(U-Th)/He Dating: Techniques, Calibrations, and Applications

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INTRODUCTION

The possibility of dating minerals by the accumulation of \(^{4}\)He from U and Th decay has been recognized for many years (e.g., Strutt 1905), but in the century since the idea was first conceived, the method has rarely been applied successfully. After several investigations of (U-Th)/He dating of various minerals (e.g., Damon and Kulp 1957; Fanale and Kulp 1962; Damon and Green 1963; Turekian et al. 1970; Bender 1973; Leventhal 1975; Ferreira et al. 1975) the technique was essentially abandoned as yielding unreliable and usually low ages, presumably as a result of diffusive \(^{4}\)He loss possibly associated with radiation damage. In 1987, Zeitler and coworkers rekindled interest in the method by proposing that in the case of apatite, \(^{4}\)He ages might be meaningfully interpreted as ages of cooling through very low temperatures. Laboratory diffusion data presented by these authors indicated a closure temperature of about 100°C, a value supported by more recent studies (Lippolt et al. 1994; Wolf et al. 1996b; Warkock et al. 1997). Consistent with this interpretation Wolf et al. (1996a) found that apatite \(^{4}\)He ages increase systematically with sample elevation in a mountain range, as expected for exhumation-induced cooling through a low closure temperature. Based on the strength of these results and additional laboratory (Farley 2000) and natural (Warkock et al. 1997; House et al. 1999; Stockl et al. 2000) constraints on \(^{4}\)He diffusivity, recent attention has focused on applications of apatite \(^{4}\)He thermochronometry. There is also renewed interest in \(^{4}\)He dating of other U- and Th-bearing minerals both for dating mineral formation and for thermochronometry. For example, Lippolt and coworkers have undertaken detailed studies of \(^{4}\)He diffusion and dating of various phases, most notably hematite formed in hydrothermal systems (Lippolt and Weigel 1988; Wernicke and Lippolt 1992; Lippolt et al. 1993; Wernicke and Lippolt 1994a,b).

Here I present an overview of recent techniques, calibrations, and applications of the (U-Th)/He dating method; Hurley (1954) provides an excellent summary of earlier work in this field. Much of this paper focuses on apatite, because the \(^{4}\)He behavior and requisite analytical techniques are better established for this phase than for other target minerals, such as zircon and titanite. Similarly, much of this paper concerns \(^{4}\)He diffusivity behavior required for thermochronometric applications, yet recent work is also considering applications to direct dating, for example, of young tephra (Farley et al. 2001).

Rationale

A wide variety of dating techniques are already well developed and widely applied. Given the methodological complexity, the limited analytical precision, and the rather small observational base presently available, it seems appropriate to consider why development and application of (U-Th)/He dating has received renewed attention. Two immediate answers come to mind. First, many materials are inappropriate for dating by existing techniques, most obviously because they are too poor in parent or radiogenic daughter isotope or too rich in non-in situ produced daughter. Thus a new technique with different requirements is potentially advantageous. This is indeed true for the (U-Th)/He
system: the sensitivity for measurement of U, Th, and He is extremely high and the background of "excess" He is low, so both young materials and those with only a trace of U and Th are potentially datable. As a result young volcanic rocks lacking sanidine for Ar/Ar dating are a potential target of the (U-Th)/He method (e.g., Graham et al. 1987; Farley et al. 2002). At most this is likely to be a niche market.

Greater interest in (U-Th)/He dating arises from the fact that He ages of various minerals can be used to delineate the cooling history of rocks through a temperature range that is only partially accessed by existing dating techniques. There is a large family of minerals potentially suitable for He thermochronometry, only a few of which have been explored. Figure 1 presents an overview of estimated He closure temperatures of several minerals (described in detail below) compared with closure temperatures of other thermochronometric methods. He dating complements existing techniques, and of particular interest is the apatite helium method, which is sensitive to temperatures substantially lower than any other method. Geologic applications which illustrate the unique uses of (U-Th)/He dating are presented at the end of this chapter.

TECHNICAL ASPECTS

He ingrowth

$^4$He nuclei ($\alpha$-particles) are produced by the series decay of $^{238}$U, $^{235}$U, and $^{232}$Th and by $\alpha$ decay of $^{147}$Sm. In essentially all minerals the overwhelming majority of radiogenic helium derives from actinide decay, so the ingrowth equation is:

$$^4\text{He} = 8 \times 10^6 \left( e^{1.89 \times 10^6 t} - 1 \right) + 7 \left( e^{1.89 \times 10^6 t/137.88} - 1 \right) + 6 \times 10^6 \left( e^{1.89 \times 10^6 t} - 1 \right)$$

Figure 1. Nominal closure temperatures of various thermochronometers showing how He systems (●) complement existing techniques (▲: Ar; △: fission track). Systems are simply ordered by closure temperature on the Y-axis. With the exception of apatite and titanite, the He closure temperatures are not yet well-known. Data sources: apatite He (Farley 2000); apatite fission track (Gallagher et al. 1998); hematite He (Wernicke et al. 1994); K feldspar multi-diffusion domain Ar (MOD) and other Ar methods (McDougall and Harrison 1999 and references therein); zircon fission track (Yamada et al. 1995); titanite fission track (Coyle and Wagner 1995), garnet He (Dunai and Roselieb 1996).
where \(^{4}\)He, U, and Th refer to present-day amounts, \(t\) is the accumulation time or He age, and \(\lambda\) is the decay constant (\(\lambda_{238} = 1.551 \times 10^{-10}\ \text{yr}^{-1}\), \(\lambda_{235} = 9.849 \times 10^{-10}\ \text{yr}^{-1}\), \(\lambda_{232} = 4.948 \times 10^{-11}\ \text{yr}^{-1}\)). The coefficients preceding the U and Th abundances account for the multiple \(\alpha\) particles emitted within each of the decay series, and the factor of \((1/137.88)\) is the present day \(^{235}\text{U}/^{238}\text{U}\) ratio. This equation assumes secular equilibrium among all daughters in the decay chain, a condition guaranteed for crystals formed more than \(-350\) kyr prior to the onset of He accumulation. For most applications this assumption is valid, but in certain cases the effects of secular disequilibrium must be considered (see below).

Equation (1) assumes the absence of initial \(^{4}\)He in the crystal being dated, and this is in general a good assumption. For example, while atmospheric Ar frequently accounts for a substantial fraction of the \(^{40}\text{Ar}\) in a K/Ar or Ar/Ar analysis (Kelley 2002, this volume), the concentration of He in the atmosphere is so low (5 ppm vs. \(-1\%\) for \(^{40}\text{Ar}\)) that trapped atmospheric He is unlikely to be important. In some cases fluid inclusions may carry crustal or mantle helium, but for U,Th-rich minerals like apatite, zircon, and titanite, the He concentration of such fluids and/or the inclusion density would have to be high to affect He ages except when the He ages are very low. Nevertheless some workers believe they have detected fluid-inclusion-hosted helium in a fewapatites (Lippolt et al. 1994; Stockli et al. 2000). The presence of helium “inherited” from some prior history, for example due to incomplete degassing of a crystal stowed into a magma chamber, is unlikely given the high diffusivity of He in most solids. However, beryl and other cyclosilicates sometimes harbor very high concentrations of “excess” helium from a poorly understood source (e.g., Toyoda and Ozima 1988). This helium probably enters the cyclosilicates through the large central channels of these minerals, making them unsuitable for He dating. A final potential source of excess He is solution from surrounding fluids into grain interiors (see Kelley 2002, this volume, for a discussion of the analogous problem in Ar geochronology). At present neither laboratory data (such as He solubilities in minerals at relevant temperatures) nor sufficient age data are available to quantitatively evaluate the significance of this phenomenon.

