Mantle fluids in the Karakoram fault: Helium isotope evidence

Highlights:

- Mantle $^3\text{He}$ occurs in hot springs for >500 km along the Karakoram fault (KKF)
- Modeled vertical transport rates imply $\geq 3.5$ Ma ascent time from the mantle up the KKF
- The KKF accessed tectonically active (Tibetan) mantle at least into Pliocene time
- Where the KKF accesses Tibetan mantle, it limits northward underthrusting of India
- No “mantle helium” signature is seen associated with the Indus-Yarlung Suture Zone

Graphical Abstract:
Mantle fluids in the Karakoram fault: Helium isotope evidence

Simon L. Klemperer*a, B. Mack Kennedyb, Siva R. Sastryc, Yizhaq Makovskyd, T. Harinarayanae and Mary L. Leechf

aDepartment of Geophysics, Stanford University, CA 94305-2215, U.S.A.,
bCenter for Isotope Geochemistry, Lawrence Berkeley National Laboratory, CA 94720, U.S.A.,
cNational Geophysical Research Institute, Hyderabad - 500007, India,
dDepartment of Marine Geosciences, University of Haifa, Haifa 31905, Israel,
e now at: Gujarat Energy Research and Management Institute, Gandhinagar - 382007, India,
fDepartment of Geosciences, San Francisco State University, CA 94132, U.S.A.
*Corresponding author. E-mail address: sklemp@stanford.edu

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Abstract
The Karakoram fault (KKF) is the 1000 km-long strike-slip fault separating the western Himalaya from the Tibetan Plateau. From geologic and geodetic data, the KKF is argued either to be a lithospheric-scale fault with hundreds of km of offset at several cm/a, or to be almost inactive with cumulative offset of only a few tens of kilometers and to be just the upper-crustal localization of distributed deformation at depth. Here we show $^{3}\text{He}/^{4}\text{He}$ ratios in geothermal springs along a 500-km segment of the KKF are 3–100 times the normal ratio in continental crust, providing unequivocal evidence that a component of these hydrologic systems is derived from tectonically active mantle. Mantle enrichment is absent along the Indus-Yarlung suture zone (ISZ) just 35 km southwest of the KKF, suggesting that the mantle fluids flow only within the KKF. Within the last few Ma, the KKF must have accessed tectonically active Tibetan mantle northeast of the “mantle suture” which we therefore locate vertically beneath the KKF, very close to the surface trace of the ISZ. Hence, in southwestern Tibet, Indian crust may not now be underthrusting substantially north of the ISZ, even though Miocene underthrusting may have placed Indian crust north of the ISZ in the lower half of the Tibetan plateau crust. This is in significant contrast to central and eastern Tibet where underthrust Indian material not only forms the lower half of the Tibetan crust but is also currently underthrusting for ~200 km north of the ISZ. Our new constraint on KKF penetration to the mantle allows an improved description of the continuously evolving Karakoram fault, as a tectonically significant yet perhaps geologically ephemeral lithospheric structure.
1. Introduction
Observation of mantle contributions to $^{3}\text{He}$ at the surface requires both a source of helium in the mantle, and a path to the surface. In Tibet the mantle source is believed to be Tibetan mantle that is tectonically active (i.e., with incipient melting and likely deforming), as opposed to cratonic Indian mantle (Hoek et al., 2000). We propose that elevated $^{3}\text{He}/^{4}\text{He}$ ratios in hot springs along the Karakoram fault (KKF) (Fig. 1) demonstrate that within the last few Ma the KKF has channeled fluids to the surface from tectonically active Tibetan mantle. The KKF therefore marked the “mantle suture”, or northern limit, at the Tibetan Moho, of actively subducting Indian lower lithosphere. Our new data showing no mantle contamination of hot springs on the Indus-Yarlung suture (ISZ) complement our new and existing data showing clear mantle enrichment along the KKF and together imply focusing of the mantle fluids by the strike-slip fault system.

1.1. The Karakoram fault controversy and the northern limit of India
The structure of Tibet as a geodynamic response to the collision of India with Asia, and the distribution of deformation, are particularly contentious in west Tibet (for the purpose of this paper, west of about 82°E). Two continental-scale strike-slip faults – the dextral KKF and the sinistral Altyn Tagh fault (ATF) (Fig. 1) – play a disputed role in the eastward extrusion of Tibet since the onset of the India-Asia collision at ~57 Ma (Leech et al., 2005). Two extreme and opposing views exist for the mechanism(s) responsible for crustal accommodation of shortening across Tibet: (1) discrete tectonic blocks, internally relatively undeformed, are extruded eastward between lithospheric strike-slip faults (e.g. Tapponnier et al., 2001; Thatcher, 2007) and (2) deformation is continuously distributed within the lithosphere via ductile flow of the lower crust and upper mantle matched by brittle failure throughout a ubiquitously faulted upper crust (e.g. England and Molnar, 1997; cf. Beaumont et al., 2006).

The relative merit of these viewpoints in western Tibet has been assessed in part from studies of slip rate on the KKF and ATF. Geologic inferences of rapid slip up to 32 mm/a (Valli et al., 2008) and large offsets up to 555 km (Replumaz and Tapponnier, 2003) or ~300 km (Rolland et al., 2009) on the KKF seem to support plate-like behavior, and an important role for the KKF as a lithospheric structure. In contrast, geodetic measurements of slow or zero modern slip on the KKF (3±5 mm/a: Jade et al., 2004; 0–6 mm/a: Wang and Wright, 2012) seem to support continuous deformation of Tibet. Geologic claims for small total offset compatible with slow slip (66-150 km: Murphy et al., 2000; 40-150 km: Phillips et al., 2004) also suggest only a minor upper-crustal role for the KKF.

At a regional scale the depth of penetration of the KKF defines the tectonic interaction between India and Tibet. Hypothesized active channel flow (e.g. Beaumont et al., 2006) of mid-to-lower crust from Tibet into the northwest Himalaya driven by gravitational potential energy of the Tibetan plateau would require that the KKF is limited to the shallow crust (Phillips et al., 2004), terminating downwards at a partial melt/channel flow zone at c. 20 km depth above underthrusting Indian crust. Alternatively the KKF may reach into the middle crust to form a southern bound to this channel flow (Leech, 2008) or penetrate the whole lithosphere and
constrain the northern limit of the underthrust Indian plate below the Tibetan Moho (Rolland et al., 2009).

Seismic imaging of western Tibet has not yet imaged the KKF, as opposed to the ATF that is widely accepted from teleseismic imaging to be of lithospheric scale (Wittlinger et al., 2004), in part because the KKF is located close to disputed international boundaries. However, the northern limit of Indian crust beneath Tibet may constrain the depth of penetration of the KKF, now or in the past. Some tomographic images of west Tibet (Li et al., 2008) allow Tibetan mantle as far south as the KKF, consistent with but not requiring a crustal-penetrating KKF. However, most seismologic estimates of the northern limit of India (Nábělek et al., 2009; Priestley et al., 2008; Rai et al., 2006; Zhao et al., 2010) (Fig. 1b) nominally preclude penetration of the KKF into the mantle because they interpret an unbroken subthrust Indian crust, and infer this Indian crust is being subducted almost horizontally at the present day. A receiver-function “doublet” of Ps converters at ~80°E (Wittlinger et al., 2004) may mark the top and bottom of underthrust Indian crust forming the lower half of the Tibetan crust northward to about the Banggong-Nujiang suture (Nábělek et al., 2009). Similar Moho depths immediately north (Wittlinger et al., 2004) and south (Rai et al., 2006) of the KKF may imply underthrusting of Indian crust and mantle as far north as the (ATF) (Rai et al., 2006), and Sp receiver-function images at ~82°E have also been interpreted as imaging Indian mantle lithosphere directly beneath Tibetan crust across the full north-south extent of west Tibet (Zhao et al., 2010) (Fig. 1b). Finally, the distribution of ‘lower crustal’ earthquakes, including those at ~77°E, has been suggested to show “underthrust Indian Shield … underlies virtually the whole of NW Tibet as far as the ATF” (Priestley et al., 2008). Most of these studies do not separately delimit the northern edges of Indian crust and Indian lithospheric mantle; do not have the resolution to distinguish a possible thin layer of Tibetan mantle separating Indian lower lithosphere from Indian upper lithosphere; and do not have the ability to distinguish whether Indian lithosphere beneath central Tibet is currently subducting or alternatively has been stranded in place north of the active subduction system. However, these interpretations of the northern limit of Indian material at Moho depths (bold red lines in Fig. 1b) span the entire north-south width of Tibet showing that current seismic studies are not yet a reliable guide to the northern limit of India in west Tibet.

