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## Chemical erosion rates in the upper Blue Nile Basin and related atmospheric CO<sub>2</sub> consumption

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► **To cite this version:**

Luc Bastian, Nathalie Vigier, Marie Revel, Gezahegn Yirgu, Dereje Ayalew, et al.. Chemical erosion rates in the upper Blue Nile Basin and related atmospheric CO<sub>2</sub> consumption. *Chemical Geology*, 2019, 518, pp.19-31. 10.1016/j.chemgeo.2019.03.033 . hal-02134565

**HAL Id: hal-02134565**

**<https://hal.science/hal-02134565v1>**

Submitted on 22 Oct 2021

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20 **Abstract**

21

22 Silicate weathering of basaltic rocks constitutes a non-negligible sink of atmospheric CO<sub>2</sub> but  
23 the role it plays in the regulation of past and future global climate is still matter of debate. In  
24 this study, silicate weathering rates for various sub-basins of the Ethiopian Traps, emplaced  
25 30 million years ago, and the corresponding atmospheric CO<sub>2</sub> consumption rates are  
26 evaluated. For this, major and trace elements were measured in the dissolved phases and in  
27 the sedimentary particles carried and deposited by the main rivers flowing through this steep  
28 region. Lithium isotopes and major elements were also measured in the extracted clay  
29 fractions in order to infer complementary information on weathering processes in this region.  
30 Clay δ<sup>7</sup>Li values correlate positively with Mg/Ti ratios, and are best explained by varying  
31 ratios of leaching versus clay formation rate.

32 Although located in a region annually submitted to monsoon, average silicate weathering rate  
33 (16.1 tons/km<sup>2</sup>/year) and CO<sub>2</sub> consumption rate (0.65x10<sup>12</sup> mol/year) are estimated to be low  
34 when compared to other basaltic regions such as the Deccan Traps, and volcanically active  
35 islands of the tropical zone. This is surprising since the concentrations of Total Dissolved  
36 Solids of the Ethiopian rivers are among the highest ones. With a 2D rainfall model that takes  
37 into account the detailed topography of the region, annual occurrence of the Monsoon, and  
38 monitoring station data, we show that runoff intensity is a key parameter that explains this  
39 difference. We determine that, at present, the weathering of the Ethiopian Traps plays a  
40 negligible role in the carbon cycle. However, simple calculations, which integrate recent  
41 knowledge on African climate variations and on weathering controls, illustrate that during the  
42 African Humid Period (14-8 kyr), a significant increase of Monsoon precipitation may have  
43 resulted in much higher weathering rates and related CO<sub>2</sub> consumption (0.91-1.5x10<sup>12</sup>  
44 mol/year). This study therefore evidences the potential importance of this region in the past,  
45 and the need to quantify more precisely the variations of the monsoon intensity and its impact  
46 on tropical watersheds for reconstructing past CO<sub>2</sub> levels.

47

48

49 **Key words**

50

51 Ethiopian Traps, silicate weathering rate, lithium isotopes, CO<sub>2</sub> consumption, basalt, African  
52 monsoon

53

## 54 **1. Introduction**

55

56 Chemical and physical erosion, which describe respectively mineral dissolution and soil  
57 formation, rock cracking and particle transport are key processes in landscape evolution.  
58 Various parameters can control weathering (chemical erosion) at a watershed scale, among  
59 them lithology, tectonics, relief, soil thickness, age, vegetation cover and climate parameters  
60 (Bayon et al., 2016; Bouchez et al., 2013; Brady et al., 1999; Clergue et al., 2015; Dessert et  
61 al., 2001; Gaillardet et al., 1999; Li et al., 2016; P Louvat and Allègre, 1998; West et al.,  
62 2005). Over geological timescale ( $>10^5$  years), silicate weathering, with organic matter burial,  
63 are major sinks of atmospheric carbon. Weathering of silicate rocks consumes atmospheric  
64  $\text{CO}_2$  and transforms it to dissolved  $\text{HCO}_3^-$  (Gaillardet et al., 1999), which further precipitates  
65 with calcium carbonate in the ocean. As  $\text{CO}_2$  is a greenhouse gas, silicate rock weathering can  
66 be a critical feedback control on the climate (Brady and Carroll, 1994; Walker et al., 1981).  
67 However, at present, the relative importance of the various continental areas and various types  
68 of lithologies having played a key role in the carbon cycle is still debated, in particular over  
69 the Cenozoic. The emplacement of the Himalayas (and its uplift) may have been critical but  
70 this is not consensual thus far. Other mountainous regions may have been of importance, in  
71 particular those rich in basaltic lithologies, and are little considered yet (Li and Elderfield,  
72 2013). Indeed, all silicate rocks do not present the same sensitivity to weathering. Among  
73 them, basaltic rocks can be weathered more easily than any other igneous ones (e.g. Dessert et  
74 al., 2003). Some studies (Dessert et al., 2003; Gaillardet et al., 1999; Louvat and Allègre,  
75 1997; P Louvat and Allègre, 1998) indicate that basaltic rocks are weathered 8 time faster  
76 than granitic rocks. Millot et al., (2002) show that the silicate weathering rate is 10 to 100  
77 times larger for basaltic regions than for granitic areas. Basalt weathering has thus been  
78 estimated as an important sink of  $\text{CO}_2$  at the global scale (Dessert et al., 2003) and it is  
79 calculated to be responsible of 30% of the present-day  $\text{CO}_2$  consumption (Gaillardet et al.,  
80 1999). The Deccan Traps were investigated in great details (e.g. Das et al., 2005; Dessert et  
81 al., 2001; Kisakürek et al., 2004; Vigier et al., 2005), as well as the Siberian Traps (e.g.  
82 Pokrovsky et al., 2005; Zakharova et al., 2005). However, some large basaltic provinces have  
83 not been studied yet. For example, weathering of the Ethiopian Traps (upper Blue Nile Basin)  
84 is not well constrained, despite the importance of this region in terms of climate, size,  
85 tectonics and age (30 Ma). A recent carbon-climate modeling study highlights that the  
86 Ethiopian Traps were the world-wide most important hotspot of silicate weathering during the  
87 Oligocene (Lefebvre et al., 2013). The invoked reason is that, during this period, the

88 Ethiopian Traps were located within the equatorial latitudes, affected by high precipitation  
89 rates and high temperatures, two parameters known to strongly increase basaltic weathering  
90 rates (Dessert et al., 2003; Godd ris et al., 2013). At present the Ethiopian Traps are  
91 influenced by monsoonal climate, with most precipitation occurring during the summer  
92 period. The monsoon intensity in North-East Africa fluctuated significantly in the past, with,  
93 for instance, particularly high peaks during high insolation periods such as the African Humid  
94 period (14.5 to 8 ka, (Lamb et al., 2018; Shanahan et al., 2015)). Numerous studies show that  
95 monsoon variations controlled physical erosion and the Blue Nile sediment flux during the  
96 Quaternary period (Revel et al., 2015, 2014, 2010). As shown by lithium isotope data for the  
97 Nile Basin (Bastian et al., 2017) and for the Himalayas (Dosseto et al., 2015), or hafnium  
98 isotopes for the Congo Basin (Bayon et al., 2012), continental silicate weathering can respond  
99 rapidly to monsoon fluctuations. Thus, the impact of the Ethiopian Traps on the atmospheric-  
100 ocean carbon system deserves to be assessed both at present and in the past.

101 In fact, from sedimentary records located in the Nile deep sea fan, the importance of physical  
102 erosion in this region, and the reactivity of sediment transport to hydro-climate change have  
103 extensively been highlighted. Neodymium isotopes measured in the silt fractions (Blanchet et  
104 al., 2014, 2013, Revel et al., 2015, 2010) of Quaternary sediments have evidenced co-  
105 variations of terrigenous flux from the Blue Nile River with monsoon intensity. More  
106 recently, (Bastian et al., 2017) have explored the use of Li isotope composition of the clay  
107 fractions to determine paleo-variations of silicate weathering in this region. Indeed, lithium  
108 isotope composition of river waters and sediments can be used to characterize the weathering  
109 intensity (Bouchez et al., 2014, 2013, Dellinger et al., 2017, 2015, 2014; Pogge von  
110 Strandmann et al., 2017). Lithium is powerful to trace silicate weathering because it is  
111 principally derived from the chemical erosion of silicate minerals (Kisakurek et al., 2005;  
112 Millot et al., 2010). During chemical erosion in soils or in weathering profiles, no significant  
113 fractionation during rock dissolution can be highlighted (Pistiner and Henderson, 2003; Ryu  
114 et al., 2014; Verney-Carron et al., 2011; Wimpenny et al., 2010). In contrast, large isotope  
115 fractionations occur during clay neoformation, in favor of the light  $^6\text{Li}$  (into octahedral sites).  
116 At a watershed scale, simple modeling taking into account most recent knowledge on the  
117 magnitude of Li isotope fractionation indicate that high clay  $\delta^7\text{Li}$  may correspond to low ratio  
118 of dissolution/clay neoformation rate (Bastian et al., 2017; Bouchez et al., 2013), or low  
119 weathering intensity (Dellinger et al., 2015).

120 A large amount of Li studies is focused on the river dissolved phases (Millot et al., 2010;  
121 Pogge von Strandmann et al., 2017; Vigier et al., 2009). However, recently, the potential of

122 river suspended particles (Dellinger et al., 2017, 2015, 2014; Millot et al., 2002), and of  
123 sediments deposited in terraces, lake or delta (Bastian et al., 2017; Dosseto et al., 2015) to  
124 trace past variations of silicate weathering has also been demonstrated (Bastian et al., 2017;  
125 Dosseto et al., 2015; Lechler et al., 2015). By studying the Nile Delta during the last 32kyr,  
126 Bastian et al (2017) show that clays deposited by the Nile River were mostly smectite formed  
127 in the Ethiopian Traps. Their  $\delta^7\text{Li}$  values co-vary with climate proxies, and in particular with  
128 precipitation variations. The controls of lithium isotope composition for clays formed in the  
129 Ethiopian Traps are therefore important to investigate further, and should help to refine our  
130 understanding of the relationships between climate and basalt weathering.

131 In this study, we target to provide chemical erosion rates and related  $\text{CO}_2$  consumption rate in  
132 the region of the Blue Nile River, the major river draining the Ethiopian Traps basins. For  
133 this, we use the first large dataset of chemical compositions of water and particles from most  
134 rivers draining the Ethiopian Traps. The objectives are (1) to study weathering rates, their  
135 controls, and their relationships with clay Li isotope composition in this region, and (2) to  
136 give an estimation of  $\text{CO}_2$  consumption rate during an intensive runoff period, well-known  
137 and well evidenced in Nile delta sedimentary cores, the African Humid Period (14.5-8kyr).

138

## 139 **2. Study area**

140

141 The Ethiopian Continental Flood Basalt (CFB) province is located in the North and West  
142 Ethiopian highlands (Fig 1 & Fig S1). It covers a surface of about 800,000  $\text{km}^2$  and has an  
143 estimated volume ranging between 1 and 1.5 million  $\text{km}^3$  (Rochette et al., 1998). They  
144 erupted 30 Ma ago over a short time period, 1-2 Ma (Hofmann et al., 1998). The lave pile  
145 thickness varies from 500 to 2000 m with maximum thickness of 3000 m in the northern part  
146 of the plateau (Mohr. P, 1988). This CFB province is mainly composed by transitional basalts  
147 of low-Ti and high-Ti affinities (Pik et al., 1999, 1998), and is covered by a rhyolite terminal  
148 formation on the western part of the plateau, along the main escarpment (Ayalew et al., 2002).  
149 The Ethiopian Traps exhibits a mean altitude of ~ 3000 m with flat under lying remnants of  
150 the original Oligo-Miocene surface. The volcanoes emplaced on the plateau present a higher  
151 elevation up to 4600 meters. The Ethiopian Highlands have been uplifted since at least 25 Ma  
152 (Pik et al., 2003) with concomitant development of the two main basins of the Nile and  
153 Tekeze rivers. The plateau presents important border escarpments with steep slopes,  
154 especially along the roughly N-S marginal graben on its Eastern side (Fig. 1). These highlands  
155 have been dissected by rapid development and propagation of the Nile and Tekeze basins

156 towards its most and early uplifted Eastern edge (Pik et al., 2003), which acted as the main  
157 tectonic drainage divide since Oligo-Miocene times. This tectonic divide is separating the  
158 two large westward basins (Nile and Tekeze) from the short and parallel eastward basins  
159 developed along the marginal graben after initiation of extension 25 Ma ago (Wolfenden et  
160 al., 2004).

