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Rare earth element and neodymium isotope tracing of sedimentary rock weathering

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Abstract

1 Chemical weathering plays an important role in sequestering atmospheric CO₂, but its
2 potential influence on global climate over geological timescales remains debated. To some
3 extent, this uncertainty arises from the difficulty in separating the respective contribution of
4 sedimentary and crystalline silicate rocks to past weathering rates in the geological record;
5 two types of rocks having presumably different impact on the long-term carbon cycle. In this
6 study, we investigate the use of rare earth element (REE) and neodymium isotopes (ϵ_{Nd}) in
7 leached iron oxide fractions of river sediments for tracing the origin of weathered rocks on
8 continents. A new index, called ‘concavity index’ (CI), is defined for measuring the degree of
9 mid-REE enrichment in geological samples, which enables the determination of the source of
10 iron oxides in sediments, such as seawater-derived Fe-oxyhydroxide phases, ancient marine
11 Fe oxides derived from the erosion of sedimentary rocks, and recent secondary oxides formed
12 in soils via alteration of crystalline silicate rocks or pyrite oxidation. Using this index, we
13 demonstrate that the ϵ_{Nd} difference between paired Fe-oxide and detrital fractions in river
14 sediments (defined here as $\Delta\epsilon_{\text{Nd Feox-Det}}$) directly reflects the relative contribution of
15 sedimentary *versus* crystalline silicate rocks during weathering. While rivers draining old
16 cratons and volcanic provinces display near-zero $\Delta\epsilon_{\text{Nd Feox-Det}}$ values indicative of dominant
17 silicate weathering (0.5 ± 1.1 ; n=30), multi-lithological catchments hosting sedimentary
18 formations yield systematically higher values (2.7 ± 1.2 ; n=44), showing that sedimentary
19 rock weathering can be traced by the occurrence of riverine Fe oxides having more radiogenic
20 Nd isotope signatures compared to detrital fractions. This assumption is reinforced by the
21 evidence that calculated $\Delta\epsilon_{\text{Nd Feox-Det}}$ values agree well with previous estimates for carbonate
22 and silicate weathering rates in large river basins.

23 Examining the influence of climate and tectonics on measured Nd isotopic compositions,
24 we find that $\Delta\epsilon_{\text{Nd Feox-Det}}$ is strongly dependent on temperature in lowlands, following an
25 Arrhenius-like relationship that reflects enhanced alteration of silicate rocks and formation of
26 secondary Fe oxides in warmer climates. In contrast, in high-elevation catchments, $\Delta\epsilon_{\text{Nd Feox-}}$
27 Det defines striking correlation with maximum basin elevation, which we also interpret as
28 reflecting the intensification of silicate weathering and associated Fe oxide formation as
29 elevation decreases, due to the combined effects of thicker soils and warmer temperature.

30 Overall, our new findings are consistent with previous assertions that the alteration of
31 sedimentary rocks prevails in high-elevation environments, while silicate weathering
32 dominates in floodplains. This novel approach combining REE and Nd isotopes opens new
33 perspectives for disentangling the weathering signals of sedimentary and crystalline silicate
34 rocks in the geologic record, which could be used in future studies to reassess the causal
35 relationships between mountain uplift, erosion and climate throughout Earth's history.

36

37 **Keywords:** World rivers; iron oxides; neodymium isotopes; Mid-REE enrichment; concavity
38 index; silicate weathering; carbonate weathering; sulphide weathering

39

40 **1. Introduction**

41 On the continents, rocks are dissolved through chemical weathering, a process that
42 consumes substantial amounts of carbon dioxide and plays a major role in regulating Earth's
43 climate through time (Walker et al., 1981; Berner et al., 1983; Kump et al., 2000). A
44 contentious view is that mountain building events during Earth's history were followed by
45 enhanced physical erosion and accelerated dissolution of crystalline silicate rocks, resulting in
46 a net drawdown of atmospheric CO₂ and global climate cooling (e.g. Raymo and Ruddiman,
47 1992; Li and Elderfield, 2013). A first difficulty with this hypothesis has been the relatively
48 small size of the atmospheric CO₂ reservoir, meaning that over millions of years, any
49 significant increase in CO₂ removal via silicate weathering would have led to an unrealistic
50 near complete depletion of carbon in the atmosphere (Berner and Caldeira, 1997).
51 Additionally, several studies have also argued for relatively invariant silicate weathering rates
52 at times when erosion of mountains was intensifying, suggesting that the evolution of
53 atmospheric CO₂ in response to mountain uplift could reflect progressive changes in the
54 susceptibility of crystalline rocks to weathering, rather than being simply proportional to
55 silicate weathering fluxes on continents (Willenbring and von Blanckenburg, 2010; Caves et
56 al., 2016; Caves Rügenstein et al., 2019).

57

58 When examining the nature of the relationships between mountain uplift, chemical
59 weathering and global climate, much complexity is brought by the fact that regions of high
60 elevation are frequently composed of continental margin sequences of old sediments rather
61 than crystalline basement (e.g. Hartmann and Moosdorf, 2012). These sequences invariably
62 contain marine carbonates and other marine authigenic phases, which weather more rapidly
63 than crystalline rocks. While mountainous regions with high denudation rates display the
64 highest silicate weathering rates on continents (e.g. West et al., 2005), field observations

65 indicate that chemical weathering fluxes from tectonically active mountain ranges are mostly
66 dominated by dissolved inputs from carbonate weathering, even in watersheds hosting
67 primarily igneous or metamorphic rocks (e.g. Stallard and Edmond, 1983; Jacobson and
68 Blum, 2003; Jacobson et al., 2003; Moore et al., 2013). Carbonate weathering is generally
69 considered to have no impact on the long-term carbon cycle. This is because the amount of
70 carbon initially consumed during the dissolution of carbonate minerals by carbonic acid is
71 quantitatively returned to the atmosphere upon subsequent carbonate precipitation in the
72 ocean. However, carbonate dissolution on continents can also proceed in presence of
73 sulphuric acid released by the oxidation of pyrite and other sulphide minerals (e.g. Spence and
74 Telmer, 2005; Calmels et al., 2007; Torres et al., 2016; Torres et al., 2017). This coupled
75 sulphide-carbonate weathering process can be particularly active in catchments hosting shales
76 and other sulphide-bearing sedimentary formations, resulting, in terms of global impact on the
77 carbon cycle, in a net release of CO₂ to the atmosphere (e.g. Stallard and Edmond, 1983;
78 Calmels et al., 2007; Torres et al., 2016; Blattmann et al., 2019). Similarly, the oxidation of
79 petrogenic organic carbon stored in sedimentary rocks also returns CO₂ to the atmosphere
80 (Bouchez et al., 2010), especially in high mountains and glacial environments (Torres et al.,
81 2016; Horan et al., 2017). On top of the current debate about the presumed importance of
82 silicate weathering to the long-term carbon cycle, the above-mentioned evidence that
83 alteration of carbonates and other sedimentary rocks dominate in high mountain environments
84 have re-emphasized the need to characterize the separate evolution of sedimentary and
85 crystalline silicate rock weathering through geologic time. To date, however, there is no
86 reliable proxy that can decipher different rock weathering signatures from the geological
87 record.
88

89 Continental detrital sediments may contain various authigenic mineral phases depending on
90 the original source rock, including iron (Fe) oxides. Authigenic Fe oxide phases precipitate
91 from water and pore fluids in the oceans, lakes and river estuaries, concentrating trace
92 elements such as neodymium and beryllium (e.g. Boyle et al., 1977), which can be used to
93 help identify the origin of weathered rocks on continents (e.g. von Blanckenburg et al., 2015;
94 Hindshaw et al., 2018; Sufke et al., 2019; Jang et al., 2020). Iron oxides are also ubiquitous
95 weathering products in soils, which can be transported as particulates in rivers prior to
96 deposition as sediments (Bayon et al., 2004; Kraft et al., 2013). Recently, a study conducted
97 in the Svalbard archipelago showed that the Fe oxide fractions leached from fjord sediments
98 surrounding glaciated catchments composed of sedimentary rocks typically displayed Nd
99 isotope signatures ($^{143}\text{Nd}/^{144}\text{Nd}$; or ϵ_{Nd} in epsilon notation) more radiogenic than
100 corresponding detrital fractions (Jang et al., 2020). In agreement, Hindshaw et al. (2018)
101 reported similar differences in ϵ_{Nd} between both the easily dissolvable component of
102 catchment rocks and their silicate residues, and dissolved and particulate riverine loads in
103 Svalbard. These data were interpreted as reflecting the preferential dissolution of marine
104 precipitates hosted by sedimentary rocks. In fact, the same suggestion had been already
105 presented more than two decades ago (Goldstein and Jacobsen, 1987), in order to account for
106 the observation that the Nd isotopic composition of filtered river waters was also more
107 radiogenic than corresponding suspended sediment loads in certain rivers. In the above
108 mentioned examples, the evidence that significant Nd isotope decoupling can occur in
109 sedimentary watersheds due to preferential alteration of marine precipitates is consistent with
110 the fact that the global Nd isotopic composition of seawater ($\epsilon_{\text{Nd}} -8.8$; Lacan et al., 2012) is on
111 average more radiogenic than the mean continental detrital input to the ocean (-11.3 ± 2.5 ;
112 Bayon et al., 2015). Sedimentary carbonate rocks dissolve indeed at a much faster rate than
113 crystalline silicate rocks, and therefore largely control the chemistry of river waters

114 worldwide and associated dissolved inputs to the ocean (Gaillardet et al., 1999; Jacobson et
115 al., 2002).

116
117 In this study, we define a novel approach based on the combined use of rare earth elements
118 (REE) and Nd isotopes in the Fe oxide fraction associated with river sediments. Our
119 approach exploits a systematic difference in Nd isotopic ratios between the Fe oxide and
120 silicate fractions of the sediment (defined here as $\Delta\epsilon_{Nd_{FeOx-Det}}$), similar to that observed
121 between filtered waters and suspended particulate matter (Goldstein and Jacobsen, 1987;
122 Hindshaw et al., 2018). This traces the relative contribution of sedimentary rock inputs with
123 an inherited old authigenic Nd isotopic signature *versus* silicate rock inputs with a
124 composition set by their silicate source. Our data set has a global reach, spanning samples
125 from different lithologies, topographies and weathering regimes. The relative amounts of
126 sedimentary- *versus* crystalline silicate- derived sources remaining in the riverine Fe oxide
127 fraction are found to display systematic trends with basin parameters, such as temperature and
128 relief that we link to different regimes of weathering, with fundamental implications for the
129 silicate weathering feedback.

