Origin of middle Miocene leucogranites and rhyolites on the Tibetan Plateau: Constraints on the timing of crustal thickening and uplift of its northern boundary

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Leucogranites play a significant role in understanding crustal thickening, melting within continental collisional belts, and plateau uplift. Field investigations show that Miocene igneous rocks from the Hoh Xil Lake area mainly consist of two-mica leucogranites and rhyolites. We studied the Bukadaban two-mica leucogranites and the Kekao Lake, Malanshan and Hudongliang rhyolites by zircon U-Pb, muscovite and sanidine ⁴⁰Ar/³⁹Ar geochronology, and whole-rock geochemical and Sr-Nd isotopic analysis. Results yielded crystallization and cooling ages for the Bukadaban leucogranites of 9.7±0.2 and 6.88±0.19 Ma, respectively. Extrusive ages of the Kekao Lake and Malanshan rhyolites are 14.5±0.8 and 9.37±0.30 Ma, respectively. All rocks are enriched in SiO₂ (70.99%–73.59%), Al₂O₃ (14.39%–15.25%) and K₂O (3.78%–5.50%) but depleted in Fe₂O₃ (0.58%–1.56%), MgO (0.11%–0.44%) and CaO (0.59%–1.19%). The rocks are strongly peraluminous (A/CNK=1.11–1.21) S-type granites characterized by negative Eu anomalies ($\delta\text{Eu}=0.18\text{--}0.39$). In also considering their Sr-Nd isotopic compositions (⁸⁷Sr/⁸⁶Sr_i=0.7124 to 0.7143; ϵNd (9 Ma)=−5.5 to −7.1), we propose that these igneous rocks were generated through dehydration melting of muscovite in the thickened middle or lower crust of northern Tibet. Melting was probably triggered by localized E-W stretching decompression in the horse tails of Kunlun sinistral strike-slip faults. Reactivation of the Kunlun strike-slip faults, accompanied by emplacement of leucogranite and eruption of rhyolite in the Hoh Xil Lake area, indicates that large-scale crustal shortening and thickening in northern Tibet mainly occurred before 15 Ma. In addition, these findings suggest that the northern Tibetan Plateau attained its present elevation (~5000 m) at least 15 Ma ago.

Hoh Xil, zircon U-Pb dating, ⁴⁰Ar/³⁹Ar dating, leucogranite, rhyolite, crustal thickening, Tibetan Plateau uplift

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Leucogranites play a significant role in understanding crustal thickening, melting within continental collisional belts, and plateau uplift. The High and North Himalayan leucogranites were generated through melting of thickened crust in the Himalayan orogen [1–14] due to the India-Asia continental collision 65–50 Ma [15,16]. Petrogenetic models have investigated the effects of different factors including: fluid infiltration [1], high radioactivity [17], shear heating [18] and tectonic decompression [19]. Most models

suggest that leucogranite generation is strongly related to previous large-scale crustal thickening. Recently, studies of Eocene to Oligocene leucogranites (44–27 Ma) from the belts of the northern Himalaya indicate that significant crustal thickening in southern Tibet possibly initiated before 44 Ma [10,14]. Gravity and isostasy studies of leucogranites indicate that the Himalayan area had or nearly reached its highest altitude prior to middle-late Eocene time [9,10,14]. Therefore, leucogranites are important and relevant to fully understanding the tectonics of the region. In contrast, little work on leucogranites has been performed on the northern

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Tibetan Plateau except in the Ulugh Muztagh area [20,21]. Rhyolites and leucogranites from the Ulugh Muztagh area were derived from melting of metapelites in the thickened crust [21]. This study reports the geochemistry and zircon U-Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of crustally derived rocks from the Hoh Xil Lake area, including not only the leucogranites of Bukadaban but also the rhyolites of Kekao Lake, Malanshan and Hudongliang, in order to improve our understanding of leucogranite petrogenesis and middle and lower crustal thickening on the northern Tibetan Plateau.

1 Geological setting and samples

The Hoh Xil-Songpan-Ganzi (HXSG) terrane is bounded by the Anyimaqen-Kunlun-Muztagh suture zone to the north and the Jinshajiang suture zone to the south (Figure 1(a)). The HXSG terrane is dominated by strongly folded Triassic flysch deposits, most of which were unconformably overlain by Cenozoic terrestrial sediments (Figure 1(b)). Neogene potassic volcanic rocks are widespread in northern Tibet (Figure 1(a)). These rocks are dominantly felsic lavas with minor basalts, and eruptive ages range from Miocene to present (18–0 Ma) [25–37]. Previous results indicated that at least three different types of magma sources may be distinguished: (1) potassic volcanic rocks derived from low-degree melting of the pre-existing enriched mantle [34] or

mixed crust-mantle layers resulting from intracontinental subduction [26,27,33]; (2) high-K calc-alkaline to adakitic volcanic rocks generated through melting of the thickened mafic lower crust [32,36]; and (3) crustally-derived igneous rocks produced by melting of metapelites at depths of 20–30 km in the thickened crust [20,21]. The occurrence of these volcanic rocks in general was spatially related to the major sinistral strike-slip Kunlun and Altyn Tagh faults or their secondary faults in the northern Tibetan Plateau [26,38]. In the Hoh Xil Lake area, volcanic rocks also were generally distributed along the South Kunlun sinistral strike-slip faults and their subordinate faults. Eruptions of volcanic rocks were associated with normal faulting in pull-apart basins or grabens along the Bukadaban-Jingyu Lake-Muztagh belts [26, 31,39,40]. Both the Jingyu Lake and Kusai Lake pull-apart basins were filled with late Miocene lacustrine and fluvial sediments, suggesting that the South Kunlun sinistral strike-slip fault was reactivated during early Miocene time [39,40].

In this study, samples were collected from areas between Hoh Xil Lake and the South Kunlun sinistral strike-slip fault (Figure 1(b)). Leucogranitic plutons are located south of Bukadaban Peak and are predominantly covered by glaciers. The Bukadaban plutons intrude an area of 50 km² and are cut by the South Kunlun sinistral strike-slip fault. Rocks are coarse-grained leucogranites (Figure 2) and are mainly composed of K-feldspar (30–40 vol.%), quartz (25 vol.%) and plagioclase (25–30 vol.%) with minor muscovite (2 vol.%)

Figure 1 (a) Cenozoic magma distributions on the Tibetan Plateau, modified after [22–24]. Ages for Cenozoic volcanic rocks are from the literature [22–24] and references therein. Ages of Miocene Himalayan leucogranites are from previous reports [1–14,18]. (b) Ages of Neogene volcanic rocks in the Hoh Xil Lake area are from previous reports [25–29] and this study. Main suture zones between major terranes: AKMS, Anyimaqen-Kunlun-Muztagh suture; JS, Jinshajiang suture; BS, Bangonghu-Nujiang suture; IYS, India-Yaluzangbu suture.

