

Elemental and Sr–Nd–Pb isotopic geochemistry of Late Paleozoic volcanic rocks beneath the Junggar basin, NW China: Implications for the formation and evolution of the basin basement

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Abstract

The basement beneath the Junggar basin has been interpreted either as a micro-continent of Precambrian age or as a fragment of Paleozoic oceanic crust. Elemental and Sr–Nd–Pb isotopic compositions and zircon Pb–Pb ages of volcanic rocks from drill cores through the paleo-weathered crust show that the basement is composed mainly of late Paleozoic volcanic rock with minor shale and tuff. The volcanic rocks are mostly subalkaline with some minor low-K rocks in the western Kexia area. Some alkaline lavas occur in the central Luliang uplift and northeastern Wulungu depression. The lavas range in composition from basalts to rhyolites and fractional crystallization played an important role in magma evolution. Except for a few samples from Kexia, the basalts have low La/Nb (<1.4), typical for oceanic crust derived from asthenospheric melts. Zircon Pb–Pb ages indicate that the Kexia andesite, with a volcanic arc affinity, formed in the early Carboniferous (345 Ma), whereas the Luliang rhyolite and the Wucuiwan dacite, with syn-collisional to within-plate affinities, formed in the early Devonian (395 and 405 Ma, respectively). Positive $\epsilon_{\text{Nd}}(t)$ values (up to +7.4) and low initial $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios of the intermediate-silicic rocks suggest that the entire Junggar terrain may be underlain by oceanic crust, an interpretation consistent with the juvenile isotopic signatures of many granitoid plutons in other parts of the Central Asia Orogenic Belt. Variation in zircon ages for the silicic rocks, different Ba, P, Ti, Nb or Th anomalies in the mafic rocks, and variable Nb/Y and La/Nb ratios across the basin, suggest that the basement is compositionally heterogeneous. The heterogeneity is believed to reflect amalgamation of different oceanic blocks representing either different evolution stages within a single terrane or possibly derivation from different terranes.

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1. Introduction

The Central Asia Orogenic Belt (CAOB) lies between the Siberian Craton to the north and the North China Craton to the south and extends for over 2000 km across northern China (Fig. 1A). It is believed to be the site of successive accretion of arc complexes in the Phanerozoic times (Coleman, 1989; Zonenshain et al., 1990; Sengör et al., 1993), which were then intruded by voluminous Paleozoic and Mesozoic granitic-syenitic rocks (Jahn et al., 2000a;

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Kovalenko et al., 2004). These magmatic rocks were formed by a complicated combination of subduction-accretion processes and late-stage extension. Two gigantic magmatic belts can be recognized in the CAOB: a northern belt extending from central-northern Mongolia to Russian Transbaikalia and a southern belt extending from Kazakhstan through northern Xinjiang (NW China) and southern Mongolia into NE China. Large quantities of isotopic data have been published on rocks of the CAOB and used to interpret the processes of Phanerozoic juvenile continental growth (e.g., Han et al., 1997; Hu et al., 2000; Jahn et al., 2000a,b; Wu et al., 2000, 2002; Chen and Jahn, 2002; Chen and Arakawa, 2005).

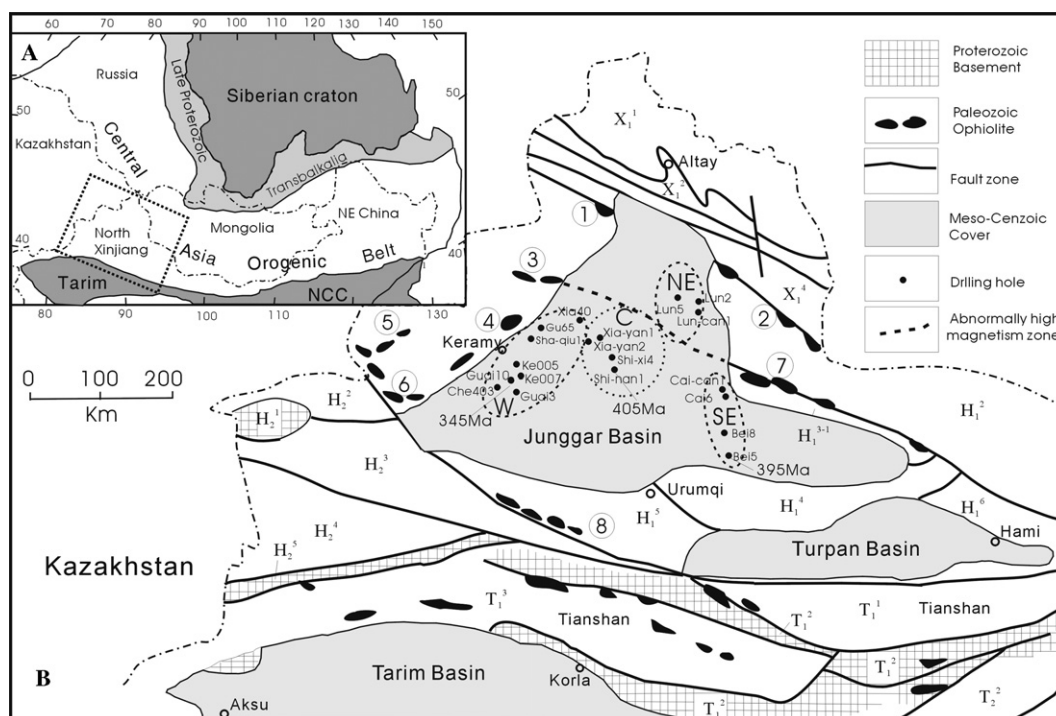


Fig. 1. Sampling locations and the ophiolite suites surrounding the Junggar basin in northern Xinjiang. (A) The locality of the study area (Junggar basin) in the Central Asia Orogenic Belt (modified after Jahn et al., 2000a). (B) Sampling locations within Junggar basin and the distribution of ophiolite suites surrounding the basin: ① Jaheba ophiolite belt; ② Armantai ophiolite belt; ③ Hongguleleng ophiolite belt; ④ Darbut ophiolite belt; ⑤ Mayilershan ophiolite belt; ⑥ Tangbale ophiolite belt; ⑦ Kalameili ophiolite belt; ⑧ Bayaangou-Dannahu ophiolite belt. The symbols are as follows: W, western Kexia area; C, central Luliang uplift; NE, the northeastern Wulungu depression; SE, southeastern Wucaiwan depressions.

The Junggar basin, a triangular feature of $\sim 200,000 \text{ km}^2$ in northern Xinjiang Province (Fig. 1B), is filled with about 10 km of continental sedimentary rocks, the oldest of which are early Permian (Xie et al., 1984; Coleman, 1989). The nature of the basement beneath the basin has long been controversial because of the paucity of stratigraphic, geochemical and geochronological data. Some workers considered the Junggar basement to represent a micro-continent of Precambrian age (e.g., Ren and Jiang, 1980; Zhang et al., 1984; Wang, 1986; Watson et al., 1987; Wu, 1987), whereas others interpreted it as a fragment of Paleozoic oceanic crust (e.g., Li et al., 1982; Jiang, 1984; Hsü, 1989; Feng et al., 1989; Chen and Jahn, 2004; Chen and Arakawa, 2005). Xiao et al. (1992) suggested that the northern Xinjiang region was formed by the convergence of three principal plates (Siberia, Kazakhstan and Tarim): the Junggar basin and Tianshan Mountains were considered to be part of the Kazakhstan plate, whereas the Altay Mountains in northernmost Xinjiang were considered to mark the southern part of the Siberian plate.

Volcanic rocks obtained from boreholes through the paleo-weathered crust of the Junggar basin in the southern belt of the CAO (see Fig. 1A) offer an ideal opportunity to probe the nature of the basement beneath the basin. In this paper, elemental and Sr–Nd–Pb isotopic geochemistry and zircon Pb–Pb ages of drill core samples are used to investigate the composition and tectonic setting of the basement rocks and thus shed light on the formation and

evolution of the basin. The boreholes are located in the western Kexia area, the central Luliang uplift, and in the northeastern Wulungu and southeastern Wucaiwan depressions. The locations are abbreviated as “Kexia” (W), “Luliang” (C), “Wulungu” (NE) and “Wucaiwan” (SE) (Fig. 1B).

2. General geologic setting

The Junggar basin is filled with thick continental sedimentary rocks not older than early Permian (XBGM, 1993) and surrounded by a number of Paleozoic ophiolite belts (Fig. 1). A variety of granitic rocks, including M-, A- and I-types, have been found around the basin (Xiao et al., 1992; Tu, 1993). Early published data suggested that most of the granitic rocks were intruded between 320 and 350 Ma (Lu and Liu, 1989; Zhou, 1989; Shen and Jin, 1993) with a few at $\sim 400 \text{ Ma}$ (Hu et al., 1997). However, recent SHRIMP results suggest that the granitoids from the Junggar terrain mostly intruded in a post-collisional, extensional regime between 300 and 280 Ma (e.g., Han et al., 1997; Chen and Jahn, 2004; Chen and Arakawa, 2005).

Two NW-trending ophiolite belts, the Armantai-Jaheba belt (① and ② in Fig. 1B) and the Hongguleleng-Kalameili belt (③ and ⑦), are present in the northeastern part of the basin. The Armantai-Jaheba belt extends eastward into Mongolia, and is composed of metamorphic peridotite,

ultramafic–mafic cumulates and intermediate-mafic volcanic rocks. A Sm–Nd whole-rock isochron age of 561 ± 41 Ma and an $\varepsilon_{\text{Nd}}(t)$ of +6.1 was obtained for the gabbros, diabases, andesitic porphyrites and plagioclase (Huang et al., 1997). The Hongguleleng–Kalameili ophiolite belt consists of metamorphic peridotite, ultramafic–mafic cumulates, and mafic volcanic rocks. The basalts yield a Sm–Nd whole-rock isochron age of 444 ± 27 Ma, with an $\varepsilon_{\text{Nd}}(t)$ of +7.2 (Xiao et al., 1992; Zhang and Huang, 1992).

Two NE-trending ophiolite belts, the Mayilashan and Darbut ophiolite belts (④ and ⑤), are located in the western part of the basin. Radiolarian fossils found in the Mayilashan belt (Wu, 1987; XBGM, 1993) have a Silurian age. A Sm–Nd isochron age of 395 ± 12 Ma and an $\varepsilon_{\text{Nd}}(t)$ of +8.9 have been obtained for the gabbros from the Darbut belt (Zhang and Huang, 1992).

Two WNW-trending ophiolite belts, the Tangbale and the Bayaangou–Danahu (⑥ and ⑧), occur in the western and southern parts of the basin. Sm–Nd isochron ages of 489 ± 53 Ma with an $\varepsilon_{\text{Nd}}(t)$ of +5.8 and 447 ± 56 Ma with an $\varepsilon_{\text{Nd}}(t)$ of +7.5 were obtained for Tangbale gabbros and basalts, respectively (Zhang and Huang, 1992). The Bayaangou–Danahu ophiolite belt, a nearly complete ophiolitic assemblage in the Tianshan area, is considered to have formed in the Late Paleozoic based on radiolarian and Condontophorida fossils (XBGM, 1993).

3. Petrography

The recovered core samples (Fig. 2A) indicate that the basement beneath the Junggar basin is mainly volcanic rock with minor shale and tuff. The volcanic rocks are basalt + basaltic andesite (basaltic trachyandesite) + andesite (trachyte) in the western Kexia area, basalt (trachybasalt) + rhyolite (bimodal?) in the central Luliang uplift, (basaltic trachyandesite) + andesite + rhyolite in the northeastern Wulungu depression, and basalt + basaltic andesite (basaltic trachyandesite) + (trachyandesite) + dacite in the southeastern Wucaiwan depression.

