

The Hulan Group: Its role in the evolution of the Central Asian Orogenic Belt of NE China

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Abstract

The Hulan Group of central Jilin Province, northeastern (NE) China, consists of felsic volcanic and sedimentary rocks that underwent sub-greenschist to amphibolite facies metamorphism and multiple phases of deformation during formation of the Central Asian Orogenic Belt. Young Nd model ages of ~ 1100 Ma obtained in this study do not support earlier models that the Hulan Group was either rifted from the North China Craton to the south or from the Jiamusi Block to the north, since both these blocks have much older Nd model ages. Detrital zircon U–Pb ages from a meta-sedimentary rock of the Hulan Group define a maximum age of deposition of 287 ± 6 Ma, not Early Palaeozoic as previously believed. Rb–Sr mineral isochron data indicate that metamorphism of the Hulan Group occurred at ~ 250 Ma. This conclusion is further supported by an age of 248 ± 4 Ma for the Dayushan granite, which intruded the Hulan Group. These new isotopic data, together with evaluation of the regional geology, indicate that final oceanic closure in the Chinese segment of the Central Asian Orogenic Belt took place in the Late Permian. The Hulan area is located along the Solonker–Xra Moron–Changchun suture that extends from Inner Mongolia across to central Jilin Province. It is this zone that marks the final ocean closure, resulting in the amalgamation of the North China and Siberian cratons.

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1. Introduction

The Central Asian Orogenic Belt (CAOB), located between the Siberian Craton in the north and the North China and Tarim cratons in the south (Fig. 1), is characterised by accretionary complexes that resulted from closure of multiple oceans (Sengör et al., 1993; Mossakovsky et al., 1994; Dobretsov et al., 1995; Dergunov, 2001; Dobretsov, 2003; Khain et al., 2003). The CAOB has been divided into the Altaids in the north, formed from the Tuva–Mongol arc (Sengör et al., 1993;

Sengör and Natal'in, 1996) and the Manchurides in the south, separated by the Solonker suture zone in Inner Mongolia. However, this zone is extremely complex (Xiao et al., 2003), with numerous linear sutures and faults extending ENE across the CAOB and the northern part of the North China Craton (NCC), and there is commonly confusion as to both the location and the nature of individual fault zones. Xiao et al. (2003), indicate that the major suture extends from Solonker to Linxi. Li and Wang (1983) and Wang and Liu (1986) extended the Solonker suture through to Changchun and due east from there. However, several authors have suggested that the main suture then swings southeastward in order to explain the distribution of ophiolites that outcrop in central and eastern Jilin Province (JBGMR, 1988; Huang, 1991; Shao and Tang, 1995).

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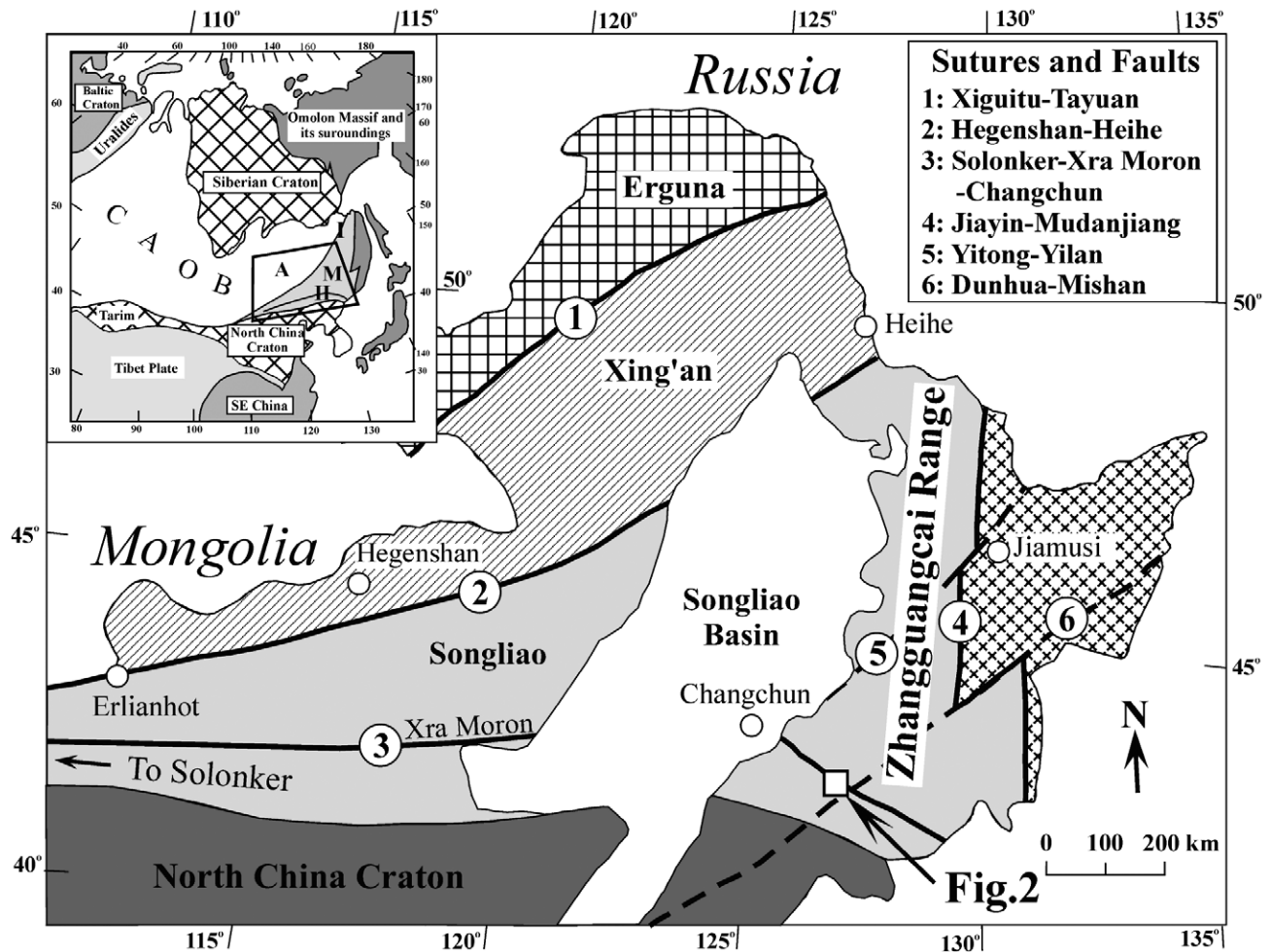


Fig. 1. Simplified geological map of NE China, showing the main tectonic subdivisions. The location of Fig. 2 is outlined. Inset shows location of the area with respect to the regional geological setting of China and Russia (outline shows location of studied area in NE China). A and M represent Altaids and Manchrides of Sengör and Natal'in (1996). (I) Hegenshan-Heihe suture zone, (II) Solonker-Xra Moron-Changchun suture zone.

Research indicates multi-stage accretion of island-arc complexes, oceanic sedimentary rocks and microcontinents in the CAOB between the North China and Siberian cratons, separated by numerous small, transient oceans (Tang, 1990; Dobretsov et al., 1995; Lamb and Badarch, 1997, 2001; Dergunov, 2001; Heubeck, 2001; Badarch et al., 2002; Khain et al., 2002; Kovalenko et al., 2004). However, controversy remains about where these oceans were located and when they were finally closed. Numerous investigators over the last decade have suggested that the Hegenshan-Heihe suture, which lies approximately 200 km north of the Solonker suture, may represent the final collision zone of the Paleo-Asian Ocean. This appeared to be supported by geochronological data, which show that the Hegenshan ophiolite is the youngest ophiolite in the area (Robinson et al., 1999; Nozaka and Liu, 2002). Alternatively, palaeontological data suggest that the Xra Moron-Changchun zone (eastern part of the Solonker suture) in central NE China could be the final suture zone (Fig. 1), since it separates Cathaysian (tropical) Permian floras in the south from Angaran (cool, temperate) species in the north

(Huang, 1983, 1991; Wang and Liu, 1986). This indicates that an extensive suture zone may extend across Inner Mongolia to Jilin Province, herein named the Solonker-Xra Moron-Changchun suture zone.

The Hulan Group in central Jilin Province lies in an important position with respect to the Solonker-Xra Moron-Changchun suture in central NE China. The purpose of this research was therefore to investigate the Hulan Group and associated granitoids to see if they provided critical evidence regarding the nature and timing of the suture, and its relation to final closure of the ocean between the NCC and Siberia, and whether or not the suture did swing southeastward to the Hulan area.

