Exhumation of the southern Pyrenean fold-thrust belt (Spain) from orogenic growth to decay

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[1] The deformation and exhumation history of an orogen reflects the interactions between tectonic and surface processes. We investigate orogenic wedge deformation, erosion, and sedimentation in the Pyrenees by (a) quantifying the spatiotemporal patterns of exhumation across the southern fold-thrust belt (FTB) with bedrock apatite fission track (AFT) thermochronology and (b) comparing the results with existing deformation, exhumation, and sedimentation chronologies. Eighteen new samples record exhumation during and after orogenesis between 90 and 10 Ma. Rocks from the range core (Axial Zone) record rapid exhumation that progresses east to west and north to south, consistent with patterns of tectonically driven uplift. Synorogenic sediments shed into piggyback basins on the southern fold-thrust belt during mountain building retain a detrital exhumation signal from the Axial Zone. In contrast, samples from other structural positions record exhumation of the thin-skinned Pyrenean thrust sheets, suggesting sediment burial and heating of sufficient magnitudes to reset the AFT system (>3 km). In some locations, exhumation of these fold-thrust structures is likely an erosional response to thrust-driven rock uplift. We identify an exhumation phase ~ 25 – 20 Ma that occurs along the central and eastern Spanish Pyrenees at the boundary between thick- and thin-skinned portions of the wedge. We suggest that this distributed exhumation event records (a) a taper response in the southern orogenic wedge to sediment loading and/or (b) a shift to wetter, stormier climate conditions following convergence-driven uplift and full topographic development. A final exhumation phase between ~ 20 and 10 Ma may record the excavation of the southern fold-thrust system following base level lowering in the Ebro Basin.

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1. Introduction

[2] Tectonic and surface processes play evolving roles in the development of linked thrust belt-foreland basin systems as they shift from active shortening to postorogenic decay [Allen, 2008]. Collisional mountain ranges are traditionally modeled in cross section as bivergent critical wedges [e.g., Beaumont *et al.*, 1992; Mugnier *et al.*, 1997] that accommodate shortening by oscillating around an equilibrium (critical) state defined by the mean topographic gradient and the basal décollement (taper) [Chapple, 1978]. Deformation, erosion, and sedimentation redistribute mass

within an evolving wedge, influencing the topographic and structural expression of convergent tectonics [Davis *et al.*, 1983; Dahlen *et al.*, 1984]. After orogenesis ceases, processes such as isostatic rebound [Gilchrist *et al.*, 1994], drainage basin reorganization [García-Castellanos *et al.*, 2003], and global-to-regional climate change [Whipple, 2009] can continue to alter the form of the mountain range. Therefore, quantifying the history of deformation, erosion, and sedimentation is important for understanding orogen evolution.

[3] The geometry of collisional orogens tends toward asymmetry across strike, with a wider pro-wedge forming over the underthrusting plate and a narrower retro-wedge forming over the overriding plate [Willett *et al.*, 1993; Sinclair *et al.*, 2005]. Orographic focusing of precipitation can enhance or even reverse this asymmetry [Willett, 1999]. Thermochronology provides a useful tool for understanding orogen evolution by tracking the thermal histories and trajectories of material that moved through an orogen [Batt and Brandon, 2002; Reiners and Brandon, 2006]. Studies often employ bedrock thermochronometer sampling on a single transect across an orogen, leading to a two-dimensional understanding of processes affecting orogen evolution [e.g., Brandon *et al.*, 1998; Fitzgerald *et al.*, 1999]. However, this approach oversimplifies the mountain range, neglecting the effects of factors such as oblique

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convergence [Whitchurch *et al.*, 2011], crustal heterogeneities [Beaumont *et al.*, 2000], and regional climate gradients in controlling three-dimensional mountain belt form and evolution.

[4] The Pyrenees are well suited to studying orogen evolution both across and along strike. Formed from Late Cretaceous to early Miocene (~80–20 Ma) continental collision of Iberia and Eurasia [Sibuet *et al.*, 2004], this presently inactive mountain range accommodated moderate but variable shortening magnitudes (<165–80 km) along strike [Muñoz, 1992; Vergés *et al.*, 1995; Teixell, 1998; Beaumont *et al.*, 2000]. Previous studies extensively investigated erosional exhumation in the range interior with a particular focus on the central Pyrenees. These studies suggest that variations in surface mass flux across the wedge and into the adjacent basins strongly impacted the development and architecture of the range [e.g., Morris *et al.*, 1998; Fitzgerald *et al.*, 1999; Sinclair *et al.*, 2005; Whitchurch *et al.*, 2011; Fillon and van der Beek, 2012]. Although the history of deformation and sedimentation within and adjacent to the southern fold-thrust belt (FTB) is well constrained [e.g., Burbank *et al.*, 1992a; Hogan and Burbank, 1996; Vergés and Burbank, 1996; Rahl *et al.*, 2011], the relationship with exhumation on the thin-skinned portion of the wedge remains poorly quantified. Zircon fission track samples from the southern Pyrenees have an inherited exhumation signal from the range interior [Filleaudeau *et al.*, 2011; Whitchurch *et al.*, 2011], but pilot apatite (U-Th)/He samples from the region are reset [Fillon *et al.*, 2010], hinting that lower-temperature thermochronometers, such as fission tracks and (U-Th)/He in apatite, may be appropriate to target exhumation patterns on the thin-skinned FTB. Recent work highlights the importance of diachronous along-strike uplift on sediment production and routing early in the range history [Whitchurch *et al.*, 2011], but the continued effects of topographic development remain unknown because the sedimentary record extends to only ~28 Ma.

[5] In this study, we use bedrock apatite fission track (AFT) thermochronology along the southern Pyrenean margin to extend constraints on spatiotemporal patterns of exhumation along strike and improve our understanding of the relationships between tectonic and surface processes during orogen development. We interpret our data set within the context of a series of balanced regional cross sections, correlate exhumation timing with magnetostratigraphic constraints on deposition and deformation, and integrate

our data with previous AFT data sets. The thermal sensitivity of AFT thermochronology allows us to distinguish between phases when the thin-skinned FTB acted as a sediment source and sink during and after orogenesis. Our results provide the first direct evidence of significant erosion of the southern FTB after known phases of structural development. Unreset to partially reset samples from the Graus-Tremp Basin support interpretations of moderate syntectonic sediment burial of the southern FTB during Ebro foreland basin closure. We identify synchronous along-strike exhumation at the boundary between the thin-skinned FTB and the range interior during the final phases of orogenesis. While previous work attributes exhumation during this time period in the central Pyrenees to adjustment in response to sediment loading [Sinclair *et al.*, 2005; Fillon *et al.*, 2012], the structural heterogeneity between the FTB in the central and eastern Pyrenees calls this mechanism into question, at least for the eastern Pyrenees. Alternatively, we suggest a potential link between full along-strike topographic development during this time period and enhanced exhumation by orographic forcing that currently results in higher precipitation and summer storminess in the eastern Pyrenees than other regions of Spain.

