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Cenozoic deformation of the Tarim Basin and surrounding ranges (Xinjiang, China): A regional overview

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ABSTRACT

With its central position between the Tibetan Plateau and the Tian Shan Range, the Tarim Basin is a key element of the Cenozoic Asian orogenic system. However, a comprehensive regional study, and more particularly the quantification of shortening through this basin and its margins, are still needed to understand its role in the Cenozoic deformation of Asia. From a compilation of previous works, together with an extensive dataset of satellite, field, seismic and well data, we provide a tectonic map of the Cenozoic structures and four balanced geological transects of the Tarim Basin and its surrounding ranges. Based on this map and these cross-sections, we characterize the Cenozoic deformation of the original Proterozoic Tarim block. From structural restorations and crustal budgets, we also quantify the compressive component of this deformation. Most of the Cenozoic compressive deformation (from ~94% to 100%) is concentrated in the ranges along the block margins. To the west, up to 78 ± 23 km and 54 +24/−18 km of crustal shortening are accommodated across the compressive Western Kunlun and Southwestern Tian Shan ranges, while to the east, up to 38.6 ± 18 km and 15 + 20/−15 km are accommodated across the transpressive Altyn Tagh and Southeastern Tian Shan ranges. A non-negligible amount of compressive deformation (up to ~6%) is also accommodated within the Tarim Basin by large basement-cored uplifts with a vergence synthetic to the deformation of the Tibetan Plateau edge. To the west, the Bachu uplift absorbs ~5 km of the total crustal shortening of the Western Kunlun thrust system, while to the south, the Tanan uplift accumulates ~0.6 km of the Altyn Tagh strike-slip system. Structural inheritance has a major influence on the Cenozoic deformation since ~33.3% to 100% of the total shortening is accommodated by reactivated basement structures inherited from the Protero-Paleozoic history of the Tarim block. Finally, we argue that the basement-cored uplifts in the centre of the basin imply a deformation transfer from the Tibetan Plateau to the Tian Shan, above a deep crustal decollement decoupling the deforming crust from an underlying rigid mantle.

1. Introduction

With the Himalaya-Tibetan Plateau orogenic system and active deformation as far as 3000 km north of the India-Asia collision zone (including the Tian Shan, Altai Range and Baikal rift system) (Fig. 1), Asia is an outstanding natural laboratory to study continental deformation. Following the pioneering studies of the late 1970s and early 1980s (e.g., Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1977; Tapponnier and Molnar, 1979; Molnar, 1984; England and Houseman, 1985), multiple investigations on the Cenozoic Asian mountain ranges, but also on the associated sedimentary basins, which are key recorders of their tectonic and climatic evolution, have been conducted to decipher the modalities of this deformation (e.g., Avouac et al., 1993; Matte et al., 1996; Métivier et al., 1998; Lavé and Avouac, 2001; Tapponnier et al., 2001; Yang and Liu, 2002; Yin et al., 2002; Guillot et al., 2003; Clark et al., 2006; DeCelles et al., 2007; Liu-Zeng et al., 2008; Charreau et al., 2009; Lu et al., 2010; Cheng et al., 2015; Cheng et al., 2016a; Coutand et al., 2016; Gourbet et al., 2015; Jolivet et al., 2018). Among these basins, the Tarim Basin occupies a remarkable central position within the orogenic system, between the Tibetan Plateau and the Tian Shan Range (Fig. 1). With a continuous Proterozoic to Present sedimentation, this wide basin is thus a first-class repository for the

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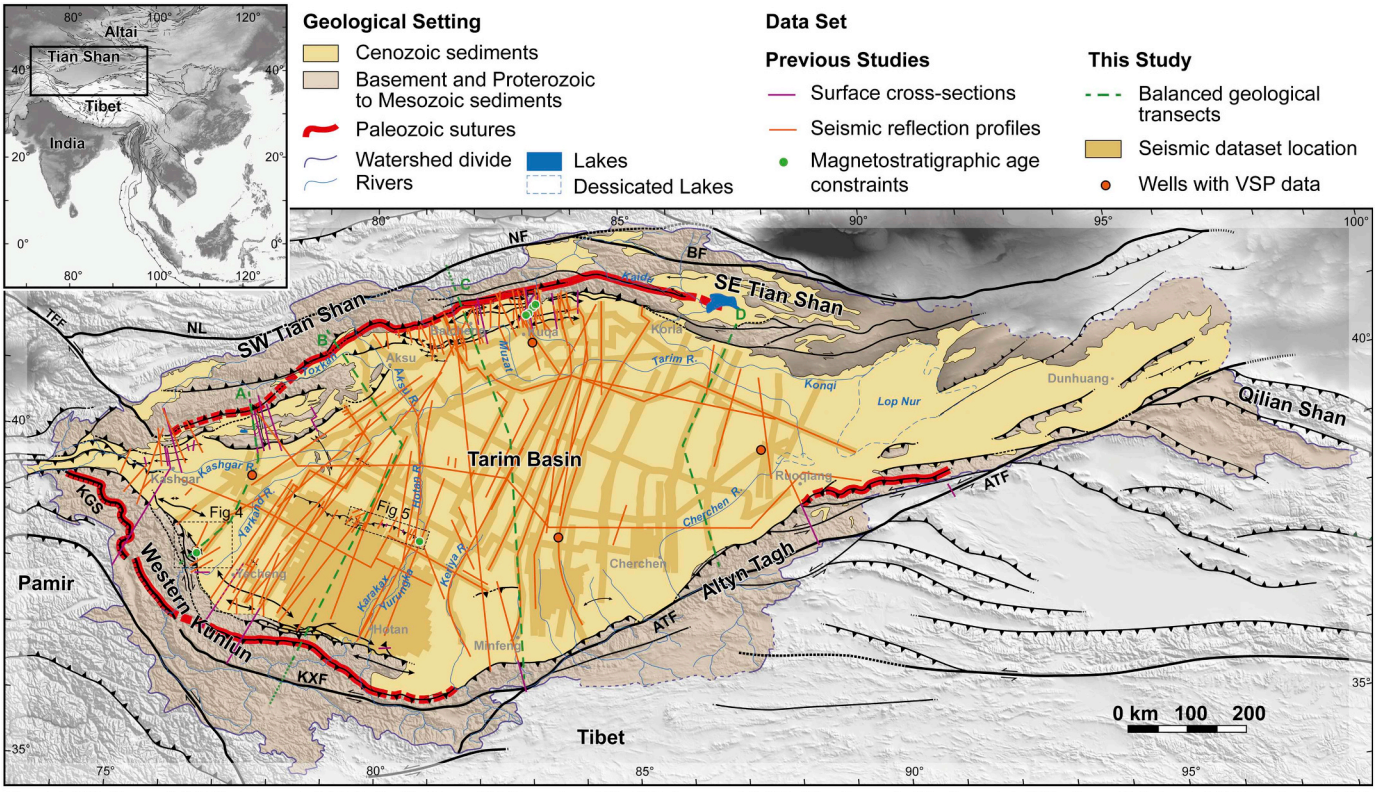

Allen et al., 1999; Blayney et al., 2019; Bosboom et al., 2014a; Burchfield et al., 1999; Chang et al., 2016a, 2017; Charreau et al., 2006, 2009; Chen et al., 2004; Cheng et al., 2016b; Cowgil, 2010; Fan et al., 2008; Fu et al., 2010; Guilbaud et al., 2017; Guo et al., 2016; He et al., 2005; He et al., 2016; Heermance et al., 2007, 2008; Huang et al., 2006, 2010; Hubert-Ferrari et al., 2007; Izquierdo-Llavall et al., 2018; Jia et al., 1998; Jiang et al., 2013; Jiang and Li, 2014; Jiang et al., 2018; Jin et al., 2008; Li et al., 1996; Li et al., 2002a; Li et al., 2012a; Li et al., 2012b; Li et al., 2013; Li et al., 2015c, 2015d, 2016c, 2016d, 2016e; Li et al., 2016b, 2018b; Lin et al., 2012b, 2012c; Lin et al., 2015a; Liu et al., 2015a; Lu et al., 2016; Matte et al., 1996; Neng et al., 2018; Pan et al., 2010; Peng et al., 2006; Qi et al., 2009; Qiu et al., 2011; Scharer et al., 2004; Sobel et al., 2006; Sun et al., 2009a, 2009b, 2017; Tang et al., 2014; Tang et al., 2017; Thompson et al., 2017; Tian et al., 2016; Tong et al., 2012; Turner et al., 2010; Wang et al., 2013; Wang and Wang, 2016; Wang et al., 2016b; Wang et al., 2017; Wei et al., 2013; Wen et al., 2017; Wu et al., 2014; Xie et al., 2017; Xu and Zhou, 2007; Yang et al., 2007a; Yang et al., 2007b; Yang et al., 2009; Yang et al., 2015; Yang et al. 2018b; Yin et al., 1998; Yin et al., 2002; Yu et al., 2014; Zeng et al., 2010; Zhang et al. 2014b; Zhao and Wang, 2016; Zheng et al., 2015

Fig. 1. Geographical and geological setting of the Tarim area and location of the data used in this study. Only the main structures outcropping at the surface are represented. References for the data from previous studies are indicated below the map. Green long-dashed lines locate the regional balanced geological transects A, B, C and D presented in Fig. 7. Green short-dashed lines at terminations of transects B and C correspond to the extended pathway of the synthetic crustal-scale cross-sections presented in Fig. 13. ATF: Altyn Tagh Fault, BF: Baluntay Fault, KFX: Karakax Fault, KGS: Kongur Shan normal fault, NF: Narat Fault, NL: Nikolaev Line, TFF: Talas Ferghana Fault, VSP: Vertical Seismic Profile. In this study, the Tarim block refers to the area comprised in between the Paleozoic ophiolitic sutures (thick red lines) displayed in the surrounding ranges, rather than to the present-day Tarim Basin only. These sutures located within the Western Kunlun, Altyn Tagh and Southern Tian Shan ranges correspond to the Oytag-Kudi, Lapeiquan-Hongliugou, and South Tian Shan sutures, respectively. Chang et al. (2016a); Chang et al. (2017); Fan et al. (2008); Jia et al. (1998); Qiu et al. (2011); Tang et al. (2017); Xu and Zhou (2007); Zheng et al. (2010). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

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geological history of the Asian continent (e.g., Li et al., 1996; Métivier et al., 1999; Yin et al., 2002; Kent-Corson et al., 2009; Sun et al., 2009b; Lin et al., 2012c; Bosboom et al., 2014a; Tang et al., 2014; Lin et al., 2015a; Liu et al., 2015a; He et al., 2016; Zhu et al., 2017; Morin et al., 2018; Wu et al., 2018; Jiang et al., 2018). Furthermore, the Tarim Basin seems to be a key element in the Cenozoic Asian deformation. Indeed, it is usually considered as an undeformed rigid block, playing the role of an heterogeneity in the strain field and/or of a secondary indenter that transmits the compressive stresses at the continental scale far to the north of the India-Asia collision zone (e.g., England and Houseman, 1985; Avouac and Tapponnier, 1993; Neil and Houseman, 1997; Mattauer et al., 1999; Wang et al., 2007; Dayem et al., 2009; Craig et al., 2012; Calignano et al., 2015).

For these reasons, numerous studies have been performed on the Cenozoic deformation and sedimentation along the margins or within the centre of this basin (e.g., Avouac et al., 1993; Matte et al., 1996; Wittlinger et al., 1998, 2004; Allen et al., 1999; Burchfiel et al., 1999; Yin et al., 2002; Scharer et al., 2004; Sun et al., 2005, 2009b; Heermance et al., 2007; Hubert-Ferrari et al., 2007; Fu et al., 2010; Turner et al., 2010; Li et al., 2012a; Jiang et al., 2013; Wei et al., 2013; Chang et al., 2014a; Cheng et al., 2016b; Li et al., 2016b; Li et al., 2016e; Lu et al., 2016; Wang and Wang, 2016; Guilbaud et al., 2017; Thompson Jobe et al., 2017; Izquierdo-Llavall et al., 2018; Jiang et al., 2018). However, they were mainly undertaken to characterize tectonic structures and their activity at a local scale. Moreover, the few studies conducted at a regional scale remain schematic or imprecise in terms of hierarchy and relationships between the structures, with sometimes no detail about which ones were activated or not during the Cenozoic (e.g., Wei et al., 2013; Tang et al., 2014; Lin et al., 2015a; Li et al., 2016e; Jiang et al., 2018). Nevertheless, it is clear that the original Proterozoic Tarim block is not totally rigid, with deformation along the margins and even in the center of the Tarim Basin (e.g., Avouac et al., 1993; Matte et al., 1996; Allen et al., 1999; Wittlinger et al., 2004; Guo et al., 2005; Heermance et al., 2007; Li et al., 2012a; Tong et al., 2012; Chang et al., 2014a; Tang et al., 2014; Fig. 15 of Wang et al., 2014a; Li et al., 2016e; Daout et al., 2018; Izquierdo-Llavall et al., 2018; Jiang et al., 2018). Yet, the amount and distribution of this deformation, as well as the type of structures accommodating it, still need to be precisely documented. A comprehensive regional study, and more particularly the quantification of shortening through this area, is therefore necessary to further discuss these issues, and to understand how the Tarim participates to the propagation of the Cenozoic compressive deformation within Asia.

In this article, we present such a comprehensive study of the Cenozoic deformation at the scale of the whole Tarim Basin and its

surrounding ranges. Thanks to a compilation of existing geological maps and of published cross-sections, together with an extensive dataset of surface (satellite images and field constraints) and subsurface (seismic profiles and well data) data, we hereafter propose a tectonic map and four balanced geological transects highlighting the structures activated during the Cenozoic in the Tarim area. On the basis of this map and these transects, we characterize and quantify the Cenozoic deformation across the whole basin and its margins, as well as its relationships with the surrounding ranges.

2. Geological setting

2.1. Tarim Basin

The Tarim Basin is located in Central Asia, in the Xinjiang Province in China (Fig. 1). It corresponds to a ~1500-km-long and ~700-km-wide active sedimentary basin with an average elevation of ~1200 m (Fig. 1). Presently, this basin is an endorheic compressive basin surrounded by mountain ranges: the Western Kunlun Range to the southwest, the Altyn Tagh Range to the southeast, and the Tian Shan to the north.

This region of Central Asia was initially structured during the Proterozoic and Paleozoic by the accretion of several blocks: the Kunlun and Qaidam blocks to the south, the Tarim block in the centre, and the Central/South Tian Shan block to the north (e.g., Windley et al., 1990; Matte et al., 1996; Sobel and Arnaud, 1999; Zhou et al., 2001; Xiao et al., 2004, 2013; Charvet et al., 2011; Lin et al., 2013; Zhang et al., 2013a; Li et al., 2018a). Prior to this accretion, the tectonic evolution of the Tarim block was dominated by extension from Neoproterozoic to Ordovician (e.g., Li et al., 1996; Zhang et al., 2013a; Tang et al., 2014; Liu et al., 2015a; Wang et al., 2015; Dong et al., 2016; Zhu et al., 2017; Ren et al., 2018; Wu et al., 2018). Then, from Ordovician to Permian, compression generalized in the area due to the closures of oceanic domains, followed by collisions between (1) the Tarim and Kunlun blocks to the southwest (from Ordovician to Devonian), (2) the Tarim and Qaidam blocks to the southeast (from Silurian to Devonian), and (3) the Tarim and Central/South Tian Shan blocks to the north (from Devonian to Permian) (e.g., Windley et al., 1990; Li et al., 1996; Matte et al., 1996; Sobel and Arnaud, 1999; Gehrels et al., 2003; Xiao et al., 2004, 2013; Charvet et al., 2011; Lin et al., 2012c; Lin et al., 2013; Zhang et al., 2013a; Tang et al., 2014; Alexeiev et al., 2015; Liu et al., 2015a, 2015b; Han et al., 2016; He et al., 2016; Li et al., 2018a). Driven by all these oceanic closures and subsequent collisions, Paleozoic ranges comprising ophiolitic sutures (Fig. 1), as well as plutonic and metamorphic rocks, formed around the Tarim block (e.g., Windley et al., 1990; Matte et al., 1996; Gao et al., 1998; Sobel and Arnaud, 1999; Cowgill et al., 2003; Gehrels et al., 2003; Xiao et al., 2005, 2013; Cowgill, 2010; Charvet et al., 2011; Glorie et al., 2011; Cao et al., 2013; Zhang et al., 2014a; Li et al., 2018a). In the contemporaneous Tarim Basin settled in the centre of the block, this Protero-Paleozoic history also induced the development of large structures known as the Bachu, Tanan, Tazhong, Tadong, Tabei, Kalpin, and Kuruk Tagh basement-cored uplifts, which were created in extension at the end of the Proterozoic-beginning of the Paleozoic, before being reactivated in compression (e.g., Li et al., 1996; Allen et al., 1999; Carroll et al., 2001; Turner et al., 2010; Lin et al., 2012b, 2012c; Zhang et al., 2013a; Gao and Fan, 2014; Tang et al., 2014; Lin et al., 2015a; Liu et al., 2015a; He et al., 2016; Zhu et al., 2017; Wu et al., 2018).

At the end of the Paleozoic, the emplacement of a Permian Large Igneous Province (LIP) affected the Tarim block (e.g., Zhang et al., 2008; Zhang et al., 2010; Zhang et al., 2012; Yang et al., 2013; Li et al., 2014a; Xu et al., 2014; Yang et al., 2016; Loury et al., 2018). Afterwards, during the Mesozoic, other blocks (Qiangtang and Lhassa blocks) were accreted further south, in the palaeo-Tibetan Plateau region (e.g., Matte et al., 1996; Yin and Harrison, 2000; Pan et al., 2012; Jolivet, 2017; and references therein). Within the Tarim region, these events produced a complex zone of deformation, with minor strike-slip, compressive or extensive reactivations of the inherited Protero-Paleozoic structures (e.g., Hendrix et al., 1992; Li et al., 1996; Sobel, 1999, 2001; Dumitru et al., 2001; Arnaud et al., 2003; Chen et al., 2003; De Grave et al., 2007; Jolivet et al., 2010, 2013; Wang et al., 2012; Chang et al., 2014b; Tang et al., 2014; Li et al., 2015a; Morin et al., 2018; Yang et al., 2018b). Finally, during the Cenozoic, India collided into the Asian continent around 55–45 Ma (e.g., Wang et al., 2014a and references therein). This last collision led to a major compressive reactivation of the mountain ranges surrounding the Tarim block, as well as of inherited structures within the Tarim Basin (e.g., Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1977, 1979; Windley et al., 1990; Avouac et al., 1993; Allen et al., 1994; Hendrix et al., 1994; Matte et al., 1996; Jolivet et al., 1999, 2001, 2010; Dumitru et al., 2001; Yin et al., 2002; De Grave et al., 2007; Turner et al., 2010; Yin, 2010 and references therein; Tong et al., 2012; Tang et al., 2014; Li et al., 2015b and references therein; Li et al., 2016e; Jiang et al., 2018; Yang et al., 2018b). In this study, because these ranges involve parts of the block itself, we refer to the Tarim block as the area comprised in between the various Paleozoic ophiolitic sutures located in the mountain belts (Fig. 1), rather than as the present-day Tarim Basin only.

Resulting from this long geological history, the Tarim Basin stratigraphy is characterized by a Proterozoic basement composed by igneous rocks, overlain by a thick, more or less continuous series of sediments from Neoproterozoic (Sinian) to Quaternary (Fig. 2) (e.g., Li et al., 1996; Carroll et al., 2001; Yin et al., 2002; Guo et al., 2005; Wang et al., 2009; Shu et al., 2011; Lin et al., 2012a, 2012c; Zhang et al., 2013a; Tang et al., 2014; Guo et al., 2015; Lin et al., 2015a; Liu et al., 2015a; He et al., 2016; Li et al., 2016d; Zheng et al., 2015; Tian et al., 2016; Zhu et al., 2017; Wu et al., 2018). During the Sinian and Paleozoic, marine calcareous and terrigenous deposits were dominant until the Carboniferous (Fig. 2) (e.g., Carroll et al., 2001; Chen and Shi, 2003; Shu et al., 2011; Lin et al., 2012c; Zhang et al., 2013a; Gao and Fan, 2014; Li et al., 2015a; Liu et al., 2015a; He et al., 2016; Zhu et al., 2017; Wu et al., 2018). In the western part of the basin, the Paleozoic sediments also contain an upper Cambrian interval of gypsiferous shales (Fig. 2), which is a major decollement level in this area (e.g., Allen et al., 1999; Lin et al., 2012a, 2012c; Li et al., 2016c, 2016e; Lu et al., 2016; Yang et al., 2018b). Within the Neoprotero-Paleozoic series, field observations, seismic profiles and well data document unconformities, together with sedimentary thickness and facies changes at the top of the basement-cored structures in the Tarim. Combined with thermochronological data, this shows that these structures were already active during the early (Sinian to Devonian) stages of the basin evolution (e.g., Allen et al., 1999; Carroll et al., 2001; Dumitru et al., 2001; Chen and Shi, 2003; Tang et al., 2008; Turner et al., 2010, 2011; Chang et al., 2012b; Lin et al., 2012b, 2012c; Gao and Fan, 2014; Lin et al., 2015a; Liu et al., 2015a; He et al., 2016; Zhang et al., 2016; Zhu et al., 2017; Wu et al., 2018).

During the Permian, a marine to continental transition occurred, with a change from calcareous to terrigenous deposits characterizing most of the Mesozoic sediments (Fig. 2) (e.g., Hendrix et al., 1992; Gao et al., 1998; Sobel, 1999; Carroll et al., 2001; Chen and Shi, 2003; Guo et al., 2015; Li et al., 2015a; Li et al., 2016d; Morin et al., 2018; Peng et al., 2018). A thick (up to ~10 km) and continuous Cenozoic sedimentary series, mainly supplied by the erosion of the surrounding mountain ranges, was then trapped in the contemporaneous compressive Tarim Basin (e.g., Yin et al., 1998, 2002; Métivier and Gaudemer, 1997; Heermance et al., 2007; Wang et al., 2009; Zheng et al., 2015; Tian et al., 2016). From the late Cretaceous to Eocene, the Tarim deposits were mostly continental and terrigenous, except in the western part of the basin where an alternation of marine transgressions/regressions occurred (e.g., Yin et al., 1998; Wei et al., 2013; Bosboom et al., 2014a, 2014b, 2017; Tian et al., 2016; Chen et al., 2018; Zhang et al., 2018; Kaya et al., 2019). In that area, the marine Paleocene-Eocene Kashgar Group is thus composed of gypsum, limestones,

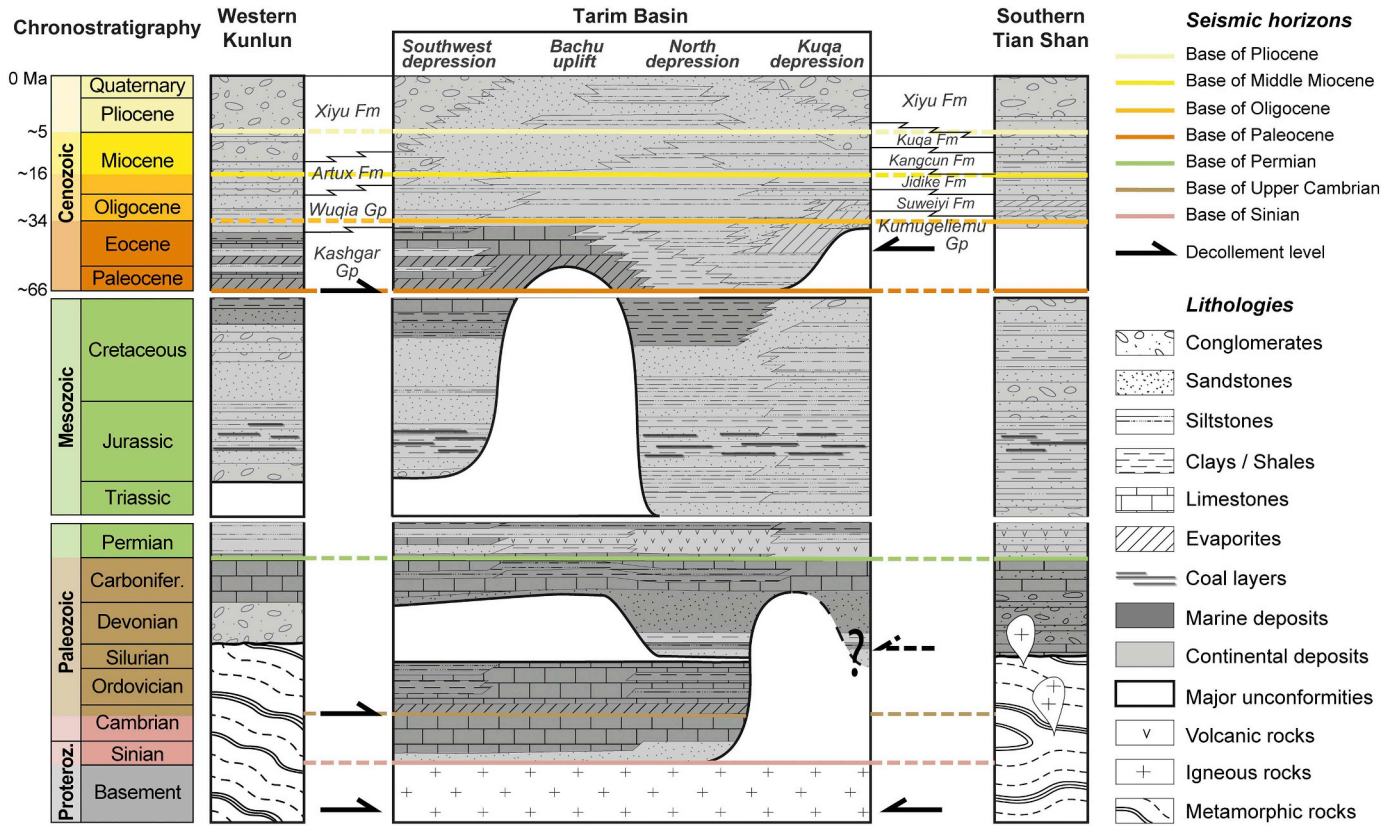


Fig. 2. Simplified chronostratigraphic chart of the Tarim Basin, Western Kunlun Range and Southern Tian Shan. Black arrows represent the main decollement levels in the Tarim area, while colour lines correspond to the seismic horizons picked for this study.

