

Tracing weathering regimes using the lithium isotope composition of detrital sediments

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- 1 Tracing weathering regimes using the lithium isotope
- 2 composition of detrital sediments
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- 9 ABSTRACT

10 Lithium (Li) isotopes are a promising tracer of chemical weathering processes for 11 both modern and ancient times. In order to improve the use of Li isotopes in the 12 sedimentary record, here we calibrate the relationship between weathering intensity and 13 detrital Li isotope composition (δ^7 Li) using the fine fraction of modern large river 14 sediments. By using independent estimates for sediment provenance to calculate the Li 15 isotope signature of the rock from which the sediments derive through weathering, we 16 show that source rock variability (in particular the relative contribution of sedimentary vs. 17 igneous rocks) must be corrected for before using Li isotopes as a weathering proxy. We 18 also show that for rivers draining mountain ranges, the contribution of particles derived 19 from sedimentary rocks to river sediments is correlated to their Li/Al ratio, making it 20 possible to use Li contents to estimate the average source rock composition. Once 21 corrected for bedrock variability, the Li isotope signature defines a negative relationship 22 with the weathering intensity (ratio between silicate weathering rate and total denudation

rate) with highest Li isotope fractionation for the highest weathering intensity.

24 Altogether, we propose a set of new relationships between weathering, erosion,

- 25 provenance and Li isotopes that can be used to quantify present-day and paleo-
- 26 weathering using detrital sediment.

27 INTRODUCTION

28 The surface of Earth evolves through the combined action of chemical weathering 29 reactions and physical denudation. The former process partly dissolves the continental 30 crust, exports cations to the ocean, and consumes atmospheric CO_2 from the atmosphere, 31 whereas physical denudation exports solid products to the ocean. Hence, over long time 32 scales, silicate weathering is one of the major processes regulating the amount of 33 atmospheric CO₂ and therefore controlling the evolution of climate (Berner et al., 1983). 34 Despite this recognition, it is still poorly known how silicate weathering has changed in 35 the past and what are the parameters controlling weathering rates and intensities over 36 long timescales. The emerging Li isotope proxy has a great potential for tracing silicate 37 weathering because Li is a soluble element predominantly hosted in silicate rocks and its 38 isotopes are fractionated during water-rock interactions (Huh et al., 2001). Although the 39 relation between the river dissolved Li isotope composition (δ^7 Li) and weathering is still 40 a matter of debate (e.g., Dellinger et al., 2015; Huh et al., 2001; Pogge von Strandmann 41 and Henderson, 2015), this has served as a basis to investigate the $\delta^7 Li$ of ancient marine 42 carbonates (e.g., Misra and Froelich, 2012; Pogge von Strandmann et al., 2013) in order 43 to unravel global past weathering conditions (Bouchez et al., 2013; Li and West, 2014; 44 Misra and Froelich, 2012).

45	Complementary to this approach, the characterization of weathering conditions
46	with Li isotopes can also be achieved with the study of detrital sediments. However,
47	several difficulties arise for using sediments $\delta^7 Li$ to trace weathering and erosion in
48	modern systems. First, detrital sediments not only contain weathered products but also
49	unweathered minerals deriving from the physical erosion of the bedrock (Dellinger et al.,
50	2014). Secondly, the δ^7 Li of river sediments is quite variable depending on the sediment
51	grain size (Dellinger et al., 2014). Finally, as on average 80% of the total Li in the
52	weathering system is contained in the solid load and only 20% in the dissolved load
53	(Misra and Froelich, 2012), terrigenous sediment $\delta^7 Li$ is usually close to that of source
54	rock (Bouchez et al., 2013), such that the source rock δ^7 Li has a large imprint on the
55	composition of the detrital material (Dellinger et al., 2014). In this contribution, we use
56	the available data from the literature to investigate how the $\delta^7 Li$ of medium to large river-
57	sediments relates to the present-day weathering regimes.
58	PRESENT-DAY RIVER SEDIMENTS DATASET
59	The data set used in this study includes some of the largest Earth river systems:

60 the Amazon, Mackenzie, Ganges and Brahmaputra (Dellinger et al., 2015, 2014), Congo 61 (Henchiri et al., 2016) and Changjiang (also known as the Yangtze River; Wang et al., 62 2015). It is complemented by new data from the Stikine and Skeena River from the 63 western Canadian cordillera, two large rivers draining predominantly volcanic rocks 64 (Gaillardet et al., 2003). All these rivers drain a surface larger than 19,000 km², and cover 65 a wide range of lithology, climate, and weathering regimes such that this data set can be 66 considered as globally representative. In the following, we characterize the modern weathering regimes by using the weathering intensity parameter "(W/D)", which is 67

68	independent of the drainage area and the discharge, and defined as the ratio between
69	chemical silicate weathering rates W (calculated as the flux of river dissolved material
70	sourced from silicate rocks only) and total denudation rate D, thereby corresponding to
71	the proportion of weathering material transported in the dissolved form by the river.
72	Within the set of rivers from this study, we distinguish two groups: (1) rivers
73	draining high elevation mountain ranges (the Andes, Himalayas, Western Canadian
74	Cordillera) characterized by moderate to high erosion rates (>100 t km ⁻² yr ⁻¹), low
75	weathering intensity (W/D < 0.1), and a kinetically limited weathering regime (West et
76	al., 2005); and (2) rivers draining lowlands with deep weathered soils, low erosion rates,
77	and high weathering intensity (W/D > 0.1) corresponding to the supply-limited
78	weathering regime (West et al., 2005).
79	Overall, for rivers draining mountain ranges, the Li concentration of river
80	sediments ranges from 10 to 102 $\mu g \ g^{-1}$ (ppm) and the $\delta^7 Li$ of the sediments
81	$(\delta^7 \text{Li}_{suspended_load}, \text{ precision better than }\%)$ from -4‰ to +5‰ (Fig. 1). The rivers draining
82	lowlands have on average lower δ^7 Li (-6.5‰ to -3‰) and Li concentration compared to
83	rivers draining mountain ranges. Interestingly, the δ^7 Li is negatively correlated to the Li
84	concentrations in large rivers draining mountain ranges, whereas no relationship is
85	observed for rivers draining the lowlands (Fig. 1).
86	REMOVING THE NON-WEATHERING EFFECTS
87	In large rivers draining mountain ranges, the δ^7 Li and Li concentration of river

- 88 sediments are strongly dependent upon grain size and lithological source (Dellinger et al.,
- 89 2014). In order to extract the weathering signal contained in river sediments, it is

- 90 necessary to remove the effect of both grain size and lithological variability on the Li
- 91 isotope composition.

92 Grain Size Effects

Fine sediments have generally lower δ^7 Li and higher Li concentration than coarse 93 94 sediments (Fig. 1). This is because they contain a higher proportion of particles sourced 95 from sedimentary rocks and a higher proportion of weathering products sourced from 96 modern soils that carry the fingerprint of the modern weathering regime (Dellinger et al., 97 2014). Therefore, similarly to element ratios (e.g., Lupker et al., 2013), large changes of 98 the δ^7 Li in sedimentary archives can simply arise from change of the sediment grain size 99 and lead to false interpretation in terms of shift in weathering conditions. In order to 100 mitigate this grain size effect, in the present study we use for each river the finest 101 sediment sample available (size $< 63 \mu m$), which for most cases has been collected at or 102 near the river channel surface. However, even these fine sediments contain a non-103 negligible fraction of unweathered material that needs to be taken into account for the 104 interpretation of their δ^7 Li data. 105 The Influence of Lithology 106 The Li isotope signature that can be measured in weathered detrital material depends upon (1) the initial δ^7 Li of river sediments, prior to chemical weathering (i.e., the 107

- 108 Li isotope signature of its source rock material, hereafter referred to as " $\delta^7 Li_{source_material}$ "),
- 109 which is a lithological effect and (2) the extent of fractionation between this initial
- 110 signature and the neoformed weathering material, which in turn depends on isotope
- 111 fractionation factors and on Li partitioning between secondary minerals and the dissolved
- 112 load (Bouchez et al., 2013). The weathering signal is contained only in the second effect.

113	This is why in the following, we consider the difference in isotope composition between
114	the fine suspended sediment and the original bedrock, $\Delta^7 Li_{fine-source} = \delta^7 Li_{fine suspended_load}$ -
115	$\delta^7 Li_{source_material}$.
116	The " $\delta^7 Li_{source_material}$ " corresponds to the weighted average of the $\delta^7 Li$ of the
117	various erosion sources: igneous, metamorphic and sedimentary rocks (hereafter referred
118	to as "shale" because shales are by far the most important type of sedimentary rock
119	contributing to the Li budget; Holland, 1984) that have distinct average Li concentration
120	and isotope composition (Dellinger et al., 2014; Sauzéat et al., 2015; Tomascak et al.,
121	2016). Indeed, the Li isotope variability of the present-day upper continental crust
122	(UCC), as characterized by using inherently integrative samples like large river sand
123	(Dellinger et al., 2014) or desert loess (Sauzéat et al., 2015), corresponds mostly to a
124	mixture between a Li-rich, low- δ^7 Li, shale end member (δ^7 Li of -0.5 ± 1 ‰ and Li/Al of
125	$1.0 \pm 0.1 \times 10^{-3}$) and a Li-poor, high- δ^7 Li, igneous end member (δ^7 Li of 3.5 ± 1.5 %),
126	Li/Al of $0.30 \pm 0.12 \times 10^{-3}$). This mixing trend can be used to determine the Li isotope
127	composition of river sediments prior to weathering ($\delta^7 Li_{source_material}$) provided that the
128	relative contribution of the different source materials to the sediments is known.
129	Quantification of the provenance of sediments can be achieved using Sr-Nd
130	isotopes and/or insoluble trace element ratios, provided that the composition of the major
131	bedrock drained by each river basin is known and that they have contrasted signatures
132	(e.g., Allègre et al., 1996). Here, we use independent tracers of provenance (Nd, Sr
133	isotopes and trace elements) in order to quantify the erosion sources of the fine sediments
134	and determine the $\delta^7 Li_{source_material}$ for fine sediments from each of the large rivers. We
135	note that the procedure of choice to characterize average source rock is specific to each

136	large river basin. In the Amazon, Congo, and Ganges-Brahmaputra basins, sedimentary,
137	igneous, and metamorphic rocks have distinct Nd isotope composition making it possible
138	to quantify precisely the provenance of fine sediments using Nd isotopes (see the GSA
139	Data Repository ¹). For the Mackenzie River, we use trace elements (Cr, Ti, Cs) ratios
140	(see the Data Repository). We emphasize that this calculation provides information on
141	source rocks for fine sediments only, which can be different for other grain sizes
142	(Bouchez et al., 2011a; Dellinger et al., 2014; Garçon and Chauvel, 2014).
143	However, in some cases like that of the Changjiang basin, neither Sr-Nd isotopes
144	nor trace elements can be used to assess the provenance of sediments because the
145	signatures of the main rock sources are under-constrained or not significantly different
146	(Chetelat et al., 2013). An alternative method is to identify another tracer that correlates
147	with an independent measure of the proportion of shale-derived particles (i.e., the amount
148	of sedimentary recycling) quantified across the rest of our river set. Among all the
149	chemical elements tested, the best correlation with the proportion of shale-derived
150	particles is obtained for the Li/Al ratio (Fig. 2). Li concentrations are normalized to Al to
151	remove dilution effect by quartz or other Li-free phases. This good relationship stems
152	mostly from the fact that (1) shales and igneous rocks have significantly distinct Li
153	contents, and (2) that chemical weathering has a minor impact on sediment Li content for
154	a majority of samples. However, we note that sediments from some lowland rivers have
155	lost a significant fraction of Li because of their higher weathering intensity (Fig. 3).
156	Using the relationship between Li and the proportion of sediments derived from erosion
157	of shales calibrated on rivers draining mountain ranges, we infer the proportion of shales