2. Regional geological setting

The CAOB consists of a series of Neoproterozoic-Phanerozoic accretionary complexes (Sengör and Natal'in, 1996; Wu et al., 2000; Khain et al., 2003). In China, the CAOB is characterised by widespread (>200,000 km²) emplacement of granites with low initial Sr ratios, positive $\epsilon_{\text{Nd}}(t)$ values

and young Nd model ages (500–1000 Ma in most cases) (Jahn et al., 2000; Wu et al., 2000). The orogenic belt is formed of a collage of microcontinental blocks (Ye et al., 1994; Wu et al., 1995): the Jiamusi Block in the southeast, the Songliao Block in the center, and the Xing'an and Erguna Blocks in the northwest, all separated by major faults (Fig. 1). Distinct lithological, stratigraphical, geochemical and geophysical characteristics have been recognised in each of these blocks (Ye et al., 1994).

The Jiamusi Block (Fig. 1) consists of three different rock series: the Mashan Complex, Heilongjiang Complex and both deformed and undeformed granites. The Mashan Complex is a granulite-facies metamorphic complex that was metamorphosed in the early Palaeozoic at ~500 Ma (Wilde et al., 1997, 2000). Late Pan-African deformed granitoids are associated with the Mashan Complex (Wilde et al., 2003). The Heilongjiang Complex in the western part of the Jiamusi Block is a blueschist-facies metamorphic complex that has been considered to represent the suture zone between the Jiamusi and Songliao Blocks (Cao et al., 1992). Recent study indicates, however, that suturing took place in the Mesozoic and was related to Pacific ocean subduction (Wu et al., 2007). Therefore, the Jiamusi block is not strictly a component of the CAOB.

The Songliao Block consists of accreted arc complexes (Wu et al., 2000), the Mesozoic to Cenozoic Songliao Basin and Mesozoic granites that make up the Zhangguangcai Range (Fig. 1). Local amphibolite facies metamorphic rocks have been identified, including the Dongfengshan Group in the north and the Hulan Group in the south (JBGMR, 1988; HBGMR, 1993). However, Phanerozoic granites occupy the majority of the block. The Songliao Basin formed in the late Mesozoic over a basement of Phanerozoic granites and Palaeozoic strata (Wu et al., 2001).

The Xing'an Block (Fig. 1) consists of voluminous Mesozoic volcanic rocks and granites, possibly overlying Precambrian basement. The Erguna Block is considered as the eastern extension of the Central Mongolian microcontinent, and is located in the northwestern most part of the region. Its geological evolution is largely unknown, as the area is heavily forested. This block is characterised by the presence of Proterozoic to Palaeozoic strata and granites (HBGMR, 1993), but no reliable geochronological data are available. Nd isotopic data for granites indicate a crustal formation age of 1680–1060 Ma, which is much older than that of the Xing'an and Songliao Blocks (Wu et al., 2003).

It is unclear when the Erguna Block collided with the Xing'an Block, although large ophiolitic complexes have been identified along the southern boundary of the former (Li, 1991). The development of Early Palaeozoic lithological assemblages in the northern part of the Xing'an Block, especially Ordovician island-arc associations and related blueschists, suggests that collision most probably occurred in the Ordovician (Yan et al., 1989; Zhang and Tang, 1989). Afterwards, the Songliao Block was accreted to

the Xing'an-Erguna composite block along the Hegen-shan-Heihe suture in the late Devonian to early Carboniferous (Ye et al., 1994). Subsequently, a regional extensional system developed in the late Palaeozoic (Tang, 1990; Wu et al., 1995).

The time of collision between the North China Craton and the composite Songliao-Xing'an-Erguna blocks (now the eastern part of the Central Mongolian microcontinent) is controversial. It has long been suggested that this collision took place in the Late Triassic (Zonenshain et al., 1985, 1990). But palaeomagnetic data suggest that they had amalgamated before the Late Permian and that the combined North China-Mongolian Block then collided with the Siberian Craton as a result of closure of the Mongol-Okhotsk ocean in the Late Jurassic (Zhao et al., 1990; Xu et al., 1997; Zorin, 1999; Kravchinsky et al., 2002).

3. The Hulan Group

The Hulan Group (Fig. 2) is a poorly exposed metamorphic complex in the CAOB, that contains the economically important Hongqiling Cu–Ni-bearing mafic-ultramafic rocks (Wu et al., 2004). It was previously considered that the Hulan Group was Early Palaeozoic in age (JBGMR, 1988), but prior to this current work there were no reliable geochronological data. The Hulan Group was metamorphosed from sub-greenschist to amphibolite facies (Lan and Shi, 1985; Meng, 1992), although little detailed work has been undertaken on either the metamorphism or the structural history.

3.1. Formations

The Hulan Group (Fig. 2) has traditionally been subdivided into the Huangyingtun, Xiaosangedingzi and Beichatun Formations (Lan and Shi, 1985; JBGMR, 1988; Chi et al., 1997). The lowermost Huangyingtun Formation is composed of tourmaline-bearing garnet-mica schist/gneiss, fine-grained biotite/hornblende-plagioclase gneiss, kyanite schist, quartzite and calc-silicate rocks (Chi et al., 1997). The Xiaosangedingzi Formation is dominated by marbles, quartzite and fine-grained gneiss (Chi et al., 1997), whereas the Beichatun Formation consists of sub-greenschist facies quartzite, phyllite, slate and limestone. *Tha mnopora* sp. and *Cladopora* sp. have been identified in limestone of the Beichatun Formation, indicating an early Palaeozoic age (Chi et al., 1997). However, the Hulan Group has undergone intense deformation, which makes this division into formations highly suspect.

Associated with the Hulan Group are a number of granitic intrusions (Fig. 2) that can be subdivided into two types. Undeformed quartz dioritic intrusions, including the Shancheshan, Yima, Hulan and Xiluoquan plutons (Fig. 2), were emplaced in the Jurassic between 178 and 171 Ma (Wu, unpublished data). The Dayushan pluton in the northeast (Fig. 2) covers an area of about 300 km² and is gneissic along its margins, although this deformation

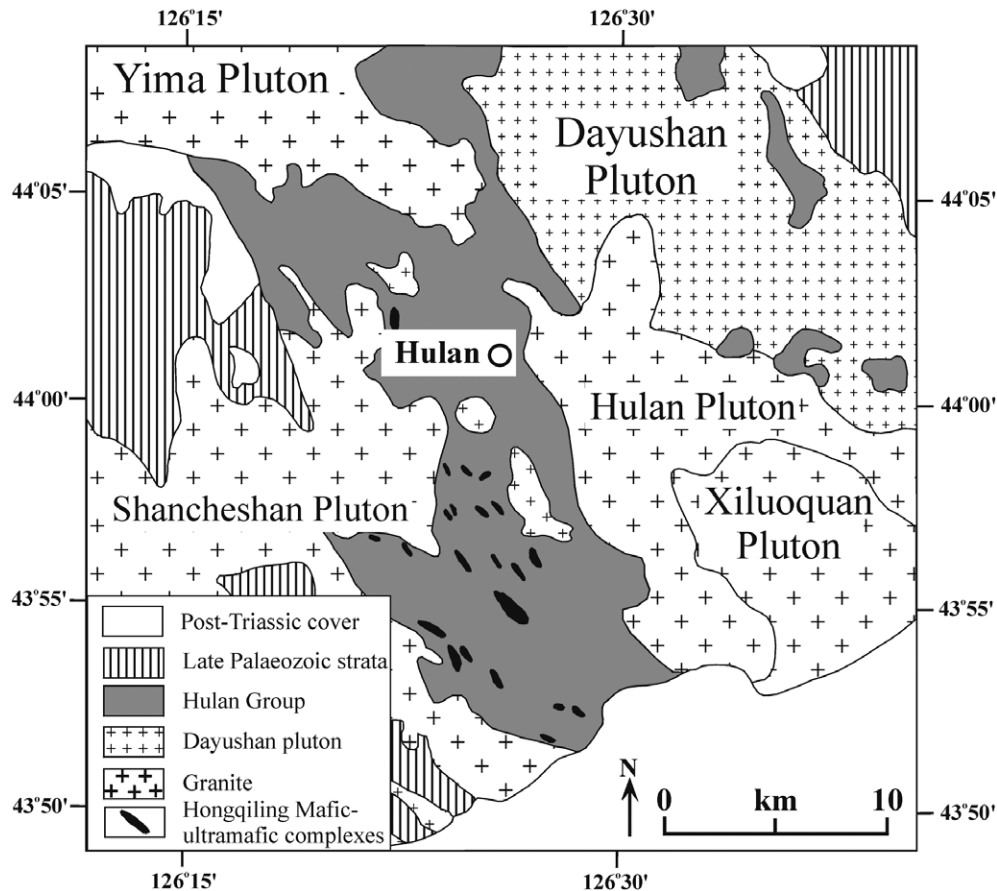


Fig. 2. Geological map of the Hulan Group and associated granites.

becomes weaker toward the interior. A geochemical study indicates that this pluton is a syn-collisional I-type and formed by partial melting of thickened mafic crust (Sun et al., 2004). Field relations suggest it was intruded into the Hulan Group (JBGM, 1988), although currently exposed contacts are sheared.

