

# Lithotectonic elements and geological events in the Hengshan–Wutai–Fuping belt: a synthesis and implications for the evolution of the Trans-North China Orogen

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**Abstract** – The Hengshan–Wutai–Fuping belt is located in the middle segment of the Trans-North China Orogen, a Palaeoproterozoic continental collisional belt along which the Eastern and Western blocks amalgamated to form the North China Craton. The belt consists of the medium- to high-grade Hengshan and Fuping gneiss complexes and the intervening low- to medium-grade Wutai granite–greenstone terrane, and most igneous rocks in the belt are calc-alkaline and have affinities to magmatic arcs. Previous tectonic models assumed that the Hengshan and Fuping gneiss assemblages were an older basement to the Wutai supracrustal rocks, but recent studies indicate that the three complexes constitute a single, long-lived Neoproterozoic to Palaeoproterozoic magmatic arc where the Wutai Complex represents an upper crustal domain, whereas the Hengshan and Fuping gneisses represent the lower crustal components forming the root of the arc. The earliest arc-related magmatism in the belt occurred at 2560–2520 Ma, marked by the emplacement of the Wutai granitoids, which was followed by arc volcanism at 2530–2515 Ma, forming the Wutai greenstones. Extension driven by widespread arc volcanism led to the development of a back-arc basin or a marginal sea, which divided the belt into the Hengshan–Wutai island arc (Japan-type) and the Fuping relict arc. At 2520–2480 Ma, subduction beneath the Hengshan–Wutai island arc caused partial melting of the lower crust to form the Hengshan tonalitic–trondhjemitic–granodioritic (TTG) suites, whereas eastward-directed subduction of the marginal sea led to the reactivation of the Fuping relict arc, where the Fuping tonalitic–trondhjemitic–granodioritic suite was emplaced. In the period 2360–2000 Ma, sporadic phases of isolated granitoid magmatism occurred in the Hengshan–Wutai–Fuping region, forming 2360 Ma, *c.* 2250 Ma and 2000–2100 Ma granitoids in the Hengshan Complex, the *c.* 2100 Ma Wangjiahui and Dawaliang granites in the Wutai Complex, and the 2100–2000 Ma Nanying granitoids in the Fuping Complex. At *c.* 1920 Ma, the Hengshan–Wutai island arc underwent an extensional event, possibly due to the subduction of an oceanic ridge, leading to the emplacement of pre-tectonic gabbroic dykes that were subsequently metamorphosed, together with their host rocks, to form medium- to high-pressure granulites. At 1880–1820 Ma, the Hengshan–Wutai–Fuping arc system was juxtaposed, intensely deformed and metamorphosed during a major and regionally extensive orogenic event, the Lüliang Orogeny, which generated the Trans-North China Orogen through collision of the Eastern and Western blocks. The Hengshan–Wutai–Fuping belt was finally stabilized after emplacement of a mafic dyke swarm at 1780–1750 Ma.

**Keywords:** Archaean, Palaeoproterozoic, magmatic arc, subduction, collision, Hengshan–Wutai–Fuping belt, Trans-North China Orogen.

## 1. Introduction

Major advancements in understanding the geological history of the North China Craton have been made in the past few years, following recognition of a major Palaeoproterozoic orogenic belt, called the Trans-North China Orogen (Fig. 1), which divides the craton into the Eastern and Western blocks (Zhao *et al.*

2001*b* and references therein). The Trans-North China Orogen is characterized by dominant Neoproterozoic to Palaeoproterozoic arc-related juvenile crust with minor reworked basement rocks, fragments of ancient oceanic crust, mélanges, high-pressure granulites, retrograded eclogites, crustal-scale ductile shear zones, and linear fold belts with sheath folds (Li *et al.* 1990; Li & Qian, 1991; Zhai, Guo & Yan, 1992; Wang *et al.* 1996, 1997; Wu & Zhong, 1998; Zhao, Cawood & Lu, 1999; Zhao *et al.* 2000*b*; Guo, O'Brien & Zhai, 2002). These

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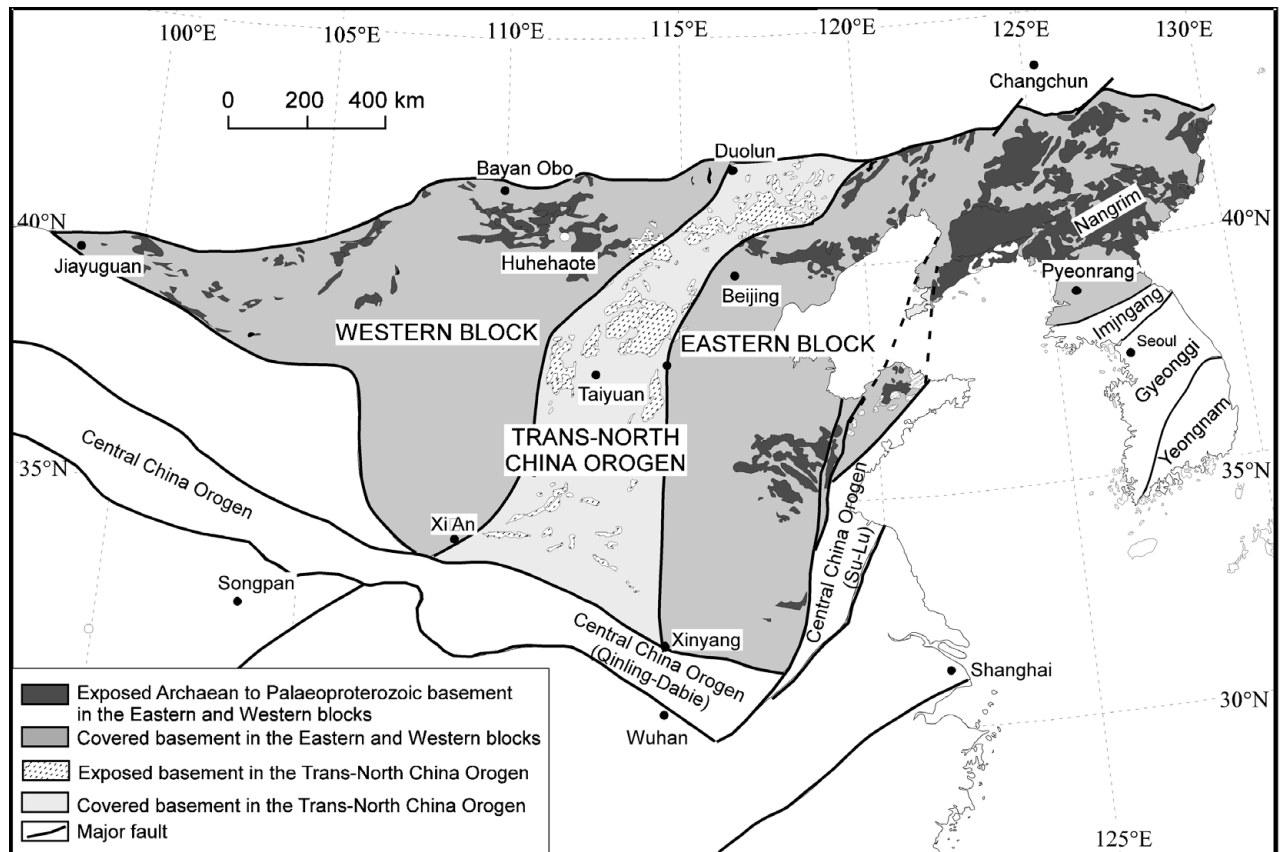


Figure 1. Three-fold tectonic subdivision of the North China Craton proposed by Zhao *et al.* (1998, 2001b).

lithotectonic elements contrast with the dominant Archaean tonalitic–trondhjemitic–granodioritic (TTG) gneiss domes surrounded by minor supracrustal rocks in the Eastern and Western blocks (Jahn & Zhang, 1984; Jahn *et al.* 1987, 1988; Jahn & Ernst, 1990; Kröner *et al.* 1998; Zhao *et al.* 1998; Ge *et al.* 2003; Wu *et al.* 2005; Xia *et al.* 2006a,b). In addition, petrographic and thermobarometric data have revealed a remarkable difference in metamorphic evolution between rocks in the Trans-North China Orogen and the Eastern and Western blocks (Zhao *et al.* 1998, 1999). The former underwent a major metamorphic event at *c.* 1.85 Ga with clockwise *P–T* paths involving isothermal decompression and suggesting a collisional environment (Zhao *et al.* 2000a), whereas the latter experienced a major metamorphic event at *c.* 2.5 Ga, with anticlockwise *P–T* paths involving isobaric cooling, possibly related to underplating of mantle-derived magmas (Zhao *et al.* 1998, 1999). These differences led Zhao *et al.* (2001b and references therein) to propose that the Trans-North China Orogen was a continental collisional belt along which the Eastern and Western blocks amalgamated to form the coherent North China Craton at about 1.85 Ga (Zhao, 2001; Zhao *et al.* 2005 and references therein). According to Zhao (2001), the western margin of the Eastern Block faced a major ocean during Neoproterozoic to Early Palaeoproterozoic times, and eastward-directed subduction of the oceanic lithosphere beneath

the western margin of the Eastern Block led to the formation of a magmatic arc that was subsequently incorporated into the Trans-North China Orogen during collision between the Eastern and Western blocks. There is now a coherent outline of timing and tectonic processes involved in the collision between the Eastern and Western blocks (Zhao *et al.* 1998, 1999; Wu *et al.* 2000; Wan *et al.* 2006), but the detailed tectonic architecture and pre-collisional history of the Trans-North China Orogen have not been well constrained, and a number of key geological issues related to the formation and evolution of this orogen are not well resolved. These issues include the relationships between the high-grade gneiss complexes and low-grade granite–greenstone terranes in the orogen, the nature of the magmatic arc system that was subsequently incorporated into the Trans-North China Orogen and the enigma of what magmatic and tectono-metamorphic events took place in the orogen following development of the 2.55 Ga arc and prior to collision of the Eastern and Western blocks at *c.* 1.85 Ga, a time span of around 700 million years.

As the largest and most lithologically representative basement exposure across the Trans-North China Orogen, the Hengshan–Wutai–Fuping belt is the most promising area for investigating the detailed magmatic, depositional, structural and metamorphic history of the orogen. Of particular significance is the presence of a low- to medium-grade granite–greenstone terrane

(Wutai Complex) intervening between two high-grade gneiss complexes (Hengshan and Fuping complexes). For these reasons, geologists from several countries have carried out extensive geological investigations in the Hengshan–Wutai–Fuping belt and obtained a wealth of new geochronological data and competing interpretations (e.g. Wilde, Cawood & Wang, 1997, 1998; Wilde, Zhao & Sun, 2002; Wilde *et al.* 2004a,b, 2005; Halls, 2000; Kröner, 2002; Kröner *et al.* 2001, 2005a,b, 2006; Kusky & Li, 2003; Passchier & Walte, 2002; Zhao *et al.* 2002b; Guan *et al.* 2002; Liu *et al.* 2000, 2002a,b, 2004a,b, 2005, 2006; Wang *et al.* 2001, 2003, 2004a,b; O'Brien, Walte & Li, 2005; P. Peng, unpub. Ph.D. thesis, Chinese Acad. Sciences, Beijing, 2005; Peng *et al.* 2005; Polat *et al.* 2005, 2006), which enable resolution of major lithotectonic elements and geological events that occurred in the belt. Here we present a summary and overview of these lithotectonic assemblages and geological events based on field relationships and recent geochronological data, especially SHRIMP U–Pb zircon ages, which provide important insights into understanding the above key issues related to the Neoproterozoic to Palaeoproterozoic evolution of the Trans-North China Orogen.

## 2. Regional setting

The North China Craton is a general term used to refer to the Chinese part of the Sino-Korean Platform or Craton. It covers about 1.5 million square kilometres in most of northern China, the southern part of northeastern China, Inner Mongolia, the Bohai Bay and the northern part of the Yellow Sea. The craton is bounded by the Early Palaeozoic Qilianshan Orogen and late Palaeozoic to Mesozoic Central Asian Orogenic Belt to the west and north, respectively, and the Mesozoic Qinling–Dabie and Su–Lu ultrahigh-pressure metamorphic belts to the south and east, respectively (Fig. 1). As mentioned above, the basement of the North China Craton has been divided into the Archaean to Palaeoproterozoic Eastern and Western blocks, separated by the Palaeoproterozoic Trans-North China Orogen (Fig. 1). Lithological, geochemical, structural, metamorphic and geochronological differences between the basement rocks of the Eastern and Western blocks and the Trans-North China Orogen have been summarized by Zhao *et al.* (2001a) and are not repeated here. Recently, the three-fold subdivision and tectonic model of the North China Craton have been further refined and modified by new structural, petrological and geochronological data obtained over the past few years (Fig. 2; Zhao *et al.* 2005). These data indicate that the Western Block formed by amalgamation of the Ordos Terrane in the south and the Yinshan Terrane in the north along the EW-trending Khondalite Belt at some time prior to the collision of the Western and Eastern blocks (Fig. 2; Zhao *et al.* 2005). The data also suggest that the Eastern Block underwent

Palaeoproterozoic rifting along its eastern continental margin in the period 2.2–1.9 Ga, and the closure of this rift basin at *c.* 1.9 Ga led to the formation of the Jiao-Liao-Ji Belt (Fig. 2; Li *et al.* 2001a,b, 2004a,b, 2005; Hao *et al.* 2004; Luo *et al.* 2004; Lu *et al.* 2006; Zhao *et al.* 2006). It remains unknown whether oceanic crust formed in the Jiao-Liao-Ji Belt during its rifting stage, though Du, Wang & Wang (1999) interpreted some ultramafic–mafic rocks as remnants of ophiolites.

