Propagation of the deformation and growth of the Tibetan–Himalayan orogen: A review

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Abstract

Long-standing problems in the geological evolution of the Tibetan–Himalayan orogen include where the India–Asia convergence was accommodated and how the plateau grew. To clarify these problems, we review the deformations and their role in the plateau's growth. Our results show that ~1630 km of shortening occurred across the Tibetan–Himalayan orogen since ~55 Ma, with more than ~1400 km accommodated by large-scale thrust belts. These thrust belts display an outward expansion from central Tibet and couple with the surficial uplift. The development of the Tibetan plateau involved three significant steps: Primitive plateau (~90–55 Ma), Proto-plateau (~55–40 Ma), and Neoteric plateau (~40–0 Ma). Several processes have collaborated to produce the Proto-plateau, including the pre-existing Primitive plateau, the India–Asia collision, and subductions of Greater India and Songpan–Ganzi beneath the Lhasa–Qiangtang terrane. Since ~40 Ma, the Proto-plateau, which was dominated by a topographic gradient, lower crustal flow and continuous India–Asia convergence, experienced three periods of rapid outward growth (~40–23, ~23–10, and ~10–0 Ma) in general. The N–S trending rifts were caused by the eastward growth of the plateau dominated by thrusting and crust flow in central Tibet, while they were the results of intense N–S shortening in Himalaya.

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http://dx.doi.org/10.1016/j.earscirev.2015.01.001
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1. Introduction

The Tibetan–Himalayan orogen, the largest mountain chain on Earth today, was considered to be the result of the continued convergence between the Indian and Eurasian continents following their initial collision approximately 50–65 Myr ago (Fig. 1) (Molnar and Tapponnier, 1975; Harrison et al., 1992; Ratschbacher et al., 1994; Yin and Harrison, 2000; DeCelles et al., 2002; Li et al., 2012; Meng et al., 2012). The deformation and its relationship to the rise of the plateau are an essential problem related to the geologic evolution of the Tibetan plateau and to continental tectonics more generally (Ratschbacher et al., 1994; Coleman and Hodges, 1995; Tapponnier et al., 2001; Wang et al., 2008a). Although the question of when and how the plateau attained its current elevation is still in dispute, it has been accepted that the deformation processes preserve the most dramatic records of the long-term evolution of the plateau (England and Houseman, 1989; DeCelles et al., 2002). Conversely, the uplifting process of the plateau would provide a first-order constraint on the dynamics of the crustal deformation (England and Molnar, 1990). Therefore, the origin of the high-elevation plateau and its relationship to crustal deformation have been studied intensively since the mid-1970s (Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1977; Allegre et al., 1984; Dewey et al., 1988; DeCelles et al., 2002; Zhu et al., 2006; Wang et al., 2008a; Yin et al., 2008; Li et al., 2012). However, how the fold–thrust belts and crustal shortening contributed to the uplift of the plateau in space and time is ambiguous. Tapponnier et al. (2001) proposed that the rise of the Tibetan plateau has undergone three main steps from central Tibet since the Eocene, the subduction of mantle lithosphere accounting for the dominant growth of the Tibet plateau toward the east and northeast. However, how the crustal shortening and exhumation history responded to this growth and whether the Himalaya experienced the analogical process is still uncertain. Moreover, the crustal shortening, uplifting history and their dynamics cannot be effectively reconciled by a growth model in eastern Tibet (western Sichuan) (Clark and Royden, 2000; Tapponnier et al., 2001; Hubbard and Shaw, 2009; Oskin, 2012; Wang et al., 2012). Although Tapponnier et al. (2001) and Wang et al. (2008a, 2014a) confirmed the growth of the plateau from central Tibet, the uplifting process and dynamics remain questionable (Murphy et al., 1997; Zhang, 2000; Kapp et al., 2003, 2005; Zhang et al., 2004; Volkmer et al., 2007; Guillot and Replumaz, 2013). This means the deformation history and mechanism of the Tibet plateau’s northward growth should be reappraised.

The E–W extension and related N–S trending rift are another notable deformation pattern of the plateau, which is widely distributed but of very low magnitude in the Himalaya and central Tibet. Some researchers
attributed the extension to the rapid uplift and gravitational collapse of the plateau (Molnar and Tapponnier, 1978; Dewey et al., 1988; England and Housenman, 1989). Other workers argued that the E–W extension in southern Tibet reflects local boundary conditions, such as oblique convergence between India and Asia (McCaffrey, 1996; McCaffrey and Nabelek, 1998), radial expansion and stretching along the Himalayan arc (Seeber and Armbruster, 1984; Armiyo et al., 1986), and concentrated compression at the Central Himalayan front (Kapp and Guynn, 2004). Therefore, a better understanding of the distribution and the correlation with contractional deformation may be the key factors to constrain the dynamics of the plateau’s extension and uplift.

The above discussions indicate that the poor constraints on the uplifting history and dynamics of the plateau can be attributed to our incomplete understanding of its deformation history and mechanism. An accurate reconstruction of the deformation history and its development in space and time is essential to assess the dynamics of the growth plateau. In this paper, we attempt to summarize the available deformational and thermochronological data in the Himalaya and the Tibetan plateau to provide new insight into the deformatioin process and the resulting uplift of the plateau. Based on our interpretation, we propose an integrated model for the growth of the Tibetan plateau and its dynamics.

2. Geological setting and models for the plateau uplift

With an area over 2.5 million km² and an average elevation of approximately 5000 m, the Tibetan–Himalayan orogen is characterized by three different geomorphic features (Fig. 1). The first is the flat plateau with an average elevation of approximately 5000 m at the central area, the majority of which has not been incised by external rivers. The second is characterized by a series of mountain chains at the margins of the plateau, with an average elevation of approximately 5500 to 6500 m, such as at the Himalaya and the Longmen Shan (Fig. 1). The third is the sedimentary basins located in and around the plateau, such as the Hoh Xil and the Lopnula Basins at the center, the Tarim and Hoxi Corridor Basins to the north, the Siwalik Basin to the south, and the Sichuan Basin to the east.

Geologically, the plateau consists of several terranes, including Qilian, Kunlun–Qaidam, Songpan–Ganzi, Qiangtang, Lhasa, and Himalaya, separated by the south Qilian, Kunlun, Jinsha, Bangong–Nujiang, and Indus–Yarlung Zangpo suture zones from north to south (Fig. 1). The E–W trending Indus–Yarlung Zangpo suture zone contains Late Jurassic to Early Cretaceous ophiolites and represents a remnant of the Neo-Tethys Ocean (Allgre et al., 1984; Dewey et al., 1988; Yin and Harrison, 2000; Hebert et al., 2012). Different opinions have been offered for the age of the India–Asia collision, varying from ~65 Ma to ~35 Ma (Aitchison et al., 2011; Ding et al., 2005; van Hinsbergen et al., 2011a, b, 2012). Although the arguments for the collision’s age are still in dispute, we prefer an age of ~55–50 Ma in our reconstruction for the following reasons: 1) the cessation of marine facies and the first appearance of arc detritus in the Himalayan foreland basin at ~55–50 Ma (Green et al., 2008; Najman et al., 2010; Wang et al., 2011); 2) the plate motion of India decreased dramatically at ~55–50 Ma (Copley et al., 2010; Klooitwijk et al., 1992; Lippert et al., 2011; Meng et al., 2012; Molnar and Stock, 2009; Patriat and Achache, 1984; van Hinsbergen et al., 2011a, b, 2012); 3) the ~55–50 Ma ultrahigh-pressure metamorphism (de Sigoyer et al., 2000; Schlup et al., 2003; Leech et al., 2005; Guillot et al., 2008); and 4) the climax of collisional magmatism that started at approximately ~55–50 Ma (Mo et al., 2008; Ji et al., 2009; Guan et al., 2012). To accurately reconstruct the convergence between India and Asia, we took ~55 Ma as the onset of collision. This implies that the total amount of convergence between India and Asia is ~3200–4000 km based on the plate reconstructions by van Hinsbergen et al. (2011a, b).

Numerous models have been proposed to explain the crustal deformation and surface uplift of the plateau. Four general categories of models have been proposed: 1) continuous deformation and crustal thickening during the Cenozoic (England and Houseman, 1989; Molnar et al., 1993); 2) underthrusting or injection of India (Argand, 1924; Powell and Conaghan, 1973; Chemenda et al., 2000); 3) intracratonic subduction of Asian lithosphere under Tibet (Willett and Beaumont, 1994; Meyer et al., 1998; Roger et al., 2000; Tapponnier et al., 2001; Guillot and Replumaz, 2013); and 4) uplift owing to dynamic processes in the lower crust and mantle (Dewey et al., 1988; Molnar et al., 1993; Clark and Rondten, 2000). Because the above models have difficulties accounting for all the known features of the plateau, some studies have proposed combinations of these mechanisms for the crustal thickening and uplift of the plateau. Up to now, the surface uplift process and pattern remain controversial: 1) the onset uplift of the plateau ranges from Late Cretaceous to Late Miocene (Harrison et al., 1992; Murphy et al., 1997; Chung et al., 1998; Rowley and Currie, 2006; Volkmer et al., 2007; Wang et al., 2008a; Rohrmann et al., 2012); and 2) the uplift pattern varies from entire uplift to progressive growth (Chung et al., 1998; Harrison et al., 1998; Ruddiman, 1998; Tapponnier et al., 2001; Wang et al., 2008a). Moreover, the uplift of the plateau has long been attributed to the India–Asia collision (Argand, 1924; Dewey et al., 1989); however, the deformation indicated the Lhasa and Qiangtang terranes were significantly shortened through folding and thrusting prior to the India–Asia collision (Kapp et al., 2003, 2005; Murphy et al., 1997; van Hinsbergen et al., 2011b, and their references therein; Volkmer et al., 2007; Zhang et al., 2012). This means the uplift of the plateau predated 55 Ma. Therefore, the temporal and spatial variability of the deformation will be significant for us to understand the growth mechanisms of the plateau.

3. Distribution of the fold–thrust belts in the Tibetan–Himalayan orogen

The India–Asia collision and associated intracratonic deformation generated numerous large-scale fold–thrust belts within the plateau and surrounding regions (Molnar and Tapponnier, 1975; Allegre et al., 1984; Dewey et al., 1988; Schelling and Arita, 1991; Ratschbacher et al., 1994; Murphy et al., 1997; Yin and Harrison, 2000; DeCelles et al., 2002; Kirby et al., 2002; Kapp et al., 2003, 2005; Wang et al., 2008a). Based on the spatial distribution, we classify these thrust belts into four thrust systems, namely, the Yarlung Zangpo–Himalaya thrust system, the Central Tibet thrust system, the Northern Tibet thrust system, and the Eastern Tibetan thrust system. The major fold–thrust belts and their ages are shown in Fig. 2 and Table 1.

3.1. Yarlung Zangpo–Himalaya thrust system

The Yarlung Zangpo–Himalaya region lies between the Indian shield to the south and the Gangdese forearc basin to the north (Figs. 1 and 2), which records the deformation histories of the Indus–Yarlung Zangpo suture and the Himalayan tectonic system (Yin, 2006). According to the tectonic location and composition, this system can be divided into the Yarlung Zangpo and Himalayan secondary systems.

The Yarlung Zangpo secondary thrust system is distributed along the Indus–Yarlung Zangpo suture and consists of two major thrust belts: the Great Counter thrust belt (GCT) to the south and the Gangdese thrust belt (GT) to the north. The south-dipping Great Counter thrust belt juxtaposed the Xigaze forearc basin's strata over the Yarlung–Zangpo ophiolites and is crosscut by the younger Great Counter thrust belt in southeastern and southwestern Tibet (Yin et al.,...
The Himalayan secondary thrust system consists of four north-dipping thrust belts: the Tethyan Himalayan fold thrust belt (THFT), the Main Boundary thrust belt (MBT), the Main Central thrust belt (MCT), and the Main Frontal thrust belt (MFT) from north to south (Fig. 2). The Tethyan Himalayan fold thrust belt is located between the Indus–Yarlung Zangpo suture and the South Tibet Detachment System (STDS). It consists of a series of folds and thrusts involving the passive continental margin sequence of India (Garzanti, 1999) and has an estimated 100–120 km since the Oligocene.

The Himalayan secondary thrust system consists of four north-dipping thrust belts: the Tethyan Himalayan fold thrust belt (THFT), the Main Boundary thrust belt (MBT), the Main Central thrust belt (MCT), and the Main Frontal thrust belt (MFT) from north to south (Fig. 2). The Tethyan Himalayan fold thrust belt is located between the Indus–Yarlung Zangpo suture and the South Tibet Detachment System (STDS). It consists of a series of folds and thrusts involving the passive continental margin sequence of India (Garzanti, 1999) and has an estimated 110–140 km of shortening (Table 2) (Ratschbacher et al., 1994; Murphy and Yin, 2003). This thrust belt initiated in the Eocene (~50 Ma) and ceased motion until approximately 11–9 Ma (Ratschbacher et al., 1994; Wiesmayer and Grasemann, 2002). The initial age of the thrusting is consistent with the collision, which implies that the Tethyan Himalayan fold thrust belt was invoked by the India–Asia collision.

