Coupling sequential restoration of balanced cross sections and low-temperature thermochronometry: The case study of the Western Carpathians

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ABSTRACT

In this paper, a new approach is applied to test a proposed scenario for the tectonic evolution of the Western Carpathian fold-and-thrust belt–foreland system. A N-S balanced section was constructed across the fold-and-thrust belt, from the Polish foreland to the Slovakia hinterland domain. Its sequential restoration allows us to delineate the tectonic evolution and to predict the cooling history along the section. In addition, the response of low-temperature thermochronometers (apatite fission-track and apatite [U-Th]/He) to the changes in the fold-and-thrust belt geometry produced by fault activity and topography evolution are tested. The effective integration of structural and thermochronometric methods provides, for the first time, a high-resolution thermo-kinematic model of the Western Carpathians from the Early Cretaceous onset of shortening to the present day. The interplay between thick- and thin-skinned thrusting exerts a discernible effect on the distribution of cooling ages along the profile. Our analysis unravels cooling of the Outer Carpathians since ca. 22 Ma. The combination of thrust-related hanging-wall uplift and erosion is interpreted as the dominant exhumation mechanism for the outer portion of the orogen. Younger cooling ages (13–4 Ma) obtained for the Inner Carpathian domain are mainly associated with a later, localized uplift, partly controlled by extensional faulting. These results, which help unravel the response of low-temperature thermochronometers to the sequence of tectonic events and topographic changes, allow us to constrain the tectonic scenario that best honors all available data.

INTRODUCTION

The sequential restoration of balanced cross sections is a powerful tool for calculating the amount of shortening, the slip rate, and the depth of the décollement surface in contractional and extensional regimes. It also is the only way to represent the evolution of the tectonic structures in time. Nevertheless, the timing of deformation cannot be inferred from kinematic models alone, particularly in cases where syntectonic deposits are not preserved and the thrust fronts are erosional. One solution is to constrain deformation events by integrating two-dimensional (2-D) kinematic modeling with low-temperature (low-T) thermochronometers. The extrapolation of the beginning of deformation events is valid when the cooling ages are strictly associated with the onset of deformation (e.g., Stockli et al., 2000; Ehlers and Farley, 2003; Stockli, 2005). In addition, sequential restoration can provide an estimate of the maximum burial experienced by different successions cropping out along the section. However, burial depth estimates need to be constrained by paleothermal indicators, such as vitrinite reflectance. This latter indicator, together with apatite fission-track (AFT) and apatite U-Th/He (AHe) data, is the main constraint for tracing burial and exhumation histories. The thermal evolution associated with thrusting has been modeled for simple tectonic structures (such as fault bend folds and hinterland dipping duplexes) by evaluating the role of topography, inclination of the thrust ramp, and amount of displacement on the predicted cooling age profile along a geological transect (Huerta and Rodgers, 2006; Lock and Willet, 2008). Here, we want to contextualize low-T thermochronometric data (AFT and AHe) in the complex kinematic restoration of the Western Carpathian fold-and-thrust belt, in which Early Cretaceous thick-skinned thrusting (Sândulescu, 1988) was followed by thin-skinned deformation of the most external part from the Eocene to the middle Miocene (Mahel’, 1974; Cieszkowski et al., 1985; Sândulescu, 1988). In this paper, we are not addressing in detail the issues concerning tectonic styles and deformation processes controlling the Carpathian orogen; rather, the main aim here is to discuss the methodology applied to a case study and, in particular, how low-T thermochronometry can be a successful tool for testing the viability of a given geological scenario. This approach allows us to convert the temperatures to depths and refine the proposed geological scenario. The temperatures are obtained with thermal modeling for several samples along the chosen transect. The depths are calculated for different times during the tectonic evolution. We use FETKIN (Almendral et al., 2014), a software dedicated to the forward modeling of thermochronometric ages. For a given kinematic restoration, integrated with thermal parameters (such as paleogeothermal gradient, thermal conductivity, and specific heat capacity), FETKIN calculates low-temperature thermochronometric ages for different thermochronometers along the present-day topographic profile of a geological cross section. The sequential restoration can be tested by com-
paring the predicted thermochronometric ages with those measured on samples collected along the topographic profile. FETKIN provides the evolution of the isotherms through time, allowing one to estimate the maximum temperatures experienced by the outcropping successions. This study follows the workflow already applied by Mora et al. (2014) on the Eastern Cordillera of Colombia. In particular, we want to evaluate the influence of fault geometry and activity, and of topography evolution, on expected thermochronometric age patterns. The final result is the best fitting between thermo-kinematic model and real data.

