Mesozoic denudation history of the lower Orange River and eastward migration of erosion across the southern African Plateau

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ABSTRACT

Topographic uplift of the southern African Plateau is commonly attributed to mantle causes, but the links between mantle processes, uplift, and erosion patterns are not necessarily straightforward. We acquired apatite (U-Th)/He (AHe) dates from eight kimberlite and basement samples from the lower reaches of the large westward-draining Orange River system with the goal of evaluating the roles of lithospheric modification and river incision on the erosion history here. Average AHe dates range from 79 to 118 Ma and thermal history models suggest that most samples are consistent with a main erosion phase at ca. 120–100 Ma, with some variability across the region indicating a complex erosion history. Major erosion overlapped with the timing of strong lithospheric thermochemical modification as recorded in xenoliths from the studied kimberlites, but the denudation pattern does not mimic the northward progression of lithospheric alteration across the study region. We attribute this area's denudation history to a combination of mantle effects, rifting, establishment of the Orange River outlet at its current location, and later faulting. When considering these results with other kimberlite-derived surface histories from an ~1000-km-long E-W transect across the plateau, an eastward-younging trend in denudation is evident. The interplay of mantle processes and the shape of the large, west-draining Orange River basin likely control this first order-pattern.

INTRODUCTION

The processes that control the topography and denudation of continental interiors are incompletely understood. While these regions are less tectonically active than those along plate boundaries, they comprise the majority of continental areas. The significance of these regions for global sediment fluxes and their importance in the coupling between erosion and climate cycles are debated (e.g., Willenbring et al., 2013; Warrick et al., 2014). Better quantification of long-term denudation patterns and insight into what drives erosion rate change are key for understanding the evolution of these regions. Buoyancy change in the mantle likely exerts stronger influence on surface uplift and topographic change here than at classic plate boundaries (e.g., Abbott et al., 1997; Pyskywvec and Mitrovica, 1998; Braun, 2010). However, the degree to which this topographic evolution is modulated by surface processes is an open question, because the relationships among mantle processes, lithospheric architecture, surface uplift, fluvial network evolution, and the erosional response is complex (e.g., Pazzaglia and Gardner, 1994; Guillou-Frottier et al., 2007; Molin et al., 2012). Southern Africa is an example of a cratonic interior region that was elevated from sea level to >1000 m elevation in post-Paleozoic time while distal from convergent plate boundaries and with little surface deformation (Fig. 1A). The timing and mechanisms of surface uplift are debated, but proposed mechanisms are mostly mantle-induced (e.g., Lithgow-Bertelloni and Silver, 1998; Nyblade and Sleep, 2003; Burke and Gunnell, 2008). Thus, it is an ideal location to study the links between mantle processes and erosion. Apatite fission-track (AFT) and more limited apatite (U-Th)/He (AHe) thermochronology studies have focused mostly on Precambrian basement samples across and seaward of the escarpment that separates the plateau from the coastal plain (e.g., Gallagher and Brown, 1999; Brown et al., 2002; Tinker et al., 2008b; Kounov et al., 2009; Flowers and Schoene, 2010; Wildman et al., 2015), with more limited work on basement samples from the plateau interior (Wildman et al., 2017). These studies broadly document two periods of intensified erosion in southern Africa at ca. 150–120 Ma and ca. 100–80 Ma, which correlate with increased sedimentation in the major offshore basins (e.g., Tinker et al., 2008a; Rouby et al., 2009; Guillocheau et al., 2012; Richardson et al., 2017; Baby et al., 2018).

Recent work using AHe on kimberlites has provided additional insight into the erosion history of the plateau interior (Fig. 1B) (Stanley et al., 2013, 2015). The abundant Mesozoic kimberlites and related rocks across the plateau have shorter, simpler thermal histories that are easier to resolve than those of the ancient cratonic basement. Moreover, the kimberlites commonly contain records of lithospheric processes active at the time of their eruption through their xenoliths and chemistries, which provides the opportunity to directly evaluate causal links between mantle and surface processes. Our past work found different temporal relationships between thermochemical modification of the lithosphere and surface erosion at...
different locations. In our off-craton Karoo study region, pervasive heating and metasomatism of the Proterozoic lithosphere (Bell et al., 2003; Kobussen et al., 2008; Janney et al., 2010) occurred simultaneously with a major phase of surface erosion that was most intense at ca. 100–90 Ma (Fig. 1; Stanley et al., 2013). This relationship suggested that lithospheric modification directly triggered surface uplift and erosion. In contrast, on the Kaapvaal craton, more subtle lithospheric modification due to the extension and metasomatism of the Proterozoic lithosphere (Bell et al., 2003; Kobussen et al., 2008; Janney et al., 2010) occurred simultaneously with a major phase of surface erosion that was most intense at ca. 120 to <60 Ma (Fig. 1B; Stanley et al., 2015). This implied that buoyancy change in the lithosphere contributed less to surface uplift here than in the off-craton region, pointing to the need for additional deeper mantle dynamic processes to explain the elevations.

Here we focus on the lower Orange River region where strong lithospheric modification at the base of a major river network provides the opportunity to examine the relative roles of lithospheric processes and river network geometry on erosion patterns (Fig. 1). Previous work on exclusively basement samples from this region inferred erosion during continental breakup and later exhumation along reactivated crustal structures driven by mantle processes (Wildman et al., 2016, 2017). Here we present new (U-Th)/He data primarily on kimberlites and related bodies of varying age in a transect across the Orange River, with a focus on deconvolving potential links between surface and lithospheric mantle processes in this region. Kimberlite-borne xenoliths in our studied pipes show that the region underwent significant lithospheric modification first in the southern portion of the study area and later in the northern part (Fig. 1B; Bell et al., 2003), so erosion patterns might mimic the pattern of
lithospheric modification if lithospheric processes dominate. Alternatively, if fluvial processes more strongly modulate the erosion patterns at the base of this major river system, we might instead observe an eastward wave of erosion reflective of river network propagation. We use the data to evaluate these hypotheses, assess how the results fit with broader erosional and lithospheric modification patterns across the plateau, and consider their implications for connections among mantle, fluvial network, and erosional processes during uplift of the southern African Plateau.

GEOLOGIC BACKGROUND OF SOUTHERN AFRICA

Topography and Orange River Evolution

Southern Africa is characterized by a distinctive topography. Elevations of the generally low-relief plateau interior average ~1000 m and in places are >3000 m (Fig. 1A). The plateau is surrounded on three sides by the high relief “great escarpment,” where the elevation drops abruptly to the coastal plain. This high topography is postulated to extend into the bathymetry of the southeastern Atlantic Ocean; together these topographically elevated areas have been termed the “African Superswell” (Nyblade and Robinson, 1994).