The Main Central thrust belt is defined by a shear zone of a few kilometers to ~10 km in thickness. It juxtaposed the Greater Himalayan metamorphic sequence southward over the Lesser Himalayan sedimentary sequence and is commonly interpreted as a major intracontinental thrust associated with the exhumation of the High Himalayan metamorphic rocks. The geochronology from the dating of 40Ar/39Ar from amphibolite-grade rocks and U–Pb zircons from a leucogranite shows that the shear deformation occurred at approximately 23–21 Ma (Hubbard and Harrison, 1989; Hodges et al., 1992; Coleman, 1998; Walker et al., 1999; Godin et al., 2001; Daniel et al., 2003; Tobgay et al., 2012). The 22–14 Ma deformation ages and 5–3 Ma metamorphic ages in this thrust belt indicate that it underwent discontinuous deformation and propagated southward (Arita et al., 1997; Yin and Harrison, 2000; Catlos et al., 2002). However, Takagi et al. (2003) considered that the younger regions of the Main Central thrust belt were locally reactivated by brittle normal faults. Because the extent, magnitude, and tectonic significance of this younger deformation along this thrust belt remain unclear (Yin, 2006), we emphasize the ~22–14 Ma activity in this study. Based on the velocity anomalies and paleolatitude evolution of India, van Hinsbergen et al. (2012) advocated the subduction of the Greater India Basin (GIB) along the Main Central thrust belt during 50–23 Ma. Although this reorganization remains disputed and needs further study, it implies that the Main Central thrust belt may have a more complicated geological history. Due to the lack of correlative stratigraphic units across the Main Central thrust belt, the total magnitude of shortening across the fault is not well constrained.

The north-dipping Main Boundary thrust belt placed metasedimentary rocks from the Lesser Himalaya over un-metamorphosed Miocene–Pleistocene clastic rocks from the Himalayan foreland basin (Yin, 2006). However, there is no overlapping or crosscutting relationship that closely defines the deformation history, so the stratigraphic and structural data were used to define the activity of the Main Boundary
thrust belt. The increasing erosion in the Lesser Himalaya at 12–10 Ma and rates of the foreland-basin fill between 11 and 9.5 Ma are considered to reflect the occurrence of the Main Boundary thrust belt (Meigs et al., 1995; Burbank et al., 1996; Huyge et al., 2001). Detailed structural and geochronological studies in Nepal show distinct phases of motion on the Main Boundary thrust belt at ~2.3 Ma (Macfarlane, 1993), which may imply that it was still active during the Pleistocene.

The Main Frontal thrust belt is regarded as the southernmost thrust belt in the Himalaya (Fig. 2), which placed the Neogene Siwalik sediments over Quaternary Gangetic plain deposits. This fault is commonly expressed as a zone of folds and blind thrusts. Sedimentation and paleomagnetic data in related foreland basin show that the thrusting is no older than 5 Ma (Baker et al., 1988; Najman, 2006). Cross sections show that the shortening between the Main Central to the Main Frontal thrust belts is 164–267 km in Bhutan, 308 km in Western Arunachal Pradesh, 45–70 km in Eastern Nepal, and 228–287 km in western Nepal (Table 2) (DeCelles et al., 2001, and their references therein; Long et al., 2012).

The above observations show that the Yarlung Zangpo–Himalaya thrust system started as early as ~50 Ma and has undergone long-term development since the India–Asia collision. These fold–thrust belts account for significant crustal shortening (Table 2) (Schelling and Arita, 1991; Schelling, 1992; Ratschbacher et al., 1994; Srivastava and Mitra, 1994; DeCelles et al., 1998, 2001; Lave and Avouac, 2000; Long et al., 2012). By combining the shortening estimates from the main thrust and tectonic units, it can be concluded that the shortening is between ~900 km (Western Nepal) and ~450 km (Pakistan and Bhutan) within the Himalaya; adding the 100–120 km shortening in the Indus–Yarlung Zangpo suture, the total shortening was at least ~550–1020 km since the India–Asia collision.

### Table 1

<table>
<thead>
<tr>
<th>Region/structure</th>
<th>Acronym</th>
<th>Age</th>
<th>Method</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Eastern Tibet thrust system</strong></td>
<td></td>
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<td></td>
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<tr>
<td>Yutungshang–Yulongshan thrust belt</td>
<td>YYT</td>
<td>12–8 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar; Rb-Sr</td>
<td>Zhang et al. (2008)</td>
</tr>
<tr>
<td>Daxueshang–Gonggashan thrust belt</td>
<td>DGT</td>
<td>~18 Ma</td>
<td>U–Pb</td>
<td>Zhang et al. (2008)</td>
</tr>
<tr>
<td>Ganzhi–Litiang thrust belt</td>
<td>GLT</td>
<td>20–16 Ma</td>
<td>Apatite fission track data</td>
<td>Lai et al. (2006)</td>
</tr>
<tr>
<td>Longmen Shan thrust belt</td>
<td>LMST</td>
<td>10–0 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar; (U–Th)/He</td>
<td>Kirby et al. (2002)</td>
</tr>
<tr>
<td><strong>Northern Tibet thrust system</strong></td>
<td></td>
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<tr>
<td>Northern Qilian Shan thrust belt</td>
<td>NQLT-1</td>
<td>8.3–0 Ma</td>
<td>Inferred from sedimentation</td>
<td>Yang et al. (2007)</td>
</tr>
<tr>
<td>Northern Qilian Shan thrust belt</td>
<td>NQLT-2</td>
<td>3–0 Ma</td>
<td></td>
<td>Tapponnier et al. (1990)</td>
</tr>
<tr>
<td><strong>Eastern Qilian Shan–Nanshan thrust belt</strong></td>
<td>EQNT</td>
<td>~29 Ma</td>
<td>Inferred from sediments and</td>
<td>Fang et al. (2003)</td>
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<td>Nan Shan thrust belt</td>
<td>NST</td>
<td>~33 Ma</td>
<td>Inferred from fission-track age</td>
<td></td>
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<tr>
<td>West Kunlun thrust belt</td>
<td>WKT</td>
<td>~23–0 Ma</td>
<td>Inferred from fission-track age</td>
<td></td>
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<tr>
<td>Mazatzage thrust belt</td>
<td>MT</td>
<td>~3 Ma</td>
<td>Biostratigraphy</td>
<td>Si et al. (2009)</td>
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<td>Kudi–Gaizi thrust belt</td>
<td>KGST</td>
<td>37–36 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Ding et al. (1996)</td>
</tr>
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<td>Main Pamir thrust belt</td>
<td>MPT</td>
<td>~20 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Solbe and Dumitru (1997)</td>
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<td>Northern Qaidam thrust belt</td>
<td>NQT</td>
<td>~49 Ma</td>
<td>Inferred from stratigraphic data of Qaidam basin</td>
<td>Yin et al. (2002)</td>
</tr>
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<td>Northern Kunlun thrust belt</td>
<td>NKT</td>
<td>~49 Ma</td>
<td>Inferred from sediments and relationship with the fault</td>
<td>Yin et al. (2002)</td>
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<td>Qimien Tagh thrust belt</td>
<td>QTG</td>
<td>~49 Ma</td>
<td>Biostratigraphy, (\text{He}^\text{fission track data})</td>
<td>Yin et al. (2002)</td>
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<td>Southern Qaidam thrust belt</td>
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<td>~35 Ma</td>
<td>Biostratigraphy, (\text{He}^\text{fission track data})</td>
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<td>SEKT</td>
<td>26–13 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Wu et al. (2009)</td>
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<td><strong>Central Tibet thrust system</strong></td>
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<tr>
<td>Fengshuo Shan–Nanqiang thrust belt</td>
<td>FT</td>
<td>56–23 Ma</td>
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<td>Tanggula Shan thrust belt</td>
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<td>~23–23 Ma</td>
<td>U–Pb and magnetostratigraphy</td>
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<td>Lugu–Rongma thrust belt</td>
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<td>~35 Ma</td>
<td>Constrained by strata</td>
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<td>Zadana–Rigangpe thrust belt</td>
<td>ZBTC</td>
<td>80–50 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Kapp et al. (2005)</td>
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<td>Gangma Co–Shuanghu thrust belt</td>
<td>GST</td>
<td>~20–27 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Kapp et al. (2005)</td>
</tr>
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<td>Southern Qiangtang thrust belt</td>
<td>SQST</td>
<td>~20–27 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Kapp et al. (2005)</td>
</tr>
<tr>
<td>Gaize–Silin Co thrust belt</td>
<td>GSCST</td>
<td>99–23 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Kapp et al. (2003)</td>
</tr>
<tr>
<td>Shiquanhe–Gaize–Amdo thrust belt</td>
<td>SGAT</td>
<td>116–23 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Kapp et al. (2003), Yin and Harrison (2000)</td>
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<tr>
<td>Coqin thrust belt</td>
<td>CQT</td>
<td>~50 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Murphy et al. (1997), Volkmer et al. (2007)</td>
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<td><strong>Yarlung Zangpo–Himalaya thrust system</strong></td>
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<td>Gangdese thrust belt</td>
<td>GT</td>
<td>30–23 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Yang et al. (1994, 1999)</td>
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<td>Great Counter thrust belt</td>
<td>GCT</td>
<td>25–10 Ma</td>
<td>$^{40}$Ar/$^{39}$Ar</td>
<td>Ratschbacher et al. (1994), Wiesmayr and Grasemann (2002)</td>
</tr>
</tbody>
</table>
| Tethyan Himalayan fold thrust belt| THFT    | ~50–10 Ma | $^{40}$Ar/$^{39}$Ar; K/Ar | Coleman (1998), Hubbard and Harrison (1989), Hodges et al. (1992), Godin et al. (2001), Walker et al. (1999), Arita et al. (1997), Tobgay et al. (2012), Daniel et al. (2003), Carle et al. (2002), Laves and Avouac, 2000; Long et al. (2012). By combining the shortening estimates from the main thrust and tectonic units, it can be concluded that the shortening is between ~900 km (Western Nepal) and ~450 km (Pakistan and Bhutan) within the Himalaya; adding the 100–120 km shortening in the Indus–Yarlung Zangpo suture, the total shortening was at least ~550–1020 km since the India–Asia collision.

### 3.2. Central Tibet thrust system

Central Tibet is bounded by the Indus–Yarlung Zangpo suture to the south and the Ayimaqin–Kunlun–Muttagh suture to the north and is made up of five tectonic units. These units are the Lhasa terrane, the Bangong–Nujiang suture, the Qiangtang terrane, the Jinchu suture, and the Songpan–Ganzi terrane from south to north (Figs. 1 and 2). A series of thrust belts and related basins have been confirmed in central Tibet, which provides effective constraints on the deformation history and
uplift of the Tibet plateau. Based on the tectonic position, these thrust systems can be divided into the Lhasa–Qiangtang and Songpan–Ganzi (Fenghuo Shan–Nangqian) secondary systems.

The southern Lhasa terrane consists mainly of Gangdese granitoids. The convergence deformation has been studied in detail in the northern part. The long-term Coqin thrust belt (CQT) records the shortening history of the Lhasa terrane. The youngest rocks involved in this thrust belt are Aptian–Albian limestones, and the thrust belt is locally overlain by gently dipping Early Tertiary volcanic tuffs. The geologic relations of the Mesozoic-Paleogene paleogeography reveals that the southern Tibet was under an extensional environment from 130 to 90 Ma and the Lhasa–Qiangtang collision was at ~90 Ma (Zhang et al., 2004, 2012, 2014); these indicate that the Coqin thrust belt has undergone intensive shortening since ~90 Ma. The thrust belt is estimated to have accommodated ~187 km of crust shortening, which is attributed to the Lhasa–Qiangtang collision (Murphy et al., 1997) and the Neo-Tethyan flat-slab subduction (Zhang et al., 2012). The termination of long-term convergence in the Lhasa terrane can be constrained by stratigraphic data. The deformation of the retro-foreland basin and Early Miocene horizontal strata in the Lhasa terrane implies that the significant folding and thrusting ceased before ~23 Ma (Kapp et al., 2007b; Wu et al., 2008), and the S–N shortening did not exceed ~15% (~40 km) between 50 and 20 Ma (van Hinsbergen et al., 2011b).

In general, the Shiquanhe–Gai–Amdo thrust belt (SGAT) and Gaize–Siling Co thrust belt (GCS) follow the trace of the Bangong–Nujiang suture and represent the deformation of the Lhasa–Qiangtang collision and reactivation of the Late Cretaceous suture during the Cenozoic (Fig. 2) (Zhang et al., 2012, 2014). The initial ages were 116–107 Ma (SGAT) (Kapp et al., 2003) and 99 Ma (GCS) (Kapp et al., 2007a).