158	contributing to the fine sediment load of the Changjiang basin river, which makes our set
159	of $\delta^7 Li_{source_material}$ estimates complete (see Data Repository).
160	Using the calculated proportion of shales and igneous rocks with the δ^7 Li of the
161	rock end members, we estimate the Li isotope composition of the rock source
162	$(\delta^7 Li_{source_material})$ for each fine river sediment sample (see the Data Repository). We note
163	that the calibration of the Li provenance tracer in detrital sediments might also serve to
164	determine the provenance and the $\delta^7 Li_{source_material}$ of ancient detrital sedimentary archives
165	for which the potential bedrock sources are usually not constrained.
166	RELATION BETWEEN THE LITHIUM ISOTOPE SIGNATURE AND
167	WEATHERING INTENSITY
168	In this section we evaluate the relation between sediment $\delta^7 Li$, Li weathering
169	index and the weathering intensity previously defined as the W/D ratio. The Li
170	weathering index of the detrital sediment can be expressed as (Li/Al)_{fine_sediments} /
171	(Li/Al) _{source_material} , where (Li/Al) _{source_material} corresponds here to the (Li/Al) ratio the
172	suspended sediments would have in the absence of any silicate weathering reactions. A
173	value of 1 means no loss or gain of Li by the solids during weathering while a value
174	lower than 1 implies loss of Li during weathering. The "(Li/Al)source_material" values are
175	determined using the rock end member compositions defined in the previous section and
176	the proportion of shales calculated above. The Li weathering index values are negatively
177	correlated to the weathering intensity (Fig. 2.b) with most of the rivers draining mountain
178	ranges having a Li weathering index higher than 0.8 (except for the Solimões and Pastaza
179	rivers), whereas rivers draining lowlands have Li weathering index values between 0.9
180	and 0.4. The higher Li depletion in sediments transported by lowland rivers is compatible

181	with the higher proportion of dissolved Li transported by these rivers (see the Data
182	Repository). This shows that significant Li depletion in sediments occurs only for very
183	intense weathering conditions (Fig. 2).
184	The relationships between Li isotope signature in modern fine river sediments
185	corrected for the source rock signature ($\Delta^7 Li_{\text{fine-source}}$) can be examined as a function of
186	the weathering regime. A significant negative relationship is observed between the
187	weathering intensity (W/D) and $\Delta^7 \text{Li}_{\text{fine-source}}$ (Fig. 3a, $r^2 = 0.75$). This correlation is
188	stronger than that between $\delta^7 \text{Li}_{\text{fine}_sediments}$ and W/D (Fig. 3b, $r^2 = 0.48$), showing that
189	source rock composition indeed introduces some scatter in this relationship. Sediments
190	from rivers characterized by high W/D have the lowest $\Delta^7 Li_{\text{fine-source}}$ whereas rivers
191	having low W/D have $\Delta^7 Li_{fine-source}$ close to zero. The $\delta^7 Li_{sed}$ and $\Delta^7 Li_{fine-source}$ also
192	correlate well with more traditional weathering proxies such as CIA ($r^2 = 0.76$, not
193	shown) or K/Th ratio ($r^2 = 0.62$, not shown). We observe no relationship between $\Delta^7 Li_{fine}$.
194	source and chemical silicate weathering rates.
195	To explain this relationship, we suggest that at high denudation rate (and hence
196	low W/D), incorporation of Li into secondary minerals is hindered by short mineral
197	residence times in soils making the dissolved $\delta^7 Li$ close to $\delta^7 Li_{source_material}$ (Bouchez et
198	al., 2013; Dellinger et al., 2015; Wang et al., 2015). Therefore, the small amount of
199	secondary minerals formed should have a low $\delta^7 Li$, which could be expected to be
200	reflected in the δ^7 Li of fine sediments (Dosseto et al., 2015). We suggest that this is not
201	observed because at very low weathering intensity (W/D < 0.01), typical of high-erosion
202	settings, the sediments carried by the rivers contain a high proportion of primary minerals
203	$(\Delta^7 \text{Li}_{\text{sed-rock}} = 0)$ compared to secondary minerals formed during the modern weathering

204	cycle, even in the fine sediments (Dellinger et al., 2014, 2015). At high weathering
205	intensity (W/D $>$ 0.1) such as in lowland areas, secondary minerals formed during the
206	present-day weathering cycle and characterized by low $\delta^7 Li_{fine_sediments}$, compose the
207	majority of the sediments and therefore control the bulk sediment Li isotope signature.
208	The contribution of unweathered debris to river fine sediment load is likely to be more
209	prominent for rivers draining fine-grained sedimentary rocks (which is the case of large,
210	lithologically mixed rivers), as these lithologies are highly erodible and produce fine
211	sediments.
212	IMPLICATIONS AND CONCLUSIONS
213	These results have important implications regarding the use of Li isotopes in
214	clastic sediments to trace weathering regimes both in the present-day and in the past.
215	First, the δ^7 Li of modern river sediments is primarily related to weathering intensity (that
216	is, in the modern Earth, mostly a function of the erosion rate; Dixon and von
217	Blanckenburg, 2012) and not to the silicate weathering rate as inferred by other studies
218	(e.g., Dosseto et al., 2015).
219	Second, the δ^7 Li of fine river sediments is sensitive only to large changes of the
220	weathering intensity (i.e., by an order of magnitude; Fig. 3a.). In particular, a $\Delta^7 Li_{fine-source}$
221	value between -1% and -3.5% can correspond to a large range of possible weathering
222	intensity comprised between 0.005 and 0.08. Indeed, in high-erosion rate settings, the
223	contribution of soil-derived weathered particles is minor compared to that of debris of
224	unweathered bedrock mobilized by mass wasting processes, even in the finest river
225	sediments fraction (Dixon and von Blanckenburg, 2012; Dellinger et al., 2014). However,
226	the δ^7 Li of clastic sediments can clearly discriminate between high weathering intensity

Publisher: GSA Journal: GEOL: Geology DOI:10.1130/G38671.1 vland areas like the Congo River basin) and intermediat

227	settings (typical of tropical lowland areas like the Congo River basin) and intermediate to
228	low weathering intensity regimes.
229	The third important implication is related to the use of Li isotopes in detrital
230	sediments as a proxy for palaeo-weathering regimes. Tracing ancient weathering
231	processes using conventional geochemical tracers in detrital sediments (e.g. chemical
232	index of alteration CIA, or chemical ratios combining a soluble and an insoluble element,
233	such as Na/Sm, U/Th or K/Al) can be challenging because these proxies are also
234	influenced by sediment provenance, grain size effects or post-depositional alteration
235	(Bouchez et al., 2011a; Carpentier et al., 2013; Li and Yang, 2010; Lupker et al., 2013).
236	Hence, some recent studies have used Li isotopes in fluvial alluvial deposits (Dosseto et
237	al., 2015), marine detrital sediments (Bastian et al., 2015) or ancient siliciclastic rocks (Li
238	et al., 2016) to determine palaeo-weathering conditions. In comparison to marine
239	carbonates, clastic sediments record the fingerprint of local weathering conditions and
240	have the advantage of not being influenced by biological fractionation ("vital effects")
241	observed during carbonate precipitation (Vigier et al., 2015).
242	The results from this study can be used as a framework for interpreting the detrital
243	$\delta^7 Li_{fine_sediments}$ and Li/Al ratio in sediment archives. It demonstrates the importance of
244	correcting for grain size and lithological variability in order to interpret the detrital Li
245	isotope composition in terms of weathering intensity. It also shows that the $\Delta^7 Li_{\text{fine-source}}$
246	value is a function of the weathering intensity only for W/D values higher than 0.05.
247	Overall, the combined use of Li/Al, δ^7 Li and other more traditional weathering proxies
248	(like CIA) has great potential for quantifying both sedimentary recycling and weathering
249	intensity in sediment archives.

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350	FIGURE CAPTIONS
351	Figure 1. Li isotope composition as a function of Li concentration for large river
352	sediments. Gray color scale applies only to rivers draining mountain ranges (circles in
353	figure) and corresponds to the logarithm of the D_{90} values (which is the grain size below
354	which 90% of the grains are found in a given sample). For lowland rivers, grain size was
355	not measured, but all lowlands sediment are surface samples, which are generally fine-
356	grained (Filizola, 2003).
357	
358	Figure 2. A: The Li/Al ratio in fine sediments as a function of the proportion of shales in
359	the source of river sediments, as calculated using Nd isotopes or trace element ratios (see
360	the Data Repository [see footnote 1]). Black dotted line corresponds to the mixing
361	between shales and igneous rocks. Gray dotted lines labeled with numbers correspond to
362	the Li weathering index values (Li/Al)sediment_load / (Li/Al)source_material for various

363	proportion of shales. The best fit of the data ($R^2 = 0.82$), excluding the lowland rivers,
364	corresponds to a Li weathering index value of 0.9. The average igneous rock and shale
365	end members as given in Dellinger et al. (2014) and Sauzéat et al. (2015) are also
366	represented. B: Li weathering index (Li/Al)sediment_load / (Li/Al)source_material in the sediments
367	as a function of the weathering intensity.
368	
369	Figure 3. A: Li isotope composition of fine large river sediments corrected from the
370	source rock ($\Delta^7 Li_{fine-source}$). B: Li isotope composition of large river fine sediments
371	$(\delta^7 Li_{sediment_load})$, as a function of the weathering intensity (W/D).
372	
373	¹ GSA Data Repository item 2017xxx, methods, supplementing text, Figures DR1–DR6
374	and Tables DR1 and DR2, is available online at
375	http://www.geosociety.org/datarepository/2017/ or on request from
376	editing@geosociety.org.







Tracing weathering regimes using the Li isotope composition of detrital sediments

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GSA Data Repository

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Figure DR1, DR2, DR3, DR4, DR5 and DR6 Table DR1 and DR2 References

References

1. Analytical methods

Here, we briefly describe the analytical methods used for measurements of the Li and Nd isotope composition of sediments.

<u>Li isotopes:</u> the digestion, column chemistry and Li isotope measurement procedure for the sediments has been described in Dellinger et al., (2014). About 20 mg of sediment were digested in HF-HNO₃ at 100°C during 24 to 48 hours. Li was then separated from the matrix by ion-exchange chromatography using a method modified from James and Palmer (2000) and described in Dellinger et al. (2014). The lithium isotope composition was measured using a MC-ICP-MS Neptune (Thermo Scientific, Bremen) at IPGP (Paris). Details on the analytical procedure are available in Dellinger et al. (2014). Accuracy and reproducibility of the isotopic measurements were checked through repeated analyses of the basalt reference materials JB-2 yielded $\delta^7 \text{Li} = +4.47\pm0.53$ ($\pm 2\sigma$, n = 30 separations and 15 digestions).

<u>Nd isotopes</u>: Nd isotope composition was measured in the same aliquot used for Li isotopes. Nd was separated from the rest of the matrix by ion-exchange chromatography using the TRU-spec and Ln-spec resins and measured by MC ICP-MS Neptune in IPGP following the method described in Cogez et al., (2015).