The Orange River currently drains most of the plateau interior (nearly 10^4 km^2), from its headwaters in the eastern highlands to where it enters the coastal plain. All paleo-drainage reconstructions for southern Africa suggest that the majority of the subcontinent has been westward-draining since the Cretaceous, likely since Gondwana breakup (Dingle and Hendey, 1984; Partridge and Maud, 1987; de Wit, 1999). Evidence for this comes from reconstructed post-135 Ma sedimentary fluxes to the adjacent marine basins, with fluxes offshore to the west that are an order of magnitude higher than to the east and south (Tinker et al., 2008a; Guillocheau et al., 2012; Baby et al., 2018b, 2018a). Additionally, large deposits of alluvial diamonds are found only in river terrace deposits to the west of the Orange River mouth prior to ca. 113 Ma (Baby et al., 2018b). These relationships indicate that the headwaters for major rivers draining to the Atlantic were persistently located on the eastern side of the continent.

Several drainage reorganization histories have been proposed for the plateau interior, all of which infer that the mouth of the proto-Orange River shifted through time (Dingle and Hendey, 1984; Partridge and Maud, 1987; de Wit, 1999; Moore, 1999; Moore et al., 2009). Recent work on the Atlantic offshore marine stratigraphy suggests that the main river depocenter was located south of the current outlet near the present-day Olifants River mouth prior to ca. 113 Ma (Baby et al., 2018b). Deposition then moved to a proto-Orange River delta slightly south of the modern Orange River outlet at 113 Ma, with the current Orange River delta becoming active by 93 Ma (Brown et al., 1995; Baby et al., 2018b). Detrital diamond studies support this drainage reconstruction. Only pre–90 Ma detrital diamonds occur between the Olifants and Orange River mouths, while post–90 Ma detrital diamonds have been found north of here (Phillips and Harris, 2009; Phillips et al., 2018). These relationships would be expected given the strong northbound current along the coast if the source of the detrital diamonds, the Orange River, had migrated north by ca. 93 Ma. Overall these observations document a northward migration of the main river stem between 113 and 93 Ma with the current outlet established at that time.

Geology, Magmatism, and Mantle Modification

Southern Africa is composed dominantly of thick, Precambrian continental lithosphere centered around the Archean Kaapvaal and Zimbabwe cratons. These cratons are surrounded by Proterozoic mobile belts and terranes (Fig. 1B). Basement terranes are variably overlain by Precambrian volcanic and sedimentary basins, as well as by the ca. 300–180 Ma Karoo Supergroup and the Cenozoic Kalahari Basin (Fig. 1B).

Southern Africa was affected by two major episodes of flood basalts magmatism during the breakup of Gondwana. The ca. 183 Ma Karoo large igneous province (LIP) centered on the east coast and basalts and associated dolerite sills are found throughout southern Africa (e.g., Marsh et al., 1997; Jourdan et al., 2005). Rocks from the ca. 132 Ma Etendeka LIP (Renne et al., 1996) are dominantly found on the west coast. Significant kimberlite and related alkaline magmatism affected southern Africa from ca. 1800 to 45 Ma (Jelsma et al., 2004). Two major episodes of kimberlite volcanism occurred in the Cretaceous at ca. 200 Ma to 110 Ma and ca. 100 Ma to 70 Ma, with a peak at ca. 90 Ma (Jelsma et al., 2004, 2009; Moore et al., 2008). These kimberlites contain a wealth of mantle and crustal xenoliths, which record snapshots of the state of the lithosphere and crustal cover at the time of eruption.

Southern Africa has one of the best-studied suites of mantle xenoliths in the world due to the abundance of diamondiferous kimberlites. Since southern African kimberlites (and related alkaline ultramafic rocks like melilitites and lamprophyres) erupted over a considerable age range, it is possible to study the evolution of the mantle lithosphere through geologic time. Thermobarometry on xenolith and xenocryst suites from both the lithospheric mantle (Konzett et al., 2000; Bell et al., 2003; Griffin, 2003; Kobussen et al., 2008) and the lower crust (Schmitz and Bowring, 2003) have been interpreted to document a lithospheric thermal pulse sweeping westward across the continent from ca. 150 Ma to ca. 70 Ma. This thermal pulse was accompanied by metasomatism and chemical modification (Griffin, 2003; Kobussen et al., 2009; Janney et al., 2010), which affected the Proterozoic off-craton lithosphere more substantially than the Archean cratonic lithosphere (Schmitz and Bowring, 2003; Bell et al., 2003). The thermal event appears earliest in xenoliths contained in the ca. 150 Ma East Griqualand pipes southeast of the Kaapvaal craton, while contemporaneous kimberlites on-craton and southwest of the craton were unaffected (Fig. 1B). By ca. 100 Ma, lithospheric thermal and metasomatic alteration reached southwest of the craton, with xenolith-derived geothermal gradients elevated by up to 100 °C (Bell et al., 2003; Kobussen et al., 2008). This thermochemical modification pulse reached our study area west of the craton toward the end of this event.

LOWER ORANGE RIVER STUDY AREA

Geology and Previous Thermochronology

Our study area is located in the western plateau in the lower reaches of the Orange River system (Figs. 1 and 2A). Here, the Orange River has carved ~600 m of relief. The Proterozoic Namaqua-Natal belt metamorphic basement is overlain by the Ediacaran to Cambrian Nama Group and the Carboniferous to Permian Dwyka Group.

Previous thermochronology studies from the western region of southern Africa mostly focused on using AFT dating of basement samples to constrain fault reactivation and erosion associated with latest Jurassic to Early Cretaceous continental breakup along this rifted margin. From the approximate latitude of the Orange River and southward, AFT data and a handful of AHe dates were used to argue for two erosional pulses in the middle and Late Cretaceous (Gallagher and Brown, 1999; Kounov et al., 2009, 2013; Wildman et al., 2015). More recent work highlighted the importance of Late Cretaceous reactivation of individual structures in locally enhancing erosion in this area (Wildman et al., 2016, 2017).
This is similar to previous AFT data from farther north along the rifted coast of northern Namibia that constrained the initiation of rapid cooling and erosion at ca. 70 Ma, with total magnitudes that varied based on location relative to reactivated structural lineaments (Raab et al., 2002). In addition, two zircon (U-Th)/He (ZH) and three zircon fission-track (ZFT) dates from several samples in the lower Orange River range from 237 to 340 Ma and 367 to 439 Ma, respectively (Kounov et al., 2013).

Kimberlites and Mantle Modification

Dated kimberlites, lamprophyres, and melilitites in our study area range from ca. 60 to 520 Ma in age (Allsopp et al., 1989; Jelsma et al., 2004; Wu et al., 2010; Griffin et al., 2014), although their non-diamondiferous character means they are less well-studied than their diamondiferous counterparts on the Kaapvaal craton. Nonetheless, the state of the lithospheric mantle is constrained at key snapshots in time at three localities: Hoedkop, Rietfontein, and Gibeon. Mantle xenolith data from these pipes can be compared with the relatively cool, cratonic Kalahari geotherm derived from mantle xenolith P-T arrays in the Kaapvaal and Zimbabwe cratons (Rudnick and Nyblade, 1999).

