The Sugetbrak basalts from northwestern Tarim Block of northwest China: Geochronology, geochemistry and implications for Rodinia breakup and ice age in the Late Neoproterozoic

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Abstract

Zircons from two samples of the Sugetbrak basalts (SB) yieldweighted mean ages of 615.2±4.8Ma and 614.4±9.1Ma. These ages,interpreted as the eruption age of the SB, providean age constraint onthe timing of theSugetbrak Formation in Sugetbrak, northwestern Tarim Block, China. These two ages suggest that the igneous activities related to the breakup of the Neoproterozoic supercontinent Rodinia lasteduntil614-615 Ma in at least the northwest Tarim Block. Geochemicalanalysis indicatesthat the SB wasgenerated in an intra-continental rifting environment.Application of the dynamic melting inversion method suggests that the Sugetbrak basalts represent 7-12% partial melts. Unlike large-volume tholeiites, these low-degree transitional basaltic melts may represent the waning stage of plume volcanism during a long-lasting continental breakup. Based on the ages of the SB and its stratigraphic relationship with the Yuermeinak diamictite, the Yuermeinakglaciation in the Sugetbrak section of the northwest Tarim Block should correlate with theTereekenglaciation in the Qurugtagh area of the northeast Tarim Block, the Nantuo glaciation in Yangtze Block, and the Elatinaglaciation in South Australia.

Keywords: Late Neoproterozoic, Tarim Block, Rodinia, SHRIMP U-Pb age, Glaciation

1. Introduction

Magmatic rocks associated with the Rodiniabreakup may record critical information onthe timing and processes of continental breakup, as they can be precisely dated, and provide cluesregardingcrustal thickness and mantle sources (Kullerud et al., 2006). TheNeoproterozoic breakupof the Rodinia supercontinent has attracted much attention recently

(Jefferson, et al., 1989; Li, et al., 1996; Veevers, et al., 1997; Wingate, et al., 1998; Li, 1999, Karlstrom, et al., 2000; Veevers, 2000； Li and Powell, 2001;Direen, et al., 2003; Li, et al., 2003; 2006; Kheraskova, et al., 2010). This event,developed in Australia, East Antarctica, Yangtze,andthe Tarim Block,is divided into four stagesin the Tarim Block: 820-800Ma, 780-760Ma, 740-735Ma and 650-635Ma (Li, et al., 1996; 2001; 2003; 2008; Li, et al., 2010; Zhang, et al., 2011; Zhang, et al., 2012). Theages of these magmatic events have been employed to constrain the division of the Neoproterozoicice ages (Ross and Villeneuve, 1997; Hoffman and Schrag, 2002; Hoffmann, et al., 2004; 2006; Zhou, et al., 2004; Condon, et al., 2005; Zhang, et al., 2005; Xu et al., 2009; Macdonald, et al., 2010).

The TarimBlock has Neoproterozoic magmatic records similar to those of the YangtzeBlock, and may have been located on the periphery of the proposed Rodiniansuperplume (Li Z.X. et al., 1999; Chen et al., 2004; Xu et al., 2005, 2009; Zhang et al., 2009; Lu et al., 2008;Shu et al., 2011).Four stages of Neoproterozoic magmatic events have been reported in the Tarim Block.The first two (820-800 Ma and 780-760 Ma) stages, represented by ultramafic-mafic complexes, adakitic granites, and mafic dykes, are interpreted to be a result of the partial melting of thickened lower crust triggered by underplating of a Rodinian superplume (Long, et al., 2011; Zhang, et al., 2009; 2011a; 2011b). The third (740-735 Ma) stage,consisting of bimodal volcanic rocks andthe fourth (650-635 Ma) stage,composed ofmafic dykes, potassic granitoids, peraluminous granite and volcanic rocks, are attributed to rifting in the Tarim Block(Xu, et al., 2005; 2009; Zhang et al., 2011; Zhu, et al., 2011). However, the presence of 615±15 Ma volcanic rocks in the Mochia-Khutuk areaof thenortheast TarimBlock (Xu, et al., 2009), and 588-619 Ma detrital zirconsin the Aksu area of the northwest TarimBlock (Zhu, et al.,

2011), may indicate that there wasa magmatic event evenyounger than thefourth event.This is useful forinvestigatingthe waning stage of the Rodiniasuperplume breakup in the Tarim Block.

In this paper we report newages of basaltsfrom the NeoproterozoicSugetbrak Formation in the Sugetbraksection of the Aksu area, NW TarimBlock, and discuss their tectonic implications for the glaciation correlation and intra-continental riftinginthe Late Neoproterozoic.

2. Geological setting

2.1.Neoproterozoic strata in the Aksu area of the northwest Tarim Block

Neoproterozoic strata occur in the Aksu area of northwest Tarim, the Quruqtagh area of northeast Tarim, and the Yecheng area of southwest Tarim.Strata containing volcanicshave been found from the Aksu area and the Quruqtagh area (Gao and Zhu, 1984; Xu et al., 2005;Wang et al., 2010a, Fig. 1B).

In the Aksu area, the Neoproterozoic strata crops out in southwest Aksu,Wushi and Sugetbrak (Fig. 1C).In the sections of southwest Aksu and Wushi, the stratainclude the Aksu Group and theoverlying Sugetbrak and Chigebrak Formations. The Aksu Group is composed of pelitic, psammitic, and mafic schists.Mafic schistsarecharacterized bygreenschists and blueschists that areconsidered to besomeof the oldest high-pressure metamorphic rocks (Liou et al., 1989, 1990; 1996; Nakajima et al., 1990; Zhang, et al., 1999).A tectonic evolution modelfor the Aksublueschist has been proposed(Zhang, et al., 2010; Zhu et al., 2011). There is a clear angular unconformity between the Aksu Group and the overlying Sugetbrak Formation in southwestAksu and Wushi (Gao, et al., 1985; Turner, 2010; Zhu, et al., 2011). The 400-450 m thick Sugetbrak

Formation is composed of red conglomerates, red fluvial sandstones and grey lacustrine mudstones with three horizons of basalts.The Chigebrak Formation is characterized by thick dolostoneand is overlainunconformably by the Cambrian Yuertus Formation (Gao et al., 1986; Turner, 2010; Zhu et al., 2011).

In the Sugetbraksection,the Neoproterozoic shows a differentsuccessioncomposed of,from bottomto top, the Qiaoenbrak, Yuermeinak,Sugetbrak and Chigebrak Formations (Fig. 2A, 2B). The Qiaoenbrak Formationis composed of epimetamorphicsandstone and siltstoneflysch,with a thickness of 1966-2094 m (Gao, et al., 1986). The glacial origin of this formation is debated. For example, Gao et al. (1993) suggestedthat it is part of a turbidite sequence. Itisunconformablyoverlain by theYuermeinak Formation and is consideredto be marineglacier deposits due to the presenceof several interbeddeddiamictites (Gao, et al., 1986). The YuermeinakFormation occurs only in the Sugetbrak area, and is characterized bydiamictites up to 61 m thick,which areinterpreted to becontinental glacier deposits (Gao, et al., 1986).The Sugetbrak Formation can be divided into the Lower and Upper members in Sugetbrak(Fig. 2B, 2C;Gao, et al., 1986; Zhan, et al., 2007). The Lower Sugetbrak Formation(108-461 m)is composed ofthin-layered red sandstone, quartz sandstone, siltstone and mudstone with a discontinuous horizon of 5-10 m thick basalt in its lower part, followed by 79 m thick volcanic rock above (Fig. 2C). The Upper Sugetbrak Formationconformably overlies the Lower member and is conformably overlain by the Chigebrak Formation. The Upper Sugetbrak Formation comprises 82 m thick red and grey thin-layered sandstone and mudstones with interbedded limestones (Fig. 2B, Gao, et al., 1986).

Fig. 1.(A):Tectonic location of Tarim Block in China;(B):Geological map of Tarim Block, West

China; (C): Geological map of the Aksu area

Fig. 2. (A): Geological map of the Sugetbrak sectionand sample positions; (B): theNeoproterozoic stratigraphic column inSugetbrak; (C): photos of the Neoproterozoic strata in Sugetbrak

2.2. Sugetbrakbasalt (SB) and samples

The Sugetbrak basalt (SB) occurs insouthwest Aksu, Wushi andSugetbrakin the Aksu area (Fig. 1C) of thenorthwest Tarim block.There are between 2 and 4horizons of thisbasalt inthe southwest Askusection, with thicknesses of 15-30 m (Wang, et al., 2010a; Zhu, et al., 2011).Two 4-5 m thick horizons of the SB are also reported in the Wushi section45 km to the northwest (Turner, 2010).In the Sugetbraksection,the SB is much thicker and has repeated occurrencescaused byanE-W extending syncline. On the north limb of the syncline,a79 m thick SB occurencecrops outat the top of the Lower member of the Sugetbrak Fm. (Fig. 2B, 2C) and shows conformity relationships with the underlying sandstone (Fig. 3A), where 7 samples of A group(05822, Y1, Y2, Y5, Y6, Y7 and Y8; Fig. 2A)werecollected. Adiscontinuous horizon of 5-10 m thick columnar-jointed basalt (Fig. 3B) occurs in the middle of the Lower member of the Sugetbrak Fm.(Fig.2B, 2C), showing a conformity relationship with underlying sandstones.On the south limb of the syncline, a continuous horizonof the SB,characterized by pillow lava, (Fig. 3D) extends for several kilometers(Fig. 3C),andshows aconformity relationship with overlying red sandstones (Fig. 3E). Here, 9samples of B group(05823, 830S2, 831S51, 83156, 83158, 83160, 83161, 83163 and 831S81)werecollected(Fig.2A).These columnar jointed basaltsand pillow lava in the SB,combinedwith sedimentary structures, such as large herringbone (Fig. 3A) and ripplebedding (Fig. 3F) in the redsandstones, suggest a very shallow continental sedimentary environment during the period of SB magmatism.About 4 km southwest ofSugetbrak, two basaltic

flows have been reported (Zhang, et al., 2012).