GEological setting

The Western Carpathians are the eastern continuation of the Alpine orogenic system (Fig. 1A), which originated from the collision between the Adriatic and Euro-Asiatic plates during the Late Cretaceous to the Miocene (Froitzheim et al., 2008). The subduction and subsequent closure of the southern branch of the Alpine Tethys (sensu Schmid et al., 2008) caused the deformation of the innermost deposits and, during the Paleogene, the imbrication of the outer successions and their relative emplacement on top of the European Platform. The complex architecture of the Carpathians can be simplified by dividing them into two tectonic domains: the Outer Carpathians and the Inner Carpathians (Fig. 1B; Książkiewicz, Poland-Slovakia)

Figure 1. (A) Geographic map showing the location of the study area. State abbreviations: AT—Austria, BA—Bosnia and Herzegovina, BG—Bulgaria, BY—Belarus, CH—Switzerland, CZ—Czech Republic, FR—France, GR—Greece, HR—Croatia, HU—Hungary, IE—Ireland, IT—Italy, NL—Netherlands, PL—Poland, PT—Portugal, RO—Romania, RS—Serbia, SP—Spain, TR—Turkey, UA—Ukraine, UK—United Kingdom. (B) Tectonic map of the Polish and Slovakian Carpathians, showing location of the modeled profile and samples used for the validation of the thermo-kinematic model.
The Outer Carpathians consist of a fold-and-thrust belt made of Upper Jurassic to Lower Miocene deposits (Książkiewicz, 1962, 1977; Bieda et al., 1963; Mahel’ and Buday, 1968; Koszarski and Słazka, 1976). It is formed by several thrust sheets (Magura, Dukla, Silesian, Subsilesian, Skole units) made of siliciclastic deposits with variable sandy/shale ratio (Fig. 2). The Inner Carpathians are formed by thick-skinned thrust sheets made of Variscan basement, with its Mesozoic cover piled up during the Austroalpine orogeny. The Mesozoic nappes are partially buried under the Paleogene deposits of the Central Carpathian Paleogene Basin (Fig. 1B). The Inner Carpathians and Outer Carpathians are separated by the Pieniny Klippen belt, a narrow belt of sheared Mesozoic to Eocene rocks, assumed to be the suture of the Vahicum Ocean (Mahel’, 1981). A debate exists about the evolution of the Carpathians and the origin of the Vahicum suture. Many authors (e.g., Birkenmajer, 1960, 1986; Picha et al., 2006; Birkenmajer, 2008) suggest the occurrence of oceanic crust between the Inner Carpathian and Outer Carpathian domains. More recent papers (Jurcewicz, 2005; Malinowski et al., 2013; Roca et al., 1995) cast doubt about the presence of oceanic crust between the Inner Carpathians and Outer Carpathians, rather suggesting that thinned continental crust floored the Pieniny Klippen belt. The stratigraphy of the Outer Carpathians, Inner Carpathians, and Pieniny Klippen belt is summarized in Figure 3, in which an attempt is made to correlate the deposits belonging to different domains.