1.2. $^3$He as a tracer of mantle fluids

Noble gases, being inert, are excellent natural tracers for fluid migration and see through complex chemical processes affecting other more reactive species (e.g. Ballentine et al., 2002). Distinct mantle, crustal, and atmospheric sources are characterized by unique noble gas compositions, so their contributions can be identified and used to constrain the fluid’s history, and helium isotopes can provide unequivocal evidence for the presence of mantle-derived volatiles. Atmospheric helium has a $^3$He/$^4$He ratio $R_A = 1.4 \times 10^{-6}$, whereas crustal helium (dominated by radiogenic $^4$He), is characterized by a $^3$He/$^4$He ratio conventionally taken to be $0.02 \times R_A$, and upper-mantle helium (enriched in primordial $^3$He) has a value most often taken as $8 \times R_A$ (Ballentine et al., 2002).

Observation of mantle helium at the surface requires both a way to extract helium from the mantle, and a path to the surface. Mantle helium enters the crustal hydrologic system either by intrusion and degassing of mantle-derived magmas or by invasion of fluids degassed from mantle
melts. In order to release its $^3$He, the lithospheric mantle must be undergoing at least incipient melting because solid-state processes are too slow to enable extraction of noble gases (Ballentine et al., 2002). Upward transport within the crust may be accomplished by focused flow along major faults (Doğan et al., 2009; Kennedy et al., 1997), disseminated flux through a deforming ductile lower crust (Kennedy and van Soest, 2007; Newell et al., 2005), or shallow emplacement of deep-sourced magmas (Hilton, 2007) into which helium is strongly partitioned.

Observations of $^3$He/$^4$He ratios >0.02•$R_A$ can be due to at least four effects other than mantle fluids:

1. Air contamination: Helium in crustal fluids is not typically impacted by contamination by air or air-saturated water because helium has a very low atmospheric abundance (~5 ppm). Nonetheless, atmospheric contributions need to be considered, particularly for samples with low helium abundance. However, because both $^{22}$Ne and $^{36}$Ar in crustal fluids are dominated by atmospheric sources one can correct for atmospheric contributions to helium using either $^4$He/$^{36}$Ar or $^4$He/$^{22}$Ne ratios (Ballentine et al., 2002). The resulting corrected $^3$He/$^4$He ratios are conventionally denoted $R_C$.

2. Radiogenic $^3$He produced by the nuclear reactions $^6$Li($n$,α)$^3$H(β)$^3$He: The conventionally assumed continental crustal $^3$He/$^4$He ratio of 0.02•$R_A$ is based on zero input of mantle fluids and a Li concentration in the rock matrix of 40–60 ppm (Andrews, 1985). Recent estimates of 24 ppm as the average Li composition of the upper crust (Rudnick and Gao, 2005) might imply a better value of 0.01•$R_A$ so that any measured values >0.01•$R_A$ would imply some mantle influx. In this paper we conservatively use the conventional value of 0.02•$R_A$.

3. $^3$He produced by the decay of $^3$H that is a residual product of atmospheric hydrogen-bomb tests: The abundance of tritium in the environment has rapidly decreased since the 1963 Partial Test Ban Treaty due to the short (12-year) half-life of $^3$H, so that recent samples are scarcely affected.

4. Re-mobilized helium from older mafic intrusions within the crust: To dominate over a direct mantle source such intrusions would have to be mafic or ultramafic (derived directly from the mantle), relatively young (such that post-eruptive decay of U and Th has not diluted the initial elevated $^3$He/$^4$He ratio), and volumetrically huge (~$10^4$ km$^3$) to provide sufficient $^3$He to influence the helium signature of geothermal systems (Hoke et al., 2000).

During transport from the mantle source with $^3$He/$^4$He $\sim$8•$R_A$, the degree to which the $^3$He/$^4$He ratio is lowered (diluted) by the addition of radiogenic $^4$He depends on the mantle helium flux into the crust, the production rate of $^4$He in the crust, and the residence age or fluid flow rate within the crust. The integrated upward flow rate for steady-state one-dimensional flow scales with the crustal thickness and can be calculated from the measured $^3$He/$^4$He ratio and $^4$He concentration in the sampled fluids, using conventional values for helium isotopic ratios in the crust (0.02•$R_A$) and mantle (8•$R_A$), and the present-day crustal production rate of $^4$He from Th and U (Kennedy et al., 1997). Conventionally, any corrected $^3$He/$^4$He ratio $R_C$•$R_A$>0.1 is considered to have an unequivocal mantle component (Ballentine et al., 2002) reflecting recent transport from the mantle.
1.3 Association of spatial patterns of $^{3}$He/$^{4}$He ratios with different tectonic styles

1.3.1 Continental strike-slip faults

In our study of the KKF, the helium isotopic signature of continental strike-slip faults is of particular interest. Plate-boundary strike-slip faults that penetrate from earth’s surface to the mantle are obvious pathways for migration of mantle fluids, and both the San Andreas fault (SAF; Kennedy et al., 1997; Kulonenoski et al., 2012) and the North Anatolian fault (NAF; Doğan et al., 2009; Güleç et al., 2002) have $R_{C}/R_{A}$ ~4 in some springs along these faults or their active splayes (Fig. 2a). However, even closely spaced springs can exhibit very different degrees of crust-mantle interaction suggesting input from different hydrologic systems. Some groundwater samples separated along the SAF by <10 km have $R_{C}/R_{A}$ values varying by a factor >20 (Kulonenoski et al., 2012), so it is customary to focus on the most-enriched samples that likely have the most geodynamic significance (envelopes of data points in Figure 2). The localization of increased $^{3}$He/$^{4}$He to within one crustal thickness (~30 km) of both the SAF and NAF, and the ratio of the highest $^{3}$He/$^{4}$He values close to the SAF and NAF to the lowest values away from these faults (~40) is a strong argument for fault-focused fluid flow penetrating the whole crust along the SAF and NAF. Simple 1D steady-state flow models imply time-averaged upward flow rates of the $^{3}$He-enriched fluid in the SAF of up to 11 mm/a (Kennedy et al., 1997) or even ~147 mm/a (Kulonenoski et al., 2012) and of up to 9 mm/a in the NAF (de Leeuw et al., 2010), suggesting high effective permeability in these intra-continental transforms.

1.3.2 Distributed intra-continental deformation

In both the western USA (Kennedy and van Soest, 2007; Newell et al., 2005) and Turkey (Güleç et al., 2002; Doğan et al., 2009; de Leeuw et al., 2010) even the lowest $^{3}$He/$^{4}$He ratios measured require that some mantle fluids are traversing the crust that is both thinner and hotter than global averages (e.g. Gilbert, 2012; Sass et al., 1997; Mutlu and Karabulut, 2011; İlkışık, 1995). Across the western USA, regions of highest $^{3}$He/$^{4}$He correspond to regions of lowest seismic wavespeed in the mantle, reflecting tectonically active and partially molten mantle (Newell et al., 2005). Far from the SAF and NAF, smaller quantities of mantle volatiles measured at the surface have traversed a ductile lower crust, without the aid of crust-penetrating faults or shallow emplacement of recent melts. This upward transport requires at least transient permeability in the lower crust. In the Basin and Range Province, western USA, Kennedy and Soest (2007) show that baseline $^{3}$He/$^{4}$He ratios increase systematically to the west, correlating to higher rates of crustal deformation (Figure 2a), and they hypothesize that the increase in total strain and specifically the northwest-oriented dextral shear component, greatly enhance average fluid flow rates. This correlation of enhanced $^{3}$He/$^{4}$He with increased (shear) strain offers a possible model for interpreting other intra-continental zones of distributed deformation.