130

131 **2. Materials and methods**

132 **2.1. World river sediments: Climatic, physical and lithological characteristics of** 133 **watersheds**

134 We analysed a total of 74 sediment samples from either rivers (n=43), deltas (n=7),
135 estuaries (tide-dominated environments; n=15), or submarine deltas (wave-dominated
136 environments; n=4). Additionally, five marine sediment samples (Congo, Niger, Nile, Sepik,
137 Gaoping) collected from more distal locations near the mouth of rivers, in deep-sea fan
138 environments (Fig 1; Table S1). The geographical coordinates of studied samples are listed in

139 Table S1. Selected samples include some of the world's largest rivers (e.g. Amazon, Congo,
140 Mississippi, Nile, Yangtze, Mackenzie, Ganges-Brahmaputra), in addition to other smaller
141 rivers associated with particular geological settings (Fig. 1). All river sediments were scooped
142 from recent river banks or correspond to suspended particulates for the case of the Congo
143 River tributaries. Other sediment samples were collected using various coring devices,
144 corresponding typically to centimetre-thick layers of core-tops or sub-surface sediment
145 horizons. All studied samples correspond either to modern or relatively recent sediments that
146 have been presumably deposited during the last few centuries.

147

148 Mean annual air temperature (MAT) for each river system were derived from the literature
149 (e.g. Pinet and Souriau, 1988; Bayon et al., 2016, 2018). Corresponding information for
150 maximum elevation in river basins was either derived from Milliman and Farnsworth (2011),
151 or determined in the geographical information system (GIS) software ArcGis (ESRI 2001,
152 ArcGis Desktop 10.3.1) using the hydrological data and maps based on shuttle elevation
153 derivatives HydroBASINS (Lehner and Grill, 2013). Finally, information on the lithological
154 composition of each river basin was extracted from the GLiM global lithological map
155 database (Table S1; Hartmann and Moosdorf, 2012).

156

157 Studied samples were classified into distinct categories depending on the lithological
158 composition of corresponding basins. A series of samples (n=24) came from rivers draining
159 igneous/metamorphic terranes, in particular from the Precambrian cratons of Africa (West
160 African and Congo Shields), North America (Canadian Shield), northern South America
161 (Guiana Shield), Fennoscandia, West Australia and Northwest Ireland. Six samples were
162 derived from rivers draining both modern (New Zealand, Reunion Island, Indonesia) and
163 ancient (Northern Ireland) volcanic provinces. The rest of studied samples (n=44)

164 corresponded to rivers draining mixed and/or sedimentary rock formations. This latter
165 category of samples was sub-classified into 2 groups: rivers draining mountains (with
166 maximum elevation > 1000 m), including most of the world's largest rivers (n=30); and rivers
167 draining lowlands, such as large Russian rivers (Volga, Don, Northern Dvina) and other small
168 rivers from e.g. France and the British Isles (n=14).

169

170 **2.2. Chemical preparation and sequential leaching procedure**

171 Sieved (<63 μm) bulk sediment samples were first treated with 5% (v/v) acetic acid (AA)
172 for removal of carbonate minerals. Easily reducible Fe-oxide fractions were then extracted
173 using a mixed solution of 15% (v/v) AA and 0.05M hydroxylamine hydrochloride (HH),
174 following a procedure adapted from Bayon et al. (2002) and Gutjahr et al. (2007). With this
175 latter leaching step, we specifically targeted the extraction of easily reducible and poorly
176 crystalline Fe-oxide phases, such as ferrihydrite and lepidocrocite (Poulton and Canfield,
177 2005). The mineral fractions dissolved during this AA-HH leaching step are referred to as
178 'HH leachates' in the discussion below. Finally, after subsequent removal of organic
179 compounds using 5% hydrogen peroxide (H_2O_2) solutions, silt-size (~4-63 μm) detrital
180 fractions were separated from the residual detritus by centrifugation, prior to digestion by
181 alkaline fusion (Bayon et al., 2015).

182

183 **2.3. Geochemical analyses**

184 Major (Mg, Al, Ca, Mn and Fe) and trace element (REE, Y, Sr) concentrations in the HH
185 leachates were determined with a Thermo Scientific Element XR sector field ICP-MS, using
186 the Tm addition method (Barrat et al., 1996). Isobaric interferences on measured REE signals
187 were corrected using oxide formation rates determined by analysing mono-elemental
188 solutions of Ba-Ce, Nd-Pr and Sm-Eu-Gd-Tb. The in-run uncertainties on all measurements

189 were generally better than 5%. The accuracy of our data was assessed by analysing seven
190 different geological certified reference materials (AGV-1; AN-G; BCR-1; DR-N; UB-N; WS-
191 E; and IF-G, i.e. an iron-ore deposit). All results were in full agreement with reference values
192 from the literature (<13%; Table S2). Neodymium isotopic measurements were performed at
193 the Pôle Spectrométrie Océan (Brest) using a Thermo Scientific Neptune multi-collector
194 ICPMS, after Nd purification by conventional ion chromatography. Mass bias corrections on
195 Nd were made with the exponential law, using $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, and $^{143}\text{Nd}/^{144}\text{Nd}$
196 corrected values were normalized to a JNdi-1 value of $^{143}\text{Nd}/^{144}\text{Nd} = 0.512115$. Repeated
197 analyses of a JNdi-1 standard solution during this study gave $^{143}\text{Nd}/^{144}\text{Nd}$ of $0.512121 \pm$
198 0.000009 (2 SD, n=33), hence corresponding to an external reproducibility of $\sim \pm 0.17\epsilon$ (2
199 SD). Epsilon Nd values (ϵ_{Nd}) were calculated using $^{143}\text{Nd}/^{144}\text{Nd} = 0.512630$ (Bouvier et al.,
200 2008).

201

202 **3. Results**

203 **3.1. Major and trace elements in the leachates**

204 For clarity, measured concentrations ($\mu\text{g/g}$) are reported relative to the initial mass of bulk
205 $<63\mu\text{m}$ sample prior to leaching (Table S3). The HH leachates are dominated by Fe and Ca,
206 except for two intensively weathered sediments from the Orinoco River Basin (i.e. Rio Caroni
207 and Caura), both characterized by comparatively slightly higher Al abundances reflecting
208 presumably the presence of gibbsite ($\text{Al}(\text{OH})_3$) in these samples. Measured Fe/Ca ratios vary
209 largely amongst studied leachates, from about 0.04 to 20 ($\mu\text{g}/\mu\text{g}$) (mean value of ~ 3). Note
210 that three sediment samples (Nile, Danube, Var) display even lower Fe/Ca ratios (< 0.02)
211 together with very high Ca concentrations (up to 20wt% of initial sample mass), which
212 indicate that they are largely dominated by carbonate material. Measured Fe/Mn ratios range
213 from ~ 1 to 230 (mean value of ~ 25), mostly reflecting the predominance of Fe- over Mn-

214 oxide phases in most samples. Rare earth elements in sediment leachates also display a large
215 range of concentrations (with Nd varying from ~ 0.2 to 28 $\mu\text{g/g}$; mean ~2.5 $\mu\text{g/g}$; Table S3)
216 and distribution patterns (Fig. 2), as normalized to the world river average silt values (WRAS;
217 Bayon et al., 2015). The average REE patterns for each river category differ significantly
218 from one group to another (Fig. 2). Note that shale-normalized REE patterns are also reported
219 as normalized to Gd in Fig. 2, using the WRAS value for Gd (5.188 $\mu\text{g/g}$), in order to remove
220 any dilution effect. The HH leachates from catchments dominated by igneous/metamorphic
221 rocks are relatively enriched in light- (LREE) relative to heavy-REE (HREE) (Fig. 2a), while
222 those from volcanic regions are characterized by marked LREE depletion (Fig 2b), both
223 displaying shale-normalized patterns that closely resemble those for corresponding detrital
224 clays (see Fig. 3 of Bayon et al., 2015). In contrast, the sediment leachates from multi-
225 lithological catchments hosting sedimentary rocks are characterized by more pronounced mid-
226 REE (MREE) enrichments relative to LREE and HREE (Fig. 2c,d).

227

228 **3.2. Neodymium isotopes**

229 Measured Nd isotopic compositions for the HH leachates are given in Table 1, together
230 with ϵ_{Nd} values for corresponding detrital silt-size fractions. Note that a total of 49 data for
231 silt-size fractions were already reported in Bayon et al. (2015). Detrital and Fe-oxide fractions
232 from studied river sediments encompass a large range of ϵ_{Nd} values from -28.7/-30.9
233 (Churchill) to +4.1/+4.4 (Brantas), respectively (Table 1; Fig. 3a). The Nd isotopic
234 composition of Fe-oxide phases is almost systematically more radiogenic than associated
235 detrital sediments, hence resulting in positive $\Delta\epsilon_{\text{Nd FeOx-Det}}$ values, except for 7 samples (i.e.
236 Kasai, Niger, Orinoco, Churchill, Kymijoki, Ume, Swilly) all derived from rivers draining old
237 crystalline basement. River sediments from these ancient igneous/metamorphic provinces and
238 volcanic regions display $\Delta\epsilon_{\text{Nd FeOx-Det}}$ close to 0 epsilon units (0.5 ± 1.1 ; $n=30$; Fig. 3b). In

239 contrast, for the majority of samples where the basin lithology contains sedimentary rocks, the
240 Nd isotopic composition (ϵ_{Nd}) of Fe oxide phases is systematically more radiogenic than
241 associated silicate residues, resulting in positive $\Delta\epsilon_{\text{Nd Feox-Det}}$ in both lowlands (2.1 ± 0.3 ;
242 $n=14$) and mountainous (2.4 ± 0.7 ; $n=30$) catchments. The observed differences in $\Delta\epsilon_{\text{Nd Feox-Det}}$
243 values are substantial and significant in magnitude; up to 6 epsilon units.

244

245 **4. Discussion**

246 **4.1. Significance of Nd isotope and REE compositions in leached sediment fractions**

247 *4.1.1 Identification of leached mineral phases*

248 An important prerequisite for interpreting measured REE and Nd isotopic compositions as
249 potential weathering signatures in riverine Fe-oxides is to identify the different mineral phases
250 dissolved during the chemical leaching. As mentioned above, the presence of relatively high
251 Ca contents in the HH leachates indicates that substantial dissolution of residual carbonates
252 (i.e. left behind the first AA extraction step) may have occurred. Using an endmember mixing
253 model based on Sr/Ca and Mg/Ca ratios (Bayon et al., 2007), the relative proportions of
254 aragonite, low-Mg calcite, high-Mg carbonates and silicate detritus in leached sediment
255 fractions can be quantitatively estimated (Fig. 4). Most samples from catchments hosting
256 igneous/metamorphic and volcanic rocks plot on a diffuse mixing trend between the silicate
257 and low-Mg calcite endmembers, indicating that the Ca budget in these leachates is mostly
258 controlled by detrital inputs (> 80%). In contrast, the HH leachates from multi-lithological
259 catchments generally extends towards the field of high-Mg carbonates characterized by low
260 Sr/Ca ratios (e.g. high-Mg calcite, dolomite, siderite). In leached sediment fractions from the
261 Ganges and the rivers draining the eastern Canadian shield (Nelson, Churchill), more than
262 60% of extracted Ca appear to be derived from dolomite (Fig. 4). By analogy, the presence of
263 high-Mg carbonates and partial dissolution of the silicate detritus could also account for some

264 of the Fe signal in the HH leachates. However, except for a few samples (e.g. Nile, Danube,
265 Mekong, Nelson, Churchill), most leachates display Fe/Mg and Fe/Al ratios that fall into the
266 general range of values for various iron oxide formations and marine ferromanganese
267 deposits, departing significantly from the high-Mg carbonate and silicate detritus endmember
268 compositions (Fig. 5). This demonstrates that a substantial Fe fraction in the HH leachates is
269 derived from the dissolution of Fe oxide phases. Importantly, the REE exhibit strong linear
270 relationships with Fe in sediment leachates (e.g. $R^2 = 0.84$ for Gd), but no correlation with Mg
271 or Al ($R^2 \sim 0$; graphs not shown). Therefore, although the chemical leaching may have
272 resulted in the dissolution of detrital and carbonate phases, the above relationships between
273 major elements implicitly suggest that the REE signal extracted with our HH solution is
274 mostly associated with Fe-oxide phases.