DIFFUSION BEHAVIOR

Knowledge of the He-retention characteristics of the phase being dated is critical for correct interpretation of (U-Th)/He data. For example, it is known that He is not retained under Earth-surface conditions in quartz (Trull et al. 1991), sanidine, and muscovite (Lippolt and Weigel 1988), so these phases have little obvious potential for any type of helium dating. However, helium is thought to be retained at the Earth’s surface in olivine (Trull et al. 1991), pyroxene (Lippolt and Weigel 1988), amphibole (Lippolt and Weigel 1988), garnet (Dunai and Roselieb 1996), non-metamict zircon (Hurley 1952; Damin and Kulp 1957), non-metamict titanite (Hurley 1952; Reiners and Farley 1999), apatite (Zeitler et al. 1987), allanite (Wolf 1997), magnetite (Fanale and Kulp 1962), hematite (Wernicke and Lippolt 1994a,b; Bahr et al. 1994) and submarine basaltic glass (Graham et al. 1987). Several of these phases, especially olivine and pyroxene, have been used extensively for cosmogenic \(^{3}\)He studies, so He retention is robustly established (Niedermann 2002, this volume). In others the conditions for retention (temperature, grain size, degree of radiation damage) have not been explored.

While in the case of rapidly cooled rocks at the Earth’s surface (e.g., volcanics) demonstration of quantitative retention at \(-25^\circ\text{C}\) is sufficient to successfully apply He dating, thermochronometry of slowly cooled rocks requires precise knowledge of how diffusivity scales with temperature. Typically, laboratory experiments are used to constrain the parameters of the Arrhenius relationship:
\[
\frac{D}{a^2} = \frac{D_0}{a^2} e^{-E_a/RT}
\]

where \(D\) is the diffusivity, \(D_0\) the diffusivity at infinite temperature, \(E_a\) the activation energy, \(R\) the gas constant, \(T\) the Kelvin temperature, and \(a\) the diffusion domain radius (Fechtig and Kalbitzer 1966). If this relationship is obeyed, then measurements of ln\(D/a^2\) as a function of reciprocal temperature will plot on a straight line with intercept ln\(D_0/a^2\) and slope \(-E_a/R\). If the measurements do not plot on a straight line, more complex behavior such as multiple diffusion mechanisms or domain sizes may be involved.

It is important to note that such laboratory measurements may not apply under natural conditions. For example, diffusion coefficients are commonly measured at temperatures far higher than are relevant in nature, so large and potentially inaccurate extrapolations are often necessary. Similarly, some minerals undergo chemical or structural transformations and possibly defect annealing during vacuum heating; extrapolation of laboratory data from these modified phases to natural conditions may lead to erroneous predictions.

Substantial effort is required to measure the He diffusivity parameters in a given phase and to determine how those parameters vary with mineral characteristics such as grain size and shape, chemical composition, and defect and/or radiation damage density. Only apatite has been studied in detail, but limited He diffusivity information is available on other minerals as well, as discussed below.

**Apatite**

Zeitler et al. (1987) initiated interest in He thermochronometry by demonstrating that apatite has a closure temperature of about 100°C (here and elsewhere closure temperatures are referenced to a cooling rate of 10°C/Myr; Dodson 1973; Kelley 2002, this volume). More recent efforts (Lippolt et al. 1994; Wolf et al. 1996b; Warnock et al. 1997; Farley 2000) confirm this approximate closure temperature, and in addition suggest:

1) He diffusion from Durango apatite (Young et al. 1969, a common “standard” apatite), as well as a variety of other apatites, obeys an Arrhenius relationship (Eqn. 2; illustrated in Fig. 2), suggesting that He diffusion from apatite is a single-mechanism thermally activated volume diffusion process, at least at temperatures <300°C. Above 300°C the Arrhenius plot is curved. The origin and significance of this curvature has been explored, but remains poorly understood; it is probably not relevant for helium diffusion in the natural setting.

2) High-precision experiments indicate activation energies between ~32 and ~38 kcal/mol. It is not clear whether this spread is real or analytical in origin; there is no persuasive evidence that this quantity is correlated with observable apatite characteristics.

3) In Durango apatite, the quantity D/a^2 varies with grain size in the manner expected if the diffusion domain is the grain itself, i.e., the quantity “a” is the physical grain dimension. He diffusion from Durango apatite is crystallographically isotropic. The relevant dimension for diffusion is thus the prism radius, as this is the shortest pathway for He loss. These characteristics have not yet been demonstrated to apply to more typical apatites, though Reiners and Farley (2001b) present He age data that support such an effect.

4) Taken together the most precise observations suggest an apatite closure temperature of 70°C in apatites of ~70-90 μm radius. Variation of the closure temperature with grain size and cooling rate based on the Durango observations is shown in Figure 3.
Figure 2. Helium diffusion Arrhenius plot with Otway Basin borehole constraint. The laboratory-measured He diffusion from this sample of Durango apatite defines an extremely linear array (filled symbols), consistent with simple thermally activated volume-diffusion. He age measurements from the Otway Basin provide a completely independent estimate of He diffusivity (open symbol), under natural conditions, which is in excellent agreement with the extrapolated laboratory data. Laboratory data are from Farley (2000). The Otway Basin constraint is from House et al. (1999).

Figure 3. Helium closure temperature ($T_c$) as a function of grain size and cooling rate. $T_c$ was calculated assuming an activation energy of 33 kcal/mol and $D_0 = 50$ cm$^2$/sec, assuming spherical geometry and including the effects of $\alpha$ ejection on He diffusion (see Farley (2000) for details and justification of geometry). The shaded region indicates ranges typically observed in nature. This figure differs slightly from the one presented by Farley (2000) because slightly different diffusion parameters were adopted.
Several studies have attempted to verify the expected diffusivity behavior in the natural setting. The most obvious method for verification is to examine the He age distribution in boreholes in which temperature is known as a function of depth. In such a setting He ages are expected to decrease rapidly downhole, defining the Helium Partial Retention Zone (HePRZ; Fig. 4). Wolf et al. (1998) discussed the shape of the HePRZ in detail. Consider the evolution of apatite He ages in a block of crust instantaneously created and subjected to a constant crustal geothermal gradient. At the Earth’s surface and at depths corresponding to temperatures below 40°C, helium is quantitatively retained and He ages track calendar time. In contrast, at temperatures above ~80°C, He is lost as rapidly as it is produced, so He ages remain essentially zero. In the region between these two limits He ages are extraordinarily temperature-sensitive, with variations of millions of years possible over very small depth increments. In this region He ages will initially increase with calendar time because ingrowth exceeds diffusive loss, but the age increase will be less than the increase in calendar time. However as the He concentration in the crystal rises, so too does the concentration gradient that drives diffusion. After a sufficiently long period a steady state He age will be achieved wherein radiogenic production and diffusive loss precisely balance. This steady state will be achieved at higher temperatures first, so while ages at depth become invariant, those at shallow depths continue to increase. Thus the precise shape of the HePRZ evolves with time. Figure 4 shows an apatite HePRZ developed over 100 Myr of holding time computed from laboratory diffusivity data. This profile is compared with the essentially analogous fission track partial annealing zone in the figure.