Figure 2 Photomicrograph of Bukadaban two-mica leucogranites. Bt, Biotite; Ms, muscovite; Kfs, K-feldspar.

and biotite (3 vol.%). Accessory minerals consist of zircon and apatite (<1 vol.%). K-feldspar is generally euhedral or subeuhedral and some large crystals can reach 2 cm in length. Plagioclase is also euhedral or subeuhedral and polysynthetic twins are uncommon. Biotite is more abundant than muscovite by volume.

The Kekao Lake rhyolites are exposed southeast of the lake in a small-scale lava dome that is ~100 m in diameter. Rocks are porphyritic rhyolites with sanidine, quartz and tourmaline phenocrysts in a microlitic groundmass. The groundmass is dominantly composed of fine-grained quartz, plagioclase and sanidine. Northeast of these rhyolites are shoshonites with K/Ar ages of 14.5 to 7.9 Ma [27,29].

The Malanshan rhyolites consist of sheeted lavas that are located south of Malanshan Peak and cover an area of 10 km². The rocks are gray in color and are composed of quartz, sanidine and biotite phenocrysts set in an isotopic groundmass.

The Hudongliang rhyolites are located northeast of Lexiewudan Lake and erupted along the secondary sinistral strike-slip fault of the South Kunlun fault system. They are dark in color and are composed of sanidine, quartz and tourmaline phenocrysts set in a glassy groundmass. Whole rock K/Ar geochronology yielded an age of ~2 Ma [28].

2 Analytical methods

Zircon crystals were separated using conventional heavy liquid and magnetic techniques. The samples, including an external zircon standard TEMORA-1 (417Ma) [41], were mounted in epoxy, polished and vacuum-coated with a 50 nm layer of gold [42]. Zircons were examined in transmitted and reflected light, and cathodoluminescence (CL) images were taken for further analysis. Zircon U-Pb analyses were processed using a SHRIMP II (Sensitive High Resolution Ion Probe) at the Institute of Geology, Chinese Academy of Geological Sciences (IGCAGS), and LA-ICPMS (Laser Ablation Inductively Coupled Mass Spectrometry) at the

Institute of Tibetan Plateau Research, Chinese Academy Sciences (ITPCAS). The SHRIMP II dating procedure followed details given in [43,44]. The data were processed offline using SQUID (version: 1.02) [45]. Common Pb was corrected against ²⁰⁷Pb rather than ²⁰⁴Pb because of low radioactive Pb abundances for the young samples [44,46]. Results are shown in Table 1. Based on CL image analysis, zircon cores of Bukadaban leucogranites were analyzed using LA-ICPMS following the details in [47]. LA-ICPMS data were processed using GLITTER (version: 4.0) [48] and common Pb corrections were made following the method described by Anderden [49]. Results are shown in Table 2. The data were processed using the ISOPLOT (version: 2.49) program [50].

Muscovite and sanidine grains were separated using conventional methods and purified by hand-picking under a binocular microscope. Samples (weight: 50 mg) were wrapped in aluminum foil to form wafers, and stacked in quartz vials with the standard FCT-1 sanidine (27.9 Ma) [51]. The position of every sample was recorded as the distance from the bottom of the vial. Then, the vials were sealed and put into a quart-sized canister. The canister was wrapped with cadmium foil to shield slow neutrons and thus prevent interface reactions during irradiation. Neutron irradiation was carried out in H8 of the 49-2 Nuclear Reactor, Beijing, China, with a neutron dose of $2.65 \times 10^{13} \text{ n}/(\text{cm}^2/\text{s})$ for 24 h. The irradiated samples were transferred into a “Christmas tree” which was high vacuumed under $1\text{--}2 \times 10^{-9} \text{ Pa}$. After a one week bake-out at 250°C, the sample wafers were dropped into a double-vacuum resistance furnace and heated at 650°C for 30 min. All samples were step-heated from 800 to 1350°C in 8–14 steps. The released gas was purified by getters and isotopic measurements were taken on the HELIX mass spectrometer at ITPCAS. K₂SO₄ and CaF₂ crystals were fused to calculate Ca and K correction factors: $(^{36}\text{Ar}/^{37}\text{Ar})_{\text{Ca}} = 2.398 \times 10^{-4}$, $(^{40}\text{Ar}/^{39}\text{Ar})_{\text{K}} = 4.782 \times 10^{-3}$, $(^{39}\text{Ar}/^{37}\text{Ar})_{\text{Ca}} = 8.1 \times 10^{-4}$. Results were processed by corrections of system blanks, zero time and mass discrimination. Ages were calculated using decay constants and the formulae from [52,53]. These data

Table 1 Zircon SHRIMP U-Pb dating results for the Kekao Lake rhyolites and the Bukadaban two-mica leucogranites ^{a)}