### Table 2
Shortening magnitudes of the major terranes and thrust belts.

<table>
<thead>
<tr>
<th>Region/structure</th>
<th>Amount</th>
<th>Deformed age</th>
<th>Method and location</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Major terranes</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Northern Pamir</td>
<td>260–300 km</td>
<td>23–0 Ma</td>
<td>Inferred from regional deformation</td>
<td>van Hinsbergen et al. (2011b), Burtman (2000)</td>
</tr>
<tr>
<td>Western Kunlun Shan</td>
<td>140–187 km</td>
<td>23–0 Ma</td>
<td>Inferred from regional deformation</td>
<td>Cowgill et al. (2003), Cowgill et al. (2003), van Hinsbergen et al. (2011b)</td>
</tr>
<tr>
<td>Qilian–Nan Shan</td>
<td>80 km</td>
<td>Eocene</td>
<td>Inferred from regional deformation</td>
<td>Guillot and Replumaz (2013), Yin and Harrison (2000)</td>
</tr>
<tr>
<td>Qaidam terrane</td>
<td>110 km</td>
<td>Post-Eocene</td>
<td>Inferred from regional deformation</td>
<td>Guillot and Replumaz (2013)</td>
</tr>
<tr>
<td>Qilian–Nanshan–Qaidam</td>
<td>140 km</td>
<td>30–0 Ma</td>
<td>Inferred from regional deformation</td>
<td>van Hinsbergen et al. (2011b), van Hinsbergen et al. (2011b), Coward et al. (1988), Spurin et al. (2005)</td>
</tr>
<tr>
<td>Songzan–Ganzi terrane</td>
<td>200 km</td>
<td>50–30 Ma</td>
<td>Inferred from regional deformation</td>
<td>van Hinsbergen et al. (2011b), Kapp et al. (2005)</td>
</tr>
<tr>
<td>Qiangtang terrane</td>
<td>120 km</td>
<td>50–23 Ma</td>
<td>Inferred regional deformation</td>
<td>van Hinsbergen et al. (2011b), Kapp et al. (2005, 2007a), DeCelles et al. (2007)</td>
</tr>
<tr>
<td>Lhasa terrane</td>
<td>250 km</td>
<td>100–50 Ma</td>
<td>Inferred regional deformation</td>
<td>van Hinsbergen et al. (2011b), Murphy et al. (1997), van Hinsbergen et al. (2011b), Kapp et al. (2007b)</td>
</tr>
<tr>
<td>Himalaya–Indus–Yarlung suture</td>
<td>~470, 480–590, 590–920, 620–640 km</td>
<td>&lt;55 Ma</td>
<td>Balanced and regional sections, in Pakistan, northern India, western Nepal, Sikkim and Bhutan, Arunachal Pradesh</td>
<td>Long et al. (2012), DeCelles et al. (2001), and their references</td>
</tr>
<tr>
<td><strong>Major thrust belts</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Longmen Shan thrust belt</td>
<td>25 km</td>
<td>10–0 Ma</td>
<td>Balanced cross section, in the central part of the Longmen Shan</td>
<td>Burchfiel et al. (2008), Hubbard and Shaw (2009)</td>
</tr>
<tr>
<td>Western Kunlun thrust belt</td>
<td>140 km</td>
<td>23–0 Ma</td>
<td>Balanced cross section, in Tianshuai</td>
<td>Cowgill et al. (2003)</td>
</tr>
<tr>
<td>Northern Qilian Shan thrust belt</td>
<td>33–56 km</td>
<td>9–0 Ma</td>
<td>Balanced cross section, in the central part of the northern Qilian Shan</td>
<td>Rumelhart (1998)</td>
</tr>
<tr>
<td>Nan Shan thrust belt</td>
<td>350 km</td>
<td>&lt;33 Ma</td>
<td>Regional deformation</td>
<td>Yang et al. (2007)</td>
</tr>
<tr>
<td>Nangqian thrust belt</td>
<td>~110 km</td>
<td>&lt;50–~23 Ma</td>
<td>Balanced cross section, in Nangqian</td>
<td>Yin and Harrison (2000)</td>
</tr>
<tr>
<td>Fenghuo Shan thrust belt</td>
<td>60–80 km</td>
<td>56–23 Ma</td>
<td>Balanced cross section, in Fenghuo Shan</td>
<td>Spurin et al. (2005)</td>
</tr>
<tr>
<td>Tangguhalan thrust belt</td>
<td>185 km</td>
<td>55–23 Ma</td>
<td>Balanced cross section, middle part of Tanggula Shan</td>
<td>Wang et al. (2002a), Yin and Harrison (2000)</td>
</tr>
<tr>
<td>Shiquanhe–Gaize–Amdo and Gaize–Siling Co thrust belt</td>
<td>250 km</td>
<td>100–25 Ma</td>
<td>Inferred from regional deformation</td>
<td>Li et al. (2012)</td>
</tr>
<tr>
<td>Cuqin thrust belt</td>
<td>&gt;187 km</td>
<td>100–50 Ma</td>
<td>Balanced cross section, in Cuqin</td>
<td>Yin and Harrison (2000)</td>
</tr>
<tr>
<td>Gangdese thrust belt</td>
<td>50–60 km</td>
<td>30–23 Ma</td>
<td>Balanced cross section, in southwest Tibet</td>
<td>Murphy et al. (1997), Yin et al. (1999), Murphy and Yin (2003), Yin and Harrison (2000)</td>
</tr>
<tr>
<td>Great Counter thrust belt</td>
<td>~38–60 km</td>
<td>25–10 Ma</td>
<td>Balanced cross section, in western Tibet</td>
<td>Murphy et al. (2009), Murphy and Yin (2003)</td>
</tr>
<tr>
<td>Tethyan Himalayan thrust belt</td>
<td>110–140 km</td>
<td>50–10 Ma</td>
<td>Balanced cross section, in central Himalaya</td>
<td>Ratschbacher et al. (1994), Murphy and Yin (2003)</td>
</tr>
<tr>
<td>Main Central thrust belt</td>
<td>97–150 km</td>
<td>23–14 Ma</td>
<td>Balanced cross section</td>
<td>Long et al. (2012)</td>
</tr>
<tr>
<td>High Himalaya (STDS–MCT)</td>
<td>138–209; 207; 140–210; 111–221 km</td>
<td>23–14 Ma</td>
<td>Balanced and regional cross sections, in Bhutan, western Arunachal Pradesh, eastern Nepal, and the western Nepal</td>
<td>Long et al. (2012), DeCelles et al. (2001), and their references therein</td>
</tr>
<tr>
<td>Lesser Himalaya (MCT–MFT)</td>
<td>164–267; 308; 45–70; 228–522 km</td>
<td>&lt;23 Ma</td>
<td>Balanced and regional cross sections, in Bhutan, western Arunachal Pradesh, eastern Nepal, and the western Nepal</td>
<td>Long et al. (2012), DeCelles et al. (2001), and their references therein</td>
</tr>
</tbody>
</table>
constrained by the age of the folded strata and volcanic flows cut by the thrusts. However, there is no direct geochronological data of the thrusting on tectonic rocks or minerals within the fault zones. Mesozoic-Paleogene paleogeography shows that the central Tibet was under an extensional environmental setting during middle Cretaceous (130–90 Ma) and the Shiquanhe–Gaize–Amdo thrust was a folded normal fault (Zhang et al., 2002, 2012, 2014). This implies that the intense thrusting and crustal shortening took place since ~90 Ma and resulted by the Lhasa–Qiangtang collision. The thrusts cut 31–25 Ma strata and covered by the Early Miocene horizontal strata imply that the significant thrusting ceased before ~23 Ma (Kapp et al., 2003, 2007a; Wu et al., 2008). The regional structural relationships and balanced cross sections indicate that the two thrust belts may have accommodated approximately 250 km or greater of crustal shortening (Table 2) (Yin and Harrison, 2000). The observations imply that the Bangong–Nujiang suture zone has undergone intense crustal shortening during the Late Cretaceous-Oligocene.

The main thrust belts in the Qiangtang terrane include the Tanggula Shan thrust belt (TTS), Gangma Co–Shanghuo thrust belt (GST), Lugu–Rongma thrust belt (LRT), Zadaona–Riganpei Co thrust belt (ZRCT), and Southern Qiangtang thrust belt (SQTT) (Fig. 2). These thrust belts occurred between the Late Cretaceous and 23 Ma and caused extensive crustal shortening (Kapp et al., 2003, 2005; Li et al., 2012). Previous estimates show that the Qiangtang terrane has undergone 400 km (50%) of shortening prior to the India–Asia collision and 120 km of shortening during 50–23 Ma (van Hinsbergen et al., 2011b). However, in the northern margin of the Qiangtang terrane, the new reconstruction indicates that the Tanggula Shan thrust belt caused 185 km of N–S shortening during 50–23 Ma. The thrust belt not only dominated the development of the Cenozoic basin but also resulted in the rapid uplift/denudation of the Tanggula Shan during the Eocene-Oligocene (Li et al., 2012). The shortening along the Tanggula Shan thrust belt during this stage may imply that the mechanism in the northern margin of the Qiangtang is different from that in the Lhasa and southern Qiangtang terranes.

The Fenghuo Shan–Nangqian secondary thrust system lies in the southern area of the Songpan–Ganzi terrane and consists of several of northwest striking thrusts (Fig. 2). These thrusts are mainly south-southwest-dipping and juxtapose Permian to Jurassic strata over north-south striking thrusts. The unconformity between the Eocene and overlying Miocene sediments indicates that the thrusts were active during 50–23 Ma. The deformation and balanced cross section suggest that the total shortening across the thrust belt is at least 60–80 km in the Fenghuo Shan (Yin and Harrison, 2000; Wang et al., 2002a). The ~260 km of shortening is comparable with the ~300 km shortening between the Pamir and Tien Shan estimated by Burtman (2000). The thrust belts within the Kunlun–Qaidam terrane consist of the Northern Kunlun thrust belt (NKT) and the Southern East Kunlun thrust belt (SEKT) (Fig. 2). These thrust belts often juxtapose the Proterozoic metamorphic and Paleozoic sedimentary rocks over the Tertiary sediments of the Qaidam Basin. Although the initiation ages are not well constrained, the thrusting and the sedimentary records in the Qaidam Basin indicate that the thrusting occurred at about the Early Eocene (~49 Ma) and has a long-term history (Table 1). For example, the Qimen Tagh thrust belt was active in ~49 Ma and in separate 5–6 Ma episodes since then (Yin et al., 2002).

3.3. Northern Tibet thrust system

Located between the Ayimaqin–Kunlun suture and the southern margin of the Tarim or Hexi Corridor Basins, the north Tibet consists of the Kunlun–Qaidam and Qilian terranes (Fig. 1). The widespread thrust belts and related foreland basins in this region record the evolution of northern Tibet during the Cenozoic. Based on the tectonic position, these thrust belts can be divided into the northwestern and northeastern secondary systems, separated by the sinistral Altyn Tagh fault (ALT).

Located to the west of the ALT, the northwestern secondary thrust system consists of the Pamir and West Kunlun thrust belts, separated by the Kashgar–Yecheng dextral strike-slip fault. The West Kunlun thrust belts lies between the Tarim and West Kunlun Shan and consists of the eastern Kudi–Gaize thrust belt (KGT), the West Kunlun thrust belt (WKT), and the Mazatage thrust belt (MT) from south to north (Fig. 2). These thrusts are south-dipping and show northward propagation in space. For example, the Kudi–Gaize thrust started at 37–36 Ma (Ding et al., 1996), the West Kunlun thrust began at ~23 Ma (Rumelhart et al., 1999; Cui et al., 2008), and the Mazatage thrust occurred ~3 Ma (Si et al., 2009). The young Mazatage thrust is the active fault that juxtaposes the Mesozoic rocks northward over the Pliocene rocks. The magnitude of the S–N shortening has been estimated at ~85 km in the eastern part and 187 km in the central part of the western Kunlun Shan (Rumelhart et al., 1999).

The Pamir thrust belts lie between the Pamir–Hindu Kush and the southwestern Tarim and was strongly affected by the India–Eurasia collision in the Cenozoic ( Molnar and Tapponnier, 1975). Age constraints show that the onset of activity along the Pamir thrust belts was at ~37 Ma, which coincides with the onset of shortening in the Western Kunlun Shan (Ding et al., 1996; Sobel and Dumitrul, 1997; Cowgill et al., 2003). Moreover, the ages of these thrusts indicate they have undergone northward growth from the Kudi–Gaize thrust (~37–36 Ma) to the Main Pamir thrust (~20 Ma) and are still active in the northern region (Sobel and Dumitrul, 1997; Burtman, 2000; Cowgill et al., 2003). The northward indentation of Pamir into Asia created the Pamir arc, which induced clockwise rotation of the western Kunlun Shan on its eastern limb and counterclockwise rotation on its western limb in western Pamir (Yin and Harrison, 2000). However, the magnitude of Cenozoic crustal shortening between the Pamir and Tien Shan is disputable, with estimations of the magnitude of overthrusting ranging between 300 and 700 km (Burtman, 2000, and their references therein). Based on kinematic constraints from the bounding faults, the Tadjik depression and palaeomagnetic constraints, van Hinsbergen et al. (2011b) concluded that the total amount of intra-Asian shortening is ~1050 km since 50 Ma south of the Pamir–Hindu–Kush region, along with ~260 km of shortening north of the Main Pamir thrust since ~23 Ma. The ~260 km of shortening is comparable with the ~300 km shortening between the Pamir and Tien Shan estimated by Burtman (2000).

3.4. Eastern Tibet thrust system

Eastern Tibet lies between the central plateau and western areas of South China (Fig. 1), dominated by E-W shortening during the
Cenozoic. Four thrust belts can be recognized based on the deformation and geochronology. From west to east, they are the Ganzi–Litang (GLT), the Daxueshan–Gonggashan (DGT), the Yupingshan–Yulongshan (YYT) and the Longmen Shan (LMST) thrust belts (Fig. 2). The 20–16 Ma ages of the Ganzi–Litang thrust and ~18 Ma age of the Daxueshan–Gonggashan thrust represent the earliest thrusting in the western part of eastern Tibet (Lai et al., 2006; Zhang et al., 2008). The NE-striking Longmen Shan thrust belt bounding the eastern margin of the Tibet Plateau has been interpreted as a foreland thrust belt during the Mesozoic. It was reactivated during the Late Miocene (12–8 Ma), as evidenced by the $^{39}$Ar/$^{39}$Ar and (U-Th)/He cooling ages in its hanging wall (Kirby et al., 2002). The thrusts remain active today, as indicated by the devastating magnitude 8.0 (Wenchuan) earthquake on 12 May 2008. Approximately 20–25 km of E–W Neogene shortening was documented in the Longmen Shan thrust belt (Burchfiel et al., 2008; Hubbard and Shaw, 2009).

4. Extension in the Tibetan–Himalayan orogen

Cenozoic extensional structures are another important deformation style of the Tibetan–Himalayan orogen, which are widely distributed in the Himalaya, the Lhasa and the Qiangtang terranes. Two sets of normal faults dominate the extensional structures of the plateau (Fig. 3; Table 3). They are the E–W striking South Tibet Detachment System (STDS) and numerous N–S striking rifts (grabens). Because of a possible linkage between the upper crustal extension and uplift of the plateau, many studies have been focused on the timing and dynamics of the extension in the plateau.