siltstones, and sandstones related to these marine incursions (Fig. 2) (e.g., Jin et al., 2003; Bosboom et al., 2011, 2014a, 2014b, 2017; Wei et al., 2013; Wang et al., 2014b; Zheng et al., 2015; Zhang et al., 2018; Blayney et al., 2019). This Paleocene-Eocene evaporitic assemblage corresponds to another major decollement level in the southwest of the Tarim (Southwest depression) (e.g., Li et al., 2016e; Lu et al., 2016). At the end of the Eocene, the western marine deposits ended while a continental sedimentation finally developed throughout the basin (e.g., Yin et al., 1998, 2002; Jin et al., 2003; Bosboom et al., 2014a; Zheng et al., 2015; Tian et al., 2016; Zhang et al., 2018). During the Oligocene and early Miocene, the sediments mostly consist in claystones, siltstones, and sandstones (Fig. 2) (e.g., Yin et al., 1998, 2002; Jin et al., 2003; Heermance et al., 2007; Wang et al., 2009; Zheng et al., 2015; Tian et al., 2016; Blayney et al., 2019). In these series however, the lacustrine Eocene-Oligocene Kumugeliemu Group and Suweiyi Formation are made up of gypsum, siltstones, sandstones, and a few conglomerates (Fig. 2) (e.g., Yin et al., 1998; Huang et al., 2006; Wang et al., 2009; Tian et al., 2016). This Eocene-Oligocene evaporitic assemblage is also a major decollement level in the north of the basin (Kuqa depression) (e.g., Chen et al., 2004; Li et al., 2012a; Izquierdo-Llavall et al., 2018; Neng et al., 2018). From Miocene to Quaternary, the gravel content in the sediments progressively increased along the margins of the Tarim until the deposition of the conglomeratic Xiyu Formation, while sandstones and siltstones still deposited in the centre of the basin (Fig. 2) (e.g., Yin et al., 1998, 2002; Jin et al., 2003; Sun et al., 2005, 2009b, 2017; Heermance et al., 2007; Jin et al., 2008; Wang et al., 2009; Chang et al., 2012a; Thompson et al., 2015; Zheng et al., 2015; Tian et al., 2016). In compressive contexts, this kind of long-term coarsening-upward sequence is typically controlled by the growth and erosion of the mountain ranges and their progressive thrusting over the adjacent basins (e.g., Puigdefabregas et al., 1986; Sinclair and Allen, 1992; Simoes and Avouac, 2006; Heermance et al., 2007; Charreau et al., 2009; Dubille and Lavé, 2015).

2.2. Western Kunlun Range

The Western Kunlun Range bounds the Tarim Basin to the southwest, between the towns of Kashgar and Hotan (Fig. 1). It is a ~700-km-long, ~100–130-km-wide and up to ~7600-m-high mountain belt at the northwestern edge of the Tibetan Plateau. Nowadays, this mountain belt corresponds to a northeastward-vergent thrust wedge that developed between the Kongur Shan normal fault and the Karakax left-lateral fault to the west and south, and the Tarim Basin to the northeast (Fig. 1) (e.g., Brunel et al., 1994; Matte et al., 1996; Wittlinger et al., 2004; Lu et al., 2016). From its overall orientation, this range can be subdivided into two parts: (1) a NNW-SSE-oriented part from Kashgar to Yecheng to the west, showing a structural and stratigraphic continuity with the northeastern Pamir, and (2) a WNW-ESE-oriented part from Yecheng to Hotan to the east.

Within this mountain belt, the southwestern boundary of the Tarim block is marked by the Oytag-Kudi suture (in red in Fig. 1) related to the closure of the Proto-Tethys Ocean and to the associated Paleozoic orogeny spanning from Ordovician to Devonian (e.g., Matte et al., 1996; Mattern and Schneider, 2000; Xiao et al., 2005; Li et al., 2018a). During the Mesozoic, a complex strike-slip to compressive history then followed this orogeny (e.g., Sobel, 1999; Arnaud et al., 2003; Cao et al., 2015). Finally, a major reactivation of the Western Kunlun Range occurred during the Cenozoic in response to the India–Asia collision (e.g., Tapponnier and Molnar, 1977; Matte et al., 1996; Sobel and Dumitru, 1997; Yin et al., 2002; Cao et al., 2015). Based on thermochronological data (e.g., Sobel et al., 1997; Cao et al., 2013, 2015; Cheng et al., 2017; Blayney et al., 2016; Yang et al., 2018a), sediment provenance and

growth strata (e.g., Yin et al., 2002; Bershaw et al., 2012; Jiang et al., 2013; Cao et al., 2014; Thompson et al., 2015; Cheng et al., 2016b, 2017; Blayney et al., 2016; Sun et al., 2016b), as well as increasing sediment accommodation and accumulation rates in the southwestern Tarim Basin (Métivier and Gaudemer, 1997; Métivier et al., 1999; Yang and Liu, 2002; Jiang and Li, 2014; Blayney et al., 2019), this reactivation seems to initiate by Eocene with subsequent Oligocene to Miocene accelerations. The present thrust wedge then formed above an intra-crustal decollement level that progressively shallows from the south to the north. Indeed, the interior of the mountain range results from the imbrication and exhumation of Protero-Paleozoic basement and Paleozoic cover, whereas the foothills are characterized by a fold-and-thrust belt of Paleozoic, Mesozoic and Cenozoic sediments (e.g., XBGMR, 1976; XBGMR, 1985; Brunel et al., 1994; Matte et al., 1996; Wittlinger et al., 2004; Li et al., 2012b, 2016b; Jiang et al., 2013; Cheng et al., 2016b; Wang and Wang, 2016; Lu et al., 2016; Guilbaud et al., 2017; Thompson Jobe et al., 2017).

The basement of the Western Kunlun is thus composed of Proterozoic to Paleozoic (Ordovician to Silurian) plutonic and metamorphic rocks that recorded the lower Paleozoic orogeny of the range (e.g., Matte et al., 1996; Mattern and Schneider, 2000; Cowgill et al., 2003; Cowgill, 2010; Cao et al., 2013), while within the Tarim Basin, the contemporaneous sedimentary series have not endured such plutonic and metamorphic events (see section 2.1) (Fig. 2). Over this basement, the Western Kunlun Paleozoic cover consists in Devonian continental conglomerates coming from the erosion of the Paleozoic mountain belt, followed by Carboniferous marine limestones (Fig. 2) (e.g., Matte et al., 1996; Mattern and Schneider, 2000; Cowgill, 2010; Cao et al., 2013). Within the foothills of the range, the stratigraphy above these units then corresponds to a Permian to Cenozoic, mostly continental and terrigenous sedimentary series, containing an upper Cretaceous-Paleogene marine interval of terrigenous, calcareous, and evaporitic deposits (Fig. 2) (e.g., Matte et al., 1996; Sobel, 1999; Yin et al., 2002; Jin et al., 2003; Sun and Jiang, 2013; Bosboom et al., 2014a, 2014b, 2017; Guo et al., 2015; Zheng et al., 2015; Chen et al., 2018). Although coarser-grained, this series is similar to the contemporaneous sediments deposited at the front of the Western Kunlun Range, along the southwestern Tarim Basin margin.

2.3. Altyn Tagh Range

The Altyn Tagh Range bounds the Tarim Basin to the southeast, from the west of Minfeng to the east of Dunhuang (Fig. 1). It is a ~1500-km-long, ~40–80-km-wide and up to ~6200-m-high, WSW-ENE-oriented mountain belt at the centre of the Tibetan Plateau northern edge. Presently, this mountain range corresponds to a northwestward-vergent strike-slip wedge that developed between the Altyn Tagh left-lateral fault to the southeast and the Tarim Basin to the northwest (Fig. 1) (e.g., Wittlinger et al., 1998; Cowgill et al., 2000; Li et al., 2002a; Yin et al., 2002; Tang et al., 2014; Lin et al., 2015b; Zhao et al., 2016).

In this mountain belt, the southeastern boundary of the Tarim block is marked by the Lapeiquan-Hongliugou suture (in red in Fig. 1) related to the closure of the Proto-Tethys Ocean and to the associated Paleozoic orogeny spanning from Silurian to Devonian (e.g., Sobel and Arnaud, 1999; Gehrels et al., 2003; Li et al., 2018a). After a complex deformation history during the Mesozoic (e.g., Sobel, 2001; Chen et al., 2003; Robinson et al., 2003; Liu et al., 2007), the Altyn Tagh Range was then reactivated during the Cenozoic due to the India–Asia collision (e.g., Tapponnier and Molnar, 1977; Yin et al., 2002; Jolivet et al., 2001; Yin, 2010). Based on thermochronological data (e.g., Jolivet et al., 1999, 2001; Sobel, 2001; Yin et al., 2002; Liu et al., 2003, 2007; Ritts et al., 2008), sediment provenance (e.g., Yue et al., 2001; Yin et al., 2002; Sun et al., 2005; Wu et al., 2012; Li et al., 2014b; Lin et al., 2015b; Lu et al., 2014; Cheng et al., 2015, 2016a), growth strata (e.g., Cheng et al., 2015, 2016a; Zhao et al., 2016), and accumulation rates (e.g., Métivier et al., 1998; Sun et al., 2005; Lu et al., 2014), this reactivation seems to initiate by Eocene with subsequent Oligocene to Miocene accelerations. The present strike-slip wedge then formed, presumably, above a deep intra-crustal decollement level. Indeed, the range interior results from the imbrication and exhumation of Protero-Paleozoic basement and Paleo-Mesozoic cover, while in the foothills, compressive structures involve the same basement and cover units overlain by Cenozoic sediments (e.g., XBGMR, 1976; XBGMR, 1985; Jolivet et al., 1999; Chen et al., 2003; Liu et al., 2007; Tang et al., 2014; Lin et al., 2015b).

The basement of the Altyn Tagh is thus composed of Proterozoic to Paleozoic (Silurian to Devonian) plutonic and metamorphic rocks that recorded the middle Paleozoic orogeny of this mountain belt (e.g., Sobel and Arnaud, 1999; Cowgill et al., 2003; Gehrels et al., 2003; Zhang et al., 2014a), whereas the contemporaneous Tarim sedimentary series have not endured such a metamorphic event (see section 2.1), similar to what is observed at the front of the Western Kunlun Range. Over this basement, the Altyn Tagh Paleo-Mesozoic cover consists in Carboniferous marine limestones and Jurassic continental and terrigenous deposits (e.g., Sobel and Arnaud, 1999; Chen et al., 2003). Within the foothills of the range, the stratigraphy above these units corresponds to a continental and terrigenous Cenozoic series in continuity with the contemporaneous sediments deposited in the Tarim Basin (e.g., Yin et al., 2002; Sun et al., 2005; Lu et al., 2014).

2.4. Southern Tian Shan

The Southern Tian Shan bounds the Tarim Basin to the north, between the towns of Kashgar and Dunhuang (Fig. 1). It is a ~2200-km-long, ~100–250-km-wide and up to ~7400-m-high mountain belt corresponding to the southern edge of the Tian Shan Range. From its overall orientation, this belt can be subdivided into two parts: (1) the WSW-ENE-oriented Southwestern Tian Shan from Kashgar to the east of Kuqa, and (2) the WNW-ESE-oriented Southeastern Tian Shan from the east of Kuqa to Dunhuang. Presently, the Southwestern Tian Shan consists in a southeastward-vergent thrust wedge developed between the Nikolaev Line and Narat fault to the northwest and the Tarim Basin to the southeast (e.g., Poupinet et al., 2002; Heermance et al., 2008; Makarov et al., 2010; Xiao et al., 2013; Alexeiev et al., 2015; Izquierdo-Llavall et al., 2018), whereas the Southeastern Tian Shan corresponds to a southwestward-vergent strike-slip wedge developed between the right-lateral Baluntay fault to the northeast and the Tarim Basin to the southwest (e.g., Allen et al., 1994; Charvet et al., 2011; Li et al., 2016a) (Fig. 1).

In these mountain belts, the northern boundary of the Tarim block is marked by the South Tian Shan suture (in red in Fig. 1) related to the closure of the South Tian Shan Ocean or back-arc basin and to the associated Paleozoic orogeny spanning from Devonian to Permian (e.g., Windley et al., 1990; Gao et al., 1998; Charvet et al., 2007, 2011; Glorie et al., 2011; Xiao et al., 2013; Alexeiev et al., 2015; Han et al., 2016). During the Mesozoic, a complex compressive to strike-slip history then followed this orogeny (e.g., Hendrix et al., 1992; Sobel, 1999; Dumitru et al., 2001; Wang et al., 2009; Jolivet et al., 2010, 2013; Glorie et al., 2011; Chang et al., 2012b; Li et al., 2015a; Morin et al., 2018; Yang et al., 2018b). Finally, a major reactivation of the Tian Shan occurred during the Cenozoic in response to the India–Asia collision (e.g., Tapponnier and Molnar, 1979; Windley et al., 1990; Avouac et al., 1993; Allen et al., 1994; Hendrix et al., 1994; De Grave et al., 2007; Dumitru et al., 2001; Jolivet et al., 2010). Based on thermochronological data (e.g., Hendrix et al., 1994; Sobel and Dumitru, 1997; Dumitru et al., 2001; Sobel et al., 2006; Jolivet et al., 2010; Glorie et al., 2011; Chang et al., 2012b; Yu et al., 2014; Jia et al., 2015), sediment provenance and growth strata (e.g., Yin et al., 1998; Heermance et al., 2007; Li et al., 2012a; Bande et al., 2017a; Izquierdo-Llavall et al., 2018; Thompson Jobe et al., 2018; Li et al., 2019), as well as increasing sediment accommodation and accumulation rates in the northern Tarim Basin (e.g., Métivier and Gaudemer, 1997; Métivier et al., 1999; Yang

and Liu, 2002; Huang et al., 2006; Heermance et al., 2007; Zhang et al., 2014b; Li et al., 2019), this reactivation seems to initiate by Oligocene with subsequent Miocene accelerations.

Following this reactivation, the Southwestern Tian Shan thrust wedge formed above an intra-crustal decollement level that progressively shallows from the north to the south. The interior of this range thus results from the imbrication and exhumation of Protero-Paleozoic basement and Paleo-Mesozoic cover (e.g., XBGMR, 1976; XBGMR, 1985; Yin et al., 1998; Xiao et al., 2013; Alexeiev et al., 2015; Jourdon et al., 2017; Loury et al., 2017). It also comprises a few compressive intra-mountainous basins filled by Cenozoic sediments (e.g., XBGMR, 1976; XBGMR, 1985; Xiao et al., 2013; Morin et al., 2019). The foothills of the Southwestern Tian Shan then present a deformation style that strongly varies laterally. In the centre, near the town of Aksu, compressive structures involving Proterozoic basement and Paleo-Cenozoic cover developed from a large basement uplift (the Kalpin uplift) inherited from the Protero-Paleozoic history of the Tarim block (e.g., XBGMR, 1976; XBGMR, 1985; Carroll et al., 2001; Turner et al., 2010). In contrast, to the west and to the east, the foothills are characterized by fold-and-thrust belts (the Kashgar-Artux, Kepingtage and Kuqa belts) involving Paleozoic, Mesozoic and Cenozoic sediments (e.g., XBGMR, 1976, XBGMR, 1985; Allen et al., 1999; Yin et al., 1998; Burchfiel et al., 1999; He et al., 2005; Heermance et al., 2008; Turner et al., 2010; Li et al., 2012a; Alexeiev et al., 2015; Li et al., 2016d; Izquierdo-Llavall et al., 2018).

Further east, the Southeastern Tian Shan strike-slip wedge could have formed above an intra-crustal or deeper decollement level. As in the Southwestern Tian Shan, the range interior indeed results from the imbrication and exhumation of Protero-Paleozoic basement and Paleo-Mesozoic cover, and it comprises a few compressive intra-mountainous basins filled by Cenozoic sediments (e.g., XBGMR, 1976, 1985; Allen et al., 1994; Charvet et al., 2011; Li et al., 2016a; Morin et al., 2019). The foothills then entirely correspond to compressive and strike-slip structures involving Proterozoic basement and Paleo-Cenozoic cover, which derive from a large inherited Protero-Paleozoic basement uplift of the Tarim (the Kuruk Tagh uplift) (e.g., XBGMR, 1976, 1985; Yin et al., 1998; Tang et al., 2008; Lin et al., 2012b).

The basement of the Southern Tian Shan interior is thus composed of Proterozoic and Paleozoic (Silurian to Devonian) metamorphic and plutonic rocks that recorded the middle-upper Paleozoic orogeny of the range (Fig. 2) (e.g., Gao et al., 1998; Charvet et al., 2007, 2011; Gao et al., 2009; Ge et al., 2012; Xiao et al., 2013; Alexeiev et al., 2015). Over this basement, the Paleozoic cover consists in Silurian-Devonian to Permian marine limestones and continental to marine, volcano-sedimentary or terrigenous deposits (Fig. 2) (e.g., Gao et al., 1998; Charvet et al., 2011; Liu et al., 2013; Xiao et al., 2013; Alexeiev et al., 2015; Li et al., 2015a; Jourdon et al., 2017). Both the basement and cover units are intruded by Paleozoic (Devonian to Permian) plutons (Fig. 2) (e.g., Gao et al., 1998; Charvet et al., 2007, 2011; Gao et al., 2009; Glorie et al., 2011; Xiao et al., 2013; Alexeiev et al., 2015). However, in the foothills of the range as in the Tarim Basin, the coeval stratigraphy is characterized by a Proterozoic igneous basement below a thick, more or less continuous series of Sinian to Permian sediments that have not endured the Paleozoic orogeny (e.g., Carroll et al., 1995, 2001; Zhou et al., 2001; Tang et al., 2008; Turner et al., 2010; Shu et al., 2011; Li et al., 2015a; Cai et al., 2018; Wu et al., 2018). Above all these units, the Southern Tian Shan cover is composed of Triassic to Cenozoic, mostly continental and terrigenous sediments containing a Paleogene lacustrine interval of terrigenous and evaporitic deposits (Fig. 2) (e.g., Hendrix et al., 1992; Yin et al., 1998; Heermance et al., 2007; Tang et al., 2008; Wang et al., 2009; Li et al., 2015a; Li et al., 2016d; Tian et al., 2016; Izquierdo-Llavall et al., 2018; Morin et al., 2018). Although coarser-grained, this series is similar to the contemporaneous sediments deposited in the basin at the front of the range.

3. Data and method

3.1. Data

To characterize the Cenozoic deformation of the Tarim Basin, we have constructed a structural map and four balanced regional geological transects. To construct this map and these transects, we first synthesized the studies that were previously published in the area (Fig. 1). Indeed, numerous data and interpretations were already available such as (1) regional and local geological maps (e.g., XBGMR, 1976, 1985; Matte et al., 1996; Burchfiel et al., 1999; He et al., 2005; Cowgill, 2010; Fu et al., 2010; Turner et al., 2010; Cao et al., 2013; Cheng et al., 2015; Wang and Wang, 2016; Guilbaud et al., 2017; Thompson Jobe et al., 2017), (2) geological cross-sections from field observations and seismic profiles (e.g., Fig. 1 and references therein), and (3) isopach and isobath maps (e.g., Lee, 1985; Li et al., 1996; Yang and Liu, 2002; Chen and Shi, 2003; Li et al., 2016e). Eventually, we also used the previous biostratigraphical and magnetostratigraphical studies in the region to constrain the ages of the sedimentary series (e.g., Yin et al., 1998, 2002; Sobel, 1999; Sun et al., 1999; Chen and Shi, 2003; Sun et al., 2005, 2009a, 2009b, 2016a, 2017; Charreau et al., 2006, 2009; Huang et al., 2006, 2010; Heermance et al., 2007; Wang et al., 2009; Chang et al., 2012a, 2014a; Bosboom et al., 2014a, 2014b, 2017; Lu et al., 2014; Zhang et al., 2014b; Zheng et al., 2015; Peng et al., 2018; Blayney et al., 2019).

In addition to these previously-published data, we used surface observations performed in the field and on satellite images (Landsat, Spot and DigitalGlobe images available through the Google Earth database), as well as on digital elevation models (SRTM30, SRTM3, SRTM1, ASTER GDEM v2, and AW3D30, with a horizontal resolution from ~900 to ~30 m) (e.g., Figs. 4 and 5). Thanks to the Sinopec Group of the China Petrochemical Corporation, we also had access to an extensive subsurface dataset comprising ~31,120 km of regional and local seismic profiles covering the entire basin (Figs. 1 and S2), together with the data of 29 wells. In particular, Vertical Seismic Profiles (VSP) of P-waves were available for four of these boreholes (red dots in Fig. 1). From these VSP, seismic velocities could be deduced and a velocity model could be proposed (Fig. 3), in order to convert the travel time of seismic interpretations into depth (see section 3.4).

3.2. Seismic interpretation

To interpret our seismic dataset, we followed a standard procedure (Nester and Padgett, 1992). First, we identified the faults based on offsets of the seismic reflectors. Then, we picked seven horizons along the seismic profiles through the whole seismic grid covering the basin: the base of the (1) Sinian (i.e. the basement-sedimentary cover boundary in the Tarim Basin), (2) upper Cambrian (i.e. the base of the main Paleozoic decollement in the region), (3) Permian, (4) Paleocene, (5) Oligocene, (6) middle Miocene, and (7) Pliocene deposits (Fig. 2). The ages of the pre-Cenozoic horizons were mainly constrained by biostratigraphical dating on outcrops and cores (e.g., Sobel, 1999; Chen and Shi, 2003; Zhang et al., 2014b; Peng et al., 2018), as well as on samples collected in the boreholes of the Sinopec Group dataset. The ages of the Cenozoic horizons were determined on the basis of previous biostratigraphical and magnetostratigraphical studies mainly performed on outcrops. For this, we took advantage of sections sampled along outcrops that can be correlated with the seismic data (Yin et al., 2002; Charreau et al., 2006, 2009; Huang et al., 2006, 2010; Peng et al., 2006; Sun et al., 2009a, 2009b, 2017; Wang et al., 2009; Bosboom

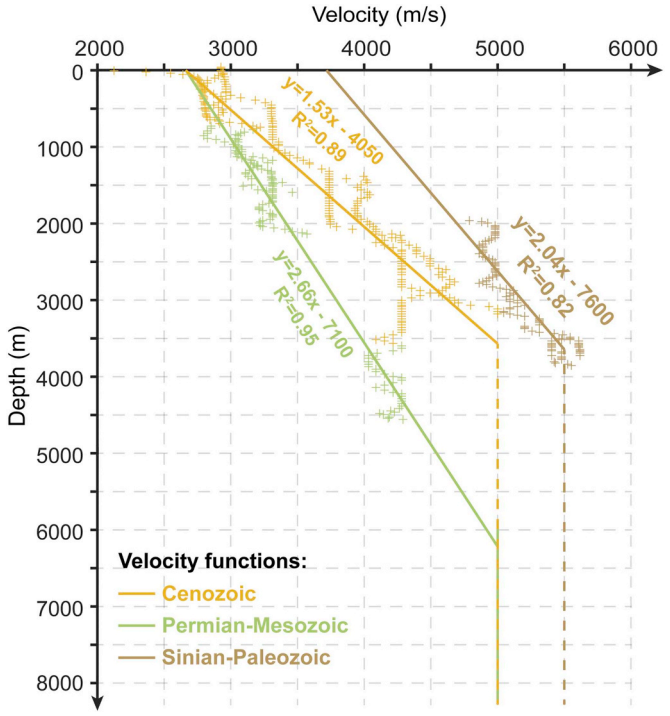


Fig. 3. Velocity functions used for time-to-depth conversion of the seismic interpretations. For each sediment package (Sinian-Paleozoic, Permian-Mesozoic and Cenozoic), these velocity functions are composed of two parts: (1) an upper part (solid line) corresponding to a single linear regression of the vertical seismic profiles of P-wave data (brown, green and yellow crosses) from four wells distributed across the Tarim Basin (red dots in Fig. 1), and (2) a lower part (dashed line) corresponding to a maximum velocity determined according to the main lithology of the sediment package (calcareous for the Sinian-Paleozoic deposits, terrigenous for the Permian-Mesozoic and Cenozoic series). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

et al., 2014a; Zheng et al., 2015; Blayney et al., 2019). These sections are located to the southwest, in the centre, and to the north of the basin (green dots in Fig. 1).