Spot	²⁰⁶ Pb _c (%)	U (μg g ⁻¹)	Th (μg g ⁻¹)	²³² Th/ ²³⁸ U	²⁰⁶ Pb (μg g ⁻¹)	²⁰⁷ Pb*/ ²⁰⁶ Pb*	±%	²⁰⁷ Pb*/ ²³⁵ U	±%	²⁰⁶ Pb*/ ²³⁸ U	±%	²⁰⁶ Pb/ ²³⁸ U age (Ma)	Discordant (%)
Kekao Lake: 2006K034 (GPS: N35°41.458', E91°31.25')													
034-1	2.2	1660	482	0.30	3.1	0.0333	23	0.0098	23	0.0021	2	13.7±0.3	2
034-2	0.5	3645	871	0.25	7.8	0.0441	6.5	0.0151	6.7	0.0025	1.6	16.0±0.3	15
034-3	2.7	993	354	0.37	1.9	0.0370	30	0.0111	30	0.0022	2.3	14.0±0.3	2
034-4	6.4	1130	222	0.20	2.4	0.0440	31	0.0138	31	0.0023	2.5	14.7±0.4	13
034-5	4.2	2091	815	0.40	4.1	0.0313	26	0.0095	26	0.0022	1.9	14.2±0.3	1
034-6	1.0	2997	700	0.24	6.2	0.0432	7.4	0.0142	7.6	0.0024	1.7	15.4±0.3	10
034-7	1.7	867	537	0.64	1.5	0.0439	16	0.0121	16	0.0020	2.7	12.9±0.4	11
034-8	1.1	1480	356	0.25	3.0	0.0444	14	0.0144	14	0.0024	1.9	15.2±0.3	17
034-9	0.3	6138	555	0.09	13.6	0.0479	3.8	0.0169	4.1	0.0026	1.6	16.5±0.3	-18
034-10	1.6	1006	599	0.62	1.8	0.0441	14	0.0127	14	0.0021	1.9	13.4±0.3	13
Bukadaban: 2006K074 (GPS: N35°58'02", E90°50'40")													
074-1	2.8	7957	2460	0.32	10.8	0.0328	16	0.0069	17	0.0015	1.9	9.9±0.2	1
074-2	0.2	18027	259	0.01	23.4	0.0482	2.3	0.0100	2.8	0.0015	1.6	9.7±0.2	-9
074-3	0.5	7918	636	0.08	12.7	0.0465	6.7	0.0119	6.9	0.0019	1.6	11.9±0.2	-50
074-4	0.5	5403	537	0.10	9.5	0.0459	3.8	0.0128	4.1	0.0020	1.6	13.1±0.2	194
074-5	4.5	2392	228	0.10	3.1	0.0500	33	0.0100	33	0.0014	2.8	9.3±0.3	-4
074-6	5.1	2074	393	0.20	2.8	0.0350	31	0.0071	31	0.0015	2.2	9.6±0.2	101
074-7	3.3	2856	100	0.04	3.1	0.0292	26	0.0049	26	0.0012	2.1	7.9±0.2	1
074-8	1.0	12835	340	0.03	17.0	0.0433	7.4	0.0091	7.6	0.0015	1.6	9.9±0.2	6
074-9	1.5	10384	259	0.03	13.7	0.0451	12	0.0094	13	0.0015	1.7	9.8±0.2	19
074-10	0.2	22014	345	0.02	31.6	0.0486	1.8	0.0112	2.4	0.0017	1.5	10.7±0.2	-8

a) Pb_c and Pb* indicate the common and radiogenic portions, respectively.

Table 2 Zircon LA-ICPMS U-Pb dating results for cores of inherited zircons from the Bukadaban two-mica leucogranites ^{a)}

Bukadaban 2006K074	Isotope ratio								Discordant (%)
	²⁰⁷ Pb*/ ²⁰⁶ Pb*	1σ	²⁰⁷ Pb*/ ²³⁵ U	1σ	²⁰⁶ Pb*/ ²³⁸ U	1σ	²⁰⁸ Pb*/ ²³² Th	1σ	
11	0.0513	0.0020	0.2864	0.0109	0.03927	0.00055	0.01435	0.00024	3
12	0.0563	0.0014	0.2462	0.0063	0.03073	0.00039	0.01302	0.00016	14
13	0.0512	0.0024	0.2178	0.0101	0.02992	0.00045	0.01527	0.00033	5
14	0.0553	0.0009	0.2615	0.0044	0.03322	0.00039	0.01557	0.00016	12
15	0.0755	0.0017	1.6393	0.0325	0.15742	0.00186	0.04737	0.00060	5
Bukadaban 2006K074	Isotope age (Ma)								²³² Th/ ²³⁸ U
	²⁰⁷ Pb*/ ²⁰⁶ Pb*	1σ	²⁰⁷ Pb*/ ²³⁵ U	1σ	²⁰⁶ Pb*/ ²³⁸ U	1σ	²⁰⁸ Pb*/ ²³² Th	1σ	
11	252	62	256	9	248	3	288	5	0.81
12	464	34	223	5	195	2	261	3	0.79
13	248	79	200	8	190	3	306	7	0.51
14	425	18	236	4	211	2	312	3	0.33
15	1082	47	985	12	942	10	935	12	0.22

a) Pb* represents radiogenic lead; common lead are corrected using measured ²⁰⁴Pb; ²⁰⁶Pb/²³⁸U ages were chosen.

were processed using ISOPLOT (version: 2.49) [50], and are shown in Table 3.

Whole-rock major and trace element and Rb-Sr and Sm-Nd isotopic analyses were processed at the Institute of

Geology and Geophysics, Chinese Academy of Sciences (IGGCAS). Major element abundances (wt.%) were determined on glass pills by an X-ray fluorescence spectrometer (XFR), and yielded <5% error. Rare earth elements (REEs)

Table 3 $^{40}\text{Ar}/^{39}\text{Ar}$ isotope dating results for the Bukadaban two-mica leucogranites and the Malanshan rhyolites ^{a)}

$T(^{\circ}\text{C})$	$(^{36}\text{Ar}/^{39}\text{Ar})_m$	1σ	$(^{37}\text{Ar}/^{39}\text{Ar})_m$	1σ	$(^{38}\text{Ar}/^{39}\text{Ar})_m$	1σ	$(^{40}\text{Ar}/^{39}\text{Ar})_m$	1σ	Age (Ma)	1σ
2006K074 (muscovite), J: 0.003333, weight: 15.0 mg, Bukadaban, GPS: N35°58'02", E90°50'40"										
750	0.01424	0.00035	0.04787	0.02039	0.01050	0.00058	6.027	0.074	10.80	1.50
850	0.00861	0.00013	0.02428	0.00496	0.01398	0.00028	3.830	0.041	7.67	0.75
930	0.00466	0.00005	0.00261	0.00290	0.01302	0.00014	2.550	0.021	7.02	0.30
1000	0.00318	0.00003	0.00123	0.00158	0.01301	0.00009	2.058	0.012	6.70	0.16
1050	0.00385	0.00002	0.00250	0.00107	0.01340	0.00003	2.262	0.002	6.73	0.10
1100	0.00335	0.00001	0.00571	0.00156	0.01318	0.00004	2.141	0.002	6.90	0.10
1150	0.00188	0.00005	0.00100	0.00554	0.01269	0.00013	1.724	0.002	6.99	0.10
1210	0.00210	0.00013	0.01011	0.00941	0.01352	0.00030	1.681	0.004	6.33	0.10
1300	0.00065	0.00158	0.11313	0.07803	0.00538	0.00158	1.984	0.006	13.13	0.19
2006K129 (sanidine), J: 0.003570, weight: 11.9 mg, Malanshan, GPS: N35°45'12.6", E90°39'57.8"										
750	0.00097	0.00100	0.29367	0.06063	0.01298	0.00135	2.509	0.070	16.54	0.46
850	0.00089	0.00006	0.03375	0.00296	0.01277	0.00009	1.944	0.002	10.24	0.14
930	0.00019	0.00003	0.02277	0.00197	0.01268	0.00004	1.594	0.001	9.37	0.13
1000	0.00035	0.00005	0.01830	0.00343	0.01309	0.00008	1.595	0.001	9.07	0.13
1110	0.00042	0.00003	0.01991	0.00317	0.01267	0.00008	1.568	0.001	8.78	0.12
1160	0.00018	0.00003	0.00473	0.00193	0.01250	0.00005	1.565	0.001	9.17	0.13
1220	0.00006	0.00004	0.00167	0.00338	0.01177	0.00007	1.571	0.002	9.64	0.13
1300	0.00006	0.00001	0.00326	0.00114	0.01224	0.00006	1.609	0.004	9.86	0.14
1450	0.00004	0.00003	0.01019	0.00381	0.01181	0.00009	1.605	0.004	9.64	0.14

a) m indicates actually measured value.

and trace elements were analyzed on an Inductively Coupled Plasma-Mass Spectrometer (ICP-MS) with uncertainties better than 5%. Whole-rock Rb-Sr and Sm-Nd isotopic analyses were conducted on a Finnigan MAT-262 mass spectrometer with isotopic spiking methods [54]. All results are shown in Table 4.