Efforts by Wolf (1997), Warnock et al. (1997) and House et al. (1999) to confirm the existence of the HePRZ were broadly successful. In three different borehole settings the apatite helium ages were found to decrease rapidly at about the predicted temperature, but problems arising from mineral inclusions, poorly known thermal histories, and other phenomena prevented a quantitative confirmation of the diffusivities extrapolated from

![Figure 4. The helium partial retention zones (HePRZ) for apatite and for titanite, and the apatite fission track partial annealing zone (FTPAZ), calculated assuming 100 Myr of isothermal holding at the temperatures indicated on the Y-axis. The right-hand axis scales temperature into sub-crustal depth assuming a geothermal gradient of 20°C/km and a mean surface temperature of 10°C. The apatite helium curve was calculated using parameters in the caption to Figure 3, for titanite, an activation energy of 44.6 kcal/mol, D0 = 50 cm/sec, and r = 250 μm were assumed (Reiners and Farley 1999). The fission track profile was calculated using AFTSolve (Ketcham and Donellick 2000) assuming kinetic parameters appropriate for Durango apatite.](image)
laboratory measurements. These data leave little doubt that the laboratory data are approximately correct, but do not constitute unassailable evidence.

House et al. (1999) recognized that problems associated with thermal history could be circumvented by taking advantage of the fact that He ages of samples held close to the closure temperature will rapidly achieve the steady state age where He production and diffusive loss are in balance (Wolf et al. 1998). Provided a sample is in steady state, the measured age and downhole temperature provide a diffusivity–temperature pair that is completely independent of the laboratory measurements. Samples thought to be in steady state in boreholes from the Otway Basin, Australia, yielded diffusivities in excellent agreement with the extrapolated laboratory data (Fig. 2). Unfortunately the downhole temperatures from these industry wells are not very well known, so some uncertainty remains.

Using a completely different approach Stockli et al. (2000) also confirmed the extrapolation of laboratory data. In the White Mountains of California, a rapidly exhumed crustal block, these authors discovered a very well defined HePRZ lying above an equally well-defined apatite FTPAZ. The FTPAZ in conjunction with an unconformity defining the pre-exhumation Earth surface was used to compute the pre-exhumation geothermal gradient. The HePRZ predicted in such a geothermal gradient using extrapolated laboratory data matched observations extremely well. This study is described more fully in the section on applications.

These studies provide compelling evidence that laboratory data adequately describe He diffusion characteristics for most apatites so far investigated. Nevertheless additional work to establish the generality of these data and the possible effects of chemical substitution and defects (Farley 2000) on He diffusivity is warranted. The strong effects of even modest substitution of C1 for F on apatite fission track annealing characteristics (Green et al. 1986) demonstrate that extreme caution is necessary when attempting to generalize kinetic behavior from a specific and limited set of observations.

Hematite

Lippolt and coworkers have undertaken several studies of the He retention characteristics of hematite. In the case of specular hematite, reasonably linear He diffusion Arrhenius plots obtained at temperatures <1000°C were interpreted as evidence for volume diffusion (Bahr et al. 1994; Lippolt et al. 1993). He diffusivity scales approximately as expected if the diffusion domain is the grain itself, and a closure temperature of ~220°C was estimated for grains of radius ~500 μm. He diffusion from botryoidal hematite plots on linear Arrhenius arrays only at temperatures below 250°C, and is apparently controlled by volume diffusion from small (0.1 to 10 μm) crystals within the larger botryoidal mass (Bahr et al. 1994; Wernicke and Lippolt 1994a). A closure temperature of 120°C was estimated from the linear part of the Arrhenius plot for botryoidal hematite grains larger than ~10 μm. This value is only a general estimate; for example, a 50°C difference in closure temperature was inferred for two zones of a single botryoidal hematite specimen (Wernicke and Lippolt 1994a). Bahr et al. (1994) attributed deviations from linearity on the Arrhenius plots of the two hematite varieties to variations in the grain size distribution of the samples. The role of grain size and chemistry on He diffusion from hematite warrants further investigation.

Titanite

Based on laboratory stepped-heating experiments, Reiners and Farley (1999) concluded that He diffusion from titanite is generally consistent with a thermally
activated volume diffusion process, although some subtleties in the diffusion characteristics remain poorly understood. An activation energy of about ~45 kcal/mol was established, and, like apatite and specular hematite, the diffusion domain was found to be the grain itself. Based on these observations, a closure temperature of ~200°C was estimated for grains of ~500 μm diameter. Insufficient data are available to evaluate possible compositional controls on He diffusion from this phase. The titanite HePRZ computed from these results is expected to lie between 150 and 200°C, substantially deeper than both the FTPAZ and the apatite HePRZ (Fig. 4).

The effects of radiation damage on He loss from radioactive minerals like zircon and titanite have long been recognized. For example, Hurley (1952) showed that He ages decrease with increasing U content in a zirrocene suite of Precambrian titanites. Hurley interpreted this observation to indicate radiation-damage induced He loss. Although he presented a method for correcting ages for this loss, it is unclear how such a correction can be applied to samples with time-varying temperature histories in which both radiation damage (and annealing) and He diffusivity vary. Based on fission track studies, radiation damage in titanite anneals over geologic timescales at temperatures above ~300°C (Choye and Wagner 1998), so the relevant measure of radiation damage is the α activity coupled with the time elapsed since cooling through this temperature. Using Hurley’s model such He loss should not strongly affect He ages until a few hundred Myr after a titanite cools below ~300°C, at least in titanites with equivalent U contents of less than 1000 ppm. This conclusion is supported by the observation that He ages of the quickly-cooled Mount Dromedary (99 Ma) and Fish Canyon Tuff (28 Ma) titanite standards indicate quantitative He retention (Reiners and Farley 1999; House et al. 2000).

These observations provide no insight to how He diffusion and closure temperature vary with radiation damage. Reiners and Farley (1999) investigated this phenomenon but saw no obvious indication of enhanced He diffusivity, not even in titanites which likely cooled below 300°C several hundred million years ago. However, in step-heating analyses of titanites with large He accumulations and dark colors (both reasonable proxies of radiation damage) unexpectedly high and erratic He release was observed in some steps, presumably related to radiation damage. A mechanistic explanation for this behavior has not yet been developed, nor is it obvious how this behavior affects He retention in the natural setting. Quantitative application of titanite thermochronometry will require careful assessment of this phenomenon.

To test the measured diffusivity data in the natural setting and to assess the role of radiation damage, Reiners et al. (2000) analyzed titanite He ages in the Proterozoic Gold Butte Block of Southern Nevada, a rapidly and very deeply exhumed crustal block. He ages define a well developed HePRZ positioned approximately at the structural position expected based on extrapolation of the laboratory data. At least for the degree of radiation damage experienced in these samples, He diffusivity is not so severely compromised as to preclude meaningful thermochronometry.

**Zircon**

Limited data suggest that He diffusion from zircon does not obey a simple Arrhenius relationship (Reiners and Farley 2001a). Instead, in most analyzed samples a distinct increase in slope and decrease in diffusivity occurs as the experiment proceeds. This observation is consistent with annealing of radiation damage or with multiple diffusion domains of different sizes, possibly induced by radiation damage. In addition, the most heavily radiation damaged zircons yield extremely erratic Arrhenius profiles. Although these data are difficult to interpret, Reiners and Farley (2000) proposed a minimum closure temperature of about 180°C for He release from zircon, with lower values arising
from radiation damage effects. He age-paleodepth relationships in the Gold Butte Block (see titanite section) also indicate a He closure temperature in zircon of about 200°C. Overall these data are not sufficient to reliably establish the He diffusivity of zircon.

Like titanite, radiation damage is known to promote He loss from zircon. For example, Damon and Kulp (1957) showed that near-quantitative He retention occurs in Sri Lankan zircons up to a total irradiation of about $6 \times 10^{15}$ $\alpha$ s/mg (equivalent to 1000 ppm$\alpha$ of U decaying for ~1 Gyr), but drops dramatically at higher dosages. Based on fission track annealing studies, radiation damage accumulates in zircons only at temperatures below ~250°C (Yamada et al. 1995), so He dating of typical zircons that have been cooler than this temperature for up to a few hundred Myr should not be greatly compromised by radiation damage. However as noted by Hurley (1954), strong zonation in U and Th in zircons may cause some areas to be severely damaged before the entire crystal becomes metamict; even local metamictization is likely to perturb He diffusion.