In the southern study area (Fig. 1), the Hoedkop ultramafic lamprophyre contains peridotite xenoliths that record metamorphic alteration and a geotherm elevated by ~100 °C at all depths over the Kalahari geotherm (Bell et al., 2003; Janney et al., 2010). This is also observed in xenoliths from the nearby Pofadder kimberlite (Shiimi and Janney, 2017). Hoedkop is dated at 79 ± 3 Ma by perovskite U-Pb (Griffin et al., 2014) and 80 ± 2 by 40Ar/39Ar on phlogopite (G.B. Kiviets, 2000, unpublished, but reported in Janney et al., 2010). This indicates that mantle modification had reached our southern study area by 80 Ma or earlier (Bell et al., 2003; Kobussen et al., 2008).

In the northeastern part of our study area (Fig. 1), eclogite and peridotite xenoliths in the Rietfontein kimberlite suggest a craton-like geotherm similar to the Kalahari geotherm (Appleyard et al., 2007; Janney et al., 2010; Shiimi and Janney, 2017). The Rietfontein pipe has generally been thought to be ca. 70 Ma (71.9 Ma, zircon U-Pb, Davis, 1977), but recent perovskite U-Pb data indicate that it is likely older (135 ± 9 Ma, Griffin et al., 2014). We favor the perovskite U-Pb date because it is a more recent measurement that includes six single-grain dates (Griffin et al., 2014) rather than only one single-grain zircon U-Pb analyses (Davis, 1977). Together the data suggest that the lithospheric mantle of the northeastern study area had not yet been modified by 135 Ma.

The Gibeon kimberlite cluster in the northern part of our study area contains granular peridotite xenoliths that reveal a Kalahari-like geotherm at depths <~100 km, while deformed peridotite xenoliths record a geotherm up to ~200 °C hotter than the Kalahari gradient at greater depths (Franz et al., 1996; Boyd et al., 2004). Xenoliths show chemical evidence for metasomatism and a thinner lithosphere than compatible with a steady-state Kalahari geotherm (Boyd et al., 2004). These observations suggest recent thinning at the time of Gibeon eruption (Bell et al., 2003; Boyd et al., 2004). The Gibeon cluster has generally been considered to be 65–75 Ma (Fig. 1; Allsopp et al., 1989), but a younger U-Pb date of 57.6 ± 0.6 Ma was recently reported for a zircon megacryst from Mukurob, Namibia, one of the Gibeon pipes (Woodhead et al., 2017). We therefore adopt 58–75 Ma as the age of the Gibeon cluster.

In summary, the mantle xenolith thermobarometry results and kimberlite emplacement ages impose three important constraints on mantle modification in our study area. The data suggest that (1) in the south the lithosphere was modified by 80 Ma, (2) in the northeast the lithospheric mantle was still unperturbed at 135 Ma, and (3) in the north the lithosphere was actively being modified at ca. 58–75 Ma (Bell et al., 2003).

Thus, the results indicate a northward progression of lithospheric heating and thermochemical modification across the lower Orange River region.

(U-Th)/He THERMOCRONOLOGY BACKGROUND

(U-Th)/He thermochronology is based on the radioactive decay of U, Th, and Sm to 4He and thermally controlled volume diffusion of the He atoms. At high temperatures He escapes rapidly from the crystal and at low temperatures it is quantitatively retained. The transition from He loss to retention occurs over a temperature range called the He partial retention zone (PRZ). Different minerals have a range of temperatures over which this transition occurs. Apatite has one of the lowest temperature sensitivities, ~30–90 °C (Farley, 2000; Shuster et al., 2006), making it useful for detecting upper crustal and erosional processes.

Thermochronology is used to constrain the possible time-temperature (T-T) paths experienced by a rock. For a single low-temperature thermochronometer there are numerous thermal histories that can produce a given date, so it is common practice to use external geologic information,
multiple types of thermochronometers, and/or clever spatial sampling strategies to better constrain thermal histories (e.g., Ehlers and Farley, 2003; Reiners and Brandon, 2006, and references therein). In the AHe system, accumulated radiation damage increases the apatite He retentivity (Shuster et al., 2006), which presents another opportunity to better decipher IT paths by dating apatites with a range of temperature sensitivities (Flowers et al., 2007, 2009). For rocks that resided near the surface long enough to accumulate damage, apatite with higher U and Th concentrations accumulate radiation damage more rapidly and will have higher effective closure temperatures than apatite with lower U and Th. Thermal histories where apatite reside in the 30–90 °C temperature window for substantial time can generate positive correlations between AHe date and effective uranium concentration (eU, [U] + 0.235[Th], Shuster et al., 2006; Flowers et al., 2007), while IT paths where the apatite cools rapidly through the PRZ yield similar AHe dates at all eU (Flowers, 2009).

Apatite from basement rocks in cratonic interiors commonly underwent very small grain sizes (Farley et al., 1996; Ketcham et al., 2011), which presents another opportunity to better decipher tT paths by dating apatites with a range of temperature sensitivities (Shuster et al., 2006), which presents another opportunity to better constrain lithospheric temperatures and cause date-eU correlations (Flowers, 2009; e.g., Flowers and Kelley, 2011; Ault et al., 2013; Emberley and Schneider, 2017).

In addition to variations in temperature sensitivity due to radiation damage, several other factors can cause dispersion in AHe dates within a single sample. These include variability in grain size (Reiners and Farley, 2001), grain fragmentation during mineral separation (Brown et al., 2013), and injection of α-particles from neighboring minerals with high-eU (Spiegel et al., 2009) or mineral coatings (Murray et al., 2014). These effects can be enhanced in samples that spent prolonged time in the PRZ or that underwent reheating and partial resetting, thus causing complex AHe data sets in settings with such thermal histories (Flowers and Kelley, 2011). He particles can travel significant distances when they are emitted, so a geometric correction is applied to account for He that is ejected out of the crystal, which adds additional uncertainty and prevents the use of very small grain sizes (Farley et al., 1996; Ketcham et al., 2011).

SAMPLES, METHODS, AND RESULTS

Strategy and Samples

Our sampling strategy aimed for a combination of kimberlites, Karoo dolerite sills, and gneissic basement rocks of differing age in a N-S transect across the Orange River at a range of elevations (Fig. 2). These samples were selected with the goal of obtaining a suite of apatite characterized by a range of chemistries and crystallization ages that would enable us to best reconstruct detailed thermal histories and regional exhumation patterns with the (U-Th)/He method. We specifically targeted the Hoedkop lamprophyre, Rietfontein kimberlite, and Gibeon kimberlite characterized by a range of chemistries and crystallization ages that would enable us to best reconstruct detailed thermal histories and regional exhumation patterns with the (U-Th)/He method. We specifically targeted the Hoedkop lamprophyre, Rietfontein kimberlite, and Gibeon kimberlite that have mantle xenolith constraints on the lithospheric geotherm and the degree of lithospheric metasomatic alteration at the time of their eruption (as described above).