3 Results

3.1 Zircon U-Pb geochronology

Zircons of Bukadaban leucogranites were colorless to light brown, and were typically ~50–150 μm in length, with length/width ratios between 5:1 and 2:1. Most crystals in the CL images exhibited light inner cores and dark rims with high U and Th concentrations (Figure 3(a)). Eight analyses of zircon rims by SHRIMP II were plotted in or near the concordia diagram, and six analyses yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 9.7 ± 0.2 Ma (MSWD=0.88) (Figure 4(a)). We interpret the age of 9.7 ± 0.2 Ma as the crystallization age for the Bukadaban leucogranites. In addition, five analyses for inherited or captured zircon cores yielded a cluster of older $^{206}\text{Pb}/^{238}\text{U}$ ages with a wide range (193 ± 3 Ma; 195 ± 2 Ma; 211 ± 2 Ma; 248 ± 3 Ma; 942 ± 10 Ma) (Figures 3(a) and 4(a)). These five zircon cores showed oscillatory zoning and had high Th/U ratios ranging from 0.22 to 0.81.

Zircons of the Kekao Lake rhyolites were colorless to

light brown, and were typically ~150–250 μm in length, with length/width ratios between 2:1 and 1:1. These crystals exhibited oscillatory zonation in the CL images (Figure 3(b)). These zircons with high Th/U ratios (>0.1) were typical of magmatic zircons. Ten zircon analyses were plotted in the concordia diagram within error margins, and gave a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 14.5 ± 0.8 Ma (MSWD=13) as the eruptive age of the Kekao Lake rhyolites (Figure 4(b)).

3.2 $^{40}\text{Ar}/^{39}\text{Ar}$ dating

Muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ analyses for the Bukadaban leucogranites yielded a plateau age of 6.88 ± 0.19 Ma (Figure 5(a)), in agreement with its inverse isochron age of 6.61 ± 0.49 Ma within error margins (Figure 5(b)). The muscovite cooling age was slightly younger than that of the zircon U-Pb crystallization age (9.7 Ma).

Sanidines from the Malanshan rhyolites were dated by $^{40}\text{Ar}/^{39}\text{Ar}$ methods to yield a plateau age of 9.37 ± 0.30 Ma (Figure 5(c)), and an inverse isochron age of 9.29 ± 0.48 Ma within large error margins. This cooling age of 9.37 ± 0.30 Ma probably dates the eruption of the Malanshan rhyolitic lavas.

3.3 Whole-rock geochemistry

The leucogranites of Bukadaban were geochemically similar

Figure 3 Zircon CL images for the Bukadaban two-mica leucogranites (a) and the Kekao Lake rhyolites (b).

Figure 4 Zircon U-Pb concordia diagrams for (a) Bukadaban two-mica leucogranites and (b) Kekao Lake rhyolites.

in composition to the rhyolites of Kekao Lake, Malanshan and Hudongliang. All rocks analyzed were characterized by high SiO_2 (70.99%–73.59%), Al_2O_3 (14.39%–15.25%) and K_2O (3.78%–5.21%), but low MgO (0.11%–0.44%), Fe_2O_3 (0.61%–1.90%) and CaO (0.47%–1.19%). All rocks were high-K calc-alkaline in the SiO_2 vs. K_2O diagram (Figure 6(a)), contained normal corundum of 1.5–2.6 vol.% (determined by CIPW calculations), and were highly peraluminous rocks with high A/CNK ratios of 1.11–1.21 (Figure 6(b)). Differences among these units were: (1) the Kekao Lake rhyolites were relatively rich in sodium ($\text{Na}_2\text{O}/\text{K}_2\text{O} > 1$), whereas the others were rich in potassium ($\text{Na}_2\text{O}/\text{K}_2\text{O} < 1$); (2) the Malanshan rhyolites had higher TiO_2 (0.36%–0.38%) and CaO (1.14%–1.19%) than the other rocks (Table 4).

Although the rocks showed different chondrite-normalized REE fractionated patterns (Figure 6(c)), all rocks were characterized by clear negative Eu anomalies ($\delta\text{Eu}=0.18$ to 0.39). The Kekao Lake and Hudongliang rhyolites contained low Ba (27–46 ppm), Sr (16–101 ppm), Nb (12.8–17.1 ppm) and Zr (20–38 ppm) abundances, and had low total REE abundances ($\Sigma\text{REE}=23$ –36 ppm) and low LREE/HREE ratios [$(\text{La}/\text{Yb})_{\text{N}}=5$ –11]. The Malanshan rhyolites and the Bukadaban leucogranites had higher Ba (219–315 ppm), Sr (117–128 ppm), Nb (117–128 ppm) and Zr (100–186 ppm) abundances, and exhibited higher total REE abundances ($\Sigma\text{REE}=139$ –236 ppm) and higher LREE/HREE ratios [$(\text{La}/\text{Yb})_{\text{N}}=46$ –63]. However, all rocks were depleted in HREE ($\text{Yb}=0.30$ to 0.73 ppm) and Y (4.1–8.5 ppm). In the trace

Figure 5 $^{40}\text{Ar}/^{39}\text{Ar}$ step-heating spectra and inverse isochron age for the Bukadaban two-mica leucogranites (a), (b) and the Kekao Lake rhyolites (c), (d).

