A Sensitivity Analysis of the Apatite (U-Th)/He Thermochronometer

R.A. Wolf, K.A. Farley, and D.M. Kass

Division of Geological and Planetary Sciences
California Institute of Technology

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Pasadena, California 91125

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Corresponding Author:

Kenneth A. Farley
farley@gps.caltech.edu
(626) 395-6005
(626) 568-0935 (Fax)
Abstract

Apatite (U-Th)/He cooling ages of crustal rocks generally reflect residence for extended periods at temperatures where helium is neither quantitatively retained nor lost by diffusion. To better characterize the response of apatite helium ages to thermal histories involving partial helium retention, we explored solutions to the helium production-diffusion equation. Under thermally static conditions the analytical solution to this equation demonstrates that the zone of partial helium retention extends from about ~45°C to ~85°C based on laboratory diffusivity measurements. This zone lies at temperatures ~35°C cooler than the analogous fission track partial annealing zone. He ages within the partial retention zone ultimately achieve a balance between He production and loss, yielding a steady state He age. Both the ultimate age and the time it takes to achieve this age are temperature dependent. For example, an apatite held at 75°C equilibrates to an age of ~2 Ma after ~17 Ma, regardless of whether equilibrium is approached from higher or lower initial He ages. For representative dynamic thermal histories we evaluated apatite He ages using a numerical solution to the ingrowth-diffusion equation. The results illustrate the sensitivity of He ages to various geologic histories and are useful for understanding He age-elevation relationships and for evaluating time-temperature paths derived from fission-track length-distributions. In addition, although helium diffuses rapidly from apatite at shallow crustal temperatures, modeling of ambient temperature fluctuations and heating by natural fires indicates
that helium ages are nearly unaffected by surficial processes.

1. Introduction

Laboratory diffusion measurements indicate that radiogenic He is retained in apatite under mean Earth surface conditions, but is lost by diffusion at only slightly higher temperatures (Zeitler et al., 1987, Lippolt et al., 1994, Wolf et al., 1996). As a consequence, apatite (U-Th)/He ages are sensitive to very low temperatures (<100°C) and in most cases reflect cooling ages (Wolf et al., 1996). The simplest way to interpret cooling ages, proposed by Dodson (1973), is to assume a constant cooling rate with time. With this assumption an age can be related to a specific closure temperature computed from the diffusion characteristics of the daughter product. For a cooling rate of 10°C/Ma, the He closure temperature in apatite is about 75°C (Wolf et al., 1996). However, at the low temperatures relevant to He in apatite, cooling histories are likely to be more complex, reflecting the tectonic and erosional processes which expose rocks on the Earth's surface. The closure temperature concept is of limited use under these conditions, and an alternative approach is required. Here we explore solutions of the radiogenic helium production/diffusion equation for a variety of geologic situations to illustrate the sensitivity characteristics of the apatite (U-Th)/He thermochronometer.

Our model of helium production and diffusion assumes that helium is produced only by the decay of uniformly distributed uranium and thorium, and that
it is lost only by volume diffusion. We assume a spherical diffusion geometry, consistent with laboratory measurements of helium diffusion from apatite (Wolf et al., 1996). The local helium concentration gradients resulting from ejection of high energy alpha particles from grain surfaces (Farley et al., 1996a) can be ignored because the diffusion domains are much smaller than the physical grain size (Wolf et al., 1996); i.e., this phenomenon does not affect most of the diffusion domains. Furthermore, because the diffusion domains are smaller than the grains themselves, our results apply to apatites in general, rather than only to those of some particular physical grain size.

Given these assumptions the helium age equation is:

\[ ^4\text{He} = 8\cdot 238 \cdot U(t)(e^{-\lambda_{238}t} - 1) + 7\cdot 235 \cdot U(t)(e^{-\lambda_{235}t} - 1) + 6\cdot 232 \cdot Th(t)(e^{-\lambda_{232}t} - 1) \]  

(1)

where \(^4\text{He}(t), U(t)\) and \(\text{Th}(t)\) are amounts present at time \(t\), and \(\lambda\)'s are the decay constants. As a consequence of radiogenic production and diffusive loss, the concentration of helium as a function of the dimensionless radial position \(r\) within the spherical diffusion domain of radius \(a\), is:
\[
\frac{\partial^4 \text{He}(r,t)}{\partial t} = \frac{D(t)}{a^2} \left[ \frac{\partial^2 \text{He}(r,t)}{\partial r^2} + \frac{2}{r} \frac{\partial \text{He}(r,t)}{\partial r} \right] + 8 \lambda_{238}^{238} \text{U}(t) + 7 \lambda_{235}^{235} \text{U}(t) + 6 \lambda_{232}^{232} \text{Th}(t) \tag{2}
\]