3.2. Deformation

Although this study is focussed on geochronology and not the structural and metamorphic evolution of the Hulan Group, it was necessary to collect specific information in and around the sample sites in order to constrain the geochronological interpretation. Three phases of deformation can be distinguished in the field on the basis of overprinting relationships. Early D_1 deformational fabrics are poorly preserved due to the later overprinting and transposition by subsequent deformational events. Evidence for this event includes small rootless intrafolial folds (F_1) and an associated early foliation within porphyroblasts in pelitic schists (Fig. 3a), defined by oriented mineral inclusions of plagioclase + quartz + muscovite + biotite $\pm$ chlorite within garnet, staurolite and biotite porphyroblasts. The main penetrative deformational event (D_2) produced

micro- to mesoscopic folds (F_2) and an axial planar foliation (Fig. 3b). A regional pervasive schistosity (S_2) is developed along the axial planes of the F_2 folds, defined by a strong preferred alignment of muscovite, chlorite and biotite. The D_3 deformation produced regional-scale NW–SE trending folds, which deflect the earlier structural features. F_3 are open folds with nearly vertical axial planes, but planar fabrics (S_3) unequivocally associated with F_3 are rarely observed. The apparent absence of S_3 foliations related to crystallisation of micaceous minerals in the F_3 hinges suggests that the D_3 phase of deformation occurred under relatively lower temperature conditions than D_2 . In addition, the D_3 deformation also produced a series of NW–SE trending faults.

3.3. Metamorphism

Pelitic schists preserve the most complete deformational and metamorphic records of the Hulan Group. Thus, we have focused on these rocks in the following discussion of the metamorphism. The metamorphic minerals in the pelitic schists include sericite, muscovite, biotite, chlorite, garnet, staurolite, kyanite, sillimanite, cordierite and plagioclase (Fig. 3a–f). Microstructures and mineral reaction

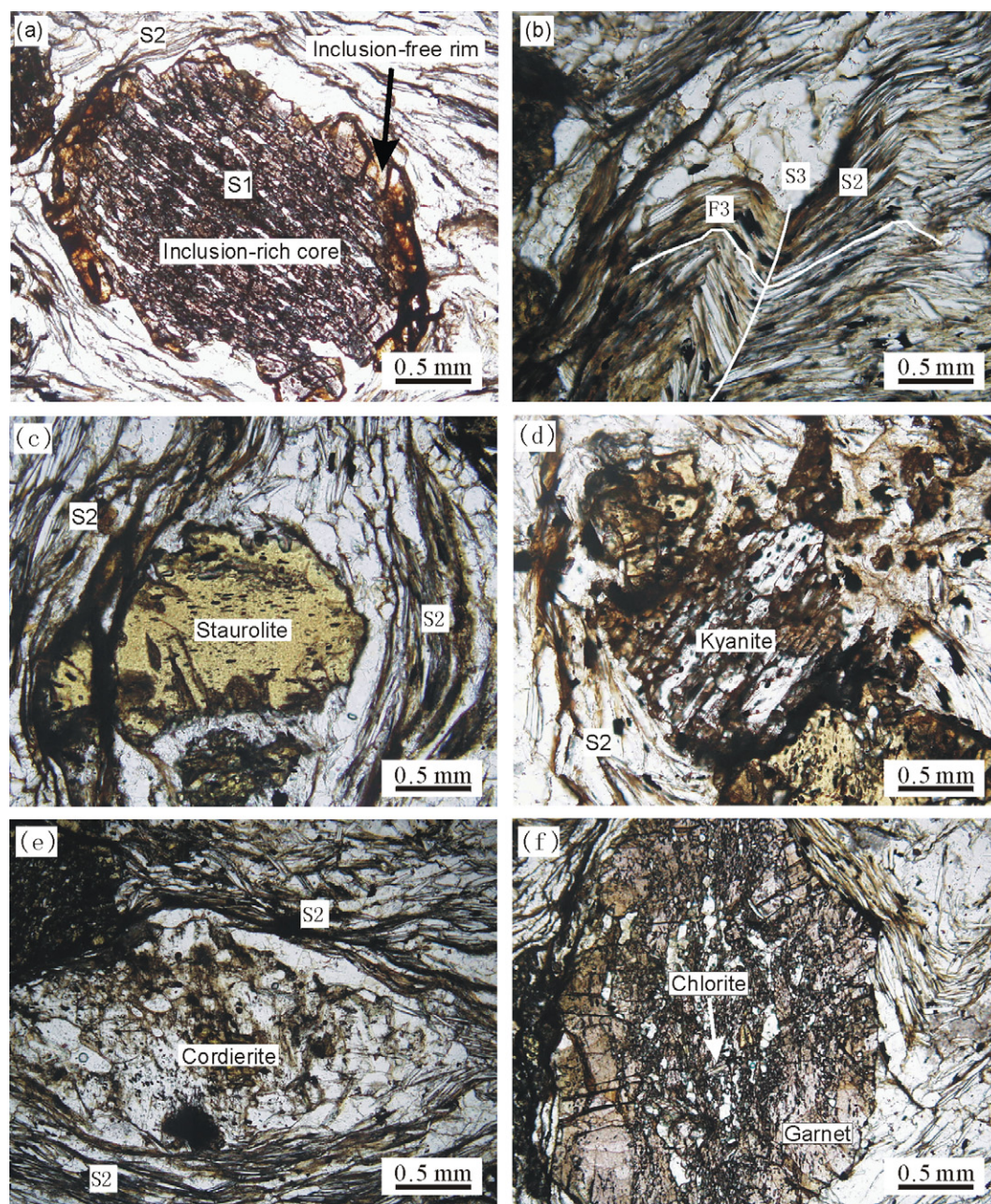


Fig. 3. Photomicrographs of one of the analysed pelitic metamorphic rocks from the Hulan Group (HYIII-4). (a) Early foliation S_1 defined by oriented mineral inclusions within a garnet porphyroblast that also shows an inclusion-free rim; (b) relationships between S_2 and S_3 . Porphyroblasts of (c) staurolite, (d) kyanite and (e) cordierite; (f) relict chlorite contained in garnet, but not present in the matrix.

relations reveal that these minerals formed during three distinct metamorphic stages (M_1 to M_3) in association with the deformational episodes (D_1 to D_3).

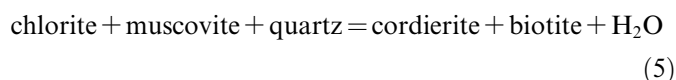
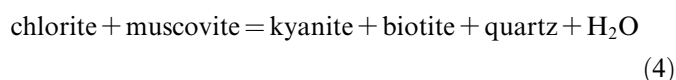
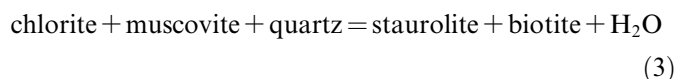
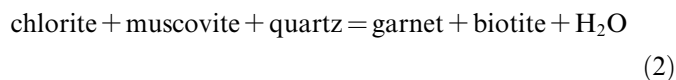
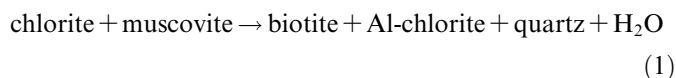
M_1 is represented by fine-grained (0.1–0.2 mm) muscovite, biotite, plagioclase, quartz and chlorite within porphyroblasts of garnet, staurolite, cordierite and kyanite. In most cases, these inclusions are oriented, defining the internal schistosity (S_1) within the porphyroblasts (Fig. 3a). S_1 is oblique to the external matrix foliation (S_2) (see Fig. 3c–e), suggesting that M_1 developed earlier than the second phase of deformation (D_2). Garnet porphyroblasts generally have an inclusion-rich core, but

rarely have thin inclusion-free rims. Such inclusion-rich garnet cores are interpreted as having formed during the D_1 deformation. Therefore, the most complete M_1 mineral assemblage identified is muscovite + biotite + plagioclase + quartz + chlorite $\pm$ garnet (inclusion-rich core). This is a typical greenschist facies assemblage, with a metamorphic temperature range of 400–500 °C over a wide range in pressure (Yardley, 1989).

The M_2 stage is characterised by the strong preferential alignment (S_2) and enlargement of the matrix-type minerals muscovite + biotite + quartz + plagioclase and the development of porphyroblasts including biotite, garnet,

staurolite, kyanite and cordierite. The matrix-type muscovite, biotite, quartz and plagioclase are larger in grain size than the corresponding M_1 mineral inclusions within porphyroblasts, suggesting that they continued to grow during M_2 . The characteristic mineral assemblage of M_2 is plagioclase + quartz + biotite + muscovite + garnet (inclusion-free rim) + staurolite + kyanite $\pm$ cordierite.

