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Magmatism during continental collision, subduction, exhumation and mountain collapse in collisional orogenic belts and continental net growth: A perspective

SONG ShuGuang[1*], WANG MengJue[1], WANG Cao[1] & NIU YaoLing[2,3]

[1] *MOE Key Laboratory of Orogenic Belt and Crustal Evolution, School of Earth and Space Sciences, Peking University, Beijing 100871, China;*
[2] *Department of Earth Sciences, Durham University, Durham DH1 3LE, UK;*
[3] *Institute of Oceanology, Chinese Academy of Sciences, Qingdao 266071, China*

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Continental orogens on Earth can be classified into accretionary orogen and collisional orogen. Magmatism in orogens occurs in every periods of an orogenic cycle, from oceanic subduction, continental collision to orogenic collapse. Continental collision requires the existence of prior oceanic subduction zone. It is generally assumed that the prerequisite of continental deep subduction is oceanic subduction and its drag force to the connecting passive-margin continental lithosphere during continental collision. Continental subduction and collision lead to the thickening and uplift of crust, but the formation time of the related magmatism in orogens depends on the heating mechanism of lithosphere. The accretionary orogens, on the other hand, have no strong continental collision, deep subduction, no large scale of crustal thrusting, thickening and uplift, and no UHP eclogite-facies metamorphic rocks related to continental deep subduction. Even though arc crust could be significantly thickened during oceanic subduction, it is still doubtful that syn- or post-collisional magmatism would be generated. In collisional orogens, due to continental deep subduction and significant crustal thickening, the UHP metamorphosed oceanic and continental crusts will experience decompression melting during exhumation, generating syn-collisional magmatism. During the orogen unrooting and collapse, post-collisional magmatism develops in response to lithosphere extension and upwelling of asthenospheric mantle, marking the end of an orogenic cycle. Therefore, magmatism in orogens can occur during the continental deep subduction, exhumation and uplift after detachment of subducted oceanic crust from continental crust, and extensional collapse. The time span from continental collision to collapse and erosion of orogens (the end of orogenic cycle) is 50–85 Myr. Collisional orogens are the key sites for understanding continental deep subduction, exhumation, uplift and orogenic collapse. Magmatism in collisional orogens plays important roles in continental reworking and net growth.

collisional orogeny, continental deep subduction, orogen unrooting and collapse, magmatism, continental net growth

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Continental orogens in the different periods of Earth's history record complicated tectonic evolution and crustal growth, which involves a complete Wilson-cycle from continental rift, ocean spreading, oceanic subduction, continental subduction/collision, to mountain building and collapse (Thomas, 1983; Torsvik et al., 1996; Handy et al., 2010; Zheng, 2012; Song et al., 2014a). All orogens are accompanied by various extent of magmatism, and we can conclude that the continental growth since the Archean is the result of

*Corresponding author (email: sgsong@pku.edu.cn)

accretion and amalgamation of various-sized continental blocks (Condie, 1998; Condie et al., 2010). Thus, magmatism in continental orogens and the continental growth is a key issue that has caught geologists' attention. Although it is generally accepted that most of the continental crust were formed during the Archean, controversy still exists in two issues: the mechanism of continental crust growth and its relationship with the Phanerozoic orogeny (Martin et al., 2005; Smithies, 2000; Condie et al., 2010). Continental subduction/collision zones are the key sites for the interaction and mass transfer between the continental crust and the upper mantle. Since the discovery of coesite and diamond in the 1980s, significant progresses have been achieved in the study of high-pressure (HP) and ultrahigh-pressure (UHP) metamorphism within the continental subduction/collision belts (Zheng, 2012).

Various extent of magmatism occurs in different stages of orogenic process from continental subduction, exhumation, to orogen collapse. The relationship between the genesis of the magmatism in the continental orogens and the continental growth still remains unclear, and has become a focused question and the frontier of the study of continental orogenesis in recent years. Numerous researchers have conducted intensive investigations into the partial melting or magmatism associated with continental subduction and UHP metamorphism within different orogens, e.g., the western Alps (Davies and von Blanckenburg, 1995), the Dabie-Sulu (Zhao et al., 2007, 2012, 2013; Xia et al., 2008; Zhang et al., 2010; Li et al., 2014), the Western Gneiss Region of Norway (Labrousse et al., 2011) and the North Qaidam (Wu et al., 2006, 2007, 2009, 2014; Yu et al., 2012, 2014; Chen et al., 2012; Song et al., 2009b, 2014a, 2014b; Wang et al., 2014). In some collisional orogens, deeply subducted continental crust experienced significant partial melting, resulting in syn-collision/exhumation magmatism, e.g., Late Triassic alkaline magmatic rocks in the Sulu (Zhao et al., 2012) and Silurian adakitic intrusions in the North Qaidam (Song et al., 2009b, 2004b). While in other collisional orogens, partial melting of deeply subducted continental crust is in low-degree and insignificant, only manifested by leucosomes and veins in UHP metamorphic rocks without sizable dykes or plutons, with the examples of the Dabie UHPM terrane (Zheng et al., 2011; Gao et al., 2012, 2013, 2014; Liu et al., 2013). Leucosomes and melt veins are also developed in the Sulu (Liu et al., 2012; Chen et al., 2013; Li et al., 2014) and the North Qaidam (Chen et al., 2012; Liu et al., 2014) UHPM terranes. Therefore, investigations of partial melting and magmatism during continental subduction/collision, exhumation and the collapse of orogens play an important role in a better understanding of the evolution of orogens, the genesis of magmas, the interaction between crust and mantle, and the continental growth. In this contribution, we summarize the types and evolutional history of continental orogens, analyze characteristics of partial melting behavior and petrogenesis of magmatic rocks in different stages of collisional orogeny, and then decipher the internal connection between the continental orogeny, magmatism and the continental growth.

1 Magmatism in continental orogens: process and definition

How to distinguish anorogenic magmatism, arc magmatism related to oceanic subduction and collisional magmatism related to continental subduction/collision/collapse is still a knotty problem when studying in some old orogenic belts. Many literatures try to use geochemical discrimination diagrams to identify the tectonic regimes in which granitic magma were generated, and then to discuss and build the tectonic framework of an area or orogeny. But they ignored the compositional inheritance of these granitic rocks from their variable sources (Zheng, 2012). It should be emphasized that the rock types and geochemical characteristics of intermediate and felsic magmatic intrusions have close relationship with their crustal and mantle sources (Chen et al., 2014). Same type of granites can be generated in different tectonic settings and different types of granites can be also formed in the same tectonic setting (Li et al., 2013; Wang et al., 2014). It must cause uncertainty and ambiguity about the tectonic evolution of ancient orogens if simply using geochemical characteristics of granitic rocks to identify tectonic settings. Therefore, the discussion about magmatism in orogens and its genesis should be based on the tectonic framework of the orogenic evolution.

In the collisional orogens with UHP metamorphic belts, the problem raised above will be easily solved. We can use the UHP metamorphic belt as an effective milestone to identify syn-/post-collisional and non-collisional magmatism. However, in the case of orogens without UHP metamorphic belt related to continental collision, for instance, the Tianshan-Xingan-Inner Mongolia orogens (also called accretionary orogen; Li et al., 2004; Windley et al., 2007; Xiao et al., 2009; Long et al., 2012), there will be confused and debated in the definition and identification for pre-collisional island arc/active continental margin magmatism, syn- or post-collisional magmatism and post-orogenic or anorogenic magmatism. Therefore, to understand the relationship between magmatism and orogenesis, we should firstly clarify the orogen types and their architecture, formation process and evolution history.

1.1 Types of continental orogenic belts

Cawood et al. (2009) summarized continental orogens into three types: (1) accretionary orogen with B-type or oceanic-type subduction; (2) collisional orogen with A-type or continental-type subduction; (3) intracratonic orogen (No A-type subduction). Li et al. (1999) classified collisional orogens in more details and Zheng (2012) also classified collisional orogens into arc-continent collision type and continent-continent collision type.

According to the definition of Cawood et al. (2009), the accretionary orogen is referred as the oceanic circum-Pacific subduction systems, including the Cordillera-Andes active continental margin accretionary belt and the trench-arc-basin system in the west part of the Pacific Ocean. The accretionary mechanism is characterized by oceanic subduction (B-type), arc magmatism and arc-continent collision. Accretionary orogens are considered as the main site for continental growth (Cawood, et al., 2009).

However, after extinction of the subducted oceanic lithosphere, two possible scenarios for interaction between the two assembling continents would be expected: (1) in the case of single-side subduction, one continent will continue to subduct and collide with the opposite continent by the drag force of subducted oceanic lithosphere; (2) in the case of double-side subduction, the drag force of subducted oceanic lithosphere disappears with no strong subduction/collision between two continents, and then a broad micro-continents-arc-ophiolite mélange belt develops resulting from the long-lived oceanic subduction and accretion (e.g., Central Asia Orogenic Belt, Windley et al., 2007; New England Orogen in Australia, Offler et al., 2011; North Qilian Orogen, Song et al., 2013), which belongs to accretionary orogen.

The best example for collisional orogens is the Cenozoic Alps-Himalaya orogen, which is formed by subduction and collision of the Indian-African-Arabian plates towards the Eurasia continent. It is also characterized by A-type or continental-type subduction belt and accompanied by UHP metamorphism terranes resulting from continent deep subduction (Ernst et al., 1995; Maruyama et al., 1996; Chopin, 2003; Liou et al., 2004, 2009a, 2009b; Song et al., 2006, 2009a, 2014a; Zheng, 2012). Figure 1 shows the distribution of continental collisional orogens with UHP metamorphic terranes on Earth.

