**Crustal deformation and exhumation within the India-Eurasia oblique convergence zone: new insights from the Ailao Shan–Red River shear zone**

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**Key Points:**

* The Diancangshan massif forms a region-scale antiform, bordered by two shear zones.
* The horizontal shearing layers can be first formed, locally domed, and then cooled before 36 Ma.
* Exhumation of middle-crust rocks during the late Oligocene and Miocene was related to the shear zone.
* The change of kinematic reversal and cooling along the shear zone since ~5 Ma imply a tectonic transformation.

**Abstract**

During the collision of India and Eurasia, regional-scale strike-slip shear zones play a key role to accommodate lateral extrusion of blocks, block rotation and vertical exhumation of metamorphic rocks as presented by deformation on the Ailao Shan-Red River shear zone (ARSZ) in the Eastern Himalayan Syntaxis region and western Yunnan. We report structural, mica Ar/Ar, apatite fission-track (AFT), and apatite (U–Th)/He (AHe) data from the Diancangshan massif in the middle segment of the ARSZ. These structural data reveal that the massif forms a region-scale antiform, bordered by two branches of the ARSZ along its eastern and western margins. Structural evidences for partial melting in the horizontal mylonites in the gneiss core document that the gneiss experienced a horizontal shear deformation in the middle crust. Muscovite Ar/Ar ages of 36 to 29 Ma from the core represent cooling ages. Muscovite Ar/Ar ages of 25 and 17 Ma from greenschist-facies mylonites along the western and southern shear zones are interpreted as recording deformation in the ARSZ. The AFT ages, ranging from 15 to 5 Ma, presenting a quiescent gap with a slow cooling/exhumation in the massif. AHe results suggest that a rapid cooling and final exhumation episode of the massif can have started before 3.2 Ma, or likely around 5 Ma, and continuing to present. The high-temperature horizontal shearing layers of the core can be first formed across the Indochina Block, locally antiformed along the tectonic boundaries, and then cooled through the mica Ar–Ar closure temperature during Eocene or early Oligocene, subsequently reworked and further exhumed by sinistral strike-slip movement along the ARSZ during the early Oligocene (ca. 29 Ma) and lasting until ca. 17 Ma, and then final exhumation of the massif by dextral normal faulting on the Weixi-Qiaohou and Red River faults along the limbs of the ARSZ since ~5 Ma. The formation of the antiform can indicate local crustal thickening in an early transpressional setting corresponding to India–Asia convergence. Large-scale sinistral ductile shear along the ARSZ in the shallow crust accommodates lateral extrusion of the Indochina Block, and further contributes to the vertical exhumation of the metamorphic massif from the late Oligocene to the middle Miocene. Furthermore, the change of kinematic reversal and associated cooling episodes along the ARSZ since the middle Miocene or early Pliocene can imply a tectonic transfer from strain localization along the major tectonic boundaries to continuous deformation corresponding to plateau growth and expansion.

1. Introduction

During the collision of India and Eurasia, the lithosphere underwent extensive intracontinental deformation, notably at the southeastern end of the Eastern Himalayan Syntaxis region (Figure 1a). In the India–Asia oblique collision zone, the deformation manifests as the regional-scale, strike-slip shearing of geological boundaries (e.g., Leloup et al., 1993, 1995; Lacassin et al., 1996; Taylor et al., 2003; Searle, 2006; Wang et al., 2006; Bracciali et al., 2016; Pellegrino et al., 2018; Liu et al., 2020) and extensive exhumation of middle-crust rocks (e.g., Jolivet et al., 2001; Leloup et al., 2001; Anczkiewicz et al., 2007; Liu-Zeng et al., 2008; Xu et al., 2015; Yang et al., 2016; Wang et al., 2017; Wang et al., 2020; Yan et al., 2021). Along the eastern and southeastern margins of the Tibetan Plateau, strike-slip movement and exhumation processes coexist, as is evident from the exposed high-grade metamorphic massifs and the presence of regional-scale strike-slip shear zones (Figure 1b) such as the Ailao Shan–Red River (ARSZ) Shear Zone, Chongshan Shear Zone, and Gaoligong Shear Zone (e.g., Peltzer and Tapponnier, 1988; Tapponnier et al., 1990; Leloup et al., 1993; Leloup and Kienast, 1993; Lee et al., 2003; Searle, 2006; Wang et al., 2006; Liu et al., 2012). The regional-scale, NW-striking ARSZ extends from the Eastern Himalayan Syntaxis and southwestern Yunnan to northern Vietnam (Figure 1b). The shear zone forms the tectonic boundary between the South China and Indochina blocks (Figure 1a–c; Leloup et al., 1993, 1995; Zhong, 2000). It is characterized by exhumation of metamorphic antiforms, sinistral ductile strike-slip shearing, and significant displacements (Figure 1a, b; e.g., Tapponnier et al., 1982; Leloup et al., 1995; Wang and Burchfiel, 1997; Wang et al., 1998; Anczkiewicz et al., 2007; Searle et al., 2010; Liu et al., 2012; Chiu et al., 2018). Gneiss antiforms within the ARSZ provide a key opportunity for investigating the evolution of intracontinental deformation and its exhumation in southeastern Tibet and Indochina during the Cenozoic.

Published structural data suggested that the ARSZ is a lithospheric-scale shear zone with a sinistral strike-slip sense of shear (e.g., Leloup et al., 1993, 1995). It has been suggested that major sinistral strike-slip movement (700 ± 200 km) occurred along the zone during the Cenozoic (Leloup et al., 1995; Liu et al., 2012). Various authors have also argued that magmatism and exhumation of the metamorphic antiforms within the shear zone occurred as a direct result of extensive sinistral strike-slip shearing during the Eocene and Miocene (Tapponnier et al., 1990; Harrison et al., 1992; Leloup et al., 1993; Chung et al., 1997; Liu et al., 2012). However, recent studies have indicated that the strike-slip shear zone roots onto a mid-crustal horizontal shear layer (Wang and Burchfiel, 1997; Jolivet et al., 2001; Searle, 2006; Anczkiewicz et al., 2007; Zhang et al., 2014; Yan et al., 2021). Based on the structural investigations along the southern segments (such as the Day Nui Con Voi zone) of the ARSZ by Jolivet et al. (2001), these authors proposed that the extensional structures in the Cenozoic evolution of the ARSZ and adjacent regions in southeastern Asia were more significant than had been previously inferred. Jolivet et al. (2001) and Anczkiewicz et al. (2007) further suggested that ARSZ-related metamorphic rocks were extruded in an overall transtensional regime, which led to the vertical extrusion of the ductile middle crust as an antiform. Namely, sinistral strike-slip fabrics were superimposed onto the subhorizontal shearing and earlier domal fabrics in these gneiss antiforms (Jolivet et al., 2001; Socquet and Pubellier, 2005; Searle, 2006; Zhang et al., 2010; Yan et al., 2021). These observations indicated that the ductile shearing layers of the middle to lower crust can play a major or dominant role during the formation and exhumation of the metamorphic antiforms within the ARSZ (Jolivet et al., 2001; Zhang et al., 2014; 2017; Chen et al., 2018; Yan et al., 2021). The investigations challenge the traditional conception that deformation and exhumation of the metamorphic complexes within the ARSZ were attributed to the sole contributions of left lateral strike-slip shearing along the ARSZ. Therefore, fundamental structure frames and formation timing of subhorizontal shearing layers, gneiss antiforms and lateral strike-slip shear zones, and exhumation mechanisms of these metamorphic rocks within the various segments of the ARSZ need to be checked further.

In this study, we present new structural observations and thermochronological data from the Diancangshan massif in the middle segment of the ARSZ. These new data reconstruct structural patterns and the late Cenozoic thermal history of the Diancangshan massif. The roles of subhorizontal shearing and lateral strike-slip shearing in the various levels of the crust during progressive shortening are discussed in the configuration of India-Eurasia continental convergence.

<Figure 1>

2. Geology of the Ailao Shan–Red River shear zone

The ARSZ is defined by four narrow, elongate metamorphic-complex massifs: The Xuelongshan massif in the northernmost segment, the Diancangshan massif in the Dali region, the Ailao Shan massif along the Red River in Yunnan, and the Day Nui Con Voi massif in Vietnam (Figure 1b; Tapponnier et al., 1982; Leloup and Kienast, 1993; Wang and Burchfiel, 1997; Jolivet et al., 2001; Leloup et al., 2001; Searle, 2006; Wang et al., 2008; Yeh et al., 2008). The Xuelongshan, Ailao Shan, and Day Nui Con Voi massifs are characterized by antiforms of variable-grade metamorphic rocks and scattered granitoid intrusions with extensive mylonitization (e.g. Jolivet et al., 2001; Leloup et a., 2001; Cao et al., 2011; Zhang et al., 2014). In the central parts of the massifs within the ARSZ, high-temperature metamorphic fabrics define the foliations and lineations (e.g. Leloup et al., 1995; Cao et al., 2016). Foliation dips are gentle or subhorizontal, and lineations are horizontal to subhorizontal, and parallel (Leloup et al., 1995, 2001) or normal (Jolivet et al., 2001; Zhang et al., 2014, 2017) to the trend of the shear zone. Within the core of the Xuelongshan massif, the mylonitic gneiss indicates under upper-greenschist- to amphibolite-facies conditions; that is, 670 °C and 4.8 kbar for the gneiss in the core and 490 °C and 4.1 kbar for the schists in the limbs (Song et al., 2001; Zhang et al., 2017). The gneiss core of the Diancangshan massif is characterized by high-temperature metamorphic fabrics, estimated by matrix mineral assemblages of garnet + biotite + kyanite/sillimanite + plagioclase + quartz and garnet + biotite + sillimanite + plagioclase + K-feldspar + quartz (720–760 °C and 8.0–9.3 kbar; Wang et al., 2016). The retrograde assemblage shows a mineral assemblage of garnet + biotite + sillimanite + plagioclase + quartz, which formed at 650–760 °C and 5.0–7.3 kbar (Wang et al., 2016). Within the Ailao Shan massif, peak metamorphic conditions are suggested to have been ~700 °C and 0.7 GPa, with retrograde conditions of ~480 °C and 0.30 GPa (Leloup et al., 1995; Nam, 1998; Leloup et al., 2001). The age of the high-temperature fabrics in the core is not precisely constrained. Leloup et al. (1995) associated all of these ductile fabrics with sinistral transpressional deformation, but other studies have proposed that the shallow- or subhorizontal-dipping foliations in the cores of the Xuelongshan, Ailao Shan, and Day Nui Con Voi massifs were derived from earlier horizontal shearing (or a mid-crustal decollement) and doming prior to sinistral strike-slip shearing (e.g., Jolivet et al., 2001; Searle, 2006; Zhang et al., 2014, 2017). Jolivet et al. (2001) also proposed two distinct deformation phases, spatially confined to specific structural levels, with different deformation regimes partitioned between the core and the limbs of the antiform. Gilley et al. (2003) and Anczkiewicz et al. (2007) suggested that upwelling of the asthenosphere due to lithospheric thinning is a plausible mechanism for the high-temperature metamorphism in the cores of the massifs within the ARSZ.