In the thin sections, no reaction relations between porphyroblasts of biotite, garnet, staurolite, kyanite and cordierite have been observed and, in most cases, these porphyroblast minerals have straight contacts, suggesting that they formed simultaneously. Since minor relict chlorite only occurs within these porphyroblasts (Fig. 3f), but is not present in the matrix, these porphyroblasts were most probably produced through one or more of the following chlorite- and muscovite-consuming metamorphic reactions (Muller and Saxena, 1977; Zhao and Cawood, 1999):



Microstructural relations also suggest that the M_2 mineral assemblages formed simultaneously with, or after, D_2 but before D_3 . For example, some garnet and staurolite porphyroblasts have mineral inclusion trails that are linear or curved and continuous with the matrix S_2 schistosity. The curved trail (“snowball structure”), indicates syntectonic growth with respect to the S_2 matrix foliation. Therefore, these garnet porphyroblasts are interpreted to have formed during D_2 .

The M_3 event represents later retrograde metamorphism, with chlorite and muscovite replacing staurolite, garnet, kyanite and cordierite porphyroblasts. Therefore, the representative mineral assemblage for M_3 is chlorite + sericite (muscovite), representing sub-greenschist facies metamorphism. These retrograde minerals do not show a preferred orientation, suggesting that they formed after the final deformation episode (post- D_3).

4. Geochronological data

4.1. Analytical techniques

Several isotopic methods have been used to constrain the protolith and metamorphic ages of the Hulan Group and crystallisation age of the associated granites.

Laser ablation ICPMS (LA-ICPMS) U–Pb zircon analyses were obtained in order to constrain the ages of detrital zircons from pelitic schist of the Hulan Group. Although the LA-ICPMS technique was only developed in the last decade (Feng et al., 1993; Hirata and Nesbitt, 1995), recent studies have shown that it is a powerful tool in geochronology (Horn et al., 2000; Li et al., 2001; Kosler et al., 2002; Kosler and Sylvester, 2003). Generally, the precision of individual spot analyses using LA-ICPMS is inferior to that of TIMS (thermal ionisation mass spectrometry) and SHRIMP, but it has been greatly improved by applying the external correction method of Ballard et al. (2001) using standard zircons. In this study, LA-ICPMS analyses were conducted using a 193 nm Elan 6100 DRC ICPMS housed at the Department of Geology, Northwest University in Xi'an. Zircon 91500 was used as an external standard and standard silicate glass NIST was used to optimize the machine: the spot diameter was 40 μm . Because of a relatively high ^{204}Hg , ^{204}Pb was not measured and so no common lead correction could be applied (Yuan et al., 2004).

U–Pb zircon isotopic analyses of the Dayushan granite were obtained using a VG 354 mass spectrometer with Daly collector at the Tianjin Institute of Geology and Mineral Resources, China Geological Survey. A fist-size sample was crushed to less than 200 μm , and mineral separations were undertaken after elutriation. Approximately 30 grains of zircon were hand-picked and homogeneous zircons with no inclusions or fractures were chosen for isotopic analyses. Dissolution of zircon and U and Pb chemical separation followed the procedures of Krogh (1973, 1982) with slight modification. ^{205}Pb – ^{235}U spike was added to the zircon samples in a 0.25 ml fluorine-plastic capsule for zircon dissolution. The final isolated U and Pb were loaded onto a Re filament with silica gel-phosphoric acid. All U and Pb data were corrected for mass fractionation. The blanks were 0.03–0.05 ng for Pb and 0.002–0.004 ng for U.

Sr and Nd isotopic data were obtained in Beijing at the Institute of Geology and Geophysics, Chinese Academy of Sciences, using a multi-collector VG 354 mass spectrometer in static mode. Detailed chemical separation and isotopic measurement procedures can be found in Qiao et al. (1987, 1990). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were normalized against the value of $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were normalized against the value of $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. During the period of data acquisition, Sr standard NBS607 gave $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 1.20032 ± 28 (2σ), and Nd standard BCR-1 yielded $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of 0.512626 ± 9 (2σ). Analytical precisions of $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ are generally better than 2% and 0.5%, respectively.

U–Pb zircon and Rb–Sr whole-rock isochron ages were calculated using the regression program ISOPLOT (Ludwig, 1999), and the errors quoted in age computation represent ± 2 standard deviations.

4.2. Protolith age of metasedimentary rocks of the Hulan Group

Rocks of the Hulan Group have not previously been well constrained in terms of their protolith ages. Nd model ages of three paragneisses and one mica schist selected from what has been mapped as the Huangyingtun Formation are shown in Table 1 and give model ages of ~ 1100 Ma that vary over a narrow range. In contrast, rocks from the North China Craton are Archaean in age and those of the Jiamusi Block are mainly Mesoproterozoic (Fig. 4) (Wu et al., 2004, 2005). Therefore, sedimentary rocks of the Hulan Group must have been derived from a source distinct from the North China Craton and Jiamusi Block.

It has long been considered that the Hulan Group was deposited in the Early Palaeozoic, based on an Rb–Sr whole-rock isochron age of 524 ± 16 Ma (JBGMR, 1988), although the analytical data were not presented. Also, some Late Palaeozoic strata in the region are unmetamorphosed, while the Hulan Group is metamorphosed from sub-greenschist to amphibolite facies, suggesting it formed earlier. To test this hypothesis, we carried out LA-ICPMS dating of detrital zircons from a pelitic schist (kyanite-bearing garnet mica schist, FW02-212) from the so-called Huanhyingtun Formation of the Hulan Group (Table 2). Cathodoluminescence (CL) images of detrital zircons show characteristic igneous oscillatory zoning, indicating their origin from an igneous protolith. This is further supported by relatively high U concentrations and Th/U ratios for these zircons (Table 2). The U–Pb zircon data (Fig. 5a) show a spread of ages along concordia from ~ 400 Ma to ~ 287 Ma; a number of more discordant grains are also present. There is a possible bimodal distribution to the data (Fig. 5b), with peaks at ~ 345 Ma and ~ 325 Ma. Thus, the zircons most likely reflect multiple sources, but from an environment with a limited age distribution, in keeping with the view that they evolved in an arc setting. The youngest zircon, with an age of 287 ± 6 Ma (Fig. 5a), defines the maximum age of deposition of the Hulan Group and establishes that deposition was not earlier than the Permian; and clearly not Early Palaeozoic as previously believed (JBGMR, 1988).

4.3. Timing of metamorphism in the Hulan Group

The Hulan Group was metamorphosed up to amphibolite facies and, under such conditions, Sm-Nd isotopic analyses of garnet should potentially date the timing of metamorphism. Unfortunately, the presence of abundant mineral inclusions within the garnet crystals we examined precludes the application of this method. Instead, we chose muscovite and biotite from three mica schist samples for Rb-Sr isotopic dating. Of the three samples (Table 1), two samples (H7III-1 and H1-1) yielded muscovite-whole rock (WR) isochron ages of 237 ± 8 Ma and 253 ± 10 Ma, respectively, and biotite-WR (-Plagioclase)

Table 1
Sr-Nd isotopic data of the Hulan Group in Jilin Province

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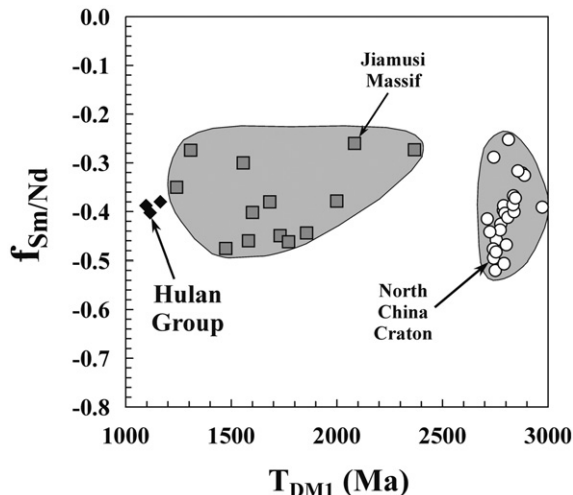


Fig. 4. Nd isotopic data for the protoliths of the Hulan Group, showing that the Nd composition of the Hulan Group is different from that of both the North China Craton to the south and the Jiamusi Block to the north. Data source: Wu et al. (2003, 2004b).

isochron ages of 216 ± 4 Ma and 217 ± 4 Ma (Fig. 6a and b). Another sample (HYT01-2) collected near the Dunhua-Mishan Fault (Fig. 1) yielded a biotite-WR Rb/Sr isochron age of 188 ± 4 Ma (Fig. 6c), much younger than the other two samples. The muscovite closure temperature of ~ 500 °C means that the muscovite-WR Rb/Sr isochron ages of 250–240 Ma most likely represent cooling ages. Notwithstanding the fact that two-point isochrons are less than ideal, owing to the weight placed on the precision of each data point, it is nonetheless clear that the ages indicate cooling from peak metamorphic conditions during the Permo-Triassic. The slightly younger biotite-WR Rb/Sr isochron ages of 216 ± 4 Ma and 217 ± 4 Ma are identical to the age of the Hongqiling Cu–Ni-bearing ultramafic-mafic intrusions dated at 216 ± 5 Ma by Wu et al. (2005), suggesting that the age of 217–216 Ma reflects thermal overprinting during emplacement of this major ultramafic-mafic igneous suite (Fig. 2). The age of 188 ± 4 Ma for sample HYT01-2, collected near the Dunhua-Mishan Fault, may record the time of faulting or else has no real geological significance.