The basalts are mostly porphyritic with phenocrysts of clinopyroxene (3–4 mm) and plagioclase (3–5 mm) set in a groundmass of fine-grained plagioclase (0.5–1 mm), clinopyroxene (~1 mm), and opaque minerals. Interstitial glass is common in the Kexia samples, but rare in those from Luliang and Wucaiwan (Fig. 2B and C), which generally contain olivine phenocrysts (3–5 mm). The trachybasalts are also porphyritic with phenocrysts of plagioclase, minor olivine and clinopyroxene. The groundmass is composed of fine-grained K-feldspar, plagioclase, clinopyroxene and glass.

The basaltic andesites in Kexia and Wucaiwan are also porphyritic with variable proportions of plagioclase (2–4 mm) and clinopyroxene (2.5–3 mm) phenocrysts. The groundmass is composed of fine-grained plagioclase (0.3–0.7 mm) and glass. Basaltic trachyandesites in Kexia, Wulungu and Wucaiwan are weakly porphyritic with sparse phenocrysts of hornblende (1–2 mm) and clinopy-

roxene (2–3 mm) in a matrix of fine-grained K-feldspar, clinopyroxene, hornblende and minor opaque oxides.

All of the andesites in the basin are strongly phyric with 8–12% of plagioclase phenocrysts set in a groundmass of fine-grained plagioclase (~0.5 mm) and glass (Fig. 2D). Compositional trachyandesites, which are found only in Wucaiwan, are weakly phyric with small phenocrysts of hornblende (2–3 mm), plagioclase (2–4 mm), minor K-feldspar (1–2 mm) and quartz (~1 mm). The groundmass consists of fine-grained (0.2–0.5 mm) amphibole, feldspar and glass, with sparse opaque minerals.

Dacites are present only in Wucaiwan. They are invariably fine-grained glassy rocks with sparse plagioclase and quartz phenocrysts (2–3 mm). Minor fine-grained quartz and plagioclase and/or K-feldspar (0.1–0.3 mm) are also present in the glassy groundmass. Compositional trachytes, found only in Kexia, are massive, nearly aphyric rocks with rare phenocrysts of K-feldspar (1–3 mm) and quartz (~1 mm) set in a matrix of feldspar minor glass and opaque minerals (Fig. 2E).

Rhyolites occur in Kexia, Luliang and Wulungu. Most are moderately phyric rocks with phenocrysts of sanidine, quartz and plagioclase (3–5 mm), set in a glassy matrix containing minor very fine-grained quartz and feldspar (<0.2 mm). Some of the rhyolites in Wulungu contain large, nearly euhedral plagioclase phenocrysts (Fig. 2F).

4. Analytical methods

Zircons were separated using standard methods and five zircon fractions were hand-picked from the least magnetic zircon population for each sample. We used the method developed by Kober (1986) involving repeated evaporation and precipitation of Pb from chemically untreated single zircons in a double-filament arrangement (Kober, 1987). Isotopic measurements were carried out on the MAT-261 mass spectrometer at China University of Geosciences (Wuhan). Detailed procedures are described in Wang (1992). The calculated ages and uncertainties are based on the means of all ratios evaluated and their 2σ mean errors. The results are listed in Table 1.

Fresh rock chips were selected from the core samples and crushed to about 1 mm. A split was then ground in an agate mill and passed through a 160 mesh sieve for geochemical and isotopic analyses. All analyses (Table 2) were also carried out at China University of Geosciences (Wuhan). Major element compositions were determined by wet chemical methods. The analytical errors for major oxides are less than 2 wt%. For trace element analyses, samples (ca. 50 mg) were weighed, dissolved in distilled HF + HNO₃ (3:1) in Savillex Teflon screw-cap capsules at 100 °C for 2 days, dried and then digested with 6 M HCl at 150 °C. A blank solution was prepared and the total procedural blanks were <50 mg for all the analysed trace elements. Three duplicates of three international standards (AGV-1, GSR-3 and DNC-1) were prepared using the same procedure to monitor analytical precision. The solu-

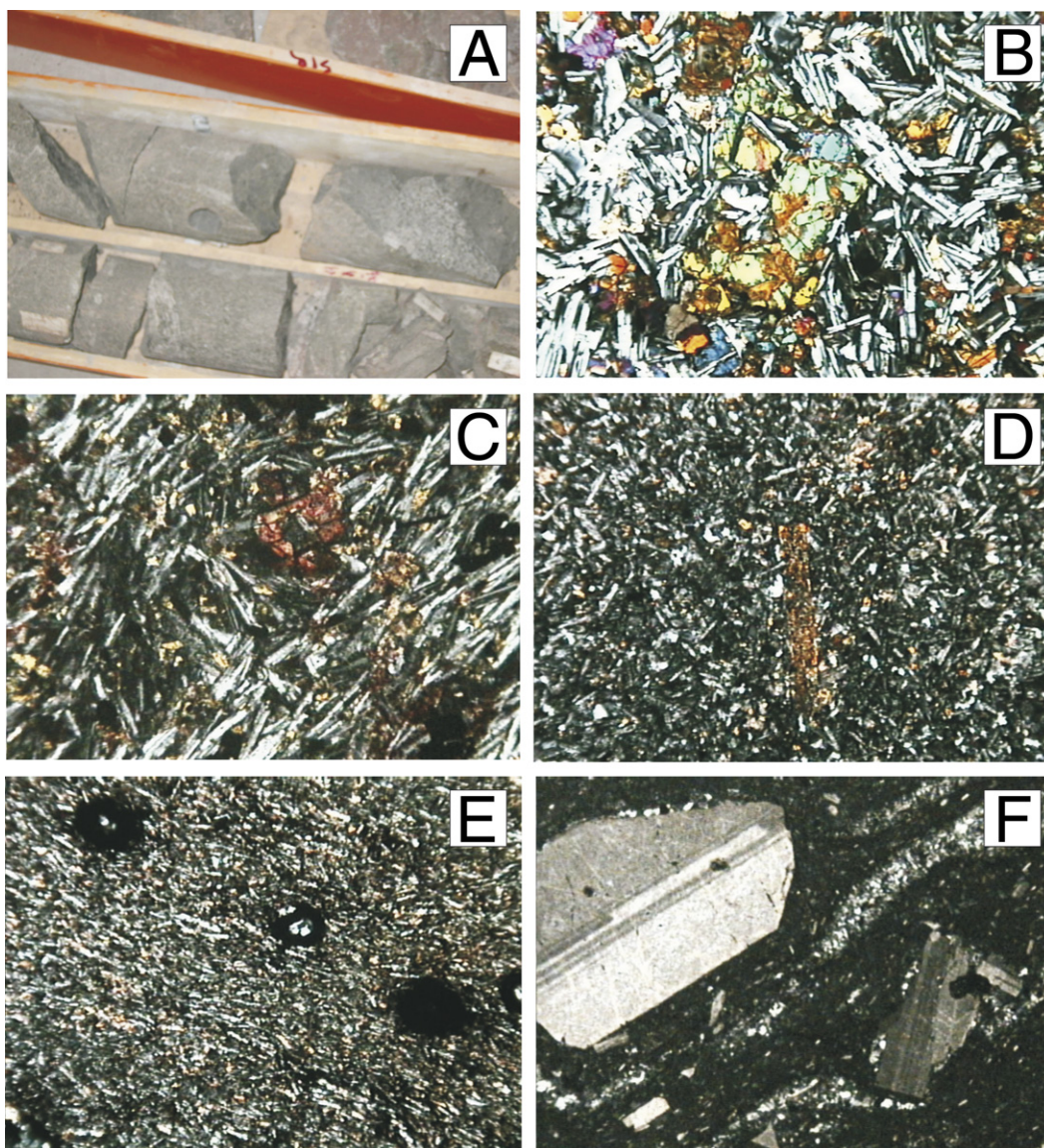


Fig. 2. Sampling and petrography of the late Paleozoic volcanic rocks. (A) Basalts from 2134 to 2138 m depth in Borehole Cai-can 1; (B) Porphyritic texture with clinopyroxene phenocrysts and groundmass minerals of fine-grained plagioclase, clinopyroxene and opaque minerals in basalt (sample k44); (C) Porphyritic texture with clinopyroxene phenocrysts and groundmass minerals of fine-grained plagioclase, clinopyroxene, opaque minerals and glass in basaltic andesite (sample k211); (D) Aphyric texture with a groundmass of fine-grained plagioclase and glass in andesite (sample k147); (E) Groundmass of predominantly feldspar, minor glass and opaque oxides in trachyte (sample k95); (F) Rhyolite with plagioclase phenocrysts and a groundmass of glass and minor very fine-grained quartz and plagioclase/K-feldspar (sample k20).

Table 1
 $^{207}\text{Pb}/^{206}\text{Pb}$ apparent ages of zircons in the volcanic rocks from Junggar basin basement

Sample	Hole no.	Depth (m)	Rock	Character	Grains	$^{207}\text{Pb}/^{206}\text{Pb}$	± 2	t (Ma)
k147	Guai 10	4933	Andesite	C, T, E, Lp	5	0.053181	0.000048	345 ± 20
k20	Shi-nan 1	3698	Rhyolite	C, T, E, Lp	5	0.054664	0.000022	405 ± 10
k32	Bei 5	3500	Dacite	C, T, E, Lp	5	0.054401	0.000018	395 ± 08

C, colorless; T, transparent; E, Euhedral; Lp, long prismatic shape.

tions were analysed using a POEMS-III ICP mass spectrometer (ICP-MS). The discrepancy among triplicates is less than 10% for all elements, and the analyses of standards are in agreement with the recommended values. The analytical procedures followed Gao et al. (1999).

Sr–Nd–Pb isotopic data were obtained using a Finnegan MAT-261 mass spectrometer. The Sr and Nd isotopic ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. The La Jolla standard yielded $^{143}\text{Nd}/^{144}\text{Nd} = 0.511846 \pm 10$ and NBS987 gave

Table 2
Elemental and Sr–Nd–Pb isotopic compositions of Carboniferous volcanic rocks beneath Junggar basin, NW China