For collisional orogens, the generally accepted orogenic process can be summarized as three steps. (1) After seafloor consumption upon continental collision, the continental portion of the same lithosphere (i.e., in the case of passive continental margins) continues to subduct beneath the opposite continent along the subduction zone by the drag force of subducted oceanic lithosphere, leading to significant crustal thickening. (2) When subducted continental crust reaches certain depth (100–300 km), the oceanic lithosphere detached from the continental lithosphere, and the low-density continental lithosphere exhumes and mountains begin to uplift. (3) In the last stage of orogens, erosion, unrooting and collapse occur, marking the end of an orogenic cycle (Song et al., 2014a). Two prerequisites are needed for the continental deep subduction: (1) prior oceanic lithosphere subduction and (2) this oceanic lithosphere shared by its continental counterpart through the passive margin (Liou et al., 2004; Song et al., 2006; 2014a). Magmatism related to collisional orogeny runs through the whole process of an orogeny.

For accretionary orogens, double-side oceanic subduction zones may exist between the two converging continents and oceanic lithosphere and continental lithosphere might be separated. Without the drag force of subducted oceanic lithosphere, there will be no deep subduction of the two converging continents, thus no strong collision and orogeny will occur. Even though the underplating of arc magmas will form a significantly thickened crust (e.g., Andean continental arc), it is still doubtful that the syn- or post-collisional magmatism would occur as the result of unrooting and collapse as do the collisional orogens.

1.2 The evolution history of collisional orogens and the timing of UHP metamorphism during continental deep subduction

The onset of continental collision should be at the converging time of two continents soon after the oceanic crust is totally consumed. It will take some time for one continent subducting beneath another to reach the UHP metamorphic conditions. It was generally believed that the onset time for the collision between the India and the Eurasia plates is around 70–65 Ma (Ding et al., 2005; Yin and Harrison, 2000, Yin, 2006). The UHP metamorphism when the India plate went down to depths of 80–100 km took place at 53–46 Ma (Kaneko et al., 2003; Leech et al., 2005; Wilke et al., 2010), with a 12–15 Myr gap from the initial collision. Some researchers suggested that the uplift of the Himalaya Orogen began at 50 Ma (An et al., 2001), which is contemporaneous with the UHP metamorphism. Up to now, the India plate is still moving northward at a speed of 3–4 cm yr^{-1}. Therefore, for the Himalaya Orogen, the orogeny might have lasted for 65–70 Myr, but it remains further clarification whether the Himalaya Orogen is still the continuation of continental collision or in the post-collision process.

With regard to some ancient collisional orogens within the continents that have ceased for a long time, it is not an easy job to figure out their evolution history. Firstly, the extinction timing of the oceanic crust between continents should be determined, which can be estimated on the basis of ages of the youngest ophiolite, arc volcanic rocks and HP metamorphic rocks. For instance, in the Qilian Orogen, the youngest arc volcanic rocks were formed at 445–440 Ma and the youngest ophiolite in back-arc basin was formed at 450 Ma and the youngest blueschist was formed at 445 Ma. Thus it is inferred that the Qilian Ocean was closed at 445–440 Ma, which is consisted with sedimentation of the Silurian remnant-sea flysch (Song et al., 2013). The Early Devonian molasse represents the timing of mountain building and uplift, later than the UHP metamorphic ages (440–420 Ma, Song et al., 2014a). The timing of termination of an orogenic cycle is also a bewildering question. In

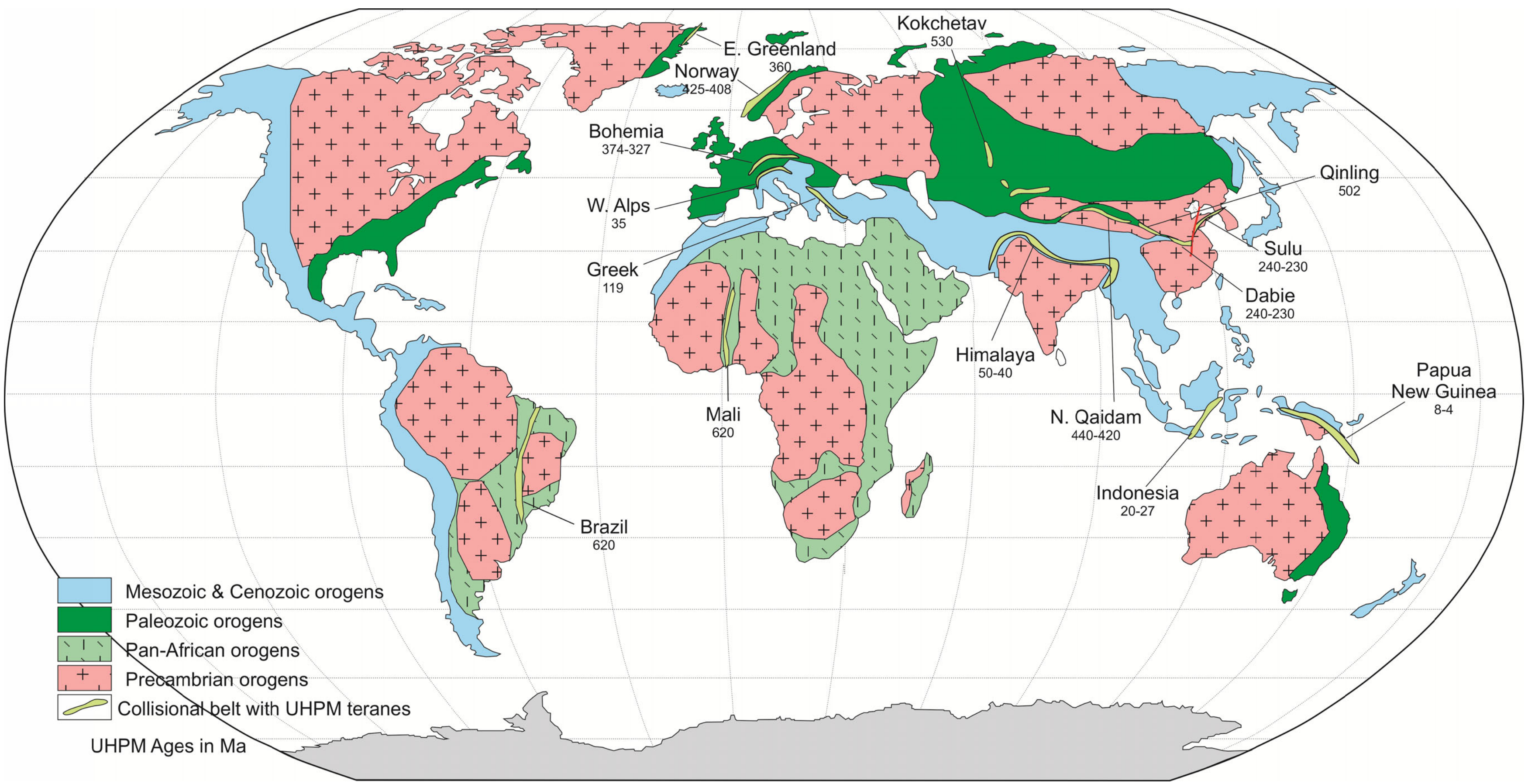

Figure 1 Global distribution of collisional orogens with UHP terranes. Revised after Liou et al. (2009a).

the last stage of collisional orogens, delamination of mountain root and collapse of mountains were the major events, which were accompanied by large scales of magmatic activity, e.g., post-collisional magmas in the Caledonian Orogen (Atherton et al., 2002, 2006) and in the North Qaidam UHPM belt (Wu et al., 2004, 2014; Wang et al., 2014). The post-orogenic magmatism, on the other hand, can be much later than continental collision. For example, the continental collision of the Dabie-Sulu orogenic belt occurred the Triassic, but the post-orogenic magmatism in the Dabie orogen occurred in the Early Cretaceous, with a period of silence (Zhao and Zheng, 2009). However, Late Triassic post-collisional magmatic rocks were found in the Sulu orogen, which could constrain the termination time of orogeny effectively.

In addition, erosion of the orogen's surface and widespread epeiric sea sedimentary cover could also serve as the marker of the end of orogeny. For example, in the Qilian-Qaidam composite orogen, the complete orogenic cycle (from oceanic subduction, continental collision, to mountain building and final collapse) began at the initial oceanic subduction (~520 Ma), and lasted for ~170 Myr to the end of collisional orogeny. However, the time span of collisional orogeny is ~80 Myr from collision (440 Ma) to collapse (360 Ma). The complete orogenic cycle of the Caledonian Orogen lasted for ~200 Myr (McKerrow et al., 2000). The collision between the Baltic and the Laurentia blocks took place at ~435–425 Ma (Torsvik et al., 1996; Torsvik et al., 2003), the UHP metamorphism in the Western Gneiss Region occurred at 415–400 Ma, and the exhumation to crustal level was at 390–375 Ma (Andersen et al., 1998; Root et al., 2004). Therefore, the Caledonian Orogen lasted for >60 Myr from collision to the end of orogeny.

As the density of continental crust is much lower than those of oceanic crust and upper mantle, subduction depths of the continental lithosphere are limited. Most of the subducted continental lithosphere can go down to 200 km, forming UHP index minerals like diamond and supersilicic garnet, e.g., the Dabie-Sulu (Xu et al., 1986; Yang et al., 1993; Ye et al., 2000), the North Qaidam (Song et al., 2004, 2005a, 2005b), the Kokchetav Massif (Sobolev et al., 1990; Ogasawara et al., 2002; Zhu et al., 2002) and the Western Gneiss Region (van Roermund et al., 1998, 2000). Some rocks may record depth up to 300 km (Doberizskaya et al., 1996; Piromallo et al., 2004; Liu et al., 2007; Spengler et al., 2006). As suggested in previous assumption, when subducted oceanic lithosphere detached from the dragged continental lithosphere, the continental crust ceased to subduct and the UHP metamorphic rocks began to exhume as the result of buoyancy (Davies and van Blanckenburg, 1995). However, according to the subduction channel model, deep subducted crustal rocks can detach from the continental crust in mantle and then exhume to crustal level or even the Earth's surface along the subduction channel driven by buoyancy and corner flow (Zheng et al., 2013), which can be documented in the *P-T-t* paths of UHP metamorphic rocks (Li et al., 2014).

