

Primary uranium sources for sedimentary-hosted uranium deposits in NE China: insight from basement igneous rocks of the Erlian Basin

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Abstract Carboniferous–Permian, Triassic and Jurassic igneous basement rocks around the Erlian Basin in northeast China have been investigated through detailed mineralogical, whole-rock geochemistry, geochronological data and Sm–Nd isotope studies. Carboniferous–Permian biotite granites and volcanic rocks belong to a calc-alkaline association and were emplaced during the Late Carboniferous–Early Permian (313 ± 1 – 286 ± 2 Ma). These rocks are characterised by positive $\varepsilon_{\text{Nd}(t)}$ (3.3–5.3) and fairly young T_{DM} model ages (485–726 Ma), suggesting a dominant derivation from partial melting of earlier emplaced juvenile source rocks. Triassic biotite granites belong to a high-K calc-alkaline association and were emplaced during the Middle Triassic (243 ± 3 – 233 ± 2 Ma). Their negative $\varepsilon_{\text{Nd}(t)}$ (–2 to –0.1) and higher T_{DM} model ages (703–893 Ma) suggest a contribution from Precambrian crust

during the magma generation processes, leading to a strong enrichment in K and incompatible elements such as Th and U. Highly fractionated magmas crystallised in U-rich biotite (up to 21 ppm U) and two-mica granites. In biotite granite, the major U-bearing minerals are uranothorite and allanite. They are strongly metamict and the major part of their uranium (90 %) has been released from the mineral structure and was available for leaching. Mass balance calculations show that the Triassic biotite granites may have, at least, liberated ~14,000 t U/km³ and thus correspond to a major primary uranium source for the U deposits hosted in the Erlian Basin.

Keywords Erlian Basin · Uranium sources · Carboniferous–Permian and Triassic igneous rocks

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Introduction

The Central Asia Uraniferous Province (CAUP) represents one of the largest metallogenic provinces in the world with uranium ore deposits located from southern Kazakhstan, Uzbekistan, Kyrgyzstan, the Transbaikal area in Russia, Mongolia and North China (OECD-NEA/IAEA 2010). With respect to other world-class provinces (Canada, Australia and Africa), which are associated with Archaean to Palaeoproterozoic basements, the study area provides an opportunity to investigate the genetic conditions of uranium deposits formed within a much younger uranium province, although some Precambrian basement rocks occur in the western part of the CAUP (Kröner et al. 2014). In northeast China, the main type of uranium deposits is sandstone-hosted, with both roll-front and tabular-shaped ore bodies (e.g. the Bayinwula deposit in the Erlian Basin, Bonnetti et al. 2015b; the Dongsheng deposit in the Ordos

Basin, Cai et al. 2007; the Qianjiadian deposit in the Songliao Basin, Li et al. 2008), which currently represents the most valuable uranium resource (>50 % of identified uranium resources in China; OECD-NEA/IAEA 2014) using in situ recovery for uranium extraction. Nie (2008) and Dahlkamp (2009) have reported the occurrence of biotite granites enriched in uranium (4–13 ppm U), relatively to the average U content in granites (i.e. $U_{\text{mean}}=3.8$ ppm; Cuney and Kyser 2008), in the basement surrounding the Erlian Basin. These granites also show elevated Th/U ratios (5.6–11.1; Dahlkamp 2009), indicating that a significant part of their uranium content has been leached out and thus may have represented a major source for the sedimentary-hosted uranium deposits in the basin. However, no geochronological constraint exists to link the uranium enrichment in these granites to a particular geodynamic event and the mineralogy and the geochemistry of uranium have not been specifically studied to determine its availability at the time of deposit formation in the Erlian Basin and in other adjacent sedimentary basins.

The purpose of this paper was therefore to characterise the potential primary uranium source rocks in the basement around the Erlian Basin that could be at the origin of the sedimentary rock-hosted deposits. Mineralogical, geochemical, and U–Pb and Sm–Nd isotopic data are used to characterise the degree of differentiation and U enrichment of the basement igneous rocks to determine their emplacement age and to describe their U-bearing minerals in order to identify the most favourable source of uranium. The study focuses on the Carboniferous–Permian and Triassic igneous rocks, which are the most abundant igneous basement rocks within the study area.

Geological setting

Eastern Asia was built during the Palaeozoic by successive subductions and collisions of intra-oceanic arcs and continental micro-blocks around the Siberia craton (Lin et al. 2008). Along the northern margin of the North China Craton (NCC), the Central Asian Orogenic Belt (CAOB), extending from Kazakhstan to the Pacific Ocean, has recorded a long history of arc and micro-continent accretion (Kröner et al. 2007; Windley et al. 2007) and is characterised by a vast distribution of Palaeozoic and Mesozoic granitic intrusions (Jahn et al. 2000). In northeast China, especially in Inner Mongolia, the CAOB is represented by a 300-km-wide NW/SE-trending belt that separates Mongolian arcs developed along the southern margin of Siberia from micro-continent and magmatic arcs belonging to the NCC (Xiao et al. 2003; Zhai and Liu 2003;

Zhai and Santosh 2011). During the Carboniferous–Early Permian, the Baolidao magmatic arc formed in response to slab rollback and retreat within a subduction zone (Xiao et al. 2003; Lin et al. 2008; Li et al. 2015). Then, these magmatic arcs and micro-continent were completely accreted to Siberia and the NCC during the Late Permian–Early Triassic, which is highlighted by a suture zone along the northern margin of the NCC (Xiao et al. 2003; Lepvrier and Maluski 2008; Lin et al. 2008). During the Jurassic (i.e. 190–150 Ma; Miao et al. 1998; Wu et al. 2000a, 2002, 2005b, c; Guo et al. 2004; Su et al. 2007), alternating transtensional and transpressional tectonic stages are responsible for the widely represented Jurassic magmatic event in northeast China (Faure et al. 2012; Charles et al. 2013). Then, intracontinental basins such as the Erlian, Yingen and Songliao basins developed (Meng 2003; Sha 2007; Charles et al. 2013) from the Late Jurassic to the Late Cretaceous during a major stage of continental extension related to the subduction of the paleo-Pacific plate (Ren et al. 2002; Charles et al. 2013).

The Erlian Basin (or Eren Basin) is located in Inner Mongolia, close to the border between China and Mongolia (Fig. 1), and covers an area of approximately 130,000 km² (Dou and Chang 2003; Wei et al. 2005). It is composed of five major sub-basins (Fig. 1): the Chuanjing Sub-basin to the west, the Wulanchabu sub-basin in the central–western part, the Manite sub-basin in the central–northern area, the Tengge'er sub-basin in the southeastern part and the Wunite sub-basin to the northeast (Dou et al. 1998; Dou and Chang 2003; Wei et al. 2005). These sub-basins are framed by a series of horsts or ranges oriented ENE–WSW: the Daxing'anling Ranges to the East, the Bayinbaolige Horst to the north, the Suolunshan Horst more to the west and, finally, the Wendurmiao Horst to the south. The Sunite Horst occupies the central part of the Erlian Basin and divides the basin into northern and southern parts (Dou et al. 1998; Wei et al. 2005).

The Erlian Basin developed during three main stages: (1) a pre-rift stage mainly characterised by Late Jurassic to Early Cretaceous volcanic rocks of the Qinganling Group (156.0–142.6 Ma; Chen and Chen 1997; Graham et al. 2001; Wang et al. 2006), which corresponds to the initiation of rifting; (2) a syn-rift stage characterised by deposition of Early Cretaceous sedimentary units (Gou et al. 1986; Cai et al. 1990; Sha 2007)—the Arshan Formation dominated by alluvial, fluvial and shallow lacustrine environments and the Tengge'er Formation dominated by deep to semi-deep lacustrine environment; and (3) a post-rift stage characterised by a switch from tectonic to thermal subsidence and the deposition of the Early Cretaceous Saihan Formation unconformably lying on syn-rift sediments (Nie et al. 2010; Bonnetti et al. 2014) and mainly composed of sandstone and minor mudstone and the Late Cretaceous Erlian Formation including a lower member dominated by sandstone and an upper member dominated by mudstone (Bonnetti et al. 2014). Sediments from both