Sample	k91	k94	k95	k97	k100	k103	k113	k146	k147	k148	k173	k178	k204	k209	k10	k16	k20	k23	k25	k29	k211
Area	Kexia														Luliang						
Hole no.	Ke005	Ke007			Guai3	Guai3	Che403	Guai10			Gu65	Gu65	Sha-qiu1	Xia40	Xia-yna2	Shi-xi4	Shi-nan1				
Depth (m)	2798	3077	3073	3070	4933	4936	1223	3907	3905	3902	780	713	3081	4980	5337	4605	3698	3601	3508	3336	4933
Rock type	BA	T	T	T	A	A	A	R	A	BTA	B	B	B	BTA	TB	R	R	R	B	TB	B
SiO ₂ (wt%)	56.6	65.2	64.8	65.0	60.0	62.4	62.0	71.2	62.2	47.9	47.8	48.4	50.4	47.3	46.7	71.5	74.7	75.9	43.4	47.5	43.6
TiO ₂	1.47	0.82	0.84	0.81	1.23	0.43	0.72	0.81	0.90	1.89	0.84	0.94	1.11	2.01	2.25	0.24	0.22	0.23	2.03	2.15	1.61
Al ₂ O ₃	13.2	15.1	15.0	15.1	12.2	14.4	14.9	12.9	14.2	15.6	18.0	19.4	18.4	16.2	14.9	14.2	11.0	10.9	15.6	15.0	14.4
Fe ₂ O ₃	2.04	1.86	1.76	1.60	7.73	3.11	1.83	0.99	4.74	4.60	5.95	4.68	5.04	7.81	7.64	0.50	2.17	2.32	8.22	7.21	9.40
FeO	5.73	2.78	3.30	3.06	0.49	0.47	3.21	2.65	0.35	5.39	2.92	4.30	3.31	4.51	3.49	1.70	0.58	0.75	3.82	5.12	1.25
MgO	2.80	1.17	1.06	1.16	3.24	1.40	1.88	1.25	0.71	3.23	4.67	3.70	2.97	6.27	5.98	0.60	0.17	0.23	7.13	6.40	5.04
CaO	5.54	1.47	1.32	1.28	6.06	7.41	2.02	1.30	3.50	6.62	8.46	10.2	7.56	5.12	7.59	0.50	1.76	0.46	8.43	5.40	10.4
Na ₂ O	4.99	6.32	6.47	6.86	4.28	3.39	3.24	5.22	5.18	4.78	3.34	2.58	3.20	5.26	4.64	3.58	2.75	2.08	3.48	4.62	3.72
K ₂ O	0.18	1.90	1.53	1.13	1.09	1.11	2.96	1.29	1.37	1.48	0.85	0.55	0.81	0.56	0.52	4.89	5.61	6.40	0.76	0.87	0.32
P ₂ O ₅	0.56	0.25	0.26	0.33	0.38	0.10	0.29	0.19	0.30	0.44	0.10	0.12	0.27	0.31	0.60	0.01	0.01	0.02	0.38	0.39	0.48
MnO	0.25	0.17	0.18	0.18	0.11	0.04	0.06	0.16	0.09	0.30	0.11	0.11	0.11	0.15	0.16	0.07	0.06	0.06	0.17	0.18	0.15
H ₂ O ⁺	3.54	1.91	1.76	1.71	2.60	4.51	4.17	1.75	2.31	4.23	6.05	3.46	5.85	3.92	3.73	1.64	0.62	0.72	4.40	4.24	5.38
H ₂ O [−]	0.24	0.14	0.19	0.12	0.22	1.00	1.62	0.24	0.30	0.80	0.78	1.06	0.36	0.50	0.58	0.21	0.06	0.08	0.78	0.64	0.78
CO ₂	1.74	0.21	1.02	0.89	0.10	0.07	0.71		3.69	2.31			0.33		0.79				1.02		3.05
Total	98.9	99.2	99.5	99.2	99.8	99.9	99.6	100.0	99.8	99.5	99.9	99.4	99.7	99.8	99.6		99.6	99.7	100.2	99.6	99.7
Mg [#]	0.43	0.37	0.32	0.35	0.58	0.56	0.45	0.42	0.33	0.43	0.59	0.51	0.48	0.58	0.60	0.36	0.16	0.18	0.62	0.57	0.61
Ni (ppm)	2.00	3.22	2.01	1.98	45.7	20.5	37.2	2.03	2.70	2.90	22.4	8.80	5.02	93.5	111	3.50	1.80	1.99	81.3	72.5	90.0
Co	15.2	17.8	47.8	28.3	44.3	24.0	21.3	19.8	22.1	21.9	59.0	68.2	59.0	118	74.1	7.00	5.80	3.10	79.3	75.1	85.4
Cr	19.6	14.1	14.1	16.5	8.50	27.5	52.2	21.3	18.7	23.3	76.4	49.8	26.7	50.5	220	21.5	12.2	8.40	113	90.2	186
Cu	10.4	8.2	10.3	8.10	59.4	10.4	41.7	1.40	11.5	14.9	57.8	96.8	2.00	27.0	49.2	7.90	3.00	1.00	20.8	16.8	34.6
Zr	211	161	98.5	119	115	315	182	203	213	205	59.2	68.4	160	186	197	201	718	602	165	175	165
Hf	6.44	5.02	4.71	4.35	3.62	8.23	5.12	6.32	6.57	6.30	1.51	1.92	3.04	4.43	6.32	6.85	17.9	18.4	4.90	5.21	4.24
Rb	29.9	33.0	55.1	22.0	3.11	10.0	86.0	25.3	19.0	10.0	28.1	14.0	21.2	10.3	17.0	192	196	171	20.0	27.0	18.0
Sr	183	259	25.0	154	127	86.0	241	346	327	339	331	484	70.0	2154	279	140	60.0	54.8	399	425	260
Ba	386	601	451	106	110	74.8	468	1678	128	1203	130	316	522	238	162	1025	75.6	73.3	242	159	121
Nb	9.00	7.10	3.90	3.90	6.30	8.20	10.1	7.20	7.20	7.20	2.70	2.00	1.00	2.00	14.0	18.4	34.1	33.9	8.60	9.40	7.80
Ta	0.20	0.31	0.46	0.39	0.20	0.20	0.24	0.20	0.64	0.63	0.20	0.54	0.56	0.25	0.36	0.25	0.20	0.20	0.63	0.49	0.28
Th	0.49	0.83	0.25	0.27	0.54	0.73	1.63	0.45	0.58	0.47	0.20	0.21	0.23	0.19	0.24	2.26	1.07	0.93	0.19	0.23	0.81
U	2.32	2.73	0.60	0.70	1.12	4.40	3.71	3.11	2.50	3.40	0.51	0.90	1.32	0.63	0.90	13.2	10.3	10.0	0.50	1.50	1.20
Y	28.5	34.8	36.1	35.1	20.9	40.2	22.8	37.2	29.3	28.3	10.8	12.4	18.2	26.7	33.0	31.6	72.6	70.2	26.8	26.4	25.1
La (ppm)	8.21	16.9	15.0	16.1	7.92	11.3	11.9	10.7	10.5	7.22	2.91	2.80	10.5	6.96	17.6	22.6	49.3	45.8	9.77	9.72	12.5
Ce	22.2	7.99	37.6	36.5	21.6	42.4	41.5	29.6	28.5	19.0	7.54	7.84	25.0	20.8	42.6	53.0	116	111	27.6	26.9	33.0
Pr	5.22	5.36	5.42	5.36	3.17	4.86	5.52	4.51	3.96	2.85	1.20	1.29	3.37	3.19	6.16	6.48	15.8	14.2	3.82	3.96	4.61
Nd	17.2	24.0	23.6	23.8	14.3	21.4	22.2	22.0	18.4	15.3	5.79	6.45	16.0	16.4	28.5	24.9	66.6	59.1	19.5	19.2	21.1
Sm	4.72	5.71	5.72	5.81	3.65	5.68	4.76	5.67	4.53	4.36	1.63	1.93	3.81	4.37	6.39	5.30	13.7	13.0	4.64	4.63	5.02
Eu	1.50	1.59	1.62	1.56	1.12	0.94	1.03	1.66	1.40	1.52	0.59	0.66	1.12	1.53	1.92	0.59	0.62	0.53	1.60	1.64	1.52
Gd	5.10	5.71	5.73	5.97	3.79	5.62	4.17	5.97	4.61	4.85	1.80	2.05	3.65	4.82	6.36	4.93	12.8	12.9	4.86	4.92	4.78
Tb	0.79	0.95	0.96	0.94	0.60	0.97	0.64	1.00	0.74	0.78	0.32	0.36	0.54	0.78	0.99	0.80	2.08	2.11	0.80	0.82	0.76
Dy	5.18	6.06	6.28	6.24	3.76	6.67	3.92	6.55	4.93	4.96	1.97	2.26	3.43	5.02	6.23	5.43	13.9	13.8	4.99	5.03	4.53
Ho	1.02	1.24	1.32	1.27	0.75	1.38	0.79	1.33	1.02	1.01	0.40	0.46	0.65	0.99	1.21	1.06	2.71	2.69	0.97	0.98	0.92
Er	2.92	3.69	3.95	3.71	2.11	4.23	2.36	3.87	3.10	2.93	1.19	1.33	1.86	2.78	3.34	3.07	7.89	7.41	2.59	2.73	2.49
Tm	0.42	0.55	0.60	0.57	0.33	0.67	0.36	0.59	0.48	0.45	0.18	0.20	0.27	0.41	0.49	0.51	1.24	1.12	0.41	0.42	0.37
Yb	2.62	3.34	3.81	3.56	1.95	4.47	2.36	3.70	2.99	2.59	1.05	1.26	1.66	2.46	2.82	3.30	8.22	6.81	2.41	2.46	2.20
Lu	0.40	0.52	0.60	0.54	0.30	0.70	0.39	0.61	0.48	0.41	0.17	0.20	0.25	0.38	0.48	0.56	1.33	1.09	0.39	0.41	0.35