Fig.3. Photographs of the SB and sedimentary rocks of the Sugetbrak Formation.

2.3. Previous geochronology for the SB

Seventeen zircons fromthe SB fromsouthwest Aksu have been dated by the LA-ICP-MS method and yieldthree ages of 1926±27 Ma, 1432±22 Ma and 807±21 Ma. These ages have been interpreted as the ages of Mesoproterozoic metamorphic zircons and Late Neoproterozoicinherited magmatic zircons (Wang et al., 2010a). The youngest age in these inherited zircons is 755 Ma, suggesting that the SB is younger than 755 Ma (Wang, et al., 2010a). Recently, Zhang et al.(2012) reported a LA-ICP-MS U-Pb age of 783.7±2.3 Ma from the SB, about 4km southwest ofSugetbrak, and interpretedthis ageas thecrystallization age of the SB.

3. Analytical procedures

Twosamples from the SB, 05822 and 05823, wereanalyzedby the SHRIMP U-Pb method for age dating. In addition, these twosamples and the 14 other SBsampleswereanalyzedto determinetheir major and trace elements compositions.

Zircons for SHRIMP analyses were separated from samples 05822 and 05823 in the SB according to magnetic properties and density, and purified by hand picking. The zircons, together with several grains of TEMORA, were cast in an epoxy mount and polished down to half section. Cathodoluminescence (CL) imaging was used to guide the SHRIMP analyses. The CL study was undertaken on a FEI-XL30SFEG electron microscope at the Department of Electronics, Peking University.

The SHRIMP U-Pb analyses wereperformed on the Beijing SHRIMP IIat the ChineseAcademy of Geological Sciences, Ministry of Land and Resource of Peoples' Republic of China. The analytical procedures followed the methodology of Williams et al (1987) and Compston et al (1992). For the zircon analyses, nine ion species of Zr_2O^+, $^{204}Pb^+$, $^{207}Pb^+$, $^{206}Pb^+$, $^{208}Pb^+$, U^+, Th^+, ThO^+, UO^+ and background weremeasured on a single electron multiplier by cyclicstepping of the magneticfield, recording the mean ion counts of every 7 scans. A primary ion beam of c. 4.5 nA, 10 kV O^{-2} and c. 20-25μm spot diameterwas used. Masses were analyzedat a mass resolution of c. 5000 (1% peakheight). Interelementfractionation in the ion emission of zircon wascorrected relative to the ANU RSES references, using the TEMORA (417 Ma, $^{206}Pb^*/^{238}U$=0.06683). Errors on individualanalysis are at the1σ levelbased on countingstatistics. The reproducibility of TEMORAwasrepeatedlymeasuredat c. 3%. The software of Ludwig SQUID1.0 and attached ISOPLOT wereused for data processing (Ludwig, 1999 ; 2001). The ages werecalculatedusing the decayconstants recommended by IUGS (1977). The weightedmeanages werequotedat a 95% confidence level. The initial lead component was corrected using measured ^{204}Pb.

Major element oxides were determined by x-ray fluorescence using glass disks at the Laboratory of Orogenic Belts and Crustal Evolution, Ministry of Education, Peking University. Analytical precision as determined on duplicate analyses was generally around 1–5%. For trace element analyses, about 50 mg sample powders were dissolved using a $HF+HNO_3$ mixture in a Teflon bomb at ~190 ℃ for 48 h. Trace elements were analyzed using a Finnigan Element IIICP-MS at the Center of Modern Analysis, Nanjing University. Analytical precision for trace elements was better than 5%. Detailed analytical procedure followedGao et al. (2003).

4. U-Pb zircon geochronology

The measured isotopic ratios and calculated ages for samples 05822 and 05823 are given in Table 1 and illustrated on Concordia plots in Fig. 4B. The measured U concentrations for these zircon grains vary from 960 to 58 ppm; Th, from 949 to 63 ppm; and Th/U ratios, from 1.59 to 0.26, which suggests that all zircons belong to a magmatic type.

Twenty-nine zircons were dated in total, and can be divided into threeage groups (Table 1).Zircons ofgroup1are~ 150 μm in size,show weakly zoned textures (Fig. 4A, 05823-11-4.1),and yield a $^{207}Pb/^{206}Pb$ age of 1756±9 Ma (Table 1, 05823-11-4.1), indicatingthe Mesoproterozoicmagmatism (Hu, et al., 2000). Most zircons of groups2 and 3 are smaller, ranging from 80to30 μm.Theyare euhedral with oscillatoryzoning,indicative of magmatic zircons (Fig. 4A).$^{206}Pb/^{238}U$ ages of twenty-fourzircons ofgroup 2 range from 676 to 864 Ma, and form several intervals of 676-691Ma (n=3), 707-743Ma (n=5), 751-790Ma (n=7) and 806-864Ma (n=9, Table 1). These ages represent inherited zircon ages andreflect multiple magmatic episodesrelated toRodinia breakup inthe Late Neoproterozoic(Zhang, et al., 2011b; Zhu et al., 2011b; Wang, et al., 2010a).Interestingly, there are four zircons in group 3 that yield $^{206}Pb/^{238}U$ ages from 617 to 612 Ma (05822-1-7.1, 05822-1A-5.1, 05823-11-2.1 and 05823-11-5.1), which give twoconcordia ages of 615.2±4.8 Ma (Fig. 4B) for sample 05822 and 614.4±9.1 Ma (Fig. 4C) for sample 05823.Because these 4 youngest zircons areeuhedral and thesmallestin size, their concordia ages of 614-615Maare interpreted as the eruption age of the SB, although we cannot completely rule out the possibility that the ~615 Ma zircons may represent the youngest inherited zircons. In the case of youngest zircon inheritance, the eruption age of the Sugetbrak basalts would be slightly

younger than ~615 Ma.

Table 1. Zircon U-Pb isotopic data

Fig. 4. Cathodoluminescence images and Zircon concordia plots

5. Geochemistry

Major and trace elements of the SB are listed in Table 2.They are characterized by high TiO_2 contents (2.2-4.0%). Their Al_2O_3 ranges from 13.9% to 17.7%. Total alkali content ranges from 2.9% to 5.4%, with $Na_2O>K_2O$. Theirhigh LOI contents, ranging from 2.7% to 8.7%, indicate alterations. Na_2O and K_2O were probably mobile during alterations, and thus cannot be employed for rock classifications using the total alkali vs. silica diagram.

Table 2. Major and trace elements of the SB

Petrologically, it is useful to determine whether the rocks are alkali basalts or tholeiites. Here we use immobiletrace elements such as Nb and Y to further evaluate their alkali character. These samples have Nb/Y ratios ranging mostly from 0.58 to 0.68. In the Zr/TiO_2 – Nb/Y diagram (Winchester and Floyd, 1977), they plot on the boundary line between sub-alkaline basalts and alkali basalts (Fig. 7A). In the Nb/Y –$Zr/(P_2O_5*10000)$ diagram, they also plot near the boundary line between alkali basalts and tholeiitic basalts (Fig. 7B). We thus regard them as transitional basalts.

Fig. 5. A: Zr/TiO_2 – Nb/Y diagram (Winchester and Floyd, 1977)

B: Nb/Y – $Zr/(P_2O_5*10000)$ diagram (Floyd and Winchester, 1975)

All samples are enriched in light rare earth elements (LREEs, Fig. 6A). They lack negative Eu anomalies, indicating that plagioclase is not a fractionatingphase.In the Spider diagram (Fig.

6B), they display positive anomalies in Ba and Pb and negative anomalies in Rb and Sr. They do not shownegative Nb anomalies.

Fig. 6.A: REE diagram; B: Spider diagram

6. Discussion

6.1 Petrogenesis and tectonic settings

6.1.1Within-plate environments

Ti, Zr, Y and Nb are immobile trace elements during zeolite to greenschistfacies metamorphism, and may be useful for the determination of tectonic environments of basaltic rocks (Pearce and Cann, 1973; Winchester and Floyd, 1975). In the Ti-Zr-Y diagram (Pearce and Cann, 1973), the lavas plot in the “within-plate” basalt field (Fig. 7A). In the Nb-Zr-Y diagram (Meschede, 1986), the samples plot on the boundary line between fields AII and C, indicating an intraplate tectonic setting (Fig. 7B). Theboundary line is the transition between intraplate alkali basalts and intraplatetholeiites.Both Ti-Zr-Y and Nb-Zr-Y diagrams suggest a within-plate environment, ruling out the possibility of a volcanic arc or ocean-ridge environment.The lack of any negative Nb anomaly in the spider diagram also argues against the island arc or continental arc environment.Therefore, the SBare intra-continental transitional basalts.Our inference of a within-plate (intra-continental) tectonic setting for the SB is largely consistent with the work by Zhang et al. (2012).