1.4. Previous helium isotope observations from Tibet and inferences on Tibetan tectonics

Observations of mantle helium at earth’s surface require at least incipient mantle melting to either degas $^{3}$He or to produce melts that transfer $^{3}$He directly into the crust (Ballentine et al., 2002). Hence helium data in Tibet (Hoke et al., 2000) have been used to probe the boundary between Indian cratonic mantle that remains far below the solidus and largely undeformed during its rapid
northward subduction (e.g. Klemperer, 2006; Priestley et al., 2008), and the orogenic mantle further north known from seismic studies to be strongly deformed, highly attenuating, and close to its melting point (e.g. Barron and Priestley, 2009; Klemperer, 2006; Zhao et al., 2010). The large south-to-north increase in seismic wave speed across the northern edge of the high Tibetan plateau both immediately beneath the Moho (Liang and Song, 2006) and in the upper mantle (Liang et al., 2012) suggests there are separate Tibetan and Asian mantle lithospheres with distinct tectono-thermal histories, and so we refer to the tectonically active mantle north of subducting Indian cratonic mantle and south of the ATF/Kunlun fault systems as “Tibetan”.

Previous helium isotope studies in Tibet focused on a region from ~86–92°E south of the Banggong-Nujiang suture zone (for the purposes of this paper, hereafter called southeast Tibet) (Hoke et al., 2000; Newell et al., 2008; Yokoyama et al., 1999; Zhao et al., 2001), including the Nagqu, Yangbajain, Tingri and Daggyai Co regions (Fig. 1). Additional papers in the Chinese literature contain helium data but lack details of other noble gases or corrections for air-contamination (e.g., Hou and Li, 2004; Zhao et al., 2002). In west Tibet, 80–81°E, helium isotopes have been reported from just three hot spring systems (Hoke et al., 2000; Zhao et al., 2002). Hoke et al. (2000) suggested the association of mantle helium with the KKF, but their Tirthapuri samples (Fig. 1) were collected from a location within 1 km of both the KKF and the Indus-Yarlung suture zone (ISZ), making it impossible to separate the effects of these structures. In this paper we use additional samples from NW India where the KKF and ISZ are separated by 35 km to show the mantle helium is associated with the KKF and is not present in the ISZ.

In southeast Tibet Hoke et al. (2000) used all the available data (Fig. 2b, blue line) to show gradual spatial variation between a “crustal helium domain” in the Himalaya in which \( R_C/R_A < 0.05 \) (blue field, Fig. 3) and a “mantle helium domain” in the northern Lhasa block in which \( R_C/R_A > 0.1 \) (yellow field, Fig. 3). Because in Tibet even the highest corrected \(^3\)He/\(^4\)He ratios were \(< R_A \), Hoke et al. (2000) suggested that between these two unambiguous domains, the “transition zone” in which \( 0.05 < R_C/R_A < 0.1 \) \(^3\)He/\(^4\)He ratios between 2.5 and 5 times the canonical crustal ratio) represents a gradually increasing mantle influx of \(^3\)He. The transition, 50–100 km north of the ISZ (near \( Y \) in Fig. 1a and 2b), is interpreted as the southern limit of tectonically active Tibetan mantle in contact with subducting India (the “mantle suture”: Hoke et al., 2000). Although existing data from the relatively few available geothermal sites could permit a very rapid transition from “crustal” to “mantle” domains, such a rapid transition has not been inferred because it would occur in the middle of the Gangdese (“Andean”) batholith in a region with no obvious Cenozoic west-east structural or magmatic features.

The majority of the \(^3\)He-enriched samples studied by Yokoyama et al. (1999), Hoke et al. (2000) and Zhao et al. (2001) come from the Yangbajain graben of the Yadong-Gulu rift system. The Yangbajain graben is also the locus of well-studied seismic brightspots (Makovsky and Klemperer, 1999) and electrical conductivity anomalies (Li et al., 2003) interpreted as aqueous fluids above magma chambers at 15 km depth within the rift. Hence Hoke et al. (2000) interpreted the mantle-derived \(^3\)He to represent degassing of volatiles from young (Quaternary) mantle-derived melts intruded into the crust, and the southern limit of the mantle helium domain as defining the southern limit of recent mantle melt extraction beneath the Tibetan plateau.
2. New observations of $^3$He/$^4$He in western Tibet

We present five new helium isotope measurements made in northwest India on and just southwest of the KKF at ~78°E, and two new isotopic analyses from west Tibet on and just northeast of the KKF at ~80°E (Fig. 1). Our data complement the only previous helium isotope data from western Tibet, three samples from two locations close to the eastern end of the KKF (Hoke et al., 2000). Our data collection and analysis followed standard procedures (Kennedy and van Soest, 2006) (Appendix). Table 1 presents our field observations and noble gas isotope data, including new R/R$_A$ values of 0.01, 0.02, and 0.02 close to the ISZ, 35–50 km southwest of the KKF; values of 0.1, 0.7, and 2.2 on the KKF; and a value of 0.19 45 km northeast of the KKF.

The sample with R/R$_A$=2.2 has unambiguous $^3$He enrichment irrespective of any atmospheric contamination. However, corrections to the measured helium isotopic ratios for contamination by air or air-saturated water are important for our samples with low absolute helium abundance and with R<R$_A$. We corrected all samples for atmospheric contributions using two methods (Table 1). The first method assumed all of the measured $^36$Ar was from air contamination (i.e. no contribution from air saturated water) and the helium isotope ratios were corrected using the measured $^4$He/$^36$Ar ratios and the known $^4$He/$^36$Ar ratio in air. This approach results in a maximum correction and, therefore, lowest corrected helium isotope ratios (Table 1). The second method used a more conventional procedure as followed by Hoke et al. (2000), Newell et al. (2008), Walla et al. (2005), and Yokoyama et al. (1999). This method uses the measured $^4$He/$^{22}$Ne ratios, assumes all of the $^{22}$Ne was derived from air and/or air saturated water and presumes that little or no He/Ne elemental solubility fractionation has occurred due to boiling or gas effervescence, relying on the similar solubility of He and Ne in water. This approach results in corrected helium isotope ratios intermediate to the He/Ar method and the measured ratios (Table 1).

Elemental fractionation of the heavier noble gases (Ne, Ar, Kr, Xe) is known in multi-phase geothermal systems due to boiling and steam separation (e.g. Pinti et al., 2013), and is expected for effervescence of gas (e.g. CO$_2$) saturated fluids. Since neither the $^36$Ar nor the $^{22}$Ne method for correcting the helium isotope ratios fully accounts for potential elemental fractionation, both methods are approximations. However, in the case of our samples, there has been only minimal noble gas fractionation as shown by fractionation factors (F values) that are all close to the anticipated values for air-saturated water (ASW) at 10°C (e.g. F($^{132}$Xe) values of 3.9 to 7.2 compared to the expected value of 3.7) (Table 1; Appendix). Indeed, the fact that the F($^{132}$Xe) values are all slightly above the F-value for 10°C ASW suggests that our sampled fluids have lost a small amount of noble gases, so that the F($^3$He) values are lower limits (He being less soluble than Ar, Kr and Xe), hence our air corrections to R/R$_A$ are over-estimated, and our estimated R$_C$/R$_A$ values are minimum values (Appendix). These minimum R$_C$/R$_A$ values are 0.01, 0.02, and 0.02 35–50 km southwest of the KKF; 0.07 0.66, and 2.24 on the KKF; and 0.19 45 km northeast of the KKF (Fig. 2b; Table 1).