\(D(t)\) is the time dependent diffusion coefficient obeying an Arrhenius relationship such that:

\[
\frac{D(t)}{a^2} = \frac{D_0}{a^2} \exp \left[ -\frac{E_a}{RT(t)} \right] \tag{3}
\]

where \(D_0\) is the diffusivity at infinite temperature and \(E_a\) the activation energy measured in laboratory experiments, \(R\) is the gas constant, and \(T(t)\) is an arbitrary thermal history. We assume a zero concentration boundary. In all of the illustrations discussed below we adopt a present day \(^{238}\text{U}/^{235}\text{U}\) ratio of 137.88 and a \(\text{Th}/\text{U}\) ratio of 1. The results are rather insensitive to the \(\text{Th}/\text{U}\) ratio because the yield and half life of these two isotopes are similar.

The diffusion characteristics of He from apatite have been investigated by several groups (Zeitler et al., 1987, Lippolt et al., 1994, Wolf et al., 1996), with somewhat varying results. For the purposes of this paper we adopt the diffusivities measured by Wolf et al. (1996), as these were measured with the highest precision.
and at temperatures requiring the smallest extrapolation to the conditions of geologic interest. Adoption of values from other workers would lead to qualitatively similar but quantitatively different results. Unless otherwise stated, we use diffusivities measured on Durango fluorapatite for our calculations. In several cases we also investigate the effects of using diffusivities obtained by Wolf et al. (1996) on apatites from broadly granitic rocks. These apatites vary with respect to chemical composition and grain size, yet, like Durango apatite, yield a narrow range of closure temperatures, 75±5°C (assuming a cooling rate of 10°C/Ma).

2. Isothermal diffusion: an analytical solution

Initially, we investigate the simplest case of production and diffusion in an isothermal apatite. By making the assumption that He production is constant with time, equation (2) can be solved analytically. This assumption is justified for timescales over which the parent abundances vary only slightly, <~3 x 10^8 yrs. For instance, over 200 Myrs, (e^{λ_{238} t} - 1) differs from λ_{238}t by 1.5%, and thorium by even less (^{235}U varies by much more, but its helium contribution can be ignored, e.g., over 200 Myrs it contributes only 0.5% of the He produced by {^{238}U decay}). Therefore, equation (1) can be rewritten as:
where \( P \) is the constant production rate. The general solution to equation (2) (Ozisik, 1989), assuming an initial homogeneously distributed concentration of helium,

\[
\frac{^4He}{P} = t' = \frac{a^2}{D} \left[ \frac{1}{15} - \sum_{n=1}^{\infty} \frac{6}{n^4 \pi^4} \exp\left( -n^2 \pi^2 \frac{D}{a^2} t \right) \right] + t''  \sum_{n=1}^{\infty} \frac{6}{n^2 \pi^2} \exp\left( -n^2 \pi^2 \frac{D}{a^2} t \right)
\]

\( ^4\text{He}^*(t=0) \), is then:

where \( t' \) is the helium age, \( t \) is the time over which isothermal diffusion and production occur, and \( t'' \) is the initial helium age \( (t'' = \frac{^4\text{He}^*(t=0)}{P}) \).

Solutions to eq. 5 for various temperatures and starting conditions are shown in Figure 1. The exponential terms in eq. 5 disappear for large values of \( (D/a^2)t \), and equilibrium between helium production and loss is ultimately achieved to yield a steady-state helium age of 0.0667\((a^2/D)\). The e-folding time for approach to equilibrium is well-approximated by the first term in the summations \( (n=1) \), and is given by \( t^* = (a^2/D)/\pi^2 \). In 5\( t^* \), the approach to equilibrium is largely complete. For example, an apatite held at 75°C will reach a steady-state He age of 2.3 Ma after being held for ~17 Myrs. In contrast the same apatite held at 25°C will ultimately
equilibrate to a He age of $\sim 10^{10}$ years but will require $>10^{11}$ years to get there. For practical purposes He retention at mean Earth surface temperatures is quantitative. The steady-state age and e-folding time are the same whether equilibrium is approached from a lower or a higher initial He age. The latter is useful for interpreting He ages in sedimentary apatites. For example, an apatite with a provenance age of 100 Ma instantaneously heated to 100°C will equilibrate to a steady state age of $\sim 0.07$ Ma after just $\sim 0.5$ Myrs, but will require nearly 200 Myrs to equilibrate to 24 Ma at 60°C.