Unfortunately, our sample from Gibeon did not yield suitable apatite crystals for dating. In total we acquired AHe data for eight samples ranging in elevation from 560 m in the Orange River valley to 865 m on the plateau surface. These samples consisted of the Rietfontein, Schuitdrift, Stolzenfels, and Ondermatjie kimberlites, the Hoedkop lamprophyre, one Jurassic Karoo dolerite sill, and two Proterozoic basement gneisses (Fig. 2A; Table 1). The kimberlites and lamprophyre have published eruption ages from Cambrian to Late Cretaceous.

Accurate eruption dates for the kimberlites are important for comparison with the AHe cooling date and for interpreting the xenolith record, but kimberlites can be challenging to date. The kimberlites, lamprophyres, and other small-volume volcanic rocks of this off-craton region are undiamondiferous and therefore less well-dated than the on-craton diamondiferous kimberlites that we targeted for much of our previous work (Stanley et al., 2015). Our recent study showed that the (U-Th)/He technique in zircon (ZHe) and perovskite (PHe) can be effective at dating kimberlites and serve as a valuable complementary method to other techniques (Stanley and Flowers, 2016). To better interpret our AHe data and inform the overall history of the region, we therefore dated two of these kimberlites, Stolzenfels and Ondermatjie, with ZHe or PHe to better constrain the eruption dates. Our other kimberlites did not yield high-quality zircon or perovskite for (U-Th)/He dating.

Methods and Results

Apatite, zircon, and perovskite crystals were separated using standard density and magnetic separation techniques. Individual crystals were selected for (U-Th)/He analysis based on crystal form and size. Apatite crystals were additionally selected to be free of inclusions under crossed-polarized light. Grains were photographed and their dimensions measured prior to loading in Nb packets for analysis. Helium measurements were made on an Australian Scientific Instruments AlphaChronTM in the University of Colorado Boulder (CU) Thermochronology Research and Instrumentation Laboratory (CU TRaIL), Boulder, Colorado, USA. The samples were then dissolved and measured for U, Th, and Sm on a Thermo Electron Corporation FinniganTM ELEMENT2 sector inductively coupled plasma–mass spectrometer at CU. Details of the experimental methods are the same as those described in Stanley et al. (2015) and Stanley and Flowers (2016). The α-electron correction of Ketcham et al. (2011) was used for all apatite and zircon crystals. For perovskite, the stopping distances of Stanley and Flowers (2016) were used with the geometries presented in Ketcham et al. (2011).

All data are reported in Table 1. Grains with α-electron corrections (Fp) corrections of less than 0.57 or eU of less than 5 ppm were excluded. A large Fp correction leads to high uncertainty, and grains with very low eU are more susceptible to bias toward older dates due to He injection from high-eU neighboring phases. In addition, two grains that are outliers from the main population are included in italics in Table 1. These are plotted but not included in sample averages or the interpretation. Average (U-Th)/He dates and the associated 1σ standard deviation are reported for samples with <20% dispersion.

Five ZHe dates for the Stolzenfels kimberlite and five PHe dates for the Ondermatjie kimberlite yield average dates of 201 ± 18 Ma and 397 ± 30 Ma, respectively (Table 1; Fig. 3). We also report 40 individual AHe dates from eight samples with 3–8 grains per sample (Table 1). Seven of the samples are characterized by <20% dispersion and have mean dates ranging from 62 ± 12 Ma to 118 ± 14 Ma, while the eighth sample (Hoedkop) has greater dispersion (35%). The sample AHe data do not show obvious correlations with elevation or proximity to the Orange River (Figs. 2 and 4). Figure 5 shows AHe date-eU plots for all samples, where individual grain dates are plotted with their 2σ analytical uncertainties. In most samples the dates do not vary strongly with eU. Exceptions are the Hoedkop lamprophyre (Fig. 5A) and Rietfontein kimberlite (Fig. 5B), which yield younger AHe dates than the other samples and show slight positive date-eU correlations with the oldest dates overlapping the published eruption age for Hoedkop.

DISCUSSION

Significance of ZHe and PHe Dates for Kimberlite Eruption Ages

The new ZHe and PHe dates better constrain the timing of kimberlite volcanism in this region. The ZHe date from the Stolzenfels kimberlite
### TABLE 1. (U-Th)/He DATA FROM SOUTHERN AFRICA

| Sample | Mass (µg) | r (µm) | Term | He (ncc) | U (ppm) | Th (ppm) | Sm (ppm) | eU | Th/U | Raw date (Ma) | F_1 | Cor. date (Ma) | 2σf (Ma) |
|--------|-----------|--------|------|----------|---------|----------|----------|----|------|--------------|-----|----------------|----------|

#### Zircon and Perovskite (U-Th)/He data

**SA13-17: Stolzenfels-Wes kimberlite, zircon:** 19.65818°E, 28.5048°S, 560 m elev.  
- Mean ZHe: 201 ± 18 Ma, 9% standard deviation  
- Published age: 60.3 ± 3.9 Ma, Perovskite U-Pb (Griffin et al., 2014)

| Sample | Mass (µg) | r (µm) | Term | He (ncc) | U (ppm) | Th (ppm) | Sm (ppm) | eU | Th/U | Raw date (Ma) | F_1 | Cor. date (Ma) | 2σf (Ma) |
|--------|-----------|--------|------|----------|---------|----------|----------|----|------|--------------|-----|----------------|----------|

#### Apatite (U-Th)/He data

**SA13-13: Hoedkop lamprophyre, 18.70811°E, 29.28263°S, 831 m elev.**  
- Mean: Date-eU trend, 62 ± 12 Ma, 19% standard deviation  
- Published ages: 135 ± 9 Ma, Perovskite U-Pb (Griffin et al., 2014); 71.9 Ma, zircon U-Pb (Davis, 1977)

| Sample | Mass (µg) | r (µm) | Term | He (ncc) | U (ppm) | Th (ppm) | Sm (ppm) | eU | Th/U | Raw date (Ma) | F_1 | Cor. date (Ma) | 2σf (Ma) |
|--------|-----------|--------|------|----------|---------|----------|----------|----|------|--------------|-----|----------------|----------|

**SA13-21: Schuitdrift kimberlite: 19.80731°E, 28.55292°S, 636 m elev.**  
- Mean: 108 ± 17 Ma, 16% standard deviation  
- Published age: No published age for this locality

| Sample | Mass (µg) | r (µm) | Term | He (ncc) | U (ppm) | Th (ppm) | Sm (ppm) | eU | Th/U | Raw date (Ma) | F_1 | Cor. date (Ma) | 2σf (Ma) |
|--------|-----------|--------|------|----------|---------|----------|----------|----|------|--------------|-----|----------------|----------|