The Mengshi sample with R$_C$/R$_A$=2.24, like the Hoke et al. (2000) Tirthapuri samples with R$_C$/R$_A$=0.30–0.38, is unequivocal evidence for mantle fluids in the KKF and/or the ISZ where the
KKF intersects the ISZ at ~81°E (Murphy et al., 2000) (Fig. 1). Our claim that the mantle fluids are associated with the KKF rather than the ISZ relies on our low observed $R_C/R_A$ values (no mantle component) on the ISZ 350 km NW of Mengshi, and an unambiguous albeit small mantle contribution to fluids on the KKF, 500 km NW of Mengshi. The high $R_C/R_A$ for Changlung makes this sample most critical in our discussion of mantle contributions to the surface fluids at 77.5°E. The very low helium abundance in our Changlung sample resulted in relatively large analytical uncertainty in the measured $^{3}$He/$^{4}$He ratio [large error bar in Fig. 2b; Table 1] and a low He enrichment factor with respect to air leading to a relatively large air correction. However, plotting $R/R_A$ against the $^{4}$He/$^{22}$Ne ratio (Fig. 3) demonstrates that this Changlung sample lies above the mixing line between canonical crustal and air end-members at the 3σ confidence level, and records a clear mantle-derived He contribution with 33 times the $^{3}$He/$^{4}$He ratio nominally expected for continental crust. Thus Changlung, Tirthapuri (Hoke et al., 2000) and Mengshi (collected by Zhao et al., 2002; isotopic data reported here) samples that were all collected on the KKF, all show a mantle signature. This contrasts with our samples from the ISZ that have <0.2% mantle contamination (Fig. 3).

As discussed in the Introduction, elevated $^{3}$He/$^{4}$He ratios could in unusual circumstances be due to radiogenic $^{3}$He produced from $^{6}$Li or by decay of tritium from atmospheric nuclear testing, or due to re-mobilisation of $^{3}$He from older mantle-sourced intrusions in the crust. We further assess these potential contributions based on the geologic setting of our samples and comparison with previous $^{3}$He studies in Tibet. One of our sampled springs (Puga) rises through up to 65 m of unconsolidated sediments (Harinarayana et al., 2006) in which borax deposits have alkali enrichment up to 4500 ppm Li derived from subjacent magma bodies (Chowdhury et al., 1974). Nonetheless this sample has the lowest $^{3}$He/$^{4}$He ratio yet reported from the Himalaya, only $0.01\cdot R_A$ (Fig. 3). We conclude that radiogenic $^{3}$He produced from $^{6}$Li is not a likely source of our observed elevated $^{3}$He/$^{4}$He ratios. Our Puga sample also has the highest F($^{4}$He) value, perhaps suggesting an association of the geothermal fluids with the inferred magma body at ~2 km depth (Harinarayana et al., 2006) and that this intrusion may be unusually radiogenic – Miocene Himalayan leucogranites are characterized by bulk uranium contents more than twice that of average granitic rocks (Cuney et al., 1984) and extraordinarily high uranium content of their zircons (e.g., Leech, 2008). The high F($^{4}$He) in the Chumathang sample suggests a similarly uraniferous granitic magma at shallow depth could be responsible for the Chumathang geothermal field. Yokoyama et al. (1999) analyzed their 1991 geothermal samples from southeastern Tibet for tritium but found no $^3$H at their detection limit of 1 tritium unit (T.U.; $^3$H/H = $10^{-18}$), and showed that this maximum level was not a significant source of $^3$He. Using the same argument as Yokoyama et al. (1999), and assuming 1 T.U. in our 2008 Changlung fluids, less than ~10-15% of the measured $^3$He could be derived from tritium.

Re-mobilisation of $^3$He from older intrusions is also an implausible source of the observed mantle helium anomaly along the KKF. Simple calculations (Hoke et al., 2000) show the size of such intrusions would be larger than the small distances across which we report large changes in $^3$He/$^4$He ratios. Although fluid inclusions in an olivine xenocryst from a ~18 Ma ultrapotassic dyke in southwestern Tibet yielded an elevated $^3$He/$^4$He ratio of 0.17$\cdot R_A$ (Hoke et al., 2000), such dykes are too volumetrically minor, too old, and too little enriched in $^3$He to be the source of the
He we measured. The samples from which we report elevated $^3\text{He}/^4\text{He}$ ratios (mantle helium domain: Changlung, intermediate domain: Panamik) were collected from the margins of the Karakoram leucogranite, a Miocene intrusion formed by crustal melting (Phillips et al., 2004). In summary, the enrichment of $^3\text{He}/^4\text{He}$ above the canonical crustal value of 0.02•$R_A$ in the Mengshi, Changlung, and Panamik samples reported here, and the Tirthapuri samples of Hoke et al. (2000), can only plausibly represent a small but real and geologically recent flux of mantle-derived fluids along the KKF.

3. Interpretation of $^3\text{He}/^4\text{He}$ ratios along and across the KKF

Hoke et al. (2000) and Zhao et al. (2002) sampled just two locations, on and north of the KKF, and Walia et al. (2005) sampled three closely spaced locations all structurally below/south of the Main Central Thrust. Our new samples, coupled with the older data, let us test for the first time the close spatial relationship of $^3\text{He}/^4\text{He}$ anomalies to the KKF (Fig. 2b).

All four spring systems yet sampled on the KKF (two by us in the Nubra Valley and two by Hoke et al. (2000) and Zhao et al. (2002) >500 km to the southeast at Tirthapuri) exhibit a measurable mantle helium signal. In contrast, all three samples just 35–50 km southwest of the KKF completely lack evidence of this mantle enrichment (Fig. 2b). The three samples 35–50 km southwest of the KKF were obtained close to the ISZ and to the active Tso Morari normal fault (Fig. 1b). Hence mantle fluids are not widely disseminated through a fractured, continuously deforming upper crust in this region, but are apparently transported from the mantle to the surface by focused flow within the KKF.

The ratio of the maximum $^3\text{He}/^4\text{He}$ values on the KKF to the maximum $^3\text{He}/^4\text{He}$ values SW of the KKF is ~100, and to $^3\text{He}/^4\text{He}$ values NE of the KKF is ~8 (red line, Fig. 2b), comparable to the equivalent ratios (~40) seen across the SAF and NAF (Doğan et al., 2009; Kennedy et al., 1997) (Fig. 2a). This enrichment in mantle $^3\text{He}$ occurs over <35 km on the southwest side of the KKF and <45 km on the northeast side (far less than the crustal thickness of c. 70 km: Rai et al. (2006)). Similarly, major enrichment in mantle $^3\text{He}$ is only present within ~25–50 km of the SAF and NAF (Doğan et al., 2009; Kennedy et al., 1997), or 1–1.5 times the respective crustal thicknesses of 25-30 km (Gilbert, 2012) and 30–35 km (Mutlu and Karabulut, 2011). The lower $R_C$ values in Tibet than in California and Turkey may correspond to the far thicker crust in Tibet (Rai et al., 2006) allowing for more dilution of the mantle helium component by radiogenic crustal helium. The 10-fold change in $R_C/R_A$ between the Panamik and Changlung hot springs that are separated by 16 km, and the similar change between the Mengshi and Tirthapuri locations that are separated by 18 km, no doubt representing different degrees of interaction of crustal and mantle fluids in separate hydrologic systems, is less than the largest changes between adjacent measured springs on the SAF, where springs separated by <10 km can differ in $R_C/R_A$ by a factor >20 (Kulongoski et al., 2012). Time-averaged upward flow rates of helium have been calculated for the SAF in central California (up to 11 mm/a: Kennedy et al., 1997) and southern California (up to 147 mm/a: Kulongoski et al., 2012), and for the NAF (up to 9 mm/a: de Leeuw et al., 2010). Using the same model corrected for the much thicker crust, we derive time-averaged
upward flow rates of 12 mm/a and 19 mm/a (for Changlung and Panamik respectively), implying a 3.5–6 Ma transit time from mantle to surface through the c. 70 km-thick crust. The high flow rate estimated for the KKF suggests near-vertical transport, as lateral transport would increase the distance traveled and hence the inferred transport rate. We conclude that an active near-vertical KKF accessed fluids from a tectonically active mantle at least until the latest Miocene (c. 6 Ma) or early Pliocene (c. 3.5 Ma), and possibly to the present day.