275

276 *4.1.2 Potential issues related to the influence of particle-seawater interactions*

277 The fundamental premise behind the use of REE and Nd isotopes in both the silicate
278 residue and Fe oxides is that they are tracers of source. Whilst the ϵ_{Nd} value and REE
279 distribution patterns of the silicate residue can be relatively unambiguously interpreted as
280 being controlled primarily by the provenance of the silicate material, REE and ϵ_{Nd} signatures
281 of the Fe-oxide fractions of river sediments may be inherited from multiple sources and hence
282 be more complicated to interpret. One particular concern is the fact that a number of samples
283 in this study (n=24) were collected in the marine environment (i.e. estuaries, submarine
284 deltas, deep-sea fans), raising the possibility that their REE and Nd isotopic compositions
285 partially reflect the influence of particle-seawater interactions and/or the presence of marine
286 Fe-oxyhydroxide phases. In estuaries, while intense REE scavenging typically takes place in
287 the low salinity regions, due to Fe-oxyhydroxide coprecipitation and coagulation of organic
288 colloids, significant REE release can also occur when salinity increases, through dissolution

289 of lithogenic suspended sediments (e.g. Sholkovitz et al., 2000; Rousseau et al., 2015). Taken
290 together with other mechanisms, such as groundwater discharge and benthic fluxes (e.g.
291 Johannesson and Burdige, 2007; Haley et al., 2017), these processes play an important role in
292 controlling the oceanic distribution of REE and can certainly influence the REE distribution
293 of labile sedimentary Fe-oxide phases at ocean margins, but also, to some extent, their Nd
294 isotope composition (e.g. Adebayo et al., 2018). Despite this potential concern, several
295 investigations have previously suggested that the sedimentary Fe-oxide fractions extracted
296 from marine depositional settings receiving high inputs of river-borne material could still
297 preserve a riverine ϵ_{Nd} signature (e.g. Bayon et al., 2004; Kraft et al., 2013; Jang et al., 2020).
298 In this study, this latter hypothesis is supported by the fact that sedimentary Fe-oxides at the
299 Niger deep-sea fan and those extracted from riverbank sediments collected along the Niger
300 mainstem, both display similar ϵ_{Nd} and $\Delta\epsilon_{Nd\ Feox-Det}$ values (Table 1). Additionally, we do not
301 observe any particular relationships between the degree of Nd isotope decoupling between
302 detrital and Fe-oxide fractions and the types of depositional environments, as inferred from
303 the similar range of $\Delta\epsilon_{Nd\ Feox-Det}$ values displayed by samples from both terrestrial (rivers,
304 deltas) and marine (estuaries, submarine deltas, deep-sea fans) environments (Fig. S1).
305 Nevertheless, the $\epsilon_{Nd\ Feox}$ value of the Congo deep-sea fan sediment (-15.6 ± 0.1) appears to be
306 slightly more radiogenic than those for the suspended particulate loads of the Kasai ($-15.8 \pm$
307 0.1) and Oubangui (-16.9 ± 0.1) rivers; i.e. the two main sediment contributors to the Congo
308 River. In this particular case, this small Nd isotope discrepancy could possibly arise from the
309 occurrence of seasonal ϵ_{Nd} variability within the Congo watershed (Allègre et al., 1996;
310 Bayon et al., 2019), complicating the comparison between ϵ_{Nd} signatures of suspended loads,
311 which represent an instantaneous snapshot of the chemical composition of river particulates,
312 and of fine-grained sediments deposited near the mouth of rivers, which integrate spatially-
313 integrated source signatures over a much longer period of time. Another exception is the case

314 of the Nile River, for which our leachate ϵ_{Nd} value (-5.4 ± 0.1) strongly differs from the
315 available dissolved ϵ_{Nd} estimate (-1.5 ; Scriviner et al., 2004). As discussed in the above
316 section 4.1.1, this sample is clearly contaminated by the presence of biogenic carbonate
317 material and, as a consequence, it probably hosts a substantial proportion of seawater-derived
318 Nd scavenged from overlying water masses having distinctive ϵ_{Nd} composition (~ -6.5 ;
319 Tachikawa et al., 2004).

320
321 Based on the above discussion, our working hypothesis in the sections below is that the
322 observed differences in both the degree of Nd isotope decoupling between detrital and Fe
323 oxide fractions and the REE distribution patterns in riverine Fe oxides can carry useful
324 information on weathering patterns in watersheds. This excludes those samples that are
325 clearly dominated by carbonate phases (i.e. Nile, Danube, Var, Nelson, Churchill; see section
326 4.1.1). Finally, note that any of the conclusions that will be yielded below regarding the
327 parameters controlling $\Delta\epsilon_{Nd_{FeOx-Det}}$ and other REE indices would still be valid if only
328 considering the ‘terrestrial’ samples (i.e. excluding those deposited in marine-influenced
329 environments).

330

331 **4.2 Tracing the sedimentary origin of river Fe-oxides with REE patterns**

332 In addition to seawater-derived Fe-oxyhydroxide phases, another potential source of Fe-
333 oxides in river sediments includes secondary oxide phases formed in soils during weathering
334 processes. This type of Fe oxides is expected to dominate in floodplains and other transport-
335 limited regions characterized by thick soil sequences, low denudation rates and intense silicate
336 weathering processes (e.g. West et al., 2005). Secondary Fe oxides may also form in response
337 to the alteration of sulphide minerals; a reaction that can further promote sulphuric acid-
338 mediated rock weathering in river catchments (e.g. Spence and Telmer, 2005; Calmels et al.,

339 2007; Blattmann et al., 2019). Finally, the alteration and/or erosion of sedimentary rocks
340 containing ancient marine Fe oxides is likely to influence the REE and Nd isotopic
341 compositions of leached oxide fractions in sediments issued from glaciated catchments (Jang
342 et al., 2020), but also in high mountainous and weathering-limited environments with high
343 denudation rates, where reduced soil cover should be accompanied with limited formation of
344 recent secondary Fe-oxides. Such preformed oxides correspond to ‘old’ authigenic phases
345 formed presumably diagenetically at the time of sediment deposition in marine basins, hence
346 inheriting the ε_{Nd} of seawater from that basin at that point in time and acquiring the REE
347 signature of corresponding reduced pore waters (e.g. Haley et al., 2017).

348

349 As reported above, the HH leachates extracted from sedimentary catchments and multi-
350 lithological river systems typically display MREE-enriched shale-normalized REE patterns,
351 while those issued from rivers draining crystalline silicate rocks exhibit REE signatures
352 mostly inherited from corresponding source rocks (Fig. 2). This can be quantified using
353 indices, such as MREE/MREE*, that are generally calculated on a linear scale (e.g. Haley et
354 al., 2004; Martin et al., 2010; Tostavin et al., 2016). In this study, the shape of the REE
355 patterns was quantified differently, assuming that the behaviour of REE is linear on a log-
356 linear plot (Lawrence and Kamber, 2005), and using shale-normalized ($_N$) Gd and the
357 geometric means of La_N and Yb_N (as illustrated in Fig. 2). This new index, referred to as
358 *concavity index* (CI), is defined as follows:

359

$$360 \quad CI = Gd_N / (La_N^6 \times Yb_N^7)^{1/13} \quad (1)$$

361

362 The concavity index can be visually represented in a $(Gd/Yb)_N$ vs $(La/Gd)_N$ plot using (Eq. 2),
363 where x represents theoretical values for CI (Fig. 6):

364

$$365 \quad (\text{La/Gd})_N = [(\text{Gd/Yb})_N \times 1/x]^{7/6} / x \quad (2)$$

366

367 A CI value of ~ 1 indicates a quasi-linear shale-normalized REE pattern, such as those
368 displayed by most detrital fractions of river sediments in Fig. 6 (see the CI=1 line), except for
369 those sediments derived from volcanic provinces. In contrast, a concavity index > 1
370 corresponds to shale-normalized REE patterns displaying downward-concave shapes,
371 increasing as the degree of MREE enrichment relative to the other REE increases. Compared
372 to the commonly used HREE/LREE vs MREE/MREE* plot (e.g. Haley et al., 2004; Martin et
373 al., 2010; Du et al., 2016), the concavity index - and its representation within the $(\text{Gd/Yb})_N$ vs
374 $(\text{La/Gd})_N$ plot - appears to be particularly well-suited for distinguishing the potential influence
375 of seawater, pore water and detrital signals in authigenic mineral phases (Fig. 6). While
376 seawater typically displays $(\text{Gd/Yb})_N$ values < 1 , sub-seafloor pore waters define a vertical
377 array in Fig. 6, with $(\text{Gd/Yb})_N$ values up to ~ 2.5 . This array corresponds to the observed
378 range of pore water REE signatures encountered in both oxic and reduced environments,
379 extending from seawater-like distribution patterns (with CI ~ 1) to pronounced MREE
380 enrichments (with CI up to ~ 2.5), respectively (e.g. Haley et al., 2004; Abbott et al., 2015).
381 Most hydrogenous Fe-Mn crusts, diagenetic nodules and Fe-oxyhydroxide phases extracted
382 from marine sediments overlap well with the field defined by pore waters (Fig. 6b), hence
383 indicating that this array is probably also representative of marine Fe oxide precipitates.

