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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 ( $\delta^7\text{Li}$ ) 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

23 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<sub>2</sub> 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 CO<sub>2</sub> 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 ( $\delta^7\text{Li}$ ) 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  $\delta^7\text{Li}$  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  $\delta^7\text{Li}$  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  $\delta^7\text{Li}$  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  $\delta^7\text{Li}$  is usually close to that of source  
54 rock (Bouchez et al., 2013), such that the source rock  $\delta^7\text{Li}$  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  $\delta^7\text{Li}$  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 km<sup>2</sup>, 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  
67 weathering regimes by using the weathering intensity parameter “(W/D)”, which is

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 ( $>100 \text{ t km}^{-2} \text{ yr}^{-1}$ ), 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  $102 \mu\text{g g}^{-1}$  (ppm) and the  $\delta^7\text{Li}$  of the sediments  
81 ( $\delta^7\text{Li}_{\text{suspended\_load}}$ , precision better than ‰) from  $-4\text{‰}$  to  $+5\text{‰}$  (Fig. 1). The rivers draining  
82 lowlands have on average lower  $\delta^7\text{Li}$  ( $-6.5\text{‰}$  to  $-3\text{‰}$ ) and Li concentration compared to  
83 rivers draining mountain ranges. Interestingly, the  $\delta^7\text{Li}$  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  $\delta^7\text{Li}$  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**

93 Fine sediments have generally lower  $\delta^7\text{Li}$  and higher Li concentration than coarse  
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  $\delta^7\text{Li}$  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  $\mu\text{m}$ ), 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  $\delta^7\text{Li}$  data.

## 105 **The Influence of Lithology**

106 The Li isotope signature that can be measured in weathered detrital material  
107 depends upon (1) the initial  $\delta^7\text{Li}$  of river sediments, prior to chemical weathering (i.e., the  
108 Li isotope signature of its source rock material, hereafter referred to as “ $\delta^7\text{Li}_{\text{source\_material}}$ ”),  
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,  $\Delta^7\text{Li}_{\text{fine-source}} = \delta^7\text{Li}_{\text{fine suspended\_load-}}$   
115  $\delta^7\text{Li}_{\text{source\_material}}$ .

116 The “ $\delta^7\text{Li}_{\text{source\_material}}$ ” corresponds to the weighted average of the  $\delta^7\text{Li}$  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- $\delta^7\text{Li}$ , shale end member ( $\delta^7\text{Li}$  of  $-0.5 \pm 1\%$  and Li/Al of  
125  $1.0 \pm 0.1 \times 10^{-3}$ ) and a Li-poor, high- $\delta^7\text{Li}$ , igneous end member ( $\delta^7\text{Li}$  of  $3.5 \pm 1.5\%$ ,  
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 ( $\delta^7\text{Li}_{\text{source\_material}}$ ) 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  $\delta^7\text{Li}_{\text{source\_material}}$  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<sup>1</sup>). 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  $\delta^7\text{Li}_{\text{source\_material}}$  estimates complete (see Data Repository).

160 Using the calculated proportion of shales and igneous rocks with the  $\delta^7\text{Li}$  of the  
161 rock end members, we estimate the Li isotope composition of the rock source  
162 ( $\delta^7\text{Li}_{\text{source\_material}}$ ) 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  $\delta^7\text{Li}_{\text{source\_material}}$  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  $\delta^7\text{Li}$ , 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  $(\text{Li}/\text{Al})_{\text{fine\_sediments}} /$   
171  $(\text{Li}/\text{Al})_{\text{source\_material}}$ , where  $(\text{Li}/\text{Al})_{\text{source\_material}}$  corresponds here to the  $(\text{Li}/\text{Al})$  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 “ $(\text{Li}/\text{Al})_{\text{source\_material}}$ ” 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 ( $\Delta^7\text{Li}_{\text{fine-source}}$ ) 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  $\Delta^7\text{Li}_{\text{fine-source}}$  whereas rivers  
191 having low W/D have  $\Delta^7\text{Li}_{\text{fine-source}}$  close to zero. The  $\delta^7\text{Li}_{\text{sed}}$  and  $\Delta^7\text{Li}_{\text{fine-source}}$  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  $\Delta^7\text{Li}_{\text{fine-}}$   
194  $\text{source}$  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  $\delta^7\text{Li}$  close to  $\delta^7\text{Li}_{\text{source\_material}}$  (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  $\delta^7\text{Li}$ , which could be expected to be  
200 reflected in the  $\delta^7\text{Li}$  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  $\delta^7\text{Li}_{\text{fine\_sediments}}$ , 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  $\delta^7\text{Li}$  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  $\delta^7\text{Li}$  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  $\Delta^7\text{Li}_{\text{fine-source}}$   
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  $\delta^7\text{Li}$  of clastic sediments can clearly discriminate between high weathering intensity

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  $\delta^7\text{Li}_{\text{fine\_sediments}}$  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  $\Delta^7\text{Li}_{\text{fine-source}}$   
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,  $\delta^7\text{Li}$  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  $(\text{Li}/\text{Al})_{\text{sediment\_load}} / (\text{Li}/\text{Al})_{\text{source\_material}}$  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  $(Li/Al)_{\text{sediment\_load}} / (Li/Al)_{\text{source\_material}}$  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 ( $\Delta^7Li_{\text{fine-source}}$ ). B: Li isotope composition of large river fine sediments  
371 ( $\delta^7Li_{\text{sediment\_load}}$ ), as a function of the weathering intensity (W/D).

372

373 <sup>1</sup>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](mailto:editing@geosociety.org).