4.1. South Tibet Detachment System

In the northern Himalaya, the N–S extension resulted in the formation of a major set of E–W striking, north-dipping normal faults and shear zones known as the Southern Tibetan Detachment (STDS) (Burg et al., 1984; Burchfiel et al., 1992). The STDS is a few kilometers thick and consists of several subparallel faults or ductile shear zones and juxtaposes the Tethys Himalayan (THS) low-grade or un-metamorphosed strata over the High Himalayan metamorphic rocks (HHM), marking one of the critical geological boundaries in the Himalaya (Fig. 3). Previous studies have shown that the STDS is composed of mylonitic gneiss and leucogranite in the lower part and brittle faults in the top part (Burchfiel et al., 1992). Based on the distance between the southernmost STDS klippe carrying Ordovician strata in the STDS hanging wall and the northernmost STDS exposure, Burchfiel et al. (1992) inferred that the magnitude of slip along the STDS is ~34 km in the central Himalaya. In the Zanskar region, the minimum slip across the STDS was estimated by Herren (1987) to be ~25 km based on the assumption that the vertical condensation of the footwall's metamorphic field gradient is solely caused by simple-shear deformation. Again, assuming simple-shear deformation, Dezes et al. (1999) estimated 35 ± 9 km of slip on the Zanskar shear zone. Considering the STDS could be caused mainly by pure-shear deformation, Yin (2006) suggested the slip estimates of Herren (1987) and Dezes et al. (1999) only as the upper bounds.

Although the timing of the STDS' initiation is not well constrained, ductile motion within the STDS and the age of a cross-cutting section suggest that the main phase of the STDS took place between 23 and 17 Ma, while the $^{40}$Ar/$^{39}$Ar cooling ages and apatite fission track ages indicate that motion along the STDS may be as young as 15–13 Ma and ceased by 11–9 Ma (Table 3; Fig. 3) (Guillot et al., 1994; Coleman and Hodges, 1995; Kumar et al., 1995; Searle et al., 1997, 1999; Dezes et al., 1999; Murphy and Harrison, 1999; Walker et al., 1999; Godin et al., 2001; Vannay et al., 2004; Viskupic et al., 2005; Zhang and Guo, 2007). These ages indicate the active period of the STDS is coeval with motion along the Main Central thrust belt and the Great Counter thrust belt.

Diverse models have been proposed to interpret the formation of the STDS; these hypotheses can be combined into three general groups: 1) the increase of potential gravitational energy, followed by collapse due to excess elevation (Burg et al., 1984; Dahlen, 1984; Royden and Burchfiel, 1987; England and Molnar, 1993; Buck and Sokoutis, 1994; Hodges et al., 2001); 2) the southward flow of the Himalayan mid-lower crust and extrusion of the orogenic wedge (Yin, 1989; Grujic et al., 1996; Nelson et al., 1996; Beaumont et al., 2001; Vannay and Grasemann, 2001); and 3) the passive roof thrust model (Webb et al., 2007; Yin, 2006). The controversy surrounding the development of the STDS can be attributed to an incomplete understanding of its deformational history and relationship with other deformation styles, such as the initial age of the extension, the correlation between the STDS and the E–W extension, and the geometry and structural relationship among the STDS, the Main Central thrust belt and the Great Counter thrust belt (Yin, 2006).

4.2. North–south trending rifts

The approximate N–S trending rifts are the most distinct active tectonics in the Himalaya and Tibet (Fig. 3), which are mainly composed of an extensional basin and a series of high-angle normal faults on their shoulders. The notable rift in the Qiangtang terrane is the NNE-trending Shuanghu rift (Fig. 3). The dating reveals that the normal faults started at 13.5 Ma, which represents the onset of the E–W extension in Qiangtang (Bliushuk et al., 2001). In the Lhasa terrane, the N–S trending rifts are the Daggyai Tso rift, Tangra Yumco rift, Xainza rift, and Yanagbajing–Lugu rift from west to east (Fig. 3). The ages of the north–south-trending dikes along the normal fault define the onset of regional E–W extension in southern Tibet at 18.3 ± 2.7 Ma (Williams et al., 2001). Compared with the timing of the thrusts, the N–S trending normal faults are significantly younger than the Cenozoic thrusts in the Lhasa and Qiangtang terranes.

A series of N–S striking rifts and high-angle normal faults are documented in the Himalaya, such as the Leo Pargil (Thiede et al., 2006; Hintersberger et al., 2010) and Gurla Mandhata (Murphy et al., 2000, 2002) in the NW Himalaya; the Thakkola, the Kung Co (Mitsuishi et al., 2012) and Ama Drime Massif (Stockli et al., 2002) in the Central Himalaya; and Yadong (Edwards and Harrison, 1997) and Cuona (Wu et al., 2005) in the NE Himalaya (Fig. 3). Although geochronologic data documenting the onset of normal faulting are sparse, the extension has apparently been a protracted process that started in the Miocene (19–16 Ma) and continued until the present day (Thiede et al., 2006; Hintersberger et al., 2010; Mitsuishi et al., 2012) (Fig. 3; Table 3). Cross-cutting relationships indicate that these normal faults cut all preexisting deformation fabrics, including the E–W striking thrusts and the STDS. For example, the N–S striking normal faults bounding the Ama Drime Massif in the Central Himalaya displaced the STDS for several kilometers (Jespj et al., 2008). The difference in strike between the STDS and the N–S trending rifts has long been interpreted as related to two different tectonic events. However, it is unclear whether the rifts are linked to the STDS or if they represent an entirely different process.

5. Discussion

5.1. Propagation of fold–thrust deformation

As discussed above, several models have been proposed for the growth of the Tibetan plateau. However, an unsolved problem regarding whether and how the thrusting and crustal shortening responded to the growing processes remains uncertain. The large-scale fold–thrust belts record the crustal shortening and the tectonic history of the plateau (Ratschbacher et al., 1994; Yin and Harrison, 2000; Guillot and Replumaz, 2013) and are crucial for understanding the deformation and plateau growth. Based on the spatial and temporal evolution of these fold–thrust belts, we propose that the crustal shortening of the
whole plateau has undergone four significant stages since the Late Cretaceous (Fig. 4), namely, the Late Cretaceous-Paleocene (~90–55 Ma), Eocene-Oligocene (~55–23 Ma), Early-Middle Miocene (~23–10 Ma), and Middle Miocene-present (~10–0 Ma). Moreover, these thrust belts display an outward propagation from central Tibet.

5.1.1. Late Cretaceous-Paleocene (~90–55 Ma)

Recent high-solution isotopic and radiolarian dating on the ophiolites has established definite Aptian–Albian and probable Cenomanian ages and reveals that the Bangong–Nujiang ocean did not close until the Late Cretaceous. Lhasa terrane was under an extensional environment from 130 to 90 Ma as indicated by intense bimodal magmatism and extensive marine sedimentation (Zhang et al., 2012, 2014 and references therein). Since the Late Cretaceous (~90 Ma), flat slab subduction of the Neo-tethys, as indicated by the adakites in the southernmost Lhasa terrane, together with the ongoing Qiangtang–Lhasa collision could have contributed to the shortening of the central Tibet and the growth of the Tibetan plateau (Zhang et al., 2012, 2014 and references therein). Therefore, we consider the earliest thrusting in central Tibet began at ~90 Ma.

During the first stage (~90–55 Ma), the fold–thrust deformation and crustal shortening mainly occurred in central Tibet, i.e., the Lhasa terrane, the Bangong–Nujiang suture, and the Qiangtang terrane (Figs. 2 and 4A; Table 1). The fold–thrust belts are E–W striking, including the Coqin Thrust belt (CQT) (~50 Ma), the Gaize–Siling Co (GSCT) (~99–23 Ma), the Shiquanhe–Gaize–Amdo (SGAT) (~23 Ma) and the Zadaona–Riganpei Co (ZRTC) (~88–50 Ma) thrust belts from south to north. The occurrence of these thrust belts is significantly earlier than the onset of the India–Asia collision (~55–50 Ma). Previous studies suggest that a substantial amount of the shortening that took place predates the India–Asia collision (van Hinsbergen et al., 2011b). The geodynamics of the crustal shortening is attributed to the Lhasa–Qiangtang collision (Murphy et al., 1997; Kapp et al., 2003, 2005; Volkmer et al., 2007) and subduction of the Neo-Tethys oceanic crust (Zhang et al., 2007, 2012, and their references therein), which implies that the Tibetan plateau is similar to the Altiplano of the Andes (Zhang et al., 2012; Lippert et al., 2014). The Tanggula Shan (TTS) and Fenghuo Shan–Nangqian thrust belts (FT) located north of the Qiangtang terrane and coeval with the onset volcanism may result from the southward subduction of the Sonpan–Ganzi terrane (see below) (Fig. 4A). This significant shortening suggests that the crust of the central plateau may have been as thick as 50–55 km at the onset of the India–Asia collision (Murphy et al., 1997; Kapp et al., 2003, 2005).

5.1.2. Eocene–Oligocene (~55–23 Ma)

The second stage (~55–23 Ma) is defined by the onset of the India–Asia collision at ~55–50 Ma and termination of the N–S shortening in central Tibet at ~23 Ma. During this stage, fold–thrust belts were widely developed from the Himalaya to Qilian Shan due to long-term convergence (Figs. 2 and 4B). Meanwhile, the activity of these fold–thrust belts dominated an obvious propagation from central Tibet during this stage.

In central Tibet, there are two structural styles related to the N–S shortening. One is the successive deformation of the thrust belts that started in the first stage, while another is newly emerging thrust belts such as the Southern Qiangtang (SQT) (~33–23 Ma), the Lugu–Rongma (LRT) (~35 Ma), and the Gangma Co–Shuanghu (GST) (~30–23 Ma) thrust belts (Figs. 2, Table 1). Tectonic reconstructions show ~40 and ~120 km N–S shortening in the Lhasa and Qiangtang terranes during 50–23 Ma, respectively (Table 2) (van Hinsbergen et al., 2011b). All of
these fold–thrust belts are overlain by unfolded Miocene sediments, indicating that N–S shortening terminated before 23 Ma in central Tibet (Wu et al., 2008; Li et al., 2012).

In the north Qiangtang and the Songpan–Ganzi terranes, the deformation and sequential filling of the Hoh Xil Basin show the Tanganla Shan thrust belt (TT) and the Fenghuo Shan–Nangan thrust belt (FT) experienced intense crustal shortening during the Eocene–Oligocene (Wang et al., 2002a; Spurlin et al., 2005) and terminated before 23 Ma (Wang et al., 2002a; Wu et al., 2008; Li et al., 2012). Widely distributed Cenozoic detritus was deposited during this stage in response to intense deformation, crustal thickening, and erosion in this region during the Eocene–Oligocene (Ding et al., 2000; Lai et al., 2001). However, compared to central Tibet, the peak of shortening in northern Qiangtang and Songpan–Ganzi terranes lagged behind that of central Tibet (>50 Ma).

In the northeastern plateau, a series of NW–SE striking fold–thrust belts were produced in the Qaidam–Qilian terrane during this stage. These thrust belts consist of the Southern Qaidam (SQT) (~35 Ma), the Southern Qaidam (QTT) (~49 Ma), the Northern Qaidam (NQT) (49 Ma), the Nan Shan (NST) (~33 Ma), and the Eastern Qilian Shan–Nanshan (EQNT) (~29 Ma) thrust belts from south to north (Figs. 2 and 4B). The ~49 Ma initiation ages of the Southern and Northern Qaidam thrust belts date the onset of thrust movement in this region, and the activity of the Nan Shan and the Eastern Qilianshan–Nanshan thrust belts indicated that the shortening has reached the Nan Shan during this period (Figs. 2 and 4B). 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5.1.4. Middle Miocene–present (~10–0 Ma)

Since the Middle Miocene (10 Ma), the thrust belts mainly developed around the edges of the Himalaya and Tibetan plateau, including the Longmen Shan thrust belt, Main Frontal thrust belt, Mazatape thrust belt and Northern Qilian Shan thrust belt. These active thrusts not only bounded the present plateau and adjacent basins in geomorphology but also produced active seismicity.

In northeastern Tibet, the Northern Qilian Shan thrust belt (NQLT) juxtaposed the Paleozoic strata northward over the Late Miocene-Quaternary sediments of the Hexi Corridor Basin (Figs. 2 and 4D). The growth strata and magnetic stratigraphy indicate that the thrusting started at approximately 9 Ma in the south (NQLT-1) and 3 Ma in the north (NQLT-2) (Tapponnier et al., 2001). Moreover, deep seismic reflection profiling reveals that the blind northward thrust developed well along the northern edge of the Hexi Corridor Basin (Gao et al., 1999). These observations imply that the thrusting expanded northwardly and the thrusts may reach the inner or northern areas of the Hexi Corridor Basin in the future.
In Western Kunlun Shan, the initiation of the northern fault within the West Kunlun thrust belt (WKt) was defined at ~5.0 Ma (Zheng et al., 2000; Sun et al., 2008) based on the deformation and growth strata of related foreland basins. The 300 km long Mazatage thrust belt (MT) parallels the West Kunlun thrust belt (WKt) and is an active thrust in the southern area of the Tarim Basin (Si et al., 2009). Teleseismic imaging also reveals that a blind thrust ramp occurs in this area (Wittlinger et al., 2004). Moreover, the seismic profiles show that the present contractional deformation is mainly concentrated in the northern part of the West Kunlun thrust belt and the Mazatage thrust belt. The Mazatage thrust belt is the result of the northward propagation of the West Kunlun thrust belt along gypsum-bearing Eocene mudstone and created fault propagation folds that are topographically expressed in the Tarim Basin. All these observations indicate that the West Kunlun thrust belt is propagating toward the Tarim Basin and the Mazatage ought to uplift in the future, extending the topographic growth northward (Figs. 2 and 4C). In the northern region of the Pamir, since the occurrence of the Main Pamir thrust at approximately 10 Ma (Burchfiel et al., 1995; Kirby et al., 2002), An estimated 20–25 km of E–W shortening has occurred in the Longmen Shan thrust belt (Burchfiel et al., 2008; Hubbard and Shaw, 2009). Meanwhile, the seismic reflection profiles from the Longmen Shan foothills to the Sichuan Basin reveal that the eastward thrusting has propagated far into the Sichuan Basin (Longquan Shan thrust), resulting in the passive uplift of the Longquan Shan (Hubbard and Shaw, 2009).