**Garnet**

Dunai and Roselieb (1996) presented the results of He solution experiments suggesting a very high closure temperature for garnet, ~600°C. No data are available to assess the role of composition on He diffusion, but given the diversity of substitutions possible in garnet some compositional effects are likely. In many cases the U and Th in garnet derives from inclusions of radioactive minerals, and this is a potential pitfall for garnet thermochronometry.

**The $\alpha$-emission correction**

For any application of (U-\text{Th})/He dating, whether as an absolute chronometer or for thermochronometry, and regardless of the material being analyzed, careful consideration of the consequences of the MeV energies of $\alpha$ decay is critical. Energetic decay and associated long $\alpha$-stopping distances greatly limit the materials suitable for He dating and present perhaps the greatest impediment to determination of ages with the sub-percent accuracy and precision of which isotopic dating techniques are commonly capable. Given its unique and as yet inadequately appreciated significance for He dating, this section elaborates the $\alpha$-stopping phenomenon and considers ways to accommodate and/or correct for it. $\alpha$ particles of the U and Th series are sufficiently energetic that they travel about 20 μm through solid matter before coming to rest. As a result, $\alpha$ decay induces a spatial separation between parent and daughter nuclei. To the extent that the parent abundance within a rock is not uniform, e.g., because distinct crystals are present, spatial variability in daughter/parent ratio will result. This unavoidably leads to the erroneous appearance of He age heterogeneity within the rock, with regions or crystals that experience a net import of $\alpha$ particles being “too old” and regions with a net export being “too young.” The effect can be substantial in small crystals and will lead to erroneous “ages” if not accommodated in some fashion.

Each $\alpha$ decay within the U and Th series has a characteristic energy and hence a characteristic and well-known (Ziegler 1977) stopping distance within a given material. As a result, an $\alpha$ particle will come to rest on the surface of a sphere centered on the site of the parent nucleus and with a radius equivalent to the stopping distance. As shown in Figure 5, there are three relevant outcomes of $\alpha$ decay in a crystal. If the parent nucleus is located more than the stopping distance away from the edge of the crystal, then that $\alpha$ particle will be retained within the crystal regardless of the trajectory the particle takes. For a parent nucleus lying within one stopping distance of the crystal boundary there is some probability that the $\alpha$ particle will be ejected. The ejection probability rises to a
maximum of about 50% on the grain edge, e.g., if the edge of the grain is considered a flat surface, then in a statistical sense half of the trajectories taken by the $\alpha$ particle will lead to ejection, and half to retention. The increase in ejection with proximity to grain edge is shown schematically in the lower part of Figure 5. It is also important to consider that decay occurring outside of the crystal can lead to implantation into the crystal of interest.

![Figure 5](image)

**Figure 5.** The effects of long $\alpha$-stopping distances on He retention. The upper figure illustrates the three relevant possibilities within a schematic crystal: $\alpha$ retention, possible $\alpha$ ejection, and possible $\alpha$ implantation. The center of the circle denotes the site of the parent U or Th nuclide, and the edge of the white circle labeled He indicates the locus of points where the $\alpha$ particle may come to rest; the arrow indicates one possible trajectory. The lower plot shows schematically how $\alpha$ retention changes from rim to core to rim along the path A-A'; exact equations defining the shape of this curve as a function of grain size were given by Farley et al. (1996).

Because this phenomenon is restricted to the outermost ~20 $\mu$m of a crystal, chemical or mechanical removal of the grain surface will eliminate the effect. However, as discussed above, the He diffusion domain in some minerals is the grain itself (Bahr et al. 1994; Reiners and Farley 1999; Farley 2000). Because the grain edge is the site of diffusional helium loss, it will tend to have a lower He concentration than the grain interior as a result of diffusive transport. Removal of the outermost portion will bias the age of the remaining crystal toward erroneously high values. As discussed by Reiners and Farley (1999), the magnitude of the effect is a sensitive function of the crystal’s thermal history. For some applications, such as dating of quickly cooled minerals (e.g., from tephra), or large crystals in which the diffusion gradient is on a longer length-scale than $\alpha$-ejection/implantation, this approach may be appropriate, but in general erroneous ages will result from surface removal when the diffusion and $\alpha$-ejection boundaries coincide.

As an alternative, Farley et al. (1996) developed a quantitative model for correcting He ages for the effects of long $\alpha$ stopping distances based on measured grain geometry and size. Several assumptions are required:

1) Implantation from the surrounding matrix is insignificant, so only $\alpha$ ejection need be
considered. In most minerals of interest for He dating, the parent nuclide concentration contrast between the mineral being dated and the host rock is so large that implantation is trivial compared to in situ produced He. In some cases, such as very U-Th poor apatites, this assumption may be violated.

2) The distribution of U and Th in the crystal being dated must be specified. Because the distribution will in general not be known, Farley et al. (1996) presented results assuming a homogeneous distribution and discussed the error introduced by various types of zonation.

In accord with intuition, the model shows that the two most important variables controlling the total fraction of alphas retained in a crystal (the “F_r” parameter, that is, the factor by which the measured age must be divided to obtain the “α-ejection-corrected” age) are the surface to volume ratio (β) of the crystal, and the α stopping distance. Crystals with small β have less “skin” affected by α ejection, and hence require smaller corrections. While each decay in the U and Th chains has a characteristic stopping distance, the mean F_r obtained by modeling each decay separately does not differ substantially from simply using a single mean stopping distance for each parent. However, because stopping distances vary significantly with the density and to a lesser extent the composition of the stopping medium, it is necessary to use different stopping distances for different minerals. Analytical and Monte Carlo results were presented that allow computation of F_r from measured dimensions for several simple grain geometries including a sphere, a cylinder, and a cube (Farley et al. 1996).

Table 1 expands on the work of Farley et al. (1996) by presenting a second-order polynomial relating F_r to β for both U and Th series decay for two systems of particular interest: a hexagonal prism of apatite density, and a tetragonal prism of zircon density. In both cases the prism terminations were ignored. The coefficients were obtained by Monte Carlo modeling as described previously (Farley et al. 1996).

A typical result of this modeling is shown in Figure 6 in which F_r is plotted as a function of prism width (defined here to be the distance between opposed apices) for an apatite hexagonal prism of length/width ratio of 3. For relatively large grain widths, F_r values are fairly constant, in the range 0.8 to 0.9, decreasing only slowly with decreasing width. However the curve becomes increasingly steep for widths less than 80 μm. The message from this plot is that in general the largest grains will have the least uncertainty on the correction, and that typical corrections for small accessory minerals will be in the range 0.65 (or even less) to 0.9.

Figure 6 and Table 1 show that α retentivity is slightly higher for 238U than for 232Th, reflecting the higher mean energy of α decays in the 232Th series. This distinction is relatively subtle but can be accommodated by computing a weighted mean of the F_r values for U and Th, where the weighting factor is the fraction of He derived from each parent. Specifically,

\[ \text{Mean F}_r = a_{235} \text{F}_{235U} + (1 - a_{235}) \text{F}_{232Th} \]

Because 235U and 232Th have very similar decay energies, the above expression associates the He from these two parents with a single F_r value. The fraction of He derived from 238U (a_{238}) can be calculated exactly from Equation (1), or approximated from the measured Th/U ratio for integration periods of less than ~200 Myr as:
Figure 6. The effects of α-ejection on He retention in an apatite hexagonal prism. $F_T$ is the total fraction of alphas retained within the crystal, assumed here to have a length/width ratio of 3. The $^{238}$U and $^{232}$Th series lie on slightly different curves because of differences in decay energy. $^{238}$U would plot on top of the $^{232}$Th curve. The curves were calculated using the equation and coefficients in Table 1.

$$A_{238} = (1.04 + 0.245 \cdot (Th/U))^4$$  \hspace{1cm} (4)

This model has been implemented in many different studies using the following procedure:

1) Grains to be analyzed are selected on the basis of good crystal morphology. In many cases it is necessary to search for grains that match the ideal geometry as closely as possible; for example, we have found that in some rocks a substantial fraction of apatites have a tabular habit that is not well-approximated by a hexagonal prism model. Grains to be analyzed together as a single aliquot are selected to be of similar size.