377

Figure 1

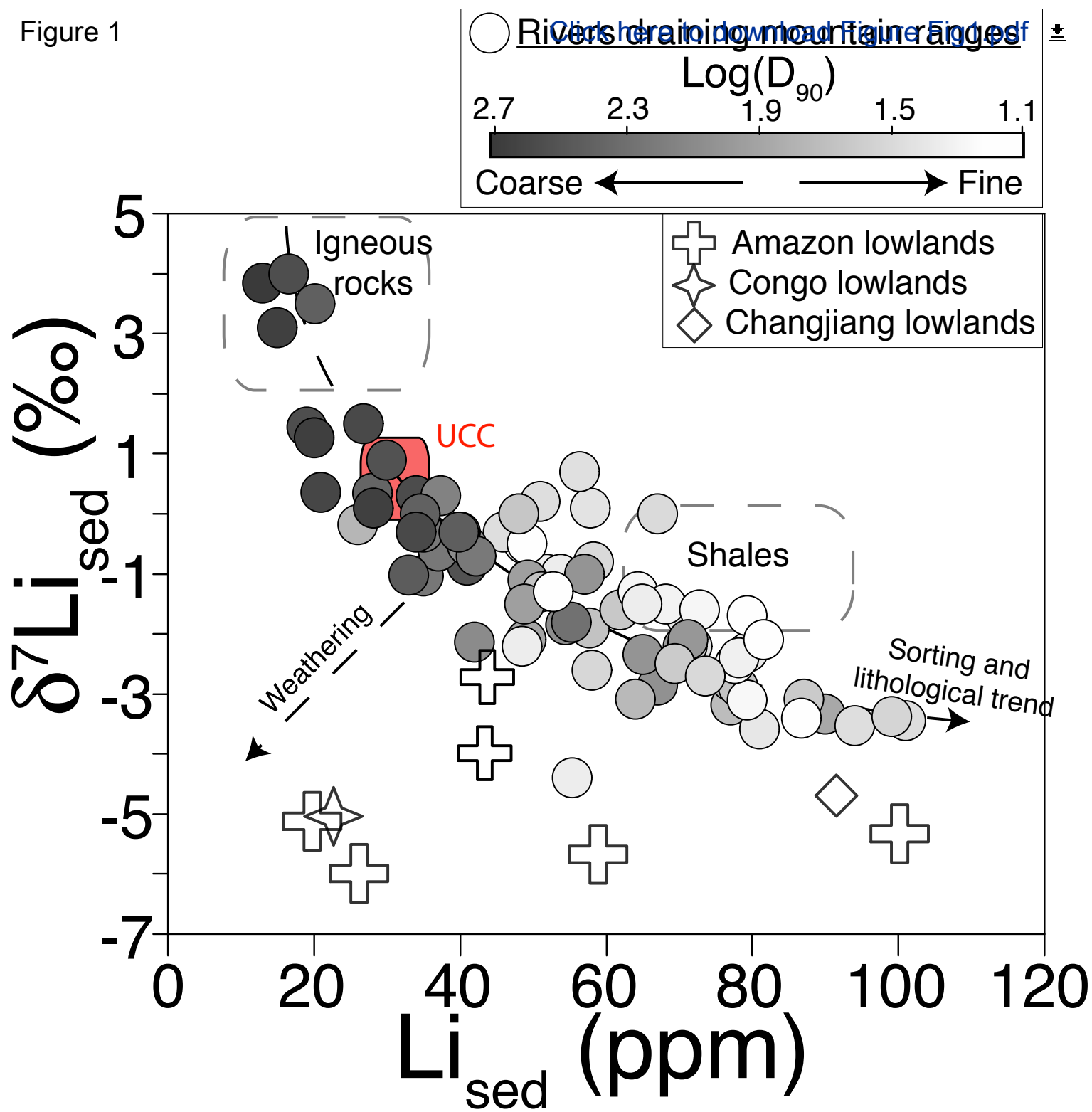


Figure 2

[Click here to download Figure Fig2.pdf](#)