In summary the comparison of spatial gradients of \(^3\)He/\(^4\)He ratios from the ISZ to the KKF with spatial gradients across the SAF and NAF (Fig. 2), and of upward flow rates within all these faults, suggests similar processes of fault-controlled fluid flow from mantle to the surface. Northeast of the KKF only two (essentially collocated) data points exist, suggesting that the KKF is a singular feature within Tibet but insufficient to refute the possibility that the KKF is merely the southwestern edge of a broader anomaly. Either hypothesis is consistent with our conclusion that the KKF is a dominant pathway for mantle fluids to reach the surface.

4. Comparison of west (77–81°E) and southeast Tibet (86–92°E) and implications for India-Asia collision

Comparison of datasets from west to east across Tibet (Fig. 2b) shows the highest \(^3\)He/\(^4\)He ratio on the KKF is nine times the highest value reported by Yokoyama et al. (1999), Hoke et al., (2000) or Newell et al. (2008) from southeast Tibet, and that the horizontal transition from “crustal” to “mantle” helium domains likely occurs over a far shorter distance at the KKF than in southeast Tibet. This short spatial transition for an order of magnitude increase in R_C/R_A is evident both at Mengshi/Tirthapuri and 500 km northwest at Changlung/Panamik (solid line, Fig. 2b). Upward transport of mantle fluids is focused in the NW Himalaya by the KKF but unfocused in southeast Tibet.

The low spatial variation in \(^3\)He/\(^4\)He ratios across southeast Tibet, when compared to results from the SAF, NAF, and KKF (Fig. 2), implies that no single fault cuts through the crust of southeastern Tibet across the helium transect. The occurrences of mantle helium in southeastern Tibet are not uniquely associated with active faults. Although several enriched samples come from the Yadong-Gulu rift, the most \(^3\)He-enriched samples from southeast Tibet come from Nagqu (Hoke et al., 2000; Yokoyama et al., 1999), outside the Yadong-Gulu rift and north of the Jiali strike-slip fault (Fig. 1) (Klemperer, 2006).

Hoke et al. (2000) interpreted the \(^3\)He enrichment in southeastern Tibet to represent degassing of volatiles from Quaternary mantle-derived melts emplaced at shallow levels in the crust, based on the geophysical evidence for magma chambers at 15 km depth (Li et al., 2003; Makovsky and Klemperer, 1999) beneath some of their samples (Fig. 4a). However, more recent observations demonstrate that other crustal magma chambers in Tibet are not associated with elevated \(^3\)He/\(^4\)He ratios. A seismic low-velocity anomaly indicative of partial melt at 15-20 km depth (Hetenyi et al., 2011) is located vertically beneath four hot springs and fumaroles that lack any mantle \(^3\)He signature at Daggyai Co just north of the ISZ (Hoke et al., 2000). Similarly, we found only crustal helium in the Puga hot springs sampled directly above a magma body attested to by heat flow, seismicity, and magnetotelluric data (Harinarayana et al., 2006) (Figs. 4b). The absence of a
mantle helium signature suggests the Puga and Daggyai Co inferred magma bodies are intra-crustal melts, and strongly suggests shallow anatexis, not mantle melting, is the origin of these shallow magma chambers.

[Figure 4 near here]

As an alternative to the Hoke et al. interpretation, the slow spatial variation of $^{3}\text{He}/^{4}\text{He}$ ratios south-to-north across southeast Tibet (Fig. 2b) may be understood via comparison with $^{3}\text{He}/^{4}\text{He}$ ratios that vary east-to-west across the western USA with similar spatial gradients (Fig. 2a). In the Basin and Range province, mantle-derived helium passes through a lithosphere that appears to lack crust-penetrating normal faults and also lacks widespread active mantle-sourced volcanism (Hilton, 2007; Kennedy and van Soest, 2007). The westward increase of the baseline trend of $^{3}\text{He}/^{4}\text{He}$ across the Basin and Range province has been correlated with the westward increase in dextral shear strain (Kennedy and van Soest, 2007). Similarly, the northward increase of $^{3}\text{He}/^{4}\text{He}$ across southeast Tibet at $\sim$86–92°E correlates with the northward increase in dextral shear (eastward escape) of Tibet with respect to India and Eurasia (Zhang et al., 2004). Thus $^{3}\text{He}/^{4}\text{He}$ ratios up to 0.25•R$_A$ in the northern Lhasa block may represent diffuse passage of fluids from Tibetan mantle upwards through continuously deforming, very weak, albeit very thick crust (Klemperer, 2006) (Fig. 4a).

In southeastern Tibet, metamorphosed Indian crust likely forms the lower half of Tibet’s thick crust for $\sim$200 km north of the Indus suture zone (Fig. 4a), as shown by the receiver-function “doublet” of positive converters (seismic wavespeed increasing downwards), of which the deeper is the Moho, on transects crossing the ISZ at 85°E and 90°E (Kind et al., 2002; Nábělek et al., 2009). The shallower converter, interpreted as the top of the underthrusting Indian lower crust (Kind et al., 2002; Nábělek et al., 2009), extends up to 100 km north of the mantle suture at 90°E as located by $^{3}\text{He}/^{4}\text{He}$ ratios and receiver-function images (Hoke et al., 2000; Kosarev et al., 1999) (Fig. 4a). If underthrusting Indian lower crust has occupied the same region over the time scales of upward transport of helium from the mantle (a few million years) then the observed $^{3}\text{He}$-enriched fluids observed in southeastern Tibet must have passed through the underthrust Indian crust, which must therefore be permeable hence likely weak and actively deforming, as opposed to cold, rigid (Priestley et al., 2008), and impermeable. This in turn suggests that the northern prolongation of underthrust Indian crust is in a physical state consistent with its participation in channel flow (Beaumont et al., 2006) (upper black arrow in Fig. 4a). A stronger component of eclogitized Indian lower crust may be subducting with the Indian lower lithosphere (Beaumont et al., 2006) (lower black arrow, Fig. 4a) where rare sub-Moho (Chen and Yang, 2004) or lower-crustal (Priestley et al., 2000) earthquakes occur.

5. Implications for the evolution of the Karakoram fault
The unequivocal observation (Hoke et al., 2000; Zhao et al., 2002; this work) of mantle-derived fluids along the KKF (Fig. 2b) strongly suggests the KKF reaches the Moho. These mantle fluids are not widely disseminated through a fractured, continuously deforming crust, because – based on our current data set – neither the ISZ nor the active Tso Morari normal fault, show any evidence of mantle helium (Figs. 2b, 3). Hence the KKF must reach significantly deeper in the
crust than these other faults, presumably to the Moho or beyond, to access tectonically active mantle lithosphere north of the India-Tibet mantle suture (Fig. 4b). The KKF may well broaden in the lower crust to >20 km width, as suggested by studies of exhumed ductile shear zones that acted as pathways for mantle fluids (e.g., Pili et al., 1997).

In southeastern Tibet, metamorphosed Indian crust likely forms the lower half of Tibet’s thick crust for ~200 km north of the Indus suture zone (e.g., Klemperer, 2008) (Fig. 4a). The same lithospheric configuration may have been present in western Tibet prior to the initiation of the KKF, and may have left Indian crust stranded north of the KKF, cut off from subducting India by a subsequently active lithospheric-scale KKF (Leech, 2008) (hatched region in Fig. 4b). Seismic interpretations of Indian crust forming the lower crust of Tibet far north of the KKF (Nábělek et al., 2009; Priestley et al., 2008; Rai et al., 2006; Zhao et al., 2010) can be understood only if this crust is not presently being underthrust to the north but is instead stranded north of the active KKF, and/or if this Indian crust is permeable hence likely weak and actively deforming. In addition, tectonically active Tibetan mantle must be present immediately below the crust of the Tibetan Plateau, hence above subducting Indian cratonic lower lithosphere (Figure 4b). Our helium data imply the mantle suture is close to the KKF as hinted at by some previous studies (dashed mantle boundary in Fig. 4b: see Zhao et al., 2010, their Fig. 2).