2. Geologic Setting

[6] In the Early Cretaceous (~125 Ma), the Iberian plate rotated counterclockwise and began converging northward with Europe (Figure 1) [Sibuet *et al.*, 2004]. Partial subduction of the Iberian lithosphere under the European plate and inversion of preexisting extensional faults accommodated shortening along the plate boundary [Vergés *et al.*, 2002]. Major Pyrenean mountain building lasted from the Late Cretaceous to early Miocene (~80–20 Ma) [Sinclair *et al.*, 2005], forming a wedge with a central Axial Zone and flanking thrust belt-foreland basin systems (Figure 2) [Muñoz, 1992; Roure *et al.*, 1989]. Beginning at ~36 Ma, the southern Ebro foreland basin lost connection with the Atlantic, causing sediment shed from the exhuming Pyrenees to fill the basin and bury the southern FTB [Coney *et al.*, 1996; Costa *et al.*, 2009]. Preserved and variably deformed sediments reach elevations of ~1700 m in the central Pyrenees and ~2100 m in the eastern Pyrenees at present [Coney *et al.*, 1996] or modeled paleoelevations of up to 2.6 km [Fillon and van der Beek, 2012]. Fluvial

Figure 1. Kinematically restored plate motions based on paleomagnetic data with Europe fixed (simplified from Sibuet *et al.* [2004]). Arrows indicate relative plate motion directions. Hatched regions show where compression occurred, and dark blue represents ocean basin.

Figure 2. Geology of northeastern Iberia. (a) Simplified geologic map for the Pyrenees and surrounding regions (modified from Vergés and Burbank [1996]). Inset shows location and regional geographic context: Ap, Apennines; CC, Catalan Coastal Ranges; P, Pyrenees. Southern thrust sheets: B, Boixols thrust sheet; C, Cadi thrust sheet; M, Montsec thrust sheet; N, Noguères Zone; P, Pedraforca; SM, Sierras Marginales. VE, vertical exaggeration. Red box is the study area, and the blue line indicates the ECORS seismic profile [Muñoz, 1992]. (b) Crustal-scale cross section along the ECORS transect (location in Figure 2a) (modified from Beaumont *et al.*, 2000). Note the bivergent wedge geometry and central duplex versus marginal thin-skinned FTB structures of the Pyrenees.

excavation of the southern FTB followed Ebro Basin capture by the Mediterranean 13–8.5 Ma [García-Castellanos *et al.*, 2003; Arche *et al.*, 2010]. Rifting related to the opening of the Valencia Trough caused uplift of the southeastern margin of the Ebro Basin and eastern Pyrenees during the Neogene and Quaternary (<~23 Ma) [Lewis *et al.*, 2000].

[7] Pyrenean thrust belt structure, deformation timing, and spatial variations in shortening are well documented and relevant to the results of this study. Seismic reflection, balanced cross sections, and geodynamic models indicate maximum shortening of ~165–147 km in the central Pyrenees along the Etude Continentale et Océanique par Reflexion et Refraction Sismique (ECORS) transect (Figure 2b) [Beaumont *et al.*, 2000; Muñoz, 1992]. Shortening decreases away from ECORS, with minimum estimates of ~125 km across the eastern Pyrenees [Vergés *et al.*, 1995] and ~80 km across the western Pyrenees [Teixell, 1998]. The ECORS deep reflection seismic survey shows that the Axial Zone basement rocks are deformed into a crustal-scale duplex structure [Muñoz, 1992]. The uppermost thrust sheet of the duplex forms the Noguères Zone that acts as the backstop for the southern, thin-skinned portion of the FTB. To the east,

the structural equivalent is the Freser antiformal stack (Figure 3) [Burbank *et al.*, 1992b; Muñoz *et al.*, 1986]. The thin-skinned portion of the FTB is widest in the central Pyrenees and consists of the Boixols, Montsec, and Sierras Marginales thrust sheets from north to south (Figures 2b and 3). Deformation across the central and eastern thin-skinned FTB lasted from the Late Cretaceous to Oligocene (~85–24 Ma). The south central thrust sheets form an imbricate fan system that generally propagated southward toward the foreland (Figures 2 and 3) [Vergés and Muñoz, 1990], but some north-vergent and out-of-sequence thrusting is evident from crosscutting stratigraphic relationships [e.g., Burbank *et al.*, 1992a; Meigs *et al.*, 1996; Meigs and Burbank, 1997; Ramos *et al.*, 2002]. In the eastern Pyrenees, the Vallfogona Thrust marks the southern boundary of the Cadi and Pedraforca thrust sheets [Vergés *et al.*, 2002]. Magnetostratigraphy and fault gouge dating constrain thrust sheet development in the eastern Pyrenees to a narrower time frame than the south central Pyrenees, from ~54 to 30 Ma [Burbank *et al.*, 1992b; Capote *et al.*, 2002; Haines, 2008].

[8] Existing thermochronometer data quantify the broad exhumation patterns across the Pyrenees (Figure 3).

Figure 3. Topography, geology, and apatite fission track thermochronology data of the central and eastern Pyrenees. This synthesis highlights new and published bedrock apatite fission track (AFT) data within 10 km of the four transects. Inset shows our sample classification scheme, based on thermal model exhumation timing (white or black rim), location relative to where the sample most recently exhumed from AFT closure temperature (circle, star, or circled star; see discussion text for details), and data source (red or blue) [Morris *et al.*, 1998; Fitzgerald *et al.*, 1999; Gibson *et al.*, 2007; Jolivet *et al.*, 2007; Gunnell *et al.*, 2009; Beamud *et al.*, 2010]. Note that AFT pooled age, shown in white boxes alongside the sample number, and exhumation timing are not correspondent for discordant samples. WT, western transect; CWT, central western transect; CET, central eastern transect; ET, eastern transect. Geologic units were generalized from Mapa Geologic de Catalunya [2003], and topography is the 30 m Advanced Spaceborne Thermal Emission and Reflection Radiometer Global Digital Elevation Model (<http://asterweb.jpl.nasa.gov/gdem-wist.asp>).

