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1 **Molybdenum isotopes in plume-influenced MORBs reveal recycling of ancient anoxic**
2 **sediments**

3

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30 **Abstract**

31 Under modern oxidising Earth surface conditions, dehydrated subducted slabs show Mo
32 isotope compositions as low as $\delta^{98/95}\text{Mo} = -1.5\text{‰}$, compared to the depleted mantle $\delta^{98/95}\text{Mo}$
33 $= -0.2\text{‰}$. Such light Mo isotope compositions reflect the redox-dependent aqueous mobility
34 of isotopically heavy Mo associated with slab dehydration. Here we analysed basaltic glasses
35 from the South-Mid Atlantic Ridge, whose parental melts are influenced by the enriched
36 Discovery and Shona mantle-plumes. We report increasingly higher $\delta^{98/95}\text{Mo}$ of up to -0.1‰
37 from the most depleted samples towards those tapping more enriched mantle sources. $\delta^{98/95}\text{Mo}$
38 values correlate positively with radiogenic Sr and Nd isotopes, which indicates recycling of
39 Proterozoic sediments with a Mo isotopic composition that was not affected by subduction-
40 related, oxic dehydration. We propose that the Mo isotope signatures were retained during
41 subduction and reflect anoxic conditions during deep-sea sedimentation in the Mid-
42 Proterozoic. Finally, Mo isotope fractionation between different terrestrial reservoirs likely
43 depends on the slab redox budget, and therefore on the timing of subduction with regard to
44 Earth's surface oxygenation.

45 **Introduction**

46 The quantification of element- and isotope fractionation in subduction zones plays an important
47 role in understanding (re)cycling mechanisms between the Earth's surface and its interior over
48 geologic time-scales. Element and isotope fractionation in this setting can be controlled by the
49 prevailing redox conditions of the subducted material. As aerobic conditions prevail at the
50 Earth's surface today, deep-sea sediments and altered oceanic crust (AOC) are largely oxidised
51 and influence the redox budget of the subducting slab (*e.g.* Evans, 2012). It is still unclear
52 whether the redox budget of slabs is directly correlated with the oxygenation of Earth's
53 atmosphere and oceans, and whether the redox state of the Earth's surface influences the
54 behaviour of redox-sensitive elements in subduction zone settings through Earth's history.

55

56 A particularly suitable element to study the relationship between oxidized and reduced
57 geochemical reservoirs is the redox-sensitive element Mo. The mobility of Mo from the
58 subducted material is controlled by the redox-state of slab-derived aqueous fluids and hydrous
59 melts (Bali et al., 2012; Skora et al., 2017). Significant Mo mobilisation and isotope
60 fractionation has been observed during subduction of oxidised lithologies, leaving behind a

61 Mo-depleted and isotopically light residual slab (*e.g.* Ahmad *et al.*, 2021; Chen *et al.*, 2019;
62 Freymuth *et al.*, 2015; König *et al.*, 2016). Under reducing conditions, only limited Mo
63 mobility is expected to occur in subducted lithologies (Bali *et al.*, 2012; Skora *et al.*, 2017) thus
64 preserving pre-subducted Mo signatures of the surface. This redox-dependent mobilisation of
65 Mo during subduction metamorphism potentially allows reconstructing the redox budget of
66 ancient subduction zones through Mo isotope compositions of mantle derived material
67 enriched by ancient recycled crustal components.

68

69 This study investigates Mo isotope systematics in a well-characterised basaltic sample suite
70 from the South Mid-Atlantic ridge (S-MAR) that shows evidence for interaction with the
71 enriched Shona and Discovery mantle plumes (Section S-1, Fig. S-1). Previous studies on these
72 samples suggested recycling of ancient oceanic crust and sediments in their mantle source with
73 an age between 1 and 2 Ga based on radiogenic isotopes, and the absence of mass-independent
74 fractionation of S isotopes (Andres *et al.*, 2002; Douglass *et al.*, 1999; Labidi *et al.*, 2013).
75 These basalts additionally feature increasingly heavier S and Se isotope compositions with
76 indicators of mantle source enrichment, and are interpreted to reflect subduction recycling of
77 reduced sediments from a redox-stratified Proterozoic ocean (Labidi *et al.*, 2013; Yierpan *et*
78 *al.*, 2020).

79 Results

80 The $\delta^{98/95}\text{Mo}$ signatures of S-MAR basaltic glass samples range between -0.24 ‰ and -0.10
81 ‰. These $\delta^{98/95}\text{Mo}$ values show correlations with $^{87}\text{Sr}/^{86}\text{Sr}$ as well as $^{143}\text{Nd}/^{144}\text{Nd}$, $^{176}\text{Hf}/^{177}\text{Hf}$
82 and $\delta^{82/76}\text{Se}$ (Fig. 1, Fig. S-4a, b). The correlations indicate that Mo isotopes trace mantle source
83 enrichment. The range in $\delta^{98/95}\text{Mo}$ of the samples is similar to that of MORBs (and seamounts)
84 from the East Pacific Rise (EPR), the Pacific-Antarctic ridge (PAR), and the Mohns Knipovich
85 ridge (MKR; Fig. 1, Fig. S-2). However, the samples from the S-MAR show a much larger
86 variability in $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ and more systematic source enrichments (Fig. 1, Fig.
87 S-2, Fig. S-4a, b).

88

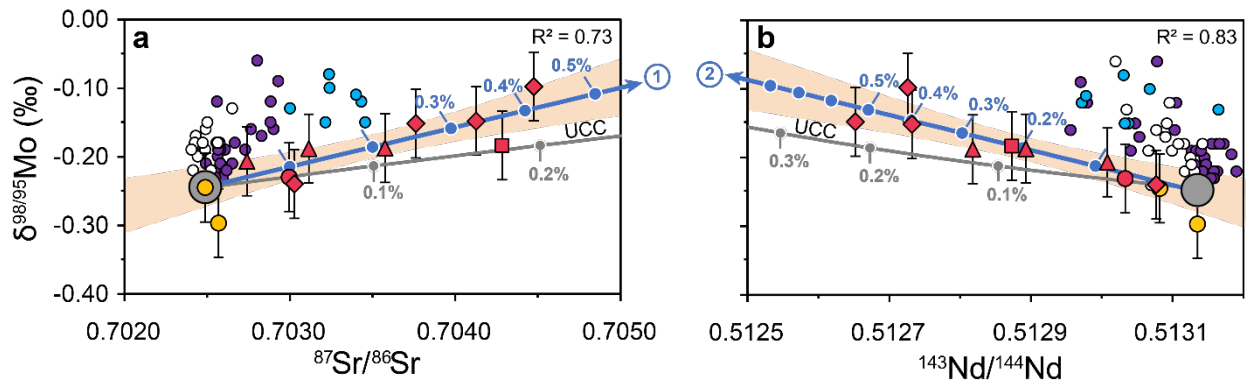
89 *Table 1: Average Mo isotope composition of MORBs from the S-MAR and PAR together with*
90 *radiogenic isotope data (Douglass *et al.*, 1999) and Se isotope compositions (Yierpan *et al.*,*
91 *2020). Individual measurements are listed in Table S-1.*

Sample	Type	$\delta^{98/95}\text{Mo}$ (‰) ^A	n	$\delta^{82/76}\text{Se}$ (‰) ^B	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$
<i>Southern Mid-Atlantic ridge</i>						
EW9309 40D-1g	Depleted N-MORB	-0.231	3	-0.18	0.702997	0.513033
EW9309 33D-1g	Discovery influenced MORB (North)	-0.098	3	-0.03	0.704475	0.512726
EW9309 28D-1g	Discovery influenced MORB (North)	-0.239	2	-0.14	0.703028	0.513077
EW9309 2D-1g	Discovery influenced MORB (South)	-0.148	2	-0.08	0.704127	0.512652
EW9309 4D-3g	Discovery influenced MORB (South)	-0.152	3	-0.04	0.703762	0.512732
EW9309 9D-3g	LOMU MORB	-0.183	2	-0.03	0.704284	0.512873
EW9309 15D-1g	Shona influenced MORB	-0.207	2	-0.13	0.702741	0.513008
EW9309 21D-1g	Shona influenced MORB	-0.188	2	-0.12	0.703115	0.512818
EW9309 22D-3g	Shona influenced MORB	-0.187	2	-0.08	0.703576	0.512893
<i>Pacific-Antarctic ridge</i>						
PAC2 DR33-1	Depleted N-MORB	-0.245	3	-0.15	0.702488	0.513082
PAC1 CV-02g	Depleted N-MORB	-0.297	3	-0.23	0.702568	0.513135

92 ^A The external reproducibility (2 SD) is 0.05 ‰ (supplementary section S-4).