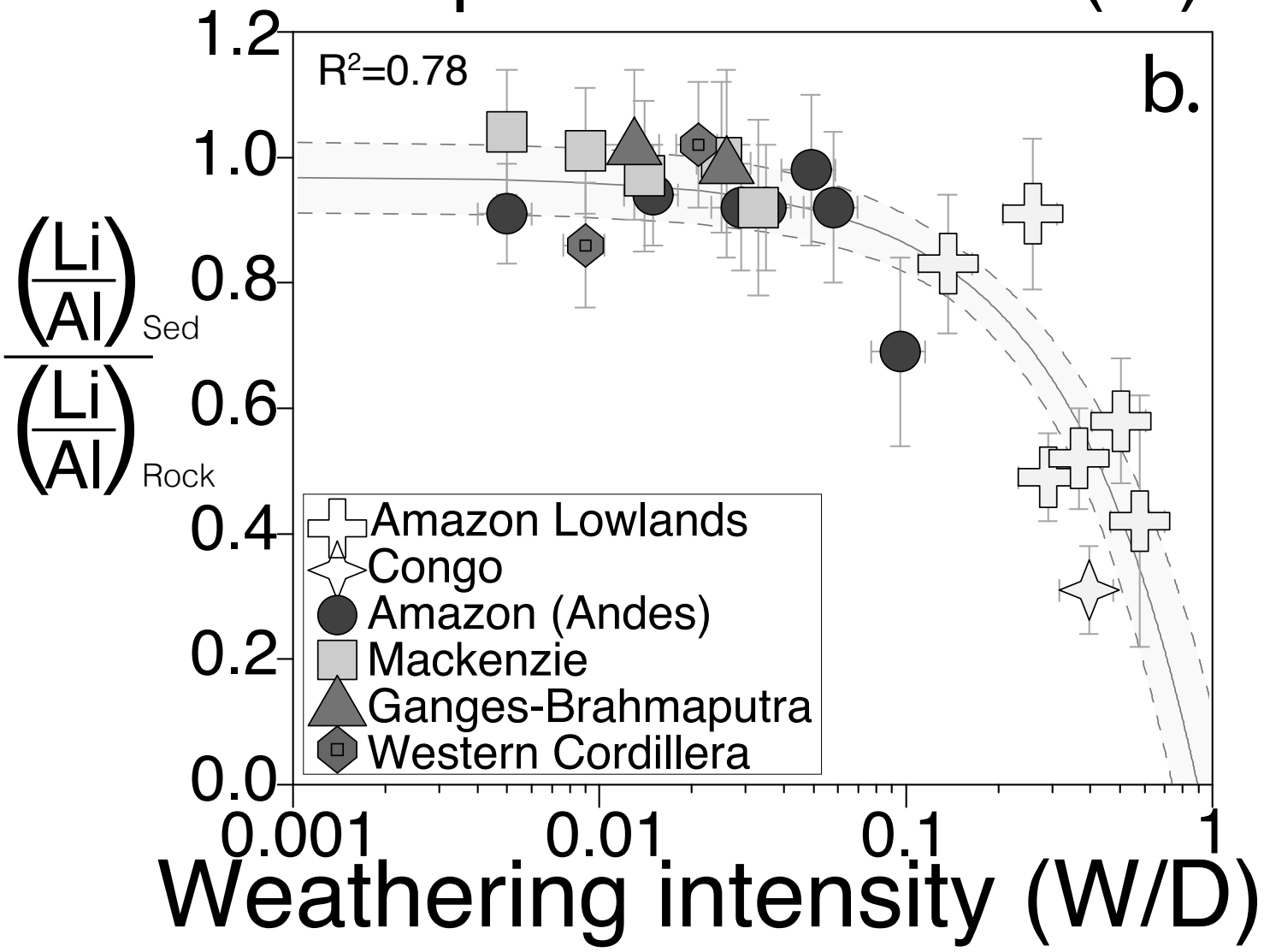
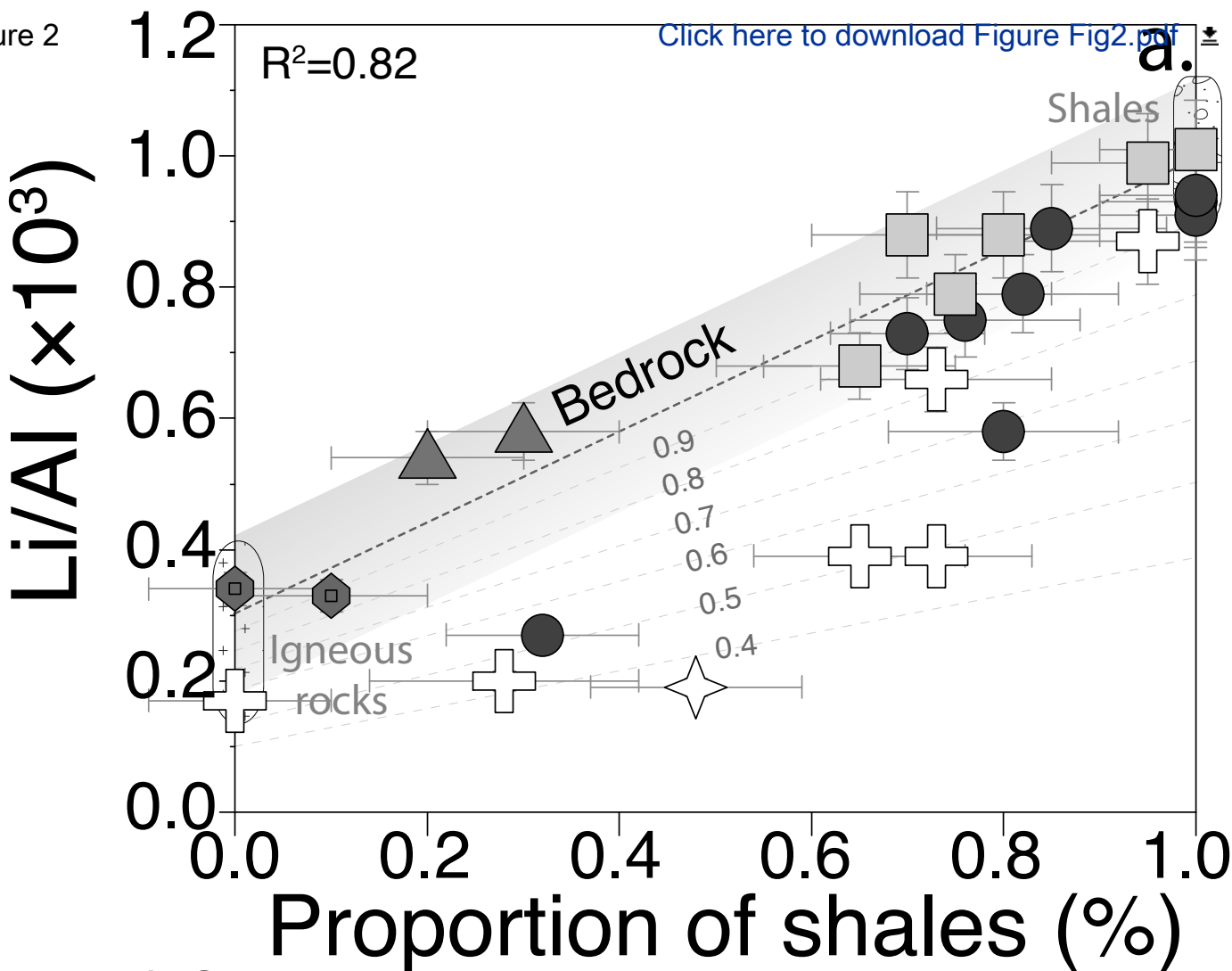
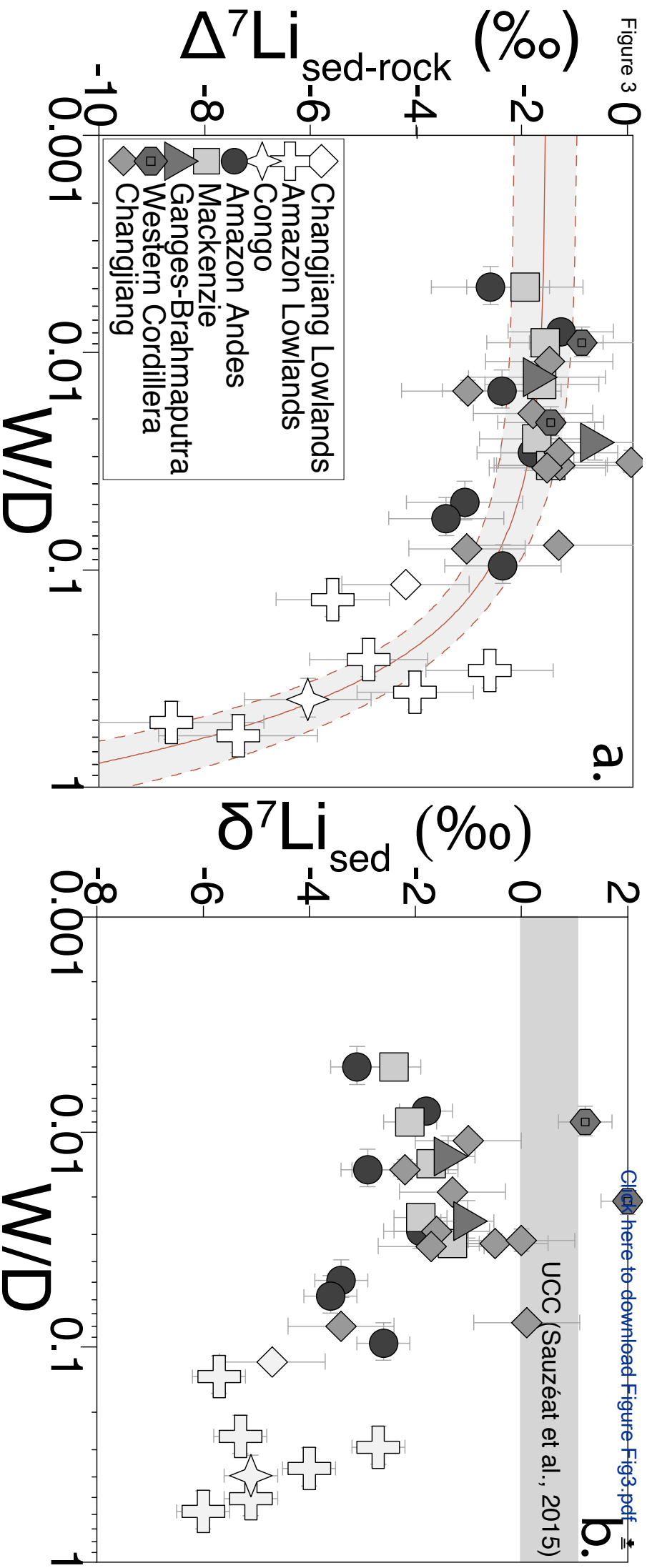