In order to ensure that our interpretation is consistent over the whole studied area, the selected seismic horizons were propagated from a seismic line to another through the entire seismic grid. The robustness of this horizon picking is confirmed by the fact that the position of our Cenozoic seismic horizons fall within a time interval of < 0.5 Myr on the dated sections outcropping several 100s of km apart in different places of the basin.

3.3. Structural mapping

In order to highlight the structures activated during the Cenozoic, we constructed a georeferenced structural map of the Cenozoic deformation through the Tarim Basin and its surrounding ranges (Figs. 6 and S1). On this map, we reported the main faults and folds affecting the Cenozoic sedimentary series. We identified and located the outcropping structures from the available surface data (i.e. the field constraints, satellite images, digital elevation models, together with the previously-published geological maps and cross-sections) (e.g., Figs. 4 and 5). When it was possible, we also verified the occurrence and location of these structures on the seismic profiles. In addition, we identified and located the structures buried below the Cenozoic sediments from the available subsurface data (i.e. the previously-published and unpublished seismic profiles and wells) (e.g., Fig. S2). Finally, we reported the position of the outcropping and buried fault tips and fold axes on the map, using different line thicknesses according to the scale of the structures, and different colours according to their rooting depth (red for the basement, orange for the Paleozoic decollements and yellow for the Cenozoic ones) (Fig. 6).

3.4. Time-to-depth conversion

To build geological cross-sections through the basin, we needed to convert our seismic interpretations (in two-way-time) into geological geometries (in depth). For this, we used the VSP data from four boreholes distributed across the basin (red dots in Fig. 1) to establish velocity models for three sediment packages: the Sinian-Paleozoic (Sinian to Carboniferous), Permian-Mesozoic (Permian to Cretaceous) and Cenozoic series (Fig. 3). Indeed, for these different packages, the variation of the interval velocities according to depth follows a similar trend in the four wells. For each series, we approximated this variation by a velocity function versus depth with two parts: (1) an upper part corresponding to a single linear regression of the velocity data (solid lines in Fig. 3), and (2) a lower part corresponding to a maximum velocity determined according to the main lithology of the sediment package (calcareous for the Sinian-Paleozoic deposits, terrigenous for the Permian-Mesozoic and Cenozoic series) (dashed lines in Fig. 3). By using this velocity model, we assumed that the three approximated velocity functions are uniform at the basin scale, i.e. we neglected the local velocity variations that can exist due to lateral lithological changes. The time-to-depth conversion based on these functions is therefore a first-order achievement, which seems to be acceptable for this study at the scale of the whole Tarim Basin. Indeed, the depths estimated for the selected horizons falls in the range of their depths assessed in previously-published structural interpretations (cross-sections, isobath and isopach maps where the horizon depths vary by a few kilometres) (e.g., Li et al., 1996; Chen and Shi, 2003; Tang et al., 2014; Li et al., 2016e; Lu et al., 2016; Wang and Wang, 2016; Izquierdo-Llavall et al., 2018; Jiang et al., 2018).

3.5. Geological transect construction

In addition to the structural map, we constructed four balanced geological transects through the basin and the outer parts of the ranges (Figs. 7–10, and S3). We built these regional cross-sections following a standard method (e.g., Woodward et al., 1989). We chose their paths perpendicular to the Cenozoic structure direction, which may change along each section (Fig. 6). As much as possible, we also located them along seismic profiles to precisely constrain the deformation in the basin (Figs. 1 and S2). Where this was not possible (e.g., where the sections are not parallel to the seismic data such as in the centre of the basin for example), the deformation could however still be constrained using the seismic profiles crossing the transects in multiple locations. As for the structural map, we used the surface geology (e.g., Figs. 4 and 5) to build these transects where structures are cropping out, in particular through the ranges at the ends of the sections. We also used the available subsurface data to constrain the structures buried below the Cenozoic sediments in the basin (e.g., Fig. S2). Finally, to balance our geological transects, we used the concepts of the typical area balancing approach (e.g., Hossack, 1979; Mitra and Namson, 1989), combined with a line-length balancing (e.g., Dahlstrom, 1969) for the Meso-Cenozoic stratigraphic boundary. The line-length and area balancing approaches refer to the basic assumption that the strata lengths and cross-sectional areas do not substantially change during deformation. The combination of these approaches allowed us to propose admissible and viable cross-sections over their whole length (Elliott, 1983), even

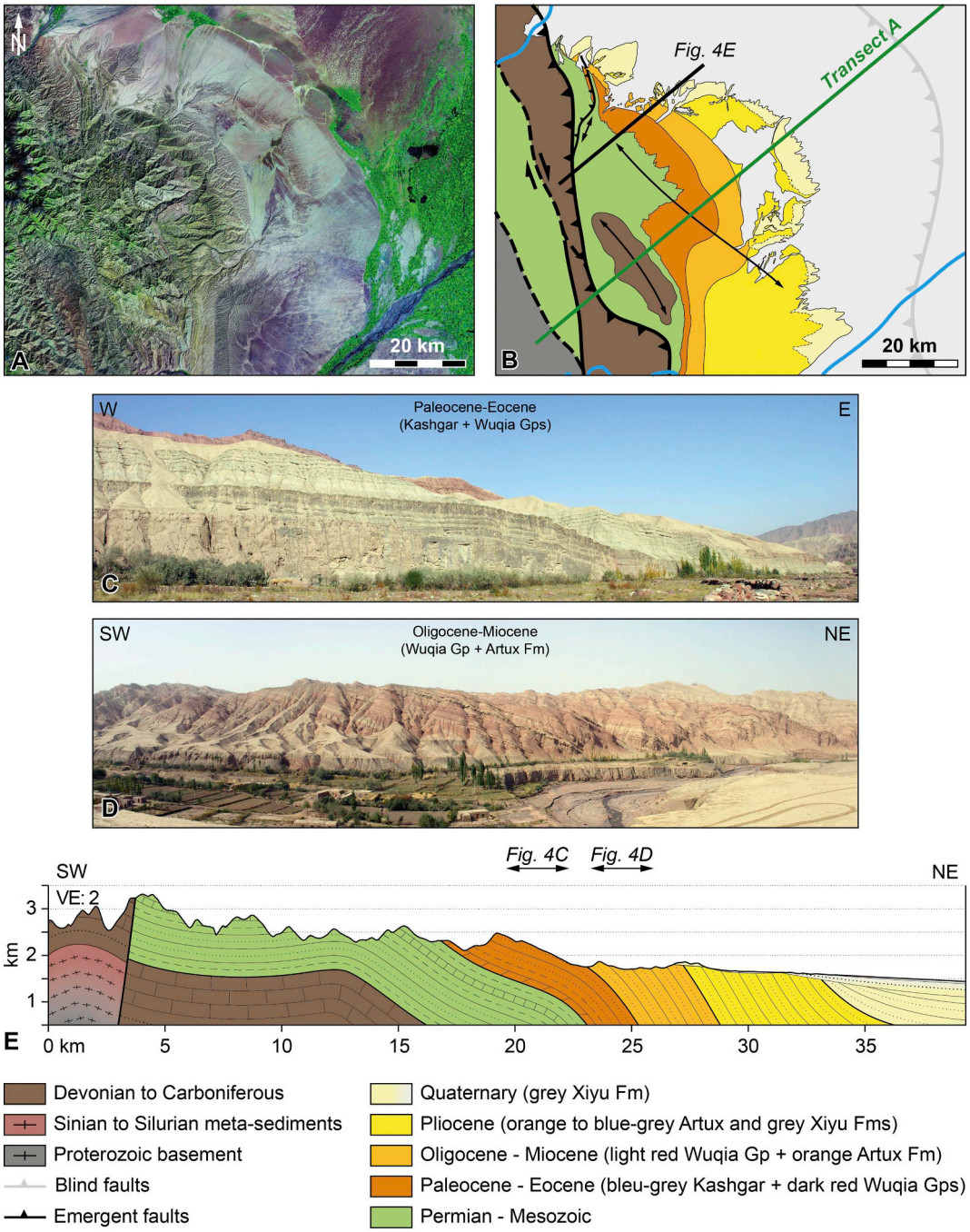


Fig. 4. Example of surface data and the associated detailed work on the Qimugen fold-thrust system (see Fig. 1 for location). (A) Satellite image (Landsat images with a 28-m resolution) of the Qimugen hills. (B) Geological and structural map of the same area constructed from field data, satellite images and seismic profiles. (C) Field picture of the Qimugen hills taken from the south. (D) Field picture of the Qimugen hills taken from the southeast. (E) Surface geological cross-section of the Qimugen hills with a vertical exaggeration (VE) of 2.

where the details of the structural geometries are not well imaged on seismic profiles (e.g., inside the duplex structures to the south of transects A and B or to the north of transect C). This implies that the produced cross-sections are restorable.

3.6. Transect restoration

In order to quantify the total Cenozoic shortening accommodated within the basin and the outer parts of the ranges, we restored the proposed balanced geological transects at the beginning of the Cenozoic (Fig. 7). This restoration was thus mainly based on the length conservation of the Paleogene basal boundary, which pre-dates the Cenozoic deformation onset. Where this boundary is lost by erosion without any older stratigraphic horizon that may be used as another marker, we minimized the eroded length by closing the structures just above the present-day topography (e.g., see the dotted lines to the south of transects C and D or to the north of transect A). Moreover, where this boundary is missing in the stratigraphy (e.g., where the lower Tertiary is lacking due to no deposit or erosion at that time), the restoration was made at the base of the oldest Cenozoic sediments existing in the area (e.g., the Middle-Upper Miocene to the north of transect A or the Plio-Quaternary to the north of transects B and D). Therefore, by following

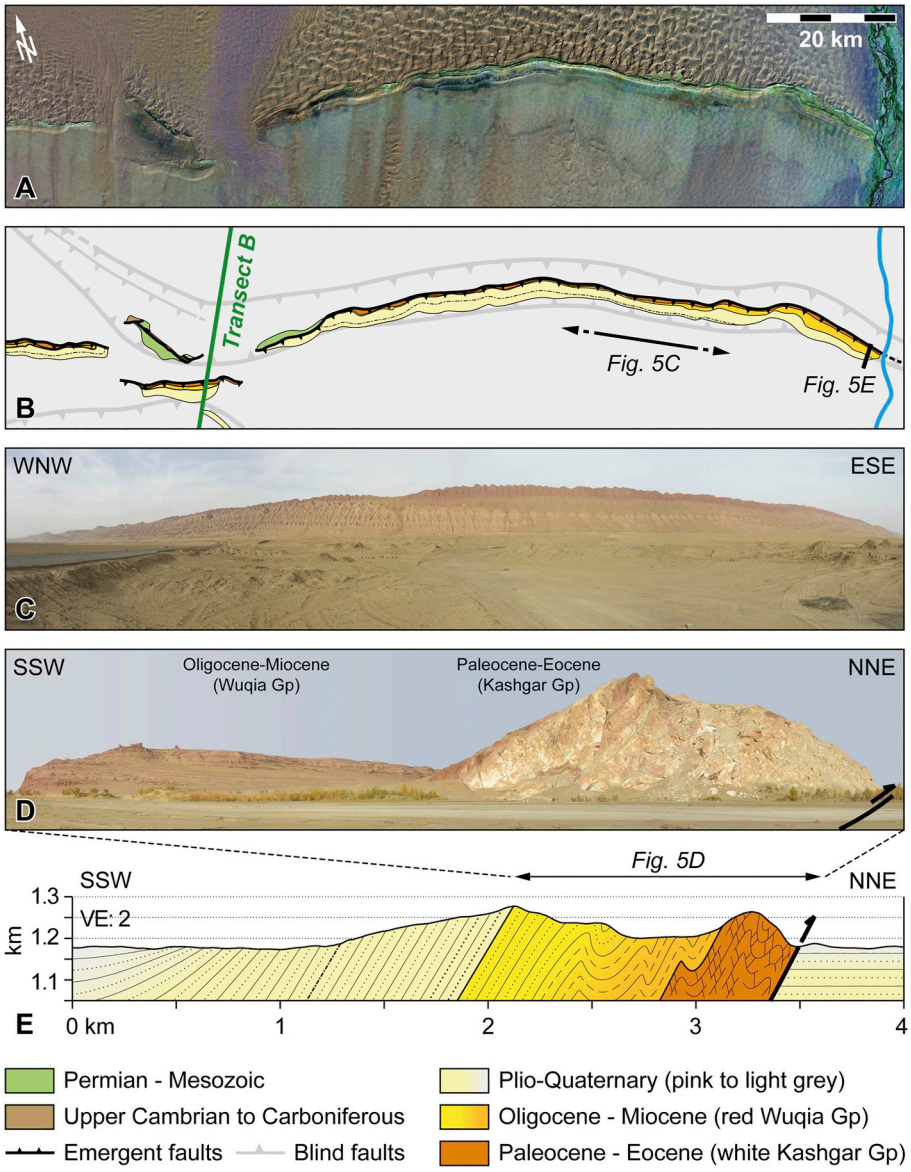


Fig. 5. Example of surface data and the associated detailed work on the eastern part of the Mazar Tagh emergent thrust ramps and deep fault system (see Fig. 1 for location). (A) Satellite image (Landsat images with a 28-m resolution) of the Mazar Tagh ridge. (B) Geological and structural map of the same area constructed from field data, satellite images and seismic profiles. (C) Field picture of the Mazar Tagh ridge taken from the south. (D) Field picture of the eastern termination of the Mazar Tagh ridge taken from the northeast. The light-colored deposits to the right correspond to the Cenozoic (Paleocene-Eocene) evaporitic decollement of the southwestern Tarim Basin, which crops out at the top of the local Mazar Tagh emergent thrust ramp. (E) Surface geological cross-section of the eastern termination of the Mazar Tagh ridge with a vertical exaggeration (VE) of 2 (modified from Pan et al., 2010).

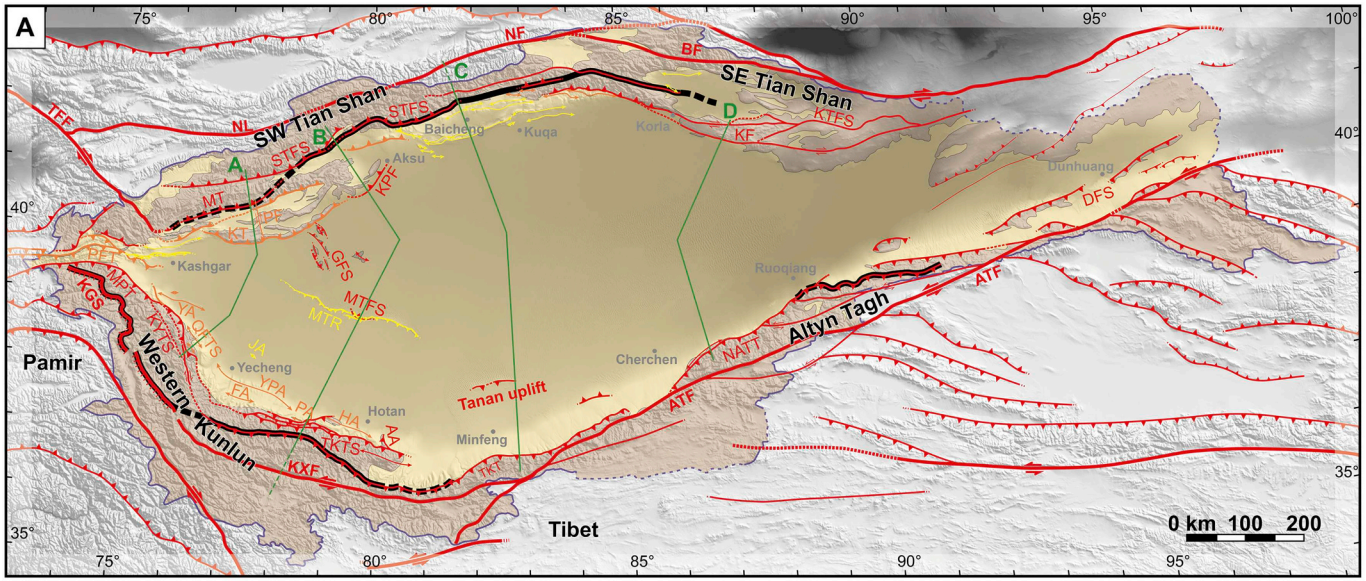
those rules, the shortening amounts deduced from our structural restoration are minimum estimates. However, to avoid too much approximation, we only performed this restoration between the main bordering faults of the Cenozoic deposits, without including the range interiors where stratigraphic markers are lacking. Eventually, the fixed reference lines used for this work are located in the basin, at the turning points of the different transects (Figs. 1 and 7).

4. Overall Cenozoic deformation

The structural map and regional geological transects constructed to document the Cenozoic deformation through the Tarim area are presented in Figs. 6 and 7. They illustrate the overall structure of the surrounding mountain ranges verging toward the basin, i.e. northward along the Tibetan Plateau edge to the south and southward along the Tian Shan to the north. These surrounding ranges correspond to well-developed thrust wedges to the west of the basin (Western Kunlun Range and Southwestern Tian Shan) and strike-slip wedges to the east (Altyn Tagh and Southeastern Tian Shan). They concentrate a large amount of deformation with thick-skinned structures involving a Protero-Paleozoic basement and Paleozoic sedimentary cover in their interiors, while their foothills are characterized by either thin- or thick-skinned structures involving a Protero-Paleozoic basement and/or Protero-Paleozoic, Mesozoic and Cenozoic sediments, respectively (Figs. 6 and 7).

In the case of the Western Kunlun and Southwestern Tian Shan compressive systems, these foothills are mainly formed of wide (10s of km) and long (100s of km) thin-skinned fold-and-thrust belts (Western Kunlun, Kashgar-Artux, Kepingtage, and Kuqa belts), except for the central part of the Southwestern Tian Shan where a thick-skinned belt locally occurs (Kalpin belt) (Figs. 6 and 7). Associated with these fold-and-thrust belts, which can accommodate a large amount of shortening (10s of km) and thickening, deep flexures (Southwest, North and Kuqa depressions) have developed in the basin with up to ~10 and ~9 km of Cenozoic sediments accumulated along the front of the Western Kunlun and Southwestern Tian Shan, respectively (Fig. 7A, 7B and 7C). In contrast, no fold-and-thrust belt and deep foreland flexure are associated with the Altyn Tagh and Southeastern Tian Shan transpressional systems (Fig. 7C and 7D). In their case, the foothills are narrow (a few 10s of km) and correspond to a few basement-involved structures accommodating a limited amount of shortening (a few kilometres at most). Along the front of these ranges, less than a few kilometres of Cenozoic sediments are usually trapped.

In addition to this deformation within the ranges, Cenozoic



AA: Aqike Anticline, ATF: Altyn Tagh Fault, ATS: Aqia Thrust System, Bayan. B: Bayanbulak Basin, BF: Baluntay Fault, CFS: Cherchen Fault System, DFS: Dunhuang Fault System, FA: Fusha Anticline, GFS: Gudongshan Fault System, HA: Hotan Anticline, JA: Jiede Anticline, KAB: Kashgar-Artux Belt, KADF: Kashgar-Artux Deformation Front, KEB: Kepingtage Belt, KF: Korla Fault, KGS: Kongur Shan normal fault, Korla B: Korla Basin, KPB: Kalpin Belt, KPF: Kalpintagh Fault; KT: Kepingtage Thrust, KTFS: Kuruk Tagh Fault System, KUB + dep: Kuqa Belt + Kuqa depression, KUDF: Kuqa Deformation Front, KXF: Karakax Fault, KYTS: Kashgar-Yecheng Transfer System, LNFS: Lop Nur Fault System, MFS: Minfeng Fault System, MPT: Main Pamir Thrust, MT: Muziduke Thrust, MTFS and MTR: Mazar Tagh deep Fault System and emergent thrust Ramps, NATT: Northern Altyn Tagh Thrust, N dep.: North depression, NF: Narat Fault, NFS: Niya Fault System, NL: Nikolaev Line, PA: Pusikai Anticline, PF: Piqiang Fault, PFT: Pamir Frontal Thrust, QFTS: Qimugen Fold-Thrust System, RFS: Ruoqiang Fault System, SE dep.: Southeast depression, SFS: Selibuya Fault System, STFS: Southern Tian Shan Fault System, SW dep.: Southwest depression, TFF: Talas Ferghana Fault, TKT: Tam Karaul Thrust, TKTS: Tiklik Thrust System, TTS: Tumuxiuke Thrust System, WFTS: Wupoer Fold-Thrust System, WKFR: Western Kunlun Frontal blind Ramp, YPA: Yecheng-Pishan Anticline, YA: Yingisar Anticline

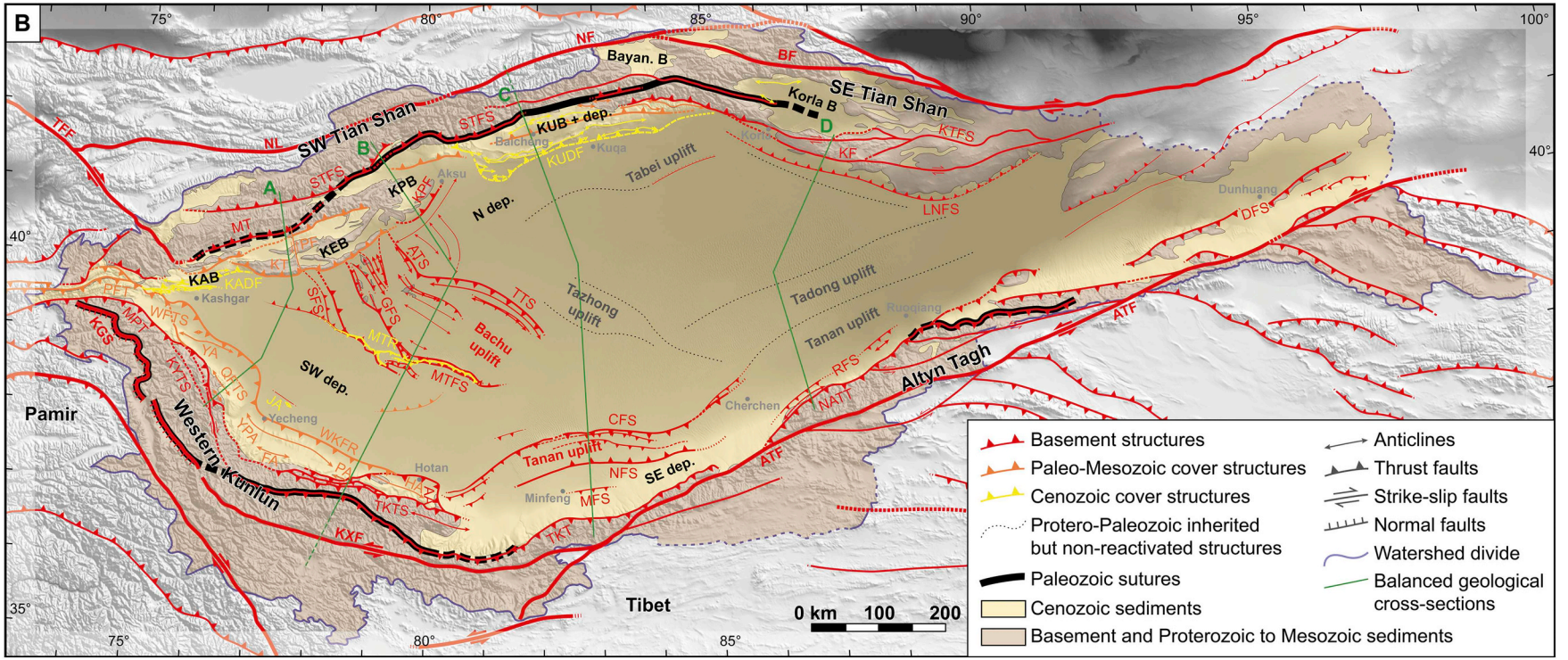


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Fig. 6. Structural map of the Cenozoic deformation of the Tarim Basin and surrounding ranges. Simplified and abbreviated structure labels are spelled out in the upper right cartoon. (A) Structures outcropping at the surface only. (B) Structures outcropping at the surface and hidden below the Cenozoic sediments of the Tarim Basin. Structures rooted in the basement are represented in red, while structures rooted into a sedimentary decollement are displayed in orange (for structures developed above Paleozoic decollements) or in yellow (for structures developed above Cenozoic decollements). Green solid lines locate the regional balanced geological transects A, B, C and D presented in Fig. 7. Green dashed lines at terminations of transects B and C correspond to the extended pathway of the synthetic crustal-scale cross-sections presented in Fig. 13. In this study, the Tarim block refers to the area comprised between the Paleozoic ophiolitic sutures (thick black lines) located in the surrounding ranges, rather than to the present-day Tarim Basin only. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

deformation also occurs within the Tarim. Wide (from ~50 to ~150 km) and long (100s of km) basement-cored uplifts are visible to the west (Bachu uplift) and south (Tanan uplift) of the basin (Figs. 6, 7B, and 7C). These basement-cored uplifts correspond to large pop-up structures bounded by conjugate thrust systems. They are clearly parallel to the ranges bordering the basin to the south (Fig. 6), with a main vergence toward the north, i.e. synthetic to the deformation of the Tibetan Plateau edge. Indeed, this vergence can be deduced from the observation of a compressive deformation accumulated on their conjugate bounding thrust systems larger to the north than to the south. Besides these basement-cored uplifts, other basement-involved structures (Tazhong, Tadong, and Tabei uplifts) exist further north and east in the basin (Figs. 6, 7C and 7D). However, they mainly deform the Protero-Paleozoic strata, with little evidence of a significant Cenozoic activity on their bounding faults. At first order, the Cenozoic deformation of the original Proterozoic Tarim block therefore mainly occurs along its edges and within its southwestern interior.