161 This tropical region is submitted to the African monsoon, with a mean annual temperature  
162 around 21°C. Thus, 90% of the annual precipitations occur during the summer month  
163 (Korecha and Barnston, 2007) (Fig 1). Precipitations display an East-West gradient in  
164 Ethiopia: rainfall increase from 200 mm/year on the Indian Ocean to 1800 mm/year in the  
165 Highlands. At present, the Blue Nile River (major affluent of the Nile River) and its  
166 tributaries drain the Ethiopian Traps. Thus, this high topography region is estimated to  
167 contribute to 60 % of the total water discharge of the main Nile and to 60 % of its terrigenous  
168 sediments (Fielding et al., 2017; Garzanti et al., 2015; Nyssen et al., 2005). In the past, during  
169 the last 30 kyr, the positioning and the intensity of the monsoon rain belt has oscillated  
170 (Shanahan et al., 2015), in relation with the precession parameter (Tuenter et al., 2003) which  
171 lead to stronger precipitation every 21 kyrs. These strong runoff variations are recorded by  
172 Ethiopian lake sediments (Amaral et al., 2013) and by the sediments transported by the Nile  
173 River and deposited in its delta (Blanchet et al., 2014; Castañeda et al., 2016; Revel et al.,  
174 2015, 2010).

175

### 176 **3. Sampling and methods**

177

#### 178 **3.1. Sampling**

179

180 All samples come from the northern Ethiopian plateau (Fig 1), which encompasses the large  
181 catchment of the Blue Nile (64 000 km<sup>2</sup>), several smaller sub-basins (1000 to 6000 km<sup>2</sup>), as  
182 well as small catchments emplaced directly on the top of the plateau (5 to 100 km<sup>2</sup>). Two  
183 sampling campaigns have been carried out in June 2004 (ETH04) and in March 2005  
184 (ETH05). At each sample site, sand, mud and river waters were collected. Waters were  
185 filtered on the field with 0.2µm filters and immediately acidified with purified nitric acid.  
186 Sands and muds were sampled on the river banks after mixing 5 different aliquots of about  
187 100g each. The difference between mud and sand is the visible granulometry: sand samples  
188 correspond to coarse sized mineral (> 2mm grain) rich-particles and the mud samples to the

189 finest grain deposits. At the laboratory, clay-size fractions were extracted from the mud  
190 samples before chemical and isotope analyses (see section 3.3.2).

191

## 192 3.2. Runoff Modeling

193

194 The hydrology of the Blue Nile upper basins is strongly controlled by seasonal precipitations,  
195 with strong rains and flooding observed during the summer monsoon, from June to September  
196 (e.g. Conway, 2009). These conditions are not appropriate for providing a reliable estimate of  
197 annual discharge and runoff for each studied zone, using classical methods. In order to  
198 provide these estimations for each sampled sub-basin, necessary for weathering rate  
199 calculations, we used a numerical model based on the IDRISI © GIS software. This model  
200 was run using the plateau topography and available public rainfall data. In order to avoid  
201 unconstrained biases linked to evapotranspiration, the output has been convincingly calibrated  
202 with current discharge data available for the upper Blue Nile basin (Conway, 1997). These  
203 calibrated discharge and runoff data for each sub-basin are presented in Table 1.

204

## 205 3.3. Analytical protocol

206

### 207 3.3.1. Chemical analyses

208

209 Major and trace element concentrations were analyzed in waters and sediments using ICP-  
210 AES and ICP-MS at the SARM National Facilities in Nancy (France)  
211 (<http://helium.crpg.cnrs-nancy.fr/SARM/index.html>). Anions have been measured using ionic  
212 chromatography Dionex ICS3000-S with AG11-HC/AS11-HC column. Corresponding  
213 uncertainties and detection limits are described in ([http://helium.crpg.cnrs-  
214 nancy.fr/SARM/pages/roches.html](http://helium.crpg.cnrs-nancy.fr/SARM/pages/roches.html)).

215

### 216 3.3.2. Lithium isotope analyses of the clay fractions

217

218 Clays were extracted from the mud samples, by decantation of the sediments in 50ml of ultra-  
219 pure water mixed with 20µl of sodium hexametaphosphate (100mg/l). This separation allows  
220 to extract particles with a diameter and a density lower than 2 µm and 1.8 respectively. Clay  
221 fractions were finely crushed into an agate mortar and 10 mg of the fine powder was  
222 mineralized using a concentrated HF/HNO<sub>3</sub>/HCl solution in a clean laboratory at LOV, using



223 ultrapure acids. The solution was evaporated and the residue dissolved in 1.00N HCl. The  
224 total dissolution of samples was systematically verified by the absence of residue after 20 min  
225 of centrifuging at 4000rpm.

226 For major and trace element analyses, an aliquot was diluted 300 times and analyzed by ICP-  
227 OES. For lithium isotope analyses, an aliquot containing  $\approx 60\text{ng}$  Li was introduced on a  
228 cationic resin column and eluted with 1.00N HCl, following the method published in (Bastian  
229 et al., 2017; Vigier et al., 2008). This separation was performed twice and allowed Li  
230 recovery and purification. For all samples, the total cationic concentration of the sample was  
231 constrained to be less than 10% of the total cationic capacity of the column resin. All seawater  
232 and rock reference materials were submitted to the exact same protocol.

233 Lithium isotope measurements were performed using a Neptune Plus (ThermoFisher) multi-  
234 collector inductively coupled plasma spectrometer (MS-ICP-MS) at the Ecole Nationale  
235 Supérieure de Lyon (CNRS National Facilities). Sample was introduced through an Aridus II  
236 desolvating system before ionization to increase sensitivity ( $1\text{V } ^7\text{Li} / \text{ppb Li}$ ). We used the jet  
237 and X cones, as described in (Balter and Vigier, 2014). To correct for instrumental mass bias,  
238 we used the standard bracketing technique, using the LSVEC standard. The total blank  
239 procedure is systematically lower than  $10\text{pg Li}$ . The accuracy of isotopic analysis was verified  
240 by measuring Li7-N reference solution (Carignan et al., 2007) and various reference materials  
241 (BE-N basalt, GH granite and seawater). Without chemical purification, an average  $\delta^7\text{Li}$  value  
242 of  $30.2 \pm 0.4\text{‰}$  (2SD,  $n=32$ ) is obtained for the pure Li7-N standard, in agreement with  
243 published and theoretical value. After chemical purification, mean  $\delta^7\text{Li}$  values are  $30.3 \pm 0.4\text{‰}$   
244 (2SD,  $n=22$ ),  $5.12 \pm 0.9\text{‰}$  (2SD,  $n=5$ ),  $1.64 \pm 0.6\text{‰}$  (2SD,  $n=3$ ) and  $31.1 \pm 0.6\text{‰}$  (2SD,  $n=6$ ), for  
245 Li7-N, BE-N basalt, GH granite and seawater respectively. All values for reference materials  
246 compare well with published values (Carignan et al., 2007; Millot et al., 2004).

247

## 248 **4. Results**

249

### 250 4.1. Runoff and discharge

251

252 As described in section 3.2, we have estimated runoff values using a specific numerical model  
253 including published data for rainfall and discharge, and regional topography. The runoff  
254 values determined for 19 sub-basins range from 190 to 273 mm/yr with a mean value of 242  
255 mm/yr. The corresponding **calculated** discharges range from  $11 \times 10^5 \text{ m}^3/\text{yr}$  (for the Ulet Wenz  
256 River) to  $1.8 \times 10^{10} \text{ m}^3/\text{yr}$  (for the Blue Nile river), with a mean value of  $19 \times 10^8 \text{ m}^3/\text{yr}$  (Table 1).

257 For the Blue Nile river, the calculated discharge is close to the annual discharge measured at  
258 the Roseires Dam Station (<http://www.grdc.sr.unh.edu/html/Polygons/P1663800.html>) :  
259  $4.7 \times 10^{10}$  m<sup>3</sup>/yr. The estimated average runoff value of 242 mm/yr for the Ethiopian Traps is in  
260 line with the value published in ((Dessert et al., 2003), 129 mm/year). These two comparisons  
261 indicate that our model estimates representative values for the runoff and river discharge.  
262 The average runoff for this region is two times lower than the estimated value for the Deccan  
263 Traps (463 mm/year). It is also lower than the one determined for the Siberian Traps (324  
264 mm/yr). Compared to small tropical basaltic areas such as Hawaii (1612 mm/yr) and the  
265 Reunion Island (2433 mm/yr), the Ethiopian runoff is also low. This is partly due to their  
266 topography, their small surface area compared to the Ethiopian Traps (800,000 km<sup>2</sup> vs 16760  
267 km<sup>2</sup> in Hawaii and 2512 km<sup>2</sup> in the Reunion Island), and the occurrence of frequent  
268 hurricanes and tropical storms. In Iceland, high runoff (1883-2432 mm/yr) values are mostly  
269 due to ice melting, which introduces water in the riverine system during the summer season  
270 (Gislason et al., 2009).

271

## 272 4.2. Major and trace elements

273

### 274 4.2.1 River waters

275

276 In waters, the dominant cations are, in increasing order of concentrations: Si, Mg, Na, and Ca,  
277 with mean values of 447 ( $\pm$  219), 1768 ( $\pm$  502), 712 ( $\pm$  332) and 883 ( $\pm$  417)  $\mu$ mol/l  
278 respectively. These concentrations are corrected from the atmospheric input using Cl  
279 concentration and oceanic X/Cl molar ratios, following the method detailed in (Berner and  
280 Berner, 1987). After corrections the concentration of Mg\*, Na\*, and Ca\* are 880 ( $\pm$  415), 68  
281 ( $\pm$  26) and 574 ( $\pm$  285)  $\mu$ mol/l respectively.

282 On average, molar Ca\*/Na\* and Mg\*/Na\* ratio are 2.5 ( $\pm$  0.9) and 0.4 ( $\pm$  0.6), respectively  
283 (Table 2). For comparison, average Ca/Na and Mg/Na in Ethiopian basalts are 3.9 ( $\pm$  1.3) and  
284 3.2 ( $\pm$  2.7) respectively, and in the Ethiopian rhyolite are 0.07 ( $\pm$  0.06) and 0.03 ( $\pm$  0.04).

285 When compared with other regions of similar lithologies, river water chemical composition  
286 displays roughly a single trend between basalts and rhyolites (Fig. 2). From this trend, it can  
287 be estimated that the dissolved contribution coming from rhyolite weathering is low (<15%)  
288 compared to basaltic rocks.