Figure 1 illustrates how helium ages change with time at different temperatures, but does not clearly demonstrate how helium ages differ with temperature for a constant holding time $t$. Such a view would be useful for considering helium age variations with depth in the crust. Figure 2 shows He ages as a function of temperature for a holding time of 50 Myrs and with no initial helium. The He ages change rapidly over a fairly narrow temperature range which we call the helium partial retention zone, HePRZ. The HePRZ is analogous to the fission track partial annealing zone (e.g., Gleadow et al., 1981; Gleadow et al., 1983; Gleadow and Fitzgerald, 1987; Wagner et al., 1989). We operationally define the HePRZ to be the range of temperatures where helium ages fall from 90% to 10% of the holding time (i.e., $0.1 \leq t/\tau \leq 0.9$). For a holding time of 50 Ma, the HePRZ is between $\sim 40^\circ$ and $70^\circ$C (Figure 2).

One consequence of defining the HePRZ in this way is that the temperature
range of the HePRZ depends on holding time. This effect can be seen most clearly when the fractional helium age \( \frac{t_f}{t} \) is plotted against holding time (Figure 3). At sufficiently low temperatures, helium is quantitatively retained, so \( \frac{t_f}{t} = 1 \) and plots horizontally in Figure 4. However, at higher temperatures, helium is only partially retained, so \( t_f \) does not increase as fast as \( t \), and the fractional helium age decreases with holding time. The 10% and 90% limits we have chosen to define the HePRZ are indicated in a shaded region in Figure 3. Figure 4 shows the range of temperatures bounding the HePRZ as a function of time. Whereas the HePRZ lies between \( \sim 50^\circ C \) and \( \sim 83^\circ C \) for a holding time of 10 Myrs, it lies between \( \sim 38^\circ C \) and \( \sim 67^\circ C \) for a holding time of 100 Myrs. In a typical 20°C/km geothermal gradient, the HePRZ resides at \( \sim 2\pm1 \) km over 10^6 to 10^8 year timescales.

These calculations assumed the diffusion coefficients we measured for Durango apatite; to determine the sensitivity of the HePRZ to the diffusivity variations measured by Wolf et al. (1996), we repeated the calculations for these apatites in Figure 5. The curves are very similar, but are shifted slightly relative to the Durango apatite. The maximum shift from the Durango curve is only about 8°C. There is no correlation between the relative position of the curves and the chemical or physical make-up of the apatites. However, these curves overlap at the limit of the reported uncertainties. Therefore, we cannot determine whether the differences among these curves are the result of real variations in the diffusivity or are an artifact of the uncertainties on the diffusion coefficient measurements.
Like apatite He ages, apatite fission track ages are sensitive to very low temperatures. We investigate the expected relationships between He and fission track ages under isothermal conditions (held for 120 Ma), specifically to compare the He partial retention zone with the fission track partial annealing zone. To simulate the fission track partial annealing zone, we use a model called AFTSolve (Ketcham et al., in review). The model HePRZ lies between ~37°C and ~65°C. The fission track partial annealing zone defined in an analogous fashion to the HePRZ resides between ~70°C and ~105°C. The HePRZ and the fission track partial annealing zone do not overlap; helium ages decline to 0 just as the fission track ages begin to decrease (Figure 6). The HePRZ lies ~35°C cooler than the partial annealing zone, and in typical geothermal gradients of ~15-35°C/km will reside ~2±1 km shallower. Thus, we predict that helium ages will always be ≤ fission track ages. We predict that in some cases (situations including as much as ~2 km of exhumation), helium ages could be almost completely reset in samples which do not exhibit the effects of fission track age reduction.