Paired zircon U-Pb geochronology and detrital fission track thermochronology in the southern thrust belt and foreland basin found that convergence, shortening, and exhumation accelerated during the Late Cretaceous (~80 Ma) [Filleaudeau *et al.*, 2011] and that oblique convergence caused topography to develop diachronously along strike, from east to west between the Late Cretaceous and Miocene (~80–20 Ma) (Figure 1) [Whitchurch *et al.*, 2011]. Axial Zone Hercynian plutons exhumed most rapidly ~35–30 Ma in both the east [Morris *et al.*, 1998] and the south [Fitzgerald *et al.*, 1999], followed by slower exhumation starting at 30 Ma. The reduced exhumation from 30 Ma is attributed to base level change in the Ebro foreland basin, where syntectonic sediment infill reduced local relief [Beamud *et al.*, 2010]. ECORS transect thermochronometer data and discrete element modeling show an asymmetric exhumation pattern in the central Pyrenees [Fitzgerald *et al.*, 1999; Sinclair *et al.*, 2005]: exhumation migrated from north to south in response to the propagation of deformation and erosion

across the pro-wedge. This asymmetric pattern caused orogenesis in the Pyrenees to end with syntectonic sediment blanketing of the southern range margin, pushing deformation from the FTB front back toward the pro-wedge hinterland, enhancing topography and accelerating erosion ~20 Ma in the Noguères Zone (Figure 3) [Sinclair *et al.*, 2005]. Remnants of this sediment blanket preserved in syntectonic basins on the southern thrust sheets record rapid cooling at ~50–40 and ~30–25 Ma associated with source region exhumation in the Axial Zone [Beamud *et al.*, 2010; Rahl *et al.*, 2011]. Samples from the lowest elevations in the piggyback basins record partial fission track annealing <10 Ma, contemporaneous with fluvial excavation following Ebro Basin capture by the Mediterranean [Beamud *et al.*, 2010]. Finally, thermokinematic modeling of the Axial Zone low-temperature thermochronometer data suggests that the Oligocene sediment cover filled in the paleotopography of the south central Pyrenees to depths of ~2 km [Fillon and van der Beek, 2012].

3. Apatite Fission Track Thermochronology

3.1. Overview

[9] Apatite fission track (AFT) thermochronology uses the formation and temperature-dependent retention of damage trails called fission tracks that accumulate in apatite grains from the fission decay of ^{238}U to reconstruct the cooling histories of rocks in the uppermost crust [Fleischer *et al.*, 1975; Gallagher *et al.*, 1998]. Apatite grains retain fission tracks at temperatures below an effective closure temperature that depends on chemical composition ($\sim 110^\circ\text{C}$ – 120°C for fluorapatite), cooling rate, and the length of time spent at elevated temperatures [Ketcham *et al.*, 1999]. An AFT age, measured from the ratio of parent to daughter (^{238}U : fission tracks) products, represents the time since the grain cooled from below its closure temperature [Dodson, 1973]. Fission tracks form with lengths of 14.5–16 μm [Gleadow *et al.*, 1986] but shorten (anneal) over geologically significant timescales within a temperature window called the partial annealing zone (PAZ) that occurs over a range of $\sim 60^\circ\text{C}$ [e.g., Gallagher *et al.*, 1998].

[10] Many (~ 20 – 40) apatite grains are analyzed in a bedrock sample, providing a distribution of fission track lengths and ages. The track length distribution preserves a record of the thermal history of the sampled rock as it cooled through the PAZ [Gleadow *et al.*, 1986]. Summing the individual grain ages within each sample yields a pooled age [Donelick *et al.*, 2005]. A chi-square test (χ^2) assesses the sample grain age variability [Galbraith, 1981]. A $P(\chi^2) > 5\%$ indicates minor grain age variability, and these samples are called concordant. A concordant sample pooled age represents the last time it cooled from closure temperature [Brandon *et al.*, 1998; Galbraith, 1981; Green, 1981]. A $P(\chi^2) < 5\%$ indicates significant grain age variability, or discordance, and the pooled age is considered less meaningful [e.g., Green, 1981]. Discordance may result from multiple component ages or heterogeneous mineral properties, especially in sedimentary rocks that can have multiple sediment sources [Tagami and O'Sullivan, 2005]. Fission track diameter (D_{par}) is a kinetic parameter that relates to chemical composition among other factors and is commonly used to determine permissible time-temperature histories when inverse thermal modeling AFT data [Donelick *et al.*, 2005; Ketcham *et al.*, 1999]. Applying a combination of component age analysis and thermal modeling is a useful approach for constraining the full range of sample cooling histories permitted by measured ages and track lengths [Barnes *et al.*, 2006, 2008].

3.2. Sample Collection and Analysis

[11] We collected bedrock samples from four across-strike transects of the central eastern Spanish Pyrenees to determine the regional exhumation patterns across the southern FTB margin (Figure 3). These transects follow the balanced cross sections of Vergés [1999]. We targeted the major FTB structures, including the Cadi, Montsec, and Sierras Marginales thrust sheets, the Noguères Zone, the Oliana anticline, and the Freser antiformal stack. We sampled Permian, Triassic, and Cretaceous sandstones and Eocene turbidites, sandstones, and conglomerates. We also sampled a Paleozoic schist north of the Freser antiformal stack and the Mount Louis-Andorra pluton exposed in the Axial Zone. We used standard techniques to isolate apatite grains

and measure fission track ages, lengths, and D_{par} values (Appendix A). The fission track ages were determined using the laser ablation inductively coupled plasma–mass spectrometry method (LA-ICP-MS) [Donelick *et al.*, 2005; Hasebe *et al.*, 2004].

3.3. Grain Age Analysis and Thermal Modeling

[12] We applied both grain age analysis and thermal modeling to aid in the interpretation of the AFT data (see details in Appendices B and C). For all discordant samples ($P(\chi^2) < 5\%$), we used binomial peak fitting [Galbraith and Green, 1990] to identify the statistically significant age components with the software RadialPlotter [Vermeesch, 2008]. We classified sample grain age distributions to assess the cooling history as reset (R), mixed reset (MR), partially reset (PR), or detrital (D) (cf. Brandon *et al.* [1998]). R and MR samples have one or more component ages that are younger than the bedrock deposition or crystallization age. A PR sample contains component ages both younger and older than the deposition or crystallization age. D samples have one or more component ages that are older than the deposition or crystallization age. In sedimentary samples, a D sample implies cooling during sediment sourcing prior to deposition, R and MR samples suggest cooling occurred after deposition and lithification, and a PR sample implies retained cooling signals from both prior to and after deposition and lithification.

[13] We used the thermal modeling software HeFTy [Ehlers, 2005; Ketcham, 2005] to constrain sample cooling histories consistent with the measured FT age, length, and D_{par} data. We simulated geologic processes that the apatite grains experienced such as unroofing, erosion and burial, and postdepositional exhumation (Figure 3, inset) by progressively incorporating time-temperature constraints on inverse models for each sample (see details in Appendix B). We report time-temperature path envelopes that encompass the common merit values of 0.5 (good model fits) and 0.05 (acceptable model fits) from a Kuiper statistical test [Ketcham, 2005]. Because our sampling locations were unaffected by Cenozoic volcanism or extension [e.g., Vergés *et al.*, 2002], we attribute recent thermochronometer-recorded cooling to erosional exhumation.