4.4. Emplacement age of the Dayushan granite

Controversy has surrounded the emplacement age of the Dayushan granite, with available isotopic data including a biotite K–Ar age of ~ 419 Ma, an apatite Pb–Pb age of ~ 400 Ma and a mineral Rb–Sr isochron age of 220.8 ± 0.8 Ma (JBGMR, 1988). Although these fail to place constraints on the emplacement age of this granite, it has generally been considered that this pluton was formed in the Early Palaeozoic (JBGMR, 1988).

In order to determine the crystallisation age of the Dayushan granite, zircons separated from sample 9923-1 were dated by the TIMS U–Pb method and the data are

listed in Table 3 and shown on a concordia diagram in Fig. 7. The weighted mean age of four analyses is 248 ± 4 Ma, identical within error to the metamorphic age (240–250 Ma) obtained by Rb–Sr techniques for the Hulan Group samples described above. Therefore, the Dayushan pluton was emplaced in the Late Palaeozoic, not in the Early Palaeozoic as was previously believed (JBGMR, 1988). This is consistent with our field observation that the Dayushan pluton intrudes the Hulan Group, whose maximum depositional age is constrained at $\sim 287 \pm 6$ Ma, the age of the youngest detrital zircon. The similarity in age of the pluton to the metamorphic age of the Hulan Group suggests it may have been emplaced during metamorphism and deformation, consistent with its syn-collisional origin obtained from a previous geochemical study (Sun et al., 2004) and its strong deformation fabrics.

5. Tectonic implications

5.1. Tectonic significance of the Hulan Group

Although there is no consensus on the appropriate orogenic model for the CAOB (Wang and Liu, 1986, 1991; Tang, 1990; Hsu et al., 1991; Sengör et al., 1993; Kovalevko et al., 1995, 2004; Xiao et al., 2003), it is commonly accepted that this belt is characterised by accretion of island-arcs. As stated earlier, the Hulan Group is in part composed of felsic paragneiss, pelitic schist and marble that show intense deformation. Fossils contained in weakly deformed limestone of the Beichatun Formation indicate these carbonates formed in the Early Palaeozoic. However, the clastic sediments were deposited in the Permian, as constrained by the present study, suggesting that the Hulan Group is composed of sediments of different ages and is thus not a true stratigraphic succession. We suggest that the Hulan Group is composed of tectonic slices that were juxtaposed during deformation along the Solonker-Xra Moron–Changchun suture, although specific structural studies need to be undertaken to confirm this view. However, when compared to the two ‘lower’ formations, the weak deformation, low metamorphic grade and preservation of fossils characterising the Beichatun Formation, attests to the great variability exhibited within the Hulan Group.

Palaeozoic strata are widely distributed in NE China. Shao and Tang (1995) have suggested that strata of this age in the Yanbian area form a tectonic melange and preliminary evidence from central Jilin Province (Jahn, 2003) suggests that certain chaotic assemblages of marble, basalt, chert and deformed pillow lava in a serpentinite matrix may also represent such a zone. The coexistence of carbonates (shallow water) and clastic sediments (possibly flysch of deep-water origin) in the Hulan Group is similar to many areas in NE China, suggesting that they may have evolved in subduction–accretion complexes.

Table 2
Zircon LA-ICPMS U–Pb analytical data of the Hulan metasedimentary rock (FW02-212)