Measured helium ages from the Otway Basin, South Australia should be one possible test of the model comparison between the HePRZ and the fission track partial annealing zone, as well as a test of the applicability of laboratory helium diffusion data to natural systems. Apatite fission track data from the Otway Basin are commonly taken as the best documented fission track partial annealing zone and an important thermal calibration in the natural environment (Gleadow et al., 1981;
Gleadow et al., 1983; Green et al., 1989). Green et al. (1989) modeled apatite fission track ages as a function of temperature based on laboratory annealing experiments and demonstrated reasonable agreement with measured fission track ages in the Otway Basin. There is only a subtle difference between the inferred thermal history for the Otway Basin sediments (Green et al., 1989) and an isothermal history held for 120 Ma. Therefore, the model HePRZ in Figure 6 should closely resemble the helium ages measured on Otway Basin sediments if laboratory diffusivity measurements are applicable in nature. Analyses of Otway Basin apatites are presently in progress.

3. Complex thermal histories: a numerical solution

When temperature and diffusivity vary with time, there is no general analytical solution to the helium production-diffusion equation (2). We used a numerical approach (the Crank-Nicolson method, Crank and Nicolson, 1947) to calculate the He concentration as a function of time for prescribed time-temperature paths. By inserting the integrated helium concentration into the age equation (1), we calculated the evolution of He age as a function of time.

3.1 Variable thermal histories and helium age-elevation profiles

To determine the sensitivity of helium ages to more dynamic time-temperature paths, we have modeled the evolution of He ages for several illustrative scenarios (Figure 7). Comparable exercises have been undertaken for apatite fission
track ages, with qualitatively similar results (Green et al., 1989; Gallagher, 1995; Ketcham et al., in review). In History 1, an apatite is cooled quickly to 15°C and held there for 40 Myrs; He retention is quantitative at this temperature, so this path represents simple closed-system evolution as might apply to a volcanic rock. All of the other histories include extended periods at temperatures where He is only partially retained. History 2 is monotonic cooling from 140°C to 15°C over 100 Myrs. The transition from total loss to total retention is apparent in this He age trajectory. In this case of constant cooling, the closure temperature can be calculated from the cooling rate to be ~63°C (Dodson, 1973). This type of history may apply to rocks experiencing erosional exhumation, in this case corresponding to a rate of 1.25°C/Ma (equivalent to 63 m/Ma in a geothermal gradient of 20°C/km). In History 3, a sample held at ~60°C for 80 Myrs is rapidly cooled at 20 Myrs to 15°C. On this path the He age approaches an equilibrium value over the first 80 Myrs, then rises linearly for the remaining 20 Myrs. This history might apply to a rock tectonically exhumed 20 Myrs ago. History 4 involves 20 Myrs of cooling followed by an isothermal period at 60°C for 60 Myrs, followed by continued cooling. The He age trajectory for this history is similar to that of History 3. Thermal histories of this type have often been proposed from modeling of fission track length distributions (Lutz and Omar, 1991; Corrigan, 1991). History 5 shows the He age trajectory of a sample experiencing heating from an initial value of 15°C at 100 Ma to a maximum value of ~60°C at 5 Ma, followed by cooling back to 15°C at 0 Ma. This He age
trajectory shows an initial rise eventually followed by decreasing ages as the sample’s He content begins to exceed the equilibrium value for the ambient temperature. This type of history could apply to a slowly subsiding sedimentary basin tectonically reactivated at 5 Ma.

Despite their different t-T paths, all five temperature histories yield apatites at 15°C with indistinguishable helium ages of 40 Ma. However the five histories would produce very different profiles of He age against structural depth. We have calculated He ages as a function of structural depth for each of the temperature paths in Figure 7 by assuming a constant geothermal gradient of 20°C/km extending below the reference t-T path (Figure 8). As shown in Figure 8, the five histories are extremely well-resolved in age-depth space. The profile for History 1 corresponds to an HePRZ evolved for 40 Myrs, and is strongly concave-up. All other histories lie to the left of this curve (i.e., to lower He ages) and are concave down, consistent with the fact that they have all experienced higher temperatures more recently. The linear cooling of History 2 yields a nearly linear age-depth profile with a slope of 65 m/Ma, identical to the modeled erosion rate that generated this t-T path. The sudden ~45°C cooling that occurs in History 3 at 20 Ma is recorded by a sharp inflection in the profile at this age; similar inflections arising from rapid cooling have been recorded by apatite fission track ages (e.g., Fitzgerald et al., 1991, 1993). Although History 4 is generally similar to History 3, it produces an age profile that differs from it by as much as 10 Ma. Finally, the sharply concave-down profile
produced by History 5 is clearly distinct from the concave up equilibrium HePRZ, and would provide strong evidence for subsidence and heating of a sedimentary basin.