384

385 In comparison, Fe oxide fractions from river sediments draining crystalline silicate rocks
386 generally depart from the field of pore waters and marine Fe oxide precipitates (Fig. 7a), with
387 volcanogenic Fe oxide fractions being clustered in the bottom-left part of the $(\text{Gd/Yb})_N$ vs.
388 $(\text{La/Gd})_N$ plot (mean CI value: 2.1 ± 0.4 ; Fig. 2b), and Fe oxides derived from igneous and

389 metamorphic provinces displaying comparatively higher $(La/Gd)_N$ values and CI values
390 ranging between ~ 2 and 1.2 (mean CI value: 1.6 ± 0.3 ; Fig. 2a). Instead, many riverine Fe
391 oxides from both lowland and mountainous catchments draining sedimentary rocks are
392 located within the field of marine authigenic Fe oxides (Fig. 7a), typically exhibiting CI
393 values > 2 (mean CI values of 2.3 ± 0.3 and 2.4 ± 0.7 , respectively; Fig. 2c,d). Based on the
394 above discussion, this observation suggests that such MREE enrichments in riverine Fe oxide
395 fractions probably trace the source signature of ancient marine Fe oxides associated with the
396 erosion and/or alteration (followed by subsequent precipitation of Fe-oxyhydroxides in soils
397 or river systems) of sedimentary rocks. Upon formation, sedimentary rocks are likely to
398 experience various diagenetic and post-depositional processes, which most probably result in
399 the preferential dissolution of easily dissolvable hydrogenous Fe oxyhydroxide phases and
400 possibly explaining why riverine oxides derived from multi-lithological catchments plot in the
401 upper part of the field defined by pore waters (Fig. 7a). An exception is the case of those HH
402 leachates extracted from the Rhine, Var and Nile sediment samples, which display markedly
403 lower CI values (~ 1.6). However, as discussed above, these HH leachates are dominated by
404 biogenic carbonates and presumably also contain a fraction of marine hydrogenous Fe-
405 oxyhydroxides, which could hence explain their particular REE distribution patterns.

406

407 Interestingly, in Fig. 7a, a number of samples from multi-lithological catchments depart from
408 the fields for pore waters and marine Fe-oxyhydroxide phases. This includes the samples from
409 the Narva and Vistula rivers, which both drain extensive sedimentary rock formations from
410 the East European platform (Table S1) that could be possibly tagged with distinctive REE
411 signatures, but also the samples from the Ganges-Brahmaputra river system and the Lower
412 River Bann in Northern Ireland, which plot in the field of volcanogenic Fe-oxides (Fig. 7a). A
413 large part of the Lower River Bann catchment is composed of Tertiary basalts (34%; Table

414 S1), meaning that partial dissolution of volcanogenic Fe-oxide and/or detrital particles is
415 likely to have occurred in this particular sample during the HH leaching step. In contrast, the
416 Ganges-Brahmaputra river basin only host minor amounts of volcanic rocks (~ 6%), hence
417 making a contamination by volcanogenic particles unlikely in these samples. Instead, we
418 speculate that the distinctive REE signatures observed in these leachates could point towards a
419 particular source composition, perhaps related to the presence of extensive metamorphosed
420 carbonate deposits within the G.-B. watershed (e.g. Jacobson et al., 2002; Bickle et al., 2005).
421 This latter hypothesis would be in agreement with the presumed evidence for the presence of
422 dolomite in the same HH leachates, as inferred from major element data (see section 4.1.1).

423

424 Another striking feature of our results is the fact that a few Fe oxide fractions display REE
425 concavity indices distinctively higher than previously investigated marine Fe-Mn
426 oxyhydroxide phases and reduced pore waters (Fig. 7a). This corresponds in particular to the
427 following samples: Fly (CI = 4.5), Mackenzie (4.1), Gaoping (3.5) and Paraná (3.2) rivers
428 (Fig. 7a). For these latter samples, we do not have any clear explanation for the occurrence of
429 such pronounced MREE enrichments. One hypothesis would be that these REE characteristics
430 relate to the oxidation of sulphide minerals in corresponding watersheds. Previous
431 investigations have indeed shown that intense sulphide weathering proceeds in at least two of
432 these river catchments (i.e. Mackenzie and Gaoping), strongly influencing the chemistry of
433 their dissolved river loads (Calmels et al., 2007; Blattmann et al., 2019). The oxidative
434 weathering of pyrite and other sulphide minerals commonly leads to the formation of
435 secondary Fe oxides, but can also result in siderite (FeCO_3) precipitation in environments
436 dominated by carbonate weathering (e.g. Lara et al., 2015). Siderite is also a common
437 diagenetic mineral in marine sediments and shales, typically characterized by strong depletion
438 of LREE (e.g. Bau and Möller, 1992; Franchi et al., 2017) and pronounced MREE

439 enrichments (with CI up to ~50; Rongemaille et al., 2011). In the Sr/Ca vs Mg/Ca plot (Fig.
440 4), the above mentioned rivers (i.e. Fly, Mackenzie, Gaoping and Paraná) do not appear to
441 cluster in the immediate proximity of the siderite endmember, hence suggesting that the
442 occurrence of high CI values in these particular samples may more directly reflect the
443 presence of secondary Fe oxides derived from sulphide-oxidation processes. While further
444 investigation would be clearly needed to more precisely identify the various mechanisms
445 controlling the distribution of REE in river Fe oxide fractions, the main message from the
446 above discussion is that the shape of shale-normalized REE patterns and corresponding
447 concavity index can be used to identify the source of leached Fe oxide phases in sediments:
448 i.e. marine Fe-Mn oxyhydroxide phases, volcanogenic Fe oxides, secondary Fe oxides related
449 to the alteration of silicate rocks, secondary Fe oxides (or other mineral phases) related to the
450 oxidation of sedimentary sulphide minerals, and ancient marine oxides associated with
451 sedimentary rocks (Fig. 7a). Importantly, a remarkable feature of our results is that the
452 occurrence of pronounced MREE enrichments (with CI > 2) represents an ubiquitous
453 characteristics of riverine Fe oxides derived from sedimentary catchments.

454

455 **4.3. Causes of non zero $\Delta\epsilon_{Nd\ Feox-Det}$**

456 For samples draining crystalline rocks (granitic, metamorphic and basaltic rocks), the Fe-
457 oxide fraction has a ϵ_{Nd} that is similar to the silicate residue (Fig. 3). This indicates that these
458 Fe-oxides have sourced their Nd, and by inference Fe, from silicate weathering of the source
459 rocks. Except maybe for the Churchill River sample, which was shown to contain substantial
460 amounts of sedimentary dolomite (see section 4.1), the observed $\Delta\epsilon_{Nd\ Feox-Det}$ variability in all
461 other rivers draining crystalline silicate basement most likely reflect subtle Nd isotopic
462 decoupling during incongruent silicate weathering (e.g. Dausmann et al., 2019). In contrast,
463 in sedimentary and multi-lithological catchments, the appreciable and systematic difference in

464 ϵ_{Nd} between the Fe oxide fraction and silicate residue demonstrates that the REE and Fe in the
465 oxides must have a distinct origin from the silicate detrital material. This observation is
466 supported by the evidence that $\Delta\epsilon_{Nd\ Feox-Det}$ display a broad correlation with the concavity
467 index in leached sediment fractions (Fig. 7b). Based on the above discussion (section 4.2),
468 this general relationship indicates that the degree of Nd isotope decoupling between paired
469 Fe-oxide and detrital fractions in river sediments is mostly controlled by the relative
470 contribution of secondary Fe oxides derived from the alteration of crystalline silicate rocks
471 (typically characterized by $\Delta\epsilon_{Nd\ Feox-Det} < 1.6$ and CI values < 2) and ancient marine Fe oxides
472 derived from the erosion and/or alteration of sedimentary rocks in watersheds (with $\Delta\epsilon_{Nd\ Feox-}$
473 $_{Det} > 1.6$ and CI values > 2). Interestingly, the observed REE differences reported above
474 between sedimentary watersheds possibly influenced by sulphide weathering (with CI > 2.5)
475 and the other high-elevation catchments (with CI < 2.5) are not accompanied by any
476 particular differences in $\Delta\epsilon_{Nd\ Feox-Det}$ values. This indicates that the difference in the Nd
477 isotopic composition of paired Fe oxide and detrital fractions probably represents a more
478 effective tracer than REE distribution patterns for identifying the respective contribution of
479 sedimentary *versus* crystalline silicate rocks to weathering, whilst REE can provide additional
480 complementary constraints on the type of Fe oxides and associated weathered rocks.

481

482 **4.4. Links between $\Delta\epsilon_{Nd\ Feox-Det}$ and carbonate weathering indices based on river** 483 **chemistry**

484 In order to further investigate whether riverine Fe-oxide phases can be used for tracing the
485 origin of weathered rocks in watersheds, we can also compare our $\Delta\epsilon_{Nd\ Feox-Det}$ values to
486 previous estimates of carbonate and silicate weathering rates in large river basins, as inferred
487 from inverse modelling of river chemistry data (Gaillardet et al., 1999). Apart from a few
488 particular exceptions, most rivers display a general correlation between $\Delta\epsilon_{Nd\ Feox-Det}$ and the

489 relative contribution of carbonate *versus* silicate rock weathering in corresponding watersheds
490 (Fig. 8). Amongst the rivers that depart significantly from these relationships, the Yangtze,
491 Seine and Rhine rivers correspond to highly industrialized and polluted river systems. As
492 previously suggested for the Seine River using Fe isotopes, it is possible that a fraction of Fe-
493 oxides hosted in these samples derives from anthropogenic activities (Chen et al., 2014).
494 Following the discussion in section 4.2, another potential contamination issue comes from the
495 preferential alteration of easily dissolvable volcanic rocks in watersheds (and also potentially
496 during the leaching process; Elmore et al., 2011), presumably accompanied by highly
497 radiogenic signatures. This effect could possibly account for the high $\Delta\epsilon_{\text{Nd Feox-Det}}$ values
498 determined in the leached sediment fractions from the Paraná and Fraser river basins, which
499 both depart from the observed relationships in Fig. 8. Although the influence of any
500 contamination by volcanogenic material in these particular samples was not suspected from
501 REE data (i.e. these two samples plotting away from the volcanogenic Fe oxide endmember
502 in Fig. 6), both catchments do indeed cover extensive areas of flood basalts and other volcanic
503 formations (~14% and 40%, respectively; Table S1), so that it is possible that their Nd
504 isotopic composition and corresponding $\Delta\epsilon_{\text{Nd Feox-Det}}$ signatures are biased towards more
505 radiogenic composition. Taken together with the findings from sections 4.2 and 4.3, the
506 general agreement observed between $\Delta\epsilon_{\text{Nd Feox-Det}}$ and rock weathering indices inferred from
507 river chemistry data provides additional evidence that leached Fe-oxide fractions from river-
508 borne sediments can be used to characterize the relative contribution of sedimentary *versus*
509 crystalline silicate rock weathering in watersheds.