Figure DR1: Maps of the river basins (elevation on the left, geology on the right) from this study, with the locations of the samples. The Geological maps are from Hartmann and Moosdorf (2013). The Changjiang samples are from Wang et al., (2015)

2. Provenance of fine river sediments

Here, we describe how the relative proportion of river sediment derived from the erosion of shales, igneous and high-grade metamorphic rocks is calculated for each river basin. The location of the samples and the geological map of each river basin are shown in Fig. (DR1). <u>Amazon River:</u> In the Amazon basin, the major rock types are sedimentary rocks in the Andes and foreland/lowlands, igneous rocks (andesites in the northern part of the Amazonian Andes and granitoids in their southern part) and granitic and metamorphic rocks in the Brazilian and Guyana shields. The Nd isotope composition (expressed here as $\varepsilon_{Nd}(0)$, the ratio in parts per 10⁴ (units) between the measured ¹⁴³Nd/¹⁴⁴Nd of the sample relative to the present-day value (0.512638) of the ¹⁴³Nd/¹⁴⁴Nd of the CHondritic Uniform Reservoir (CHUR)) is a reliable tracer of the provenance of river sediments in the Amazon basin because the major rock types have distinct ε_{Nd} values (Allègre et al., 1996; Roddaz et al., 2005; Viers et al., 2008). In the following we consider that river sediments in the Amazon Basin result from a binary mixture between two end members, such that the following mixing equation can be written:

$$\varepsilon_{Nd}(0)_{sed} = \varepsilon_{Nd}(0)_{sha} \times \gamma_{sha} + \varepsilon_{Nd}(0)_{Ign} \times (1 - \gamma_{sha})$$

With $\varepsilon_{Nd}(0)_{sed}$, $\varepsilon_{Nd}(0)_{sha}$ and $\varepsilon_{Nd}(0)_{Ign}$, respectively the Nd isotope composition of river sediment, shale and igneous rock. The proportion of shales-derived particles (" γ_{sha} " in mass %) can be calculated as:

$$\gamma_{sha}^{Nd} = \gamma_{sha} \times \frac{\left[Nd\right]_{sha}}{\left[Nd\right]_{sed}}$$

As the average Nd concentration of shale and igneous rocks is not significantly different (Condie, 1993), we consider that $\gamma_{sha} = \gamma_{sha}^{Nd}$. As explained below, the composition of the igneous end member ($\epsilon_{Nd}(0)_{Ign}$) varies between the different sub-basins, while the composition of the shale end member ($\epsilon_{Nd}(0)_{sha}$) is considered homogenous throughout the Amazon River basin.

The Beni River drains exclusively shales from the Bolivian Andes and we use the $\varepsilon_{Nd}(0)$ of its sediments to estimate the average $\varepsilon_{Nd}(0)$ value of Andean sedimentary rocks. The $\varepsilon_{Nd}(0)$ of Beni sediments ranges between -12.2 and -12.9, in good agreement with the range of $\varepsilon_{Nd}(0)$ values measured in Andean sedimentary rocks (-10 to -13; Pinto, 2003; Viers et al., 2008) and Neogene South Amazonian foreland basin sediments from Roddaz et al., (2005). Similarly, the

Yata river that drains exclusively the lowlands Tertiary sedimentary rocks (originated from the erosion of the region of the Andes) have a $\varepsilon_{Nd}(0)$ of -11.6, in good agreement with the $\varepsilon_{Nd}(0)$ value measured in Beni river sediments. Therefore, we assume a $\varepsilon_{Nd}(0)_{sha}$ of -12.0 ± 1.0 for the shale end member. The same value of -12.0 was also assumed for the metasedimentary endmember of the Madre de dios river basin by Basu et al., (1990). In addition, we consider that this value is also representative of the shales from the northern Andes. Indeed, the $\varepsilon_{Nd}(0)$ of Solimões sediments defines a trend with the Al concentration (grain size proxy) that tends toward a similar shale $\varepsilon_{Nd}(0)$ value (Fig. DR2) and Neogene North Amazonian foreland basin sediments from Roddaz et al., (2005) also plot on the same trend.



Figure DR2: $\varepsilon_{Nd}(0)$ as a function of the Al concentration in sediments. Solimões and Beni data are from Bouchez et al., (2011a), Dellinger et al., (2015) and Viers et al., (2008). Data for foreland basin sediments are from Roddaz et al., (2005). The grey band corresponds to the estimated $\varepsilon_{Nd}(0)$ of the mean shale end-member in the Amazon basin.

Regarding the igneous rock end member, the Nd isotope composition measured in Andean Neogene granitoid rocks from the Sierra Blanca in Peru have $\varepsilon_{Nd}(0)$ generally between -3 to 0 (Petford and Atherton, 1996). Pemian-Triassic and Carboniferous granitoids in southern Peru have a median $\varepsilon_{Nd}(0)$ of -4.7±0.6 (Miskovic, 2008). Accordingly, bed sediments from the

Solimões River draining the Peruvian and Ecuadorian Andes, and for which the major and trace element signature reflect an igneous rock provenance (Bouchez et al., 2011a; Dellinger et al., 2014), have $\varepsilon_{Nd}(0)$ ranging from -6.5 to -2.0. However, volcanic rocks in Ecuador (mostly andesites) have a higher $\varepsilon_{Nd}(0)$ ranging between +2 and +6 with (Bryant et al., 2006), and bed sediments from the Pastaza River derived from the erosion of andesite in Ecuador has a $\varepsilon_{Nd}(0)$ of +1.9 (Dellinger et al., 2015a). This high $\varepsilon_{Nd}(0)$ indicates that Pastaza sediments predominantly derive from the erosion of more recent (Quaternary) andesite. In the following, we use a $\varepsilon_{Nd}(0)_{Ign}$ of -3.0 ± 2.0 for the igneous end member of all the Andean-fed rivers, except for the Pastaza River for which we use a $\varepsilon_{Nd}(0)_{Ign}$ value of +2.0±1.0.



Fig. DR3: Sm/Nd as a function of the ε_{Nd} of lowland river sediments. Shield rock data are from Santos et al., (2000). The Tapajós and Pastaza rivers have a $\varepsilon Nd(0)$ and Sm/Nd close to the signature of shield rocks whereas Urucara and Negro river sediments contain shale-derived sediments. The shale end-member corresponds to the average composition of Beni river sediments.

Regarding the shield regions, the range of $\varepsilon_{Nd}(0)$ measured in rock samples from the Rio Negro, Central Amazon and Tapajos-Parima provinces is very large (-20 to -30) and positively correlated to the Sm/Nd ratio as a result of magmatic differentiation (Allègre et al., 1996; Basu et al., 1990; Hoorn et al., 2009; Roddaz et al., 2005; Santos et al., 2000). The Sm/Nd ratio of rivers draining the shield range between 0.15 to 0.20, and for a given Sm/Nd ratio, lowlands rivers generally have higher $\varepsilon_{Nd}(0)$ than cratonic rocks (Fig. DR3) and plot on a line trending towards the shale end member (defined as the mean Sm/Nd and $\varepsilon_{Nd}(0)$ of Beni River sediments). Only the Tapajós and Trombetas river sediments have a Nd isotope composition similar to that of the cratonic rocks in the Amazon basin whereas the Rio Negro and Urucara rivers have a $\varepsilon_{Nd}(0)$ intermediate between cratonic rocks and Andean-derived shales, consistently with the findings of Allègre et al., (1996). We consider therefore the Trombetas River composition as representative of the mean Nd isotope composition of the shield end member ($\varepsilon_{Nd}(0)_{Ign} = -20.0\pm 2.0$).

<u>Congo River</u>: The lithology of the Congo Basin is mainly composed of Archean and Proterozoic TTG along the basin border, and of sedimentary rocks in the center. To our knowledge, there is no available Nd isotope data on Congo rocks in the literature. However, Allègre et al. (1996) report $\varepsilon_{Nd}(0)$ values for the major Congo tributaries. The $\varepsilon_{Nd}(0)$ range from -12.9 for the Lobaye River to -18.2 for the Likouala. We also measured the $\varepsilon_{Nd}(0)$ on a lateritic soil profile sampled 100 km downstream from Bangui (Henchiri et al., 2016) and developed on TTG rocks. The Nd isotope values range from -18.1 to -18.9 (See table DR2) and is similar to the value of the Likouala.



Fig. DR4: Crustal Na/Sm for Congo rivers inferred by Gaillardet et al. (1995) as a function of the sediment ε_{Nd} . The Nd isotope composition of the cratonic end member is estimated to be -18.5 ± 0.5 based on the above trend, the Na/Sm of UCC (Rudnick and Gao, 2014) and the Nd

isotope composition of a soil profile developed on cratonic rocks in Bangui (this study). The Nd isotope composition of the shales end-member is estimated to be -12 ± 0.5 by projecting the typical Na/Sm value of global Phanerozoic shales (from Condie, 1993) on the mixing trend. This value is also compatible with the Nd isotope composition of the Lobaye River.

Interestingly, the $\varepsilon_{Nd}(0)$ of Congo tributaries sediments is negatively correlated to the Na/Sm ratio of the Congo tributaries continental crust (sediments + dissolved) inferred by Gaillardet et al., (1995) (Fig. DR4). This crustal Na/Sm ratio can be used to distinguish TTG (high Na/Sm) from shales (low Na/Sm resulting from Na depletion during ancient weathering processes). The Likouala has a TTG-like Na/Sm signature whereas the Lobaye has a signature close to that of shales. Altogether, the range of variation of $\varepsilon_{Nd}(0)$ in Congo sediments can be explained at first order as resulting from a mixture between a TTG igneous end member of high $\varepsilon_{Nd}(0)_{Ign}$ around -19 and a metasedimentary end member of low $\varepsilon_{Nd}(0)_{sha}$ (around -12), as for the Amazon. Using eqs. (1) and (2), we calculate that the proportion of shale-derived particles in the Congo River sediments is 40±10%. We note that the correlation between $\varepsilon_{Nd}(0)$ and Na/Sm is scattered and therefore that this method for determining the provenance of Congo River sediments gives only a first-order estimate of the proportion of shale-derived particles.

Ganges-Brahmaputra rivers: In the Ganges-Brahmaputra system, the source of sediments has been characterized using Nd and Sr isotopes by previous studies (Galy and France-Lanord, 2001; Lupker et al., 2013; Singh et al., 2008; Singh and France-Lanord, 2002). The Ganges sediments are mostly derived from the erosion of the high-grade metamorphic rocks from the High Himalaya Crystalline (HHC, about 70%) and the rest of the sediments from the low-grade metasedimentary rocks of the Lesser Himalayas (LH, 30%). For the Brahmaputra, Singh and France-Lanord (2002) estimate that in average, between 50 and 70% of the sediments are derived from the erosion of the HHC, about 10% from the LH and the rest from the erosion of the Trans-Himalayan Batholith (igneous rocks). However, the provenance of the Brahmaputra sediments can be quite variable depending upon the depth and the sampling date (Lupker et al., 2013). As the Nd and Sr isotope composition of the Brahmaputra sediment sample from this study (BR8210) has been measured by Lupker et al., (2013), we can use these data along with the estimate of the Himalayan rock end-members (Galy and France-Lanord, 2001; Singh et al., 2008; Singh and France-Lanord, 2002) to determine the provenance for this sample. We calculate that

for this Brahmaputra sample, the proportion of LH, and therefore of shales-derived sediments, is $20\pm10\%$.

Mackenzie River: The upper crust drained by the main Mackenzie tributaries is composed by a mixture between shales and igneous rocks having granodioritic to granitic composition (Driver et al., 2000; Reeder et al., 1972). The Nd and Sr isotope composition has not been measured on sediments from the Mackenzie tributaries. However, a large dataset of shale chemical composition from the Liard River basin is available in Ross and Bustin, (2009). Dellinger, (2013) shows that some insoluble elements (Cr, Ti, Eu, Cs) have significantly distinct concentration between shales and granitic rocks. It is therefore possible to use these element contents to calculate the proportion of shale-derived and igneous rock-derived particles in the fine sediments of the Mackenzie rivers (Fig. DR.5). The Peel River that drain almost exclusively shales have sediments having chemical signature similar to the mean composition of shales from Ross and Bustin (2009). Other rivers have lower Al-normalized concentrations showing that they contain a larger, yet small fraction of igneous-derived material compared to the Peel. We estimate the composition of the shale end member using the data from Ross and Bustin (2009), Garzione et al., (1997) and Morrow, (1999). For the shale end member, we estimate that $Ti/Al = 0.057 \pm 0.002$ (median value $\pm 2\sigma/\sqrt{n}$, n=101), Cr/Al = 1.54 $\pm 0.12 \times 10^{-3}$ (n=106) and Cs/Al = 0.10 $\pm 0.01 \times 10^{-3}$ (n=50). For the granitic end member, we use data from Driver et al., (2000) and Heffernan, (2004): Ti/Al = 0.030 ± 0.005 (n=60), Cr/Al = $0.13\pm0.29\times10^{-3}$ (n=23) and median Cs/Al = $0.050\pm0.005\times10^{-3}$ (n=50). We calculate that the Mackenzie tributaries contain between 70 and 100% of shale-derived sediments.