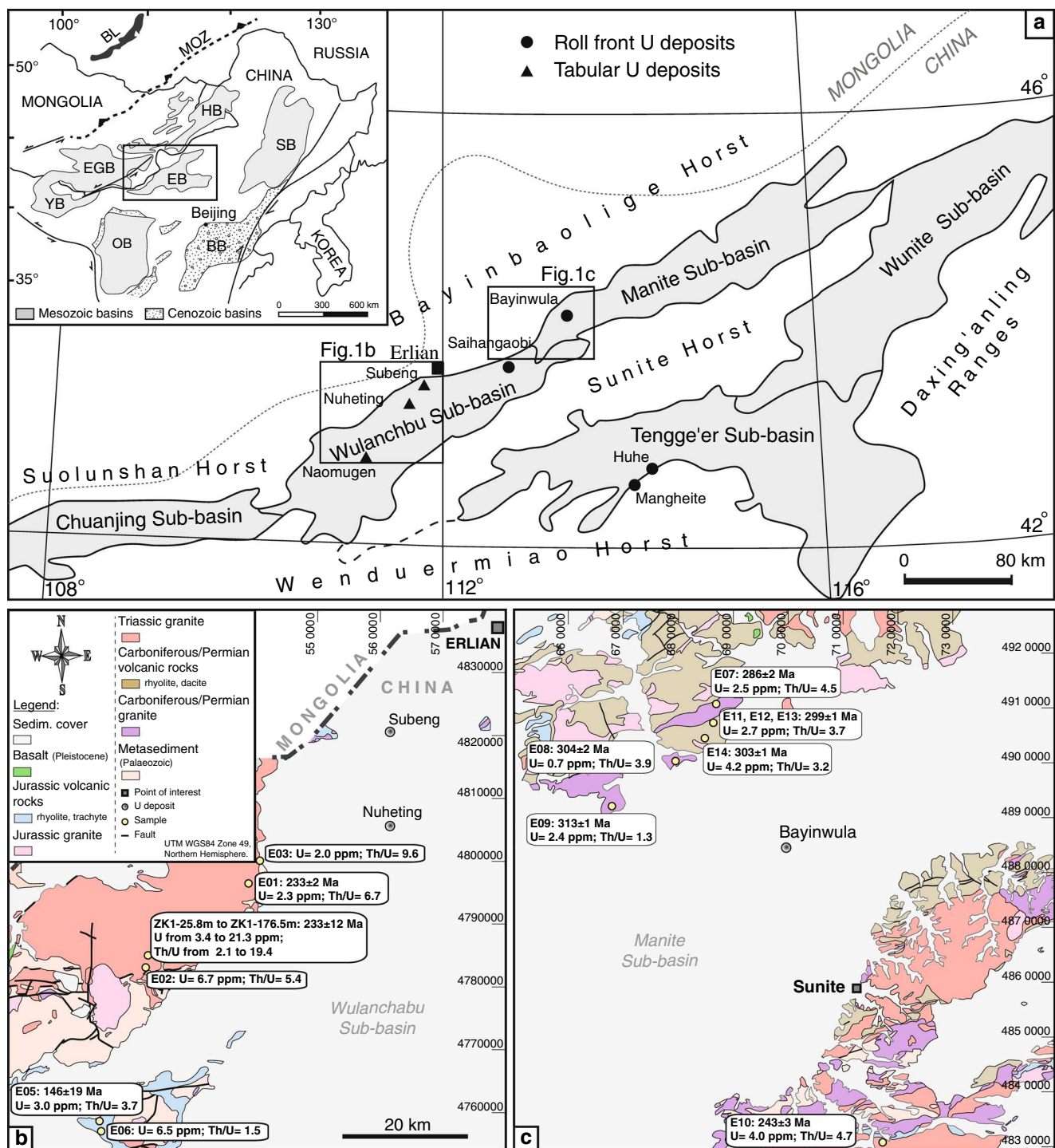


Fig. 1 Geological maps of the Erlian Basin and the study areas. **a** Structural map of the Erlian Basin showing the distribution of different types of uranium deposits. *U* uranium, *BB* Bohai Basin, *EB* Erlian Basin, *EGB* East Gobi Basin, *HB* Hailar Basin, *OB* Ordos Basin, *SB* Songliao Basin, *YB* Yingen Basin, *BL* Baikal Lake, *MOZ* Mongol–Okhotsk Zone. Modified after Bonnetti et al. (2014), CNNC (unpublished data),

Dahlkamp (2009), Dou et al. (1998), Meng (2003) and Wei et al. (2005). **b** Geological map of the central part of the Wulanchabu sub-basin. **c** Geological map of the western part of the Manite sub-basin. Modified after CNNC (unpublished data). Age, *U* content and *Th/U* value per sample are indicated when they have been analysed

formations were deposited in fluvial and shallow lacustrine environments (Currie and Eberth 1993; Lin et al. 2001; Van Itterbeeck et al. 2005, 2007; Bonnetti et al. 2014).

The Erlian Basin lies on a folded and metamorphosed basement corresponding to the southern margin of the Xing'an or Xing Meng Mongolian Orogenic Belt (Lin et al. 2001; Dou

and Chang 2003; Wei et al. 2005), which represents the south-eastern part of the CAOBS that was accreted during the Late Permian (Xiao et al. 2003). Then, a south-directed thrust and fold belt developed during the Triassic–Jurassic (Davis et al. 2001; Xiao et al. 2003; Faure et al. 2012). This basement is mainly composed of: (1) Carboniferous–Permian to Early Cretaceous igneous rocks including intermediate to felsic granitic plutons (Miao et al. 1998; Wu et al. 2000a, 2002, 2005a, b, c, 2007; Xiao et al. 2003; Guo et al. 2004; Ge et al. 2007; Su et al. 2007; Yang et al. 2007; Zhang et al. 2010; Zhou et al. 2010; Li et al. 2012) and mafic to felsic volcanic rocks (Ge et al. 2007; Ying et al. 2010a, b; Zhou et al. 2010) and (2) Palaeozoic sedimentary or metasedimentary units (Lin et al. 2008; Zhou et al. 2010). From north to south, there is a parallel alternation of horsts and half-grabens, which could be related to a ‘Basin and Range’ structural pattern (Wernicke et al. 1988; Ren et al. 1998; Meng et al. 2003).

The selected areas for this study (Fig. 1) correspond to the central part of the Wulanchabu sub-basin (Fig. 1b) and the western part of the Manite sub-basin (Fig. 1c). Both areas were selected because basement rocks are well exposed and because they surround parts of the basin that host uranium deposits (e.g. the Nuheting deposit, Bonnetti et al. 2015a; the Bayinwula deposit, Bonnetti et al. 2015b; Fig. 1). The study was carried out in the Bayinbaolige Horst and in the northern part of the Sunite Horst.

Materials and methods

Twenty-one samples (Fig. 1), including 15 samples of granite and six samples of volcanic rocks were collected from outcrops and in the ZK1 drill hole (Fig. 1b). Basement igneous rocks were sampled at the western margin of the Wulanchabu Sub-basin (Fig. 1b) and at the northern and southern margins of the Manite sub-basin (Fig. 1c). Whole-rock geochemistry was performed by inductively coupled plasma mass spectroscopy/optical emission spectroscopy (Carignan et al. 2001) at the Service d’Analyses des Roches et des Minéraux – Centre de Recherches Pétrographiques et Géochimiques (CRPG) in Nancy, France. The mineralogical study was carried out at the GéoRessources laboratory using optical microscopy (transmitted and reflected light) and at the Service Commun de Microscopies Electroniques et de Microanalyses of the Lorraine University (Nancy, France) using: (1) a Hitachi S-4800 scanning electron microscope (SEM) and (2) a CAMECA SX100 electron probe microanalyser (EPMA). The U–Pb isotope dating on zircon was performed using the CAMECA IMS 1270 ion microprobe facility, at CRPG (Nancy). The analysed masses were: 203.5 (background noise, measured for 4 s), Zr_2O (2 s), ^{204}Pb (10 s), ^{206}Pb (4 s), ^{207}Pb (40 s), ^{208}Pb (10 s), ^{238}U (2 s),

$^{238}\text{U}^{16}\text{O}$ (2 s) and $^{238}\text{U}^{16}\text{O}_2$ (2 s). Each analysis consists of 16 iterative cycles over each mass. The mass resolution was about 5000, which is sufficient to separate the molecular interferences. It is important to note that the quantity of analysed material was very low, which constitutes one limitation of this method: the analytical uncertainty will be greater with younger samples. The age calculations were based on the isotopic ratios corrected for background noise and common lead (using ^{204}Pb). The U and Pb abundances are calculated on the basis of the Zr_2O and UO_2 correlation for the standard and the isotope ratio $^{238}\text{U}/^{206}\text{Pb}$ (Compston et al. 1984). The $^{207}\text{Pb}/^{206}\text{Pb}$ ratios are directly derived from each spot of analysis. The standard used for determination of the correlation line, following Compston et al. (1984), was the 91500 zircon dated at 1064 Ma (Wiedenbeck et al. 1995). U–Th–Pb chemical dating on magmatic monazites was performed by EPMA equipped with five wavelength-dispersive spectrometers using an acceleration voltage of 20 kV and a beam current of 100 nA. Counting times (peak + background) were 30 s for Th, 30 s for U and 80 s for Pb. The detailed analytical method and the age calculation procedure are presented in Cocherie and Albarede (2001). Six drill core samples belonging to the sediments of the Erlian and Saihan formations were also selected to perform U–Th–Pb chemical dating on monazite detrital grains. Sample DH3-54.6m belongs to the sediments of the Erlian Formation at the Nuheting deposit (Fig. 1b; see drill hole location and sample information in Bonnetti et al. 2015a) and samples DH1-111.6m, DH2-111.8m, DH2-121.3m, DH2-146.2m and DH3-118.0m belong to the sediments of the Saihan Formation at the Bayinwula deposit (Fig. 1c; see drill hole location and sample information in Bonnetti et al. 2015b). Sm–Nd isotopic values were determined on crushed whole-rock samples by isotope dilution. All analyses were carried out at the Geosciences Rennes Laboratory, at the University of Rennes 1. Samples were spiked with a ^{150}Nd – ^{149}Sm mixed solution and dissolved in HF–HNO₃. Rare earth elements (REEs) were separated using BioRad AG 50W × 8H⁺ 200–400 mesh cationic resin. Sm and Nd were separated and collected by passing the solution through a further set of ion exchange columns loaded with Ln spec Eichrom resin. Sm and Nd were loaded with HNO₃ reagent onto double Re filaments and analysed in a Finnigan MAT262 multicollector mass spectrometer in static mode. In each analytical session, the unknowns were analysed together with the Ames Nd-1 Nd standard, which, during the course of this study, yielded an average of 0.511964 (standard deviation = 7×10^{-6}). All analyses were adjusted to a nominal $^{143}\text{Nd}/^{144}\text{Nd}$ value of 0.511850 for the La Jolla standard. Mass fractionation was monitored and corrected using the value $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. Procedural blanks analysed during the period of these analyses were ~190 pg and are considered to be negligible compared to the total quantity of Nd in the samples.

Results

Both Carboniferous–Permian and Triassic igneous rocks (ESM Fig. S1) are well represented in the basement surrounding the Wulanchabu and Manite sub-basins (Fig. 1). These igneous rocks, especially granite, are strongly weathered and present a late set of fractures that may have played a significant role in the weathering process. Drill core observations on the ZK1 drill hole (Fig. 1b) showed that the weathering affected the granites vertically for several tens of metres below the surface. Nevertheless, non-weathered sections of these granites can be observed in the deeper part of the ZK1 drill hole. Jurassic volcanic rocks located at the base of the Wulanchabu sub-basin have also been studied. The main mineralogical, geochemical and chronologic characteristics of the selected samples are presented in Table 1.

Petrography and mineralogy

Carboniferous–Permian biotite granites

Carboniferous–Permian biotite granites are mainly isotropic with a granular to microgranular texture and do not show evidence of deformation. Their accessory mineral assemblage consists of allanite, zircon, apatite, titanite, and Fe/Ti and Ti oxides (Table 1). In outcrop samples, allanite is always strongly metamict, altered and presents a destruction rim, indicating high initial radioactive element concentration within the mineral. However, its present uranium content is below 0.01 wt% UO_2 . In the destruction rims of allanite, secondary epidote, monazite and hematite are newly formed. Uranium concentration in zircon is generally below the detection limit of the EPMA and only some rare altered Fe/Ti oxides show a significant U content (a few hundreds of parts per million UO_2 ; ESM Table S1). Therefore, in the Carboniferous–Permian biotite granites, allanite seems to represent the main U-bearing accessory mineral.

Carboniferous–Permian volcanic rocks

Whole-rock geochemical data (ESM Fig. S1b) indicate two different types of Carboniferous–Permian felsic volcanic rocks: (1) dacite and (2) rhyolite. Both present a trachytic texture and have a low proportion of glass in the matrix. No U-bearing accessory mineral was identified.