289 The dominant anions are, in increasing order of concentrations: HCO<sub>3</sub><sup>-</sup>, SO<sub>4</sub><sup>2-</sup>, Cl<sup>-</sup>, NO<sub>3</sub><sup>-</sup>, and  
290 F, with mean values of 3170 (406 to 6039) 202 (15 to 880), 156 (15 to 348), 71 (4 to 292) and

291 10 (5 to 21)  $\mu\text{mol/l}$  respectively. The anion concentrations are typical of those measured in  
292 other rivers draining basaltic regions (e.g. Dessert et al., 2001).

293

#### 294 4.2.2 River bed particles

295

296 Sands are enriched in  $\text{SiO}_2$  compared to mud samples (Table 3), ( $51.9 \pm 10.6\%$  on average vs  
297  $43.2 \pm 2.6\%$  for muds) and  $\text{Na}_2\text{O}$  ( $2.2 \pm 0.8\%$  vs  $1.1 \pm 0.5\%$ ), while depleted in  $\text{Al}_2\text{O}_3$  ( $11.7$   
298  $\pm 3.4\%$  vs  $13.0 \pm 3.5\%$  in mud),  $\text{CaO}$  ( $7.0 \pm 3.4\%$  vs  $4.7 \pm 2.1\%$  in mud) and  $\text{Fe}_2\text{O}_3$  ( $11.4 \pm$   
299  $3.4\%$  vs  $13.5 \pm 1.5\%$ ).  $\text{TiO}_2$  and  $\text{MgO}$  ( $5.7 \pm 4.0\%$  vs  $5.8 \pm 3.1\%$ ) concentrations are similar  
300 between sands and muds: ( $\approx 3.1 \pm 1.4\text{ wt}\%$  and  $\approx 5.8 \pm 3.1\text{ wt}\%$  respectively). Si/Al ratio is  
301 lower by  $\approx 30\%$  in muds compared to sands, which is consistent with occurrence of  
302 aluminosilicate minerals (clays/smectite) in the collected muds. The chemical composition of  
303  $<2\mu\text{m}$  clay fractions extracted from mud samples, contains, on average,  $24.2\%$  ( $\pm 4.7\%$ )  
304  $\text{Al}_2\text{O}_3$ ,  $22.2\%$  ( $\pm 1.8\%$ )  $\text{Fe}_2\text{O}_3$ , and  $15.1\%$  ( $\pm 6.9\%$ )  $\text{MgO}$ . We note that the mud samples  
305 collected in river banks are not representative of the clay fraction separated at the laboratory  
306 and rather reflects a mixing between clays and primary phases. Thus, clay fractions ( $<2\mu\text{m}$ )  
307 were preferentially used for assessing weathering processes in this region.

308 Some trace elements measured in the solid phases can be used to study chemical erosion  
309 degree or intensity (Louvat and Allègre, 1997; P Louvat and Allègre, 1998). In the diagram  
310 Sr/Th versus U/Th (Fig 3, Table 2 and 3), when compared to bedrocks, Ethiopian river waters  
311 display elevated U/Th and Sr/Th ratios (respectively  $>2$  and  $> 8405$ ) while the solid phases  
312 display low U/Th and Sr/Th values (respectively  $<0.3$  and  $<87.8$ ). Sr and U are two mobile  
313 elements during water-rock interactions and are usually depleted in the solid phase, in contrast  
314 with Th, which is an immobile element that can easily adsorb, except in organic-rich  
315 environments (Dupré et al., 1999). Figure 3 also indicates that the river particles are mostly  
316 derived from basaltic rocks, as previously described, and that the influence from rhyolite  
317 weathering is minimal, except for one sample (Nazero Wenz), which is located in a rhyolite  
318 rich area (Figure S1).

319

#### 320 4.3 Lithium isotope compositions of clay size ( $<2\mu\text{m}$ ) fractions

321

322 Ethiopian bedrock samples exhibit a mean  $\delta^7\text{Li}$  value of  $5 \pm 0.5\text{‰}$  ( $n=3$ ) (Bastian et al., 2017)  
323 (Fig. 4), which compares well with the range published for intraplate basalts. Compared to  
324 these rocks, or to tholeiitic basalts ( $3.1$  to  $5.2\text{‰}$ , (Elliott et al., 2006)), and to OIB basalts ( $3.1$

325 - 5.2 ‰, (Genske et al., 2014)) clays deposited by Ethiopian rivers are significantly enriched  
326 in <sup>6</sup>Li: clay δ<sup>7</sup>Li values range from 0.3‰ (± 0.5‰) to 4‰ (± 0.5‰) , with a mean value of  
327 2.0‰ (± 1.2‰). This is consistent with preferential enrichment in light lithium isotope during  
328 smectite neoformation, as seen experimentally (Vigier et al., 2008) and in basaltic soils (Huh  
329 et al., 2004; Kisakürek et al., 2004; Pistiner and Henderson, 2003; Ryu et al., 2014).  
330 In the literature, a large range of δ<sup>7</sup>Li values is observed in suspended particles (SPM) of  
331 rivers draining basaltic regions e.g. in Iceland (with values from -1.3 to 7.5‰, (von  
332 Strandmann et al., 2006)) and in the Azores archipelago (3.5 to 9.5‰, (von Strandmann et al.,  
333 2010)). Also in soils, the range of published δ<sup>7</sup>Li for bulk soils samples is large (-15 to 10‰,  
334 (Clergue et al., 2015; Pistiner and Henderson, 2003; Ryu et al., 2014; Schmitt et al., 2012; von  
335 Strandmann et al., 2012)). These observed variations are possibly due to mixing between clay  
336 and primary minerals, because these samples are not necessarily representative of clay pure  
337 endmember (Dellinger et al., 2014). Here, clays were extracted from river sediments in order  
338 to minimize the influence of primary minerals.

339

## 340 **5. Discussion**

341

### 342 **5.1. Assessing weathering processes at the basin scale**

343

344 As described in section 4.2, chemical and isotopic compositions of river waters and particles  
345 can give indications on weathering rates and/or intensity at the watershed scale. Some  
346 elements are particularly mobile during basalt alteration (e.g. Sr, K, Na), and preferentially  
347 lost from weathering profiles after a short period of time (Chadwick et al., 1999; Ryu et al.,  
348 2014). Thus, as described in section 4.2, Sr and U are depleted in sands and mud samples,  
349 while enriched into the dissolved phases (Fig. 2) of the corresponding rivers. Also, Na, Ca and  
350 Mg concentrations are typical of waters draining basaltic rocks (Fig. 3). However, using  
351 elementary ratios, or elementary concentrations, it is often difficult to distinguish the role of  
352 lithology from the one of dissolution or leaching process, because of the chemical diversity of  
353 the drained bedrocks. As described in previous works (Pistiner and Henderson, 2003; Vigier  
354 et al., 2008; Wimpenny et al., 2010), lithium isotopes provide a complementary mean to  
355 examine chemical erosion processes (Bastian et al., 2017; Dellinger et al., 2015). Bastian et  
356 al., (2017), have analyzed 55 clay fractions extracted from Nile deep sea fan core sediments.  
357 Their δ<sup>7</sup>Li values vary between 4 and -1.2 ‰ during the last 32,000 years. In their study, clay  
358 δ<sup>7</sup>Li were related to the ratio between dissolved Li flux and the flux of Li incorporated in the

359 secondary phases ( $F_{lea}^{Li}/F_{sp}^{Li}$ ). In rivers currently draining the Ethiopian Traps, clay fractions  
 360 display variable  $\delta^7Li$  values, from 0.3‰ to 4.0‰, a range that is similar to the 0-32 kyr clays  
 361 deposited in the Nile delta. Clay  $\delta^7Li$  and Mg/Ti ratios display a positive correlation (shown  
 362 in fig. 4), which suggests that both proxies are impacted by similar processes during basalt  
 363 weathering. This is not surprising since  $Li^+$  essentially substitutes to  $Mg^{2+}$  during smectite  
 364 formation (Decarreau et al., 2012; Vigier et al., 2008), and smectite is the most abundant  
 365 phase carried by rivers draining the Ethiopian Traps (Garzanti et al., 2015).

366 Using a simple modeling, based on mass balance, clay  $\delta^7Li$  values have been shown to be  
 367 essentially explained by the ratio of leaching versus clay formation rates, and thus reflect  
 368 different weathering intensities (Bastian et al., 2017; Bouchez et al., 2013; Vigier et al., 2009;  
 369 Vigier and Godd ris, 2015; von Strandmann et al., 2012). The behavior of Li isotopes at the  
 370 scale of a watershed or of a continent has been described with the following equations,  
 371 considering a simple mass balance between bedrock, rivers and soils at the watershed scale. It  
 372 is defined such that all the lithium released by rock dissolution is either incorporated into  
 373 secondary phases formed in weathering profiles or released into superficial waters (and *in fine*  
 374 in rivers).

375

$$376 \quad F_{lea}^{Li}\delta_{lea} = F_{sp}^{Li}\delta_{sp} + F_{riv}^{Li}\delta_{riv} \quad \text{equation (1)}$$

377

378 where  $F_{lea}^{Li}$ ,  $F_{sp}^{Li}$ , and  $F_{riv}^{Li}$  are the Li flux related to dissolution, secondary phase precipitation  
 379 and soil water outflow respectively.  $\delta_{lea}$ ,  $\delta_{sp}$ ,  $\delta_{riv}$  are the Li isotopic signature ( $\delta^7Li$  value) of  
 380 the dissolved phase, the secondary phase, the river water and the bedrock, respectively. Since  
 381 basalt dissolution is isotopically congruent,  $\delta_{lea} = \delta_{Bedrock}$  (Verney-Carron et al., 2011;  
 382 Wimpenny et al., 2010). The isotope fractionation between dissolved phase and the secondary  
 383 phases is noted  $\Delta_{land}$ . After simplification, the isotope composition of the secondary phases is  
 384 given by:

385

$$386 \quad \delta_{sp} = \frac{\Delta_{land}}{F_{lea}^{Li}/F_{sp}^{Li}} + \delta_{BedRock} - \Delta_{land} \quad \text{Equation (2)}$$

387

388 This equation states that if  $F_{sp}^{Li}$  is close to  $F_{lea}^{Li}$ , then  $\delta_{sp} = \delta_{BedRock}$ . If  $F_{sp}^{Li}$  is low compared  
 389 to  $F_{lea}^{Li}$ , then  $\delta_{sp} = \Delta_{land}$ .

390

391 The positive correlation between clay  $\delta^7\text{Li}$  and Mg/Ti cannot be explained by a binary mixing  
392 between bedrocks and Li-rich clays, as illustrated in Figure 4. Thus, a similar mass balance  
393 between rocks and weathering products, as for  $\delta^7\text{Li}$ , was applied to clay Mg/Ti. For solving it,  
394 we assumed that no Ti is released into water during dissolution (due to the high immobility of  
395 this element during water/rock interactions) ( $\text{Ti}_{rlea} = 0$ ). Thus Mg/Ti ratio of secondary  
396 phases can be described as:

397

$$398 \quad \left(\frac{\text{Mg}}{\text{Ti}}\right)_{sp} = \frac{1}{\frac{F_{lea}^{Mg}}{F_{sp}^{Mg}} \cdot \frac{\text{Ti}_{clay}^{mean}}{\text{Mg}_{BedRock}}} \quad \text{Equation (3)}$$

399

400 The relation between the Li and the Mg flux becomes:

401

$$402 \quad \frac{F_{lea}^{Li}}{F_{sp}^{Li}} \cdot \beta = \frac{F_{lea}^{Mg}}{F_{sp}^{Mg}} \quad \text{Equation (4)}$$

403 with  $\beta$  reflecting the Li/Mg ratio of the clay fraction (normalized to the source rock one).