Ductile shear senses determined through macro- and microstructural analyses indicate that these three dome-like gneissic massifs were overprinted by intense sinistral ductile strike-slip shearing (e.g., Jolivet et al., 2001; Socquet and Pubellier, 2005; Anczkiewicz et al., 2007; Liu et al., 2007; Searle et al., 2010; Zhang et al., 2014). The mylonitic foliation of the strike-slip shear zone is generally steep, and the lineation is usually horizontal or subhorizontal, with both being nearly parallel to the trend of the gneiss massifs (e.g., Leloup et al., 2001; Jolivet et al., 2001). Evidence for sinistral strike-slip shear was investigated extensively by Tapponnier et al. (1986) and Leloup et al. (1993, 1995, 2001), who documented consistent kinematics within the ARSZ. The amount of sinistral displacement was estimated at about 700 ± 200 km, with slip rates of 3 to 5 cm/yr (Leloup et al., 1995). Leloup et al. (2001) inferred that sinistral shearing movement occurred under high-temperature amphibolite-facies conditions (700 °C) and propagated down into greenschist-facies conditions (~480 °C and 3 kbar) for the subsequent mylonitization in the shear zone. However, Jolivet et al. (2001) suggested that sinistral strike-slip shear occurred in greenschist-facies conditions along the margins of the antiforms, such as the Xuelongshan antiform (Zhang et al., 2017) and the Day Nui Con Voi antiform (Jolivet et al., 2001). In the Diancangshan and Ailao Shan massifs, sinistral strike-slip shear is suggested to have been coeval with the emplacement of leucogranitic veins/sills (Leloup et al., 1995; Cao et al., 2011). The U–Pb dating results of zircon, monazite, and titanite from deformed granitoids within the Ailao Shan segment have yielded ages ranging from 34 to 21 Ma, interpreted as the timing of the sinistral strike-slip movement (Schärer et al., 1994; Zhang and Schärer, 1999; Gilley et al., 2003). These syn-kinematic veins have yielded U–Pb ages of 26–22 Ma in the Ailao Shan massif (Leloup et al., 1995, 2001) and 30–21 Ma in the Diancangshan massif (Leloup et al., 2001; Cao et al., 2011), implying that the sinistral strike-slip shear took place during the middle Oligocene and early Miocene (Harrison et al., 1996; Searle, 2006; Cao et al., 2011). Numerous 40Ar/39Ar dating results obtained for the ARSZ have yielded a wide range of ages. 40Ar/39Ar hornblende and muscovite ages in the shear zone are 27–19 Ma and 20–17 Ma, respectively, in the Diancangshan segment (Cao et al., 2011). Biotite ages cluster around 20–17 Ma in the Diancangshan segment (e.g., Harrison et al., 1996; Leloup et al., 2001). These ages, ranging from 34 Ma to 17 Ma, have been interpreted as activity ranges of the sinistral strike-slip movement along the ARSZ. Gneiss cores in the Diancangshan and Ailao Shan zones cooled rapidly to ~300 °C during 22–17 Ma before the cessation of sinistral strike-slip shearing (Schäerer et al., 1994; Searle, 2006; Chen et al., 2018).

Kinematics reversed along the ARSZ with dextral strike-slip movement commencing on the Weixi-Qiaohou and Red River faults since the late Miocene (Leloup et al., 2001). The Weixi-Qiaohou fault extends along the northeastern side of the Xuelongshan massif and the southwestern side of the Diancangshan massif, and is featured by dextral strike-slip movement with a normal component of faulting (Figure 1b and 2a). The Red River fault lies on the northeastern sides of the Diancangshan and Ailao Shan massifs, and on the southwestern side of the Day Nui Con Voi massif, generating triangular facets on the range fronts and dextral-laterally deflected tributaries along strike (Allen et al., 1984; Leloup et al., 1993; Wan et al., 1997; Leloup et al., 2001; Schoenbohm et al., 2005, 2006; Socquet and Pubellier, 2005; Wang et al., 2016; Cao et al., 2016). The Red River fault is characterized by dextral transtension with shortening structures in the central restraining bend area (Leloup et al., 1993; Wang et al., 2016). The onset timing for kinematic reversal of the Red River fault is controversial. K-feldspars 40Ar/39Ar data display complex age spectra, yielding ages of 10–6 Ma (Harrison et al., 1996), 9–4 Ma (Cao et al., 2011), and ca. 5 Ma (Leloup et al., 1993) in the eastern side of the Diancangshan. These data indicate rapid cooling commencing at ca. 10 Ma or 5 Ma, which has been interpreted to result from the normal component of dextral strike-slip fault movement and has been suggested as the initiation timing of slip reversal along the Red River fault (Leloup et al., 1993). However, the current publications imply that the most recent phase of a low-magnitude accelerated cooling event is inferred to have started during 14-10 Ma along the Ailao Shan (Wang et al., 2020), with a later cooling event at ~8 Ma along the Diancangshan and at ~5 Ma along the Xuelongshan (Leloup et al., 2001; Wang et al., 2016; 2020; Zhang et al., 2017). These various cooling episodes in the mid-late Miocene to Pliocene has been interpreted to result from dextral strike-slip faulting with dip-slip components along these brittle faults (Bergman et al., 1997; Gilley et al., 2003; Wilson and Fowler, 2011; Hoke et al., 2014; Wang et al., 2016; Hoke, 2018; Wang et al., 2020).

3. Structures and kinematics of the Diancangshan massif

The ARSZ in the Dali region, Yunnan, is defined by the Diancangshan massif (Figures 1 and 2a–b). This massif is a NW–SE-trending metamorphic zone measuring 100 km long and up to about 10 km wide, extending from Qiaohou to Midu (Figure 2a–b). It is bounded to the NE by the NE-dipping Red River fault, and to the SW by the SW-dipping Weixi–Qiaohou fault (Figures 2 and 3a–d). The massif is composed of a variety of rock types, including paragneisses, augen gneisses, gneissic migmatite, amphibolites, leucogranites, garnet mica schists, quartzites, and marbles. These rocks commonly exhibit various mylonitic fabrics. Sedimentary basins are developed along the Red River and Weixi–Qiaohou faults, and are filled with detrital material of Neogene to Quaternary age (BGMRYP, 1987; Socquet and Pubellier, 2005; Liu et al., 2007; Figure 2a).

<Figure 2>

The southern end of the Diancangshan massif is well exposed. Because of difficult access in the field, the inner part of the Diancangshan was particularly challenging to investigate. Only one incomplete section (section Q in figure 19 of Leloup et al., 1995) across the massif had been published previously (Leloup et al., 1993, 1995). However, two recent roads across the massif have provided nearly continuous exposures, which allowed us to investigate structures and kinematics across the entire massif (cross-sections A–A’ and B–B’ in Figures 2b and 3a–b). In our investigation, two other incomplete cross-sections through the southern part of the massif were also studied (Figures 2b and 3c–d).

The first-order structure of the Diancangshan massif is defined as a large-scale NW–SE-trending open antiform, which presents as an elongate, elliptical dome-like structure (Figure 2b). The antiformal massif includes a gneiss core and is flanked by shear zones and normal faults (Figures 2a–b and 3a–d). The eastern and western margins of the gneiss core are mantled by two steeply dipping shear zones, namely, the eastern and western branches of the ARSZ. To the outside, the antiform is bounded by the steeply dipping Red River and Weixi–Qiaohou faults (Figures 2b and 3a–d). The southeastern boundary of the antiform is a low-angle normal fault, the ~3-km-long SSE-dipping Xiaguan fault (Figures 2 and 3e). Along this normal fault zone, the southeastern margin of the Diancangshan antiform consists of a shallow-dipping, strongly mylonitized belt (Figure 2b) (Leloup et al., 1993).

Different structures and kinematics are recorded in the various tectonites of the core and margins of the Diancangshan massif. We present the structural domains separately as follows.

**3.1. The gneiss core**

The core of the antiform contains gneiss and migmatite (Figures 2b and 3a–e). Throughout the entire core of the Diancangshan antiform, there is widespread evidence for varying degrees of partial melting, such as migmatization, numerous leucogranitic veins/sills, quartz veins, and aplites in the various gneisses (Figure 4a–g). The gneiss and migmatite are extensively deformed, with a gently to moderately dipping foliation and roughly NW–SE-trending, NE–SW-trending or N–S-trending stretching/mineral lineations on the major, high-temperature foliations (Figure 3a–d). Numerous leucogranitic veins/sills have been overprinted by varying degrees of solid-state deformation in concordance with the dominant foliation in the gneiss, especially in the mylonitic gneiss layers (Figure 4d–e, g, and h–k). The geometry of the foliation is thus similar in shape to that of a gneiss antiform, in which the gneissic foliations are shallowly outward dipping (Figures 2b, 3, and 4a–e). The mineral lineation is primarily defined by aligned micas and strain shadows of plagioclase and quartz porphyroblasts (Figure 4b). However, the trends of these mineral lineations are markedly different from the sinistral strike-slip-related lineation in the limbs of the antiform. Unlike the eastern and western branches of the ARSZ, shear senses associated with these mineral lineations are scarce in the gneiss core and are hard to interpret. This can be attributed to tectonic disruption of the core of the antiform, including gneiss foliations that are extensively disrupted or commonly cut by variously dipping normal faults (Figures 3a–b and 4c). All these structures indicate that the main phase of subhorizontal ductile shearing with amphibolite-facies took place in the presence of partial melt in the gneiss core of the Diancangshan massif.

<Figure 3>

< Figure 4>

**3.2. Western and eastern margins**

Two ductile shear zones, the western and eastern branches of the ARSZ, bound the gneiss core along the western and eastern sides of the Diancangshan massif (Figure 5a–f and 6a–c). The western branch of the ARSZ is a NW–SE-trending, 1–2 km wide, and ~70 km long mylonitic belt (Figure 3a–d). The belt is truncated locally, and cut by brittle faults (Figure 2b). The northern tip intersects with the Weixi–Qiaohou fault (Figure 2a–b). The southern end is cut by the Xiaguan normal fault (Figure 2b; Wintsch and Yeh, 2013). Foliation planes in the mylonites are defined by sillimanite, muscovite, and biotite, as well as strongly flattened quartzofeldspathic minerals (Figure 6a–c). The mylonites in the western branch of the ARSZ present a pervasive NW–SE-striking mylonitic foliation (Figures 2b and 3a–d) and a well-developed SE- or S-plunging stretching lineation (Figure 5c–e). The mylonitic foliation dips steeply ~250° SW with lineation plunging 10°–15° SE or S (Figure 5c–e). The mylonitic rocks are extensively intruded by leucogranitic veins or sills parallel to mylonitic foliation (Figure 5c). On the outcrops and thin sections, quartz gains have been dynamically recrystallized by bulging, subgrain rotation, grain boundary migration processes (Figure 6d–e). Plagioclase is characterized by plastic elongation, undulose extinction, and core–mantle structures (Figure 6e). Grains with bulging are common, and are similar in size to subgrains around the plagioclase porphyroclasts, which suggests that new grains developed by subgrain rotation and local grain boundary migration during ductile shearing (Figure 6e). Kinematic indicators at all scales, including asymmetric plagioclase porphyroclasts, mica fish, and asymmetric sub-recumbent folds, consistently suggest sinistral strike-slip shear (Figure 6a–e). Mesoscopic and microstructural kinematic indicators are consistent with sinistral strike-slip shear in the eastern branch of the ARSZ (Figure 6b and f–g). The deformation microstructures indicate that the mylonites were developed under middle-greenschist-facies conditions with deformation temperatures of at least 400 °C during sinistral strike-slip shear along the western shear zone.

To the southward, the western branch of the shear zone gradually changes trend to WNW–ESE in the Xiaguan region (Figure 2b). Mylonitic gneisses and schists show well-developed foliations dipping to the SW or SE and mineral stretching lineations plunging SSE (Figures 2b, 7, and 8a–e). In the mylonites, recrystallized, elongated quartz ribbons are common (Figure 8c–d). Small strain-free, fine-grained subgrains around polygonal quartz grains indicate a component of subgrain rotation during dynamic recrystallization (Figure 8f–g). Two types of feldspar grain can be identified in quartzo-feldspathic mylonites, namely porphyroclasts and newly developed fine grains. Plagioclase porphyroclasts show fish or flattened augen shapes, and display inhomogeneous or undulose extinction in the core (Figure 8f). Core–mantle structures in feldspars are developed widely in the mylonites, indicating the presence of subgrains (Figure 8g). Some feldspar porphyroclasts show brittle fracturing (Figure 8f). These microstructures suggest that feldspars underwent ductile deformation at middle- or upper-greenschist (400–500 °C) conditions during shearing deformation. Asymmetric rotated feldspar porphyroclasts, S–C structures, and asymmetric core–mantle structures show a top-to-the-SSE or -SE sense of shear (Figure 8c–g).

On the eastern margin of the Diancangshan antiform, the eastern branch of the ARSZ is 0.5–1 km in width, generally dips steeply to the NE or ENE (60°–90°), and is cut by the NW–SE-striking Red River fault to the east and by the Xiaguan normal fault to the south (Figures 2a–b and 3a–d). Various rocks, including calcareous quartzites, marbles, micaceous quartzites, pelitic schists, mafic metavolcanics, and augen granitoids, display mylonitic fabrics. The mylonitic rocks typically show steeply dipping, NW–SE-striking foliation and subhorizontal mineral stretching lineation parallel to the strike of the shear zone (Figures 3a–d and 5a–b). Stretching lineations are defined by elongated quartz and feldspar rods and trend SE with shallow to subhorizontal plunges (Figure 5b). In the mylonite, quartz grains are characterized by strongly elongated single crystals in ribbons (Figure 6f–g). Plastic elongation, undulose extinction, subgrains, and feldspar core–mantle structures are also well developed in the mylonitic granites (Figure 6f–g). Muscovite crystalsin these rocks are typically intergrown with fine-grained quartz or albite, or locally with chlorite (Figure 6c, f–g). Some plagioclase porphyroclasts display brittle fracturing (Figure 6g). In addition, feldspar commonly exhibits a wavy interface with intergrown muscovite and quartz (Figure 6g), indicative of albite replacement by muscovite. Chlorite flakes also tend to be oriented parallel to the foliation (Figure 6g). The common coexistence of chlorite and K-feldspar indicates that the rocks were metamorphosed and deformed under greenschist-facies conditions. Kinematic criteria associated with these microstructural fabrics suggest sinistral shearing movement at middle- or upper-greenschist-facies conditions (400–500 °C) within the eastern shear zone of the ARSZ (Figure 6b, f–g).