Figure 3



## Tracing weathering regimes using the Li isotope composition of detrital sediments

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### GSA Data Repository

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

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

Table DR1 and DR2

References

### 1. Analytical methods

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

Li isotopes: 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-HNO<sub>3</sub> 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  $\delta^7\text{Li} = +4.47 \pm 0.53$  ( $\pm 2\sigma$ ,  $n = 30$  separations and 15 digestions).

Nd isotopes: 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 Cogež et al., (2015).

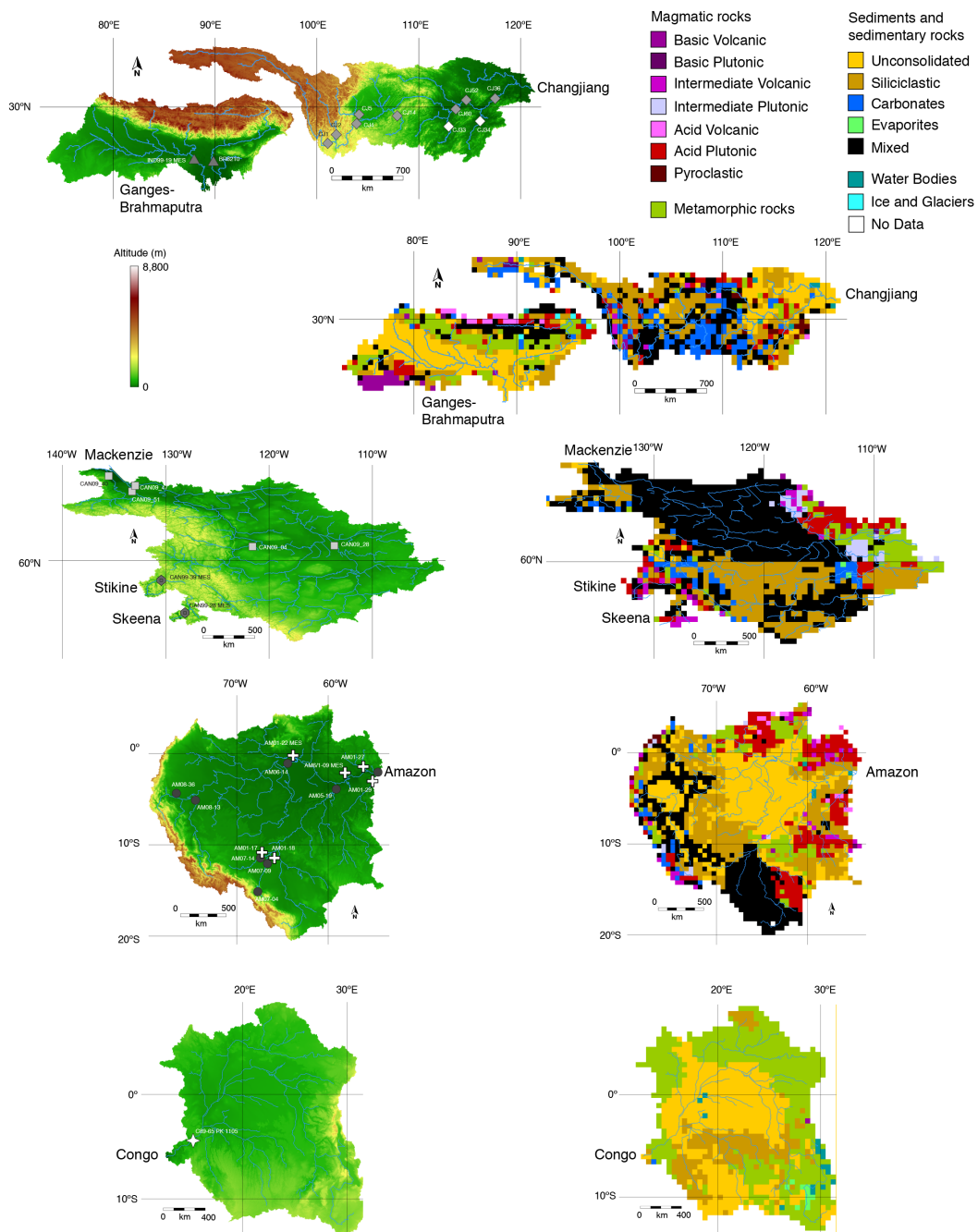


Figure DR1: Maps of the river basins (elevation on the left, geology on the right) from this study, with the locations of the samples. The Geological maps are from Hartmann and Moosdorf (2013). The Changjiang samples are from Wang et al., (2015)



## 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).

Amazon River: 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  $\epsilon_{Nd}(0)$ , the ratio in parts per 10<sup>4</sup> (units) between the measured  $^{143}\text{Nd}/^{144}\text{Nd}$  of the sample relative to the present-day value (0.512638) of the  $^{143}\text{Nd}/^{144}\text{Nd}$  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  $\epsilon_{Nd}$  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:

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

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

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

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  $\epsilon_{Nd}(0)$  of its sediments to estimate the average  $\epsilon_{Nd}(0)$  value of Andean sedimentary rocks. The  $\epsilon_{Nd}(0)$  of Beni sediments ranges between  $-12.2$  and  $-12.9$ , in good agreement with the range of  $\epsilon_{Nd}(0)$  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  $\epsilon_{Nd}(0)$  of  $-11.6$ , in good agreement with the  $\epsilon_{Nd}(0)$  value measured in Beni river sediments. Therefore, we assume a  $\epsilon_{Nd}(0)_{sha}$  of  $-12.0 \pm 1.0$  for the shale end member. The same value of  $-12.0$  was also assumed for the metasedimentary end-member 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  $\epsilon_{Nd}(0)$  of Solimões sediments defines a trend with the Al concentration (grain size proxy) that tends toward a similar shale  $\epsilon_{Nd}(0)$  value (Fig. DR2) and Neogene North Amazonian foreland basin sediments from Roddaz et al., (2005) also plot on the same trend.

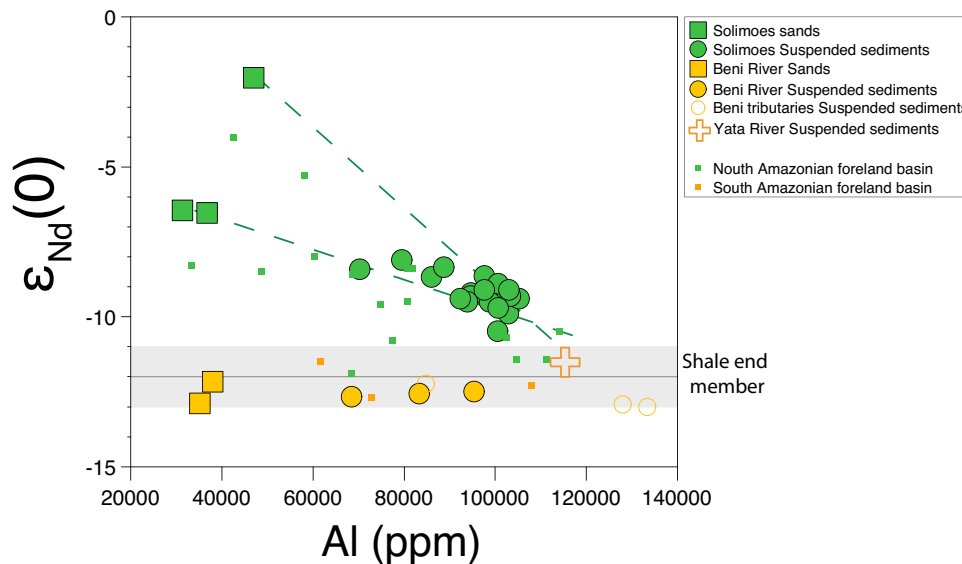
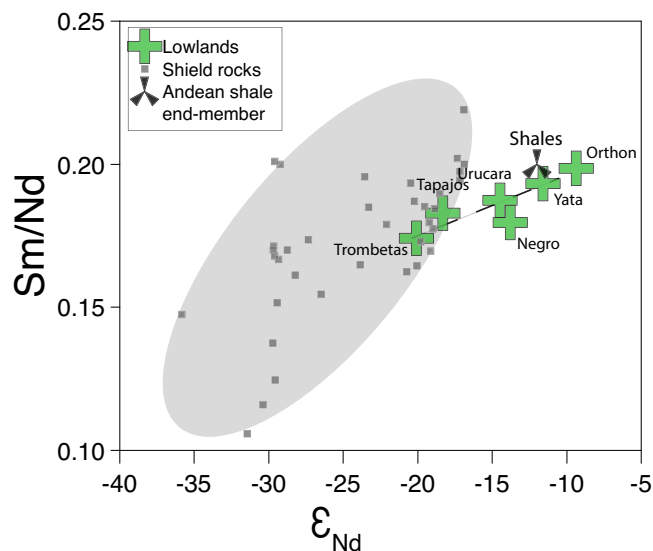


Figure DR2:  $\epsilon_{Nd}(0)$  as a function of the Al concentration in sediments. Solimões and Beni data are from Bouchez et al., (2011a), Dellinger et al., (2015) and Viers et al., (2008). Data for foreland basin sediments are from Roddaz et al., (2005). The grey band corresponds to the estimated  $\epsilon_{Nd}(0)$  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  $\epsilon_{Nd}(0)$  generally between  $-3$  to  $0$  (Petford and Atherton, 1996). Permian-Triassic and Carboniferous granitoids in southern Peru have a median  $\epsilon_{Nd}(0)$  of  $-4.7 \pm 0.6$  (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  $\epsilon_{Nd}(0)$  ranging from  $-6.5$  to  $-2.0$ . However, volcanic rocks in Ecuador (mostly andesites) have a higher  $\epsilon_{Nd}(0)$  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  $\epsilon_{Nd}(0)$  of  $+1.9$  (Dellinger et al., 2015a). This high  $\epsilon_{Nd}(0)$  indicates that Pastaza sediments predominantly derive from the erosion of more recent (Quaternary) andesite. In the following, we use a  $\epsilon_{Nd}(0)_{Ign}$  of  $-3.0 \pm 2.0$  for the igneous end member of all the Andean-fed rivers, except for the Pastaza River for which we use a  $\epsilon_{Nd}(0)_{Ign}$  value of  $+2.0 \pm 1.0$ .