The Trans-North China Orogen, a nearly NS-trending zone, about 1200 km long and 100–300 km wide (Fig. 3), is separated from the Eastern and Western blocks by the Xingyang–Kaifeng–Shijiazhuang–Jianping Fault and the Huashan–Lishi–Datong–Duolun Fault, respectively. Both faults strike NS in the central and southern parts and NE in the north. The presence of voluminous mantle-derived basalts exposed along the two faults suggests that they are deep structures, possibly reaching into the lower crust or upper mantle (Ren, 1980). The main lithotectonic features of the Trans-North China Orogen include: (1) dominant Neoproterozoic to Palaeoproterozoic arc-related juvenile crust with minor reworked basement rocks (Kröner *et al.* 1988; Sun, Armstrong & Lambert, 1992; Zhao *et al.* 2000a); (2) linear structural belts defined by strike-slip ductile shear zones, large-scale thrusting and folding, and transcurrent tectonics (Li & Qian, 1991; Zhang, Dirks & Passchier, 1994; Zhao *et al.* 1999); (3) sheath folds and mineral lineations (Wu & Zhong, 1998); (4) high-pressure granulites and retrograde eclogites (Zhai, Guo & Yan, 1992; Guo *et al.* 1999, 2001; Guo, O'Brien & Zhai, 2002; Guo, Sun & Zhai, 2005; Guo & Zhai, 2001; Zhao *et al.* 2001a; Zhang *et al.* 2006); (5) clockwise metamorphic *P–T* paths involving near-isothermal decompression (Liu, Shen & Geng, 1996; Zhao *et al.* 2000a; Guo, Sun & Zhai, 2005); (6) ancient oceanic fragments and ophiolitic mélange (Bai, 1986; Li *et al.* 1990; Bai, Wang & Guo, 1992; Wang *et al.* 1996, 1997; Wu & Zhong, 1998); (7) syn- or post-tectonic granites; and (8) post-collisional mafic dyke swarms (Halls *et al.* 2000; P. Peng, unpub. Ph.D. thesis, Chinese Acad. Sciences, Beijing, 2005; Peng *et al.* 2005). Most of these lithotectonic elements are classical indicators of collision tectonics.

The Hengshan–Wutai–Fuping belt is situated in the middle segment of the Trans-North China Orogen (Fig. 3) and consists of three distinct tectonic complexes: the upper amphibolite to granulite facies Fuping and Hengshan complexes in the southeast and northwest, respectively, separated by the greenschist- to lower amphibolite-facies Wutai Complex (Fig. 4), which is interpreted as a typical granite–greenstone belt (Bai, 1986; Bai, Wang & Guo, 1992; Tian, 1991). The Hengshan Complex is bounded in the northwest by the valley of the Sanggan River and is separated from the Wutai Complex by the broad valley of the Hutuo River and thus the nature of the Hengshan–Wutai boundary is unclear. The Fuping Complex was previously considered

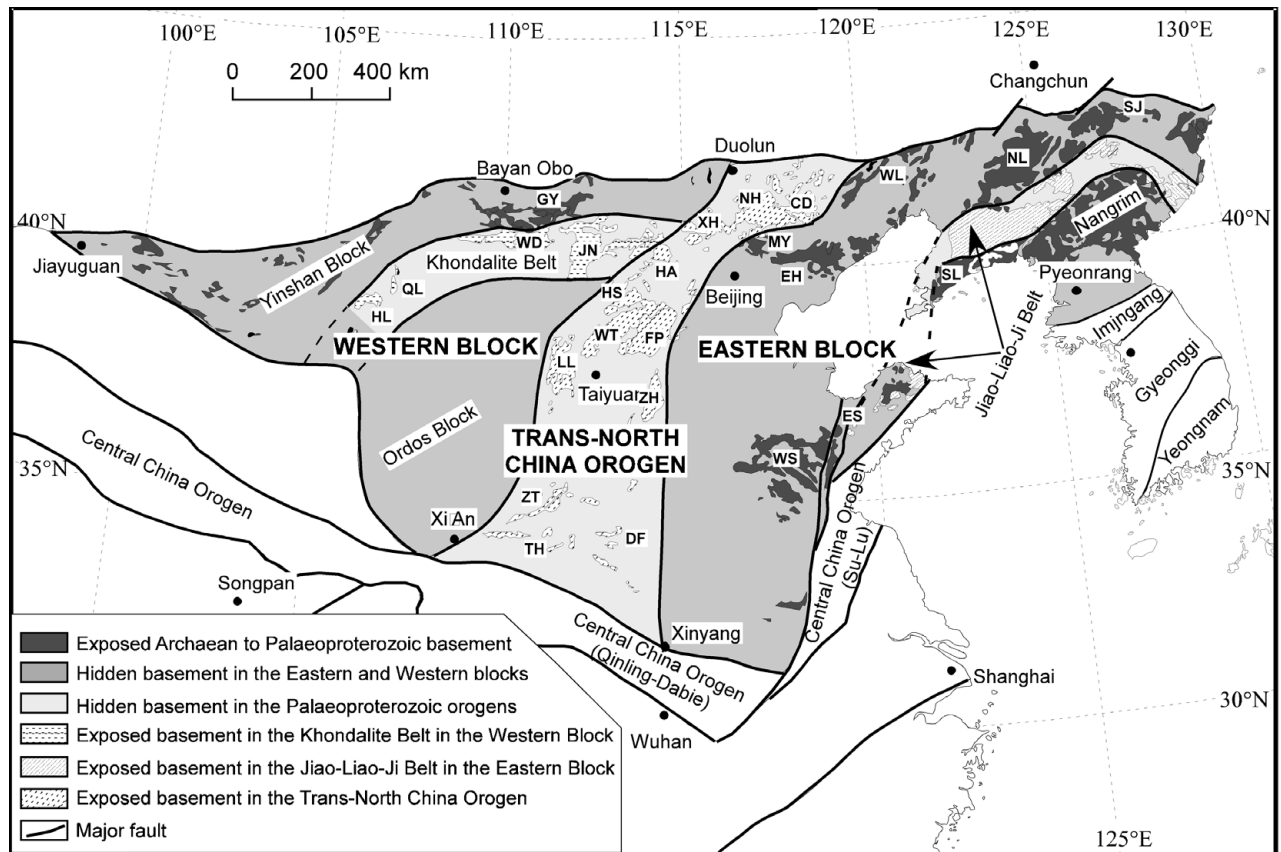


Figure 2. Tectonic subdivision of the North China Craton, modified by Zhao *et al.* (2005). Abbreviations for metamorphic complexes: CD – Chengde; DF – Dengfeng; EH – Eastern Hebei; ES – Eastern Shandong; FP – Fuping; GY – Guyang; HA – Huai’an; HL – Helanshan; HS – Hengshan; JN – Jining; LL – Lüliang; MY – Miyun; NH – Northern Hebei; NL – Northern Liaoning; QL – Qianlishan; SJ – Southern Jilin; SL – Southern Liaoning; TH – Taihua; WD – Wulashan–Daqingshan; WL – Western Liaoning; WS – Western Shandong; WT – Wutai; XH – Xuanhua; ZH – Zanhuang; ZT – Zhongtiao.

to be unconformably overlain by the Wutai Complex along the ‘Tiebao unconformity’ (Bai, 1986; Wu *et al.* 1989; Tian, 1991; Bai, Wang & Guo, 1992), but recent research revealed that the so-called ‘Tiebao unconformity’ is a regional-scale ductile shear zone, named the Longquanguan Ductile Shear Zone by Li & Qian (1991), and thus the nature of the boundary between the Fuping and Wutai complexes is a tectonic feature.

### 3. Major rock units of the Hengshan–Wutai–Fuping belt

#### 3.a. Hengshan Complex

The Hengshan Complex is composed of four main rock units (Fig. 4): (1) the Hengshan granitoid gneisses, (2) the Hengshan mafic granulites, (3) the Yixingzhai gneisses, and (4) the Zhujiayang supracrustal assemblage (Tian, 1991; Zhao *et al.* 2001a).

The Hengshan granitoid gneisses are strongly deformed layered orthogneisses of dioritic–tonalitic–trondhjemitic–granodioritic–granitic composition. These rocks are extensively migmatized, and some of the migmatized zones show evidence of *in situ*

melting and advanced anatexis, generating reddish granites. The compositional layering in the gneisses ranges from dark, hornblende-rich dioritic to tonalitic compositions to K-feldspar-dominated leucocratic granitoid varieties. Major element data show that the Hengshan granitoid gneisses belong to a high-alumina calc-alkaline suite (Li & Qian, 1994), and geochemical data suggest that they were emplaced in a magmatic arc environment (Fig. 5a; Kröner *et al.* 2005a,b).

Within the Hengshan granitoid gneisses there are discontinuous lenses or layers of amphibolite and high- and medium-pressure mafic granulite, ranging from 0.1 to 5 m in width and 0.1 to 50 m in length and interpreted as boudinaged gabbroic dykes (Kröner *et al.* 2005a,b, 2006). In the field, the high-pressure granulite lenses can be distinguished from the surrounding medium-pressure granulites by the presence of coarse-grained textures and lack of brown orthopyroxene in the former. The long axes of the granulite lenses are parallel to the regional foliation of the Hengshan granitoid gneisses. In some places, the interiors of amphibolites and mafic granulites preserve a relict igneous gabbroic/doleritic texture. There can be no doubt that these amphibolites and mafic granulites are remnants of mafic dykes that

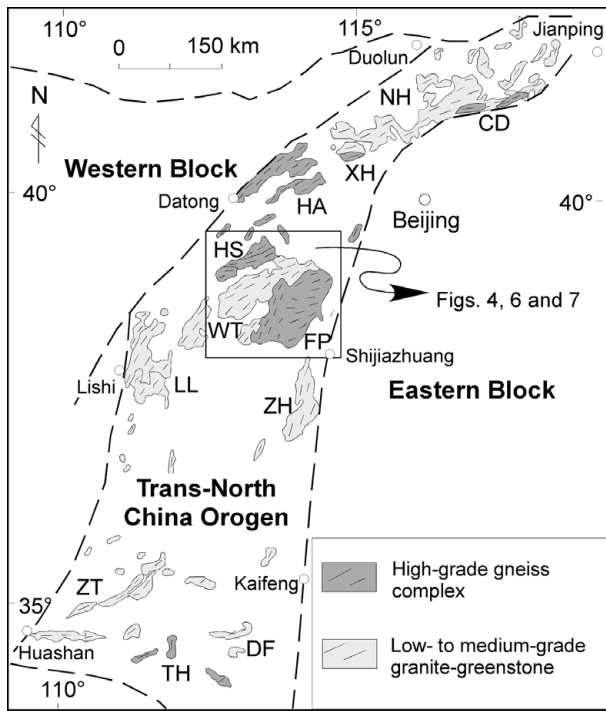


Figure 3. Simplified tectonic map showing the distribution of metamorphic complexes in the Trans-North China Orogen (after Zhao *et al.* 2000a). Area of Figures 4, 6 and 7 is outlined. Symbols and abbreviations as in Figure 2.

originally intruded into the granitoid rocks, as can still be observed at a few localities in low-strain zones (Kröner *et al.* 2005a,b, 2006).

The Yixingzhai gneisses occur in the southern part of the Hengshan Complex. They are chemically similar to the Hengshan granitoid gneisses, belonging to a high-alumina calc-alkaline suite and forming in a magmatic arc setting (Kröner *et al.* 2005a,b), but are only metamorphosed to greenschist to lower amphibolite facies and are less deformed, locally preserving igneous textures.

The Zhujiayang assemblage consists predominantly of amphibolite, felsic gneiss, mica schist, banded iron formation (BIF) and quartzite, which occur largely along two nearly EW-trending belts that cut across the regional layering/foliation of the Hengshan granitoid gneisses and Yixingzhai gneisses (Fig. 4). These rocks are characterized by strong ductile deformation and ubiquitous mylonite fabrics and are distinctively lower in metamorphic grade than the other supracrustal rocks within the Hengshan granitoid gneisses. Some Chinese workers have considered the Zhujiayang assemblage to be the equivalent of the Wutai greenstones in the Hengshan area (Tian, 1991; Li & Qian, 1994).

### 3.b. Wutai Complex

The Wutai Complex consists of Neoproterozoic to Palaeoproterozoic granitic plutons and metamorphosed

volcanic and sedimentary rocks, traditionally named the Wutai and Hutuo 'groups' in the Chinese literature.

The granitoid plutons in the Wutai Complex can be largely divided into three distinct types: the Wutai, Dawaliang and Fengkuangshan granites. The Wutai granites are the major components of the granitoid plutons in the Wutai Complex, including the Chechang–Beitai, Dazhaikou, Duyu, Ekou, Shifu, Lanzhishan, Guangmingshi and Wangjiahui bodies. They are composed of strongly deformed diorite–tonalite–trondhjemite–granodiorite suites with penetrative foliations, and thus are considered to be pre-tectonic granitoids (Tian, 1991). Geochemical data suggest that the Wutai granites were emplaced in a magmatic arc environment (Fig. 5a; Liu *et al.* 2004b, 2005; Kröner *et al.* 2005b). The Dawaliang-type granites, represented by the Dawaliang body and the younger phase of the Wangjiahui body, are composed of weakly deformed porphyric syenogranites. The Fengkuangshan-type granites, including the Fengkuangshan, Lianhuashan and Pingxingguan bodies, are composed of undeformed A-type granites with a massive structure, interpreted as post-orogenic or anorogenic granites (Tian, 1991).

