

# Origin of plateau uplift using the $^4\text{He}/^3\text{He}$ apatite thermochronometry and the $^{13}\text{C}$ - $^{18}\text{O}$ carbonate paleothermometry

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## 1 Introduction

Why do continental plateaus rise? A variety of geodynamic hypotheses have been advanced, many of them with specific reference to the Colorado Plateau, which poses the question particularly well: Like most of the North American craton, the plateau is a Paleozoic-Mesozoic platform that on average subsided slowly for nearly 500 Myr while sedimentation maintained its elevation near sea level (Hunt, 1956); unlike most of the craton, at some time after 80 Ma, via net uplift of both rocks and the topographic surface, the plateau attained its current mean elevation of about 1900 m (Pederson, 2002) with little internal strain (<1%) of the upper part of the crust. Post-Cretaceous erosional unroofing of the plateau ranges from zero in preserved early Tertiary basins on the perimeter of the plateau to as much as 4500 m in the Upper Granite Gorge of the Grand Canyon (Dumitru et al., 1994; Kelley et al., 2001; Naeser et al., 2001; Naeser, 1989). How and when did the lithosphere acquire so much buoyancy, and what is the relationship between topographic uplift, rock uplift and erosional unroofing?

Proposed mechanisms for intraplateau uplift are numerous and in many cases not mutually exclusive. A summary table of proposed mechanisms published in 1980 enumerated 14 mechanisms described among 25 references (Table 3 in McGetchin et al., 1980), and a similar synthesis in 1985 focused on the Colorado Plateau alone listed 11 mechanisms and 20 references (Table 1 in Morgan and Swanberg, 1985). A recent summary of mechanisms (Roy, 2005) broadly ascribed them to three categories (1) Late Cretaceous to Early Tertiary uplift related to Sevier and Laramide contractile deformation from 80 to ~40 Ma, adding buoyancy by thickening of the crust, thinning of the upper mantle, or rarifying the upper mantle through the introduction of volatiles; (2) mid-Tertiary uplift related to the demise of a Laramide flat slab, where buoyancy is added to the upper mantle by mechanical thinning or chemical modification, and (3) late Tertiary 'epeirogenic' uplift associated with regional extensional tectonism, either by convective removal of lithosphere or heating from below.

Potential Laramide crustal thickening mechanisms would supply buoyancy through lateral redistribution of deep crustal material by top-west basal shear (Bird, 1979), underplating of crust during Laramide flat-slab subduction, or by lateral flow in a weak deep crustal channel from the Sevier orogen to the west (McQuarrie and Chase, 2000). Convective removal of the mantle lithosphere (England and Houseman, 1988) during Laramide shortening could also lead to late Cretaceous or early Tertiary uplift. Chemical modification of the lithosphere may also play a role. In particular the hydration of the lithosphere from volatiles derived from the underthrust Laramide slab would predict early Tertiary uplift (Humphreys et al., 2003).

Mechanical attenuation of the mantle lithosphere after 40 Ma could have occurred in a variety of ways. Foundering of a Laramide slab that had already tectonically attenuated (but had not completely removed) the plateau lithosphere during Laramide underthrusting, and the replacement of the Laramide slab with asthenosphere, predict uplift to have been concentrated at the onset of mid-Tertiary magmatism at ~35 Ma (Spencer, 1996). Post-Laramide chemical modification through melt extraction would predict a mid-Tertiary pulse of uplift along the volcanic margins of the province relative to the interior (Roy et al., 2005). This mechanism underscores the possibility that there may be spatial as well as temporal variation in the amount and mode of uplift on the plateau (Roy, 2005).

Late Tertiary conductive re-equilibration due to heating the lithosphere from below (e.g. (Thompson and Zoback, 1979), perhaps aided by a mantle plume (Parsons et al., 1994), provides a simple

mechanism for uplift that predicts protracted unroofing, perhaps over tens of millions of years. Convective instability or delamination could have occurred at any time after 40 Ma (e.g. Bird, 1979; Humphreys, 1995), and has been identified as an important driver of elevation change west of plateau beneath the Sierra Nevada region in Pliocene and Quaternary time (Jones et al., 2004; Zandt et al., 2004). In contrast to conductive mechanisms it predicts uplift on a timescale of a few million years.

As can be ascertained from this brief discussion, interest in this problem is both longstanding and deep because the plateau's rise contains a fundamental, and still cryptic, lesson in how the lithosphere works. In contrast to tectonic subsidence histories, which can be estimated fairly accurately using marine stratigraphy, estimates of continental uplift require using proxies that are by comparison less robust. Thermochronology is essential for constraining unroofing histories, and this information has been used to infer uplift on the qualitative premise that uplift events correspond with erosional unroofing events. Although unroofing is an essential parameter in isostatic balance calculations, the thermal structure of the uppermost crust is not sensitive to elevation gain or loss (e.g. Hodges, 1991). However, a number of methods that constrain paleo-elevation using proxies for atmospheric temperature and pressure have recently been developed, including leaf physiognomy in fossil plants (Wolfe et al., 1998a; Forest et al., 1999), vesicle size in basalt flows (Sahagian et al., 2002ab), and O and C isotopic compositions in carbonates and clays (Poage and Chamberlain, 2002; Eiler et al., 2005; Ghosh, 2005; Horton et al., 2004).