<Figure 5>

<Figure 6>

**3.3. Relationship between brittle and ductile structures in the southern margin**

The contact between the high-temperature metamorphic rocks of the massif and overlying low-grade metasediments is best exposed at the southernmost edge of the antiform. This contact is a structural boundary, marked by the Xiaguan low-angle normal fault (Figures 2b and 7). The normal fault dips 20°–35° to the SSE or SE, and is commonly associated with a 30–50-m-thick cataclastic deformation zone. Cleavages and fracture planes in the cataclastic deformation zone dip at 15°–40° to the SE (Figures 7, and 9a–b). Striations trend parallel to the high-temperature stretching lineation in the mylonitic gneiss of the footwall and plunge toward the SE (Figures 7, 8a-b, and 9b). Cataclastic deformation is superimposed upon earlier schistose fabrics along the southern margin of the antiform (Figures 7, 8a–g, 9a, and 10). Microstructural analyses reveal randomly oriented fractured K-feldspar porphyroclasts within a fine-grained matrix of quartz, plagioclase, and minor chlorite and muscovite (Figure 10a–e). The feldspar clasts exhibit a range of sizes, and some have elliptical or rounded shapes, suggesting comminution (Wintsch and Yeh, 2013). At all scales of observation, tension fractures are commonly filled with fine grains of quartz and albite (Figure 10b–f). Most of the fine quartz grains exhibit undulose extinction, indicating subgrain development in the quartz-rich matrix. Together these overprinting textures and microstructures in the cataclasis can demonstrate several oscillations between brittle and viscous deformation, all at lower greenschist facies conditions (at depths of 12–15 km) (Wintsch and Yeh, 2013).

Structurally below the cataclastic deformation zone, gneisses, schists, and plutonic rocks are extensively mylonitized, with SSE- or SE-plunging mineral lineations (as described above; Figures 7, 8a–b, and 9a). Downdip stretching lineations, together with extensional crenulation cleavages, S–C structures, and asymmetric “drag”-type folds, are indicative of sinistral transtensional deformation at the southeastern tip of the massif (Figures 7, 8c–g, and 9a) (Wintsch and Yeh, 2013, and our own field investigations).

Mesozoic rocks exposed to the west and southwest of the antiform are moderately to strongly deformed; open upright folds and brittle faults and fractures increase in frequency toward the antiform (Figures 3). Along the southernmost margin of the antiform, the hangingwall of the Xiaguan fault contains evidence of brittle deformation only (Figure 9b). Normal faults with variable dips (20°–60°) and lengths of 0.1–0.5 or 1 km are common.

<Figure 7>

<Figure 8>

<Figure 9>

<Figure 10>

4. Geochronology and thermochronology

**4.1. Methods and sample descriptions**

To better resolve the Cenozoic thermal history of the Diancangshan massif, we utilized a variety of thermochronometers, including 40Ar/39Ar in mica, AFT, and AHe. The AFT and AHe thermochronometers are used to constrain the timing and magnitude of cooling through a given temperature window within the shallow crust (60–120 ℃ for AFT and 40–80 ℃ for AHe, corresponding to depths of 2–4 km; e.g., Reiners and Farley, 2001; Reiners and Brandon, 2006).

We present 40Ar/39Ar mica data from six samples of the mylonitic gneiss and granitic mylonite as well as AFT and AHe data from nine mylonite samples (Tables 1–5). Sampling locations are shown in Figure 2b. Samples for AFT and AHe analyses were collected over a 4-km-long composite transect spanning 2500 m of vertical relief along the massif and from additional locations at the southern margin of the range. To avoid 4He loss due to reheating events associated with igneous emplacement, sampling locations were positioned far from known large-scale intrusions (e.g., Ehlers and Farley, 2003). Thus, the analyzed samples should record cooling due only to erosion or tectonic exhumation. Three grains were analyzed for each sample (Table 3).

Minerals were separated using standard crushing, sieving, magnetic, and heavy liquid techniques. Muscovite 40Ar/39Ar dating was performed using a conventional step-heating technique in a high-frequency furnace at the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGGCAS), Beijing, China. The AHe and AFT analyses were performed at the Arizona Radiogenic Helium Dating Laboratory (ARHDL) and the fission-track laboratory, respectively, at the University of Arizona (Tucson, USA), using the protocols proposed by Reiners (2005). Estimated analytical uncertainties are 2%–9% (1σ) for apatite AHe ages, reflecting the reproducibility of replicate analyses of laboratory-standard samples. The external-detector method, using uranium-poor muscovite as the detectors, was employed for AFT analyses. Full methodological details and microstructures (Figure S1) for these dated samples are presented in Table S1 in the Appendix.

**4.2. Samples and analytical results**

4.2.1. Microstructures and muscovite 40Ar/39Ar step-heating ages

Based on microscopic investigation, size fractions of >100 μm were collected for porphyroclasts or matrix muscovites. The matrix muscovites were too fine-grained to be separated from the samples; therefore, most dated muscovite grains were porphyroclasts from the mylonite samples. In Figure 11a, plateau ages are presented in preference to isochron ages, as on isochron plots, the representative points cluster around the 39Ar/40Ar intercept, implying an imprecise 40Ar/36Ar ratio. Six muscovite separate 40Ar/39Ar analyses yielded different (within 2σ analytical error) weighted mean plateau ages (WMPAs) and inverse isochron ages (IIAs). The results and interpreted muscoviteages are summarized in Tables 1 and 2.

The two samples (DCS-GL-069 and DCS-GL-090) were collected from two mylonitic gneisses in the core of the Diancangshan massif (Figures 2b and 11b). At the sampling sites, the mylonitic gneisses are characterized by a gently dipping foliation, a NW–SE-trending stretching lineation, and a top-to-the-SE sense of shear. The fine-grained orthogneiss is composed of K-feldspar, quartz, biotite, and muscovite (Table 2; Figure S1a, b). Micas (20%–25% by mode) are strongly deformed and oriented to define the mylonitic foliation. Most mica grains in these two samples range between 50 and 500 μm in length, and are distributed in two patterns (Figure S1a, b). The main pattern is represented by masses of segmented mica grains, which are not optically continuous but resemble porphyroclasts in outline. This pattern of mica exhibits mica fish structures with aspect ratios of >2:1, and defines the primary shearing C-plane of the S–C structures (Figure S1a, b). The other mica pattern is characterized by finer mica grains (<50 μm) and neo-crystallized micas oriented parallel to the S-plane (Figure S1a). Extensively stretched quartz and feldspar grains feature undulose extinction and irregular grain boundaries that formed in response to bulging, subgrain rotation, and grain-boundary migration recrystallization (Figure S1a, b; Table S1). Considering these microstructures, deformation temperatures for the dated mylonites are estimated to have been 500 °C or higher. The age spectrum for muscovite in sample DCS-GL-069 defines a weakly developed staircase-like shape, with ages increasing from 26–30 Ma for low temperatures to 36.0 ± 0.8 Ma in the last significant steps. Intermediate degassing temperatures exhibit an age of 37.0 ± 1.0 Ma (Figure 11a). This spectrum relates to the closure of the system at an age of 36 Ma, followed by Ar loss. The plateau age roughly agrees with the total fusion age (35.5 ± 0.4 Ma) and inverse isochron age (34.3 ± 1.5 Ma), within errors. The age spectrum of the muscovite for DCS-GL-090 also shows a weakly developed staircase-like shape, but displays (for 95% of released 39Ar) increasing ages from 27 Ma to a first integrated age of 28.4 ± 0.5 Ma and a second at 29.7 ± 0.6 Ma. The plateau age of 29.7 ± 0.6 Ma agrees largely with its total fusion age (29.0 ± 0.3 Ma) and inverse isochron age (29.5 ± 1.3 Ma), within errors. The age difference between two muscovite grains (two sampling sites DCS-GL-069 and DCS-GL-090), corresponding to different sampling locations within the subhorizontal layer of the antiform, exceeds 5 Myr.

The two samples (DCS-GL-049 and DCS-GL-050) were collected from granitic mylonite with the vertical foliation and subhorizontal lineation on the western sinistral strike-slip shear zone, ~11 km to the north of the Xiaguan fault (Figures 2b and 11b). Sample DCS-GL-49 is a coarse-grained augen mylonite, whereas DCS-GL-050 has mylonitic microstructures with a fine-grained banded quartz–feldspar foliation and well-oriented mica and feldspar porphyroclasts (Figure S1c, d). In thin sections, micro-shear zones are typically enriched in muscovite. These zones exhibit very fine-grained (generally <250 μm in length; 15%–20% of the rock) aggregates of mica at the expense of feldspar or biotite grains (Table 2). Most muscovite fragments are distributed in two patterns. The main pattern is characterized by fine neo-crystallized muscovite grains (<50 μm), lying with variable orientation to the main shear planes (Figure S1c). The microstructures of these grains reveal that progressive shearing deformation is accompanied by the increasing alteration of feldspar grains into muscovite. Some neo-crystallized muscovite grains have been rotated, then oriented into the shearing planes (Figure S1d). The other pattern of muscovite fragments is characterized by thin, individual, foliation-parallel grains that are distributed throughout the fine quartz matrix (Figure S1c, d). Chloritization occurs commonly at the microscopic scale. Microstructural analyses imply progressive deformation within a single deformation phase of sinistral strike-slip shearing (Figure S1c, d). Dynamic recrystallized quartz grains and polycrystalline quartz aggregates (50–250 μm in size) with an average aspect ratio of >3:1 probably developed predominantly by subgrain rotation recrystallization (Figure S1c, d). Myrmekite (intergrowth of quartz and plagioclase) replaces microcline (K-feldspar) in the mylonitic felsic layers (Figure S1c). Based on these microstructures, deformation temperatures for the dated mylonites are estimated at 400–450 °C (Table S1). The 40Ar/39Ar analyses yield well-defined plateau ages of 24.7 ± 0.3 and 25.6 ± 0.3 Ma, respectively, during intermediate- and high-temperature heating steps (Figure 11a). These ages are defined by more than 95% of total 39Ar release and have 40Ar/36Ar initial ratios of 289–304. The corresponding IIAs are consistent with their plateau ages (Table 2). The age difference between two muscovite fragments, corresponding to different sampling sites within the sinistral strike-slip shear zone (Figure 2b), does not exceed 1.5 Myr.

The two samples (DCS-GL-076 and DCS-GL-104) are mylonites from the southern margin of the massif, close to the Xiaguan fault (Figures 2b and 11b). They exhibit roughly E–W-striking, moderately or gently dipping mylonitic foliations with gently plunging stretching lineations, and contain quartz-feldspathic ribbons, asymmetric feldspar clasts, and thin deformed micas (25%–30% by mode) that indicate top-to-the-SE extensional shearingwithin these two sampling sites (Figures S1e, f, and 8f). The mica-rich layers in these samples define a south-dipping mylonitic foliation that is parallel to the plane of the Xiaguan fault. In thin sections, the major type of the muscovite appears as fine, neo-crystallized fragments (generally <150 μm in size) or aggregates developed at the expense of biotite and feldspar, resembling porphyroclasts. These muscovite fragments are commonly rotated, reoriented, and distributed parallel or at variable angles to the main shearing foliation (Figure S1f). Some segmented and sheared coarse muscovite grains (resembling porphyroclasts) are oriented parallel to the neo-crystallized mica layers (Figure 8f). Extensively stretched quartz grains (200–600 μm in size with an aspect ratio of >6:1) are characterized by core–mantle structures, undulose extinction in the core grains, and irregular grain boundaries that formed in response to bulging and subgrain rotation recrystallization (Figures S1e, f, and 8f). Elongate subgrains are also common in the deformed quartz and plagioclase crystals (Figure 8f). Considering the microstructures, deformation temperatures for the dated mylonites are estimated at 400–450 °C (Table S1). The 40Ar/39Ar analyses yield well-defined plateau ages of 23.3 ± 0.3 (DCS-GL-076) and 17.1 ± 0.2 Ma (DCS-GL-104) during intermediate- and high-temperature heating steps (Figure 11a). The obtained ages correspond to 95% of total 39Ar release, and initial 40Ar/36Ar ratios of 304 and 293 (similar to the present atmospheric 40Ar/36Ar ratio), respectively. The corresponding IIAs are consistent with their plateau ages (Figure 11a; Table 2). The age difference between two muscovite grains, corresponding to different sampling sites (Figure 2b), is ~5 Myr.