As the indicator of continental deep subduction, the time spans of UHP metamorphism in different UHP metamorphic belts and orogens are surprisingly similar. The duration of UHP metamorphism in most UHP metamorphic belts is less than 20 Myr and possibly has a positive correlation with the exposed size of UHP metamorphic rocks (Zheng et al., 2012, 2013). The UHP metamorphism in the Dabie-Sulu UHP metamorphic belts, one of the largest exposed UHP metamorphic zone in the world, lasted for only 15–20 Myr (Liu et al., 2006; Zheng et al., 2009; Liu and Liou, 2011).

Isotopic geochronological study of coesite-bearing gneisses, eclogites and diamond-bearing garnet peridotites shows that the UHP metamorphism of the North Qaidam occurred no later than ~420 Ma (Song et al., 2005b, 2006; Mattinson et al., 2006). Mingled by the HP-UHP metamorphic rocks produced in previous oceanic subduction, ages of eclogites and gneisses in the North Qaidam UHP metamorphic belt are complicated and some researchers believed that the duration of UHP metamorphism in the North Qaidam UHP metamorphic belt is more than 60 Myr (Zhang et al., 2005; Mattinson et al., 2006). However, detailed investigations of petrology and U-Pb zircon geochronology reveal that two stages of eclogite-facies metamorphism can be identified: the earlier stage (>440 Ma) eclogite-facies metamorphism related to oceanic subduction and the later stage (435–420 Ma) eclogite-facies metamorphism related to continental subduction (Song et al., 2014a). The UHP metamorphism related to continental subduction lasted for about 15 Myr.

1.3 Determination of syn- and post-collisional magmatism

Pitcher (1983) systematically summarized the relationship between the granitic rock types and their tectonic settings, pointing out that granitic rocks can be generated in (1) island arc and continental arc (pre-collisional); (2) continental collision; (3) post-collisional collapse; (4) anorogenic environment (continental rift, ocean island and mid-ocean ridge). Many geologists tried to use the geochemical indicators of granitic rocks to discriminate and distinguish the tectonic setting in which granitic rocks were generated, leading to some terms of granitic rocks, such as ocean ridge granite (ORG), volcanic arc granite (VAG), within plate granite (WPG), syn-collisional granite (syn-COLG) and post-collisional granite (post-COLG) (Pearce et al., 1984, 1990; Rogers and Greenberg, 1990). However, misleading will arise if we only use geochemical indicators to discriminate tectonic settings of granitic rocks (Figure 2).

Continental deep subduction and crustal thickening are two major mechanisms for generation of magmas during collisional orogeny. On the basis of the collisional orogenic process, we can understand the generation of magmatism

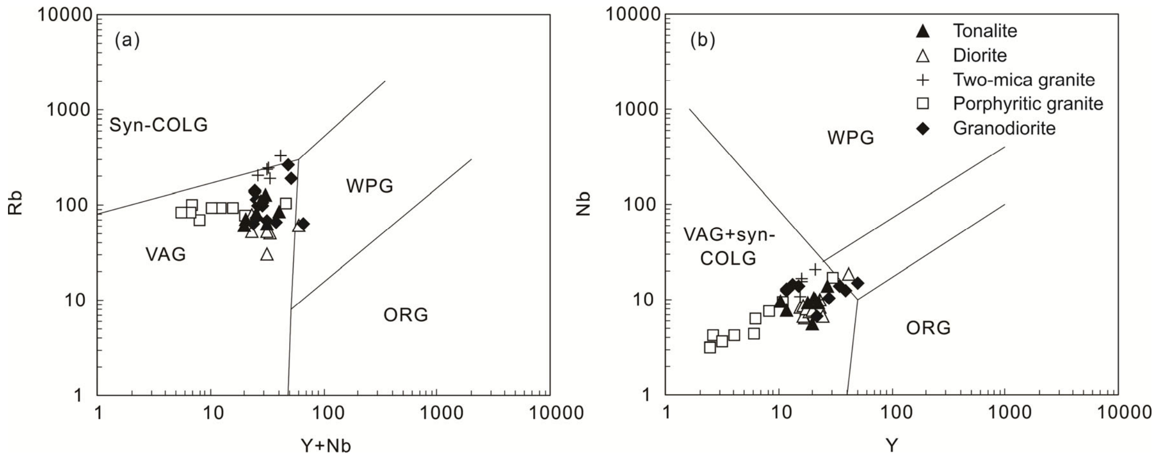

Figure 2 Tectonic discrimination for the post-collisional magmatism by using Rb vs. Y+Nb (a) and Nb vs. Y (b) of Pearce et al. (1984). The post-collisional magmatism in the North Qaidam UHPM belt formed at 405–360 Ma, much later than the UHPM ages. Almost all analyses plot in the VAG or VAG+syn-COLG filed. Data from Wang et al. (2014).

from three aspects as following. (1) After consumption of subducted oceanic lithosphere, two continents began to collide and one is dragged by the earlier subducted oceanic crust beneath the other, leading to the superimposition, thickening and UHP metamorphism. If breakoff occurred between continental and oceanic lithospheres, the deep subducted oceanic and continental crust would exhume and experience decompression melting to generate magmatic rocks, as exampled by the North Qaidam (Song et al., 2014b) and the Sulu (Zhao et al., 2012) UHPM terranes. (2) During continental collision, large scale of thrust and strike-slip would take place and side-way flow and extrusion would occur in mid-lower crustal level. But the genetic relationship between the mid-high pressure granulite-facies metamorphism and anatexis in the mid-lower crustal level and continental subduction remains to be further clarified. Granitic intrusions and veins formed during continental thrust and strike-slip, but their scale is minor and they are derived from melting of felsic crust and sediments, e.g. the Greater Himalayan Orogen (Harrison et al., 1997, 2005; Harris et al., 2004; Zhang et al., 2015a, 2015b; Song et al., 2010). (3) When continental subduction is ceased, delamination and unrooting occurred, giving rise to orogen collapse and upwelling of the asthenosphere, and resulting in generation of complicated magmas by interaction between mantle and continental crustal melts. Among the post-collisional magmatic rock assemblages in North Qaidam, there are a large number of mafic dykes and gabbroic diorites from decompression melting of the asthenosphere and I-type granite from mixing between crust and mantle, suggesting that the post-collisional stage is an important period of interaction between crust and mantle (Wang et al., 2014).

Actually, due to the complexity of the orogenic process, without some key indicators such as the subduction timing of oceanic crust (the youngest arc volcanic rock), the closing timing of ocean basin (e.g., the youngest ophiolite) and HP-UHP metamorphism related to oceanic and continental subduction, it is fairly difficult to discriminate magmatism in different stages of the orogeny. Definition and determination of items such as syn-orogenic, post-orogenic, syn-collisional and post-collisional are of considerable obscurity and confusion. Post-collisional or post-orogenic magmatism are frequently used to refer magmatism that postdates the main collisional orogenic event.

When considering the relationship between magmatism and collisional orogeny, it is essential to further clarify the definition of syn-orogenic, post-orogenic, syn-collisional, post-collisional and non-collisional, on the basis of Liégeois (1998). As illustrated in Figure 3, "syn-orogenic" magmatism refers to magmas generate in the whole orogenic cycle from oceanic subduction, continental collision to orogen/mountain collapse. "Post-orogenic" magmatism refers to magmas occurred after the end of an orogenic cycle, which belongs to the category of within plate magmatism, with no correction to the orogenic event, or superimposed by other orogenic events. Besides, categories for orogenic and anorogenic belts should be also clarified. For example, with regard to Cenozoic tectonic events in the Himalaya, the Tibetan Plateau is not strictly a Cenozoic orogen, but a region strongly affected by the Himalaya Orogen. Magmatism within the Tibetan Plateau should be classified as within plate magmatism, with large scale of extremely enriched potassic and ultrapotassic volcanic rocks (Miller et al., 1999; Chung et al., 2005; Zhao and Zheng, 2009; Liu et al., 2014b). The source of the potassic components might be related to the enrichment of mantle wedge induced by subduction of Paleo-tethys and Meso-tethys oceanic crusts (Zheng, 2012).

(i) Syn-orogenic. Magmatism covers all stages of an orogenic cycle from the beginning to the end, including the pre-collisional island arc and continental arc magmatism associated with oceanic subduction, and the syn- and post-collisional magmatism during continental collision and orogen collapse.

Types		Period	Mechanism/ setting	Source	Rock type	Example
Syn−orogenic	Pre-collisional		Oceanic subduction	Mantle wedge Subducted oceanic crust Arc crust Continent marginal crust	Tholeiitic to calc-alkaline Arc volcanics Adakites Intermediate to felsic intrusions	Circum-Pacific North Qilian
	Syn-collisional	5–10 Myrs	Continental collision & subduction	Previous subducted oceanic crust Subducted continental crust	Adakites Low Sr/Y sodic volcanics Tonalites S-type granites	Himalaya North Qaidam Western Gneiss Region
		5–20 Myrs	UHP metamorphism			
		20–30 Myrs	Exhumation	Interaction between slab and mantle in the subduction channel	Alkaline mafic magmatic rocks	Dabie-Sulu
	Post-collisional	30–40 Myrs	Unroot and collapse of orogens	Continental crust Asthenoshphere (crust-mantle interation)	I-, S-, A-type granites High-K volcanic rocks	North Qaidam Caledonian Tibet (WPB affected by orogenesis)
Post-orogenic Anorogenic			Superimposition of orogens Continental rift (within plate)	Continental lithosphere Asthenosphere	Calc-alkaline, alkaline-calc and alkaline volcanics I-, S-, A- type granites	Xingan-Inner Mongolia Dabie-Sulu (Superimposed by the subduction of the Pacific Ocean)

Figure 3 Classification for orogeny and related magmatism.

