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Review article

Low-temperature thermochronology applied to ancient settings

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ABSTRACT

Low-temperature thermochronometric dating techniques are commonly used to reveal and quantify the spatial and temporal pattern of cooling and exhumation in many active mountain belts. These methods illuminate the relationship between rock exhumation, and the processes that bring rock to the surface, including climate-driven surface processes. Thermochronological studies on rocks in ancient settings (Precambrian and Paleozoic) are far more complicated, and in general have largely failed to reveal exhumation processes in orogenic belts, mainly due to imprecision of the technique, complications from radiation damage, and subsequent resetting. However, in a few locations, remnants of the exhumation record of ancient orogens may be studied with low-temperature dating techniques and in these cases we can gather first-order information on tectonics, thermal events, and exhumation processes that have occurred in the deep past. Here we provide an overview of the most common low-temperature thermochronology methods of fission-track and (U–Th)/He dating of apatite and zircon and discuss methodological challenges in their application to old settings. We review several case studies from the Northern Appalachians, central Argentina, the Canadian Shield, central Wyoming, and the European Variscides where thermochronology data have been successful in reconstructing geologic processes in deep-time that have affected the upper crust. With these examples we provide sampling strategies and analytical approaches that partly circumvent issues related to radiation damage in old mineral grains that result in changes in the retention of helium and fission tracks.

Contents

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1. Introduction

Low-temperature thermochronometric dating methods such as fission track (FT) and (U–Th)/He are widely used to quantify the timing, rates, and spatial distribution of rock cooling and exhumation related to processes affecting the upper crust (e.g. Garver et al., 1999, 2005; Reiners and Brandon, 2006; Blythe et al., 2007). Exhumation results in the vertical motion of rock with respect to the Earth's surface, and the processes that remove overlying rock are driven by tectonic and climate processes (Ring et al., 1999). Low-temperature thermochronology studies are ideally suited to regions characterized by significant exhumation that has occurred in the late Mesozoic to Cenozoic time, including extensional tectonic settings (e.g. Stockli et al., 2000; Armstrong, 2003; Ehlers, 2003) and contractional tectonic settings at subduction zones and collisional belts (e.g. Garver and Kamp, 2002; Garver et al., 2005; Enkelmann et al., 2009; Cao et al., 2013).

Thermochronology has revealed important and otherwise subtle features of orogenic belts and as such these studies have allowed for significant advancement of understanding how mountain belts form and evolve though time. For example, development of orographic barriers results in enhanced erosion of rocks which drives erosional exhumation, and dramatic changes in cooling rates in orogenic belts that have been quantified in many areas such as the Himalayan Range (e.g. Wang et al., 2010), the Coast Mountains and Cascades in the northwestern North America (e.g. Reiners et al., 2002a; Shuster et al., 2005; Ehlers et al., 2006), New Zealand (e.g. Garver and Kamp, 2002) and many other orogenic belts. The much dryer lee side of mountain ranges tends to experience little erosion and thus minor rock exhumation is expressed in thermochronologic data that are usually characterized by older cooling ages with large spatial and temporal variations.

Rocks of ancient orogens exposed at the surface today commonly comprise high-grade metamorphic and plutonic rocks that have been exhumed from the middle and lower crust. Investigations of exhumation processes in these older orogenic settings commonly employ higher temperature dating techniques such as $^{40}Ar/^{39}Ar$ analysis of potassium bearing minerals like biotite, hornblende, muscovite, or U/Pb dating on zircon, titanite, and monazite (e.g. Eide et al., 1994; Hacker and Gans, 2005; Hacker et al., 1995, 1998; Rötzler and Romer, 2001; O'Brien and Rötzler, 2003). Combined with geochemical, structural, and petrological data, the temperature–pressure–time (T–P–t) paths are derived to constrain burial and exhumation models that explain the motion of rock units from deep structural levels into the middle crust (e.g. Chemenda et al., 1995; Vanderhaeghe et al., 1999; Vanderhaeghe et al., 2003; Godin et al., 2006; Kroner and Goerz, 2010; Sizova et al., 2012). However, the timing of exhumation, the processes that affected rocks at the near surface, and the role of surface processes on rock exhumation during ancient orogeny are usually unknown, partly because the rocks that hold this record have long-since been removed by erosion, or thermally overprinted. An important opportunity is that a fragmentary part of this exhumation may be preserved in bedrock or flanking sedimentary sequences that record erosion of the orogen.