4. Results

4.1. Overview

[14] We report AFT data for 18 bedrock samples from Cambrian through Eocene sandstones, conglomerates, schists, and granodiorites exposed across the southern Axial Zone and FTB of the Spanish Pyrenees (Figure 3). Sample grain age and track length yields range from maximum (40 ages, 200+ track lengths) to very poor (3 ages, 5 track lengths). Sample pooled ages range from Late Cretaceous to early Miocene (76.8–20.1 Ma) with mean track lengths from 15.1 to 12.9 μm , implying fast to moderate cooling through the PAZ (Table 1). The pooled ages, identical within error to the central ages (Appendix C), suggest that the sampled bedrock cooled during Pyrenean orogenesis. However, since most samples are discordant, we used thermal modeling and grain age analysis to evaluate this simple interpretation. There is no correlation between D_{par} and either track lengths or grain ages within our samples, indicating that sample discordance does not result from

Table 1. Apatite Fission Track Data

Sample	Lat	Long	Elevations (m)	Fm Age (Ma)	n	N_c (tracks)	D_{par} (μm)	Area (cm^2)	$\Sigma(PQ)$ (cm^2)	$1\sigma \Sigma(PQ)$ (cm^2)	ξ_{MS}	$1\sigma \xi_{MS}$	$P(\chi^2)$ (%)	Pooled Age (Ma $\pm 1\sigma$)	C. Age Class	MTL $\pm 1\sigma$, μm (N_r tracks)
<i>Eastern Transect (ET)</i>																
ET1	42.27	2.16	823	99–54.8	40	356	1.48	1.08E–03	1.11E–04	8.55E–07	14.309	0.328	0.0	22.9 $\pm$ 2.6	PR	13.84 $\pm$ 2.82 (98)
ET2	42.29	2.16	854	99–54.8	27	120	1.25	5.81E–04	5.27E–05	9.68E–07	17.633	0.431	0.7	20.1 $\pm$ 3.8	PR	14.66 $\pm$ 2.02 (52)
ET3	42.32	2.17	934	543–443	8	35	1.57	1.26E–04	8.03E–06	3.34E–07	14.279	0.328	5.8	31.0 $\pm$ 11	–	13.32 $\pm$ 2.90 (25)
<i>Central Eastern Transect (CET)</i>																
CET1	42.13	1.9	637	49–41.3	31	197	1.28	4.19E–04	5.52E–05	6.97E–07	16.821	0.415	0.4	30.0 $\pm$ 4.6	R	13.01 $\pm$ 02.96 (156)
CET2	42.28	1.85	1052	248–206	38	734	1.36	7.21E–04	2.70E–04	2.84E–06	17.491	0.428	0.2	23.7 $\pm$ 2.2	MR	13.60 $\pm$ 2.84 (182)
<i>Central Western Transect (CWT)</i>																
CWT1	42.06	1.31	472	37–33.7	40	939	1.55	1.09E–03	1.53E–04	1.15E–06	14.249	0.329	0.0	43.6 $\pm$ 3.6	PR	14.24 $\pm$ 3.10 (205)
CWT2	42.18	1.32	520	99–54.8	5	61	1.13	9.56E–05	1.99E–05	6.24E–07	17.396	0.426	41.8	26.7 $\pm$ 7.2	–	13.25 $\pm$ 2.99 (12)
CWT3	42.29	1.37	592	248–206	40	1643	1.52	1.74E–03	4.22E–04	3.20E–06	14.198	0.33	0.0	27.6 $\pm$ 2.0	R	14.18 $\pm$ 3.54 (202)
CWT4	42.3	1.37	578	290–248	40	155	1.56	1.03E–03	3.78E–05	2.57E–07	14.046	0.332	3.6	28.7 $\pm$ 4.8	R	14.06 $\pm$ 3.62 (26)
<i>Western Transect (WT)</i>																
WT1	41.88	0.88	286	248–206	3	16	2.91	4.27E–05	2.17E–06	3.97E–08	13.957	0.334	48.8	51.2 $\pm$ 25.8	–	15.12 $\pm$ 0.46 (5)
WT2	42.01	0.88	336	248–206	8	119	1.44	9.12E–05	1.87E–05	2.91E–07	13.964	0.334	0.0	44.3 $\pm$ 8.4	MR	14.67 $\pm$ 3.38 (24)
WT3	42.06	0.89	364	144–165	39	3075	1.27	1.13E–03	3.38E–04	3.74E–06	16.99	0.418	0.0	76.8 $\pm$ 5.0	PR	12.95 $\pm$ 3.60 (204)
WT4	42.16	0.88	504	99–54.8	17	205	1.36	3.11E–04	3.51E–05	5.66E–07	17.113	0.421	0.0	49.7 $\pm$ 7.6	MR	12.85 $\pm$ 3.06 (13)
WT5	42.15	0.81	766	54.8–33.7	15	190	1.37	3.38E–04	2.73E–05	6.25E–07	17.183	0.422	0.0	59.5 $\pm$ 9.6	D	13.09 $\pm$ 2.56 (37)
WT6	42.16	0.78	1039	54.8–33.7	37	2422	1.29	1.45E–03	4.37E–04	5.16E–06	17.3	0.424	0.0	47.7 $\pm$ 3.2	PR	13.93 $\pm$ 2.95 (200)
WT7	42.33	1.07	611	248–206	40	953	1.56	1.43E–03	2.44E–04	1.48E–06	13.995	0.333	0.0	27.3 $\pm$ 2.2	MR	14.49 $\pm$ 2.80 (201)
<i>Axial Zone (AZ)</i>																
AZ1	42.51	1.57	1593	354–248	40	557	1.42	1.40E–03	1.72E–04	1.40E–03	14.147	0.331	0.0	22.8 $\pm$ 2.2	MR	14.17 $\pm$ 3.82 (133)
AZ2	42.51	1.55	1134	354–248	40	471	1.41	1.44E–03	1.37E–04	1.44E–03	14.097	0.332	0.0	24.2 $\pm$ 2.6	MR	13.64 $\pm$ 3.42 (136)

Fm, Formation; n , number of grain measured; D_{par} , mean maximum diameter of fission track etch figures parallel to the c axis; N_c , number of spontaneous tracks counted; Area, grain area analyzed; $\Sigma(PQ)$, area-weighted $^{238}\text{U}/^{43}\text{Ca}$, summed over n grains per sample; σ , standard deviation; ξ_{MS} , zeta calibration factor; $P(\chi^2)$, chi-square probability; MTL, mean track length measured from N_r tracks.