$^{87}\text{Rb}/^{86}\text{Sr}$	0.0229	0.1714	0.0831	0.2996	0.2782	0.1222	0.0222	0.0406	0.0061	0.0740	3.0552	7.2926	0.0765	0.1075
$^{87}\text{Sr}/^{86}\text{Sr}$	0.703977	0.704695	0.705033	0.704759	0.705058	0.704309	0.703824	0.703919	0.705054	0.704742	0.714046	0.723882	0.704876	0.705530
$\pm 2\sigma$	0.000037	0.000044	0.000044	0.000033	0.000028	0.000050	0.000038	0.000032	0.000029	0.000038	0.000043	0.000032	0.000045	0.000023
$^{147}\text{Sm}/^{144}\text{Nd}$	0.17565	0.15059	0.15667	0.16639	0.17739	0.17852	0.18045	0.15141	0.17280	0.15384	0.13950	0.14692	0.16054	0.16013
$^{143}\text{Nd}/^{144}\text{Nd}$	0.513017	0.512967	0.512931	0.512993	0.513017	0.512967	0.512983	0.512882	0.513075	0.512916	0.512806	0.512871	0.512992	0.512990
$\pm 2\sigma$	0.000005	0.000012	0.000016	0.000007	0.000007	0.000016	0.000008	0.000008	0.000011	0.000007	0.000007	0.000006	0.000009	0.000010
T_{DM} (Ga)	0.53	0.44	0.59	0.51	0.56	0.79	0.77	0.66	0.28	0.60	0.71	0.64	0.45	0.46
T_{DM2} (Ga)	0.42	0.41	0.49	0.43	0.43	0.52	0.50	0.55	0.32	0.53	0.64	0.57	0.43	0.43
$\varepsilon_{\text{Nd}}(t)$	7.4	6.4	5.7	6.9	7.4	6.4	6.7	4.8	8.5	5.4	3.3	4.5	6.9	6.9
$f_{\text{Sm}/\text{Nd}}$	−0.11	−0.23	−0.20	−0.15	−0.10	−0.09	−0.08	−0.23	−0.12	−0.22	−0.29	−0.25	−0.18	−0.19
$\varepsilon_{\text{Nd}}(0)$	8.3	8.5	7.5	8.3	8.3	7.2	7.5	6.8	9.6	7.6	6.2	7.1	8.8	8.7
$^{206}\text{Pb}/^{204}\text{Pb}$	18.077	18.203	18.309	18.220	17.958	18.056	18.144	18.269	17.985	18.251	18.488	18.158	18.200	18.210
$\pm 2\sigma$	0.004	0.004	0.009	0.003	0.005	0.009	0.006	0.006	0.011	0.009	0.009	0.007	0.012	0.009
$^{207}\text{Pb}/^{204}\text{Pb}$	15.435	15.446	15.473	15.449	15.432	15.463	15.510	15.514	15.372	15.567	15.524	15.517	15.482	15.485
$\pm 2\sigma$	0.002	0.001	0.003	0.001	0.002	0.003	0.003	0.003	0.004	0.004	0.005	0.004	0.005	0.006
$^{208}\text{Pb}/^{204}\text{Pb}$	37.819	38.022	38.156	37.941	37.639	38.847	38.013	38.013	37.432	38.363	38.557	38.343	37.892	38.869
$\pm 2\sigma$	0.002	0.002	0.004	0.002	0.003	0.004	0.003	0.003	0.005	0.005	0.006	0.005	0.006	0.007
$(^{87}\text{Sr}/^{86}\text{Sr})_{\text{initial}}$	0.703865	0.703859	0.704628	0.703298	0.703701	0.703713	0.703716	0.703721	0.705024	0.704318	0.696549	0.682118	0.704438	0.704914
$(^{206}\text{Pb}/^{204}\text{Pb})_{\text{initial}}$	17.279	17.996	18.066	17.147	16.794	17.878	17.833	17.812	17.770	17.882	13.044	14.065	17.997	17.591
$(^{207}\text{Pb}/^{204}\text{Pb})_{\text{initial}}$	15.392	15.435	15.460	15.392	15.370	15.454	15.493	15.490	15.361	15.547	15.226	15.293	15.471	15.451
$(^{208}\text{Pb}/^{204}\text{Pb})_{\text{initial}}$	37.764	37.994	38.126	37.891	37.587	38.825	37.990	37.987	37.411	38.331	38.256	38.220	37.867	38.838
Sample	k1	k3	k4	k7	k8	k32	k44	k46	k49	k53	k62	k86	k119	
Area	Wulungu					Wucaiwan								
Hole no.	Lun2	Lun-can1	Lun-can1	Lun5	Lun5	Bei5	Cai-can1					Bei8	Cai6	
Depth (m)	2088	4639	4973	3365	3364	3500	2134	2138	2224	2652	2981	2349	1310	
Rock type	R	A	BTA	BTA	BTA	D	B	B	TA	BTA	BA	BA	BA	
SiO ₂ (wt%)	71.5	56.0	52.0	55.4	52.8	71.6	48.1	47.5	58.2	54.5	53.1	53.1	53.6	
TiO ₂	0.22	1.04	0.97	1.65	2.31	0.27	1.54	1.48	0.90	1.14	1.32	0.89	1.30	
Al ₂ O ₃	14.1	16.0	17.0	18.2	17.3	13.5	17.7	17.4	16.9	15.5	16.5	16.9	15.9	
Fe ₂ O ₃	0.50	1.38	1.26	0.99	1.71	2.50	9.81	8.84	4.76	6.67	3.82	5.27	6.26	
FeO	1.59	7.17	5.92	6.79	9.36	0.58	1.43	1.96	1.20	1.23	3.52	1.25	1.75	
MgO	0.22	2.50	7.18	2.42	4.20	0.90	3.60	4.26	2.40	2.60	2.44	4.90	3.96	
CaO	0.55	3.35	5.58	6.04	4.40	1.71	8.49	9.37	4.15	6.63	6.88	6.82	6.62	
Na ₂ O	4.51	1.68	4.04	4.74	5.16	2.74	3.79	3.52	5.30	4.47	3.95	4.12	3.44	
K ₂ O	4.95	2.89	1.54	1.08	0.96	2.89	1.22	1.00	2.04	1.54	1.22	0.89	1.67	
P ₂ O ₅	0.04	0.18	0.14	0.42	0.42	0.68	0.96	0.75	1.06	0.82	0.36	2.08	2.27	
MnO	0.06	0.10	0.14	0.15	0.10	0.11	0.15	0.15	0.10	0.16	0.10	0.10	0.16	
H ₂ O ⁺	0.64	4.84	3.32	1.55	0.83	2.00	1.99	2.27	2.68	2.85	6.15	2.68	2.54	
H ₂ O [−]	0.08	0.52	0.42	0.14	0.10	0.68	0.96	0.75	1.06	0.82	0.36	1.08	1.27	
CO ₂	0.52	1.09				0.22	0.73	0.95		1.03		0.22	0.01	
Total	99.5	98.8	99.4	99.5	99.6	100.4	100.4	100.2	100.7	100.0	99.8	100.2	100.8	
Mg [#]	0.18	0.37	0.66	0.38	0.43	0.48	0.52	0.56	0.56	0.52	0.45	0.70	0.60	
Ni (ppm)	1.60	13.8	29.5	10.7	35.4	7.10	18.5	16.0	1.13	7.10	38.0	40.6	45.9	
Co	3.90	29.5	84.6	40.8	63.1	10.2	56.5	54.3	26.7	36.0	42.3	28.6	43.8	
Cr	22.0	36.0	240	29.5	64.1	23.0	46.6	40.7	21.7	26.9	13.0	70.6	76.6	
Cu	7.60	47.0	32.6	41.7	62.0	19.4	31.7	23.1	22.5	20.3	37.8	47.0	36.7	
Zr	374	125	105	340	274	237	113	111	160	131	287	172	285	
Hf	13.2	5.40	3.93	8.75	9.02	6.84	3.61	3.92	4.52	4.50	6.21	4.32	7.83	
Rb	128	103	14.0	23.0	3.00	102	28.1	29.0	27.2	36.0	22.3	10.0	54.0	
Sr	41.5	135	1026	632	246	155	529	518	329	42.0	730	693	367	
Ba	215	466	2960	592	297	540	343	290	724	402	511	283	443	

(continued on next page)

Table 2 (continued)

Sample	k1	k3	k4	k7	k8	k32	k44	k46	k49	k53	k62	k86	k119
Area	Wulungu					Wucuiwan							
Hole no.	Lun2	Lun-can1	Lun-can1	Lun5	Lun5	Bei5	Cai-can1					Bei8	Cai6
Depth (m)	2088	4639	4973	3365	3364	3500	2134	2138	2224	2652	2981	2349	1310
Rock type	R	A	BTA	BTA	BTA	D	B	B	TA	BTA	BA	BA	BA
Nb	31.2	10.7	8.60	11.6	15.6	11.4	7.40	7.60	11.8	8.20	10.1	2.00	11.2
Ta	0.42	0.23	0.45	0.81	0.76	0.20	0.81	0.20	0.58	0.20	0.30	0.23	0.48
Th	4.69	0.52	0.36	0.52	0.55	1.16	0.21	0.20	1.08	0.75	0.51	0.21	1.26
U	10.8	2.00	2.30	2.90	2.40	4.50	1.01	0.70	4.00	2.10	1.40	1.30	2.00
Y	51.4	28.2	18.0	33.3	41.3	21.8	19.2	19.0	18.0	18.5	32.2	13.4	31.7
La (ppm)	62.3	12.1	8.40	19.0	19.1	18.3	9.66	10.2	17.8	11.4	22.0	8.42	18.5
Ce	131	28.8	20.2	49.5	49.2	39.7	24.9	25.6	38.7	26.6	51.3	21.7	48.8
Pr	15.6	3.76	2.48	6.70	6.60	5.10	3.15	3.19	4.58	0.57	6.68	3.08	6.52
Nd	58.0	18.3	12.1	30.0	30.7	22.0	15.3	15.5	18.5	15.0	28.4	13.9	26.7
Sm	10.7	4.50	2.93	6.97	7.23	4.31	3.57	3.53	3.79	3.51	6.16	2.96	5.96
Eu	0.52	1.14	1.10	1.85	1.97	0.71	1.23	1.19	1.09	1.10	1.65	0.92	1.63
Gd	9.05	4.68	3.06	6.39	7.71	3.80	3.63	3.54	3.52	3.46	5.79	2.70	5.78
Tb	1.42	0.76	0.52	1.01	1.23	0.65	0.59	0.58	0.49	0.53	0.90	0.38	0.90
Dy	8.93	5.23	3.37	6.27	7.79	4.05	3.67	3.57	3.34	3.90	5.74	2.50	5.58
Ho	1.82	1.04	0.64	1.22	1.54	0.82	0.73	0.68	0.64	0.67	1.14	0.48	1.12
Er	5.16	3.06	1.80	3.33	4.20	2.35	1.99	1.91	1.79	1.93	3.32	1.31	3.13
Tm	0.83	0.51	0.28	0.51	0.61	0.37	0.32	0.29	0.25	0.29	0.49	0.19	0.48
Yb	4.56	3.29	1.65	3.21	3.71	2.36	1.81	1.75	1.69	1.77	3.01	1.14	2.93
Lu	0.76	0.56	0.26	0.52	0.59	0.39	0.29	0.28	0.26	0.28	0.49	0.17	0.47
$^{87}\text{Rb}/^{86}\text{Sr}$	7.1280	1.3440	0.0571	0.0954	0.1398	1.1242		0.1318		0.1760	0.0674		
$^{87}\text{Sr}/^{86}\text{Sr}$	0.735527	0.711752	0.704960	0.703673	0.704972	0.708353		0.704212		0.704479	0.703969		
$\pm 2\sigma$	0.000044	0.000047	0.000029	0.000019	0.000057	0.000024		0.000033		0.000043	0.000029		
$^{147}\text{Sm}/^{144}\text{Nd}$	0.11904	0.15901	0.15998	0.14600	0.16028	0.13127		0.15530		0.14669	0.12867		
$^{143}\text{Nd}/^{144}\text{Nd}$	0.512853	0.512860	0.512903	0.512953	0.512961	0.512902		0.512897		0.512846	0.512826		
$\pm 2\sigma$	0.000007	0.000011	0.000011	0.000007	0.000008	0.000007		0.000008		0.000008	0.000007		
T_{DM} (Ga)	0.48	0.81	0.70	0.44	0.54	0.46		0.66		0.69	0.58		
T_{DM2} (Ga)	0.48	0.64	0.57	0.43	0.48	0.45		0.56		0.61	0.56		
$\varepsilon_{\text{Nd}}(t)$	4.2	4.3	5.2	6.1	6.3	5.1		5.1		4.1	3.7		
$f_{\text{Sm}/\text{Nd}}$	−0.39	−0.19	−0.19	−0.26	−0.19	−0.33		−0.21		−0.25	−0.35		
$\varepsilon_{\text{Nd}}(0)$	8.2	6.3	7.1	8.7	8.2	8.5		7.2		6.6	7.2		
$^{206}\text{Pb}/^{204}\text{Pb}$		18.314	18.328	18.072	18.147	18.200		18.218		18.448	18.075		
$\pm 2\sigma$		0.007	0.010	0.008	0.009	0.006		0.012		0.009	0.008		
$^{207}\text{Pb}/^{204}\text{Pb}$		15.492	15.495	15.519	15.604	15.473		15.488		15.534	15.414		
$\pm 2\sigma$		0.002	0.004	0.004	0.004	0.002		0.006		0.005	0.003		
$^{208}\text{Pb}/^{204}\text{Pb}$		38.019	38.079	38.049	38.345	37.874		37.868		38.238	37.659		
$\pm 2\sigma$		0.003	0.005	0.004	0.005	0.003		0.007		0.006	0.004		
$(^{87}\text{Sr}/^{86}\text{Sr})_{\text{initial}}$	0.695211	0.704150	0.704637	0.703133	0.704181	0.702074		0.703476		0.703496	0.703593		
$(^{206}\text{Pb}/^{204}\text{Pb})_{\text{initial}}$	17.508	17.400	16.906	17.176			17.940		17.608	17.523			
$(^{207}\text{Pb}/^{204}\text{Pb})_{\text{initial}}$	15.448	15.444	15.455	15.551			15.473		15.488	15.384			
$(^{208}\text{Pb}/^{204}\text{Pb})_{\text{initial}}$	37.951	38.032	37.982	38.273			37.842		38.141	37.594			

Rock name: B, basalt; BA, Basaltic andesite; A, Andesite; D, Dacite; R, Rhyolite; TB, Trachybasalt; BTA, Basaltic trachyandesite; TA, Trachyandesite; T, Trachyte.