The Longmen Shan thrust belt (LMST), which bounds the Longmen Shan and Sichuan Basin, represents folding and thrusting in the eastern margin of the plateau (Figs. 2 and 5B). The $^{40}\text{Ar}/^{39}\text{Ar}$ and (U–Th)/He cooling ages in its hanging wall, combined with the deposition of Late Miocene conglomerates in the western Sichuan Basin, define the onset of the thrusting at approximately 10 Ma (Burchfiel et al., 1995; Kirby et al., 2002). An estimated 20–25 km of E–W shortening has occurred in the Longmen Shan thrust belt (Burchfiel et al., 2008; Hubbard and Shaw, 2009). Meanwhile, the seismic reflection profiles from the Longmen Shan foothills to the Sichuan Basin reveal that the eastward thrusting has propagated far into the Sichuan Basin (Longquan Shan thrust), resulting in the passive uplift of the Longquan Shan (Hubbard and Shaw, 2009).

In the Himalaya region, the activity of the Main Boundary and the Main Frontal thrust belts illustrates that the folding and thrusting have propagated to the Himalayan front (Figs. 2 and 4D). Although the Main Boundary thrust belt may have begun between 12 and 11 Ma, significant changes in the sedimentation patterns of related foreland basins indicate that it has undergone intense deformation since 11 Ma (Burbank et al., 1996). The Main Frontal thrust belt placed the Siwalik Group over the Quaternary sediments, which shows that thrusting has expanded to the southern boundary of the Himalayan range since 5 Ma. The Holocene slip rate in the Main Frontal thrust belt is estimated at approximately 21 ± 1.5 mm/yr (Lave and Avouac, 2000), similar to the GPS results across the Nepal Himalaya with rates of 18 ± 2 mm/yr and 15 ± 5 mm/yr (Jouanne et al., 1999; Larson et al., 1999), which means that it accommodates most of the present-day convergence.

From the above analyses, it can be clearly seen that the development of the fold–thrust belts and crustal shortening represents outward migration from central Tibet. Prior to the India–Asia collision, contractional deformation mainly occurred in the Qiangtang and Lhasa terranes. With the India–Asia collision and subsequent convergence, the thrust belts have been propagating to the margins of the orogen. The active thrust belts around the Tibetan–Himalayan orogen mark ongoing thrusting.

5.2. Deformation and uplift of the Tibetan–Himalayan orogen

Deformation and topography are inextricably linked. Although the local folding and thrusting have undergone an outward propagation as discussed above, it is still uncertain how crustal shortening responded to the outward propagation of the contractional deformation and whether the topographic uplift of the plateau was led by the folding and thrusting.

5.2.1. Shortening and crustal thickening

According to the available palaeomagnetic data from the Indian Ocean, the Cenozoic Mongolian volcanic rocks and the Chinese Paleogene sediments, most researchers concur on their estimation of a total
convergence between India and stable Asia between 2000 and 3600 km since the India–Asia collision (Molnar and Tapponnier, 1975; Dewey et al., 1989; Le Pichon et al., 1992; Replumaz and Tapponnier, 2003; Dupont-Nivet et al., 2010; Guillot and Replumaz, 2013). However, the amount of convergence depends on the onset of the collision (van Hinsbergen et al., 2011b). According to the apparent polar wander paths of India and Eurasia, Dupont-Nivet et al. (2010) estimated 2900 ± 600 km of subsequent latitudinal convergence between India and Asia, divided into 1100 ± 500 km within Asia and 1800 ± 700 km within India since the onset of the collision at ~46 Ma. By comparing the positions of P-wave anomalies in the deep mantle with the present-day position of the Indian front, Replumaz et al. (2013) estimated that India has indented Asia for approximately ~2200 km since 45 Ma. Based on marine magnetic anomalies and the Eurasia–North America–Africa–India plate circuit, van Hinsbergen et al. (2011a, 2011b) assigned ~2400 and ~3200 km of post-50 Ma India–Asia convergence for the western and eastern Himalayan syntaxes. Moreover, the convergence is up to 3200–4000 km for the India–Asia collision at 55 Ma (van Hinsbergen et al., 2011b). However, where this convergence was accommodated (thrusting/subduction or extrusion) and how much crustal shortening to the south or north of the Indus–Yarlung Zangbo suture remains disputed (Replumaz and Tapponnier, 2003; Dupont-Nivet et al., 2010; Replumaz et al., 2010, 2013; van Hinsbergen et al., 2011b; Sun et al., 2012; Guillot and Replumaz, 2013). To estimate the crustal shortening of the Himalaya and Tibet plateau, we summarize the available shortening data in Table 2.

South of the Indus–Yarlung Zangbo suture, the deformation and the shortening since the India–Asia collision are relatively well-known. From the western to eastern Himalayan syntaxes, the magnitudes of shortening are ~470 km in Pakistan, 480–590 km in north-western India, 590–920 km in western Nepal, 620–640 km in Sikkim and Bhutan, and 580 km in Arunachal Pradesh (92°E) (DeCelles et al., 2002; Long et al., 2012, and their references therein; van Hinsbergen et al., 2011b; Webb, 2013). The total percent of shortening varies, approaching 64% near the western syntaxes, increasing to a maximum of 84% in western Nepal, and decreasing to 75–66% in Bhutan and Arunachal Pradesh (Long et al., 2012). However, one cross section from western Nepal shows as much as 920 km of shortening; most shortening estimates converge to 500–600 km. To effectively utilize the previous data and obtain an accurate amount from the Himalaya to Northern Tibet, we carry out a estimation along 94°E, which shows the total shortening to the south of the Indus–Yarlung Zangbo suture is 654 km (Long et al., 2012); subtracting ~34 km of extension along the STDS (Burchfiel et al., 1992), the actual magnitude is 620 km. The above shortening magnitude is less than the size of Greater India (~3000 km) (van Hinsbergen et al., 2011b, 2012). The magnitude of the deficits suggests that large volumes of India may have been subducted below Asia without leaving a shortening record in the Himalaya.

Central Tibet (Lhasa and Qiangtang terranes) was significantly shortened through folding and thrusting during the Late Cretaceous–Paleogene (Table 2). The shortening of the Qiangtang and Lhasa terranes were ~400 km and ~250 km prior to the India–Asia collision (van Hinsbergen et al., 2011b, and their references therein). This implies that central Tibet accommodated ~650 km of N–S shortening before the India–Asia collision. Field evidence from a continuous E–W belt of Late Cretaceous thrusts and associated basins in western and central Tibet shows that central Tibet underwent minimal reactivation by thrusting during the Late Paleogene (Murphy et al., 1997; Kapp et al., 2003, 2005). That is consistent with the 120 km (~25%) shortening by thrusting in the southern Qiangtang terrane and Bangong–Nujiang suture between 50 and 20 Ma (van Hinsbergen et al., 2011b). However, the new data from the northern Qiangtang suggest that the Tanggula Shan thrust resulted in 185 km of crustal shortening during 52–~23 Ma (Li et al., 2012). Thus, it can be inferred that the total amount of shortening in the Qiangtang terrane and Bangong–Nujiang suture must exceed 300 km after 55 Ma. Moreover, the intense shortening in the Tanggula Shan thrust and the homochronous sediments in the Hoh Xil Basin led Wang et al. (2008a) to propose that the Tanggula Shan was the northern boundary of the Proto-plateau during the Oligocene.

Assuming consistent shortening (50%) in the Yushu–Nangqian and Fenghuo Shan regions, van Hinsbergen et al. (2011b) obtained 200 km of N–S shortening for the entire Songpan–Ganzi terrane between 50 and 30 Ma. However, balanced cross sections across the Fenghuo Shan and Yushu–Nangqian show 60–80 km (43%) (Wang et al., 2002a) and ~110 km (~43%) (Spurlin et al., 2005), respectively. If we extrapolate the estimated 43% shortening from our traverse to the entire width of the ~200-km-wide Songpan–Ganzi terrane, our minimum estimated Cenozoic shortening across the entire Songpan–Ganzi terrane along the longitude 94°E is 150 km. Our result is consistent with the 150 km shortening estimated by Roger et al. (2010).

The sinistral Altyn Tagh fault (ATF) bounds Northern Tibet into northwestern and northeastern parts. Displacement along the ATF is transferred into the fold–thrust belts of the Western Kunlun Shan in the western part and the Kunlun Shan–Qaidam–Qilian Shan–Nan Shan in the eastern part. In the same way, the Kashgar–Yecheng dextral strike–slip fault (KYF) transferred convergence along the North Pamir thrust to shortening in the Western Kunlun Shan. Age constraints for the onset of activity of the KYF are inferred to coincide with the 37–23 Ma onset of shortening in the Western Kunlun Shan (Sobel and Dumitru, 1997; Cowgill et al., 2003; van Hinsbergen et al., 2011b).

For the Western Kunlun Shan, different balanced cross sections show a similar S–N shortening magnitude of 140–187 km (Rumelhart et al., 1999; Cowgill et al., 2003; van Hinsbergen et al., 2011b). However, the amount of shortening within the northeastern part of the ALF is a matter of debate because the exact location and distribution of shortening in the Qaidam–Qilian Shan–Nan Shan region remains enigmatic. Yin and Harrison (2000) estimated ~280 ± 30 km of left-lateral slip along the eastern part of the ATF; the minimum amount of shortening accommodated by Nan Shan thrust belt is approximately 270 km. Adding the 80 ± 20 km of shortening between the Qilian and Nan Shan thrust belts before the activation of the ATF, the total amount of shortening within the Qilian–Nan Shan thrust belt is ~350 km during the Cenozoic. Guillot and Replumaz (2013) suggested 190 km and 330 km of shortening during the Eocene and post-Eocene in the Qaidam–Nan Shan region–Qilian Shan. Assuming for simplicity that all motion along the ATF was accommodated south of the northern Qilian Shan front, and applying bulk shortening evenly over the area from the Eastern Kunlun Shan to the Qilian Shan, van Hinsbergen et al. (2011b) suggested that 140 km of S–N shortening has been accommodated within the Qaidam–Qilian Shan–Nan Shan since 30 Ma. This explanation commendably reconciles the amount of shortening in the western and eastern parts of the ATF. However, we think that the amount of shortening was undervalued in the eastern part of the ATF for the following reasons: 1) The amount of shortening after 30 Ma was 140 km, but the shortening before this stage was neglected. The deformation has shown that the contractional structure in the Qilian Shan and Eastern Kunlun Shan occurred in the Early Eocene, immediately after the onset of the India–Asia collision (Yin and Harrison, 2000; Jolivet et al., 2001; Yin et al., 2008). Moreover, based on the recovery of the seismic profile and active thrusts in northern Qilian Shan, it has been estimated that the cumulative shortening reached 150 km in the last 10 Ma, producing a shortening rate of 1.5 cm/yr, which is similar to the sinistral slip rates along the Kunlun, Haimayan and eastern branch of the Altyn Tagh strike–slip faults (Yang et al., 2007; Guillot and Replumaz, 2013). 2) The evolution of the Qaidam basin and its relationship with the sinistral ATF indicate that ~110 km and 60 km of shortening has been accommodated within Qaidam during and after the Eocene (Fig. 2), though the shortening magnitude decreases eastward (Guillot and Replumaz, 2013; Horton et al., 2002; Li et al., 2012; Meyer et al., 1998; Tapponnier et al., 1990; Yin et al., 2008). Considering the shortening in the main thrust belts and basins, we assume that the total
shortening accommodated within the Qaidam–Qilian Shan–Nan Shan region is 520 km (350 km in thrusts and 170 km in related basins).

Our shortening estimate within the northwestern Tibetan plateau is larger than in the Western Kunlun Shan (140–187 km) since the onset of the ATF (~30 Ma). Similarly, the total amount of shortening since ~23 Ma was 260–300 km (even 700 km) in northern Pamir (Burman, 2000, and their references therein; van Hinsbergen et al., 2011b). Although these estimates were made using structural, palaeomagnetic and lithological data analyzed separately rather than as a whole, the corresponding old suture and strata allow us to deduce an at least 100–150 km shortening discrepancy between the north Pamir and Western Kunlun Shan. The discrepancy among the Pamir, Western Kunlun Shan and northwestern Tibet can be attributed to the continental subduction below the Western Kunlun Shan (Gao et al., 2000). These reflectors can extend to a depth of ~100 km and crosscut the south-dipping Cenozoic subduction of the Tarim. The cross-cutting relationship shows that the north-dipping reflectors, younger than the south-dipping reflectors, should also be Late Cenozoic in age. This geometrical relationship indicates that the north-dipping thrust system has absorbed most of shortening of the ATF and KYF (Yin and Harrison, 2000).

Based on the above shortening analyses, the following can be inferred:

(1) The Lhasa and Qiangtang terranes were significantly shortened through folding and thrusting during the Late Cretaceous–Paleogene. The total shortening of the central Tibet was ~650 km prior to the India–Asia collision (van Hinsbergen et al., 2011b). Moreover, field investigation shows the central Tibet underwent minimal reactivation by thrusting since Late Paleogene (Kapp et al., 2003, 2005). All of these imply that central Tibet has undergone intensive shortening before the India–Asia collision.