Figure 4. Structure of the eastern study transect (ET, location in Figure 3) (simplified from Vergés [1999]), numbered bedrock apatite fission track sample locations (symbol description in Figure 3, inset), and selected permissible time-temperature histories modeled with HeFTy [Ketcham, 2005]. Thermal envelopes show the range of good (dark red) and acceptable (light red) model fits. The white line is the best fit model path. Black boxes are modeling constraints (see methods for details). Horizontal gray lines highlight the partial annealing zone. Dashed vertical lines represent sample component ages determined using RadialPlotter [Vermeesch, 2008] (Appendix C). The track length distribution (red bars) and best fit model (line) are shown in the right panels.

variable kinetic properties but instead most likely reflects multiple age populations, although U loss could also play a role [Brandon *et al.*, 1998; Ketcham *et al.*, 1999]. All samples except one (WT1) have low mean D_{par} values ($<1.6 \mu\text{m}$), suggesting that they contain thermally sensitive calcian-fluorapatite [Carlson *et al.*, 1999; Donelick *et al.*, 2005] with fission tracks that survive heating to $\sim 100^\circ\text{C}$ and partially anneal between $\sim 100^\circ\text{C}$ and 40°C (Figures 4–7) [Reiners and Brandon, 2006].

[15] Below, we summarize our results along the four FTB transects from east to west in the direction of increased shortening [Vergés *et al.*, 2002], followed by the Andorra-Mount Louis pluton in the Axial Zone.

Samples are numbered from south to north, and representative thermal model results are shown in Figures 4–7. We focus on the onset timing of most recent rapid cooling predicted by the statistically “good” thermal model fits [Ketcham, 2005].

4.2. Eastern Transect

[16] We report results from three samples along the eastern transect (ET) (Figure 4). Two samples (ET1 and ET2) are from Late Cretaceous–Paleocene sandstone units of the Garumnian Formation exposed in the Freser antiformal stack. They have pooled ages of 23–20 Ma and moderate mean track lengths (13.8–14.7 μm). Both samples are discordant

Figure 5. Structure of the central eastern transect (CET, location in Figure 3), bedrock apatite fission track sample locations, and associated thermal modeling results. See Figure 4 caption for details.

Figure 6. Structure of the central western transect (CWT, location in Figure 3), bedrock apatite fission track sample locations, and associated thermal modeling results. See Figure 4 caption for details.

and partially reset. Despite their apparent similarities, thermal modeling indicates that sample ET1 cooled rapidly through 100°C between 25 and 20 Ma, whereas sample ET2 produced no model fits. Sample ET3 is from quartz-rich schist exposed on the hanging wall of the Ribes-Camprodon fault, north of the Freser antiformal stack. This sample has poor data quality (8 grains, 35 track lengths), and modeling suggests that rapid cooling began from ~90°C 38–18 Ma.

4.3. Central Eastern Transect

[17] We report two samples from the Cadi thrust sheet to the south and north of the Pedraforca thrust sheet on the central eastern transect (CET) (Figure 5). Sample CET1 is from the middle Eocene Campdevanol Formation on the Vallfogona thrust hanging wall. It has a pooled age of 30 ± 4.6 Ma and a mean track length of 13.0 ± 3.0 μm. CET1 is reset and slowly cooling through PAZ temperatures after 38–24 Ma. The northern sample, CET2, is from the

lower Triassic Buntsandstein Formation [Gradstein *et al.*, 2005] and has a pooled age of 23.7 ± 2.2 Ma and a mean track length of 13.6 μm. This sample has mixed reset ages, and acceptable thermal model fits indicate recent cooling through ~110°C 27–21 Ma.

4.4. Central Western Transect

[18] We present four samples along the central western transect (CWT) (Figure 6). Sample CWT1 is from a conglomerate exposed on the southern limb of the Oliana anticline. The depositional age (33.7–37 Ma) of this sample is well constrained by high-resolution magnetostratigraphy [Burbank *et al.*, 1992a], and it has a pooled age of 43.6 ± 3.6 Ma and a mean track length of 14.2 ± 3.1 μm. Component age analysis shows that this sample is partially reset, and thermal modeling constrains rapid cooling from PAZ temperatures to a narrow window of ~18–16 Ma. Sample CWT2 is from the Cretaceous sandstone of the Garumnian Formation and has a pooled age of 26.7 ± 7.2 Ma and a

Figure 7. Structure of the western transect (WT, location in Figure 3), bedrock apatite fission track sample locations, and associated thermal modeling results. Structurally equivalent to the ECORS profile. See Figure 4 caption for details.

mean track length of $13.3 \pm 2.9 \mu\text{m}$. Modeling suggests this sample cooled from $\sim 90^\circ\text{C}$ between 34 and 15 Ma. Samples CWT3 and CWT4 are from Triassic and Permian sandstones in the footwall of the Boixols back thrust and have similar pooled ages (27.6–28.7 Ma) and mean track lengths (14.1 – $14.2 \mu\text{m}$). Both samples are fully reset. Sample CWT3 cooled below 90°C between 25 and 24 Ma. CWT4 is less constrained but cooled between 36 and 18 Ma.

4.5. Western Transect

[19] We report seven samples on the western transect (WT), which is equivalent to the ECORS profile (Figures 2 and 7). The southernmost sample (WT1) is from the Triassic Keuper Formation exposed in the hanging wall of the Sierras Marginales thrust sheet. This sample is concordant, but because it has poor FT data quality (3 ages, 16 track lengths), permissible model paths define a broad time period (64–32 Ma) for rapid cooling from closure temperatures. Sample WT2 is from the Triassic Keuper Formation exposed on the Montsec thrust sheet. This mixed reset sample has a pooled age of 44.2 ± 8.4 Ma, a mean track length of $14.7 \pm 3.4 \mu\text{m}$, and rapidly cooled 18–13 Ma from $\sim 100^\circ\text{C}$. Sample WT3 is from the Cretaceous Marbre sandstone on the Montsec thrust sheet. It is partially reset and cooled slowly through the PAZ 90–45 Ma. Samples WT4–WT6 form an ~ 500 m elevation transect within the Graus-Tremp Basin on the Montsec thrust sheet. These samples have Paleocene-to-Eocene pooled ages (47.7–59.5 Ma) and mean track lengths that increase up section, suggesting that the lower elevation samples from the profile spent more time in PAZ temperatures. WT4, a mixed reset sample from the Late Cretaceous–Paleocene Garumnian Formation at the profile base, began cooling 61–37 Ma. WT5, a detrital sample from Eocene sandstone in the middle of the profile, began cooling 71–53 Ma. Partially reset sample WT6, from the highest elevation, rapidly cooled from above PAZ temperatures between 52 and 41 Ma. Thermal models of WT5 and WT4 indicate slow recent cooling through PAZ temperatures. Finally, the northernmost sample (WT7) is from a Triassic sandstone in the Noguera Zone and is mixed reset. WT7 cooled rapidly through closure temperature 29–23 Ma.