The $^3$He-enriched fluids at the surface today require that the KKF accessed the mantle at 6 to 3.5 Ma (estimated transit time of mantle fluids to surface), and possibly also earlier and later. This age is consistent with the most recent assessments of KKF initiation at 25–15 Ma (Leech, 2008; Phillips et al., 2004; Styron et al., 2011; Valli et al., 2008) in the central segment near the BNS, and southward propagation of the KKF to its southern tip after 13 Ma (Styron et al., 2011; Valli et al., 2008). Our estimated upward flow rates are also consistent with geologic observations that the northern KKF became inactive in the Late Pliocene, following southwest propagation of the ATF along the Longmu Co-Gozha Co fault system to meet the KKF at ~34°N (Robinson, 2009) (Fig. 1b). The $^3$He-enriched fluids we collected in the Nubra Valley at 35°N originated in the mantle between 6 and 3.5 Ma but traveled up a segment of the KKF that may no longer be active nor form a continuing barrier to northward subduction of India. The earthquakes 30–80 km northeast of the KKF at 85–95 km depth that have very diverse focal mechanisms and are only recorded north of 36°N (Priestley et al., 2008) (Fig. 1b), may mark a return to the underthrusting of Indian crust (Fig. 4a) following a limited period of strike-slip truncation of the MHT (Fig. 4b). In contrast, the $^3$He-enriched fluids collected at Mengshi (Zhao et al., 2002) and Tirthapuri (Hoke et al., 2000) are on two separate strands of the KKF (separated by 20 km in a northeast direction) both still active based on satellite imagery (Murphy et al., 2000; Murphy and Burgess, 2006).

The most significant difference between our cross-sections in Fig. 4 is in the northern limit of active underthrusting of Indian crust beneath Tibet. It is unclear whether this represents control by the KKF, or whether along-strike changes in the Himalayan belt give rise to these differences in crustal structure that in turn help control evolution of the KKF. In numerical thermomechanical models of channel flow extrusion of the Himalaya from beneath the Tibetan Plateau, the distance that Indian crust underthrusts Asian crust north of the ISZ is sensitive to small changes in vertical crustal strength profiles (Jamieson et al., 2006) and to cross-strike changes in
crustal strength (Beaumont et al., 2006). For a range of models that show well-developed channel flow, Indian crust may underthrust 200 km north of the ISZ (cf. Fig. 4a), or not underthrust at all (cf. Fig. 4b) (Beaumont et al., 2006; Jamieson et al., 2006). Because numerical models with modestly different starting scenarios simulate very different channel flow behaviors, it is hard to test the channel flow hypothesis from simple observations of crustal structure, and a large range of scenarios for tectonic development is possible. Thus even if the KKF now limits (or until recently limited) active northward underthrusting of Indian crust into the lower crust of southwestern Tibet (Fig. 4b), channel flow extrusion of the Himalaya from beneath the southernmost Tibetan Plateau (Fig. 4a) could still be viable southwest of the KKF. Initiation of the KKF, perhaps as a response to oblique plate convergence in the NW Himalaya (Styron et al., 2011), must have been contemporaneous with modification of, but not necessarily termination of, channel flow extrusion of the Himalaya.

6. Summary
Our measurements of $^{3}$He/$^{4}$He levels in geothermal springs lets us map the configuration of mantle lithospheric types beneath the Tibet plateau in the geologically recent past and sheds light on deep crustal processes within Earth’s largest collisional orogen. Despite the relatively few hot springs yet sampled, we see the close association of the KKF with a mantle component of geothermal fluids that requires the KKF to access tectonically active Tibetan mantle. Our cross-section, consistent with all available geophysical and geological data, suggests the KKF may have acted as a plate-boundary transcurrent fault in latest Miocene and early Pliocene time. Because this orogenic system is rapidly evolving, our geochemical data cannot show whether the KKF is still acting as a crust-penetrating fault today.

Acknowledgements
Fieldwork was supported by NSF grant EAR-0409939 and NGRI. Laboratory measurements at Lawrence Berkeley National Laboratory were supported by the Director, Office of Energy Research, Basic Energy Sciences, Chemical Sciences Division under Contract No. DE-AC02-05CH11231. Reviews by J.P. Avouac, D. Grujic, P. Kapp, and an anonymous reviewer, and editorial comments from B. Marty, greatly improved this paper.

Appendix. Supplementary data
Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl........
References


Mantle fluids in the Karakoram fault: Helium isotope evidence

Table 1: Sample locations, characteristics, and noble gas isotope analyses of geothermal occurrences in west Tibet and northwest Himalaya

<table>
<thead>
<tr>
<th>Location</th>
<th>Type of sample</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation</th>
<th>Temperature</th>
<th>pH</th>
<th>Conductivity</th>
<th>Geologic setting</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kasol (IND2)</td>
<td>2002</td>
<td>32° 57' 58&quot; 77° 08' 45&quot;</td>
<td>3183</td>
<td>76</td>
<td>7.53</td>
<td>22</td>
<td>0.079</td>
<td>Mesozoic Karakoram leucogranite, formed by crustal melting (Philips et al., 2004)</td>
</tr>
<tr>
<td>Rampur quartzite, Lesser Himalaya series</td>
<td>205</td>
<td>72.5</td>
<td>7.08</td>
<td>0.055</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Notes:

- Locations and elevations measured with handheld Garmin GPS.
- Temperature, conductivity and pH measured with handheld Hanna Instruments 98130 digital meter.
- Chloride measured with high chloride Quantab test strips.

New Mass Spectrometer Data

<table>
<thead>
<tr>
<th>Location</th>
<th>R/Ar</th>
<th>R/Ne</th>
<th>F(He)</th>
<th>F(Ne)</th>
<th>F(Kr)</th>
<th>F(Xe)</th>
<th>R/ArNe</th>
<th>R/XeAr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kasol (IND2)</td>
<td>0.08</td>
<td>0.04</td>
<td>0.03</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>0.08</td>
<td>0.04</td>
</tr>
<tr>
<td>Rampur quartzite, Lesser Himalaya series</td>
<td>0.055</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Notes:

- All analyses reported here were performed at Lawrence Berkeley National Laboratory's Center for Isotope Geochimistry.
- R/Ar = the measured (36Ar/3He) ratio normalized to the ratio in air (1.936-08).
- R/Ne = (36Ne/3He) ratio normalized to the ratio in air (1.936-08) after correction for atmospheric contamination as measured by (1) the F(He) method: assumes all Ar is from air contamination and R corrected accordingly, resulting in a maximum correction and (2) the F(He) method: uses the F(36Ne/3He) ratio and assumes the contamination from both air and saturated water. All Ne is assumed atmospheric.
- F(He) = air normalized relative abundance: (F(He) = [36He/3He]air) +/- F(He) inaccuracy.
- F(Ne) = radiogenic Ar/He ratio in the sample = [36He/36Ar] - 1.22262/6.32262, and relative Ar and Ne extraction efficiency from source minerals.

Uncertainties:

- All reported uncertainties are analytical errors (1 standard deviation).
- Errors associated with the reproducibility of a given measurement are folded in as part of normalization to air values.
- Lower limit uncertainties for an isotope ion current measurement are proportional to 1/square root of the number of ions counted.