<Figure 11>

<Table 1>

Table 1 40Ar/39Ar step-heating results.

|  |  |  |  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- | --- | --- | --- |
| Step | Laser Power(%) | 40Ar/39Ar | 37Ar/39Ar | 36Ar/39Ar | 40Ar\*/39Ark | 40Ar\*(%) | 39Ark(%) | Age(Ma) |
| **Sample: DCS-GL-049 J=0.00333800 ± 0.00001669** |
| 1 | 1.60 W | 7.976269 | 327.301205 | 0.014943 | 3.56096 | 44.64 | 0.51 | 21.37± 2.19 |
| 2 | 1.70 W | 5.578637 | 346.103677 | 0.005071 | 4.08005 | 73.14 | 0.45 | 24.46± 1.59 |
| 3 | 1.80 W | 6.567875 | 54.219546 | 0.008581 | 4.03209 | 61.39 | 3.02 | 24.18± 0.79 |
| 4 | 1.90 W | 6.516513 | 24.466644 | 0.008052 | 4.13694 | 63.48 | 6.69 | 24.80± 0.68 |
| 5 | 2.00 W | 5.587064 | 17.861521 | 0.004996 | 4.11057 | 73.57 | 9.35 | 24.64± 0.47 |
| 6 | 2.10 W | 5.207514 | 16.943917 | 0.003754 | 4.09796 | 78.69 | 9.52 | 24.57± 0.39 |
| 7 | 2.20 W | 5.570831 | 13.659380 | 0.004956 | 4.10628 | 73.71 | 11.88 | 24.62± 0.46 |
| 8 | 2.30 W | 5.454454 | 11.292259 | 0.004506 | 4.12276 | 75.59 | 14.65 | 24.72± 0.43 |
| 9 | 2.40 W | 5.260815 | 13.607601 | 0.003834 | 4.12756 | 78.46 | 12.67 | 24.74± 0.40 |
| 10 | 2.50 W | 5.143676 | 17.728580 | 0.003522 | 4.10280 | 79.76 | 8.96 | 24.60± 0.39 |
| 11 | 2.60 W | 5.245849 | 12.938071 | 0.003734 | 4.14217 | 78.96 | 13.07 | 24.83± 0.39 |
| 12 | 2.70 W | 5.336937 | 18.351471 | 0.004117 | 4.12006 | 77.20 | 9.22 | 24.70± 0.42 |
| **Sample: DCS-GL-050 J=0.00333700 ± 0.00001669** |
| 1 | 1.60 W | 7.223320 | 295.097537 | 0.011554 | 3.80894 | 52.73 | 0.63 | 22.84± 1.85 |
| 2 | 1.70 W | 6.405501 | 59.062902 | 0.007209 | 4.27519 | 66.74 | 3.23 | 25.62± 0.73 |
| 3 | 1.80 W | 6.233006 | 21.697118 | 0.006531 | 4.30283 | 69.03 | 8.36 | 25.78± 0.58 |
| 4 | 1.90 W | 5.401082 | 22.269585 | 0.003745 | 4.29432 | 79.51 | 8.67 | 25.73± 0.41 |
| 5 | 2.00 W | 5.584749 | 17.687032 | 0.004334 | 4.30392 | 77.07 | 10.58 | 25.79± 0.44 |
| 6 | 2.10 W | 5.109592 | 12.338490 | 0.002723 | 4.30465 | 84.25 | 15.80 | 25.79± 0.34 |
| 7 | 2.20 W | 5.471619 | 13.113649 | 0.004100 | 4.25990 | 77.85 | 14.68 | 25.52± 0.41 |
| 8 | 2.30 W | 4.916334 | 13.796112 | 0.002227 | 4.25795 | 86.61 | 14.28 | 25.51± 0.32 |
| 9 | 2.40 W | 4.971328 | 21.034218 | 0.002380 | 4.26779 | 85.85 | 9.06 | 25.57± 0.35 |
| 10 | 2.50 W | 4.789540 | 36.472011 | 0.001759 | 4.26961 | 89.14 | 5.53 | 25.58± 0.36 |
| 11 | 2.60 W | 4.604058 | 40.703833 | 0.001229 | 4.24053 | 92.10 | 4.87 | 25.41± 0.35 |
| 12 | 2.70 W | 5.171815 | 45.694837 | 0.001948 | 4.59600 | 88.87 | 4.32 | 27.52± 0.40 |
| **Sample: DCS-GL-069 J=0.00333600 ± 0.00001668** |
| 1 | 1.60 W | 8.244402 | 303.246202 | 0.012931 | 4.42344 | 53.65 | 1.59 | 26.49± 1.88 |
| 2 | 1.70 W | 8.179018 | 160.697984 | 0.010501 | 5.07587 | 62.06 | 2.88 | 30.36± 1.25 |
| 3 | 1.80 W | 9.279721 | 55.618014 | 0.010462 | 6.18814 | 66.68 | 8.48 | 36.95± 0.94 |
| 4 | 1.90 W | 8.529666 | 47.893395 | 0.007602 | 6.28298 | 73.66 | 9.78 | 37.51± 0.76 |
| 5 | 2.00 W | 10.191465 | 145.528377 | 0.013156 | 6.30343 | 61.85 | 3.02 | 37.63± 1.39 |
| 6 | 2.10 W | 9.498187 | 163.641860 | 0.011342 | 6.14631 | 64.71 | 3.01 | 36.70± 1.27 |
| 7 | 2.20 W | 9.216143 | 202.841113 | 0.010360 | 6.15462 | 66.78 | 2.31 | 36.75± 1.35 |
| 8 | 2.30 W | 7.724529 | 25.645784 | 0.006137 | 5.91077 | 76.52 | 18.87 | 35.31± 0.61 |
| 9 | 2.40 W | 7.743516 | 24.434910 | 0.005793 | 6.03164 | 77.89 | 19.09 | 36.02± 0.59 |
| 10 | 2.50 W | 7.493514 | 29.859589 | 0.005125 | 5.97891 | 79.79 | 16.29 | 35.71± 0.57 |
| 11 | 2.60 W | 6.775284 | 31.772214 | 0.003672 | 5.68986 | 83.98 | 14.68 | 34.00± 0.49 |
| **Sample: DCS-GL-076 J=0.00333500 ± 0.00001668** |
| 1 | 1.60 W | 5.292594 | 285.874162 | 0.005555 | 3.65081 | 68.98 | 1.16 | 21.88± 1.34 |
| 2 | 1.70 W | 5.249005 | 95.199943 | 0.004762 | 3.84167 | 73.19 | 3.42 | 23.02± 0.66 |
| 3 | 1.80 W | 5.329786 | 47.711457 | 0.004855 | 3.89499 | 73.08 | 6.96 | 23.34± 0.51 |
| 4 | 1.90 W | 4.976638 | 28.539425 | 0.003706 | 3.88110 | 77.99 | 11.53 | 23.26± 0.41 |
| 5 | 2.00 W | 4.954487 | 28.670359 | 0.003555 | 3.90376 | 78.79 | 11.68 | 23.39± 0.40 |
| 6 | 2.10 W | 4.841121 | 22.804841 | 0.003170 | 3.90411 | 80.64 | 14.74 | 23.39± 0.36 |
| 7 | 2.20 W | 4.999861 | 31.054493 | 0.003628 | 3.92762 | 78.55 | 10.38 | 23.53± 0.42 |
| 8 | 2.30 W | 4.829784 | 29.343559 | 0.003192 | 3.88629 | 80.47 | 10.80 | 23.29± 0.39 |
| 9 | 2.40 W | 5.220564 | 33.356359 | 0.004389 | 3.92326 | 75.15 | 9.54 | 23.51± 0.46 |
| 10 | 2.50 W | 5.272558 | 40.310275 | 0.004641 | 3.90091 | 73.99 | 7.77 | 23.37± 0.50 |
| 11 | 2.60 W | 5.041271 | 45.111589 | 0.003869 | 3.89784 | 77.32 | 7.13 | 23.36± 0.47 |
| 12 | 2.70 W | 4.715043 | 63.911980 | 0.002933 | 3.84820 | 81.62 | 4.89 | 23.06± 0.48 |
| **Sample: DCS-GL-090 J=0.00333400 ± 0.00001667** |
| 1 | 1.60 W | 7.393590 | 316.603679 | 0.011764 | 3.91801 | 52.99 | 1.20 | 23.47± 1.97 |
| 2 | 1.70 W | 6.292189 | 150.138881 | 0.005994 | 4.52114 | 71.85 | 2.56 | 27.05± 1.05 |
| 3 | 1.80 W | 6.611638 | 59.633187 | 0.006403 | 4.71939 | 71.38 | 6.45 | 28.23± 0.85 |
| 4 | 1.90 W | 5.499509 | 25.299182 | 0.002326 | 4.81206 | 87.50 | 15.25 | 28.78± 0.49 |
| 5 | 2.00 W | 5.528430 | 38.195801 | 0.002750 | 4.71556 | 85.30 | 10.38 | 28.21± 0.54 |
| 6 | 2.10 W | 5.412700 | 29.613097 | 0.002212 | 4.75877 | 87.92 | 12.91 | 28.47± 0.49 |
| 7 | 2.20 W | 5.521507 | 34.423968 | 0.001851 | 4.97432 | 90.09 | 10.84 | 29.74± 0.50 |
| 8 | 2.30 W | 5.913253 | 28.258278 | 0.003164 | 4.97817 | 84.19 | 13.57 | 29.77± 0.56 |
| 9 | 2.40 W | 6.087324 | 39.780577 | 0.003725 | 4.98637 | 81.91 | 9.63 | 29.82± 0.61 |
| 10 | 2.50 W | 5.561400 | 53.575014 | 0.001838 | 5.01816 | 90.23 | 7.12 | 30.00± 0.55 |
| 11 | 2.60 W | 5.612770 | 86.869700 | 0.002206 | 4.96067 | 88.38 | 4.35 | 29.66± 0.66 |
| 12 | 2.70 W | 5.665281 | 64.497631 | 0.002451 | 4.94090 | 87.21 | 5.74 | 29.55± 0.61 |
| **Sample: DCS-GL-104 J=0.00333400 ± 0.00001667** |
| 1 | 1.60 W | 7.435864 | 303.011681 | 0.016252 | 2.63321 | 35.41 | 0.90 | 15.81± 1.97 |
| 2 | 1.70 W | 7.424124 | 96.790119 | 0.015276 | 2.90995 | 39.20 | 2.93 | 17.46± 1.32 |
| 3 | 1.80 W | 5.111796 | 62.967369 | 0.007827 | 2.79878 | 54.75 | 4.47 | 16.80± 0.75 |
| 4 | 1.90 W | 5.835663 | 28.336578 | 0.010297 | 2.79269 | 47.86 | 10.66 | 16.76± 0.80 |
| 5 | 2.00 W | 4.907161 | 11.127891 | 0.006916 | 2.86321 | 58.35 | 26.01 | 17.18± 0.54 |
| 6 | 2.10 W | 4.401164 | 30.474551 | 0.005123 | 2.88714 | 65.60 | 9.27 | 17.32± 0.47 |
| 7 | 2.20 W | 4.595205 | 27.990818 | 0.005839 | 2.86973 | 62.45 | 10.26 | 17.22± 0.50 |
| 8 | 2.30 W | 4.253652 | 31.805962 | 0.004670 | 2.87354 | 67.55 | 8.86 | 17.24± 0.44 |
| 9 | 2.40 W | 4.298398 | 27.318683 | 0.004958 | 2.83297 | 65.91 | 10.12 | 17.00± 0.44 |
| 10 | 2.50 W | 4.521714 | 37.944412 | 0.005841 | 2.79559 | 61.83 | 7.55 | 16.78± 0.53 |
| 11 | 2.60 W | 4.333259 | 60.734275 | 0.004994 | 2.85747 | 65.94 | 4.75 | 17.15± 0.54 |
| 12 | 2.70 W | 4.458765 | 65.850757 | 0.005582 | 2.80924 | 63.00 | 4.22 | 16.86± 0.58 |