Fig. 7. A: Ti-Zr-Y diagram (Pearce and Cann, 1973);

B: Nb-Zr-Y diagram (Meschede, 1986)

6.1.2. Low-degree small-volume melts

The high Ti contents in all samples reflect low degree partial melts. We have used the dynamic melting inversion (DMI) method (Zou and Zindler, 1996; Zou, 1998; Zou et al., 2000) to estimate the degrees of partial melting more quantitatively. The DMI method uses variations of between-magma concentration ratios for two incompatible elements with different partition coefficients, and does not require assumptions about mantle rock concentrations or ratios. Two immobile trace elements are selected with different partition coefficients, Nb and Zr. Note that Nbis more incompatible than Zr. The concentration ratios between a low-degree melt sample (sample 831S41) and a high-degree melt sample (sample 83161) are 1.51 (=24.63/16.27) for Nb and 1.31 (=200/152) for Zr. Samples 831S41 and 83161 are selected because sample 831S41 has the highest incompatible element concentrations (representing the lowest degree of partial melting) while sample 83161 has the lowest incompatible element concentrations (indicative of the highest degree of melting) in B Group. The degree of partial melting calculated using DMI by solving a system of simultaneous non-linear equations is 7.4% for sample 831S41,and 10.9% for sample 83161, using bulk partition coefficients of 0.01 for Nb and 0.04 for Zr. If we still select sample 831S41 as a low-degree melt, but select another sample such as 83160 as a high-degree melt (another sample with low incompatible element concentrations in B Group), then the low-degree/high-degree concentration ratios are 1.52 (=24.63/16.27) for Nband 1.33 (=200/150) forZr, and the partial melting degrees using DMI are 7.9% for sample 831S41 and 11.8% for sample 83160. Thus, these transitional basalts were produced at 7-12%mantle partial melting.

Note that in the calculations we use concentration ratios instead of elemental abundances, because concentration ratios are insensitive to subsequent fractional crystallizations.

Unlike large-volume tholeiitesproduced in the main stage of plume volcanism, these low-degree, small-volume melts might represent melts formed at the onset stage or the waning stage of a plume during continental breakup. We prefer to considerwaning stage of plume volcanism for the creation of the SB because of its younger age, while Zhang et al. (2012) prefer the onset stage of plume volcanism, owing to the different age results for the SB (784 Ma in Zhang et al. (2012) vs. 614-615 Ma from this study). As mentioned above, we regard the 784 Ma age as the age for inherited zircons (see discussion in section 6.2).

6.2. Implications for the rift succession of the Sugetbrak Formation and Neoproterozoic igneous activities

Because the studied samples were collected in the horizons of the SB between the Lower and Upper Sugetbrak members, the new age representsthe boundary age between the upper and lower members, further providing a time constraint on the timing oftheSugetbrak Formation in the Sugetbrak section. Recently, the Sugetbrak Formation in the southwestAksu and Wushi areas has been interpreted as fluvial and lacustrine faciessediments (Turner, 2010; Wang et al., 2010a). Combinedwith several horizons of basalts that record episodic volcanism, these sedimentary faciesarethoughtto indicate the development of a Neoproterozoic rift system in the Aksu area (Turner, 2010; Wang et al., 2010a).Hence, our new age represents the development age of this rift system in the northwest Tarim Block.

Though four phases of Neoproterozoic igneous activities (820-800Ma, 780-760Ma,

740-735Ma and 650-635Ma) have been recognized in the TarimBlock(Zhang et al., 2011), ournew age(615 Ma) is even younger,and may represent thelast stage of the Neoproterozoic rifting relating to thebreakup of the Rodinia Supercontinent in the northwest Tarim Block. It has been noticed that there aresimultaneousvolcanic rocks in the Qurugtagh area of thenortheast Tarim Block, where an age of 615±6 Ma from andesitic lava has been reported (Xu et al., 2009). These data suggest that the igneous activities related to the breakup of the Neoproterozoic Rodinia supercontinent lasteduntil at least 615 Ma.

Six detrital zircon ages from the red sandstone of the Sugetbrak Formationin southwestAksurange from 619 to 588 Ma and give a weighted average age of 602±13 Ma (Zhu, et al., 2011), which isyounger thanthe ageof the SB from Sugetbrak. The age difference betweentheseplaces might be due to the fact that the Neoproterozoic succession in southwest Aksuis incomplete, which has been indicated by the lacunas of the Qiaoenbrak and Yuermeinak Formations, and of thebottom part ofthe Lower Sugetbrak Formation (Fig. 8). Turner (2010) pointedout that small, isolated depocentres and pre-existing topography might causelocal variety of basal conglomerates in the Aksu area. Therefore,the age of 602±13 Ma might represent the maximal depositional age of the Upper member of theSugetbrak Formation (Fig. 9). Obviously, the direct evidence for the age of the Sugetbrak Formation depends on the eruption age of the SB in the southwest Aksu section. According to available geochronological data in the southwest Aksu and Sugetbraksections, a revised correlation of the Neoproterozoic strata between southwest Aksu, Wushi and Sugetbraksections is proposed in Fig. 8.

Recently, Zhang et al (2012) reported a LA-ICP-MS U-Pb zircon age of 783.7±2.3 Ma from the SB about 4 km southwest ofSugetbrak, and interpreted it as a crystallization age of the SB. We

regard 784 Ma as the age for inherited zircons instead of the crystallization age for the SB, because the zircons from Zhang et al. (2012) are large crystals (most zircons > 80 μm), similar to group 2 zircons of this study and the inherited zircons of Wang et al. (2010a). In addition, Zhu et al. (2011) have reported several detrital zircon ages ranging from 619 to 588 Ma from the red sandstone of the Upper member of the Sugetbrak Formation, which implies that it is impossible for the Lower member of this formation to be 784 Ma old. Geologically, if this age of 784 Ma is correct, then the underlying Yuermeinak glacier deposits would be older than 784 Ma. However, no Neoproterozoic ice age older than 784 Ma has been reported in the world (Hoffman and Schrag, 2002; Xu, et al., 2009; Macdonald, et al., 2010).

Fig. 8. Correlation of the Neoproterozoic strata between southwest Aksu, Wushi and Sugetbrak areas

6.3. Implications for thecorrelation of the Yuermeinak glaciation

Ages, distribution, and correlationsof the Neoproterozoic glaciations around the worldare still controversial(Kaufman et al., 1997; Hoffman et al., 1998; Kennedy et al.,1998; Hoffman and Schrag, 2002; Jiang et al., 2003). Two glaciations (Sturtian and Marinoan, Young, 1995; Kennedy et al., 1998), three glaciations (Sturtian, Marinoan and post-Marinoan, Knoll, 2000, Xiao et al., 2004; Halverson, 2006; McCay et al., 2006)orfourglaciations (Kaigas, Sturtian,Marinoan or Elatina and Gaskiers, Hoffman and Schrag, 2002; Hoffmann et al., 2004;Macdonald et al., 2010)have been proposed.On the other hand, four glaciations have been reported in the northeast Tarim Block, including Bayisi (730 Ma), Altungol, Tereeken (725-615 Ma) and Hankalchough

(615-542 Ma) glaciations (Kou et al., 2008; Xu et al., 2008; Xu et al., 2009). However, the age of the Yuermeinakglaciation in the northwestTarim Block,and its correlation,werepreviously unknown.

According to the Neoproterozoic succession in theSugetbraksection(Fig. 9), the Lower member of Sugetbrak Formation containing the SBconformably overlies the Yuermeinak Formationdiamictite.Because there is only about400 m of continental sedimentary succession between the SB and Yuermeinak Formation, the occurrence of the Yuermeinak glaciation is not much earlier than 614-615 Ma. This is similartothe Tereeken glaciation in the Quruqtagh area of the northeast Tarim Block, where there is only the ZhamoketiFormation turbidite between the Tereekendiamictite and volcanic layerwith an age of 615±6 Ma (Xu et al., 2009).It is thereforereasonableto correlatethe Yuermeinake glaciationwith the Tereeken glaciation in the Quruqtagh area.Based on aTIMS U-Pb age of 635.2±0.6 Ma for the Nantuo glaciation (Condon et al., 2005) and an age of 615±6 Ma for the volcanic rocks above the Tereeke glaciation,previous studies havesuggested that the Tereekenglaciation can be correlated with the Nantuoglaciation in the Yangtze Blockand the Elatina glaciation in Australia (Xu et al., 2009). If this is correct,then the Yuermeinak glaciation should also correlate with them (Fig. 10).