(ii) Syn-collisional. During the continental collision and subduction, the previous subducted oceanic crust and continental crust experience decompression melting in the processes of exhumation and crust thickening. Magmas in these processes would occur at the same time with, or 20–30 Myr later than the UHP metamorphism. They are produced by partial melting of crustal materials (including subducted oceanic and continental crust), including tonalites (with or without adakite signatures) and peraluminous granites, without input of mantle materials. Besides, crustal melts generated during the exhumation of deep subdutcted continental crust could metasomatize the overlying lithospheric mantle, producing mafic magmas (Zhao et al., 2012).

(iii) Post-collisional. It commonly refers to magmatism after the major collisional event, but there is confusion in the definition in some literatures. Liegeois (1998) defined post-collisional magmatism as 'formed in an intracontinental environment with the major ocean closed but still with large horizontal terrane movements including thrust, rotation, extrusion, strike-slip along mega-shear zones, lithospheric delamination, and opening of small oceanic plate and rifting'. This definition corresponds to late-collisional magmatism in the last stage of continental orogenic cycle. Magmatism should occur in the relaxation period after the main orogenic period. In the earlier stage, large scale horizontal movements took place along the plate boundaries, and in the later stage, unrooting and collapse of orogens and extension of lithosphere occurred with interaction between mantle and crust. The rock types formed in these periods include (1) potassic and high-K calc-alkaline derived from the metasomatic upper mantle, (2) peraluminous granites generated by partial melting of upper continental crust, (3) I-type granites, granodiorites and diorites formed by partial melting of lower continental crust and interaction between mantle and crust, with mafic dykes from asthenospheric and lithospheric mantle. Generally the post-collisional magmatism lasts for around 10–40 Myr, and marks the end of the orogenic cycle.

(iv) Post-orogenic. It commonly refers to magmatism after the orogeny and with no correlation to orogenic process, only because it is located within the framework of an orogen. The genetic mechanisms are complicated, mainly resulting from the superimposition of tectonic movements and mantle plume activity in abjection regions, e.g., interaction of three main tectonic regimes in the Xing'an-Inner Mongolian Orogen from Paleozoic to Mesozoic, superimposition of the Pacific plate subduction on the Paleozoic orogen in South China and Dabie-Sulu-Qinling orogenic belt, and so on. The rock types are variable and depend on their source compositions (asthenospheric and lithospheric mantle and continental crust), including calc-alkaline to alkaline volcanic rocks and granitic rocks with compositions of normal, potassic-ultrapotassic with LILE-enrichment. These magmas have no essential difference from anorogenic magmatic rocks in terms of geochemical compositions.

(v) Anorogenic. It refers to within plate magmatism related to rift or mantle plume activity within a stable continent.

It should be emphasized that the "post-collisional" and the "post-orogenic" are two different concepts. In the Dabie-Sulu-Qinling orogenic belt, the 140–120 Ma magmatic rocks (Ma et al., 1998; Zhao and Zheng, 2009; Zhang et al., 2010) are not "post-collisional", but correlates with the widespread Mesozoic magmatism in Eastern China related to the subduction of the Pacific Plate (Niu, 2005, 2014; Li et al., 2007; Zhao et al., 2013).

2 Classification and formation mechanism of magmatism in continental collision belts

2.1 Magmatism during continental collision, deep subduction and exhumation of UHP terranes

In the Himalayan orogen, Linzizong volcanic rocks are thought to represent typical syn-collisional magmatism (Mo et al., 2008; Niu et al., 2013). These intermediate-felsic volcanic rocks are composed of andesite, dacite, and rhyolite with dominantly sodic compositions (N_2O>K_2O), and formed in 64–40 Ma, coeval with Asian-Indian collision. They have positive $\varepsilon_{Nd}(t)$, indicating a juvenile crustal source. Niu et al. (2013) suggested they are melts of previously subducted oceanic crust basalt during continental collision before slab breakoff.

Combining HP/HT melting experiment with natural melt veins in the UHP metamorphic zone, Labrousse et al. (2011) indicated that the earliest trondhjemitic dykes in UHP rocks in Western Gneiss Region (WGR) are resulted from melting of eclogite under conditions of T=750°C, P>2.5 GPa, coeval with the UHP metamorphism. Granitic gneiss in the Sulu UHP terrane also records anatexis during 237–223 Ma, which overlaps the age of UHP metamorphism (Li et al., 2014). These studies provide evidence for synchronous UHP rock melting and deep subduction. More melts generate in the exhumation stage (Chen et al., 2012, 2013; Yu et al., 2012; Zhao et al., 2012; Liu et al., 2014a), which may, in turn, promote further exhumation of UHP rocks (Zheng et al., 2011; Chen et al., 2013).

Tonalite and trondhjemitic veins and intrusions in South Dulan belt maintain obvious adakitic characters (Song et al., 2009, 2014b; Yu et al., 2012). Zircon and rutile U-Pb dating yielded ages of 435–410 Ma (Song et al., 2014b; Yu et al., 2012; Zhang et al., 2014), overlapping the UHP metamorphic ages (435–420 Ma). Whole rock Nd isotopes and zircon Hf isotopes show that most samples are derived from depleted crustal source ($\varepsilon_{Nd}(t)$>0, $\varepsilon_{Hf}(t)$>5), indicating they are melts of previous subducted oceanic crust during exhumation. Melting occurred during transition from UHP eclogite phase to HP granulite phase metamorphism, and crystallization take place under conditions of T=850–930°C, P=2.0–1.8 GPa. Therefore, deep subduction of continent is capable of inducing exhumation of previously subducted oceanic crust, and partial melting induced by decompression, heating, or dehydration of lawsonite/zoisite produces adakitic tonalite and trondhjemite with depleted isotopic feature (Figure 4). Further exhumation leads to rising and melting of subducted lower continental crust, which gives rise to adakitic melts with enriched Nd-Sr isotopic feature (Zhang et al., 2015).

It needs to note that the syn-collisional adakitic rocks that

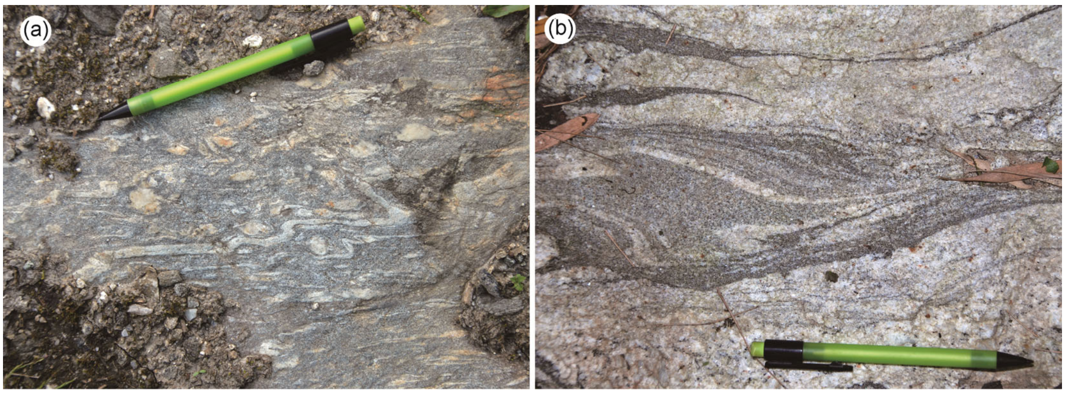

Figure 4 Partial melting of meta-pelite with strong deformation under high-strain conditions, eastern Himalaya.

are coeval with, or postdate UHP metamorphism, either derived from oceanic crust or thickened lower mafic continental crust, have been identified within a variety of UHP belts, e.g. adakites in Agvanis terrane, northeastern Turkey (51 Ma, Topuz et al., 2011), adakites in Tethys Himalaya (43–41 Ma, Zeng et al., 2011), and the earliest trondhjemitic dykes in WGR (Labrousse et al., 2011). These adakites resemble Archean TTGs and play an important role in growth of continental crust. In the Dabie-Sulu orogenic belt, the overloaded lithospheric mantle of the North China Craton can be altered by syn-exhumation felsic melts in the subduction channel and then become the mantle sources of post-collisional, mafic-ultramafic magma (Dai et al., 2011, 2012; Zhang et al., 2012; Yang et al., 2012a, b; Zhao et al., 2011, 2012, 2013). Therefore, the diversity of magmatism in continental collision belts calls for worldwide further research.

To sum up, partial melting and magmatism can follow the continental deep subduction and UHP metamorphism through decompression and dehydration of water-bearing minerals, and early subducted oceanic crust is most likely the dominant source of the sodic tonalite-trondhjemite.

2.2 Magmatism related to slab detachment and anatexis of thickened crust

Traditional hypothesis suggested that exhumation of UHP terranes and syn-exhumation magmatism are closely associated with slab breakoff (Davies and von Blanckenberg, 1995). As the depth of subduction increases, regional extension in continent-ocean transitional zone and strength reduction occur as consequences of dense difference between continental and oceanic lithosphere and asthenosphere thermal erosion, which lead to detachment of the dense, eclogitized oceanic lithosphere from the continental lithosphere, and exhumation of the latter, or whole-scale subduction reversal. Upwelling asthenosphere causes partial melting of upper metasomatized lithospheric mantle, forming high-K calc-alkaline series or shoshonitic series magmatism. Upward movement of mantle-derived magma could also induce partial melting of lower continental crust, or even upper continental crust, forming bimodal magma suite. Such magmas are often associated with HT metamorphism and upraise of lithosphere, and are linearly distributed along large strike-slip faults, reflecting similar distribution of the deep heat source, thus are different from magmatism induced by mantle plume. Numerical modeling indicates that the time interval between slab breakoff and continental collision is controlled by original thermal structure, rheology of subducted lithosphere, and the intensity of asthenosphere thermal erosion. Slab breakoff occurs 5–30 Myr after the continental collision, while magmatism would last for 10–20 Myr (Gerya et al., 2004).