(2) Since 55 Ma, approximately 1630 km (94°E) of shortening occurred between India and Eurasia (namely, 340 km within Lhasa and Qiangtang, 150 km in Songpan–Ganzi, 520 km in the northwestern Tibetan plateau, and 620 km within the Himalaya), divided into 1010 km within the Tibetan plateau and 620 km within the Himalaya. Our estimation about the Tibetan plateau is much more than the reconstructions of 600–750 km by van Hinsbergen et al. (2011b) and Johnson (2002). This is mainly because of the bigger shortening estimation from the Tanggula Shan and Qaidam–Qilian Shan–Nan Shan region.

(3) Our reconstruction provides direct implications for the size and subduction of Greater India. Plate circuits constrained by marine magnetic anomalies and fracture zone-based reconstructions of the Indian and Atlantic oceans suggest that the amount of India–Asia convergence since 55 Ma is 3200–4000 km (van Hinsbergen et al., 2011a, b). Combining 1630 km of intra-Asian shortening, the 250 km SE-ward extrusion (Fyhn et al., 2009; Hall et al., 2008; van Hinsbergen et al., 2011b) and 4000 km convergence (~94°E) allows us to deduce that most of the convergence was accommodated by the northward subduction of Greater India, which is probably the main cause of the crustal thickening and uplifting of central Tibet in the Cenozoic.

(4) Fold–thrust belts with relatively narrow zones have accommodated most of the N–S shortening since 55 Ma, which includes ~620 km within the Himalayan fold–thrust belt, 88–110 km within the Great Counter and the Gangdese thrust belts, 120 km within the Shiquanhe–Gaize–Amdo and southern Qiangtang thrust belts (van Hinsbergen et al., 2011b), 185 km within the Tanggula thrust belt, 60–80 km in the Fenghuo Shan–Nanqian thrust belt, 270 km within the Nan Shan thrust belt, and approximately 56 km within the Northern Qilian Shan thrust belt (Table 2). This suggests that at least 1400 km (of 1630 km) of shortening has been absorbed by the large-scale fold–thrust belts. All these estimates are minimal because they do not include penetrative strain or small-scale folds and faults, which could significantly increase the total shortening. Our results imply that the thrust belts played a crucial role in the crustal thickening of the plateau, and a small magnitude of convergence took place along the main strike-slip faults and extrusion.

5.2.2. Development of the thrusting and uplift of the Tibetan–Himalayan orogen

The above discussions show that the large-scale fold–thrust belts and crustal shortening display an outwards propagation form central Tibet. Moreover, paleoaltimetry shows that the central Tibet (Lhasa and Qantang) has obtained its current elevation probably as early as Eocene (Rowley and Currie, 2006; Wang et al., 2008a; Rohrmann et al., 2012). This was also confirmed by paleoenvironmental change in northeastern Tibetan plateau during 38–36 Ma (Dupont-Nivet et al., 2008; Hoorn et al., 2012). However, whether the growth of the Proto-plateau corresponds to the thrusting is an essential problem for understanding the growing mechanism of the plateau. Low-temperature thermochronology is an effective method by which to restrict the uplifting and exhumation history of the plateau. In this part, we will discuss the relationship between the deformation and exhumation histories of the plateau by thermochronological data.

5.2.2.1. Central Tibetan plateau.

In the central Tibetan plateau, 40Ar/39Ar biotite ages and K-feldspar total ages show that the duration of the exhumation is 68 Ma to 120 Ma (Fig. 6) and the rapid exhumation from the mid-crust began at ca. 100 Ma (Rohrmann et al., 2012). Apatite fission track (AFT) ages vary from 23 to 116 Ma with a concentration between 40 and 80 Ma (Figs. 6 and 7) (Hetzel et al., 2011; Rohrmann et al., 2012). Zircon (U–Th)/He (ZHe) ages range from 61 to 91 Ma and apatite (U–Th)/He (AHe) ages vary from 15 to 67 Ma with the majority varying from 34 to 67 Ma (Hetzel et al., 2011; Rohrmann et al., 2012). The younger AHe ages in central Tibet (19–10 Ma) were probably caused by faulting and river erosion because these samples were collected from the hanging walls of small thrusts and an externally draining river valley (Kapp et al., 2005). The above thermochronological dates indicate the exhumation of central Tibet mainly occurred between ~100 and 23 Ma and experienced two stages of intensive exhumation (~100–61 and ~52–40 Ma) (Figs. 5 and 6). Another interesting phenomenon in central Tibet is the cooling ages, which decreased gradually from the central Lhasa–Qiangtang (~100–40 Ma) to the northern margin of the Qiangtang (55–23 Ma) (Figs. 6 and 7). In the Tanggula Shan–Tuotuohe region, the AFT ages (30–55 Ma) from the hanging wall of the Tanggula Shan thrust belt (TTS) are synchronous with the sedimentation of the Tuotuohe foreland basin (Wang et al., 2008a; Li et al., 2012). Immediately north of the Tanggula Shan, rapid sediment accumulation in the Hoh Xil Basin was simultaneous with the exhumation (~40 Ma) of the hanging wall of the Fenghuo Shan thrust belt (Wang et al., 2008a). Moreover, the peaks of detrital zircon ages in Hoh Xil basin show quite similar characteristic features with the Qiangtang and Lhasa terranes, indicating that they were derived from the central Tibetan plateau (Dai et al., 2012). All above results imply the exhumation would be northward and invoked by the fold–thrust belts.

5.2.2.2. Northeastern Tibetan plateau. Feldspar 40Ar/39Ar multi-domain diffusion modeling shows that the exhumation of the central Kunlun Shan (near Golmud) occurred between 41 and 21 Ma (Mock et al.,
Apatite helium age/depth transects from the east Kunlun range and South Qaidam Basin indicate initial cooling events in the hanging wall of the South Qaidam thrust belt (SQT) at 35 Ma (Clark et al., 2010). This is consistent with the rapid sedimentation in southern Qaidam basin. Moreover, the sedimentary and deformation records indicate that the Hoh Xil Basin and the Qaidam Basin were two parts of a single basin (Yin et al., 2008), which is in agreement with the Neogene related basins. Sedimentation rates increased in southern Hexi Corridor Basin at about 8–7 Ma, and with a more rapid increase at about ~3.66 Ma. These observations indicate that the northern margin of Qilian Shan began to uplift since 8–6.6 Ma (Fang et al., 2005). The tectonic position, exhumation history, and structure of the Hexi Corridor foreland basin provide strong evidence that the exhumation was induced by thrusting of the Northern Qilian Shan thrust belt, and the tectonic uplift reached northern Qilian at about ~10 Ma due to the northward thrusting.

5.2.2.3. Northwestern Tibetan plateau. In the Western Kunlun Shan, the AFT age from the hanging wall of the West Kunlun thrust (WKT) shows that it experienced rapid cooling during 20–18 Ma, 12–8 Ma, and 5–4.8 Ma (Fig. 6) (Wang et al., 2003). These rapid exhumation stages are consistent with the activity of the West Kunlun thrust and sharply increased sedimentation rates at the southern margin of the Tarim Basin (Zheng et al., 2000; Wang et al., 2003; Sun et al., 2008). This indicates that the uplift and exhumation of the western Kunlun Shan was caused by the folding and thrusting of the West Kunlun thrust belt.

In the north Pamir, the 2He ages show that exhumation mainly during 35–38 Ma, 12–20 Ma, and 10–2 Ma, and the AFT ages are concentrated on 4–2 Ma (Sobel and Dumitru, 1997; Sobel et al., 2011). Based
on $^{40}$Ar/$^{39}$Ar thermochronology from the western Pamir, Lukens et al. (2012) suggested that the deep, extensive exhumation at ~13–21 Ma is consistent with a higher magnitude of shortening in the Pamir, which also resulted in the upper crustal deformation and the rapid removal of crustal material from the hinterland. Moreover, slip along the Kashgar–Yecheng fault (KYF) accommodated relative motion between Tarim and the Pamir. New thermochronologic data obtained along the Tashkurgan–Yarkand River gorge indicates that the magnitude of exhumation decreases to the east across the eastern flank of the Pamir, and slow dextral slip along the easternmost strand of the KYF since 5–3 Ma (Sobel et al., 2011). This deceleration was attributed to the substantial increase of northward motion of Tarim rather than a significant decrease of the northward velocity of the Pamir (Sobel et al., 2011). These observations coincide with the northward expansion of the thrusting and onset of the Mazar Tagh thrust belt (~3 Ma). The above facts demonstrate the exhumation and uplift dominated by thrusting and shortening in Western Kunlun Shan and Pamir. Moreover, based on the present topographic and tectonic relationships, the Mazar Tagh thrust belt (MT) can be expected to uplift and mark the northwestern boundary of the plateau in the future (Wang et al., 2014a).

5.2.2.4. Himalaya. Numerous thermochronological data have been obtained to define the uplift and denudation of the Himalaya, including apatite and zircon fission tracks, $^{40}$Ar/$^{39}$Ar ages, and (U–Th)/He ages (Fig. 6). A notable fact is that the thermochronological ages vary significantly in space, with most of them younger than 23 Ma (Figs. 5 and 6). Based on 255 published apatite and zircon fission tracks and white mica $^{40}$Ar/$^{39}$Ar ages, Thiede et al. (2009) quantified the rates and exhumation along the strike of the Lesser, High, and Tethyan Himalaya. In the Tethyan Himalaya, the exhumation range is from 33 to 5 Ma, with rapid denudation at 17–5.7 Ma in the central Himalaya (Liu et al., 2005) and 15–3 Ma in the southwestern Himalaya (Sorkhabi et al., 1996; Sobel and Dumitr, 1997; Thiede et al., 2005). The exhumation rates for the northern High Himalaya (hanging wall of the Main Central thrust) were high (2–3 mm/a) during 23–19 and 3–0 Ma and low (0.5–0.7 mm/a) during 19–3 Ma. In the footwall of the Main Central thrust, however, high exhumation rates of 2–3 mm/a existed since 11 Ma (Thiede et al., 2009). The initial denudation of the Lesser Himalaya began at 10 Ma and rapid denudation at ~5 Ma (Patel et al., 2007; Herman et al., 2010; Nadin and Martin, 2012).

McQuarrie et al. (2014) linked exhumational variability in space and time to the evolving geometry of the Himalayan fold–thrust belt in western Bhutan. By combining deformation and thermochronologic data, they concluded that structural and chronologic data exhibit a general north to south progression of Himalayan deformation, the strongest control on exhumation magnitude and variability is fold–thrust belt geometry. Most notably, the rapid exhumation in the hanging and footwall of the Main Central thrust is in accordance with the initiation of this thrust belt and the Main Boundary thrust (Figs. 4, 6). Moreover, the uplift history of the Himalaya has been documented in the NW India–Nepal foreland basin and Siwalik foreland basin. The Oligocene unconformity and sharp increase in sedimentation rate from the Miocene to Pliocene illustrate that intense uplift and denudation of the High Himalaya began at 23 Ma (Najman, 2006; Yin, 2006). All these observations show the Himalaya underwent a southward denudation from the Tethyan Himalaya to the Lesser Himalaya, which is also consistent with the deformation history of major thrusts and formation of the foreland basin (Fig. 7). For example, the initiation of denudation in the Lesser Himalaya is coeval with the onset of the Main Boundary thrust and foreland basin, which indicates that the thrusting results in both uplift and sedimentation.

5.2.2.5. Eastern Tibetan plateau. Different thermochronological methods have been implemented to document the uplift and mechanism of eastern Tibet, but the processes and dynamics are still in dispute. Based on $^{40}$Ar/$^{39}$Ar and (U–Th)/He data, Kirby et al. (2002) emphasized that Neogene denudation prevailed in the Longmen Shan. However, because Paleogene–Neogene cooling ages exist in the eastern Songpan–Ganzi and the western Longmen Shan (Fig. 6), most thermochronology and thermal models concluded that the Longmen Shan experienced several exhumation stages during the Late Cenozoic (Xu and Kamp, 2000; Kirby et al., 2002; Clark et al., 2005; Godard et al., 2009; Wilson and Fowler, 2011; Wang et al., 2012). For example, Xu and Kamp (2000) emphasized rapid denudation at circa 22, 7 and 2 Ma. Wang et al. (2012) proposed two pulses of rapid exhumation in the hanging wall of the Longmen Shan thrust (30–25 and 15–0 Ma). Considering the local structure and deep incision of large rivers, more and more researchers emphasized that the rapid exhumation and high topography of the Longmen Shan occurred during 10–12 Ma (Xu and Kamp, 2000; Kirby et al., 2002; Clark et al., 2005; Wang and Meng, 2009; Oskin, 2012; Wang et al., 2012). Moreover, the oldest alluvial deposits within the Chengdu Plain (~12–8 Ma) indicate that the Longmen Shan was established since the Late Miocene and caused by the Longmen Shan thrust belt (LMST) (Kirby et al., 2002; Clark et al., 2005; Zhang
et al., 2008; Wang and Meng, 2009; Wang et al., 2012). The significant difference in the cooling ages between the hanging wall and footwall of the Longmen Shan thrust and the balanced geologic cross-sections suggest that the thrusting is a primary driver for the uplift and topography of the Longmen Shan (Hubbard and Shaw, 2009).