Fig. DR5: Cr/Al as a function of the Cs/Al and Ti/Al ratio for Mackenzie River fine sediments. The igneous and shales end members are also represented (as discussed in the section 2 of the data repository).

<u>Western Cordillera</u>: Most of the bedrock of the Canadian western cordillera rivers is formed of volcanic rocks, with a minor contribution of shales. The Sr isotope composition of both dissolved and suspended material for the Skeena and Stikine rivers is unradiogenic (<0.707; Gaillardet et al., 2003) so we consider that the proportion of shales is lower than $10\pm10\%$.

<u>Changjiang</u>: For the Changjiang, we determined the proportion of shales by using the Li concentration in sediments and the relation between the Li concentration and proportion of shales defined by other rivers (see section "The influence of lithology" in the main text). We note that samples CJ33 and CJ34 correspond to rivers draining lowland areas (Wang et al., 2015). Hence, the provenance method based on Li might not be applicable to these rivers. However, their weathering intensity is quite low compared to lowland rivers of the Amazon. In addition, the Li content of the sediments is higher than 85 ppm indicating that even if they have lost a significant proportion of Li, their main source is the erosion of shales (in agreement with the lithology in these two river basin; Wang et al., 2015). Based on the uncertainty on the trend between Li concentration and proportion of shales defined by other rivers, the estimated uncertainty on the proportion of shale-derived particles for Changjiang samples is $\pm 15\%$.

3. Calculation of the $\delta^7 \text{Li}_{\text{rock}}$ and $(\text{Li}/\text{Al})_{\text{rock}}$

To calculate the Li isotope composition of the bedrock, $\delta^7 \text{Li}_{\text{rock}}$ (which is the $\delta^7 \text{Li}$ that the sediments would have in absence of chemical weathering) and (Li/Al)_{rock} values, we can use the proportion of shales calculated before (γ_{sha}) in combination with the mixing equation in Dellinger et al., (2014):

$$\left(\frac{\text{Li}}{\text{Al}}\right)_{\text{rock}} = \left(\frac{\text{Li}}{\text{Al}}\right)_{\text{sha}} \times \gamma_{\text{sha}} + \left(\frac{\text{Li}}{\text{Al}}\right)_{\text{Ign}} \times (1 - \gamma_{\text{sha}})$$

and

$$(\delta^{7}\text{Li})_{\text{rock}} = (\delta^{7}\text{Li})_{\text{sha}} \times \left(\frac{(\text{Li}/_{\text{Al}})_{\text{sha}}}{(\text{Li}/_{\text{Al}})_{\text{sed}}}\right) \times \gamma_{\text{sha}} + (\delta^{7}\text{Li})_{\text{Ign}} \times \left(\frac{(\text{Li}/_{\text{Al}})_{\text{Ign}}}{(\text{Li}/_{\text{Al}})_{\text{sed}}}\right) \times (1 - \gamma_{\text{sha}})$$

With the shale end member composition being $\delta^7 \text{Li}$ of $-0.5\pm1\%$ and Li/Al of $1.0\pm0.1\times10^{-3}$ and the igneous end member $\delta^7 \text{Li}$ of $+3.5\pm1.5\%$, Li/Al of $0.30\pm0.12\times10^{-3}$ (Dellinger et al., 2014; Sauzéat et al., 2015). For the Congo River, Henchiri et al., (2016) show that the TTG of the Congo basin have possibly high $\delta^7 \text{Li}$ (+6 to +7‰) due to their Archean age. Therefore, we use a higher $\delta^7 \text{Li}_{ign}$ end member of $+5\pm2\%$ for the Congo River.

For the Ganges-Brahmaputra (G-B) system, we note that G-B sands, which are derived mostly from the erosion of high-grade metasedimentary rocks, have a distinct composition (δ^7 Li of 0 ± 1‰, Li/Al of 0.40 ± 0.10 × 10⁻³; Dellinger et al., 2014) relative to igneous and sedimentary rocks. Hence, for the Ganges-Brahmaputra, we consider a mixing between a shale end member and a high-grade metamorphic end member having a δ^7 Li of 0±1‰ and a Li/Al ratio of 0.40±0.10×10⁻³ (Dellinger et al., 2014).

4. Comparison between the Li depletion in sediments and the fraction of Li transported into the dissolved load

As Li is a soluble element, Li concentration in river sediments is sensitive to weathering. In particular, for a given proportion of shales, lowland rivers have a lower Li/Al ratio than rivers draining mountain ranges (Fig. 2). Yet, this Li-depletion in lowland rivers could be explained by i) the loss of Li by weathering processes and/or ii) an incorrect estimation of the bedrock (*i.e.* the proportion of shales and/or the composition of the rock end members). Following the approach of Gaillardet et al., (1999), we can test these two hypotheses using a steady state assumption. At steady state, the fraction of Li transported as dissolved should be equal to the Li depletion (Li weathering index (Li/Al)_{sed}/(Li/Al)_{rock}, see main text) in the sediments. The proportion of Li riverine export occurring as dissolved (w^{Li} as defined by Bouchez et al., 2013) can be calculated using the Li concentration in the dissolved and suspended load and the concentration of suspended sediment transported by the rivers.

$$w^{\text{Li}} = \frac{1}{1 + \frac{[\text{Li}]_{\text{sed}}}{[\text{Li}]_{\text{diss}}} \times [\text{SPM}]}$$

With $[Li]_{sed}$, $[Li]_{diss}$ and [SPM] being respectively the Li concentrations in sediment, and dissolved load, and the suspended sediment concentration. The comparison between $(1-w^{Li})$ and the weathering index (Fig. DR6) shows that at first order, there is a relatively good agreement between the amount of Li depletion in the sediments and the fraction of dissolved Li export, within uncertainties.



Figure DR6: Proportion of solid Li transported by the river $(1-w^{Li})$ as a function of the Li weathering index calculated as $(Li/Al)_{sed} / (Li/Al)_{rock}$. The 1:1 line corresponds to the line for which $(1-w^{Li})$ is equal to $(Li/Al)_{sed} / (Li/Al)_{rock}$, meaning that the fraction of Li dissolved export is compatible with that of the solid export, a requisite for steady state denudation.

This calculation does not take into account the change in Li concentration in the sediments with depth. Indeed, in large rivers draining mountain ranges, the grain size distribution of the suspended sediment load is generally bimodal with a fine mode composed of clay and silt particles and a coarse mode composed of sands (Bouchez et al., 2011a; Dellinger, 2013; Lupker et al., 2012). The fine sediments (< 63 μ m) are relatively homogeneously distributed in the water

column while coarse sediments (size > 63 µm) concentration strongly increases from the surface toward the bottom of the river channel (Bouchez et al., 2011b; Galy et al., 2007; Lupker et al., 2011). As coarse sediments are mostly composed of quartz, the Li concentration in sediments decreases toward the bottom of the channel because of dilution by increasing quartz amount. This implies that the $(1-w^{Li})$ values calculated from surface sediments are underestimated. However, as shown by Dellinger et al., (2015), the difference is relatively small (less than 0.05 units of w^{Li} values). In addition, we note that in lowland rivers suspended sediments collected at any depth contain a negligible amount of coarse (> 63µm) particles (Meade, 1985). Therefore, the conclusions about the agreement between $(1-w^{Li})$ values and $(Li/Al)_{sed} / (Li/Al)_{rock}$ stands whether these values are depth-integrated or calculated with surface sediments only.

Sample name	Basin	River	Location	Longitude	Latitude	References for Li isotope data	Surface	Discha rge	D ₉₀	AI	Li	¹⁴³ Nd/ ¹⁴⁴ Nd	ε _{Nd} (0)	δ ⁷ Li	Li/Al (×10 ⁻³)	CIA	[Li] _{diss}	TDS _{Si}
							(km²)	(m ³ /s)	(µm)	(µg/g)	(µg/g)			(‰)	(g/g)		(mmol/L)	(mg/L)
Amazon Lowlar	nds																	
AM01-17	Amazon	Orthon	Riberalta	66°02'11,0" W	10°48'39,9" S	Dellinger et al., (2014, 2015)	33485	475		89835	59	0.512158	-9.37	-5.7	0.66	82	0.31	19.6
AM01-18	Amazon	Yata	Mouth	65°35'00,2" W	10°58'44,8" S	Dellinger et al., (2014, 2015)	20770	222		115390	100	0.512043	-11.61	-5.3	0.87	86	0.08	10.4
AM01-27	Amazon	Trombetas	Mouth	N.D.	N.D.	Dellinger et al., (2014, 2015)	250000	6200		115312	20	0.511609	-20.08	-5.1	0.17	92	0.037	7.8
AM01-22 MES	Amazon	Negro	Paricatuba	N.D.	N.D.	Dellinger et al., (2014, 2015)	691000	29000		113016	44	0.511930	-13.81	-2.7	0.39	87	0.031	4.3
AM6/1-09 MES	Amazon	Urucara	Mouth	N.D.	N.D.	Dellinger et al., (2014, 2015)	150000	3000		110691	43	0.511895	-14.50	-4.0	0.39	87	0.049	8.7
AM01-29	Amazon	Tapajos	Mouth	N.D.	N.D.	Dellinger et al., (2014, 2015)	490000	13500		129350	27	0.511698	-18.33	-6.0	0.20	93	0.038	11.5
Congo																		
C89-63 PK 1105	Congo	Congo	Mouth	N.D.	N.D.	Henchiri et al., (2016), this study	3700000	38026			29	0.511796	-16.43	-6.2			0.10	15.0
C89-65 PK 1105	Congo	Congo	Mouth	N.D.	N.D.	Henchiri et al., (2016), this study	3700000	38026		117769	23	0.511833	-15.70	-5.1	0.19	92	0.10	15.0
Amazon Andes																		
AM01_08	Amazon	Alto Beni	Mouth	67°40'18' W	15°07'24" S	Dellinger et al., (2014, 2015)	21197	750		84800	79	0.512011	-12.24	-1.8	0.93	74	1.62	26.4
AM07_04	Amazon	Beni	Rurrenabaque	67°32'05,2" W	14°28'26,8" S	Dellinger et al., (2014, 2015)	69980	2153	44	95289	87	0.511998	-12.49	-3.1	0.91	76	1.28	17.4
AM07_09	Amazon	Beni	Riberalta	66°05'53,8" W	10°58'44,8" S	Dellinger et al., (2014, 2015)	114379	3772	50	83319	78	0.511994	-12.56	-2.9	0.94	76	0.57	14.9
AM05_19	Amazon	Madeira	Foz Madeira	58°48'30,1" W	03°27'00,9" S	Dellinger et al., (2014, 2015)	1325000	31250	36	111822	99	0.512097	-10.55	-3.4	0.89	78	0.18	12.8
AM06_65	Amazon	Amazonas	Obidos	55°30'18,6" W	01°56'05,3" S	Dellinger et al., (2014, 2015)	4618750	182200	27	107370	81	0.512141	-9.70	-3.6	0.75	78	0.14	11.4
AM08_13	Amazon	Ucayali	Jenaro Herrera	73°40'28" W	04°54'28" S	Dellinger et al., (2014, 2015)	352593	12090	50	85664	58	0.512158	-9.37	-1.9	0.73	68	0.66	15.6
AM06_14	Amazon	Solimoes	Manacapuru	60°33'11,0" W	03°19'41,0" S	Dellinger et al., (2014, 2015)	2148000	103000	35	100535	58	0.512101	-10.48	-2.6	0.58	76	0.14	13.1
AM08_36	Amazon	Pastaza	Embouchure	N.D.	N.D.	Dellinger et al., (2014, 2015)	39000	2770	61	97031	26	0.512512	-2.45	-0.2	0.27	68	0.14	24.2
AM07_14	Amazon	Madre de Dios	Riberalta	66°05'54" W	10°57'46" S	Dellinger et al., (2014, 2015)	124231	5602	35	94447	75	0.512106	-10.38		0.79	77	0.22	14.6
Mackenzie																		
CAN09_51	Mackenzie	Red Arctic	Tsiigehtchic	133°46'44.3" W	67°25'34.2" N	Dellinger et al., (2014)	18600	159	23	79022	78.1	0.512066	-11.16	-2.4	0.99	77	0.89	8.0
CAN09_04	Mackenzie	Liard	Fort_Simpson	121°17'6.2" W	61°50'8.6" N	Dellinger et al., (2014)	275000	2450	49	64955	51.2	0.511904	-14.31	-1.7	0.79	73	0.63	8.7
CAN09_47	Mackenzie	Mackenzie	Tsiigehtchic	133°43'23.8" W	67°27'30.1" N	Dellinger et al., (2014)	1680000	8990		76477	67.5	0.512013	-12.19	-1.9	0.88	76	0.57	9.5
CAN10_32	Mackenzie	Middle Channel	Inuvik	134°4'49.8" W	68°24'33.3" N	Dellinger et al., (2014)	1805200		14	90608	79.3	0.511979	-12.86	-1.7	0.88	79		
CAN09_40	Mackenzie	Peel	Fort_McPherson	134°52'8.6" W	68°19'56.6" N	Dellinger et al., (2014)	70600	686	18	80981	81.5	0.512062	-11.23	-2.1	1.01	77	0.71	8.8
CAN09_28	Mackenzie	Slave	Fort_Smith	111°53'25.7" W	60°0'59.5" N	Dellinger et al., (2014)	616400	3200	17	77069	52.8	0.512014	-12.18	-1.3	0.68	76	0.60	12.0
Changjiang																		
CJ1	Changjiang	Jinshajiang	Panzhihua	101°37.551' E	26°34.503' N	Wang et al., (2015)	285556	1309	24	80546	48.5			-2.2	0.60	69	2.07	18.6
CJ4	Changjiang	Jinshajiang	Yibin	104°33.385' E	28°42.164' N	Wang et al., (2015)	485000	8239	22	82252	48.7			-0.5	0.59	73	0.97	18.5
CJ14	Changjiang	Changjiang	Chongqing	N.D.	N.D.	Wang et al., (2015)	867000	10945	24	84970	53.8			-1.0	0.63	71	0.78	13.4
CJ60	Changjiang	Changjiang	Yueyang	113°18.739' E	29°38.079' N	Wang et al., (2015)	1250000	15844			76.6			-1.3			0.37	14.2
CJ52	Changjiang	Changjiang	Wuhan	114°29.876' E	30°40.747' N	Wang et al., (2015)	1488000	22340	45	83606	61.9			-1.6	0.74	72	0.44	16.9
CJ36	Changjiang	Changjiang	Datong	117°38.324' E	30°46.085' N	Wang et al., (2015)	1705000	28266	26	98071	70.9			-1.7	0.72	76	0.27	16.7
CJ2	Changjiang	Yalongjiang	Panzhihua	101°49.200' E	26°40.127' N	Wang et al., (2015)	128444	1857		161917	41.3			0.1	0.26	63	0.68	17.0
CJ5	Changjiang	Minjiang	Yibin	104°33.721' E	28°48.779' N	Wang et al., (2015)	133000	2732	33	83956	67.0			0.0	0.80	67	0.45	18.3
CJ33	Changjiang	Xiangjiang	Changsha	112°56.795' E	28°07.364' N	Wang et al., (2015)	94700	2405	15	108398	86.7			-3.4	0.80	82	0.16	13.3
CJ34	Changjiang	Ganjiang	Nanchang	115°52.053' E	28°40.206' N	Wang et al., (2015)	80900	2104		103092	88.6			-4.7	0.86	80	0.25	22.4
Ganges-Brahma	aputra																	
IND99-19 MES	Ganges	Ganga	Harding bridge	89°1'28.739" E	24°3'10.44" N	Dellinger et al., (2014)	1050000	15622		94322	55.0	0.511725	-17.81	-1.02	0.58		0.49	29.8
BR8210	Brahmaputra	Brahmaputra	Jamuna bg.	N.D.	N.D.	Dellinger et al., (2014)	580000	16161		87063	47	0.511897	-14.45	-1.38	0.54	66	1.08	11.8
Western cordille	era																	
CAN99-28 MES	Skeena	Skeena	Kitwanga	N.D.	N.D.	This study	42200	962		83647	28.7			2.0	0.34	64	0.05	7.9
CAN99-39 MES	Stikine	Stikine	Iskut	N.D.	N.D.	This study	29300	755		75071	24.5			1.2	0.33	63	0.10	7.2