Fig. 10Correlation of the Neoproterozoicglaciations

7. 8. Conclusion

This study reached the following conclusions:

(1)Small euhedralzircons from two samples of the Sugetbrak basalts (SB) of the Sugetbrak Formationyieldweighted mean ages of 615.2±4.8Ma and 614.4±9.1Ma.These ages are interpreted

as the eruption age of the SB, thereby providing an age constraint on the timing oftheSugetbrak Formation in Sugetbrak, northwestTarim Block. These two ages suggest that the igneous activities related to the breakup of the Neoproterozoic Rodinia supercontinent lasteduntil at least614-615 Ma in the northwest Tarim Block.

(2)Geochemicalanalysis indicatesthat the SB was generated in an intra-continental rifting environment.Application of the dynamic melting inversionmethod suggests thatthe degree of partial melting of the SB ranges from 7% to 12%. Unlike large-volume tholeiites, these low-degree transitional basaltic melts may represent the waning stage of plume volcanism during a long-lasting continental breakup.

(3)Based on the new ages of the SB and its relationship with the Yuermeinak diamictite, the Yuermeinakglaciation in Sugetbrak of the northwest Tarim Block should correlate with theTereekenglaciation in the Qurugtagh area of the northeast Tarim Block, the Nantuo glaciation in theYangtze Block and the Elatinaglaciation in Australia.

Acknowledgements

We are grateful toanonymous reviewers for their constructive reviews that significantly improved the quality of this paper. We also thank Biao Song for his help in SHRIMP analyses of zircons, Chen Peng, Wei Wei, and Tong Qinlongfor field assistance, and Li Huaikun and Wang Yanyang for drawing figures. Ross T. Tucker and Erin Summerlin from Auburn University provided helpful in-house reviews. This study was supported by grants from the National Natural Science Foundation of China (40972126, 41121062).

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Figure Captions

Fig. 1. (A): Tectonic location of the Tarim Block in China; (B): Geological map of the Tarim Block, WestChina, showing the distribution of Neoproterozoic and the study area. Legend—PH: Phanerozoicrocks; NP: Neoproterozoic rocks; MP: Mesoproterozoic rocks; PP: Palaeoproterozoic rocks; AR:
Archaean rocks; NPG: early Neoproterozoic granitoids; F: faults; IF: inferred Faults; Tillite: Neoproterozoic tillite; Q: Quaternary desert and sedimentary deposits (modified after Lu, et al., 2008); (C): Geological map of the Aksu area (modified after Gao, et al., 1985; Turner, 2010).

Fig. 2. (A): Geological map of the Sugetbrak area and sample positions (modified after BGMRED ofXinjiang, 1993); (B): the Neoproterozoic stratigraphic column of the Sugetbrak area; (C): Photographof the Neoproterozoic succession in the Sugetbrak area.Notice the positions of SB sample 05822.

Fig.3. Photographs of the SB and sedimentary rocks of the Sugetbrak Formation. (A): largeherringbone and conformity relationships between the SB and underlying sandstone; (B):columnar joints of the SB; (C): Continuous distribution of the SB on the south limb of a syncline; (D): pillow lava in the SB; (E): conformity relationship between the SB and overlying red sandstones;(F): ripple bedding in red sandstone of the Lower member of the Sugetbrak Fm.

Fig. 4. A: Cathodoluminescence images of zircons from the SB.Notice the zircons can be dividedinto 3 groups according to their sizes; B: Concordia plot of zircons from sample 05822 of the SB;
C: Concordia plot of zircons from sample 05823 of the SB.

Fig.5. A: Zr/TiO_2 – Nb/Y diagram (Winchester and Floyd, 1977); B: Nb/Y – $Zr/(P_2O_5*10000)$diagram (Floyd and Winchester, 1975)

Fig. 6. A: REE diagram of the SB; B: Trace element spider diagram of the SB.

Fig. 7. A: Ti-Zr-Y diagram (Pearce and Cann, 1973); B: Nb-Zr-Y diagram (Meschede, 1986)

Fig. 8. Correlation of the Neoproterozoic between the southwest Aksu, Wushi and Sugetbrak sections in the northwest Tarim Block.

Fig. 9. Correlation of Neoproterozoic glaciations between the northeast Tarim, northwest Tarim and Yangtze Blocks. Data from: Yin et al., 2003; Hoffmann et al., 2004; Zhou et al., 2004; Chu et al., 2005; Condon et al., 2005; Zhang et al., 2005; Kendall et al., 2006; Xu et al., 2009.

Table

Table 1.SHRIMP zircon U-Pb data for the Sugetbrak basalts in Sugetbrak area, northwest Tarim Block

Spot	U(ppm)	Th(ppm)	$^{232}Th/^{238}U$	$^{206}Pb_c$	$^{206}Pb^*$	$^{206}Pb/^{238}U$ Age	$^{207}Pb/^{20}$
22-1-1.1	246	147	0.62	0.13	28.9	825.4 ± 8.9	849
22-1-2.1	287	179	0.64	0.30	33.0	808.0 ± 8.4	808
22-1-3.1	355	412	1.20	0.16	37.3	743.4 ± 7.6	737
22-1-4.1	451	209	0.48	0.14	49.6	774.6 ± 7.9	771
22-1-5.1	180	205	1.17	0.26	18.4	722.2 ± 8.1	671
22-1-6.1	159	146	0.95	0.27	17.5	776.1 ± 8.8	779
22-1-7.1	240	164	0.71	0.63	20.9	617.3 ± 6.8	641
22-1-8.1	58	63	1.13	0.85	6.10	744 ±11	797
22-1-9.1	960	536	0.58	2.44	104	750.8 ± 7.3	774
2-1-10.1	171	132	0.80	0.25	19.5	804.2 ± 8.8	825
2-1-11.1	517	425	0.85	0.75	64.0	861.9 ± 8.7	808
822-12.1	63	67	1.11	0.36	6.96	780 ±10	639
822-13.1	155	104	0.69	16.79	18.6	707 ±16	590
2-1A-1.1	72	83	1.19	0.30	7.57	743.3 ±7.5	833
2-1A-2.1	284	83	0.30	0.06	32.5	805.9 ±4.2	820
2-1A-3.1	71	82	1.19	0.67	7.78	773.0 ±8.0	703
2-1A-4.1	87	89	1.06	1.57	9.48	758.9 ±7.6	712
2-1A-5.1	442	111	0.26	0.13	38.0	614.9 ±2.6	671
2-1A-6.1	361	416	1.19	0.24	40.6	790.4 ±3.7	816
2-1A-7.1	832	305	0.38	0.51	103	863.6 ±3.2	841
3-11-1.1	233	151	0.67	0.48	27.3	821.2 ± 9.5	760
3-11-2.1	1102	816	0.77	0.69	95.7	616.6 ± 6.6	694
3-11-3.1	568	741	1.35	0.32	54.1	676.2 ± 7.5	657
3-11-4.1	415	200	0.50	0.12	110	1,735 ±17	1,755.6
3-11-5.1	623	949	1.58	0.22	53.4	612.3 ± 6.5	663
3-11-6.1	762	558	0.76	0.37	73.7	685.6 ± 7.3	689
3-11-7.1	140	216	1.59	0.92	16.9	839 ±11	698
3-11-8.1	588	261	0.46	0.20	57.3	691.4 ± 7.3	778
3-11-9.1	461	184	0.41	0.15	52.8	806.8 ± 8.8	801

Errors are 1-sigma; Pb_c and Pb^* indicate the common and radiogenic portions, respective
Error in Standard calibration was 0.20%(not included in above errors but required whe
(1) Common Pb corrected using measured ^{204}Pb. (2) Common Pb corrected by assuming 206Pb
(3) Common Pb corrected by assuming $^{206}Pb/^{238}U$-$^{208}Pb/^{232}Th$ age-concordance.