In the western part of eastern Tibet, the AFT ages and thermal history from the hanging wall of the Ganzi–Litang thrust belt (GLT) show that this area has undergone rapid exhumation during 20–16 Ma (Lai et al., 2006) prior to the Longmen Shan. Moreover, the deformation history of the thrusting belts reveals an eastward migration from the Ganzi–Litang thrust belt to Longmen Shan thrust belt (Figs. 2 and 5) (Lai et al., 2006; Zhang et al., 2008). Considering the exhumation and deformation, we infer that the Ganzi–Litang thrust belt resulted in the rapid uplift of the Yudon block during 20–16 Ma and marked the Miocene margin of the plateau. With the stress migrating eastward, the thrusting reached the Daxue Shan and Longmen Shan during 20–12 Ma (Fig. 5A).

Under the control of the progressively eastward thrusting, the Longmen Shan underwent rapid uplift and exhumation since 12–10 Ma (Fig. 5B). This means that thrusting and its eastward migration played a crucial role in the uplift of eastern Tibet.

In summary, the folding–thrusting and thermochronological data indicate that the outward thrusting is coeval with the expanding exhumation (Fig. 7), which implies that the topographic growth of the plateau would be outward and invoked by thrusting and related upper crustal shortening.

5.3. Growth dynamics of the Tibetan–Himalayan orogen

5.3.1. Genetic mechanism of the Proto-plateau

The above analyses indicate that the thrusting, crustal shortening, and exhumation have undergone an outward expansion from central Tibet since the Late Mesozoic. The paleoaltimetry and thermochronological data indicate that the Lhasa and Qiangtang terranes gained their elevations at ~40 Ma (Proto-plateau) (Rowley and Currie, 2006; Dupont-Nivet et al., 2008; Hoorn et al., 2012; Rohrmann et al., 2012; Ding et al., 2014; Wang et al., 2014a). However, the shortening and uplifting mechanisms of the Proto-plateau are ambiguous, presenting an obstacle for understanding the growth of the entire plateau.

Previous studies have illustrated that the Lhasa and Qiangtang terranes were significantly shortened through folding and thrusting during the Late Cretaceous–Paleogene (Murphy et al., 1997; Kapp et al., 2003, 2005; DeCelles et al., 2007; Volkmer et al., 2007; van Hinsbergen et al., 2011b). Shortening reconstruction indicates that the central Tibet accommodated ~650 km of shortening (~50%) prior to the India–Asia collision, which is much larger than the post-collision shortening (~340 km). This significant crustal shortening and uplift have been attributed to the Lhasa–Qiangtang collision and probably thickened the crust to ~45–55 km and generated 3–4 km of elevation (Murphy et al., 1997; DeCelles et al., 2002; Kapp et al., 2003, 2005; Volkmer et al., 2007). Murphy et al. (1997) proposed that the southern Tibetan plateau had attained 3–4 km elevation at ~99 Ma by Lhasa–Qiantang collision. However, new geological and paleontological evidence reveals that the Bangong–Nujiang ocean did not close until the Late Cretaceous (~90 Ma) (Zhang et al., 2004, 2012, 2014). This indicates that the Qiangtang–Lhasa collision could have contributed to the growth of the Tibetan plateau since 90 Ma. Moreover, the 90–75 Ma high-K and adakitic magmatism in northern Lhasa and southern Qiangtang verify that the central Tibet had undergone crustal thickening and subsequent delamination during Late Cretaceous (Li et al., 2013; Wang et al., 2014b). This uplift was coincident with the first stage (~100–61 Ma) of accelerated exhumation in central Tibet (Hetzel et al., 2011; Rohrmann et al., 2012; Wang et al., 2014b). If this is correct, it implies that the Lhasa–Qiangtang collision alone aroused the uplift of central Tibet before the India–Asia collision. Nevertheless, the geological and magmatic evidence reveals that a high elevation Andean-type arc occurred along the Gangdese during Late Cretaceous–Paleogene as a result of the flat slab subduction of the Neo-Tethys ocean and related magmatism (Wen et al., 2008; Zhang et al., 2007, 2012, and their references therein). This suggests that the Neo-Tethys oceanic slab contributed to the crustal shortening and uplift of the Gangdese mountains. Reconstruction of the latitude history of the Lhasa terrane suggests that Tibetan plateau is similar to the Altiplano of the Andes in that most of the plateau developed at subtropical latitudes above an oceanic subduction zone prior to India–Asia collision (Lippert et al., 2014). This viewpoint is also supported by palaeogeomorphic results from the Linzhou Basin (Ding et al., 2014). Therefore, we support that the Qiangtang–Lhasa collision in combination with the ongoing subduction of the Neo-Tethys Ocean led to the earliest uplift of central Tibet (Zhang et al., 2012); thus, we term the high elevation prior to the India–Asia collision as the Primitive plateau.

The second stage of accelerated exhumation (~52–40 Ma) in central Tibet lags shortly behind the onset of the India–Asia collision and corresponds to the Eocene–Oligocene folding and thrusting in central Tibet (Table 2). This exhumation is generally attributed to the India–Asia collision or the northward subduction of Greater India (DeCelles et al., 2002; Rowley and Currie, 2006; Li et al., 2012; Meng et al., 2012). Combining our shortened reconstruction with the geophysical data allows us to conclude that the subduction of Greater India probably played a key role in crustal thickening and uplifting. The main evidence is as follows: 1) central Tibet has undergone minor shortening (~160 km, except northern Qiangtang) since the India–Asia collision (Table 2); 2) the southern margin of Asia remained almost fixed from ~50 to 34 Ma (Meng et al., 2012); 3) the size of Greater India is more than ~3000 km including the minor eastward extrusion (250 km) (van Hinsbergen et al., 2011b); and 4) mantle tomographic and seismic studies suggest that the Greater Indian slab extended approximately half-way across the Tibetan plateau, resulting in lower crustal thickening and uplift of Tibet (DeCelles et al., 2002; Replumaz et al., 2010; Zhao et al., 2011, and references therein).

In addition, if the Proto-plateau was created by the India–Asia collision and subsequent convergence, a long, narrow, high orogen with high range geomorphy would have been formed. However, the topography of the Proto-plateau with a width of ~600 km and an elevation of ~5000 m (Tapponnier et al., 2001; Wang et al., 2008a, 2014a) is quite different from a typical orogen. Moreover, the intensive crustal shortening (~185 km) in the northern margin of the Qiangtang terrane during the Eocene cannot be attributed to the India–Asia collision because the stress could not have reached this region soon after the initial collision between India and Asia (Yin and Harrison, 2000). Therefore, another mechanism must be taken into consideration to explain the abnormal geomorphic and deformational styles of the Proto-plateau. In the northern Qiangtang, widespread potassium-rich lavas and subordinate sodium-rich basalts originating from the melting of the enriched lithospheric mantle domains were generated from ~50 to 35 Ma. The geochemical composition reflects the melting of the lithospheric mantle contaminated by melts derived from a metasedimentary crust (Ding et al., 2007). The tectonic location and petrogenesis reflect magmatism caused by the southward subduction of the Songpan–Ganzi terrane (Roger et al., 2000; Tapponnier et al., 2001; Ding et al., 2007; Wang et al., 2008b). Furthermore, seismic tomography has identified a high-wavespeed anomaly (1100–900 km in depth) in the north of the Tethyan anomaly that is interpreted as a remnant of the Songpan–Ganzi slab (Guillot and Replumaz, 2013; Hafenschield et al., 2006; Replumaz et al., 2010). All geophysical and geologic observations indicate that the Songpan–Ganzi terrane was initially subducted beneath the Qiangtang terrane approximately 50–45 Ma (Meyer et al., 1998; Guillot and Replumaz, 2013).

Therefore, we suggest that formation of the Proto-plateau was primarily due to the northward and southward subductions of the Greater India and Songpan–Ganzi terranes, respectively, beneath the Lhasa–Qiangtang terrane. The northern margin of the Proto-plateau was
located in the Tanggula Shan range during the Eocene. Our conclusion agrees with the paleogeomorphic result indicating that the Hoh Xil had an elevation of 2000 m during the Late Eocene (Cyr et al., 2005).

5.3.2. Dynamic mechanism for the growth of the Tibetan–Himalayan orogen

To date, no single geophysical or geologic model can accurately explain the formation of the Tibetan–Himalayan orogen. A comprehensive understanding of the growth of the plateau must take different observations into consideration. In the following discussion, we highlight the importance of the crustal rheological behaviors, topographic gradient of the Proto-plateau, and continental subduction for the growth of the Proto-plateau.

The above analyses show that the Proto-plateau formed –40 Ma with an average elevation of ~5000 m and a thickened crust of 50–60 km (Tapponnier et al., 2001; Rowley and Currie, 2006; Wang et al., 2008a, 2014a). The viscosity-depth profile of the crust plays an important role in continental deformation and is a dominant factor in controlling the mode of crustal deformation (Ranalli and Murphy, 1987; Royden, 1996; Watts and Burov, 2003). Three-dimensional viscous Newtonian models of the Tibetan plateau suggest that the thickened crust led to the transformation of rheological behaviors and produced a moderately weak or extremely weak lower crust in Tibet (Royden et al., 1997; Shen et al., 2001). Rey et al. (2010) proposed that an elevated plateau (~5000 m) required an initial Moho temperature of approximately 560 °C. Thus, upper crustal deformation is strongly decoupled from that of the lower crust and mantle, and the partitioning of strain observed at the surface may not extend to the lower crust. Although motion in the lower crust (channel flow) is not essential for plateau development, it indicates that the deformations between the lower and upper crusts are essentially discontinuous (Shen et al., 2001). The above numerical simulations implicitly suggest that the coexistence of the upper crustal shortening and lower crustal flow was the primary mechanism for the deformation and growth of the plateau.

As mentioned above, when the Proto-plateau gained its present elevation ~40 Ma (Rowley and Currie, 2006; Wang et al., 2008a; Rohrmann et al., 2012; Ding et al., 2014), its northern and southern areas were still low (Wang et al., 2008a). Simulations have shown a plateau at high elevation with steep topographic gradient across its margins, corresponding to a large lateral pressure aroused by the gravitational spreading of the high plateau. If the topographic gradient is averaged over distances longer than the flexural length scale of the lithosphere, it will exert a first-order control on crustal deformation (Shen et al., 2001); this has been illustrated for several actively deforming regions including the Himalayan region, and it was concluded that the modern topography of the Himalayans is capable of explaining most of the active deformation (Holt et al., 1991; Holt and Haines, 1993; England and Molnar, 1997; Holt, 2000). However, correlation among the topographic gradient, deformation, and growth of the Proto-plateau was not discussed in previous studies.

Geologic data and reflection seismic profiling also show clear evidence for the subduction of the lithospheric mantle of North and South China beneath the plateau (Meyer et al., 1998; Tapponnier et al., 2001; DeCelles et al., 2002; Zhao et al., 2011; Guillot and Replumaz, 2013; Guo et al., 2013). Several models emphasize the relative importance of continental subduction in the thickening and growth processes of Tibet (Tapponnier et al., 2001; DeCelles et al., 2002; Guillot and Replumaz, 2013). Most of these models consider that the post collisional convergence between India and Asia is accommodated by thrusting and folding within the continental subduction belts. Meyer et al. (1998) and Tapponnier et al. (2001) considered that the sequential south-directed subduction of Songpan–Ganzi and Qaidam led to the crustal shortening and northeast growth of the plateau and that the northern Qilian Shan may become an upcoming continental subduction. This consistent with our results suggesting that active deformation in this area is largely a result of north–south convergence, as well as with recent GPS results (King et al., 1997; Chen et al., 2000). All of these observations indicate that continental subduction has contributed to the crustal shortening, thickening and growth of the plateau, although the causation and function of the subduction remain unclear.

Based on the above analyses, we propose that three main factors were involved in the Cenozoic crustal deformation and rapid growth of the Proto-Tibetan plateau; our interpreted sequence of plateau growth is as follows:

1. As the occurrence of the Proto-plateau was about ~40 Ma, an extremely weak lower crust formed as a result of the transformation of rheological behaviors, while the upper crust was relatively cold and consequently rigid. The deformation is controlled by the folding and thrusting at the upper crust, while, the lower weak crust displays ductile shear, thickening and flow. Meanwhile, due to the steep topographic gradient and related lateral pressure, a stress front was produced along the margins of the Proto-plateau and exhibited a spreading tendency toward the lowland (Fig. 8A).

2. With ongoing convergence between the Indian and Asian continents, the stress front spread rapidly to the lowland via thrusting and folding, giving rise to upper crustal shortening and surface uplift. Dominated by the shortened upper crust and the outward migration of the Proto-plateau’s lower crust, the crust of the lowland was thickened and eventually uplifted. Thus, a new topographic gradient was produced in front of this area (Fig. 8B). Because of the decoupling, the crustal shortening and thickening between the upper and lower crusts may be inconsistency during this stage.

3. If the upper crust deformation and growth of the Proto-plateau were provoked by the topographic gradient, the altitude would gradually decrease from central Tibet. However, when the stress front and flowing lower crust met a tectonic boundary (old suture or edges of a different terrane), the propagation would be impeded in this area due to the strong resistance from another terrain. The tremendous accumulation of upper crustal shortening and upwelling lower crust would cause the topographic bulge (Fig. 8C). Increasing amounts of crustal shortening and topographic bulge in this area would have forced the subduction of the lowland mantle lithosphere (Fig. 8D). Subsequently, the intense convergence between two continents would result in the elevation reaching a critical level such as the Longmen Shan.