Table DR1: Data for all the clastic river sediment samples from this study. The "references" column indicate the source of Li isotope and concentration data used for this study. CIA values have been calculated using similar formula as in McLennan, (1993). Silicate weathering rates (calculated as $Q \times ([Na]_{sil}+[K]_{sil}+[Ca]_{sil}+[Mg]_{sil})$, with "[X]_{sil}" the silicate-derived dissolved concentration of the element X and "Q" the discharge) and erosion rates are from multiple sources (Armijos et al., 2013; Dellinger et al., 2015b; Gaillardet et al., 1997; Guyot et al., 1996; Moquet et al., 2011; Wittmann et al., 2010). [Li]_{diss} corresponds to the dissolved Li concentration, TDS_{sil} to the total dissolved load deriving from silicates weathering (in mg/L). Sediment concentration corresponds to the mean measured suspended sediment concentration for each river (in mg/L).

Sample name	River	Weathering Intensity (W/D)	Proportion of shales	δ ⁷ Li _{rock}	∆ ⁷ Li _{sed-rock}	1-w ^{Li}	(Li/Al) _{sed} / (Li/Al) _{rock}	References
				(‰)	(‰)			
Amazon Lowlar	nds							
AM01-17	Orthon	0.137	0.73 ± 0.12	-0.10 ± 0.89	-5.58 ± 1.07	0.77 ± 0.09	0.83 ± 0.11	Dellinger et al., (2014, 2015)
AM01-18	Yata	0.259	0.95 ± 0.14	-0.36 ± 0.93	-4.90 ± 1.12	0.84 ± 0.07	0.91 ± 0.12	Dellinger et al., (2014, 2015)
AM01-27	Trombetas	0.504	0.00 ± 0.10	3.50 ± 1.50	-8.63 ± 1.75	0.38 ± 0.14	0.58 ± 0.10	Dellinger et al., (2014, 2015)
AM01-22 MES	Negro	0.291	0.73 ± 0.10	-0.11 ± 0.90	-2.61 ± 1.20	0.68 ± 0.15	0.49 ± 0.07	Dellinger et al., (2014, 2015)
AM6/1-09 MES	Urucara	0.367	0.65 ± 0.11	0.04 ± 0.90	-4.02 ± 1.10	0.66 ± 0.12	0.52 ± 0.08	Dellinger et al., (2014, 2015)
AM01-29	Tapajos	0.581	0.28 ± 0.14	1.37 ± 1.11	-7.37 ± 1.50	0.45 ± 0.18	0.42 ± 0.20	Dellinger et al., (2014, 2015)
Congo								
C89-63 PK 1105	Congo	0.396	0.35 ± 0.10			0.51 ± 0.13		Henchiri et al., (2016), this study
C89-65 PK 1105	Congo	0.396	0.48 ± 0.11	1.00 ± 1.11	-6.05 ± 1.20	0.45 ± 0.12	0.31 ± 0.07	Henchiri et al., (2016), this study
Amazon Andes								
AM07 04	Beni	0.005	1.00 ± 0.10	-0.50 ± 1.00	-2.59 ± 1.12	0.97 ± 0.03	0.91 ± 0.08	Dellinger et al., (2014, 2015)
AM07_09	Beni	0.015	1.00 + 0.10	-0.50 ± 1.00	-2.38 + 1.12	0.95 + 0.03	0.94 + 0.08	Dellinger et al. (2014, 2015)
AM05_19	Madeira	0.049	0.85 + 0.12	-0.30 + 1.00	-3.08 + 1.10	0.00 ± 0.00	0.98 + 0.12	Dellinger et al. (2014, 2015)
AM06_65	Amazonas	0.058	0.76 + 0.12	-0.2 + 1.00	-3.43 + 1.09	0.94 + 0.03	0.00 ± 0.12	Dellinger et al. (2014, 2015)
AM08_13	Licavali	0.020	0.70 ± 0.02	-0.10 + 0.89	-1.80 + 1.05	0.87 + 0.06	0.02 + 0.10	Dellinger et al. (2014, 2015)
AM06_14	Solimooo	0.025	0.00 ± 0.00	0.10 ± 0.05	-1.00 ± 1.00	0.07 ± 0.00	0.60 ± 0.10	Dellinger et al., (2014, 2015)
AM09 26	Bootozo	0.090	0.80 ± 0.12	1 10 + 0.02	1 20 ± 1.10	0.00 ± 0.00	0.09 ± 0.15	Dellinger et al., (2014, 2015)
AM07_14	Fasiaza Madra da Dian	0.025	0.02 ± 0.10	1.10 1 0.93	-1.20 ± 1.14	0.05 ± 0.02	0.02 + 0.10	Dellinger et al., (2014, 2015)
Maakanzia	Maure de Dios	0.035	0.02 ± 0.10			0.55 ± 0.05	0.52 ± 0.10	Denniger et al., (2014, 2015)
CANOD 51	Rod Arotio	0.002	0.05 + 0.10	0.47 ± 0.01	1.04 + 1.10	0.05 ± 0.02	1.04 + 0.10	Dollinger et al. (2014)
CANO9_31	Lierd	0.003	0.35 ± 0.10	-0.47 ± 0.91	-1.54 ± 1.10	0.95 ± 0.03	1.04 1 0.10	Dellinger et al., (2014)
CAN09_04	Liaru	0.007	0.75 ± 0.10	-0.04 ± 0.91	-1.02 ± 1.08	0.87 ± 0.06	0.97 ± 0.12	Dellinger et al., (2014)
CAN09_47	Middle Chennel	0.016	0.60 ± 0.10	-0.20 ± 0.91	-1.72 ± 1.06	0.67 ± 0.06	1.00 ± 0.12	Dellinger et al., (2014)
CANTO_32	Middle Charnel	0.000	0.70 ± 0.10	-0.04 ± 0.91	-1.05 ± 1.05		1.10 ± 0.12	Dellinger et al., (2014)
CAN09_40	Peel	0.006	1.00 ± 0.10	-0.50 ± 0.91	-1.56 ± 1.10	0.94 ± 0.03	1.01 ± 0.10	Dellinger et al., (2014)
CANU9_28	Slave	0.020	0.65 ± 0.10	0.15 ± 0.91	-1.45 ± 1.07	0.82 ± 0.08	0.92 ± 0.14	Dellinger et al., (2014)
Changjiang								
CJ1	Jinshajiang	0.008	0.50 ± 0.15	0.82 ± 1.13	-3.02 ± 1.26	0.80 ± 0.09		Wang et al., (2015)
CJ4	Jinshajiang	0.016	0.50 ± 0.15	0.78 ± 1.16	-1.28 ± 1.20	0.79 ± 0.09		Wang et al., (2015)
CJ14	Changjiang	0.003	0.58 ± 0.15	0.48 ± 1.03	-1.48 ± 1.20	0.92 ± 0.04		Wang et al., (2015)
CJ60	Changjiang	0.006	0.84 ± 0.15	0.49 ± 1.02	-1.79 ± 1.13	0.96 ± 0.03		Wang et al., (2015)
CJ52	Changjiang	0.012	0.68 ± 0.15	-0.3 ± 0.99	-1.29 ± 1.11	0.92 ± 0.04		Wang et al., (2015)
CJ36	Changjiang	0.010	0.78 ± 0.15	-0.2 ± 0.99	-1.52 ± 1.10	0.95 ± 0.03		Wang et al., (2015)
CJ2	Yalongjiang	0.032	0.38 ± 0.15	1.40 ± 1.20	-1.30 ± 1.63	0.64 ± 0.12		Wang et al., (2015)
CJ5	Minjiang	0.019	0.74 ± 0.15	-0.1 ± 0.97	0.07 ± 1.09	0.92 ± 0.04		Wang et al., (2015)
CJ33	Xiangjiang	0.014	0.92 ± 0.15	-0.4 ± 0.93	-3.04 ± 1.10	0.92 ± 0.04		Wang et al., (2015)
CJ34	Ganjiang	0.023	1.00 ± 0.15	-0.50 ± 1.00	-4.20 ± 1.20	0.90 ± 0.06		Wang et al., (2015)
Ganges-Brahm	aputra							
IND99-19 MES	Ganga	0.026	0.30 ± 0.10	-0.40 ± 0.95	-0.62 ± 1.00	0.95 ± 0.03	0.99 ± 0.15	Dellinger et al., (2014)
BR8210	Brahmaputra	0.013	0.20 ± 0.10	0.33 ± 1.10	-1.71 ± 1.30	0.85 ± 0.07	1.02 ± 0.12	Dellinger et al., (2014)
Western cordill	era							
CAN99-28 MES	Skeena	0.016	0.00 ± 0.10	3.50 ± 1.10	-1.45 ± 1.00	0.97 ± 0.03	1.02 ± 0.10	This study
CAN99-39 MES	Stikine	0.007	0.10 ± 0.10	2.06 ± 1.20	-0.86 ± 1.00	0.96 ± 0.03	0.86 ± 0.10	This study

Table DR1 (next): The parameter "1-w^{Li}" corresponds to the proportion of Li transported in the solid form by the river. The parameter " $(Li/Al)_{sed} / (Li/Al)_{rock}$ " is the ratio between the Li/Al of the sediments and the Li/Al of the rock.