Triassic biotite granites

Triassic biotite granites present a granular to microgranular texture and may show ductile deformation characterised by grain boundary migrations of quartz crystals mainly developed in shear zones at the thin section scale (Fig. 2c). Their

Table 1 Sample description from the basement igneous rocks surrounding the Wulanchabu and Manite sub-basins

Sample no.	Rock facies	Geological time	Age (Ma)	Main mineral assemblage	Accessory mineral assemblage	Alteration mineral assemblage	U content (ppm)	Th/U	Geochemical key features
E05, E06	Rhyolite, trachyte	Jurassic	146 ± 19	Quartz, K-feldspar, plagioclase, microliths	Zircon, monazite, xenotime, Fe/Ti and Ti oxide	Epidote, sericite, Fe oxide	3.0 and 6.5	3.7 and 1.5	High-K calc-alkaline association, slightly to significantly peraluminous, weathered
ZK1-25.8m–ZK1-176.5m, E01–E03	Biotite granite	Triassic	233 ± 12, 233 ± 2	Quartz, K-feldspar, plagioclase, biotite, hornblende	Uraniothorite, allanite, titanite, zircon, apatite, Fe/Ti and Ti oxide	Chlorite, epidote, REE carbonate, pyrite, Fe oxide	2.0–21.3	2.1–19.4	High-K calc-alkaline association, slightly metaluminous to slightly peraluminous, weathered at the outcrop and at shallow depth, non-weathered at depth >80 m
E10	Two-mica granite	Triassic	243 ± 3	Quartz, K-feldspar, plagioclase, biotite, muscovite	Monazite, xenotime, zircon, apatite	Chlorite, epidote, quartz II, muscovite II	4.0	4.7	High-K calc-alkaline association, peraluminous, greisenised, weathered
E04, E07–E09	Biotite granite, granodiorite	Carboniferous/Permian	286 ± 2, 304 ± 2, 313 ± 1	Quartz, K-feldspar, plagioclase, biotite, hornblende	Allanite, titanite, zircon, apatite, Fe/Ti and Ti oxide	Chlorite, epidote, monazite, Fe oxide, carbonate	0.7–3.1	1.3–6.0	Low- to medium-K calc-alkaline association, slightly metaluminous to slightly peraluminous, weathered
E11–E14	Rhyolite, dacite	Carboniferous/Permian	299 ± 1, 303 ± 1	Quartz, K-feldspar, plagioclase, microliths	Zircon, Ti oxide	Epidote, sericite, Fe oxide	1.7–4.2	2.8–4.1	Low- to medium-K calc-alkaline association, slightly to significantly peraluminous, weathered

Presented are the type of rock (see ESM Fig. S1), the geological time and age determination when analysed (see Figs. 5 and 6), the main/accessory/alteration mineral assemblage (see Fig. 2), the whole-rock U content, the Th/U ratio and the geochemical key features (see Figs. 3 and 4)

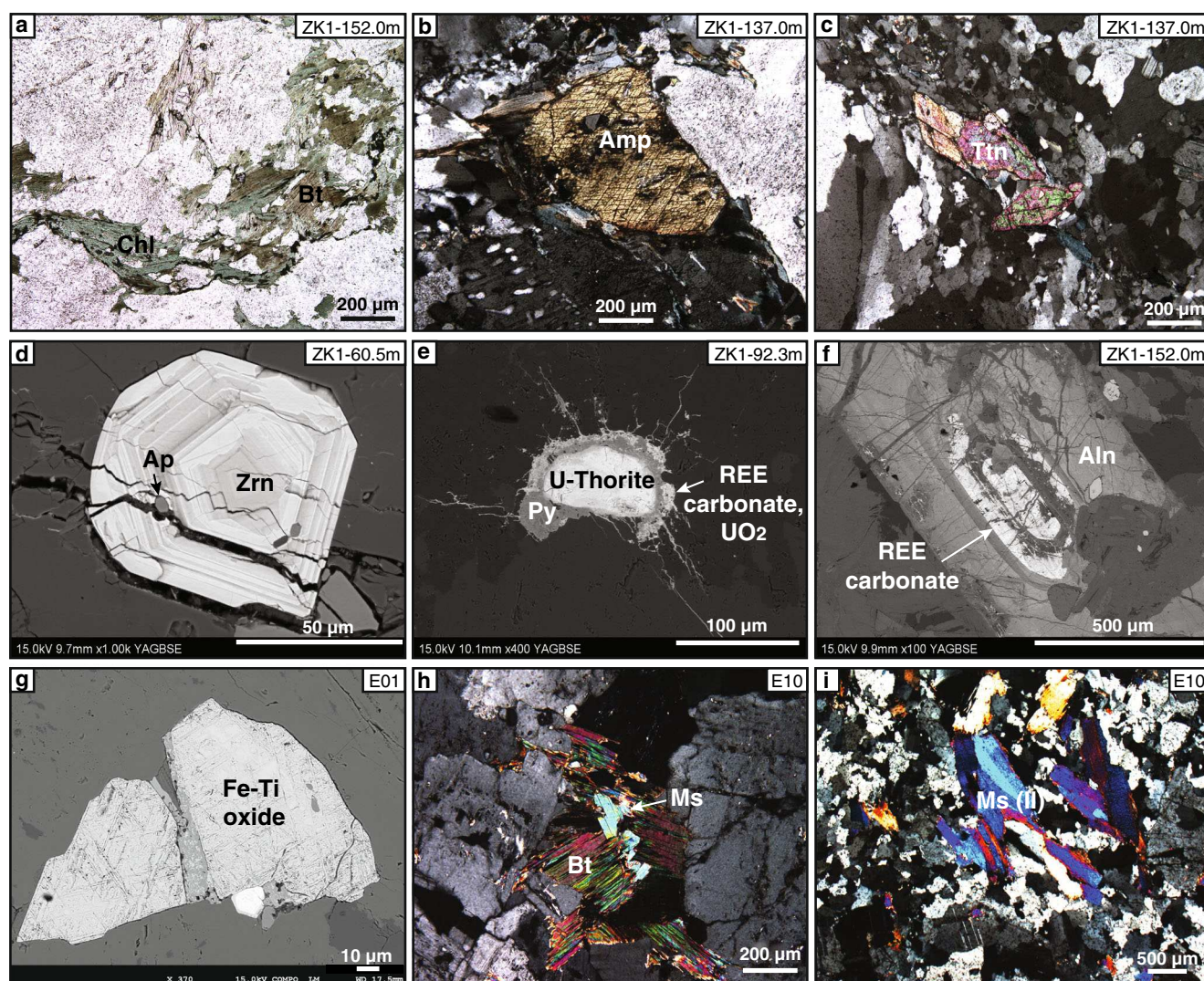


Fig. 2 Major and accessory minerals in biotite and two-mica granites of the basement surrounding the Wulanchabu and Manite sub-basins. **a** Biotite and chlorite (respectively *Bt* and *Chl*). **b** Amphibole (*Amp*). **c** Titanite (*Ttn*). **d** Zircon and apatite (respectively *Zrn* and *Ap*). **e** Metamict U-thorite showing a destruction rim associated with pyrite

(*Py*), rare earth elements (*REE*) carbonate and uranium oxide. **f** Metamict allanite (*Aln*) associated with REE carbonate. **g** Fe-Ti oxide. **h** Biotite associated with muscovite (respectively *Bt* and *Ms*). **i** Secondary quartz and muscovite (*MsII*) in the greisenised part of the two-mica granite (next to the location of sample E10)

accessory mineral assemblage (Table 1) is titanite (Fig. 2c), zircon (Fig. 2d), uranothorite (Fig. 2e), allanite (Fig. 2f), Fe/Ti and Ti oxides (Fig. 2g) and apatite (Fig. 2d). Zircon has a low to moderate U content (up to 1.57 wt% UO_2 ; ESM Table S1), allanite has up to 0.38 wt% UO_2 (ESM Table S1), and U-thorite presents a variable U concentration from <1 wt% up to 13.25 wt% UO_2 (ESM Table S1) depending on its degree of alteration and metamictisation. Some altered Fe-Ti oxides, characterised by a strong depletion of their Fe and Mn contents, contain significant uranium concentrations (up to 0.09 wt% UO_2 ; ESM Table S1). The major U-bearing accessory minerals such as U-thorite and allanite both present destruction rims due to alpha radiation. In the destruction rim of allanite, a REE carbonate of bastnäsite type and secondary epidote are newly formed (Fig. 2f). In the destruction rim of

U-thorite, pyrite, uranium oxide, REE carbonate and galena have been identified (Fig. 2e). In the non-weathered part of the biotite granites (at several tens of metres depth), these U-bearing accessory minerals are metamict and their U contents have been partly to totally released from the mineral structures into microfractures, but not leached out of the rock, as attested by their Th/U ratios remaining close to 4. In the weathered part of these granites, the destruction rims of these two major U-bearing accessory minerals are associated with Fe oxide. Their uranium concentrations are very low, generally below the EPMA detection limit (U-thorite tends to present a composition close to the thorite end member; ESM Table S1). The initial U content of these minerals has been leached out by oxidising fluid, as shown by the high Th/U ratios of the corresponding whole rocks (5.4–19.4; ESM Table S3).

Triassic two-mica granites

The Triassic two-mica granite presents a granular texture (Fig. 2h) and is weakly deformed. U-bearing accessory minerals such as monazite and xenotime have U contents of a few thousands of parts per million and 1–2 wt% UO_2 , respectively (ESM Table S2). Part of the granite has been greisenised (Fig. 2i) with the replacement of the feldspars by secondary quartz and muscovite. No evidence of alteration of these U-bearing minerals has been observed. Uraninite may have been present in such granite (Cuney and Friedrich 1987), but has not been detected here.

Jurassic volcanic rocks

Whole-rock geochemical data (ESM Fig. S1b) indicate two different types of Jurassic felsic volcanic rocks: (1) a trachyte and (2) a rhyolite. Both show a trachytic to microgranular texture and have a low proportion of glass in the matrix. U contents in accessory minerals such as monazite and xenotime never exceed 1 % and no evidence of alteration of these minerals has been observed.

Whole-rock geochemistry

The different types of sampled igneous rocks are highly felsic (ESM Fig. S1 and Table S3). Incompatible trace element ratios (ESM Table S5) are high to very high and characterise highly fractionated magmas compared to the reference ratios presented in Saunders et al. (1988) and Weaver (1991). Moreover, in spider diagrams, these igneous rocks are characterised by high but variable Ba, U and Th positive anomalies (ESM Fig. S2 and Tables S3 and S4) and by REE patterns enriched in light REE relatively to heavy REE.