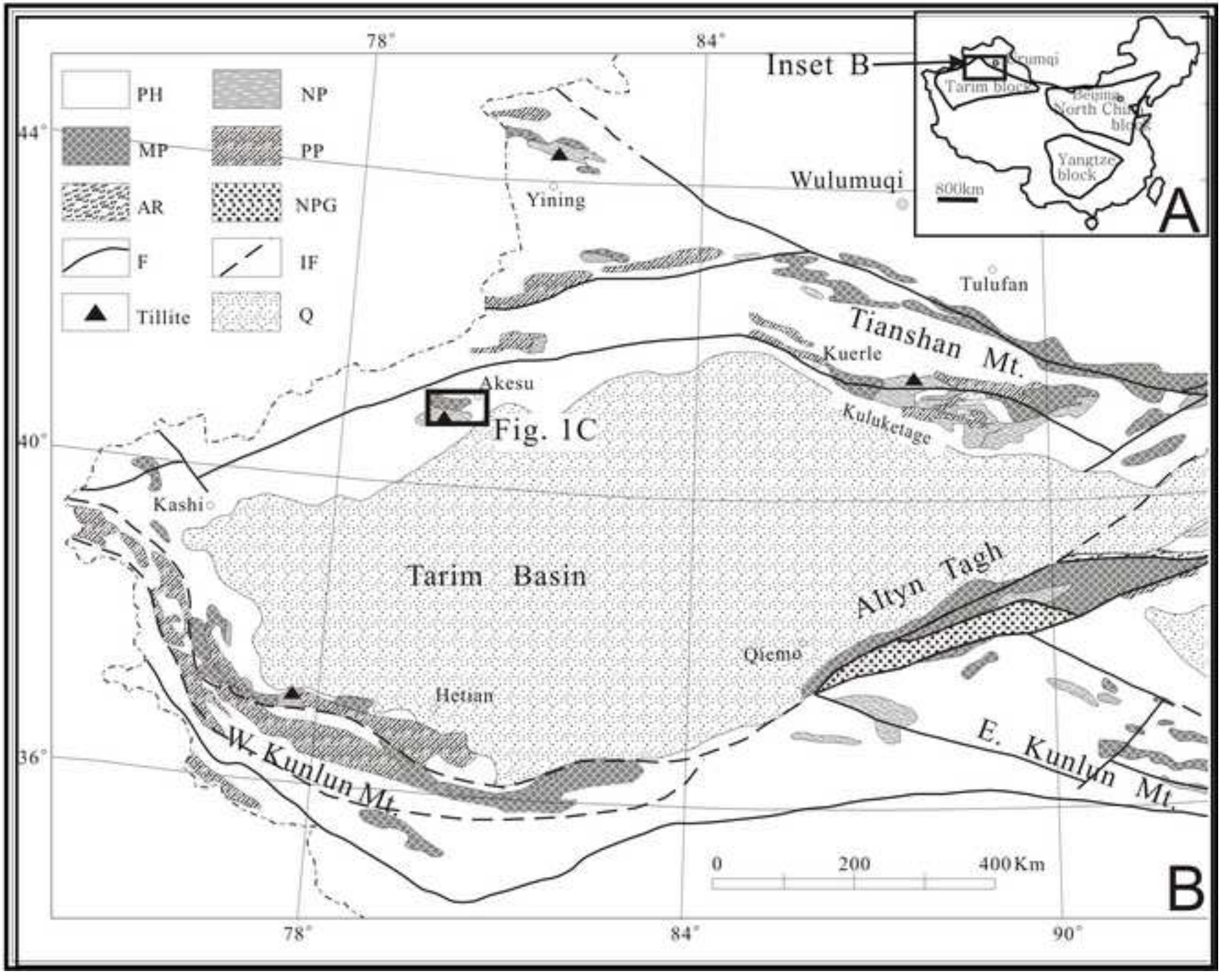
Table

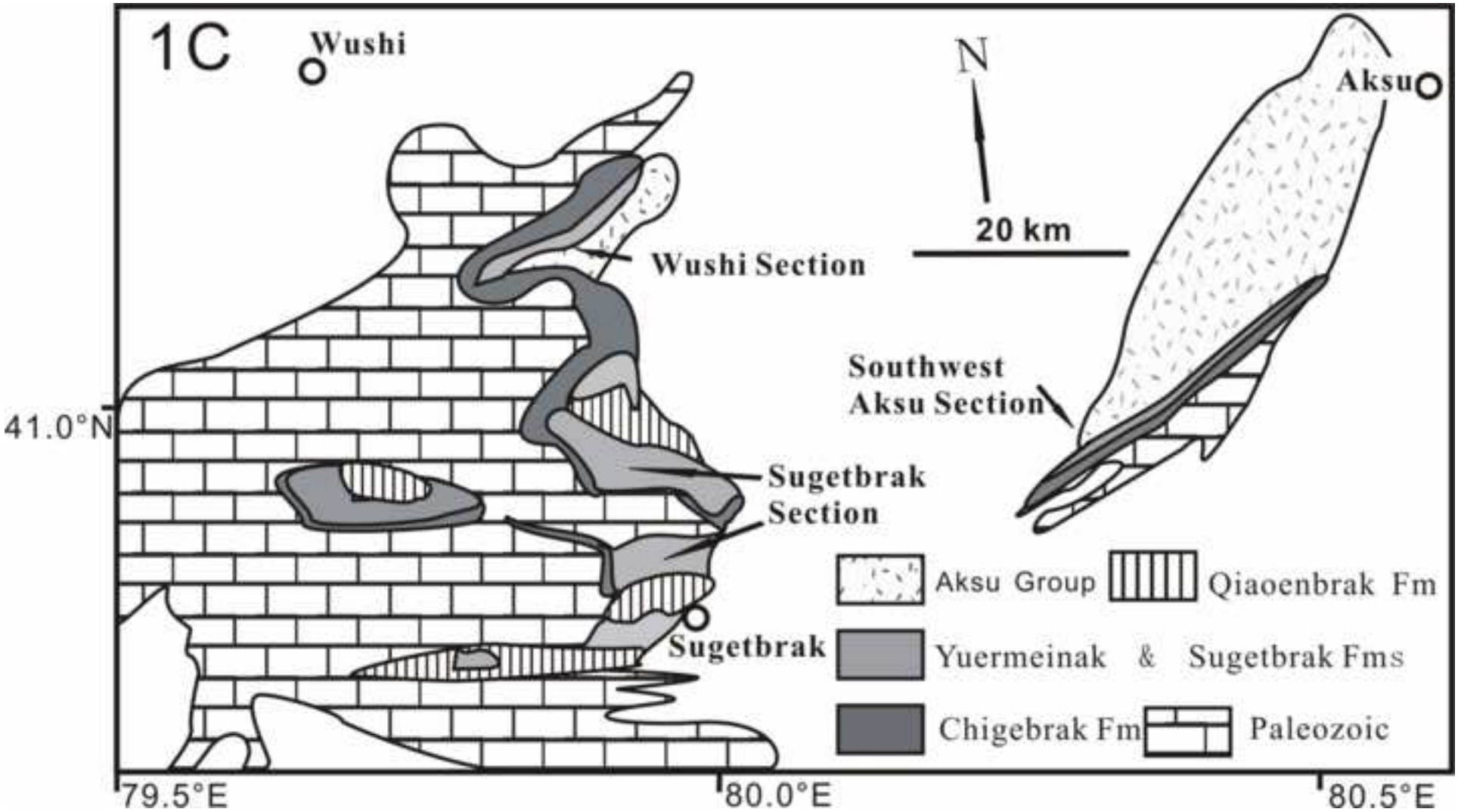
Table 2 Major and trace element abundances for the Sugetbrak basalts in Sugetbrak area, west Tarim Bl