The post-orogenic processes that have affected the thermal field in the upper crust commonly complicate the study of older orogenic sequences by low-temperature thermochronology due to thermal resetting. The thermal sensitivity of these dating methods to temperatures of ∼40–300 °C (see Table 1) results in common thermal resetting due to subsequent heating and small temperature variations in the crust. The same is the case for the cooling-ages retained in the minerals of synorogenic sediments, which commonly record the thermal history of the basin that experienced heating due to subsidence and burial, followed by cooling due to later uplift and basin inversion.

Table 1
Most commonly used thermochronologic systems, their closure temperature and depths of closure isotherm.

Method	Closure temperature (°C)	Closure depths (km)
Apatite (U–Th)/He	55–75	1.5–3.3
Zircon (U–Th)/He	160–200	5–9.5
Apatite fission track	100–120	3–5.5
Zircon fission track	210–300	9.7–14.5

Note: closure depths calculated using a geothermal gradient of 20–30 °C/km, and 10 °C surface temperature.

Thus a central challenge in understanding the low-temperature thermal history of old orogenic belts is finding sequences that have a record of the orogen, but have escaped subsequent heating. In some cases, the history of orogenic cooling may be preserved over hundreds of million years and can be studied using low-temperature thermochronology. In these special cases, we are able to derive first-order information about exhumation processes that may be used to evaluate not only the role of certain structures, but also paleo-climate, paleo-geography, and ancient surface processes. In this paper, we aim to provide the methodological background for the most commonly used low-temperature thermochronology methods and the challenges regarding their applications in ancient tectonic settings. We then present case studies where low-temperature thermochronology has been used successful in revealing ancient geologic histories. We hope that these examples provide strategies and approaches that encourage the investigation of upper-crustal processes in ancient mountain belts by means of low-temperature thermochronology, but we highlight some important limitations.

2. Low-temperature thermochronology

Low-temperature thermochronology employs radiometric-dating techniques that are characterized by thermally sensitive daughter products that accumulate and are retained in minerals at relative low temperatures (40–300 °C) over geologic times. FT analysis and (U–Th)/He dating of the minerals apatite and zircon are commonly and widely used (Table 1). Both methods are primarily based on radioactive decay of uranium. The α-decay of uranium and thorium causes accumulation of helium in minerals, which forms the basis for the (U–Th)/He analysis (Zeitler et al., 1987; Farley, 2002), whereas the FT analysis is based on the fission decay of uranium that forms a linear damage zone in the mineral lattice (Fleischer et al., 1975).

Fission tracks are formed continuously at all temperatures and depths, but they annealed more or less instantaneously if mineral is brought to high temperatures. For apatite this temperature of continuous annealing is above ∼100–120 °C (Green et al., 1986) and for zircon it is above ∼250–300 °C (Tagami et al., 1998; Tagami, 2005). One unique aspect of FT dating is that there is a temperature of partial annealing where the process of track annealing is slow enough to be geologically significant, and this is typically referred to as the Partial Annealing Zone (PAZ). At temperatures ranging from 60 to 120 °C and 210 to 300 °C tracks are partially annealed in apatite and zircon, respectively, resulting in the shortening of the track length (Fig. 1).

An analogous concept is recognized with the retention of helium in what is referred to as the Partial Retention Zone (PRZ). Helium is a noble gas and is quickly lost by volume diffusion at high temperatures. For apatite this is above ∼80–90 °C (Farley, 2000) and for zircon this is ∼180–200 °C (Reiners et al., 2002b). At temperatures ranging from ∼40–80 °C and ∼160–200 °C helium is partially retained in apatite and zircon, respectively (Fig. 1). Within the higher-temperature part of the PAZ or PRZ is the closure