2) Grains are measured using a reticle in a binocular microscope. In the case of apatite, the prism diameter and length are measured.

<table>
<thead>
<tr>
<th>Geometry-Mineral</th>
<th>$a_1$</th>
<th>$a_2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Apatite hexagonal prism</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$^{238}$U series</td>
<td>-5.13</td>
<td>6.78</td>
</tr>
<tr>
<td>$^{232}$Th series</td>
<td>-5.90</td>
<td>8.99</td>
</tr>
<tr>
<td>Zircon tetragonal prism</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$^{238}$U series</td>
<td>-4.31</td>
<td>4.92</td>
</tr>
<tr>
<td>$^{232}$Th series</td>
<td>-5.00</td>
<td>6.80</td>
</tr>
</tbody>
</table>

$F_T = 1 + a_0 \beta + a_1 \beta^2$. $\beta$ is the surface to volume ratio; for a hexagonal prism given by $\beta = 2.31L + 2R/(RL)$ where $R$ is half the distance between the opposed apices and $L$ is the length. For a tetragonal prism $\beta = (4L + 2W)/(LW)$ where $L$ is the length and $W$ is the width of the prism. $^{238}$U parameters are essentially identical to $^{232}$Th.
3) An $F_T$ value is computed for the grain based on the grain's dimensions and geometry and the $\alpha$ ejection model described above. In the case of apatite, many grains break perpendicular to the c-axis during sample preparation, such that measured lengths are shorter than existed in the rock. As demonstrated by Farley et al. (1996), $F_T$ is not very sensitive to the original length; at Caltech it is standard practice to multiply the observed length of all grains by a factor of 1.5 to account for breakage unless the grains are demonstrably unbroken.

4) The mean $F_T$ of the entire population of grains in the aliquot is computed, weighting each grain by its mass contribution to the aliquot. The weighting is based on observed grain dimensions. This weighting implicitly assumes that grains contribute helium in proportion to their mass. If grains of very different sizes have very different U or Th contents, this weighting will be incorrect. Hence the effort in step 1 to pick grains of a common size. Indeed we observe that many apatite and zircon samples show large grain-to-grain heterogeneity in U and Th concentration, presumably reflecting the compositional evolution of magmas during crystallization. Thus this precaution is important.

This technique works well for minerals that retain their characteristic size and morphology when separated from the host rock, as both apatite and zircon commonly do. It is less successful for minerals that shatter, leaving no indication of original grain morphology or dimensions. This problem is particularly acute for titanite, which in our experience is very commonly broken during sample preparation.

Experiments at Caltech show that $F_T$ corrections in the range of $\sim$0.65 to $\sim$0.85 can be reproduced by individual observers to better than a few percent. Systematic observer bias in $F_T$ determinations has not been observed. At smaller grain sizes, where the variation in $F_T$ with grain size becomes increasingly steep (Fig. 6), errors in $F_T$ will become increasingly large. We avoid analyzing grains with $F_T < 0.65$.

The biggest problem with this approach to correcting $\alpha$ ejection effects is the assumption of a uniform parent nuclide distribution; U- and Th-zoned crystals violate this assumption. It is useful to consider how large an error zonation can induce in the final computed He age. Consider an apatite hexagonal prism of typical dimensions yielding an $F_T$ value of 0.75. If instead of a homogenous distribution, 100% of the U and Th is located more than one stopping distance from the grain boundary, then the true fraction of alphas retained would be unity, and the model $F_T$-corrected age would be 33% greater than the true age. Alternatively, if all of the parent is located on the prism rim, then the fraction of alphas retained would by $\sim$0.5. In this case the model $F_T$-corrected He age would be $\sim$33% younger than the true age. For all other degrees and styles of zonation, the error introduced will be between these two bounds, so the maximum error is $\pm$33%. Farley et al. (1996) considered several different scenarios and concluded that only extreme zonation will produce large errors in $F_T$ correction. For example, an 11-fold linear decrease in parent abundance from core to rim yielded an $F_T$ value differing from the homogeneous case by just 10%.

The degree to which U and Th are zoned in phases relevant for He dating has not been extensively explored, but in zircons and to a lesser extent apatites zonation is generally thought to be fairly common. Of greatest concern are situations in which the parent concentration of the outermost 20 $\mu$m is dramatically different from the remainder of the grain; oscillatory zonation is no more problematic than monotonic zonation that affects the surface in a comparable fashion. In the case of apatite, rare earth elements continue to diffuse even at fairly low temperatures (Watson et al. 1985). If this observation applies to U and Th as well, then slowly cooled apatites may in general have fairly uniform parent distributions. Further documentation of the magnitude and style of
zonation in phases of interest such as apatite, titanite and zircon would be useful for evaluating how large an uncertainty to place on the ejection-correction component of He age determinations.

A final point to consider is that because $\alpha$ ejection and diffusion occur at the same boundary—at least in specular hematite (Bahr et al. 1994), apatite (Farley 2000) and titanite (Rainers and Farley 1999), the He gradient that drives diffusion will be more rounded than diffusion alone would produce. For example, in a quickly cooled sample that has experienced $\alpha$ ejection, the He concentration profile would look similar to Figure 2. In contrast, without $\alpha$ ejection, there would be a step function right at the grain boundary. As a result, diffusive transport will be slightly reduced; for most cooling scenarios the effect is equivalent to just a few degree increase in closure temperature (Farley 2000). Note however that during laboratory diffusion experiments, this rounding may significantly bias estimates of diffusion coefficients made when the first few percent of the helium is released (Farley 2000).

**Analytical procedures, accuracy, precision and mineral standards**

Once grains have been selected for analysis and measured for $\alpha$ ejection correction, a (U-Th)/He age determination involves $^4$He extraction and measurement, and U and Th measurement on either the same or a separate aliquot. Analytical techniques, especially for the small amounts of $^4$He obtained from most samples, have evolved rapidly in the last five years, and so are described in some detail here.

**Helium extraction.** Early work by us and others involved the analysis of separate aliquots for He and for U and Th. While this approach permits straightforward vacuum fusion of the grains being analyzed for helium, it has the serious drawback that aliquots must be large enough to average out any grain-to-grain variability in U-Th-He content. Without a detailed study of a given sample it is hard to estimate how large an aliquot will ensure homogeneity but aliquots of at least a few mg are likely required. Preparation of such large aliquots can be very tedious, hence we have developed several techniques for analyzing parent and daughter on the same aliquot. The single-aliquot technique has the additional advantage that the aliquot need never be weighed: the age is computed directly from the relative parent and daughter abundances using Equation (1).

Initially we loaded grains for analysis into stainless steel capsules that could be retrieved from the furnace after outgassing of He at temperatures lower than the melting point of the mineral being dated. This technique, described in detail in the Appendix, is particularly well suited to apatite, which can be quantitatively outgassed of $^4$He in 20 min at 950°C. However the technique is awkward or unsuccessful when high temperatures (above the melting point of the steel capsule) are required for complete He extraction, e.g., from zircon and titanite. Blank levels for the furnace technique average about 0.6 fmol of $^4$He, sufficiently low for the analysis of aggregates of about ten crystals in typical apatite samples as young as a few million years. However this blank is too high to permit routine single crystal dating of apatite unless the crystal is unusually large or has a He age of more than a few tens of million years.