**SA13-31: Karoo dolerite, 20.31352°E, 26.74322°S, 848 m elev.**  
- Mean: 108 ± 20 Ma, 18% standard deviation  
- Age: ca. 181–184 Ma (Jourdan et al., 2008)

| Sample | Mass (µg) | r (µm) | Term | He (ncc) | U (ppm) | Th (ppm) | Sm (ppm) | eU | Th/U | Raw date (Ma) | F_1 | Cor. date (Ma) | 2σf (Ma) |
|--------|-----------|--------|------|----------|---------|----------|----------|----|------|--------------|-----|----------------|----------|

(continued)
TABLE 1. (U-Th)/He DATA FROM SOUTHERN AFRICA (continued)

| Sample* | Mass (µg) | r (µm) | Term* | He (ncc) | U (ppm) | Th (ppm) | Sm (ppm) | eUd (ppm) | Th/U | Raw date (Ma) | F₂ⁿ | Cor. date (Ma) | 2σ f (Ma) |
|---------|-----------|--------|-------|----------|---------|---------|---------|-----------|------|--------------|------|---------------|-----------|

**Apatite (U-Th)/He data (continued)**

SA13-15: Ondermatjie kimberlite: 19.43458°E, 28.36494°S, 681 m elev.

| a1  | 2  | 43  | 1T   | 0.25  | 18.4   | 23.6   | 5.9    | 24.0    | 1.3  | 56  | 0.67  | 83  | 7  |
| a2  | 2  | 47  | 0T   | 0.47  | 25.0   | 31.0   | 8.5    | 32.3    | 1.2  | 68  | 0.70  | 97  | 6  |
| a6  | 5  | 54  | 0T   | 0.23  | 2.5    | 11.7   | 10.3   | 5.3     | 4.7  | 68  | 0.72  | 94  | 3  |
| a7  | 2  | 50  | 2T   | 0.10  | 3.1    | 11.5   | 11.2   | 5.8     | 3.7  | 76  | 0.70  | 108 | 8  |
| a9  | 1  | 33  | 0T   | 0.09  | 11.5   | 13.9   | 3.4    | 14.7    | 1.2  | 56  | 0.58  | 96  | 4  |
| a10 | 2  | 45  | 2T   | 0.08  | 4.1    | 13.0   | 15.9   | 7.1     | 3.2  | 54  | 0.67  | 81  | 5  |
| a12 | 2.1 | 37.2 | 0T   | 0.074 | 7.3    | 0.9    | 1.7    | 7.5     | 0.1  | 7   | 0.63  | 11  | 0.6

Mean: 93 ± 10 Ma. 11% standard deviation
Published Age: 515 ± 6 Ma, Perovskite U-Pb (Wu et al., 2010)

SA13-22: Biotite gneiss: 19.81271°E, 28.55596°S, 628 m elev.

| a1  | 2  | 48  | 2T   | 0.99  | 40.5   | 71.6   | 6.9    | 57.4    | 1.8  | 59  | 0.70  | 85  | 2  |
| a2  | 1  | 37  | 0T   | 0.59  | 70.2   | 116.3  | 13.5   | 97.5    | 1.7  | 60  | 0.62  | 96  | 4  |
| a3  | 1  | 34  | 2T   | 0.35  | 52.2   | 74.7   | 8.4    | 69.8    | 1.4  | 47  | 0.58  | 81  | 4  |
| a4  | 2  | 47  | 1T   | 1.18  | 57.0   | 105.1  | 18.1   | 81.7    | 1.8  | 65  | 0.70  | 94  | 3  |
| a5  | 1  | 40  | 2T   | 0.48  | 45.1   | 68.9   | 8.2    | 61.3    | 1.5  | 49  | 0.65  | 75  | 3  |
| a6  | 4  | 57  | 1T   | 1.75  | 40.4   | 66.6   | 7.6    | 56.1    | 1.6  | 67  | 0.75  | 89  | 2  |

Mean: 87 ± 8 Ma, 9% standard deviation
Age: ca. 1.2 Ga, Proterozoic Namaqua-Natal belt basement

SA13-19: Biotite gneiss, 19.75778°E, 28.39045°S, 865 m elev.

| a1  | 1  | 39  | 0T   | 0.25  | 22.9   | 7.0    | 46.5   | 24.5    | 0.3  | 80  | 0.64  | 122 | 5  |
| a2  | 1  | 38  | 2T   | 0.30  | 18.7   | 4.9    | 33.8   | 19.9    | 0.3  | 87  | 0.63  | 135 | 3  |
| a3  | 1  | 41  | 0T   | 0.28  | 20.4   | 5.1    | 35.6   | 21.6    | 0.2  | 75  | 0.66  | 113 | 3  |
| a4  | 0.9 | 34.8 | 2T   | 0.192 | 12.1   | 6.1    | 36.9   | 13.5    | 0.5  | 127 | 0.60  | 208 | 112|
| a5  | 1  | 40  | 0T   | 0.17  | 15.7   | 3.0    | 34.1   | 16.4    | 0.2  | 66  | 0.68  | 102 | 4  |

Mean: 118 ± 14 Ma, 12% standard deviation
Age: ca. 1.2 Ga, Proterozoic Namaqua-Natal belt basement

**Note:** Rows in italics indicate two grains that are outliers from the main population and not included in averages. Cor.—corrected; elev.—elevation; N/A—not applicable.

*Mean and 1s standard deviation of corrected dates reported for samples with <20% standard deviation. No mean reported for samples with >20% standard deviation.
*aEquivalent spherical radius (r), the radius of a sphere with the same surface area to volume ratio.
*bType of grain terminations: 2T—whole grain; 1T—one tip broken off; 0T—both tips broken off; F—interior fragment (no F₂ correction applied).
*deU: effective uranium concentration, weights U and Th for their alpha productivity, computed as \([U] + 0.235 \times [Th]\).
*eFT is alpha-ejection correction of Ketcham et al. (2011).
*fAnalytical uncertainty based on U, Th, He, and grain length measurements.

Figure 3. Data-eU plots for ZHe data from the Stolzenfels kimberlite (A) and PHe data from the Ondermatjie kimberlite (B), southern African Plateau. Data can be found in Table 1. Error bars represent the 2σ analytical uncertainty (in some cases smaller than the marker).
of 201 ± 18 Ma (Fig. 3A) is substantially older than the published U-Pb perovskite date of 60.3 ± 3.9 Ma (Griffin et al., 2014). While perovskite U-Pb is generally the preferred dating method for kimberlites, potential complexities include high common Pb contents and multiple age domains that can make the ages less reliable (see Stanley and Flowers, 2016, for a summary of kimberlite dating techniques). In this particular case we favor our ZHe date both because it is a collection of five dates from individual grains yielding consistent results and because the AHe dates from this sample are reproducible and older than the perovskite U-Pb date (AHe average 107 ± 11 Ma).