5. Southwest deformation

To the southwest of the Tarim Basin, the northeastward-vergent Western Kunlun thrust system can be divided into four structural parts with, from the southwest to the northeast: (1) the inner range from the Kongur Shan normal fault and Karakax left-lateral fault to the Main Pamir thrust, Kashgar-Yecheng transfer system, and Tiklik thrust system, (2) the outer range from the Main Pamir thrust, Kashgar-Yecheng transfer system and Tiklik thrust system to the Western Kunlun frontal blind ramp, (3) the Mazar Tagh thrust sheet and Southwest depression from the Western Kunlun frontal blind ramp to the Mazar Tagh emergent thrust ramps, and (4) the Bachu uplift from the Selibuya-Mazar Tagh deep fault systems to the Aqia-Tumuxiuke thrust systems (Figs. 6, 7A, 7B, 8A, 8B and 9).

5.1. Western Kunlun inner range

The inner range is several 10s to ~100 km wide, with typical elevations between ~2500 and ~5500 m. It is composed of exhumed imbricates of the Western Kunlun Protero-Paleozoic (Sinian to Silurian) basement and Paleozoic (Devonian to Carboniferous) cover (see section 2.2). In this system, the inner range is a continuous belt stretching from the Pamir to the west to the Altyn Tagh to the east. There, its northern part ends in a large basement-involved anticline and syncline to the southeast of Hotan (Fig. 6). Due to limited field access and subsurface data, its detailed structure is poorly constrained, but connecting the major faults within the basement to a deep basal decollement seems a reasonable hypothesis (Figs. 7A, 7B, 8A and 8B), as previously proposed by Matte et al., 1996, Wittlinger et al., 2004 or Guilbaud et al., 2017. Toward the northeast, the range interior is separated from the outer range by the Main Pamir thrust to the northwest and by the Tiklik thrust system to the southeast, connected by the Kashgar-Yecheng transfer system along the eastern edge of the Pamir salient (Fig. 6). These emergent thrust and strike-slip zones also correspond to the main bordering faults of the southwestern Cenozoic series of the Tarim Basin.

5.2. Western Kunlun outer range

At the front of the inner range, the outer range is also several 10s to ~100 km wide, with typical elevations between ~1500 and ~2500 m. It consists in a fold-and-thrust belt that mainly developed below the Cenozoic sediments of the Tarim Basin (Figs. 6, 7A, 7B, 8A, 8B, S2A and S2B). This belt follows the range from the Pamir to the west to the southeast of Hotan to the east where it abruptly ends into a lateral thrust ramp (Fig. 6). Several cross-sections have already been proposed through this outer range with various and contradictory interpretations on its structure (e.g., Brunel et al., 1994; Matte et al., 1996; Li et al., 2012b, 2016b; Jiang et al., 2013; Wang et al., 2013; Wei et al., 2013; Tang et al., 2014; Cheng et al., 2016b, 2017; Lu et al., 2016; Wang and Wang, 2016; Wang et al., 2016a; Guilbaud et al., 2017; Thompson Jobe et al., 2017). From our surface and subsurface data, we derived a regionally-consistent structural model settling the issue of these interpretations (see Supplementary information S1 for further details and discussion). We concluded that the surface and subsurface data provide clear evidence for 10s-of-km-long, emergent or blind folded thrust slices made of the Western Kunlun Protero-Paleozoic (Sinian to Silurian) basement and its Paleo-Mesozoic (Devonian to Cretaceous) cover at the rear, whereas 10s- to 100s-of-km-long duplex structures developed within the Tarim Paleo-Mesozoic (Cambrian to Cretaceous) sedimentary cover at the front (Figs. 7A, 7B, 8A, 8B, S2A and S2B).

These structures imply a basal decollement level that progressively shallows from the southwest to the northeast. At the rear of the outer belt, the major thrusts necessarily root into a deep decollement within the basement since they exhumed basement units (Figs. 7A, 7B, 8A and 8B). This decollement probably connects with the one below the range interior as classically inferred or proposed for most compressional orogens (e.g., Muñoz, 1992; Burkhard and Sommaruga, 1998; Lavé and Avouac, 2001). The northeastward change from the thick-skinned thrust slices to the thin-skinned duplexes then requires the presence of a ramp connecting this deep decollement to an intermediate one at the base of the deformed sedimentary cover (Figs. 7A, 7B, 8A, and 8B) (see also Matte et al., 1996; Wittlinger et al., 2004; Wang et al., 2013; Li et al., 2016b, 2018b; Lu et al., 2016; Cheng et al., 2017; Guilbaud et al., 2017). According to our regional seismic interpretation constraints by local borehole observations, this intermediate decollement is located at a depth of about 10 to 13 km below sea level (b.s.l.), in the upper Cambrian evaporitic assemblage of the Tarim sedimentary series (see also Wang et al., 2013; Tang et al., 2014; Lu et al., 2016; Guilbaud et al., 2017). Remarkably, the stratigraphy of the Paleozoic terrains is different on both sides of the ramp connecting the deep and intermediate decollements (Figs. 7A, 7B, 8A and 8B). To the southwest, this stratigraphy is characteristic of the Paleozoic Western Kunlun Range with Ordovician to Silurian metamorphic rocks covered by Devonian to Permian deposits (see Section 2.2), whereas to the northeast, it is typical of the Paleozoic Tarim Basin with Cambrian to Permian sediments (see Section 2.1) (Fig. 2). Therefore, the deep ramp to the southwest of the outer range seems to be a major inherited Paleozoic structure reactivated during the Cenozoic.

To the front of the outer belt, the formation of duplexes within the Tarim Paleo-Mesozoic series requires the activation of a series of ramps connecting the intermediate decollement within the upper Cambrian evaporitic assemblage to a shallower one (Figs. 7A, 7B, 8A and 8B). According to the seismic profiles, together with field and well constraints, this shallow decollement is located at a depth of about 6 to 9 km b.s.l. at the front of the duplexes, in the Paleogene evaporitic assemblage at the base of the Cenozoic deposits (see also Li et al.,

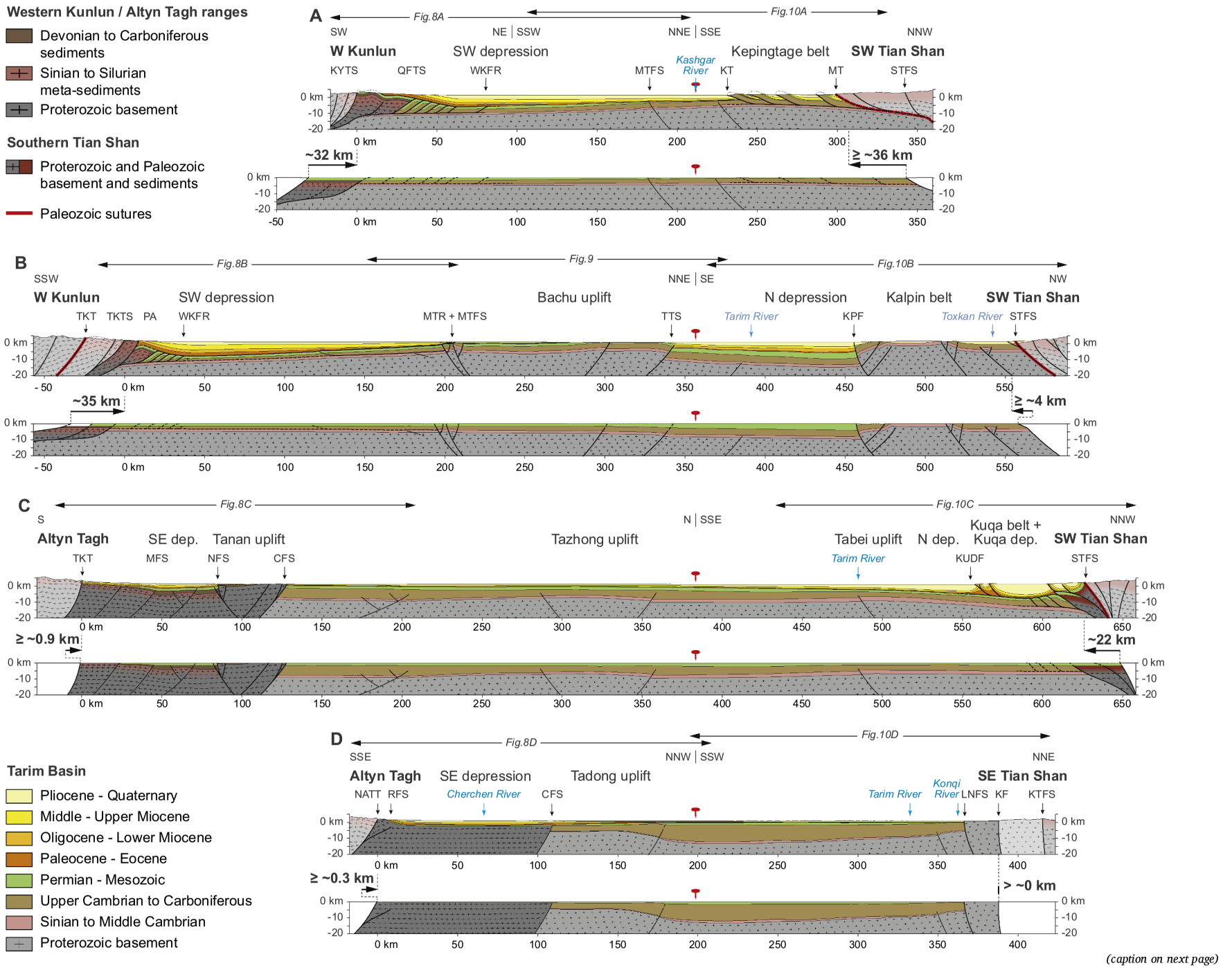

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Fig. 7. Regional balanced geological transects across the Tarim Basin and surrounding ranges (see Figs. 1 or 6 for location, Fig. 6 for abbreviation labels, and Fig. S3 for a 1:2,000,000-scale version of this figure). Below each transect, a restored state at the beginning of the Cenozoic is presented. For these restorations, the fixed reference lines (red pins) are located at the turning point in the centre of sections, and the lighter thrust slices in the range interiors are not considered. Shortening amounts (black bold numbers) measured on each side of the fixed reference lines are indicated to the northern and southern ends of the restored transects.

2012b, 2016b; Jiang et al., 2013; Wang et al., 2013; Tang et al., 2014; Cheng et al., 2017; Lu et al., 2016; Wang and Wang, 2016; Guilbaud et al., 2017; Thompson Jobe et al., 2017). Below this decollement, the stratigraphy of the Paleo-Mesozoic sedimentary cover seems to be the same (Cambrian to Cretaceous deposits) across the ramps connecting the Paleozoic and Cenozoic decollements. Therefore, these shallow ramps at the front of the outer range likely correspond to structures created during the Cenozoic.

At the top of the blind basement-involved thrust slices and of duplexes of the outer belt, the Cenozoic sedimentary series is folded into 10s-of-km-wide and up to ~180-km-long anticlines and synclines (Figs. 4, 6, 7A, 7B, 8A and 8B). Assuming an area conservation during deformation to keep the cross-sections balanced, the geometry of the underlying duplexes can be reconstructed from the overall envelope of these folds, which is usually well visible on the seismic profiles (Fig. S2A and S2B). These structures are thus interpreted as a stack of several (from four to six) horses, comprised between a floor thrust within the Paleozoic decollement level and a roof thrust within the Cenozoic one (Figs. 7A, 7B, 8A and 8B). This duplex style is typical of the frontal deformation in the central part of the Western Kunlun outer range, such as for the Qimugen fold-thrust system, the Yecheng-Pishan anticline or the Pusikai anticline (see also Jiang et al., 2013; Wang et al., 2013; Wei et al., 2013; Tang et al., 2014; Li et al., 2016b; Lu et al., 2016; Wang and Wang, 2016; Guilbaud et al., 2017).

Near the lateral terminations of the belt, structures of the outer range vary slightly from this pattern. To the northwestern termination (Yingisar anticline and Wupoer fold-thrust system), duplexes are also described in the Paleo-Mesozoic cover, but they are associated with emergent ramps breaching their roof thrust or rooted in the Cenozoic decollement at their top (e.g., Li et al., 2012b, 2013; Thompson et al., 2015; Thompson Jobe et al., 2017, 2018; Cheng et al., 2016b; Wang and Wang, 2016; Wang et al., 2016a). To the southeastern termination, the Hotan anticline corresponds to a ramp anticline involving the Western Kunlun basement thrusted over the Cenozoic decollement (e.g., Jiang et al., 2013; Cheng et al., 2017). In the centre of the fold, this thrust sheet is subsequently folded by a duplex structure that developed in the footwall of the basement ramp, within the Tarim Pre-Cenozoic sedimentary cover. However, in spite of these geometrical variations, the main structural features of the outer range are similar almost everywhere: (1) a deep ramp rooted in the basement to the south, at the transition between the Western Kunlun and Tarim Paleozoic domains, and (2) a duplex structure in the Paleo-Mesozoic Tarim sedimentary series to the north, with an emergent or blind leading shallow ramp. The only exception corresponds to the lateral ramp of the Aqike anticline at the eastern edge of the belt (Fig. 6). Indeed, this anticline corresponds to a basement-involved breached duplex with a tailing emergent ramp and no cover duplex at its front (Yang et al., 2009).

Although largely covered by the Cenozoic deposits coming from the Western Kunlun erosion, the duplexes of the outer belt localize the topographic front of the mountain range. They usually form smooth topographic features uplifted by several 100s of m above the present-day Tarim sedimentation level (Fig. 6). In the northwestern and southeastern ends of the range, this topographic front corresponds to emergent ramps, i.e. to an emergent deformation front. However, along most of the range, the Western Kunlun frontal ramp is a blind structure since it corresponds either to the leading blind ramp of a duplex (Qimugen fold-thrust system, Yecheng-Pishan anticline or Pusikai anticline) or to the blind front of a ramp anticline (Hotan anticline) (Figs. 7A, 7B, 8A and 8B). This raises the question of how the shortening accumulated within the outer belt is accommodated and transferred upward to the Cenozoic sedimentary pile. In the NNW-SSE-oriented part of the range, at the level of the Qimugen fold-thrust system, there is no evidence of deformation, and thus of slip transfer, in the Cenozoic deposits of the foreland, which are still attached to the underlying series (Figs. 6, 7A, 8A and S2A). This implies the existence of a triangular zone at the front of this fold-thrust system, with a passive roof thrust at its top within the Cenozoic decollement. It means that the Cenozoic deposits are detached above the underlying duplex, and that their shortening is accommodated by an emergent backthrust at their base (see also Wang and Wang, 2016; Wei et al., 2013). In contrast, to the east, in the WNW-ESE-oriented part of the range, such passive roof duplexes cannot exist due to the basement-involved imbricates emerging at the rear of the outer belt, which render back slip into the Cenozoic decollement impossible (Figs. 7B and 8B). Moreover, there is evidence of deformation, and thus of slip transfer, within the Cenozoic deposits of the foreland, such as in the case of the Jiede anticline or of the Mazar Tagh emergent thrust ramps located in the Tarim Basin, to the north of the Western Kunlun topographic front (Figs. 6, S2B, 7B and 8B) (see for example Wang et al., 2013; Wei et al., 2013; Wang et al., 2014a; Li et al., 2016b; Lu et al., 2016; Guilbaud et al., 2017; Jiang et al., 2018). Accordingly, in the eastern part of the range, the shortening accumulated within the outer belt appears to be transferred toward the foreland rather than toward the hinterland, with the activation of the Cenozoic decollement within the basin.

5.3. Mazar Tagh thrust sheet and Southwest depression

The Mazar Tagh is a striking topographic feature of the western central part of the Tarim Basin corresponding to an up to ~5-km-wide and ~220-km-long single to multiple ridge, raising a few 100s of m above the sedimentation level of the basin (Fig. 5A, 5C and 5D). This ridge stretches ~150–200 km to the northeast of the Western Kunlun topographic mountain front, parallel to the eastern part of the range (Fig. 6). Several contradictory cross-sections have been proposed through the Mazar Tagh area (e.g., Wittlinger et al., 2004; Yang et al., 2007a; Pan et al., 2010; Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; Yang et al., 2015; Guo et al., 2016; Li et al., 2016b; Li et al., 2016e; Lu et al., 2016; Guilbaud et al., 2017; Xie et al., 2017; Jiang et al., 2018). From our surface and subsurface data, we derived a regionally-coherent structural model in agreement with some of these previous studies (see Supplementary information S1 for further details and discussion).

According to our surface and subsurface dataset, this topographic high is clearly controlled by northeastward-vergent emergent thrust ramps synthetic to the Western Kunlun deformation and rooting at the base of the Cenozoic deposits (Figs. 5B, 5D, 5E, 7B, 8B, 9 and S2B) (see also Yang et al., 2007a; Pan et al., 2010; Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; Yang et al., 2015; Guo et al., 2016; Li et al., 2016b; Li et al., 2016e; Lu et al., 2016; Guilbaud et al., 2017; Xie et al., 2017; Jiang et al., 2018). Indeed, the Paleogene evaporitic assemblage acting as a major decollement level in the outer range extends northeastward into the Tarim, at least until the Mazar Tagh where it outcrops at the surface (Fig. 5D and 5E). This suggests that the slip transferred into this Cenozoic decollement at the front of the Yecheng-Pishan, Pusikai, and Hotan anticlines of the outer belt is released at the Mazar Tagh emergent thrust ramps, with erosion of the Cenozoic deposits of their hangingwalls. In agreement with some of the previous studies (Pan et al., 2010; Li et al., 2016e; Lu et al., 2016; Guilbaud et al., 2017; Li et al., 2018b), we thus argue that to the east of the central bend of the range, these ramps correspond to the present-day emergent deformation front located in the basin, ~150–200 km away from the

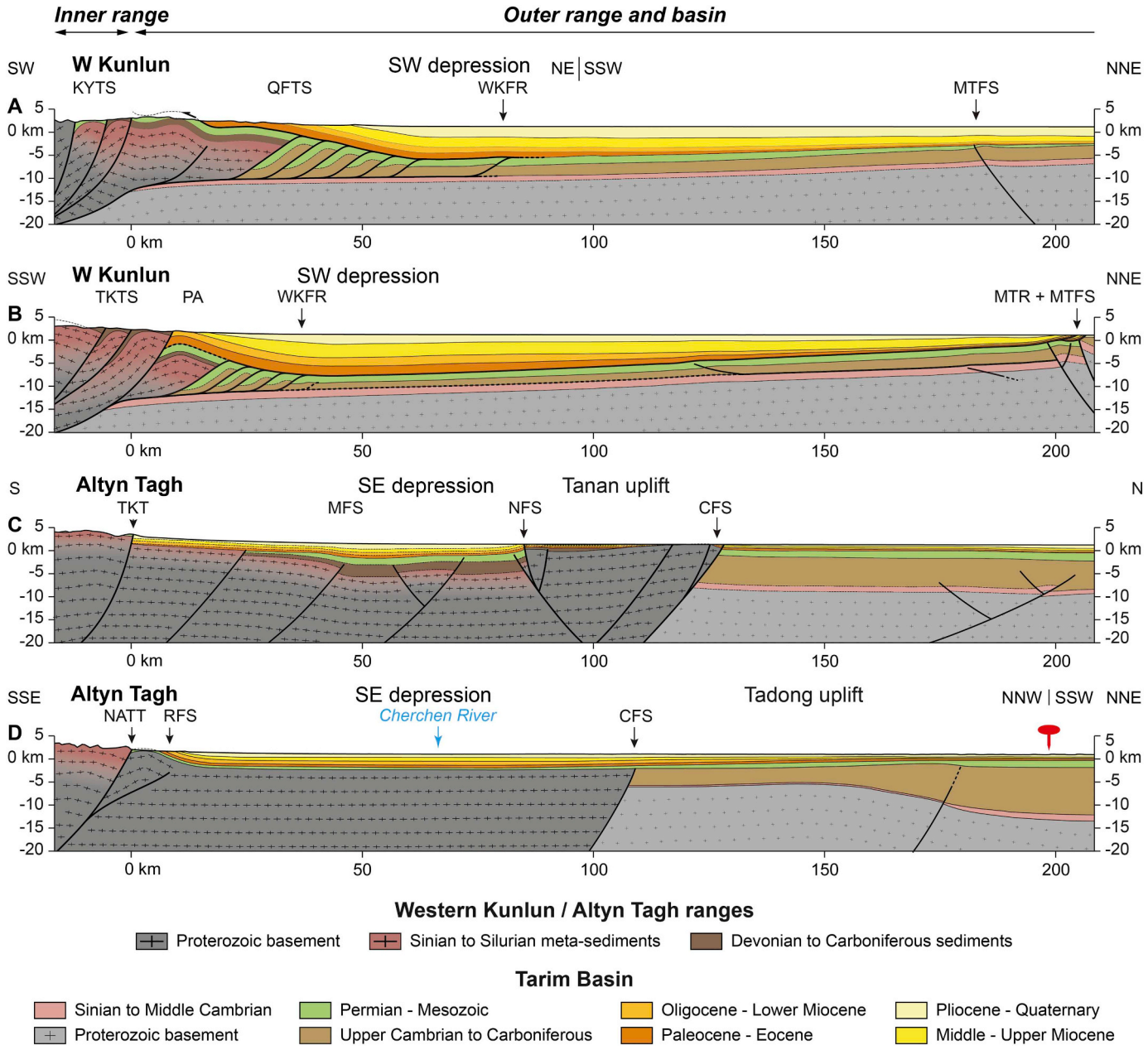

Fig. 8. Details of the regional balanced geological transects across the southern part of the Tarim Basin and the Western Kunlun and Altyn Tagh ranges (see Fig. 7 for location and Fig. 6 for abbreviation labels).

topographic mountain front, with a large frontal thrust sheet without any major internal deformation.

From the surface and subsurface data, structures involving the Proterozoic basement and Protero-Paleozoic to Mesozoic (Sinian to Jurassic) cover of the Tarim Basin also exist below the shallow Mazar Tagh deformation (Figs. 5A, 5B, S2B, 7B, 8B and 9) (see also Yang et al., 2007a; Pan et al., 2010; Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; Guo et al., 2016; Li et al., 2016b; Li et al., 2016e; Lu et al., 2016; Guilbaud et al., 2017; Xie et al., 2017; Jiang et al., 2018). Despite a complex and sometimes poorly-imaged geometry, these structures appear as a series of fault-related anticlines, often with backthrusts, or of minor pop-ups associated with steep thrust ramps. Regardless of a certain degree of symmetry, their main vergence is clearly toward the southwest since the base of the Cenozoic series is systematically higher to the northeast than to the southwest of these structures (Figs. 7B, 8B, 9 and S2B) (see for example Yang et al., 2007a; Pan et al., 2010; Tong et al., 2012; Wang et al., 2014a; Yang et al., 2015; Guo et al., 2016; Li et al., 2016b, 2018b; Li et al., 2016e; Lu et al., 2016; Guilbaud et al., 2017; Xie et al., 2017; Jiang et al., 2018). This unambiguously indicates an uplift of their northeastern compartments due to a southwestward thrust movement on ramps antithetic to the Western Kunlun deformation. At depth, the Mazar Tagh deep fault system needs to connect to a deep decollement within the basement. At their tips, the Mazar Tagh emergent thrust ramps seem to be correlated with south-facing steps in the Cenozoic decollement, indicating that the location of these shallow ramps was probably controlled by the hangingwall uplift of the underlying deep structures (Figs. 6, 7B, 8B, 9 and S2B).