404 As shown in Figure 4, some basins are characterized by low clay  $\delta^7\text{Li}$  and low Mg/Ti values,  
405 (ie Jara Wenz (ETH05-11), Borkena Wenz (ETH05-13), Robi Wenz (ETH05-4), Meri Wenz  
406 (ETH05-28), Tikur Wenz (ETH05-19)). For these small basins, the weathering model  
407 indicates that soil formation rate ( $F_{sp}$ ) is low compared to the leaching rate ( $F_{lea}$ ). Other areas  
408 display high clay  $\delta^7\text{Li}$  values associated to high Mg/Ti ratios (Ketching Abeba Wenz  
409 (ETH05-25), Gereb Ara Wenz (ETH05-40), Tekeze (ETH0535), Hormat Wenz (ETH05-47),  
410 Ala-Wiha (ETH05-20), Nazero Wenz (ETH05-6)).

411 Following our modeling, the various  $\delta^7\text{Li}$  and Mg/Ti values are best explained by variations  
412 of the weathering intensity, leading also to a large range of clay Li/Mg ( $\beta$  parameter). For a  
413 large part, the  $\beta$  values are greater than 1 (Fig. 4). This indicates that Li is more retained in the  
414 weathering profiles than Mg (compared to source rocks). This is consistent with observations  
415 made in various Hawaiian weathering profiles where kaolinite prevail (Ryu et al., 2014).

416 Thus, the trend between clay  $\delta^7\text{Li}$  and clay Mg/Ti is likely due to variable weathering  
417 conditions. However, no clear relationship with the runoff, the topography or vegetation can  
418 be highlighted for this group of rivers. Since Mg is involved in the vegetation cycle, and Li is  
419 not, perhaps a loss of Mg in relation with recent plant and root activity can also explain part of

420 this variability. More work would certainly need to be undertaken in the most vegetated areas  
421 of this region since no convincing correlation between  $\delta^7\text{Li}$  or Mg/Ti and physical parameters  
422 such as the slope, runoff, vegetation and land use can be highlighted here. This suggests that,  
423 at present, basalt weathering and soil production rate in the Blue Nile Basin are variable in  
424 space, and that this variation is likely to be a complex function of several parameters.

425

## 426 **5.2. Weathering rates and controls in the Upper Blue Nile Basin**

427

428 As discussed above, the relatively large range of clay Mg/Ti and  $\delta^7\text{Li}$  reflects a significant  
429 variation of weathering conditions in the Ethiopian Traps region while the lithology is  
430 dominantly basaltic. It is possible to use the chemical composition of river waters in order to  
431 quantify present-day weathering rates (Table 2) at the scale of each corresponding watershed  
432 (e.g. Gaillardet et al., 1999; Stallard and Edmond, 1983). The Total Dissolved Solid (TDS) is  
433 defined by the sum of major mobile cations (Ca, Mg, K and Na) concentrations in the  
434 dissolved phases of rivers. These concentrations are corrected from the atmospheric input  
435 using Cl concentration and oceanic X/Cl molar ratios, following the method detailed in (Das  
436 et al., 2005). TDS\* values (\* indicate corrected from atmospheric inputs) estimated from  
437 concentrations measurement in the dissolved phase of the Blue Nile Basin rivers range from  
438 9.5 to 111.0 mg/l, with a mean value of 67 mg/l. In the diagram TDS\* versus temperature, the  
439 Blue Nile tributaries plot along the same trend defined by all other basaltic regions studied  
440 thus far (Fig. 5a). This first indicates the good representative nature of our sampling and  
441 methods. The average TDS\* for the Ethiopian Traps is as high as the one determined for the  
442 Deccan Traps (54 mg/l) and for Java Island (43 mg/l) (Dessert et al., 2003; Li et al., 2016),  
443 which are also characterized by elevated annual temperature (27°C and 24.8 °C respectively).  
444 These three regions display river TDS\* values significantly greater than in Hawaii (7.4 mg/l)  
445 and in Iceland (8.8-10.3 mg/l), displaying both lower surface temperatures on average (16 and  
446 2 °C respectively). This trend confirms that, for basaltic basins, temperature is a key  
447 parameter which controls river TDS\* (Dessert et al., 2003; Li et al., 2016).

448 Using the TDS\* values, the estimated annual discharge, and the watershed area, it is possible  
449 to determine silicate weathering rates. These rates in the Ethiopian Traps basins range from  
450 2.6 to 24.4 tons/km<sup>2</sup>/year, with a mean value of 16.1 tons/km<sup>2</sup>/year (Table 2). For the Blue  
451 Nile River, sampled at the output of the Ethiopian Traps (see map Fig. 1), the silicate  
452 weathering rate is 11.3 tons/km<sup>2</sup>/year. The mean value estimated for chemical erosion rate in  
453 the Ethiopian Traps, based on 17 data (Table 2), is consistent with the positive correlation

454 previously observed with the runoff (see Fig. 5b, (Dessert et al., 2003)). In contrast to TDS\*,  
455 weathering rates in Ethiopia are low compared to other regions of similar lithology. For  
456 instance, the weathering rate in the Deccan Traps is 25.1 tons/km<sup>2</sup>/year, 48.4 tons/km<sup>2</sup>/year  
457 for La Reunion, and 152 tons/km<sup>2</sup>/year for Java (Dessert et al., 2003). Lower runoff in the  
458 Ethiopian Traps (242 mm/years) compared to these regions is certainly a key factor. However,  
459 other parameters may also play a role, in particular temperatures, the age of the basaltic rocks  
460 (Li et al., 2016), the soil age (Ryu et al., 2014; Vigier et al., 2005), and formation of glassy  
461 material in volcanically active area (e.g. Java Island and Iceland) (Gislason et al., 2009;  
462 Gislason and Oelkers, 2003; Vigier et al., 2009).

463 Overall, both river particles and water composition can provide valuable information on  
464 chemical erosion rate in the various basins. However here there is no correlation between  
465 river particles and water chemical compositions. This is likely to be due to a recent  
466 decoupling between soil (or weathering profile) formation and river aqueous phase. Indeed,  
467 large soil reservoir where clays have formed can encompass different weathering conditions,  
468 issued from past climate variations, and may therefore not be representative of the current  
469 state. Recent observations in the Nile deltaic sediments show that both chemical and physical  
470 erosion responded rapidly to hydro-climate change in the last 30kyr (Bastian et al., 2017).  
471 This argues for a residence time of clay-rich sediments that is not above a few thousand years  
472 at most, in coherence with elevated topography of the Traps where clays mainly come from.  
473 The observed decoupling between clays and waters must then have occurred just recently,  
474 perhaps due to intensification of anthropic activities in this region (Lanckriet et al., 2015)

475

### 476 **5.3. CO<sub>2</sub> consumption and SO<sub>4</sub><sup>2-</sup> concentration due to silicate weathering**

477

478 From river water chemical compositions, it is possible to estimate the CO<sub>2</sub> consumption rates  
479 related to silicate weathering. In the literature, several methods have been used, using  
480 corrections from source rocks in mixed lithology basins, calcium concentration in silicate  
481 rivers (Goddéris et al., 2013), or HCO<sub>3</sub><sup>-</sup> concentrations in carbonate-free watersheds (e.g. Das  
482 et al., 2005). Since the Ethiopian Traps are essentially constituted of silicate rocks, with little  
483 vegetated areas, major cations concentrations in rivers have only been corrected for  
484 atmospheric inputs, as described in section 5.1. HCO<sub>3</sub><sup>-</sup> concentrations are determined using  
485 the following electronic mass balance:

486

$$487 \quad [\text{HCO}_3^-] = 2[\text{Ca}^{2+}] + [\text{K}^+] + 2[\text{Mg}^{2+}] - 2[\text{SO}_4^{2-}] - [\text{F}^-] - [\text{NO}_3^-] \text{ Equation (5)}$$



488

489 where all concentrations are corrected for atmospheric inputs (see section 4.2.1). The average  
490 pH of the Ethiopian rivers (7.5-8.2, (MERID, 2005)) makes it to possible to assume that  
491  $\text{HCO}_3^-$  is the dominant anion compared to  $\text{CO}_3^{2-}$ . Thus, in order to determine the  $\text{CO}_2$   
492 consumption related to silicate weathering, we consider that 1 mole of  $\text{HCO}_3^-$  exported by  
493 rivers corresponds to 1 mole of consumed atmospheric  $\text{CO}_2$ .  $\text{CO}_2$  consumption rates are then  
494 calculated using the river discharge and the basin surface area (for specific rates). Some recent  
495 studies (Calmels et al., 2007; Li et al., 2008; Torres et al., 2016) show that sulfate mineral  
496 dissolution produces sulfuric acid that can enhance bedrock and mineral dissolution, in  
497 particular in sedimentary or carbonate rich context. Any silicate weathering due to sulfuric  
498 acid may lead to an overestimation of the  $\text{CO}_2$  consumption rate if this effect is not considered  
499 into the calculations. In the rivers studied here, dissolved sulfate concentrations do not  
500 correlate to dissolved nitrate but rather with elements that are mobile during silicate  
501 dissolution (for example Ca and Mg). This suggests that sulfates are not significantly related  
502 to fertilizer addition but may mostly derive from weathering and oxidizing processes. This is  
503 not surprising since this region is not intensively cultivated, and is also supported by small  
504 amounts of nitrate in waters (see section 4.2.1). The use of Equation (5) to calculate  $\text{HCO}_3^-$   
505 allows us to avoid a potential bias due to sulfuric acid in the determination of  $\text{CO}_2$   
506 consumption rate.

507 Table 2 show that specific  $\text{CO}_2$  consumption rate due to basalt dissolution in Ethiopian Traps  
508 vary from  $0.11 \times 10^6$  mol/km<sup>2</sup>/yr (for the Ulet-Wenz basin) to  $1.33 \times 10^6$  mol/km<sup>2</sup>/yr (for the  
509 Ketchiin Abeba Wenz), and is  $0.76 \times 10^6$  mol/km<sup>2</sup>/yr on average. The  $\text{CO}_2$  consumption rate  
510 cover a large range, possible due to local – small scale - effects. Thus, for the comparison  
511 with literature data, we use the two values estimated for the Blue Nile River, which integrates  
512 most waters draining the Ethiopian Traps. The Blue Nile River was sampled at two different  
513 seasons (ETH04-2; Juin-04 and ETH05-52; Mars -05). The two samples lead to similar  $\text{CO}_2$   
514 consumption rate, of 0.49 and 0.52 mol/km<sup>2</sup>/yr respectively, but are low compared to other  
515 basaltic regions. For comparison, this flux is estimated to be  $1.26 \times 10^6$  mol/km<sup>2</sup>/yr for the  
516 Deccan Traps and  $6.41 \times 10^6$  mol/km<sup>2</sup>/yr for Java Island (Dessert et al., 2003). The  $\text{CO}_2$   
517 consumption rate due to silicate weathering in the Blue Nile basin is slightly higher, than the  
518 value published by (Dessert et al., 2003) for the Ethiopian Traps, i.e.  $0.15 \times 10^6$  mol/km<sup>2</sup>/yr.  
519 When using the Dessert et al., (2003) theoretical equation  $f\text{CO}_2 = R_f \times 323.44 \exp(0.0642 T)$   
520 where  $f\text{CO}_2$ ,  $R_f$  and  $T$  are respectively the specific atmospheric  $\text{CO}_2$  consumption

521 (mol/km<sup>2</sup>/year), the runoff (mm/year) and the temperature (°C), the corresponding  
522 consumption rate is also slightly lower (0.16x10<sup>6</sup> mol/km<sup>2</sup>/year) than our calculated value.