*Fig. DR3: Sm/Nd as a function of the  $\epsilon_{Nd}$  of lowland river sediments. Shield rock data are from Santos et al., (2000). The Tapajós and Pastaza rivers have a  $\epsilon_{Nd}(0)$  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  $\epsilon_{Nd}(0)$  measured in rock samples from the Rio Negro, Central Amazon and Tapajós-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  $\epsilon_{Nd}(0)$  than cratonic rocks (Fig. DR3) and plot on a line trending towards the shale end member (defined as the mean Sm/Nd and  $\epsilon_{Nd}(0)$  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  $\epsilon_{Nd}(0)$  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 ( $\epsilon_{Nd}(0)_{Ign} = -20.0 \pm 2.0$ ).

Congo River: 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  $\epsilon_{Nd}(0)$  values for the major Congo tributaries. The  $\epsilon_{Nd}(0)$  range from  $-12.9$  for the Lobaye River to  $-18.2$  for the Likouala. We also measured the  $\epsilon_{Nd}(0)$  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.

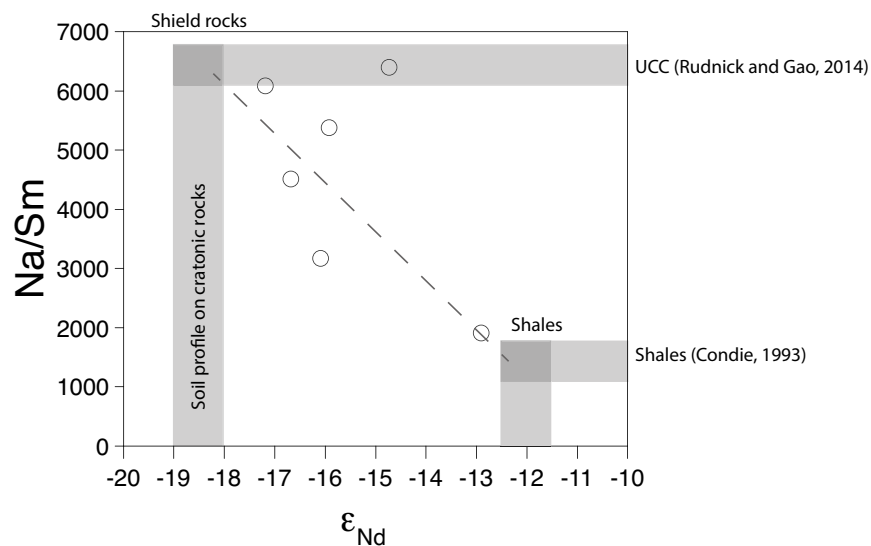


Fig. DR4: Crustal Na/Sm for Congo rivers inferred by Gaillardet et al. (1995) as a function of the sediment  $\epsilon_{Nd}$ . The Nd isotope composition of the cratonic end member is estimated to be  $-18.5 \pm 0.5$  based on the above trend, the Na/Sm of UCC (Rudnick and Gao, 2014) and the Nd

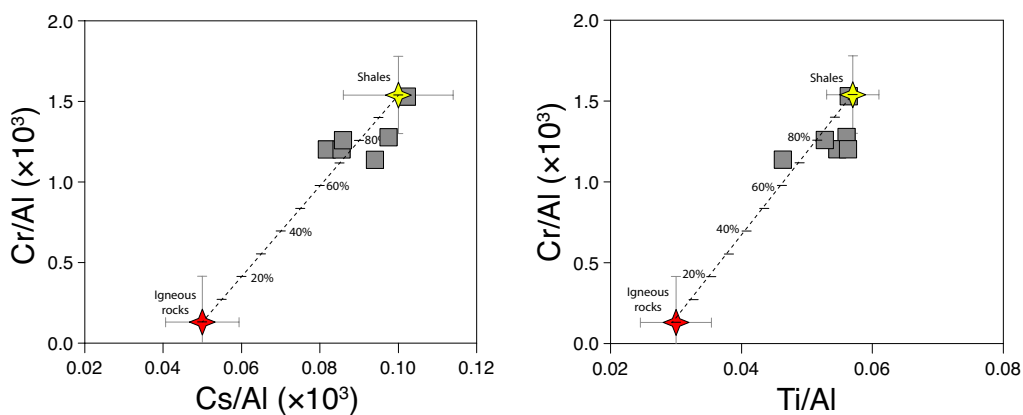
*isotope composition of a soil profile developed on cratonic rocks in Bangui (this study). The Nd isotope composition of the shales end-member is estimated to be  $-12 \pm 0.5$  by projecting the typical Na/Sm value of global Phanerozoic shales (from Condie, 1993) on the mixing trend. This value is also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

Interestingly, the  $\epsilon_{Nd}(0)$  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  $\epsilon_{Nd}(0)$  in Congo sediments can be explained at first order as resulting from a mixture between a TTG igneous end member of high  $\epsilon_{Nd}(0)_{Ign}$  around  $-19$  and a metasedimentary end member of low  $\epsilon_{Nd}(0)_{sha}$  (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  $40 \pm 10\%$ . We note that the correlation between  $\epsilon_{Nd}(0)$  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  $20\pm 10\%$ .

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), Garziona 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):  $Ti/Al = 0.030\pm 0.005$  ( $n=60$ ),  $Cr/Al = 0.13\pm 0.29\times 10^{-3}$  ( $n=23$ ) and median  $Cs/Al = 0.050\pm 0.005\times 10^{-3}$  ( $n=50$ ). 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).*

Western Cordillera: 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  $10\pm 10\%$ .

Changjiang: 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  $(\text{Li}/\text{Al})_{\text{rock}}$  values, we can use the proportion of shales calculated before ( $\gamma_{\text{sha}}$ ) 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  $-0.5 \pm 1\%$  and  $\text{Li}/\text{Al}$  of  $1.0 \pm 0.1 \times 10^{-3}$  and the igneous end member  $\delta^7\text{Li}$  of  $+3.5 \pm 1.5\%$ ,  $\text{Li}/\text{Al}$  of  $0.30 \pm 0.12 \times 10^{-3}$  (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  $\delta^7\text{Li}_{\text{ign}}$  end member of  $+5 \pm 2\%$  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 ( $\delta^7\text{Li}$  of  $0 \pm 1\%$ ,  $\text{Li}/\text{Al}$  of  $0.40 \pm 0.10 \times 10^{-3}$ ; 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  $\delta^7\text{Li}$  of  $0 \pm 1\%$  and a  $\text{Li}/\text{Al}$  ratio of  $0.40 \pm 0.10 \times 10^{-3}$  (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  $\text{Li}/\text{Al}$  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  $(\text{Li}/\text{Al})_{\text{sed}}/(\text{Li}/\text{Al})_{\text{rock}}$ , see main text) in the sediments. The proportion of Li riverine export occurring as dissolved ( $w^{\text{Li}}$  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.



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

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.