$^{87}\text{Sr}/^{86}\text{Sr} = 0.710240 \pm 11$. The total procedural blank is less than $2\text{--}5 \times 10^{-10}$ g for Sr and 5×10^{-10} g for Nd. Analytical errors for Sr and Nd isotopic ratios are given as 2σ . The $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios were calculated using the Rb, Sr, Sm and Nd abundances obtained by ICP-MS. Detailed procedures for the Sr–Nd isotopic measurements are given in Gao et al. (1999). Nd model ages (T_{DM}) were calculated using a linear isotopic ratio growth equation (Peucat et al., 1989), i.e., $T_{\text{DM}} = \lambda\lambda^{-1} \ln \{1 + ((^{143}\text{Nd}/^{144}\text{Nd}) - 0.51315)/((^{147}\text{Sm}/^{144}\text{Nd}) - 0.2137)\}$. Pb was separated by anion exchange in HCl–Br columns using a silica gel–phosphoric acid loading technique. Total procedural blanks for Pb were in the range of 0.2–0.4 ng. During the course of this study, 13 measurements of the common-lead standard NBS981 gave average values of $^{206}\text{Pb}/^{204}\text{Pb} = 16.931 \pm 3$ (2σ), $^{207}\text{Pb}/^{204}\text{Pb} = 15.480 \pm 4$ and $^{208}\text{Pb}/^{204}\text{Pb} = 36.731 \pm 6$. Duplicate analyses indicate a reproducibility better than 0.1%. The uncertainties are $<0.1\%$ at the 95% confidence level for $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$. Details of the Pb isotopic measurement procedure are given in Zhang et al. (1997). The initial Pb and Sr isotopic compositions of the studied rocks at the time of magmatism were calculated from the measured whole-rock Pb and Sr isotopic compositions using the observed Th, U, Rb and Sr contents and ages.

5. Results

5.1. Zircon Pb–Pb dating

Zircons from three intermediate-silicic rocks, including andesite (k147), rhyolite (k20) and dacite (k32) from Kexia, Luliang and Wucuiwan, respectively, were selected for zircon Pb–Pb dating. All of the zircon grains are colorless, transparent, euhedral prisms with length/width ratios of 2.5–3.5, indicating a magmatic origin, and at least 5 zircon grains have been dated for each sample. Ages of 345, 405 and 395 Ma were obtained, respectively, for the three samples and correspond to the late Paleozoic.

5.2. Major elements

The volcanic rocks from the basement of the Junggar basin are predominantly sub-alkaline in character and range from basalt through basaltic andesite, andesite, and dacite to rhyolite. Minor alkaline and low-K sub-alkaline varieties include trachybasalts, basaltic trachyandesites and trachyte (Fig. 3). The subalkaline rocks from Kexia have a compositional range of 47.3–71.2 wt% SiO_2 and 0.71–6.27 wt% MgO, those from Wulungu have 52.0–71.5 wt% SiO_2 and 0.22–7.18 wt% MgO and those from Wucuiwan have 47.5–71.6 wt% SiO_2 and 0.90–4.90 wt% MgO. The alkaline samples from Luliang samples are subdivided into two groups: 43.4–47.5 wt% SiO_2 for basalt/trachybasalt, and 71.5–75.9 wt% SiO_2 for rhyolite (e.g., samples k16, k20 and k23).

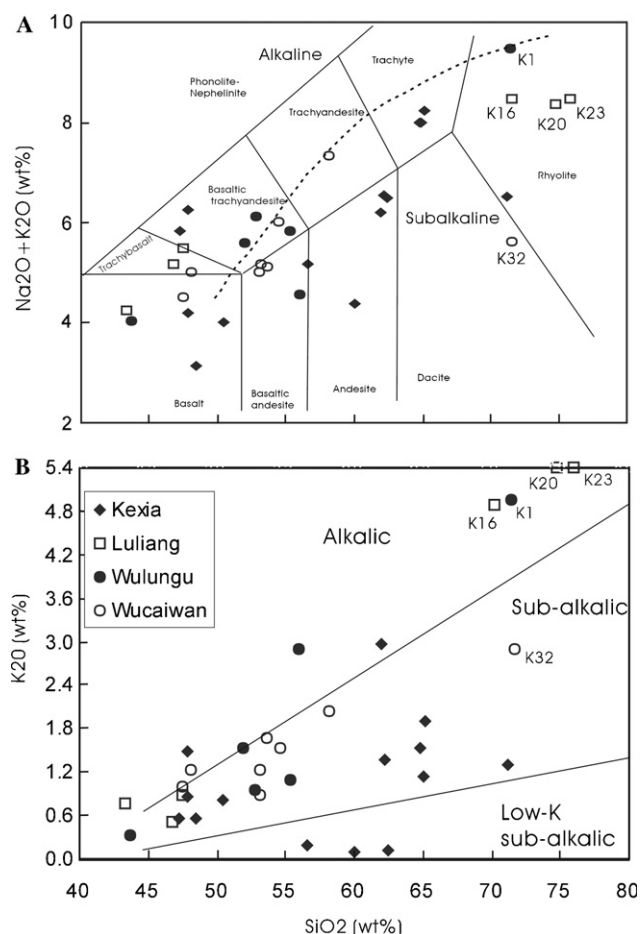


Fig. 3. Plots of SiO_2 versus $\text{Na}_2\text{O} + \text{K}_2\text{O}$ (A) and K_2O (B) for late Paleozoic volcanic rocks. (A) After Le Bas et al. (1986); (B) modified after Irvine and Baragar (1971).

A clear differentiation trend is observed for intermediate-silicic rocks (e.g., $\text{SiO}_2 \geq 53$ wt% and $\text{MgO} \leq 4.0$ wt%) from all four areas in diagrams of MgO vs. other major elements (Fig. 4). SiO_2 and $\text{K}_2\text{O} + \text{Na}_2\text{O}$ correlate negatively, whereas TFeO (total $\text{FeO} = \text{FeO} + \text{Fe}_2\text{O}_3$), CaO and TiO_2 correlate positively with MgO at <4.0 wt%. However, correlations between MgO and other major elements vary from location to location for the mafic rocks. For instance, TFeO and TiO_2 correlate negatively with MgO in the Wulungu basalts but not in the others. Alumina show no obvious trends with MgO in any of the basalts. Phosphorus is relatively rich (mean 1.12 wt%) in the Wucuiwan basalts and shows a weak positive correlation with MgO.

5.3. Trace elements

The volcanic rocks from all four areas have relatively wide ranges of Ni (i.e., 1.98–93.5 ppm, 1.80–111 ppm, 1.60–35.4 ppm and 1.13–45.9 ppm for samples from Kexia, Luliang, Wulungu and Wucuiwan, respectively). In general, Ni shows a positive correlation with MgO (Fig. 4).

The $\sum\text{REE}$ values span ranges of 37.6–148 ppm, 110–385 ppm, 76.8–362 ppm and 73.2–169 ppm, respectively,

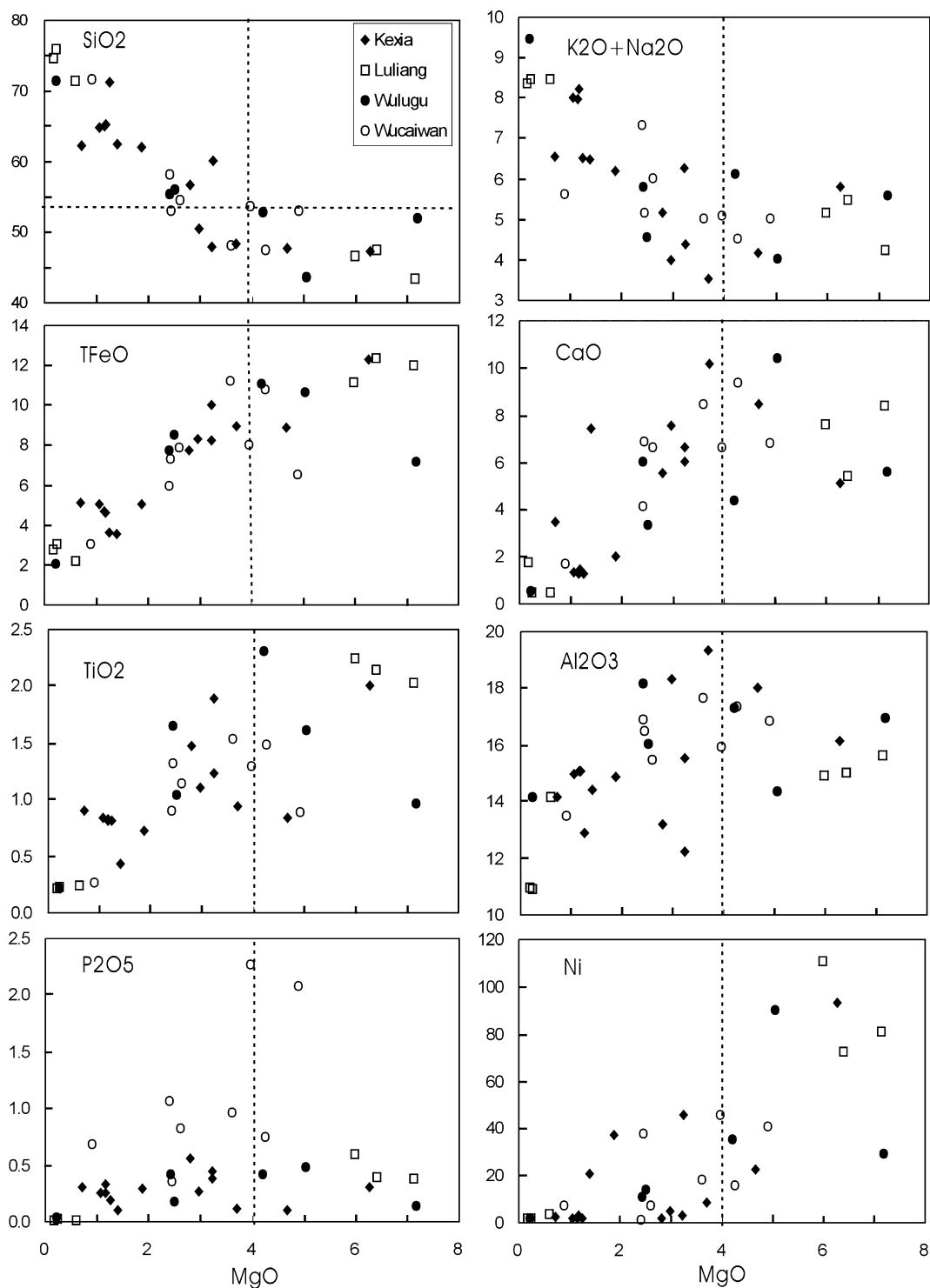


Fig. 4. Plots of MgO versus other major elements and Ni for late Paleozoic volcanic rocks.

for the Kexia, Luliang, Wulugu and Wucaiwan samples. Rhyolites from Kexia have the lowest $\sum\text{REE}$ (14.9 ppm) and $(\text{La}/\text{Yb})_N$ ratio (1.9), whereas the alkaline samples from Luliang have the highest $\sum\text{REE}$ (164–385 ppm) and $(\text{La}/\text{Yb})_N$ ratio (3.9–4.5). The mafic-intermediate rocks, including the basalt + basaltic andesite and basaltic trachyandesite + trachyandesite + trachyte assemblages, show slightly LREE-enriched chondrite-normalized REE

patterns (Fig. 5A–D) and lack significant Eu anomalies. Values of δEu (defined as $2\text{Eu}_N/(\text{Sm}_N + \text{Gd}_N)$) range from 0.8 to 1.1, while all the silicic rocks have low δEu values (e.g., 0.1–0.5) (Fig. 5E–H), and show positive subsection correlations with $\text{Mg}^\#$ (Fig. 6A).