Carboniferous–Permian igneous rocks

The Carboniferous–Permian igneous rocks are characterised by a large variation of the plagioclase/K-feldspar proportions ($-195 \leq P \leq -18$; Fig. 3) associated with a moderate variation of the quartz content ($129 \leq Q \leq 211$; Fig. 3). Such variations are typical of a low- to medium-K calc-alkaline association. These rocks are slightly metaluminous to significantly peraluminous ($-9 \leq A \leq 58$; Fig. 4) and leucocratic ($28 \leq B \leq 78$; Fig. 4). Therefore, the Carboniferous–Permian igneous rocks belong to a low to moderate calc-alkaline metaluminous magmatism.

Triassic igneous rocks

The biotite granites present a moderate variation of the K-feldspar/plagioclase proportions, with higher K-feldspar

contents ($-110 \leq P \leq -22$; Fig. 3), associated with a moderate variation of the quartz content ($130 \leq Q \leq 188$; Fig. 3) typical of a high-K calc-alkaline association. These granites are slightly metaluminous to significantly peraluminous ($-17 \leq A \leq 17$; Fig. 4) and leucocratic ($12 \leq B \leq 48$; Fig. 4). Therefore, the Triassic biotite granites likely belong to a high-K calc-alkaline metaluminous magmatism. They are more enriched in incompatible elements than the Carboniferous–Permian igneous rocks, as shown by the stronger positive anomalies in Ba, light REE, U and Th in the spider diagrams and REE patterns (ESM Fig. S2). The two-mica granite has *P* and *Q* parameters (39 and 265, respectively; ESM Table S6) much higher than the average granite composition, typical of greisenisation. It is significantly peraluminous ($A=23$) and highly felsic ($B<20$).

Jurassic volcanic rocks

The rhyolite has a high proportion of quartz ($Q=316$) and K-feldspar relatively to plagioclase ($P=59$), is highly felsic ($B=23$) and significantly peraluminous ($A=35$; Figs. 3 and 4). The trachyte has a low proportion of quartz ($Q=117$) and a high proportion of K-feldspar relatively to plagioclase ($P=-43$) and is slightly peraluminous ($A=5$). The *Q* and *P* values of both samples indicate that the Jurassic volcanic rocks most probably belong to a high-K calc-alkaline association, although it is very difficult to characterise confidently the magmatic series for these volcanic rocks with only two samples. Similarly to Carboniferous–Permian igneous rocks, Jurassic volcanic rocks have lower positive anomalies in Ba, light REE, U and Th in the spider diagrams and REE patterns than anomalies for Triassic igneous rocks.

Geochronology

Carboniferous–Permian and Triassic granites as well as Carboniferous–Permian and Jurassic volcanic rocks occurring in the basement around the Wulanchabu and Manite sub-basins (Fig. 1) were dated by U–Pb isotopes on zircon. Monazites of the Triassic two-mica granite occurring in the northern part of the Sunite Horst (Fig. 1c) and detrital monazite grains from post-rift sediments of the Erlian Basin were dated using the U–Th–Pb chemical method.

U–Pb isotopic dating on zircon

Prior to dating, zircon crystals were first analysed by the EPMA and only the non-metamict crystals and the most homogeneous zones have been selected for in situ dating. Thirty zircons from the biotite granites and volcanic rocks (ESM Fig. S3) were dated by the ion microprobe (ESM Table S7 and Fig. 5). The 39 analytical points are represented in Concordia diagrams (Fig. 5). A Concordia age (Ludwig

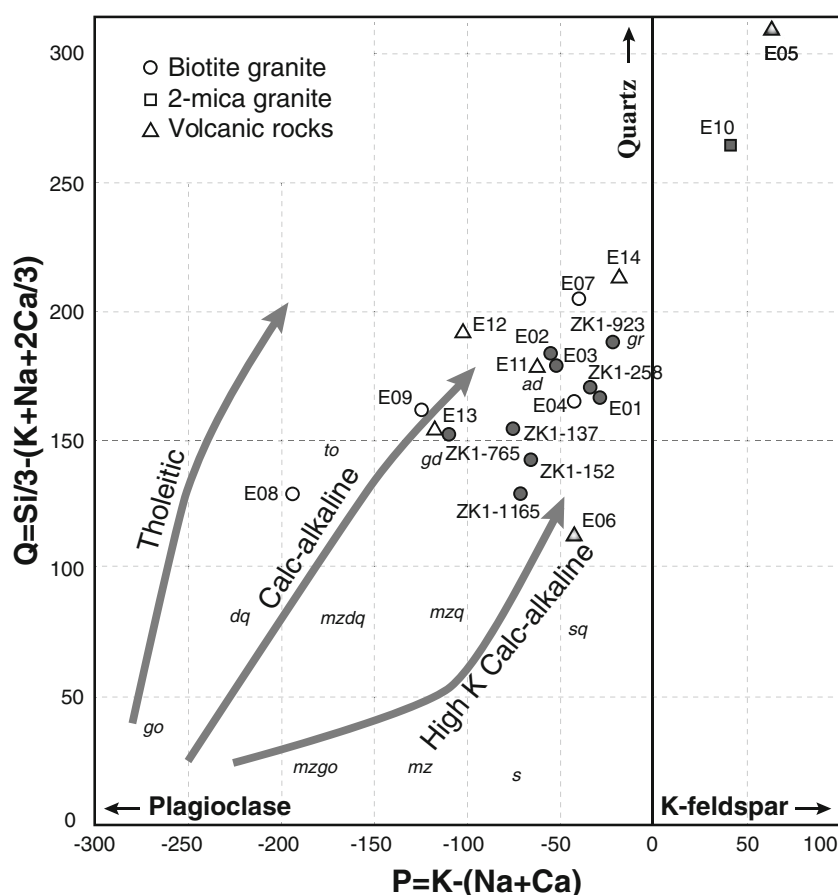


Fig. 3 Q - P diagram showing the variation of the quartz content relative to the variations of proportions of K-feldspar and plagioclase in basement igneous rocks surrounding the Erlian Basin (modified after Debon and Lefort 1983, 1988; see Q and P values in ESM Table S6). Q = chemical-mineralogical parameter (in millications) estimating the quartz content of a rock by subtracting silica bound to the feldspar from the total silica content of the rock. P = chemical-mineralogical parameter (in millications) representing the proportion of K-feldspar (positive)

relative to plagioclase (negative). *gr* granite, *ad* adamellite, *sq* quartz syenite, *s* syenite, *gd* granodiorite, *to* tonalite, *mzq* quartz-monzonite, *mz* monzonite, *mqz* quartz-monzodiorite, *mqz* monzo gabbro-diorite, *go* gabbro. Data points in white belong to Carboniferous-Permian igneous rocks; data points in dark grey belong to Triassic igneous rocks; data points in shade of grey belong to Jurassic volcanic rocks

1998) was calculated for the samples showing concordant dates. Because of the greater quantity of ^{206}Pb compared to of ^{207}Pb , the $^{206}\text{Pb}/^{238}\text{U}$ ratios (and the corresponding dates) have smaller error than $^{207}\text{Pb}/^{235}\text{U}$ ratios (ESM Table S7). For this reason, only the $^{206}\text{Pb}/^{238}\text{U}$ dates are used to calculate the weighted average ages for the samples. The $^{206}\text{Pb}/^{238}\text{U}$ age dispersion is similar to the individual errors, which typically range from ± 1 to ± 7 Ma (ESM Table S7). This indicates good reproducibility of the analyses.

All selected zircons are idiomorphic and do not show pre-magmatic cores (ESM Fig. S3). No inherited zircons were observed in the different samples. Zircon crystals from biotite granites are generally 100–300 μm long and 50–150 μm wide. They show a core and well-developed euhedral concentric rims of magmatic growth (ESM Fig. S3a, b, c, f, g). Zircon crystals from volcanic rocks are generally 50–150 μm long and 50–100 μm wide. They present a core with limited rim zoning of magmatic growth (ESM Fig. S3d, e, h).

Carboniferous-Permian biotite granites give Concordia ages of 286 ± 2 Ma on sample E07 (five data points on three crystals giving five concordant dates; Fig. 5a), 304 ± 2 Ma on sample E08 (seven data points on four crystals giving seven concordant dates; Fig. 5b) and 313 ± 1 Ma on sample E09 (five data points on four crystals giving five concordant dates; Fig. 5c). Carboniferous-Permian volcanic rocks also give Concordia ages of 299 ± 1 Ma on sample E11 (five data points on four crystals giving five concordant dates; Fig. 5d) and 303 ± 1 Ma on sample E14 (four data points on three crystals giving four concordant dates; Fig. 5e). Both biotite granites and volcanic rocks were emplaced during a time span of 30 Ma during the Late Carboniferous-Early Permian. Triassic biotite granites give a date of 233 ± 12 Ma on sample ZK1-152.0m (four data points on four crystals giving two concordant dates; Fig. 5f) and a Concordia age of 233 ± 2 Ma on sample E01 (five data points on four crystals giving five concordant dates; Fig. 5g). They were emplaced during