The Late Caledonian magma in Scotland and the north part of Ireland (Late Granites, 435–390 Ma, Atherton and Ghani, 2002) and the Tertiary magma in Alps (Davies and von Blanckenburg, 1995) are thought to represent typical slab-breakoff-related magma assemblages. The Late Caledonian magmatism includes two magma series: (1) The early series consists of mantle-derived mica/amphibole-rich, high-K calc-alkaline magmas that may be evolved from mica lamprophyre with high Mg#, Ba, Sr, K/Na, La/Yb, and depleted in HFSE (Nb, Ti) and HREE, which is considered to represent primary mantle melts. Mica-lamprophyre may experience fractional crystallization during upward transfer, forming a magmatic suite containing mica peridotite, amphibole pyroxenite, mica-bearing diorite, quartz diorite, granodiorite and granite. (2) The contemporary or later magmatism consists of I-type granitoids that are derived from an essentially igneous source. Comparing to these mantle-derived magmas, they exhibit higher Rb, lower Ba and Sr. Geochemistry and Sr-Nd isotopes of Tertiary magmatism in Alps (Davies and von Blanckenburg, 1995) suggested that partial melting of continental crust is coeval with melting of metasomatized mantle, and mantle-derived magmas contain crustal material from the subducted continental crust. Besides these features above, the Miocene slab-breakoff-induced mantle melts in Karakoram, which have lasted for ~15 Myrs, also show evolutionary trend from enrichment of non-radiogenic Nd-Hf isotopes to depletion, indicating the range of partial melting expands towards deep mantle with slab breakoff (Mahéo et al., 2002).

Recent modelling shows that the depth of slab breakoff in the large scale orogenic belt often exceeds 200 km (van Hunen et al., 2011), coinciding with UHP metamorphism of continental crusts. Asthenosphere is incapable of melting at such depth unless decompression occurs (Davies and von Blanckenburg, 1995). Slab breakoff in shallow level(<70 km), which may be induced by subduction of mid ocean ridge or juvenile oceanic crust, leads to decompression melting of upwelled asthenosphere and genesis of magma with depleted mantle signature (van de Zedde et al., 2001). Such magma is contemporary with, or little later than, the alkaline magma derived from metasomatized mantle, and accompanied by large scales of granitoids from melting of continental crust, with example of the Silurian magma suite in Newfoundland Appalachia (Whalen et al., 2006).

Slab breakoff is not the requisite for exhumation of UHPM rocks. Decoupling of the less dense upper crust from the lower crust/mantle lithosphere could also trigger exhumation of UHP rocks (Chemanda et al., 1995, 2000; Zheng et al., 2013). Being dragged downward by oceanic lithosphere, the continental crust is weakened by strain, fluid alteration and partial melting, and then detached from underline continental lithosphere, moving upward under control of positive buoyancy and regional tectonic regime, forming metamorphic core complex (Warren et al., 2008; Beaumont et al., 2009). Exhumation generally lasts for 10–30 Myrs in large orogenic belts and less than 10 Myrs in small-scale orogenic belts (Zheng et al., 2009; Kylander-

Clark et al., 2012). The difference of this model from slab breakoff is that exhumation do not necessarily mean whole-scale subduction reversal. Experiments show that the strength of the subducting continental crust is intensively reduced while partial melting proceeding (Rosenberg et al., 2005). Trondhjemitic leucosomes in UHP gneisses in WGR were formed by fluid-bearing melting of crustal rocks under peak metamorphic condition, which is considered to trigger strength reduction and exhumation (Labrousse et al., 2011).

Due to absent of heating from asthenosphere, the dominant magma is peraluminous leucogranite resulting from decompression melting of thickened continental crust, occurs as leucosome in migmatite, or small-scale dyke/body in metamorphic rocks. Multistage exhumation of the thickened crust gives rise to multiple partial melting under different condition. For example, the 45–40 Ma intrusive rocks in the Himalayan Orogen mainly occur in Tethys Himalayan core complexes, including adakitic granites (high Sr/Y, negative $\varepsilon_{Nd}(t)$, Zeng et al., 2011) and S-type leucogranites (Aikman et al., 2008), representing melts from amphibolites at the middle/lower crust level and supracrustal rocks, respectively. In the High Himalaya Crystalline (HHC), metamorphism and anatexis of HP granulite and related magmatism have lasted a period of ~20 Myr from 32 to 12 Ma (Harrison et al., 1997; Ding et al., 2001; Searle et al., 2003; Rubatto et al., 2013; Zhang et al., 2014a, 2014b). These magmatic rocks are mainly strongly-peraluminous leucogranites derived from granulite-facies dehydration melting of pelite (Figure 5). The early-stage melts are result from muscovite dehydration, and the late-stage melts are from breakdown of garnet (Imayama et al., 2012), reflecting long-lived melting of continental crust from its deep to shallow levels (Zhang et al., 2014b). Early melts often exhibit shear deformation, and are crosscut by later melts. The ages of emplacement could be determined by monazite or zircon of different stages with core-mantle-rim structures in melts (Streule et al., 2010; Gordon et al., 2012), which could also help calculating exhumation velocity in combination with their melting *P*-*T* conditions.

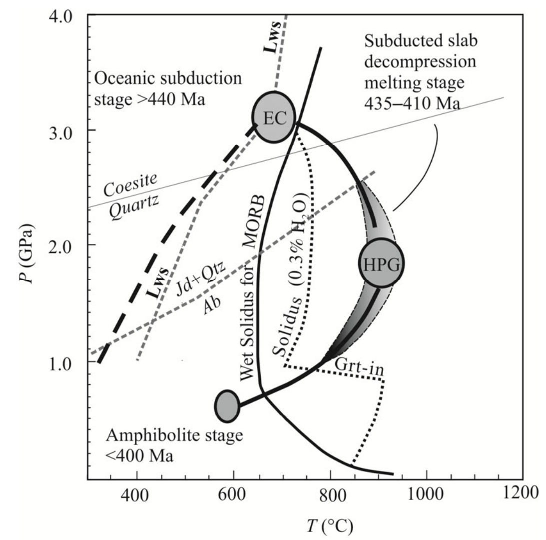

Figure 5 *P-T-t* path for generation of adakite melts in the Dulan Terrane, N. Qaidam UHPM belt. After Song et al. (2014b).

2.3 Post-collisional magmatism related to unrooting and collapse of orogens

Unrooting of an orogen is the process of detachment or delamination of low part of the thickened orogenic lithosphere from its upper part due to mineral phase transition, change in rheology, which is also affected by gravity and mantle convection (McKenzie, 1978; Bird, 1979; Houseman et al., 1981; Fleitout and Froidevax, 1982; Marotta et al., 1998). Unrooting leads to tectonic transition from compression to extension, i.e. forming of extensional structure, elevation of heat flow, regional uplift and collapse, etc. As summarized by Marotta et al. (1998), two tectonic models are representative: (1) lithospheric mantle is peeled away from the overlying crust owing to the intrusion of asthenospheric material by convective erosion of thermal boundary layer (Figure 6(a)), and (2) delamination, or whole-scale detachment of mantle lithosphere (with or without lower crust) from crust because of its gravitational instability (Figure 6(b)) (McKenzie, 1978; Houseman et al., 1981, 1997). Magmatism with high-K magmas from metasomatized lithospheric mantle is dominant in the latter scenario. Delamination, however, triggers shallow upwelling and decompression melting of asthenosphere, leading to genesis of mafic-ultramafic magma with depleted mantle signature in association with coeval crustal melts (Wang et al., 2014).

Unrooting occurred in southern Tibetan plateau in between 25 and 10 Ma under the effect of Himalayan orogeny. Magmatism related to this event includes high-K volcanics with OIB feature (Turner et al., 1996; Miller et al., 1999; Williams et al., 2001), peraluminous granites (Mo et al., 2007), and adakites (Chung et al., 2003), and was considered to be the consequences of convective erosion of thermal boundary layer. High-K porphyritic volcanics in southwestern Tibet plateau are phlogopite/pyroxene-bearing mafic magmas with minor fractionation. They exhibit high K_2O (>3.5 wt%), Ni and LREE, and enrichment in LILE relative to HFSE, indicating an enriched continental lithospheric mantle source (Zhao et al., 2009). Coeval high-K calc-alkaline magmas are widespread in southern Tibet and are mainly porphyritic dacite-rhyolite or their intrusive equivalents. Their adakitic signatures, e.g. high Sr/Y, La/Yb and low Y, Yb, HREE, imply a source of lower continental crust where garnet remains in residuum. Sr-Nd-Pb isotope compositions show possible mantle participation, which could also provide heat for melting of crust (Hou et al., 2004; Guo et al., 2007). However, such isotope signature could also indicate a juvenile oceanic crust source (Zheng, 2012; Niu et al., 2013; Song et al., 2014b).