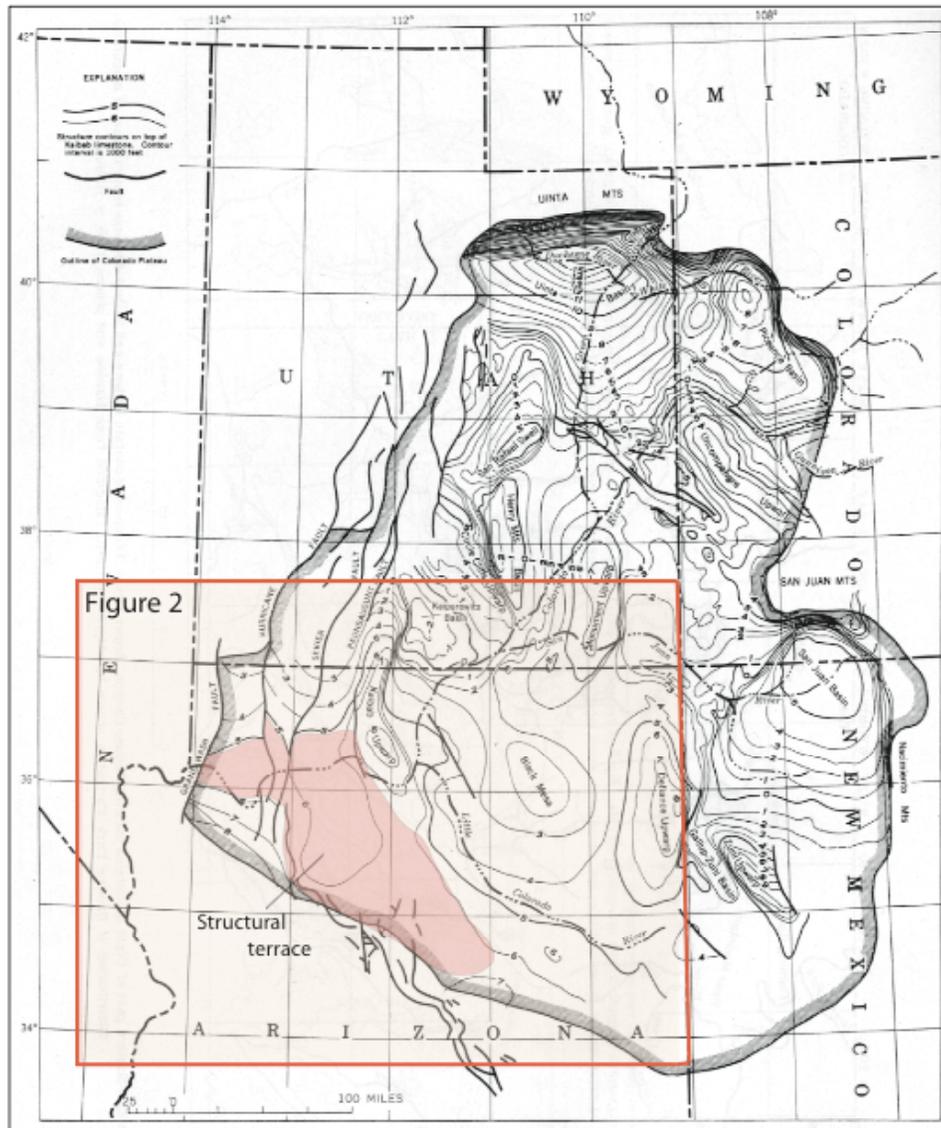
In this proposal, we outline a strategy to investigate the both the unroofing and surface uplift histories of the southeastern portion of the plateau, using two very new but reasonably well tested innovations in He thermochronometry and C-O isotope thermometry. The first, the  $^4\text{He}/^3\text{He}$  apatite thermochronometer, constrains the timing and rate of low-temperature cooling events (Farley et al., 2005; Shuster and Farley, 2004), to temperatures significantly lower (down to 40 °C) than those accessible by conventional whole crystal (U-Th)/He apatite ages (~70 °C for typical cooling rates). This method is therefore sensitive to near-surface (last 1000 to 2000 m) erosional events such as the excavation of deep canyons. The second technique, the  $^{13}\text{C}$ - $^{18}\text{O}$  carbonate paleothermometer, simultaneously constrains both the carbonate growth temperature and the O isotopic composition of the waters from which it was derived, and is therefore useful as a paleoaltimeter (Eiler et al., 2005; Ghosh, 2005). This method requires fewer assumptions regarding climate change and seasonality than conventional two-phase carbonate-water oxygen isotope thermometry, and permits more effective isolation of the altitudinal contribution to this isotopic record.

## **2 Tectonic Setting**

The Colorado Plateau lies in the foreland of Cordilleran orogen, and has escaped all but relatively mild Phanerozoic deformations that more profoundly affected neighboring areas, especially in the late Paleozoic ('Ancestral Rockies' deformation), late Cretaceous/early Tertiary (contractile Sevier orogen to the west, Laramide orogen to the north and east) and late Tertiary (Basin and Range/Rio Grande rift extension).

Stratigraphically, Paleoproterozoic basement and nonconformably overlying, younger Proterozoic strata are unconformably overlain by a Paleozoic platform section about 1000 m thick (Beus and Billingsley, 1989). The Paleozoic is disconformably overlain by Mesozoic deposits 2000 to 3000 m thick (Hintze, 1993). The Cretaceous interior seaway blanketed the plateau, leaving marine deposits as young as Turonian (ca. 90 Ma) in northeastern Arizona, and as young as Campanian (<~80 Ma) across central and eastern Utah (Nations, 1989). Intra- and post-Sevier/Laramide deposits along the western, northern and eastern flank of the plateau are up to several thousand meters thick, but age-equivalent strata exposed along the southwestern margin of the plateau (known informally as the Rim gravels; Figure 2) are generally only a few hundred meters thick (Elston and Young, 1991). Immediately southwest of the plateau, a broad region of Precambrian crystalline rocks are unconformably overlain by either Late Cretaceous or mid- to Late Tertiary volcanic and sedimentary strata. The Rim gravels record Early Tertiary unroofing and northeastward fluvial transport away from Laramide uplands to the southwest (Potochnik, 2001; Young, 1979).

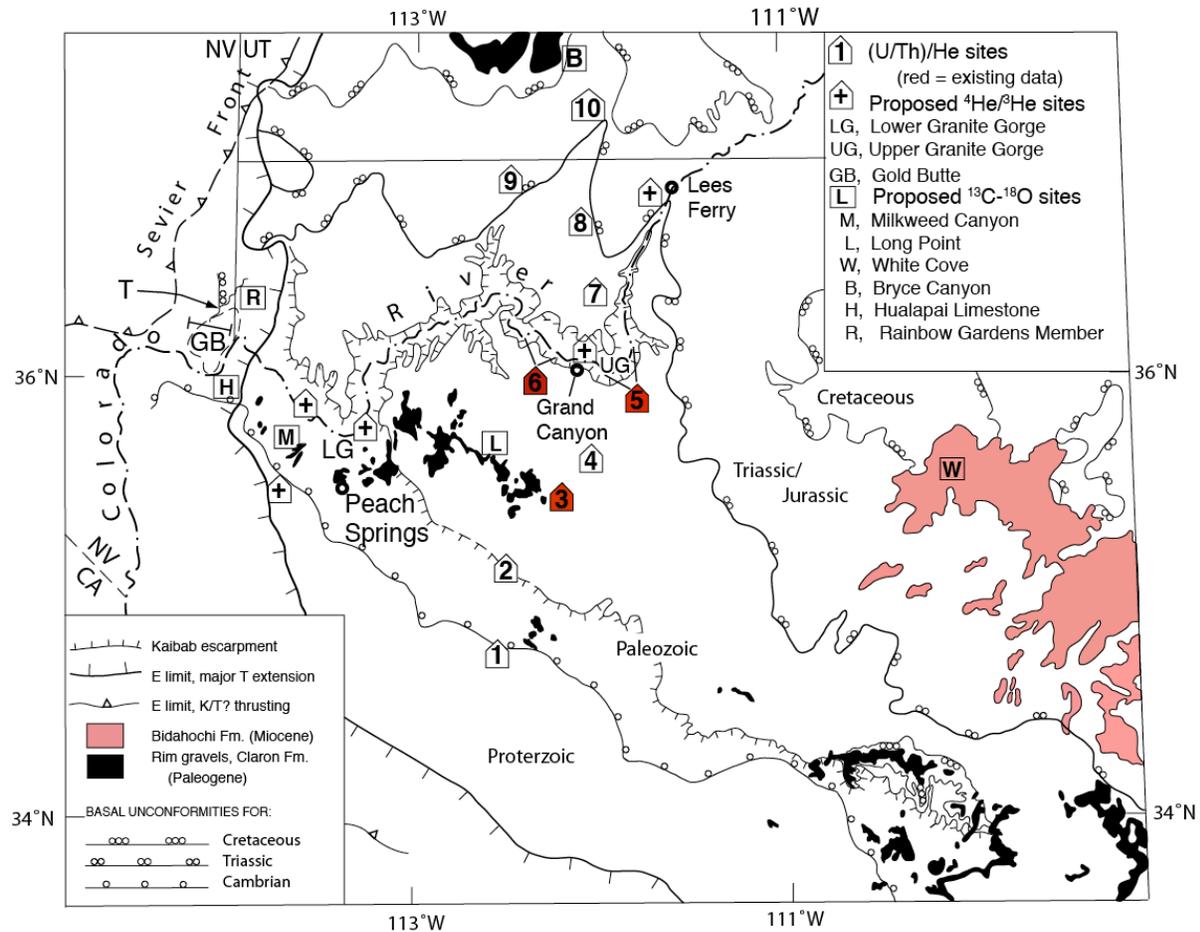
On the margins of the plateau, Sevier/Laramide age deposits are overlain by extensive areas of Oligocene through Recent volcanic strata (omitted from Figure 2), while toward the center of the plateau they are intruded by small, isolated silicic plutons mostly of Oligocene or Miocene age. Along the southwestern margin of the plateau, Oligocene and younger volcanic and sedimentary deposits record a reversal of drainage to the southwest, toward the former Laramide uplands, presumably as a result of mid-