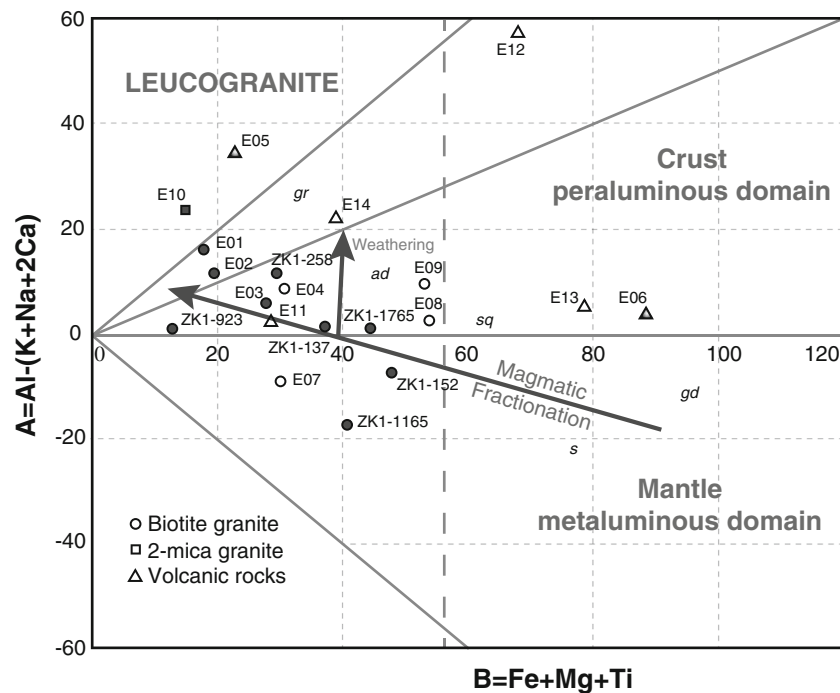


Fig. 4 A – B diagram showing the variations of the aluminous index (A parameter) relative to a differentiation index (B parameter) of basement igneous rocks surrounding the Erlian Basin (modified after Debon and Lefort 1983, 1988; see A and B values in ESM Table S6). A = chemical–mineralogical parameter (in millications) giving the excess of alumina not bound to the feldspars. B = chemical–mineralogical parameter (in

millications) used as a differentiation index for igneous rocks. *gr* granite, *ad* = adamellite, *sq* quartz syenite, *s* syenite, *gd* granodiorite. Data points in white belong to Carboniferous–Permian igneous rocks; data points in dark grey belong to Triassic igneous rocks; data points in shade of grey belong to Jurassic volcanic rocks

the Middle Triassic. Jurassic volcanic rocks were dated on sample E05 at 146 ± 19 Ma (four data points on four crystals giving two concordant dates; Fig. 5h) and thus have been emplaced during the initiation of the rifting stage of the Erlian Basin (i.e. pre-rift stage).

U–Th–Pb chemical dating on monazite

Twelve monazite crystals from the Triassic two-mica granite (sample E10) were dated with the EPMA (176 determinations). All the analytical results are reported in ESM Table S8. Monazites show a pronounced magmatic zoning highlighted by the distribution of their U, Th, Pb and Ce contents (Fig. 6a). The monazite crystals yield a chemical U–Th–Pb age of 243 ± 3 Ma, indicating that the two-mica granite emplaced during the Middle Triassic. In the diagram presented in Fig. 6b, the intercepts of the regression line with the U/Pb and Th/Pb axes give the U–Pb and Th–Pb ages ($325 + 61$ – 111 and $234 + 13$ – 11 Ma, respectively). Note that these ages are calculated on the regression line where the error is the largest. In addition, the dates obtained for all analyses give a cumulated frequency histogram with Gaussian distribution (Fig. 6b).

Some detrital monazites from post-rift sediments belonging to the Early Cretaceous Saihan and Late Cretaceous Erlian formations were also dated with the EPMA. The chemical

compositions of these monazites are presented in ESM Table S9 and the analytical results of the dating are reported in ESM Table S10. The most homogeneous spots were selected for the analyses. Two monazite grains give Carboniferous weighted average chemical U–Th–Pb dates (e.g. 322 ± 9 Ma; ESM Fig. S4b), three grains give Triassic dates (e.g. 238 ± 39 Ma; ESM Fig. S4d), two grains give Jurassic dates (e.g. 160 ± 8 ; ESM Fig. S4f) and one grain gives a Cretaceous date (104 ± 6 Ma; ESM Fig. S4h). Therefore, the monazites likely derive from the erosion of basement igneous rocks that were emplaced during these different time periods.

Sm–Nd isotopes

Eleven samples of Carboniferous–Permian and Triassic igneous rocks as well as Jurassic volcanic rocks from the basement of the Erlian Basin yield $^{143}\text{Nd}/^{144}\text{Nd}_i$ ratios ranging from 0.512241 to 0.512523. ϵNd_t values broadly range from -2.0 to 5.3 (Table 2 and Fig. 7). Two ϵNd_t signatures can be characterised (Table 2): (1) ϵNd_t shows positive values from 3.3 up to 5.3 for the Carboniferous–Permian igneous rocks that were analysed and (2) ϵNd_t shows negative values for the Triassic biotite granites (from -2.0 to -0.1) and a positive value for the Triassic two-mica granite (2.4; Table 2). Two populations of model ages for these igneous rocks were calculated (Table 2 and Fig. 7): (1) the

Carboniferous–Permian biotite granites and volcanic rocks yield T_{DM} model ages ranging from 485 to 726 Ma (mean = 612 Ma) and (2) the Triassic granites yield T_{DM} model ages ranging from 703 to 893 Ma (mean = 812 Ma). However, as T_{DM} model ages show values varying largely and significantly different from the emplacement ages of the studied igneous rocks, they most likely correspond to average ages reflecting variable mixtures of materials derived from the mantle at different times (Arndt and Goldstein 1987). Jurassic volcanic rocks show a slightly positive value of $\epsilon Nd(t)$ (0.8; Table 2) and a T_{DM} model age at 728 Ma.

Uranium leaching

The Carboniferous–Permian igneous rocks have low Th and U contents respectively ranging from 2.8 to 18.4 ppm and from 0.7 to 4.2 ppm (ESM Table S3). Their Th/U ratios range from 1.3 to 6.0 (Fig. 8 and ESM Table S3), which are close to the values of the mean crust (i.e. mean crust Th/U = 4; Cuney and Kyser 2008) and indicate that only a minor part of their uranium content has been leached out. Contrariwise, the Triassic igneous rocks show high Th and U contents (respectively 11.2–84.8 ppm and 2.0–21.3 ppm; Fig. 8 and ESM Table S3). Their Th/U ratios range from 2.1 to 19.4 (ESM Table S3). The high ratios (>4; Fig. 8) indicate that a significant part of the uranium has been leached out from the rocks.

As the U contents of the samples belonging to the Triassic granites are significantly higher than the U contents of the Carboniferous–Permian igneous rocks, our observations were focused on the U-bearing minerals occurring in the Triassic granites and especially on the non-weathered samples from the ZK1 drill hole. An EPMA profile on a U-thorite section (Fig. 9) shows that the core of the mineral has still preserved uranium contents higher than 1 wt% (although fresh U-thorite may have up to 30 wt% UO_2 when in equilibrium with uraninite; Cuney and Friedrich 1987), but the external envelope has uranium contents below 0.01 wt% UO_2 . The U-thorite presents a destruction rim within which uranium oxide has been identified and radiating fractures in the enclosing minerals. Therefore, the metamictisation of uranothorite crystals has led to the release of U from the mineral structure into their destruction rims and surrounding fractures and was made available for leaching.

The evolution of the Th and U contents and Th/U ratios of the U-rich Triassic biotite granite along the ZK1 drill cores is presented in Fig. 10. Two main facies can be differentiated: (1) from 0 to 105 m, the granite facies is highly fractionated (Th content > 50 ppm; Fig. 10) and shows a low ferromagnesian mineral content and high U concentrations in the non-weathered part (up to 21.3 ppm in sample ZK1-92.3m; Fig. 10 and ESM Table S3) and (2) from 105 to 180 m, the granite facies is less fractionated ($11.2 \text{ ppm} \leq \text{Th} \leq 30.4 \text{ ppm}$; Fig. 10 and ESM Table S3) and shows higher ferromagnesian

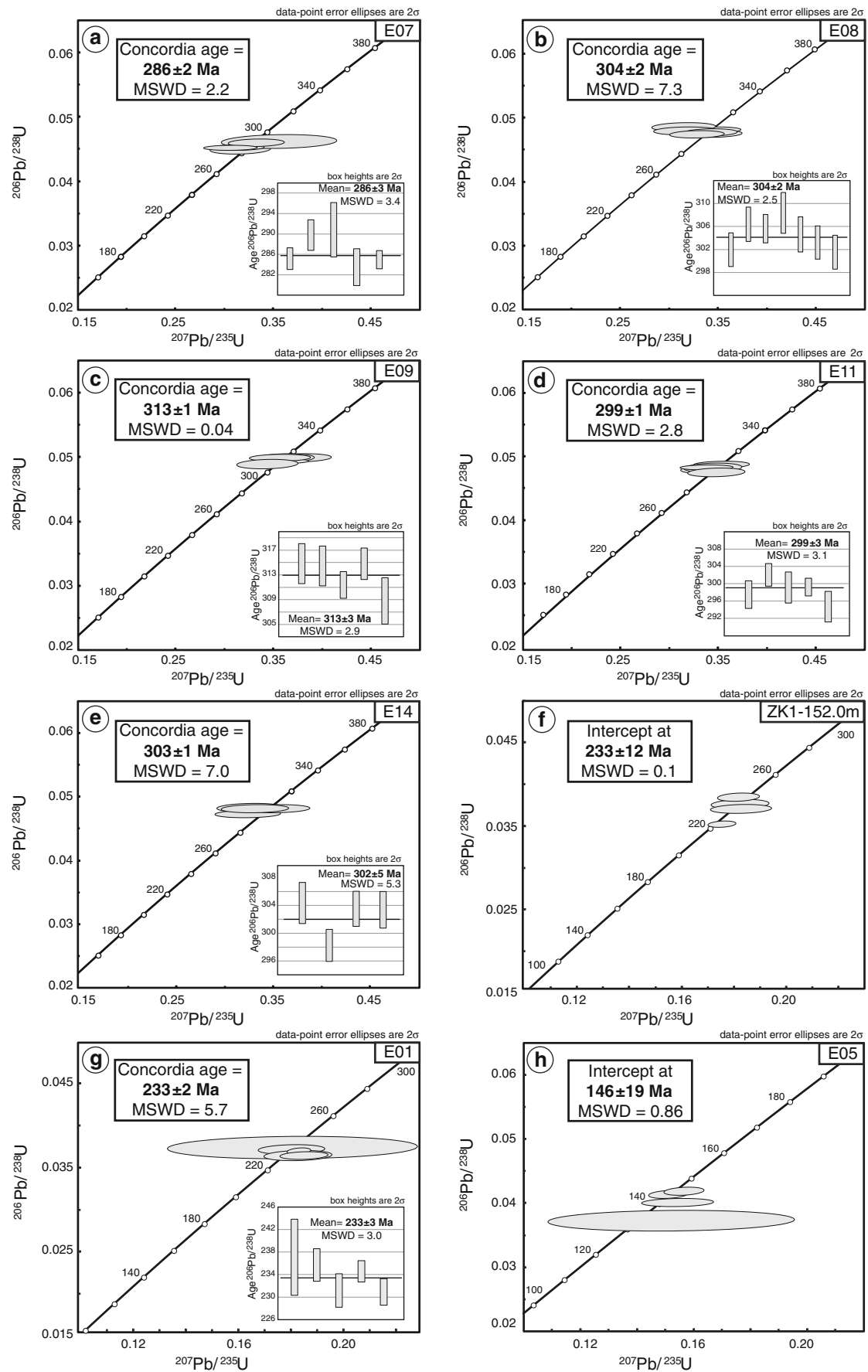
mineral content and moderate to high U concentrations ($3.4 \text{ ppm} \leq \text{U} \leq 13.5 \text{ ppm}$; Fig. 10 and ESM Table S3). In the non-weathered part of the granite (82–180 m), the U and Th contents are high ($3.4 \text{ ppm} \leq \text{U} \leq 21.3 \text{ ppm}$, $U_{\text{mean}} = 11.1 \text{ ppm}$; Fig. 10 and ESM Table S3), whilst Th/U ratios are lower than 4. Both major U-bearing minerals, U-thorite and allanite, are metamict and their uranium has been partly to totally released from the mineral structure, but preserved within the rock, as attested by the low Th/U ratios of the whole rocks. In the weathered part of the granite (0–82 m), the U contents are of 2–4 ppm U only, whereas the Th contents remain very high (Fig. 10), leading to high Th/U ratios (up to 19.4 for sample ZK1-25.8m; Fig. 10 and ESM Table S3); allanite and uranothorite are strongly altered, and the uranium has been nearly totally leached out from their mineral structure (Fig. 10) and from the rock, as attested by the very high Th/U ratios of the corresponding whole rocks.