Zr and Nb contents are 59.2–315 ppm and 1.00–10.1 ppm, and 111–285 ppm and 2.00–11.4 ppm for the Kexia and Wucaiwan rocks, respectively, which are slightly

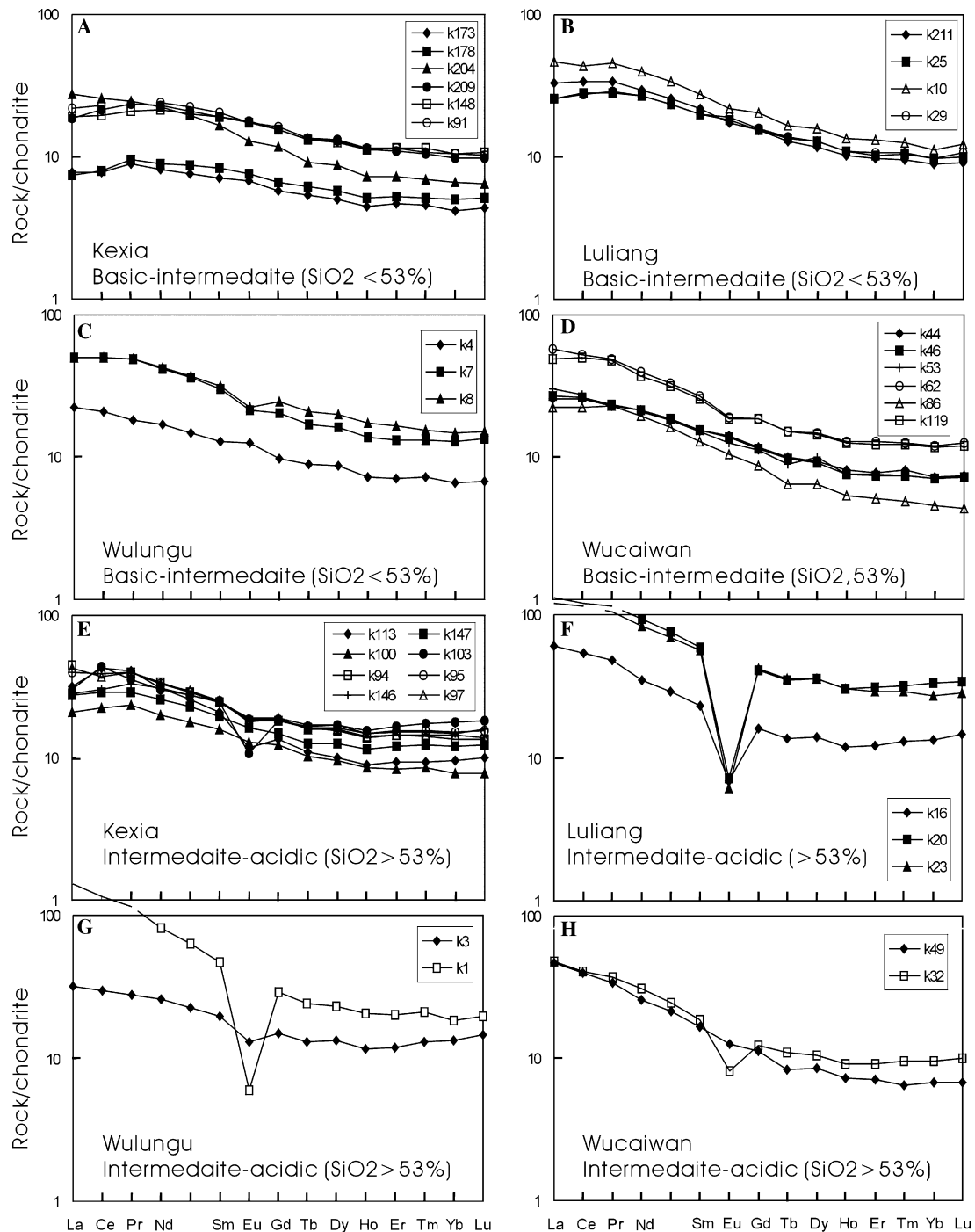


Fig. 5. Chondrite-normalized REE patterns of late Paleozoic volcanic rocks. Chondrite normalization values from Sun and McDonough (1989).

lower than those of the Luliang (165–718 ppm Zr and 105–374 ppm Nb) and Wulungu samples (80–34.1 ppm Zr and 8.60–31.2 ppm Nb). The Kexia volcanic rocks have the lowest Nb/Y, and the widest range of La/Nb (Fig. 6B).

As shown in Fig. 7, most of the mafic-intermediate rocks have high contents of highly incompatible trace elements relative to normal mid-oceanic ridge basalt (MORB). The mafic-intermediate rocks from Kexia have pronounced positive Ba anomalies and slightly positive P anomalies. Nb anomalies in these rocks vary widely from highly neg-

ative (sample k204) to slightly positive (sample k91, a basaltic andesite), and Ti anomalies vary from weakly positive to negative (sample k91) (Fig. 7A). The Luliang basalts and trachybasalts show pronounced negative Th anomalies (Fig. 7B), whereas the Wulungu basaltic trachyandesites exhibit strong positive Ba anomalies and weak negative Th anomalies (Fig. 7C). The Wucuiwan mafic-intermediate rocks display pronounced positive P anomalies (except for sample k62, a basaltic andesite). They also show positive Ba anomalies (Fig. 7D).

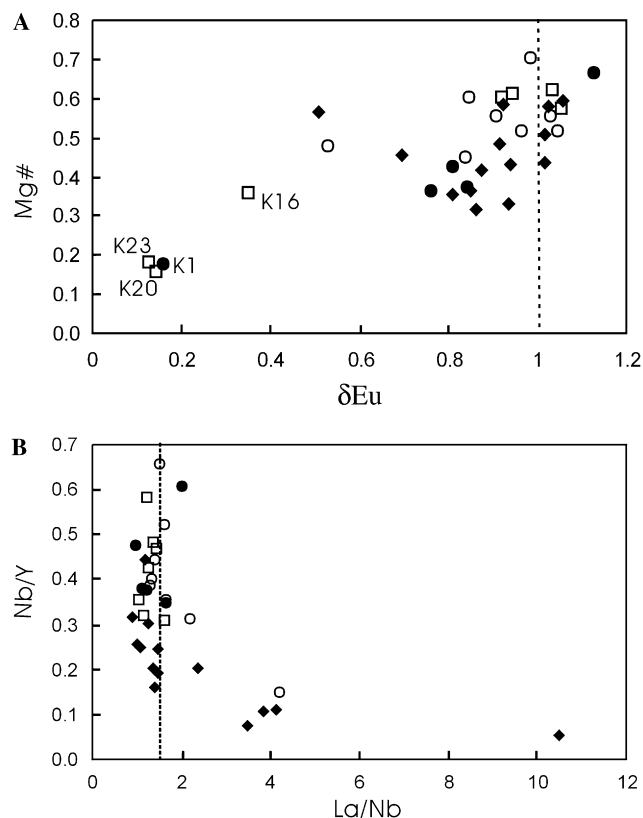


Fig. 6. Plots δEu versus $\text{Mg}^\#$ (A) and La/Nb versus Nb/Y (B) for late Paleozoic volcanic rocks. Symbols are the same as in Fig. 3. Vertical dashed lines represent $\delta\text{Eu} = 1$ (A) and $\text{La}/\text{Nb} = 1.4$ (B).

The intermediate-silicic rocks all have negative Ti anomalies (except sample k100, a Kexia andesite). The rocks from Kexia have Ba anomalies ranging from strongly positive (sample k146, a rhyolite) to none, whereas P anomalies range from obviously negative (sample k103, an andesite) to positive (sample k100, an andesite). Zr anomalies ranging from negative (sample k95) to positive (sample k103) (Fig. 7E). In the Luliang samples, P anomalies are strongly negative (Fig. 7F), whereas samples from Wulungu have both positive and negative anomalies (Fig. 7G), and those from Wucaiwan have positive anomalies (Fig. 7H).

5.4. Sr–Nd–Pb isotopes

Sr and Nd isotope data for 9 samples from Kexia show narrow ranges of $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7039–0.7050) and $\varepsilon_{\text{Nd}}(t)$ (+4.8 to +8.5), regardless of the rock type. In contrast, samples from Wucaiwan and especially from Luliang and Wulungu have widely varying $^{87}\text{Sr}/^{86}\text{Sr}$ ratios but narrow ranges of $\varepsilon_{\text{Nd}}(t)$, e.g., 0.7039–0.7083 and +3.7 to +5.2 in Wucaiwan samples, 0.7047–0.7234 and +3.3 to +6.9 in Luliang samples, and 0.7036–0.7355 and +4.2 to +6.3 in Wulungu samples. The highest $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are from rhyolites (e.g., samples k16 and k23 from Luliang and sample k1 from Wulungu), but these have low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Fig. 8A) because of their high Rb/Sr ratios. Similarly, all of the volcanic rocks have low and limited ranges of

initial $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$. They plot to the left of the ‘Geochron’ line and trend to the ‘LCC’ field (the average lower continental crust), far away from ‘UCC’ (average upper continental crust), ‘FOZO’ (focal zone) and ‘C’ (common) areas (Fig. 8B–D).

6. Discussion

6.1. Fractional crystallization of the asthenosphere-derived melts

The incompatible element distribution in mid-ocean ridge basalts (MORB) and ocean island basalts (OIB) differs from those in arc-related basalts, with the latter showing a characteristic depletion in Nb and Ta compared to large ion lithophile elements (LILE) such as K, Rb, Th and Ce (Mahoney et al., 1993). MORB and OIB lavas, which appear to come from the asthenosphere and from mantle plumes, do not show Ta–Nb depletion, and in some instances may show enrichment in these elements. Thus, Condie (1999) distinguished MORB and OIB lavas from arc lavas based on La/Nb ratios (e.g., <1.4 for OIB and MORB and >1.4 for arc basalts).

The Ni distribution in basalts is a very sensitive indicator of magma evolution. The Ni contents of OIB and MORB at a Mg number of 40 (noted as Ni_{40}) are usually >25 ppm and typically in the range of 50–100 ppm, whereas at the same Mg number, subduction-related basalts have Ni contents of <25 ppm (Condie, 1999). In OIB and MORB, Ni concentrations remain elevated even in fractionated magmas, whereas in arc-related basalts, Ni falls rapidly with the degree of fractionation (Arndt, 1991). Thus, Ni_{40} is regarded as a sensitive indicator of the differing fractional crystallization histories of basalts from different tectonic settings.