Laser extraction is a more elegant alternative that has recently been adapted for He dating (House et al. 2000). Even the most refractory minerals can be heated to the melting point using a laser of appropriate wavelength. In addition, laser-based methods have far lower He blanks compared with resistance heating. Unfortunately laser heating of bare mineral grains is known to cause volatilization and loss of U, and in some cases Th, from the fused sample. This presumably occurs when the most intensely heated part of the sample melts, causing it to couple even better with the laser. Local thermal runaway and boiling results. Loss of U or Th precludes analyzing parent and daughter on the same
aliquot. For example, Reiners and Farley (1999) found that, even at very low power from a Nd-YAG laser, fusion of bare titanite grains to a glassy bead was associated with substantial U loss. The U is presumably vaporized from the sample and deposited somewhere in the vacuum chamber. A similar result was obtained by Stuart and Persano (1999) upon CO₂ laser heating of bare apatite.

House et al. (2000) overcame this problem by wrapping the sample in a very small (1 mm × 1 mm) Pt foil envelope which is heated by a Nd-YAG laser. The Pt is a sufficiently good heat conductor that uniform heating results, as indicated by a uniform incandescence. Using an optical pyrometer the temperature of the sample can be estimated and controlled with a feedback loop to the laser output. As expected, the laser blanks are far better than those on the resistance furnace; holding an empty Pt-foil envelope at 1350°C for 30 min yields about 0.02 fmol of He. Because the technique is far faster (many samples can be loaded in the laser vacuum chamber and pumped simultaneously) and has a lower blank, we now routinely outgasapatites, zircons, and titanites with this technique. The blank is sufficiently low that single crystal dating of nearly all apatite samples is possible. Except for the method by which the sample is heated, the laser procedure follows the resistance heating procedure (see the Appendix).

U and Th are analyzed by isotope dilution inductively-coupled plasma mass spectrometry performed on the same aliquot analyzed for He (Appendix).

Accuracy and precision. Based on reproducibility of pure standard gases and aqueous standard solutions, the overall analytical precision of He ages determined by this procedure should be about 2% (2σ, excluding errors in α-ejection correction) when ages are well above blank levels. Most of this uncertainty arises from the He measurement. In actual practice, we obtain He ages that reproduce to about 6% (2σ), demonstrating some natural variability within grain populations.

Accuracy of He ages is as important as precision for most geologic interpretations. Obtaining accurate U and Th abundances is a routine procedure in, e.g., U-Pb geochronology laboratories. Standard solutions accurate to better than a few per mil can be obtained by dilution of commercial ICPMS standards, which routinely carry a stated accuracy of 1 per mil. Obtaining precise and accurate He aliquots is more challenging, requiring knowledge of both pipette volume and the partial pressure of He in the standard tank. At Caltech the volume of the standard He pipette was determined by manometric comparison to a primary calibrated volume (determined by weighing after filling with water or Hg). He pressure was determined by capacitance manometry corrected for thermal transpiration effects. The accuracy of a He aliquot calibrated in this fashion is probably ~1%.

The He dating method as described above is an absolute dating technique, based on fundamental measured quantities rather than by reference to independently dated mineral standards. In this regard it is different from other low temperature thermochronometers like Ar/Ar and fission track counting. Nevertheless it is useful to have mineral standards to verify both analytical calibrations and the degree to which the system faithfully acts as a chronometer. Unfortunately identifying appropriate standards is not easy, particularly for a low temperature system like He-in-apatite. The key requirement is that the mineral of interest cooled through both He closure and closure of the independent chronometer at the same time. Like apatite fission track standards, appropriate standards for He dating include rapidly cooled volcanic and very shallow plutonic samples. However at least in the case of apatite, He ages are very susceptible to post-eruption diffusive He loss that may be problematic. For example, consider the case of an apatite of 55 µm radius, erupted 30 Ma and immediately buried under 750 m of overburden. Given a mean annual temperature of 15°C and a geothermal gradient of 27°C/km, this sample will sit at 35°C.
Modeling based on He diffusivity data shown in Figure 2 indicates that such a sample would yield a He age of just 27.6 Ma, almost 10% lower than the "expected age" obtained from other techniques such as Ar/Ar and fission track counting. If such a sample were recently exposed and used as a He age standard, problems would clearly result. Thus standards for He dating, at least for apatite, must be carefully selected. No consensus yet exists on the most appropriate specimens for such standards.

He ages are available for only a few "standards." At Caltech the mean of several dozen age determinations on Durango apatite is 32.0 Ma, with a standard deviation of the population of 1 Ma (Farley 2000; House et al. 2000 and Farley, unpublished). This age is in excellent agreement with the accepted age of this apatite (31.6±1 Ma) (Jonckheere et al. 1993). Wramack et al. (1997) reported a slightly younger He age (27.5 Ma) for this apatite, while Wolf et al. (1996b) reported a slightly older value. Note that Durango apatite crystals are so large (cm) that the ejection-affected rim can be excluded from analysis, and furthermore that the closure temperature of these large grains means that even substantial post-eruptive burial is not an issue. At present Durango apatite is perhaps the most appropriate age standard, although its unusual size and chemistry (e.g., high Th/U ratio) make it atypical of most unknowns to be dated. Reiners and Farley (1999) reported a titanite He age for Fish Canyon Tuff of 30.1±2 Ma, while House et al. (2000) reported a value of 27.9±1 Ma for aliquots of the same material. Unpublished work at Caltech reveals significantly lower He ages and large scatter from Fish Canyon apatite, possibly because of burial. Reiners and Farley (2001a) reported a zircon He age of 28.1±2.8 for Fish Canyon Tuff. These values are similar to the accepted age of 28.5±0.1 for the Fish Canyon Tuff (Schmitz and Bowring 2001).

**Mineral inclusions.** In our experience at Caltech the single biggest difficulty in He dating of apatite is the presence of small U-Th-rich inclusions within the dated grains. This difficulty was first noted by Lippolt et al. (1994) and was further described by House et al. (1997). The most common inclusions are zircon and monazite, but we have occasionally encountered xenotime and allanite. We have also observed inclusions of quartz, feldspar, pyrite and graphite(?), but these are unlikely to carry sufficient U and Th to be a problem.

Inclusions cause several different problems for (U-Th)/He dating. Although some effect might be expected from differing He closure temperatures of host and inclusion, for inclusions smaller than ~15 \(\mu\)m diameter, essentially all He will be ejected into the host mineral, so only the He diffusion characteristics of the host are important (Farley et al. 1996). However, localization of U and Th may bias the ejection effect and may also modify the diffusion behavior by changing the He concentration gradient. More importantly, many inclusion phases, especially zircon, survive the dissolution technique we use on apatite (see the Appendix). Hence the inclusions will contribute He to the analysis, but not U and Th. As a result, some He will be "parentless," and anomalously old ages will result. These anomalously old ages will tend to be irreproducible because the inclusions are not in equal abundance in the analyzed aliquots.