**Deformation History**

For the purpose of this work, we propose a conservative scenario in which the Outer and the Inner Carpathian successions were deposited in sedimentary domains floored by thinned continental crust (Fig. 4A). As we cannot quantify the width of the oceanic basin postulated to be represented by the Pieniny Klippen belt, which could have been very narrow or even nonexisting, our undeformed stage includes an undefined original separation between the Inner and Outer Carpathian paleogeographic realms. Early Cretaceous shortening involved the southern part of the sedimentary basin (Voigt and Wagreich et al., 2008), producing the reactivation of Mesozoic normal faults as reverse faults (Fig. 4B). Thick-skinned thrusting propagated northward during the Late Cretaceous (Maluski et al., 1993) to Paleocene.
Figure 3. Correlation chart for the main tectono-stratigraphic units of the Western Carpathians. The successions are not represented with true thickness. Lithological descriptions are based on Gross et al. (1993); Samuel and Fusan (1992); Ślączka et al. (2006); Uchman (2004); and Voigt et al. (2008). PKB—Pieniny Klippen belt.
Figure 4. Two-dimensional (2-D) forward modeling of the balanced cross section from the Early Cretaceous to the present day, performed using Move package, developed by Midland Valley Ltd. Displacement values used as input data come from the sequential restoration. Vertical simple shear and fault-parallel flow algorithms were used to restore/forward model the normal faults and the reverse faults, respectively. The flexural slip algorithm was applied to simulate the flexure of the lower plate. Shortening rate ($v_s$) and extensional rate ($v_{ext}$) are indicated for each time step of the sequential restoration. For unit colors see legend in Figure 2.
During the Oligocene–middle Miocene, ok et al., Cretaceous shaly deposits (Nemčok et al., 2006; Sotáková et al., 2001). Imbrication within the Inner Carpathians belt proceeded during the Late Cretaceous–Paleocene. The flexural subsidence affecting the European Platform as a result of the Inner Carpathians emplacement produced a large foreland basin that was filled in its proximal part with olistoliths and olistostromes produced by reworking of Mesozoic successions. The olistoliths, the southern provenance of which has been delineated by Roca et al. (1995), originated from the subaerial exposure and subsequent erosion of the Inner Carpathians successions (Fig. 4C). The Mesozoic olistoliths, contained in an Upper Cretaceous–Paleocene matrix, are the main components of the so-called Pieniny wildflysch (Plašienka and Mikuš, 2010). Although the olistostromes are said to be associated with the tectonic emplacement of oceanic slivers by, for example, Picha et al. (2006), other authors infer the occurrence of continental crust beneath the Pieniny wildflysch (as was already suggested by Jurewicz, 2005) and a sedimentary rather than tectonic origin of the mélangé forming the Pieniny Klippen belt. The Pieniny wildflysch overthrust the Outer Carpathians successions during the middle Eocene (Fig. 4D; Bromowicz, 1999). This tectonic episode marked the end of thick-skinned shortening and the onset of thin-skinned thrusting involving the Outer Carpathians domain. During the Oligocene, thrusting propagated northward, detaching the Outer Carpathians successions along the Upper Cretaceous shaly deposits (Nemčok et al., 1999). During the Oligocene–middle Miocene, thrusting and subsequent erosion of the Outer Carpathians successions occurred contemporaneously with sedimentation of the Podhale wedge-top basin deposits south of the Pieniny Klippen belt. During the middle–late Miocene, shortening affected the basement, inverting the inherited Mesozoic normal faults cutting through the lower plate (Oszczypko et al., 2006). The stacking of thrust sheets increased the gravitational instability of the Carpathian accretionary wedge, leading to the nucleation of normal faults, some of them reactivating reverse structures (Fig. 4F).