Discussion

Carboniferous–Permian vs. Triassic igneous rocks as a major source for uranium

Both the Carboniferous–Permian and Triassic igneous rocks from the basement surrounding the Wulanchabu and Manite sub-basins are dominantly represented by strongly felsic rocks variably enriched in K, Th, U and other incompatible elements. The Carboniferous–Permian biotite granites and felsic volcanic rocks correspond to a low- to medium-K calc-alkaline magmatism (transitional I to A type), whereas the Triassic biotite granites belong to a high-K calc-alkaline magmatism (i.e. A2-type granites according to Eby 1992). The difference in the nature of the magmatism is also further supported by the Th/Yb vs. Ta/Yb diagram (ESM Fig. S5) and their accessory mineral assemblage. The Triassic biotite granites contain uranothorite along with allanite, titanite, zircon and apatite reflecting much higher Th and U contents than in the Carboniferous–Permian igneous rocks, which do not contain uranothorite. The Carboniferous–Permian and most Triassic igneous rocks are both derived from metaluminous magmas. The slightly peraluminous signature of some samples may result from either high degree of fractionation of Ca-rich pyroxene and amphibole (i.e. highly fractionated calc-alkaline granites), or from a small degree of peraluminous crustal material assimilation, or from incipient hydrothermal alteration of the feldspars, or a combination of these (Cuney and Kyser 2008). In northeast China, the Late Permian marks the final

Fig. 5 a–h Results of ion probe U–Pb dating on zircon from basement igneous rocks occurring at the margins of the Wulanchabu and Manite sub-basins. The results are represented in Concordia and weighted average diagrams



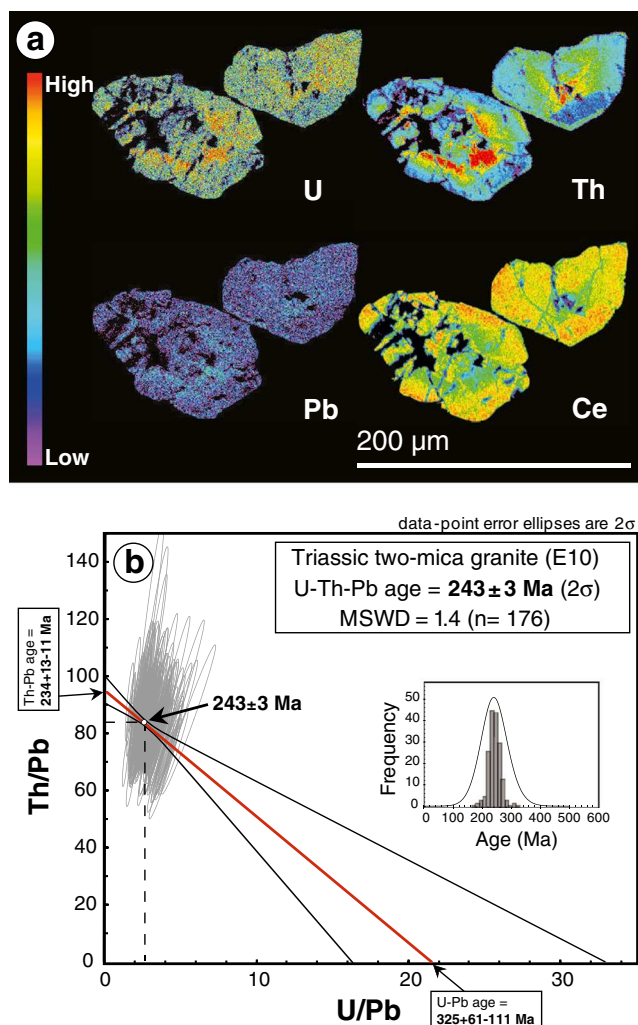


Fig. 6 U, Th, Pb and Ce mapping and U–Th–Pb isochron age of monazites from the Triassic two-mica granite (sample E10) occurring in the northern part of the Sunite Horst. **a** U, Th, Pb and Ce mapping performed by EPMA. **b** Th/Pb vs. U/Pb plot (ellipse errors are 2σ). U–Th–Pb age is calculated at the centroid (i.e. weighted mean point, represented by the *white dot*) of the best-fit line. The error is calculated on the hyperbolas, close to the centroid, where the dimension is the narrowest. The best regression line (*heavy red line*) has a slope indicating the theoretical isochron age. The error envelope (*black lines*) corresponds to the two symmetrical hyperbolas

accretion stage of magmatic arcs and micro-continents in the southeastern part of the CAOB (Xiao et al. 2003; Lin et al. 2008), which is followed by a collision during the Early–Middle Triassic (Li et al. 2013, 2014). The Late Carboniferous–Early Permian biotite granites and volcanic rocks from the study are characterised by positive εNd_t (3.3–5.3) and fairly young T_{DM} model ages (485–726 Ma), suggesting that they may dominantly derive from the partial melting of juvenile source rocks that have been emplaced earlier in the lower crust (De Paolo 1988; Peucat et al. 1988; Turpin et al. 1990; Jahn et al. 2000; Wu et al. 2000b). These rocks emplaced simultaneously with the Baolidao magmatic

arc that was formed within an oceanic subduction context (Xiao et al. 2003; Lin et al. 2008). However, their calc-alkaline signature could also be inherited from the melting of proto-arc material in this area (Li et al. 2015). To the opposite, the Middle Triassic biotite granites emplaced within a collisional setting (Li et al. 2013, 2014) are characterised by negative εNd_t (–2 to –0.1) and higher T_{DM} model ages (703–893 Ma). This suggests the contribution of Precambrian crustal material during magma generation (De Paolo 1988; Peucat et al. 1988; Turpin et al. 1990; Jahn et al. 2000; Wu et al. 2000), in accordance with their enrichment in incompatible elements such as K, Th and U. Further incompatible element enrichment has resulted from a high degree of crystal fractionation, giving the high-K calc-alkaline signature and the high enrichment in U of these granites. Therefore, the Carboniferous–Permian igneous rocks and especially the biotite granites probably represent a minor source of uranium for the deposits hosted in the sediments of the Erlian Basin because of their low U contents and Th/U ratios. Contrarily, the Triassic granites seem to constitute one of the major sources of uranium for the U deposits owing to their high U contents (up to 21.3 ppm) and their high Th/U ratios (up to 19.4), indicative of major uranium leaching from the weathered part of these granites. The implication of the Triassic granites as a uranium source is further corroborated by detrital U-thorite crystals within sediments of the Erlian Formation in the Erlian Basin (Bonnetti et al. 2015a).

Potential volcanic uranium source

Trachyte and rhyolite located at the base of the Erlian Basin were emplaced during the Late Jurassic (146 ± 19 Ma) and belong to the Qinganling Group (Chen and Chen 1997; Wang et al. 2006). These rocks are slightly to significantly peraluminous and most probably belong to a high-K calc-alkaline magmatic association, as indicated by Ren et al. (2002). These Late Jurassic volcanic rocks mark the initiation of the rifting stage in the Erlian Basin and other intracontinental basins of northeast China. In this region, Late Mesozoic volcanism is widely represented, especially in the Daxing'anling Ranges (Wang et al. 2006; Ying et al. 2010a, b) separating the Erlian Basin to the west from the Songliao Basin to the east. Although the U content of these rocks is not very high (3.0 and 6.5 ppm) and the U-bearing minerals are mainly non-altered monazite and xenotime, this volcanism may have produced tuffs and ashes that have been deposited as interlayers within the sediments of the Early Cretaceous formations in the Erlian Basin (e.g. in the Arshan and Tengge'er formations; Graham et al. 2001) and in sediments of other intracontinental basins in northeast China, until the end of the Early Cretaceous with successive volcanic events at 163–160, 147–140, 125–120 and 116–113 Ma in the Daxing'anling Ranges (Wang et al. 2006). Therefore, pyroclastic volcanic material could be

Table 2 Whole-rock Sm–Nd isotopic data for biotite granite, two-mica granite and volcanic rocks from the northern margins of the Wulanchabu and Manite sub-basins

Samples	Facies	Age (Ma)	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	Error ($\times 10^{-6}$)	$^{143}\text{Nd}/^{144}\text{Nd}_i$	$\varepsilon\text{Nd}_{(t)}$	T_{DM}
ZK1-25.8m	Biotite γ	—	4.6	29.7	0.094142	0.512383	5	0.512241	−2.0	872
ZK1-137.0m	Biotite γ	—	4.1	24.3	0.102934	0.512425	5	0.512270	−1.4	882
ZK1-152.0m	Biotite γ	233	3.3	25.6	0.077377	0.512448	5	0.512332	−0.2	703
ZK1-176.5m	Biotite γ	—	3.3	20.3	0.099082	0.512486	6	0.512336	−0.1	780
E02	Biotite γ	—	2.1	12.3	0.103638	0.512421	6	0.512265	−1.5	893
E04	Biotite γ	—	5.9	27.4	0.129951	0.512676	5	0.512421	3.3	726
E05	Rhyolite	146	4.3	22.8	0.113638	0.512597	6	0.512486	0.8	728
E07	Biotite γ	286	5.5	23.4	0.141605	0.512802	4	0.512523	5.3	593
E08	Biotite γ	304	2.1	14.3	0.090673	0.512697	6	0.512519	5.2	485
E10	2-mica γ	243	3.5	16.7	0.127057	0.512650	6	0.512451	2.4	745
E14	Rhyolite	303	8.2	32.6	0.151106	0.512814	6	0.512518	5.2	645