Based on lithologies and metamorphic grades, the Wutai 'Group' is conventionally subdivided into three subgroups: Shizui, Taihuai and Gaofan subgroups. The Shizui Subgroup consists of peridotite, oceanic tholeiite, dacite, rhyolite, chert, banded iron formation (BIF), sandstone, siltstone, shale, calc-silicate rocks and minor limestone metamorphosed in lower amphibolite facies. Of these, the peridotites, oceanic tholeiites and cherts are considered to represent relict oceanic crust, whereas most sediments are interpreted as continental margin or back-arc basin deposits (Li *et al.* 1990; Bai & Dai, 1998; Wu & Zhong, 1998). The Taihuai Subgroup comprises felsic volcanic rocks and tholeiites metamorphosed in greenschist facies. On the Pearce (1976) major element discrimination diagram (Fig. 5b), most of the felsic and tholeiitic volcanics in the Wutai 'Group' have affinities to modern volcanic-arc assemblages or MORB (Bai, 1986; Wang *et al.* 2004b). On the Th/Ta v. La/Yb diagram (Condie, 1994, 1997, 2005), most of the amphibolite samples from the Wutai 'Group' plot in the arc field, though a few samples are located in the NMORB field and its adjacent areas (Fig. 5c; Wang *et al.* 2004b). The Gaofan Subgroup contains conglomerates, quartz greywackes, siltstones, limestones, and minor mafic to felsic volcanic rocks metamorphosed from subgreenschist to greenschist facies, which are interpreted as having developed in an intra-arc basin and/or a retro-arc foreland basin (Zhao *et al.* 2001b). Recent SHRIMP U–Pb zircon ages indicate that the Shizui, Taihuai and Gaofan subgroups are not a single lithostratigraphic sequence, the 'Wutai Group', as previously considered; instead, they are similar-aged volcanic-sedimentary formations that are

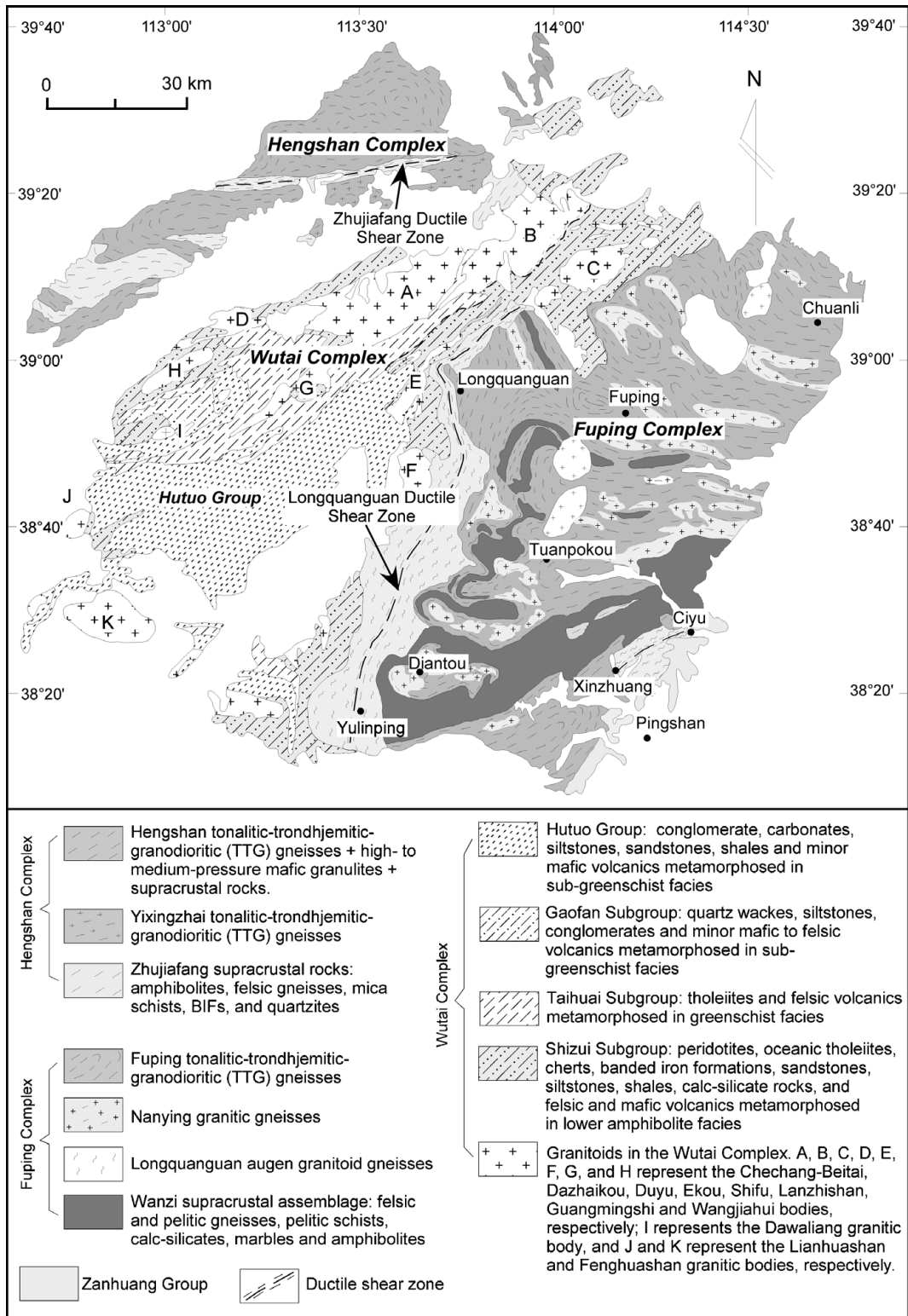


Figure 4. Simplified geological map of the Hengshan, Wutai and Fuping complexes.

structurally disrupted and juxtaposed along a series of NE–SW-trending ductile shear zones (see Section 3.c).

The Hutuo Group is considered to be the youngest unit in the Hengshan–Wutai–Fuping belt and to unconformably overlie the Wutai Group, although in places it is tectonically interleaved with the Wutai sequence. The group ranges upward from an extensive

basal sequence of conglomerate into quartzite and other clastic sediments, slate, dolomite and marble, with another unit of conglomerate at the top of the sequence. Rocks of the Hutuo Group are mostly metamorphosed in greenschist facies, although there may be an eastward increase in grade, reaching upper greenschist to amphibolite facies.

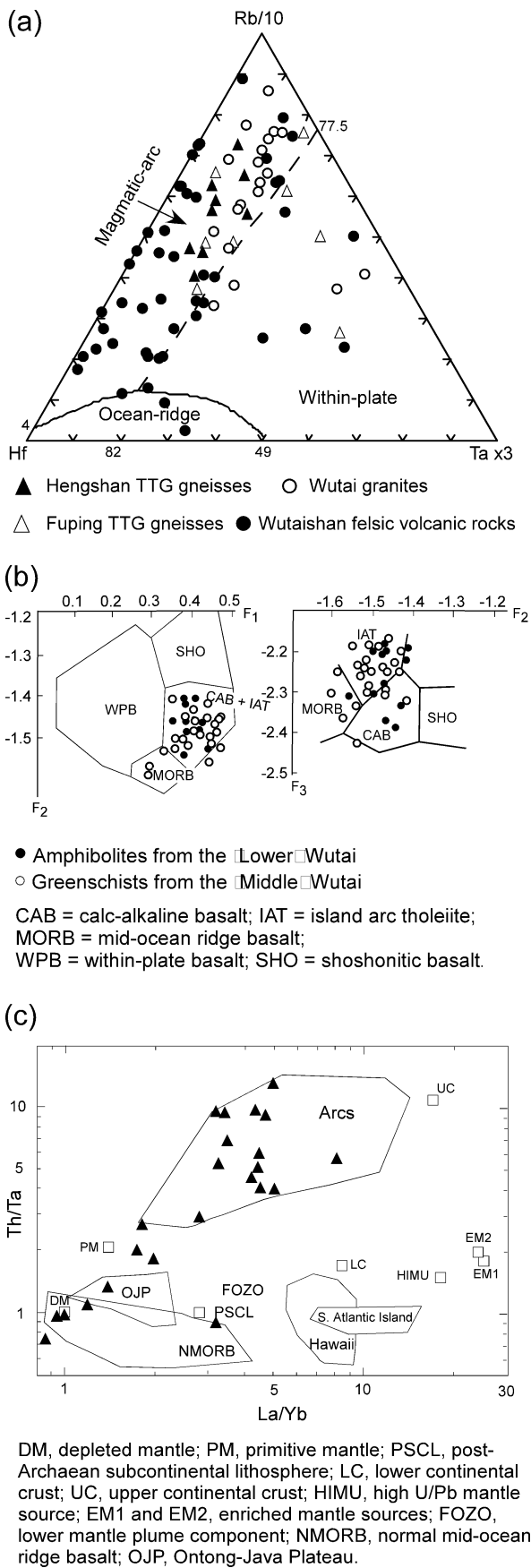


Figure 5. (a) Hf-Rb/10-Ta × 3 discrimination diagram (after Harris, Pearce & Tindle, 1986) for Hengshan TTG gneisses (Kröner *et al.* 2005b; Liu *et al.* 2005), Wutai volcanic and

3.c. Fuping Complex

The Fuping Complex comprises four distinct rock units, herein called the Fuping granitoid gneisses, Longquanguan augen gneisses, Wanzi supracrustal assemblage, and Nanying granitic gneisses (Fig. 4; Liu *et al.* 2000; Zhao *et al.* 2002b).

The Fuping granitoid gneisses make up about 60 % of the complex and consist of medium-grained dioritic, tonalitic, trondhjemitic, granodioritic gneisses enclosing mafic granulite, amphibolite and hornblende gneiss that have undergone a complex history of upper amphibolite- to granulite-facies metamorphism and polyphase deformation. Petrological and geochemical data suggest that, like the Hengshan granitoid gneisses, the Fuping granitoid gneisses formed in a magmatic arc environment (Fig. 5a; Wang, Li & Liu, 1991).

The Longquanguan augen gneisses are mainly exposed along the Longquanguan ductile shear zone and are composed predominantly of coarse-grained to porphyritic granodioritic and monzogranitic gneisses and mylonitized granitic pegmatites containing K-feldspar phenocrysts, most of which have been intensely deformed to form augen. Enclosed in the Longquanguan augen gneisses are amphibolite and hornblende gneiss enclaves, similar to those in the Fuping granitoid gneisses. The Longquanguan augen gneisses display a tectonic contact with the Wanzi supracrustal assemblage and the Fuping gneisses (Wu *et al.* 1989; Li & Qian, 1991).

The Wanzi supracrustal assemblage forms a 100 km long, NE-SW-trending belt in the southern part of the complex (15 km wide) that swings northward to the central part of the complex, where it is extensively folded (Fig. 4). The supracrustal rocks are metamorphosed to amphibolite facies and comprise felsic and pelitic gneisses, pelitic schists, calc-silicate rocks, pure and impure marbles and amphibolites (Zhao *et al.* 2000a,b). Also associated with the supracrustal rocks are several small sillimanite-bearing granites which are considered to represent S-type granites derived from partial melting of pelitic gneisses and felsic paragneisses (Zhao *et al.* 2000a,b).

The Nanying granitic gneisses only occur within the Fuping granitoid gneisses and are dominated by medium- to fine-grained, weakly foliated, magnetite-bearing monzogranitic gneiss with minor granodioritic gneiss (Fig. 4; Zhao *et al.* 2000a,b). In addition to

granitoid rocks (Wang *et al.* 2004b; Liu *et al.* 2005) and Fuping TTG gneisses (Liu *et al.* 2005). (b) Major element discrimination diagram (after Pearce, 1976) for Wutai amphibolites (Lower Wutai) and greenschists (Middle Wutai), based upon functions F<sub>1</sub>, F<sub>2</sub> and F<sub>3</sub> (Bai, Wang & Guo, 1992). (c) Th/Ta v. La/Yb diagram (Condie, 1994, 1997, 2005) for the greenschist to amphibolite facies basaltic volcanics from the Wutai Complex (Wang *et al.* 2004b).

compositional differences, the Nanying gneisses are more massive in structure and more homogeneous in composition than the Fuping granitoid gneisses. In the field where contact relations are preserved, the Nanying granitic gneisses are clearly intrusive into the Fuping granitoid gneisses, but their relatively weak foliation is consistently parallel to the strong, penetrative foliation of the Fuping granitoid gneisses, suggesting that they most likely underwent the same deformational event that resulted in the development of the regional foliation in the Fuping Complex.

#### 4. Major lithotectonic elements and geological events

In the last few years, the SHRIMP zircon dating technique, combined with single-grain zircon evaporation and mineral Sm–Nd and Ar/Ar methods, has been widely applied to date the major rock units of the Hengshan–Wutai–Fuping belt, producing a large number of precise age data for the mountain belt (see references in Table 1). These ages, in conjunction with structural, metamorphic and geochemical considerations, lead us to recognize the following major litho-tectonic assemblages and geological events in the Hengshan–Wutai–Fuping belt.