Table 4 Whole-rock major (%) and trace (ppm) element and Sr-Nd isotope results for the crustally-derived igneous rocks from the Hoh Xil Lake area^{a)}

Location	Kekao Lake				Bukadaban				Malanshan				Hudongliang			
Sample	2006K033	2006K033-1	2006K033-2	2006K034	2006K074	2006K074-1	2006K074-2	2006K074-3	2006K129	2006K130	2006K129-1	2006K129-2	HDL-1	HDL-2	HDL-3	HDL-4
Major elements																
SiO ₂	73.59	72.44	72.91	73.54	72.11	72.03	73.07	71.92	72.03	71.55	71.54	70.99	72.28	72.12	72.23	72.24
TiO ₂	0.06	0.07	0.06	0.06	0.21	0.19	0.21	0.22	0.36	0.37	0.37	0.38	0.08	0.08	0.08	0.08
Al ₂ O ₃	15.25	15.16	15.06	15.43	15.09	14.72	14.39	14.79	14.85	14.82	14.85	15.02	14.96	14.96	14.99	14.96
TF ₂ O ₃	0.58	1.04	0.98	0.61	1.34	1.89	1.87	2.11	1.35	1.56	1.90	1.89	1.78	1.75	1.64	1.77
MnO	0.08	0.09	0.10	0.09	0.01	0.02	0.02	0.02	0.00	0.01	0.01	0.02	0.07	0.07	0.07	0.07
MgO	0.11	0.11	0.11	0.11	0.31	0.31	0.32	0.34	0.41	0.44	0.44	0.44	0.11	0.11	0.11	0.11
CaO	0.62	0.75	0.65	0.59	0.79	0.80	0.71	0.77	1.14	1.19	1.15	1.18	0.47	0.47	0.47	0.47
Na ₂ O	4.47	4.53	4.53	4.52	3.47	3.34	3.42	3.34	3.20	3.22	3.48	3.32	3.97	4.05	4.03	3.96
K ₂ O	3.78	3.96	3.84	3.81	5.14	5.27	5.00	5.21	5.12	5.06	5.21	5.22	4.63	4.64	4.65	4.66
P ₂ O ₅	0.29	0.29	0.29	0.31	0.30	0.28	0.28	0.29	0.30	0.30	0.30	0.31	0.37	0.37	0.37	0.37
LOI	0.75	0.90	0.80	0.70	1.07	1.00	1.00	1.10	1.02	1.07	1.04	1.10	0.70	0.78	0.94	0.72
Total	99.6	99.3	99.3	99.8	99.8	99.8	100.3	100.1	99.8	99.6	100.3	99.9	99.4	99.4	99.6	99.4
CIPW																
Q	33	30	31	32	31	32	33	32	32	32	29	30	31	31	31	31
Or	24	24	23	24	32	32	30	32	32	32	32	32	28	28	28	29
Ab	39	40	39	39	30	29	30	29	28	28	30	29	35	35	35	35
An	1.3	3.9	3.3	1.1	2.1	4.1	3.6	3.9	1.8	4.0	5.8	6.0	2.4	2.4	2.4	2.4
C	2.2	2.0	2.3	2.3	2.0	2.2	2.2	2.4	1.8	1.7	1.5	1.9	2.6	2.5	2.5	2.6
Na ₂ O/K ₂ O	1.2	1.1	1.2	1.2	0.7	0.6	0.7	0.6	0.6	0.6	0.7	0.6	0.9	0.9	0.9	0.8
A/CNK	1.21	1.15	1.17	1.22	1.19	1.17	1.18	1.19	1.15	1.15	1.11	1.14	1.21	1.19	1.20	1.21
A/NK	1.33	1.33	1.34	1.37	1.38	1.28	1.29	1.32	1.31	1.34	1.32	1.36	1.30	1.28	1.28	1.29

(To be continued on the next page)

(Continued)