**Figure 1. Map showing geomorphic outline of the Colorado Plateau, the Colorado River and its major tributaries, and structure contours on the top of the Paleozoic section. Relatively flat structural terrace underlain by Permian Kaibab Limestone discussed in text shown in red. Modified from Hunt (1956).**

Tertiary crustal extension (Peirce, 1979; Young, 1989). On the plateau northeast of the rim, the thin (~100 m), areally extensive Bidahochi Formation (Figure 2) records a prolonged episode of fluvial and lacustrine (?) aggradation from ~16 to 6 Ma (Dallegge, 2001). Just west of the plateau in the Basin and Range province where the Colorado River enters Lake Mead (Figure 2), lacustrine sedimentation occurred both before and after a major pulse of mid-Miocene extension, including the Rainbow Gardens

Member of the Horse Spring Formation (~24 to 16 Ma) and the Hualapai Limestone (~11 to 6 Ma) (Faulds, 2001; Lamb et al., 2005; Spencer, 2001).



**Figure 2. Map showing selected tectonic elements and sampling localities along the southwestern margin of the Colorado Plateau discussed in text.**

Structurally, in the northern plateau the Laramide orogeny created a dome-and-basin structural pattern with wavelength of 100 to 150 km and amplitude of 1500 to 2000 m. The amplitude ranges up to 3000 m along the north and east margins where it is bordered by the much larger Laramide uplifts of the Rocky Mountains (Figure 1), where amplitudes are as large as 6000 m (e.g. (Smithson et al., 1979). Structural relief on the pre-Laramide section effectively dies out southwestward. Within northernmost Arizona, structural relief on pre-Laramide strata varies by less than  $\pm 1000$  m across the Defiance and Kaibab uplifts; within 100 km of the southeastern plateau rim, relief varies by less than  $\pm 600$  m (Figure 1).

To first order, the structure of the southwestern portion of the plateau is a NE-dipping homocline with an average dip of  $\sim 0.4^\circ$  across a strike length of 500 km (Figures 1 and 2). In the vicinity of the Grand Canyon, the homocline is interrupted by a broad (ca. 15,000 km<sup>2</sup>) structural terrace residing at an elevation of 1500 to 2000 m (Figure 1), underlain by the Permian Kaibab Limestone and discontinuous exposures of the lower part of the Triassic Moenkopi Formation. Throughout much of this region, the basal Cambrian unconformity (overlain by the Tapeats Sandstone) resides at an elevation of about 600 m,

near the elevation of the modern Colorado River. The terrace is deeply incised by the Colorado and locally interrupted by normal faults and monoclinical warps of Tertiary and Laramide age, respectively (Kelley et al., 2001; Pederson et al., 2002).

To the northeast of the structural terrace, progressively younger Mesozoic and Cenozoic formations are exposed in a series of sinuous cuestas (the “Great Rock Staircase”) beginning with the Vermillion Cliffs, which is underlain by basal Mesozoic sandstones of the Moenkopi Formation. The terrace is interrupted by the Kaibab uplift (Figure 1), a north-trending anticline composed of a number of monoclinical flexures. At its structural crest it raises the Kaibab Limestone to elevations near 3000 m (Huntoon, 1989). The deepest incision of the Colorado (~1,600 m) occurs in the Grand Canyon Village area (Figure 2), where the basal Tapeats and Kaibab lie at elevations of about 1100 m and 2200 m, respectively.

To the southwest of the terrace, along the modern southwest-facing erosional escarpment on the Paleozoic (Mogollon Rim) the basement ascends gradually to the surface (Figure 2). In the Peach Springs area where the modern erosion surface cuts across the Paleozoic section, several modern canyons up to 350 meters deep (Milkweed, Hindu and Peach Springs) were previously filled with Rim gravels (Music Mountain Formation and younger strata (Billingsley, 1999), which lie in buttress unconformity with Paleozoic along the sides of the canyons (Elston and Young, 1991; Young, 1989, 2001).

West of the terrace, the plateau is cut by the southern part of Hurricane fault zone and other minor faults, and is abruptly truncated by major normal faults of the highly extended Basin and Range province in the Lake Mead area (Brady et al., 2000), giving rise to the Gold Butte crustal section (Figure 2), a relatively intact, east-tilted fault block with some 10 to 15 km of structural relief (Fitzgerald et al., 1991; Fryxell et al., 1992; Reiners, 2000). The easternmost valley in the Basin and Range (Grand Wash Trough) is underlain by a section of upper Miocene nonmarine deposits that include the Hualapai Limestone mentioned above (Lucchitta, 1979). These deposits lie in buttress unconformity against the imposing western margin of the plateau (Grand Wash Cliffs), suggesting that the differentiation of the plateau and Basin and Range had occurred by ~11 Ma (Lucchitta, 2001; Lucchitta, 1989).

As is clear from Figure 1, the Colorado Plateau is currently drained toward the southwest by the Colorado River and its tributaries, and on the basis of widespread Pliocene river gravels along the lower part of the river derived in part from central Utah, there is strong consensus that a throughgoing Colorado existed at least as far back as latest Miocene or early Pliocene time (Hunt, 1956, 1969; Longwell, 1946; Lucchitta, 1979), or ~5 Ma (Spencer, 2001). But how much older is it, and what is the relationship between excavation of the Grand Canyon and plateau uplift?

It has long been noted that at Lees Ferry (Figure 2), the Colorado lies at about 1000 m above sea level in a terrain of very little local relief, and cuts improbably southwestward through a ~1000 to 1500 m high topographic barrier to form the Grand Canyon. The river’s incision of the high southwest margin of the plateau is most simply explained by having the river predate the development of the relief either as an antecedent or superimposed stream (Hunt, 1969; Lucchitta, 1989). But where the river currently exits the plateau at Grand Wash, the stratigraphic character and geochemistry of the Hualapai Limestone and related deposits record an internally drained basin, ruling out the existence of a through-going Colorado drainage in late Miocene time (Longwell, 1946; Lucchitta, 1979; Lucchitta, 1989). As discussed above, the western margin of the plateau was in existence through late Miocene time, raising the question of whether the Colorado existed before the Pliocene, and if it did exist, how it found its way off of the plateau. Following paleohydrological evidence summarized by (Hunt, 1969), most workers regard the integration of the Colorado River drainage above Lees Ferry to have been completed no later than mid-Miocene time, and therefore the course of the river below Lees Ferry from ~16 to 5 Ma is problematic.