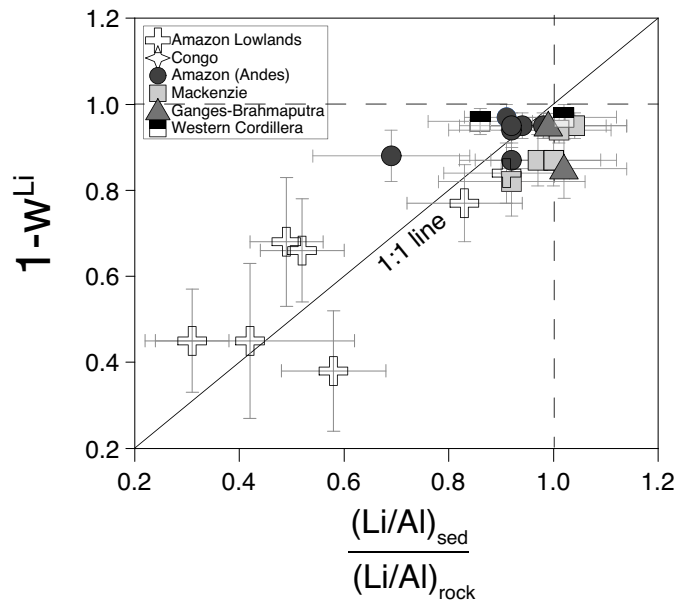


Figure DR6: Proportion of solid Li transported by the river ( $1-w^{Li}$ ) as a function of the Li weathering index calculated as  $(Li/Al)_{sed} / (Li/Al)_{rock}$ . The 1:1 line corresponds to the line for which  $(1-w^{Li})$  is equal to  $(Li/Al)_{sed} / (Li/Al)_{rock}$ , meaning that the fraction 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 ( $< 63 \mu m$ ) are relatively homogeneously distributed in the water

column while coarse sediments (size > 63  $\mu\text{m}$ ) 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  $(1-w^{\text{Li}})$  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  $w^{\text{Li}}$  values). In addition, we note that in lowland rivers suspended sediments collected at any depth contain a negligible amount of coarse (> 63 $\mu\text{m}$ ) particles (Meade, 1985). Therefore, the conclusions about the agreement between  $(1-w^{\text{Li}})$  values and  $(\text{Li}/\text{Al})_{\text{sed}} / (\text{Li}/\text{Al})_{\text{rock}}$  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	Discharge	D <sub>90</sub>	Al	Li	<sup>13</sup> Nd/ <sup>14</sup> Nd	ε <sub>Nd</sub> (0)	δ <sup>7</sup> Li	Li/AI (×10 <sup>-3</sup> )	CIA	[Li] <sub>diss</sub>	TDS <sub>sil</sub>
							(km <sup>2</sup> )	(m <sup>3</sup> /s)	(μm)	(μg/g)	(μg/g)		(‰)	(‰)	(g/g)	(mmol/L)	(mg/L)	(mg/L)
<b>Amazon Lowlands</b>																		
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	
<b>Congo</b>																		
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	
<b>Amazon Andes</b>																		
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°49'30.1" 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°30'18.8" W	01°56'05.3" S	Dellinger et al., (2014, 2015)	4618750	182200	27	107370	81	0.512141	-8.70	-3.6	0.75	78	0.14	11.4
AM08_13	Amazon	Ucayali	Janaro Herrera	79°40'28" W	04°54'28" S	Dellinger et al., (2014, 2015)	352593	12690	50	85684	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
<b>Mackenzie</b>																		
CAN09_51	Mackenzie	Red Arctic	Tsigiehtic	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°11'7.2" 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	Tsigiehtic	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°44'49.8" 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
<b>Changjiang</b>																		
CJ1	Changjiang	Jinshajiang	Panzhuhua	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	Chongqing	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	Panzhuhua	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
<b>Ganges-Brahmaputra</b>																		
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
<b>Western cordillera</b>																		
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]<sub>sil</sub>” 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]<sub>diss</sub> corresponds to the dissolved Li concentration, TDS<sub>sil</sub> 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	$\delta^{Li}_{rock}$ (‰)	$\Delta^{Li}_{bed-rock}$ (‰)	$1-w^{Li}$	$(Li/Al)_{sed} / (Li/Al)_{rock}$	References
<b>Amazon Lowlands</b>								
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)
<b>Congo</b>								
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
<b>Amazon Andes</b>								
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	0.95 ± 0.03	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	0.92 ± 0.12	Dellinger et al., (2014, 2015)
AM08_13	Ucayali	0.029	0.70 ± 0.08	-0.10 ± 0.89	-1.80 ± 1.05	0.87 ± 0.06	0.92 ± 0.10	Dellinger et al., (2014, 2015)
AM06_14	Solimoes	0.096	0.80 ± 0.12	-0.24 ± 0.96	-2.36 ± 1.10	0.88 ± 0.06	0.69 ± 0.15	Dellinger et al., (2014, 2015)
AM08_36	Pastaza	0.32	0.32 ± 0.10	1.10 ± 0.93	-1.28 ± 1.14			Dellinger et al., (2014, 2015)
AM07_14	Madre de Dios	0.035	0.82 ± 0.10			0.95 ± 0.03	0.92 ± 0.10	Dellinger et al., (2014, 2015)
<b>Mackenzie</b>								
CAN09_51	Red Arctic	0.003	0.95 ± 0.10	-0.47 ± 0.91	-1.94 ± 1.10	0.95 ± 0.03	1.04 ± 0.10	Dellinger et al., (2014)
CAN09_04	Liard	0.007	0.75 ± 0.10	-0.04 ± 0.91	-1.62 ± 1.08	0.87 ± 0.06	0.97 ± 0.12	Dellinger et al., (2014)
CAN09_47	Mackenzie	0.016	0.80 ± 0.10	-0.20 ± 0.91	-1.72 ± 1.08	0.87 ± 0.06	1.00 ± 0.12	Dellinger et al., (2014)
CAN10_32	Middle Channel		0.70 ± 0.10	-0.04 ± 0.91	-1.65 ± 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)
CAN09_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)
<b>Changjiang</b>								
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)
<b>Ganges-Brahmaputra</b>								
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)
<b>Western cordillera</b>								
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	$\epsilon_{Nd}$	$^{143}Nd / ^{144}Nd$
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  $\epsilon_{Nd}(0)$  are calculated by normalizing to the CHUR value (0.512638). For details about the samples, see Henchiri et al., (2016).