Ophiolitic basalts have La/Nb and Ni_{40} characteristics ranging from arc-like to MORB-like (Condie, 1997), depending on whether ocean ridges are associated with subduction zones in back arc basins (e.g., back arc spreading centers; Alabaster et al., 1982; Pearce et al., 1984), or if an ocean ridge approaches a subduction zone such as the Chile Ridge (Karsten et al., 1996). In these cases, even MORB can carry a subduction geochemical imprint (Condie, 1999). As shown in Fig. 9A, basaltic rocks from the Junggar basin basement have very low Ni_{40} (<22 ppm), showing an arc basalt affinity. However, most of the Junggar basin lavas have low La/Nb ratios (e.g., <1.4), (Fig. 9B), exhibiting the geochemical characteristics of enriched MORB and OIB (Fig. 7) (Fig. 10A). SiO_2 shows a negative correlation with $\text{Mg}^\#$ and Co, a positive correlation with Zr, Rb, Nb, Th, U and Y, and no obvious correlation with Sr and Ba (Fig. 11). These trends (also see Fig. 4) suggest that fractional crystallization played an important role in magma evolution (see Fig. 9A). Therefore, the basalts from the Junggar basin basement are believed to represent trapped Paleozoic oceanic crust (e.g. Hsü, 1989; Feng et al., 1989; Chen and Arakawa, 2005)

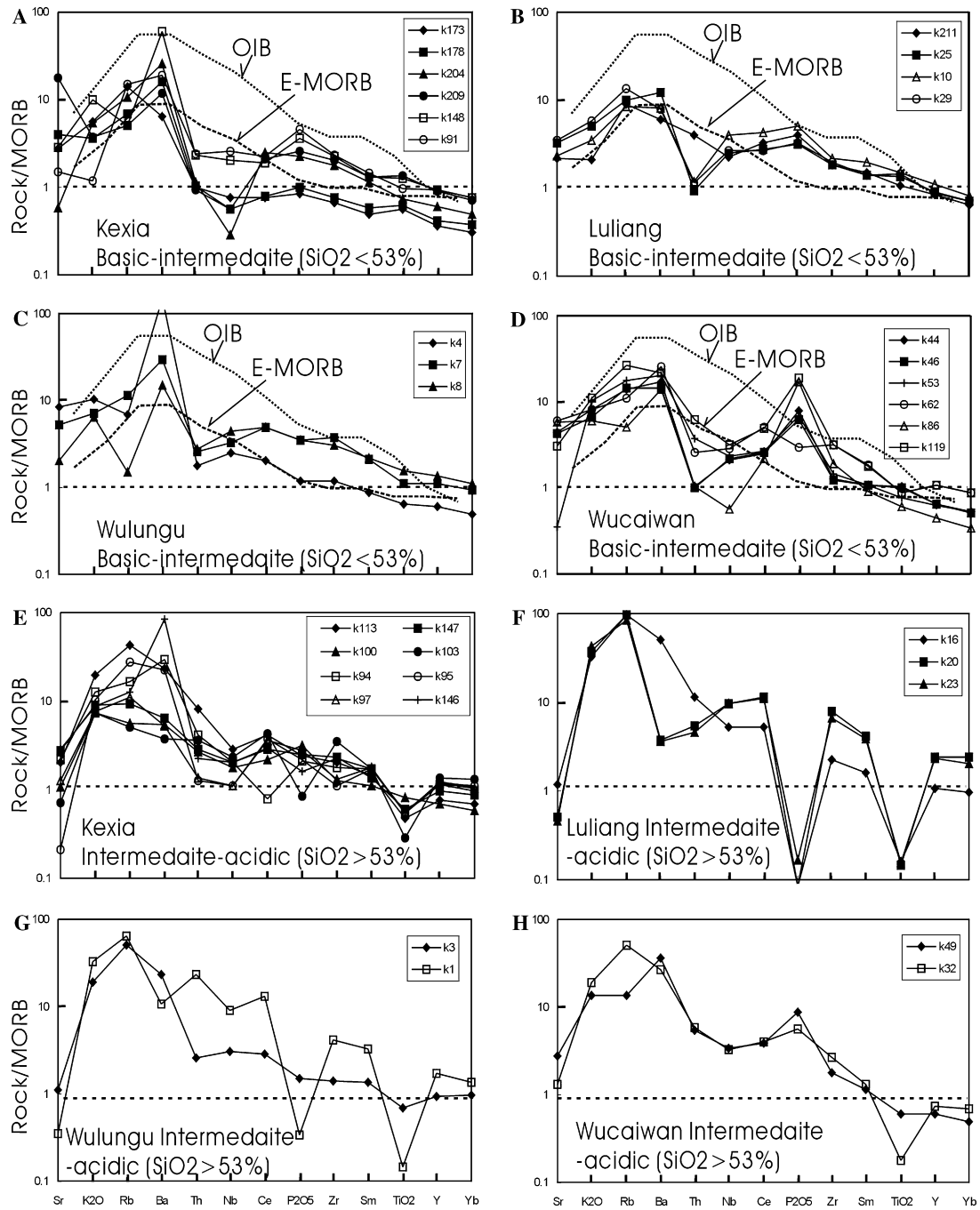


Fig. 7. MORB-normalized spidergrams for late Paleozoic volcanic rocks. The symbols are as follows: MORB, (normal) mid-ocean ridge basalt; E-MORB, enriched mid-ocean ridge basalt; OIB, ocean island basalt (after Sun and McDonough, 1989).

derived from asthenospheric melts. This interpretation is supported by the high $\varepsilon_{\text{Nd}}(t)$ and low initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of the basalts (Fig. 8) and their relationship to the surrounding Paleozoic ophiolite zones (Fig. 1B).

6.2. Silicic rocks derived from the juvenile Phanerozoic crust

Granitic rocks are the main components of continental crust, and their ε_{Nd} values are reliable indicators of their magma sources. Juvenile crust (e.g. MORB and island arc volcanic rocks) has positive ε_{Nd} and low $^{87}\text{Sr}/^{86}\text{Sr}$ iso-

topic ratios. In contrast, old continental crustal rocks generally have negative ε_{Nd} and high $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios, with the ε_{Nd} values becoming more negative with increasing crustal age. Therefore, Phanerozoic granitoids that originated from recycling of juvenile crust and by direct extraction from depleted mantle sources are expected to have initial Nd and Sr isotopic ratios very different from those of continental origin. This has been used to identify and determine the proportion of juvenile components in Phanerozoic granitoids (DePaolo, 1981; Jahn et al., 2000a).

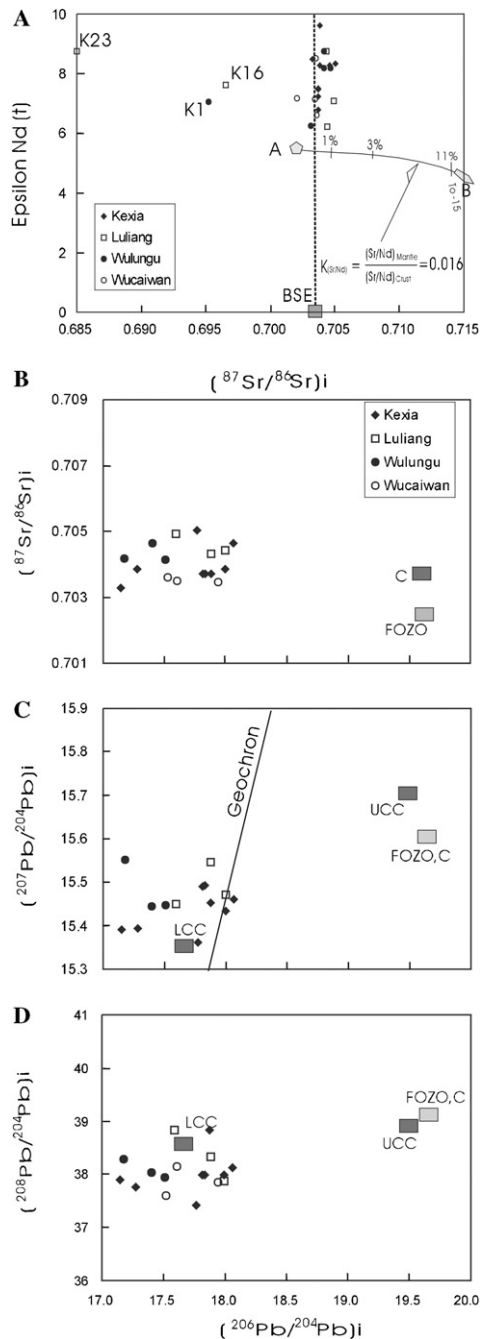


Fig. 8. Plots of (A) $\epsilon_{\text{Nd}}(t)$ versus $(^{87}\text{Sr}/^{86}\text{Sr})_{\text{initial}}$ and $(^{206}\text{Pb}/^{204}\text{Pb})_{\text{initial}}$ versus (B) $(^{87}\text{Sr}/^{86}\text{Sr})_{\text{initial}}$, (C) $(^{207}\text{Pb}/^{204}\text{Pb})_{\text{initial}}$ and (D) $(^{208}\text{Pb}/^{204}\text{Pb})_{\text{initial}}$ for late Paleozoic volcanic rocks. The curve (A–B) in Fig. 8A represents a two-component mixing line. The line is constructed using a highly differentiated mantle magma in the Altai Mountains granites (pole A) and crustal material represented by Sinian-Cambrian greenschist from the Altai Mountains (pole B). Pole A correspond to a very uniform $\epsilon_{\text{Nd}}(t) = +6.0$ and $^{87}\text{Sr}/^{86}\text{Sr} = 0.703$, with Sr = 6 ppm and Nd = 65 ppm; Pole B correspond to $\epsilon_{\text{Nd}}(t) = -15.0$ and $^{87}\text{Sr}/^{86}\text{Sr} = 0.722$, with Sr = 110 ppm and Nd = 20 ppm. The calculation was assumed under $K_{\text{Sr}/\text{Nd}} = (\text{Sr}/\text{Nd})_{\text{mantle}}/(\text{Sr}/\text{Nd})_{\text{crust}} = 0.016$ (Han et al., 1997). The symbols are as follows: BSE, primitive mantle (bulk silicate Earth); UCC and LCC, average upper and lower continental crust, respectively; FOZO ('focal zone'), where the composition is defined by the point of convergence, in three-dimensional isotope diagrams, of linear data arrays for individual ocean islands; C ('common') is a similar point of convergence of isotopic arrays for MORB and defines the component (Hofmann, 1997). The Geochron marks the locus of possible primitive mantle values assuming an overall age of the mantle of 4.5 Ga (Galer and O'Nions, 1986).

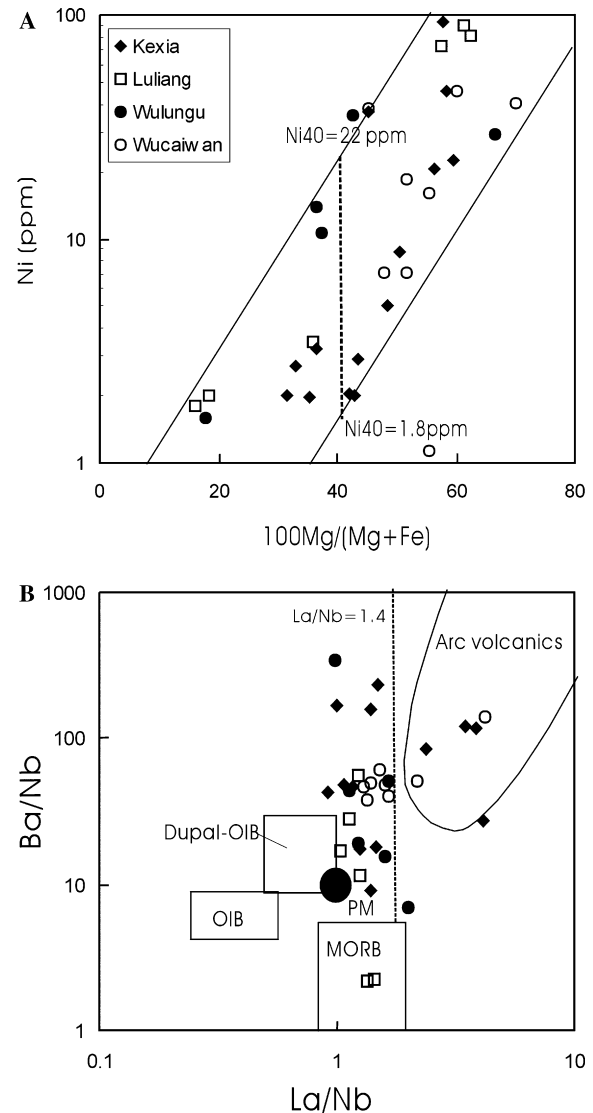


Fig. 9. Plots of (A) Ni versus $\text{Mg}^\#$, and (B) Ba/Nb versus La/Nb ratios for late Paleozoic volcanic rocks. Data source: Arc volcanics after Jahn et al. (1999) and Fan et al. (2004). The symbols are as follows: PM, primitive mantle; OIB, oceanic island basalt; MORB, mid-ocean ridge basalt; Dupal OIB, ocean island basalt with a Dupal anomaly.