Location	Kekao Lake				Bukadaban				Malanshan				Hudongliang			
Sample	2006K 033	2006K 033-1	2006K 033-2	2006K 034	2006K 074	2006K 074-1	2006K 074-2	2006K 074-3	2006K 129	2006K 130	2006K 129-1	2006K 129-2	HDL -1	HDL -2	HDL -3	HDL -4
Trace elements																
Sc	1.4	1.4	1.1	1.3	1.8	1.6	1.7	1.7	2.7	2.9	3.2	3.1	2.0	1.8	1.8	1.7
V	2.4	1.8	1.8	2.5	7.2	7.4	7.1	7.8	21.1	23.0	23.3	23.5	2.1	2.4	2.2	1.1
Co	1.8	0.6	0.4	1.3	1.8	1.1	1.2	1.9	3.2	3.4	2.7	2.8	1.0	1.1	1.1	1.1
Ni	1.8	1.2	0.4	3.9	10.9	4.8	2.2	2.1	2.3	43.8	4.3	4.9	2.1	2.0	1.9	2.0
Ga	34	36	34	34	24	28	28	28	25	26	27	28	31	32	32	31
Rb	1118	1002	942	1081	387	428	397	421	372	382	405	413	815	828	850	817
Sr	53	54	41	101	119	127	128	124	117	125	126	128	16	16	17	16
Y	4.4	4.7	4.1	4.5	5.0	4.9	5.0	5.0	8.5	7.9	8.4	8.5	5.7	5.7	5.8	5.6
Zr	34	24	20	28	107	100	124	114	176	167	186	181	38	38	38	37
Nb	47.6	54.8	51.5	48.1	14.8	16.3	16.7	17.1	12.8	13.3	14.2	15.2	36.7	37.0	37.6	36.2
Cs	240	250	248	225	50	68	61	67	14	14	16	17	140	142	145	139
Ba	46	39	27	50	219	237	232	232	279	315	314	312	39	41	38	36
Hf	1.7	1.3	1.2	1.5	3.6	3.1	3.8	3.5	5.3	5.0	5.1	5.0	1.7	1.8	1.8	1.7
Ta	17.4	19.0	17.7	17.8	3.2	3.6	3.3	3.4	1.5	1.5	1.5	1.7	8.0	8.1	8.2	8.0
Pb	18.3	18.8	17.0	18.1	13.6	11.5	10.5	10.6	34.0	35.2	35.7	35.8	24.8	24.9	25.7	24.4
Th	5.1	4.3	3.6	6.3	34.4	5.6	1.1	2.2	45.4	50.7	0.0	0.1	11.0	10.9	11.3	10.7
U	13.3	5.0	5.4	17.1	15.6	31.4	34.4	33.0	6.9	8.1	48.6	46.9	39.0	39.6	40.6	39.4
La	5.4	5.2	5.2	6.4	31.4	33.7	30.3	29.8	47.8	49.5	51.3	50.0	6.9	7.0	7.1	6.8
Ce	9.3	10.2	9.9	11.9	61.9	71.8	65.6	63.8	100.5	104.5	110.6	108.2	15.3	15.3	15.7	15.0
Pr	1.1	1.1	1.0	1.3	7.6	8.0	7.4	7.2	12.0	12.3	12.3	12.1	1.7	1.7	1.8	1.7
Nd	3.5	3.7	3.4	4.2	26.8	28.2	26.4	25.6	42.8	45.7	44.4	44.0	6.0	6.0	6.3	5.9
Sm	0.68	0.66	0.61	0.79	4.75	4.82	4.62	4.52	7.18	7.57	7.58	7.52	1.36	1.31	1.39	1.32
Eu	0.08	0.06	0.05	0.07	0.34	0.32	0.29	0.31	0.56	0.57	0.57	0.57	0.08	0.09	0.09	0.07
Gd	0.53	0.54	0.49	0.59	3.12	3.01	2.84	2.95	4.63	4.79	4.78	4.67	1.14	1.12	1.13	1.12
Tb	0.10	0.10	0.08	0.11	0.36	0.31	0.32	0.32	0.54	0.54	0.50	0.49	0.20	0.21	0.21	0.20
Dy	0.63	0.64	0.58	0.69	1.35	1.22	1.23	1.28	2.02	2.06	1.99	2.01	1.05	1.07	1.11	1.04
Ho	0.14	0.13	0.11	0.14	0.19	0.15	0.17	0.16	0.31	0.31	0.29	0.29	0.17	0.17	0.17	0.16
Er	0.44	0.43	0.40	0.45	0.44	0.34	0.39	0.38	0.77	0.77	0.72	0.75	0.42	0.45	0.44	0.42
Tm	0.09	0.09	0.07	0.09	0.07	0.04	0.05	0.05	0.11	0.11	0.10	0.10	0.06	0.07	0.07	0.06
Yb	0.67	0.73	0.65	0.67	0.41	0.30	0.33	0.32	0.70	0.66	0.63	0.68	0.40	0.42	0.42	0.42
Lu	0.11	0.10	0.09	0.11	0.05	0.04	0.04	0.04	0.10	0.10	0.09	0.09	0.05	0.06	0.06	0.06
ΣREE	23	24	23	28	139	152	140	137	220	229	236	231	35	35	36	34
δEu	0.39	0.29	0.29	0.33	0.27	0.25	0.25	0.26	0.30	0.29	0.29	0.29	0.20	0.22	0.21	0.18
(La/Yb) _N	5	5	5	7	51	76	61	63	46	51	55	49	12	11	11	11
Nb/Ta	3	3	3	3	5	5	5	5	9	9	9	9	5	5	5	5
Rb/Sr	21	18	23	11	3	3	3	3	3	3	3	3	50	51	51	52
Sr-Nd isotopic data																
⁸⁷ Rb/ ⁸⁶ Sr	68.4			28.9	9.4				8.6	9.6						
⁸⁷ Sr/ ⁸⁶ Sr	0.723029			0.716120	0.713r649				0.714255	0.714166						
1σ	0.000016			0.000012	0.000011				0.000013	0.000012						
⁸⁷ Sr/ ⁸⁶ Sr _i	0.7143			0.7124	0.7125				0.7132	0.7129						
¹⁴⁷ Sm/ ¹⁴⁴ Nd	0.112			0.110	0.112				0.108	0.108						
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512336			0.512325	0.512342				0.512291	0.512267						
1σ	0.000014			0.000013	0.000012				0.000012	0.000014						
εNd (0)	-5.9			-6.1	-5.8				-6.8	-7.2						
εNd (9 Ma)	-5.8			-6.0	-5.7				-6.7	-7.1						
T _{2DM} (Ga)	1.29			1.31	1.29				1.36	1.40						

a) $T_{2DM} = (1/\lambda) \ln \{ 1 + [(^{143}\text{Nd}/^{144}\text{Nd})_s - (^{143}\text{Nd}/^{144}\text{Nd})_{DM} - ((^{147}\text{Sm}/^{144}\text{Nd})_s - (^{147}\text{Sm}/^{144}\text{Nd})_{DM}) \times (e^{\lambda t} - 1)] / [(^{147}\text{Sm}/^{144}\text{Nd})_c - (^{147}\text{Sm}/^{144}\text{Nd})_{DM}] \}$. $\lambda = 6.54 \times 10^{-11}$, $(^{147}\text{Sm}/^{144}\text{Nd})_c = 0.118$, $(^{147}\text{Sm}/^{144}\text{Nd})_{DM} = 0.21357$, $(^{143}\text{Nd}/^{144}\text{Nd})_{DM} = 0.513151$, $(^{147}\text{Sm}/^{144}\text{Nd})_{CHUR} = 0.1967$, A/CNK = molar $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$, N represents chondrite normalized.

Figure 6 (a) SiO₂ vs. K₂O diagram; (b) A/CNK vs. A/NK plot; (c) chondrite-normalized REE patterns; (d) initial Sr-Nd isotope diagram. Chondrite-normalized data are from [55]; data of mafic shoshonites from the Jingyu Lake area are from [31,34]; data of K-rich adakitic volcanic rocks are from [26,32,36]; data of crustally-derived igneous rocks in the Ulugh Muztagh area are from [21]; data of Miocene Himalayan leucogranites are from [1–5]; Sr-Nd isotopic data of Late Triassic flysch sediments are from [56]; A/CNK=molar Al₂O₃/(CaO+Na₂O+K₂O), A/NK=molar Al₂O₃/(Na₂O+K₂O).

element spider diagram (not shown), all rocks were enriched in LILE after normalization by primitive mantle, but were depleted in Nb, Ta, P and Ti. The REE patterns are identical to those of crustally-derived igneous rocks from the Ulugh Muztagh area, but are entirely different from shoshonitic or adakitic volcanic rocks in the Hoh Xil Lake area (Figure 6(c)).

Isotopic results are given in Table 4. Experimental results yielded a range of ⁸⁷Sr/⁸⁶Sr ratios from 0.713649 to 0.723029 and ¹⁴³Nd/¹⁴⁷Nd ratios from 0.512267 to 0.512342, respectively. The initial ⁸⁷Sr/⁸⁶Sr isotopic ratios for these rocks were recalculated to account for post-crystallization decay due to their high Rb/Sr ratios of 8.6–68.4 [57]. After age-corrected calculations (9 Ma), initial ⁸⁷Sr/⁸⁶Sr ratios ranged from 0.7124 to 0.7143. In addition, εNd (9 Ma) values for these rocks ranged from –5.7 to –7.1, with depleted-mantle model ages (*T*_{2DM}) of 1.21–1.27 Ga.