To the west of the central bend of the range, the offset on these Mazar Tagh emergent thrust ramps decreases so that these ramps progressively disappear westward, with no more evidence of a wide thrust sheet or of any other deformation within the Cenozoic foreland deposits (Figs. 7A, 8A and S2A). On the other hand, a thrust ramp is still visible in the seismic profiles through the Proterozoic basement and Protero-Paleozoic to Mesozoic cover of the Tarim Basin, in continuity with the

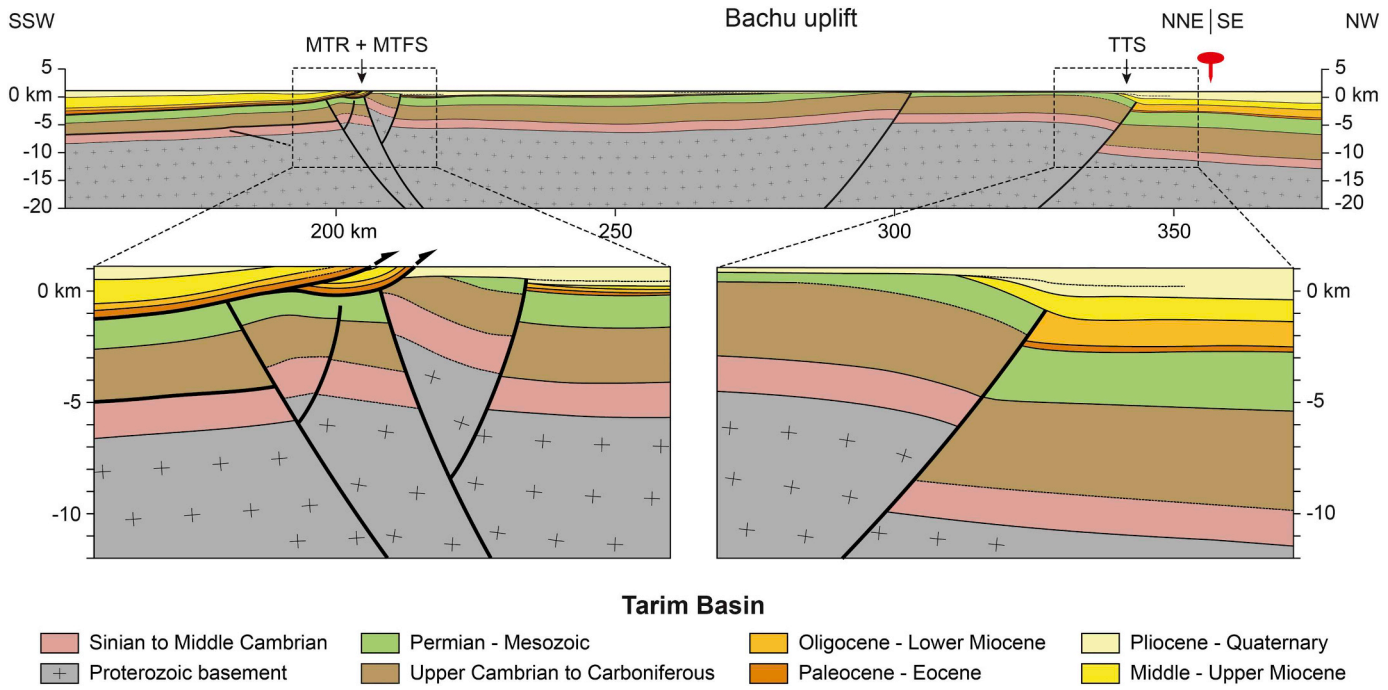

Fig. 9. Details of the regional balanced geological transect B across the central part of the Tarim Basin (see Fig. 7 for location and Fig. 6 for abbreviation labels).

WNW-ESE Mazar Tagh deep fault system (Figs. 6, 7A, 8A and S2A). However, their Cenozoic activity remained limited as indicated by very smooth fault-propagation folds at their tips without any offset of the overlying Cenozoic deposits.

In the Tarim Basin, the area extending from the outer range to the Mazar Tagh deep fault system is known as the Southwest depression (Figs. 6, 7A and 7B). In this area, the Tarim basement and sedimentary cover are tilted southward and the Cenozoic deposits thicken toward the Western Kunlun Range to reach their maximum thickness at the front of the outer belt (Figs. 7A, 7B, 8A, 8B, S2A and S2B). This overall geometry of the southwestern Tarim is thus typical of a flexural basin associated with the shortening and thickening accumulated in the Western Kunlun thrust system. It corresponds to the region of maximal accommodation space in the Tarim Basin with up to ~10 km of Cenozoic sediments trapped close to the cities of Kashgar and Yecheng, i.e. where the Western Kunlun converge with the Tian Shan and at the front of the central bend of the range, respectively. In this foreland system, the Western Kunlun outer belt corresponds to the wedge-top depozone, while the area comprised between the front of this belt and the Mazar Tagh deep fault system represents the foredeep (after the terminology of DeCelles and Giles, 1996), with the specificity that the whole Cenozoic sediments of the foredeep have been detached from the underlying series to form the Mazar Tagh thrust sheet to the east of the central bend of the mountain range.

5.4. Bachu uplift

To the north of the Tarim Southwest depression, the Bachu uplift is a ~150-km-wide and ~250-km-long basement-involved structure forming a structural high of the Proterozoic basement and Protero-Paleozoic to Mesozoic (Sinian to Triassic) sedimentary cover of the basin, below the Cenozoic sediments (Figs. 6, 7B, 9 and S2B). From its overall geometry documented by seismic profiles, together with field and well data, this basement-cored structure corresponds to a large pop-up bounded by two conjugate thrust fault systems necessarily rooted in a deep decollement within the basement below the Tarim. Bounding this basement high to the southwest, the Selibuya and Mazar Tagh deep fault systems are composed of a series of fault-related anticlines and small pop-ups associated with steep southwestward-vergent thrust ramps (Figs. 6, 7B, 9 and S2B) (see also Wang et al., 2014a; Yang et al., 2018b; Guo et al., 2016; Li et al., 2016e; Jiang et al., 2018). To the northeast, the Aqia and Tumuxiuke fault systems display large folds related to thrust ramps with a clear northeastward vergence and a hangingwall uplift of a few kilometres, which is larger than the one of the Selibuya and Mazar Tagh fault systems (Figs. 6, 7B, 9 and S2B) (see also Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; Guo et al., 2016; He et al., 2016; Li et al., 2016e; Jiang et al., 2018). Consequently, the main vergence of the Bachu pop-up appears to be toward the northeast, i.e. synthetic to the Western Kunlun deformation. In addition, this structure clearly developed parallel to the Western Kunlun Range, with a NNW-SSE orientation to the west changing to a WNW-ESE one to the east. To the southwest of the Bachu uplift, a thrust ramp also exists in lateral continuity with the WNW-ESE Mazar Tagh deep fault system (Figs. 6, 7A, 8A and S2A), but its Cenozoic activity remained limited compared to the one of the NNW-SSE Selibuya faults slightly further north (Tong et al., 2012; Li et al., 2016e; Yang et al., 2018b). Finally, there is no evidence of flexure associated with this large basement structure, neither at its front to the northeast, nor at its rear to the southeast.

In the western part of the Bachu uplift, a strong internal deformation occurs in between the bounding fault systems (Fig. 6). There, this large basement pop-up is relatively symmetric with several internal conjugate steep thrusts rooted in the basement and emerging at the surface (Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; Li et al., 2016e; Jiang et al., 2018; Yang et al., 2018b). Some of these thrusts (Gudongshan Fault System) join the Mazar Tagh deep fault system in the central bend of the Bachu structure, where a relay zone occurs between several fault segments (Fig. 6). However, the internal deformation of the Bachu high tends to decrease eastward of its central bend, with only a few thrust ramps and their associated folds, as well as smooth undulations (Fig. 6). Concomitantly, the Bachu pop-up becomes more asymmetric and acquires the geometry of a large fault-propagation fold with a backlimb broken by the Mazar Tagh deep fault system (Tong et al., 2012; Guo et al., 2016; He et al., 2016; Li et al., 2016e). Finally, at the eastern termination of the Bachu uplift, the offset of the conjugate Tumuxiuke and Mazar Tagh deep faults rapidly decreases

eastward, together with the associated basement-cored folding (Fig. 6).

Based on seismic profiles and well data, previous studies have demonstrated that the Bachu uplift is an inherited Protero-Paleozoic structure that was probably formed in extension before to be reactivated in compression during the early stages of the Tarim evolution (e.g., Carroll et al., 2001; Chen and Shi, 2003; Lin et al., 2012b, 2012c; Gao and Fan, 2014; Liu et al., 2015a; He et al., 2016; Wu et al., 2018). The restoration of this uplift at the beginning of the Cenozoic supports this former activity. Indeed, the compressional offsets accumulated by the Pre-Cenozoic deposits along the Bachu bounding and inner faults are not totally reset by this restoration (Fig. 7B). During the Cenozoic, these faults were again reactivated in compression. The Bachu uplift has then developed to the north of the foredeep represented by the Southwest depression, taking the place of the forebulge of the Western Kunlun foreland system (after the terminology of DeCelles and Giles, 1996). Indeed, initially at the surface at the beginning of the Cenozoic, as indicated by a major angular unconformity, the Cenozoic deposits then gradually onlapped and overlapped this reactivated uplift, which is now covered by a few 100s of m of Plio-Quaternary sediments (Figs. 7B and 9) (see also Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; He et al., 2016; Li et al., 2016e; Jiang et al., 2018).

5.5. Deformation amount

Along the two transects crossing the Western Kunlun thrust system, we quantified the shortening from the main faults bounding the southwestern Tarim Cenozoic series (i.e. from the Kashgar-Yecheng transfer system and Tiklik thrust system at the rear of the outer range) to the front of the Western Kunlun compressive system (i.e. to the front of the Bachu uplift) (see Figs. 6, 7A and 7B for the fixed reference lines set at the turning points in the centre of sections A and B). Accordingly, the shortening values and proportions given below do not account for the deformation accommodated within the inner range. In its outer range and foreland only, the Western Kunlun thrust wedge accommodates a large amount of Cenozoic compressive deformation, with a total shortening of ~32 and ~35 km measured west and east of the central bend of the system, respectively (Fig. 7A and 7B). These shortening estimates are in the range of the other values published from previous cross-section restorations of the outer range, which are comprised between 24.6 and 54 km (Jiang et al., 2013; Cheng et al., 2016b, 2017).

To the west, in the NNW-SSE oriented part of the range (transect A), most of the measured shortening concentrates in the outer range, with ~31,5 km (~98.4% of the ~32 km of total shortening) accommodated within the outer belt and ~0.5 km (~1.6%) within the basin. In the outer belt, ~9.5 km (~29.7% of the ~32 km of total shortening) are accommodated by the basement-involved thrust slices and ~22 km (~68.7%) by the cover duplex. Compared to these values, the shortening of ~0.5 km (~1.6%) accommodated within the basin on the Mazar Tagh deep fault system is almost negligible. However, it should be noted that our section A was built through the western end of the Tarim Basin to cross perpendicularly the various structures intersected from the Western Kunlun to the Tian Shan. Built perpendicularly to the Western Kunlun Range only, it would have crossed the western part of the Bachu uplift where a few additional kilometres of shortening seem to have been accommodated (e.g., Tong et al., 2012; Tang et al., 2014; Li et al., 2016e; Jiang et al., 2018; Yang et al., 2018b).

To the east, in the WNW-ESE oriented part of the range (transect B), the shortening is distributed between the outer range and the Bachu uplift, with ~30 km (~85.7% of the ~35 km of total shortening) accommodated within the outer belt and ~5 km (~14.3%) within the basin. In the outer belt, ~16.5 km (~47.1% of the ~35 km of total shortening) are accommodated by the basement-involved thrust slices and ~13.5 km (38.6%) by the cover duplex. Consequently, these ~13.5 km of shortening had to be transferred to the Mazar Tagh emergent thrust ramps through the shallow Cenozoic decollement. Across the Bachu uplift, ~2 km (~5.7% of the ~35 km of total shortening) are accommodated by the Mazar Tagh deep fault system and ~3 km (~8.6%) by the Tumuxiuke thrust system. The resulting ~5 km of shortening accommodated by the whole Bachu uplift is thus in agreement with the values of a few kilometres published earlier (Li et al., 2016e).

According to these estimates, the shortening accumulated in the Western Kunlun thrust wedge outward from the inner range is of the same order of magnitude to the west and to the east of the central bend of the mountain belt. The structure and high topography of the range interior, together with the large flexure of the Southwest depression, indicate that a large amount of shortening must also be accommodated by the internal part of the system. However, in the absence of stratigraphic markers, it was not possible to restore and quantify this shortening from the geological transects.

In addition, previous investigations demonstrated that a significant amount of strike-slip movements is accommodated on major strike-slip fault systems in the range, such as the Karakax fault or the Kashgar-Yecheng transfer system (Ding et al., 2004; Cowgill, 2010). Strike-slip motion could also have occurred in the basin, for example in the western part of the Bachu uplift or along the Mazar Tagh deep fault system where very steep faults and relatively symmetric structures exist. However, our 2D restoration does not allow for quantifying such a deformation that takes place outside of the planes of our sections.

6. Southeast deformation

To the southeast of the Tarim Basin, the northwestward-vergent Altyn Tagh strike-slip system can be divided into three structural parts with, from the southeast to the northwest: (1) the inner range from the left-lateral Altyn Tagh fault to the Tam Karaul and Northern Altyn Tagh thrusts, (2) the outer range and Southeast depression from the Tam Karaul and Northern Altyn Tagh thrusts to the front of the Minfeng, Ruoqiang, and Dunhuang fault systems, and (3) the Tanan uplift (Figs. 6, 7C, 7D, 8C, 8D, S2C and S2D).

6.1. Altyn Tagh inner range

The inner range is several 10s to ~100 km wide, with typical elevations between ~2500 and ~5500 m. It is composed of exhumed imbricates of the Altyn Tagh Protero-Paleozoic (Sinian to Silurian-Devonian) basement and Paleo-Mesozoic (Carboniferous and Jurassic) cover (see section 2.3). However, this inner range is discontinuous and only developed to the southwest and in the centre of the Altyn Tagh system (Fig. 6). Indeed, except to the southwest where it is in continuity with the southeastern part of the Western Kunlun interior, the Altyn Tagh inner range disappears where its frontal thrusts (the Tam Karaul and Northern Altyn Tagh thrusts) laterally join and merge into the major left-lateral Altyn Tagh fault at the rear. This fault organization pattern is typical of transpressive wedges. As for the Western Kunlun Range, the detailed structure of the Altyn Tagh inner range is poorly constrained, but connecting the major faults to a deep basal decollement within the basement seems a reasonable hypothesis (see also Wittlinger et al., 1998 and Tang et al., 2014). To the northwest, the range interior is separated from the outer range by the Tam Karaul and Northern Altyn Tagh emergent thrusts (Fig. 6). With the Altyn Tagh fault, these thrusts correspond to the main bordering faults of the southeastern Cenozoic series of the Tarim Basin.

6.2. Altyn Tagh outer range and Southeast depression

At the front of the inner range, the outer range is several 10s to ~60 km wide, with typical elevations between ~1500 and ~2500 m. To our knowledge, no specific cross-section has ever been proposed so far for this outer range. However, contrary to the Western Kunlun compressive system bordering the Tarim Basin to the southwest, the

Altyn Tagh transpressive system does not have any fold-and-thrust belt in its outer part. According to the surface and subsurface data, the outer range is mainly composed of several-10s-of-km-long basement-involved structures made of the Altyn Tagh Protero-Paleozoic basement and its Paleo-Mesozoic to Cenozoic (Carboniferous to Quaternary) cover (Figs. 7C, 7D, 8C, 8D, S2C and S2D). These structures are related to northwestward-vergent, blind or emergent thrust ramps forming the Minfeng, Ruoqiang and Dunhuang fault systems. These ramps necessarily root into a deep decollement within the basement, which is probably in continuity with the one below the inner range.

With no sediments older than the Carboniferous (see also XBGMR, 1985; Li et al., 1996; Tang et al., 2014; He et al., 2016), the Paleozoic stratigraphy of the outer belt is similar to the one of the inner range (e.g., Sobel and Arnaud, 1999; Chen et al., 2003). In the outer range, unconformities are also frequent in the Paleo-Mesozoic strata or at the base of the Cenozoic deposits in the hanging wall of the thrust ramps (Figs. 7C, 8C and S2C) (see also XBGMR, 1985; Li et al., 1996; Tang et al., 2014; He et al., 2016). Moreover, the restoration of the section at the beginning of the Cenozoic does not totally reset the compressional offsets accumulated by the older sediments along these faults (Fig. 7C). All this implies that the latter are inherited structures reactivated during the Cenozoic. In the Altyn Tagh outer range, the whole Cenozoic deformation is associated with these inherited faults since there is no evidence of younger structures rooted in a decollement level in the Paleozoic or Cenozoic sedimentary series.

Associated with these reactivated thrusts, the Paleo-Mesozoic and Cenozoic sediments are gently folded into 10s-of-km-wide and up to 100s-of-km-long anticlines and synclines. To the southwest, the Cenozoic deposits related to the erosion of the Altyn Tagh cover these folds and the topographic mountain front is located behind, at the front of the inner range. To the northeast, they form smooth topographies uplifted several 100s of m above the present-day Tarim sedimentation level and localize the topographic mountain front.

In the Tarim Basin, the area at the front of these folds is known as the Southeast depression (Figs. 6, 7C, and 7D). In this area, the Tarim basement and sedimentary cover are gently tilted southeastward, with the Cenozoic deposits slightly thicker at the front of the Altyn Tagh outer belt than further northwest in the basin (Figs. 7C, 7D, 8C, 8D, S2C and S2D). This overall geometry of the southeast Tarim thus indicates a slight flexure associated with the shortening and thickening accumulated in the Altyn Tagh strike-slip system. However, with ~3 km of trapped Cenozoic sediments at most, the generated accommodation space remains moderate. This suggests a limited compressive deformation accommodated across the Altyn Tagh Range, which is somehow expected when compared to a thrust wedge such as the Western Kunlun Range for example. Moreover, the southwestern part of the Altyn Tagh foredeep is broken by the Tanan basement uplift, which probably explains the limited accommodation space in this foreland (Strecker et al., 2012).

6.3. Tanan uplift

To the north of the Tarim Southeast depression, the Tanan uplift (also known as the Qiemo uplift) is a ~50-km-wide and ~600-km-long basement-involved structure forming a structural high of Proterozoic basement and Paleo-Mesozoic (Carboniferous to Cretaceous) cover, below the Cenozoic sediments of the basin (Figs. 6, 7C, 8C and S2C) (see also Lin et al., 2012b, 2012c; Tang et al., 2014; Li et al., 2015c; Liu et al., 2015a; He et al., 2016). From its overall geometry documented by seismic profiles, together with field and well data, this basement-cored structure corresponds to a large pop-up bounded by two conjugate fault systems necessarily rooted in a deep decollement within the basement below the Tarim. To the south, the Niya fault system is composed of a few conjugate thrust faults that locally delimitate a small pop-up, at the tip of a southward-vergent steep thrust ramp (Figs. 6, 7C, 8C and S2C) (see also Tang et al., 2014; Li et al., 2015c). To the north, the Cherchen fault system displays fault-related folds and steep thrust ramps with a clear northward vergence and a hangingwall uplift of up to ~1.5 km, which is larger than that of the Niya fault system (Figs. 6, 7C, 8C and S2C) (see also Lin et al., 2012b, 2012c; Tang et al., 2014; Li et al., 2015c; Liu et al., 2015a; He et al., 2016; Jiang et al., 2018). Consequently, the Tanan pop-up has clearly developed parallel to the Altyn Tagh Range over its entire length (Fig. 6), and its main vergence appears to be toward the northwest, i.e. synthetic to the Altyn Tagh deformation. However, in spite of this clear northwestward vergence, there is no evidence of flexural subsidence at the front of the Tanan uplift (Figs. 7C, 8C and S2C).

In its central part, where the two conjugate bounding fault systems are present, this large pop-up is relatively symmetric (Figs. 6, 7C, 8C and S2C) (see also Li et al., 2015c). However, where the Niya fault system vanishes laterally, the Tanan pop-up becomes more asymmetric and acquires the geometry of a large fault-propagation fold (Liu et al., 2015a; Jiang et al., 2018). Finally, at the lateral terminations of the Tanan uplift, the offset of the Cherchen fault system decreases and the associated basement-cored fold progressively flattens toward the eastern part of the Tarim Basin to the northeast and the eastern end of the Western Kunlun Range to the southwest (Fig. 6). In this area, the Hotan-Qira normal fault system is a series of conjugate normal faults described and interpreted by Avouac and Peltzer (1993) as related to the subsidence of the Tarim Basin loaded by the Western Kunlun mountain range. However, seismic profiles in this area (Fig. 1) led us to re-evaluate these faults as extrado normal faults above the southwestward vanishing basement-cored fold of the Tanan uplift (see also Li et al., 2018b).

Based on seismic profiles, together with field and well data, previous studies have demonstrated that the Tanan uplift is an inherited Protero-Paleozoic compressive structure active during the early stages of the Tarim evolution (e.g., Lin et al., 2012b, 2012c; Gao and Fan, 2014; Lin et al., 2015a; Liu et al., 2015a; He et al., 2016). Here again, the restoration of this uplift at the beginning of the Cenozoic supports a former activity since the compressional offsets accumulated by the Pre-Cenozoic deposits along the Tanan bordering faults are not totally reset (Fig. 7C). As for the Bachu uplift, these faults were then reactivated in compression during the Cenozoic. However, although well visible on seismic profiles all along the Altyn Tagh Range, the Cherchen fault system has not been reactivated significantly all along its entire length. Indeed, at the eastern end of the basin, this fault system does not yet present any geometrical evidence for a significant cumulative Cenozoic deformation (Figs. 7D, 8D and S2D) (see also Morin et al., 2019). Yet, reactivation of this part of the fault system could be currently ongoing since some clues of recent activity exist at its eastern termination (Chang et al., 2014a).

Where it is reactivated, the Cherchen fault system now corresponds to the present-day hidden to emergent deformation front of the Altyn Tagh wedge, located in the basin ~100–150 km away from the topographic mountain front. Remarkably, this structure also localizes a major stratigraphic change between the Altyn Tagh Paleozoic domain with Silurian to Devonian metamorphic and plutonic rocks covered by Carboniferous deposits (see section 2.3) to the south, and the Tarim Paleozoic domain with Cambrian to Permian sediments (see section 2.1) to the north (Figs. 7C and 8C) (e.g., Lin et al., 2012b, 2012c; Tang et al., 2014; Li et al., 2015c; Liu et al., 2015a; He et al., 2016). Accordingly, the Cherchen fault system is a major inherited structure equivalent to the deep basement ramp below the Western Kunlun outer belt, where a similar change between two stratigraphic domains is localized. Together, these structures therefore represent a major tectonic palaeo-boundary between the Paleozoic orogenic systems of the Tibetan edge to the south and the contemporaneous Tarim Basin to the north.

This tectonic palaeo-boundary was thus reactivated during the Cenozoic all along the northern edge of the Tibetan Plateau. In the foreland of the Western Kunlun compressive system, the deformation

also propagated as far north as the Bachu uplift (Figs. 6, 7B and S2B), whereas in the foreland of the Altyn Tagh transpressive system, there is no evidence of Cenozoic activity further north than the Tanan uplift (Figs. 6, 7C and S2C), even though some inherited Protero-Paleozoic structures, known as the Tazhong and Tadong uplifts, also exist in the central part of the basin (Figs. 6, 7C, 7D, S2C and S2D) (see also Lin et al., 2012b, 2012c; Tang et al., 2014; Lin et al., 2015c; Liu et al., 2015a; He et al., 2016). Indeed, there is little evidence of a significant Cenozoic activity on the bounding faults of these basement highs, and if deformation occurs in this part of the basin, it remains very limited. In this undeformed area of the Tarim, the Cenozoic deposits are typically less than ~2 km thick.

6.4. Deformation amount

Along the two transects crossing the Altyn Tagh strike-slip system, we quantified the shortening from the main faults bounding the southeastern Tarim Cenozoic series (i.e. from the Tam Karaul and Northern Altyn Tagh thrust faults at the rear of the outer range) to the centre of the basin (see Figs. 6, 7C and 7D for the fixed reference lines set at the turning points in the centre of sections C and D). Accordingly, the shortening values and proportions given just below do not account for the deformation accommodated within the inner range. In its outer range and foreland only, the Altyn Tagh strike-slip wedge accommodates a small amount of Cenozoic compressive deformation, with a total shortening of ~0.9 and ~0.3 km measured to the west and to the east of the system, respectively (Fig. 7C and 7D).

To the west (transect C), the measured shortening is distributed between the outer range and the Tanan uplift, with ~0.3 km (~33.3% of the above ~0.9 km of total shortening) accommodated within the outer belt and ~0.6 km (~66.7%) within the basin. Across the Tanan uplift, ~0.2 km (~22.2% of the ~0.9 km of total shortening) are accommodated by the Niya fault system and ~0.4 km (~44.5%) by the Cherchen thrust and associated faults. To the east (transect D), the entire measured ~0.3 km shortening concentrates in the outer range and is accommodated by the local basement thrust ramps.