Previously recorded results shown in Figures 1, 2, 3

Notes:

- The springs named as Tirthapuri by Hoke et al. are c. 18 km ENE of the monastery and hot-springs conventionally given that name.
- Although Hoke et al. (2000) did not publish uncertainties and their original data are no longer available, their results come from the well-established noble-gas laboratories in Amsterdam and Rochester that would be remarkable if the Hoke et al. data points from Tirthapuri and Shiquanhe, that are reported to have 3–5% mantle fluid component, all actually represent fluids with only crustal helium, or have published many other data with small uncertainties (e.g. Hilton et al., 2000; Poreda and Farley, 1992). The similarity between the Hoke et al. analyses of two closely adjacent hot springs at Tirthapuri, in n.r. = not reported: neither Hoke et al. nor Walia et al. publish any uncertainties or error bars, hence none are shown in Figures 2b and 3.
Figure Captions

Fig. 1. $^{3}He/^{4}He$ sample locations and faults in Tibet, and previous estimates of the northern limit of India beneath western Tibet. (a) Shaded relief index map of Tibet and major faults: KKF–Karakoram fault, MFT–Main Frontal thrust, MCT–Main Central thrust, ISZ–Indus-Yarlung suture zone, BSZ–Banggong-Nuijiang suture zone, JSZ–Jinsha suture zone, ATF–Altyn Tagh fault (Karakax fault). N: Nagqu; Y, T, D: Yangbajain, Tingri, and Daggyai Co graben systems; Y is location of mantle suture of Hoke et al. (2000). (b) $^{3}He$ data locations (diamonds–this study; circles–previous studies in Tibet (Hoke et al., 2000) and the Himalaya (Walia et al., 2005) superimposed on tectonic map (Murphy et al., 2000). Yellow shading: $R_C/R_A>0.05/R_A$. White symbols: no mantle signature. Letters identifying geothermal sites (C, P, G, Ch, Pu, S, L, M, T, Ku, Ka, Ma) are spelled out in Fig. 1c used in Figs. 2 and 3. Bold red lines: previous interpretations of northern limit of Indian lithosphere at the Moho from S-receiver functions (Zhao et al., 2010), from P-receiver functions (Nábělek et al., 2009; Wittlinger et al., 2004), from northern limit of deep earthquakes (at 85-95 km depth–black stars) (Priestley et al., 2008), and from body-wave tomography (Li et al., 2008). LGF–Longmu Co-Gozha Co fault system. TM and ZB: Tso Morari and Zada basin (Leo Pargil) normal faults. (c) Satellite image of KKF (same area and scale as (b)). All hot springs or fumaroles with $R_C/R_A>0.05•R_A$ (colored yellow) lie on or immediately north of the KKF (C, P, L, S, M, T).

Fig. 2. Characteristic $^{3}He/^{4}He$ patterns in different tectonic environments, and $^{3}He/^{4}He$ measurements from Tibet. (a) Data from geothermal areas across intra-continental plate boundary strike-slip faults and the Basin and Range extensional province (BRP) at the same scale. Orange line: envelope of highest $^{3}He/^{4}He$ values (orange circles) measured at each distance relative to the San Andreas fault (SAF) (Kennedy et al., 1997; Kulengoski et al., 2012), i.e., those springs that presumably penetrated most deeply and scavenged most mantle helium (note, broad $^{3}He/^{4}He$ peak is due to high mantle helium component also present on the Holocene Ortigalita splay of the SAF system). Pink line: envelope of $^{3}He/^{4}He$ values measured close to the North Anatolian fault (NAF) (individual data points not shown for clarity) (Doğan et al., 2009). Rapid spatial gradients of $^{3}He/^{4}He$ across SAF and NAF suggest these plate boundary strike-slip faults tap mantle sources. Red line: envelope of $^{3}He/^{4}He$ values measured across the KKF (from Fig. 2b), at same scale. Turquoise line: baseline trend of $^{3}He/^{4}He$ ratios across BRP showing a long wavelength increase from east to west inferred to represent mantle fluids penetrating the ductile lithosphere via deformation-enhanced permeability (Kennedy and van Soest, 2007), related to westward-increasing strain and NW shear component shown by dashed turquoise line: GPS velocity in N35°W direction with respect to North America (Hammond and Thatcher, 2004). BRP values are shown from east to west, to more easily compare to southeast Tibet (Fig. 2b). (b) Red: line: highest $^{3}He/^{4}He$ values in west Tibet measured at each distance southwest/northeast of the KKF (upper scale). Data points labeled as in Fig. 1; circles–from Hoke et al. (2000) and Walia et al. (2005) (no error bars available) and diamonds–this work, analytic errors smaller than symbols, except as shown for Changlung and Langjiu. Yellow shading: samples enriched in mantle helium; white shading: samples lacking any mantle helium. Blue line: envelope of highest $^{3}He/^{4}He$ values in southeast Tibet measured at each distance south/north of the ISZ (lower scale) and blue dots: data points from Hoke et al. (2000), and Yokoyama et al. (1999) (no error bars.
available). Y: Yangbajain geothermal area. Dashed blue line: GPS velocity of southeast Tibet relative to Eurasia in the direction 110°, orthogonal to the India-Asia convergence direction (Zhang et al., 2004) plotted against distance north of the ISZ to show increasing eastward lateral escape velocity of Tibet. Grey boxes: inferred mantle suture (the southern limit of Tibetan mantle in contact with subducting India), 50-100 km north of the ISZ in southeast Tibet (Hoke et al., 2000), and at the KKF in west Tibet (this paper). Upper and lower distance scales in (b) are offset 35 km to align our samples from ISZ in NW India (Gaik, Chumathang) with ISZ in southeast Tibet. In (a) and (b), definition of crustal and mantle helium domains from Hoke et al. (2000); distance scale and GPS velocity scale are identical, but $^3\text{He}/^4\text{He} R_C/R_A$ scale changes by a factor of 2.5.

**Fig. 3.** Helium isotope signature of all Karakoram fault, western Tibet and western Himalaya geothermal springs with published He and Ne data (this study–diamonds, Hoke et al., 2000–circles) (see Fig. 2 and Table 1 for additional data for which only He are available (this study; Walia et al., 2005)). Data are superimposed on mixing curves of three end-member components (mantle $R/R_A=8$ and $^4\text{He}/^{22}\text{Ne}=1000$, crust $R/R_A=0.02$ and $^4\text{He}/^{22}\text{Ne}>1000$, and air $R/R_A=1$ and $^4\text{He}/^{22}\text{Ne}≈3$). $2\sigma$ uncertainty ellipses are shown where available (this study (ellipse is smaller than symbol for Mengshi); not Hoke et al., 2000). The displacement of the samples with mantle enrichment (yellow) from the mixing line between the crustal end-member and air or air-saturated water requires excess mantle $^3\text{He}$. Yellow field: mantle-enriched samples from southeastern Tibet (Yokoyama et al., 1999; Hoke et al., 2000). Blue field: samples from “crustal helium domain” in southeastern Tibet (Hoke et al., 2000; Newell et al., 2008). Grey field: compositional gap between pure crustal and mantle-enriched fluids in which no data are reported from Tibet or the Himalaya (this study; Hoke et al., 2000; Newell et al., 2008; Walia et al., 2005; Yokoyama et al., 1999). Definition of “crustal” and “mantle” helium domains from Hoke et al. (2000).