<Table 2>

Table 2 Mineral assemblage and muscovite 40Ar/39Ar geochronological data for mylonitic samples from the DCS massif.

|  |  |  |  |  |  |  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- |
| **Sample** | **Rock type** | **Distance to XGDF****(km)** | **Elevation****(m)** | **Mineral assemblage** | **AM** | ***J*-value** | **WMPA****(Ma±2σ)** | **IIA****(Ma±2σ)** | **TFA****(Ma±2σ)** | **40Ar/36Ar****intercept** | **MSWD** |
| DCS-GL-104 | Migmatitic mylonite | 0.30 | 1650 | Kfs+Qtz+Pl+Ms+Bi+Chl | Ms | 0.003334 | 17.10±0.20 | 17.20±0.60 | 17.10±0.30 | 293.0±15.9 | 0.62 |
| DCS-GL-076 | Granitic mylonite | 0.35 | 1780 | Qtz+Kfs+Ms+Chl+Bi | Ms | 0.003335 | 23.30±0.30 | 23.30±0.90 | 23.30±0.30 | 304.1±38.2 | 0.43 |
| DCS-GL-050 | Granitic mylonite | 11 | 1900 | Kfs+Pl+Qtz+Bi+Ms+Chl | Ms | 0.003337 | 25.60±0.30 | 25.50±0.40 | 25.70±0.30 | 304.0±14.6 | 0.37 |
| DCS-GL-049 | Granitic mylonite | 11 | 1850 | Kfs+Pl+Qtz+Bi+Ms | Ms | 0.003338 | 24.70±0.30 | 24.80±0.60 | 24.70±0.30 | 289.9±19.3 | 0.32 |
| DCS-GL-069 | Mylonitic gneiss | 41 | 2500 | Kfs+Pl+Qtz+Ms+Bi±Hbl | Ms | 0.003336 | 36.20±0.70 | 34.30±1.50 | 35.50±0.40 | 340.4±33.6 | 2.19 |
| DCS-GL-090 | Mylonitic gneiss | 42 | 2480 | Pl+Kfs+Qtz+Ms+Bi+Chl | Ms | 0.003334 | 29.20±0.50 | 29.50±1.30 | 29.00±0.30 | 276.8±74.3 | 6.71 |
| *Note.* WMPA = weighted mean plateau age; IIA = inverse isochron age; TFA = total fusion age; AM = analyzed mineral; Kfs = K-feldspar; Pl = Plagioclase; Qtz = Quartz; Chl = Chlorite; Bi = Biotite; Ms = Muscovite. XGDF: Xiaguan low-angle normal fault. |

4.2.2. AFT and AHe ages

Apatite Fission-track analysis was undertaken on samples from the Diancangshan massif to complement the Ar dataset and constrain the low-temperature cooling history.

We obtained four AFT ages along the E–W vertical transect, which are scattered over a wide range from 15 to 6 Ma (Figure 12a; Table 3). The AFT ages in the N–S transect also show a wide range from 10 to 3 Ma (Figure 12b; Table 3). These AFT ages are younger than 15 Ma, and also younger than the muscovite 40Ar/39Ar ages determined for the rocks from the massif. There is no evidence for a systematic correlation between age and elevation, as would be expected from a terrain that had undergone a low to moderate uniform rate of denudation, that is, the cooling pattern is not attributable to variable depths of erosion.

Within our obtained AHe data, we rejected single-grain ages that were clearly distinct from the remaining analyses used in the mean age calculation (Table 4). There is some dispersion in the AHe ages, which can be attributed to several factors, including uranium zonation (Hourigan et al., 2005), radiation damage, and variation in crystal size (Reiners and Farley, 2001; Reiners et al., 2017). However, most AHe ages are young, and plotting eU (effective uranium concentration) against these single-grain ages does not display a clear correlation (Figure S2), indicating that any radiation damage in these samples is minor. Five mean AHe ages obtained from the N–S transect exhibit a narrow age range from 3.16 to 2.31 Ma (Figure 12b; Tables 4 and 5). However, for two samples taken from low elevations (DC1506-003 and DC1506-002), the AHe ages are close to or overlap with the AFT ages (Table 4; Figure 12b), which may indicate very rapid cooling through both system-closure-temperature windows. Our data from the E–W vertical transect include four mean AHe ages, ranging from 3.17 to 2.72 Ma (Figure 12a; Tables 4 and 5). This narrow range of AHe ages over a relatively narrow elevation range (1300 m) indicates that a period of rapid cooling commenced at or before ca. 2.7 Ma and lasted for ~1 Myr, resulting in ~60 °C of cooling.

<Figure 12>

<Table 3>

Table 3. Results of AFT analyses of samples from the DCS antiform.

|  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- |
| **Rock type** | **Sample ID** | **Mass****(μg)** | **U****(ppm)** | **Th****(ppm)** | **Sm****(ppm)** | **eU****(ppm)** | **4He (Ca)****(nmol/g)** | **Fta** | **Rsb (μm)** | **Raw Date****(Ma)** | **1σ****(Ma)** | **Corr Date****(Ma±1σ)** | **Mean Age****(Ma)** | **Elevation****(m)** | **Latitude****Longitude** |
| **The E-W vertical transect** |
| Granitic mylonite | DC1511-001\_Ap1 | 3.76 | 5.62  | 11.14 | 83.05  | 8.24  | 0.09  | 0.749 | 57.15  | 2.02  | 0.12  | 2.73±0.16 |  | 2600.00 | 25o 38’ 55.13” |
| DC1511-001\_Ap2 | 3.22 | 6.19  | 14.48 | 80.79  | 9.60  | 0.10  | 0.727 | 52.17  | 1.93  | 0.11  | 2.70±0.15 |  |  | 100o 08’ 31.23” |
|  | DC1511-001\_Ap3 | 4.63 | 8.23  | 16.52 | 78.12  | 12.11  | 0.30  | 0.769 | 62.53  | 4.61  | 0.21  | 6.06±0.28 |  |  |  |
|  | **DC1511-001** |  |  |  |  |  |  |  |  |  |  |  | **2.72±0.01** |  |  |
| Granitic mylonite | DC1511-005\_Ap1 | 3.77 | 15.07  | 50.45 | 101.69  | 26.93  | 0.28  | 0.737 | 54.30  | 1.90  | 0.09  | 2.63±0.12 |  | 2710.00 | 25o 39’ 36.53” |
| DC1511-005\_Ap2 | 3.44 | 9.60  | 30.38 | 101.81  | 16.74  | 0.20  | 0.745 | 56.00  | 2.23  | 0.11  | 3.06±0.15 |  |  | 100o 07’ 47.33” |
|  | DC1511-005\_Ap3 | 4.95 | 9.01  | 24.60 | 86.50  | 14.79  | 0.22  | 0.773 | 63.49  | 2.67  | 0.07  | 3.51±0.09 |  |  |  |
|  | **DC1511-005** |  |  |  |  |  |  |  |  |  |  |  | **3.07±0.21** |  |  |
| Granitic mylonite | DC1506-013\_Ap1 | 2.56 | 3.86  | 1.95 | 154.31  | 4.32  | 0.36  | 0.731 | 52.88  | 14.85  | 0.56  | 20.24±0.77 |  | 3855.00 | 25o 39’ 37.12” |
| DC1506-013\_Ap2 | 1.35 | 4.40  | 4.79 | 149.46  | 5.52  | 0.06  | 0.671 | 42.46  | 2.11  | 0.26  | 3.16±0.39 |  |  | 100o 06’ 26.28” |
|  | DC1506-013\_Ap3 | 1.69 | 8.68  | 17.71 | 168.97  | 12.84  | 0.13  | 0.673 | 42.72  | 1.87  | 0.12  | 2.83±0.18 |  |  |  |
|  | **DC1506-013** |  |  |  |  |  |  |  |  |  |  |  | **3.00±0.10** |  |  |
| Mylonitic gneiss | DC1506-011\_Ap1 | 1.93 | 4.76  | 9.77 | 34.08  | 7.06  | 0.09  | 0.690 | 45.23  | 2.22  | 0.15  | 3.29±0.22 |  | 3960.00 | 25o 39’ 48.24” |
| DC1506-011\_Ap2 | 0.75 | 8.98  | 18.00 | 118.34  | 13.21  | 0.15  | 0.625 | 36.55  | 2.12  | 0.19  | 3.48±0.32 |  |  | 100o 05’ 51.16” |
|  | DC1506-011\_Ap3 | 0.94 | 7.79  | 15.41 | 83.50  | 11.41  | 0.11  | 0.629 | 37.07  | 1.69  | 0.14  | 2.75±0.24 |  |  |  |
|  | **DC1506-011** |  |  |  |  |  |  |  |  |  |  |  | **3.17±0.18** |  |  |
| **The N-S vertical transect** |
| Granitic mylonite | DC1506-005\_Ap1 | 5.21 | 18.13  | 33.34 | 98.89  | 25.97  | 0.30  | 0.774 | 63.81  | 2.11  | 0.07  | 2.76±0.09 |  | 1659.00 | 25o 33’ 41.99” |
| DC1506-005\_Ap2 | 5.23 | 1.64  | 1.21 | 22.59  | 1.92  | 0.01  | 0.784 | 67.19  | 1.05  | 0.17  | 1.35±0.21 |  |  | 100o 07’ 16.97” |
| DC1506-005\_Ap3 | 4.21 | 1.73  | 0.21 | 33.13  | 1.78  | 0.02  | 0.760 | 59.90  | 2.15  | 0.21  | 2.82±0.28 |  |  |  |
|  | **DC1506-005** |  |  |  |  |  |  |  |  |  |  |  | **2.31±0.39** |  |  |
| Granitic mylonite | DC1506-003\_Ap1 | 9.10 | 48.47  | 2.41 | 87.66  | 49.04  | 0.58  | 0.820 | 81.32  | 2.20  | 0.07  | 2.68±0.08 |  | 1865.00 | 25o 34’ 30.35” |
| DC1506-003\_Ap2 | 10.34 | 20.83  | 2.23 | 28.72  | 21.36  | 0.21  | 0.820 | 81.15  | 1.80  | 0.06  | 2.20±0.07 |  |  | 100o 08’ 46.35” |
| DC1506-003\_Ap3 | 7.69 | 74.72  | 6.11 | 90.80  | 76.15  | 0.83  | 0.800 | 72.82  | 2.02  | 0.06  | 2.53±0.08 |  |  |  |
|  | **DC1506-003** |  |  |  |  |  |  |  |  |  |  |  | **2.47±0.12** |  |  |
| Granitic mylonite | DC1506-002\_Ap1 | 1.60 | 23.60  | 5.27 | 84.93  | 24.84  | 0.25  | 0.674 | 42.80  | 1.84  | 0.09  | 2.75±0.13 |  | 1950.00 | 25o 34’ 11.82” |
| DC1506-002\_Ap2 | 1.13 | 25.38  | 8.21 | 108.82  | 27.31  | 0.33  | 0.653 | 39.96  | 2.20  | 0.11  | 3.39±0.17 |  |  | 100o 11’ 07.43” |
| DC1506-002\_Ap3 | 1.40 | 26.70  | 5.16 | 90.51  | 27.91  | 0.31  | 0.667 | 41.79  | 2.02  | 0.09  | 3.04±0.13 |  |  |  |
|  | **DC1506-002** |  |  |  |  |  |  |  |  |  |  |  | **3.06±0.15** |  |  |
| Granitic mylonite | DC1404-061\_Ap1 | 3.35 | 19.91  | 73.85 | 281.02  | 37.27  | 0.40  | 0.728 | 52.27  | 1.95  | 0.06  | 2.74±0.09 |  | 2340.00 | 25o 36’ 39.18” |
| DC1404-061\_Ap2 | 3.39 | 16.20  | 50.70 | 236.49  | 28.12  | 0.30  | 0.725 | 51.62  | 1.92  | 0.06  | 2.71±0.09 |  |  | 100o 05’ 05.13” |
| DC1404-061\_Ap3 | 4.48 | 20.33  | 61.11 | 212.05  | 34.69  | 0.43  | 0.746 | 56.34  | 2.25  | 0.07  | 3.08±0.09 |  |  |  |
|  | **DC1404-061** |  |  |  |  |  |  |  |  |  |  |  | **2.84±0.10** |  |  |
| Granitic mylonite | DC1404-058\_Ap1 | 15.95 | 18.10  | 1.04 | 116.35  | 18.35  | 0.30  | 0.842 | 93.15  | 3.03  | 0.09  | 3.60±0.11 |  | 2500.00 | 25o 36’ 57.53” |
| DC1404-058\_Ap2 | 10.47 | 10.57  | 2.09 | 70.97  | 11.06  | 0.15  | 0.817 | 79.96  | 2.46  | 0.08  | 3.02±0.10 |  |  | 100o 05’ 08.62” |
|  | DC1404-058\_Ap3 | 7.95 | 18.26  | 2.94 | 102.99  | 18.95  | 0.24  | 0.807 | 75.68  | 2.31  | 0.07  | 2.86±0.09 |  |  |  |
|  | **DC1404-058** |  |  |  |  |  |  |  |  |  |  |  | **3.16±0.18** |  |  |