523 In order to determine the relative role of the Blue Nile Basin in the present-day carbon cycle,  
524 one needs to estimate its total CO<sub>2</sub> consumption flux (in mol/year). The average CO<sub>2</sub>  
525 consumption rate is calculated to be 0.65x10<sup>12</sup> mol/year. This is significantly higher than in  
526 Iceland (0.070 x10<sup>12</sup> mol/year), the Reunion Island (0.006 x10<sup>12</sup> mol/year) or the Azores  
527 (0.0004 x10<sup>12</sup> mol/year) (see Fig. 6). However, this value remains lower than for the Deccan  
528 (0.392 x10<sup>12</sup> mol/year) or than SE Asia (1.033 x10<sup>12</sup> mol/year). The upper Blue Nile represent  
529 a CO<sub>2</sub> consumption close to the global average ((Gaillardet et al., 1999), see Fig. 6), and only  
530 7.5 % of the total CO<sub>2</sub> consumed by silicate weathering on Earth. It corresponds to 7.8.10<sup>3</sup>Gt  
531 of carbon consumed by year, which is the equivalent of 0.4% of the total human carbon  
532 emission per year. Our results demonstrate that, at present, the Ethiopian Traps weathering  
533 has a negligible impact on the carbon cycle, and thus on global climate.

534 In the past, this could however be different. First, as described in the Introduction, the  
535 modeling study from Lefebvre et al., (2013), show that, during the Oligocene, the Ethiopian  
536 Traps was likely the most important region on Earth in terms of silicate weathering. More  
537 recently, since 100kyr, the North East African region is known to have been submitted to  
538 extreme variations of precipitation and runoff. The monsoon rain belt oscillated together with  
539 the precession cycle of 21,000 years. This induced succession of “humid” and “arid” periods  
540 that affected sediment flux and physical erosion in this region. This is evidenced in lakes  
541 where arid periods are characterized by hiatus or paleo-soils in the sedimentary records and  
542 water level variations (for lake Turkana, lake Abhé (Forman et al., 2014; Garcin et al., 2009;  
543 Hoelzmann et al., 2010)). This is also observed in the marine cores of the Nile deep-sea-fan,  
544 with much higher sedimentation rates during humid periods (Revel et al., 2015, 2014, 2010).

545 Based on both observations and modeling, it is possible to provide first extrapolation of our  
546 results to some critical period of the Quaternary period. Indeed, recent studies illustrate that  
547 during this time, silicate weathering rates were able to respond rapidly to climate change  
548 (Bastian et al., 2017; Dosseto et al., 2015; Rothacker et al., 2018). This region was submitted  
549 to large changes of the African Monsoon intensity (Shanahan et al., 2015) while runoff likely  
550 plays a key role on basalt weathering and related CO<sub>2</sub> consumption ((Goddéris et al., 2013),  
551 see also Fig. 5b and 8). Since 6 kyr, the climate in the Ethiopian Traps is characterized by  
552 relatively low monsoonal precipitation and low runoff. However, between 15 and 8 kyr,  
553 during the African Humid Period (AHP), this region was submitted to much higher  
554 precipitation rates (Blanchet et al., 2013; Revel et al., 2014). Climate modeling studies, using

555 LMDZ and ORCHIDEE models, determined an increase of precipitation in the Ethiopian  
556 Traps of 1.4 to 2.5 times larger during the AHP ((Marzin and Braconnot, 2009) ; Vadsaria and  
557 Ramstein, Pers. Comm.). This corresponds to a runoff ranging between 340 and 600  
558 mm/years. By applying this factor, CO<sub>2</sub> consumption rate during African Humid Period may  
559 thus have increased up to values of 0.91-1.5x10<sup>12</sup> mol/yr (brown bar in Fig. 6), which  
560 corresponds to the most elevated current rates. These estimations are in line with the study of  
561 (Lefebvre 2013), which showed that the importance of this region in the carbon cycle during  
562 the Oligocene period. However, it is important to note that during the Oligocene, the global  
563 temperature was also higher, while for the AHP, the runoff increase in North Africa is likely  
564 not related to a significant increase of temperature (De Boer et al., 2013).  
565 These results also suggest a potential increase of global CO<sub>2</sub> consumption during more humid  
566 periods due to increase of silicate weathering. Indeed, other large basaltic regions (e.g.  
567 Deccan, Parana) are also located in the tropical zone influenced by monsoon. This is likely a  
568 key for understanding past rapid climate change and the influence of tropical basins on the  
569 carbon cycle.

570

## 571 **6. Conclusion**

572

573 This study examines the chemical composition of river water and sediments in the Ethiopian  
574 Traps in order to characterize silicate weathering and corresponding CO<sub>2</sub> consumption in this  
575 region. Lithium isotopic composition of the clay fractions was also examining and present a  
576 positive correlation with clay Mg/Ti ratios, in relation with variable weathering conditions in  
577 this region.

578 The average weathering rate (16.1 tons/km<sup>2</sup>/yr) and CO<sub>2</sub> consumption rate (0.65.10<sup>12</sup> mol/yr)  
579 determined for the Ethiopian Traps are lower than for other basaltic regions (e.g. Deccan or  
580 SE Asia). This is bet explained by the low runoff values observed in Ethiopia at present (242  
581 mm/yr), despite regular occurrence of the monsoon. This demonstrates that, at present, the  
582 Ethiopian Traps weathering has a negligible impact on the carbon cycle, and thus on climate.

583 In the recent or the remote past, such as during the African Humid Period (from 14.500 to  
584 8.000 years BP) or during the Oligocene (30 Ma) when this region was under equatorial  
585 latitudes, the runoff was increased to much higher values that modelers have estimated. Based  
586 on these results and weathering models, we find comparable results for CO<sub>2</sub> consumption rate  
587 for both periods. The determined CO<sub>2</sub> consumption rate for the African Humid Period range

588 between  $0.91$  and  $1.5 \cdot 10^{12}$  mol/yr. This indicates that during humid periods the Ethiopian  
589 traps could have played a significant role on carbon cycle and on climate.

590 Overall, this study brings out the key role of a region that is understudied thus far in terms of  
591 weathering and climate paleo-variations. More generally, the influence on global climate of  
592 tropical basins influenced by monsoon would deserve to be assessed more precisely, in  
593 particular those of basaltic lithology.

594

595

596

### 597 **Acknowledgments**

598

599 This work was supported by BQR Geoazur, CNRS INSU-SYSTER grants and ANR  
600 INTOCC. We thank P. Telouk (LGLTPE) for his assistance with the ENS-Lyon Neptune  
601 *plus*. We thank also P.H Blard and B. Marty (CRPG) for their help on the field and fruitful  
602 discussions.

603

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605

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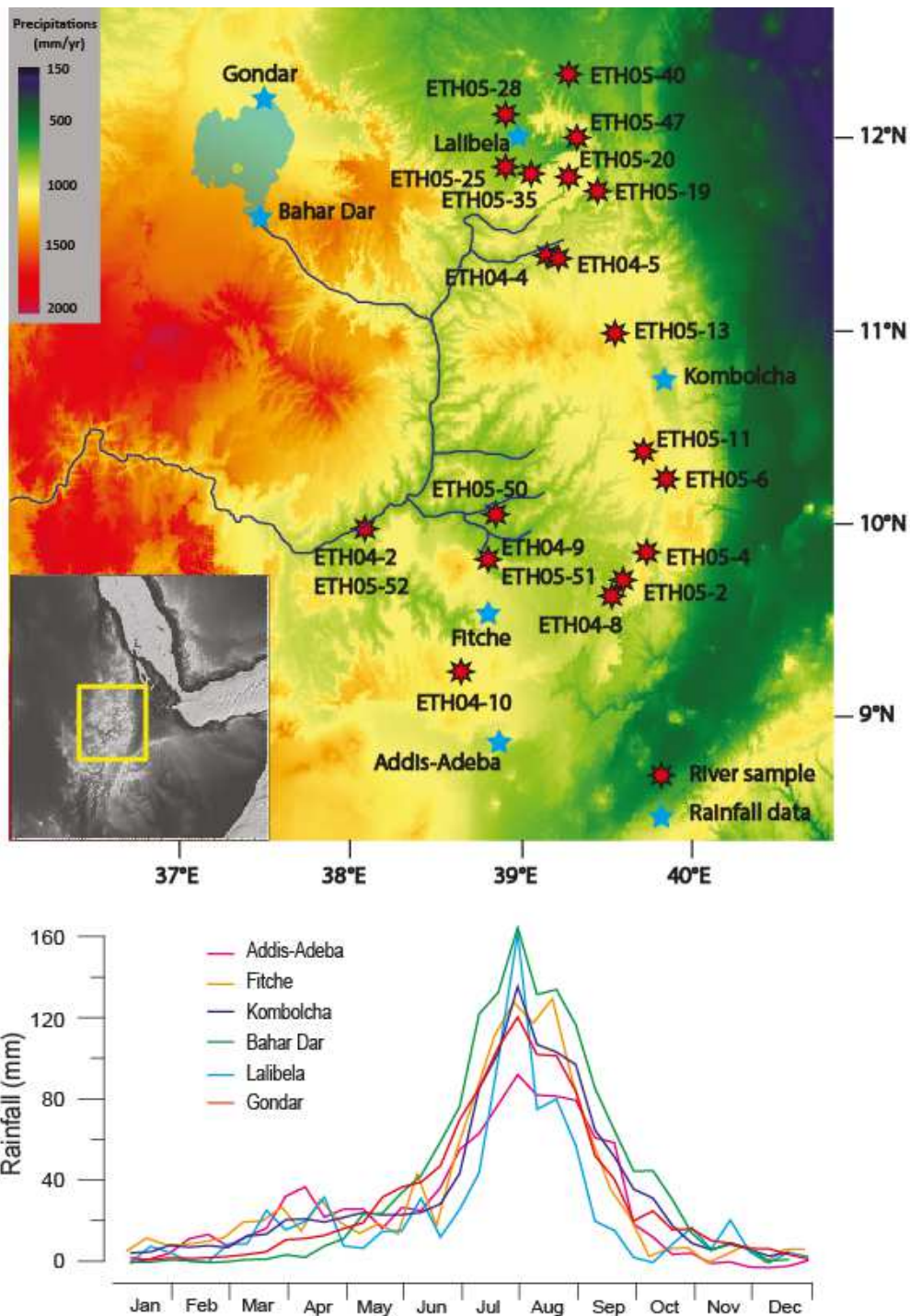
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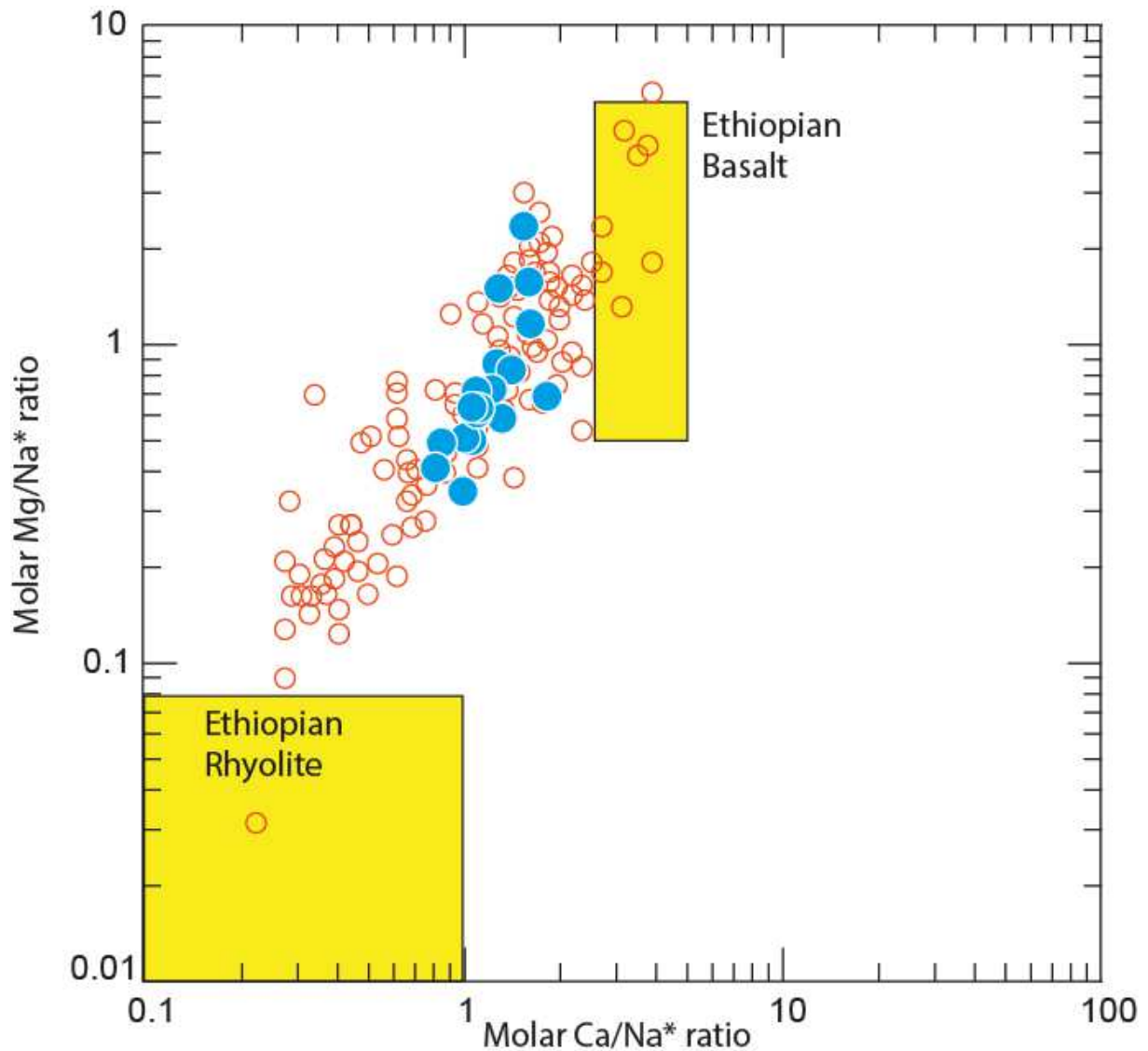
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938 Fig 1: A. Map showing samples localization and the studied area. Sample name on the map are: 1:  
 939 Jara Wenz (ETH05-11); 2:Borkena Wenz (ETH05-13) ; 3: Robi Wenz (ETH05-4) ; 4: Meri Wenz  
 940 (ETH05-28) ; 5: Tikur Wenz (ETH05-19) ; 6: Ketchin Abeba Wenz (ETH05-25) ; 7: Gereb Ara Wenz  
 941 (ETH05-40) ; 8: Tekeze (ETH05-35) ; 9: Hormat Wenz (ETH05-47) ; 10: Ala-Wiha (ETH05-20) ; 11:  
 942 Nazero Wenz (ETH05-6) .B. Measured rainfall (mm/week) temporal variations in various places of  
 943 the Ethiopian Traps. Reported values corresponds to the average of rainfall between 1961 and 1997  
 944 for Gondar, Bahar Dar, Kombolcha and Addis Ababa, between 1991 and 1997 for Fiche and between  
 945 1994 and 1997 for Lalibellathe.