Collapse magmatism found in the west part of Anatolia

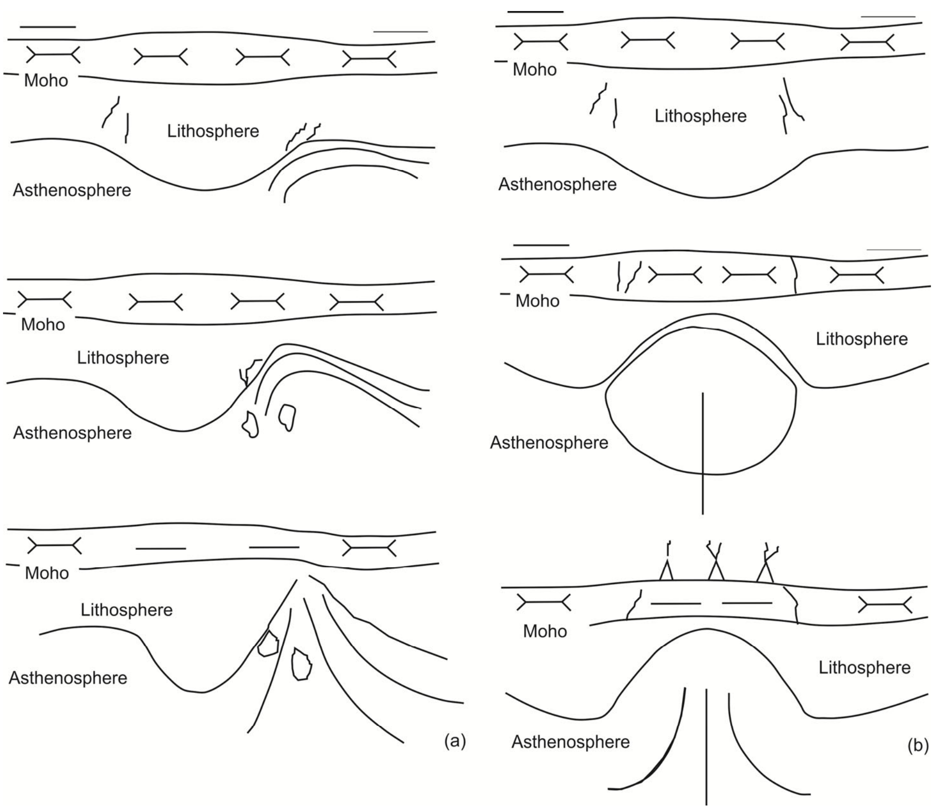

Figure 6 Schematic representation of lithospheric unrooting according to the model of Bird (1979) (a) and the model of Houseman et al. (1981) (b). After Marotta et al. (1998).

Orogen is shoshonitic lamprophyre and exhibits high MgO, Mg#, Ni, Cr, Sr (Prelevic et al., 2012). ^{87}Sr enrichment and depletion in non-radiogenic Nd-Hf isotopes reflect the early metasomatic mantle source by subduction zone fluids. From north to south and from early to late, mantle-derived magmas exhibit gradual transition from enriched to depleted mantle signatures, indicating a mantle process of lithospheric mantle "asthenospherization" owing to lithosphere-asthenosphere interaction (Prelevic et al., 2012). Although participation of asthenospheric mantle melts could also lead to depletion in radiogenic Nd-Hf isotopes, such melts are depleted in LILE and LREE and are not in consistent with collapse magmatism in Anatolia (Zheng, 2012).

Mantle-derived, Middle-Late Permian mafic volcanics in Iberian Variscan Orogen show obvious source transition from metasomatized mantle to depleted mantle, reflected by the elevation in radiogenic Nd isotope of the late volcanics. Such transition could be a consequence of delamination after which asthenosphere became source of magma (Gutierrez-Alonso et al., 2011), or simply reflect mantle metasomatism by melts from crustal source with geochemical diversity (Zheng, 2012).

46–25 Ma bimodal volcanics/subvolcanics in Lut-Sistan, eastern Iran are calc-alkaline or high-K calc-alkaline (Pang et al., 2013). Most samples are sodic and exhibit HFSE and HREE of E-MORB signature, while they are depleted in Th, Nb-Ta-Ti, P and exhibit a wide range of Sr-Nd isotope ratios (I_{Sr}=0.7042 to 0.7065; $\varepsilon_{Nd}(t)$=−4.9 to 5.5), resembling those of Linzizong volcanics. These volcanics are considered to be products of delamination and asthenosphere upwelling (Pang et al., 2013), or, as mentioned above, reflect mantle metasomatism by melts from crustal source with geochemical diversity (Zheng, 2012).

Post-collisional, unrooting-related magmatism occurred in Dulan terrane, North Qaidam (Wang et al., 2014) includes early granites with intense right-declined REE patterns and LILE enrichment relative to HFSE, indicating crustal origin. Such magmatism was followed by MME-bearing granodiorite-granite generated by mixing of mantle and crustal melts, and granodiorite exhibits slight negative HFSE anomaly. The latest magma is quartz diorite with high Mg# (66–73), slightly right-declined REE patterns, insignificant HFSE depletion, $\varepsilon_{Hf}(t)$ up to 12, and ubiquitous intrusion of diabase dykes, combined of which indicating a mantle origin with slight contamination of upper crust. Mantle-derived magma triggered partial melting of juvenile crust, giving rise to granites with MREE depletion. In $\varepsilon_{Nd}(t)$-I_{Sr} diagram (Figure 7), the majority of post-collisional magmas are plotted on the mixing line of upper continental crust and MORB, and show transition from depletion of ^{143}Nd to enrichment with time, indicating an increasing tendency of mantle material during the process of lithospheric extension and asthenosphere upwelling at late stage of the orogenic cycle.

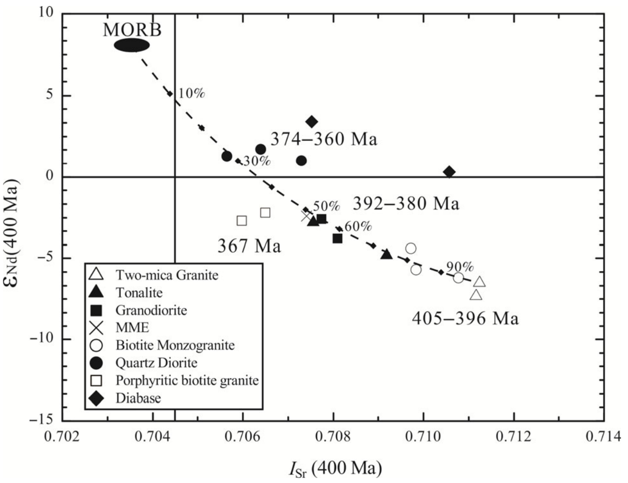

Figure 7 Diagram of $\varepsilon_{Nd}(t)$ versus I_{Sr} for the intrusive rocks in the Dulan UHPM terrane. Modified from Wang et al. (2014).

3 Time scale of magmatism in typical continental collision belts

3.1 European Caledonian orogenic belt

European Caledonian orogenic belt includes eastern Greenland, north part of Ireland, Scotland, and northwestern Scandinavia. The closure of Iapetus Ocean and collision of Baltic-Avalon and Lauren Continent occurred at 435–425 Ma (Torsvik et al., 1996, 2003). Collision and eclogite-facies metamorphism lasted to 400 Ma, followed by 400–390 Ma rapid extensional collapse of the orogen. In Western Gneiss Region, peak UHP metamorphism occurred at 415–400 Ma, and exhumed to crustal level at 390–375 Ma (Andersen et al., 1998; Root et al., 2004; Hacker et al., 2010). The syn-exhumation anatexis ages of UHP rocks are 410–400 Ma (Labrousse et al., 2011; Gordon et al., 2013). Late Caledonian calc-alkaline granites in Scotland and the north part of Ireland emplaced at 435–380 Ma (Atherton et al., 2002, 2006), lasted for ~55 Myr, representing magmatism from closure of Iapetus Ocean and onset of collision to orogen collapse.

3.2 North Qaidam UHP belt

According to records of oceanic-subduction-related arc volcanics, ophiolites, and metamorphic rocks in North Qilian-Laji Mountains, the initial collision of Qilian-Qaidam occurred at ~440 Ma, followed by deep subduction of continental crust and UHP metamorphism at 435–420 Ma (Song et al., 2014). Collisional magmatism is ubiquitous in UHP terranes, e.g. Dulan, Xitieshan, Lvliangshan. Syn-collisional adakitic magmas in Dulan terrane emplaced at 435–410 Ma (Yu et al., 2012; Song et al., 2014b), which was derived from partial melting of subducted oceanic crust (Song et al., 2014b). Exhumation-related, decompression melting of continental crust occurred at 405–390 Ma, and unrooting magmatism at 386–360 Ma (Wu et al., 2006, 2007, 2009, 2014; Wang et al., 2014). Syn- and post-collisional magmatism in Xitieshan and Luliangshan have three major age sets: ~430 Ma, 410-400 Ma, and 375–370 Ma (Meng et al., 2005; Wu et al., 2007; Yu et al., 2014). Therefore, in North Qaidam UHP belt, magmatism from continental collision to orogen collapse lasted for a period of ~75 Myr.

3.3 European Variscan belt

European Variscan belt crops out as several terranes in southern and middle Europe, including middle-northern Pyrenees and Iberian, French Massif Central, Vosges-Black Forest, middle-southern Alps, northern Apennine, Corsica-Sardinia, and Bohemian Massif. Collision of multiple blocks occurred at late Silurian-early Permian, concentrated in early Carboniferous. Ages of peak metamorphism of UHP rocks in French Massif Central are 420–400 Ma, and ages of retrogression and regional uplift are 380–360 Ma (Lardeaux et al., 2001; Berger et al., 2010), clearly predate the Variscan UHP metamorphism, but coeval with the Caledonian UHP metamorphism. UHP terranes in Erzgebirge, Bohemian Massif and Alps record eclogite-facies UHP metamorphism of 360–353 Ma and following 355–348 Ma retrogression (Schmadicke et al., 1995). Syn-tectonic migmatization at 345–310 Ma and post-collisional crustal melting at 310–260 Ma are identified in several terranes (Gerdes et al., 2000; Dallagiovanna et al., 2009; Maino et al., 2012), crustal anatexis at e.g. 345–325 and 300 Ma in French Massif Central, which could possibly related to thinning of orogenic lithosphere (Faure et al., 2010). Syn- or late-Variscan magmas in ages of 345, 338 and 305 Ma were found in Corsica (Paquette et al., 2003). 345, 338 and 305 Ma syn- or late-Variscan magmas were found in Corsica (Paquette et al., 2003). 314–306 Ma syn-tectonic magmatism in Iberian was correspondent to thinning of continental crust (Valle Aguado et al., 2005). Fernández-Suárez et al. (2000) identified four stages of magmatism in Iberian related to collision and collapse of the orogen: (1) syn-D2 325 Ma tonalite-granodiorite-monzogranite; (2) 320–310 Ma leucogranite; (3) widespread 295–290 Ma tonalite-granodiorite-monzogranite with minor mafic-intermediate intrusive rocks; and (4) minor 290–285 Ma leucogranite. The first two stages of magmatism are related to partial melting of thickened continental crust, and the rest reflect collapse and upwelling of asthenosphere. In conclusion, the age of Variscan syn- and post-collisional magmatism occurred in a period of ~85 Ma from 345 to 260 Ma.