This paradox has resulted in a remarkable diversity of proposals for the pre-Pliocene paleohydrology of the southwestern part of the plateau, thoroughly discussed in a 33-article proceedings volume from a symposium in 2000 dedicated to the topic (Young and Spamer, 2001), and ably summarized in a recent popular book published by the Grand Canyon Natural History Association (Ranney, 2005). Proposed ages of initiation of the Colorado range from Late Cretaceous to Pliocene, and proposed pre-Pliocene courses range from central Utah to the Rio Grande Valley. The development of

Sevier/Laramide uplands, the reversal of drainage along the Mogollon Rim at 20 to 30 Ma induced by mid-Tertiary extension in central Arizona, the sharp truncation of the western margin by extension from 16 to 11 Ma, and the opening of the Gulf of California at 6 Ma are all important events that are cited as playing a role in integrating the Colorado River drainage.

Despite the lack of consensus on the Colorado's paleohydrology, it seems clear that (1) the timing of excavation of Mesozoic rocks from atop the structural terrace, (2) the timing of incision of the Grand Canyon below the structural terrace, and (3) the elevation history of Cenozoic lake beds both on and immediately west of the plateau will place important new constraints on the relationship between uplift, erosional unroofing and tectonic events.

### 3 Methods

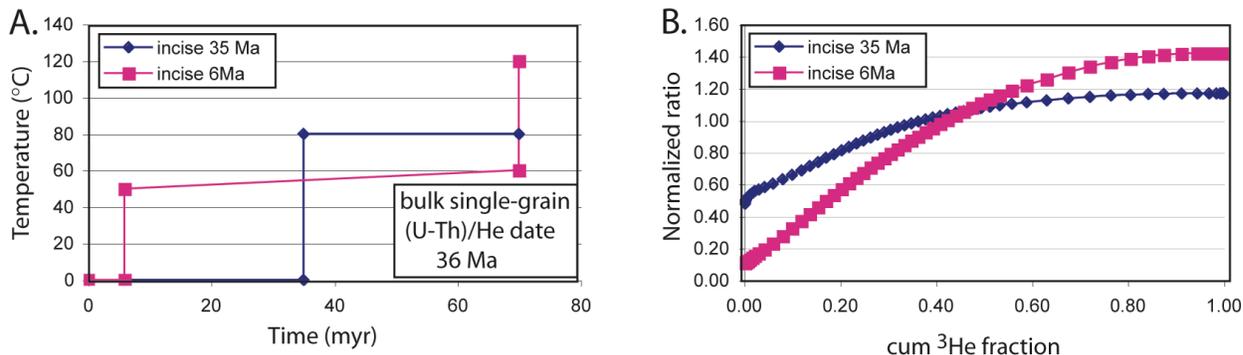
***<sup>4</sup>He/<sup>3</sup>He Thermochronometry.*** (U-Th)/He apatite dates are widely used to constrain the timing of near surface cooling due to denudational processes (Ehlers and Farley, 2002). Whole crystal or “bulk” (U-Th)/He apatite ages reflect the combined effects of He loss due to diffusion and He ingrowth due radioactive decay of U and Th, integrated over the cooling history of the sample. If a rock is known to have cooled monotonically, the measured date reflects the time of cooling through ~60-70 °C (Farley, 2000). However, if a rock has spent a significant time in the He partial retention temperature range (70-30 °C), a measured age does not directly constrain the cooling path; the interpretation is non-unique. Additional information regarding the cooling history over the 70 to 30 °C temperature range is recorded by the spatial distribution of <sup>4</sup>He within a crystal (Shuster and Farley, 2004, 2005). For example, an apatite that instantaneously cooled will have a square concentration profile (modified only by alpha-ejection; (Farley, 1996) owing to the fact that there was no opportunity for diffusive He loss. In contrast, an apatite that resided in the 70-30 °C temperature window for sufficient time for diffusive He loss to occur will be characterized by a rounded <sup>4</sup>He profile. The recently developed <sup>4</sup>He/<sup>3</sup>He method allows determination of the <sup>4</sup>He concentration profile in a crystal (Shuster and Farley, 2004, 2005). This technique is sensitive to diffusion length scales of just a few microns, and can place constraints on cooling to temperatures as low as 30 °C. Until now, unique (U-Th)/He constraints on the timing of an erosional event required that the episode be of sufficient magnitude (2300 to 3500 m of denudation for geothermal gradients of 20-30 °C/km) such that acquisition of apatite ages from rocks in a vertical transect would reveal the timing of unroofing. However, this new technique provides the potential to constrain lower magnitude denudational events based on information that can be accessed in apatite crystals from a single sample.

The timing of Grand Canyon incision is a classic problem that is an obvious target for (U-Th)/He thermochronometry. However, this approach has not yet been successfully implemented because it requires the acquisition of apatite data from a vertical transect through the stratigraphic section in rocks that either lack high-quality apatite (e.g. limestones) or from detrital grains that may be problematic for the (U-Th)/He method. We have obtained new single-grain (U-Th)/He apatite ages from three crystalline basement samples and three Dox sandstone samples in the Upper Granite Gorge (Figure 2). These ages generally range from 30 to 40 Ma. They are ~30 to 40 Myr younger than apatite fission track ages from the same area (Kelley et al., 2001), indicating that if the canyon existed at that time, was not incised to a level below the Jurassic at 35 Ma.

This information alone does not differentiate between an old (ca. 35 Ma) versus young (ca. 6 Ma) age for incision, but application of the <sup>4</sup>He/<sup>3</sup>He method to these crystals would clearly distinguish these end-member histories. Figure 3A depicts two different time-temperature histories, reflecting old or young incision, that result in the same bulk (U-Th)/He age of 36 Ma. However, the spatial distribution of <sup>4</sup>He in the grain due to these thermal histories are dramatically different, and can be clearly differentiated by the evolution of the <sup>4</sup>He/<sup>3</sup>He ratio (normalized to the bulk ratio) during step heating (Figure 3B). Thus, the application of the new <sup>4</sup>He/<sup>3</sup>He technique to the basement samples from which we have already acquired (U-Th)/He apatite data may allow us to more precisely constrain the timing of Grand Canyon incision.

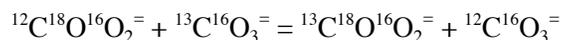
Although the actual history may lie between these end-member examples, there should be a diagnostic signal that can shed new light on this problem.