4.6. Andorra-Mount Louis Pluton

[20] Two Axial Zone granodiorite samples from the Andorra-Mount Louis pluton have similar pooled ages (24.2–22.8 Ma), mean track lengths (14.2 – $13.6 \mu\text{m}$), and D_{par} values ($1.4 \mu\text{m}$). Both samples began to cool rapidly 25–20 Ma.

5. Discussion

5.1. Unroofing, Burial, and Exhumation

[21] The AFT analyses and thermal modeling results reveal that the rocks exposed in the southern Pyrenees have experienced a variety of cooling histories and trajectories through the upper crust. For sedimentary bedrock, AFT-recorded cooling could represent exhumation of the sedimentary rock parcel itself or an inherited (detrital)

exhumation signal from the sediment source region. We use the term unroofing when discussing inherited exhumation and the term exhumation when discussing postdepositional exhumation (Figure 3, inset) and differentiate between the two possibilities by comparing the timing of exhumation with wedge-top deposition (Figure 8). When exhumation precedes or is contemporaneous with deposition, the apatite grains retain an inherited signal of erosion from the Axial Zone source region because burial and associated heating were not sufficient to reset the AFT system (Figures 3–8, red stars). This is the case for four of our samples from the Montsec thrust sheet and all of the Graus-Tremp Basin samples (Figure 8). Granitic cobbles from related syntectonic basins preserved on the Montsec and Boixols thrust sheets and Noguera Zone have a similar inherited unroofing signal (Figures 3 and 8, blue stars) [Beamud et al., 2010; Rahl et al., 2011]. When sediment burial and heating is sufficient to partially or completely reset the AFT system, exhumation is recorded after deposition. The remaining sedimentary bedrock sampled from Cadi thrust sheet, Oliana anticline, Sierras Marginales thrust sheet, and Montsec thrust sheet experienced this type of postdepositional exhumation (Figures 3–8, red encircled stars).

5.2. Variation in Exhumation Magnitudes

[22] The spatial variability in the magnitude of exhumation across the southern Pyrenees is evident from the exposure of both sedimentary bedrock that retains an inherited exhumation signal and that experienced sufficient reexhumation to erase this inherited signal. The present geothermal gradient in the foreland basin is $30^\circ\text{C}/\text{km}$ to $35^\circ\text{C}/\text{km}$, and rapid deposition or sediment blanketing effects could have caused the paleogeothermal gradient to be either lower or higher, respectively [Fillon and van der Beek, 2012]. Therefore, we used $30^\circ\text{C}/\text{km}$ as a paleogeothermal gradient estimate [after Beamud et al., 2010], a closure temperature of 100°C because of the low mean D_{par} in our samples, and a 20°C surface temperature to suggest an average depth to closure of ~ 2.7 km. Estimating that the lower limit of the PAZ extends to 40°C , our data suggest that exhumation magnitudes were greater than ~ 0.6 km in order to expose partially reset samples, but did not exceed ~ 2.7 km in the interior of the Montsec thrust sheet (samples WT3–WT6). Our estimation agrees with previous structural estimates suggesting that 0.5–1 km of material was removed from the central Pyrenees wedge top after major episodes of deformation (post-28 Ma) based on the current wedge geometry, taper, and exposure of syntectonic conglomerates [Meigs and Burbank, 1997].

5.3. Spatiotemporal Patterns of Synorogenic Exhumation

[23] The Late Eocene was a time of abrupt change in the southern Pyrenees. Thrusting and duplex formation rapidly exhumed the south central Axial Zone ~ 35 Ma [Muñoz, 1992; Morris et al., 1998; Fitzgerald et al., 1999; Beaumont et al., 2000; Sinclair et al., 2005]. Ebro foreland basin closure [Burbank et al., 1992a; Vergés and Burbank, 1996] caused drainage system reorientation from dominantly orogen parallel to orogen transverse (Figure 9) [Whitchurch et al., 2011]. On the thin-skinned FTB, only

Figure 8. Simplified cross sections (from Vergés [1999]) along the four studied transects shown in Figures 3–7 synthesizing existing exhumation and sedimentation timing data across the central eastern Spanish Pyrenees [Beamud *et al.*, 2010; Fitzgerald *et al.*, 1999; Gibson *et al.*, 2007; Gunnell *et al.*, 2009; Jolivet *et al.*, 2007; Morris *et al.*, 1998; Rahl *et al.*, 2011; Sinclair *et al.*, 2005]. Overlap between timing of rapid cooling and deposition means the samples retain an inherited exhumation signal from the Axial Zone source region. Onset of rapid cooling from samples in this study is shown from the acceptable (red) and good (dark red) fit thermal model envelopes (e.g., Figures 4–7). Cross sections extend southward from the North Pyrenean Fault to the Ebro Basin. See Figure 3, inset, for sample symbol descriptions. Gray bars highlight the time windows shown in Figure 9.

the syntectonic basin sediments are known to preserve records of pre-35 Ma unroofing (Figures 3–8, white rimmed symbols). AFT data with a pre-35 Ma unroofing signal are clustered toward the north and east in the Axial Zone (Figures 3 and 9a).

[24] The overall distribution of exhumation before 35 Ma reflects patterns of along- and across-strike structural development during Pyrenean orogenesis (Figure 9a). Exhumation started in the eastern Pyrenees in response to oblique convergence driving rock uplift and topographic growth (Figure 1). Exhumation subsequently decreased in the eastern Pyrenees, preserving a record of older exhumation as uplift shifted along strike to the central Pyrenees during the early to middle Eocene

~55–40 Ma [Whitchurch *et al.*, 2011]. While exhumation in the central Pyrenees began before 35 Ma, asymmetric convergence between the pro-wedge and the retro-wedge drove exhumation southward over time [Sinclair *et al.*, 2005], removing most of the pre-35 Ma unroofing signal from the southern Axial Zone. Subsidence within the Graus-Tremp Basin and equivalent syntectonic basins generated enough accommodation space to preserve the earlier erosion signal in detrital apatites from the Axial Zone source region [see also Beamud *et al.*, 2010; Rahl *et al.*, 2011; this study]. In this way, the hinterland and thin-skinned FTB components of the orogenic wedge functioned together to redistribute mass on the surface in response to plate convergence.

Figure 9. Summary of the synorogenic to postorogenic evolution of the central eastern Spanish Pyrenees. The Axial Zone develops diachronously from east to west [Whitchurch *et al.*, 2011] and north to south in the central Pyrenees [Fitzgerald *et al.*, 1999]. Closure of the Ebro Basin ~35 Ma causes syntectonic sediments to bury the southern thrust sheets, forcing the wedge to deform internally to rebuild taper and accommodate shortening. Fluvial excavation follows when the basin connects with the Mediterranean ~13–8.5 Ma [Garcia-Castellanos *et al.*, 2003]. N-S cross sectional view of the structural evolution of the southern orogenic wedge and total shortening estimates for the ECORS transect from Beaumont *et al.* [2000]. For the key to AFT sample location symbols, see Figure 3. Axial Zone antiformal stack thrust sheets: N, Nogueres; O, Orri; R, Rialp. Southern thrust sheets: B, Boixols; M, Montsec; SM, Sierras Marginales.