4 Discussion

4.1 Petrogenesis

All rocks analyzed in this study were characterized by high SiO₂ (70.99%–73.59%), Al₂O₃ (14.39%–15.25%) and K₂O

(3.78%–5.21%), but low MgO (0.1%–0.44%), Fe₂O₃ (0.61%–1.90%) and CaO (0.47%–1.19%) content. The rocks were composed of more than 1 vol.% corundum (Table 4) and were strongly peraluminous (Figure 5(b)), with an affinity to S-type granites [58]. These rocks are geochemically identical to those of crustally-derived igneous rocks from the Ulugh Muztagh area [21] and of leucogranites from the High Himalayan belts [1–5].

It remains theoretically possible to derive the studied rhyolitic and leucogranitic rocks from a less siliceous magma by intense fractional crystallization (FC) [27,28]. These rocks were characterized by higher initial ⁸⁷Sr/⁸⁶Sr ratios, but lower εNd (9 Ma) values that were distinguishable from shoshonitic or adakitic rocks nearby (Figure 6(d)). In addition, the rhyolites and leucogranites had lower Nb/Ta ratios than those of the shoshonites or adakites (Figure 7(a)). As expected, Nb/Ta ratios cannot be significantly changed by rock-forming minerals through fractional crystallization. In addition, Figure 6(d) revealed that leucogranites and rhyolites exhibited lower εNd (9 Ma) values than those of Triassic flysch wall-rocks. Thus, it seems unlikely that leucogranites and rhyolites were generated through FC or assimilation fractional crystallization (AFC) processes during mafic magma ascent and emplacement. Inherited zircons

Figure 7 (a) SiO₂ vs. Nb/Ta; (b) Rb/Sr vs. Rb/Ba diagram [59]; (c) An-Ab-Or plot [60], experimental melts are from [61]; (d) Rb/Sr vs. Ba diagram [3], Ms(FP): Fluids-present muscovite melting, Ms(FA): Fluids-absent (dehydration) muscovite melting, Bi(FA): Fluids-absent biotite melting.

with older ages were probably plucked from wall-rocks during magma ascent and emplacement, and/or sourced from metasedimentary protoliths (Figure 3). Thus, we propose that the rocks were probably produced by partial melting of metasedimentary rocks because their geochemical characteristics were identical to those of crustally-derived igneous rocks of Ulugh Muztagh [21]. Also, a clay-rich source is likely based on the high Rb/Sr and Rb/Ba ratios [59] in Figure 7(b).

Leucogranitic melts can result from one of the following three potential reactions in metapelites [61–63]: (1) fluids-present melting of muscovite at <750°C, $9\text{Ms} + 15\text{Pl} + 7\text{Qtz} + x\text{H}_2\text{O} = 31\text{Melt}$; (2) fluids-absent muscovite dehydration melting at 750–800°C, $22\text{Ms} + 7\text{Pl} + 8\text{Qtz} = 5\text{Kfs} + 5\text{Als} + 2\text{Bt} + 25\text{Melt}$; and (3) biotite breakdown after muscovite exhaustion at >850°C, $\text{Bt} + \text{Als} + \text{Pl} + \text{Qtz} = \text{Grt} + \text{Kfs} + \text{Melt}$. The leucogranites and rhyolites from the Hoh Xil Lake area were probably generated through muscovite dehydration melting without fluid infiltration, based on the following. First, experimental results revealed that melting of metapelites at high pressures (>8×10⁸ Pa) would accelerate the breakdown of plagioclase and quartz to generate melts of trondjemitic compositions [61], which are clearly different from the Hoh Xil leucogranites and rhyolites of granitic compositions in

the An-Ab-Or diagram (Figure 7(c)). Second, trace element Rb-Sr-Ba modeling suggests that melts with low Rb/Sr ratios (<1.5), in general, result from fluids-present melting [3]. Samples from the Hoh Xil Lake area exhibited high Rb/Sr ratios ranging from 3 to 52, indicating melting without fluids infiltration. Finally, the negative correlation between Rb/Sr ratios and Ba (Figure 7(d)) suggests that these rocks were probably derived from muscovite dehydration melting of metapelites [3].

Neogene shoshonitic lavas from the Hoh Xil area locally contained xenoliths and xenocrysts captured from the upper mantle or lower crust [23,27,39,64]. Siliciclastic granulites record melting of the lower crust and their protoliths were probably sourced from pelites or graywackes [64]. Unfortunately, xenoliths from volcanoes were not large enough to separate minerals for geochronology. Thus, there were no constraints on the timing of lower crust anatexis or metamorphism [27,39,64]. Based on the eruptive ages of shoshonites with xenoliths, it has been proposed that lower crustal melting began at least 15 Ma [39,64], which is consistent with the oldest age of 14.5–9.4 Ma for leucogranites and rhyolites from the Hoh Xil Lake area. Also, this indicates that lower crustal anatexis could date back to approx-

imately 15 Ma.

The metasedimentary granulites were dominated by (1) garnet+orthopyroxene+plagioclase+K-feldspar+quartz±biotite±sillimanite±hercynite±monazite±zircon [64]. The most aluminous sample was composed of (2) cordierite+sillimanite+orthopyroxene+quartz [64]. The group (1) metasedimentary granulites contained minerals that were identical to experimental melting work on the metapelites [61]. Leucogranites and rhyolites from the Hoh Xil Lake area were depleted in HREE and Y, probably because of garnets in the residue. Negative Eu anomalies were the result of abundant relic plagioclase and/or K-feldspar. These resultant minerals seemed to be in agreement with metasedimentary granulites [64]. The Kekao Lake rhyolites enrichment in Na may indicate more biotite in the residue [61]. The high CaO Malanshan rhyolites probably resulted from melting at deeper depths, thus incorporating more plagioclase into the melts at high pressures. The high TiO₂ Malanshan rhyolites may have contained Ti-Fe oxides, such as ilmenites or spinels captured from the residue during magma segregations. In general, REE behaviors were controlled by accessory phases, such as apatite and monazite [65,66]. The Bukadaban leucogranites and the Malanshan rhyolites were richer in total REE and LREE abundances compared to those of the Kekao Lake and Hudongliang rhyolites, indicating more apatite and monazite dissolution during anatexis.

Previous studies have revealed that High Himalayan leucogranites result from melting of metapelites in the thickened continental crust [1–5]. Models have investigated the effects of fluid infiltration [1], high radioactivity [17], shear heating [18] and tectonic decompression [19,61]. The high Rb/Sr ratios of the leucogranites and rhyolites preclude the possibility that crustal melting in the Hoh Xil area was accompanied by fluid infiltration (Figure 7(d)). Low conductive and high radioactive heating rates in the thickened Triassic flysch sediments have been proposed by [21] to generate crustally derived igneous rocks from the Ulugh Muztagh area by melting of metapelites. However, radioactive heating production rates may have been overestimated because abundances of heat-producing elements in Triassic flysch deposits (U<3 ppm; Th<11 ppm; K₂O<2.7 wt.%) found by [56] were significantly lower than the values used by [21] (U=10 ppm; Th=12 ppm; K₂O=3.6 wt.%). Contributions of heat into the middle and lower crust by mafic magma injections were minor because of the rarity of mafic materials near the Hoh Xil Lake area.