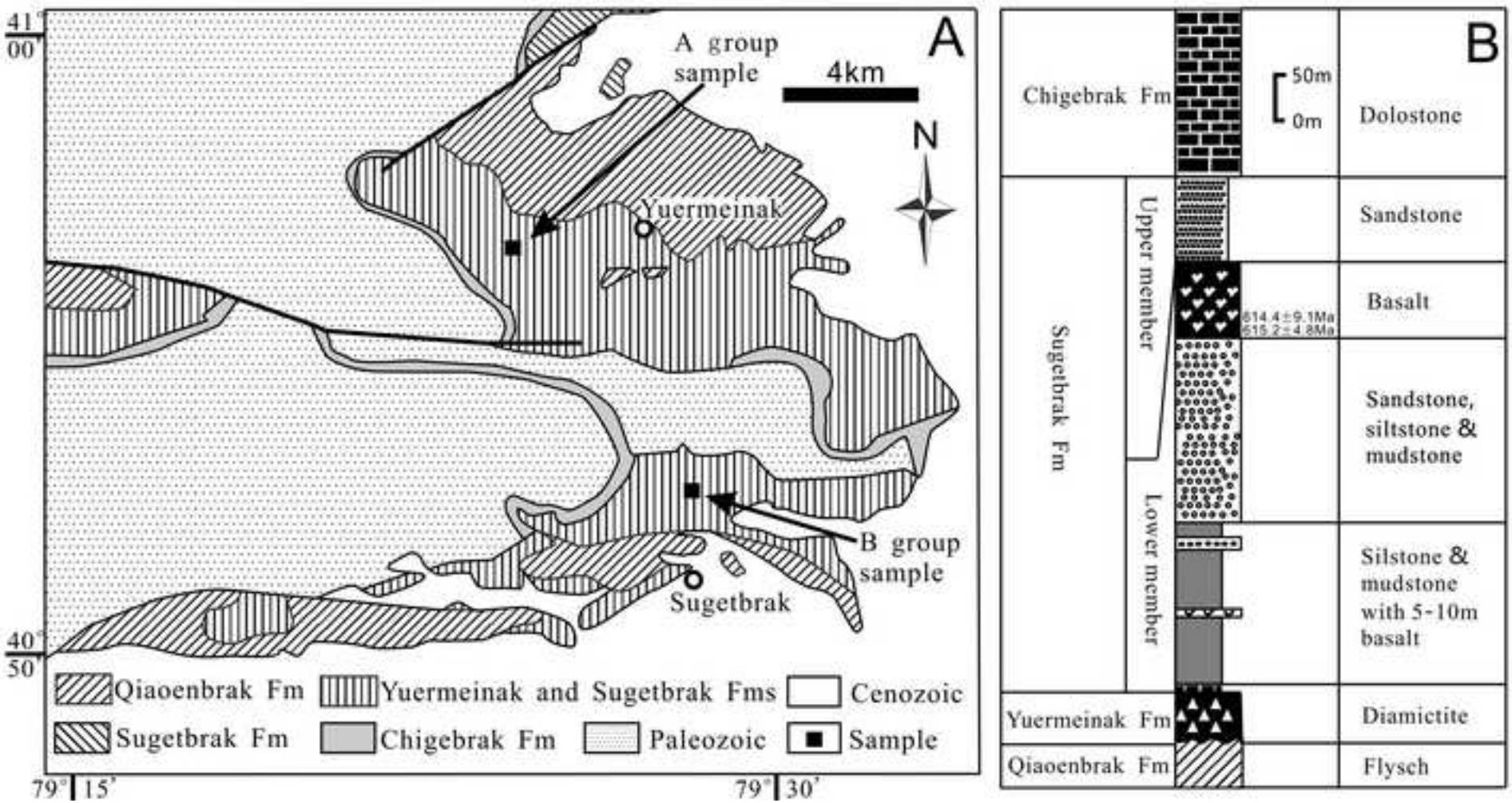
Sample	831S41	831S51	830S2	83156	83158	83160	83161	83163
SiO_2	44.35	43.45	45.40	44.38	45.47	45.44	46.81	44.37
TiO_2	4.002	3.494	3.763	3.124	2.865	2.215	2.216	2.489
Al_2O_3	13.91	14.91	13.92	14.9	15.32	16.95	16.73	17.73
MgO	4.52	5.11	5.3	5.88	5.33	5.29	5.19	6.44
Fe_2O_3	16.47	13.76	16.46	16.34	15.38	13.22	13.5	12.52
MnO	0.241	0.643	0.251	0.236	0.2	0.131	0.125	0.16
CaO	8.00	8.21	8.00	5.45	7.30	7.18	5.98	4.84
K_2O	0.28	0.62	0.66	1.23	1.18	1.41	1.67	0.59
Na_2O	4.01	3.16	2.93	3.3	2.48	3.36	3.59	3.75
P_2O_5	0.66	0.598	0.635	0.481	0.457	0.452	0.462	0.468
LOI	3.57	6.06	2.72	4.8	4.07	4.45	3.83	6.9
Total	100.01	100.02	100.04	100.13	100.05	100.09	100.11	100.26
Ti	24519	21911	23322	19024	17451	13702	13690	15193
Rb	6.56	12.68	10.31	17.23	17.42	22.93	29.61	8.86
Sr	718	474	474	342	383	464	442	335
Y	37.2	33.2	35.2	33.9	33.3	26.3	26.9	27.6
Zr	200	183	192	190	188	150	152	158
Nb	24.63	22.58	23.23	21.49	21.07	16.24	16.27	17.64
Cs	0.16	0.41	0.28	0.15	0.07	0.11	0.12	0.30
Ba	121	292	444	831	889	696	705	327
Hf	4.79	4.38	4.80	4.55	4.47	3.71	3.68	3.87
Ta	1.60	1.42	1.54	1.38	1.35	1.03	1.02	1.11
Pb	4.02	14.93	4.01	1.70	2.67	3.46	3.58	6.64
Th	2.99	2.72	2.81	2.16	2.11	1.75	1.77	1.83
U	0.68	0.66	0.62	1.27	0.54	0.41	0.40	0.45

Table2 （continued）

Sample	831S41	831S51	830S2	83156	83158	83160	83161	83163
La	37.42	33.39	34.29	28.31	26.41	26.43	26.29	26.38
Ce	74.41	67.07	68.73	55.70	52.86	51.41	51.42	51.79
Pr	9.22	8.53	8.81	7.31	7.17	6.65	6.60	6.84
Nd	39.80	36.03	36.61	32.03	30.63	27.98	28.56	29.78
Sm	8.08	7.36	7.65	6.66	6.59	5.80	5.80	6.17
Eu	2.66	2.46	2.52	2.32	2.30	2.21	2.15	2.40
Gd	8.62	7.95	8.27	7.32	7.09	6.04	6.28	6.41
Tb	1.10	1.00	1.06	0.98	0.93	0.80	0.78	0.81
Dy	7.01	6.40	6.65	6.37	6.02	4.99	5.15	5.17
Ho	1.44	1.30	1.36	1.30	1.24	1.04	1.03	1.06
Er	3.74	3.45	3.63	3.31	3.37	2.77	2.76	2.86
Tm	0.51	0.46	0.48	0.46	0.45	0.36	0.36	0.38
Yb	2.81	2.68	2.76	2.69	2.68	2.11	2.19	2.19
Lu	0.44	0.42	0.43	0.45	0.42	0.33	0.34	0.34


PH
NP
MP
PP
AR
NPG
F
IF
Tillite
Q
78°
84°
90°
44°
40°
36°
Inset B
Urumqi
Tarim block
Beijing
North China block
Yangtze block
800km
A
Wulumuqi
Yining
Tulufan
Tianshan Mt.
Kuerle
Akesu
Fig. 1C
Kuluketage
Kashi
Tarim Basin
Altyn Tagh
Qiemo
Hetian
W. Kunlun Mt.
E. Kunlyn Mt.
0
200
400Km
B


1C
Wushi
N
Aksu
20 km
Wushi Section
Southwest
Aksu Section
Sugetbrak
Section
41.0°N
Sugetbrak
Aksu Group
Qiaoenbrak Fm
Yuermeinak & Sugetbrak Fms
Chigebrak Fm
Paleozoic
79.5°E
80.0°E
80.5°E


A
41°
00'
40°
50'
79°
15'
79°
30'
A group sample
4km
N
Yuermeinak
B group sample
Sugetbrak
Qiaoenbrak Fm
Yuermeinak and Sugetbrak Fms
Cenozoic
Sugetbrak Fm
Chigebrak Fm
Paleozoic
Sample
B
Chigebrak Fm
50m
0m
Dolostone
Sugetbrak Fm
Upper member
Lower member
Sandstone
Basalt
614.4±9.1Ma
615.2±4.8Ma
Sandstone, siltstone & mudstone
Silstone & mudstone with 5-10m basalt
Yuermeinak Fm
Diamictite
Qiaoenbrak Fm
Flysch


Chigebrak Fm
Upper Sugetbrak Fm
Continuous SB(79m)
05822
Lower Sugetbrak Fm
Uncontinuous thin SB(5-10m)
Yuermeinak Fm
Qiaoenbrak Fm


A


B


C
Sugetbrak Basalt


D


E


F

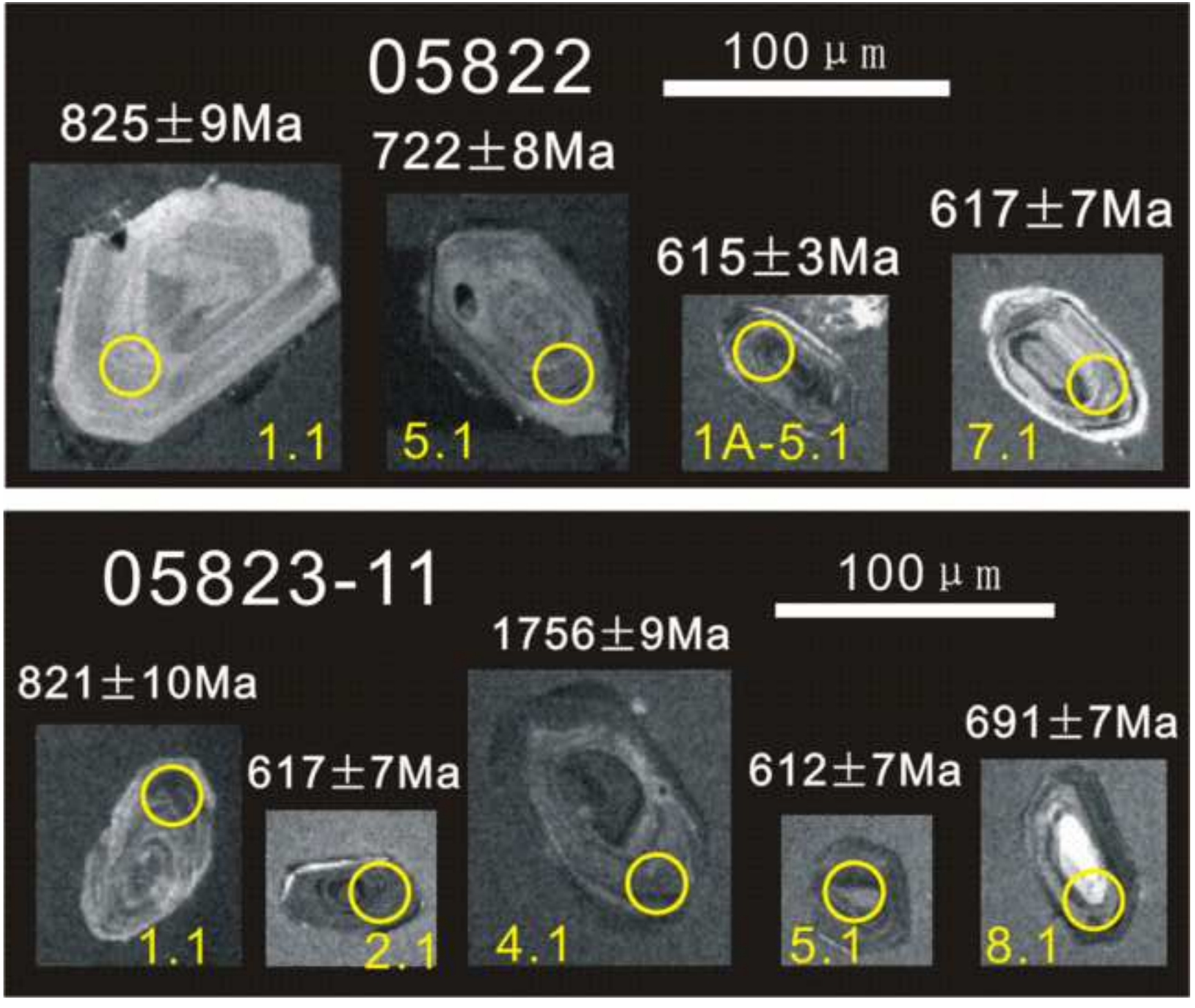

05822
100 μm
825±9Ma
722±8Ma
615±3Ma
617±7Ma
1.1
5.1
1A-5.1
7.1
05823-11
100 μm
821±10Ma
617±7Ma
1756±9Ma
612±7Ma
691±7Ma
1.1
2.1
4.1
5.1
8.1