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Sample name	Basin	River	Location	Longitude	Latitude	References for Li isotope data	Surface (km <sup>2</sup> )	Discharge (m <sup>3</sup> /s)	D <sub>50</sub> (µm)	AI (µg/g)	LI (µg/g)	<sup>13</sup> Nd/ <sup>144</sup> Nd ε <sub>Li</sub> (‰)	δ <sup>7</sup> Li (‰)	L/Al (×10 <sup>-3</sup> )	Cl/Al [L] <sub>class</sub> (mmol/L)		
<b>Amazon Lowlands</b>																	
AM01-17	Amazon	Orthon	Fiberalta	66°02'1.0" W	10°48'36.9" S	Dellingner et al., (2014, 2015)	33485	475		89835	59	0.512158	-9.37	-5.7	0.66	82	0.31
AM01-18	Amazon	Yata	Mouth	65°35'00.2" W	10°58'44.8" S	Dellingner et al., (2014, 2015)	20770	222		115390	100	0.512043	-11.61	-5.3	0.87	86	0.08
AM01-27	Amazon	Trombetas	Mouth	N.D.	N.D.	Dellingner et al., (2014, 2015)	250000	6200		115312	20	0.511609	-20.08	-5.1	0.17	92	0.037
AM01-22 MIES	Amazon	Negro	Paricatuba	N.D.	N.D.	Dellingner et al., (2014, 2015)	691000	29000		113016	44	0.511930	-13.81	-2.7	0.39	87	0.031
AM6/1-09 MIES	Amazon	Urucara	Mouth	N.D.	N.D.	Dellingner et al., (2014, 2015)	150000	3000		110891	43	0.511895	-14.50	-4.0	0.39	87	0.049
AM01-29	Amazon	Tapajos	Mouth	N.D.	N.D.	Dellingner et al., (2014, 2015)	490000	13500		129350	27	0.511698	-18.33	-6.0	0.20	93	0.038
<b>Congo</b>																	
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		92	0.10
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
<b>Amazon Andes</b>																	
AM01_08	Amazon	Alto Beni	Mouth	67°40'18" W	15°07'24" S	Dellingner et al., (2014, 2015)	21197	750		84800	79	0.512011	-12.24	-1.8	0.93	74	1.62
AM07_04	Amazon	Beni	Furranabaque	67°32'05.2" W	14°28'26.8" S	Dellingner et al., (2014, 2015)	69980	2153	44	95289	87	0.511998	-12.49	-3.1	0.91	76	1.28
AM07_09	Amazon	Beni	Fiberalta	66°05'53.8" W	10°58'44.8" S	Dellingner et al., (2014, 2015)	114379	3772	50	83319	78	0.511994	-12.56	-2.9	0.94	76	0.57
AM05_19	Amazon	Madeira	Foz Madeira	58°48'30.1" W	03°27'00.9" S	Dellingner et al., (2014, 2015)	1325000	31250	36	111822	99	0.512097	-10.55	-3.4	0.89	78	0.18
AM06_65	Amazon	Amazonas	Obidos	55°30'18.6" W	01°56'05.3" S	Dellingner et al., (2014, 2015)	4618750	182200	27	107370	81	0.512141	-9.70	-3.6	0.75	78	0.14
AM08_13	Amazon	Ucayali	Jenaro Herrera	73°40'28" W	04°54'28" S	Dellingner et al., (2014, 2015)	352533	12090	50	85664	58	0.512158	-9.37	-1.9	0.73	68	0.66
AM06_14	Amazon	Sollinos	Manacapuru	60°33'11.0" W	03°19'41.0" S	Dellingner et al., (2014, 2015)	2148000	103000	35	100535	58	0.512101	-10.48	-2.6	0.58	76	0.14
AM08_36	Amazon	Pastaza	Embochura	N.D.	N.D.	Dellingner et al., (2014, 2015)	39000	2770	61	97031	26	0.512512	-2.45	-0.2	0.27	68	0.14
AM07_14	Amazon	Madre de Dios	Fiberalta	66°05'54" W	10°57'46" S	Dellingner et al., (2014, 2015)	124231	5602	35	94447	75	0.512106	-10.38		0.79	77	0.22
<b>Mackenzie</b>																	
CAN09_51	Mackenzie	Red Arctic	Tsigitchic	133°46'44.3" W	67°25'34.2" N	Dellingner et al., (2014)	18600	159	23	79022	78.1	0.512066	-11.16	-2.4	0.99	77	0.89
CAN09_04	Mackenzie	Liard	Fort Simpson	121°17'6.2" W	61°50'8.6" N	Dellingner et al., (2014)	275000	2450	49	64955	51.2	0.511904	-14.31	-1.7	0.79	73	0.63
CAN09_47	Mackenzie	Mackenzie	Tsigitchic	133°43'23.8" W	67°27'30.1" N	Dellingner et al., (2014)	1680000	8990		76477	67.5	0.512013	-12.19	-1.9	0.88	76	0.57
CAN10_32	Mackenzie	Middle Channel	Inuvik	134°44.9" W	68°24'33.3" N	Dellingner 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	Dellingner et al., (2014)	70800	686	18	80981	81.5	0.512062	-11.23	-2.1	1.01	77	0.71
CAN09_28	Mackenzie	Slave	Fort Smith	111°53'25.7" W	60°0'59.5" N	Dellingner et al., (2014)	6164000	3200	17	77069	52.8	0.512014	-12.18	-1.3	0.68	76	0.60
<b>Changliang</b>																	
CJ1	Changliang	Jinshatlang	Panzhuhua	101°37'55.1" E	26°34'50.3" N	Wang et al., (2015)	285556	1309	24	80546	48.5		-2.2	0.60	69	2.07	
CJ4	Changliang	Jinshatlang	Yibin	104°33'38.5" E	28°42'16.4" N	Wang et al., (2015)	485000	8239	22	82252	48.7		-0.5	0.59	73	0.97	
CJ14	Changliang	Changliang	Chongqing	N.D.	N.D.	Wang et al., (2015)	867000	10845	24	84970	53.8		-1.0	0.63	71	0.78	
CJ60	Changliang	Changliang	Yueyang	113°18'7.99" E	29°38'07.9" N	Wang et al., (2015)	1250000	15844		76.6			-1.3			0.37	
CJ62	Changliang	Changliang	Wuhan	114°29'8.76" E	30°40'7.47" N	Wang et al., (2015)	1705000	22340	45	83606	61.9		-1.6	0.74	72	0.44	
CJ36	Changliang	Changliang	Dalong	117°38'3.24" E	30°46'0.85" N	Wang et al., (2015)	1488000	28266	26	98071	70.9		-1.7	0.72	76	0.27	
CJ2	Changliang	Yalongliang	Panzhuhua	101°49'20.0" E	26°40'1.27" N	Wang et al., (2015)	128444	1857		161917	41.3		0.1	0.26	63	0.68	
CJ5	Changliang	Mingliang	Yibin	104°33'7.21" E	28°48'17.9" N	Wang et al., (2015)	133000	2732	33	83956	67.0		0.0	0.80	67	0.45	
CJ33	Changliang	Xiangliang	Changsha	112°56'7.95" E	28°07'3.64" N	Wang et al., (2015)	94700	2405	15	108398	86.7		-3.4	0.80	82	0.16	
CJ34	Changliang	Ganliang	Nanchang	115°52'0.53" E	28°40'2.06" N	Wang et al., (2015)	80900	2104		103092	88.6		-4.7	0.86	80	0.25	
<b>Ganges-Brahmaputra</b>																	
IND99-19 MIES	Ganges	Ganga	Harding bridge	89°1'28.739" E	24°31'0.44" N	Dellingner et al., (2014)	1050000	15622		94322	55.0	0.511725	-17.81	-1.02	0.58		0.49
BR8210	Brahmaputra	Brahmaputra	Jamuna bg.	N.D.	N.D.	Dellingner et al., (2014)	580000	16161		87063	47	0.511897	-14.45	-1.38	0.54	66	1.08
<b>Western cordillera</b>																	
CAN9-28 MIES	Sikine	Sikine	Kiwanga	N.D.	N.D.	This study	42200	962		83647	28.7		2.0	0.34	64	0.05	
CAN9-39 MIES	Sikine	Sikine	Iskut	N.D.	N.D.	This study	29300	755		75071	24.5		1.2	0.33	63	0.10	