93 ^B The external reproducibility (2 SD) is 0.08 ‰, except for EW9309 2D-1g and EW9309 9D-

94 3g where it is 0.04 ‰ (Yierpan et al., 2020).



depleted { ● PAR N-MORB ◆ Discovery anomaly ● EPR
MORB { ● S-MAR N-MORB ■ LOMU anomaly ○ PAR
● ambient depleted mantle ▲ Shona anomaly ● MKR

1.5 Ga pelagic sediment endmember:

① $\delta^{98/95}\text{Mo} = 0.78 \pm 0.20$ ‰; [Mo] \approx 0.76 $\mu\text{g/g}$; $^{87}\text{Sr}/^{86}\text{Sr} = 0.7203$

② $\delta^{98/95}\text{Mo} = 0.12 \pm 0.06$ ‰; [Mo] \approx 2.98 $\mu\text{g/g}$; $^{143}\text{Nd}/^{144}\text{Nd} = 0.5117$

95

96 *Figure 1: Covariation diagram of MORBs: (a) $\delta^{98/95}\text{Mo}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$; (b) $\delta^{98/95}\text{Mo}$ vs.*
97 *$^{143}\text{Nd}/^{144}\text{Nd}$. Mixing of a 1.5 Ga old pelagic sediment endmember (best-fit parameters from*
98 *linear regression) with the ambient depleted mantle (See Table S-2 for mixing parameters).*
99 *Mixing with upper continental crust (UCC) is plotted for comparison. External reproducibility*
100 *on each isotope value is considered for regression, and shaded area indicates 95% CI error*
101 *envelope. PAR samples are excluded from regression. Error bars indicate 2SD standard*
102 *deviation external reproducibility.*

103 **Discussion**

104 **Origin of enriched mantle component**

105 Combined element- and isotope systematics of S-MAR samples strongly suggest that
106 secondary mantle melting processes (such as sulphide melt segregation and fractional
107 crystallization), or seawater alteration, are unlikely the causes for the observed Mo isotope
108 variations. This is because all these processes would have obliterated any correlation between
109 Mo isotope composition and radiogenic isotopes (see supplementary section S-2 for further
110 discussion). Low degree melting of mantle plume material has been suggested to affect
111 geochemical signatures of investigated samples (le Roux et al., 2002), and might have affected
112 their $\delta^{98/95}\text{Mo}$ (e.g. Chen et al., 2022). We argue that this process did not result in a first order
113 modification of the enriched source signatures of our samples. Their distinct radiogenic isotope
114 compositions, their covariation with stable Se and S signatures, and trace element systematics
115 are independent of partial melting variations, suggesting the plume source was enriched prior
116 low-degree mantle melting (le Roux et al., 2002; Labidi et al., 2013; Yierpan et al., 2020).
117 Moreover, the absence of a covariation of $\delta^{98/95}\text{Mo}$ and Pb isotopes (Fig S-4c, d) does not
118 support the influence of an ancient low degree melt (Chen et al., 2022). Rather, a recycled
119 sediment component can explain the Pb isotope signature of LOMU-affinity samples (cf.
120 Douglass et al., 1999; Andres et al., 2002). In addition, mantle melting degrees of 2.5 % (le
121 Roux et al., 2002) cannot explain the observed variations of $\delta^{98/95}\text{Mo}$ indicating that heavy Mo
122 is likely a characteristic primary signature of the enriched plume material (see supplementary
123 section S-2 for further details). Previous studies attributed this correlation to the influence of a
124 recycled component, such as delaminated subcontinental lithospheric mantle, lower continental
125 crust, or subducted sediment ($\pm\text{AOC}$) (Andres et al., 2002; Douglass et al., 1999; le Roux et
126 al., 2002). In line with previous investigations showing covariations between other redox-
127 sensitive stable Se-S isotope systematics and radiogenic isotopes (Labidi *et al.*, 2013; Yierpan

128 *et al.*, 2020), we suggest that a sediment contribution to the S-MAR mantle source is the most
129 likely scenario for the observed heavy Mo isotope enrichment in our samples (see
130 supplementary section S-3 for further discussion). This sedimentary source was previously
131 inferred to have a mid-Proterozoic age (1 to 2 Ga; Andres *et al.*, 2002; Douglass *et al.*, 1999;
132 Labidi *et al.*, 2013; Yierpan *et al.*, 2020).

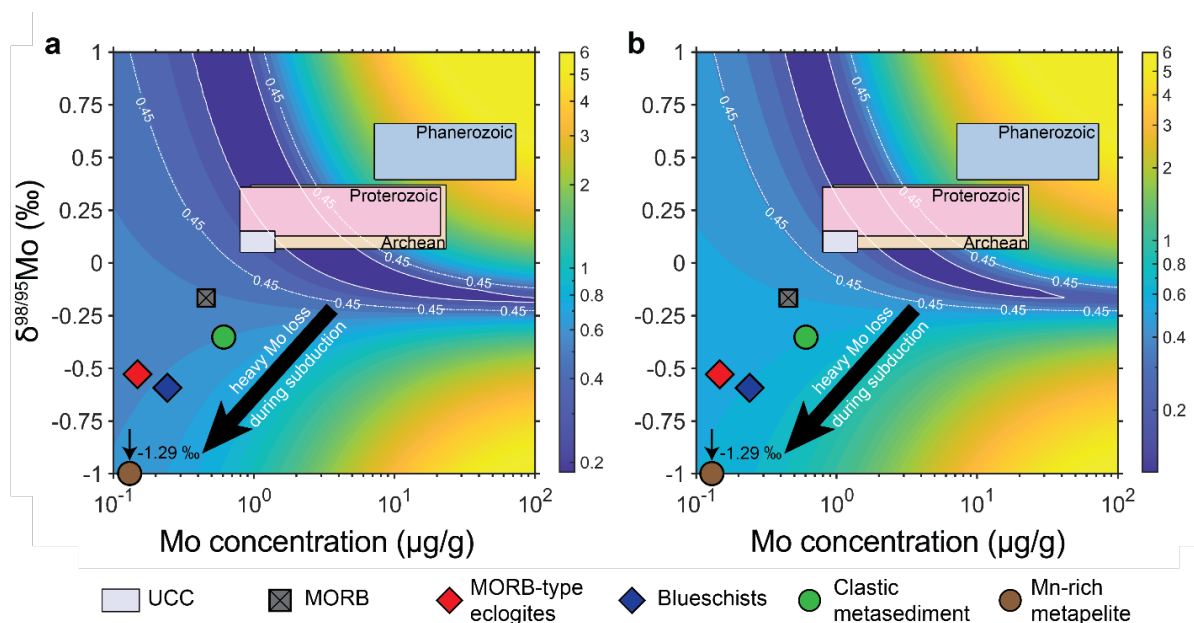
133 **Mobility of Mo during subduction: The role of fO_2**

134 Recycled Proterozoic sediments that enriched the S-MAR mantle source with isotopically
135 heavy Mo are in stark contrast to inferred, isotopically light Mo in dehydration residues of
136 subducting Phanerozoic metasediments (Ahmad *et al.*, 2021). The trend towards higher
137 observed (co-) variations of $\delta^{98/95}\text{Mo}$ with increasing degrees of mantle source enrichment
138 within the S-MAR suite therefore implies recycling of a sedimentary $\delta^{98/95}\text{Mo}$ or even total Mo
139 budget unaffected by dehydration and melting during subduction. This may be reconciled with
140 experimental studies showing immobility of Mo in low fO_2 -bearing slab fluids and hydrous
141 slab melts (Bali *et al.*, 2012; Chowdhury *et al.*, 2022; Skora *et al.*, 2017). These experiments
142 were conducted at subduction zone P-T and at reducing conditions, and showed that Mo^{4+} is
143 immobile in fluids in the presence of rutile (Bali *et al.*, 2012) and mobilization of Mo is
144 inefficient in melts due to increased partitioning of Mo^{4+} relative to Mo^{6+} into the residual
145 phases such as sulphides or rutile (Chowdhury *et al.*, 2022; Skora *et al.*, 2017). These findings
146 are also consistent with subduction of OM-rich black shales in the Lesser Antilles arc, where
147 lavas south of Martinique exhibit high $\delta^{98/95}\text{Mo}$ along with lower Mo/Ce (Freymuth *et al.*, 2016;
148 Gaschnig *et al.*, 2017), suggesting the minute contribution of unfractionated slab-derived Mo
149 to the mantle sources relative to melts originating from more oxidising sediments. Therefore,
150 significant loss of heavy Mo during subduction metamorphism did not occur in the enriched
151 mantle source component. This would have resulted in a preferential loss of more incompatible
152 (isotopically heavy) Mo^{6+} during melting and dehydration (*e.g.* McCoy-West *et al.*, 2019; Chen
153 *et al.*, 2019) and would shift $\delta^{98/95}\text{Mo}$ of the residual subducted material towards lighter values.
154 The interpretation is in line with the fO_2 -sensitive stable isotope systematics of S and Se (Table
155 1), which indicate negligible mobilisation and isotope fractionation during subduction (Labidi
156 *et al.*, 2013; Yierpan *et al.*, 2020). It is noteworthy that in some cases, that do not apply here,
157 sediments may also buffer subduction zone fluids towards oxidising conditions, such as Fe-
158 and Mn-rich (meta)sediments that show a high metamorphic fO_2 (Ague *et al.*, 2022 and
159 references therein). Yet, our S-MAR data can be attributed to an immobile behaviour of Mo
160 with unchanged $\delta^{98/95}\text{Mo}$ under reducing conditions, which is in sharp contrast with the

161 mobility of Mo under oxidising conditions where prograde subduction metamorphism can
 162 cause Mo mobilisation and alter the primary slab Mo isotope signature.