##### 4.a. Remnants of an old (2.7–2.8 Ga) crustal component

The oldest known basement components in the Hengshan–Wutai–Fuping belt are rare medium-grained felsic gneisses that occur interleaved with the younger granitoid gneisses but are difficult to recognize as separate units in the field (Wilde, Cawood & Wang, 1997; Guan *et al.* 2002; Kröner *et al.* 2005a). In the Fuping Complex, xenocrystic zircons from a medium-grained hornblende gneiss enclosed in the Fuping granitoid gneisses were SHRIMP-dated by Guan *et al.* (2002) at  $2708 \pm 8$  Ma (Fig. 6, Table 1). A sillimanite paragneiss from the Wanzi Supracrustal assemblage contains two ancient detrital zircons of ages  $2827 \pm 8$  Ma and  $2686 \pm 8$  Ma, of which the former, as far as we are aware, is the oldest single-grain zircon age so far reported from the Hengshan–Wutai–Fuping belt (Fig. 6, Table 1). In the Wutai Complex, Wilde, Cawood & Wang (1997) found four xenocrystic zircons in the  $2553 \pm 8$  Ma Lanzhishan granites that yielded a weighted mean SHRIMP  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $2702 \pm 10$  Ma. Wilde (2002) also recognized  $2763 \pm 10$  Ma and  $2660 \pm 7$  Ma old xenocrystic zircons in the  $2534 \pm 10$  Ma Longquanguan augen gneiss (Fig. 6), which is now considered to be part of the Wutai Complex (see Section 4.b). In addition, Wilde *et al.* (2004a) found two xenocrystic zircons with a weighted mean SHRIMP  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $2702 \pm 10$  Ma from a  $2529 \pm 10$  Ma meta-andesite of the Zhuangwang Formation. In the Hengshan Complex, a grey trondhjemitic gneiss sample collected from a migmatite domain in Changchenggou ('Great

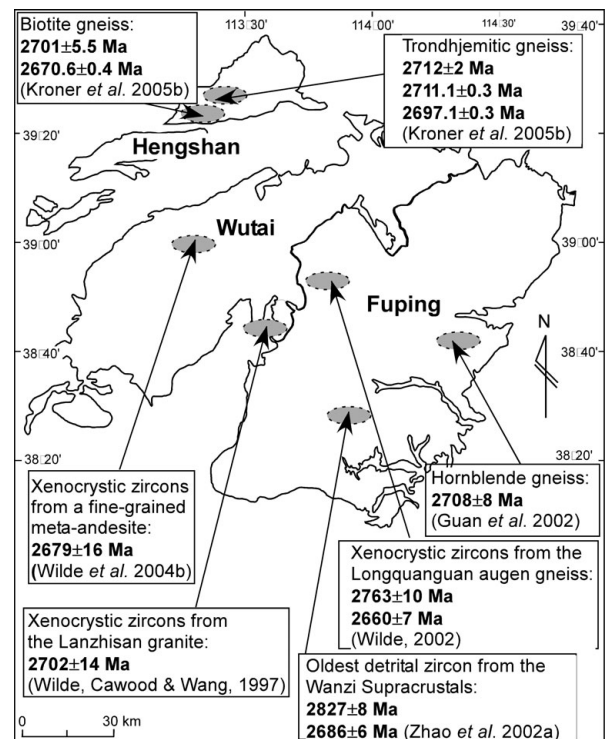


Figure 6. Distribution of 2.7–2.8 Ga xenocrystic zircons or gneisses in the Hengshan–Wutai–Fuping belt.

Wall Valley' in English) and a grey biotite gneiss sample collected from Dashiyu ('Big Stone Valley' in English) were SHRIMP-dated by Kröner *et al.* (2005b) at  $2712 \pm 2$  Ma and  $2701 \pm 5.5$  Ma, respectively (Fig. 6). Kröner *et al.* (2005b) also used the single-grain evaporation technique to date a foliated grey gneiss from Changchenggou and a fine-grained biotite gneiss from Dashiyu at  $2697.1 \pm 0.3$  Ma and  $2670.6 \pm 0.4$  Ma, respectively. Taken together, these data, although sparse, confirm the existence of *c.* 2.7 Ga crustal components in the Hengshan–Wutai–Fuping region, now tectonically interleaved with the younger gneisses and most probably representing remnants of an older crustal domain whose history and evolution is difficult to reconstruct.

##### 4.b. 2560–2520 Ma: emplacement of Neoproterozoic Wutai granites and Longquanguan augen gneisses

The Wutai granites include the Ekou, Chechang–Beitai, Lanzhishan, Shifu, Guangmingshi and Wangjiahui plutons, of which the oldest is the Ekou pluton dated at  $2566 \pm 13$  Ma and  $2555 \pm 6$  Ma (Wilde, Cawood & Wang, 1997), and the youngest is the Wangjiahui pluton dated at  $2517 \pm 12$  Ma (Wilde *et al.* 2005). As the largest exposure of the Wutai granites, the Chechang–Beitai pluton is composed of tonalites and granodiorites. Four samples collected from different phases of the pluton yielded SHRIMP zircon ages of  $2552 \pm 11$ ,  $2551 \pm 5$ ,  $2546 \pm 6$  and  $2538 \pm 6$  Ma (Table 1; Wilde *et al.* 2005), suggesting that the pluton was emplaced



Table 1. Summary of isotopic ages for the Hengshan, Wutai and Fuping complexes

| Sample  | Description   | Age (Ma)     | Method | Sources                      |
|---|---|--------------|--------|------------------------------|
| <b>Ages of remnants of 2.7–2.8 Ga old continental crust</b>   |   |              |        |                              |
| <i>Hengshan Complex</i>   |   |              |        |                              |
| 990843  | Grey trondhjemitic gneiss (Changchenggou)             | 2712 ± 2     | SHRIMP | Kröner <i>et al.</i> (2005b) |
|   | Same sample   | 2711.1 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 990838  | Well foliated grey biotite gneiss (Dashiyu)           | 2701 ± 5.5   | SHRIMP | Kröner <i>et al.</i> (2005b) |
| 980811  | Grey granodioritic gneiss (Changchenggou)             | 2697.1 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 980824  | Finely layered, fine-grained biotite gneiss (Dashiyu) | 2670.6 ± 0.4 | EVAP   | Kröner <i>et al.</i> (2005b) |
| <i>Wutai Complex</i>  |   |              |        |                              |
| 95-PC-114   | Xenocrystic zircons from a meta-andesite (Ekou)       | 2679 ± 16    | SHRIMP | Wilde <i>et al.</i> (2004b)  |
| 95-PC-94  | Xenocrystic zircons from the Lanzhisan granite        | 2702 ± 14    | SHRIMP | Wilde <i>et al.</i> (1997)   |
| <i>Fuping Complex</i>   |   |              |        |                              |
| 95-PC-65  | Xenocrystic zircons from the Longquanguan granite     | 2763 ± 10    | SHRIMP | Wilde (2002)                 |
|   | Same sample   | 2660 ± 7     | SHRIMP | Wilde (2002)                 |
| FP50  | Xenocrystic zircons from a biotite gneiss (Diebuan)   | 2708 ± 8     | SHRIMP | Guan <i>et al.</i> (2002a)   |
| FP260   | Oldest detrital zircons from the Wanzi Supracrustals  | 2826 ± 8     | SHRIMP | Zhao <i>et al.</i> (2002a)   |
|   | Same sample   | 2686 ± 6     | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| <b>Ages of late Archaean Wutai granitoids and Longquanguan augen gneisses (2560–2520 Ma)</b>                  |   |              |        |                              |
| <i>Ekou Pluton</i>  |   |              |        |                              |
| 95-PC-34  | Deformed, pink, medium-grained granitoid (Ekou)       | 2566 ± 13    | SHRIMP | Wilde <i>et al.</i> (1997)   |
| 95-19   | deformed, pink, medium-grained granitoid (Ekou)       | 2555 ± 6     | SHRIMP | Wilde <i>et al.</i> (1997)   |
| <i>Chechang–Beitai pluton</i>   |   |              |        |                              |
| WC7   | Foliated, coarse-grained tonalite (Shahe)             | 2552 ± 11    | SHRIMP | Wilde <i>et al.</i> (2005)   |
| 95-PC-6B  | Foliated, coarse-grained tonalite (Taipinggou)        | 2551 ± 5     | SHRIMP | Wilde <i>et al.</i> (2005)   |
| WC6   | Deformed, fine-grained granodiorite (Taipinggou)      | 2546 ± 6     | SHRIMP | Wilde <i>et al.</i> (2005)   |
| WC5   | Deformed, coarse-grained granodiorite (Taipinggou)    | 2538 ± 6     | SHRIMP | Wilde <i>et al.</i> (2005)   |
| <i>Lanzhishan Pluton</i>  |   |              |        |                              |
| 95-PC-94  | Deformed, coarse-grained granitoid (Changjiagtan)     | 2553 ± 8     | SHRIMP | Wilde <i>et al.</i> (1997)   |
| 95-PC-96  | Deformed, coarse-grained granitoid (Changjiagtan)     | 2537 ± 10    | SHRIMP | Wilde <i>et al.</i> (1997)   |
| <i>Shifo Pluton</i>   |   |              |        |                              |
| 95-PC-98  | Deformed medium-grained monzogranite (Nanliang)       | 2531 ± 4     | SHRIMP | Wilde <i>et al.</i> (2005)   |
| <i>Guangmishi Pluton</i>  |   |              |        |                              |
| 95-PC-76  | Deformed, medium-grained granitoid (Guangmishi)       | 2531 ± 5     | SHRIMP | Wilde <i>et al.</i> (2005)   |
| <i>Wangjiahui Pluton (Grey Phase)</i>   |   |              |        |                              |
| 95-PC-62  | Deformed medium-grained granodiorite (Shi Gang)       | 2520 ± 9     | SHRIMP | Wilde <i>et al.</i> (2005)   |
| 95-PC-63  | Deformed medium-grained granodiorite (Shi Gang)       | 2517 ± 12    | SHRIMP | Wilde <i>et al.</i> (2005)   |
| <i>Longquanguan augen gneisses</i>  |   |              |        |                              |
| WL12  | Sheared porphyritic granitoid (Longquanguan)          | 2543 ± 7     | SHRIMP | Wilde <i>et al.</i> (1997)   |
| WN11  | Sheared tonalitic granitoid (Yushuwan Village)        | 2541 ± 14    | SHRIMP | Wilde <i>et al.</i> (1997)   |
| WL9   | Sheared porphyritic granitoid (Longquanguan)          | 2540 ± 18    | SHRIMP | Wilde <i>et al.</i> (1997)   |
| <b>Ages of late Archaean Wutai ‘Group’ (2530–2515 Ma)</b>   |   |              |        |                              |
| 95PC-114  | Meta-andesite, Zhuangwang ‘Formation’ (Ekou)          | 2529 ± 10    | SHRIMP | Wilde <i>et al.</i> (2004b)  |
| 96PC-119  | Meta-andesite, Zhuangwang ‘Formation’ (Ekou)          | 2513 ± 8     | SHRIMP | Wilde <i>et al.</i> (2004b)  |
| 95-PC-115   | Meta-andesite, Baizhiyan ‘Formation’ (Ekou)           | 2524 ± 10    | SHRIMP | Wilde <i>et al.</i> (2004b)  |
| WT13  | Meta-rhyolite, Hongmenyan ‘Formation’ (S–T)           | 2533 ± 8     | SHRIMP | Wilde <i>et al.</i> (2004b)  |
| WT17  | Meta-rhyodacite Hongmenyan ‘Formation’ (S–T)          | 2524 ± 8     | SHRIMP | Wilde <i>et al.</i> (2004b)  |
| WT9   | Meta-dacite, Hongmenyan ‘Formation’ (S–T)             | 2523 ± 9     | SHRIMP | Wilde <i>et al.</i> (2004b)  |
| WT12  | Meta-rhyodacite, Hongmenyan ‘Formation’ (S–T)         | 2516 ± 10    | SHRIMP | Wilde <i>et al.</i> (2004b)  |
| 95-PC-55c   | Meta-rhyolite, Gaofan ‘Subgroup’ (Xiazhuang)          | 2528 ± 6     | SHRIMP | Wilde <i>et al.</i> (2004b)  |
| <b>Ages of late Archaean to Palaeoproterozoic Hengshan, Yixingzhai and Fuping TTG gneisses (2520–2480 Ma)</b> |   |              |        |                              |
| <i>Hengshan TTG gneisses</i>  |   |              |        |                              |
| 980814  | Dioritic gneiss (Changchenggou)                       | 2479 ± 3     | SHRIMP | Kröner <i>et al.</i> (2005b) |
|   | Same sample   | 2478.2 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 990803  | Dioritic gneiss with melt patches (Dashiyu)           | 2475 ± 2     | SHRIMP | Kröner <i>et al.</i> (2005b) |
| 990859  | Dioritic gneiss (Xiaoshiyu)                           | 2506 ± 5     | SHRIMP | Kröner <i>et al.</i> (2005b) |
|   | Same sample   | 2504.6 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 980802  | Homogeneous tonalitic gneiss (Dashiyu)                | 2500.5 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 990871  | Tonalitic gneiss, small side valley of Ruyuegou       | 2502.3 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| HG1   | Tonalitic gneiss (Dashiyu)                            | 2520 ± 15    | SHRIMP | Wilde <i>et al.</i> (unpub.) |
| HG5   | Garnetiferous trondhjemitic gneiss (Yanmenguan)       | 2520 ± 10    | SHRIMP | Kröner <i>et al.</i> (2005b) |
| HG6   | Trondhjemitic gneiss, roadcut near Yanmenguan         | 2526 ± 12    | SHRIMP | Kröner <i>et al.</i> (2005b) |
| HG7   | Trondhjemitic gneiss, A roadcut near Yanmenguan       | 2507 ± 4     | SHRIMP | Kröner <i>et al.</i> (2005b) |
| 990845  | Foliated trondhjemitic gneiss (Xiaoshiyu)             | 2504.4 ± 0.4 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 980838  | Trondhjemitic gneiss (Ruyuegou)                       | 2503.0 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 990847  | Grey, unveined trondhjemitic gneiss (Yanmenguan)      | 2524 ± 8     | SHRIMP | Kröner <i>et al.</i> (2005b) |
|   | Same sample   | 2521.7 ± 0.4 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 990854  | Trondhjemitic gneiss (Xiaoshiyu)                      | 2504.6 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 980809  | Foliated pegmatitic red granite-gneiss (Dashiyu)      | 2501 ± 3     | SHRIMP | Kröner <i>et al.</i> (2005b) |
|   | Same sample   | 2503.5 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 980825  | Fine-grained, strained granite-gneiss (Xiaoshiyu)     | 2496.3 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 980833  | Granitic gneiss (Xiaoshiyu)                           | 2492.4 ± 0.3 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 990873  | Foliated granite gneiss from migmatite (Guaner)       | 2499 ± 6     | SHRIMP | Kröner <i>et al.</i> (2005b) |
|   | Same sample   | 2497.6       | EVAP   | Kröner <i>et al.</i> (2005b) |
| 980803  | Homogeneous felsic gneiss (Dashiyu)                   | 2502.3 ± 0.6 | EVAP   | Kröner <i>et al.</i> (2005b) |