3.4 Dabie-Sulu orogen

The UHP metamorphic rocks in the Dabie-Sulu orogenic belt were formed by deep subduction of the South China Block after its collision with the North China Craton, with

UHP eclogite-facies metamorphic ages of 245–220 Ma. Zircons in UHP orthogneisses in the Sulu orogen record ages of 237 and 223 Ma (Li et al., 2014), corresponding to syn-subduction and syn-exhumation anatexis, respectively (Li et al., 2014), but no coeval magmatism is found. Late-collisional, alkaline magmatism occurred at 212–201 Ma (Zhao and Zheng, 2009). Some low-degree partial melting events with Tertiary ages are also found in the Dabie UHP terrane (Xia et al., 2008; Gao et al., 2012, 2013; Liu et al., 2013; Chen et al., 2014, 2015), while no coeval magmas, such as dykes or plutons are found (Zhao and Zheng, 2009). Previous studies regarded 210 Ma in some gabbro-monzodiorite as the age of post-collisional slab breakoff (Ma et al., 1998), but subsequent studies of geochronology have demonstrated that these intrusive rocks formed in the early Crataceous, rather than in the late Triassic. Syn-collisional magmatism in the eastern section of the Sulu orogen mainly consists of an alkaline assemblage of gabbro-quartz monzonite-granite with $\varepsilon_{\rm Nd}(t)$ ranging from −13.8 to −16.4 (Guo et al., 2005; Zhao et al., 2012), and its emplacing age is earlier than 203–201 Ma amphibolite-facies retrogression in this area (Liu et al., 2009). These magmatic rocks were considered as a consequence of partial melting of lithospheric mantle and subducted continental crust during exhumation (Zhao et al., 2012). Liu et al. (2012) identified two stages of anatexis leucocratic dykes with ages of 219–215 and 156–151 Ma, representing melting event during exhumation of the UHP terrane and collapse of the orogen, respectively. To sum up, the ages of partial melting and syn-collisional magmatism in the Sulu orogen related to continental subduction and exhumation are 237–201 Ma, and those related to orogenic extension and collapse are in the late Jurassic (160–150 Ma), with for a total time span of ~85 Myr.

The 140–120 Ma bimodal magmas are ubiquitous in the Dabie-Sulu orogenic belt and adjacent regions, including early 140–130 Ma adakitic felsic intrusive rocks (Xu et al., 2007; Wang et al., 2007; He et al., 2011) and late 130–120 Ma mafic-ultramafic and felsic intrusive rocks (Zhao et al., 2005, 2007). The former intrusions are considered to be partial melts of thickened lower continental crust, while the latter intrusions are derived from metasomatized lithospheric mantle and deeply subducted continental crust, respectively (Zhao and Zheng, 2009; Zhao et al., 2013). However, these magmatic rocks may be identical to the coeval Yanshanian magmatism widespread in eastern China, which may represent the products of magmatism in response to the westward subduction of the Pacific plate (Zhao et al., 2013). They do not belong to the independent products of Dabie post-collisional tectonic events (Ratschbacher et al., 2000; Niu, 2005, 2014).

3.5 Alps orogen

Alps is also a composite orogen recording evolution from oceanic subduction to continental collision. According to the available geochronological data, the ages of HP and UHP metamorphism seems confused (Berger et al., 2008). HP metamorphism in Sesia-Lanzo, western Alps occurred in 79–65 Ma (Rubatto et al., 1999, 2011). Ages of HP/UHP metamorphism in Piemont-Liguria are 62–38 Ma (Berger et al., 2008). UHP eclogites in Ophilites in Zermatt-Saas have ages of 44 Ma (Rubatto et al., 1998). Sm-Nd isochron age of garnet peridotite and eclogite in Swiss Alps is 40 Ma (Becker, 1993). UHP Age of eclogite in Alpe Arami is 35.8 Ma (Gebauer et al., 1997), and age of eclogite in Dora Maira is 35 Ma (Rubatto et al., 2001). Collision-related mafic rocks and granites have ages of 42–25 Ma with a relatively short time span of 17 Myr. Davies and von Blanckenburg (1995) considered these magmatic rocks to be partial melts of continental lithosphere during exhumation, which is 15 Myr later than the initial subduction of continental crust.

3.6 Himalaya orogen

Continuous northward movement of Indian plate and intracontinental thrust made an over-thickened continental crust of 65–80 km in the Himalaya orogen (Brandon et al., 1986; Holt et al., 1990). Collision between Indian and Eurasian continents occurred at 70–65 Ma, and the UHP metamorphic ages, the time when continental crust subducted to 80–100 km, is about 53–46 Ma (Leech et al., 2005), while no younger UHP rocks are found. In the entire High Himalaya-Tethys Himalaya-Gondese magmatic belt from Nepal to southern Tibetan Plateau, the earliest collision-related magmatism is the Linzizong volcanics with ages of 64–40 Ma, which were considered to represent syn-collisional partial melting of oceanic crust or juvenile continental crust (Mo et al., 2008; Niu et al., 2013). The 45–40 Ma magmatic rocks mainly crops out in core complex in Tethys Himalaya, including adakitic granites (Zeng et al., 2011) and S-type leucogranites (Aikman et al., 2008), representing partial melts of middle/lower crust amphibolite and supracrustal rocks, respectively. In High Himalaya, magmatism continued from 32 to 12 Ma (Harrison et al., 1997; Harris et al., 2004; Searle et al., 2003; Imayama et al., 2012; Rubatto et al., 2013). Metamorphism and anatexis of HP granulites in eastern Himalaya syntaxis initiated at 40 Ma and ended to 8 Ma, lasting for ~32 Myr (Ding et al., 2001; Zhang et al., 2014b). These magmatic rocks are mainly leucogranite from the thickened crust; their strongly peraluminous feature indicates a dominant pelite source formed by dehydration of micas under middle-to high-pressure (middle to lower crust level) granulite-facies conditions, reflecting long-lived crustal thickening and anatexis (Zhang et al., 2014b). Syn-collisional magmatism started from 64 to 8 Ma, lasting for 56 Myr. Because continuous plate convergence is still proceeding, current magma activity is not capable of reflecting the overall syn- and post-collisional magmatism in the Himalayan Orogen.

Figure 8 shows time spans of syn- and post-collisional magmatism in typical collisional orogenic belts. Except for short time in the Cenozoic Alps Orogen (~17 Myr), most last for 60–85 Myr, but no longer than 100 Myr.

4 Magmatism in continental orogen and the net growth of continental crust

4.1 Orogenic adakites and relations with Archean TTGs

It is generally accepted that adakites were generated in Cenozoic arc settings. They were produced by partial melting of subducted/subducting basaltic oceanic crust in some unique tectonic settings, e.g., initial subduction, low-angle subduction, slab window related to slab breakoff and subduction of oceanic plateau. They are characterized by relatively high SiO_2 (>56 wt.%), Al_2O_3 (>5 wt.%), Na_2O (Na_2O>K_2O), Sr contents and high Sr/Y 和$(La/Yb)_N$ ratios (Defant et al., 1990; Castillo, 2006; Martin et al., 2014), which are distinct from typical calc-alkaline arc volcanic rocks (Martin, 1999; Martin et al., 2005). However, investigations of some rocks with adakitic geochemical features suggest that adakitic rocks can be generated by fractionation of basaltic parental magmas at high pressure, which are not direct products of slab melting (Castillo, 2006, 2012; Macpherson, 2006).

Recently, some adakitic rocks were also reported in non-arc settings. These adakitic rocks mainly occur in continental magmatic arcs, collisional orogens and intracratonic settings, which were believed to be produced by partial melting of mafic lower crust (Chung et al., 2003; Zeng et al., 2011; Castillo et al., 2012). It should be noted that in collisional orogens the earlier subducted oceanic crust would experience decompression melting during continental subduction and exhumation, generating adakitic rocks (e.g., Qilian and North Qaidam; Chen et al., 2012; Song et al., 2014b). In contrast, adakitic rocks in intracratonic settings were mainly sourced from partial melting of mafic lower crust (e.g., the North China Craton; Chen et al., 2013; Wang et al., 2015). Nd and Hf isotopic compositions of adakitic rocks are important markers to distinguish their sources: depleted oceanic crust or ancient lower crust. However, Zheng (2012) suggested that typical adakites with negative anomalies of Nb, Ta and Ti cannot be accomplished directly by partial melting of oceanic basaltic crust, but require partial melting of the mantle wedge metasomatized by slab fluids.