163 **Recycled sediments from an anoxic deep-ocean**

164 The $\delta^{98/95}\text{Mo}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ covariations (Fig. 1) in the S-MAR data combined with the
 165 previously established model of linear $\delta^{34}\text{S}$ - $\delta^{82}\text{Se}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ relationship (Labidi
 166 et al., 2013; Yierpan et al., 2020) allows extrapolation of $\delta^{98/95}\text{Mo}$ and [Mo] to constrain the
 167 nature of the recycled sediment. Extrapolation of the linear regression to a model composition
 168 of 1.5-Ga-old sediment ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7203$, $^{143}\text{Nd}/^{144}\text{Nd} = 0.5117$; Fig. 1, see Table S-2 for
 169 details) yields two vastly different endmember Mo signatures with $\delta^{98/95}\text{Mo}$ of $0.78 \pm 0.20 \text{ ‰}$
 170 and $0.12 \pm 0.06 \text{ ‰}$, and [Mo] of $\sim 0.76 \text{ } \mu\text{g/g}$ and $2.98 \text{ } \mu\text{g/g}$, respectively (see blue mixing lines
 171 in Fig. 1a+b). This indicates that the linear extrapolation might not sufficiently constrain the
 172 Mo composition of the recycled sediment source. Considering the present-day $^{87}\text{Sr}/^{86}\text{Sr}$ and
 173 $^{143}\text{Nd}/^{144}\text{Nd}$ model composition of 1.5-Ga-old subducted sediment, the isotopic variations of
 174 the samples represents only a small fraction of the mixing space between the ambient depleted
 175 mantle and recycled sediment. Therefore, for given radiogenic Nd and Sr signatures of the
 176 sedimentary component, different $\delta^{98/95}\text{Mo}$ and [Mo] combinations can potentially result in a
 177 minimum least squares error between the mixing model and the S-MAR data. As both variables
 178 cannot be independently constrained, a misfit function (supplementary section S-5) was applied
 179 to calculate the sedimentary $\delta^{98/95}\text{Mo}$ and [Mo] that represent the measured $\delta^{98/95}\text{Mo}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ -
 180 $^{143}\text{Nd}/^{144}\text{Nd}$ covariation best (Fig. 2).



181

182

183 *Figure 2: Least squares error ($\Delta^{98/95}\text{Mo}$ in ‰; up to 6 ‰) between calculated mixing lines and*
184 *analysed samples. Variables are (a) the $\delta^{98/95}\text{Mo}$ and [Mo] values derived from the $\delta^{98/95}\text{Mo}$*
185 *vs. $^{87}\text{Sr}/^{86}\text{Sr}$ and (b) $\delta^{98/95}\text{Mo}$ vs. $^{143}\text{Nd}/^{144}\text{Nd}$ relationship (Fig. 1). Literature values for*
186 *potential (concentration averaged) recycled lithologies, and anoxic sediments sorted by age*
187 *intervals are shown for comparison. See Table S-2 and supplementary section S-5 for*
188 *references, mixing parameters, and further details.*

189

190 Least squares errors for variable $\delta^{98/95}\text{Mo}$ and [Mo] values for a given 1.5 Ga old sediment with
191 $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ endmember values (blue field in Fig. 2) show a similar pattern for
192 both radiogenic isotope systems, which indicates that a Mid-Proterozoic pelagic sediment is
193 likely to be a valid endmember. Best fits with minimum errors overlap with the lower 1s of
194 Proterozoic OM-rich sediment data from the literature (Ye *et al.*, 2021; Table S-2) and can also
195 be achieved with a recycled sediment contribution close to that of the UCC (~ 0.15 ‰; Willbold
196 and Elliott, 2017 and references therein) with slightly higher $\delta^{98/95}\text{Mo}$ and/or [Mo] values. This
197 points towards a minor authigenic Mo enrichment from seawater or a residual enrichment of
198 Mo (*e.g.* Kendall *et al.*, 2017) and implies that $\delta^{98/95}\text{Mo}$ has not been affected by oxic conditions
199 during slab dehydration due to subduction of reducing lithologies (Fig. 2).

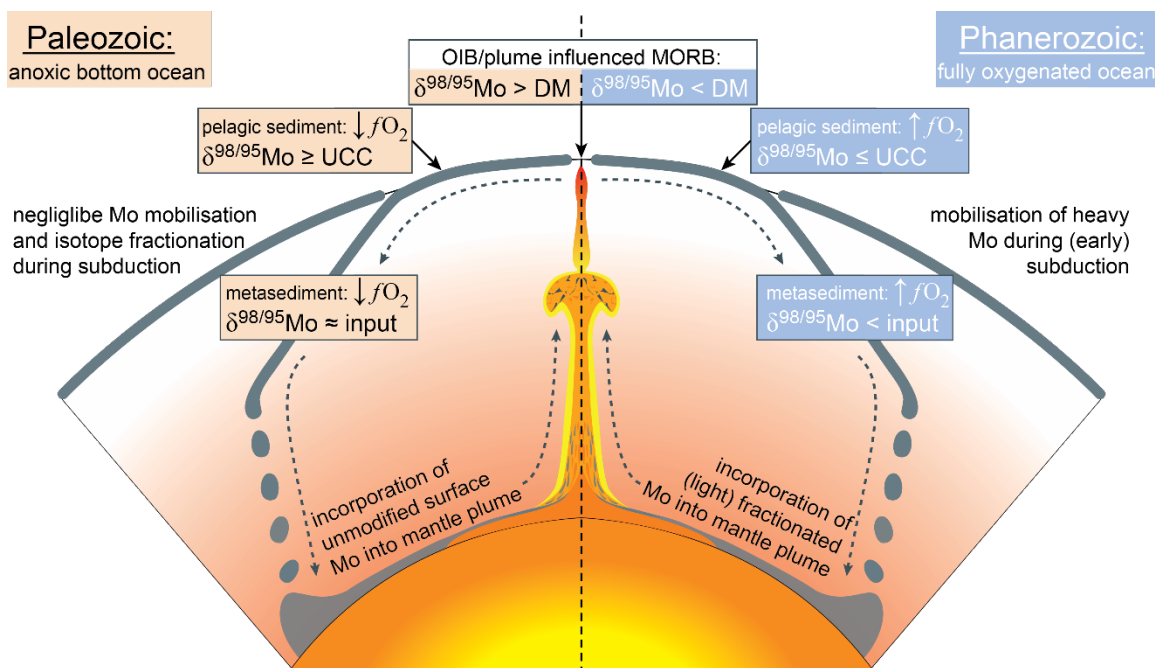
200

201 These findings support the notion that the deep-ocean remained anoxic until the beginning of
202 the Phanerozoic (*e.g.* Poulton and Canfield, 2011; Stolper and Keller, 2018). The extent of the
203 biological pump in the Proterozoic ocean, where primary productivity in oxygenated surface
204 ocean was dominated by cyanobacteria, was lower compared to modern oceans and higher
205 primary surface productivity was restricted to marine environment close to continents (Laakso
206 and Schrag, 2019). This limited the OM-flux to the Proterozoic deep-ocean and therefore
207 authigenic Mo accumulation from seawater. However, deep-ocean anoxic conditions increased
208 OM-preservation and burial efficiency (Burdige, 2007). Furthermore, under anoxic conditions
209 with an overall low concentration of dissolved SO_4^{2-} and MoO_4^{2-} , neither a significant Mo
210 enrichment from seawater into the sediment nor a significant Mo mobilisation during fluid
211 alteration is expected (Lyons *et al.*, 2014). The overall lower OM-input to the deep-ocean and
212 smaller oceanic Mo reservoir can therefore explain the small authigenic heavy Mo contribution
213 to the recycled Mid-Proterozoic deep-sea sediment (Fig. 2). With respect to the much shorter
214 ocean residence time of Mo relative to the average lifetime of an oceanic crust, the S-MAR
215 enriched endmember is likely representative for overall reducing conditions during subduction

216 an average subducted sedimentary Mo signature, which provides a context of deep-ocean redox
 217 conditions.