Figure4BC

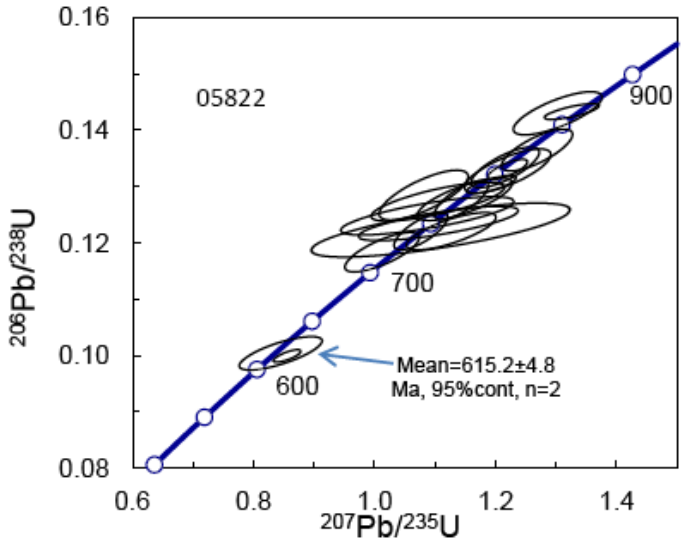


Fig.4B

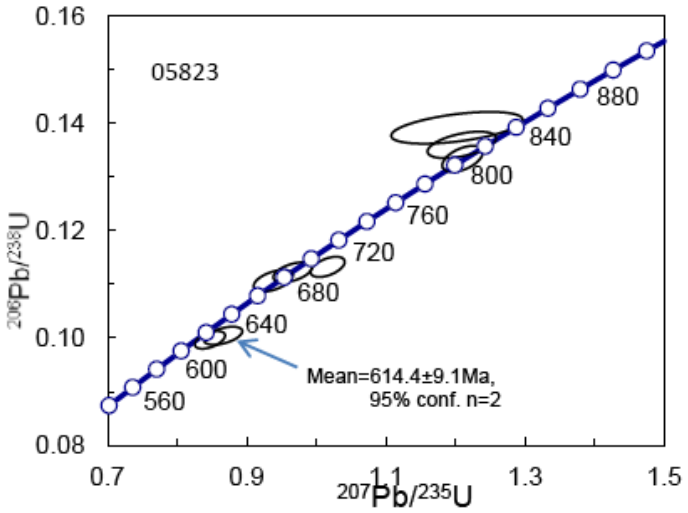


Fig.4C

Figure5AB

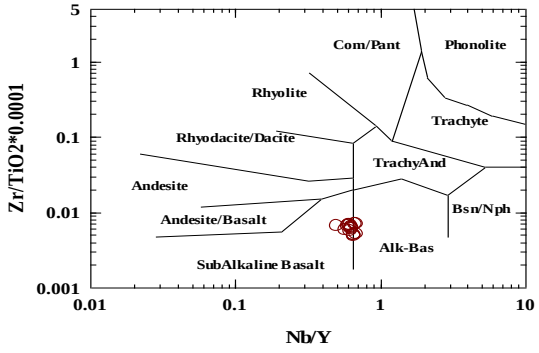


Fig.5A

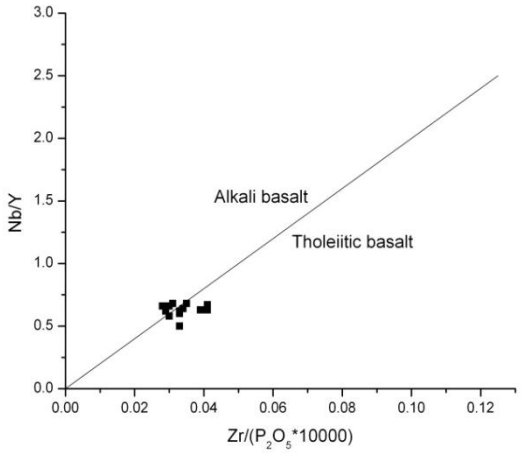


Fig.5B

Figure6AB

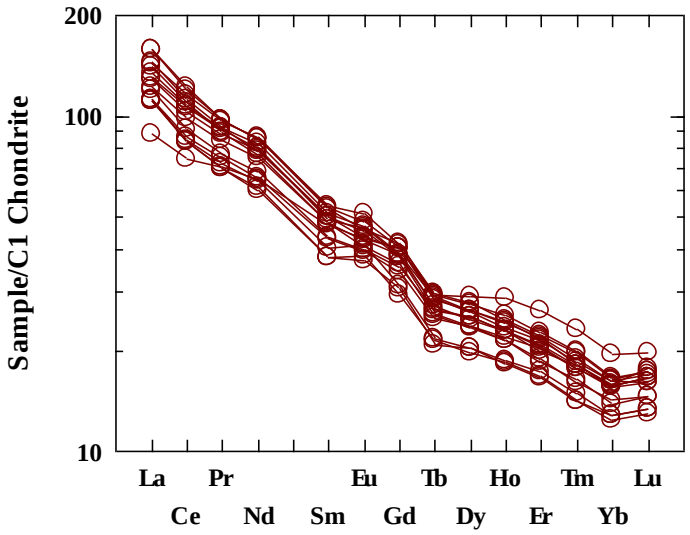


Fig.6A

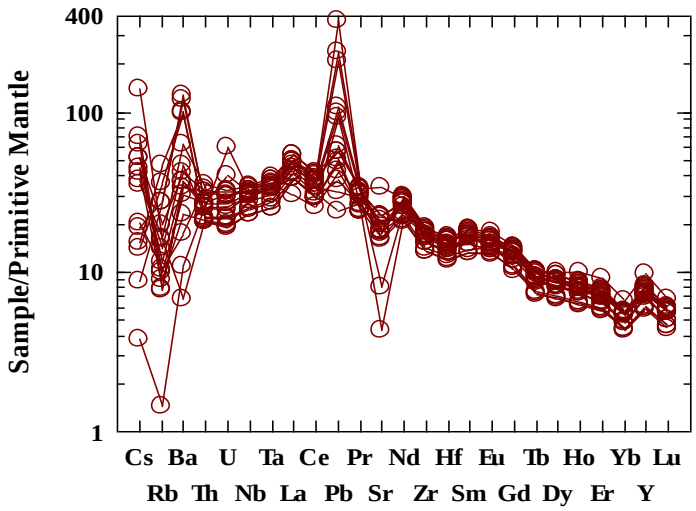


Fig.6B

Figure7AB

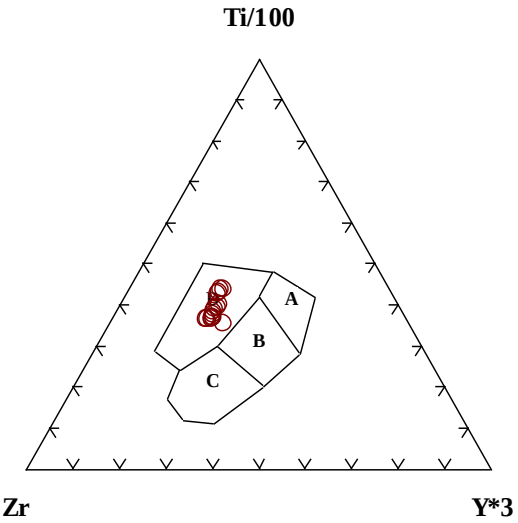


Fig.7A

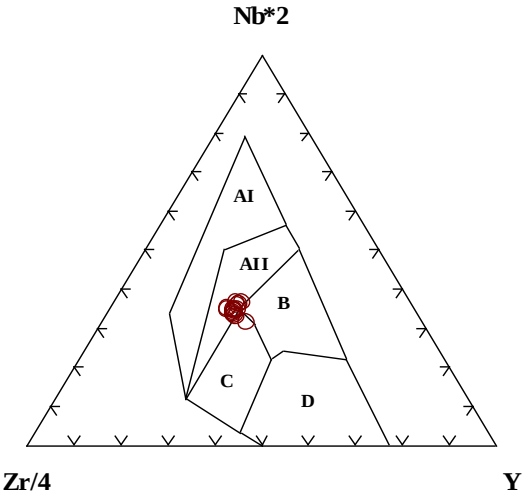
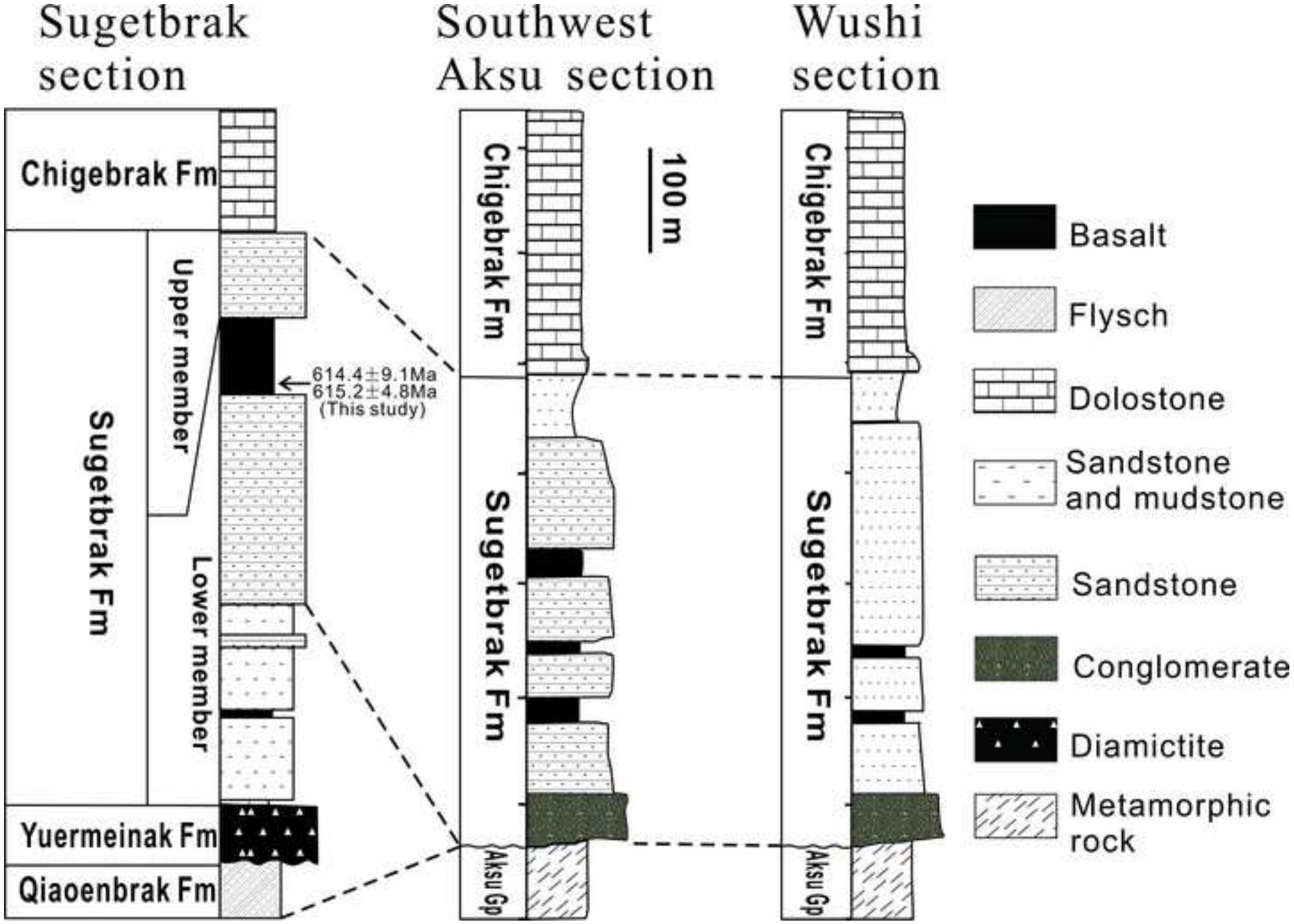
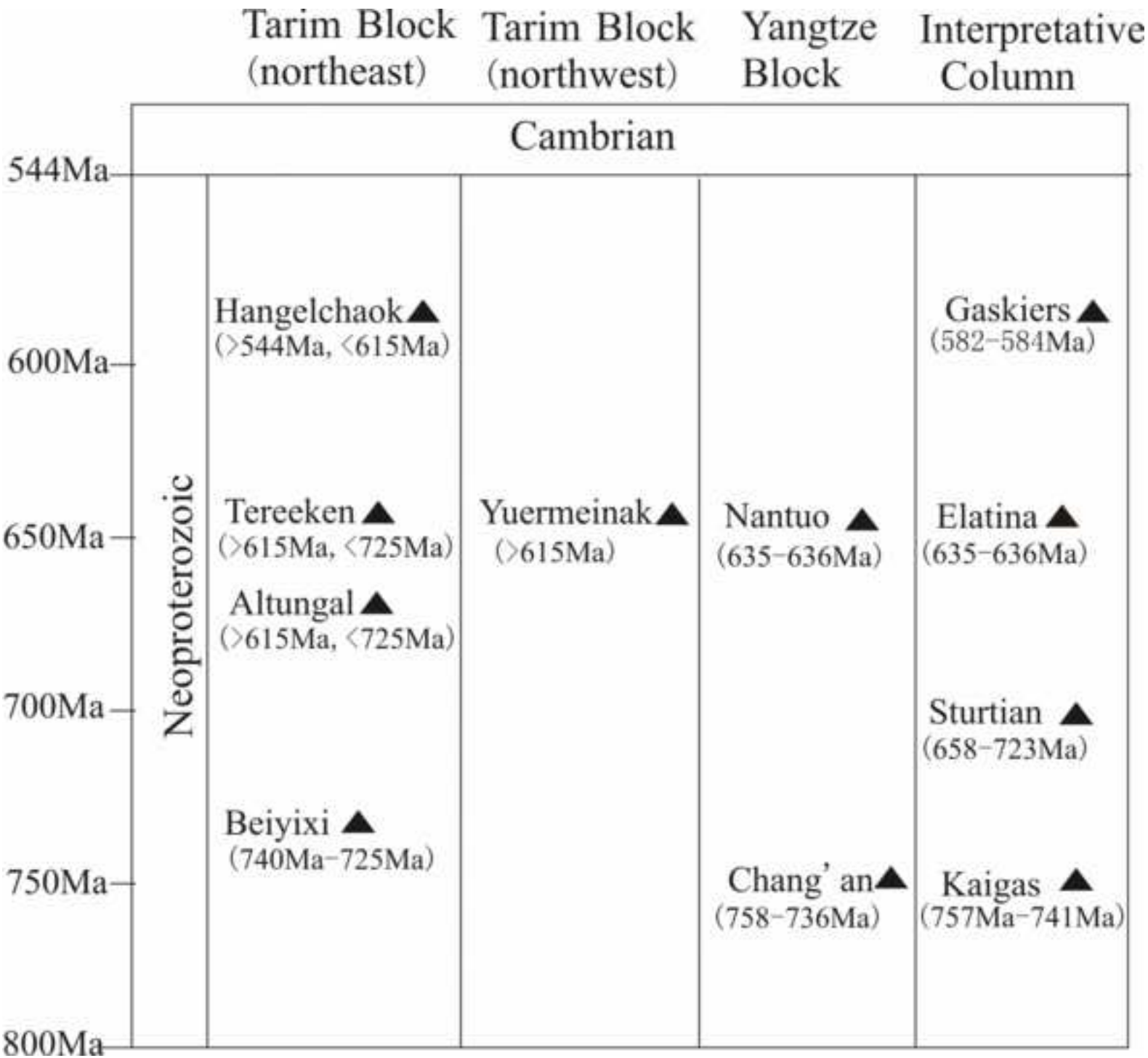


Fig.7B


Sugetbrak section
Southwest Aksu section
Wushi section
Chigebrak Fm
Upper member
Sugetbrak Fm
Lower member
Yuermeinak Fm
Qiaoenbrak Fm
614.4±9.1Ma
615.2±4.8Ma
(This study)
100 m
Aksu Gp
Basalt
Flysch
Dolostone
Sandstone and mudstone
Sandstone
Conglomerate
Diamictite
Metamorphic rock


Tarim Block (northeast)
Tarim Block (northwest)
Yangtze Block
Interpretative Column
Cambrian
544Ma
600Ma
650Ma
700Ma
750Ma
800Ma
Neoproterozoic
Hangelchaok
(>544Ma, <615Ma)
Tereeken
(>615Ma, <725Ma)
Altungal
(>615Ma, <725Ma)
Beiyixi
(740Ma-725Ma)
Yuermeinak
(>615Ma)
Nantuo
(635-636Ma)
Chang' an
(758-736Ma)
Gaskiers
(582-584Ma)
Elatina
(635-636Ma)
Sturtian
(658-723Ma)
Kaigas
(757Ma-741Ma)