**<sup>13</sup>C-<sup>18</sup>O carbonate paleothermometry.** Previous stable isotope paleothermometry studies have used the δ<sup>18</sup>O value of authigenic minerals to obtain information on past surface temperatures and surface waters, and thereby infer the paleoelevation of the earth's surface (Chamberlain and Poage, 2000). However, this approach requires important assumptions regarding climatic change and seasonality. The <sup>13</sup>C-<sup>18</sup>O isotope paleothermometer is a new technique that constrains carbonate growth temperatures based on the temperature-dependent “clumping” of the <sup>13</sup>C and <sup>18</sup>O bonds in the solid carbonate phase alone,



**Figure 3. A) Two time-temperature paths consistent with speculation of an old (ca. 35 Ma) versus a young (ca. 6 Ma) origin for the Grand Canyon. Both thermal histories generate single-grain (U-Th)/He apatite dates of 36 Ma, consistent with our (U-Th)/He data for basement samples in the upper Grand Canyon. B) Evolution of the <sup>4</sup>He/<sup>3</sup>He ratio (normalized to the bulk ratio) during step heating of single apatite grains that evolved along the thermal histories depicted in A). These data reflect the spatial distribution of <sup>4</sup>He in the apatite grains, and are clearly very sensitive to differences in the cooling paths. In this example, both the edge dates and the shape of the ratio evolution diagram can be used to clearly distinguish an older Grand Canyon incision event from a more recent one. For reference, a typical error bar would be about 2x the size of the symbols in 2B.**

independent of any solid-fluid equilibria (Ghosh et al., in press-a). Carbonate ion groups that contain both a <sup>13</sup>C and an <sup>18</sup>O atom (i.e., <sup>13</sup>C<sup>18</sup>O<sup>16</sup>O<sub>2</sub><sup>=</sup>) have lower zero point energies than their isotopically ‘normal’ and singly-substituted relatives (i.e., <sup>12</sup>C<sup>16</sup>O<sub>3</sub><sup>=</sup>, <sup>13</sup>C<sup>16</sup>O<sub>3</sub><sup>=</sup>, <sup>12</sup>C<sup>17</sup>O<sup>16</sup>O<sub>2</sub><sup>=</sup> and <sup>12</sup>C<sup>18</sup>O<sup>16</sup>O<sub>2</sub><sup>=</sup>), leading to a thermodynamic driving force that promotes the bonding of rare isotopes with each other as opposed to being randomly dispersed throughout the mineral lattice. This effect can be described using an isotope exchange reaction among isotopologues of the carbonate ion:

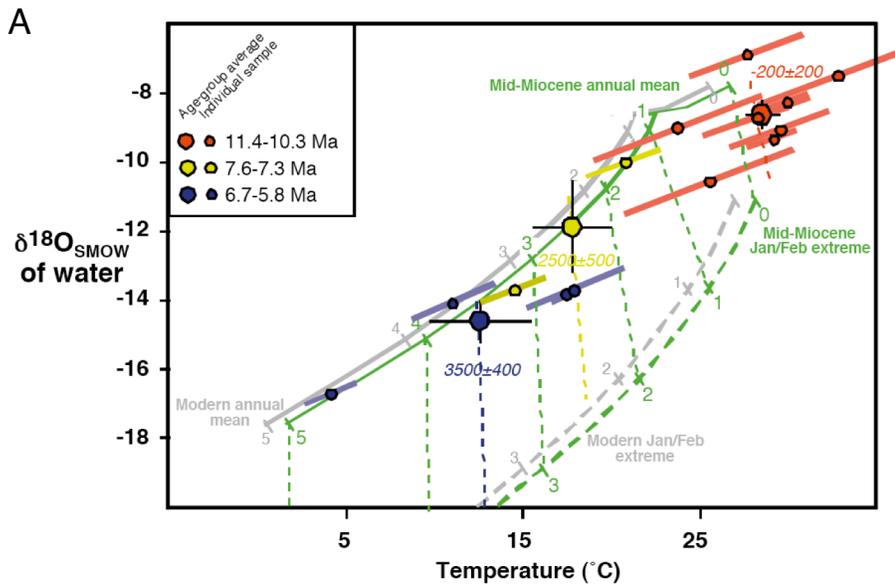


The equilibrium constant for this reaction increases with decreasing temperature, and can be determined by digesting a carbonate mineral in phosphoric acid and measuring the δ<sup>18</sup>O, δ<sup>13</sup>C, and abundance of mass-47 isotopologues (mostly <sup>13</sup>C<sup>18</sup>O<sup>16</sup>O) in product CO<sub>2</sub>. From these data, one can calculate the enrichment of mass-47 isotopologues in product CO<sub>2</sub> relative to the ‘stochastic’, or random, distribution of all C and O isotopes among all possible isotopologues. This enrichment, termed the Δ<sub>47</sub> value, is proportional to the equilibrium constant for the above reaction in reactant carbonate and varies with carbonate growth temperature by the relation:

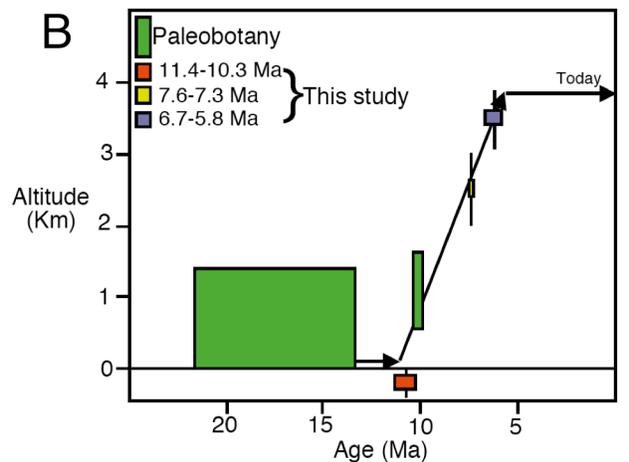
$$\Delta_{47} = 59200/T2 - 0.02$$

where Δ<sub>47</sub> is in units of per mil and T is temperature in Kelvin (Ghosh et al., in press-a).

In contrast to conventional methods, this thermometer does not require knowledge of the  $\delta^{18}\text{O}$  of the waters from which the mineral grew or the isotopic composition of other phases with which the mineral may have isotopically exchanged. This thermometer can therefore more effectively distinguish the effects of altitude, climate, latitude and seasonality because it independently and simultaneously 1) constrains the growth temperature of soil carbonate that can be compared to a known elevational temperature gradient, 2) determines the  $\delta^{18}\text{O}$  of the waters from which the carbonate grew that can be compared to an elevational dependence of  $\delta^{18}\text{O}$  of the waters, and 3) thus allows examination of the correlation between soil temperature and the  $\delta^{18}\text{O}$  of the waters in a suite of samples. This latter point is a