510

511 **4.5. Temperature-dependency of $\Delta\epsilon_{\text{Nd Feox-Det}}$ in lowlands**

512 There is plentiful evidence for the dependence of mineral dissolution rates on climate and
513 tectonics (e.g. Velbel, 1993; White et al., 1999; West et al., 2005; West, 2012; Gaillardet et

514 al., 2019). As a consequence, the climatic and geomorphic characteristics of studied
515 catchments can also be used to provide information on the factors controlling the distribution
516 of Nd isotopes in riverine Fe oxides, and to further investigate their potential utility as
517 weathering proxies. Below, we examine the relationships between $\Delta\epsilon_{\text{Nd Feox-Det}}$ and: i) the
518 mean annual temperature (MAT); ii) the maximum basin elevation; and iii) the percentage
519 area of carbonaceous sedimentary rocks in studied catchments, which, to a first
520 approximation, can be used as proxies for the climate, geomorphic and lithological parameters
521 in catchments. In rivers draining lowlands, measured ϵ_{Nd} differences between paired Fe-oxide
522 and detrital sediment fractions define a strong exponential anti-correlation ($R^2 = 0.62$; $n=11$;
523 Fig. 9a) with temperature, but no particular relationships with maximum elevation (Fig. 9b)
524 and lithology (Fig. 9c). An exponential dependency of $\Delta\epsilon_{\text{Nd Feox-Det}}$ on temperature is strongly
525 suggestive of a relationship similar to the Arrhenius dependency of silicate mineral
526 dissolution rates (ϖ) of the form: $\varpi = A e^{-E_a/RT}$, where A is a pre-exponential factor, E_a the
527 activation energy and R the gas constant. Silicate weathering rates increase exponentially
528 with temperature as demonstrated in previous experimental and field-based investigations
529 (White et al., 1999; Li et al., 2016). This visible Arrhenius dependency is what would be
530 predicted for transport-limited weathering regimes, where the temperature dependency of
531 silicate mineral dissolution occurs at a faster rate than the supply of detrital material through
532 denudation and erosion. In low-elevation regions and other transport-limited weathering
533 environments, the above relationship between $\Delta\epsilon_{\text{Nd Feox-Det}}$ and MAT hence probably reflects
534 the combination of two factors: 1) the increased rate of secondary Fe oxide formation as both
535 temperature and silicate weathering intensity increases; and 2) the dissolution of carbonaceous
536 rocks and associated ancient marine Fe oxides, which occur preferentially between 10 and
537 15°C in river catchments (Gaillardet et al., 2019).

538

539 **4.6. Elevation-dependency of $\Delta\epsilon_{\text{Nd Feox-Det}}$ in mountainous catchments**

540 For mixed lithology mountainous basins, measured $\Delta\epsilon_{\text{Nd Feox-Det}}$ values do not display any
541 correlation with temperature (Fig. 10a), but define instead a striking positive relationship with
542 maximum basin elevation ($R^2 = 0.75$; $n=21$; Fig. 10b), where the catchments with the highest
543 elevations display the most positive $\Delta\epsilon_{\text{Nd Feox-Det}}$ values. Note that those samples potentially
544 influenced by: i) anthropogenic contamination (Yangtze, Rhine; see section 4.4); ii) the
545 dissolution of silicate detritus (Mekong; Fig. 5); iii) the dissolution of volcanogenic material
546 in catchments draining large areas of volcanic rocks (Fraser: 40%; Chubut: 31%; Sefid Rud:
547 29%; Paraná: 14%; Table S1); or iiiii) the presence of high amounts of carbonates in
548 corresponding leachates (Nile, Danube, Nelson; section 4.1.1), were excluded from the
549 regression analysis.

550
551 One hypothesis is that the observed relationship between $\Delta\epsilon_{\text{Nd Feox-Det}}$ and maximum elevation
552 in corresponding watersheds simply reflects the overabundance of sedimentary rocks in
553 mountainous regions relative to crystalline silicate rocks. Assuming that limestones and
554 dolostones preferentially outcrop in high-elevation regions rather than in lowlands, it would
555 make sense that the proportion of ancient marine Fe oxides in river sediments increases as
556 elevation increases in watersheds. However, neither $\Delta\epsilon_{\text{Nd Feox-Det}}$ (Fig. 10c), maximum
557 elevation (Fig. S2a), nor carbonate weathering indices (Fig. S2b) appear to correlate with the
558 percentage areal coverage of carbonaceous sedimentary rocks in studied watersheds, hence
559 making this assumption unlikely. This apparent lack of lithological control on $\Delta\epsilon_{\text{Nd Feox-Det}}$ and
560 other weathering indices in studied watersheds is coherent with the fact that mineral
561 dissolution rates in high-mountain environments are primarily controlled by climate and
562 geomorphic parameters (Riebe et al., 2001; Jacobson et al., 2003; West et al., 2005; West,
563 2012). To some extent, it could also indicate that a substantial proportion of carbonate

564 weathering in high-elevation watersheds is not derived from carbonaceous sedimentary rocks,
565 but from various trace carbonate phases associated for example with siliciclastic sediments
566 (e.g. Hartmann et al., 2014).

567

568 Instead, we hypothesise that in mountainous catchments the observed difference in ϵ_{Nd} is large
569 when a significant fraction of the ancient marine Fe-oxides remains in the sediment, but
570 becomes gradually reduced as the degree of weathering of the detrital fraction increases, when
571 'young' secondary oxides derived from modern silicate weathering progressively replace or
572 overprint the ancient marine Nd signal preserved in riverine Fe oxide fractions. Areas of high
573 topography typically have high denudation rates and are commonly weathering-limited as
574 defined by water chemistry (e.g. West et al., 2005; West et al., 2012). This means that the
575 rate of supply of particulate material in these high-elevation environments is more rapid than
576 silicate mineral dissolution. Because the rate of silicate mineral dissolution is low relative to
577 the supply of particulate matter, there is little scope to reset the ϵ_{Nd} of Fe oxides through
578 recrystallization and addition of Nd through silicate mineral dissolution. As elevation
579 decreases, denudation rates also decrease, making the proportion of silicate mineral
580 dissolution higher relative the supply of particulate matter. In this case, the ϵ_{Nd} of Fe-oxides
581 becomes progressively reset by supply of Nd from silicate dissolution and, as a consequence,
582 $\Delta\epsilon_{Nd\ Feox-Det}$ decreases. This hypothesis would be in full agreement with results obtained
583 earlier for various river basins in Switzerland and New Zealand, which have clearly shown
584 that silicate weathering rates significantly decrease with increasing elevation (Drever and
585 Zobrist, 1992; Jacobson and Blum, 2003; Jacobson et al., 2003; Moore et al., 2013).

586

587 Given the temperature dependency that is observed in lowlands, it is possible that the
588 observed relationship between maximum basin elevation and $\Delta\epsilon_{Nd\ Feox-Det}$ in mountainous

589 regions (Fig. 10b) could be partially driven by temperature in upstream catchment areas (i.e.
590 differing from the mean annual temperatures listed in Table 1, which correspond to the mean
591 average T°C at the entire basin scale). It is well known that temperature decreases with
592 elevation in mountains. Although this relationship may vary significantly, as a function of
593 moisture for example, the corresponding temperature gradient (i.e. the so-called adiabatic
594 thermal lapse rate) is generally about 0.6°C per 100 m (Meyer, 1992). In lowlands, the
595 observed range of temperature (~ 20°C) between sub-Arctic (e.g. Northern Dvina) and sub-
596 tropical (Fitzroy) catchments corresponds to about 3 $\Delta\epsilon_{\text{Nd Feox-Det}}$ units (Fig. 9). To a first
597 approximation, the same range of temperature would represent about a 3500 m difference in
598 elevation in mountainous regions. Interestingly, this would be fully coherent with the
599 observed evolution of $\Delta\epsilon_{\text{Nd Feox-Det}}$ with elevation in Fig. 10, ranging from ~1 to 4 between
600 about 2000 m and 5500 m. By analogy, this observation would suggest that temperature also
601 plays a role in controlling the evolution of $\Delta\epsilon_{\text{Nd Feox-Det}}$ in mountainous regions, through direct
602 effect on the relative contribution of sedimentary *versus* crystalline silicate rock to total
603 weathering rates. In contrast to high-elevation catchments, where cold temperatures inhibit
604 the dissolution of silicate minerals, the gradual shift towards warmer temperatures as
605 elevation decreases would be accompanied by intensifying silicate weathering and, as a
606 consequence, enhanced precipitation of secondary Fe oxides, resulting in turn in river
607 sediments having lower $\Delta\epsilon_{\text{Nd Feox-Det}}$.

608

609 **4.7. Final implications and future perspectives**

610 One important implication of our results is that high mountains and glacial environments
611 are dominated by alteration of sedimentary rocks, whereas silicate weathering prevails in low-
612 elevation environments. This finding is entirely consistent with previous works, which
613 proposed a similar decoupled weathering pattern between ‘cold’ tectonically active high

614 mountain ranges - where carbonates and other sedimentary rocks are preferentially weathered,
615 and 'warm' floodplains - characterised by more intense silicate weathering (e.g. Stallard and
616 Edmond, 1983; Jacobson and Blum, 2003; Jacobson et al., 2003; Moore et al., 2013; Torres et
617 al., 2016; Yu et al., 2020). In contrast with the above-mentioned studies, mostly based on
618 water chemistry data, our new approach that combines the use of REE and Nd isotopes in
619 leached Fe oxide fractions of river-borne sediments could be potentially applicable to
620 sediment records, thereby providing a novel means for reconstructing the evolution of
621 sedimentary *versus* crystalline silicate weathering during past mountain building events, and
622 its links to the global carbon cycle and climate. This would require a careful selection of
623 well-suited sedimentary records located near the mouth of large river systems (e.g. Amazon
624 margin, Bay of Bengal) and characterized by high sedimentation rates of terrigenous material.
625 An important prerequisite for applying this new approach to sediment records will also be to
626 investigate whether leached Fe oxide phases truly correspond to river-borne oxides and assess
627 the extent to which early diagenesis and other post-depositional processes may affect the
628 preservation of pristine REE and Nd isotope signatures.

629

630 **5. Concluding remarks**

631 This study reports on a novel approach for disentangling silicate and non-silicate
632 weathering signals in the sedimentary record based on rare earth elements and neodymium
633 isotopes. In modern river sediments, the Nd isotopic difference between paired iron oxide and
634 detrital fractions ($\Delta\epsilon_{Nd_{Feox-Det}}$) can provide qualitative information on the relative contribution
635 of sedimentary *versus* crystalline silicate rocks during continental weathering, in line with
636 previous estimates based on river chemistry data. In order to better distinguish between Fe
637 oxides derived from the erosion and/or alteration of sedimentary rocks in river catchments and
638 secondary Fe oxides related to silicate weathering or oxidative weathering of sulphide

639 minerals in soils, a new REE index is also proposed: the concavity index (CI), which provides
640 a quantitative measure of the shape of shale-normalized patterns and corresponding MREE
641 enrichments. Overall, our global survey for Nd isotopes and REE in river sediments
642 demonstrates a clear link between $\Delta\epsilon_{\text{Nd Feox-Det}}$, temperature and maximum elevation in river
643 basins, which reflects a gradual temperature-dependent shift from silicate- to carbonate-
644 dominated weathering regimes, from lowlands to high-mountain environments, respectively.