According to these estimates, the shortening accumulated within the Altyn Tagh strike-slip wedge outward from the inner range is very limited. Morevover, it decreases from the southwest to the northeast to become almost negligible at the eastern termination of the system. The structure and high topography of the inner range indicate that most of the shortening must rather be accommodated in this internal part of the system. However, even though we were not able to restore and quantify this deformation given the lack of stratigraphic markers in the range interior, the slight flexure associated with the Altyn Tagh wedge suggests a limited cumulative shortening and thickening, as expected for such a strike-slip system. Indeed, previous studies demonstrated that a large amount of strike-slip deformation is accommodated on major strike-slip faults in the range such as the Altyn Tagh fault (e.g., Wang, 1997; Cowgill et al., 2000, 2003; Sobel, 2001; Yue et al., 2001; Meng et al., 2001; Yin et al., 2002; Gehrels et al., 2003; Ding et al., 2004; Cheng et al., 2015, 2016a). Strike-slip motion may also have occurred in the basin, along the Niya and Cherchen bounding fault systems of the Tanan uplift, where very steep thrusts and relatively symmetric structures exist (see also Avouac and Peltzer, 1993; Yin et al., 2002). However, our 2D restoration does not allow for quantifying these movements, which take place outside of the planes of our sections.

7. Northwest deformation

To the northwest of the Tarim Basin, the southeastward-vergent Southwestern Tian Shan thrust system can be divided into three structural parts with, from the northwest to the southeast: (1) the inner range from the Nikolaev Line and Narat fault to the Muziduke thrust and Southern Tian Shan fault system, (2) the outer range from the Muziduke thrust and Southern Tian Shan fault system to the Kashgar deformation front, Kepingtage thrust, Kalpintagh fault, and Kuqa deformation front, and (3) the Southwest and North depressions at the front of the Kashgar deformation front, Kalpintagh thrust and Kuqa deformation front (Figs. 6, 7A, 7B, 7C, 8A, 8B, 10C, S2B and S2C).

7.1. Southwestern Tian Shan inner range

The inner range is several 10s to ~200 km wide, with typical elevations between ~2500 and ~7400 m. It is composed of exhumed imbricates of the Southern Tian Shan Proterozoic basement, Paleozoic (Devonian to Permian) plutons, and Paleozoic-Mesozoic (Silurian to Cretaceous) cover (see section 2.4). It also comprises a few compressive intra-mountainous basins filled by Cenozoic sediments, such as the Bayanbulak Basin for example (Fig. 6). In the Southwestern Tian Shan system, the inner range is a continuous belt crossed by the Talas Fergana fault to the southwest and in continuity with the Southeastern Tian Shan interior to the northeast (Fig. 6). Despite numerous surface and subsurface studies, its detailed structure remains controversial. To the southwest, connecting the major basement faults, such as the Muziduke thrust and Southern Tian Shan fault system, to a deep basal decollement seems a reasonable hypothesis (Figs. 7A, 7B, 10A and 10B) (see also Buslov et al., 2007; Gao et al., 2013; Jia et al., 2015). However, alternative interpretations have been proposed with these major faults crossing the entire crust or branching onto the Southern Tian Shan fault system (e.g., Allen et al., 1999; Loury et al., 2017; Makarov et al., 2010; Xiao et al., 2013). To the northeast, the Southern Tian Shan fault system is much steeper and possibly cuts through the whole crust since its location at the surface correlates with a step in the Moho at depth (Figs. 7C and 10C) (Li et al., 2007; Qi et al., 2009; Wen et al., 2017). Alternatively, this major basement fault system has also been suggested to connect at depth into a deep basal decollement (e.g., Yin et al., 1998; Poupinet et al., 2002). Toward the southeast, the range interior is then separated from the outer range by the Muziduke thrust to the southwest and the Southern Tian Shan fault system to the northeast (Fig. 6). These emergent thrust zones also correspond to the main bordering faults of the northwestern Cenozoic series of the Tarim Basin.

Remarkably, in the western and central parts of the Southwestern Tian Shan, the stratigraphy of the Paleozoic terrains is different on both sides of the Muziduke thrust and Southern Tian Shan fault system (Figs. 7A, 7B, 10A and 10B). To the northwest, this stratigraphy is characteristic of the Southern Tian Shan Paleozoic Range with Silurian to Devonian metamorphic and plutonic rocks covered by Devonian to Permian deposits (see section 2.4), whereas to the southeast, it is typical of the Paleozoic Tarim Basin with Cambrian to Permian sediments (see section 2.1) (Fig. 2). Therefore, the Muziduke thrust and Southern Tian Shan fault system seem to be major inherited Paleozoic structures that have been reactivated during the Cenozoic. Such a major stratigraphic change also occurs in the northeastern part of the Southwestern Tian Shan, but it seems to be located slightly to the south, within the outer range.

7.2. Southwestern Tian Shan outer range and Kuqa depression

At the front of the inner range, the outer range is also several 10s to ~100 km wide, with typical elevations between ~1000 and ~2500 m. It consists in four different fold-and-thrust belts that developed below and above the Cenozoic sediments of the Tarim Basin (Figs. 6, 7A, 7B, 7C and S2C). These four belts present different styles of deformation all along the range. They are presented below from the southwest to the northeast.

7.2.1. Kashgar-Artux fold-and-thrust belt

To the southwestern end of the Southwestern Tian Shan outer range, the Kashgar-Artux fold-and-thrust belt is a ~150-km-long belt located between the Muziduke thrust and the Kashgar deformation front

(Fig. 6). Several cross-sections have already been proposed across this belt (e.g., Scharer et al., 2004; Sobel et al., 2006; Heermance et al., 2007, 2008). The surface and subsurface data provide clear evidence for 10s-of-km-long, emergent or blind, folded thrust slices made of the Tarim Paleo-Mesozoic (Cambrian to Cretaceous) sedimentary cover at the rear, whereas several-km-wide and 10s-of-km-long fault-propagation or detachment folds developed within the Tarim Cenozoic (Paleogene to Quaternary) deposits at the front.

Together with the basement-involved imbricates of the inner range, these structures imply a basal decollement level that progressively shallows from the northwest to the southeast. Indeed, in the range interior, the major thrusts probably root in a deep decollement within the basement. At the transition between the inner and outer ranges, the change from thick-skinned to thin-skinned thrust slices that operates within the Paleo-Mesozoic cover then requires the presence of a ramp connecting this deep basement decollement to an intermediate one, most probably within the upper Cambrian evaporitic formation found in the western Tarim sedimentary series (e.g., Heermance et al., 2007, 2008). However, the poor quality of the seismic profiles and the lack of deep well data do not allow to precisely constrain the depth of this Paleozoic decollement in this part of the basin.

Toward the front of the outer belt, the formation of thrusts and folds within the Tarim Cenozoic cover then required the activation of a second ramp connecting this intermediate decollement at the base of the deformed Paleo-Mesozoic series to a shallow one at its top. According to the seismic profiles, together with field and well constraints, this shallow decollement corresponds to the Paleogene evaporitic assemblage located at the base of the Cenozoic sediments, at a depth of about 6 to 9 km b.s.l. in this area (e.g., Scharer et al., 2004; Sobel et al., 2006; Heermance et al., 2007, 2008). In the Kashgar-Artux belt, there is no evidence for a structural inheritance such as stratigraphic changes or non-restorable offsets across faults. Accordingly, all the structures developed in this belt likely correspond to structures created during the Cenozoic.

Although partially covered by the Cenozoic deposits coming from the Southwestern Tian Shan erosion, the frontal folds of the Kashgar-Artux belt form smooth topographies uplifted several 10s to 100s of m above the present-day Tarim sedimentation level (Fig. 6). Along this belt, the local topographic front of the mountain range is thus localized by the detachment fold associated with the Kashgar deformation front, beyond which the Cenozoic deposits of the foreland have remained undeformed and attached to the underlying series.

7.2.2. Kepingtage fold-and-thrust belt

In the southwestern part of the Southwestern Tian Shan outer range, the Kepingtage fold-and-thrust belt is a ~230-km-long belt comprised between the Muziduke thrust to the northwest or Southern Tian Shan fault system to the northeast, and the Kepingtage thrust (Fig. 6). Several cross-sections have already been proposed through the Kepingtage belt (e.g., Yin et al., 1998; Allen et al., 1999; Turner et al., 2010; Tang et al., 2014; Yang et al., 2018b). In agreement with the previous interpretations, we concluded that the surface and subsurface data provide clear evidence for a-few-10s-of-km-wide and 10s-of-km-long thrust sheets made of the Tarim Paleozoic (Cambrian to Permian) sedimentary cover, with fault-propagation folds related to emergent thrust ramps at their front (Figs. 7 and 10A). These thrust sheets also include Cenozoic (Paleogene to Quaternary) sediments deposited in up to ~4-km-deep piggy-back basins. Associated with these compressive structures, a set of strike-slip faults perpendicular to the thrust sheets also exists in the Kepingtage belt, including the left-lateral Piqiang fault (Fig. 6) (see Allen et al., 1999; Turner et al., 2010, 2011; Yang et al., 2018b).

As in the Kashgar-Artux belt, the transition from the basement imbricates of the inner range to the sedimentary cover thrust sheets of the Kepingtage belt implies a basal decollement level that progressively shallows from the northwest to the southeast. Indeed, the major thrusts of the range interior probably root in a deep decollement within the basement, while the change from thick-skinned to thin-skinned thrust slices that involves the Paleozoic and Cenozoic cover requires the presence of a ramp connecting this deep decollement to the intermediate one within the upper Cambrian evaporitic formation of the western Tarim sedimentary series (Fig. 2) (e.g., Allen et al., 1999; Turner et al., 2010). According to the seismic profiles and field constraints, this Paleozoic decollement is located here at a depth of about 4 to 6 km b.s.l. (Figs. 7A and 10A) (see also Allen et al., 1999; Turner et al., 2010; Yang et al., 2018b). In the Kepingtage belt, the whole deformation is accommodated by the thrust sheets that grew above this decollement level, since there is no evidence of structures rooted in a shallower one within the Cenozoic sedimentary series.

Based on field data, previous studies have documented unconformities, as well as SW-NE thickness and facies changes, in the Paleozoic sedimentary series of the Kepingtage belt (e.g. Allen et al., 1999; Turner et al., 2010, 2011; Chang et al., 2012b). Moreover, the strike-slip faults offsetting the belt are aligned with the faults present in the northwestern part of the Bachu uplift to the southeast. These observations suggest that the Kepingtage area was the place of a large inherited structure, in continuity with the one of the Bachu uplift, which has also been reactivated during the Cenozoic (e.g. Allen et al., 1999; Carroll et al., 2001; Turner et al., 2010, 2011; Yang et al., 2018b). However, the reactivated strike-slip faults appear to be perpendicular to the compressive structures of the Kepingtage belt, for which there is no evidence of a structural inheritance such as stratigraphic changes or non-restorable offsets across the faults. Therefore, the shallow thrusts rooting into the Paleozoic decollement likely correspond to faults created during the Cenozoic above a large basement-involved structure, whose inheritance rather controls the occurrence of orthogonally-oriented strike-slip deformation (e.g., Turner et al., 2010, 2011; Yang et al., 2018b).

Although partially covered by the Cenozoic deposits resulting from the erosion of the Southwestern Tian Shan, the frontal part of the Kepingtage thrust sheets form sharp topographies uplifted several 100s of m to more than ~1 km above the present-day Tarim sedimentation level (Fig. 6). In this area, the local topographic mountain front is thus localized by the Kepingtage thrust ramp emerging at the front of the belt.

7.2.3. Kalpin belt

In the central part of the Southwestern Tian Shan outer range, the Kalpin belt is a ~80-km-long belt located between the Southern Tian Shan fault system and the Kalpintagh fault (Fig. 6). As far as we know, there has been no published cross-section across this belt. According to the surface and subsurface data, this belt is composed of 10s-of-km-long, emergent and folded thrust slices and small pop-ups related to a few steep thrust ramps. These structures involved the Tarim Proterozoic basement and its Protero-Paleozoic (Sinian to Permian) cover (Figs. 7B and 10B). They also individualize two basins containing less than ~2 km of Cenozoic (Plio-Quaternary) deposits. At depth, these structures possibly connect to a deep decollement within the basement. In the Kalpin belt, like in the Kepingtage one, there is no evidence of structures rooting into a shallow decollement level within the Cenozoic sedimentary series.

According to field data, the Kalpin area, which is known as the Kalpin uplift, appears to be a lateral extension of the large inherited Kepingtage basement structure (e.g. Allen et al., 1999; Carroll et al., 2001; Turner et al., 2010; Chang et al., 2012b). Consequently, in contrast with the neoformed compressive structures of the thin-skinned Kepingtage fold-and-thrust belt, the basement-involved structures of the thick-skinned Kalpin belt are probably related to inherited basement faults reactivated during the Cenozoic, such as possibly for the frontal Kalpintagh thrust fault.

Although partially covered by the Cenozoic deposits coming from the erosion of the Southwestern Tian Shan, the structures of the Kalpin belt now form sharp topographies uplifted several 100s of m up to

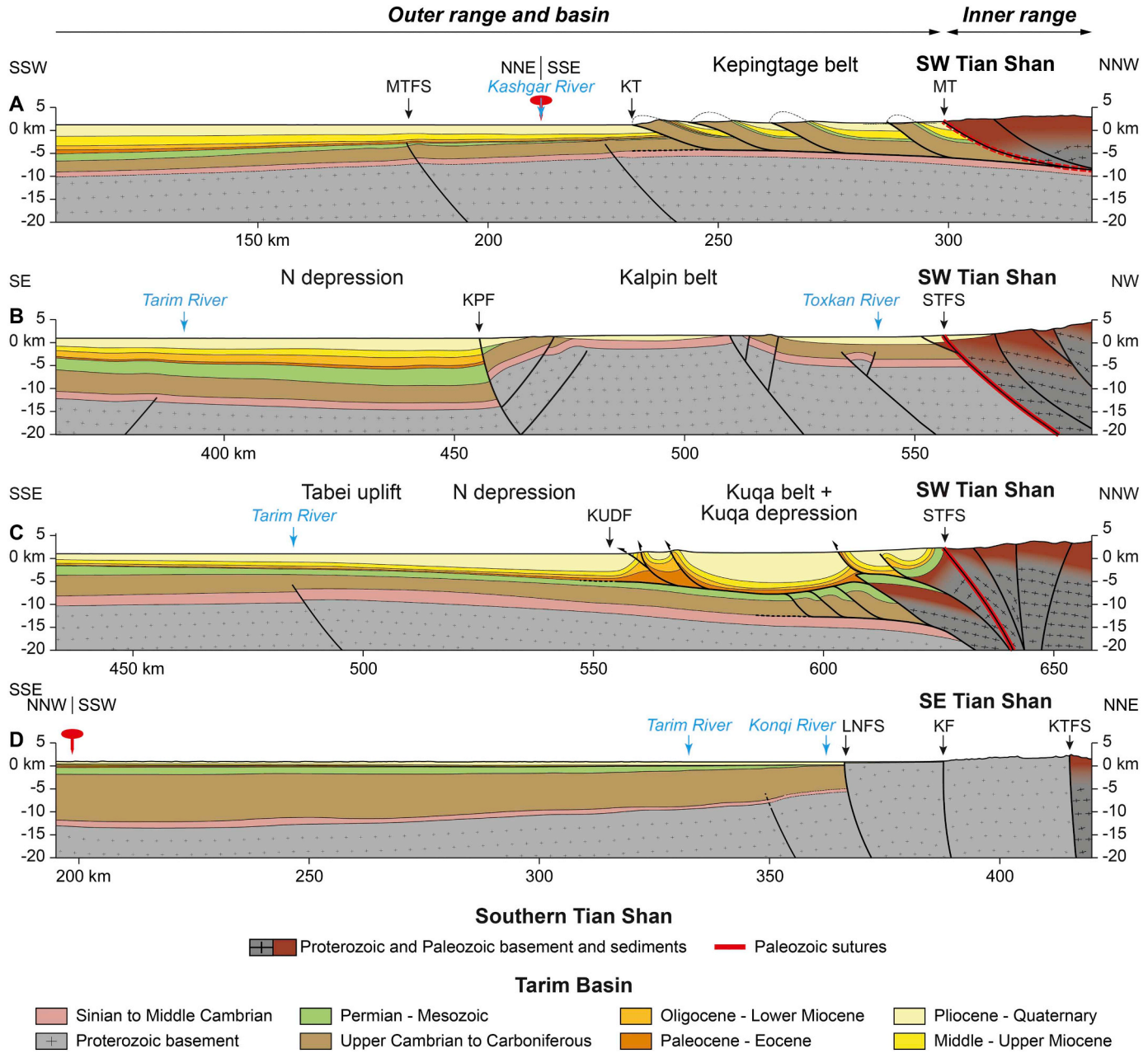

Fig. 10. Details of the regional balanced geological transects across the northern part of the Tarim Basin and the Southern Tian Shan Range (see Fig. 7 for location and Fig. 6 for abbreviation labels).

~1.5 km above the present-day Tarim sedimentation level (Fig. 6). In this area, the local topographic front of the mountain range is localized by the Kalpintagh fault emerging at the front of the belt.

7.2.4. Kuqa fold-and-thrust belt and depression

In the northeastern part of the Southwestern Tian Shan outer range, the Kuqa fold-and-thrust belt is a ~370-km-long belt located between the Southern Tian Shan fault system and the Kuqa deformation front (Fig. 6). Numerous contradictory cross-sections have been proposed through this belt (e.g. Burchfiel et al., 1999; Chen et al., 2004; He et al., 2005; Qi et al., 2009; Li et al., 2012a; Wu et al., 2014; Zhang et al., 2014b; Li et al., 2016d; Wen et al., 2017; Zhao and Wang, 2016; Wang et al., 2017; Izquierdo-Llavall et al., 2018; Neng et al., 2018). From our surface and subsurface dataset, we derived a regionally-consistent structural model settling the issue of these contradictory interpretations (see Supplementary information S1 for further details and discussion). At the rear of the belt, the surface and subsurface data provide evidence for 10s-of-km-long, emergent or blind, folded thrust slices constituted of the Southern Tian Shan Proterozoic basement, Paleozoic (Devonian to Permian) plutons and Paleo-Mesozoic (Silurian to Cretaceous) cover. At the front, 10s-of-km-long duplex structures involve the Tarim Paleo-Mesozoic (Cambrian? to Cretaceous) sedimentary cover (Figs. 7C, 10C and S2C). The surface and subsurface data also reveal several-km-wide and up to ~200-km-long fault-propagation or detachment folds that developed within the Tarim Cenozoic (Paleogene to Quaternary) deposits at the top and the front of the basement-involved thrust slices and Paleo-Mesozoic cover duplexes (Figs. 7C, 10C and S2C). Finally, these folds individualize a few piggy-back basins, such as the ~30-km-wide and ~130-km-long Kuqa depression where ~9 km of Cenozoic sediments are stacked nearby the city of Baicheng (Fig. 6, 7C and 10C).

These structures imply a basal decollement level that progressively shallows from the north to the south. At the rear of the Kuqa belt, the

major thrusts necessarily connect to a deep decollement within the basement or directly onto the Southern Tian Shan fault system. The southeastward change from the thick-skinned thrust slices to the thin-skinned duplexes then requires the presence of a ramp transmitting slip from depth to an intermediate decollement at the base of the deformed Paleo-Mesozoic sedimentary cover (Figs. 7C and 10C). However, in this part of the basin, the quality of the seismic profiles and the lack of deep well data do not allow for precisely constraining the age, lithology and depth of this intermediate decollement that remains unclear (compare the cross-sections of Chen et al., 2004; He et al., 2005; Qi et al., 2009; Li et al., 2012a; Wen et al., 2017; Wang et al., 2017; Izquierdo-Llavall et al., 2018; Neng et al., 2018; and Supplementary information S1 for more details on the different interpretations).

Nonetheless, the stratigraphy of the Paleozoic terrains appears to be different on both sides of the ramp connecting the deep and intermediate decollements (Figs. 7C and 10C). To the northwest, this stratigraphy is characteristic of the Southern Tian Shan Paleozoic domain with Silurian to Devonian metamorphic and plutonic rocks directly covered by Devonian to Permian deposits (see section 2.4) (Fig. 2). To the southeast, the Proterozoic basement of the Tarim Basin could be topped by a lower Paleozoic series, even if the latter is truncated by an unconformity located at the base of the overlying Mesozoic sediments (Fig. 2). The occurrence within this Paleo-Mesozoic cover of a sedimentary level acting as an intermediate decollement in the Kuqa belt, but which does not exist at the surface further north, is an additional indication of a significant stratigraphic change between these two Paleozoic domains with Southern Tian Shan vs. Tarim stratigraphic affinities. Therefore, the ramp connecting the deep and intermediate decollements to the northwest of the Kuqa belt seems to be a major inherited Paleozoic structure reactivated during the Cenozoic, similar to the Muziduke thrust and the Southern Tian Shan fault system further west. Together, these structures thus represent a major structural palaeo-boundary between the Paleozoic orogenic systems of the Southern Tian Shan to the north and the contemporaneous Tarim Basin to the south.

At the front of the Kuqa belt, the formation of duplexes within the Tarim Paleo-Mesozoic cover and of thrusts and folds within the Cenozoic deposits requires the activation of blind ramps connecting the intermediate decollement to a shallow one between the two sedimentary series (Figs. 7C, 10C and S2C). Based on the field constraints, seismic profiles, and well data, this shallow decollement is located at a depth of about 5 to 8 km b.s.l. at the front of the duplexes, within the Paleogene evaporitic assemblage of the Cenozoic deposits (see also Chen et al., 2004; He et al., 2005; Li et al., 2012a; Wu et al., 2014; Tian et al., 2016; Izquierdo-Llavall et al., 2018; Neng et al., 2018). Below this decollement, the interpretation and reconstruction of the underlying duplexes may be problematic because of a strong component of salt tectonics, as well as a seismic imaging degraded by the presence of evaporites (e.g., Chen et al., 2004; He et al., 2005; Li et al., 2012a; Wu et al., 2014; Li et al., 2016d; Tian et al., 2016; Wang et al., 2017; Izquierdo-Llavall et al., 2018; Neng et al., 2018). However, the stratigraphy of the Paleo-Mesozoic cover seems to be the same across the ramps connecting the Paleozoic and Cenozoic decollements. Therefore, these shallow ramps in the centre of the Kuqa belt likely correspond to structures created during the Cenozoic.

Above the basement-involved thrust slices and Paleo-Mesozoic cover duplexes, the slip is transferred toward the top and the front of the Kuqa belt through the Cenozoic decollement. Within the Cenozoic deposits, long fault-propagation or detachment folds developed in association with emergent or blind thrust ramps, often with a strong component of salt tectonics (Figs. 7C and 10C) (e.g. Chen et al., 2004; He et al., 2005; Hubert-Ferrari et al., 2007; Li et al., 2012a; Wu et al., 2014; Li et al., 2016d; Tian et al., 2016; Zhao and Wang, 2016; Wang et al., 2017; Izquierdo-Llavall et al., 2018; Neng et al., 2018). These folds are located just at the front of the structures involving the basement and the Paleozoic cover below the Cenozoic decollement, or a few 10s of km further south. In this case, these most frontal folds individualize the Kuqa depression that corresponds to a piggy-back basin developed within a wide syncline at their rear (Figs. 6, 7C and 10C).

Although largely covered by the Cenozoic deposits resulting from the erosion of the Southwestern Tian Shan, these frontal folds localize the topographic front of the mountain range. At the surface of the basin, they usually form smooth topographies uplifted several 100s of m above the present-day Tarim sedimentation level (Fig. 6). Along the belt, the topographic front is thus localized by the detachment folds at the front of the Kuqa belt. Further southeast, the Cenozoic deposits of the foreland have remained undeformed and attached to the underlying series.

7.3. Southwest and North depressions

The Tarim Basin at the front of the Southwestern Tian Shan outer range can be considered in terms of different sub-areas, which are presented below from the southwest to the northeast.

The Tarim Basin sub-area located at the front of the Kashgar-Artux belt corresponds to the northwestern termination of the Southwest depression (Fig. 6). There, this depression corresponds to the region where the accommodation space is maximal in the Tarim Basin, with up to ~10 km of Cenozoic sediments nearby Kashgar. Indeed, the added loads of the Western Kunlun and Southwestern Tian Shan thrust wedges probably explain the very strong flexural subsidence of this specific area. In this foreland system, the Western Kunlun outer belt and the Kashgar-Artux belt correspond to two facing wedge-top depozones, and the Southwest depression located between these belts to a triangular foredeep, according to the terminology of DeCelles and Giles (1996).

At the front of the Kepingtage belt, there is no such depression (Figs. 7A and 10A). Indeed, this belt is thrusted over the western part of the Bachu uplift, which clearly prevents the development of a flexural foredeep (Fig. 6). Moreover, the Bachu high seems to persist and be also active below the Kepingtage belt (see also Allen et al., 1999; Turner et al., 2010), possibly interfering with its deformation. In particular, this may explain why the Kepingtage wedge-top depozone (terminology of DeCelles and Giles, 1996) is mainly developed above the sedimentation level, with deeply-eroded anticlines at the front of the thrust sheets and shallow piggy-back basins at their top.