Considering the tectonic setting in the Hoh Xil Lake area, leucogranitic and rhyolitic melt generation was probably triggered by E-W extensional decompression. Previous studies have suggested that the reactivation of the South Kunlun sinistral strike-slip fault had begun by the Early Miocene, and resulted in generating a cluster of normal faults and pull-apart basins in its horse tails leading to E-W extension [38–40]. Reactions of muscovite breakdown triggered by extensional decompression resulted in deep melt-

ing of metapelites in the thickened middle or lower crust. Normal faulting or pull-apart basin development in the upper crust may have been critical to allow crustal melts to migrate from their sources, and to be emplaced or erupt at the surface [67]. E-W tectonic extension in the Hoh Xil area initiated at approximately 15 Ma [39], roughly coincident with N-S rifting time in southern Tibetan Plateau. A minimal age of 13.5 Ma for the onset of normal faulting was obtained by a mineral Rb/Sr isochron and ⁴⁰Ar/³⁹Ar dating in the Shuanghu rift located in southern Qiangtang terrane [68]. Additionally, outcrops of 18–13 Ma N-S trending dikes indicated that local extension in the Lhasa terrane could date back to 18 Ma [69]. Hydrothermal mica in a normal fault of the Thakkhola rift returned an age of ~14 Ma, indicating that E-W extension began prior to 14 Ma ago in northern Nepal [70]. The N-S rifting on the southern Tibetan Plateau generally has been interpreted as a result of extensional gravitational collapses or eastern extrusions along major strike-slip faults [71].

The Tibetan Plateau crust has nearly twice the thickness of a normal continent, which provides a necessary prerequisite for leucogranite generation. Graben and rift basins developed because of extensional effects in the horse tails along the large strike-slip faults in Tibetan Plateau [68–71]. In the Hoh Xil area, the distributions of Cenozoic volcanic rocks were clearly controlled by the South Kunlun sinistral strike-slip fault and its secondary faults [27]. Pull-apart or rift basins were well developed in the Hoh Xil area [38–40]. Rhyolite eruption and leucogranite emplacement were spatially associated with these faults in the Hoh Xil Lake area. This association further constrains the likelihood that leucogranitic and rhyolitic melts were generated through extensional decompression melting in the Hoh Xil Lake area.

4.2 Implications for timing of crustal anatexis and plateau uplift in northern Tibet

The ⁴⁰Ar/³⁹Ar geochronology obtained by previous studies suggest that leucogranite cooling ages range from 10.5 to 8.4 Ma, and that rhyolites had an eruptive age of 4.0±0.1 Ma [20]. These results indicate that crustal melting occurred from 10 to 4 Ma in northern Tibet [20]. The ages of leucogranites and rhyolites in this study from the Hoh Xil Lake area ranged from 14.5 to 9.4 Ma. This suggests that crustal melting may have initiated at 15 Ma.

Prior to the 65–50 Ma India-Asia collision [15,16], crustal thickening of the Tibetan Plateau was dominantly characterized by distributed shortening [72,73]. Also, the Hoh Xil crust was suspected to have been affected by compressional stress, producing significant crustal shortening and thickening. Geologic investigations have shown that the shortening rate for Paleogene strata of the Fenghuoshan and Yaxicuo groups reaches 43% [74], and that the Triassic flysch sediments thrust over the Paleogene terrigenous deposits [39] (Figure 1(b)). The reactivation of the South

Kunlun sinistral strike-slip fault during the Early Miocene [39,40], and the extrusive and intrusive ages of 14.5–9.4 Ma, indicate that NE-SW compression was suddenly replaced by localized E-W extension in the Hoh Xil area [39]. Furthermore, the ages of magma emplacement and eruption indicate that large-scale crustal thickening and shortening ceased prior to ~15 Ma. The N-S shortening could not have been completely accommodated by the E-W striking South Kunlun sinistral strike-slip fault. Thus, thickening was transferred to the northern Tibetan Plateau by the NE striking Altyn Tagh strike-slip fault. This resulted in the gradual development of thrust faults along the Akazishan, Danghenanshan and Qilianshan belts in northern Tibetan Plateau [73,75,76].

The timing and style of plateau uplift remains poorly constrained by several factors, such as rift basins [68,70], foreland basins [73,75–77], shoshonites [69,78], apatite fission track [79], paleobotany [80] and oxygen isotopes [81]. Leucogranite formation typically lags behind crustal thickening and topographic uplift. Thus, in this study, crustal melts of 14.5–9.4 Ma from the Hoh Xil Lake area are regarded as new evidence to define the timing of uplift on the northern Tibetan Plateau. Our geochronology results indicate that the Hoh Xil and Kunlun areas may have attained at or near their present elevations before at least 15 Ma. This is contemporaneous with the formation of the Miocene N-S rifts [68,70] and dikes [69] in southern Tibet. Together with the paleoelevation defined by fossil leaves [80], we propose that the entire Tibetan Plateau possibly may have reached or nearly reached its present elevation before 15 Ma.

Furthermore, the ages of 14.5–9.4 Ma for leucogranite emplacement and rhyolite eruption in the Hoh Xil area provide new constraints on the timing of plateau uplift and crustal shortening and thickening in northern Tibet. Large-scale crustal thickening and plateau uplift in northern Tibet occurred before 15 Ma, suggesting that the Hoh Xil Lake area probably attained its present high elevation before the Middle Miocene.

5 Conclusions

(1) A zircon U-Pb crystallization age of 9.7 ± 0.2 Ma was obtained for the Bukadaban leucogranites. The extrusive ages of Kekao Lake and Malanshan rhyolites are 14.5 ± 0.8 and 9.37 ± 0.30 Ma, respectively. Together with previous results, we propose that crustal anatexis in northern Tibet could be at least up to 15 Ma in age.

(2) Crustal melts were generated in the Hoh Xil Lake area by decompression melting, initiated by localized E-W extension developed in horse tails of Kunlun sinistral strike-slip faults.

(3) The timing of crustal thickening and plateau uplift are constrained by the crystallization and eruption ages of crus-

tal melts in the Hoh Xil area. Large-scale crustal thickening and plateau uplift in northern Tibet occurred before 15 Ma, suggesting that the Tibetan Plateau attained its present altitude before 15 Ma.

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