The PHe date of 397 ± 30 Ma for the Oondermatjie kimberlite (Fig. 3B) is younger than the published perovskite U-Pb date of 515 ± 6 Ma (Wu et al., 2010). We favor the Cambrian U-Pb date as representing the eruption age over our Devonian PHe date because the U-Pb date is the weighted average of 20 reproducible measurements and because the perovskite closure temperature for Pb is presumed to be higher than for He (PHe closure temperature likely >300 °C; Stanley and Flowers, 2016). PHe is probably less reliable than U-Pb in perovskite for older kimberlites because He is more likely to be lost than Pb due to mid-to-upper crustal tectonic and thermal processes after eruption. The discrepancy between the kimberlite U-Pb and PHe dates for Ondermatjie suggests that heating partially reset the PHe system in the early Paleozoic and that temperatures subsequently were <300 °C. This is consistent with ZFT dates of 350–450 Ma from the area that limit temperatures to <200–300 °C since that time (Kounov et al., 2013).

**Erosion History Along the Lower Orange River**

Our new AHe data for samples along the lower Orange River help constrain the cooling and erosion history along a NNE-SSW transect through this region. The majority of the samples yield Cretaceous dates, with the exception of several individual dates from the Hoedkop and Rietfontein pipes that are <65 Ma. Plots of AHe date versus distance and AHe date versus elevation for our sample transect display no systematic patterns (Figs. 4A and 4B). Nor do obvious patterns emerge when previously published AHe data for Precambrian samples from this same region (Kounov et al., 2013; Wildman et al., 2017) are added to these plots (Figs. 4C and 4D; colored by eU, where such data are available, to reveal dispersion
Figure 5. (A–H) AHe date-eU plots and thermal history modeling results for all samples, southern African Plateau. Date-eU plots show kimberlite and lamprophyre emplacement ages and associated uncertainties as blue bars, eU bins used in the models as vertical dashed lines, and the date-eU pattern predicted by the best fit time-temperature (tT) path as the black curve. Errors are plotted as the 2σ analytical uncertainties, and in some cases are smaller than the symbol. Gray points are outliers not included in the modeling or interpretation. The tT plots show the thermal history constraints as dashed boxes, the weighted mean path in black, the best-fit path in dark gray, good-fit paths in gray, and acceptable-fit paths in light gray. The tT models for the Ondermatjie kimberlite and two Precambrian basement samples start at 450 Ma, but the plots are truncated at 230 Ma for visual clarity. Additional details of model inputs and their rationale are in Table 2 and full model outputs are in the Data Repository material (see text footnote 1). Elev—elevation; Temp—temperature.
For kimberlite and Karoo dolerite samples, the only geologic constraints applied to the thermal history were that the rock had to be >120 °C at the time of eruption/emplacement (Ketcham, 2005). For kimberlite and Karoo dolerite samples, the only geologic constraints applied to the thermal history were that the rock had to be >120 °C at the time of emplacement and at surface temperature of 20 °C today. Additional constraints were imposed on the Precambrian basement and Cambrian Ondermatjie kimberlite samples. The detailed data and constraints used for each model, and their rationale, are summarized in Table 2 (after Flowers et al., 2015) and the full model parameters and outputs can be found in the GSA Data Repository1.

The Hoedkop ultramafic lamprophyre (831 m) and Rietfontein kimberlite (835 m) are located at similar elevations at the southern and northern ends of the sample transect. The AHe dates for Hoedkop (Fig. 5A) are slightly dispersed but overlap the eruption age of 79 ± 3 Ma (Griffin et al., 2014). This suggests that the AHe date records cooling at the time of eruption and the sample underwent little erosive cooling after eruption, which is corroborated by the tT models (Fig. 5A). Assuming that the Rietfontein

![GSA Data Repository Item 2020089, full HeFTy model parameters and output used to make Figures 5 and 6, is available at http://www.geosociety.org/datarpository/2020, or on request from editing@geosociety.org.]

possibly due to radiation damage. The lack of systematic relationships either indicates that there are no significant spatial differences in the timing of cooling, or that the patterns are obscured by other causes of data dispersion, such as radiation damage and/or the sample crystallization age.

We used the HeFTy thermal history modeling software (Ketcham, 2005) to further evaluate any spatial patterns in the timing of cooling, and to determine if the patterns are obscured by other causes of data dispersion, such as radiation damage and/or the sample crystallization age.

| TABLE 2. THERMAL HISTORY MODEL INPUT FOR SOUTHERN AFRICAN SAMPLES |
|---------------------------------------------------------------|
| **1. Thermochronologic data**                                 |
| **Samples and data used in simulations**                      |
| **Sample** | **Rock type** | **eU bins for AHe data** | **Emplacement age used in model** |
|-------------|---------------|--------------------------|----------------------------------|
| Mesozoic rocks |               |                          |                                   |
| SA13-13     | Hoedkop lamprophyre | <15ppm (2 grains), >15 ppm (2 grains) | 79 ± 3 Ma (Griffin et al., 2014) |
| SA13-30     | Rietfontein kimberlite | <15ppm (3 grains), >15 ppm (2 grains) | 70–140 Ma (72 Ma, Davis, 1977; 135 Ma, Griffin et al., 2014) |
| SA13-21     | Schutdrift kimberlite | 1 bin (3 grains) | 80-220 Ma (age unknown, range encompasses ages of other kimberlites in the region) |
| SA13-17     | Stolzenfels-Wes kimberlite | <65ppm (3 grains), >65 ppm (2 grains) | 201.0 ± 18 Ma (ZHe, this study) |
| SA13-31     | Karoo dolerite | <15 ppm (4 grains), >15 ppm (1 grain) | 183 ± 3 Ma (Jourdan et al., 2008) |
| Paleozoic and older rocks |           |                          |                                   |
| SA13-15     | Ondermatjie kimberlite | <15ppm (4 grains), >15 ppm (2 grains) |                                   |
| SA13-19     | Basement gneiss | 1 bin (5 grains) |                                   |
| SA13-22     | Basement gneiss | <65 ppm (3 grains), >65 ppm (3 grains) |                                   |
| **AHe data treatment: Dates, uncertainties, and other relevant constraints** |
| Treatment: Samples were binned into groups with similar eU values, and the mean of each group was modeled. |
| Error (Ma) applied in modeling: The 1s sample standard deviation of each bin was applied if >10%. If <10%, then 10% was applied. |
| r (um): Mean equivalent spherical radius of each bin. |
| eU (ppm): Mean U and Th for each bin. |
| eU zonation: None assumed. |
| **2. Additional geologic information**                        |
| **Constraint** | **Explanations and data source** |
| Mesozoic rocks |                                       |
| (1) >120 °C at time of eruption/emplacement | Kimberlite and dolerite emplaced as hot magma. Eruption dates for each sample from part 1. Present-day approximate surface temperature. |
| (2) 20 °C at present day |                                       |
| Paleozoic and older rocks |                                |
| (1) >200 °C between 350 and 450 Ma | Based on PHe dates from the Ondermatjie kimberlites in this study (397 ± 30 Ma), as well as regional zircon fission-track dates from Kounov et al. (2013) that range from 362 to 439 Ma. |
| (2) <80 °C around 300 Ma | Dwyka sediments are preserved nearby, and the start of Dwyka deposition was at ca. 300 Ma (Catuneanu et al., 2005). |
| (3) >110 °C between 180 and 110 Ma | Based on regional thermal history models of basement AFT and AHe data from Wildman et al., (2017). Regional apatite fission-track dates are between 59 and 129 Ma (Wildman et al., 2017; Kounov et al., 2013) suggesting that they were reset in the Cretaceous. Karoo basin burial peaked at ca. 183 Ma with the eruption of the flood basalts. |
| (4) 20 °C at present day | Present-day approximate surface temperature. |
| **3. System- and model-specific parameters**                  |
| He kinetic model: RDAAM (radiation damage accumulation and annealing model; Flowers et al., 2009). Statistical fitting criteria: Goodness of fit (GOF) values >0.5 for “good” fits, >0.05 for “acceptable” fits. The good-fits also must have a minimum GOF of 1/(N+1) where N is number of statistics used (Ketcham et al., 2005). |
| Modeling code: AHe uses HeFTy v1.9.3. |
| Number of time-temperature paths attempted: 30,000 for all models. |
kimberlite was emplaced at 135 ± 9 Ma (based on its perovskite U-Pb date; Griffin et al., 2014), then its average AHe date of 62 ± 10 Ma indicates substantial post-eruptive cooling. If instead the Rietfontein kimberlite was emplaced at 72 Ma (Davis, 1977) then the cooling history for Rietfontein is similar to that of Hoedkop with little post-eruptive cooling due to erosion. Both the Hoedkop and Rietfontein samples contain low eU (<20 ppm) apatites that define a weak, positive correlation between date and eU (Figs. 5A and 5B). For Rietfontein, this pattern allows for later, low magnitude Cenozoic cooling in the tT simulation result (Fig. 5B).