TDS <sub>SI</sub>	Sample name	River	Cation silicate Weatherin g rate ( $\mu\text{km}^2/\text{yr}$ )	Total silicate Weatherin rate (W) ( $\mu\text{km}^2/\text{yr}$ )	Sediment concentrat ion ( $\text{mg/L}$ )	Erosion rate (E) ( $\mu\text{km}^2/\text{yr}$ )	Denudation rate (D) ( $\mu\text{km}^2/\text{yr}$ )	Weathering Intensity (W/D)	Proportion of shales	$\delta^{18}\text{O}_{\text{rock}}$ (‰)	$\Delta^{18}\text{O}_{\text{basalt/rock}}$ (‰)	$1-w^{\text{LI}}$	(L/A) <sup>bas</sup> / (L/A) <sup>rock</sup>
<b>Amazon Lowlands</b>													
19.6	AM01-17	Orthon	2.8	8.8	123	55	64	0.137	0.73 ± 0.12	-0.10 ± 0.89	-5.58 ± 1.07	0.77 ± 0.09	0.83 ± 0.11
10.4	AM01-18	Yata	0.8	3.5	30	10	14	0.259	0.95 ± 0.14	-0.36 ± 0.93	-4.90 ± 1.12	0.84 ± 0.07	0.91 ± 0.12
7.8	AM01-27	Trombetas		6.1	8	6	12	0.504	0.00 ± 0.10	3.50 ± 1.50	-8.63 ± 1.75	0.38 ± 0.14	0.58 ± 0.10
4.3	AM01-22 MIES	Negro		5.7	10	14	20	0.291	0.73 ± 0.10	-0.11 ± 0.90	-2.61 ± 1.20	0.68 ± 0.15	0.49 ± 0.07
8.7	AM01-09 MIES	Urucara		5.5	15	10	15	0.367	0.65 ± 0.11	0.04 ± 0.90	-4.02 ± 1.10	0.66 ± 0.12	0.52 ± 0.08
11.5	AM01-29	Tapajos		10.0	8	7	17	0.581	0.28 ± 0.14	1.37 ± 1.11	-7.37 ± 1.50	0.45 ± 0.18	0.42 ± 0.20
<b>Congo</b>													
15.0	C89-63 PK 1105	Congo		5.3	25	8.1	13	0.396	0.35 ± 0.10			0.51 ± 0.13	
15.0	C89-65 PK 1105	Congo		5.3	25	8.1	13	0.396	0.48 ± 0.11	1.00 ± 1.11	-6.05 ± 1.20	0.45 ± 0.12	0.31 ± 0.07
<b>Amazon Andes</b>													
26.4	AM01_08	Alto Beni	20.5	29.5	3403	3800	3829	0.008	1.00 ± 0.10	-0.50 ± 1.00	-1.26 ± 1.00		
17.4	AM07_04	Beni	8.9	16.9	3234	3140	3157	0.005	1.00 ± 0.10	-0.50 ± 1.00	-2.59 ± 1.12	0.97 ± 0.03	0.91 ± 0.08
14.9	AM07_09	Beni	5.8	15.5	980	1036	1036	0.015	1.00 ± 0.10	-0.50 ± 1.00	-2.38 ± 1.12	0.95 ± 0.03	0.94 ± 0.08
12.8	AM05_19	Madelira	3.2	9.5	248	184	194	0.049	0.65 ± 0.12	-0.30 ± 1.00	-3.08 ± 1.10	0.95 ± 0.03	0.98 ± 0.12
11.4	AM06_65	Amazonas		14.2	186	231	245	0.058	0.76 ± 0.12	-0.15 ± 1.00	-3.43 ± 1.09	0.94 ± 0.03	0.92 ± 0.12
15.6	AM08_13	Ucayali		16.9	527	570	587	0.029	0.70 ± 0.08	-0.10 ± 0.89	-1.80 ± 1.05	0.87 ± 0.06	0.92 ± 0.10
13.1	AM06_14	Sollinosa		19.8	124	188	207	0.096	0.80 ± 0.12	-0.24 ± 0.96	-2.36 ± 1.10	0.88 ± 0.06	0.69 ± 0.15
24.2	AM08_36	Pastaza	31.7	54.3					0.32 ± 0.10	1.10 ± 0.93	-1.28 ± 1.14		
14.6	AM07_14	Madre de Dios	7.5	20.7	401	570	591	0.035	0.82 ± 0.10			0.95 ± 0.03	0.92 ± 0.10
<b>Mackenzie</b>													
8.0	CAN09_51	Red Arctic	1.37	2.2	1454	392	395	0.005	0.95 ± 0.10	-0.47 ± 0.91	-1.94 ± 1.10	0.95 ± 0.03	1.04 ± 0.10
8.7	CAN09_04	Liard	1.19	2.4	595	167	170	0.014	0.75 ± 0.10	-0.04 ± 0.91	-1.62 ± 1.08	0.87 ± 0.06	0.97 ± 0.12
9.5	CAN09_47	Mackenzie	1.03	1.6	377	64	65	0.025	0.80 ± 0.10	-0.20 ± 0.91	-1.72 ± 1.08	0.87 ± 0.06	1.00 ± 0.12
	CAN10_32	Middle Channel							0.70 ± 0.10	-0.04 ± 0.91	-1.65 ± 1.05		1.10 ± 0.12
8.8	CAN09_40	Peel	1.73	2.7	961	295	297	0.009	1.00 ± 0.10	-0.50 ± 0.91	-1.56 ± 1.10	0.94 ± 0.03	1.01 ± 0.10
12.0	CAN09_28	Slave	1.14	2.0	348	57	59	0.033	0.65 ± 0.10	-0.15 ± 0.91	-1.45 ± 1.07	0.82 ± 0.08	0.92 ± 0.14
<b>Changjiang</b>													
18.6	CJ11	Jinshajiang	1.4	2.7	1211	175	178	0.015	0.50 ± 0.15	0.82 ± 1.13	-3.02 ± 1.26	0.80 ± 0.088	
18.5	CJ4	Jinshajiang	4.8	9.9	538	289	299	0.033	0.50 ± 0.15	0.78 ± 1.16	-1.28 ± 1.20	0.79 ± 0.088	
13.4	CJ14	Changjiang	1.7	5.4	1226	488	494	0.011	0.58 ± 0.15	0.48 ± 1.03	-1.48 ± 1.20	0.92 ± 0.04	
14.2	CJ60	Changjiang	1.8	5.7	740	296	302	0.019	0.84 ± 0.15	0.49 ± 1.02	-1.79 ± 1.13	0.96 ± 0.03	
16.9	CJ52	Changjiang	3.3	8.0	572	271	279	0.029	0.68 ± 0.15	-0.31 ± 0.99	-1.29 ± 1.11	0.92 ± 0.04	
16.7	CJ36	Changjiang	2.4	8.7	475	249	257	0.034	0.78 ± 0.15	-0.18 ± 0.99	-1.52 ± 1.10	0.95 ± 0.03	
17.0	CJ2	Yalongjiang	7.8	20.5	205	93	101	0.077	0.38 ± 0.15	1.40 ± 1.20	-1.30 ± 1.63	0.64 ± 0.117	
18.3	CJ5	Mingjiang	7.1	11.9	552	358	370	0.032	0.74 ± 0.15	-0.07 ± 0.97	0.07 ± 1.09	0.92 ± 0.04	
13.3	CJ33	Xiangjiang	1.7	10.6	153	122	133	0.080	0.92 ± 0.15	-0.36 ± 0.93	-3.04 ± 1.10	0.92 ± 0.04	
22.4	CJ34	Ganjiang	3.2	18.4	169	138	157	0.117	1.00 ± 0.15	-0.50 ± 1.00	-4.20 ± 1.20	0.90 ± 0.055	
<b>Ganges-Brahmaputra</b>													
29.8	IND99-19 MIES	Ganga		14	1129	530	544	0.026	0.30 ± 0.10	-0.40 ± 0.95	-0.62 ± 1.00	0.95 ± 0.029	0.99 ± 0.15
11.8	BR8210	Brahmaputra		10	921	810	820	0.013	0.20 ± 0.10	0.33 ± 1.10	-1.71 ± 1.30	0.85 ± 0.069	1.02 ± 0.12
<b>Western cordillera</b>													
7.9	CAN99-28 MIES	Skeena	4.2	5.7	362	261	266	0.021	0.00 ± 0.10	3.50 ± 1.10	-1.45 ± 1.00	0.97 ± 0.03	1.02 ± 0.10
7.2	CAN99-39 MIES	Silkine	4.1	5.9	755	614	620	0.009	0.10 ± 0.10	2.06 ± 1.20	-0.86 ± 1.00	0.96 ± 0.03	0.86 ± 0.10