METHODS

The balanced cross section has been integrated with paleothermal and low-temperature thermochronometric data (Fig. 2). AFT, AHe, and zircon U-Th/He (ZHHe) ages (Andreucci et al., 2013; Anczkiewicz et al., 2013; Králiková et al., 2014, and references therein) are the main constraints for tracing the thermal history of each sample projected onto our cross section. Thermal models were performed using HeFTy (Ketcham, 2005) and, together with the illite-smectite (Šrodoň et al., 2006) and vitrinite reflectance (R; Andreucci, 2013; Wagner, 2011) data, allowed us to infer the amount of maximum burial for each sample. Temperature values were converted into burial depths using a constant geothermal gradient of 18 °C/km (Fig. 5; Andreucci et al., 2013; Hurai et al., 2006; Schwierzewska, 2005). Once we constructed and sequentially restored the balanced cross section, we chose 10 steps of the restoration to be exported as ASCII files into FETKIN. The following main steps were considered: (1) 145 Ma as the initial undeformed stage before the Early Alpine orogeny (Froitzheim et al., 2008); (2) 70 Ma as the end of the imbrication of the Inner Carpathian Mesozoic cover (3) 28 Ma as the time of Pieniny Klippen belt thrusting on top of the Outer Carpathian Oligocene deposits; (4) 15 Ma as the end of thrusting (Nemčok et al., 2006); and (5) 0 Ma as the final setting. Further intermediate steps were chosen at 56, 38, 23, 22, and 20 Ma. The age was assigned to each step taking into account the youngest deposits preserved at the footwall of the more recent fault. The sequential restoration was further calibrated by processing each reconstructed step with FETKIN (Almond et al., 2014). This software solves the transient advection-diffusion equation in two dimensions (Carslaw and Jaeger, 1986). Starting from the velocity field generated from the kinematic restoration, FETKIN calculates the temperature distribution honoring the structural setting at each time step. Applying an iterative workflow, we changed erosion rate, paleotopography, and fault geometry in order to achieve the best fit between modeled and measured data. No changes of the thermal parameters and timing of the deformation were applied, as these parameters are reasonably well constrained in the literature (Andreucci et al., 2013, and references therein). Once we assigned the topographic profiles, geothermal gradient (18°/km; Andreucci et al., 2013; Schwierzewska, 2005), thermal conductivity (2.2 W/m·°C), density (2.7 g/cm³), and specific heat (1000 kcal/kg·°C) to the horizons, we defined the top and bottom boundary conditions. The mean sea-level temperature was set at 24 °C. The depth of the lower boundary was located at 44 km below sea level (bsl), and the assigned temperature was 774 °C. The final result is the modeling of the isotherms at each time step of the kinematic restoration. The paleogeothermal gradient is the main thermal variable that could affect the validity of our thermo-kinematic model. We assume a relatively low paleogeothermal gradient in the light of published results on samples from this thrust

![Figure 5. Balanced cross section, showing sample location and maximum burial obtained by thermal modeling of apatite fission track (AFT) and U-Th/He (AHe) data (Andreucci et al., 2013; Králiková et al., 2014) integrated with illite-smectite values and vitrinite reflectance data (Andreucci, 2013; Šrodoň et al., 2006; Wagner, 2011). For unit colors see legend in Figure 2.](https://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/4/367/3039785/367.pdf)
path highlights the validity of the methods. For unit colors see legend in Figure 2.

Figure 6. Comparison between the temperature-time (t-T) path resulting from HeFTy inverse modeling best-fit path (black line) and the thermal path modeled with FETKIN (blue line) for the same location. The represented thermal histories are those best-constrained by AFT, AHe, and vitrinite reflectance. The red boxes represent the depositional constraints, and the red bars correspond to the end of thrusting. The good match between the two paths highlights the validity of the methods. For unit colors see legend in Figure 2.
files (width and elevation of the relief) in order to achieve the best fit between the predicted AFT and AHe ages and the measured ones.

TESTING TWO DIFFERENT CASE HISTORIES

We compared two different scenarios (case A and case B; Fig. 7) differing in the paleotopographic evolution for the last 20 m.y. and the geometry of the Sub-Tatra normal fault, which represents the most relevant extensional structure in terms of displacement. The implications of the two different cases are discussed in terms of topography development, subsidence, and fault geometry. The final result is a thermo-kineomatic model in which the selected geological scenario has been integrated with an admissible thermal history.

Case A

In case A (Fig. 7A), we simulated synthrusting erosion affecting the innermost part of the Outer Carpathians during early Oligocene to early Miocene times. The eroded deposits were carried downslope to the outer zones, accumulating on top of the Magura thrust hanging-wall block during the early Miocene. The inferred mean erosion rate is 1.10 mm/yr, resulting from erosion enhanced by thrusting. The northward propagation of thrusting and the related flexure of the lower plate, caused by the tectonic load of the accretionary wedge, produced the accommodation space later filled by middle Miocene deposits, while the highest part of the chain was still affected by erosion. After the end of thrusting at ca. 15 Ma (Nemčok et al., 2006), several normal faults developed as a response to the internal gravitational instability of the orogen. In this first case, the Sub-Tatra fault, controlling the northern boundary of the Liptov Basin, has been interpreted as a listric normal fault detaching along the sole thrust of the orogenic system (at ~7 km bsl in the area).
Case B

In case B (Fig. 7B), synthrusting erosion is assumed to have affected both the Magura thrust 2 and Magura thrust hanging-wall blocks earlier than the previous case. The deposition of the early Miocene synthrusting deposits occurred only in the outermost part of the wedge, in front of the high-displacement Magura thrust. In this scenario, early Oligocene–early Miocene erosion affecting the Magura thrust 2 hanging-wall block would have occurred at a lower mean erosion rate (0.64 mm/yr). Subsidence of the European Platform is inferred to have slowed down during the early–middle Miocene, causing the erosion of the Oligocene deposits of the Magura thrust hanging-wall block and the deposition of Miocene successions in the frontal part of the section and in the Podhale Basin located in the inner part (Ludwinia̜k, 2010, and references therein).