It is important to note that only a few samples taken from the uppermost ~1 km suffice to distinguish these profiles; samples drawn from several km of structure are not required. In contrast, 1 km of structure would be insufficient to distinguish these paths using apatite fission track ages of typical precision. Although the exact details of the fission track age variations are very dependent on the annealing model chosen, in general at least 2 km would be required. For example using Laslett’s calibration of Durango apatite annealing (Laslett et al. 1987), and AFTSolve software (Ketcham et al., in review), age-depth profiles for histories 3 and 4 become >10% different in age only at depths of ~2.2 km.

3.2 Comparison with fission track length distributions

The only commonly used dating technique that is sensitive to the same low temperatures as He in apatite is fission track length distribution. The distribution of fission track lengths constrains the t-T path that an apatite has experienced, but typically a family of cooling curves satisfies the distribution equally well (Lutz and Omar 1991, Corrigan, 1991). Because apatite He ages are sensitive to the same range of temperatures as are fission track lengths, and can be measured on the same samples, they can be used to reduce the number of acceptable histories (Farley et al.,
1996b). An example of the family of time-temperature paths that might be
generated by a track-length-derived thermal model is shown in Figure 9. These
paths yield distinct apatite helium ages, with differences in age of more than 25 Ma.
Thermal histories which yield model helium ages that do not agree with measured
ages may be rejected.

The advantage of this approach is that it relies on two different phenomena
that are modeled in completely independent ways. The pair of techniques would
provide important cross-validation of the resulting thermal histories provided the
techniques are properly calibrated. However such calibration clearly requires
further work. For example, extrapolation of laboratory measurements of fission
track length shortening to natural conditions remains problematic (Green et al.,
1989). Similarly, there are only a handful of He diffusion measurements from
apatite, and these are somewhat discrepant (Zeitler et al., 1987, Lippolt et al., 1994,
Wolf et al., 1996).

3.3 Sensitivity to measured diffusivities

To illustrate the sensitivity of He ages to even modest variations in He
diffusivity, we have modeled ages for apatites subjected to the thermal histories in
Figure 7, but instead of using the diffusion characteristics of Durango apatite (as we
have so far), we have repeated the calculation for the diffusivities we recently
published on a range of apatites (Wolf et al., 1996). The quick cooling in History 1
precludes any dependence of age on He diffusivity. However for the other histories the differences can be substantial. The He age ranges are as follows: History 2, 36-44 Ma; History 3, 31-55 Ma; History 4, 29-53 Ma; and History 5, 22-57 Ma. The differences are largest for the histories involving the longest residence in the HePRZ; because History 5 includes the longest duration within the HePRZ, it provides the greatest amplification of differences in diffusivity. The apatites analyzed by Wolf et al. (1996) have He closure temperatures that are the same within error (75±5°C assuming 10°C/Ma cooling rate), yet the model He ages resulting from History 5 differ by as much as 30 Ma. It should be stressed that these diffusivities are the same within laboratory error, and should not be taken to indicate true variability in diffusion parameters. These results again demonstrate that some applications of He thermochronometry may require a very accurate knowledge of the diffusion coefficients of the actual samples dated, especially if the model thermal histories involve long durations in the HePRZ.

3.4 The Effect of Ambient Temperature Variations on Helium Ages

Forest and range fires and even extremes in Earth surface temperatures are thought to cause diffusive loss of cosmogenic He from some minerals (Bierman and Gillespie, 1991; Trull et al., 1995). To assess whether radiogenic He in apatite is susceptible to these effects, we modeled the diffusive loss associated with extreme Earth surface conditions. We start by modeling the effects of natural fires. Following
Bierman and Gillespie (1991), we assume a rather extreme case that fire heats a rock surface to 600°C for 10 minutes. For simplicity we take the surface heating as instantaneous, and by using a thermal diffusivity of 0.01 cm²/s and by assuming that heat travels from a plane source in one direction, we calculate temperature as a function of depth in the rock (Carslaw and Jaeger, 1959). We then make the extreme assumption that the resulting thermal profile is held for the entire 10 minute duration and use the relationship between time, diffusivity and fractional loss published by Crank (1975) to determine how much He is retained as a function of position (Figure 10). For this calculation we assume an initially constant He profile within the diffusion domains. At the rock surface, the fire’s effect is large: more than 30% of the He is lost. However the retention rises rapidly in the rock, reaching unity at just 3 cm depth. Because granite tends to spall after a fire, exposing fresh surfaces, and because typical sampling involves chipping away the outer weathered rind, we believe it is unlikely that apatite helium ages will be substantially modified by the effects of natural fires.