The Schuitdrift kimberlite (636 m) in the middle of the transect is part of the Ariemsvei kimberlite cluster. This cluster is generally thought to be middle to Late Cretaceous in age (Jelsma et al., 2004), but the exact age of Schuitdrift is unknown. Its average AHe date of 108 ± 17 Ma imposes a minimum age on emplacement and could either represent cooling associated with eruption or post-eruptive cooling/exhumation (Fig. 5C). This kimberlite lacks low eU apatites, so the lowest temperature portion of its history is not well-determined (Fig. 5C).

The two Jurassic-aged samples, the Stolzenfels kimberlite (560 m, ZHe age of 201 ± 18 Ma) and a Karoo dolerite sill (SA13-31, 848 m, ca. 181–184 Ma, Jourdan et al., 2008), yield overlapping AHe dates of 107 ± 11 Ma and 108 ± 17 Ma that postdate emplacement (Figs. 5D and 5E). Consequently, tT models for both samples show a ca. 120–100 Ma cooling phase (Figs. 5D and 5E). The simulation for Stolzenfels allows additional later cooling but does not require it.

The Cambrian Ondermatjie kimberlite and two Precambrian basement samples also yield Cretaceous AHe dates that record Cretaceous cooling. The Ondermatjie kimberlite (681 m) and a basement gneiss at similar elevation (SA13-22, 628 m) yield overlapping AHe dates of 93 ± 10 Ma and 87 ± 8 Ma (Figs. 5F and 5G). At slightly higher elevation, another basement gneiss (SA13-19) gives a slightly older AHe date of 118 ± 14 Ma (Fig. 5H). The thermal history models show substantial Cretaceous cooling of all three samples, with the higher elevation basement sample cooling 10–40 m.y. earlier than the other two (Figs. 5F, 5G, and 5H).

Taken together, the relationship between eruption ages and AHe dates for our samples suggests that there was significant cooling and erosion in Early to mid-Cretaceous time that occurred prior to eruption of the 70–80 Ma kimberlites. The shape of the cooling paths varies between samples and with eruption age (Fig. 5). None of the good-fit tT paths reach surface temperature before 125 Ma. All samples cooled below ~70 °C by ca. 75 Ma, with the exception of the Hoedkop lamprophyre that erupted around this time and the Rietfontein kimberlite (Figs. 5A and 5B). Although most samples can be explained by a main cooling phase from ca. 120 to 100 Ma, both older (150–110 Ma; SA13-19, Fig. 5H) and younger (100–70 Ma; SA13-30, SA13-22, Figs. 5B and 5G) cooling is permitted by some samples. Variability in cooling was also detected by previous AFT and AHe study of basement in this region, from which two phases of Cretaceous erosion starting at 130–120 Ma and 110–70 Ma were inferred (Wildman et al., 2017).

**Evaluating the Role of Lithospheric Thermochemical Modification on the Erosion History of the Lower Orange River**

Our AHe data set includes results for four kimberlites and one ultramafic lamprophyre with crystallization ages as young as Late Cretaceous. These data complement previous work exclusively on ancient basement samples in this region (Wildman et al., 2016, 2017), but our focus on kimberlites allows us to more comprehensively evaluate temporal and causal links with lithospheric thermochemical modification processes as recorded by mantle xenoliths in the same and nearby pipes. This approach is like that in our past kimberlite thermochronology studies across the plateau in both the off-craton Karoo (Stanley et al., 2013) and on-craton Kaapvaal (Stanley et al., 2015) regions. Here we can assess if erosion in our lower Orange River study area overlaps with the broad timing of lithospheric modification, and if it mimics the northward progression of lithospheric modification across the region.

Most of our lower Orange River samples are consistent with major erosion at ca. 120–100 Ma, with both earlier and later erosion implied by some samples. The main ca. 120–100 Ma unroofing phase thus overlaps broadly with the timing of lithospheric modification. Xenoliths in the Hoedkop lamprophyre indicate that mantle metasomatism and thinning had occurred by 80 Ma, while Rietfontein kimberlite xenoliths indicate that it had not started by 135 Ma, pinning the probable timing between these dates. Lithospheric modification in the Karoo region to the east is more tightly constrained to between 117 and 103 Ma (Fig. 1B; Bell et al., 2003; Kobussen et al., 2008; Janney et al., 2010). Since lithospheric modification likely progressed from SE to NW (Bell et al., 2003), our Orange River study area located northwest of the Karoo (Fig. 1B) was probably not affected prior to 117 Ma. This limits likely thermochemical modification of Proterozoic lithosphere in the lower Orange River to 117–80 Ma. Metasomatism of the off-craton lithosphere here was substantial and associated with lithospheric thinning (Janney et al., 2010). The broad coincidence of this lithospheric event with major surface erosion from ca. 120 to 100 Ma suggests that lithospheric modification may have contributed to surface uplift and denudation. These relationships are similar to those documented in our off-craton Karoo study area where regional erosion was simultaneous with strong lithospheric modification (Stanley et al., 2013), but the timing of the mantle thermochemical event is less tightly constrained in the lower Orange River.