Sample name	٤ _{Nd}	¹⁴³ Nd/ ¹⁴⁴ Nd
Betou 1A	-18.30	0.511700
Betou 1B	-18.13	0.511709
Betou 1C	-18.16	0.511707
Betou 2A	-18.44	0.511693
Betou 2B	-18.92	0.511668

Table DR2: Nd isotope composition of the Soil profile samples from the Congo river basin. The $\varepsilon_{Nd}(0)$ are calculated by normalizing to the CHUR value (0.512638). For details about the samples, see Henchiri et al., (2016).

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CAN99-39 ME	CAN99-28 ME	BR8210	Ganges-Brahi	CJ34	CJ33	CJ5	CJ2	CJ36	CJ52	CJ60	CJ14	CJ4	CJ1	Changjiang	CAN09_28	CAN09_40	CAN10_32	CAN09_47	CAN09_04	CAN09_51	Mackenzie	AM07_14	AM08_36	AM06_14	AM08_13	AM06_65	AM05_19	AM07_09	AM07_04	AM01_08	Amazon Ande	C89-65 PK 110	C89-63 PK 110	Congo	AM01-29	AM6/1-09 MES	AM01-22 MES	AM01-27	AM01-18	AM01-17	Amazon Lowl		Sample name
S Stikine	3 Skeena	Brahmaputra	naputra Gannes	Changjiang		Mackenzie	Mackenzie	Mackenzie	Mackenzie	Mackenzie	Mackenzie		Amazon	ŝ	15 Congo	15 Congo		Amazon	Amazon	Amazon	Amazon	Amazon	Amazon	ands		Basin																	
Stikine	Skeena	Brahmaputra	Ganna	Ganjiang	Xiangjiang	Minjiang	Yalongjiang	Changjiang	Changjiang	Changjiang	Changjiang	Jinshajiang	Jinshajiang		Slave	Peel	Middle Channel	Mackenzie	Liard	Red Arctic		Madre de Dios	Pastaza	Solimoes	Ucayali	Amazonas	Madeira	Beni	Beni	Alto Beni		Congo	Congo		Tapajos	Urucara	Negro	Trombetas	Yata	Orthon			River
lskut	Kitwanga	Jamuna bg.	Harding bridge	Nanchang	Changsha	Yibin	Panzhihua	Datong	Wuhan	Yueyang	Chongqing	Yibin	Panzhihua		Fort_Smith	Fort_McPhersor	Inuvik	Tsiigehtchic	Fort_Simpson	Tsiigehtchic		Riberalta	Embouchure	Manacapuru	Jenaro Herrera	Obidos	Foz Madeira	Riberalta	Rurrenabaque	Mouth		Mouth	Mouth		Mouth	Mouth	Paricatuba	Mouth	Mouth	Riberalta			Location
N.D.	N.D.	N.D.	89°1'28 739" F	115°52.053' E	112°56.795' E	104°33.721' E	101°49.200' E	117°38.324' E	114°29.876' E	113°18.739' E	N.D.	104°33.385' E	101°37.551' E		111°53'25.7" W	1 134°52'8.6" W	134°4'49.8" W	133°43'23.8" W	121°17'6.2" W	133°46'44.3" W		66°05'54" W	N.D.	60°33'11,0" W	73°40'28" W	55°30'18,6" W	58°48'30,1" W	66°05'53,8" W	67°32'05,2" W	67°40'18' W		N.D.	N.D.		N.D.	N.D.	N.D.	N.D.	65°35'00,2" W	66°02'11,0" W			Longitude
N.D.	N.D.	N.D.	24°3'10 44" N	28°40.206' N	28°07.364' N	28°48.779' N	26°40.127' N	30°46.085' N	30°40.747' N	29°38.079' N	N.D.	28°42.164' N	26°34.503' N		′ 60°0′59.5" N	68° 19'56.6" N	68°24'33.3" N	/ 67°27'30.1" N	61°50'8.6" N	/ 67°25'34.2" N		10°57'46" S	N.D.	03°19'41,0" S	04°54'28" S	01°56′05,3" S	03°27'00,9" S	10°58'44,8" S	14°28'26,8" S	15°07'24" S		N.D.	N.D.		N.D.	N.D.	N.D.	N.D.	10°58'44,8" S	10°48'39,9" S			Latitude
This study	This study	Dellinger et al., (2014)	Dellinger et al (2014)	Wang et al., (2015)		Dellinger et al., (2014)		Dellinger et al., (2014, 2015)		Henchiri et al., (2016), this study	Henchiri et al., (2016), this study		Dellinger et al., (2014, 2015)			References for Li isotope data																											
29300	42200	580000	1050000	80900	94700	133000	128444	1705000	1488000	1250000	867000	485000	285556		616400	70600	1805200	1680000	275000	18600		124231	39000	2148000	352593	4618750	1325000	114379	08669	21197		3700000	3700000		490000	150000	691000	250000	20770	33485		(km²)	Surface
755	962	16161	15600	2104	2405 15	2732 33	1857	28266 26	22340 45	15844	10945 24	8239 22	1309 24		3200 17	686 18	14	0668	2450 49	159 23		5602 35	2770 61	103000 35	12090 50	182200 27	31250 36	3772 50	2153 44	750		38026	38026		13500	3000	29000	6200	222	475		(m ³ /s) (µn	Dischar D _{gc} ge
75071	83647	87063	04399	103092	108398	83956	161917	98071	83606		84970	82252	80546		77069	80981	90608	76477	64955	79022		94447	97031	100535	85664	107370	111822	83319	95289	84800		117769			129350	110691	113016	115312	115390	89835		(μg/g) (r	à
24.5	28.7	47 0	лл О О	88.6	86.7	67.0	41.3	70.9	61.9	76.6	53.8	48.7	48.5		52.8 0	81.5 0	79.3 0	67.5 0	51.2 0	78.1 0		75 0	26 0	58 0	58 0	81 0	0 66	78 0	87 0	79 0		23 0	29 0		27 0	43 0	44 0	20 0	100 0	59 0		(g/g/l)	Ľ.
		.511897	511705												.512014	.512062	.511979	.512013	.511904	.512066		.512106	.512512	.512101	.512158	.512141	.512097	.511994	.511998	.512011		.511833	.511796		.511698	.511895	.511930	.511609	.512043	.512158			43Nd/144Nd
		-14.45	-17 81												-12.18	-11.23	-12.86	-12.19	-14.31	-11.16		-10.38	-2.45	-10.48	-9.37	-9.70	-10.55	-12.56	-12.49	-12.24		-15.70	-16.43		-18.33	-14.50	-13.81	-20.08	-11.61	-9.37			ε _{Nd} (0)
1.2	20	-1.38	-1 03	-4.7	-3.4	0.0	0.1	-1.7	-1.6	-1.3	-1.0	-0.5	-2.2		 β.	2.1	-1.7	-1.9	-1.7	-2.4			-0.2	-2.6	-1.9	-ა .6	-3.4	-2.9		-1.8		-5.1	-6.2		- 6 .0	-4.0	-2.7	5.1	-β- 3	-5.7		(‰)	δ ⁷ Li
0.33	0.34	0.54	0 58	0.86	0.80	0.80	0.26	0.72	0.74		0.63	0.59	0.60		0.68	1.01	0.88	0.88	0.79	0.99		0.79	0.27	0.58	0.73	0.75	0.89	0.94	0.91	0.93		0.19			0.20	0.39	0.39	0.17	0.87	0.66		(g/g)	Li/AI (×10 ⁻³)
63	64	66		80	82	67	63	76	72		71	73	69		76	77	79	76	73	77		77	68	76	68	78	78	76	76	74		92			93	87	87	92	86	82			CIA
0.10	0.05	1.08	0 49	0.25	0.16	0.45	0.68	0.27	0.44	0.37	0.78	0.97	2.07		0.60	0.71		0.57	0.63	0.89		0.22	0.14	0.14	0.66	0.14	0.18	0.57	1.28	1.62		0.10	0.10		0.038	0.049	0.031	0.037	0.08	0.31		(mmol/L)	[Li] _{diss}