[25] Significant exhumation occurs during periods of active deformation on the FTB from 35 to 25 Ma (Figure 9b). For example, rapid cooling beginning between 34 and 29 Ma on the Vallfogona thrust hanging wall (Figure 3) is likely an erosional exhumation response to thrust-driven rock uplift because it is coincident with final fault motion 36–30 Ma [Vergés and Burbank, 1996]. Emplacement of the Sierras Marginales thrust sheet occurred during this time period as well [Meigs et al., 1996], but the potential cooling paths modeled for our hanging wall sample fail to distinguish a narrow period of recent exhumation. It is notable that deformation and exhumation occurs across the outboard structures, despite previous findings suggesting that during this period, there is a shift from rapid denudation of the central Axial Zone [Metcalf et al., 2009] to the onset of topographic development in the western Pyrenees [Whitchurch et al., 2011].

[26] Between 25 and 20 Ma, exhumation concentrates at the boundary between the thick-skinned Axial Zone and the thin-skinned southern FTB (Figure 9c). In the eastern Noguères Zone, rapid bedrock cooling began 29–21 Ma. Upper Eocene syntectonic conglomerates onlap thrusts in the Noguères Zone and authigenic illite in a minor overturned proximal thrust has an age of ~56 Ma [Rahl et al., 2011], so it is clear that thrusting ceased prior to exhumation. Samples from the Freser antiformal stack and from north of the Pedraforca thrust sheet record contemporaneous exhumation between 25–20 and 27–21 Ma, respectively. These structures occupy a similar structural position to those in the Noguères Zone at the transition between the Axial Zone and the southern FTB (Figure 8). Previous studies recognized rapid exhumation during this time period in the central Pyrenees and attributed it to antiformal stacking and underplating as a response to sediment loading on the thin-skinned FTB [Sinclair et al., 2005]. Our results demonstrate that rapid exhumation at this time extends at least 100 km farther eastward.

[27] Late-stage (25–20 Ma) synorogenic exhumation in the Pyrenees could result from either exhumation commensurate with tectonic deformation and stable climate, or exhumation coincident with deformation and enhanced by climate change. Antiformal stack formation may have occurred synchronously along strike (Figure 9), coevally or immediately following the last phase of shortening on the southern thrust sheets after ~24.7 Ma [Meigs et al., 1996]. Sediment sourced from the Axial Zone buried and stabilized the outboard FTB structures [Coney et al., 1996], possibly causing deformation to shift toward the hinterland [Jolivet et al., 2007; this study]. This would have allowed the orogenic wedge to rebuild taper through underplating and thrusting, accommodating the final phase of convergence [Beaumont et al., 2000]. Existing thermochronometer data sets and numerical models support this idea, suggesting that rock uplift from antiformal stacking drove rapid exhumation in the south central Pyrenees during the early Miocene [Sinclair et al., 2005; Gibson et al., 2007; Jolivet et al., 2007]. However, the diachronous development of the Axial Zone as well as variations in the along-strike structure and width of the southern FTB call into question a similar sediment loading response in the eastern Pyrenean wedge. There, in contradiction to the predictions of numerical models [Fillon et al., 2012], syntectonic

sediment is preserved at higher elevations than in the central Pyrenees [Coney et al., 1996] and in greater thicknesses in the nearest foreland basin [Sinclair et al., 2005], yet the thrust sheets of the FTB are much narrower than the central Pyrenees.

[28] Alternatively, a regional or global shift to more erosive climate conditions could have initiated a pulse of rapid exhumation in the Axial Zone. Global climate in the Miocene is thought to be relatively warm and arid, peaking with the middle Miocene climatic optimum [Zachos et al., 2001]. However, recent investigations into the major element geochemistry of early Miocene paleosols from the north central Ebro Basin suggest that the regional climate was subhumid to humid, with considerably more precipitation than the earlier arid to semiarid climate during the late Oligocene [Cabrera et al., 2002; Hamer et al., 2007]. Uplift of the western Pyrenees by 30 Ma suggests that significant topography existed across the entire range [Whitchurch et al., 2011]. At present, regional climate modeling shows that orography is the principal control on precipitation patterns in Spain that favor high-altitude precipitation and sheltered lowlands [Sotillo et al., 2003]. The eastern Pyrenees in particular are subjected to high-frequency, convective rainstorms during the summer unlike elsewhere in the Mediterranean basin because of the coupling of orographic cooling and wet air masses produced by the Mediterranean Sea [Callado and Pascual, 2005]. The presence of an evaporative lacustrine environment in the Ebro Basin through the Miocene [García-Castellanos et al., 2003] may have also enhanced orographic precipitation and, thus, erosion along the southern range front. A shift to a stormier, more erosive climate following full along-strike topographic development of the Pyrenees [Whitchurch et al., 2011] poses another explanation for our observation of uniform along-strike exhumation ~25–20 Ma at the transition between the Axial Zone and the thin-skinned southern FTB. We suggest that climate-enhanced feedback between erosion and antiformal stack development during this time period is a more plausible explanation for synchronous exhumation than a taper adjustment response to sediment loading because of the structurally heterogeneous form of the FTB.

5.4. Postorogenic Exhumation

[29] Two samples from the central FTB record the youngest, postorogenic exhumation in the Pyrenees (Figure 9d). Syntectonic conglomerates on the eastern margin of the central thrust sheets indicate Montsec thrust motion and folding of the Oliana anticline ~40–25 Ma [Burbank et al., 1992a; Vergés and Muñoz, 1990]. Middle Eocene syntectonic conglomerates onlapping the Montsec thrust itself west of sample WT2 also bracket deformation older than ~45 Ma [Meigs and Burbank, 1997]. However, model results show that the Oliana anticline and Montsec thrust sheet hanging wall both experienced rapid cooling that began during the Miocene (<23 Ma) (Figures 6 and 7). This suggests a period of erosion that occurs after significant episodes of active shortening. Several postorogenic factors could be responsible for this exhumation. Basin modeling [García-Castellanos et al., 2003], seismic imaging of the Ebro delta [Urgeles et al., 2010], and paleotopographic signals within thermochronometer data [Fillon and van der Beek, 2012] imply that capture of the

paleo-Ebro River by the Mediterranean Sea (at ~13–8.5 Ma) caused regional base level lowering and triggered excavation of the sediment blanket that buried both structures. Bedrock on the eastern Montsec thrust sheet began rapidly exhuming between 21 and 10 Ma and may corroborate these conclusions with direct evidence from the excavated region. However, the Oliana anticline exhumation began 18–16 Ma, well before regional base level lowering. Thermokinematic modeling of thermochronometer data from the Axial Zone ruled out climate change as an important factor in the postorogenic evolution of the Pyrenees [Fillon and van der Beek, 2012], yet our findings suggest that increased storminess may have enhanced erosion along the southern range front and could have been an impetus for exhumation. Finally, late Oligocene to early Miocene tilting near the southeast margin of the Ebro Basin, caused by lithospheric extension and thinning, coeval with opening of the Valencia Trough, may also enhance exhumation along large areas paralleling the range [Lewis et al., 2000]. More detailed sampling and combined AFT and (U-Th)/He dating in apatite (closure temperature ~70°C [Farley, 2000]) could test if this early Miocene exhumation event is robust.