Table 1. Continued.

| Sample  | Description   | Age (Ma)     | Method | Sources                      |
|---|---|--------------|--------|------------------------------|
| 990821  | Finely layered, fine-grained biotite gneiss (Dashiyu) | 2526 ± 04.7  | SHRIMP | Kröner <i>et al.</i> (2005b) |
| <i>Yixingzhai TTG gneisses</i>  |   |              |        |                              |
| 96PC153   | Homogeneous tonalitic gneiss, near Yixingzhai         | 2513 ± 15    | SHRIMP | Wilde (2002)                 |
| 96PC154   | Quartz dioritic gneiss, a roadcut near Yixingzhai     | 2499 ± 4     | SHRIMP | Wilde (2002)                 |
| <i>Fuping TTG gneisses</i>  |   |              |        |                              |
| FG1   | Layered tonalitic gneiss, a roadcut near Xicaokou     | 2523 ± 14    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP50  | Foliated, hornblende-rich tonalitic gneiss (Diebuan)  | 2520 ± 20    | SHRIMP | Guan <i>et al.</i> (2002)    |
| FP54  | Trondhjemitic gneiss collected from Pingyang          | 2513 ± 12    | SHRIMP | Guan <i>et al.</i> (2002)    |
| FP217   | Trondhjemitic gneiss collected from Tuanpokou         | 2499 ± 9.5   | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP216   | Granodioritic gneiss collected from Tuanpokou         | 2486 ± 8     | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP08  | Granodioritic gneiss collected from Xicaokou          | 2475 ± 8     | SHRIMP | Guan <i>et al.</i> (2002)    |
| FP236   | Mylonitized monzogranitic gneiss (near Ciyu)          | 2510 ± 22    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP224   | Mylonitized pegmatitic dyke (Xinzhuang)               | 2507 ± 11    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| <b>Ages of Palaeoproterozoic (2360–2000 Ma) granitoids</b>  |   |              |        |                              |
| <i>Hengshan Complex</i>   |   |              |        |                              |
| 980806  | Fine-grained granitic orthogneiss (Dashiyu)           | 2358.7 ± 0.5 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 990850  | Layered trondhjemitic gneiss (Dashiyu)                | 2329.7 ± 0.6 | EVAP   | Kröner <i>et al.</i> (2005b) |
| HG4   | Pegmatitic granite-gneiss (Dashiyu)                   | 2331 ± 36    | SHRIMP | Kröner <i>et al.</i> (2005b) |
| 980881  | Pegmatite cutting older gneisses (Xiaoshiyu)          | 2248.5 ± 0.5 | EVAP   | Kröner <i>et al.</i> (2005b) |
| 980844  | Red anatectic granite (Changchenggou)                 | 2113 ± 8     | SHRIMP | Kröner <i>et al.</i> (2005b) |
|   | Same sample   | 2112.3 ± 0.6 | EVAP   | Kröner <i>et al.</i> (2005b) |
| <i>Wutai Complex</i>  |   |              |        |                              |
| D2  | Coarse-grained Dawaliang porphyritic granite          | 2176 ± 12    | SHRIMP | Wilde (2002)                 |
| 95-PC-50  | Pink Phase of the Wangjiahui granite                  | 2117 ± 17    | SHRIMP | Wilde <i>et al.</i> (2005)   |
| 95-PC-51  | Pink Phase of the Wangjiahui granite                  | 2116 ± 16    | SHRIMP | Wilde <i>et al.</i> (2005)   |
| 95-PC-60  | Pink Phase of the Wangjiahui granite                  | 2084 ± 16    | SHRIMP | Wilde <i>et al.</i> (2005)   |
| <i>Fuping Complex</i>   |   |              |        |                              |
| FP188-2   | Nanying granitoid gneisses, 10 km south of Fuping     | 2077 ± 13    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP204   | Nanying granitoid gneisses near Dianfang              | 2024 ± 21    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP30  | Nanying granitoid gneisses collected from Gangnan     | 2045 ± 64    | SHRIMP | Guan <i>et al.</i> (2002)    |
| <b>Ages of Palaeoproterozoic pre-tectonic mafic dykes (now amphibolite or high-pressure granulites)</b> |   |              |        |                              |
| <i>Hengshan Complex</i>   |   |              |        |                              |
| Ch020901  | High-pressure granulite facies mafic dyke (Dashiyu)   | 1915 ± 4     | SHRIMP | Kröner <i>et al.</i> (2006)  |
| Ch020902  | High-pressure granulite facies mafic dyke (Dashiyu)   | 1914 ± 2     | SHRIMP | Kröner <i>et al.</i> (2006)  |
| <i>Wutai Complex</i>  |   |              |        |                              |
| 02SX009   | Amphibolite facies mafic dyke (Hengling)              | 2147 ± 5     | SHRIMP | Peng (2005)                  |
| <b>Ages of the Palaeoproterozoic Hutuo Group and Wanzi supracrustal assemblage</b>                      |   |              |        |                              |
| <i>Hutuo Group in the Wutai Complex</i>   |   |              |        |                              |
| HTG-10  | A felsic tuffaceous rock at Taihuai: an older group   | 2180 ± 4     | SHRIMP | Wilde <i>et al.</i> (2004a)  |
|   | A younger population from the same sample             | 2087 ± 9     | SHRIMP | Wilde <i>et al.</i> (2004a)  |
| <i>Wanzi Supracrustal assemblage in the Fuping Complex</i>  |   |              |        |                              |
| FP62  | A paragneiss collected from Pingyang                  | 2097 ± 46    | SHRIMP | Guan <i>et al.</i> (2002)    |
| FP249   | The youngest concordant zircon from a paragneiss      | 2109 ± 5     | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP260   | One young discordant zircon from a paragneiss         | 2097 ± 6     | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP014   | Upper intercept age of igneous zircons cores          | 2099 ± 22    | LA-ICP | Xia <i>et al.</i> (2006c)    |
| FP037   | Upper intercept age of igneous zircons cores          | 2110 ± 30    | LA-ICP | Xia <i>et al.</i> (2006c)    |
| FP053   | Upper intercept age of igneous zircons cores          | 2112 ± 23    | LA-ICP | Xia <i>et al.</i> (2006c)    |
| <b>Ages of metamorphism of the Hengshan, Wutai and Fuping complexes (1880–1800 Ma)</b>                  |   |              |        |                              |
| <i>Hengshan Complex</i>   |   |              |        |                              |
| 990803  | One euhedral zircon from a dioritic gneiss (Dashiyu)  | 1881 ± 8     | SHRIMP | Kröner <i>et al.</i> (2005b) |
| M068  | Meta-zircons from high-pressure granulite (Dashiyu)   | 1850 ± 3     | SHRIMP | Kröner <i>et al.</i> (2006)  |
| HG2   | Meta-zircons from high-pressure granulite (Dashiyu)   | 1867 ± 23    | SHRIMP | Kröner <i>et al.</i> (2006)  |
| Ch990839  | Coarse-grained pegmatitic melt (Xiaoshiyu)            | 1851 ± 5     | SHRIMP | Kröner <i>et al.</i> (2006)  |
|   | Same sample   | 1856.1 ± 0.6 | EVAP   | Kröner <i>et al.</i> (2006)  |
| HG1   | Meta-zircons from a granitic gneiss (Dashiyu)         | 1872 ± 17    | SHRIMP | Kröner <i>et al.</i> (2006)  |
| Ch980871  | Meta-zircons from a retro-eclogite (Xiaoshiyu)        | 1881.3 ± 0.4 | EVAP   | Kröner <i>et al.</i> (2006)  |
| Ch990853  | Meta-zircons from a retrograded eclogite (Dashiyu)    | 1859.7 ± 0.5 | EVAP   | Kröner <i>et al.</i> (2006)  |
| Ch990886  | Meta-zircons from a mafic granulite (Dashiyu)         | 1850.9 ± 0.4 | EVAP   | Kröner <i>et al.</i> (2006)  |
| Ch990848  | Meta-zircons from high-pressure granulite (Dashiyu)   | 1885.6 ± 0.4 | EVAP   | Kröner <i>et al.</i> (2006)  |
| <i>Wutai Complex</i>  |   |              |        |                              |
| Unknown   | Garnet amphibolite from the Jingangku Formation       | 1851 ± 9     | Sm–Nd  | Wang <i>et al.</i> (2001)    |
| S2010-2-1   | Meta-monazites from a kyanite schist (Jingangku)      | 1822 ± 14    | EPMA   | Liu <i>et al.</i> (2006)     |
| S2010-2-1   | Meta-monazites from a kyanite schist (Jingangku)      | 1833 ± 8     | EPMA   | Liu <i>et al.</i> (2006)     |
| SZ10  | Meta-monazites from a kyanite schist (Jingangku)      | 1847 ± 62    | EPMA   | Liu <i>et al.</i> (2004a)    |
| <i>Fuping Complex</i>   |   |              |        |                              |
| FG1   | Meta-zircons from a tonalitic gneiss (Xicaokou)       | 1802 ± 43    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP217   | Meta-zircons from a trondhjemitic gneiss              | 1875 ± 43    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP216   | Meta-zircons from granodioritic gneiss (Tuanpokou)    | 1825 ± 12    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP249   | One meta-zircon rim from a paragneiss (Jiaan)         | 1821 ± 42    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP260   | One meta-zircon rim from a paragneiss (Diaoyutai)     | 1891 ± 6     | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP188-2   | Meta-zircons from a Nanying granitoid gneisses        | 1826 ± 12    | SHRIMP | Zhao <i>et al.</i> (2002a)   |
| FP204   | Meta-zircons from a Nanying granitoid gneisses        | 1850 ± 9.6   | SHRIMP | Zhao <i>et al.</i> (2002a)   |

Table 1. Continued.

| Sample  | Description   | Age (Ma)     | Method  | Sources                    |
|---|---|--------------|---------|----------------------------|
| FP30  | Two meta-zircon rims from a gneissic granite        | 1825 ± 18    | SHRIMP  | Guan <i>et al.</i> (2002)  |
| FP08  | Meta-zircons from a granodioritic gneiss (Xicaokou) | 1817 ± 26    | SHRIMP  | Guan <i>et al.</i> (2002)  |
| FP053   | Meta-zircon rims from a paragneiss (Wuyuezhai)      | 1863 ± 30    | LA-ICP  | Xia <i>et al.</i> (2006c)  |
| FP014   | Meta-zircon rims from a paragneiss (Tuanpokou)      | 1822 ± 20    | LA-ICP  | Xia <i>et al.</i> (2006c)  |
| <b>Ages of Post-orogenic (1780–1750 Ma) mafic dyke swarms</b> |   |              |         |                            |
| <i>Hengshan Complex</i>                                       |   |              |         |                            |
| GU12  | Unmetamorphosed mafic dyke (Near Tuling)            | 1769.1 ± 2.5 | SGD     | Halls <i>et al.</i> (2000) |
| <i>Wutai Complex</i>  |   |              |         |                            |
| 03WT08  | Unmetamorphosed diabase dyke (Wutai County)         | 1754 ± 71    | SGD     | Peng (2005)                |
| <i>Fuping Complex (Taihangshan)</i>                           |   |              |         |                            |
| 99JX-16   | Unmetamorphosed mafic dyke (Canyansi)               | 1765.3 ± 1.1 | W–Ar–Ar | Wang <i>et al.</i> (2004a) |
| 99JX-65   | Unmetamorphosed mafic dyke (Huangbeiping)           | 1774.7 ± 0.7 | W–Ar–Ar | Wang <i>et al.</i> (2004a) |
| 99JX-71   | Unmetamorphosed mafic dyke (Hujian'an)              | 1780.7 ± 0.5 | W–Ar–Ar | Wang <i>et al.</i> (2004a) |

SHRIMP – Sensitive high resolution ion microprobe; EVAP – Single grain evaporation. EPMA – Electron Probe Microanalysis; LA-ICP – LA-ICP-MS; SGD – Single grain dissolution; W–Ar–Ar – Whole rock  $^{40}\text{Ar}$ – $^{39}\text{Ar}$ . Meta-zircon – metamorphic zircons; S–T – Shahe–Taihuai Road. Note: In Kröner *et al.* (2005a,b, 2006a), Dashiyou, Xiaoshiyou and Changchengou were translated into Big (or Large) Stone Valley, Little (or Small) Stone Valley and Great Wall Valley, respectively.

at about 2550–2540 Ma, slightly later than the Ekou pluton. The Lanzhishan pluton is divisible into older and younger phases with SHRIMP U–Pb zircon ages of  $2553 \pm 8$  and  $2537 \pm 10$  Ma, respectively (Wilde, Cawood & Wang, 1997). SHRIMP U–Pb zircon data also reveal that the Guangmingshi and Shifou plutons were simultaneously emplaced at *c.* 2531 Ma and probably belong to the same pluton. In summary, the above SHRIMP U–Pb zircon ages confirm the end-Archaeon (2560–2520 Ma) emplacement of the Wutai granites which reflect the earliest arc-related magmatic event in the Hengshan–Wutai–Fuping belt.