218 **Implications for the sedimentary Mo subduction cycle**

219 Due to the anoxic conditions in the Proterozoic deep ocean, oxidised species of major and
 220 minor elements like Fe, S and Mn, were absent in deep-sea sediments, thus lowering their redox
 221 budget/oxidising capacity compared to present-day marine lithologies (*e.g.* Ague *et al.*, 2022;
 222 Evans, 2012). This may also explain the preserved $\delta^{98/95}\text{Mo}$ of a reduced Proterozoic sediment
 223 component recycled into the S-MAR mantle source, in contrast to Neo-Proterozoic, deep
 224 mantle recycling of low $\delta^{98/95}\text{Mo}$ into mantle plume sources (see also Ma *et al.*, 2022). This is
 225 in line with Samoan OIBs, where $\delta^{98/95}\text{Mo}$ signatures are interpreted as a mixture of isotopically
 226 heavy terrigenous sediments ($\delta^{98/95}\text{Mo} \approx \text{UCC}$) and isotopically light dehydrated mafic oceanic
 227 crust, which reflect the influence of a distinct pool of Mid-Proterozoic recycled ocean crust
 228 (Gaschnig *et al.*, 2021). Altogether, this indicates that changing Earth surface redox conditions
 229 have influenced the $f\text{O}_2$ conditions during subduction and the mobility of sedimentary Mo (and
 230 by analogy that of other redox-sensitive elements) and hence, the Mo isotope budget between
 231 different Earth (silicate) reservoirs. This emphasises the time- and condition-related variations
 232 in Mo mobility during subduction on our planet (Fig. 3). We therefore conclude that the Mo
 233 isotope signature of plume-influenced volcanic rocks can be used to reconcile the redox
 234 conditions during ancient surface deposition of deep-sea sediments (*cf.* Gaschnig *et al.*, 2021)
 235 as well as during subduction-related prograde metamorphism and the inception of modern
 236 subduction.



238 *Figure 3: Illustration of the subducted sedimentary Mo cycle during the Paleozoic (left) and*
239 *Phanerozoic (right). Deep-sea sediment carry variable redox budgets influencing Mo mobility*
240 *and hence isotope fractionation during subduction over Earth's history (see text). UCC =*
241 *upper continental crust. DM = Depleted Mantle.*

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335

Molybdenum isotopes in plume-influenced MORBs reveal recycling of ancient anoxic sediments

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Supplementary Information

The Supplementary Information includes:

- Section S-1: Geochemical background of the samples
- Figure S-1
- Section S-2: Potential controls on the Mo isotope variability of S-MAR basalts
- Section S-3: Origin of enriched mantle
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Section S-1: Geochemical background of the samples

The studied samples are fresh basaltic glasses that were dredged from the South-Mid Atlantic Ridge (S-MAR) during the R/V Maurice Ewing cruise EW93-09 (Douglass et al., 1999, 1995). All samples were well-characterized in terms of radiogenic isotopes (Sr, Nd, Hf, and Pb) (Andres et al., 2002; Douglass et al., 1999), S isotopes (Labidi et al., 2013), Se isotopes (Yierpan et al., 2020), noble gases (Moreira et al., 1995; Sarda et

al., 2000) and major and trace element abundances (Kelley et al., 2013; le Roux et al., 2002c, 2002b, 2002a). This makes the selected sample suite ideal for investigating their Mo isotope signature.

The most prominent feature in the S-MAR is the localised interaction between the ambient asthenospheric mantle and the Discovery and Shona mantle plumes (Fig. S-1). A variety of recycled components have been suggested to be incorporated in these mantle plumes based on radiogenic and stable isotope systematics (Fig. S-2; see references above).

Two additional samples from the Pacific-Antarctic Ridge (PAR) complement existing PAR-MORB Mo isotope data (Bezard et al., 2016). This setting is dominantly sourced by depleted mantle being devoid of any deep mantle plume influence (*e.g.* Hamelin *et al.*, 2011) with radiogenic Sr and Nd isotope signature similar to the ambient depleted mantle (ADM) constrained for the S-MAR suite (Table S-1; Andres et al., 2002; Douglass et al., 1999).

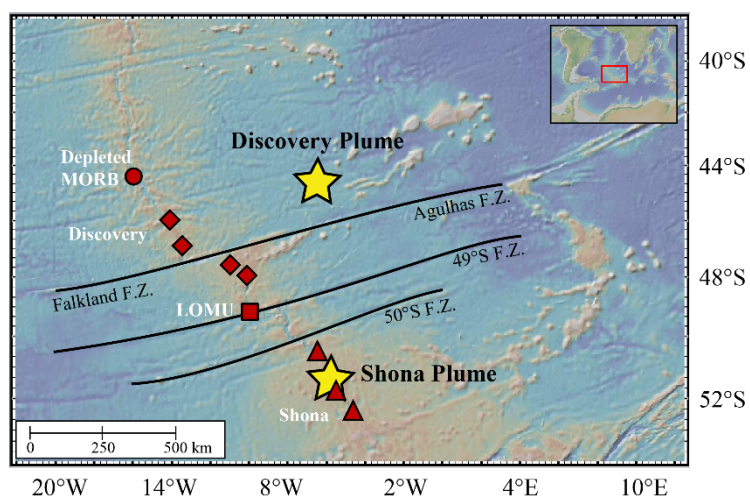


Figure S-1: Sample locations along the S-MAR together with the Discovery and Shona plume, as well as tectonic fracture zones (F.Z.). Figure made with GeoMapApp (www.geomapapp.org).

Section S-2: Potential controls on the Mo isotope variability of S-MAR basalts

Observed correlations between $\delta^{98/95}\text{Mo}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ or $^{176}\text{Hf}/^{177}\text{Hf}$ (Fig. 1, S-4a) exclude post-emplacment seawater alteration effects on the Mo isotope composition, as seawater incorporation is insensitive to relatively fluid-immobile Nd or Hf (*cf.* Labidi et al., 2013). This is also supported by Cl/K ratios close to the range of global MORB (Table S-1) (Labidi et al., 2013). Also unlikely is isotope fractionation due to segregation of an isotopically heavy sulphide melt at low melting degrees, which results into an isotopically light silicate melt (Liang et al., 2017). The studied samples were sulphide-saturated during melting (Labidi *et al.*, 2013 and references therein), however, there is no correlation between $\delta^{98/95}\text{Mo}$ and Mo/Ce with Ce being

similarly incompatible during mantle melting, but Mo being more compatible in presence of a segregated sulphide melt. Furthermore, a potential sulphide melt segregation would have removed any relationship between $\delta^{98/95}\text{Mo}$, and radiogenic isotopes (cf. Labidi *et al.*, 2013). This covariation together with lack of covariations between $\delta^{98/95}\text{Mo}$ and MgO or Mg#, overall high MgO contents (Table S-1) suggests that also fractional crystallisation during differentiation is not the cause of the observed Mo isotope variation (*e.g.* Voegelin *et al.*, 2014). This is in line with other tholeiitic systems (*e.g.* Bezard *et al.*, 2016; Yang *et al.*, 2015).