Spot	U (ppm)	Th (ppm)	Th/U	Pb (ppm)	Isotopic ratios (1 σ)			Age (Ma, 1 σ)	
					$^{207}\text{Pb}/^{206}\text{Pb}$	$^{207}\text{Pb}/^{235}\text{U}$	$^{206}\text{Pb}/^{238}\text{U}$	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$
FW02-212-01	124	74	0.60	9	0.0638 $\pm$ 46	0.518 $\pm$ 36	0.0589 $\pm$ 11	369 $\pm$ 07	424 $\pm$ 24
FW02-212-02	311	91	0.29	17	0.0665 $\pm$ 39	0.460 $\pm$ 26	0.0501 $\pm$ 09	315 $\pm$ 05	384 $\pm$ 18
FW02-212-03	103	49	0.48	7	0.0592 $\pm$ 68	0.428 $\pm$ 47	0.0524 $\pm$ 15	330 $\pm$ 09	362 $\pm$ 34
FW02-212-04	317	107	0.34	19	0.0529 $\pm$ 30	0.364 $\pm$ 20	0.0499 $\pm$ 08	314 $\pm$ 05	315 $\pm$ 15
FW02-212-05	296	119	0.40	18	0.0535 $\pm$ 34	0.375 $\pm$ 23	0.0508 $\pm$ 08	320 $\pm$ 05	324 $\pm$ 17
FW02-212-06	282	264	0.94	21	0.0538 $\pm$ 37	0.401 $\pm$ 27	0.0541 $\pm$ 10	339 $\pm$ 06	342 $\pm$ 19
FW02-212-07	197	112	0.57	14	0.0599 $\pm$ 33	0.462 $\pm$ 25	0.0559 $\pm$ 09	351 $\pm$ 05	385 $\pm$ 17
FW02-212-08	105	55	0.52	7	0.0531 $\pm$ 49	0.402 $\pm$ 36	0.0550 $\pm$ 12	345 $\pm$ 07	343 $\pm$ 26
FW02-212-09	267	154	0.57	18	0.0518 $\pm$ 43	0.370 $\pm$ 30	0.0519 $\pm$ 11	326 $\pm$ 07	320 $\pm$ 22
FW02-212-10	169	81	0.48	12	0.0791 $\pm$ 49	0.593 $\pm$ 35	0.0544 $\pm$ 10	341 $\pm$ 06	473 $\pm$ 22
FW02-212-11	160	91	0.57	11	0.0542 $\pm$ 31	0.419 $\pm$ 23	0.0561 $\pm$ 09	352 $\pm$ 05	355 $\pm$ 16
FW02-212-12	123	102	0.83	10	0.0546 $\pm$ 56	0.430 $\pm$ 43	0.0572 $\pm$ 15	358 $\pm$ 09	363 $\pm$ 31
FW02-212-13	162	77	0.48	11	0.0546 $\pm$ 66	0.401 $\pm$ 47	0.0532 $\pm$ 15	334 $\pm$ 09	342 $\pm$ 34
FW02-212-14	69	42	0.60	6	0.0537 $\pm$ 97	0.347 $\pm$ 61	0.0468 $\pm$ 20	295 $\pm$ 12	302 $\pm$ 46
FW02-212-15	164	79	0.48	11	0.0538 $\pm$ 30	0.409 $\pm$ 22	0.0552 $\pm$ 08	346 $\pm$ 05	348 $\pm$ 16
FW02-212-16	482	414	0.86	34	0.0621 $\pm$ 22	0.445 $\pm$ 15	0.0520 $\pm$ 06	327 $\pm$ 04	374 $\pm$ 11
FW02-212-17	228	130	0.57	17	0.0700 $\pm$ 47	0.506 $\pm$ 33	0.0524 $\pm$ 10	330 $\pm$ 06	416 $\pm$ 22
FW02-212-18	111	62	0.55	8	0.0547 $\pm$ 52	0.427 $\pm$ 39	0.0566 $\pm$ 13	355 $\pm$ 08	361 $\pm$ 28
FW02-212-19	193	55	0.28	12	0.0517 $\pm$ 41	0.324 $\pm$ 25	0.0455 $\pm$ 09	287 $\pm$ 06	285 $\pm$ 19
FW02-212-20	138	89	0.65	11	0.0732 $\pm$ 48	0.565 $\pm$ 35	0.0560 $\pm$ 11	351 $\pm$ 07	455 $\pm$ 23
FW02-212-21	185	128	0.70	13	0.0538 $\pm$ 57	0.395 $\pm$ 41	0.0534 $\pm$ 14	335 $\pm$ 09	338 $\pm$ 30
FW02-212-22	189	78	0.41	13	0.0533 $\pm$ 37	0.414 $\pm$ 28	0.0564 $\pm$ 10	354 $\pm$ 06	352 $\pm$ 20
FW02-212-23	93	56	0.60	6	0.0532 $\pm$ 60	0.407 $\pm$ 45	0.0555 $\pm$ 15	348 $\pm$ 09	347 $\pm$ 32
FW02-212-24	206	61	0.29	12	0.0526 $\pm$ 31	0.377 $\pm$ 22	0.0520 $\pm$ 08	327 $\pm$ 05	325 $\pm$ 16
FW02-212-25	78	48	0.61	6	0.0922 $\pm$ 57	0.691 $\pm$ 41	0.0543 $\pm$ 11	341 $\pm$ 07	533 $\pm$ 25
FW02-212-26	448	396	0.88	32	0.0531 $\pm$ 24	0.378 $\pm$ 16	0.0516 $\pm$ 07	324 $\pm$ 04	325 $\pm$ 12
FW02-212-27	221	119	0.54	16	0.0663 $\pm$ 47	0.479 $\pm$ 33	0.0524 $\pm$ 11	329 $\pm$ 06	398 $\pm$ 22
FW02-212-28	214	125	0.59	16	0.0546 $\pm$ 35	0.449 $\pm$ 28	0.0596 $\pm$ 10	373 $\pm$ 06	377 $\pm$ 19
FW02-212-29	185	77	0.41	12	0.0629 $\pm$ 39	0.487 $\pm$ 29	0.0562 $\pm$ 10	352 $\pm$ 06	403 $\pm$ 20
FW02-212-30	200	84	0.42	14	0.0531 $\pm$ 35	0.403 $\pm$ 26	0.0551 $\pm$ 09	346 $\pm$ 06	344 $\pm$ 18
FW02-212-31	220	120	0.55	15	0.0532 $\pm$ 24	0.417 $\pm$ 18	0.0569 $\pm$ 07	357 $\pm$ 05	354 $\pm$ 13
FW02-212-32	178	78	0.44	11	0.0565 $\pm$ 40	0.381 $\pm$ 26	0.0489 $\pm$ 09	308 $\pm$ 06	328 $\pm$ 19
FW02-212-33	242	115	0.48	15	0.0558 $\pm$ 36	0.391 $\pm$ 24	0.0508 $\pm$ 09	319 $\pm$ 05	335 $\pm$ 18
FW02-212-34	80	52	0.65	8	0.1312 $\pm$ 81	0.988 $\pm$ 57	0.0546 $\pm$ 13	343 $\pm$ 08	698 $\pm$ 29
FW02-212-35	186	103	0.55	12	0.0525 $\pm$ 51	0.387 $\pm$ 36	0.0535 $\pm$ 13	336 $\pm$ 08	332 $\pm$ 27
FW02-212-36	124	38	0.31	7	0.0597 $\pm$ 45	0.403 $\pm$ 30	0.0490 $\pm$ 10	308 $\pm$ 06	344 $\pm$ 21
FW02-212-37	113	44	0.39	7	0.0577 $\pm$ 47	0.439 $\pm$ 35	0.0552 $\pm$ 12	346 $\pm$ 07	369 $\pm$ 25
FW02-212-38	90	52	0.58	7	0.0536 $\pm$ 50	0.472 $\pm$ 43	0.0639 $\pm$ 15	399 $\pm$ 09	393 $\pm$ 30
FW02-212-39	105	68	0.65	8	0.0620 $\pm$ 56	0.440 $\pm$ 39	0.0515 $\pm$ 13	324 $\pm$ 08	370 $\pm$ 27
FW02-212-40	595	383	0.64	37	0.0545 $\pm$ 25	0.383 $\pm$ 17	0.0510 $\pm$ 07	321 $\pm$ 04	329 $\pm$ 12
FW02-212-41	189	93	0.49	13	0.0544 $\pm$ 25	0.429 $\pm$ 19	0.0572 $\pm$ 08	359 $\pm$ 05	362 $\pm$ 14
FW02-212-42	191	113	0.59	14	0.0539 $\pm$ 55	0.372 $\pm$ 37	0.0501 $\pm$ 13	315 $\pm$ 08	321 $\pm$ 27
FW02-212-43	160	70	0.44	10	0.0528 $\pm$ 28	0.399 $\pm$ 20	0.0548 $\pm$ 08	344 $\pm$ 05	341 $\pm$ 15
FW02-212-44	209	106	0.51	14	0.0578 $\pm$ 45	0.397 $\pm$ 30	0.0498 $\pm$ 10	313 $\pm$ 06	339 $\pm$ 22
FW02-212-45	222	98	0.44	15	0.0558 $\pm$ 25	0.419 $\pm$ 18	0.0545 $\pm$ 07	342 $\pm$ 04	355 $\pm$ 13
FW02-212-46	236	123	0.52	16	0.0558 $\pm$ 39	0.371 $\pm$ 25	0.0482 $\pm$ 09	304 $\pm$ 05	321 $\pm$ 19
FW02-212-47	248	146	0.59	17	0.0535 $\pm$ 32	0.412 $\pm$ 24	0.0558 $\pm$ 09	350 $\pm$ 05	350 $\pm$ 17
FW02-212-48	117	61	0.52	9	0.0931 $\pm$ 59	0.731 $\pm$ 44	0.0570 $\pm$ 12	357 $\pm$ 07	557 $\pm$ 26
FW02-212-49	106	68	0.64	8	0.1050 $\pm$ 46	0.821 $\pm$ 34	0.0567 $\pm$ 09	356 $\pm$ 05	608 $\pm$ 19
FW02-212-50	157	70	0.45	13	0.0886 $\pm$ 38	0.717 $\pm$ 29	0.0586 $\pm$ 09	367 $\pm$ 05	549 $\pm$ 17
FW02-212-51	123	38	0.31	7	0.0540 $\pm$ 53	0.355 $\pm$ 34	0.0477 $\pm$ 11	300 $\pm$ 07	308 $\pm$ 25
FW02-212-52	113	43	0.38	7	0.0635 $\pm$ 64	0.431 $\pm$ 42	0.0492 $\pm$ 13	310 $\pm$ 08	364 $\pm$ 30
FW02-212-53	89	51	0.57	7	0.0831 $\pm$ 57	0.624 $\pm$ 41	0.0545 $\pm$ 12	342 $\pm$ 07	492 $\pm$ 25

Although the age of peak metamorphism of the Hulan Group has not been precisely dated, the retrograde cooling age at $\sim 500^\circ\text{C}$ has been constrained at ~ 250 Ma, as defined by the muscovite-whole rock Rb–Sr isochron ages of the mica schist and the zircon U–Pb age of the intrusive Dayushan pluton.

5.2. Metamorphic P – T path of the Hulan Group

Much of the recent research in metamorphic petrology has shown that modern field- and thermodynamics-based metamorphic investigations, in combination with lithological, structural and geochronological considerations, can be

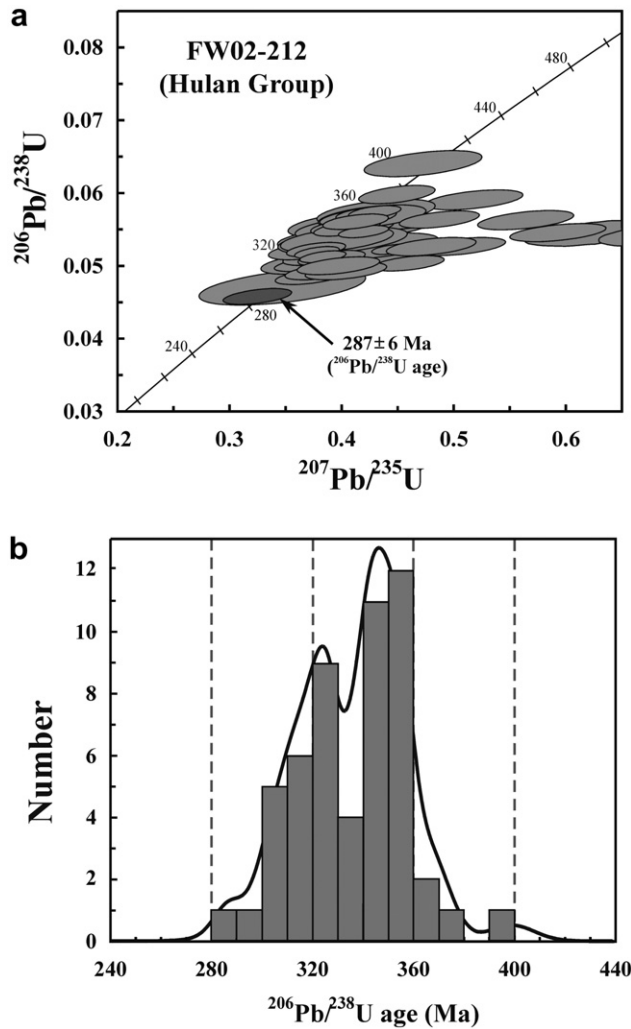


Fig. 5. (a) U–Pb concordia plot for detrital zircons from pelitic schist (FW02-212) from the Hulan Group, (b) cumulative probability diagram showing age variation.