The Longquanguan augen gneisses are restricted to the Longquanguan ductile shear zone and were previously interpreted as the youngest lithological unit in the Fuping Complex (Liu *et al.* 1985). However, recent SHRIMP U–Pb zircon ages reveal that the Longquanguan augen gneisses were emplaced at *c.* 2540 Ma (Wilde, Cawood & Wang, 1997), broadly simultaneous with the Wutai gneissic granites, but older than the 2520–2480 Ma Fuping TTG gneisses (see Section 4.d). Furthermore, magmatic zircons of the Longquanguan augen gneisses are not only identical in age to those of the Lanzhishan Granite, but both granitoids contain inherited zircons and zircon cores giving ages of *c.* 2700 Ma. Therefore, the Longquanguan augen gneisses most likely represent the equivalents of the Wutai granites but were structurally reworked during the development of the Longquanguan ductile shear zone. Similarities in fabric and petrographic features as well as the close spatial distribution between the Longquanguan augen gneisses and some of the Wutai granites (e.g. Lanzhishan granite) support this conclusion.

#### 4.c. 2530–2515 Ma: formation of the Neoarchaeon Wutai ‘Group’

The volcano-sedimentary rocks of the Wutai ‘Group’ have long been considered to be either older than, or coeval with, the Wutai granites (Tian, 1991; Bai, Wang &

Guo, 1992). However, SHRIMP U–Pb zircon ages show that the volcanic rocks of the Wutai ‘Group’ formed at 2530–2515 Ma (Wilde *et al.* 2004a), later than most of the Wutai granites. Traditionally, the Wutai ‘Group’ was subdivided into the lower, middle and upper sequences, represented by the Shizui, Taihuai and Gaofan subgroups, respectively (Bai, 1986; Bai, Wang & Guo, 1992; Tian, 1991). However, SHRIMP zircon ages reveal no difference between volcanic rocks previously considered to occupy different stratigraphic levels within the Wutai ‘Group’ (Table 1). For example, zircons from a felsic tuff of the Gaofan Subgroup provided a weighted mean  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $2528 \pm 6$  Ma, virtually the same as the age of  $2529 \pm 10$  Ma obtained for a felsic tuff of the Shizui Subgroup (Wilde, 2002). A meta-andesite, collected from the Zhuangwang Formation of the Shizui Subgroup, yielded a SHRIMP U–Pb zircon age of  $2513 \pm 8$  Ma (Wilde *et al.* 2005), which is the youngest age so far obtained for the Wutai ‘Group’. Five rhyolite or rhyodacite samples collected from the Taihuai Subgroup yielded SHRIMP U–Pb zircon ages ranging from  $2533 \pm 8$  Ma to  $2516 \pm 10$  Ma (Wilde *et al.* 2005), suggesting that their eruption was largely coeval with that of the volcanic rocks in the Shizui and Gaofan subgroups. These results, together with field observations, suggest that the Wutai ‘Group’ is not a simple lithostratigraphic succession but constitutes a lithotectonic assemblage of volcano-sedimentary rocks that formed in the period 2530–2515 Ma. The volcanic rocks of the Wutai ‘Group’ have a close affinity to modern volcanic arc assemblages (Bai, 1986; Wang *et al.* 2004b; Polat *et al.* 2005) and may thus document the earliest arc-related volcanism in the Hengshan–Wutai–Fuping belt.

#### 4.d. 2520–2480 Ma: emplacement of Neoarchaeon to Palaeoproterozoic Hengshan, Yixingzhai and Fuping TTG gneisses

Granitoid gneisses are the major components of the Hengshan and Fuping complexes, making up 70–80 %

of the total exposures of both the complexes (Fig. 4). Conventionally, the high-grade Hengshan and Fuping complexes were assumed to be older than the low-grade Wutai Complex (Tian, 1991; Bai, Wang & Guo, 1992), and the Hengshan and Fuping TTG gneisses were considered to have been emplaced prior to intrusion of the Wutai granitoids (Tian, 1991; Bai, Wang & Guo, 1992). However, recent SHRIMP and evaporation zircon dating does not support these assumptions. As shown in Table 1, SHRIMP U–Pb ages for magmatic zircons indicate that the Fuping TTG gneisses were emplaced between 2523 and 2486 Ma (Guan *et al.* 2002; Zhao *et al.* 2002b). Similarly, SHRIMP and evaporation zircon ages show that the Hengshan TTG gneisses were also emplaced between 2524 and 2479 Ma (Table 1; Kröner *et al.* 2005a,b). An important conclusion from these data is that the Hengshan and Fuping tonalitic–trondhjemitic–granodioritic gneisses are younger than the Wutai granitoids; the latter were emplaced during Neoproterozoic times, whereas the former were emplaced in the period between Neoproterozoic and Early Palaeoproterozoic times. This conclusion contrasts with previous models that assumed the high-grade Fuping and Hengshan complexes to constitute an older basement below the lower grade Wutai Complex.

The amphibolite facies Yingxingzhai TTG gneisses in the southern Hengshan area have long been considered to be younger than the granulite facies Hengshan TTG gneisses in the northern area, but recent SHRIMP zircon ages do not support this assumption. Zircons from two Yingxingzhai TTG gneiss samples were SHRIMP-dated by Wilde (2002) at  $2513 \pm 5$  Ma and  $2499 \pm 4$  Ma (Table 1), which are similar to the age range (2526–2480 Ma) of the Hengshan TTG gneisses (Table 1; Kröner *et al.* 2005a,b). Therefore, the Yingxingzhai TTG gneisses most likely represent the upper crustal equivalent of the Hengshan TTG gneisses (Kröner *et al.* 2005a,b). This conclusion is consistent with the available geochemical data, which indicate that both the Hengshan and Yingxingzhai TTG gneisses belong to calc-alkaline suites emplaced in a magmatic arc environment (Fig. 4; Kröner *et al.* 2005a,b).

#### 4.e. 2350–2050 Ma: intermittent emplacement of Palaeoproterozoic granitoids

Palaeoproterozoic granitoids in the Hengshan–Wutai–Fuping belt have not been recognized until recently, following acquisition of new SHRIMP and evaporation zircon ages which indicate the widespread presence of Palaeoproterozoic granitoids in the belt. In the Wutai Complex, Palaeoproterozoic granitoids are represented by the Dawaliang pluton and the pink phase of the Wangjiahui pluton, of which the former yielded a SHRIMP zircon age of  $2176 \pm 12$  Ma, whereas three samples of the latter were dated at  $2117 \pm 17$ ,

$2116 \pm 16$  and  $2084 \pm 20$  Ma, respectively (Wilde *et al.* 2005). In the Fuping Complex, the Palaeoproterozoic granitoids are represented by the Nanying granitic gneisses which widely occur in the Fuping TTG gneisses (Fig. 4). SHRIMP data show that the Nanying gneissic granites were emplaced in the period 2077–2024 Ma (Table 1; Guan *et al.* 2002; Zhao *et al.* 2002b), slightly later than the Dawaliang granites. In the Hengshan Complex, similar-aged gneissic granites have been identified by Kröner *et al.* (2005a,b) using the SHRIMP and single-grain evaporation techniques (Table 1). They recognized three phases of Palaeoproterozoic gneissic granitoids that were emplaced at *c.* 2360 Ma, 2330 Ma and 2250 Ma, respectively (Table 1). It is particularly important to note that these Palaeoproterozoic granitoids in the Hengshan and Fuping complexes contain the same deformational features as the older Hengshan and Fuping granitoid gneisses and thus unambiguously demonstrate that the main deformational event is not Archaean but Palaeoproterozoic in age.

#### 4.f. 2150–1915 Ma: emplacement of pre-tectonic mafic dykes

Pre-tectonic gabbro and diabase dykes are represented by numerous boudins and lensoid layers of mafic granulites or amphibolites in the Hengshan, Wutai and Fuping complexes. In some places, and in particular in low-strain zones such as those south of the Zhujiafang shear zone (Fig. 4), primary igneous fabrics (e.g. ophitic or subophitic texture) are well preserved in the amphibolite to granulite facies dykes (O'Brien, Walte & Li, 2005; Kröner *et al.* 2005a, 2006). In most cases, and especially in high-strain zones such as the areas north of the Zhujiafang shear zone (Fig. 4), the pre-tectonic gabbroic/diabase dykes occur as high-pressure granulites or retrograded eclogites, and strong ductile deformation has later rotated these dykes into parallelism with the layering in the enclosing gneisses and caused boudinage (Kröner *et al.* 2005a, 2006). In all of these cases the amphibolitic or granulitic layers and boudins are part of original gabbroic/diabase dykes that indicate an extensional event. In the Hengshan Complex, igneous zircons from two pre-tectonic mafic dykes (high-pressure granulites) yielded SHRIMP  $^{207}\text{Pb}/^{206}\text{Pb}$  ages of  $1915 \pm 4$  Ma and  $1914 \pm 2$  Ma (Kröner *et al.* 2006), indicating that emplacement of the dykes occurred shortly before a major collisional (tectono-thermal) event at *c.* 1850 Ma (see Section 4.h). The deformed and metamorphosed mafic dykes are also found in the Wutai Complex, and one amphibolite facies mafic dyke yielded a SHRIMP  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $2147 \pm 5$  Ma (P. Peng, unpub. Ph.D. thesis, Chinese Acad. Sciences, Beijing, 2005; Table 1). These mafic dykes were most probably emplaced in back-arc or intra-arc basin environments.

#### 4.g. 2100–1860 Ma: deposition of the late Palaeoproterozoic Hutuo Group and Wanzi supracrustal assemblage

The Hutuo Group is considered to be the youngest unit in the Hengshan–Wutai–Fuping belt and unconformably overlies both the Wutai and Fuping complexes (Tian, 1991). However, at several localities, the Hutuo rocks are tectonically interleaved with the Wutai ‘Group’, and at some localities the Hutuo Group shows the same deformational and metamorphic patterns as the Upper Wutai ‘Subgroup’. This implies that deposition of the Hutuo Group rocks places a maximum age on the timing of the major deformational event in the Hengshan–Wutai–Fuping belt. A felsic tuffaceous rock collected from a metamorphosed sequence of volcanic rocks and sediments of the Hutuo Group contains two zircon populations with SHRIMP  $^{207}\text{Pb}/^{206}\text{Pb}$  ages of  $2180 \pm 5$  and  $2087 \pm 9$  Ma, respectively (Wilde *et al.* 2004b). The older age is within error of the age of the Dawaliang granite in the Wutai Complex and is considered to be derived from a similar crustal magmatic source. The younger age is within the error of reported ages for the Nanying granitic gneisses in the adjacent Fuping Complex and is interpreted to reflect the timing of Hutuo volcanism. The age of  $2087 \pm 9$  Ma also means that the Hutuo rocks must have been deformed and metamorphosed after this time, which further supports the conclusion, obtained from other recent studies (Kröner *et al.* 2005a, 2006), that the main tectonism in the Hengshan–Wutai–Fuping belt occurred in the Palaeoproterozoic era and not in the late Archaean era as previously considered (Tian, 1991).

The Wanzi supracrustal assemblage in the Fuping Complex was previously considered to have been deposited between 2800 and 2560 Ma, constrained by conventional multigrain U–Pb zircon ages (Liu *et al.* 1985). However, recent SHRIMP data do not support this conclusion. Zhao *et al.* (2002b) obtained SHRIMP  $^{207}\text{Pb}/^{206}\text{Pb}$  zircon ages of  $2502 \pm 7$  Ma and  $2507 \pm 14$  Ma from two Al-rich gneiss samples of the Wanzi supracrustal assemblage, interpreted as the crystallization ages of the igneous protolith. These authors also obtained a concordant zircon age of  $2109 \pm 5$  Ma from one Al-rich gneiss sample, which suggests that the protoliths of the Wanzi assemblage must have been deposited after *c.* 2109 Ma. This conclusion is supported by Guan *et al.* (2002), who obtained a SHRIMP zircon age of  $2097 \pm 46$  Ma from a fine-grained paragneiss of the Wanzi assemblage. Most recently, Xia *et al.* (2006c) applied the LA-ICP-MS technique to date the Wanzi supracrustal rocks and obtained three upper concordia intercept ages of  $2099 \pm 22$  Ma,  $2110 \pm 30$  Ma and  $2112 \pm 23$  Ma for igneous zircon cores separated from a sillimanite paragneiss (Table 1). They also obtained metamorphic ages of 1863–1822 Ma for zircon rims that formed

due to overgrowth or recrystallization. These data establish that (1) deposition of the Wanzi supracrustal assemblage occurred at some time between 2100 and 1860 Ma, and (2) that sedimentation of the Wanzi assemblage in the Fuping Complex was largely coeval with volcanism in the Hutuo Group of the Wutai Complex.

