature xenololiths crosses the fertile-mantle solidus at 250 km depth.
Table 1

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<td>2.06 [80,81]</td>
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* With respect to collisional axis of orogen.
A) Stretching

B) Adiabatic Decompression

Figure 1
C)

Two Stage

Stage I  Formation/Maintenance

Stage II  Drainage

Figure 1
Figure 2a
Figure 2b
Figure 5