**Figure 4. A) Plot of  $\delta^{18}\text{O}_{\text{SMOW}}$  value of water in equilibrium with soil carbonate nodules vs. the growth temperatures of those nodules. Small symbols are individual samples. Large symbols are averages for the 11.4-10.3 Ma, 7.6-7.3 Ma and 6.7-5.8 Ma age groups. Error bars for individual measurements are based on external precision in  $\delta^{18}\text{O}$  for acid digestion analyses of carbonates, and consider the effect of errors in T on estimated  $\delta^{18}\text{O}_{\text{SMOW}}$  values of water. Error bars for age groups are  $\pm 1$  standard error of the population. Gray curves show the mean annual trend (solid curve) and trend of Jan/Feb extremes (dashed curve) for the relationships between surface temperature and  $\delta^{18}\text{O}_{\text{SMOW}}$  of meteoric water. These curves are contoured for altitude in km. The similar green curves plot the expected location of the mean annual and Jan/Feb extremes in the mid-Miocene, based on inferred changes in the latitude of Bolivia and low-latitude climate change. Fine dashed lines connecting the mid-Miocene mean annual trend and Jan/Feb extreme trend show the slopes of seasonal variations in T and  $\delta^{18}\text{O}$  of water at a fixed altitude (we infer these were the same in the Miocene as today). Paleoelevations of age-group averages are estimated by their intersections with this set of altitude contours, as indicated by the red, yellow and blue dashed lines. B) Constraints on the uplift history of the Altiplano, based on new  $^{13}\text{C}$ - $^{18}\text{O}$  isotope paleothermometry (blue boxes, Ghosh et al., in press-b, with error bars indicating  $\pm 1$  standard error) and previous paleobotanical constraints.**



critical advantage of this approach, because variations of  $\delta^{18}\text{O}$  with temperature due to altitude change may contrast strongly with variations due to temperature change at a single altitude (climatic effects), seasonality and changes in latitude. In other words, the independent T and  $\delta^{18}\text{O}$  constraints may permit evaluation of whether or not any measured temperature variations in a suite of samples result from elevation change.

This new technique has been used to constrain mid- to late Miocene surface uplift rates in the Altiplano plateau in the Bolivian Andes (Ghosh, in press-b). The stable isotopic compositions of pedogenic soil carbonates varying in age between 11.4 and 5.8 Ma at modern elevations of  $\sim 3900$  m yield a relationship between carbonate growth temperature and the  $\delta^{18}\text{O}$  of parental waters that parallels the modern mean-annual surface temperature and annual weighted average meteoric water trend (Figure 4A), which is opposite in slope and an order of magnitude lower than the variation due to seasonal temperature variations. The modern mean-annual trend was shifted to account for second-order latitudinal and climate change effects during the mid-Miocene, with consideration of seasonality and evaporative effects on carbonate growth (Figure 4A). The  $\delta^{18}\text{O}/\text{T}$  slope in Figure 4A is  $0.34 \text{ ‰}/^\circ\text{C}$ , indistinguishable from the slope defined by the mean annual altitude gradient in surface temperature and  $\delta^{18}\text{O}$  of meteoric water in the Altiplano and surrounding areas, and contrasts with slopes in these dimensions associated with low-latitude climate variations (ca.  $0 \text{ ‰}/^\circ\text{C}$  for temperatures between 29 and 12  $^\circ\text{C}$ ), latitude variations (ca.  $3 \text{ ‰}/^\circ\text{C}$ ; or  $0.6 \text{ ‰}$  per degree of latitude), or seasonality at any one altitude (ca.  $-5 \text{ ‰}/^\circ\text{C}$ ), but varying with altitude; see dashed lines in Figure 4A).

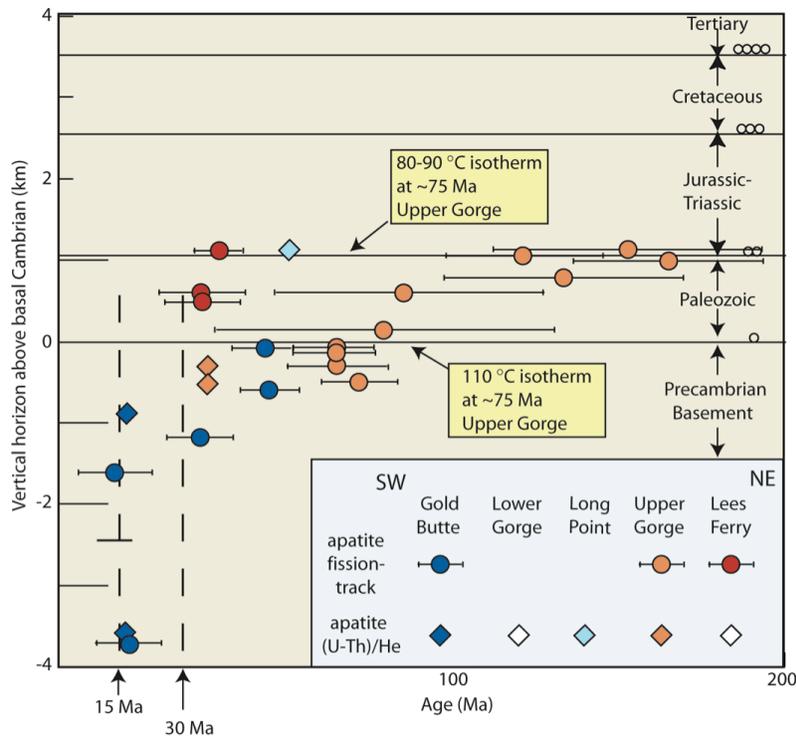
This information is used to estimate an Altiplano surface uplift rate of  $1.03 \pm 0.014$  mm/yr between  $\sim 10.3$  and 6.7 Ma (Figure 4B). Application of this technique to carbonate soils, lacustrine sediments and fossils in the Colorado Plateau, accounting for seasonality, climate and latitudinal shifts, and fractionation, may similarly provide useful constraints on elevation history. Uncertainties in measuring temperature with this thermometer are  $\sim 2^\circ\text{C}$ , and thus we would generally elevation error of  $\sim \pm 400$  m. Although our expected signal is only 50% of the Andean values, it is still well above these uncertainties.

#### 4 Proposed Research

**Unroofing History.** Our strategy for refining the Tertiary unroofing history of the region is two-fold. First, we will complete a N-S transect of apatite (U-Th)/He (AHE) ages in an effort to track first-order differences in cooling history across the structural terrace. This should provide an accurate history of the stripping of Mesozoic cover. This will include a 10-site traverse between the Long Point area and the Vermillion Cliffs using Permian and Triassic sandstones (Figure 2). Second, we will examine when incision of the Grand Canyon occurred, and whether it was rapid or slow, using  $^4\text{He}/^3\text{He}$  thermochronometry. This will include two samples in the Lower Granite Gorge, one on either side of the intersection of Peach Springs Canyon with the Grand Canyon, to test the hypothesis that the Paleocene Peach Spring Canyon followed the course of the Grand Canyon upstream from this intersection, while the lowermost section of the canyon was carved much later. It will also include one basement sample southwest of Peach Springs to evaluate the timing and rate of denudation in the Laramide uplands just southwest of the plateau. We will also apply this technique to the Upper Granite Gorge and the Lees Ferry area. Although we have already sampled much of this material, and we have full access to an existing archive of mineral separates from the Upper and Lower Granite Gorges (see Bowring letter, Section I), we anticipate that we will need at least one two-week sampling trip in both Year 1 and Year 2 of the project.