In many cases inclusions in apatite can be detected during the grain selection process. At Caltech apatites to be dated are inspected at ~120× under a binocular microscope using transmitted light and crossed-polarizers. When the apatites are taken to extinction even tiny inclusions of phases like zircon stand out—in some cases entire grains appear to be shot full of inclusions. These grains are easily removed prior to analysis. In rare cases this technique has been found inadequate, usually because the inclusions are oriented parallel to the c-axis (possibly from exsolution of monazite) and are extinct at the same time as the apatite host. In these cases the re-extract test (see the Appendix) and age irreproducibility are sufficient to identify problem samples.
Unlike apatite, microscopic detection of inclusions is very difficult in dark or opaque minerals such as hematite or hornblende, and in minerals with high indices of refraction such as zircon and titanite. Insufficient evidence exists to evaluate the prevalence of problem inclusions in such phases.

**INTERPRETATION OF HE AGES AND EXAMPLES**

**He cooling ages**

The simplest way to interpret He cooling ages is to associate them with the time of cooling of the sample through a specific closure temperature, as defined by Dodson (1973). However, this approach is restricted to t-T paths involving monotonic cooling. A more robust way to interpret He ages has been described by Wolf et al. (1998), who presented a numerical forward model which yields a He age given an arbitrary t-T path and measured diffusivity parameters. This is analogous to computational models for interpreting Ar cooling ages (Kelley 2002, this volume).

A wide variety of t-T paths can yield a given He age, so a unique interpretation of a single He age is not generally possible without additional information. Age-elevation profiles can be used to more tightly constrain cooling histories (Wolf et al. 1998), as can comparison with other techniques, such as fission track methods (Stockli et al. 2000). For the latter to be successful, further efforts must be made to ensure compatibility of the thermal calibrations of the dating techniques. For example, a recent study suggests that cooling histories derived from apatite fission track-length models (Gallagher 1995) are inconsistent with apatite He ages (House et al. 1999). Such a discrepancy is not surprising given the comparative insensitivity of fission tracks to temperatures at which He ages are extremely sensitive.

There are two additional ways to obtain more detailed cooling information from a single sample: analysis of He ages of crystals of different sizes, and examination of the He concentration profile within an individual crystal. Consider three very different cooling histories: instantaneous cooling at 7 Ma, cooling at 10°C/Myr for the past 10 Myr, and a 100-Myr isothermal period at 65°C. All three histories yield apatite He ages of ~7 Ma in grains of 65 µm radius. How can these histories be distinguished?

1) **Age-grain size relationships.** Because the diffusion domain in apatite is the grain itself (Farley 2000), as the grain size decreases, the He diffusivity increases (i.e., "a" in D/a varies as the minimum grain dimension). Grains of different sizes are sensitive to different temperatures and have different closure temperatures. As shown in Figure 7, the three histories yield distinctly different He age-grain size relationships. In the case of instantaneous cooling, all He ages are the same because cooling through the range of closure temperatures happens instantly. In the sample cooling at 10°C/Myr, the ~12°C variation in closure temperature from 35 to 105°C radius causes He ages to vary by ~1.2 Myr. In the isothermal case, He ages achieve equilibrium between diffusive loss and production, and hence are enormously sensitive to grain size. Provided high precision ages can be obtained on grains spanning a fairly large size range, this technique can be used to narrow down allowed thermal histories. Such work also presupposes that the apatites lack fast sub-grain He diffusion pathways such as cracks; this may not always be the case (see Kelley 2002, this volume, for a discussion of fast pathways for Ar loss). In apatites from the Bighorn Mountains of Wyoming, Reiners and Farley (2001b) found a strong correlation between grain size and He age which they attributed to this effect and used to elaborate the thermal history of the mountain range.
Figure 7. Additional cooling history information from grain size variation of He age (upper panel) and He concentration gradient (lower panel). Three different cooling histories—instantaneous cooling at 7 Ma, constant cooling from 10 Ma to present at 10°C/Myr, and isothermal for 100 Myr at 65°C, yield the same helium age of ~7 Ma for apatite grains of 65 μm radius. The upper panel shows that the three histories yield distinct relationships between He age and grain size; the ability to actually distinguish among the histories depends on the range of grain sizes available for analysis, as well as analytical precision. The lower panel shows that each history also yields a characteristic He concentration profile within an apatite grain that might be distinguished using step-heating experiments. Ages in the upper panel were computed using the apatite diffusion parameters listed in Figure 3; ages were assigned an arbitrary uncertainty of 5%. The concentration profiles in the lower panel were computed for a grain of 65 μm radius. For both panels the α ejection effect was included and corrected for in the computations.

2) He-concentration profile. The concentration gradient within a crystal is a sensitive function of the thermal history experienced by the sample. As shown in Figure 7, the three thermal histories yield concentration profiles that are very different. In the case of instantaneous cooling, no diffusional rounding is observed at all (only the α-
ejection effect is present). The sample cooling at $10^\circ$C/Myr shows limited diffusional rounding, while the isothermal sample shows the extreme rounding characteristic of a balance between radiogenic production and diffusive loss. In particular note the differences in the profiles over the outermost 5% of the grain. If these profiles could be measured, detailed information on cooling history could be obtained. The shape of the concentration profile can to some extent be deduced from step-heating experiments (Albarede 1978), but it is not yet clear whether such an approach is practical.

**Some case studies**

Applications of (U-Th)/He thermochronometry have only recently come under investigation, so it is too early to predict how the technique may ultimately be used. Compared to other techniques, the two obvious attractions of the He method are sensitivity to previously inaccessible temperature ranges, and comparatively high precision. By combining He ages on various phases with other methods such as feldspar multidomain thermochronometry (Lovera et al. 1989) and fission track techniques (Gallagher 1995; Gallagher et al. 1998), more detailed and robust cooling histories may be obtained than previously available. In the case of apatite, the He closure temperature of $70^\circ$C is lower than that of any other known technique, so apatite He ages provide unique information on the final stages of cooling. In addition, in some cases He ages provide better age precision and better temperature control than existing techniques. This is particularly true in samples that have experienced recent cooling and hence have young ages.

The following are summaries of several studies that illustrate these applications.

**Sensitivity to smaller degrees of exhumation.** Compared to apatite fission track techniques commonly used to assess the timing and rate of mountain range exhumation, the apatite (U-Th)/He method is sensitive to temperatures $\sim 25^\circ$C lower, or about 1 km shallower in the crust. As a consequence, smaller degrees of total exhumation are required to expose the thermochronometric record of the exhumation process. In the case of rapid cooling, such exhumation may reveal an exhumed HePRZ, in which the lower break-in-slope indicates the base of the HePRZ and yields the age of the onset of exhumation (see Fitzgerald and Gleadow 1990 for a discussion of the analogous and now well understood concept in fission track dating). For example, as indicated in Figure 6, in a typical geothermal gradient of $20^\circ$C/km the base of the apatite HePRZ will be brought to the surface after $\sim 3.5$ km of exhumation. In contrast, $\sim 5$ km of exhumation are required to reveal the base of the FTPAZ.

This application has been explored by Stockli et al. (2000) in the Basin and Range province of the Western US. In the White Mountains of Eastern California, He ages decrease with structural paleodepth from 55 Ma to just a few Ma, in precisely the pattern expected for an exhumed HePRZ (Fig. 8) established between $\sim 55$ Ma and 12 Ma. Extending for $\sim 1$ km below the HePRZ the He ages are invariant at 12 Ma (Fig. 7). This invariance occurs because these initially sub-HePRZ apatites had zero age while the HePRZ developed, then were rapidly exhumed and cooled at 12 Ma. At the base of the section the He ages again decline, recording more recent cooling. Apatite fission track ages capture the 12-Ma event, but only in the structurally deepest samples; were the total exhumation only slightly less, the timing of the event would be very difficult to estimate from fission track ages alone.