6. Conclusions

[30] This study documents Cenozoic exhumation across the fold-thrust belt of the southern Pyrenees with 18 new bedrock apatite fission track (AFT) samples. Comparison between the AFT cooling histories and sample deposition timing differentiates between sedimentary bedrock retaining an inherited unroofing signal from the range interior (Axial Zone) and recording exhumation after incorporation into the Pyrenean thrust sheets. Graus-Tremp syntectonic piggyback basin sediments record Axial Zone exhumation ~70–40 Ma, consistent with detrital exhumation signals present in AFT samples from other nearby piggyback basins [Beamud et al., 2010; Rahl et al., 2011]. AFT samples from other southern Pyrenean structures are partially or fully reset, recording FTB exhumation in excess of ~3 km. The Cadi thrust sheet exhumed between 34 and 29 Ma, contemporaneous with its final phase of deformation 36–30 Ma. Reset cooling ages from the Oliana anticline and southern edge of the Montsec thrust sheet reflect post-tectonic exhumation 20–10 Ma, perhaps in response to fluvial incision from abrupt foreland base level change and/or regional tilting coeval with Valencia Trough extension. Basement rocks at the transition from the cover thrust sheets to the crustal-scale duplex of the Axial Zone record exhumation along the central to eastern FTB at 25–20 Ma. This rapid exhumation may result from antiformal stack development in response to either (a) sediment loading, previously noted in the Noguères Zone of the central Pyrenees [Sinclair et al., 2005], and/or (b) a shift to more erosive climate conditions in the early Miocene. We favor the latter explanation because of the high degree of structural heterogeneity between the central and the eastern FTB.

Appendix A: Analytical Procedures

[31] All AFT analyses were performed at Apatite to Zircon, Inc. Samples were crushed and sieved to sand-sized particles, then apatite grains were isolated using standard

gravimetric and magnetic mineral separation techniques [Donelick et al., 2005]. Grains were mounted in epoxy resin, cured at 90°C for 1 h, and then polished to expose the internal surfaces of the grains. Mounts were then immersed in 5.5 M HNO₃ for 20 s at 21°C to reveal natural fission tracks that intersected the polished grain surface. Samples were irradiated with ²⁵²Cf to facilitate fission track length measurement [Donelick and Miller, 1991]. Fission track lengths and crystallographic orientation were measured using a digitizing tablet interfaced with a computer at 2000x magnification under unpolarized light. Only natural, horizontal, confined fission tracks with clearly visible ends were measured. The mean D_{par} was determined for each grain by measuring up to four etch pit diameters. Fission track grain ages were calculated using a modified form of the radioactive decay equation and the ratio of the number of fission tracks present to the concentration of ²³⁸U measured using laser ablation inductively coupled plasma–mass spectrometry (LA-ICP-MS) [Donelick et al., 2005; Hasebe et al., 2004]. This decay equation includes a zeta calibration factor, which was determined by analyzing Durango and Fish Canyon apatite (30.6 ± 0.3 Ma from Cerro de Mercado, Durango, Mexico, and 27.9 ± 0.7 Ma from the San Juan Mountains, CO, U.S.) at the beginning and end of each LA-ICP-MS session for its U/Ca ratio.

Appendix B: Thermal Modeling

[32] We performed inverse thermal modeling of our AFT data using the software HeFTy version 1.7.4 [Ketcham, 2005]. We used the multikinetic annealing model of Ketcham et al. [1999] and included D_{par} values. We projected the track lengths to the crystallographic *c* axis [Donelick et al., 1999] and included ²⁵²Cf irradiation.

[33] Open-ended modeling was performed with all sample data as one kinetic population with a starting temperature of (1) 200°C at a time that is 50 Ma older than deposition and (2) 20°C at present (after Barnes et al. [2006, 2008]). We ran this model to assess how distinct the recent cooling history is without bias from user-defined constraints. We then simulated source region exhumation and subsequent incorporation of apatite grains into sedimentary bedrock samples by forcing the time-temperature paths to travel from depth (200°C) 50 My before sample deposition to surficial conditions (10°C–30°C) during deposition, then to reheat (50°C–180°C) between deposition and 1 Ma before returning to the surface (20°C) by the present. By comparing the two models, we could confirm that the refined model better constrains the same cooling event as the open-ended model, ensuring that we imposed realistic modeling parameters that did not remove the intrinsic cooling history. We designated each time-temperature cooling path segment to be monotonic for simplicity. For maximum flexibility [see Ketcham, 2005], thermal history segments between each imposed *t*-*T* constraints were designated as episodic style, random spacing, with the largest number of vertices (halved 5 times). We enforced a maximum slope of 40°C/My, assuming that cooling rates on the FTB structures will be similar to or lower than those in the Axial Zone [e.g., Metcalf et al., 2009]. We ran each inversion with a Monte Carlo search and 50,000 attempted paths.

ET1

Central value = 25.3 +/- 6 (2se)

Dispersion = 60 %

ET2

Central value = 26.2 +/- 6.59 (2se)

Dispersion = 37.7 %

Central value = 34.8 +/- 6.45 (2se)

Dispersion = 27.5 %

CET2

Central value = 24.1 +/- 2.62 (2se)

Dispersion = 20.6 %

CWT1

Central value = 45.9 +/- 10.3 (2se)

Dispersion = 61.0 %

CWT3

Central value = 28.4 +/- 2.22 (2se)

Dispersion = 16.3 %

Figure C1. Central and component ages determined for our samples using RadialPlotter [Vermeesch, 2009]. The radial plot displays the single grain age as the degree of rotation along the curved axis, and the analytical uncertainty as the distance from the curved axis. Component ages are shown as the radiating lines.

Appendix C: RadialPlotter Results

[34] We determined AFT central and component ages with the Java application RadialPlotter version 4.4 [Vermeesch, 2009]. We generated radial plots for our LA-ICP-MS data from single grain ages using the Other input, not the Fission Tracks input that is designed for the external detector method. We linearly transformed our data and used an automated mixing model to generate the radial plots shown in Figure C1.

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