0 CS9-63 PK 1105 Congo 5.3 25 8.1 0 CS8-65 PK 1105 Congo 5.3 25 8.1 4 AM07_08 Alto Beni 20.5 29.5 3403 360 4 AM07_09 Beni 3.2 9.5 3403 360 4 AM05_19 Amazona 3.2 9.5 24.8 314 4 AM06_19 Amazona 3.2 9.5 24.8 102 6 AM07_14 Madeira 3.2 9.5 20.7 401 570 6 AM07_14 Madrie de Dios 7.5 20.7 401 570 7 CAN09_47 Mackenzie 1.03 1.6 377 64 CAN09_40 Pael 1.14 2.0 348 57 CAN09_40 Pael 1.14 2.0 348 57 CAN09_40 Slave 1.14 2.0 348 57 CLAN09_40	TDS _{si} (mg/L) 10.4 7.8 4.3 8.7 11.5	Sample name Amazon Lowlands AM01-17 AM01-18 AM01-27 AM01-22 MES AM01-29 MES AM01-29 MES AM01-29	River Orthon Yata Trombetas Negro Urucara Tapajos	Cation Silicate y rate (t/km²/yr) 2.8 0.8	Total Silicate Weathering (t/km ² /yr) (t/km ² /yr) (t/km ² /yr) (t/km ² /yr) 5.5 5.7 5.5 5.7 5.5	Sediment concentrat ion (mg/L) 123 30 30 30 30 12 30 30 15 8 8	Erc 10 10 7	e (E) n²/yr)	sion Denudation ₂ (E) rate (D) <u>m²/yr)</u> (<i>t/km²/yr</i>) 64 14 12 20 15 17	sion Denudation rate (D) Weathering Intensity (W/D) 64 0.137 14 0.259 12 0.504 20 0.291 15 0.367 17 0.581			$ \begin{array}{c c c c c c c c c c c c c c c c c c c $
4:3 PK 11105 Congo 5.3 25 8.1 azon Andes Alto Beni 20.5 5.3 25 8.1 azon Andes Alto Beni 20.5 29.5 34.03 3800 91_08 Beni 20.5 29.5 34.03 3800 97_09 Beni 20.5 29.5 34.03 3800 97_09 Madeira 3.2 9.5 28.3 16.9 32.34 31.40 98_36 Pastaza 3.2 9.5 28.4 16.9 32.34 31.40 98_36 Pastaza 31.7 54.3 16.9 52.7 570 98_36 Madre de Dios 7.5 20.7 401 570 98_36 Stard 1.37 2.2 14.4 382 109_0247 Madre de Dios 7.5 20.7 401 570 103_2 Peel 1.37 2.7 401 570 109_03 Slave 1.4 2	Con	01-29 1 go	Tapajos		10.0	8	7		17	17 0.581	17 0.581 0.28 ± 0.14	17 0.581 0.28 ± 0.14 1.37 ± 1.11	17 0.581 0.28 ± 0.14 1.37 ± 1.11 -7.37 ± 1.50
MM07_08 Alto Beni 20.5 29.5 34.03 3800 MM07_09 Beni 8.9 16.9 32.34 31.40 3 MM07_09 Beni 3.2 9.5 24.6 18.4 31.40 3 MM07_09 Beni 3.2 9.5 24.8 18.4 31.40 3 MM05_19 Amazonas 3.2 9.5 24.8 18.4 18.4 31.40 3 MM06_14 Canazonas 31.7 54.3 12.4 18.8 2 34.01 57.0 2 MM06_14 Pastaza 31.7 54.3 12.4 18.8 2 34.01 57.0 2 SAM09_04 Diard 1.19 2.4 59.5 16.7 1 SAM09_14 Middle Channel 1.7 2.7 961 29.5 2 2 3 3 3 3 3 3 3 3 3 3 3 3 3 <		089-63 PK 1105 089-65 PK 1105 4 mazon Andes	Congo Congo		5.3 3	25 25	8.1		ယယ	3 0.396 0.396	3 0.396 0.35 ± 0.10 3 0.396 0.48 ± 0.11	3 0.396 0.35 ± 0.10 3 0.396 0.48 ± 0.11 1.00 ± 1.11	3 0.396 0.35 ± 0.10 3 0.396 0.48 ± 0.11 1.00 ± 1.11 -6.05 ± 1.20
AM07_04 Beni 8.9 16.9 3234 3140 3 AM05_19 Madeira 3.2 9.5 980 1020 1 AM06_13 Ucayali 8.5 16.9 3234 3140 3 AM06_13 Ucayali 8.5 16.9 52.7 570 5 AM07_14 Madre de Dios 7.5 20.7 401 570 5 AM07_14 Madre de Dios 7.5 2.2 14.54 392 3 CAN09_04 Liard 1.19 2.4 595 167 1 CAN09_02 Mackenzie 1.03 1.6 377 64 6 CAN09_28 Slave 1.14 2.0 348 57 5 CAN09_28 Slave 1.14 2.0 348 57 5 CAN09_28 Slave 1.7 5.4 126 289 2 CJ1 Liard Changilang 1.7 121		AM01_08	Alto Beni	20.5	29.5	3403	3800	ω	829	829 0.008	829 0.008 1.00 ± 0.10	829 0.008 1.00 ± 0.10 -0.50 ± 1.00	829 0.008 1.00 ± 0.10 -0.50 ± 1.00 -1.26 ± 1.00
AM05_19Madeira3.29.52.481841AM06_65Amazonas14.214.21862.312AM06_13Ucayali8.516.95.275.705AM06_36Pasiaza31.754.31241862.312AM06_36Pasiaza31.754.31241862.312AM06_36Pasiaza31.754.31241862.312AM06_36Pasiaza31.754.31241882AM07_14Madre de Dios7.520.74015705Mackenzie1.372.214543923CAN09_47Mackenzie1.031.63.77646CAN09_40Peel1.732.79612.955CAN09_40Peel1.742.712111751CAN09_40Pael1.742.712111751CAN09_40Peel1.732.75382892CJ1Uinshajiang1.42.712111751CJ4Uinshajiang1.75.43.22712CJ2Changjiang3.38.05722.142CJ3Changjiang1.710.61531221CJ3Changjiang1.710.61531221CJ3Changjiang1.710.61531221 <tr< td=""><td></td><td>AM07_04 AM07_09</td><td>Beni Beni</td><td>5.8 9</td><td>16.9 15.5</td><td>3234 980</td><td>3140 1020</td><td> ω</td><td>157 036</td><td>0.005 0.015 0.015</td><td>157 0.005 1.00 ± 0.10 036 0.015 1.00 ± 0.10</td><td>$\begin{array}{rrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrr$</td><td>157 0.005 1.00 ± 0.10 -0.50 ± 1.00 -2.59 ± 1.12 036 0.015 1.00 ± 0.10 -0.50 ± 1.00 -2.38 ± 1.12</td></tr<>		AM07_04 AM07_09	Beni Beni	5.8 9	16.9 15.5	3234 980	3140 1020	ω	157 036	0.005 0.015 0.015	157 0.005 1.00 ± 0.10 036 0.015 1.00 ± 0.10	$\begin{array}{rrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrr$	157 0.005 1.00 ± 0.10 -0.50 ± 1.00 -2.59 ± 1.12 036 0.015 1.00 ± 0.10 -0.50 ± 1.00 -2.38 ± 1.12
AM06_65 Amazonas 14.2 186 221 AM06_13 Ucayali 8.5 16.9 527 57.0 AM06_14 Solimoes 11.7 54.3 124 188 221 AM06_14 Solimoes 31.7 54.3 124 188 221 AM06_14 Madre de Dios 7.5 20.7 401 570 5 AM06_26 Red Arctic 1.37 2.2 1454 392 3 CAN09_47 Mackenzie 1.03 1.6 377 64 6 CAN09_40 Peel 1.73 2.7 961 295 26 CJ1 Jinshajiang 1.4 2.7 1211 175 1 CJ4 Jinshajiang 1.7 5.4 8.9 28 2 CJ3 Changjiang 1.7 5.4 1226 48 289 2 CJ3 Changjiang 1.7 1.6 1226 249 2<		AM05_19	Madeira	3.2	9.5	248	184		94	94 0.049	94 0.049 0.85 ± 0.12	94 0.049 0.85 ± 0.12 -0.30 ± 1.00	94 0.049 0.85 ± 0.12 -0.30 ± 1.00 -3.08 ± 1.10
AM06_14 Solimose Compose <		AM06_65	Amazonas	о л	14.2 16.9	186 527	231 570	ט ת	45 87	45 0.058 87 0.029	45 0.058 0.76 ± 0.12 87 0.029 0.70 ± 0.08	45 0.058 0.76 ± 0.12 -0.15 ± 1.00	45 0.058 0.76 ± 0.12 -0.15 ± 1.00 -3.43 ± 1.09 87 0.029 0.70 ± 0.08 -0.10 ± 0.89 -1.80 ± 1.05
AM06_36 Pastaza 31.7 54.3 Mackenzie Madre de Dios 7.5 20.7 401 570 5 Mackenzie Red Arctic 1.37 2.2 1454 382 3 CAN09_0114 Mackenzie 1.37 2.2 1454 382 3 CAN09_0132 Middle Channel 1.9 2.4 585 167 1 CAN09_240 Peel 1.73 2.7 961 295 2 2 377 64 6 CAN09_240 Peel 1.73 2.7 961 295 2 2 2 377 64 6 CAN09_240 Peel 1.73 2.7 1211 175 1 CJ1 Jinshajiang 1.4 2.7 1211 175 1 CJ4 Changjiang 1.7 5.8 8.0 5.7 2.0 2.0 2.0 2.0 2.0 2.0 2.0 2.0 2.0		AM06_14	Solimoes		19.8	124	188	N	07	07 0.096	07 0.096 0.80 ± 0.12	07 0.096 0.80 ± 0.12 -0.24 ± 0.96	07 0.096 0.80 ± 0.12 -0.24 ± 0.96 -2.36 ± 1.10
AM07_14 Madre de Dios 7.5 20.7 401 570 Mackenzie Red Arctic 1.37 2.2 1454 392 392 CAN09_51 Red Arctic 1.19 2.4 595 167 1 CAN09_47 Mackenzie 1.03 1.6 377 64 62 CAN09_47 Mackenzie 1.03 1.6 377 64 62 CAN09_40 Peel 1.73 2.7 961 295 62 64 66 CAN09_28 Slave 1.14 2.0 348 57 5 5 CJ1 Jinshajiang 1.4 2.7 1211 175 1 CJ4 Changjiang 1.8 5.7 740 288 28 CJ32 Changjiang 1.7 1.8 5.7 740 29 2 CJ42 Yalongjiang 1.7 1.6 152 23 23 2 2 CJ3		AM08_36	Pastaza	31.7	54.3						0.32 ± 0.10	0.32 ± 0.10 1.10 ± 0.93	0.32 ± 0.10 1.10 ± 0.93 -1.28 ± 1.14
Mackenzie Red Arctic 1.37 2.2 1454 392 CAN09_51 Liard 1.19 2.4 595 167 1 CAN09_47 Mackenzie 1.03 1.6 377 64 6 CAN09_47 Middle Channel 1.73 2.7 961 295 2 CAN09_40 Peel 1.74 2.0 348 57 5 CAN09_28 Slave 1.14 2.0 348 57 5 CAN09_28 Jinshajiang 1.4 2.7 1211 175 1 CAN09_28 Slave 1.73 5.4 2.7 1211 175 1 CJ1 Jinshajiang 1.8 5.7 740 289 2 2 1 2 1 2 1 2 1 2 1 2 1 2 1 2 1 2 1 2 1 2 1 2 2 1		AM07_14	Madre de Dios	7.5	20.7	401	570	сл	91	0.035	91 0.035 0.82 ± 0.10	91 0.035 0.82 ± 0.10	91 0.035 0.82 ± 0.10
CANU9_51 Hed Arctic 1.37 2.2 1454 392 CANU9_47 Mackenzie 1.03 1.6 377 64 66 CANU9_47 Mackenzie 1.03 1.6 377 64 66 CANU9_40 Peel 1.73 2.7 961 295 67 6 CANU9_28 Slave 1.14 2.0 348 57 5 CANU9_28 Slave 1.14 2.0 348 57 5 CJ1 Jinshajiang 1.4 2.7 1211 175 1 CJ4 Changjiang 1.8 5.7 740 289 2 CJ2 Changjiang 3.1 7.8 205 93 1 CJ3 CJ3 Kangjiang 1.7 16.6 155 21 2 CJ3 CJ3 Kangjiang 1.7 1.6 155 23 2 2 CJ3 Kangjiang 1.7		Mackenzie			1								