<Table 4>

Table 4. Apatite (U–Th)/He replicate analyses from the DCS antiform

|  |  |  |  |  |  |  |  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- |
| **Sample ID** | **Latitude** | **Longitude** | **Elevation****(m)** | **Rock****type** | **Grains****(N)** | **ρs****(104com-2)** | **ρi****(106com-2)** | **ρs/ρi** | **U****(ppm)** | **D****(%)** | **P(**χ2**)****(%)** | **Central Age****(Ma±1σ)** |
| **The N-S vertical transect** |
| DC1404-058 | 25o 36’ 57.53”N | 100 o 05’ 08.62”E | 2500 | GM | 24 | 4.433 | 0.8675 | 0.0511 | 11.7 | 19 | 95 | 10.16±2.22 |
| DC1404-061 | 25o 36’ 39.18” N | 100 o 05’ 05.13” E | 2340 | GM | 21 | 5.484 | 2.518 | 0.0218 | 34.4 | 0 | 99 | 4.51±1.00 |
| DC1506-002 | 25o 34’ 11.82” N | 100 o 11’ 07.43” E | 1950 | GM | 23 | 3.975 | 2.825 | 0.0141 | 38.9 | 0 | 100 | 2.67±0.56 |
| DC1506-003 | 25o 34’ 30.35” N | 100 o 08’ 46.35” E | 1865 | GM | 22 | 7.023 | 2.771 | 0.0186 | 52.4 | 1 | 96 | 3.28±0.61 |
| DC1506-005 | 25o 33’ 41.99” N | 100 o 07’ 16.97” E | 1659 | GM | 22 | 4.951 | 1.414 | 0.0350 | 19.8 | 0 | 100 | 7.31±1.47 |
| **The E-W vertical transect** |
| DC1506-011 | 25 o 39’ 48.24” N | 100 o 05’ 51.16” E | 3960 | MG | 17 | 8.726 | 1.177 | 0.0742 | 16.6 | 25 | 21 | 14.96±3.32 |
| DC1506-013 | 25 o 39’ 37.12” N | 100 o 06’ 26.28” E | 3855 | GM | 24 | 2.215 | 0.6332 | 0.0350 | 9 | 35 | 33 | 6.44±1.83 |
| DC1511-005 | 25 o 39’ 36.53” N | 100 o 07’ 47.33” E | 2710 | GM | 23 | 4.846 | 1.353 | 0.0358 | 20 | 0 | 97 | 6.72±1.23 |
| DC1511-001 | 25 o 38’ 55.13” N | 100 o 08’ 31.23” E | 2600 | GM | 24 | 5.699 | 1.291 | 0.0442 | 19 | 0 | 99 | 8.19±1.38 |
| *Note.* Standard (**ρs**) and induced (**ρi**) track densities were measured using external mica detectors. Ages were calculated using the zeta method (351.6 ± 30.9 A.F.) for dosimeter glass. Grains (N) = the number of grains counted; P(χ2) = the Chi-square probability that the single-grain ages represent one population; D = age dispersion;. GM = granitic mylonite; MG = mylonitic gneiss. |

<Table 5>

Table 5. Composite ages of samples from the DCS antiform.

|  |  |  |  |  |
| --- | --- | --- | --- | --- |
| **Sample ID** | **Elevation** **(m)** | **AHE** |  | **AFT** |
| **Replicates (n)** | **Mean ages (Ma±1σ)** |  | **Central Age (Ma±1σ)** |
| The N-S vertical transect |
| DC1506-005 | 1659 | 2 | 2.31±0.39 | 7.31±1.47 |
| DC1506-003 | 1865 | 3 | 2.47±0.12 | 3.28±0.61 |
| DC1506-002 | 1950 | 3 | 3.06±0.15 | 2.67±0.56 |
| DC1404-061 | 2340 | 3 | 2.84±0.10 | 4.51±1.00 |
| DC1404-058 | 2500 | 3 | 3.16±0.18 | 10.16±2.22 |
| The E-W vertical transect |
| DC1511-001 | 2600 | 2 | 2.72±0.01 | 8.19±1.38 |
| DC1511-005 | 2710 | 3 | 3.07±0.21 | 6.72±1.23 |
| DC1506-013 | 3855 | 2 | 3.00±0.10 | 6.44±1.83 |
| DC1506-011 | 3960 | 3 | 3.17±0.18 | 14.96±3.32 |

5. Discussion and tectonic implications

5.1. Interpretation of thermochronological data

The 40Ar/39Ar, AFT, and AHe ages obtained for the Diancangshan massif differ significantly; therefore, they relate to different aspects of the geodynamic evolution of the massif.

Within the core of the massif, the microstructural characteristics of quartz and feldspar grains suggest that subhorizontal ductile shear of the core occurred at temperatures of at least 500 °C. Deformation at a temperature exceeding 600 °C has also been inferred from P–T estimations from amphibole and plagioclase grains (Leloup et al., 2001; Gilley et al., 2003), and microstructural analyses in the gneiss core (Cao et al., 2010). The estimated deformation temperatures of the dated porphyroclastic muscovite samples (Table S1 in the Appendix) are higher than the closure temperature range of muscovite (typically 350–400 °C for muscovite; Dunlap, 1997; McDougall and Harrison, 1999). Therefore, the plateau ages of 36.0 ± 0.8 and 29.7 ± 0.6 Ma for muscovite from DCS-GL-069 and DCS-GL-090, respectively, are interpreted here as the cooling ages following formation of the subhorizontal foliation. The plateau age of 36.0 ± 0.8 Ma from the core is the oldest cooling age obtained for the massif. The plateau age of 29.7 ± 0.6 Ma for muscovite DCS-GL-090, close to DCS-GL-069 sampling site, is also interpreted as the age at which the muscovite cooled through it closure temperature after the subhorizontal ductile shearing. The age difference between two muscovite samples, corresponding to different sampling sites (Figure 2b), is significant (~5 Myr), perhaps reflecting an inhomogeneous cooling or exhumation history associated with the ARSZ. Thus, it is inferred that the obtained plateau age of 29.7 ± 0.6 Ma of muscovite may be close to the onset timing of shear deformation associated with the sinistral strike-slip movement.

The plateau ages of 24.7 ± 0.3 and 25.6 ± 0.3 Ma for muscovite from DCS-GL-049 and DCS-GL-050, respectively, from the shear zone can be interpreted here as close to the cooling age of the ARSZ strike-slip shear. The estimated deformation temperature (ranging from 400 to 500 °C) for the two samples is close to the upper limit of the closure temperature range of muscovite. Therefore, we infer that the plateau ages of 24.7 ± 0.3 and 25.6 ± 0.3 Ma for the two samples are also close to the time of deformation associated with the ARSZ shearing movement along the massif. These ages are similar to the muscovite 40Ar/39Ar ages (23.6 to 22.8 Ma) obtained by Cao et al. (2011) for mylonite samples in the shear zone of the massif. Mica ages are remarkably consistent throughout the sinistral shear zone and indicate that presently exposed rocks all cooled through 400–350 °C at 25–22 Ma, corresponding to contemporaneous strike-slip shearing. The older muscovite ages (such as 29.7 ± 0.6 Ma) from the samples of the core and younger muscovite ages for the strike-slip shear zones in the margin of the massif can indicate that inhomogeneous cooling events associated with the sinistral strike-slip shear or strain partitioning across and along the ARSZ.

The plateau ages of 23.3 ± 0.3 and 17.1 ± 0.2 Ma for muscovite from DCS-GL-076 and DCS-GL-104, respectively, from the extensional shearing zone at the southern margin of the massif, are interpreted as cooling ages or deformation ages very close to the timing of sinistral or top-to-the-SSE or -SE extensional shearing, as deformation temperature of extensional shear movement (ranging from 400 to 500 °C) is roughly equal to the closure temperature of muscovite. The age difference between these two samples is significant, about 5.2 Myr, perhaps reflecting different times of activity on distinct shear zones, or inhomogeneous shear heating within one extensional deformation system (Kramar et al., 2001; Zhu et al., 2005; Rolland et al., 2009). The difference between the two muscovite ages also indicates a low cooling rate during or after the ASRZ strike-slip shearing.

The muscovite 40Ar/39Ar ages range from late Eocene to early Miocene. The oldest cooling ages obtained from the core of the Diancangshan massif are 36 Ma. It is concluded on the basis of these 40Ar/39Ar ages that horizontal tectonic planes in the deep crustal layers formed before the early Oligocene, well before the sinistral strike-slip movement (25–17 Ma). The weakly concordant ages of 25.6–24.7 Ma for sinistral strike-slip shearing and 23.3–17.2 Ma for extensional shearing (Figures 11b), taken from widely separated localities, imply essential exhumation and cooling of this structural level through ~400 °C across the massif.

 In contrast to the Ar data, the AFT results are extremely dispersed, without correlation between age and altitude (Figure 12). Figure 12a shows an age vs. altitude diagram for AFT data of the E–W transect. Central AFT ages along the E–W transect range from 8.19–6.72 Ma at 2600–2700 m to 14.96–6.44 Ma at elevations approaching 4000 m (Figure 12a; Table 4). Central AFT ages for the southern part of the massif range from 10 to 5 Ma (except for samples DC1506-002 and DC1506-003). However, in spite of a large altitude difference between the top and the base of the Diancangshan for the E–W transect (>1300 m), the pattern doesnot define a clear, positively sloped trend for these AFT data. Instead, three adjacent ages (except for sample DC1506-011 at the top) overlap significantly within 2σ error limits and within a narrow range from 8–6 Ma (Figure 12a). As stated above in section E-W, the AFT data show no correlation between age and altitude, which suggests that simple erosion is not responsible for the distribution of ages. Their dispersion and absence of altitude/age relationship could rather indicate a low exhumation rate between 15 and 5 Ma, with most samples close to the PAZ and differences in median ages related to local conditions (e.g., differences in chemistry, location of the samples, or thermal anomalies). This would roughly fit the history already suggested by Leloup et al. (1993; 2001) and Harrison et al. (1996) with a pause in exhumation shortly after the end of sinistral strike-slip shearing along the ARSZ (ca. 17-15 Ma), and the late exhumation phase (ca. 5-0 Ma) related to dextral normal faulting reactivation of the Red River fault.

All adjacent AHe ages overlap significantly within a narrow range, from 3 to 2 Ma (Figure 12a) in the E–W transect. A similar pattern is observed for AHe ages in the N–S transect. This pattern indicates a high cooling and exhumation rate during the middle and late Pliocene (between ~3 and 2 Ma) for the southern parts of the Diancangshan massif. This interpretation is also supported by the two AFT ages (3.28 Ma for DC1506-003 and 2.67 Ma for DC1506-002) obtained for the N–S transect (Table 5), which correspond to AHe ages of 2.47 and 3.06 Ma, respectively (Figure 12b). The two AFT ages are identical within error to the AHe ages (Figure 12b and Table 5). The similar age distribution for the AFT and AHe ages is strongly indicative of rapid cooling since ~3 Ma (Viola and Anczkiewicz, 2008). Our AHe results clearly document a rapid cooling between ~ 3 and 2 Ma. It represents the final episode of exhumation of the Diancangshan massif that can start before 3.2 Ma, or likely ~5 Ma as proposed by Leloup et al. (1993; 2001). This phase exhumation of the massif can be related to dextral strike-slip movement of the Weixi-Qiaohou and Red River faults with normal faulting along the southwestern and northeastern edges of the Diancangshan massif (Leloup et al., 1993).