The existing thermochronological database for the southwestern Colorado plateau includes a roughly horizontal AFT (apatite fission track) transect down the Colorado River from Lees Ferry to Grand Wash (Naeser et al., 1989; 2001; Kelley et al., 2001), a vertical AFT profile in the Grand Canyon Village area (Dumitru et al., 1994), and a “vertical” profile of the Gold Butte block (tilted at 15 Ma) just west of the plateau (Fitzgerald et al., 1991), that also includes an AHE transect beneath the basal Cambrian unconformity (Reiners et al., 2000). AFT data from the river-level profile, while showing relatively large analytical uncertainties show a marked overall increase in age from northwest to

southeast, from 30-40 Ma near Lees Ferry up to 70-80 Ma in the Lower Granite Gorge, consistent with the idea of SW to NE stripping of the Mesozoic cover (Naeser et al., 1989). Vertical transect data are summarized in Figure 5. The AFT data in the Grand Canyon Village area suggest the base of a ~75 Ma partial annealing zone is preserved within the Paleozoic section, with complete Laramide annealing below the basal Cambrian unconformity. These data suggest that the 110°C isotherm lay near the base of the Paleozoic at the end of the Cretaceous, and the 80 to 90°C isotherm lay near the top of Paleozoic (Dumitru et al., 1994). These observations suggest that the Kaibab uplift developed near the end of the Cretaceous, stripping some (but not all) of the younger cover from the Paleozoic and preserving the base



**Figure 5. Summary of age vs. stratigraphic position for apatite fission-track and apatite (U-Th)/He systems in the southwestern Colorado Plateau, from Dumitru et al., (1994), Fitzgerald et al. (1991), and Reiners et al. (2000). (U-Th)/He apatite dates from the SW plateau margin and Upper Granite Gorge are new data. White diamonds represent the locations of samples for future (U-Th)/He apatite analysis. Additional samples between these listed locations will be analyzed as needed for comprehensive reconstruction of the low temperature denudational history across the region. These and future data provide the framework for targeting the optimal samples for  $^4\text{He}/^3\text{He}$  apatite thermochronometry (sample locations shown in Figure 1).**

of the fossil partial annealing zone (Dumitru et al., 1994). Ages within and SW of the Kaibab uplift are generally Early Tertiary or Late Cretaceous. The mid-Tertiary ages in the Lees Ferry area and possibly one age from the Upper Granite Gorge (~30 to 40 Ma) suggest much younger stripping of the Mesozoic there, perhaps controlled by post-annealing vertical motions in both Laramide and post-Laramide time (Naeser et al., 2001) (Kelley, 2001). At Gold Butte, AFT ages just below the basal Cambrian unconformity are similar to, or perhaps slightly younger than, ages in the Grand Canyon Village area, consistent with this pattern (Figure 5). At greater structural depth, the base of a ~15 Ma partial annealing zone is preserved 1 km below the unconformity. AHE ages from these structural levels, as expected, are also 15 Ma (Figure 5).

These data indicate that both the Laramide and mid-Miocene geothermal gradients were between 20 and 30°C/km (Fitzgerald et al., 1991; Dumitru et al., 1994), and that the projected base of the AHE

partial retention zone ( $\sim 75^\circ\text{C}$  for most apatites and cooling rates) lay above the Paleozoic section. Relatively short confined track lengths (12  $\mu\text{m}$ ) observed for structurally deep samples suggest slight reheating during the Tertiary (Dumitru et al., 1994); age and track length modeling from Upper Granite Gorge samples suggest a rapid pulse of Late Tertiary cooling which does not appear to have affected the Lower Granite Gorge (Kelley, 2001). Thus our  $^4\text{He}/^3\text{He}$  samples from these two localities have strong potential to replicate these observations. Thus, based on these data the general thermal structure of the region is appropriate for gauging the timing of removal of the Mesozoic cover using AHE from upper Paleozoic or lowermost Mesozoic samples.

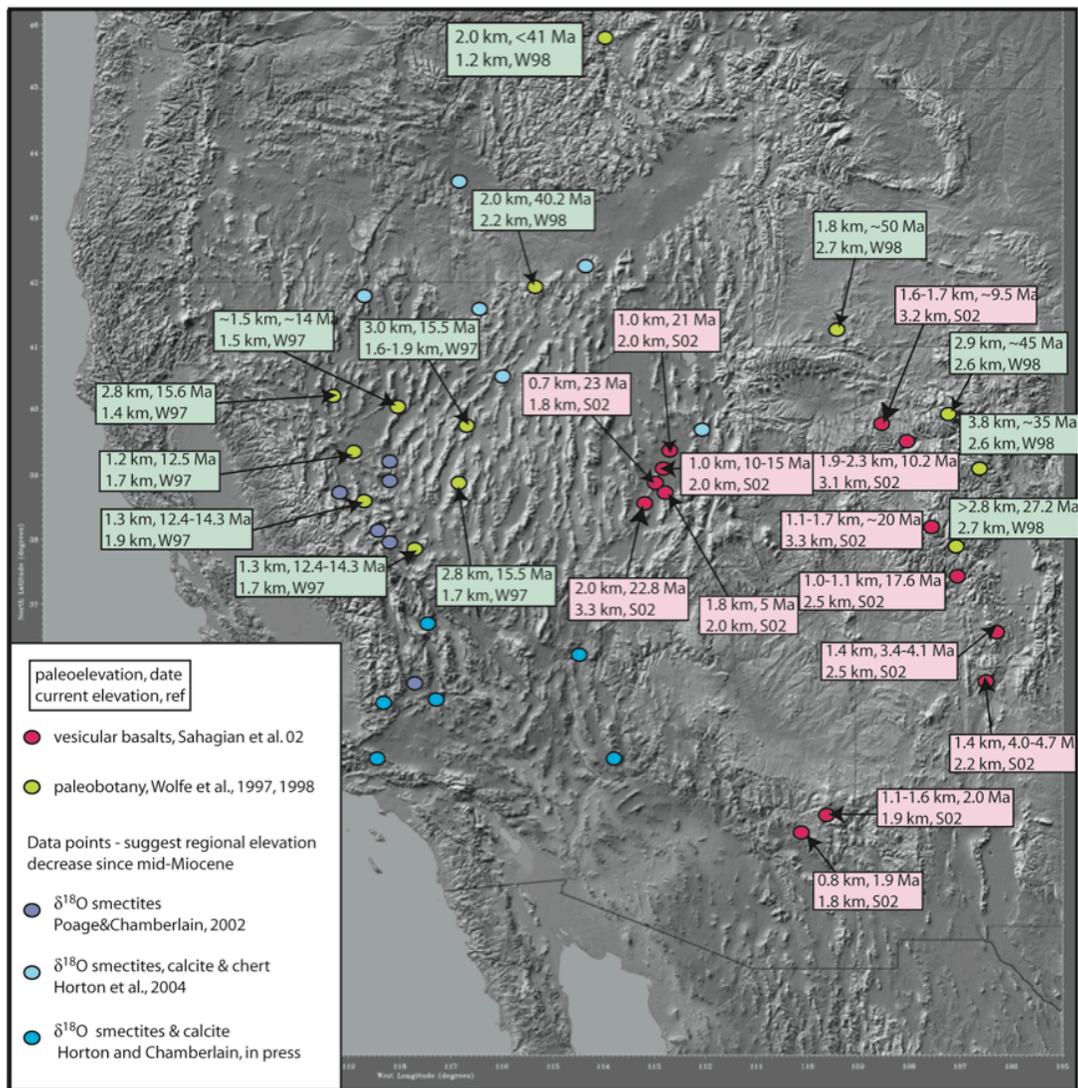
In a feasibility study for this project, we have dated three samples using the conventional AHE techniques (red samples in Figure 2). Two samples in the Upper Granite Gorge, one within and one east of the Kaibab uplift, yielded identical ages of 35 Ma, suggesting that any thermal signature from the Laramide uplift had been removed by mid-Oligocene time. Another analysis from sandstones in the Triassic Moenkopi Formation near Long Point (Figure 2) yielded an age of 55 Ma. Given the position of this sample just below the unconformity with the Rim gravels that are presumably no younger than late Eocene ( $\sim 48$  Ma), this age suggests a relatively narrow window of time to strip a relatively thick Mesozoic section (1500 to 2000 m) and deposit the Rim gravel. Thus far our preliminary results are consistent with the trend in fission track data, and demonstrate the feasibility of obtaining robust ages from detrital apatites in Mesozoic sandstones. We recognize that on the basis of our initial results our strategy may change somewhat in the course of our study. The sampling strategy shown in Figure 2 is a minimum for what we believe will place meaningful new quantitative constraints on the timing, amount and rate of Cenozoic unroofing across the region.