The late Oligocene–early Miocene mean erosion rate calculated for the Magura thrust is almost the same as for the previous case (0.42 mm/yr compared to 0.48 mm/yr in case A). Middle Miocene deepening of the Magura front and its exhumation in more recent time (8 Ma) resulted in overall lower erosion rates. The inferred Langhian to present-day mean erosion rates range from 0.12 mm/yr (case A) to 0.07 mm/yr (case B). Furthermore, in case B, the Sub-Tatra fault has been interpreted as a deep basement structure cutting through the lower plate and detaching at a depth of ~17 km bsl.

**DISCUSSION**

Comparing the AFT age profiles resulting from the two previously described scenarios (red diamonds for case A and blue circles for case B; Figs. 7C and 7D), some important differences can be highlighted. The change in topography in the hanging wall of the Magura thrust, combined with subsidence (more pronounced for case A), reduces the cooling ages in this specific area. Partially reset AFT ages predicted for case A are considerably younger than those predicted for case B, in some cases by more than 80 m.y. No major changes in the predicted AFT ages occur in the hanging wall of the Magura thrust 2, although topography evolution is different for the two cases. In the hanging wall of the Sub-Tatra fault the variation in the predicted AFT ages is mainly due to the change of the depth to detachment; the shallower is the detachment (case A), the older are the AFT ages predicted along the hanging wall. The change in the geometry of the fault appears to have a more pronounced influence on the AHe cooling ages (Fig. 7D). For a deeper detachment (case B, blue circles), the AHe ages are considerably younger. No major changes are recorded along the AHe profile for the Outer Carpathians, except for the Magura thrust hanging wall, where the lower-relief modeled topography combine with the higher tectonic subsidence for case A to produce a younging of the cooling ages in this sector.

Case B represents the scenario best approximating the measured thermochronometric data. The resulting thermo-kinematic model provides a complete framework of AFT (Fig. 8) and AHe (Fig. 9) cooling ages along the studied transect. The predicted AFT age profile (Fig. 8) highlights two areas that experienced temperatures not exceeding ~120 °C. These are: (1) the foreland basin together with the outermost successions of the Outer Carpathians fold-and-thrust belt (at a distance from 0 to 24 km along the section), and (2) the Oligocene deposits of the Liptov Basin (from 100 to 108 km). A partial reset of the AFT cooling ages is predicted for the central part of the Podhale Basin (from 70 to 80 km), while totally reset AFT ages from the limbs of the Podhale syncline indicate cooling through the partial annealing zone (e.g., Wagner, 1979b) during the middle–late Miocene. Cooling ages predicted by FETKIN match well with the AFT ages obtained by Anczkiewicz et al. (2013). Furthermore, this thermo-kinematic model is consistent with the trend of increasing temperature toward the back limb of the...
Podhale syncline suggested by Środoń et al. (2006) and Wagner (2011). The partial mismatch between observed and predicted AFT ages in the central part of this basin is probably due to an underestimation of the thickness of the eroded Oligocene–Miocene successions, as the data projected onto the section are from partially reset samples located farther to the west (Anczuckiewicz et al., 2013; Botor et al., 2006, 2011).

AFT cooling ages ranging between 14 and 22 Ma are predicted for the Outer Carpathians sector between 36 and 68 km, in line with the cooling ages observed by Andreucci et al. (2013). Here, the predicted age profile is controlled by thrust-related uplift and erosion, with the oldest ages being located in the hanging wall of the Magura thrust 2 fault. The youngest ages are from the footwall of the normal fault bounding the Mszana Dolna tectonic window to south, whereas cooling ages become older moving to the Magura thrust front. Younger exhumation ages are predicted for the crystalline basement of the Inner Carpathians, this being consistent with the middle–late Miocene cooling documented by Burchart (1972), Kráľ (1977), and Králíková et al. (2014). In the Inner Carpathians, the cooling age profile is locally controlled by normal faulting. The youngest ages are located at the footwall of the Sub-Tatra fault (Gross, 1973; Gross et al., 1980) and are mainly associated with the coeval middle–late Miocene regional uplift controlling the cooling of the whole Inner Carpathians domain.