645

646

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664 **References**

- 665 Adebayo, S.B., Cui, M., Hong, T., Johannesson, K.H., Martin, E.E., 2018. Rare earth
666 elements geochemistry and Nd isotopes in the Mississippi River and Gulf of Mexico
667 mixing zone. *Front. Mar. Sci.* 5, 166.
- 668 Abbott, A.N., Haley, B.A., McManus, J., Reimers, C.E., 2015. The sedimentary flux of
669 dissolved rare earth elements to the ocean. *Geochim. Cosmochim. Acta* 154, 186-200.
- 670 Allègre, C.J., Dupré, B., Nègrel, P., Gaillardet, J., 1996. Sr-Nd-Pb isotope systematics in
671 Amazon and Congo River systems: constraints about erosion processes. *Chem. Geol.*
672 131, 93-112.
- 673 Barrat, J.A., Keller, F., Amossé, J., Taylor, R.N., Nesbitt, R.W., Hirata, T., 1996.
674 Determination of rare earth elements in sixteen silicate reference samples by ICP-MS
675 after Tm addition and ion exchange separation. *Geostand. Newslett.* 20, 133-139.
- 676 Basak, C., Martin, E.E., Kamenov, G.D., 2011. Seawater Pb isotopes extracted from Cenozoic
677 marine sediments. *Chem. Geol.* 286, 94-108.
- 678 Bau, M., Möller, P., 1992. Rare earth element fractionation in metamorphogenic
679 hydrothermal calcite, magnesite and siderite. *Min. Petrol.* 45, 231-246.
- 680 Bayon, G., German, C.R., Boella, R.M., Milton, J.A., Taylor, R.N., Nesbitt, R.W., 2002. An
681 improved method for extracting marine sediment fractions and its application to Sr and
682 Nd isotopic analysis. *Chem. Geol.* 187, 179-199.
- 683 Bayon, G., German, C.R., Burton, K.W., Nesbitt, R.W., Rogers, N., 2004. Sedimentary Fe–
684 Mn oxyhydroxides as paleoceanographic archives and the role of aeolian flux in
685 regulating oceanic dissolved REE. *Earth Planet. Sci. Lett.* 224, 477-492.
- 686 Bayon, G., Pierre, C., Etoubleau, J., Voisset, M., Cauquil, E., Marsset, T., Sultan, N., Le
687 Drezen, E., Fouquet, Y., 2007. Sr/Ca and Mg/Ca ratios in Niger Delta sediments:

688 implications for authigenic carbonate genesis in cold seep environments. *Mar. Geol.* 241,
689 93-109.

690 Bayon, G., Burton, K.W., Soulet, G., Vigier, N., Dennielou, B., Etoubleau, J., Ponzevera, E.,
691 German, C.R., Nesbitt, R.W., 2009. Hf and Nd isotopes in marine sediments: Constraints
692 on global silicate weathering. *Earth Planet. Sci. Lett.* 277, 318-326.

693 Bayon, G. et al., 2015. Rare earth elements and neodymium isotopes in world river sediments
694 revisited. *Geochim. Cosmochim. Acta* 170, 17-38.

695 Bayon, G., Skonieczny, C., Delvigne, C., Toucanne, S., Bermell, S., Ponzevera, E., André, L.,
696 2016. Environmental Hf-Nd isotopic decoupling in World river clays. *Earth Planet. Sci.*
697 *Lett.* 438, 25-36.

698 Bayon, G., Delvigne, C., Ponzevera, E., Borges, A.V., Darchambeau, F., De Deckker, P.,
699 Lambert, T., Monin, L., Toucanne, S., André, L., 2018. The silicon isotopic composition
700 of fine-grained river sediments and its relation to climate and lithology. *Geochim.*
701 *Cosmochim. Acta* 229, 147-161.

702 Bayon, G., Schefuss, E., Dupont, L., Borges, A.V., Dennielou, B., Lambert, T., Mollenhauer,
703 G., Monin, L., Ponzevera, E., Skonieczny, C., André, L., 2019. The roles of climate and
704 human land-use in the late Holocene rainforest crisis of Central Africa. *Earth Planet. Sci.*
705 *Lett.* 505, 30-41.

706 Berner, R.A., Lasaga, A.C., Garrels, R.M., 1983. The carbonate-silicate geochemical cycle and
707 its effect on atmospheric carbon dioxide over the past 100 million years. *Am. J. Sci.* 283,
708 641-683.

709 Berner, R.A., Caldeira, K., 1997. The need for mass balance and feedback in the geochemical
710 carbon cycle. *Geology* 25, 955-956.

711 Bickle, M.J., Chapman, H.J., Bunbury, J., Harris, N.B., Fairchild, I.J., Ahmad, T., Pomiès, C.,
712 2005. Relative contributions of silicate and carbonate rocks to riverine Sr fluxes in the
713 headwaters of the Ganges. *Geochim. Cosmochim. Acta* 69, 2221-2240.

714 Blattmann, T. M., Wang, S. L., Lupker, M., Märki, L., Haghypour, N., Wacker, L., Chung, L.-
715 H., Bernasconi, S.M., Plötze, M., Eglinton, T.I., 2019. Sulphuric acid-mediated
716 weathering on Taiwan buffers geological atmospheric carbon sinks. *Sci. Rep.* 9, 1-8.

717 Bouchez, J., Beyssac, O., Galy, V., Gaillardet, J., France-Lanord, C., Maurice, L., Moreira-
718 Turcq, P., 2010. Oxidation of petrogenic organic carbon in the Amazon floodplain as a
719 source of atmospheric CO₂. *Geology* 38, 255-258.

720 Bouvier, A., Vervoort, J.D., Patchett, P.J., 2008. The Lu–Hf and Sm–Nd isotopic composition
721 of CHUR: constraints from unequilibrated chondrites and implications for the bulk
722 composition of terrestrial planets. *Earth Planet. Sci. Lett.* 273, 48–57.

723 Boyle, E.A., Edmond, J.M., Sholkovitz, E.R., 1977. The mechanism of iron removal in
724 estuaries. *Geochim. Cosmochim. Acta* 41, 1313-1324.

725 Calmels, D., Gaillardet, J., Brenot, A., France-Lanord, C., 2007. Sustained sulphide oxidation
726 by physical erosion processes in the Mackenzie River basin: climatic perspectives.
727 *Geology* 35, 1003-1006.

728 Caves, J.K., Jost, A.B., Lau, K.V., Maher, K., 2016. Cenozoic carbon cycle imbalances and a
729 variable silicate weathering feedback. *Earth Planet. Sci. Lett.* 450, 152–163.

730 Caves-Rugenstein, J.K., Ibarra, D.E., von Blanckenburg, F., 2019. Neogene cooling driven by
731 land surface reactivity rather than increased weathering fluxes. *Nature* 571, 99-102.

732 Chen, J. B., Busigny, V., Gaillardet, J., Louvat, P., Wang, Y.N., 2014. Iron isotopes in the
733 Seine River (France): Natural versus anthropogenic sources. *Geochim. Cosmochim. Acta*
734 128, 128-143.

735 Dausmann, V., Gutjahr, M., Frank, M., Kouzmanov, K., Schaltegger, U., 2019. Experimental
736 evidence for mineral-controlled release of radiogenic Nd, Hf and Pb isotopes from
737 granitic rocks during progressive chemical weathering. *Chem. Geol.* 507, 64-84.

738 Drever, J.I., Zobrist, J., 1992. Chemical weathering of silicate rocks as a function of elevation
739 in the southern Swiss Alps. *Geochim. Cosmochim. Acta* 56, 3209-3216.

740 Elderfield, H., Sholkovitz, E.T., 1987. Rare earth elements in the pore waters of reducing
741 nearshore sediments. *Earth Planet. Sci. Lett.* 82, 280-288.

742 Elmore, A.C., Piotrowski, A.M., Wright, J.D., Scrivner, A.E., 2011. Testing the extraction of
743 past seawater Nd isotopic composition from North Atlantic deep sea sediments and
744 foraminifera. *Geochem. Geophys. Geosyst.* 12(9).

745 Franchi, F., Rovere, M., Gamberi, F., Rashed, H., Vaselli, O., Tassi, F., 2017. Authigenic
746 minerals from the Paola Ridge (southern Tyrrhenian Sea): Evidences of episodic methane
747 seepage. *Mar. Petrol. Geol.* 86, 228-247.

748 Gaillardet, J., Dupré, B., Louvat, P., Allègre, C.J., 1999. Global silicate weathering and CO₂
749 consumption rates deduced from the chemistry of large rivers. *Chem. Geol.* 159, 3-30.

750 Gaillardet, J., Calmels, D., Romero-Mujalli, G., Zakharova, E., Hartmann, J., 2019. Global
751 climate control on carbonate weathering intensity. *Chem. Geol.* 118762.

752 Goldstein, S.J., Jacobsen, S.B., 1987. The Nd and Sr isotopic systematics of river-water
753 dissolved material: Implications for the sources of Nd and Sr in seawater. *Chem. Geol.*
754 46, 245-272.

755 Gutjahr, M., Frank, M., Stirling, C.H., Klemm, V., Van de Flierdt, T., Halliday, A.N., 2007.
756 Reliable extraction of a deepwater trace metal isotope signal from Fe–Mn oxyhydroxide
757 coatings of marine sediments. *Chem. Geol.* 242, 351-370.

758 Haley, B.A., Klinkhammer, G.P., McManus, J., 2004. Rare earth elements in pore waters of
759 marine sediments. *Geochim. Cosmochim. Acta* 68, 1265-1279.

760 Haley, B.A., Du, J., Abbott, A.N., McManus, J., 2017. The impact of benthic processes on
761 rare earth element and neodymium isotope distributions in the oceans. *Front. Mar. Sci.* 4,
762 426.

763 Hartmann, J., Moosdorf, N., 2012. The new global lithological map database GLiM: A
764 representation of rock properties at the Earth surface. *Geochem. Geophys. Geosyst.* 13.

765 Hartmann, J., Moosdorf, N., Lauerwald, R., Hinderer, M., West, A.J., 2014. Global chemical
766 weathering and associated P-release—The role of lithology, temperature and soil
767 properties. *Chem. Geol.* 363, 145-163.

768 Himmler, T., Haley, B.A., Torres, M.E., Klinkhammer, G.P., Bohrmann, G., Peckmann, J.,
769 2013. Rare earth element geochemistry in cold-seep pore waters of Hydrate Ridge,
770 northeast Pacific Ocean. *Geo-Mar. Lett.* 33, 369-379.

771 Hindshaw, R.S., Aciego, S.M., Piotrowski, A.M., Tipper, E.T., 2018. Decoupling of dissolved
772 and bedrock neodymium isotopes during sedimentary cycling. *Geochem. Persp. Lett.* 8,
773 43-46.

774 Horan, K., Hilton, R.G., Selby, D., Ottley, C.J., Gröcke, D.R., Hicks, M., Burton, K.W., 2017.
775 Mountain glaciation drives rapid oxidation of rock-bound organic carbon. *Sci. Adv.* 3,
776 e1701107.

777 Jacobson, A.D., Blum, J.D., Walter, L.M., 2002. Reconciling the elemental and Sr isotope
778 composition of Himalayan weathering fluxes: insights from the carbonate geochemistry
779 of stream waters. *Geochim. Cosmochim. Acta* 66, 3417-3429.

780 Jacobson, A.D., Blum, J.D., 2003. Relationship between mechanical erosion and atmospheric
781 CO₂ consumption in the New Zealand Southern Alps. *Geology* 31, 865-868.

782 Jacobson, A.D., Blum, J.D., Chamberlain, C.P., Craw, D., Koons, P.O., 2003. Climatic and
783 tectonic controls on chemical weathering in the New Zealand Southern Alps. *Geochim.*
784 *Cosmochim. Acta* 67, 29-46.