directed towards understanding the tectonic setting and processes that were operative during metamorphism. For example, clockwise, especially isothermal decompressional, P – T paths are considered to be related to collision environments (Brown, 1993), whereas anticlockwise, especially isobaric cooling, P – T paths are related to the intrusion and underplating of mantle magmas, which can occur in continental magmatic arc regions, hot spots related to mantle plumes or rift environments (Bohlen, 1991). Because rocks of the Hulan Group may have been assembled by juxtaposition of once disparate sequences (see above) that may have undergone different metamorphic conditions, albeit during the same tectonic event, it is difficult to assign a P – T path with any confidence based on the limited sampling that we undertook. For instance, the development of cordierite, kyanite and staurolite during M_1 is unlikely to be a stable assemblage in these rocks. Nonetheless, we have attempted, in a very cursory way, to estimate a likely P – T path based on our limited data set.

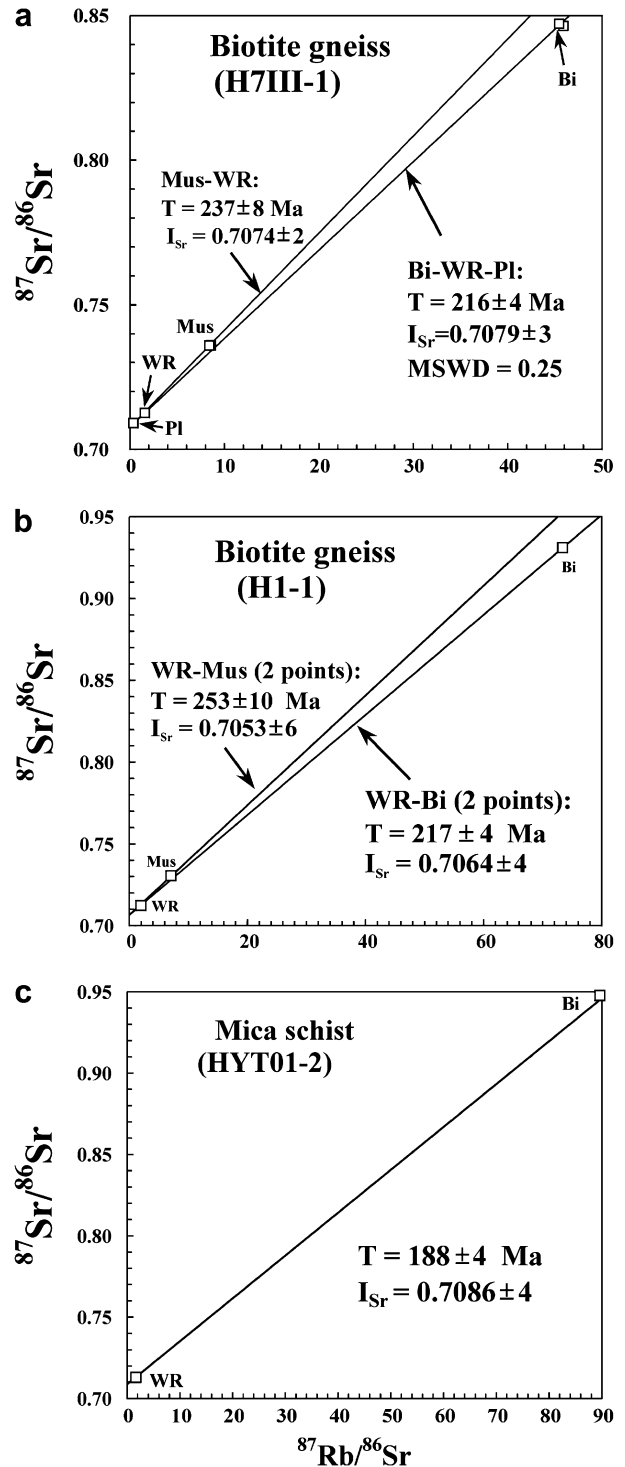


Fig. 6. Rb–Sr isochron diagrams for the mica schists from the Hulan Group; (a) H7III-1, (b) H1-1 and (c) HYT01-2.

Fig. 8 is a P – T pseudosection of the KFMASH grid built up for a common pelitic AFM composition from Powell et al. (1998). It shows a prograde sequence of mineral assemblages similar to those observed in the pelitic schists of the Hulan Group. For example, the domains garnet + biotite + chlorite, staurolite + garnet + biotite, kyanite + garnet + biotite, and biotite + chlorite, each of

Table 3
Zircon U–Pb isotopic data of the Dayushan granite (9923-1)

Spot	U (ppm)	Pb (ppm)	Isotopic ratio					Age(Ma)		
			$^{206}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{206}\text{Pb}$	$^{207}\text{Pb}/^{235}\text{U}$	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{206}\text{Pb}$	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$	$^{207}\text{Pb}/^{206}\text{Pb}$
1	304	13	931	0.1438	0.279 ± 09	0.0392 ± 07	0.0516 ± 14	248 ± 04	250 ± 08	267 ± 07
2	122	5	1294	0.1817	0.274 ± 13	0.0395 ± 12	0.0504 ± 16	250 ± 07	246 ± 11	212 ± 07
3	94	5	143	0.1554	0.254 ± 28	0.0379 ± 21	0.0487 ± 43	240 ± 13	230 ± 26	134 ± 12

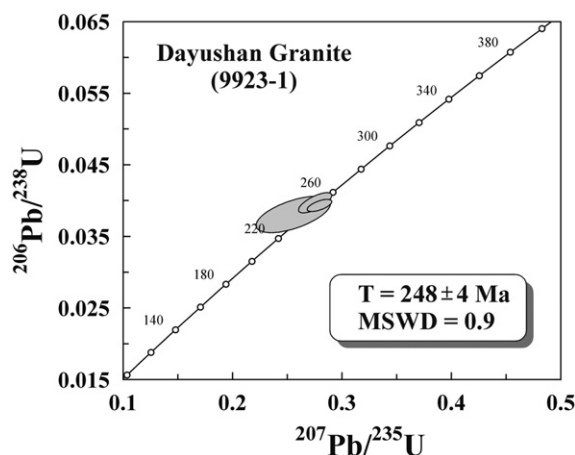


Fig. 7. U–Pb concordia diagram for zircons from the syn-collisional Dayushan granite (sample 9923-1).

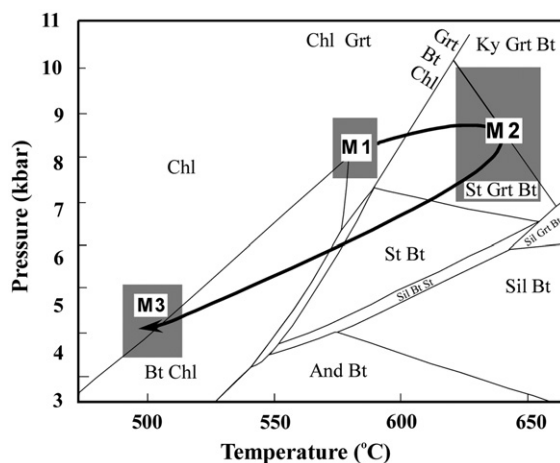


Fig. 8. Qualitative P – T path inferred for pelitic schists of the Hulan Group. Mineral abbreviations are the same as in Kretz (1983).

which has muscovite + quartz + H_2O in excess, can be used to show the relationship of the M_1 , M_2 and M_3 assemblages in the pelitic schists of the Hulan Group. The transition from M_1 to M_2 shows an increase in P and T , whereas the transition from M_2 to M_3 represents a decrease in P and T (Fig. 8). Taken together, a clockwise P – T path is inferred for the Hulan pelitic schists (Fig. 8).

Although a quantitative P – T path cannot be defined from this grid because of the lack of sufficient critical reactions and mineral compositional information, the qualitatively inferred P – T path of the Hulan Group is similar to that of Phanerozoic orogenic belts (Brown, 1993). As

reasoned above, a clockwise P – T path generally accords with a metamorphic event that is related to continent–continent collision. The coeval increase of temperature and pressure in the period from M_1 to M_2 reflects crustal thickening during the early stage of collision, and the coeval pressure and temperature decrease suggests a post-peak exhumation process after collision. Therefore, the metamorphic history of the Hulan Group provides important information about the collision between the Siberian and North China cratons. Although the age of peak metamorphism of the Hulan Group has not been precisely dated, the retrograde cooling age has been constrained at ~ 250 Ma, as defined by the muscovite-whole rock Rb–Sr isochron ages of the mica schist and the zircon U–Pb age of the syn-tectonic Dayushan pluton.