To assess whether extremes in Earth surface conditions can affect apatite helium ages, we investigated the effects of high ambient temperatures as measured in Death Valley, California. This site was selected because it routinely records some of the highest air temperatures on Earth. The Western Regional Climate Center (1997) reports the daily temperature fluctuations averaged over the last quarter century (Figure 11). The annual mean temperature at this site is 25°C, and daily
highs range from 19°C in the winter months up to 45°C in the summer.

The temperature fluctuations experienced by a rock depend on its proximity to the Earth’s surface. At the surface the rock experiences the full diurnal oscillation, possibly with an enhancement above the daily high as a result of direct solar heating. However, the diurnal and yearly oscillations are rapidly damped with depth below the surface. Following Carslaw and Jaeger (1959), the daily oscillation of the diurnal cycle is damped to the daily mean value by ~85 cm depth, and the annual cycle is damped to the mean annual temperature by ~15 m depth.

At the Death Valley mean annual temperature of 25°C (experienced by a rock at >15 m subsurface), Figure 1 shows that He retentivity is quantitative. To investigate the effects of the oscillating temperatures experienced closer to the surface, we compute the mean diffusivity over the annual cycle. Using this diffusivity and the fractional loss equation (Crank, 1975) we calculate how long it takes to lose 2% of an apatite’s He (2% being an arbitrary estimate of significant He loss). Forapatites shallow enough to experience the full annual cycle but not the diurnal cycle (e.g., around 1 m subsurface, the curve labeled “daily mean” in Figure 11), the mean diffusivity is equivalent to a temperature of 31.3°C. At this effective temperature 2% He loss occurs after ~2 Ma (strictly speaking some He ingrowth occurs over this time period, but including this effect using our numerical solution to equation (2) does not alter this observation). For samples right at the surface, tracking the curve labeled “diurnal cycle” in Figure 11, the effective temperature is
only slightly higher, 31.5°C, and again 2% loss requires 2 Ma. Because erosion rates are typically >1m/Ma, it is unlikely that even these extreme temperatures significantly modify He ages.

An additional effect that would cause greater He loss is direct solar heating of the rock surface. The temperature increase associated with such heating is difficult to estimate in a general way, as it is very sensitive to specific geometric and radiative properties of the rock. In a study of granitic boulders in the Mojave desert, Roth (1964) found that rock surfaces exceeded the air temperature typically by no more than 20°C. Adopting this figure as a constant offset added to the daytime temperatures yields a reasonable upper limit to the temperature of rock surfaces in Death Valley (curve labeled "diurnal cycle + solar heating" in Figure 11). Under these conditions the mean diffusivity is equivalent to a temperature of 48°C, and 2% loss occurs in just over 100 ka. In environments where temperatures and insolation are high and erosion is slow, this effect may require consideration.

4. Conclusions

In general apatite He ages cannot be interpreted to reflect the time elapsed since the crystals cooled through the closure temperature (~75°C) because the requisite assumption of a constant cooling rate is unlikely to be valid in the relevant temperature range. Apatite He ages more likely reflect thermal histories involving varying amounts of time in the region where He retention is very temperature
sensitive. We refer to this region as the helium partial retention zone, and it resides between ~45° and ~85°C. Under isothermal conditions a steady state is achieved between radiogenic He production and diffusive loss; both the steady-state He age and the time to achieve the steady state are temperature dependent and can be predicted directly from the He diffusion parameters.

The He partial retention zone does not overlap the fission track partial annealing zone, indicating that the apatite (U-Th)/He technique should record lower temperature histories and younger ages than the apatite fission track method. We predict that in some cases He ages may be essentially zero in apatites with comparatively high fission track ages. Like fission track ages, He age-elevation profiles likely convey rich information on the cooling history of crustal blocks (c.f. Wolf et al. 1997). Because He ages are sensitive to lower temperatures (hence shallower depths) than fission track ages, in some cases far less structural relief may be required to distinguish among various cooling histories using the He method.