785 Jang, K., Bayon, G., Han, Y., Joo, Y.J., Kim, J.H., Ryu, J.S., Woo, J., Forwick, M.,
786 Szczuciński, W., Kim, J.H., Nam, S.I., 2020. Neodymium isotope constraints on chemical
787 weathering and past glacial activity in Svalbard. *Earth Planet. Sci. Lett.* 542, 116319.

788 Jochum, K.P., Nohl, U., Herwig, K., Lammel, E., Stoll, B., Hofmann, A.W., 2005. GeoReM:
789 a new geochemical database for reference materials and isotopic standards. *Geostand.*
790 *Geoanal. Res.* 29, 333-338.

791 Johannesson, K.H., Burdige, D.J., 2007. Balancing the global oceanic neodymium budget:
792 evaluating the role of groundwater. *Earth Planet. Sci. Lett.* 253, 129-142.

793 Lacan, F., Tachikawa, K., Jeandel, C., 2012. Neodymium isotopic composition of the oceans:
794 A compilation of seawater data. *Chem. Geol.* 300, 177-184.

795 Lara, R.H., Monroy, M.G., Mallet, M., Dossot, M., González, M.A., Cruz, R., 2015. An
796 experimental study of iron sulfides weathering under simulated calcareous soil
797 conditions. *Environ. Earth Sci.* 73, 1849-1869.

798 Lawrence, M. G., Greig, A., Collerson, K.D., Kamber, B.S., 2006. Rare earth element and
799 yttrium variability in South East Queensland waterways. *Aquat. Geochem.* 12, 39-72.

800 Lehner, B., Grill, G., 2013. Global river hydrography and network routing: baseline data and
801 new approaches to study the world's large river systems. *Hydrol. Process.* 27, 2171–
802 2186.

803 Li, G., Elderfield, H., 2013. Evolution of carbon cycle over past 100 million years. *Geochim.*
804 *Cosmochim. Acta* 103, 11-25.

805 Li, G., Hartmann, J., Derry, L.A., West, A.J., You, C.F., Long, X., Zhan, T., Li, L., Li, G.,
806 Qiu, W., Li, T., 2016. Temperature dependence of basalt weathering. *Earth Planet. Sci.*
807 *Lett.* 443, 59-69.

808 Martin, E.E., Blair, S.W., Kamenov, G.D., Scher, H.D., Bourbon, E., Basak, C., Newkirk, D.
809 N., 2010. Extraction of Nd isotopes from bulk deep sea sediments for paleoceanographic
810 studies on Cenozoic time scales. *Chem. Geol.* 269, 414-431.

811 Meyer, H.W., 1992. Lapse rates and other variables applied to estimating paleoaltitudes from
812 fossil floras. *Palaeogeog. Palaeoclimatol. Palaeoecol.* 99, 71-99.

813 Milliman, J.D., Farnsworth, K.L., 2011. *River Discharge to the Coastal Ocean, A global
814 synthesis.* Cambridge University Press, 392 pp.

815 Molina-Kescher, M., Frank, M., Hathorne, E.C., 2014. Nd and Sr isotope compositions of
816 different phases of surface sediments in the South Pacific: Extraction of seawater
817 signatures, boundary exchange, and detrital/dust provenance. *Geochem. Geophys.
818 Geosyst.* 15, 3502-3520.

819 Moore, J., Jacobson, A.D., Holmden, C., Craw, D., 2013. Tracking the relationship between
820 mountain uplift, silicate weathering, and long-term CO₂ consumption with Ca isotopes:
821 Southern Alps, New Zealand. *Chem. Geol.* 341, 110-127.

822 Porcelli, D., Andersson, P.S., Baskaran, M., Frank, M., Björk, G., Semiletov, I., 2009. The
823 distribution of neodymium isotopes in Arctic Ocean basins. *Geochim. Cosmochim. Acta,*
824 73, 2645-2659.

825 Poulton, S.W., Canfield, D.E., 2005. Development of a sequential extraction procedure for
826 iron: implications for iron partitioning in continentally derived particulates. *Chem. Geol.*
827 214, 209-221.

828 Raymo, M.E., Ruddiman, W.F., 1992. Tectonic forcing of late Cenozoic climate. *Nature* 359,
829 117-122.

830 Riebe, C.S., Kirchner, J.W., Granger, D.E., Finkel, R.C., 2001. Strong tectonic and weak
831 climatic control of long-term chemical weathering rates. *Geology* 29, 511-514.

832 Rongemaille, E., Bayon, G., Pierre, C., Bollinger, C., Chu, N.C., Fouquet, Y., Riboulot, V.,
833 Voisset, M., 2011. Rare earth elements in cold seep carbonates from the Niger delta.
834 Chem. Geol. 286, 196-206.

835 Rousseau, T.C., Sonke, J. E., Chmeleff, J., Van Beek, P., Souhaut, M., Boaventura, G.,
836 Seyler, P., Jeandel, C., 2015. Rapid neodymium release to marine waters from lithogenic
837 sediments in the Amazon estuary. Nature Commun. 6, 7592.

838 Scrivner, A.E., Vance, D., Rohling, E.J., 2004. New neodymium isotope data quantify Nile
839 involvement in Mediterranean anoxic episodes. Geology 32, 565-568.

840 Sholkovitz, E.R., Piegras, D.J., Jacobsen, S.B., 1989. The pore water chemistry of rare earth
841 elements in Buzzards Bay sediments. Geochim. Cosmochim. Acta 53, 2847-2856.

842 Sholkovitz, E.R., Shaw, T.J., Schneider, D.L., 1992. The geochemistry of rare earth elements
843 in the seasonally anoxic water column and porewaters of Chesapeake Bay. Geochim.
844 Cosmochim. Acta 56, 3389-3402.

845 Sholkovitz, E., Szymczak, R., 2000. The estuarine chemistry of rare earth elements:
846 comparison of the Amazon, Fly, Sepik and the Gulf of Papua systems. Earth Planet. Sci.
847 Lett. 179, 299-309.

848 Spence, J., Telmer, K., 2005. The role of sulfur in chemical weathering and atmospheric CO₂
849 fluxes: evidence from major ions, $\delta^{13}\text{C}_{\text{DIC}}$, and $\delta^{34}\text{S}_{\text{SO}_4}$ in rivers of the Canadian
850 Cordillera. Geochim. Cosmochim. Acta 69, 5441-5458.

851 Stallard, R.F., Edmond, J.M., 1983. Geochemistry of the Amazon: 2. The influence of
852 geology and weathering environment on the dissolved load. J. Geophys. Res. Oceans 88,
853 9671-9688.

854 Sűfke, F., Gutjahr, M., Gilli, A., Anselmetti, F.S., Glur, L., Eisenhauer, A., 2019. Early stage
855 weathering systematics of Pb and Nd isotopes derived from a high-Alpine Holocene lake
856 sediment record. Chem. Geol. 507, 42-53.

857 Tachikawa, K., Roy-Barman, M., Michard, A., Thouron, D., Yeghicheyan, D., Jeandel, C.,
858 2004. Neodymium isotopes in the Mediterranean Sea: comparison between seawater and
859 sediment signals. *Geochim. Cosmochim. Acta* 68, 3095-3106.

860 Torres, M.A., West, A.J., Li, G., 2014. Sulphide oxidation and carbonate dissolution as a
861 source of CO₂ over geological timescales. *Nature* 507, 346-349.

862 Torres, M.A. et al., 2016. The acid and alkalinity budgets of weathering in the Andes–
863 Amazon system: Insights into the erosional control of global biogeochemical cycles.
864 *Earth Planet. Sci. Lett.* 450, 381-391.

865 Torres, M.A., Moosdorf, N., Hartmann, J., Adkins, J.F., West, A.J., 2017. Glacial weathering,
866 sulphide oxidation, and global carbon cycle feedbacks. *Proc. Nat. Acad. Sci. U.S.A.* 114,
867 8716-8721.

868 Tostevin, R., Shields, G.A., Tarbuck, G.M., He, T., Clarkson, M.O., Wood, R.A., 2016.
869 Effective use of cerium anomalies as a redox proxy in carbonate-dominated marine
870 settings. *Chem. Geol.* 438, 146-162.

871 Velbel, M.A., 1993. Temperature dependence of silicate weathering in nature: How strong a
872 negative feedback on long-term accumulation of atmospheric CO₂ and global greenhouse
873 warming? *Geology* 21, 1059–1062.

874 von Blanckenburg, F., Bouchez, J., Ibarra, D.E., Maher, K., 2015. Stable runoff and
875 weathering fluxes into the oceans over Quaternary climate cycles. *Nat. Geosci.* 8, 538-
876 542.

877 Walker, J.C.G., Hays, P.B., Kasting, J.F.A, 1981. negative feedback mechanism for the
878 long-term stabilization of Earth's surface temperature. *J. Geophys. Res.* 86, 9776–9782.

879 West, A. J., Galy, A., Bickle, M., 2005. Tectonic and climatic controls on silicate weathering.
880 *Earth Planet. Sci. Lett.* 235, 211-228.

881 West, A.J., 2012. Thickness of the chemical weathering zone and implications for erosional
882 and climatic drivers of weathering and for carbon-cycle feedbacks. *Geology* 40, 811-814.
883 White, A.F., Blum, A.E., Bullen, T.D., Vivit, D.V., Schulz, M., Fitzpatrick, J., 1999. The
884 effect of temperature on experimental and natural chemical weathering rates of granitoid
885 rocks. *Geochim. Cosmochim. Acta* 63, 3277-3291.
886 Willenbring, J.K., von Blanckenburg, F., 2010. Long-term stability of global erosion rates and
887 weathering during late-Cenozoic cooling. *Nature* 465, 211-214.
888 Wilson, D.J., Piotrowski, A.M., Galy, A., Clegg, J.A., 2013. Reactivity of neodymium
889 carriers in deep sea sediments: Implications for boundary exchange and
890 paleoceanography. *Geochim. Cosmochim. Acta* 109, 197-221.
891 Yu, Z., Colin, C., Bassinot, F., Wan, S., Bayon, G., 2020. Climate-driven weathering shifts
892 between highlands and floodplains. *Geochem. Geophys. Geosyst.*, e2020GC008936.

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894

895 **Figure captions**

896 **Figure 1.** The location of studied sediment samples and corresponding river basins. Yellow
897 and green diamonds correspond to rivers from cratonic areas (associated with
898 igneous/metamorphic crystalline basement rocks) and volcanic watersheds, respectively.
899 Black circles indicate rivers draining coastal plains, lowlands and uplands; and white
900 triangles, those flowing from mountain environments with an elevation higher than 1000
901 meters above sea-level. Note that the numbers corresponds to the samples listed in Table
902 1. The digital elevation model is derived from the ETOPO1 Global relief Model
903 (<https://www.ngdc.noaa.gov/mgg/global/global.html>). Digital hydrographic information
904 for rivers and watersheds is derived from HydroBASINS
905 (<http://www.hydrosheds.org/page/overview>; Lehner and Grill, 2013).