**Paleoaltimetry.** For this component of our investigation, our strategy will be to sample the freshest carbonate possible from key stratigraphic sections both on and adjacent to the Plateau, as summarized in Table 1 and Figure 2. Although our sampling region is restricted to key units on the southwestern margin of the plateau, if time permits we would envisage broadening the scope somewhat to include Paleocene and Eocene lacustrine units along the northern margin of the plateau. Our target localities include analyzing samples from both Early and Late Tertiary carbonate units on the plateau, including the Rim gravels ( $\sim 55$  to 48 Ma) and the Bidahochi Formation (16 to 6 Ma; Table 1), which we hope will differentiate between a Laramide, mid-Tertiary, and Late Tertiary uplift hypotheses. In addition, we will analyze samples from the late Oligocene/early Miocene Rainbow Gardens Member of the Horse Spring Formation, where it lies just above an angular unconformity that cuts northward across the lower half of the Mesozoic section (point labeled "T" on Figure 2), which will give a pre-extension (24 to 16 Ma) measurement along the westernmost margin of the plateau. We will also sample the Haulapai Limestone and possibly the Bouse Formation in the lower Colorado River trough, for the purposes of evaluating whether extension resulted in a net lowering of elevation in the central Basin and Range as suggested by Horton et al. (2004). In many instances fresh, well characterized carbonate samples already exists in collections from these formations and will not need to be sampled in the field (see Patchett letter, Section I), however we anticipate the need to visit the field to collect some of the material for analysis.

Previous work in paleoaltimetry in the western US is summarized in Figure 6. As can be seen, in general these determinations have focused on the Rocky Mountains and Basin and Range provinces, with a dearth of estimates from the southern Colorado Plateau region. The estimates based on leaf physiognomy and stable isotope techniques generally support the idea of high elevation throughout Tertiary time. An important exception to this overall trend, which includes the only published data from the southern part of the plateau, comes from basalt vesicle studies of (Sahagian et al., 2002a), which suggest a general acceleration of uplift in Late Tertiary time, and a rather dramatic increase in elevation in the last two million years in samples closest to our study area. The overall trend is controversial (Libarkin and Chase, 2003; Sahagian et al., 2003), and the subject is actively under investigation by the Sahagian group using very young basalts (1000 kyr) near Flagstaff, Arizona. We believe the results of

this study will provide an important complementary dataset to ongoing basalt vesicle studies in the region.

Previous stable isotope studies, for which we do not show absolute elevation on Figure 6, exploit the observation that the  $\delta^{18}\text{O}$  isotopic composition of precipitation is correlated with altitude to infer paleoelevation. A compilation of data from mountain belts around the globe quantified this effect as  $\sim 0.28$  per mil/100m, with no significant differences in isotopic lapse rates in most regions except at extreme latitudes (Poage and Chamberlain, 2001). Because of this relationship, these data are most effectively used to determine the net paleoelevation change rather than absolute elevation, which is sensitive to variability in precipitation and seasonality at a given elevation (Poage and Chamberlain, 2001). Second order effects of climate change and latitude must still be addressed. In the case of the western U.S., it has remained at approximately the same latitude through the Cenozoic (Smith et al.,



**Figure 6. Shaded relief map of the western U.S. showing the locations of previously published paleoaltimetry datasets. The vesicular basalt and paleobotanical locations include the current elevation, age, and paleoelevations of previous data (Wolfe et al., 1997; Wolfe et al., 1998b; Sahagian et al., 2002). The stable isotope data suggest regional elevation decrease since the mid-Miocene, with details on trends in the stratigraphic sections at the locations shown given in the listed references (Poage and Chamberlain, 2002; Horton et al., 1994; Horton and Chamberlain, in press).**

1981), and hence will have a similarly small effect as in the Andes. We will estimate the effect of Cenozoic climactic changes on the carbonate isotopic record using previous studies of these effects (Zachos et al., 2001; Wolfe, 1994).

**Table 1.  $^{13}\text{C}$ - $^{18}\text{O}$  Paleoaltimetry Targets**

Location	Material to be Sampled	Timing Constraints	Current Elev (m)	References
<u>Hualapai Limstone</u>				
Top of section	Limestone, root and stem casts, ostracods	6 Ma	630 m	Spencer et al., 2001
Bottom of section	Limestone, root and stem casts, ostracods	11 Ma	330 m	Faulds et al., 2001
<u>Bidahochi Formation</u>				
Upper member	Tufa	6 Ma	1920	Dallegge et al., 2001
Middle member	Vertebrate fossil material at White Cone faunal locality	8.5 to 6 Ma	1860	Dallegge et al., 2001
Lower member	Lacustrine sediments (5-20 cm thick marl horizons), freshwater ostracods, vertebrate fossils at Cheto Mine	15.8 Ma to ~12 Ma	1730-1830	Dallegge et al., 2001; Love, 1989; Gross et al., 2001; Repenning and Irwin, 1954
<u>Rainbow Gardens Member of the Horse Spring Formation</u>				
Upper member	Root casts, carbonate beds	26 to 18 Ma	330-430	Beard, 1996
Middle member	Root casts, bivalves, gastropods, ostracodes, lacustrine sediments	26 to 18 Ma	290-330	Beard, 1996
<u>Claron Formation</u>				
	Ostracodes, gastropods, lacustrine limestones	Late Paleocene to Eocene	~2400 m	Goldstrand and Eaton, 2001; Goldstrand, 1989
<u>Rim Gravels (Long Point)</u>				
	Viviparid gastropods	Paleocene/early Eocene	1770-1800	Young and Hartman, 1984; Elston et al., 1989
<u>Rim Gravels (Milkweed Canyon)</u>				
West Water Formation	Lacustrine limestones	Paleocene/early Eocene	1345-1325	Billingsley et al., 1999
Music Mountain Formation	Fossils in arkosic sedimentary unit	Paleocene/early Eocene	1325-1175	Billingsley et al., 1999

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