Recent studies reveal that a substantial proportion of the Phanerozoic crust is isotopically juvenile, such as the Canadian Cordilleran batholiths (Samson and Patchett, 1991), the Peninsular Range and Sierra Nevada batholiths (DePaolo et al., 1991), the Lachlan and New England Foldbelts of eastern Australia (Collins, 1998), and the Central Asia Orogenic Belt (Sengör et al., 1993; Han et al., 1997; Jahn et al., 2000b). Previous work has shown that post-collisional granitoids from western Junggar (Feng et al., 1989; Xian et al., 2002; Chen and Arakawa, 2005), northern Junggar (Han et al., 1997) and eastern Junggar (Chen and Jahn, 2004) are also characterized by highly depleted isotopic signatures. All of the late Paleozoic silicic rocks studied from the Junggar basin basement have positive ϵ_{Nd} values (up to +7.4) and low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Table 2 and Fig. 8), suggesting that the entire

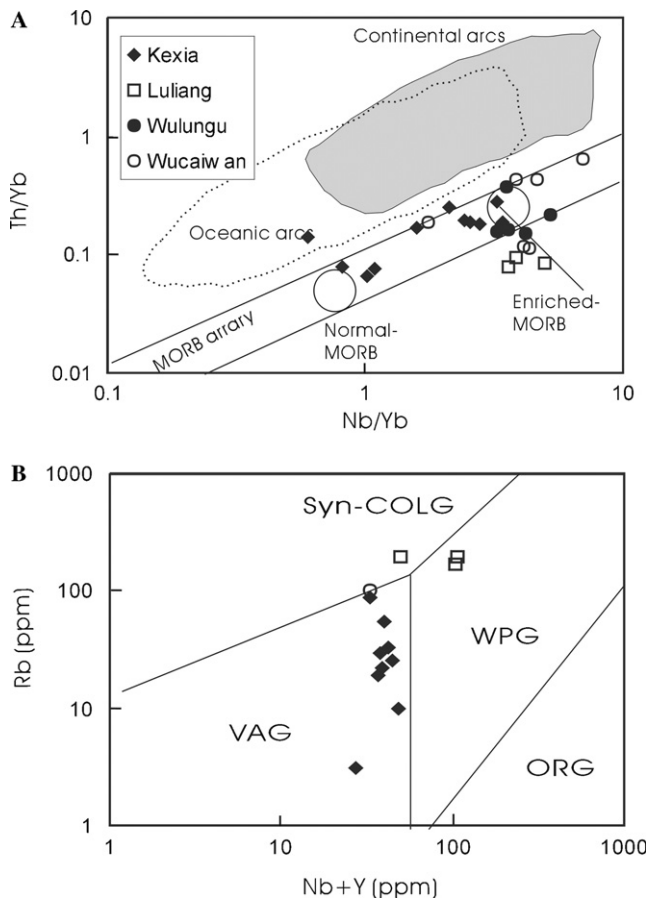


Fig. 10. Plots of (A) Th/Yb versus Nb/Yb, and (B) Rb versus Nb + Y for late Paleozoic rocks. (A) Graph of Th/Yb versus Nb/Yb ratios modified after Pearce and Peate (1995). Normal and enriched MORB values are after Sun and McDonough (1989); oceanic and continental arcs are after Shinjo et al. (1999) and Xu et al. (2003). (B) Plots of Rb versus Nb + Y after Pearce et al. (1984). The symbols are as follows: syn-COLG, syn-collisional granites; VAG, volcanic arc granites; WPG, within-plate granites; ORG, ocean-ridge granites.

Junggar terrane may be dominated by juvenile crust. The crust may have been derived from asthenospheric melts followed by fractional crystallization in a post-collisional setting, or represent the product of melting of Paleozoic oceanic crust due to underplating of Mesozoic magmas (Han et al., 1999). This is consistent with the juvenile isotopic signature shown by many granitoid plutons in other parts of the Central Asia Orogenic Belt, such as those in NE China (Wu et al., 2000; Wei et al., 2002), Russian Transbaikalia (Wickham et al., 1996), and Kazakhstan (Heinhorst et al., 2000). These facts suggest that the Central Asia Orogenic Belt, including the Junggar basement, is probably one of the most important sites of juvenile crustal formation in the Phanerozoic (Jahn et al., 2000a).

6.3. Heterogeneous Phanerozoic basement beneath the Junggar basin

Different zircon ages (Table 1), different Ba, P, Ti, Nb or Th anomalies in mafic rocks (Fig. 7), variable Nb and

Zr contents (Table 2), and variable Nb/Y and La/Nb ratios (Fig. 6B) in different areas all suggest that the Junggar basin is underlain by heterogeneous basement. The heterogeneity is also recorded in the intermediate-silicic rocks. For instance, the silicic rocks from Kexia have compositions similar to volcanic arc granites, whereas samples from Luliang and Wucaiwan are transitional between syn-collisional and within-plate granites as indicated by their Rb and/or Nb + Y contents (Fig. 10B). The heterogeneity most likely reflects different stages of evolution of the basement beneath the Junggar basin, which resulted in different rock assemblages with related geochemistry.

The triangular Junggar basin is bordered by several ophiolite belts that mainly trend NW, WNW and NE (Fig. 1B). The NW-trending ophiolites include the Armantai-Jaheba ophiolite belt (561 Ma; Huang et al., 1997) and the Kalameli ophiolite belt, which lies south of the Altay Mountains and east of the Junggar basin. The Hongguleleng ophiolite belt (444 Ma; Zhang and Huang, 1992) trends NW in the western part of the Junggar basin and is regarded as the western extension of the Kalameli belt, from which it is separated by a highly magnetic zone ($>2000 \times 10^6$ CGSM). Thus, the Wulungu depression northeast of the Junggar basin is located between the Armantai-Jaheba and Kalameli-Hongguleleng ophiolite belts, and is regarded as a fragment of oceanic crust (intra-oceanic arc; He et al., 1994). Northeast-trending ophiolites include the Silurian Mayilashan belt (XBGM, 1993) and the Devonian Darbut belt (395 Ma; Zhang and Huang, 1992). The WNW-trending ophiolites include the early-late Paleozoic Tangbale belt and the middle-late Paleozoic Bayaangou-Dannahu belt, which lie in the southwestern part of the Junggar basin and in the Tianshan area, respectively (Zhang and Huang, 1992; XBGM, 1993). The early Devonian magmatism recorded in rhyolite and dacite of the central Luliang uplift (405 Ma; sample k20) and southeastern Wucaiwan depression (395 Ma; sample k32) is slightly younger than the NW-trending Kalameli-Hongguleleng ophiolite belt. In addition, early Carboniferous andesitic magmatism in the western Kexia area (345 Ma; sample k147) is younger than the Devonian Darbut and Silurian Mayilashan ophiolites. The different tectonic history of the areas around the Junggar basin (e.g. Cui et al., 1996; Wang, 1996; Zhang and Wang, 1996) implies that the ophiolites were derived from oceanic lithosphere during different stages of evolution and from different terranes. Therefore, we interpret the basement of the Junggar basin as the amalgamated product of oceanic crustal blocks derived from different terranes and in different stages of evolution (Fig. 10B).

7. Conclusions

1. The basement of the Junggar basin is composed mainly of late Paleozoic volcanic rocks with minor shale and tuff. The volcanic rocks are predominantly subalkaline

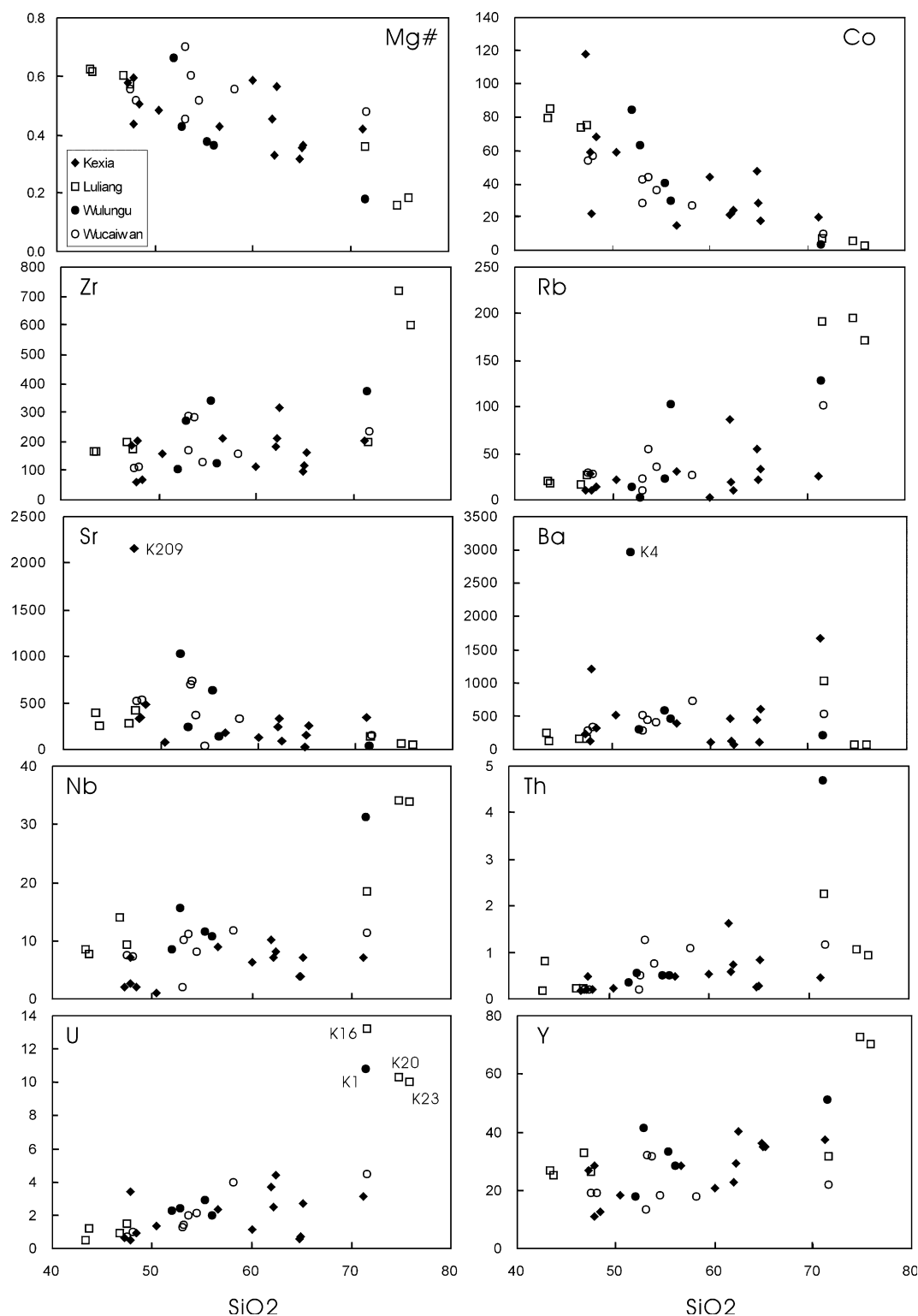


Fig. 11. Plots of SiO_2 versus $\text{Mg}^\#$ and trace elements for late Paleozoic volcanic rocks. These plots suggest that fractional crystallization played an important role in magma evolution.

in character and some in the Kexia area are low-K varieties. A few alkaline lavas occur in the central Luliang uplift and northeastern Wulungu depression. The basaltic rocks mainly represent trapped late Paleozoic oceanic crust produced from asthenospheric melts.

2. Positive ϵ_{Nd} values (up to +7.4) and low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the silicic rocks suggest that they formed by fractional crystallization of asthenosphere-derived melts in a post-collisional setting, or are products of melting of oceanic crust due to underplating of Mesozoic magmas.

The juvenile basement is widespread, but heterogeneous, and underlies the entire Junggar region. The heterogeneity may reflect amalgamation of oceanic crustal blocks during different stages of evolution involving different terranes.

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