906

907 **Figure 2.** Shale-normalized (WRAS; Bayon et al., 2015) REE patterns for riverine Fe oxides
908 from (a) igneous and metamorphic crystalline basements; (b) volcanic provinces; (c) low-
909 elevation regions (< 1000m) draining sedimentary and multi-lithological catchments
910 regions; and (d) mountainous regions (> 1000m) draining sedimentary and multi-
911 lithological catchments. For clarity, note that the REE data for leached sediment fractions
912 (reported relative to the initial mass of bulk sediment) were also normalized Gd_{WRAS}.
913 The thin grey lines correspond to individual samples, while the thicker line with symbols
914 represent the average shale-normalized pattern for each sample category. Also reported
915 are the concavity indices (CI) for each category of rivers, calculated using the weighted
916 geometric means of shale-normalized La and Yb (see text for details).

917

918 **Figure 3.** The distribution of Nd isotopes in paired Fe-oxide and silicate detrital fractions of
919 world river sediments. In (a) the dotted line corresponds to a 1:1 relationship between
920 Nd isotopes (expressed using the epsilon notation ϵ_{Nd}) in Fe-oxide and detrital fractions.
921 In (b) $\Delta\epsilon_{Nd\ Feox-Det}$ represents the Nd isotopic difference measured between paired Fe-
922 oxide and detrital silicate fractions. River sediments from crystalline basements (yellow
923 diamonds) and volcanic watersheds (green diamonds) display near zero $\Delta\epsilon_{Nd\ Feox-}$
924 $_{Det}$ values (mean 0.5 ± 1.1), indicating that Fe-oxides are mostly derived from alteration
925 of silicate rocks. The other studied sediment samples (black circles: coastal plains and
926 lowlands; white triangles: mountains) come from watersheds hosting mixed lithologies,
927 and are systematically shifted towards positive $\Delta\epsilon_{Nd\ Feox-Det}$ values, which we interpret
928 here as mainly reflecting the signature of ancient Fe oxides of marine origin derived from
929 the alteration and/or erosion of sedimentary rocks.

930

931 **Figure 4.** Relationship between Sr/Ca and Mg/Ca ratios in leached (Fe-oxide bearing)
932 sediment fractions. The endmember mixing model is adapted from Bayon et al. (2007),
933 using the following endmember compositions: aragonite (Sr: 10,500 ppm; Mg: 180 ppm;
934 Ca: 39%; Bayon et al., 2007), low-Mg calcite (Sr: 1,200 ppm; Mg: 250 ppm; Ca: 42.2%;
935 Bayon et al., 2007), detrital silicate (Sr: 127 ppm; Mg: 9,980 ppm; Ca: 6,735 ppm;
936 WRAS values; Bayon et al., 2015) and dolomite (Sr: 100 ppm; Mg: 11.5%; Ca: 22.5%;
937 compilation of literature data). The blue dots on mixing lines correspond to 20%
938 contribution increments from each two endmembers. The symbols for leached sediment
939 fractions correspond to river sediments from igneous/metamorphic terranes (yellow
940 diamonds), volcanic watersheds (green diamonds), low- (black circles) and high-(red
941 triangles) elevation sedimentary catchments. The black crosses correspond to
942 corresponding silt-size silicate fractions.

943

944 **Figure 5.** Relationships between (a) Sr/Ca and Fe/Mg and (b) Sr/Ca and Fe/Al in leached
945 sediment fractions. For symbols; see the caption of Fig. 6. The blue squares correspond to
946 potential end-members for detrital sediments (WRAS; Bayon et al., 2015) and dolomite
947 (compilation of literature data). The departure of leached sediment fractions from the
948 above endmember Fe/Mg and Fe/Al compositions indicates that Fe in these samples is
949 mostly associated with Fe-oxide mineral phases. The blue fields correspond to the
950 observed range of elemental ratios for various certified reference materials corresponding
951 to terrestrial iron formations (FeR-1, Fe-R2, FeR-4, IF-G) and marine Fe-rich precipitates
952 (OOPE601, OOPE603, NOD-A-1, GSPN-1, GL-O) [data derived from the GEOREM
953 web portal; Jochum et al., 2005].

954

955 **Figure 6.** Shale-normalized $(\text{Gd}/\text{Yb})_{\text{N}}$ versus $(\text{La}/\text{Gd})_{\text{N}}$ plot in (a) pore waters, seawater and
956 detrital fractions of river sediments; and (b) marine Fe-oxyhydroxide phases. The dashed
957 lines correspond to theoretical values of the concavity index (CI) of corresponding shale-
958 normalized REE patterns, calculated geometrically from La and Yb on a log-linear REE
959 pattern using the following equation $(\text{La}/\text{Gd})_{\text{N}} = [(\text{Gd}/\text{Yb})_{\text{N}} \times 1/x]^{7/6} / x$; where x
960 represents theoretical CI values. Data for sub-seafloor pore waters (Elderfield and
961 Sholkovitz, 1987; Sholkovitz et al., 1989; Sholkovitz et al., 1992; Haley et al., 2004;
962 Bayon et al., 2011; Himmler et al., 2013; Abbott et al., 2015), detrital river sediments
963 (Bayon et al., 2015), diagenetic and hydrogenous Fe-Mn deposits (Bau et al., 2014) and
964 leached Fe-oxyhydroxide phases from marine sediments (Bayon et al., 2004; Martin et
965 al., 2010; Basak et al., 2011; Wilson et al., 2013; Molina-Kescher et al., 2014; Casse et
966 al., 2019) are shown for comparison. The references used for seawater values are listed in
967 Table S4. The coloured field in panel (b) represents the range of sub-seafloor pore water
968 data. Note that the colour gradient from yellow to blue indicates changing pore water
969 REE compositions from oxic to suboxic conditions, respectively (e.g. Haley et al., 2004;
970 Abbott et al., 2015).

971

972 **Figure 7.** Using the concavity index (CI) and Nd isotopes for discriminating the source of Fe-
973 oxides in sediments. (a) Shale-normalized $(\text{Gd}/\text{Yb})_{\text{N}}$ versus $(\text{La}/\text{Gd})_{\text{N}}$ plot in leached Fe-
974 oxide fractions of river sediments. The dashed lines correspond to theoretical values of
975 the concavity index (CI) of corresponding shale-normalized REE patterns, which enable
976 to distinguish between several types of Fe-oxides in sediments: marine Fe-oxyhydroxide
977 phases, volcanogenic Fe-oxides, secondary Fe-oxides (formed via alteration of silicate
978 rocks), secondary Fe-oxides (formed presumably via oxidative weathering of sedimentary
979 sulphide minerals), ancient marine Fe oxides derived from the alteration and/or erosion of

980 sedimentary rocks. The coloured field in represents the range of marine Fe-oxyhydroxide
981 phases inferred from pore water data. Note that the colour gradient from yellow to blue
982 indicates changing pore water REE compositions between hydrogenous and diagenetic
983 Fe-oxides formed under oxic and suboxic conditions, respectively. (b) REE concavity
984 index (CI) versus $\Delta\epsilon_{Nd\ Feox-Det}$ in leached Fe oxide fractions. The broad correlation
985 between CI and $\Delta\epsilon_{Nd\ Feox-Det}$ indicates that the Nd isotope decoupling in paired Fe-oxide
986 and detrital fractions in river sediments is mostly controlled by the relative contribution
987 of secondary Fe oxides derived from the alteration of crystalline silicate rocks (when
988 $\Delta\epsilon_{Nd\ Feox-Det} < 1.6$ and CI values < 2) and ancient marine Fe oxides derived from the
989 erosion and/or alteration of sedimentary rocks in watersheds (when $\Delta\epsilon_{Nd\ Feox-Det} > 1.6$ and
990 CI values > 2).

991

992 **Figure 8.** Comparison of $\Delta\epsilon_{Nd\ Feox-Det}$ with carbonate-silicate weathering indices in major
993 world river basins. The estimates for carbonate-silicate weathering rates reported here for
994 comparison were calculated by inverse modelling of river chemistry data (Gaillardet et
995 al., 1999). (a) $\Delta\epsilon_{Nd\ Feox-Det}$ versus percentage relative contribution of carbonate
996 weathering to total weathering rates. (b) $\Delta\epsilon_{Nd\ Feox-Det}$ versus carbonate-silicate weathering
997 ratio (logarithmic scale). The observed general relationships suggest that $\Delta\epsilon_{Nd\ Feox-Det}$ can
998 be used as a proxy for estimating the relative contribution of sedimentary (carbonate-rich)
999 versus crystalline silicate rock weathering in modern and past environments. The symbols
1000 with dotted contours correspond those samples that depart from the observed
1001 relationships due to preferential alteration of basaltic rocks (Paraná and Fraser) or
1002 anthropogenic influence (Seine, Rhine, Yangtze).

1003

1004 **Figure 9.** Relationships between $\Delta\epsilon_{\text{Nd Feox-Det}}$ and (a) mean annual temperature (MAT), (b)
1005 maximum basin elevation and (c) percentage contributions of carbonate and mixed
1006 sedimentary rocks in low-elevation river basins (with maximum elevation < 1000 m). The
1007 observed exponential correlation between $\Delta\epsilon_{\text{Nd Feox-Det}}$ and MAT indicate that temperature
1008 controls the relative contribution of sedimentary *versus* crystalline silicate rock
1009 weathering in low-elevation watersheds. The dotted circles in the panel (a) corresponds to
1010 samples that were excluded in the regression analysis, due to potential anthropogenic
1011 contamination (Seine) or dissolution of volcanogenic particles in watersheds hosting
1012 substantial basaltic outcrops (Lower River Bann: 34%; Moyola: 32%).

1013

1014 **Figure 10.** Relationships between $\Delta\epsilon_{\text{Nd Feox-Det}}$ and (a) mean annual temperature, (b)
1015 maximum basin elevation and (c) percentage contributions of carbonaceous sedimentary
1016 rocks in high-elevation river basins (with maximum elevation > 1000 m). The observed
1017 correlation between $\Delta\epsilon_{\text{Nd Feox-Det}}$ and maximum basin elevation reflects a gradual shift
1018 from carbonate-dominated to silicate-dominated weathering regimes from high mountain
1019 environments to lowlands. The dotted triangles in the panel (a) correspond to those
1020 samples that were excluded from the regression analysis due to potential anthropogenic
1021 contamination (Yangtze, Rhine), dissolution of volcanogenic particles in catchments
1022 draining large areas of volcanic rocks (Fraser: 40%; Chubut: 31%; Sefid Rud: 29%;
1023 Paraná: 14%), dissolution of silicate detritus (Mekong) or due to the inferred presence of
1024 high amounts of carbonates in corresponding leachates (Nile, Danube, Nelson). Note that
1025 both exponential and linear best fit lines return similar coefficients of correlation ($R^2 =$
1026 0.75 and 0.72, respectively). However, the presumed temperature-dependency of this
1027 relationship suggests that exponential regression analysis may be more appropriate here.

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