Elemental systematics of S-MAR basalts suggest mixing of ambient mantle with low-degree melts ($F = 2\text{--}3\%$; le Roux *et al.*, 2002a). Previously, elevated $\delta^{98/95}\text{Mo}$ in lavas from the EPR (devoid of a mantle plume component) were interpreted to reflect mixing of depleted MORB with an isotopically heavy endmember of magmatic origin without contribution of recycled components (Chen *et al.*, 2022). The authors argued that this heavy endmember is an ancient ($> 1\text{ Ga}$) low-degree melt ($F = 0.2\%$) of depleted mantle material that led to an initial fractionation of $\delta^{98/95}\text{Mo}$ (due to different magmatic compatibilities of Mo^{4+} and Mo^{6+} ; McCoy-West *et al.*, 2019), similar incompatible elements (*e.g.* Nb/Zr) and further parent-daughter ratios of radiogenic isotopes. This explains the observed correlation of $\delta^{98/95}\text{Mo}$ and radiogenic isotopes of Pb, Nd, and Sr (Chen *et al.*, 2022). In the S-MAR sample suite, the trace element ratios are accompanied by radiogenic isotope systematics attributable to an enriched mantle 1 (EM1) signature requiring a higher initial Rb/Sr, lower Sm/Nd and Lu/Hf, and similar U/Pb compared to DMM (Fig. S-2). These signatures, however, are primary and cannot be explained by low-F melting of plume mantle material (le Roux *et al.*, 2002). Furthermore, plume-material is pre-enriched in isotopically heavy S and Se prior to low-F melting (Labidi *et al.*, 2013; Yierpan *et al.*, 2020). The question remains, if heavy Mo is primary or if it is generated during low-F mantle melting. If we consider a low-F melting of $F=2.5\%$ melting (le Roux *et al.*, 2002) of plume material ($T = 1200\text{ }^\circ\text{C}$; $\text{Mo}^{6+}/\Sigma\text{Mo} = 0.99$), we would expect a maximum isotope shift of $\Delta^{98/95}\text{Mo}_{\text{max}}$ of $\sim 0.085\text{ }‰$ between pyroxenite melt and DMM ($F=15\%$). A mixture of both of these endmembers with 20% pyroxenite contribution in our heaviest sample EW9309 33D-1g (le Roux *et al.*, 2002) would lead to a maximum $\Delta^{98/95}\text{Mo}_{\text{max}}$ of $0.059\text{ }‰$ (melting model from Wang & Becker (2018)). This value is likely lower ($\Delta^{98/95}\text{Mo}_{\text{max}} = 0.040\text{ }‰$) when considering lower N-MORB like [Mo] of pyroxenite melts, as there is no covariation of $\delta^{98/95}\text{Mo}$ and Mo/Ce. Moreover, these pyroxenites are formed along the plume-adiabat, which is hotter than the N-MORB adiabat, lowering further the isotopic fractionation induced by low-F melting. Furthermore, we do not observe a correlation of $\delta^{98/95}\text{Mo}$ with Pb isotopes, which is inconsistent with the increase of U/Pb during ancient low-F melting (Chen *et al.*, 2022). In contrast, we observe LOMU-samples that carry an isotopic heavy signature, which is consistent with an *e.g.* Proterozoic sediment contribution (*e.g.* Rehkämper & Hofmann, 1997; Douglass *et al.*, 1999; Andres *et al.*, 2003). As correlations between radiogenic isotopes and $\delta^{34/32}\text{S}$, $\delta^{82/76}\text{Se}$, and $\delta^{98/95}\text{Mo}$ (*e.g.* Fig. 1, S-4b)

cannot be explained by magmatic processes or post-emplacement alteration (c.f. Labidi *et al.*, 2013; Yierpan *et al.*, 2020), this relationship is in strong support of recycling of enriched mantle components with $\delta^{98/95}\text{Mo}$ > depleted mantle.

Section S-3: Origin of enriched mantle

Different recycled components (*e.g.* sub-continental mantle, lower continental crust, oceanic crust, and sediment) in the mantle source of the S-MAR (Andres *et al.*, 2002; Douglass *et al.*, 1999; Labidi *et al.*, 2013; le Roux *et al.*, 2002c; Yierpan *et al.*, 2020) can have an influence on $\delta^{98/95}\text{Mo}$ of the samples. Sub-continental mantle can be excluded as a significant component, since worldwide peridotites have a MORB-like $\delta^{98/95}\text{Mo}$ ($-0.206 \pm 0.05\%$; Liang *et al.*, 2017). An aqueous fluid expelled from subducted crust can influence the mantle composition leading to elevated $\delta^{98/95}\text{Mo}$ and [Mo] (*e.g.* Ahmad *et al.*, 2021); however, we observe correlations between $\delta^{98/95}\text{Mo}$ and aqueous fluid-immobile Nd and Hf isotopes (Fig. 1, Fig. S-4a) and no covariation is observed with fluid proxies, such as Ba/Th ($R^2 = 0.1$). There is no direct estimate of $\delta^{98/95}\text{Mo}$ of the lower continental crust (LCC), however, cumulates which possibly reside in the LCC, likely incorporate preferentially light Mo isotopes (Nebel-Jacobsen *et al.*, 2021; Wille *et al.*, 2018) and are unsuitable candidates to explain Mo signatures in the basalt samples. Apart from a clear EM1 affinity, two samples analysed here show a trend towards HIMU (Fig. S-2), thus a signature of recycled AOC. However, $^{206}\text{Pb}/^{204}\text{Pb}$ ratios do not correlate with $\delta^{98/95}\text{Mo}$ and most radiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ values are accompanied with depleted mantle-like $\delta^{98/95}\text{Mo}$ (Fig. S-4c). This leads to the conclusion that a significant influence of recycled AOC is unlikely.

Section S-4: Analytical Methods

Molybdenum purification and isotope analysis was conducted in the clean laboratory facilities at the Institute of Geological Sciences, University of Bern. For all samples, enough powder was weighed to have 25 – 50 ng Mo for analysis. The material was spiked with an enriched isotope tracer solution (^{97}Mo - ^{100}Mo double spike). The samples were then dissolved in concentrated single-distilled acids in the following sequence: (i) concentrated HF-HNO₃ (3:1 mixture), (ii) concentrated HNO₃, (iii) 6 M HCl in Savillex™ Teflon vials. The dissolved samples were processed through anion- and cation exchange columns to obtain a clean Mo separate (Wille *et al.*, 2013). In the first step, samples were taken up in 4 M HCl and passed through Dowex 1X8, 200-400 mesh, anion resin. In a second step, the sample material was taken up in 0.5 M HCl and passed through Dowex 50WX8 200-400 mesh, cation exchange resin. Stable Mo isotope measurements were carried out on a Neptune Plus MC-ICPMS coupled with an Aridus II desolvating nebulizer with an uptake rate of 100 – 150 $\mu\text{l}\cdot\text{min}^{-1}$ (details are provided in Ahmad *et al.*, 2021). Six Mo isotopes (^{94}Mo , ^{95}Mo , ^{96}Mo , ^{97}Mo , ^{98}Mo and

^{100}Mo) were measured as well as ^{99}Ru and ^{101}Ru to monitor potential isobaric interference. We used resistors with $10^{-11}\Omega$ to analyse all isotopes except ^{101}Ru , which was measured using a $10^{-12}\Omega$ resistor. For analyses, we have used a combination of ‘H’ Ni sampler cone and ‘X’ Ni skimmer cone and obtained ~ 120 V/ppm on ^{95}Mo . We used a double spike correction method based on an iterative calculation procedure (see Siebert *et al.*, 2001). The total procedural blank was between 0.22 to 0.65 ng. We present all data in the δ -notation and relative to NIST SRM 3134 in ‰ (Goldberg *et al.*, 2013; Greber *et al.*, 2012) with an interference correction based on ^{99}Ru :

$$\delta^{98/95}\text{Mo} = \left(\frac{\left(\frac{^{98}\text{Mo}}{^{95}\text{Mo}} \right)_{\text{Sample}}}{\left(\frac{^{98}\text{Mo}}{^{95}\text{Mo}} \right)_{\text{NIST SRM 3134}}} - 1 \right) * 1,000 \quad (\text{S-1})$$

Interference corrected $\delta^{98/95}\text{Mo}$ ratios based on ^{99}Ru and ^{101}Ru correction were compared to ensure accurate determination of Ru interference. Background correction was obtained by bracketing samples with measurements of 0.5 M HNO_3 carrier solution on-peak. Sample and background analysis consisted of 80 and 30 cycles, respectively, with a signal integration time of 4.194 s for each cycle. Repeated measurements of the standard NIST SRM 3134 and an in-house J&M standard solution lot 602332B gave an isotopic difference of $\Delta^{98/95}\text{Mo} = 0.269 \pm 0.017$ ‰ (2 SD, $n = 14$), in agreement with (Ahmad *et al.*, 2021; Greber *et al.*, 2012). Solution standards were measured at a concentration of 25 ppb. Individually digested and chemically purified whole-rock reference materials AGV-2, BHVO-2 and W-2a yielded a $\delta^{98/95}\text{Mo}$ of -0.19 ± 0.03 ‰ (2 SD, $n = 3$), -0.11 ± 0.04 ‰ (2 SD, $n = 3$) and -0.09 ± 0.05 ‰ (2 SD, $n = 3$), respectively (Table S-1). The values are within uncertainty of those determined previously (*e.g.* Burkhardt *et al.*, 2014; Willbold *et al.*, 2016; Zhao *et al.*, 2016). Individual measurements of whole rock reference materials (Table S-1) are within the long-term 2SD external reproducibility of ± 0.05 ‰ as determined by previous measurements of BHVO-2 (Ahmad *et al.*, 2021). Including these previous BHVO-2 measurements (-0.09 ± 0.05 ‰; $n = 10$), we consider ± 0.05 ‰ as the long-term 2SD external reproducibility of our sample measurements. All samples were measured multiple times (after individual digestions and chemical separation) and obtained $\delta^{98/95}\text{Mo}$ and Mo concentration values were averaged for data presentation and interpretation (Table S-1). All individual sample measurements are within ± 0.05 ‰ compared to their average $\delta^{98/95}\text{Mo}$, except PAR sample PAC CV-02g.

Table S-1: Average and individual Mo isotope and elemental data of analysed MORBs from S-MAR and PAR. MgO and calculated Cl/K from (Labidi et al., 2013; le Roux et al., 2002b). Individual measurements represent separate sample digestions and chromatographic separations.*Internal precision on a sample run (over 80 cycles) is reported as 2 standard error (2SE).