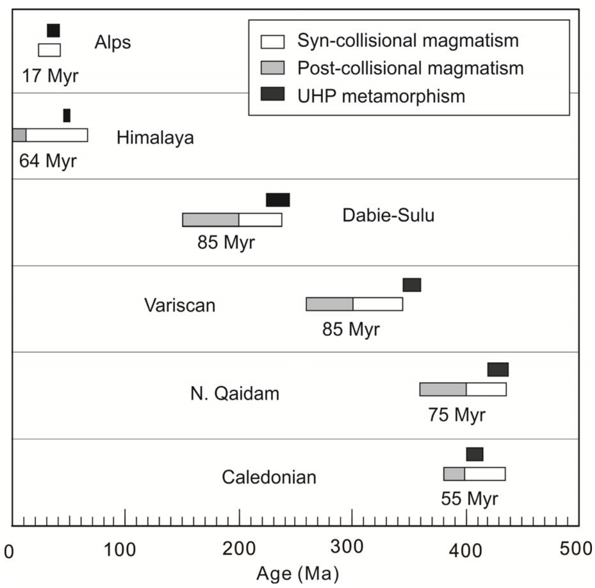

Figure 8 Time spans of magmatism and UHP metamorphism in typical collisional orogens.

TTGs are a suite of felsic rocks, including tonalites, trondhjemites and granodiorites, which are the main components of Archean continental crust formed during 4.0–2.5 Ga (Jahn et al., 1981; Rudnick et al., 2004). TTGs mainly occur as grey gneisses in the Archean high-grade terranes and granitic gneisses in the Archean granite-greenstone belts. TTGs have relatively high SiO_2 (>64 wt.%, usually ~70 % or higher) and Na_2O (K_2O/Na_2O<0.5) (Martin et al., 2005), with Mg# (~0.43), Ni (14 ppm) and Cr (29 ppm) slightly lower than adakites. Based on Al_2O_3 contents, TTGs can be classified into high-Al and low Al series (Barker and Arth, 1976) and recent studies show that most TTGs belong to the high Al series (Martin, 1994; Champion et al., 2003). The most unique geochemical feature of TTGs is their high Sr and Eu, low Yb and Y concentrations and strongly fractionated REE patterns with accordingly high Sr/Y and $(La/Yb)_N$ ratios, which means that they were partial melts of mafic rocks at high pressures while garnet and amphibole remained as residues in the source. Like adakites, TTGs have negative anomalies of Nb, Ta and Ti but some have no positive Sr anomalies, implying that in some cases, plagioclase might also be left in the residue (Martin, 1999).

For the genesis of TTGs, it is widely recognized that they were produced by partial melting of mafic rocks at relatively high pressures (P>10 kbar) (Drummond et al., 1990; Martin, 1994). However, controversy still exists about the exact tectonic settings and P-T conditions for the partial melting of mafic rocks.

Slab melting. Because of the geochemical similarities such as high Sr/Y and $(La/Yb)_N$, TTGs have been suggested to be generated by partial melting of subducted oceanic crust just like Cenozoic adakites (Martin, 1999). Even though the depth for the generation of TTGs is assumed to be over the stability of garnet, it is still debatable whether the residue is garnet amphibolite or eclogite. Foley et al. (2002) conducted partial melting experiments to investigate HFSE partition between TTG melts and residues and proposed that the high Zr/Sm and low Nb/Ta ratios of TTGs are consistent with garnet amphibolite residues. However, Rapp et al. (2003) proposed that the high Zr/Sm and low Nb/Ta ratios of TTGs are controlled by the sources and it is appropriate for the generation of TTGs from partial melting of mafic rocks with eclogite residues.

Melting of tectonically thickened lower crust in arc settings. Smithies (2000) proposed that there are significant compositional differences between TTGs and Cenozoic adakites and it is not plausible to take Cenozoic adakites as analogue of TTGs and attribute TTGs as the products of slab melting. The generation of TTGs might be through partial melting of thickened arc crust like the adakitic rocks in Andes. The thickening of arc crust was achieved by thrusting during subduction. Based on thermodynamic calculation and geochemical modeling, Nagel et al. (2012) put forward that partial melts of tectonically thickened lower crust in arc settings at 10–14 kbar is much more similar to TTGs than slab melts.

Partial melting of mafic lower crust related to mantle plume. It is speculated that partial melting of mafic rocks at the base of oceanic plateau (mantle plume-related) could generate TTGs, e.g., TTGs in the Minto Block, Superior Province (Bédard, 2006). Dacites in Iceland were produced in the same way (Willbold et al., 2009). However, the partial melting of mafic rocks at the base of oceanic plateau won't be deep enough to produce melts with HREE depletion. Therefore, this mechanism can only explain a few cases of TTGs and cannot serve as a general model for the generation of TTGs (Moyen et al., 2012).

Actually, the geochemistry of TTGs is complicated and variable. Moyen (2011) systematically summarized that TTGs can be classified into three types: high-pressure, mid-pressure and low-pressure on the basis of their compositions. The high-pressure TTGs were related to partial melting of rutile-bearing eclogites; the low-pressure TTGs were related to partial melting of plagioclase-bearing garnet amphibolites with compositions distinct from high-pressure TTGs; the mid-pressure TTGs were related to partial melting of garnet amphibolites with minor or without plagioclase. Correspondingly, based on previous experiments, Moyen et al. (2006) proposed that TTGs can be formed over a large pressure ranges of 10–25 kbar, which is also reflected by the compositional diversity of TTGs. Also the fractionation of amphibole and plagioclase from the TTG parental magmas can also contribute to the compositional diversity of TTGs (Moyen et al., 2012).

4.2 Magmatism in continental collision orogens and the net growth of continental crust

Continental crust covers almost 40% of the Earth's surface and is an important reservoir of incompatible elements. The origin, growth and evolution of continental crust have long been the focus and is still a debatable issue (Rudnick, 1995; Harrison et al., 2005; Iizuka et al., 2007), especially the rate and mechanism of continental growth (Taylor, 1977; Armstrong, 1981; Allègre et al., 1984; Taylor et al., 1995; Niu et al., 2013). It is widely acknowledged that the continental growth has lasted from the Early Archean to present-day, and was associated with several giant episodic magmatic events (McCulloch et al., 1994; Condie, 1998). The estimates of the continental growth rate in various models are different but the Archean is believed to be an important period of continental growth (Armstrong, 1981; Allègre et al., 1984; Taylor et al., 1995; Belousova et al., 2010). Whole rock Nd and zircon Hf isotopes of granites and Os isotopes of mantle xenoliths reveal that at least 70% of continental crust were formed between 3.5–2.5 Ga (Condie et al., 2009, 2010) and most of them were TTGs. However, the average composition of continental crust is not similar to TTGs, implying that the contribution of TTGs to the continental crust is not significant (Niu et al., 2013).

As there is the similarity between trace element compositions of continental crust and arc magmas, such as depletion of fluid-immobile elements and relative enrichment of fluid-mobile elements, it is widely accepted that arc magmatism is the main mechanism for the continental growth and continental crust is growing with time (Taylor, 1967; Taylor et al., 1985; McCulloch et al., 1994; Arculus, 1981). The net continental growth was thought to be accomplished in island arc and continental arc, with continental collisional orogens as the site for the reworking of the continental crust (Cawood et al., 2009; Zheng, 2012). However, two aspects should be noted. (1) The overall composition of arc crust is basaltic/mafic (Arculus, 1981; Pearcy et al., 1990), including tholeiitic/calc-alkaline volcanic rocks, boninites and high-Mg andesites with minor felsic volcanic rocks, and they were sourced from partial melting of mantle wedge (Gill, 1981). The overall composition of continental crust is andesitic/intermediate (Rudnick and Gao, 2003) and there is some discrepancies between the compositions of arc crust and continental crust (e.g., Sr and Nb/Ta). (2) During subduction, the erosion of arc crust by slab and the subduction recycling of arc crustal sediments are almost equivalent to the contribution of arc magmatism to arc crust (Scholl et al., 2007; Niu et al., 2009; Stern et al., 2010) and in some extreme cases, there is no net continental growth in arcs. Therefore Niu and his collaborators challenged this traditional viewpoint of continental growth and proposed that the arc magmatism is not an ideal model for the continental growth (Niu et al., 2009, 2013).

The Archean crust is dominated by tonalites and trondhjemites (TT). High *P-T* experiments, field observations and geochemical modeling show that there is obvious compositional similarities between Archean TT and Cenozoic adakites, and the TT is the partial melts of the first differentiated mafic crust with compositions similar to oceanic crust or oceanic plateau, rather than partial melting of upper mantle (Martin et al., 2005). As discussed above, magmatism in continental collision orogens is the reworking of the deep subducted continental crust (Zhao et al., 2004, 2007; Zhao and Zheng, 2009; Zhang et al., 2010; Wang et al., 2014) and only a few mafic magmatic rocks are partial melts of the overlying mantle wedge metasomatized by melts of deep subducted continental crust (Zhao et al., 2013). Investigations of the Linzizong syn-collisional volcanic rocks in Tibetan Plateau and syn-collisional adakitic rocks in the North Qaidam UHP metamorphic belt reveal that: (1) the composition of syn-collisional magmatic rocks is similar to that of the continental crust; (2) these syn-collisional magmatic rocks are characterized by depleted mantle isotopic signatures. The sources of these syn-collisional magmatic rocks are juvenile materials from mantle. Thus magmatism in continental orogens has been considered as the most appropriate mechanism for the net continental growth (Niu et al., 2007, 2009, 2013; Mo et al., 2008; Song et al., 2014b). The continental crust has experienced weathering and erosion since its formation, and recycled into the mantle through subduction and delamination (Rudnick, 1995). Therefore, when considering continental growth, it is essential to take both the addition of juvenile crustal materials and the recycling of ancient crustal materials and their balance into account (Hawkesworth et al., 2009; Niu et al., 2009; Dhuime et al., 2012). When the addition of juvenile crustal materials surpasses the recycling of ancient crustal materials, there will be real growth of continental growth, i.e., net continental growth (Niu et al., 2009, 2013). From this standpoint, identification of the recycling of ancient oceanic crust in the deep continental subduction might be the key to understand net continental growth in continental orogens (Niu et al., 2013; Dai et al., 2014; Song et al., 2014b).

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