$^{143}\text{Nd}/^{144}\text{Nd}_i$ initial and $\varepsilon\text{Nd}_{(t)}$ were calculated using the equation of Michard et al. (1985), where $^{143}\text{Nd}/^{144}\text{Nd}_i$ corresponds to the calculated initial Nd isotope ratio of the given mantle reservoir from which a sample was separated; $\varepsilon\text{Nd}_{(t)}$ represents at time (t) the $^{143}\text{Nd}/^{144}\text{Nd}$ deviation from the CHUR composition; T_{DM} is a model age calculated relative to the depleted mantle corresponding to the theoretical time at which a sample separated from the given mantle reservoir

γ granite

present within the sandstones hosting the roll-front U deposits in the Erlian Basin (e.g. sandstones of the Aptian–Albian Saihan Formation) and would represent an additional source of uranium as U (Walton et al. 1981) and other elements such as Zr can be released from volcanic glass during

devitrification (Forbes et al. 1984) and leached out when oxidising groundwater percolate through the host sediments to form the roll-front uranium deposits. Such a contribution is supported by the presence of abundant volcanic fragments in the sandstones of the Saihan Formation reported by Nie et al. (2010) and the high Zr contents in coffinite from the Cenozoic Bayinwula deposit (up to 1.71 wt% ZrO_2 ; Bonnetti et al. 2015b). Nevertheless, more investigations in the sandstones, such as analyses of melt inclusions in detrital quartz, would be necessary to further support this hypothesis.

Uranium mass balance calculation for Triassic biotite granites

Mass balance calculation was carried out on drill core samples from the two main facies of Triassic biotite granites selected in the non-weathered section of the ZK1 drill hole (Th/ $U_{\text{mean}} = 2.6 < 4$) and containing the main U-bearing accessory minerals U-thorite and allanite (samples ZK1-92.3m, ZK1-116.5m, ZK1-137.0m and ZK1-152.0m). The mass balance calculation showing the relative contribution of the U-bearing accessory minerals (U-thorite, allanite and zircon; see EPMA results in ESM Table S1) to the whole-rock U content is presented in Table 3. It shows that zircon is a negligible host for uranium, which only accounts for about 1.0 ppm of the whole-rock uranium content and does not show any texture of metamictisation, and thus corresponds to a refractory uranium source. The present U amounts preserved in allanite and uranothorite from the whole rock are respectively 0.7 and 0.4 ppm. As the average U content in the non-weathered granite is 12.4 ppm, and no other significant U-bearing mineral

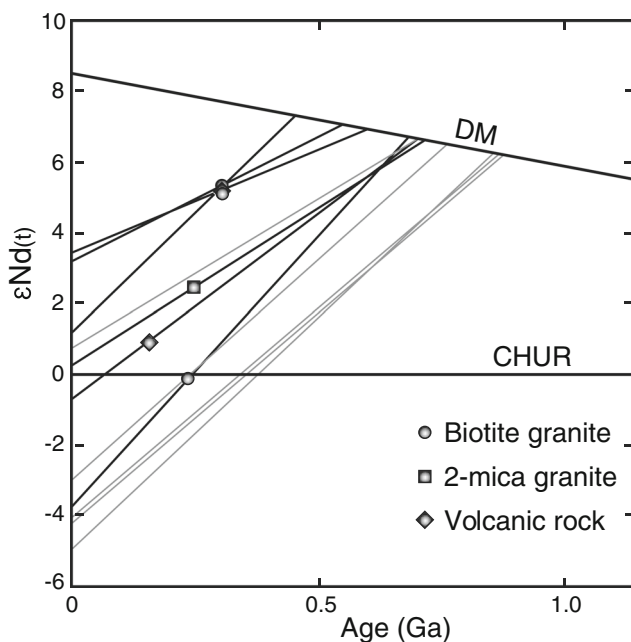
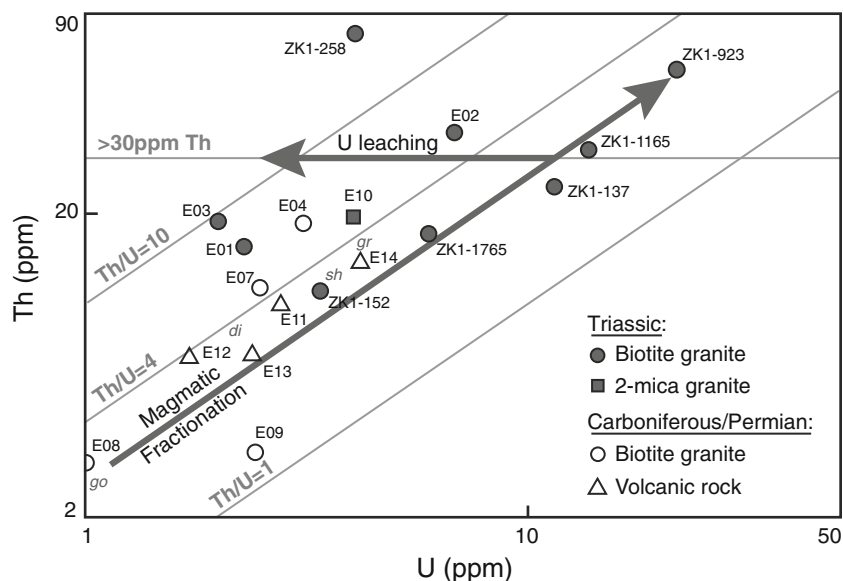


Fig. 7 $\varepsilon\text{Nd}_{(t)}$ evolution diagram for basement igneous rocks surrounding the Wulanchabu and Manite sub-basins. $\varepsilon\text{Nd}_{(t)}$ data points are plotted for samples with determined emplacement age. For other samples, the slope of the lines was calculated using the determined age of the rock unit they belong to. The depleted mantle (DM) model is from De Paolo (1988). CHUR chondritic uniform reservoir

Fig. 8 Th vs. U diagram showing the magmatic fractionation trend and the trend of U leaching for Carboniferous–Permian and Triassic igneous rocks occurring at the margins of the Wulanchabu and Manite sub-basins (modified after Debon and Lefort 1983, 1988). *gr* granite, *sh* shale, *di* diorite, *go* gabbro



phase has been identified, this means that the major part (90 %) of the uranium has been released from metamict accessory U-bearing minerals and is available for leaching. If we take as reference the maximum U content measured in the best-preserved U-thorite (11.7 wt% U), which is still below the possible initial U content of a non-metamict U-thorite (i.e. up to 28.0 wt% U; Cuney and Friedrich 1987), the calculated U content which was hosted by U-thorite in the whole rock is 5.2 ppm, and if we take the maximum U content which may be hosted in U-thorite, it would represent 12.3 ppm of the whole rock. Therefore, most of the uranium contained in the Triassic biotite granites was hosted in accessory minerals and

only a minor part of it may have been hosted in other minerals such as uraninite. However, no evidence of this last U mineral has been observed. Finally, if we consider 44 ppm of U-thorite at 11.7 wt% U as a minimum value occurring in the non-weathered Triassic biotite granites, a cubic kilometre of this granite can liberate $\sim 14,000$ t U/km³, and if we consider the maximum U content which can be hosted in U-thorite, a cubic kilometre of this granite can release up to $\sim 33,000$ t U/km³. Therefore, a significant proportion of uranium must have been leached by oxidising water during weathering, as attested by the high Th/U ratios $\gg 4$ in near-surface drill core and in outcrop samples.

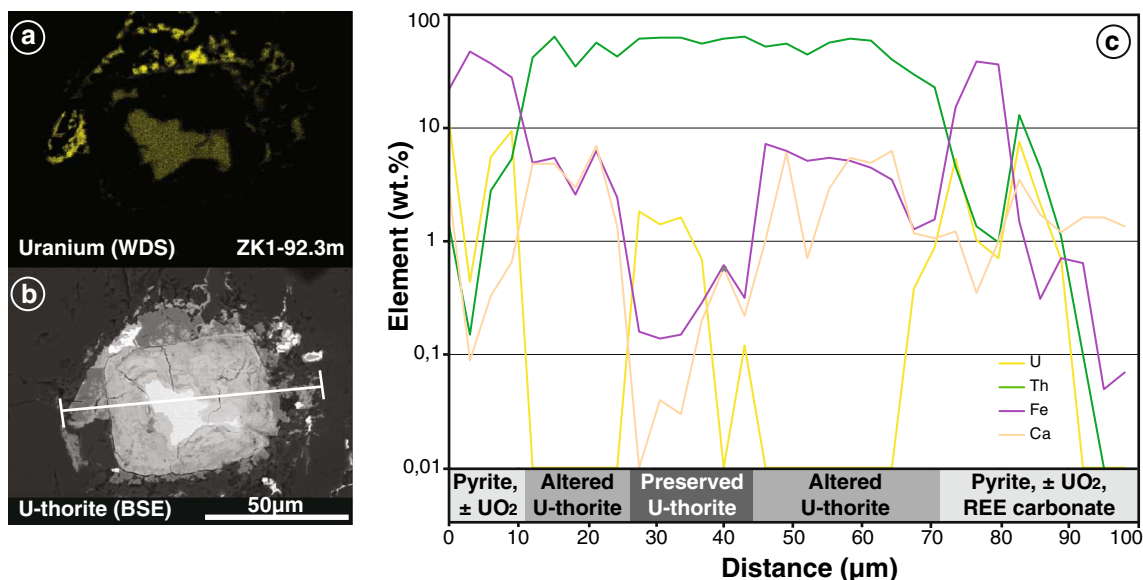


Fig. 9 EPMA profile on U-thorite from non-weathered Triassic biotite granite showing U release into the rock due to metamictisation. **a** U mapping on a metamict U-thorite section by SEM-WDS. **b** Photograph of metamict U-thorite section by SEM with the location of the EPMA

profile. **c** EPMA profile showing the variation of U, Th, Fe and Ca contents of the U-thorite in (b) and the mineral phases occurring in the destruction rim