5.3. Final closure of the ocean between NE China and Siberia

Recent studies show that the history of oceans in the CAOB between the North China and Siberian cratons can be traced back to the Mesoproterozoic (Khain et al., 2002). In the Neoproterozoic and Palaeozoic, a number of large ocean basins existed (Dobretsov et al., 1995; Khain et al., 2002, 2003), although there is controversy surrounding where these oceans were located and when they closed. One school of thought proposes that the Hegenshan–Heihe suture (between the Xing'an and Songliao blocks) in Inner Mongolia (Fig. 1) represents the final closure of the ocean between the North China and Siberian cratons and that it formed during the Late Devonian–Early Carboniferous (Zhang and Tang, 1989; Tang, 1990; Guo, 1991). This appears to be supported by palaeomagnetic data that suggest the North China Craton was amalgamated with the Mongolian continent initially along the Hegenshan–Heihe belt and that this composite continental block then finally collided with the Siberian Craton along the Mongol–Okhotsk belt (Muller et al., 1991; Zorin, 1999; Kravchinsky et al., 2002). However, other workers have argued that the Solonker–Xra Moron–Changchun suture (between the Songliao Block and the North China Craton) represents the final collision zone (Li and Wang, 1983; Wang and Liu, 1986, 1991; Huang, 1983, 1991).

More recently, Nozaka and Liu (2002) have proposed that suturing along the Hegenshan–Heihe belt did not occur until the Mesozoic. Their hornblende K–Ar data suggest that the Hegenshan ophiolite was emplaced at about 130–110 Ma (and not at 403 ± 27 Ma as previously

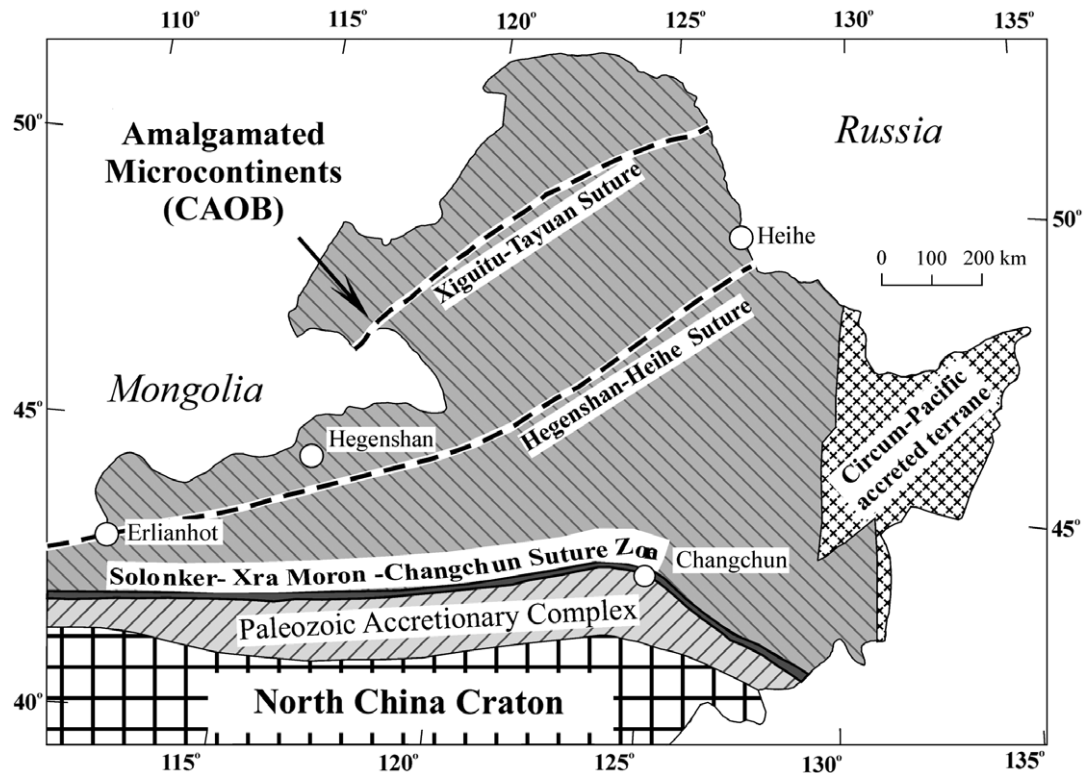


Fig. 9. Tectonic sketch map showing collision between the North China Craton and combined microcontinents of the Central Asian Orogenic Belt (CAOB).

considered by Bao et al., 1994), coeval with the eruption of Mesozoic volcanic rocks in the Great Xing'an Range. However, this interpretation needs to be further investigated owing to the possible effects of later disturbance (Zhou et al., 2003). In fact, a large number of Mesozoic ages have been obtained from Palaeozoic volcanic and granitic rocks in this region, and are generally interpreted as reflecting younger overprinting due to Late Mesozoic thermal events (IMBGM, 1990; HBGMR, 1993). The existence of 290–260 Ma, post-orogenic A-type granites north of the Hegenshan ophiolites further suggests that suturing here occurred before the Permian (Hsu et al., 1991; Muller et al., 1991; Cao, 1993; Hong et al., 1995; Robinson et al., 1999; Wu et al., 2002; Zhou et al., 2003).

Based on the available palaeontological evidence, the presence of island-arc volcanic rocks (Wang and Liu, 1986), and our new isotopic and metamorphic data from the Hulan area, we favour the interpretation that the Solonker-Xra Moron-Changchun suture marks the final closure of the ocean basin between North China and the Mongolian terranes (Fig. 9). Although the ophiolite near Xra Moron in the Solonker-Xra Moron-Changchun zone yields a Sm-Nd isotopic age of 665 ± 46 Ma (Chen et al., 1991), older than the Sm-Nd isochron age of 403 ± 27 Ma obtained for the Hegenshan ophiolite (Bao et al., 1994), this age most probably represents the rock-forming age of the oceanic crust. Alternatively, the above ages are

suspect due to their large errors. Wang and Fan (1997) found Permian Radiolarians in cherts along the Solonker-Xra Moron-Changchun suture zone, much younger than the Middle-Late Devonian or Early Palaeozoic Radiolarians discovered in ophiolites along the Hegenshan-Heihe suture zone (He and Shao, 1983; Robinson et al., 1999). This suggests that suturing along the Solonker-Xra Moron-Changchun zone probably occurred in the Late Permian, coeval with the collisional event identified in central Jilin Province. This is consistent with the recent interpretation of Xiao et al. (2003), who consider that the Solonker segment of the suture represents the southern termination of the CAOB. Therefore, the Hulan Group of central Jilin Province most likely represents the eastern extension of the Solonker-Xra Moron-Changchun suture zone and defines the major collision along which the ocean finally closed at ~ 250 Ma and the North China Craton was accreted to the Mongolian terrane. Considering recent palaeogeographic reconstructions (Heubeck, 2001; Zhou et al., 2001), the Tianshan Belt (former site of the Turkestan Ocean) in northwestern China may represent the western extension of the Solonker-Xra Moron-Changchun suture zone (Fig. 9). However, the final closure along the Tianshan Belt possibly occurred at some time earlier than that in NE China (Heubeck, 2001; Zhou et al., 2001), suggesting that closure was diachronous and moved progressively from west to east.

6. Conclusions

A comprehensive study on the Hulan Group in central Jilin Province of northeastern China leads to the following conclusions:

- (1) The Hulan Group consists of felsic paragneiss, pelitic schist and marble, with some undeformed limestone. Young Nd model ages for the schist indicate that the rocks are not a rifted fragment from the Archean North China Craton in the south nor from the Proterozoic Jiamusi Block in the north;
- (2) U–Pb isotopic analysis of detrital zircons from a kyanite-garnet schist indicate a maximum deposition age of 287 ± 6 Ma, which is inconsistent with the Early Paleozoic fossils contained in the weakly deformed limestone of the Beichatun Formation, suggesting that the Hulan Group is not a stratigraphic succession, but rather is a composite tectonic domain;
- (3) Parts of the Hulan Group have undergone sub-greenschist to amphibolite facies metamorphism and multiple phases of deformation. Rb–Sr mineral isochron ages indicate that the metamorphism occurred at ~ 250 Ma, which is much younger than tectonothermal events recorded in other zones of the Central Asian Orogenic Belt. It is considered that this time therefore represents the final closure of the ocean between the North China and Siberian cratons, leading to their amalgamation.

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