Fission tracks are shortened within the He partial retention zone, so He ages should provide quantitative tests of track-length-derived thermal models. Comparison of the two can be used to validate laboratory calibrations of the two techniques. Once the two techniques are adequately cross-calibrated, He ages can be used to reduce the number of permissible thermal histories generated by inversion of the track-length distributions. For some applications it may be necessary to determine He diffusion parameters to higher precision than is now routine.
Although helium diffuses rapidly from apatite at fairly low temperatures, natural fires and extreme Earth surface temperatures should not significantly affect helium ages except under unusual circumstances.
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Figure Captions

Figure 1.
Helium age evolution as a function of time ($t'$ vs. $t$) for apatites held isothermally at various temperatures (labeled). In the upper panel the apatites are assumed to have no initial He. In the lower panel the apatites have an initial He age of 100 Ma (uniformly distributed within the grain). Unlabeled curves represent 25° and 35°C. Calculated from equation 5.

Figure 2.
Apatite He age as a function of temperature for an isothermal holding time of 50 Myrs. The helium partial retention zone (HePRZ) is defined to be the region in which He ages range from 10% to 90% of the holding time, i.e., in the temperature range where He age changes most rapidly with temperature (shaded region). Calculated from equation 5 with no initial helium.

Figure 3.
Apatite fractional helium age ($t'/t$) as a function of isothermal holding at various temperatures assuming no initial helium (from equation 5). Because the fractional age decreases with holding time at a given temperature, the temperatures which bound the helium partial retention zone also decrease with time.
Figure 4.
The range of temperatures bounding the helium partial retention zone as a function of isothermal holding time. Contours represent constant lines of $t_i/t$. The HePRZ lies between 64 and 98°C for a holding time of 1 Myr, and between 38 and 68°C for 100 Myr holding time. The right hand axis shows the depth of the HePRZ in a typical geothermal gradient of 20°C/km: ~2 ±1 km over 10 to 100 Myr timescales.

Figure 5.
The isothermal helium partial retention zone plotted for the range of measured diffusion parameters (Wolf et al., 1996) and a 50 Myr holding time. Within uncertainties propagated from the measured diffusion parameters, the curves actually overlap.

Figure 6.
The isothermal helium partial retention zone compared with the isothermal fission track partial annealing zone (modeled using AFTSolve, Ketcham et al., in review) held for 120 Ma. The HePRZ is ~35°C cooler than the FTPAZ, and the two zones do not overlap. These curves are predicted to apply to apatites in the Otway Basin of Australia (see text).
Figure 7.

Apatite He age evolution curves (dotted lines) for several representative time-temperature paths (solid lines) using the Crank-Nicolson model. All paths yield a He age of 40 Ma.

Figure 8.

He age as a function of structural depth in a 20°C/km geothermal gradient for the time-temperature histories shown in Figure 7. The histories can easily be resolved with ~ 1 km of structural relief; similar resolution using apatite fission track ages would require more than 2 km.

Figure 9.

Comparison of He ages with thermal models derived from apatite fission track length distributions. The upper panel shows the family of allowed thermal histories derived from an inversion of a typical track length distribution (Lutz and Omar, 1991; Corrigan, 1991). The bottom panel shows the evolution of helium ages for each of these thermal histories. The final He ages for these curves differ by more than a factor of two. Thermal histories which produce model helium ages that disagree with measured helium ages can be rejected provided the two techniques are properly calibrated.
Figure 10.
The effects of fire on He retentivity of apatite. The solid line shows the temperature profile in a rock heated to 600°C for 10 minutes. The dashed line shows the fraction of He retained by apatites subjected to such heating. While grains on the surface will lose more than 30% of their He, by just 3 cm depth no significant loss occurs. As a result of natural surface spalling and removal of weathered rind prior to mineral separation, we conclude that apatites affected by fire are unlikely to be present in samples analyzed for He age.
Figure 11.

The average temperature history of Death Valley, California, from the Western Regional Climate Center (1997). The upper panel shows the reported daily high and low temperatures and the daily and annual mean temperatures as a function of calendar day. The lower panel shows the diurnal temperature cycle and an estimate of the temperature enhancement arising from direct solar heating of rock surfaces for a short portion of the year. The temperature history experienced by a rock depends on its proximity to the surface. Based on these temperature variations we conclude that even the full diurnal cycle is unlikely to significantly affect He ages. However under conditions where apatites may experience intense direct solar heating and where exposed surfaces are eroded slowly, some He loss may occur (see text).
Figure 1
Figure 2
Figure 5
Figure 7
Daily Mean

Daily High

Annual Mean

Daily Low

Diurnal Cycle + Solar Heating

Diurnal Cycle

Day of Year

Figure 11