#### 4.h. 1880–1820 Ma: Late Palaeoproterozoic continent–continent collision

In the Chinese literature, the high-grade Hengshan and Fuping complexes and the low-grade Wutai Complex have long been considered to be the products of two different tectono-metamorphic events, called the Fuping (*c.* 2.5 Ga) and Wutai (2.4–2.3 Ga) ‘movements’, respectively (Bai, 1986; Wu *et al.* 1989; Tian, 1991; Bai, Wang & Guo, 1992; Bai & Dai, 1998). This conclusion was based on a few ‘unconformities’, conventional multigrain zircon geochronology, and the common misconception that high-grade metamorphic rocks were older than low-grade ones. However, new geochronological data do not verify either the Fuping or Wutai ‘movement’. In the Hengshan and Fuping complexes, SHRIMP U–Pb zircon studies combined with cathodoluminescence imaging and U–Th chemistry confirm the existence of metamorphic zircons in most lithologies (Guan *et al.* 2002; Zhao *et al.* 2002b; Kröner *et al.* 2005a,b, 2006). These zircons occur either as overgrowth rims surrounding older magmatic cores or as single, metamorphic grains, and are structureless, highly luminescent, multifaceted and very low in Th and U contents. These features make them distinctly different from the igneous zircons that are generally characterized by oscillatory zoning and comparatively high Th and U contents. Most metamorphic zircons from both Neoproterozoic and Palaeoproterozoic rocks in the Fuping and Hengshan complexes yielded similar concordant ages in the range 1880 to 1820 Ma (Table 1), which are 700 Ma to 150 Ma younger than their igneous zircon cores. In the low- to medium-grade Wutai Complex, although metamorphic zircons have not yet been observed, a garnet–amphibole–whole rock Sm–Nd isochron age of  $1851 \pm 9$  Ma was obtained for a garnet amphibolite collected from the Lower Wutai ‘Subgroup’ (Table 1; Wang *et al.* 2001). Most recently, Liu *et al.* (2004b, 2006) applied the electron probe microanalysis (EPMA) monazite dating technique to determination of metamorphic ages of the Wutai Complex. They obtained  $\text{ThO}_2^*/\text{PbO}$  ages of  $1833 \pm 8$ ,  $1822 \pm 14$  and  $1846 \pm 64$  Ma for three monazite grains separated from a garnet–kyanite mica schist from the lower part (called the Jinganku Formation) of the ‘Shizui Subgroup’ (see Fig. 4). These data are consistent with the above mineral Sm–Nd isochron age of  $1851 \pm 9$  Ma. Taken together, these results indicate that the main regional metamorphism of the Hengshan, Wutai and

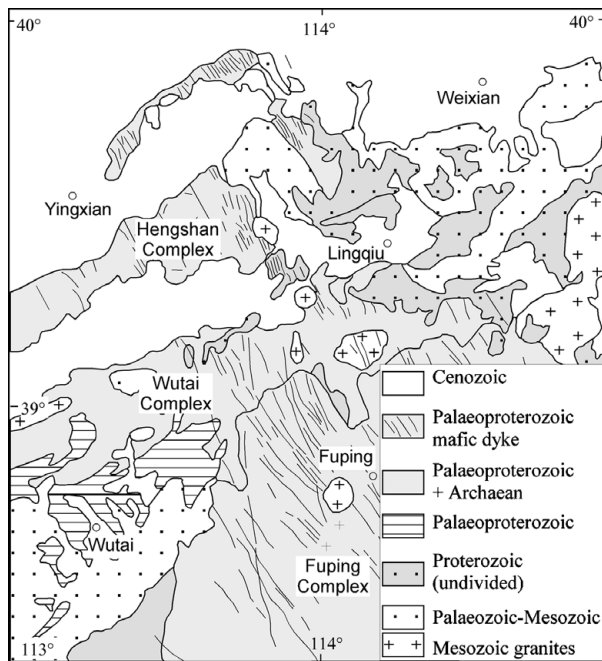


Figure 7. Spatial distribution of post-tectonic mafic dykes in the Hengshan–Wutai–Fuping belt (after Zhao *et al.* 2001b).

Fuping complexes occurred during the 1880–1820 Ma ‘Lüliang Movement’. Based on these metamorphic ages, we conclude that this major tectonothermal event at *c.* 1.85 Ga is related to collision between the Eastern and Western blocks to form the North China Craton.

#### 4.i. 1780–1750 Ma: emplacement of post-orogenic mafic dyke swarm

Late Palaeoproterozoic mafic dykes are widespread in the Hengshan–Wutai–Fuping belt, with a predominant NW–SE to NNW–SSE trend and, on a regional scale, constitute a dyke swarm (Fig. 7). Generally, they dip steeply, cutting both the Archaean and Palaeoproterozoic basement rocks and are covered by Mesoproterozoic and younger strata. Individual dykes range in width from 10 to 50 m to a maximum of about 100 m, and in length from 10 to 40 km, to a maximum of about 100 km. Most dykes are unmetamorphosed and undeformed, with chilled contacts. However, a few dykes were metamorphosed to upper amphibolite facies, indicated by garnet and amphibole reaction rims surrounding igneous plagioclase and clinopyroxene grains. Using the single-grain dissolution technique, Halls *et al.* (2000) obtained an upper concordia intercept zircon age of  $1769.1 \pm 2.5$  Ma for an unmetamorphosed diabase dyke in the Hengshan Complex. Applying the same technique, Peng (P. Peng, unpub. Ph.D. thesis, Chinese Acad. Sciences, Beijing, 2005) obtained a similar upper intercept age of  $1754 \pm 71$  Ma for an undeformed and unmetamorphosed mafic dyke in the Wutai Complex. The post-tectonic mafic dykes in the Fuping Complex have not been well dated,

but Wang *et al.* (2004a) obtained whole-rock  $^{40}\text{Ar}$ – $^{39}\text{Ar}$  ages of  $1765.3 \pm 1.1$  Ma,  $1774.7 \pm 0.7$  Ma and  $1780.7 \pm 0.5$  Ma for mafic dykes in the Zanzhuang Complex, which is 50 km south of the Fuping Complex (Fig. 3), and the two together constitute the Taihang Mountain range. All of these data suggest that emplacement of undeformed and largely unmetamorphosed mafic dykes in the Hengshan–Wutai–Fuping belt occurred in the period 1780–1750 Ma. Controversy has surrounded the origin of these mafic dykes, with one school of thought believing that they were the products of a post-collisional extensional event (Zhao *et al.* 2001b), whereas others argued that they were produced from a mantle plume event that led to break-up of the hypothetical Palaeo-Mesoproterozoic supercontinent Columbia (Zhai, Bian & Zhao, 2000; P. Peng, unpub. Ph.D. thesis, Chinese Acad. Sciences, Beijing, 2005).

## 5. Discussion

Major lithotectonic assemblages and geological events in the Hengshan–Wutai–Fuping belt, as listed in Table 1, enable us to place rigorous constraints on a number of important geological issues regarding the Neoproterozoic to Palaeoproterozoic evolution of this belt and the Trans-North China Orogen.

### 5.a. Relationships between the high-grade Hengshan and Fuping gneiss complexes and low-grade Wutai granite–greenstone terrane

Two contrasting tectonic models have previously been proposed for the evolution of the Hengshan, Fuping and Wutai complexes. One suggests that the first two developed as a single continental block that underwent Neoproterozoic rifting associated with formation of the Wutai greenstones and closed upon itself in the Palaeoproterozoic era (Tian, 1991; Yuan & Zhang, 1993), while the other model proposes that the entire terrane is a Neoproterozoic continent–arc–continent collision system, in which the Hengshan and Fuping complexes represent two exotic Archaean microcontinental blocks, and the Wutai granite–greenstone represents an intervening island arc (Bai, 1986; Li *et al.* 1990; Bai, Wang & Guo, 1992; Wang *et al.* 1996; Polat *et al.* 2005). Although these two models propose different tectonic settings and processes for the evolution of these complexes, they both assume that the Hengshan and Fuping gneisses are a crystalline basement to the Wutai granite–greenstone terrane. However, this assumption is not supported by the isotopic age data summarized above. As shown in Table 1, the Hengshan, Wutai and Fuping complexes are not remarkably different in age and all formed during Neoproterozoic to Palaeoproterozoic times, although the Wutai granitoids are slightly older than major lithologies of the Hengshan and Fuping complexes. These three complexes

seem to have experienced a common Neoproterozoic to Palaeoproterozoic evolution, with the low- to medium-grade Wutai Complex representing an upper crustal calc-alkaline volcano-plutonic assemblage, probably generated above a subduction zone in a magmatic arc setting, whereas the Hengshan and Fuping complexes may be the lower crustal root zone of the arc. Thus, the tectonic evolution of the Hengshan–Wutai–Fuping belt may not be related to local interaction of the three complexes, either through closure of a Wutai rift (Tian, 1991; Yuan & Zhang, 1993) or collision between a Wutai arc and the putative Hengshan and Fuping microcontinental blocks (Bai, 1986; Bai, Wang & Guo, 1992; Li *et al.* 1990; Wang *et al.* 1996; Polat *et al.* 2005). Instead, we suggest that the three complexes represent elements of a single Neoproterozoic to Palaeoproterozoic, long-lived magmatic arc that was subsequently incorporated into the Trans-North China Orogen during collision between the Eastern and Western blocks at around 1.85 Ga (Zhao *et al.* 2000b, 2001b, 2002a; Wilde, Zhao & Sun, 2002).

### 5.b. Nature of the Hengshan–Wutai–Fuping magmatic arc

Available geochemical data for rocks of the Hengshan, Fuping and Wutai complexes support an arc derivation (Geng & Wu, 1990; Li *et al.* 1990; Bai, Wang & Guo, 1992; Sun, Armstrong & Lambert, 1992; Wang *et al.* 1996; Liu *et al.* 2002b, 2004a,b, 2005; Wang *et al.* 2004b; Kröner *et al.* 2005a,b), with minor ultramafic to mafic rocks in the Lower Wutai that have been interpreted as remnants of ancient oceanic crust (Bai, Wang & Guo, 1992; Wang *et al.* 1996). Bai, Wang & Guo (1992) showed that the majority of amphibolites from the Lower Wutai and greenschists from the Middle Wutai have an affinity to calc-alkaline basalts or arc tholeiites (Fig. 5b). Wang *et al.* (2004b) also showed that the majority of amphibolites and greenschists in the Wutai ‘Group’ were derived from island arc-type basalts, though minor amounts have an affinity to MORB-type and back-arc basin basalts (Fig. 5c). Based on major and trace element features and Nd isotopes, Kröner *et al.* (2005a,b) showed that the Hengshan and Fuping TTG gneisses are characterized by a wide range in SiO<sub>2</sub> content, high Na<sub>2</sub>O, Ba, Sr and low Y and HREE, and their selective enrichment in LIL elements and depletion in Nb, Ta and Ti can be derived from magmatic precursors with a strong mantle signature, modified by a subduction component and variable contributions from older crust. A particularly diagnostic discriminant for the tectonic setting is the Rb–Hf–Ta triangular plot for felsic to intermediate magmatic rocks, in which arc-derived assemblages occupy a field distinct from those generated in within-plate and ocean floor settings (Harris, Pearce & Tindle, 1986), though this diagram is useful only when Rb has not been mobilized. Figure 5a shows such a plot for the Hengshan TTG gneisses (Kröner *et al.* 2005a,b;

Liu *et al.* 2004a), Wutai volcanic and granitoid rocks (Wang *et al.* 2004b; Liu *et al.* 2004b) and Fuping TTG gneisses (Liu *et al.* 2004a), clearly supporting their arc derivation. However, controversy still remains as to the nature of this magmatic arc. In most previous models, the Wutai Complex was considered to have formed in a Mariana-type arc system that was initiated by intra-oceanic subduction (Li *et al.* 1990; Bai, Wang & Guo, 1992; Wang *et al.* 1996). Based on the presence of minor MORB-type basalts and adakite-featured felsic volcanic rocks in the ‘Wutai Group’, Wang *et al.* (2004b) also preferred an intra-oceanic subduction process for the formation of the 2.6–2.5 Ga Wutai granite–greenstone terrane. However, the Mariana-type arc model is not supported by recent age data, which reveal the widespread existence of remnants of 2.66–2.82 Ga old continental crust in the Hengshan, Wutai and Fuping complexes. This is indicated by *c.* 2.7 Ga medium-grained grey biotite or hornblende gneisses that occur interlayered with the younger Hengshan and Fuping TTG gneisses (Guan *et al.* 2002; Kröner *et al.* 2005b), 2.67–2.76 Ga xenocrystic zircons from the Wutai and Longquanguan granitoid gneisses (Wilde, Cawood & Wang, 1997; Wilde *et al.* 2004a; Wilde, 2002), and 2.68–2.82 Ga detrital zircons from the Wanzi supracrustal rocks (Zhao *et al.* 2002b). More importantly, Wilde *et al.* (2004a) found a group of  $2679 \pm 16$  Ma old xenocrystic zircons in a  $2529 \pm 10$  Ma meta-andesite from the Zhuangwang ‘Formation’ that is considered part of the Shizui ‘Subgroup’. These data suggest that the Hengshan–Wutai–Fuping belt was not an intra-oceanic arc but seems to have developed on a continental margin. The available data do not allow us to determine further whether it was a Japan-type island arc or an Andean-type continental margin magmatic arc. Although minor MORB-type basalts have been reported from the Wutai Complex (Wang *et al.* 2004b), it remains uncertain whether these erupted in a back-arc basin or a marginal basin environment.

### 5.c. Trans-North China Orogen: a long-lived (*c.* 700 Ma) accretionary magmatic arc?

As shown in Table 1, much of the juvenile crust of the Hengshan–Wutai–Fuping belt was accreted in the period 2560–2480 Ma, though a number of magmatic accretionary events occurred during Palaeoproterozoic times. However, more than 90 % of juvenile crust in other complexes (e.g. the Lüliang, Zhongtiao, Northern Hebei complexes; Fig. 3) in the Trans-North China Orogen was added in the period 2300–1900 Ma (Sun *et al.* 1993; Sun, 1997; Yu, Wang & Wang, 1997; Geng *et al.* 2000; Tian *et al.* 2005). This suggests that the Trans-North China Orogen records magmatic–tectonic evolution lasting for nearly 650 Ma. This is in contrast to most Phanerozoic collision belts where pre-orogenic sedimentation and magmatism are

followed, within tens of millions of years, by orogenic deformation and metamorphism. The long-lived nature of the Trans-North China Orogen has been regarded as a major argument against the 1.85 Ga collision model, since few convergent continental-margin arcs in the world sat undisturbed for *c.* 700 Ma before being deformed and metamorphosed during accretionary and collisional events (Kusky & Li, 2003). However, such long-lived arcs have been documented or inferred in many other orogenic belts (Rivers & Corrigan, 2000), as summarized below. Kröner, Klemd & Zhao (2006) discussed the evolution of two such long-lived continental collisional belts, the Central Zone of the Limpopo belt in southern Africa and the Hengshan Complex of the Trans-North China Orogen, both of which consist predominantly of Archaean granitoid gneisses and minor shallow-water sediments with minor additions of early Palaeoproterozoic granites, and which were subjected to intense deformation and metamorphism in the late Palaeoproterozoic era, 500–700 Ma after their generation in arc-related settings. As pointed out by Kröner, Klemd & Zhao (2006), in both cases continental collision has been inferred, but there is no record of pre-collisional continental margin evolution and ocean closure; instead both terranes show evidence of having been generated in magmatic arc environments. These authors suggested that these terranes were rifted off from unknown Archaean active continental margins during Palaeoproterozoic times, similar to present-day Japan, and were then displaced before becoming attached, through terrane collision, to different Archaean blocks with which they share no common history.