The AHe cooling age profile (Fig. 9) shows an overall pattern consistent with the AFT profile. The foreland and the outer part of the Carpathian fold-and-thrust belt (from 0 to 24 km) and the northern part of the Liptov Basin (from 100 to 106 km) are not reset. As with the AFT profile, a narrow zone in the frontal part of the Magura thrust is also not reset (from 32 to 35 km), whereas in its hanging wall, the AHe ages are all reset. This means that this entire zone experienced temperatures higher than ~60 °C.

A thermochronometer sensitive to very low temperatures, such as the AHe, is particularly effective at unraveling the influence of thrusting on cooling ages (Ehlers and Farley, 2003). Therefore, in the Outer Carpathians, the relationship between the age profile and the development of the structures forming the fold-and-thrust belt is clearer than in the AFT profile. The age gaps recorded along the thermochronometric age profile are indicative of high-displacement thrust faults. Meaningful examples can be found: (1) at a distance of 26 km along the cross section, where the younger age (22 Ma) indicates that the hanging wall of the Magura thrust moved, at least, up to that point; (2) at 51 km, where the younger ages mark the location of the Magura thrust 2 footwall ramp, with the oldest ages located in its hanging wall; and (3) at 100 km, where a significant age gap is associated with the Sub-Tatra fault, with its 10 km of dip-slip displacement. According to our model best fitting all available thermochronometric data, totally reset AHe cooling ages range between 8 and 20 km, with the youngest being located in the Tatra Mountains region (Fig. 9).

The validity of this model is additionally supported by comparing the t-T paths resulting from HeFTy and the ones obtained from FETKIN for selected samples on the topographic line. The FETKIN t-T paths are more detailed because they are strictly connected with the sequential restoration. Although it does match the general pattern modeled by HeFTy fairly well, in some cases, the maximum burial is underestimated (i.e., PL 27, PL 37, PL 25). The exhumation of PL 25 and PL 24 starts before the end of thrusting, and it is older than the one computed by HeFTy. These differences influence the age of the first stage of cooling, making them older, but they provide a good match with the measured data.

CONCLUSIONS

This work represents a successful integration of thermal modeling and kinematic restoration applied to the case study of the Carpathian fold-and-thrust belt–foreland system. FETKIN has been a key tool in predicting thermochronometric ages and calculating t-T paths along a sequentially restored balanced cross section. It allowed us to select, validate, and refine the most suitable structural model for which predicted thermochronometric ages match quite well with the real data. The obtained thermo-kinematic model implies that exhumation in the
Outer Western Carpathians was coeval with the abrupt increase of shortening rate during early Oligocene to early Miocene times, consistent with thrusting being the driving mechanism in the exhumation of the Outer Carpathians, as suggested by Andrecucci et al. (2013). Our model is also in agreement with the in-sequence thrust propagation across the Outer Carpathians sedimentary basin suggested by Roca et al. (1995). Furthermore, the z-T paths traced for some samples along the section (Fig. 6) indicate that the Outer Carpathians did not experience temperatures higher than ~120 °C during their thermal history. Slightly higher temperatures are recorded in the crystalline basement cropping out in the Inner Carpathian region, in which cooling during the middle–late Miocene was locally controlled by the dip-slip extensional offset along the Sub-Tatra fault. The comparison between two possible tectonic scenarios points out that the development of tectonic structures—rather than the evolution of the topographic relief through time—exerts the main control on the modeled thermochronometric ages. The case of the Sub-Tatra fault, which is interpreted in this study as a deeply rooted basement structure rather than a shallow-detaching listric fault, reveals that thermochronometry can be a useful tool in constraining the geometry of structures at depth. Our results suggest that integration of AFT analyses with a thermochronometer sensitive to very low temperatures—such as the AHe system—is particularly effective in enhancing structural interpretation and restoration in fold-and-thrust belts.

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