Finally, the Tarim Basin sub-area located at the front of the Kalpin and Kuqa belts is known as the North depression (Fig. 6). In this area, the Tarim basement and sedimentary cover are tilted northward so that the Cenozoic deposits thicken toward the Southwestern Tian Shan and reach their maximum thickness at the front of its outer belt (Figs. 7B, 7C, 10B, 10C, S2B and S2C). This overall geometry of the northern Tarim is thus typical of a flexural basin, associated here with the shortening and thickening accumulated in the Southwestern Tian Shan thrust system. It corresponds to a region of high accommodation space, with up to ~7 km of Cenozoic sediments to the south of Aksu, i.e. at the front of the Kalpin belt (Figs. 7B, 10B and S2B). Moreover, to the northeast, the North depression is bounded by the Kuqa belt, which in turn surrounds the Kuqa depression where up to ~9 km of Cenozoic deposits are trapped close to the city of Baicheng (Figs. 7C and 10C). In this foreland system, the Kuqa belt and its associated depression correspond to the wedge-top depozone, the North depression at the front of this belt represents the foredeep, and the undeformed central area of the basin constitutes the forebulge (terminology of DeCelles and Giles, 1996).

At the transition between these foredeep and forebulge, an inherited Protero-Paleozoic basement-cored structure called the Tabei uplift exists. However, there is little evidence of a significant Cenozoic activity on the faults associated with this basement high, except to its very eastern part. There, its southern bounding fault appears to have been reactivated as a normal fault, most probably in response to the stresses

generated by the flexural bending of the northern Tarim Basin. However, this Cenozoic deformation occuring in the northern part of the basin remains very limited (Figs. 7C, 10C and S2C).

7.4. Deformation amount

Along the transects crossing the Southwestern Tian Shan thrust system, we quantified the shortening from the main faults bounding the northwestern Tarim Cenozoic series (i.e. from the Muziduke thrust and Southern Tian Shan fault system at the rear of the outer range) to the centre of the basin (see Figs. 6, 7A, 7B and 7C for the fixed reference lines set at the turning points in the centre of sections A, B and C). Accordingly, the shortening values and proportions given just below do not account for the deformation accommodated within the inner range. In its outer range only, the Southwestern Tian Shan thrust wedge already accommodates a large amount of Cenozoic compressive deformation, with a total shortening of ~36 and ~22 km measured to the southwest and to the northeast, i.e. through the Kepingtage and Kuqa belts, respectively (Fig. 7A and 7C). These shortening amounts are in the range of the other values published from previous cross-section restorations of these systems, which range from 23 to 50 km for the Kepingtage belt (e.g., Yin et al., 1998; Allen et al., 1999; Yang et al., 2007b; Turner et al., 2010), and from ~10 to ~42 km for the Kuqa one (e.g., Yin et al., 1998, Burchfiel et al., 1999; Li et al., 2012a; Zhang et al., 2014b; Tian et al., 2016; Wen et al., 2017; Izquierdo-Llavall et al., 2018). Unfortunately, in the absence of a preserved Mesozoic-Cenozoic stratigraphic marker, it was not possible to properly quantify the Cenozoic deformation accumulated in the central part of the Southwestern Tian Shan compressive system, i.e. through the Kalpin belt, since eroded material cannot be reconstructed. The proposed restoration therefore only provides a very low and probably meaningless minimum estimate of ~4 km for the total shortening across this belt.

In the Southwestern Tian Shan, the whole quantified shortening is accommodated in the outer range with no shortening within the basin. To the southwest, in the Kepingtage belt (transect A), the ~36 km of total shortening are a minimum estimate due to the eroded front of the cover thrust sheets, which accommodate 100% of the deformation. To the northeast, in the Kuqa belt (transect C), ~19 km (~86.4% of the ~22 km of total shortening there) are accommodated by the basement-involved thrust slices and ~3 km (~13.6%) by the cover duplexes.

According to these estimates, the shortening accumulated within the Southwestern Tian Shan thrust wedge outward from the inner range decreases from west to east. The structure and high topography of the range interior, together with the large flexure of the Southwest and North depressions, indicate that a large amount of shortening must also be accommodated in this internal part of the system. However, the absence of stratigraphic markers impedes here the quantification of this deformation using a structural restoration approach.

In addition to shortening, strike-slip deformation may be accommodated on strike-slip fault systems in the range (e.g., Yin et al., 1998; Zubovich et al., 2010). Strike-slip motion may also have occurred on faults at the margins of the basin, for example on the steep Kalpintagh fault (Li et al., 2015d). However, our 2D restoration does not allow for quantifying such movements, which take place outside of the planes of our sections.

8. Northeast deformation

To the northeast of the Tarim Basin, the southward-vergent Southeastern Tian Shan strike-slip system can be divided into two parts with, from the north to the south: (1) the inner range from the right-lateral Baluntay fault to the right-lateral Korla fault, and (2) the outer range from the Korla fault to the Lop Nur fault system (Figs. 6, 7D, 10D and S2D).

8.1. Southeastern Tian Shan inner range

The inner range is several 10s to ~140 km wide, with typical elevations between ~1000 and ~4800 m. It is composed of exhumed imbricates of the Southern Tian Shan Proterozoic basement, Paleozoic (Devonian to Permian) plutons, and Paleozoic-Mesozoic (Silurian to Jurassic) cover (see section 2.4). These terrains are in continuity with the Southwestern Tian Shan interior further west (Fig. 6). A few intramountainous basins filled by Cenozoic sediments, such as the Korla Basin for example, are also included in this part of the range (Fig. 6). There, the steep inherited Paleozoic faults have been mainly reactivated as right-lateral strike-slip faults that delimitate some ridges separating the basins (e.g., XBGMR, 1976; XBGMR, 1985; Yin et al., 1998; Allen et al., 1994; Li et al., 2016a), a deformation style typical of a transpressive system. In this system, the major faults appear to cut through the whole crust since their location at the surface correlates with steps in the Moho at depth (Li et al., 2016a). To the south, this inner range is separated from the outer range by the Korla fault. However, this strike-slip fault zone does not always correspond to the northeastern boundary of the Cenozoic series of the Tarim Basin, which is most often located further south (Fig. 6).

In this part of the Southern Tian Shan, the main stratigraphic change between the Paleozoic range and the Paleozoic Tarim Basin is localized on the Kuruk Tagh fault system (Figs. 7D and 10D) (e.g., XBGMR, 1976, 1985; Charvet et al., 2011). Accordingly, this fault zone is a major inherited structure equivalent to the thrust ramps at the font of the basement imbricates of the Southwestern Tian Shan outer belt. The Kuruk Tagh fault system thus corresponds to the eastward extension of the major deformation palaeo-boundary between the Paleozoic orogenic systems of the Southern Tian Shan to the north and the contemporaneous Tarim Basin to the south.

8.2. Southeastern Tian Shan outer range

At the front of the inner range, the outer range is a few to several 10s of km wide, with typical elevations between ~800 and ~1000 m. To our knowledge, no cross-section across this outer range has ever been published. However, like the Altyn Tagh transpressional system bordering the Tarim Basin to the southeast, the Southeastern Tian Shan strike-slip wedge does not have any fold-and-thrust belt in its outer part. According to the surface and subsurface data, the outer range is composed of 100s-of-km-long basement-involved structures made of the Tarim Proterozoic basement and its Protero-Paleozoic to Cenozoic (Sinian to Quaternary) cover (Figs. 7D and 10D). These structures are related to emergent or blind faults, such as the Lop Nur fault system. Although very steep to nearly vertical, these faults generally have a north block uplifted by a few 10s to 100s of m, which indicates a southward vergence (Figs. 7D, 10D and S2D). At depth, all these structures possibly connect to a deep basal decollement within the basement (e.g., Charvet et al., 2011).

Unconformities, together with thickness and facies changes, were documented from surface observations in the Protero-Paleozoic (Sinian to Devonian) sedimentary cover of the outer range, such as in the Tarim Basin to the south (e.g., Tang et al., 2008; Lin et al., 2012b; Wu et al., 2018; Morin et al., 2019). These observations imply that the Lop Nur fault system and the associated basement-cored uplift, which is known as the Kuruk Tagh uplift, are inherited Protero-Paleozoic structures of the Tarim Basin reactivated during the Cenozoic. In the Southeastern Tian Shan outer range, the whole Cenozoic deformation is associated with these inherited faults, since there is no evidence of younger structures rooting into a Paleozoic or Cenozoic decollement level in the sedimentary series at their front.

The large basement-involved structure of the outer range now forms a smooth topography uplifted by a few 100s of m above the present-day Tarim sedimentation level, and localizes the mountain front (Fig. 6). Located several kilometres further south, the Lop Nur fault system,

which corresponds to the present-day deformation front of the Southeastern Tian Shan range, is hidden by onlapping Plio-Quaternary deposits (Figs. 6, 7D, 10D and S2D). At the front of this fault system, there is no evidence of a significant flexural subsidence, with less than ~1 km of Cenozoic sediments trapped in this part of the basin (Figs. 7D, 10D and S2D) (see also Morin et al., 2019). There, some inherited Protero-Paleozoic structures also exist, but there is little evidence of a significant Cenozoic activity on their bounding faults, so that any eventual Cenozoic deformation in this area would have remained very limited. However, toward the very eastern end of the Tarim Basin, other likely inherited structures may have been slightly reactivated (Fig. 6).

8.3. Deformation amount

Along the transect crossing the Southeastern Tian Shan strike-slip system, we quantified the shortening from the main bordering fault of the northeastern Tarim Cenozoic series (i.e. from the Korla fault in the outer range) to the centre of the basin (see Figs. 6 and 7D for the fixed reference lines set at the turning points in the centre of section D). Accordingly, the shortening values given just below do not account for the deformation accommodated within the inner range. In its outer range only, the Southeastern Tian Shan strike-slip wedge accommodates a very limited amount of Cenozoic compressive deformation, with a possible total shortening of less than several 10s of m. However, in the absence of a preserved Mesozoic-Cenozoic stratigraphic marker, it was not possible to properly quantify the Cenozoic shortening, since eroded material cannot be reconstructed. Nevertheless, this shortening is entirely accommodated by basement-involved structures.

Even if the compressive deformation accumulated within the outer range of the Southeastern Tian Shan strike-slip wedge is poorly constrained, it seems to be limited. Moreover, the structure and low topography of the inner range suggest that the compressive deformation accommodated across this internal part of the system should also be limited. Even though the absence of stratigraphic markers impedes any restoration and quantification of the deformation within the inner range, the non-existence of a flexure associated with the Southeastern Tian Shan supports limited shortening and thickening, typical of such a strike-slip wedge. Instead, a large amount of strike-slip movements has likely been accommodated on major strike-slip faults of the range, such as the Kuruk Tagh fault system or the Korla fault. Strike-slip motion may also have occurred in the basin, along the very steep Lop Nur faults for instance. However, 2D restoration does not allow for quantifying such deformation, which takes place outside of the plane of our section.

9. Discussion

9.1. Neoformed structures and particular case of the Mazar Tagh thrust sheet

In the outer part of the ranges surrounding the Tarim Basin, a significant fraction of the deformation (up to 100%) is accommodated by structures neoformed during the Cenozoic (Table 1). However, these neoformed structures are restricted to the sedimentary cover (thin-skinned deformation) at the front of the Western Kunlun Range and Southwestern Tian Shan, where the amount of compressive deformation is most important (Fig. 11). In contrast, neoformed structures are absent at the front of the Altyn Tagh or Southeastern Tian Shan transpressive systems. In addition, the development of these structures is also strongly correlated with the occurrence of Paleozoic and Cenozoic favorable decollement levels in the sedimentary cover.

Among these neoformed structures, the Mazar Tagh thrust sheet, which is located at the front of the Western Kunlun, has remarkable dimensions (Figs. 6, 7B, 8B and 11). With a north-south width of ~150–200 km and an east-west length of ~300 km, it could even be the

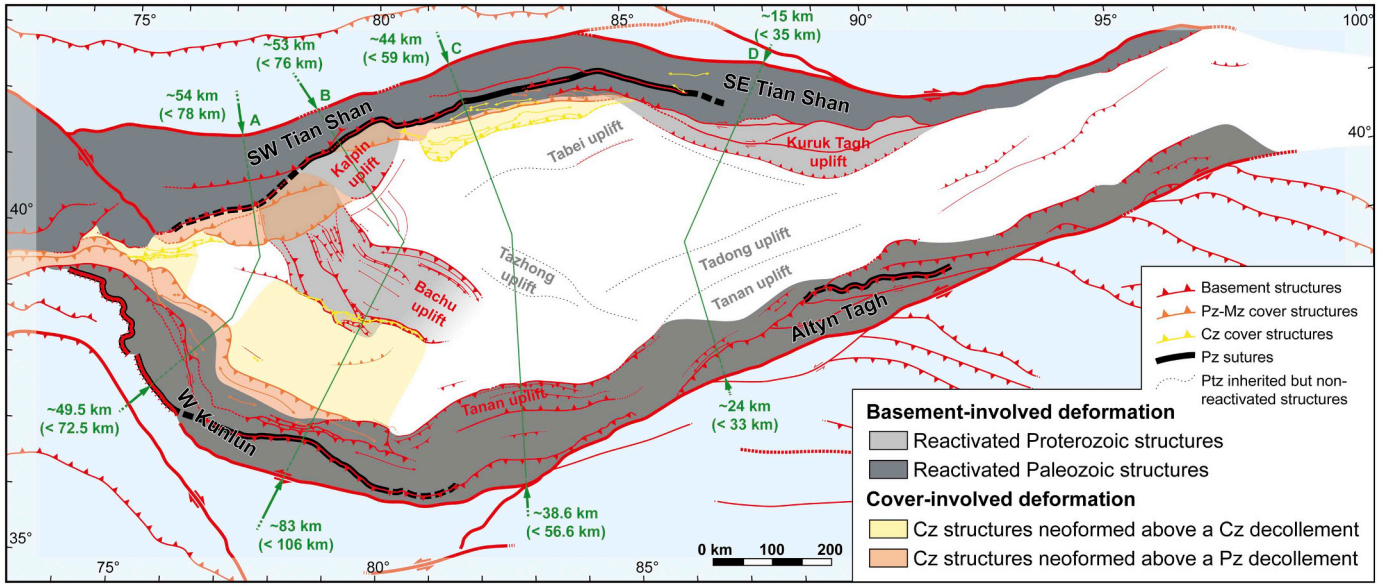

Fig. 11. Tectonic map of the Cenozoic deformation of the Tarim Basin and surrounding ranges. Structures rooted in the basement are represented in red, while structures rooted into a sedimentary decollement are displayed in orange (for structures developed above a Paleozoic decollement) or in yellow (for structures developed above a Cenozoic decollement). Areas comprising Proterozoic and (Protero-)Paleozoic inherited basement structures reactivated during the Cenozoic orogeny are indicated in light and dark grey, respectively. Regions with Cenozoic neoformed cover structures are shown in orange (for structures developed above a Paleozoic decollement) or in yellow (for structures developed above a Cenozoic decollement). Green solid lines locate the regional balanced geological transects A, B, C and D presented in Fig. 7. Green long-dashed lines at terminations of these transects correspond to the extended pathway of the simplified crustal-scale cross-sections presented in Fig. S4, which are used to assess the shortening accommodated across the ranges from crustal budgets (see Fig. 12 and Supplementary information S2 for methodology). Green arrows and values at the ends of these cross-sections represent and scale with the most probable (solid line) and possible maximum (short-dashed line and bracketed numbers) values for the total Cenozoic shortening estimated along these transects (Table 1). Ptz: Proterozoic, Pz: Paleozoic, Mz: Mesozoic, Cz: Cenozoic. In this study, the Tarim block refers to the area comprised between the Paleozoic ophiolitic sutures (thick black lines) located in the surrounding ranges, rather than to the present-day Tarim Basin only. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

largest active thrust sheet in the world without internal deformation (see also Wang et al., 2014c; Suppe et al., 2015; Guilbaud et al., 2017). Consequently, the whole foredeep sedimentary pile (i.e. the sediments of the Southwest depression) is detached at the base of the Cenozoic series and transported northwards. The Mazar Tagh thrust sheet is therefore oversized compared to classical thrust sheets. To the first order, the size of a thrust sheet is controlled by the ratio between its internal resistance (depending on its thickness) and its basal friction (depending on the nature and thickness of its basal decollement level) (e.g., Liu et al., 1992; Mandal et al., 1997; Smit et al., 2003; Hatcher, 2004). Accordingly, the gypsiferous Paleocene-Eocene decollement level should thus be very efficient (i.e. with a low basal friction) to allow for the development of the Mazar Tagh thrust sheet (Suppe et al., 2015). However, this is probably not the only explanation for such an oversized thrust sheet. Indeed, the large thickness (up to ~10 km) of sediments trapped in the Southwest depression may have also drastically increased the rigidity of this thrust sheet, leading to the emergence of its frontal ramps far away from the range topographic front, where flexure, accommodation and sedimentation are reduced (e.g., Mugnier et al., 1997; Fillon et al., 2013).

Finally, an obvious spatial correlation exists between the location of the Mazar Tagh emergent thrust ramps rooting into the Cenozoic decollement and the Mazar Tagh deep faults below it (Fig. 6). In fact, these ramps appear to emerge toward the surface from existing structural steps in this decollement, caused by the hangingwall uplift of the underlying basement-cored structures (Figs. S2B, 7B, 8B and 9). Therefore, the deformation of the Cenozoic decollement by the deep Mazar Tagh thrusts most probably controlled the subsequent localization of the shallow ramps. Potentially, the Mazar Tagh thrust sheet could have been even wider if this basement structures did not exist.

9.2. Structural inheritance

In the ranges surrounding the Tarim Basin, metamorphic and stratigraphic evidences indicate that the basement-cored structures (thick-skinned deformation) are inherited from the orogens that developed around the Tarim block during the Paleozoic (e.g., Windley et al., 1990; Allen et al., 1993, 1994; Matte et al., 1996; Jolivet et al., 2010, 2013; Glorie et al., 2011; Macaulay et al., 2014; this study). Within the inner ranges, these inherited structures absorb all the deformation, while in the outer ranges they accommodate from 0 to 100% of the shortening (Fig. 11; Table 1). Within the basin, at the front of the Altyn Tagh Range, the Tanan uplift also corresponds to an inherited structure from the large Paleozoic orogenic system of the southern margin of the Tarim block. Together with the most frontal basement ramp below the Western Kunlun outer belt, this structure corresponds to a major structural palaeo-boundary between the Paleozoic ranges to the south and the contemporaneous Tarim Basin to the north (Fig. 11). To the north, the structural palaeo-boundary between the Paleozoic Tian Shan and the Tarim Basin also presents evidence of a Cenozoic activity along its whole length. Accordingly, the Paleozoic structures all around the Tarim block have been reactivated during the Cenozoic orogeny (Fig. 11).

In the outer part of the Tian Shan and within the basin, structures inherited from the Protero-Paleozoic history of the Tarim block were also reactivated (e.g., Allen et al., 1999; Carroll et al., 2001; Turner et al., 2010, 2011; Chang et al., 2012b; this study). The Kalpin and Kuruk Tagh uplifts are now part of the Southwestern and Southeastern Tian Shan outer ranges, respectively (Fig. 11). On the other hand, in the western part of the Tarim Basin, the Bachu uplift was rejuvenated in the foreland of the Western Kunlun (Fig. 11).

The Cenozoic reactivation of all those Paleozoic and Protero-Paleozoic structures confirms the major role of structural inheritance in continental deformation, as already widely demonstrated in Asia (e.g., Molnar and Tapponnier, 1981; Windley et al., 1990; Matte et al., 1996; De Grave et al., 2007; Jolivet et al., 2010; Jolivet et al., 2013; Tang et al., 2014). However, all the inherited Protero-Paleozoic structures in the basin have not been reactivated. If the Tanan and Bachu uplifts show evidence of a significant Cenozoic deformation, such is not the case for the Tazhong, Tadong and Tabei uplifts (Figs. 6, 7 and 11). The location and orientation of the Tanan and Bachu structures may have played a decisive role in their reactivation, with a compressive vs. strike-slip main deformation component similar to the one of the local edge of the Tibetan Plateau. Indeed, the Tanan uplift, which is probably reactivated with a prevailing strike-slip component, is oriented parallel to the Altyn Tagh and located just to the front of this range. The Bachu uplift, which is reactivated with a prevailing compressive component, is oriented parallel to the Western Kunlun Range and developed where the forebulge should have been located in the foreland of this orogenic system. In contrast, the Tazhong, Tadong and Tabei uplifts were not reactivated even though their orientations are comparabale to the ones of the surrounding ranges, probably because they are located further from the Tibetan Plateau edge.

In spite of all the reactivated structures in the Tarim basin and surrounding ranges, the experimental and numerical models designed to understand the Cenozoic deformation of Asia usually only focus on the influence of stress boundary conditions and lithospheric rheology, but neglect structural inheritance (e.g. Tapponnier et al., 1982; Vilotte et al., 1984; England and Houseman, 1985; Davy and Cobbold, 1988; Peltzer and Tapponnier, 1988; Burg et al., 1994; Neil and Houseman, 1997; Wang et al., 2007; Dayem et al., 2009; Calignano et al., 2015). Ideally, these various parameters should be all considered in the modelling. In the future, testing the effect of different stress boundary conditions on the reactivation of the inherited basement structures in the area could be a step forward, as a comprehensive regional study is now available to compare with model results.

9.3. Strike-slip faults and potential deformation partitioning

This study mainly deals with the compressive deformation in the Tarim region since 2D restoration does not allow for quantifying strike-slip movements. However, strike-slip deformation may occur in various places in the ranges (e.g., Wang, 1997; Cowgill et al., 2000, 2003; Sobel, 2001; Yue et al., 2001; Meng et al., 2001; Yin et al., 2002; Gehrels et al., 2003; Ding et al., 2004; Cowgill, 2010; Cheng et al., 2015, 2016a; Wang et al., 2016b) and in the basin, such as on the fault systems bounding the Tanan and Bachu uplifts (e.g., Avouac and Peltzer, 1993; Yin et al., 2002; Li et al., 2016e; Yang et al., 2018b; this study). Indeed, the central part of the Tanan uplift corresponds to a symmetric basement pop-up bounded by steep ($> 50°$) thrusts (Figs. 6, 7C and 8C), an overall geometry characteristic of strike-slip structures (e.g., Casas et al., 2001; McClay and Bonora, 2001 and references therein; Woodcock and Rickards, 2003). The Bachu uplift also corresponds to an asymmetric (to the east) or symmetric (to the west) basement pop-up with steep ($> 60°$) bounding thrusts, in particular in the case of the Seyibula or Mazar Tagh deep fault systems (Figs. 6, 7B and 9). Vertical faults may even be observed in these systems (Li et al., 2016e). The existence of these potential strike-slip structures may indicate deformation partitioning through the Bachu uplift, with a prevailing compressive component to the north (on the Aqia-Tumuxiuke thrust systems) and a prevailing strike-slip one to the south (on the Seyibula-Mazar Tagh deep fault systems) (Fig. 6). As for the structure orientation, this configuration is similar to the one of the Western Kunlun Range where deformation is partitioned between compressive structures to the front, and the Kashgar-Yecheng transfer system and the Karakax strike-slip fault to the rear.

Additionally, a few studies proposed a prolongation of the Talas Fergana strike-slip fault (Shache-Yangdaman fault) inside the Tarim Basin, in front of the Pamir Range, southeastward down to the city of Yecheng (Wei et al., 2013; Blayney et al., 2016, 2019; Bande et al., 2017b). However, we find no evidence of deformation associated with such a fault prolongation, neither at the surface, nor on the seismic data in spite of some available seismic profiles crossing the supposed fault location (Figs. 1, S2A, 7A and 8A). Indeed, the Shache basement fault below the Jiede anticline nearby Yecheng has a limited lateral extension (Fig. 6), since it is not observed on the seismic lines available laterally. Consequently, this structure cannot be a prolongation of the Talas Fergana fault. A horsetail splay termination of this fault into the Southwestern Tian Shan thrusts, as suggested by Jia et al. (2015), seems to be a more reasonable interpretation, with regards to our dataset in the basin.

Finally, the sharp eastern edges of the Western Kunlun Range and Bachu uplift in map view could suggest the occurrence of a major NNE-SSW strike-slip fault linking both terminations. However, we did not observe any evidence for such a fault in the numerous seismic lines crossing the area (Fig. 1). The eastern extremities of the Western Kunlun Range and Bachu uplift rather correspond to a progressive, but locally abrupt, eastward decline of their structures (Fig. 6).