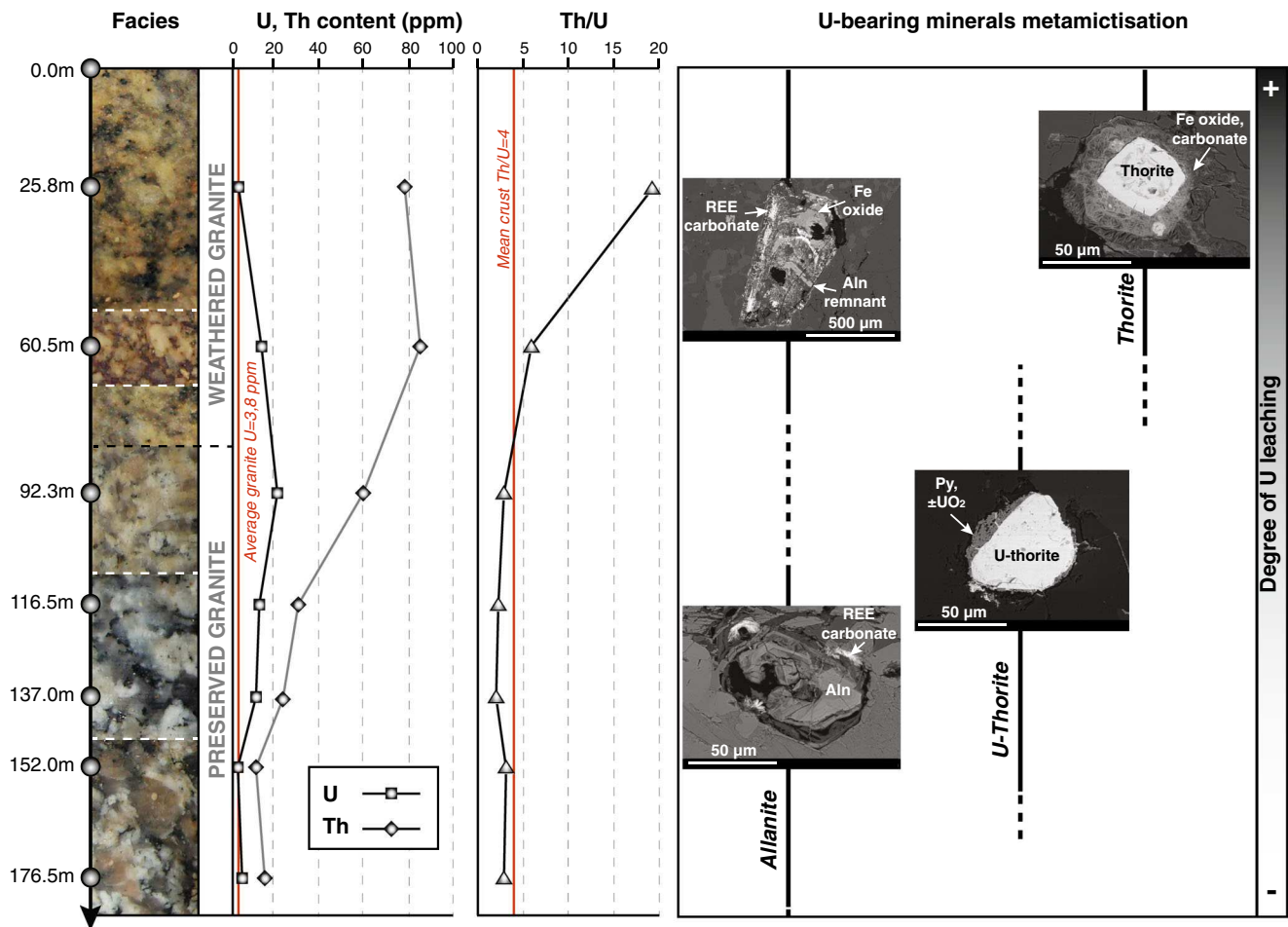


Fig. 10 Evolution of the granite facies, U and Th contents, Th/U ratios and the degree of leaching for major U-bearing accessory minerals according to depth of the ZK1 drill hole crossing a Triassic biotite granite located at the northern margin of the Wulanchabu Sub-basin.

Regional implications for sedimentary-hosted U deposits

In northeast China, several sandstone-hosted uranium deposits were formed within different Late Mesozoic sedimentary basins such as the Bayinwula deposit in the Erlian Basin (Bonnetti et al. 2015b), the Dongsheng deposit in the Ordos Basin (Cai et al. 2007), the Qianjiadian deposit in the Songliao Basin (Li et al. 2008) and other occurrences in the Tamusu area in the Yingen Basin (Wang 2009; He et al. 2010). The Carboniferous–Permian and Triassic igneous rocks constitute a major component of the basement of these basins (Wu et al. 2001, 2002; Zhang et al. 2010; Li et al. 2012). Triassic granites that represent a major uranium source for the mineralisation in the Erlian Basin also have been identified in the basement of the neighbouring basins (Orolmaa et al. 2008; Zhang et al. 2010; Li et al. 2012). A-type Jurassic and Cretaceous granites are also widely distributed in northeast China (Wu et al. 2000a, 2002, 2005b, c; Su et al. 2007) and in the basement of these sedimentary basins. Jurassic granites could represent an additional source of uranium if they contain U-bearing accessory

Photographs of U-bearing accessory minerals were taken on thin sections from samples ZK1-25.8m to ZK1-176.5m. *Aln* allanite, *Py* pyrite, *REE* rare earth elements

minerals such as allanite and U-thorite that may liberate their U during metamictisation. However, Cretaceous granites likely represent a more limited source of uranium as even though they may contain such U-bearing minerals, their U contents would be partially available for leaching because of a shorter time for the metamictisation process. The Late Jurassic–Early Cretaceous volcanic material from the Daxing'anling Ranges during the sedimentation of the Erlian Basin and other surrounding basins may have also contributed to significant U pre-concentrations within the sediments. In the Erlian Basin, the Late Palaeozoic to Early Cretaceous ages of detrital monazites from post-rift sediments indicate that these sediments are largely derived from the erosion of Carboniferous–Permian, Triassic and other igneous rocks from the basement surrounding the basin. This means that the U-rich Triassic biotite granites were outcropping in the source area, providing the material during the deposition of post-rift sediments in the Erlian Basin. As these post-rift sediments contain significant amounts of syndimentary to diagenetic U concentrations as background concentrations in the reduced sandstones (e.g. sandstones of the

Table 3 Mass balance calculation showing the relative contribution of the U-bearing accessory minerals to the whole-rock U content in non-weathered Triassic biotite granite

U-bearing accessory minerals	U content in mineral (ppm)	Element content in mineral (wt%)	Element content in whole rock (ppm)	Mineral content in whole rock (ppm)	U contribution in whole rock (ppm)
Zircon (refractory)	3400	Zr = 48.0	Zr = 148.0	308	1.0
Allanite (metamict)	900	La = 4.2	La = 34.0	809	0.7
U-thorite (metamict)	9000	Th = 57.0	Th = 25.0	44	0.4
U-thorite (max. value)	280,000	Th = 57.0	Th = 25.0	44	12.3
Total U content remaining in non-refractory accessory minerals					1.1
Whole-rock average U content					12.4
90 % of the U contents in non-refractory accessory minerals have been released					

Calculation was performed assuming that all La content of the non-weathered granite was initially contained in allanite and that all Th content (31.5 ppm) is contained in both minerals allanite and U-thorite. The Th amount contained in allanite being 6.5 ppm and the Th content in U-thorite represents 25.0 ppm of the whole rock. The maximum possible U content for U-thorite that was used is presented in Cuney and Friedrich (1987)

Saihan Formation; Bonnetti et al. 2015b) or even as synsedimentary–diagenetic mineralisation (e.g. as in the Nuheting deposit hosted in the Erlan Formation; Bonnetti et al. 2015a), this also means that the U-bearing minerals from the Triassic granites were metamict at the time of the sedimentation of the Saihan and Erlan formations and that it was therefore possible to mobilise their uranium towards the basin. Finally, epigenetic roll-front deposits formed during the Cenozoic (e.g. the Bayinwula deposit is 44 ± 5 Ma; Nie 2008), when oxidising meteoric waters, may have first leached part of the uranium directly from the basement source rocks, then infiltrated in permeable sandstones where they were able to remobilise part of the synsedimentary to diagenetic U concentrations. Therefore, the same source rocks may have provided uranium at two stages: (1) during the sedimentation of the basins and (2) when the meteoric oxidising waters infiltrated the permeable sandstones to form the epigenetic U deposits. As a comparison, strongly peraluminous two-mica granites were identified as the main source rocks for the formation of granite-related uranium deposits in the European Variscan Orogenic Belt (French Massif Central and Erzgebirge Massif in Germany; Leroy 1984; Cuney et al. 1989; Förster 1999; Tartèse et al. 2012; Ballouard et al. 2015). In these granites, uraninite occurred as one of the main U-bearing accessory minerals (Friedrich et al. 1987; Förster 1999) and was easily leached by percolation of oxidising fluids.

Conclusions

Carboniferous–Permian, Triassic and Jurassic igneous rocks are widespread in the CAOB and represent a major proportion of the basement around the Erlan Basin and other sedimentary basins in northeast China. Carboniferous–Permian granites and volcanic rocks correspond to a low- to medium-K calc-alkaline association (transitional I to A type), dominantly derived from the partial melting of juvenile source rocks and

emplaced during the Late Carboniferous–Early Permian. Triassic granites belong to a high-K calc-alkaline association (A2 type) and emplaced during the Middle Triassic. These granites most likely derived from the partial melting of juvenile source rocks with a significant contribution of Precambrian crustal material, explaining their enrichment in incompatible elements (especially U), which have been further enriched by a high degree of crystal fractionation. This enrichment in incompatible elements resulted in the crystallisation of U/Th/REE-rich accessory minerals such as U-thorite and allanite. These U-bearing minerals were metamict at the time of the sedimentation of the Saihan and Erlan formations and their uranium has been released from the mineral structures into microfractures of the rocks. This uranium has been subsequently leached during the erosion of these granites and groundwater percolation in the weathered sections. Mass balance calculations indicate that the U-rich Triassic biotite granites could have released from $\sim 14,000$ to $33,000$ t U/km³. An additional contribution may have been provided by felsic volcanism during sedimentation. Therefore, U-rich igneous rocks such as Triassic biotite granites in the basement around the Erlan Basin and other intracontinental sedimentary basins in northeast China represent a major source of uranium, explaining largely the U amounts trapped as (1) synsedimentary to diagenetic concentrations in the host sediments and as (2) epigenetic sandstone-hosted deposits, tabular or of roll-front type.

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