Sample	Type	Individual measurement			Mean			MgO (wt. %)	Cl/K
		$\delta^{98/95}\text{Mo}$ (‰)	2SE* (‰)	Mo ($\mu\text{g/g}$)	$\delta^{98/95}\text{Mo}$ (‰)	2SD (‰)	Mo ($\mu\text{g/g}$)		
	<i>Southern Mid-Atlantic ridge</i>								
EW9309 40D-1g	Depleted N-MORB	-0.226	0.015	0.254	-0.231	0.064	0.254	8.30	0.11
		-0.201	0.022	0.255					
		-0.265	0.020	0.254					
EW9309 33D-1g	Discovery influenced MORB (North)	-0.100	0.013	1.25	-0.098	0.048	1.25	7.83	0.04
		-0.073	0.020	1.25					
		-0.121	0.013	1.25					
EW9309 28D-1g	Discovery influenced MORB (North)	-0.191	0.018	0.168	-0.239	0.136	0.169	8.22	0.06
		-0.287	0.018	0.170					
EW9309 2D-1g	Discovery influenced MORB (South)	-0.122	0.020	0.665	-0.148	0.074	0.663	6.24	0.03
		-0.174	0.014	0.661					
EW9309 4D-3g	Discovery influenced MORB (South)	-0.136	0.015	0.539	-0.152	0.077	0.528	7.60	0.04
		-0.124	0.020	0.527					
		-0.195	0.016	0.517					
EW9309 9D-3g	LOMU MORB	-0.167	0.017	0.654	-0.183	0.045	0.654	8.67	0.06

EW9309 15D-1g	Shona influenced MORB	-0.199	0.015	0.654					
		-0.199	0.028	0.363	-0.207	0.023	0.364	7.81	0.08
EW9309 21D-1g	Shona influenced MORB	-0.215	0.019	0.364					
		-0.179	0.027	0.587	-0.188	0.024	0.586	7.10	0.05
EW9309 22D-3g	Shona influenced MORB	-0.196	0.013	0.585					
		-0.155	0.019	1.03	-0.187	0.090	1.03	4.83	0.07
		-0.219	0.017	1.03					
PAC2 DR33-1	<i>Pacific-Antarctic ridge</i> Depleted N-MORB	-0.225	0.018	0.470	-0.245	0.062	0.462	6.63	-
		-0.228	0.028	0.461					
PAC1 CV-02g	Depleted N-MORB	-0.281	0.016	0.454					
		-0.249	0.016	0.213	-0.297	0.116	0.215	7.74	-
		-0.280	0.031	0.213					
		-0.361	0.021	0.218					
AGV-2	<i>Rock reference materials</i> Andesite	-0.192	0.015	2.13	-0.194	0.042	2.14	-	-
		-0.211	0.019	2.11					
		-0.179	0.012	2.19					
BHVO-2	Basalt	-0.092	0.015	3.43	-0.108	0.032	3.68	-	-
		-0.099	0.017	4.64					
		-0.132	0.015	2.96					
W-2a	Diabase	-0.100	0.018	0.462	-0.087	0.053	0.452	-	-
		-0.104	0.021	0.455					
		-0.057	0.010	0.439					

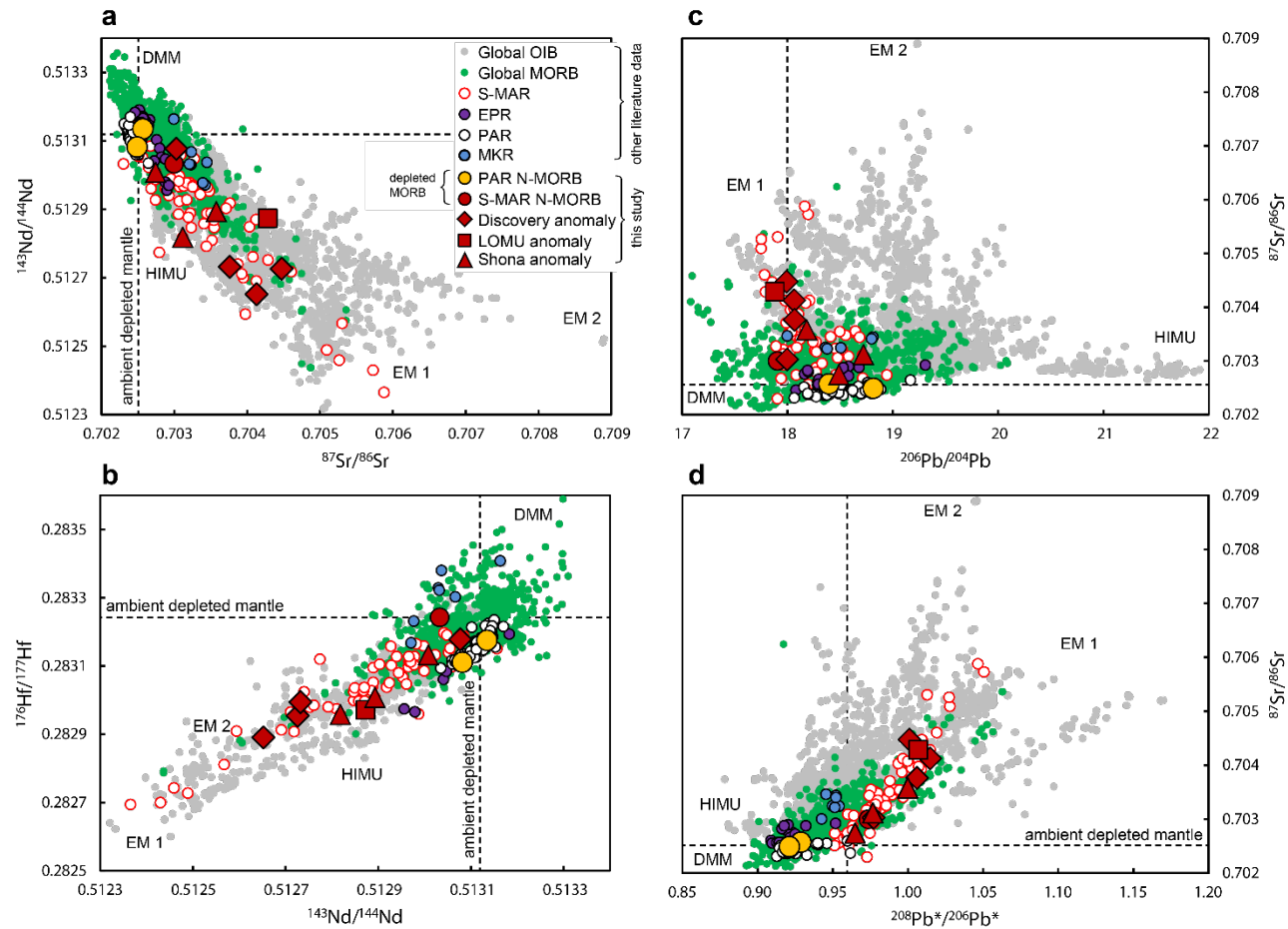


Figure S-2: Radiogenic isotope data for the studied S-MAR and PAR glasses modified after Yierpan *et al.*, (2020). (a-d) Shown for comparison are a global compilation of MORB and OIB (Stracke, 2012 and references therein). MORBs (and seamounts) near the East Pacific Rise (EPR) (Chen *et al.*, 2021 and references therein), Mohns-Knipovich ridges (MKR) (Bezard *et al.*, 2016), and PAR (Hamelin *et al.*, 2011 and references therein) are plotted for comparison. A more detailed precompilation of S-MAR basalts can be found in Labidi *et al.*, (2013). Dashed lines indicate the composition of the S-MAR ambient depleted mantle (Andres *et al.*, 2002; Douglass *et al.*, 1999). (d) $^{208}\text{Pb}^*/^{206}\text{Pb}^*$ exemplifies the time-integrated Th/U ratios (cf. Stracke, 2012).