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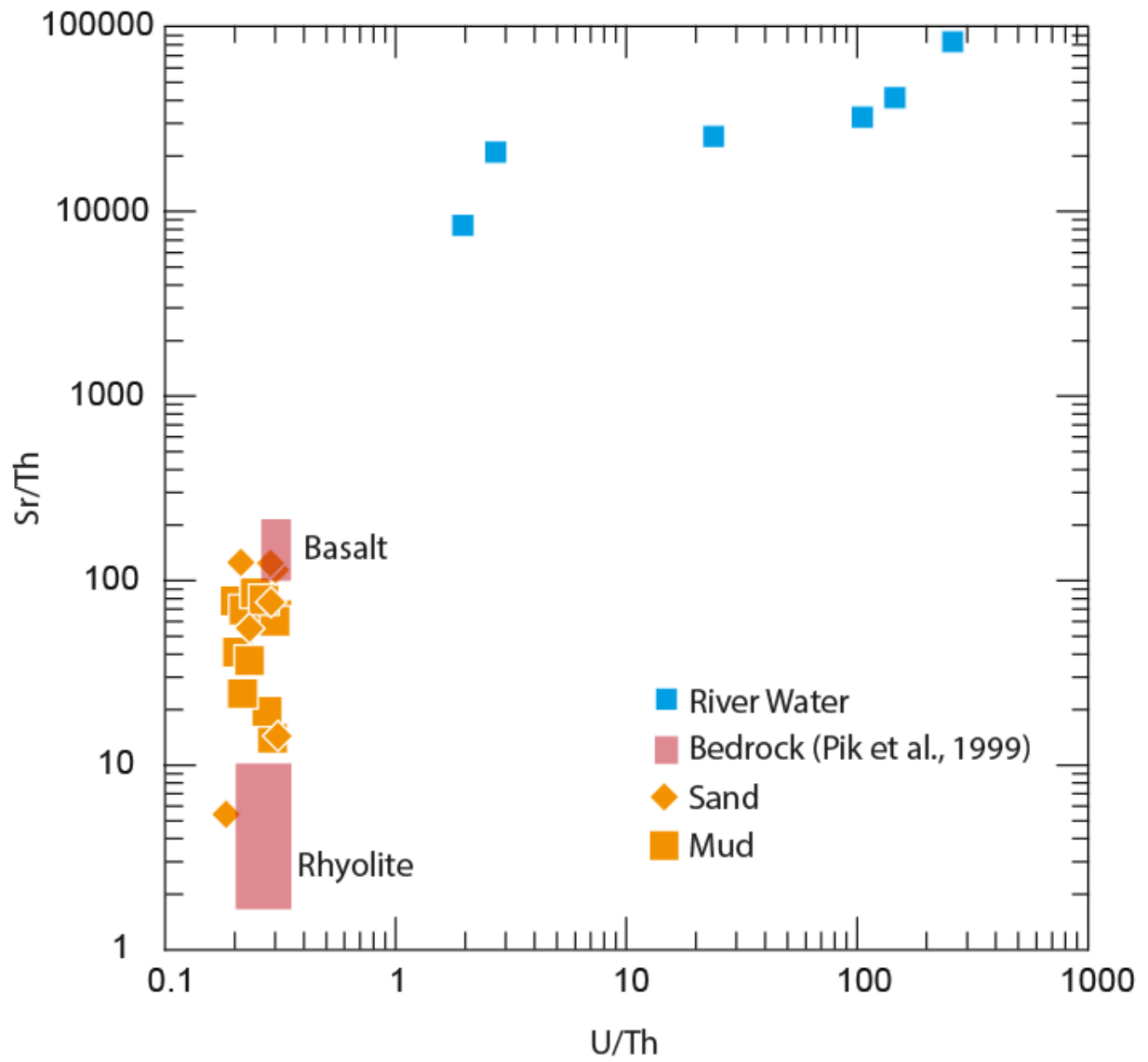


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949 *Fig 2: Mg/Na molar ratio as a function of Ca/Na molar ratio, for river dissolved phases. Both ratios*  
950 *are corrected from atmospheric inputs as described in section 4.2.1. The blue circles are for samples*  
951 *analyzed in this study. Orange circles are for other basaltic regions reported in Dessert et al., (2003).*  
952 *Bedrock data come from (Pik et al., 1999). (Benedetti et al., 1994; Dessert et al., 2001;*  
953 *Gaillardet et al., 1999; Gislason et al., 1996; Leemans and Cramer, 1991; Louvat, 1997;*  
954 *Pascale Louvat and Allègre, 1998; Louvat and Allègre, 1997; Meybeck, 1986; Nègre et al.,*  
955 *1993; Nègre and Deschamps, 1996; Riotte et al., 2003)*

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957



958 *Fig 3: Sr/Th as a function of U/Th (wt ratios) in the dissolved phases and solid phases of the sampled*  
 959 *ivers. Bedrock values come from Pik et al., 1999.*