CANU9_14 Liar 2.4 955 167 CANU9_24 Mackenzie 1.19 2.4 955 167 CANU9_24 Mackenzie 1.03 1.6 377 64 6 CANU9_240 Peel 1.73 2.7 961 295 2 CANU9_28 Slave 1.14 2.0 348 57 5 CJ1 Jinshajiang 1.4 2.7 1211 175 1 CJ4 Jinshajiang 1.4 2.7 1211 175 1 CJ4 Jinshajiang 1.4 2.7 740 296 2 CJ4 Changjiang 1.7 5.4 1226 488 4 CJ52 Changjiang 3.1 7.8 205 93 1 CJ36 Changjiang 1.7 10.6 153 122 1 CJ37 Ganges-Brahmaputra Ganjiang 3.2 18.4 169 138 1 <td></td> <td>CAN09_51</td> <td>Red Arctic</td> <td>1.37</td> <td>2.2</td> <td>1454</td> <td>392</td> <td>ω</td> <td>95</td> <td>0.005</td> <td></td> <td>0.005 0.95 ± 0.10 -0.47 ± 0.91</td> <td>$95 0.005 0.95 \pm 0.10 -0.47 \pm 0.91 -1.94 \pm 1.10$</td>		CAN09_51	Red Arctic	1.37	2.2	1454	392	ω	95	0.005		0.005 0.95 ± 0.10 -0.47 ± 0.91	$95 0.005 0.95 \pm 0.10 -0.47 \pm 0.91 -1.94 \pm 1.10$
CANNOS_47 Middle Channel 1.03 1.6 377 64 CANNOS_48 Slave 1.73 2.7 961 295 2 CANOS_28 Slave 1.14 2.0 348 57 5 CANOS_28 Slave 1.14 2.0 348 57 5 CJ1 Jinshajiang 1.4 2.7 1211 175 1 CJ4 Jinshajiang 1.4 2.7 1211 175 1 CJ4 Jinshajiang 1.7 5.4 1226 488 4 CJ4 Dinshajiang 1.7 5.4 1226 488 4 CJ52 Changjiang 3.3 8.0 572 271 2 CJ38 Changjiang 7.1 11.8 525 363 3 CJ3 Ganges-Brahmaputra Gangiang 1.7 10.6 153 122 1 CJ34 Ganga 3.2 18.4 169		CAN09_04	Liard	1.19	2.4	595	167	<u> </u>	70	70 0.014	70 0.014 0.75 ± 0.10	70 0.014 0.75 ± 0.10 -0.04 ± 0.91	70 0.014 0.75 \pm 0.10 -0.04 \pm 0.91 -1.62 \pm 1.08
CAN09_40 Peel 1.73 2.7 961 295 CAN09_28 Slave 1.14 2.0 348 57 5 CAN09_28 Slave 1.14 2.0 348 57 5 CAN09_28 Jinshajiang 1.4 2.7 1211 175 1 CJ1 Jinshajiang 1.4 2.7 1211 175 1 CJ4 Jinshajiang 1.4 2.7 5.4 1226 488 2 CJ4 Linshajiang 1.7 5.4 1226 289 2 CJ52 Changjiang 1.7 5.4 1226 289 2 CJ52 Changjiang 7.1 11.8 5.7 740 296 3 CJ33 Xiangjiang 7.1 11.9 552 358 3 Ganges-Brahmaputra Ganjang 3.2 18.4 169 138 1 IND99-19 MES Ganga 2.7 10		CAN09_47 CAN10_32	Mackenzie Middle Channel	1.03	1.6	377	64	6	5	0.025	55 0.025 0.80 ± 0.10 0.70 ± 0.10	35 0.025 0.80 ± 0.10 -0.20 ± 0.91 0.70 ± 0.10 -0.04 ± 0.91	35 0.025 0.80 ± 0.10 -0.20 ± 0.91 -1.72 ± 1.08 0.70 ± 0.10 -0.04 ± 0.91 -1.65 ± 1.05
CAN09_28 Slave 1.14 2.0 348 57 Changjiang Jinshajiang 1.4 2.7 1211 175 CJ1 Jinshajiang 1.4 2.7 1211 175 CJ4 Jinshajiang 1.4 5.4 1226 488 29 CJ4 Changjiang 1.7 5.4 1226 289 271 2 CJ52 Changjiang 3.3 8.0 5.7 245 295 2 CJ38 Changjiang 7.1 11.9 552 358 2 CJ34 Minjang 1.7 106 153 122 2 CJ34 Gangas-Brahmaputra Ganga 1.1 1128 530 128 BR8210<		CAN09_40	Peel	1.73	2.7	961	295		297	297 0.009	297 0.009 1.00 ± 0.10	297 0.009 1.00 ± 0.10 -0.50 ± 0.91	297 0.009 1.00 ± 0.10 -0.50 ± 0.91 -1.56 ± 1.10
Changjiang Jinshajiang 1.4 2.7 1211 175 CJ4 Jinshajiang 1.4 2.7 1211 175 CJ4 Jinshajiang 1.7 5.4 1226 488 CJ44 Changjiang 1.8 5.7 740 296 CJ40 Changjiang 1.8 5.7 740 296 CJ60 Changjiang 3.3 8.0 572 271 CJ22 Changjiang 3.1 7.8 205 93 CJ36 Changjiang 1.7 10.6 153 122 CJ37 CJ38 Xiangjiang 1.7 10.6 153 122 CJ34 Ganga 1.7 10.6 153 122 122 CJ34 Ganga 1.7 11.9 530 138 122 Ganges-Brahmaputra Ganga 1 1129 530 138 14 1129 530 530 530 541 14 <td></td> <td>CAN09_28</td> <td>Slave</td> <td>1.14</td> <td>2.0</td> <td>348</td> <td>57</td> <td></td> <td>59</td> <td>59 0.033</td> <td>59 0.033 0.65 ± 0.10</td> <td>59 0.033 0.65 ± 0.10 0.15 ± 0.91</td> <td>59 0.033 0.65 ± 0.10 0.15 ± 0.91 -1.45 ± 1.07</td>		CAN09_28	Slave	1.14	2.0	348	57		59	59 0.033	59 0.033 0.65 ± 0.10	59 0.033 0.65 ± 0.10 0.15 ± 0.91	59 0.033 0.65 ± 0.10 0.15 ± 0.91 -1.45 ± 1.07
5 CJ4 Jinshajiang 4.8 9.9 538 289 4 CJ14 Changjiang 1.7 5.4 1226 488 4 2 CJ60 Changjiang 1.7 5.4 1226 488 4 2 CJ60 Changjiang 1.8 5.7 740 296 2 4 CJ22 Changjiang 3.3 8.0 5.72 2.71 2 4 CJ36 Changjiang 3.1 7.8 205 93 2 5 CJ33 Xiangjiang 1.7 10.6 153 122 3 6 CJ34 Ganges-Brahmaputra 14 169 138 122 3 4 CJ34 Ganga 11 1129 530 122 3 8 IND99-19 MES Ganga 14 1129 530 12 8 IND99-28 MES Stein 10 921 810 8 8 Brahmaputra 10 921 810 8 <	<i></i>	Changjiang CJ1	Jinshaliang	1.4	2.7	1211	175		178	0.015	178 0.015 0.50 ± 0.15	178 0.015 0.50 ± 0.15 0.82 ± 1.13	178 0.015 0.50 ± 0.15 0.82 ± 1.13 -3.02 ± 1.26
1 CJ14 Changliang 1.7 5.4 1226 488 2 CJ60 Changliang 1.8 5.7 740 296 296 2 CJ22 Changliang 3.3 8.0 572 271 2 7 CJ36 Changliang 2.4 8.7 475 249 2 3 CJ2 Yalongliang 7.1 11.9 552 358 2 3 CJ35 Minjang 7.1 11.9 552 358 122 3 CJ34 Ganjiang 3.2 18.4 169 138 122 4 CJ34 Ganjiang 3.2 18.4 169 138 122 4 CJ34 Ganga 14 1129 530 5 5 8 IND99-19 MES Brahmaputra 10 921 810 8 8 BR270 Brahmaputra 10 921 810 8 4 CAN99-28 MES Stikine 4.1 5.9 755 614	01	CJ4	Jinshajiang	4.8	9.9	538	289	D D	999	0.033	299 0.033 0.50 ± 0.15	299 0.033 0.50 ± 0.15 0.78 ± 1.16	299 0.033 0.50 ± 0.15 0.78 ± 1.16 -1.28 ± 1.20
9 CJS2 Changjiang 3.3 8.0 572 271 9 CJ36 Changjiang 3.3 8.0 572 271 7 CJ36 Changjiang 3.1 7.8 8.7 475 249 2 0 C.J2 Yaongjiang 3.1 7.8 205 93 - 3 C.J3 Xiangjiang 1.7 10.6 152 328 - 3 C.J34 Ganjang 3.2 18.4 169 138 - 4 C.J34 Ganjang 3.2 18.4 169 138 - 8 IND99-19 MES Ganga 14 11.29 530 6 8 IND99-28 MES Skeena 4.2 5.7 362 261 2 CAN99-39 MES Stikine 4.1 5.9 755 614 6	40	CJ14 C.160	Changjiang	1.7	5.4	1226 740	488 296	- N	194 302	194 0.011 302 0.019	194 0.011 0.58 ± 0.15 102 0.019 0.84 ± 0.15	$\begin{array}{rrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrrr$	494 0.011 0.58 ± 0.15 0.48 ± 1.03 -1.48 ± 1.20 302 0.019 0.84 ± 0.15 0.49 ± 1.02 -1.79 ± 1.13
7 CJ36 Changjiang 2.4 8.7 475 249 0 C.J2 Yalongjiang 3.1 7.8 205 93 3 C.J3 Minjiang 7.1 11.9 552 358 3 3 C.J3 Xiangjiang 1.7 10.6 153 122 4 C.J34 Ganjang 3.2 18.4 169 138 4 C.J34 Ganjang 3.2 18.4 169 138 6 IND99-19 MES Ganga 14 1128 530 10 8 IND99-28 MES Skeena 4.2 5.7 362 261 261 CAN99-39 MES Stikine 4.1 5.9 755 614 14	9	CJ52	Changjiang	3.3	8.0	572	271		279	279 0.029	279 0.029 0.68 ± 0.15	279 0.029 0.68 ± 0.15 -0.31 ± 0.99	279 0.029 0.68 ± 0.15 -0.31 ± 0.99 -1.29 ± 1.11
O C./2 Yalongliang 3.1 7.8 205 93 3 CuJS Minjiang 7.1 11.9 552 358 35 3 CuJS Xiangjiang 1.7 10.6 1552 358 36 4 CuJ4 Ganjaing 3.2 18.4 169 138 Ganges-Brahmaputra Ganga 3.2 18.4 169 138 8 IND99-19 MES Ganga 14 1128 530 362 8 BR292-19 MES Brahmaputra 10 921 810 93 8 BR292-28 MES Skeena 4.2 5.7 362 261 4 CAN99-39 MES Stikine 4.1 5.9 755 614 41	.7	CJ36	Changjiang	2.4	8.7	475	249		257	257 0.034	257 0.034 0.78 ± 0.15	257 0.034 0.78 ± 0.15 -0.18 ± 0.99	257 0.034 0.78 ± 0.15 -0.18 ± 0.99 -1.52 ± 1.10
3 Cu39 Minijarij 7.1 11.9 532 530 5	0.0	CJ2	Yalongjiang	3.1	7.8	205	93		101	101 0.077	101 0.077 0.38 ± 0.15	101 0.077 0.38 \pm 0.15 1.40 \pm 1.20	101 0.077 0.38 \pm 0.15 1.40 \pm 1.20 \pm 1.30 \pm 1.63
4 CJ34 Ganjiang 3.2 18.4 169 138 Ganges-Frahmaputra Ganga 14 1129 530 18 8 IND99-19 MES Ganga 14 1129 530 18 8 IND99-19 MES Ganga 14 1129 530 18 8 IND99-19 MES Brahmaputra 10 921 810 8 8 IND99-28 MES Skeena 4.2 5.7 362 261 2 2 CAN99-39 MES Stikine 4.1 5.9 755 614 0	ω	CJ33	Xiangjiang	1.7	10.6	153	122		133	0.080	133 0.080 0.92 ± 0.15	133 0.080 0.92 ± 0.15 -0.36 ± 0.93	133 0.080 0.92 ± 0.15 -0.36 ± 0.93 -3.04 ± 1.10
IND99-19 MES Ganga 14 1129 530	2.4	CJ34 Ganges-Brahmaputra	Ganjiang	3.2	18.4	169	138		157	0.117	157 0.117 1.00 ± 0.15	157 0.117 1.00 ± 0.15 -0.50 ± 1.00	157 0.117 1.00 ± 0.15 -0.50 ± 1.00 -4.20 ± 1.20
o DPDC/ID Distribution 10 92.1 010 Western cordiliera Skeena 4.2 5.7 362 261 CAN99-39 MES Stikine 4.1 5.9 755 614	9.8	IND99-19 MES	Ganga		14	1129	530		544	544 0.026 600	544 0.026 0.30 ± 0.10	544 0.026 0.30 ± 0.10 -0.40 ± 0.95	544 0.026 0.30 ± 0.10 -0.40 ± 0.95 -0.62 ± 1.00
CAN99-26 MES Stikine 4.1 5.9 755 614 6		Western cordillera	<u>.</u>	2	n J	2000	2021	۰ .	50	0000			
	10 0	CAN99-39 MES	Stikine	4.1 4	5.9	755	614	σr	20	20 0.009	20 0.009 0.10 ± 0.10 20 0.009 0.10 ± 0.10	20 0.009 0.10 ± 0.10 2.06 ± 1.20	20 0.009 0.10 ± 0.10 2.06 ± 1.20 -0.86 ± 1.00