Section S-5: Misfit model

The “linear” Mo-Sr(-Nd) isotope sample array (Fig. 1) represents only a small segment relative to the sediment endmember (Table S2), suggesting that a hyperbolic mixing curve could also account for the observed correlation. Mixing of two components (sediment & ambient depleted mantle endmember) with different Mo-Sr(-Nd) isotope compositions and concentrations will result in a hyperbola (cf. Vollmer, 1976) of the form

$$Ax + Bxy + Cy + D = 0 \quad (\text{S-2})$$

where

$$A = a_{DM} b_{SED} y_{DM} - a_{SED} b_{DM} y_{SED} \quad (\text{S-3})$$

$$B = a_{SED} b_{DM} - a_{DM} b_{SED} \quad (\text{S-4})$$

$$C = a_{DM} b_{SED} x_{SED} - a_{SED} b_{DM} x_{DM} \quad (\text{S-5})$$

$$D = a_{SED} b_{DM} x_{DM} y_{SED} - a_{DM} b_{SED} x_{SED} y_{DM} \quad (\text{S-6})$$

and

$$x = {}^{87}\text{Sr}/{}^{86}\text{Sr} \text{ or } {}^{143}\text{Nd}/{}^{144}\text{Nd}$$

$$y = \delta^{98/95}\text{Mo}$$

$$a_{SED} = \text{conc. of Mo}$$

$$b_{SED} = \text{conc. of } {}^{86}\text{Sr} (\sim \text{Sr}) \text{ or } {}^{144}\text{Nd} (\sim \text{Nd})$$

} of the sediment endmember with the isotopic composition (x_{SED}/y_{SED})

$$a_{DM} = \text{conc. of Mo}$$

$$b_{DM} = \text{conc. of } {}^{86}\text{Sr} (\sim \text{Sr}) \text{ or } {}^{144}\text{Nd} (\sim \text{Nd})$$

} of the depleted mantle endmember with the isotopic composition (x_{DM}/y_{DM}).

By inserting literature values for Mo, Sr, and Nd concentrations and isotope compositions from the literature (Table S-2) into above equations, this leads to two unknowns of the sediment endmember, a_{SED} and y_{SED} . By varying $a_{SED-MIN}$ to $a_{SED-MAX}$ and $y_{SED-MIN}$ to $y_{SED-MAX}$, a total least squares error can be determined for each $\delta^{98/95}\text{Mo}^{exp}$ generated at a given ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ or ${}^{143}\text{Nd}/{}^{144}\text{Nd}$ against the observed sample points

$$\sum_{i=a_{SED-min}}^{a_{SED-max}} \sum_{j=y_{SED-min}}^{y_{SED-max}} \sqrt{\left(\delta^{98/95}\text{Mo}_{i,j}^{exp} - \delta^{98/95}\text{Mo}^{sample}\right)^2} \quad (\text{S-7})$$

where $a_{SED-MIN} = 0.1 \mu\text{g/g}$, $a_{SED-MAX} = 100 \mu\text{g/g}$, $y_{SED-MIN} = -1 \text{‰}$, and $y_{SED-MAX} = 1 \text{‰}$.

In the obtained misfit plot (Fig. 2), the best agreement between potential sediment endmember and measured samples is given by the least squares error (blue) of the misfit function. The white dashed line indicates the sample amount multiplied by the external 2SD reproducibility to obtain a conservative upper limit of model compositions. The white solid line indicates the contour line of the least squares error obtained (see error bar in Fig. 2a+b) added by $\Delta^{98/95}\text{Mo} \approx 0.1 \text{‰}$. For comparison, the compositions of OM-rich sediments sorted by

age intervals (Ye *et al.*, 2021; Table S-2), UCC (Freymuth *et al.*, 2015; Greaney *et al.*, 2020; Rudnick and Gao, 2014; Voegelin *et al.*, 2014; Willbold and Elliott, 2017; Yang *et al.*, 2017), MORB (Bezard *et al.*, 2016; Chen *et al.*, 2021; Gale *et al.*, 2013), pelagic Mn-rich and clastic metasediments (Ahmad *et al.*, 2021), and blueschists and MORB-type eclogites (Ahmad *et al.*, 2021; Chen *et al.*, 2019) are plotted (Fig. 2).

Additional scenarios considered for misfit calculations, such as 1 Ga and 2 Ga model sediments, are plotted in Fig S-3 and show least squares errors similar to Fig. 2. This indicates that a Mid-Proterozoic UCC-like sediment with minor authigenic Mo enrichment is a likely sediment endmember (see main text).

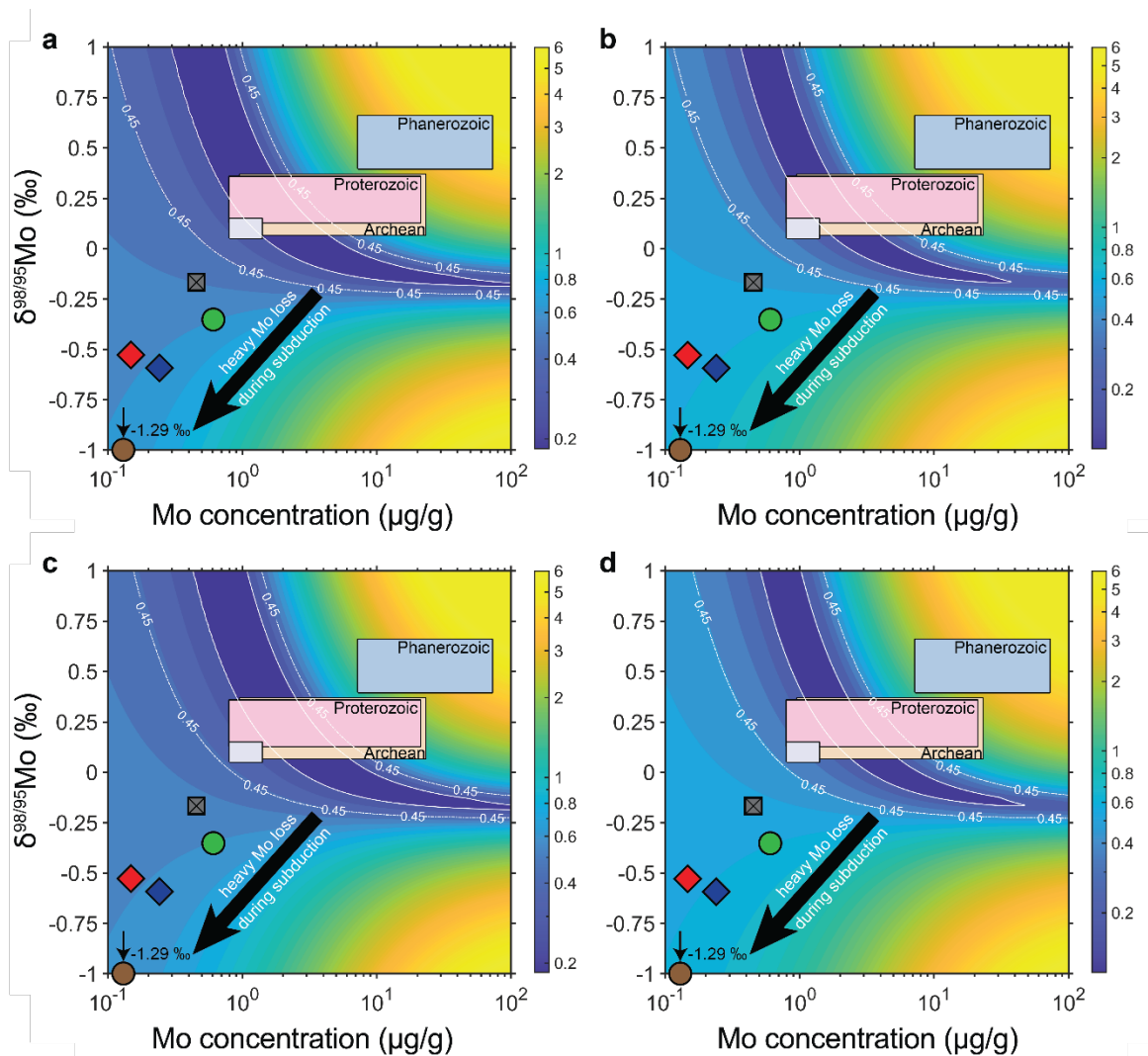


Figure S-3. Misfit calculation based on mixing parameters reported in Fig. 2 and adjusted time-integrated Sr- and Nd isotope composition of the model sediment : a) $^{87}\text{Sr}/^{86}\text{Sr}$ (1 Ga) = 0.7173; b) $^{143}\text{Nd}/^{144}\text{Nd}$ (1 Ga) = 0.5115; c) $^{87}\text{Sr}/^{86}\text{Sr}$ (2 Ga) = 0.7233; d) $^{143}\text{Nd}/^{144}\text{Nd}$ (2 Ga) = 0.5119.

Table S-2 Parameters from the two-component mixing model (Fig. 2) and compositions of marine sediments.