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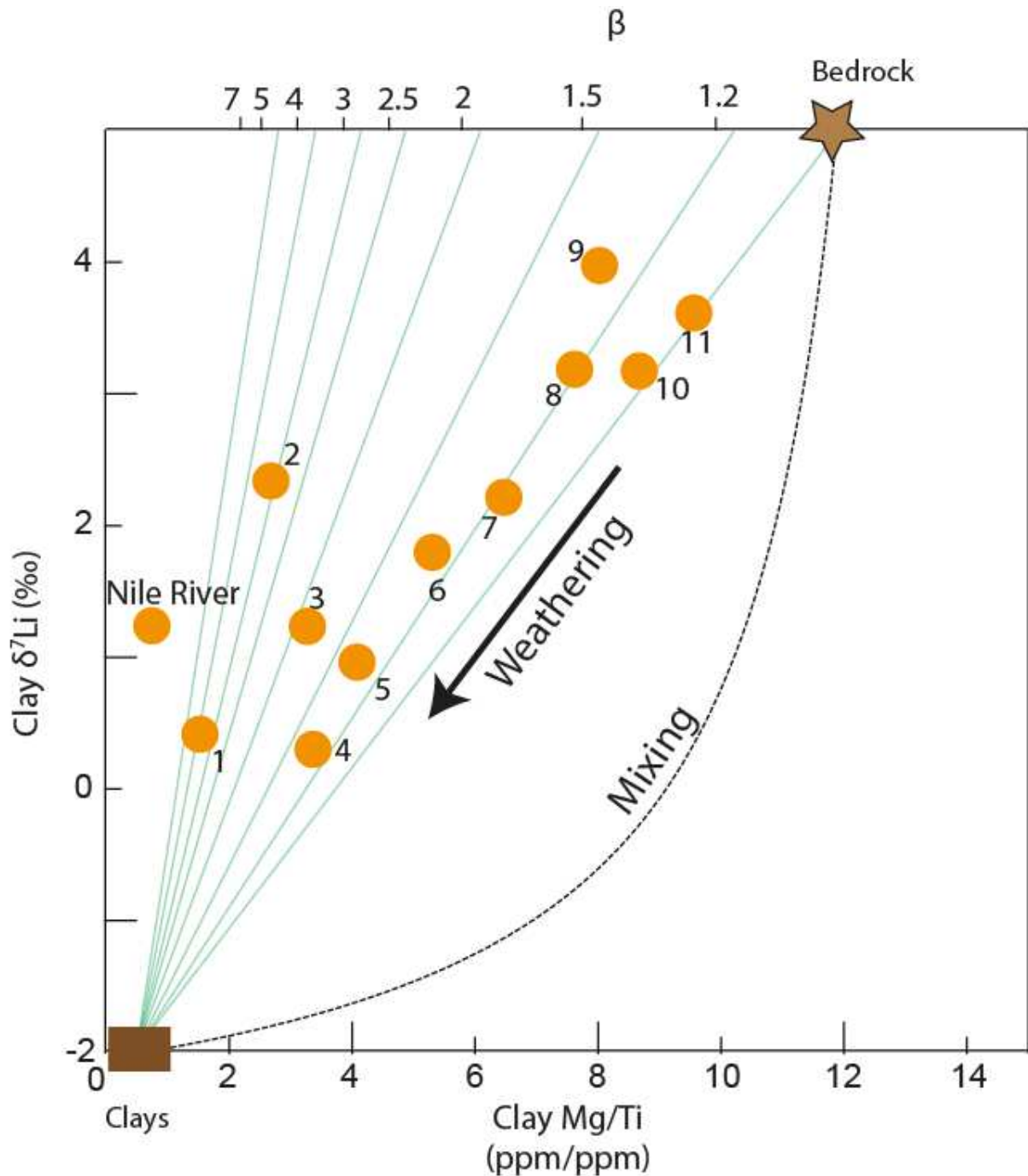
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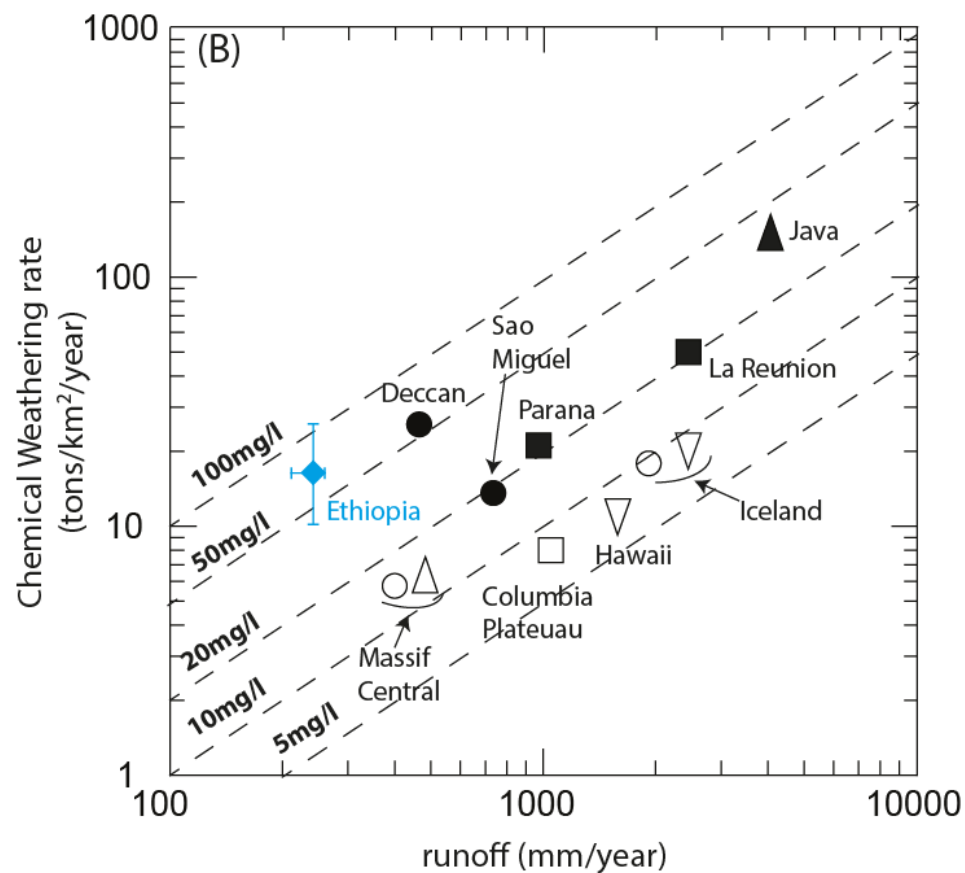
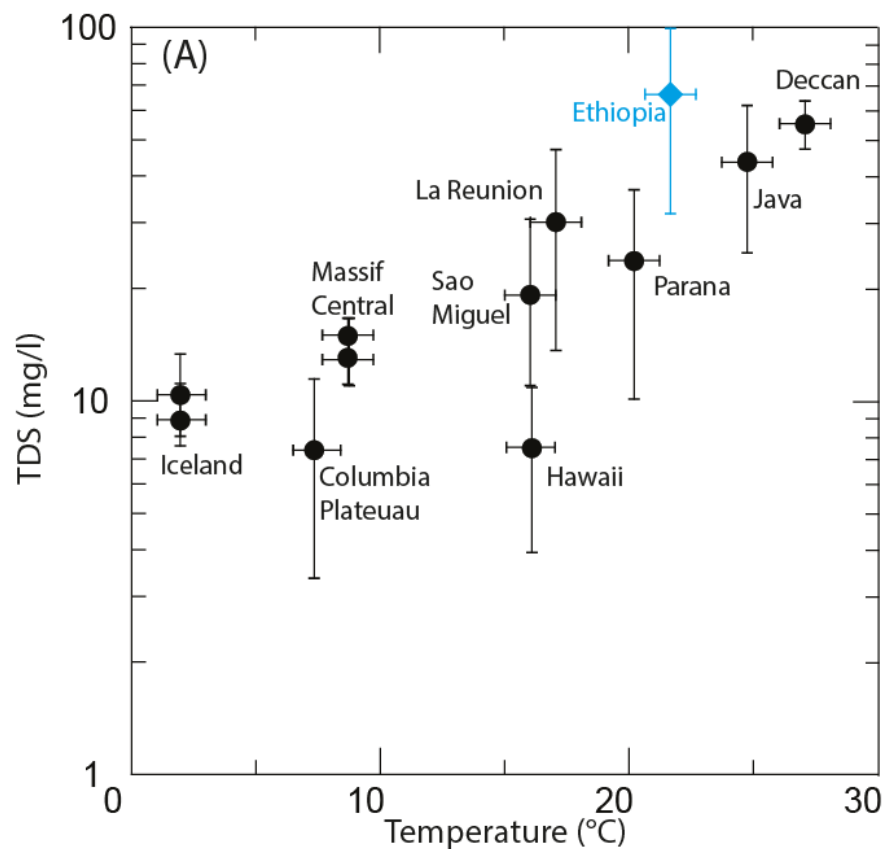
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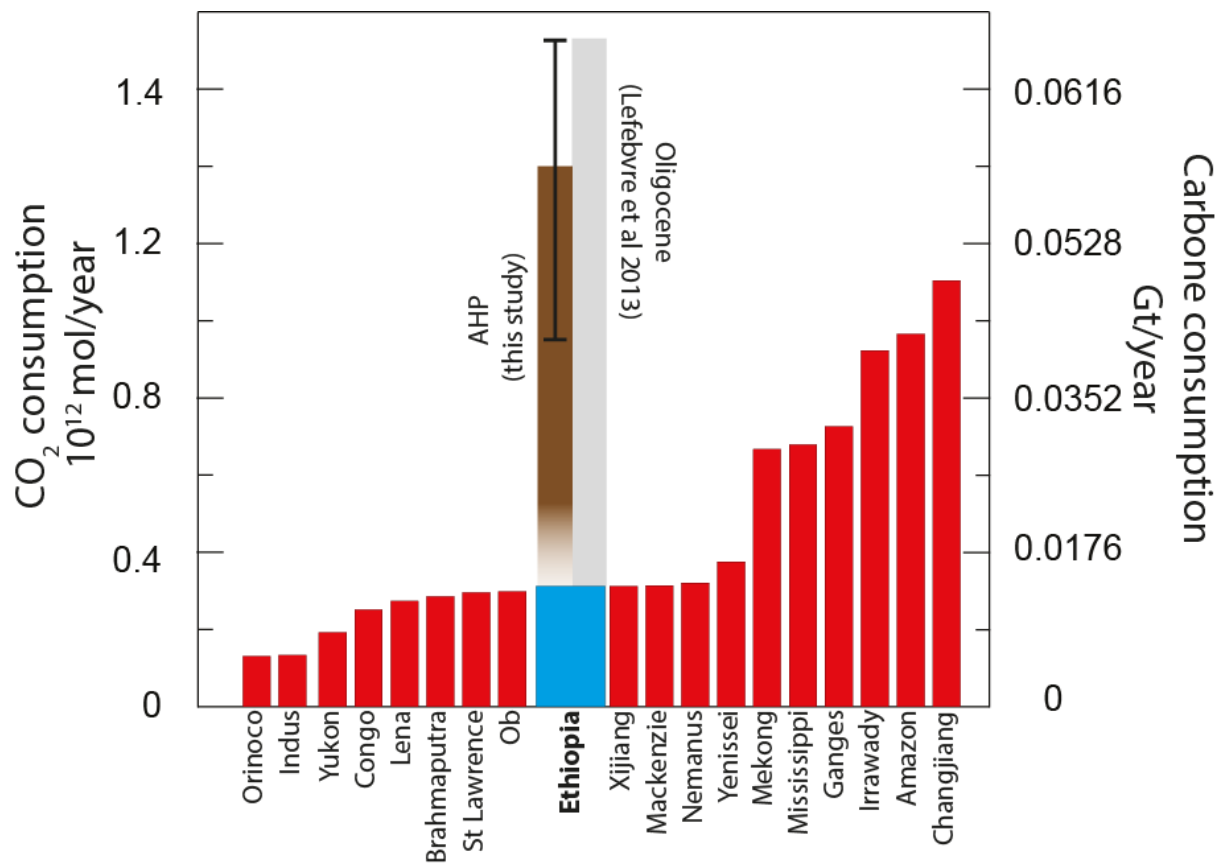
965  
 966 Fig 4: Clay  $\delta^7\text{Li}$  as a function of clay Mg/Ti for the main Blue Nile Basin rivers (blue  
 967 diamonds). Green lines correspond to the result of the model described in section 5.1. for  
 968 different  $\beta$  values. The dotted line corresponds to the theoretical mixing curve between  
 969 bedrock and isotopically fractionated clay endmembers. Mixing model conditions: Bedrock  
 970  $\delta^7\text{Li} = 5\text{‰}$ ,  $[\text{Li}] = 10 \text{ ppm}$  (this study); clays  $\delta^7\text{Li} = -2.3\text{‰}$  (clay  $\delta^7\text{Li}$  measured in the most  
 971 recent Nile delta fan sediments,  $[\text{Li}] = 100 \text{ ppm}$  (Bastian, 2017; Bastian et al., 2017)).  
 972 River numbers are the same as in figure 1.





973

974 Fig 5: (A) TDS\* (Total Dissolved Solids corrected for atmospheric inputs) as a function of temperature. The blue diamond corresponds to the  
 975 Ethiopian Traps (average based on 17 data, see Table 3), and the black circles come from the compilation of (Dessert et al., 2003) (Benedetti et  
 976 al., 1994; Dessert et al., 2001; Gaillardet et al., 1999; Gislason et al., 1996; Leemans and Cramer, 1991; Louvat, 1997; Pascale Louvat and  
 977 Allègre, 1998; Louvat and Allègre, 1997; Meybeck, 1986; Nègre et al., 1993; Nègre and Deschamps, 1996; Riotte et al., 2003) and  
 978 corresponds to most other basaltic areas studied thus far. (B) Plot of chemical erosion rate as a function of runoff (average based on 17 data,  
 979 see Table 3). The blue diamond reflects the Ethiopian Traps (this study) and the other symbols come from (Dessert et al., 2003) and display other  
 980 basaltic basins with active zones (in black) and inactive zone (in white). The dashed lines correspond to the TDS\* concentrations.