If lithospheric modification alone influenced the erosion history, then we might expect the AHe data to young northward because mantle modification occurred earlier in the southern study area (between 117 and 80 Ma, as described above) than to the north where Gibeon pipe xenoliths indicate the lithosphere was actively undergoing modification at 75–58 Ma (Fig. 1B; Bell et al., 2003; Boyd et al., 2004; Janney et al., 2010). However, neither our AHe data nor previously published results suggest a systematic northward progression of erosion. Thus, despite the broad overlap in the timing of erosion and lithospheric modification, the relationships imply additional influences on the denudation patterns across the region.

**An Eastward-Migrating Wave of Erosion across the Southern African Plateau and Its Causes**

Thermochronologic data from the lower Orange River region suggest a multiphase unroofing history. The Cretaceous AHe dates from this area are far more heterogeneous over short wavelengths than in our AHe data sets from elsewhere on the plateau (Stanley et al., 2013, 2015). This is compatible with Late Cretaceous movement on faults causing spatial complexity in the thermochronologic data patterns (Wildman et al., 2017), and implies multiple influences on the erosion history. Opening of the south Atlantic Ocean at ca. 130 Ma (Collier et al., 2017), establishment of the mouth of the Orange River in its current location from 112 to 93 Ma (de Wit, 1999; Baby et al., 2018b), deeper mantle dynamic effects (e.g., Lithgow-Bertelloni and Silver, 1998), as well as lithospheric modification (e.g., Bell et al., 2003) all may have influenced denudation here.

Distinctly different relationships between surface erosion patterns and lithospheric mantle evolution emerge from our kimberlite AHe studies in three different plateau regions. Unlike the more complex erosion history in our current off-craton study area in the lower Orange River, the off-craton Karoo study region farther east underwent a simpler history of substantial erosion at 100–90 Ma coincident with pervasive lithospheric modification.
and thinning (Fig. 1B; Stanley et al., 2013). And our kimberlite transect across the southern Kaapvaal craton revealed an eastward migrating wave of erosion from ca. 120–60 Ma associated with relatively weak modification of the Archean lithosphere (Fig. 1B; Stanley et al., 2015).

To better evaluate the large-scale trends in this broader data set across an ~1000-km-long region, Figure 6 indicates the time when the weighted mean tT path from the thermal history models for Mesozoic kimberlites and related igneous rocks in an E-W transect across the plateau cool through 70 °C (tT models in Fig. 5 of this paper and those in Stanley et al., 2013, 2015). We only plot results for those kimberlites with post-eruption AHe dates that capture cooling associated with major erosion. The younger pipes that yield AHe dates coincident with pipe eruption are excluded because they do not record the erosion event owing to kimberlite emplacement after major erosion occurred. Figure 6 shows that the western, off-craton lower Orange River samples record the earliest cooling/erosion, with slightly later cooling/erosion of the off-craton Karoo samples, followed by a general eastward younging of cooling across the craton. The timing of lithospheric modification is also shown on Figure 6. In the off-craton regions and western craton, cooling overlaps with lithospheric modification, while dates trend younger to the east on the craton where modification was mild.

We suggest that the timing and nature of mantle processes as well as the location within the drainage network critically influenced the eastward progression of erosion across the plateau. Westward tilting of the continent initially established a large westward-draining river system (Braun et al., 2014). Latest Jurassic to Early Cretaceous rifting on the west coast and the onset of seafloor spreading ca. 130 Ma (Collier et al., 2017) initiated Early Cretaceous cooling and erosion here, followed by a pulse of erosion that propagated up the river network. The west-to-east cooling/erosion patterns in Figure 6 are broadly congruent with predictions of the stream power model (Howard and Kerby, 1983), where a continental uplift event triggers a wave of incision that starts at the outlet of the river network, migrates upstream (e.g., Rosenbloom and Anderson, 1994; Whipple and Tucker, 1999), and then decelerates to the diminishing upstream drainage area (e.g., Berlin and Anderson, 2007).

Lithospheric modification, possibly driven by deeper dynamic processes, may have further enhanced the eastward erosional pattern. Uplift associated with strong thermochemical alteration in the lower Orange River region likely promoted rapid mid-Cretaceous erosion here, could have helped drive mid- to Late Cretaceous exhumation in the west along faults (Wildman et al., 2017), and may even have contributed to pushing the mouth of the Orange River northward between ca. 112 and 93 Ma (de Wit, 1999; Baby et al., 2018). Similarly, lithospheric-induced surface uplift in the off-craton Karoo region enhanced denudation (Stanley et al., 2013) and would have further promoted rapid eastward propagation of erosion. In contrast, the wave of enhanced erosion and incision slowed in the upstream, on-craton region to the east (as indicated by a more dramatic eastward younging in the timing of cooling of samples here; Fig. 6) owing to less lithospheric-induced surface uplift and a smaller drainage area. A thicker package of Karoo basalt in the east may have added a secondary lithologic control. Thus, the interaction of mantle-derived uplift and river network evolution likely exerted key control on the Cretaceous uplift, denudation, and topographic history of the southern African Plateau.

**CONCLUSIONS**

New AHe dates from kimberlites, other shallow intrusions, and basement rocks from the lower Orange River region document a regional phase of cooling/erosion dominantly from ca. 120 to 100 Ma prior to the eruption of the youngest kimberlites, with some spatial complexity in the cooling/erosion history across the study area. The 120–100 Ma erosional phase overlaps the time of strong lithospheric modification here but does not mimic the spatial patterns in detail, indicating that surface uplift and erosion from this process alone cannot explain the data. Mantle processes and the northward shift of the Orange River mouth to its current location, as well as previously proposed continental breakup and fault reactivation activity (Wildman et al., 2016, 2017), may all have influenced the Cretaceous surface history of the study area. When data from kimberlites in this region are compared with those from across the plateau (Stanley et al., 2013, 2015), an eastward younging of Cretaceous cooling is apparent (Fig. 6). We suggest westward tilting of the continent triggered an erosional wave that started in the west and propagated across the continent (Braun et al., 2014). Uplift due to lithospheric modification west of the craton likely induced more rapid erosion here than on-craton. The progression of the main phase of erosion from west to east suggests that the large, westward-draining river network played a key role in controlling erosion patterns across the plateau. Together, kimberlite AHe data from the lower Orange River region and across the southern African Plateau imply that geomorphic processes, especially the large-scale drainage network geometry, temper the denudational response to mantle processes. Interplay between mantle and surface processes has a strong control on erosion patterns and rates in continental interiors.