5.2. Structural and thermal evolution

Recognition of the Diancangshan antiform provides critical evidence for interpreting the structure of the Diancangshan massif and the evolution of the ARSZ (Yan et al., 2021 and this study). Leloup et al. (1993, 1995) interpreted the massif as a sinistral transpressional zone and did not regard the regional-scale antiform as a key structure in the massif. The horizontal foliation in the gneiss core had been presumed to be due either to the vertical flattening (kinematically linked to transtension) of an earlier, originally vertical foliation, or to the preservation of a relic of an earlier, extensively reworked structure within the ARSZ (Leloup et al., 1993, 1995, 2001). Current publication also inferred that the lateral subhorizontal middle to lower crustal flow contributed to the horizontal foliation in the Diancangshan antiform (Yan et al., 2021). Those authors also argued that the cooling phases of the metamorphic massifs within the ARSZ were related entirely to the strike-slip shear activity of the shear zone during the Oligocene–Miocene (Leloup et al., 1993, 2001; Wang et al., 1998; Cao et al., 2011; Yan et al., 2021). However, our 40Ar/39Ar geochronological data from the core of the Diancangshan massif, which range from 36 to 29 Ma, record the oldest 40Ar/39Ar cooling ages of the massif. We alternatively suggest that the domal core could have already cooled to about 350–400 °C by the late Eocene or early Oligocene. The data show that the central portion of the massif was already high in the crust when the greenschist-facies strike-slip shear activity of the ARSZ occurred. Therefore, the new 40Ar/39Ar data obtained further document that the horizontal foliation of the Diancangshan core rocks can be first formed, antiformed at depth and experienced an early cooling, then subsequently reworked and further exhumed by a steep sinistral strike-slip shear zone along the margins of the Diancangshan. Namely, the simple shear-dominated strain is strongly localized in the limbs of the Diancangshan massif. The horizontal shear foliations may reflect horizontal shearing layers in the deep crust, consistent with horizontal shearing movement at ca. 33 Ma in the Day Nui Con Voi massif documented by Jolivet et al. (2001) (Figure 13a). Similar horizontal shearing layers in the deep crust have been investigated within other exhumed high-strain zones in the Indochina Block, including the Xuelongshan (Zhang et al., 2014, 2017), Diancangshan (Yan et al., 2021), and Day Nui Con Voi massifs (Jolivet et al., 2001; Anczkiewicz et al., 2007; Yeh et al., 2008), and the Gaoligong shear zone (e.g., Wang and Burchfiel, 1997; Socquet and Pubellier, 2005; Zhang et al., 2010, 2017). It is possible that the horizontal fabrics preserved in the cores of the gneiss antiforms were continuous, forming a regional-scale horizontal shearing plane in the deep crust across the Indochina Block, prior to exhumation and strike-slip shearing of the ARSZ.

Later sinistral shearing along the ARSZ would have extensively disrupted their initial continuity and kinematic during the prolonged exhumation and shearing of the Diancangshan massif. During this period, exhumation of the metamorphic antiform occurred because of the oblique component of slip along the sinistral shear zone (Leloup et al., 1993, 2001) and/or of the extensional component of shear at the southern edge of the massif. On the basis of our new data and previous publications, and provided that the available constraints on the initiation and duration of the sinistral strike-slip shear movement are correct, the uplift and exhumation of the massif related to sinistral shear movement of the ARSZ are inferred to have started during the early Oligocene (~29 Ma) and lasted until ca. 17 Ma (Figure 13b–c).

The AFT ages obtained during this study range from 15 to 5 Ma, indicating that another cooling interval started during the mid-Miocene and ended at ~5 Ma in the Diancangshan massif. This interval is characterized by a slow cooling rate than the preceding and succeeding cooling phases. These data defined a quiescent interval (quiescent gap) between the two stages of cooling from ca. 15 to 5 Ma at the Diancangshan (Leloup et al., 1993). The quiescent gap can be indicative of very slow exhumation through or stasis in the Diancangshan massif during this interval. Leloup et al. (1993) interpreted that this quiescent gap to correspond to a period during which both no significant unroofing occurred and the geothermal flux remained essentially constant.

Prior to this study, the timing of the onset of the latest stage of rapid cooling and associated brittle normal faulting was estimated to be ca. 4.7 or 5 Ma in the Diancangshan, based on the average plateau age derived from K-feldspar (Leloup and Kienast, 1993). Bergman et al. (1997) published an AFT age of 2.7 ± 0.6 Ma for the footwall of the Weixi–Qiaohou fault, which suggested rapid Pliocene cooling through 60–120 °C. In this study, the southern part of the Diancangshan massif yielded the youngest AHe ages, ranging from ~3 to 2 Ma. We suggest that this part of the massif had already cooled to about 80 °C by the Pliocene, indicating rapid final exhumation and cooling of the Diancangshan massif. The high-angle brittle faults (such as the Weixi–Qiaohou and Red River faults) flanking each limb of the massif are characterized by dextral strike-slip shearing and normal faulting. We interpret these brittle faults as representing a dextral transtensional setting in the region of the Diancangshan and Ailao Shan, which resulted to the final and current exhumation (e.g., Peltzer and Tapponnier, 1988; Wang and Burchfiel, 1997; Viola et al., 2003; Morley, 2007; Wang et al., 2016).

5.3. Kinematic transition in the crust

Previous publications had suggested that the early transpressional deformation in western Yunnan and northern Vietnam may have been the cause of local crustal thickening along the Yulong–Habashan (Lacassin et al., 1996), Xuelongshan (Zhang et al., 2014, 2017), Diancangshan (Wang and Burchfiel, 1997; Gilley et al., 2003; Liu et al., 2014; Yan et al., 2021; and this study), and Ailao Shan antiforms (Wang and Burchfiel, 1997). The local crustal thickening could have occurred during the Eocene or before the late Oligocene in response to India–Asia oblique collision (e.g., Lacassin et al., 1996; Wang and Burchfiel, 1997; Royden et al., 1997; Clark and Royden, 2000; Jolivet et al., 2001; Gilley et al., 2003; Clark, 2011). The continuation of oblique convergence between India and Asia imposed overall shortening in the Indochina Block (Wang and Burchfiel, 1997), which was accommodated by NW-trending structures. This shortening caused horizontal ductile layers to dome to form regional-scale anticlines with approximately NW-striking fold axes in the deep crust along the boundary between the South China and Indochina blocks (Figure 13a). Therefore, the Xuelongshan, Diancangshan, Ailao Shan, and Day Nui Con Voi metamorphic massifs, defining a regional gneiss antiform zone, represent distributed deformation in a portion of the Indochina Bock during the Cenozoic.

Local thickening along domal antiforms is one of the likely causes of the metamorphism and melting of deep crustal rocks in southwestern Yunnan and northern Vietnam (Lacassin et al., 1996; Chung et al., 1997; Jolivet et al., 2001; Anczkiewicz and Thirlwall, 2003; Searle, 2013; Zhang et al., 2017). The early phase of doming structures may have lasted until the late Eocene (ca. 36 Ma) for Yulong Shan (Lacassin et al., 1996), until the early Oligocene (ca. 32 Ma) for the Xuelongshan massif (Zhang et al., 2017), until ca. 36 or 29 Ma at the Diancangshan massif in the present study, and until ca. 33 Ma for the Day Nui Con Voi massif (Jolivet et al., 2001), when the tectonic regime was transformed from a compressive setting to an extensional or transtensional setting (Figure 13b; Jolivet et al., 2001). We propose that, during the Eocene and early Oligocene, exhumation of high-temperature rocks of the Diancangshan antiformal massif occurred as a result of the transition to a sinistral transtensional or extensional setting (Figure 13b).

The transition from transpression to transtension can be attributed to a single mechanism, namely lithospheric thinning (Chung et al., 1997).Upwelling of the asthenosphere due to lithospheric thinning can explain the early exhumation of the gneiss massifs during the Eocene and early Oligocene (Figure 13b) (Anczkiewicz et al., 2007). As proposed by Anczkiewicz et al. (2007), progressive dilation within the ARSZ accommodated infilling by mid-crustal gneisses. In the subsequent deformation (between 25-17 Ma or early since 27 Ma), the releasing bend of the ARSZ acting as a sinistral transtensional shear zone further facilitated and accommodated the vertical ascent of the Diancangshan massif along the ARSZ (Figure 13c; e.g., Leloup et al., 1993, 1995).

Jolivet et al. (2001) and Searle (2006) proposed that the ARSZ does not cut through the entire lithosphere; we follow this interpretation. Our data suggest that the sinistral strike-slip shear deformation along the ARSZ did not affect the deep crust at least before the early Oligocene in western Yunnan. Instead, the sinistral strike-slip shear zone remained localized in the upper crust (giving rise to greenschist-facies conditions) above a deep crustal domain in which a horizontal shear layer or antiformal structure was able to dominate (Jolivet et al., 2001; Searle, 2006; Anczkiewicz et al., 2007; Zhang et al., 2014).

The currently published low-temperature thermochronological data have presented a northwestward propagation of kinematic reversal and cooling events since the mid-late Miocene along the various metamorphic massifs within the ARSZ (Leloup et al., 1993; seeing Figure 7 in Wang et al., 2020). Dextral strike-slip movement with a dip-slip component along the Red River fault can have begun at 14-10 Ma in the Ailao Shan region (Wang et al., 2016; 2020). Dextral normal faulting along the eastern and western limbs of the Diancangshan massif initiated at ~5 Ma by Leloup et al. (1993) and at ~3 Ma in this study, or at ~8 Ma by Wang et al. (2020). To the northwest, dextral transtension of the Weixi-Qiaohou fault along the Xuelongshan massif began at ~5 Ma (Wang et al., 2020). Assuming E–W-directed plane strain, the present-day south-directed global positioning system (GPS) velocities and ENE–WSW extensional strain field (relative to a stable Eurasia) in the study area are consistent with regional E–W extension since the Pliocene (Figure 13d) (Gan et al., 2007). The cooling trend yielded by our AHe and AFT data suggests that the most recent uplift and cooling phase occurred under similar kinematic conditions to those that dominated during Pliocene–Quaternary E–W-directed extension in the Eastern Himalaya Syntaxis region and western Yunnan (Figures 13d). This pulse is also coeval along the Xianhuihe–Xiaojiang fault and with the eastward propagation of deformation from Tibet (Clark et al., 2005; Liu-Zheng et al., 2008; Wang et al., 2012), which formed the Erhai, Midu, Fengqing, Lijiang–Heqing, Binchuan, and Qiaohou basins (BGMRYP, 1987; Wan et al., 1997), suggesting along-strike continuity in the timing of vertical exhumation of crust. We interpret the change in structural style/kinematics along the large-scale strike-slip shear zone at this stage to reflect a tectonic transfer from strain localization along the structural boundaries to continuous deformation corresponding to plateau growth and expansion.

<Figure 13>

6. Conclusion

Structural and kinematic observations and thermochronological data for the Diancangshan massif allow us to establish the deformation type and a geodynamic scenario for the evolution of the ductile fabrics of the ARSZ, as follows:

1. The Diancangshan metamorphic massif, a major segment of the ARSZ, is framed by a NW–SE-trending large-scale antiform. The antiform is dominated by gneisses and migmatitic gneisses in the core, which is bounded by two sinistral ductile shear zones, developed under greenschist-facies conditions, at the eastern and western sides of the massif. The eastern branch of the ARSZ shows E-dipping foliations with a horizontal mineral lineation and a sinistral sense of shear, and records middle- to upper-greenschist-facies conditions. The west branch of the ARSZ has W-dipping foliations with a sinistral sense of shear, and preserves middle- to upper-greenschist-facies conditions.
2. Thermal histories derived from integrated muscovite 40Ar/39Ar, AFT, and AHe dating place important constraints on the timing of the kinematics and cooling of the massif. 40Ar/39Ar thermochronological data, ranging from 36 to 29 Ma for muscovite samples from the core, record cooling ages when the muscovites passed through their closure temperatures. The gneiss core of the massif yielded the oldest 40Ar/39Ar ages, on which basis we suggest that the core had already cooled to about 350–400 °C by the late Eocene or early Oligocene. Muscovite 40Ar/39Ar ages in the ranges 25–24 Ma and 23–17 Ma are considered to mark the further cooling and activity time of sinistral strike-slip and extensional shearing of the ARSZ. Therefore, these data indicate that the Diancangshan antiform was already at a shallow level of the crust when greenschist-facies strike-slip shear deformation of the ARSZ occurred. The thermal histories of the massif reflected two stage exhumations of the Diancangshan massif before the middle Miocene.
3. The southern part of the Diancangshan yielded AFT ages of 15–5 Ma, from which we infer that this part had already cooled to ~100 °C by the middle to late Miocene. These data define a quiescent gap, which can be indicative of very slow exhumation through or stasis in the Diancangshan massif during this interval. The youngest AHe ages, ranging from ~3 to 2 Ma, indicate that rapid final exhumation and cooling of the Diancangshan massif since the Pliocene. This phase exhumation of the massif can be related to dextral strike-slip movement of the Weixi-Qiaohou and Red River faults associated to normal faulting along the edges of the massif.
4. During the Eocene, deformation was dominated by horizontal shearing and antiformal doming within the deep crust, characterized by local crustal thickening in the Indochina Block, corresponding to continental collision between India and Asia. During the Eocene and early Oligocene, the high-temperature rocks of the Diancangshan antiform were exhumed and cooled as a consequence of a transtensional or extensional tectonic regime in southeastern Asia. Sinistral strike-slip shearing in the upper crust, characterized by strain localization along shear zones/faults, favored further exhumation of metamorphic rocks within the regional-scale strike-slip shear zones from late Oligocene to mid-Miocene. We suggest the change in structural style/kinematics along the large-scale strike-slip shear zone at Pliocene or slightly earlier to reflect a tectonic transfer from strain localization along the regional-scale tectonic boundaries to continuous deformation corresponding to plateau growth and expansion.