981

982 *Fig 6: Total CO<sub>2</sub> consumption in mol/year (Gaillardet et al., 1999) and annual CO<sub>2</sub>*  
 983 *consumption in GtC/yr due to silicate weathering for the largest rivers system. The blue bar*  
 984 *corresponds to the current CO<sub>2</sub> consumption rate estimated in Ethiopia based on our*  
 985 *measurements in river waters. The brown bar indicate the CO<sub>2</sub> consumption rate estimated*  
 986 *for the African Humid Period (AHP, see text for more details, and the grey one is for the*  
 987 *Oligocene and was published by (Lefebvre et al., 2013).*

988

989 *Table 1: Characteristics of the studied watersheds.*

990

991 *Table 2: Chemical composition of river waters, Total Dissolved Solid (TDS), silicate*  
992 *weathering rate and related CO<sub>2</sub> consumption rate for the major basins of the Ethiopian*  
993 *Traps.*

994 *Table 3: Chemical composition of sand, mud and clay samples (clay fractions have been*  
995 *physically separated following the procedure described in part 3.3.2)*

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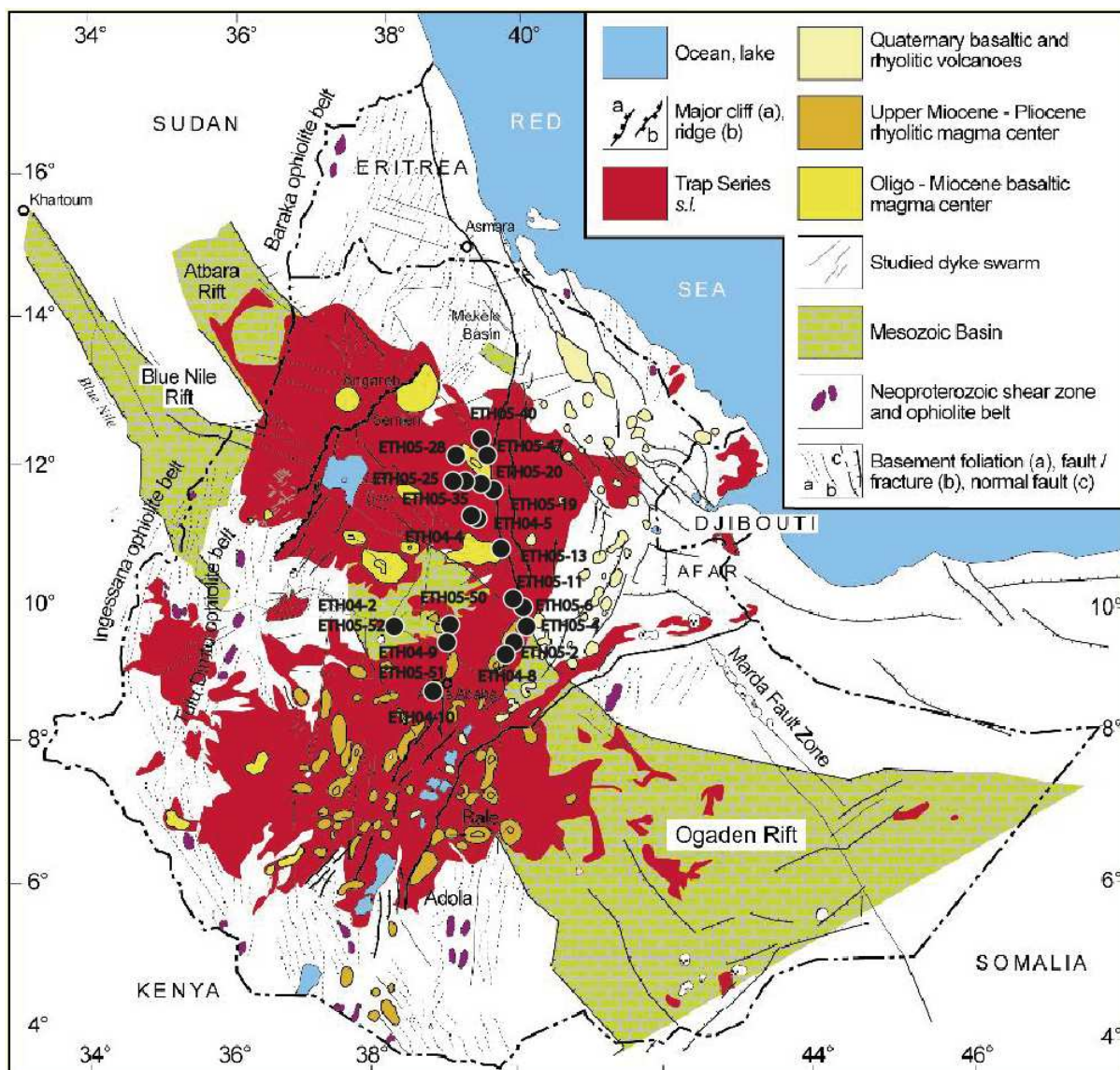
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1004 Fig S1: Ethiopian geological map. The black circles indicate our sampling sites.  
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1006 ([https://esdac.jrc.ec.europa.eu/images/Eudasm/Africa/images/maps/download/afr\\_etgm.jpg](https://esdac.jrc.ec.europa.eu/images/Eudasm/Africa/images/maps/download/afr_etgm.jpg))

1007

*Table 1: Characteristics of the studied watersheds.*

Sample number	Sample name	Elevation (m)	Mean elevation (m)	Area (km <sup>2</sup> )	Annual discharge (m <sup>3</sup> /year)	calculated runoff (mm/an)
ETH04-2	Nil (Juin-04)	1158	2277	64214	1.8E+10	273
ETH 05 -52	Nil (Mars-05)	1158	2277	64214	1.8E+10	273
ETH04-9	Jemma Juin-04)	1290	2516	5878	1.4E+09	232
ETH05-51	Jemma (Mars-05)	1290	2516	5878	1.4E+09	232
ETH 05 -50	Wench't Wenz	1384	2520	2853	7.2E+08	252
ETH 05 -13	Borkena Wenz	1948	2658	24	6.5E+06	268
ETH 05 -19	Tikur Wiha (Woldia)	1729	2500	171	3.9E+07	227
ETH04-4	Wegel Tena	1651	2625	945	2.2E+08	234
ETH04-5	Bashillo	1649	2719	2668	7.1E+08	266
ETH05-40	Gereb Ara Wenz	1496	2243	58	1.1E+07	190
ETH05-20	Ala Wiha (Godg Wabit)	2028	2874	75	1.6E+07	218
ETH05-47	Hormat Wenz	1523	2346	134	2.7E+07	200
ETH 05 -25	Ketchin Abeba Wenz	1998	2848	220	4.8E+07	220
ETH 05 -28	Meri Wenz	2005	2591	131	2.6E+07	198
ETH 05 -35	Tekeze amont	2133	3020	493	1.1E+08	219
ETH04-10	plateau-N-Addis	2486	2697	56	1.5E+07	265
ETH04-8	Plateau E-Addis	2880	3267	32	8.5E+06	267
ETH 05 -2	Ulet Wenz	3034	3250	4	1.1E+06	270
ETH 05 -4	Robi Wenz	1389	2071	219	5.5E+07	252
ETH 05 -11	Jara Wenz	1656	2575	77	2.0E+07	265
ETH 05 -6	Nazero Wenz	1422	1879	171	4.4E+07	259

Table 2: Chemical composition of river waters, Total Dissolved Solid (TDS), silicate weathering rate and CO<sub>2</sub> consumption rate for the major basins of the Ethiopian Traps

Sample number	Sample name	Si (μmol/l)	Mg (μmol/l)	Ca (μmol/l)	Na (μmol/l)	K (μmol/l)	Al (μmol/l)	Sr (μg/l)	Th (μg/l)	U (μg/l)	F <sup>-</sup> (μmol/l)	Cl <sup>-</sup> (μmol/l)	NO <sub>3</sub> <sup>-</sup> (μmol/l)	SO <sub>4</sub> <sup>2-</sup> (μmol/l)	Mg <sup>+</sup> (μmol/l)	Ca <sup>+</sup> (μmol/l)	Na <sup>+</sup> (μmol/l)	K <sup>+</sup> (μmol/l)	F <sup>+</sup> (μmol/l)	NO <sub>3</sub> <sup>+</sup> (μmol/l)	SO <sub>4</sub> <sup>2+</sup> (μmol/l)	HCO <sub>3</sub> <sup>+</sup> (μmol/l)	TDS* (μmol/l)	TDS* (mg/l)	Total chemical weathering (tons/km <sup>2</sup> /year)	CO <sub>2</sub> (10 <sup>6</sup> mol/km <sup>2</sup> /year)
ETH04-2	Nil (Juin-04)	142	304	530	494	70	3405	135	0.005	0.1	8.5	76	43	146	296	529	427	68	8.5	43	146	1808	1320	40.9	11.2	0.49
ETH 05 -52	Nil (Mars-05)	154	341	584	414	61	398	174	< L.D.	0.2	8.9	70	4	171	334	582	352	60	8.9	4	171	1897	1328	41.9	11.4	0.52
ETH04-9	Jemma Juin-04)	332	323	688	650	141	3946	162	0.011	0.6	12.2	116	100	252	311	686	548	139	12.2	100	252	2075	1683	53.1	12.4	0.48
ETH05-51	Jemma (Mars-05)	660	684	1173	959	98	93	371	< L.D.	0.5	17.7	127	14	269	671	1171	847	96	17.7	14	269	4070	2785	86.5	20.1	0.95
ETH 05 -50	Wench't Wenz	631	585	1035	1219	92	264	257	< L.D.	0.6	14.2	165	8	402	568	1032	1073	89	14.2	8	402	3551	2762	83.3	21.0	0.90
ETH 05 -13	Borkena Wenz	465	499	1334	734	59	26	265	< L.D.	0.3	8.5	345	292	125	464	1327	429	52	8.5	292	125	3546	2273	76.4	20.5	0.95
ETH 05 -19	Tikur Wiha (Woldia)	667	789	1110	688	70	198	237	< L.D.	0.9	9.7	138	115	98	775	1107	566	68	9.7	115	98	4089	2516	78.9	17.9	0.93
ETH04-4	Wegel Tena	639	775	1141	901	65	3625	236	0.006	0.8	9.1	273	77	272	748	1135	659	60	9.1	77	272	3880	2602	81.2	19.0	0.91
ETH04-5	Bashillo	498	663	1022	940	60	4823	227	0.007	0.7	10.1	164	16	182	647	1019	795	57	10.1	16	182	3809	2518	77.1	20.5	1.02
ETH 05 -25	Ketchin Abeba Wenz	526	1565	1357	1055	71	62	488	< L.D.	0.9	8.8	348	132	212	1530	1350	748	65	8.8	132	212	6039	3692	111.0	24.4	1.33
ETH 05 -28	Meri Wenz	473	1365	1386	869	69	458	672	< L.D.	0.5	9.2	184	11	880	1346	1382	706	65	9.2	11	880	4466	3500	106.9	21.2	0.89
ETH 05 -35	Tekeze amont	453	1768	1170	761	48	101	360	< L.D.	1.2	7.3	298	129	205	1739	1164	497	43	7.3	129	205	5826	3442	102.0	22.4	1.28
ETH04-10	plateau-N-Addis	197	169	304	273	56	443	115	< L.D.	0.0	7.7	52	154	22	164	303	228	55	7.7	154	22	1014	750	23.5	6.2	0.27
ETH04-8	Plateau E-Addis	153	88	160	147	51	3359	53	0.006	0.0	6.4	54	21	15	83	159	99	50	6.4	21	15	582	391	12.6	3.4	0.16
ETH 05 -2	Ulet Wenz	199	59	133	102	20	2225	44	0.002	0.0	5.1	15	26	27	57	133	89	20	5.1	26	27	406	299	9.5	2.6	0.11
ETH 05 -4	Robi Wenz	539	250	721	734	73	307	114	< L.D.	0.4	12.2	56	59	70	244	720	684	72	12.2	59	70	2478	1720	53.3	13.4	0.62
ETH 05 -6	Nazero Wenz	877	581	1161	1155	100	68	174	0.002	0.5	21.3	168	8	90	565	1158	1007	97	21.3	8	90	4354	2826	87.1	22.5	1.13