	$\delta^{98/95}\text{Mo}$ (‰)	[Mo] ($\mu\text{g/g}$)	$^{87}\text{Sr}/^{86}\text{Sr}$	Sr ($\mu\text{g/g}$)	$^{143}\text{Nd}/^{144}\text{Nd}$	Nd ($\mu\text{g/g}$)
<i>End-members</i>						
Ambient depleted mantle	$-0.245 \pm 0.05^{\text{A}}$	0.025 ± 0.007	0.70250 or 0.702488 ± 0.000032	9.80 ± 1.86	0.51312 or 0.513135 ± 0.00002	0.713 ± 0.05
1.5 Ga old recycled pelagic sediment	$0.78 \pm 0.20^{\text{B}}$ or $0.12 \pm 0.06^{\text{B}}$	0.76^{B} or 2.98^{B}	0.7203	300 ± 17	0.5117	$85 \pm 5.2^{\text{C}}$
<i>Marine sediments^D</i>						
Archean	0.218 ± 0.150 (M=19, N =417)	$4.70^{+18.4}_{-3.75}$ (M=19, N =744)				
Proterozoic	0.244 ± 0.114 (M=51, N =696)	$4.10^{+17.0}_{-3.30}$ (M=53, N =1262)				
Phanerozoic	0.528 ± 0.132 (M=48, N =882)	$22.9^{+49.7}_{-15.7}$ (M=48, N =1154)				

The mixing trends are generated using compositions of the Pacific/ambient depleted mantle or the most depleted MORB: PAC2 DR33-1 & PAC1 CV-02g (in italics). Uncertainties on mixing endmember elemental concentrations (Labidi et al., 2013; Plank, 2014; Rehkämper and Hofmann, 1997; Salters and Stracke, 2004; Yierpan et al., 2020) and on [Mo]- $\delta^{98/95}\text{Mo}$ values of the marine sediment compilation (Ye et al., 2021) are all 1s (except [Mo] uncertainty in depleted mantle, see Salters and Stracke, 2004).

Pelagic sediment and ambient mantle endmember isotope composition (Andres et al., 2002; Labidi et al., 2014; le Roux et al., 2002c; Rehkämper and Hofmann, 1997; Yierpan et al., 2020) are reported with 1s uncertainties, (except $\delta^{98/95}\text{Mo}$ uncertainty, which is reported as 2s). M = number of age-averaged data, N = number of different sediment samples from the literature.

^A The Mo isotope composition of the Pacific-Antarctic Ridge (PAR) basalts can be used as the depleted mantle endmember, because their Sr and Nd isotope composition match the S-MAR ambient depleted mantle isotopic composition determined previously (Andres et al., 2002; Douglass et al., 1999). The Mo isotope composition of the depleted N-MORB DR33-1g ($\delta^{98/95}\text{Mo} = -0.245 \pm 0.05 \text{‰}$) matches the light isotope composition of Phanerozoic Gorgona komatiites ($\delta^{98/95}\text{Mo} = -0.18$ to -0.25‰ , McCoy-West *et al.*, 2019) and the lighter end of PAR basalts measured previously ($\delta^{98/95}\text{Mo} = -0.06$ to -0.24‰ , excluding the three heavy outliers; Bezard *et al.*, 2016). Because $\delta^{98/95}\text{Mo}$ values of S-MAR basalts (and MORBs from EPR and PAR, Fig. 1) extend towards slightly lower values (but still within error) compared to the mean depleted mantle estimate (*e.g.*, McCoy-West *et al.*, 2019), we use this lower end of observed literature values for the ambient depleted mantle. The depleted N-MORB CV-02g ($\delta^{98/95}\text{Mo} = -0.297 \pm 0.05 \text{‰}$), however, represents an outlier being lighter than this range (but still within error) and is excluded in this model.

^B Obtained best-fit parameters from linear correlation between Mo-Sr isotope and Mo-Nd isotope (*italic font*) trend and the Mo, Sr, and Nd contents of depleted mantle and 1.5 Ga pelagic sediment (see above). Note the discrepancy between both $\delta^{98/95}\text{Mo}$ and $[\text{Mo}]$ obtained from both relationships (Fig. 1).

^C Assigned relative error from GLOSS-II (Plank, 2014) as there was no uncertainty considered for pelagic sediment Nd content.

^D Marine sediment values of different eons are log-normal means of sediments sorted by age intervals (> 1 Myr) from the literature (see compilation of Ye *et al.*, 2021).

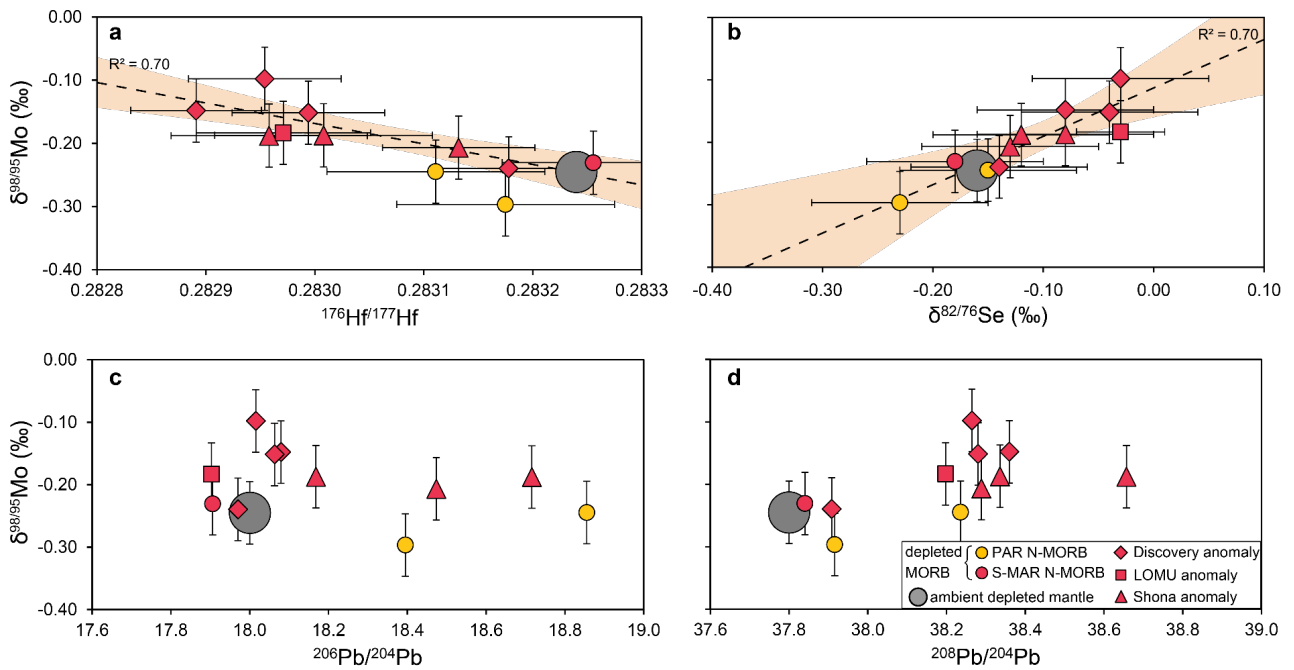


Figure S-4: Covariation diagram of S-MAR samples with: (a) $\delta^{98/95}\text{Mo}$ vs. $^{176}\text{Hf}/^{177}\text{Hf}$; (b) $\delta^{98/95}\text{Mo}$ vs. $\delta^{82}\text{Se}$ (c) $\delta^{98/95}\text{Mo}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$; (d) $\delta^{98/95}\text{Mo}$ vs. $^{208}\text{Pb}/^{204}\text{Pb}$. Error bars indicate 2SD standard deviation external reproducibility. External reproducibility on each isotope value is considered for regression in (a) and (b), and shaded area indicates 95% CI error envelope. PAR samples are excluded from the regression. Radiogenic isotopes, and Se isotopes for this sample set together with ambient depleted mantle values are from (Andres et al., 2002; Douglass et al., 1999; Yierpan et al., 2020).

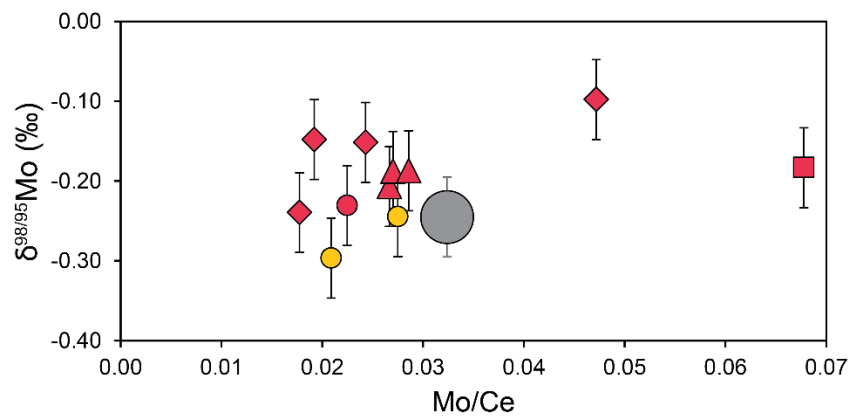


Figure S-5: Covariation diagram of S-MAR samples with $\delta^{98/95}\text{Mo}$ vs. Mo/Ce. Ce contents of samples and depleted mantle Mo/Ce are from (Kelley et al., 2013; Salters & Stracke, 2004).

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