**Paleotopography.** Apatite He ages are sensitive to temperatures found at depths of between 1 and 3 km within the Earth's crust (Fig. 4). At these depths the crustal temperature field is strongly influenced by the position of the free cooling surface, such
that isotherms track the overlying topography. The amplitude of relief on the isothermal surfaces depends on the wavelength of the overlying topography, decreasing rapidly with decreasing wavelength. The amplitude also decreases with increasing crustal depth and hence temperature. This phenomenon has been recognized as a complexity for interpreting low temperature cooling ages in terms of exhumation (Stuwe et al. 1994; Mancktelow and Grasemann 1997), but House et al. (1998) turned it to advantage to investigate the history of topography, in the Sierra Nevada of California. These authors proposed that if the wide and deeply incised canyons transverse to the range were an ancient feature, then He ages on an orogen-parallel transect at a constant elevation might be affected. In the limit of exhumation through steady-state topography, as the range is exhumed, rocks at a given elevation lying below deep canyons will cool before rocks lying below major ridges. As a result, He ages would be older in samples presently exposed near canyon bottoms than in samples from the same elevation between canyons.

Figure 9 shows the results obtained by House et al. (1998) on a horizontal transect at 2000-m elevation from the southern and central Sierra Nevada. Apatite He ages in the vicinity of the San Joaquin and Kings rivers are older by almost 20 Myr than samples from the intervening ridge locations. The simplest interpretation of these results is that the modern river drainages are in much the same place as when these He ages were set 60 to 80 million years ago. Modeling by House et al. (1998) suggests that the amplitude of the He age variation is consistent with mean canyon-ridge relief of about 3 km by 80 Ma.

The importance of this work is that it demonstrates that paleotopography, as distinct from exhumation, may be obtained from apatite He cooling ages. While other thermochronometers may also record this effect, because the relief on isotherms declines
rapidly with increasing temperature, it is most pronounced in the apatite (U-Th)/He system. As discussed by House et al. (1998), a variety of complications must be considered in choosing a suitable study locality, in developing an appropriate sampling strategy, and in interpreting the results, but such a tool has great potential value in understanding long-term landscape evolution.

**High-precision ages even on young samples.** In principle (U-Th)/He ages of fairly high precision (few percent) can be obtained even on samples that are very young. This suggests two additional applications for (U-Th)/He dating: absolute dating of very young volcanic units, and obtaining precise cooling histories of recently exhumed rocks. In the case of young volcanic units such as tephas, existing techniques such as $^{14}$C, U-series disequilibrium and $^{40}$Ar/$^{39}$Ar have limited ability to date samples in the range $\sim$50 kyr to $\sim$1.5 Myr, yet this age range is of considerable interest in a variety of disciplines including hominid evolution and climate history. In U-Th rich minerals such as zircon, He ingrowth is so rapid that high precision ages are possible in samples as young as 50 kyr. For example, 1 mg of zircon with 200 ppm each of U and Th will carry more than 300 times typical helium blank levels using laser extraction in 50 kyr, and a single zircon crystal of $\sim$10 $\mu$g mass will exceed blank by 10 times in just 175 kyr.

However, an important issue to be dealt with in tephra dating is that young crystals might acquire He prior to establishment of secular equilibrium in the $^{238}$U system. In particular, zircons likely crystallize with a significant $^{230}$Th/$^{232}$U deficit. As a result, each decay of $^{238}$U initially produces less than 8 $\alpha$ particles (Eqn. 1) until secular equilibrium is achieved. The magnitude of the discrepancy introduced by secular disequilibrium (relative to Eqn. 1) varies systematically with eruption age, the Th/U ratio of the crystal, and the magma residence time prior to eruption (Farley et al. 2002). For example, in the case of a typical zircon, the $^{230}$Th deficit would cause He ages computed from Equation (1) to underestimate the eruption age by 50% in very young samples (10 kyr) declining to $\sim$10% by 1 Myr. Farley et al. (2002) presented equations for accommodating secular disequilibrium, and demonstrated that analysis of samples from a single eruption spanning a range of Th/U ratios can be used to identify and eliminate secular disequilibrium effects on computed eruption ages. These authors presented measurements of apatite and zircon separates of a New Zealand Tephra indicating an eruption age of 330 $\pm$ 10 kyr. Although this application is in its infancy, even with the uncertainties introduced by secular disequilibrium the technique appears promising.

In a similar fashion, cooling ages obtained from the (U-Th)/He system compare favorably in precision, especially at young ages, with alternative techniques such as apatite fission track counting. For example, Spotila et al. (2001) showed that apatite He ages of $\sim$1.5 Ma can be obtained with a precision of 100 kyr (2$\sigma$) on a rapidly exhumed crustal slice lying between two strands of the San Andreas fault where it passes through San Gorgonio Pass in Southern California. In a 1 km vertical transect, 8 samples yielded He ages ranging from 1.39$\pm$0.10 (2$\sigma$) to 1.64$\pm$0.11 Ma. The high precision on these ages constrains not only the onset of rapid exhumation of the slice, but places fairly restrictive limits on the exhumation rate. Precision on both of these geologic quantities is critical for establishing the mechanism responsible for exhumation along this major strike-slip fault.

**FUTURE PROSPECTS**

Work over the past five years suggests a variety of techniques and applications for (U-Th)/He dating. Some of these, such as apatite He thermochronometry, have been applied in a range of settings and have yielded geologically reasonable and scientifically interesting results. Other techniques, while tantalizing, are not yet sufficiently developed
for a critical evaluation. The coming decade will provide a much clearer view of which applications and techniques "work," and which suffer insurmountable problems.

Beyond the (U-Th)/He method itself, it will become increasingly important to link these new methods to existing techniques. For example, detailed cooling histories through the temperature range ~110 to 45°C should be possible using combined apatite fission track (Gallagher et al. 1998) and (U-Th)/He methods. Similarly, titanite He ages can logically be used to evaluate and refine cooling histories derived from K-feldspar $^{40}$Ar/$^{39}$Ar multidomain modeling (Lovera et al. 1989). Before these linkages can be made it is critical to establish that the various techniques are accurately intercalibrated, an exercise that will likely require both laboratory investigations and analysis of well-understood natural systems.

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REFERENCES

Hurley PM (1952) Alpha ionization damage as a cause of low He ratios. EOS Trans Am Geophys Union 33:174-183
Lippolt HJ, Weigel E (1988) \(^{3}H_{e}\) diffusion in \(^{39}Ar\) retentive minerals. Geochim Cosmochim Acta 52: 1449-1458
Reiners PW, Farley KA (2001b) Influence of crystal size on apatite (U-Th)/He thermochronology: an example from the Bighorn mountains, Wyoming. Earth Planet Sci Lett 188:413-420
Spottiswoode JA, Farley KA, Yule JD, Reiners PW (2001) Near-field convergence along the San Andreas fault zone in southern California, based on exhumation constrained by (U-Th)/He dating. J Geophys Res 106:26731-26746
Stuart FM, Persano C (1999) Laser melting of apatite for (U-Th)/He chronology: Progress to date. EOS Trans Am Geophys Union 80:F1169
Toyoda S, Ozima M (1983) Investigations of excess \(^{3}He\) and \(^{39}Ar\) in beryl by laser extraction technique. Earth Planet Sci Lett 70:69-76
Trull TW, Kurz MD, Jenkins WJ (1991) Diffusion of cosmogenic \(^{3}He\) in olivine and quartz: implications for surface exposure dating. Earth Planet Sci Lett 103:241-256
Wernicke RS, Lippolt HJ (1994a) \(^{4}He\) age discordance and release behavior of a double shell botryoidal hematite from the Schwarzwald, Germany. Geochim Cosmochim Acta 58:421-429
Young E, Myers A, Munson E, Conklin N (1969) Mineralogy and geochemistry of fluorapatite from Cerro de Mercado, Durango, Mexico: U S Geol Surv Prof Paper 650-D:84-D93