**Fig. 4.** Comparative cross-sections of southeastern and western Tibet at present day. Both sections are true scale, from MFT across the Himalayan orogenic wedge (MHT: Main Himalayan Thrust, STD: South Tibet Detachment), to Tibetan terranes (Lhasa and Qiangtang blocks) and aligned on the ISZ. (a) South-north at 90°E (INDEPT H transect (Klemperer, 2006, 2008); also representative of Hi-CLIMB transect at 85°E (Nábělek et al., 2009)); (b) SW-NE crossing the KKF at ~79°E. Red arrows are mantle helium flux; solid grey lines are major faults, crustal boundaries and magma chambers taken directly from geophysical studies referenced in figure, projected along strike where necessary; dashed grey lines and question marks identify boundaries and magma chambers inferred but not yet imaged. Green shading denotes Indian crust and its sedimentary cover; receiver-function “doublet” corresponds to northern tongue of “Indian middle crust” in (a) and “stranded Indian crust” in (b). Black arrows schematically mark presumed material transfer, of Indian upper crust upwards across the MHT as anticipated by channel flow models, and eclogitised Indian lower crust downwards across the Moho to be subducted into the deep mantle. Stars are deep earthquakes close to the crust-mantle boundary.
Mantle fluids in the Karakoram fault: Helium isotope evidence
Mantle fluids in the Karakoram fault: Helium isotope evidence

Figure 2
Mantle fluids in the Karakoram fault: Helium isotope evidence

Figure 3
Mantle fluids in the Karakoram fault: Helium isotope evidence

(a) INDEPTH transect @ 90°E (west Tibet analog prior to initiation of Karakoram Fault)

(b) Karakoram Fault @ 79°E

Figure 4
Appendix  Data collection and analysis

Sampling protocols

Samples of non-condensable gas or water were collected from springs and fumaroles. For springs that were either non-boiling or without a free gas phase, a tube was submerged in the spring, as close to the spring inlet as possible, and water was pulled through the sampling apparatus using a peristaltic hand pump. After several volumes of water had passed through the system a sample of the liquid was collected for analysis of the dissolved gases. When sampling boiling springs or springs with a free gas phase, water displacement method was used to capture the gas phase. First water was pulled through the sampling apparatus using the hand pump and then the vapor/gas phase was captured, using an inverted funnel submerged in the spring, and forced through the water-filled sampling apparatus. Along the flow line, the total fluid (liquid and gas) encounters an inverted “Y-shaped” tube that separates the liquid and gas phases, allowing for collection of the non-condensable gases at ambient conditions. Fumarole gases were captured and forced through the sampling apparatus using an inverted funnel sealed around the funnel-ground contact. The gas and vapor are allowed to flow through the sampling apparatus in order to adequately purge the sampling lines of air before collecting a sample. While purging the line, the gas is discharged into a water reservoir to minimize back diffusion of air gases.

Analytical method

In all cases, a 9.8 cc sample of gas or liquid was collected at ambient conditions in a Cu-tube cold-welded at each end using bolt-driven clamps. To insure sample integrity, the clamps remained in place until the sample was ready for analysis and attached to the sample preparation vacuum line that is in series with the noble-gas mass spectrometer. Sample preparation and the
noble gas analyses were conducted in the RARGA (Roving Automated Rare Gas Analysis) laboratory at the Lawrence Berkeley National Laboratory. Sample preparation, analytical techniques and the instrumentation employed in the RARGA laboratory, are identical to those described by Kennedy and van Soest (2006). The Cu-tube is opened by re-rounding a cold weld at one end of the tube, allowing the sample gas to expand into the sample preparation line for processing. First, water vapor is condensed in a flow-through trap cooled externally using a methanol-liquid N$_2$ slurry. Following removal of water vapor, the line pressure was measured to calculate the amount of non-condensable gases in the sample. Then CO$_2$ and other reactive gases (N$_2$, H$_2$, CO, etc.) are chemically removed by exposure to a stream of evaporating Ti-metal. After removal of the reactive gases, an aliquot of the remaining purified noble gas fraction is isolated for determination of absolute and relative abundances. The rest of the noble gas fraction (~95%) is trapped on activated coconut charcoal cooled to ~35 K, from which each noble gas can be thermally separated and analyzed individually for its isotopic composition. The sample preparation line and mass spectrometer performance are calibrated using aliquots of air and a Berkeley helium standard with an isotopic composition of 2.4 $R_A$ (where $R_A$ is the $^3$He/$^4$He ratio in air: $1.39\times10^{-6}$).

**Corrections for contamination by air**

Helium in crustal fluids is not typically impacted by contamination by air or air-saturated water because helium has a very low atmospheric abundance (~5 ppm). Nonetheless, atmospheric contributions need to be considered whether they occur during spring recharge or during sampling, particularly for samples with low helium abundance. These corrections are most relevant to sample Changlung, which has the highest measured helium isotopic composition ($R/R_A = 0.725$) and lowest $F(\He)$ (air-normalized relative abundance, 2.417 compared to values of ~30–8500 for the other samples, see Table 1). Furthermore, the high $R/R_A$ for Changlung makes this sample most critical in our discussion of mantle contributions to the surface fluids.
In the main text, we discuss correction of $R/R_A$ values for air contamination using two methods. In the first air-correction method we assumed that all $^{36}\text{Ar}$ was from air contamination and corrected the measured $R/R_A$ accordingly. This results in a maximum correction to the measured $^{3}\text{He}/^{4}\text{He}$ ratio so gives a minimum $R_C/R_A$ (Changlung $R/R_A = 0.725\pm0.205$ becomes $R_C/R_A = 0.531\pm0.349$). In the second method we used the He/Ne ratio and assumed air contributions came from both air and air-saturated water. The only other assumption is that the He/Ne ratio is not significantly altered by gas loss (e.g. boiling, etc.) due to the similar solubility of He and Ne. Using this neon method, Changlung $R/R_A = 0.725\pm0.205$ is corrected to $R_C/R_A = 0.663\pm0.321$. Given these large analytical uncertainties, to further strengthen our case for the presence of mantle (or excess) $^3\text{He}$ in the Changlung sample, we provide a third correction method that relies on all of the air-derived noble gases ($^{22}\text{Ne}$, $^{36}\text{Ar}$, $^{84}\text{Kr}$ and $^{132}\text{Xe}$).

The Supplementary Figure plots of $F(^{132}\text{Xe})$ vs. $F(^{22}\text{Ne})$ and $F(^{84}\text{Kr})$ vs. $F(^{22}\text{Ne})$ show that all our samples (including Changlung) have experienced a combination of gas loss and consequent elemental fractionation (due to boiling or bubble formation) and contamination with air. Our model for fluid evolution presumes that our samples have initial $F$ values for the air-derived noble gases equal to air-saturated water (ASW). The geothermal fluids undergo a small amount of elemental fractionation with a decrease in $F(^{22}\text{Ne})$ and increase in $F(^{132}\text{Xe})$ and $F(^{84}\text{Kr})$, either by batch gas loss (black dashed line) or Rayleigh distillation (black dotted line) or some combination of the two processes. The resulting residual liquids are then contaminated by some fraction of air (solid red arrows) to reach their final $F$ values (diamonds) close to but just above the ASW-Air mixing line (red dotted line). Assuming the sample compositions reflect mixing between two components (air, and liquid residual to gas loss) and using the interpolated end-member compositions for the residual liquid, the fraction of $^{36}\text{Ar}$ from air can be readily calculated. Once the fraction of $^{36}\text{Ar}$ from air is calculated, we calculate the corresponding fraction of $^{4}\text{He}$ from air.
and correct the measured $R/R_A$ values for air contamination. The advantage of this approach is that a more realistic value for $F(^{22}\text{Ne})$ in the water end-member is used, as opposed to assuming the value for air-saturated water, resulting in a more accurate estimate of the fraction of $^{36}\text{Ar}$ from air. For Changlung the corrected $R_C/R_A$ value by this method is 0.678 (using an $F(^{22}\text{Ne})$ value for the residual water of ~0.150).

In all three cases, we are left with the conclusion that Changlung has excess $^3\text{He}$ with respect to anticipated radiogenic (or crustal) values. This conclusion is consistent with and evident from our plot of $R/R_A$ (measured) vs. $^4\text{He}/^{22}\text{Ne}$ (Figure 3, in the main text).

Supplementary Figure caption:

Top: plot of $F(^{132}\text{Xe})$ vs. $F(^{22}\text{Ne})$. Bottom: plot of $F(^{84}\text{Kr})$ vs. $F(^{22}\text{Ne})$. Diamonds: observed values for Changlung–C, Chumathang–Ch, Gaik–G, L–Langjiu, M–Mengshi, and Puga–Pu. (Panamik not shown, as heavy noble-gas data were not measured for this sample.) Red crosses: compositions of air, and air-saturated water at 0°C and 10°C. Black dashed line: batch gas loss; dotted line: a Rayleigh process, both representing liquid compositions residual to gas loss from the 10°C ASW composition. Red solid lines: mixing lines between residual liquid and air. Red dashed line: mixing line between ASW and air.