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Figure 1. Regional tectonics and topography of the southeastern margin of the Tibetan Plateau, modified from Tapponnier et al. (1990), Leloup et al. (1995), Ding et al. (2005), Searle (2006), Yin (2006), and Xu et al. (2015). (a)–(b) Maps showing major Cenozoic strike-slip shear zones/faults in southeastern Asia. Major shear zones include the Xuelongshan, Diancangshan, Ailao Shan, and Day Nui Con Voi shear zones, collectively forming the ARSZ, as well as the Chongshan and Gaoligong shear zones. The dextral strike-slip Red River fault extends along the northeastern margin of the ARSZ. (c) Seismic tomographic image across the Indochina and South China blocks (Liu et al., 2000), illustrating the relationship between major strike-slip shear zones/faults and lithospheric structures. Location of the seismic tomographic image marked in Figure 1a. M = Moho, L = base of the lithosphere, ARSZ = Ailao Shan–Red River Shear Zone, CSSZ = Chongshan Shear Zone, RRF = Red River fault. Blue and red correspond to areas of relatively high and low seismic velocity, respectively.

Figure 2. Geological map showing the major structures of the Diancangshan massif and adjacent region. (a) Simplified map of the main structural elements of the massif and surrounding region, based on our own mapping, interpretation of the 1:200,000 geological map of the Dali region (BGMRYP, 1987), and previous publications (Leloup et al., 1995; Searle et al., 2010; Liu et al., 2012). (b) Structures along the Diancangshan massif, showing the antiformal core and the sinistral strike-slip zones and normal faults bounding the massif. Mapping of these structures was based on field observations and topographic data obtained from digital elevation models.

Figure 3. Structural geometry, fabrics, and kinematics along four geological cross-sections across the Diancangshan massif (see Figure 2b for locations). (a)–(d) First-order gneissic antiform bound to the east and west by the sinistral strike-slip shear zones of the ARSZ. The accompanying stereoplots show tectonic foliation orientations (great circles) and mineral stretching lineations (black dots) observed within the antiform. All stereoplots are equal-area, Schmidt net, lower-hemisphere projections. (e) Simplified conceptual model illustrating spatial relationships between various metamorphic rock units and structural and kinematic features in the footwall and hanging wall of the Xiaguan low-angle normal fault.

Figure 4. Typical structures of migmatitic gneisses in the core of the Diancangshan antiform. (a)–(c) Gently to moderately dipping foliation of migmatitic gneisses. (d) Folded migmatite with subhorizontal layers in the host gneisses. (e) Mylonitic gneisses and leucogranitic veins with a gently dipping foliation. (f) Folded stromatic migmatite with leucosome, mesosome, and melanosome (collapse and rockfall at field site DCS-GL-89). (g) Mylonitic gneiss with a gently dipping foliation. (h)–(k) Microstructures showing that the foliation is defined by alternating quartzofeldspathic-rich and mica-rich layers. (h) Microstructure and mineral mapping obtained using a TESCAN Integrated Mineral Analyzer (TIMA). A top-to-the-SE sense of shear is indicated by asymmetric orthoclase porphyroblasts, mica fish, and S–C fabrics. (i)–(k) Optical photomicrographs of quartz and feldspar microstructures. All photomicrographs were obtained under cross-polarized light in lineation-parallel (XZ) thin sections. (i)–(j) Deformed quartz and plagioclase porphyroclasts in the mylonitic gneiss of the core, showing high-temperature grain-boundary migration (GBM) microstructures in quartz, including bulging (BG) and subgrain rotation (SGR) recrystallization in quartz and plagioclase grains. (i) Plagioclase grains showing crenulate margins that are indented against quartz grains (red arrows in sample DCS-GL-69); (j) Microboudinaged plagioclase porphyroclast from the banded gneiss showing the inter-boudin dilation space infilled with fine-grained quartz and feldspar grains. (k) Deformed quartz porphyroclasts showing deformation bands and chessboard subgrain patterns. Note: TIMA is a system in which scanning electron microscopy is linked with energy dispersive spectrometry, enabling rapid, automated, quantitative mineralogical analysis.

Figure 5. Mylonitic structures within sinistral shear zones in the western and eastern flanks of the Diancangshan antiform. (a)–(b) Mylonitic structures within the eastern flank of the antiform, near Dali. The dominant foliation within the mylonitized granite and schist dips steeply toward the NE or ENE and contains a subhorizontal stretching lineation. (c)–(f) Mylonitic structures in granitic gneiss within the western flank of the antiform. Mylonitic foliation dips steeply to the SW and contains a subhorizontal stretching lineation that plunges toward the SE. Inset stereoplots are lower-hemisphere, equal-area projections displaying mylonitic foliation and lineation data collected at the field sites.

Figure 6. Structures and kinematics of tectonites in the shear zones. (a)–(b) Mica-rich layers and quartzofeldspathic porphyroclasts defining the steep foliation in the western and eastern shear zones of the Diancangshan antiform. (c) Three-dimensional view of the steeply dipping mylonitic foliation with a subhorizontal lineation in the shear zone. (d) Asymmetric core–mantle structures of quartz porphyroclasts indicating a sinistral sense of shear in the western shear zone. (e) Microstructures showing extensive elongated quartz grains with bulging and subgrain rotation recrystallization, fracturing plagioclase crystals with undulose extinction and deformation lamellae in the western shear zone. (f)–(g) Elongated quartz porphyroclasts showing undulose extinction and irregular grain boundaries formed in response to bulging and subgrain rotation recrystallization in the eastern shear zone. These kinematic features indicate a sinistral sense of shear. The structures were identified at outcrops and in oriented sections parallel to the stretching lineation and normal to the foliation. Sampling sites are shown in Figure 2b.

Figure 7. Structural map of the southernmost edge of the Diancangshan massif, showing the location of the Xiaguan low-angle normal fault (XGLF). Structures were mapped using a combination of field observations and topographic data from digital elevation models. Stereoplots are lower-hemisphere, equal-area projections illustrating the dominant structural fabrics and trends in the mapping area, as measured at locations along the Xier River. Line H–H’ marks the location of the cross-section shown in Figure 9a.

Figure 8. Ductile deformation within the mylonitic granite and schist in the footwall of the Xiaguan low-angle normal fault. (a)–(d) Foliation in the mylonitic granite dipping to the south, subparallel to the dip of the Xiaguan fault. Kinematic indicators indicate a top-to-the-SSE sense of shear. (c)–(g) Structures and kinematics in the southern margin of the Diancangshan massif. Asymmetric feldspar or quartz porphyroclasts and folded leucogranitic veins within the mylonitic granite demonstrating top-to-the-SSE shear. (f)–(g) Typical fabrics of dynamic recrystallization in quartz. Relicts of large old quartz grains show undulose extinction and elongated shapes. New grains formed by bulging and subgrain rotation recrystallization. These observed kinematic features indicate a top-to-the-SSE sense of shear. The structures and kinematics were identified at outcrops and in oriented sections (under cross-polarized light), parallel to the stretching lineation and normal to the foliation. Inset stereoplot is a lower-hemisphere, equal-area projection.

Figure 9. Structural and kinematic characteristics of the Xiaguan low-angle normal fault. (a) Structures as observed along a NW-SE-oriented slip-parallel profile. Photographs and diagrams show variations in the characteristics of the footwall cataclasites. The density of shear bands increases progressively toward the fault. Location of section H–H’ shown in Figure 7. (b) Photograph and structural interpretation of an outcrop of the Xiaguan low-angle normal fault near Xiaguan. The fault separates weakly metamorphosed sedimentary rocks in the hangingwall (upper unit) from cataclasites and mylonitized granite in the footwall (lower unit). Steeply dipping normal faults in the upper unit root onto the low-angle fault surface. Inset stereoplots are lower-hemisphere, equal-area projections displaying structural data from the upper and lower units.

Figure 10. Typical cataclastic microstructures within deformed porphyritic granite in the footwall of the Xiaguan low-angle normal fault. K-feldspar porphyroclasts have been pervasively fractured under tension, resulting in subangular to subrounded clasts in a matrix of quartz, albite, and minor mica and chlorite. Fractures are filled with quartz, albite, mica, and minor chlorite. Photomicrographs were obtained using an SEM-based backscattered electron detector (BSED) and a TIMA. Sampling location corresponds to that of image DC15-06-005 in Figure 9a, and oriented thin sections were cut parallel to the XZ plane. (a) Cataclastic flow of fine-grained quartz. (b) K-feldspar fragments separated from the matrix by a continuous foliation of muscovite ± chlorite + quartz. (c) BSED image showing the replacement of biotite by intergrowths of chlorite and rutile. Oriented biotite grains (medium white) and foliated chlorite (gray) define a C-plane fabric. (d)–(f) Fragmented K-feldspar (orthoclase) porphyroclasts surrounded by quartz and partially replaced by albite. Inset in (d) is a cathodoluminescence image of microbreccia comprising angular clasts of K-feldspar. Clusters of K-feldspar grains are aligned subparallel to C-planes (dashed lines), which are defined by bands of quartz, mica, and albite. Albite and quartz also occur in orthoclase boudin necks, thereby cementing the orthoclase fragments.

Figure 11. Muscovite 40Ar/39Ar data for samples from the Diancangshan massif. (a) Spectra and inverse isochron plots for muscovite 40Ar/39Ar data. (b) Spatial distribution of obtained ages within the footwall of the Xiaguan low-angle normal fault.

Figure 12. Elevation-age relationships for AFT and AHe data from two transects across the Diancangshan massif. (a) Sampling elevation plotted against sample age along the E–W sampling transect. (b) Sampling elevation plotted against sample age along the N–S transect in the footwall of the Xiaguan low-angle normal fault.

Figure 13. Proposed three-dimensional tectonic model of the Diancangshan antiform along the ARSZ during the Cenozoic. (a) Emplacement of horizontal shearing corresponding to NE–SW oblique India–Asia collision since 55 Ma. Formation of a NW-striking antiformal structure along the tectonic boundary between the Indochina and South China blocks, corresponding to regional transpression (Lacassin et al., 1996; Jolivet et al., 2001). Green arrows mark NE–SW regional compression. (b) Transformation from transpression to transtension during the late Eocene or early Oligocene in southeastern Asia, causing the exhumation of antiformal ductile crust (Jolivet et al., 2001; Anczkiewicz et al., 2007). Green arrows mark NE–SW regional compression. (c) Development of a NW-striking sinistral strike-slip ductile shear zone since 25 Ma or early 27 Ma in a regional sinistral transtensional setting. Continued sinistral strike-slip shearing, combined with local extension at the releasing bend zone, which caused further exhumation of the massif from 25 to 17 Ma. (d) The diagram presents a three-dimensional interpretation of the structure in the Eastern Himalayan Syntaxis region and adjacent regions since since ~3.2 Ma, showing relationships between active dextral normal faulting, lower-crustal flow, and the present-day strain field and movement velocities across the Eastern Himalayan Syntaxis region and the DCS region. The present-day strain field and movement velocities across these regions were derived from global positioning system (GPS) data (Gan et al., 2007). The southeast-directed GPS velocity field and ~ENE–WSW-oriented extensional field are bounded by the Xianshuihe–Xiaojiang Fault and the Red River Fault or Nujiang Fault. Variations in buoyancy and the degree of orogen-parallel flow in the lower crust can significantly influence differential surface uplift and resulting landforms, as well as the structural and kinematic characteristics of upper-crustal faults (based on the numerical and analog experiments of Cruden et al., 2006).