The above model can properly provide an explanation for the following questions. The crustal shortening (Tapponnier et al., 2001; Hubbard and Shaw, 2009; Robert et al., 2010) and lower-crustal flow (Clark and Royden, 2000; Burchfiel et al., 2008; Oskin, 2012; Wang et al., 2012) are long-standing questions on uplift of the Longmen Shan and the eastern Tibet. The channel flow requires no horizontal shortening at the surface, because the upper crust is largely uplifted by the lower-crustal motion, but not laterally displaced ( Hubbard and Shaw, 2009; Yin, 2010). However, our studies show that the thrusting and upper crustal shortening experienced an eastward migration along eastern Tibet; the crust was probably shortened by ~100 km during the Mio-Pliocene ( Royden et al., 1997). Moreover, balanced geologic cross-sections suggest that crustal shortening is a primary driver for uplift and topography in the Longmen Shan on the flanks of the plateau ( Hubbard and Shaw, 2009). If the Longmen Shan is produced by thrusting and crustal shortening, it cannot reasonably explain the limited shortening rate (~3 mm yr⁻¹) and lack of typical foreland basin in front of the fold-and-thrust belt ( Chen et al., 2000; Hubbard and Shaw, 2009). Moreover, the recent deep seismic reflection profile across eastern Tibet reveals that crustal-scale deformation participated in the eastward extrusion and uplift of the Longmen Shan ( Guo et al., 2013); this implies that neither channel flow nor upper crustal shortening can alone explain the
uptilt of the eastern plateau. On the basis of our model, with the eastward expansion of the Proto-plateau, the upper crust exhibits thrusting and shortening via topographic gradient and related lateral pressure, while the weak lower crust was locally soft, favoring homogeneous thickening and crustal flow. The resistance of the westward subduction of South China led to intensive crustal shortening and upwelling of the lower crust along with rapid surficial uplifting (Fig. 5). Moreover, because of the inconsistency and discontinuity between the upper and lower crusts, multiple exhumation stages were produced in front of the Longmen Shan (Xu and Kamp, 2000; Kirby et al., 2002; Clark et al., 2005; Wang et al., 2012).

In addition, the massive accumulation of stress along the major tectonic boundaries indicates that the upper crustal shortening distributes mainly along large thrust belts, which is in accordance with our estimation that most of the crustal shortening was accommodated by large thrust belts. Moreover, reconstructions of the relative positions of the Indian and Asian plates show that India's convergence rate with Eurasia significantly decreased during 23–20 and 11–10 Ma (Molnar and Stock, 2009; Copley et al., 2010), although the cause is still ambiguous. It is notable that the slowing convergence rate is coincident with the intensive activities of large thrust belts in Himalaya, Northern Qilian and eastern Tibet (Fig. 2) such as the Main Central thrust belt (~23 Ma), Main Boundary thrust belt (~11–10 Ma), Northern Qilian Shan (~9 Ma), and Longmen Shan thrust belts (12–10 Ma). Thus, it can be inferred that the impeded propagation and tremendous accumulation of shortening were the dominant cause of the slowing convergence rate.

5.3.3. Extensional structures and growth of the plateau

As discussed above, various models have been proposed to explain the causes of the N–S trending rifts, but the mechanism is still in dispute. A reasonable model must take all factors into consideration, including whether the rifts in the central Tibet are linked to the Himalaya or represent different, possibly coeval processes that generate similar structures and geomorphic features.

In Himalaya, age constraints show that the Main Central thrust belt (MCT) and the Southern Tibetan Detachment (STDS) mainly developed between 23 and 12 Ma (Tables 1 and 3). Previous research indicates that the onset of S–N rifting occurred 13–9 Ma (Edwards and Harrison, 1997; Murphy et al., 2002; Stockli et al., 2002; Wu et al., 2005), significantly before the STDS. Nevertheless, the younger ages of the STDS in the Kula (~12 Ma) (Edwards and Harrison, 1997) and Zanskar (~15–12 Ma) (Godin et al., 2001) imply that the two extensional structures are contemporaneous (Table 3). This relationship is reconfirmed by the onset of E–W extension in Kung Co rift (19 Ma) (Mitsuishi et al., 2012). This indicates that the two extensions are interrelated and likely represent different manifestations of the same tectonic events. Considering the above factors and the tectonic regime during the Miocene, we infer that the N–S shortening and the progressive contraction induced the occurrence of the E–W extension, while the STDS results from the passive roof thrust in the hanging wall of the Main Central thrust (Fig. 9). In addition, deformation and thermochronology indicate that the two extensions are coincident with the rapid exhumation of the High Himalayas (Thiede et al., 2009). Three-dimensional finite element modeling suggests that significant crustal extension occurred only when the plateau had reached ~75% of its present elevation (Liu and Yang, 2003). Based on the above observations, the two extensions result from crustal shortening and represent the southward growth of the plateau.

The E–W extension in central Tibet was quite different from similar tectonics in Himalaya. First, the onset of the rifts (~18–13 Ma) in central Tibet significantly lagged behind the N–S shortening (~23 Ma) (Tables 1 and 3) and the occurrence of the Proto-plateau. This indicates that central Tibet reached its present elevation prior to the E–W extension (Rowley and Currie, 2006; Dupont-Nivet et al., 2008; Wang et al., 2008a). Simulations show that the eastward movement of the upper and lower crustal shortenings occurred in the eastern plateau significantly after 20 Ma (Shen et al., 2001), coeval with the onset of the normal faults. Deformational ages also reveal that the timing of the initial E–W extension in central Tibet (~18–13 Ma) agrees with the onset of thrusting (20–16 Ma) in eastern Tibet (Lai et al., 2006; Zhang et al., 2008). Moreover, the distance of the extension (~100 km) in central

Fig. 8. Tectonic mechanism to account for the growth of the Proto-plateau. (A) With the formation of the Proto-plateau, a extremely weak lower crust, topographic gradient and related lateral pressure produced along the margins of the Proto-plateau and exhibited a spreading tendency toward lowland. (B) Dominated by the deformations in upper and lower crust, together with outward flow of the Proto-plateau’s lower crust, the lowland crust was thickened and eventual uplift. A new topographic gradient was produced. (C) Impeded by tectonic boundary (old suture), the upper crustal shortening and lower crust flow cause the topographic bulge. (D) Increasing of crustal shortening and topographic bulge forced the subduction of the lowland mantle lithosphere.
Tibet (van Hinsbergen et al., 2011b) is identical to that of the Miocene crustal shortening (~100 km) in eastern Tibet (Royden et al., 1997). Taking the above observations into consideration, we can conclude the following. Influenced by the topographic gradient of the Proto-plateau, the materials of Proto-plateau moved eastward by thrusting and lower crust flow, which caused E–W extension in central Tibet. Due to the velocity gradient within the upper crust, the N–S trending rifts in central Tibet are characterized by half grabens (Fig. 9).

5.4. Integrated model for the growth of the Tibetan–Himalayan orogen

Based on our reconstruction of deformation and regional evolution, we summarize the development of the Tibetan–Himalayan orogen as follows:

1. Occurrence and development of the Primitive plateau (~90–55 Ma). During the Late Cretaceous–Palaeocene (Fig. 10A), with the Lhasa–Qiangtang collision and northward subduction of the Neo-Tethys oceanic lithosphere, the crust of central Tibet was tectonically shortening and thickened and thrusting (Wen et al., 2008; Zhang et al., 2007, 2012). Although the elevation is ill-defined, deformation and volcanism in central Tibet indicate that central Tibet underwent crustal thickening during this period, resulting in the thickening and uplift of the Lhasa–Qiangtang crust to ~45–55 km and 3–4 km, respectively (Murphy et al., 1997; DeCelles et al., 2002; Kapp et al., 2003, 2005; Volkmer et al., 2007; Wang et al., 2008b, 2014b; Li et al., 2013). Meanwhile, plate circuit reconstruction shows that there was a wide Neo-Tethys ocean between India and Asia (Fig. 10A).

2. Formation of the Proto-plateau (55–40 Ma). With the India–Asia collision and subsequent subductions of Greater India, the crust of central Tibet was intensively shortened and thickened. In Tethyan Himalaya, the development of the Tethyan–Himalayan fold thrust belt marks the crustal shortening of northern Greater India (Fig. 10B) (Ratschbacher et al., 1994; DeCelles et al., 2002; Wiesmaery and Grasemann, 2002). Meanwhile, the initiation of the Tanggula Shan thrust belt, exhumation, and potassium-rich lavas demonstrate the onset of underthrusting in the Songpan–Ganzi terrane (Tapponnier et al., 2001; Wang et al., 2002a, 2008a; Li et al., 2012; Guilbot and Replumaz, 2013) (Fig. 10B). The India–Asia collision and subductions of Greater India and Songpan–Ganje within the Lhasa–Qiangtang terrane resulted in the eventual uplift of the Proto-plateau (5000 m) at approximately 40 Ma (Rowley and Currie, 2006; Wang et al., 2008a, 2014a).

3. Rapid growth of the Proto-plateau and formation of the Neotectonic plateau (40 Ma–present). Dominated by the topographic gradient and continuous convergence between India and Asia, the Proto-plateau experienced an outward propagation by crustal shortening and lower crust flow. Three significant growth gradations can be discriminated during this period:

During 40–23 Ma, the northward thrusting and folding led to the uplift of Kunlun Shan at 35–30 Ma and Qilian Shan at 33–24 Ma in northern Tibet. Meanwhile, the Main Pamir thrust belt (MPT) and the West Kunlun thrust belt (WKT) resulted in the uplift of northern Pamir and western Kunlun Shan (Sobel and Dumitru, 1997; Sobel et al., 2011). In the Himalaya, shortening concentrated in the Indus–Yarlung Zangpo suture and Tethyan Himalaya (Fig. 10C). The 33–24 and ~23 Ma exhumations in Tethyan Himalaya and northern High Himalaya imply a southward expansion and tectonic uplift during this period (Copeland et al., 1995; Liu et al., 2005). Simultaneously, the deformation and cooling ages imply that the eastward expansion had reached the west Ganzi–Litang in eastern Tibet (Lai et al., 2006; Zhang et al., 2008). Altogether, at about ~23 Ma, the boundary of the plateau may have reached Nan Shan and western Kunlun in the north, Ganzi–Litang in the east, and the Tethyan–Himalayan area in the south (Fig. 10C).

From 23 to 10 Ma, with convergence between the plateau and the surrounding terranes, the shortening was concentrated mainly in the margins of the pre-existing plateau and led to the continuous growth of the plateau (Fig. 10D). In eastern Tibet, the topographic gradient and lower crust flow not only gave rise to the eastward moving of the plateau but also aroused the generation of the N–S trending rifts in central Tibet. The thrusting and folding reached the Daxueshan–Gonggashan and lead to the tectonic uplift of this region. In Himalaya, the Main Central thrust belt (MCT) played a key role during this period. It not only accommodated most of the N–S shortening but also induced the development of the E–W and N–S extensions. Rapid exhumation rates in the High Himalaya and subsidence in the India–Nepal foreland basin indicate that the High Himalaya was completely uplifted during this period (Yin, 2006). Meanwhile, simultaneous deformation and sedimentation in western Kunlun and the Hexi Corridor Basin demonstrate that the northern boundary of the plateau reached northern Qilian and southern Tarim, respectively (Fig. 10D).

Since 10 Ma, the crustal shortening and tectonic uplift was concentrated on the edges of the orogen including Longmen Shan, Lesser Himalaya and northern Qilian Shan (Fig. 10E). The ages of thrust belts and sediments in related basins show that these regions experienced rapid exhumation since the Late Miocene (~10 Ma). The active thrust belts around the orogen mark the ongoing growth of the plateau. Combining our model with the surficial velocity vectors of recent GPS, we deduce that the plateau will be larger in the future and that the margins of the plateau will reach Shillong–Madhipura in the south, Longquan Shan (east of Chengdu) in the east, Mazatage in the northwest, and Heishan in the northeast in the upcoming 10 Ma (Fig. 10 F).
Our deformation restoration suggests that at least ~1630 km of shortening (along 94°E) has taken place between India and Asia since ~55 Ma, with ~1010 km and ~620 km of this accommodated by Asia and Himalaya, respectively. More than ~1400 km (of 1630 km) of shortening was absorbed by large-scale fold–thrust belts that have undergone an outward propagation from central Tibet, leading to upper crustal shortening and surface uplift of the plateau. The formation of the Himalayan orogen and Tibetan plateau experienced three significant processes: the Primitive plateau, Proto-plateau, and Neoteric plateau. The Primitive plateau occurred in central Tibet during the Late Cretaceous-Paleocene (~90–55 Ma) as a result of the Lhasa–Qiangtang collision and northward subduction of the Neo-Tethys oceanic
lithosphere. During 55–40 Ma, the India–Asia collision and the subduction of Greater India and Songpan–Ganzi terranes beneath the Hhasa–Qiangtang terrane resulted in the eventual uplift of the Proto-plateau (at an elevation of 5000 m). Since ~40 Ma, under the control of the topographic gradient and continuous convergence between India and Asia, the Proto-plateau underwent an outward propagation due to upper crustal shortening and lower crustal flow that produced the Neoantic plateau. The active thrust belts and related basins around the Tibetan–Himalayan orogen represent the ongoing growth of the orogen, and the upcoming orogen is larger than present. The N–S trending ridges were caused by the eastward growth of the plateau dominated by thrusting and crust flow in central Tibet, while they were the results of intense N–S shortening in Himalaya.

Acknowledgments

We thank Editor Prof. Gillian R. Foulger and three referees (Zhang K.J., van Hinsbergen and Douwe) for their constructive and helpful comments which greatly helped in improving our manuscript. We thank Zhongpeng Han, Ming Xu, Jun Meng, Aorigje Zhout, Guanglin Cai, Xi Chen, Licheng Wang, Yushuai Wei and Lidong Zhu, who participated in many field expeditions in the Himalaya and Tibet; these studies provided the fundamental basis for this paper. We also thank Alan R. Carroll and Michael L. Wells for comments and improvement of the manuscript. The research was financially supported by the National Natural Science Foundation of China (41172129 and 40672086), the National Key Project based on U–Th/He thermochronometry and terrestrial cosmogenic nuclide methods. Geomorphology 107 (3–4), 285–299.