9.4. Shortening

In the basin and the outer part of the surrounding ranges, we quantified the Cenozoic shortening from structural restorations (Fig. 7; Table 1). However, the associated estimates correspond to the deformation accommodated from the margins of the basin to its centre, and therefore consist in minimum values with respect to the total

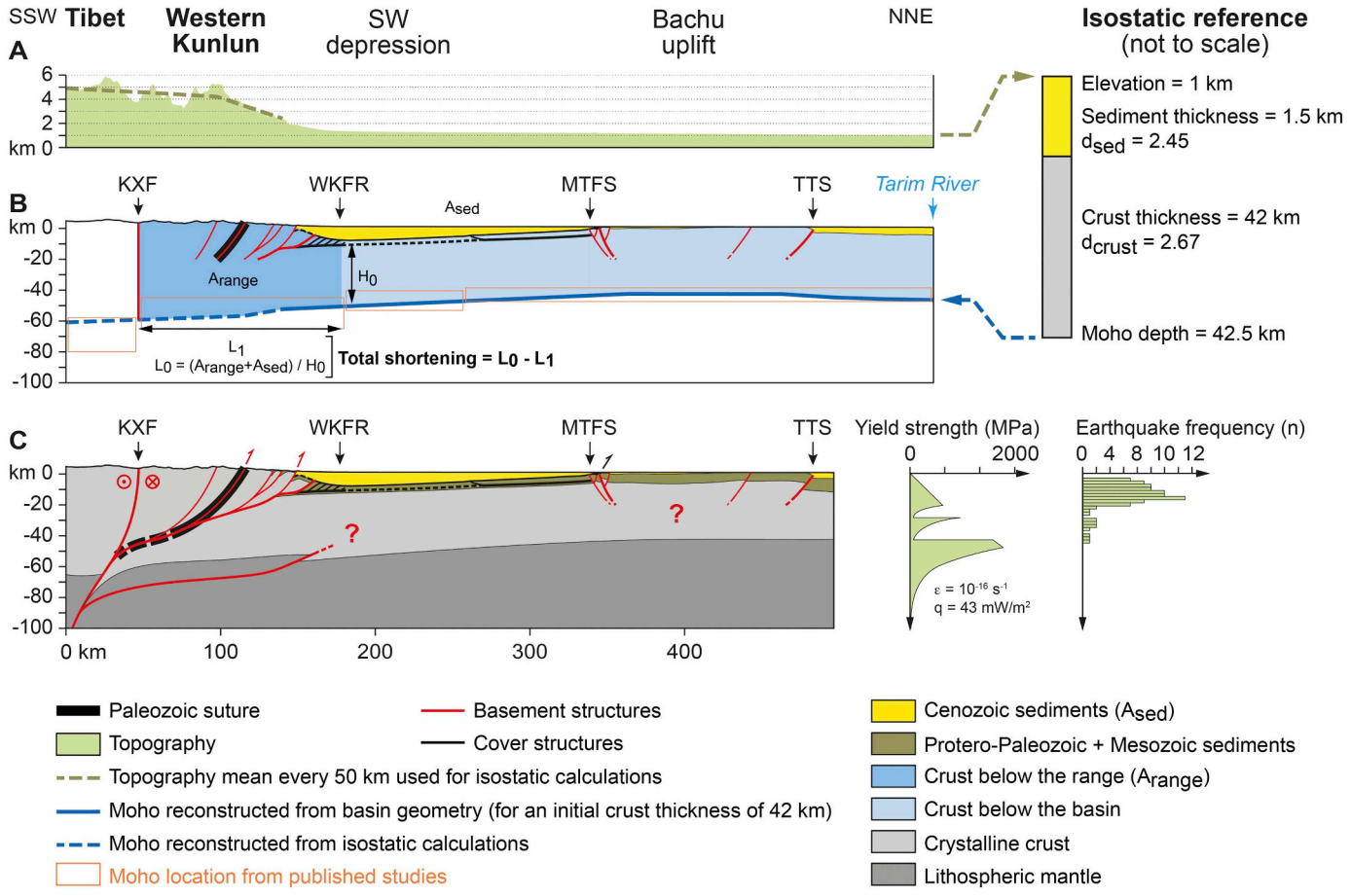

Fig. 12. Components for the crustal-scale cross-sections presented in Figs. 13 and S4 (see Fig. 6 for abbreviation labels). (A) Example of an elevation profile extracted from the SRTM3 DEM and averaged every 50 km in the range (green dashed line). This average topography corresponds to the main input for assessing the Moho depth below the ranges by isostatic calculations. (B) Example of a simplified crustal-scale cross-section used to estimate the shortening accommodated across the ranges surrounding the Tarim Basin from crustal budgets (see Supplementary information S2 for methodology). To assess the Moho depth below the ranges by isostatic calculations, a reference in a stable place (i.e. with no Cenozoic deformation or flexure) is required. This stable isostatic reference was taken in the centre of the basin, between kilometres 300 and 350 along transect C, where there is no structures activated during the Cenozoic (Figs. 7C and S4C1). A_{range}: crustal area below the range, A_{sed}: area of the Cenozoic sediments coming from the range erosion (i.e. crustal area of the range lost by erosion), H_0: initial crustal thickness at the beginning of Cenozoic, L_1: present-day length of the range, L_0: initial crustal length at the beginning of Cenozoic, d: density. (C) Intermediate state of construction for the synthetic crustal-scale cross-sections. Structures from the surface down to 20 km deep are constrained by the regional balanced geological transects (Fig. 7B). Structures below 20 km deep are inferred from the isostatic calculations and the basin geometry achieved in this study, and from the deep geophysical data available in the area (e.g., Li et al., 2002b; Jiang et al., 2004; Wittlinger et al., 2004). Examples of a lithospheric yield strength profile (from Zhang et al., 2013b) and of earthquake frequency according to depth (from Sloan et al., 2011) below the Tarim Basin. Ɛ: deformation rate, q: heat flux, n: number of earthquakes. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 1
Shortening estimates across the Tarim Basin and surrounding ranges.

Range			Shortening (km)				
			1	2	3	4	5
			Entire range (crustal budg.)	Inner range (= 1 - 3)	Outer range (restor.)	Basin (restor.)	Total (= 1 + 4)
N Tibet	W Kunlun	*Transect A (A1)*	49 +23/-17.5	~17.5 +23/-17.5	~31.5 (inh. 9.5 + neo. 22)	~0.5	~49.5 +23/-17.5
		Transect B (B1)	78 ± 23	~48 ± 23	~30 (inh. 16.5 + neo. 13.5)	~5	~83 ± 23
	Altyn Tagh	*Transect C (C1)*	38 ± 18	~37.7 ± 18	≥ ~0.3 (inh. 0.3 + neo. 0)	≥ ~0.6	~38.6 ± 18
		Transect D (D1)	24 ± 9	~23.7 ± 9	≥ ~0.3 (inh. 0.3 + neo. 0)	0	24 ± 9
S Tian Shan	SW Tian Shan	*Transect A (A2)*	54 +24/-18	~18 +24/-18	≥ ~36 (inh. 0 + neo. 36)	0	54 +24/-18
		Transect B (B2)	53 ± 23	/	≥ ~4 (inh. 4 + neo. 0)	0	53 ± 23
		Transect C (C2)	44 ± 15	~22 ± 15	~22 (inh. 19 + neo. 3)	0	44 ± 15
	SE Tian Shan	*Transect D (D2)*	15 +20/-15	/	> ~0	0	15 +20/-15

Shortening values are provided for (1) the entire (inner plus outer) ranges, (2) the inner ranges, (3) the outer ranges, (4) the basin, and (5) the entire orogenic systems. Shortening values for the entire ranges (1) are derived from crustal budgets (Figs. 12 and S4; Supplementary material S2). Those for the outer ranges (3) and for the basin (4) are derived from structural restorations (Fig. 7). For the outer ranges, the contributions to the shortening of the (Protero-)Paleozoic inherited (inh.) and Cenozoic neoformed (neo.) structures are indicated in brackets. Shortening values for the inner ranges (2) correspond to the differences between the shortenings across the entire ranges (1) and those accumulated in their outer parts (3). Finally, the total shortening amounts accommodated across the whole orogenic systems (5) are obtained by adding the shortening values accommodated across the entire ranges (1) and in the basin (4). Values given for shortenings derived from crustal budgets correspond to the most probable values according to our regional balanced geological transects (Cenozoic sediment thicknesses) and to other published parameters (densities and sediment porosity) used for calculations. Uncertainties given for these values correspond to possible errors on the crustal areas below the range (coming from the densities) and on the crustal areas lost by erosion (coming from the sediment porosity) (see Supplementary information S2 for more explanations).

shortening accumulated in the ranges. Indeed, our restorations do not account for the shortening absorbed across the inner ranges. In order to assess the total deformation across the ranges, we used a crustal budget approach (for the methodology see Fig. 12A and 12B, as well as the Supplementary information S2 and references therein). This approach is based on the conservation of the crustal matter during deformation. Practically, it corresponds to an area balancing of the crust of the ranges between the present-day and initial states of deformation (e.g., Ménard et al., 1991; Avouac et al., 1993; Teixell, 1998). Eventually, the compressive deformation across the inner ranges can be inferred from the difference between the shortening estimates across the entire and the outer ranges (Table 1).

In the Western Kunlun compressive system, we estimate a total crustal shortening of ~49.5 +23/−17.5 km to the west (transect A) and ~83 ± 23 km to the east (transect B) (Figs. 11, S4A1 and S4B1; Table 1). This compressive deformation is essentially accommodated within the entire (inner and outer) range when compared to the basin, with ~99% (49 +23/−17.5 km) and ~94% (78 ± 23 km) of this total shortening concentrated across the range for transects A and B, respectively (Table 1). These estimates are within the wide range of values, from ~42 to ~272 km, already suggested for this orogen (Lyon-Caen and Molnar, 1984; Rumelhart et al., 1999; Cowgill et al., 2003). However, the highest of these suggested shortening values are based on the probably oversimplified assumption that the Pamir-Western Kunlun Range was initially straight and subsequently acquired its present-day bend during the Cenozoic orogeny (see Chen et al., 2018 and references therein). In the inner range, ~35.4% (17.5 +23/−17.5 km) and ~57.8% (48 ± 23 km) of the total shortening is absorbed to the west (transect A) and to the east (transect B), respectively (Table 1). In addition, ~63.6% (~31.5 km) and ~36.2% (~30 km) of the total shortening are accumulated within the outer range to the west (transects A) and to the east (transect B) (Fig. 7A and 7B; Table 1). The different shortening amounts absorbed across the inner range, when the shortening amount within the outer range is similar, could thus reflect varying proportions of strike-slip vs. compressive deformation in the interior of the mountain belt, with more strike slip deformation in its NNW-SSE part than in its WNW-ESE one. In the Tarim Basin, a significant amount of deformation is also accommodated by the Bachu uplift that absorbs ~6% (~5 km) of the total compressive deformation in the Western Kunlun system along transect B (Fig. 7B; Table 1).

Across the Altyn Tagh transpressive system, we estimate a total crustal shortening of ~38.6 ± 18 km to the west (transect C) and 24 ± 9 km to the east (transect D) (Figs. 11, S4C1 and S4D1; Table 1). Here again, this compressive deformation is essentially accommodated within the entire (inner and outer) range when compared to the basin, with ~98.5% (38 ± 18.0 km) and 100% (24 ± 9 km) of this total shortening accommodated in the range for transects C and D, respectively (Table 1). The main part of this compressive deformation is absorbed within the inner range, with ~97.7% (37.7 ± 18 km) and ~98.8% (23.7 ± 9 km) concentrated across transects C and D, respectively (Table 1). The outer range only accumulates ~0.8 to ~1.2% (~0.3 km) of the total shortening (Fig. 7C and 7D; Table 1). To the west, a limited amount of deformation is also accommodated within the basin where the Tanan uplift accommodates ~1.5% (~0.6 km) of the total compressive deformation across the Altyn Tagh system (Fig. 7C; Table 1). To our knowledge, these values are the first shortening estimates provided for this strike-slip wedge.

Across the Southwestern Tian Shan compressive system, we estimate a total crustal shortening of 54 +24/−18 km to the west (transect A), 53 ± 23 km in the centre (transect B), and 44 ± 15 km to the east (transect C) (Figs. 11, S4A2, S4B2 and S4C2; Table 1). This compressive deformation is totally accommodated within the range when compared to the basin (Table 1). Many other shortening values, ranging from ~80 to ~280 km, have been proposed for the Tian Shan (e.g., Molnar and Tapponnier, 1975; Chen et al., 1991; Avouac et al., 1993; Allen et al., 1994). However, these values correspond to estimates of shortening across the whole range (Northern, Central and Southern Tian Shan), and cannot be therefore easily compared with our results. In the inner range, ~33.3% (18 +24/−18 km) and ~50% (22 ± 15 km) of the total shortening are accumulated to the west (transect A) and to east (transect C), respectively (Table 1). In the outer range, at least ~66.7% (~36 km) and ~50% (~22 km) of the compressive deformation is

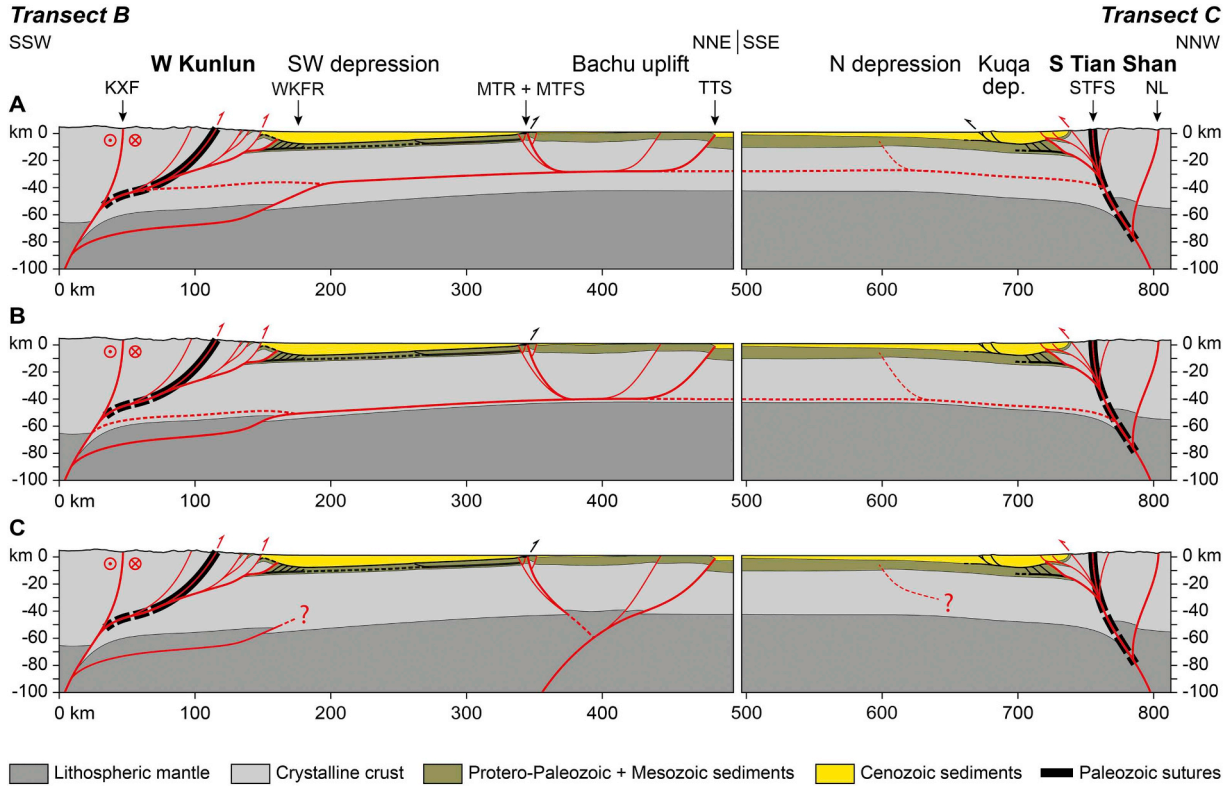

Fig. 13. Possible synthetic crustal-scale cross-sections across the Tarim Basin, Western Kunlun Range and Southern Tian Shan (see Fig. 6 for location and abbreviation labels). (A) and (B) Cross-sections with the basement-cored uplifts in the basin rooted into a crustal decollement (in the middle or at the base of the crust, respectively). These interpretations are favoured due to their consistency with the geological and geophysical data available so far in the area (see text for more explanations). (C) Alternative cross-section with the large basement-cored uplifts in the basin rooted into the mantle. In this study, the Tarim block refers to the area comprised between the Paleozoic ophiolitic sutures (thick black lines) located in the surrounding ranges, rather than to the present-day Tarim Basin only.

absorbed across the Kepingtage and Kuqa belts, respectively, to the west (transect A) and to the east (transect C) of the Southwestern Tian Shan (Fig. 7A and 7C; Table 1).

In the Southeastern Tian Shan transpressive system, we estimate a total crustal shortening of 15 +20/−15 km (transect D) that may be entirely accommodated in the inner range (Figs. 11, S4D2; Table 1). Indeed, in the outer range, the compressive deformation seems to be negligible, with less than several 10s of m of shortening (Fig. 7D; Table 1). To our knowledge, these values are the first estimates of shortening provided for the Southeastern Tian Shan strike-slip system.

In this study, we do not have enough constraints to quantify the shortening on each side of the Paleozoic sutures in the ranges surrounding the Tarim Basin, and thus to isolate the exact contribution of compressive deformation affecting the original Proterozoic Tarim block. However, based on our shortening assessments, we find that several 10s of km of Cenozoic compressive deformation have been accommodated on the margins of the Tarim block. Nevertheless, this deformation is lower in the eastern strike-slip systems (Altyn Tagh and Southeastern Tian Shan) than in the western compressive ones (Western Kunlun and Southwestern Tian Shan). Accordingly, regarding this geological evidence and the quantified shortening, a continental crust dipping down to ~80 km below the Western Kunlun Range (e.g., Lyon-Caen and Molnar, 1984; Wittlinger et al., 2004; this study) is plausible. In contrast, a continental subduction implying a crust dipping down to ~100 km below the Altyn Tagh Range (Wittlinger et al., 1998) seems unlikely given the limited number of compressive structures and the associated very low shortening.

Finally, in addition to the Cenozoic deformation on the Tarim block margins, some basement-involved deformation also occurs within the western and southern parts of this block. Indeed, in spite of an overall flat topography due to an important sediment filling, a non-negligible amount of compressive deformation (up to a few % of the total shortening) is also accommodated within the Tarim Basin across the Bachu and Tanan uplifts. Accordingly, the Tarim block, including its central part occupied by the present-day Tarim Basin, does not correspond to a totally rigid and undeformable heterogeneity within the Asian compressive system.

9.5. *Crustal structure and implications for the deformation of Central Asia*

In order to propose a crustal-scale cross-section across the Tarim Basin and the surrounding mountain ranges, we combined our geological transects with Moho depths inferred from the deep geophysical data (gravimetry, tomography, seismology, rheology, etc.) available in the area, the basin geometry and isostasic calculations (Figs. 12 and S4; Supplementary information S2 and references therein). Below the ranges (e.g., below the Western Kunlun and Southern Tian Shan), we connected the major faults (e.g., the Karakax fault and Southern Tian Shan fault system) with well-documented offsets of the Moho (e.g., Jiang et al., 2004; Wittlinger et al., 2004; Li et al., 2007). However, in the centre of the basin, the deep geophysical data do not precisely constrain the Moho geometry below the Bachu and Tanan basement-

cored uplifts, raising the question of where they root at depth (Fig. 12C). The involvement of the basement in these structures and their large wavelength (respectively ~150 km and ~50 km of width) suggest that they are large-scale structures that may be deeply rooted into the middle crust, the base of the crust or into the mantle (Fig. 13).

Such large foreland basement-cored uplifts exist in other orogenic systems (e.g., Himalayas, Laramides, Andes, Alps, etc.), where their rooting depth has also been discussed and investigated (e.g., Lacombe and Mouthereau, 2002; Yin et al., 2010; Yeck et al., 2014; Yonkee and Weil, 2015; Lacombe and Bellahsen, 2016). Usually, these structures are interpreted as developing above a mid-crustal decollement level (e.g., Lacombe and Mouthereau, 2002; Yeck et al., 2014; Yonkee and Weil, 2015; Lacombe and Bellahsen, 2016). Based on these other case examples and on the rheology of the local continental lithosphere (Fig. 12C), we thus propose that the Bachu and Tanan uplifts also root into a decollement level within the crust (Fig. 13A and 13B). Indeed, some geophysical interpretations (yield strength profiles and earthquake frequency at depth) suggest the existence of two possible decollement levels below the Tarim Basin, in the middle and at the base of the crust (Fig. 12C) (Sloan et al., 2011; Zhang et al., 2013b; Liu et al., 2015b; Chang et al., 2016b; Deng and Tesauro, 2016). The Bachu and Tanan uplifts may therefore deform above one of these crustal decollements (Fig. 13A and 13B). The main northward structural vergency of these basement highs would then indicate that the activated decollement transfers deformation from the northern edge of the Tibetan Plateau to the centre of the basin.

However, in the Himalayan system, the Shillong Plateau is a foreland basement-cored uplift that may root into the mantle, as suggested by underlying possible Moho offsets (Yin et al., 2010). To investigate this possibility for the Bachu and Tanan basement uplifts, new deep geophysical data are necessary to specifically constrain the local Moho geometry. Indeed, accurate geophysical images of these areas are still lacking (e.g., Fig. 3 of Wittlinger et al., 2004). Nevertheless, rooting these structures directly into the mantle raises the problem of the foreland-ward transfer of the deformation associated with the Moho offset imaged below the outer range of the Western Kunlun (Wittlinger et al., 2004) (Fig. 13C). Consequently, we favor our previous interpretations of a crustal decollement level below the Tarim.

These interpretations imply a decoupling between the deforming crust and an underlying rigid mantle. This is in agreement with the tomographic studies proposing a cold and rigid mantle below the basin (Deng and Tesauro, 2016; Deng et al., 2017). Moreover, the deformation pattern of the Tarim block can be compared with results of experimental models studying the influence of the depth of a rigid lithospheric heterogeneity on continental deformation (Calignano et al., 2015). Indeed, the deformation localization around the Tarim Basin and the associated flexures are in agreement with models simulating a strong upper mantle (Figs. 5, 8c and 9b of Calignano et al., 2015). Therefore, the geological, geophysical and experimental observations all suggest that the rigidity of the Tarim block is supported by the mantle, while deformation is transmitted above a crustal decollement below the Tarim Basin, from the Tibetan Plateau to the Tian Shan and further north.

10. Conclusion

On the basis of a structural map and of four balanced geological transects, we characterized the Cenozoic deformation across the Tarim Basin and the surrounding ranges. From structural restorations and crustal budgets, we also quantified the compressive component of this deformation. This comprehensive study thus provides a consistent regional overview of how the Tarim block participates to the Cenozoic deformation of Asia.

To the west, up to 83 ± 23 and 54 +24/−18 km of total crustal shortening is accommodated across the Western Kunlun and Southwestern Tian Shan compressive systems, respectively. On the other hand, to the east, only up to 38 ± 18 and 15 +20/−15 km of total crustal shortening is accommodated across the Altyn Tagh and Southeastern Tian Shan transpressive systems, respectively. The crustal shortening is therefore higher in the thrust wedges to the west, than in the strike-slip ones to the east. Accordingly, continental subduction is plausible below the Western Kunlun Range, whereas it seems unlikely below the Altyn Tagh.

Most of this Cenozoic compressive deformation of the Tarim block (from ~94% to 100%) is concentrated within the surrounding ranges. Indeed, a shortening of up to 78 ± 23 km (with ~30 km in the outer range only) is accommodated in the Western Kunlun Range, 54 +24/−18 km (~36 in the outer range) in the Southwestern Tian Shan, 38 ± 18 km (~0.3 in the outer range) in the Altyn Tagh Range, and 15 +20/−15 km (~0 km in the outer range) in the Southeastern Tian Shan. In spite of an overall flat topography due to an important sediment filling, a non-negligible amount of compressive deformation is also accommodated within the Tarim Basin by large basement-cored uplifts with a vergence synthetic to the deformation of the Tibetan Plateau edge. To the west, the Bachu uplift absorbs ~6% (~5 km) of the total crustal shortening of the Western Kunlun compressive system. To the south, the Tanan uplift accumulates ~1.5% (~0.6 km) of the total crustal shortening of the Altyn Tagh transpressive system. Strike-slip deformation may have also occurred on the bordering faults of these uplifts, but these displacements still need to be quantified. Consequently, the Tarim block, including its central part occupied by the present-day Tarim Basin, does not correspond to a totally rigid and undeformable heterogeneity within the Asian compressive system.

In the ranges around the Tarim Basin, thin-skinned structures neoformed during the Cenozoic orogeny accommodate a significant fraction of the compressive deformation (up to ~66.7%). However, the main part of the deformation (from ~33.3% to 100%) is rather accommodated by reactivated thick-skinned structures inherited from the Protero-Paleozoic history of the Tarim block. The structural inheritance thus appears to be at least as important as the stress boundary conditions or the lithospheric rheology in controlling the Cenozoic deformation of the Tarim area.

Finally, in the centre of the basin, the Bachu and Tanan uplifts probably root into a deep crustal decollement level. This decollement implies decoupling between the deforming crust and an underlying rigid mantle, in agreement with the geological and geophysical data available in the region, and in line with experimental observations. Therefore, the rigidity of the Tarim block seems to be supported by the mantle, while deformation is transmitted above a decollement in the crust from the Tibetan Plateau to the Tian Shan and further north.

Supplementary data to this article can be found online at https://doi.org/10.1016/j.earscirev.2019.102891.

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