Similar Palaeo-Mesoproterozoic long-lived magmatic arcs have also been found in southeastern Laurentia, southern Baltica, central Australia and western Amazonia (Karlstrom *et al.* 2001; Bingen *et al.* 2002; Brewer *et al.* 2002). In southeastern Laurentia and southern Baltica, a 1.8 to 1.3 Ga magmatic arc zone extends from Arizona through Colorado, Michigan, southern Greenland, Scotland, Sweden and Finland to western Russia, bordering the present southern margin of North America, Greenland and Baltica (Gower, Ryan & Rivers, 1990; Karlstrom *et al.* 2001). Petrological and geochemical studies indicate that this large magmatic arc zone includes dominantly juvenile volcanogenic sequences and granitoid suites resembling those of present-day island arcs and active continental margins (Nelson & DePaolo, 1985; Bennet & DePaolo, 1987) and may represent subduction-related episodic outgrowth along the continental margin of a Palaeo-Mesoproterozoic supercontinent (Karlstrom *et al.* 2001; Zhao *et al.* 2002a, 2004).

The Central Asian Orogenic Belt and the Kunlun Orogens are good examples of Phanerozoic long-lived magmatic arcs (Enkin *et al.* 1992; Sengör, Natalin & Burtman, 1993; Heubeck, 2001; Windley *et al.* 2002, 2007; Xiao *et al.* 2003, 2004, 2005; Jahn *et al.* 2004).

The Central Asian Orogenic Belt, also called the Altaids (Sengör, Natalin & Burtman, 1993), is situated between the Siberian and Sino-Korean-Tarim cratons, and encompasses an immense area (*c.* 1000 km wide and *c.* 7000 km long) from the Urals in the west, through Kazakhstan, NW China, Mongolia, NE China to the Okhotsk Sea in the Russian Far East (Jahn *et al.* 2004). Available data show that the Central Asian Orogenic Belt was the world's largest site of Phanerozoic juvenile crust that developed from early Palaeozoic to early Mesozoic times (Sengör, Natalin & Burtman, 1993; Windley *et al.* 2002, 2007; Xiao *et al.* 2003, 2004; Jahn *et al.* 2004). A present-day example of long-lived magmatic arcs is the Andes, where the Pacific plate has been subducting under the west coast of South America for *c.* 500 million years since the Cambrian (Howell, 1995; Dalziel, 1997; Rivers & Corrigan, 2000). The Japanese arc has also existed since the Early Permian (Howell, 1995). These examples demonstrate that such long-lived arc magmatism is not unique, though the mechanism maintaining such long-lived arc magmatism remains enigmatic. Rivers & Corrigan (2000) argued that long-lived magmatic arcs are generally large geological features and may have faced a major open ocean, like the present-day Andes and Japanese arcs. This is consistent with our present model for the Trans-North China Orogen inferring a major ocean between the Eastern and Western blocks from before 2.55 Ga until its final closure at 1.85 Ga.

#### 5.d. Palaeoproterozoic magmatic events and their regional significance

Most of the crust in the Hengshan–Wutai–Fuping belt was formed during Neoproterozoic times, including the 2560–2520 Ma Wutai and Longquanguan granitoids, 2530–2515 Ma Wutai greenstone-type volcanics, and 2520–2480 Ma Hengshan, Yixingzhai and Fuping grey TTG gneisses (Table 1). These lithologies occupy more than 70% of the total exposure of the Hengshan–Wutai–Fuping belt (Fig. 4). As discussed above, all these rocks have affinities to modern magmatic arcs and are considered to have formed in a long-lived Japan-type or Andes-type arc.

In the Chinese literature, all lithologies in the Hengshan–Wutai–Fuping belt, except for the Hutuo Group, were considered to be Archaean in age (Bai, 1986; Tian, 1991; Bai, Wang & Guo, 1992). However, the zircon ages summarized above and in Table 1 reveal the widespread presence of Palaeoproterozoic granitoids in the Hengshan, Wutai and Fuping complexes. The presence of these ductilely deformed Palaeoproterozoic granitoid gneisses in the Hengshan and Fuping complexes indicates that the major deformation in these complexes is not Archaean but Palaeoproterozoic in age. This is inconsistent with the models of Kusky & Li (2003) and Polat *et al.* (2005),



which suggested that the collisional event leading to major deformation in these complexes occurred at *c.* 2.5 Ga.

Another important Palaeoproterozoic igneous event revealed by isotopic studies is the emplacement of pre-tectonic mafic dykes which were boudinaged and transformed to high- to medium-pressure mafic granulites in the Hengshan and Fuping complexes and garnet-bearing amphibolites in the Wutai Complex. The precursors of these mafic granulites have been considered to be Archaean in age because Tian *et al.* (1992) obtained a controversial Sm–Nd whole-rock isochron age of  $2818 \pm 86$  Ma for these rocks from the Hengshan Complex. However, the new ages for igneous zircons from two metamorphosed mafic dykes in the Hengshan Complex (Table 1) undoubtedly indicate that they were emplaced in the Palaeoproterozoic era.

## 6. A tectonic scenario for the evolution of the Hengshan–Wutai–Fuping belt

On the basis of lithological, structural, metamorphic, geochemical and geochronological data summarized in this paper, we propose the following scenario for the evolution of the Hengshan–Wutai–Fuping belt, as shown in Figure 8.

- (1) During Neoproterozoic times, the Hengshan–Wutai–Fuping region was part of an Andean-type arc along the western margin of the Eastern Block, which was separated from the Western Block by a major ocean, with subduction of the oceanic lithosphere beneath the western margin of the Eastern Block (Fig. 8a). At 2560–2520 Ma, the release of water from the down-going slab caused partial melting of the lower crust in the overriding plate, producing large volumes of granitoid magma that were emplaced into the upper levels of the crust to form the Wutai and Longquanguan granitoid rocks (Fig. 8a).
- (2) At 2530–2515 Ma, subduction of oceanic lithosphere and further release of fluids caused partial melting in the overriding mantle wedge, leading to underplating of mafic magmas in the lower crust and widespread mafic to felsic volcanism, forming part of the Wutai greenstone assemblage (Fig. 8b). Extension driven by widespread mafic to felsic volcanism led to the development of a back-arc basin or marginal sea which divided the region into the Hengshan–Wutai island arc (Japan-type) and the Fuping relict arc. Minor MORB-type basalts and ultramafic rocks from the Wutai greenstone assemblage were formed in the back-arc basin or marginal sea (Fig. 8b).
- (3) At 2520–2480 Ma, subduction beneath the Hengshan–Wutai island arc triggered partial melting of the lower crust to form large amounts of TTG magmas that crystallized into the Hengshan and Yixingzhai TTG suites (Fig. 8c). Meanwhile, eastward-directed subduction of oceanic lithosphere of the marginal sea led to reactivation of the Fuping relict arc, where the Fuping TTG suite was emplaced (Fig. 8c).
- (4) In the Palaeoproterozoic era (2360–2000 Ma), several phases of granitoid magmatism occurred in both the Hengshan–Wutai island arc and the Fuping reactivated arc (Fig. 8d), represented by the 2360 Ma, *c.* 2250 Ma and 2000–2100 Ma granitoid gneisses in the Hengshan Complex (Kröner *et al.* 2005a,b), *c.* 2100 Ma Wangjiahui and Dawaliang granites in the Wutai Complex (Wilde *et al.* 2005), and 2100–2000 Ma Nanying granitoid gneisses in the Fuping Complex (Zhao *et al.* 2002b; Guan *et al.* 2002; Liu *et al.* 2002b, 2005).
- (5) At *c.* 1920 Ma, the Hengshan–Wutai island arc underwent an extensional event, possibly due to subduction of a mid-ocean ridge, leading to the emplacement of a major mafic dyke swarm (Fig. 8e). These dykes were subsequently metamorphosed to medium- to high-pressure granulites or retrograded eclogites in the Hengshan Complex and some amphibolites in the Wutai Complex.
- (6) At 1880–1820 Ma, closure of the ocean between the Eastern and Western blocks led to continent–arc–continent collision (Fig. 8f). This collision induced large-scale thrusting, tectonic imbrication and isoclinal folding (Fig. 8f) and transported part of the Hengshan and Fuping lithologies into deep crustal levels where granulite and eclogite facies metamorphism occurred, whereas the Wutai Complex underwent greenschist- to lower amphibolite-facies metamorphism at higher crustal levels. Following peak metamorphism, the thickened crust underwent exhumation and near-isothermal decompression, forming widespread symplectic textures in the rocks. Further exhumation and inflow of fluids along ductile extensional shear zones led to cooling and retrogression, exemplified by widespread migmatization and *in situ* melting in the granitoid gneisses.
- (7) The last magmatic event in the Hengshan–Wutai–Fuping belt was the emplacement of a 1780–1750 Ma mafic dyke swarm, probably as a result of orogenic collapse or post-orogenic extension.

Other metamorphic complexes in the Trans-North China Orogen document a Neoproterozoic to Palaeoproterozoic history essentially similar to that of the Hengshan–Wutai–Fuping belt. For example, the Huai'an, Xuanhua, Chengde and Taihua complexes (Fig. 3) can be compared to the Hengshan or

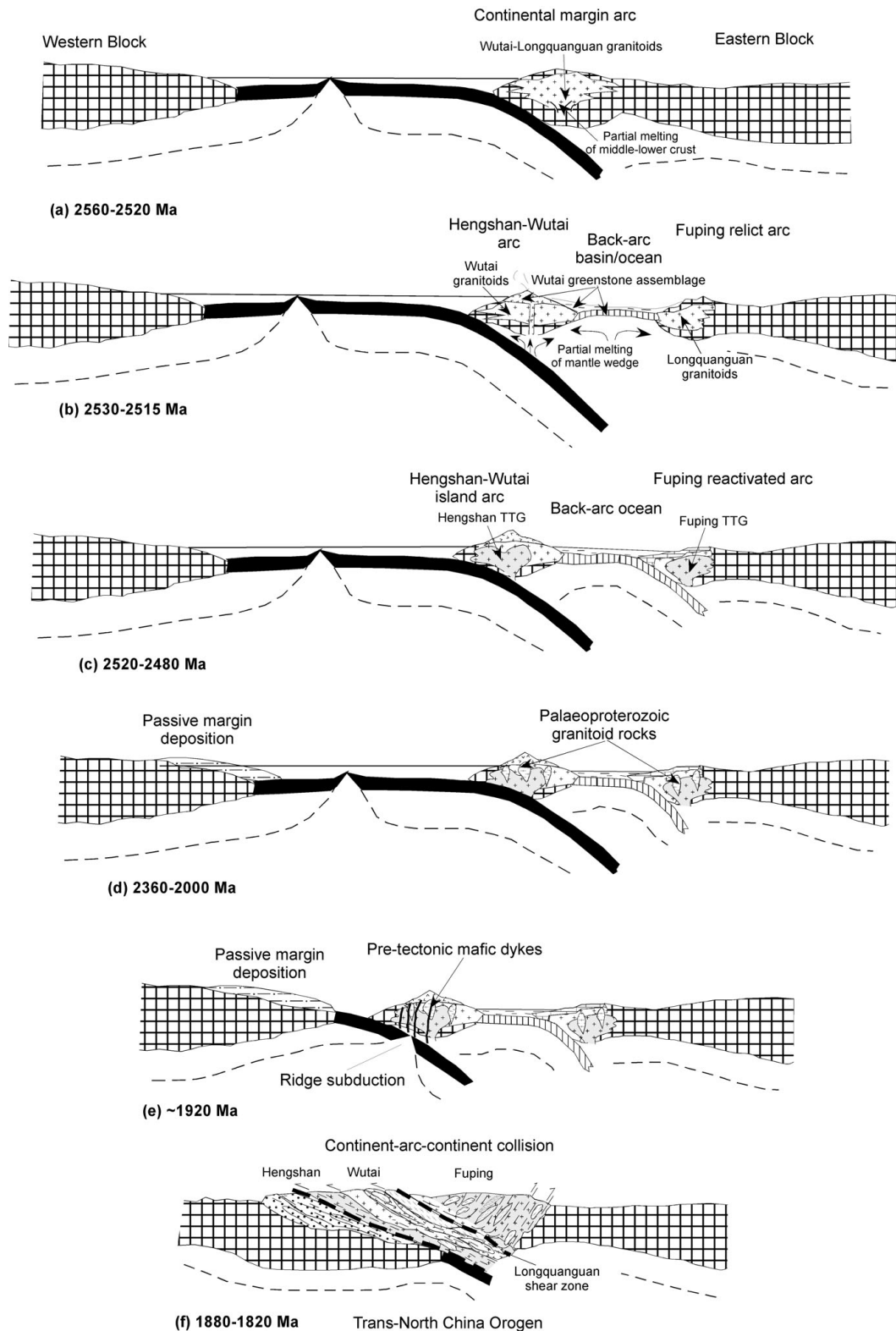


Figure 8. Schematic and speculative sections showing proposed geological evolution of the Hengshan–Wutai–Fuping belt and amalgamation of the North China Craton.

Fuping complexes, whereas the Zhongtiao, Dengfeng, Zanhuang and Northern Hebei complexes (Fig. 3) are similar to the Wutai Complex (Zhao *et al.* 2000a). Therefore, the magmatic, structural and metamorphic

history of the Hengshan–Wutai–Fuping belt as summarized in this paper provides important insights for understanding the tectonic evolution of the Trans-North China Orogen.

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