^{184}Os-^{180}W decay: a new chronometer in geocosmochemistry

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Alpha decay of ^{184}Os to ^{180}W has been theoretically predicted, but was previously not observed in particle counting experiments [1]. Sufficiently precise measurements to detect ^{184}Os-decay in natural materials have long been impossible due to the low abundance of ^{180}W (~0.12%). However, recent advances in ICP mass spectrometry have permitted measurements at precisions better than one part in 10,000, by which non-terrestrial abundances of ^{180}W in some iron meteorites could be resolved [2]. Here, we investigate whether these ^{180}W-anomalies can be explained by α-decay of ^{184}Os, by using combined ^{180}W and Os-W concentration measurements. Osmium and W concentrations were determined by isotope dilution in 11 iron meteorites and in one H5 ordinary chondrite (Pultusk). Tungsten isotope compositions were analysed on splits of the same samples by the Neptune multi-collector ICPMS at Cologne-Bonn (for details of the analytical procedure, see [3]). The abundances of ^{180}W in our samples correlate significantly with ^{184}Os/^{184}W ratios (MSWD = 1.3) with a slope of m = 0.000295 ± 0.000060. We propose that this correlation represents a combined isochron for iron meteorites and ordinary chondrites (age: ~4.565 Ga). The slope of this isochron corresponds to a decay constant value for $\lambda^{184}\mathrm{Os}(\alpha)$ of $6.46 \pm 1.34 \times 10^{-14}$ a^{-1}. The calculated half-life is $1.12 \pm 0.23 \times 10^{13}$ years. This value is in good agreement with theoretical predictions (e.g., [4,5]), and is only slightly lower than the minimum estimate from particle counting.

Importantly, the now confirmed ^{184}Os-^{180}W decay system may constitute a viable new chronometer and tracer for geological processes that fractionate Os from W, such as core formation and silicate differentiation. This is illustrated by a measured ^{180}W-deficit in terrestrial basalts relative to chondrites by 1.16 ± 0.69 parts in 10,000, consistent with core formation ~4.5 Ga ago.

[1] Sperlein and Wolke (1976) *J. Inorg. Nucl. Chem.* **38**, 27-29. [2] Schulz et al (2012) *EPSL* **362**, 246-257. [3] Peters *et al* (2013) *LPI* 1719, 2073. [4] Medeiros *et al* (2006) *J. Phys. G: Nucl. Part. Phys* 32 B23 [5] Gangopadhyay (2009), *J. Phys. G: Nucl. Part. Phys.* 36, 095105.

Deglacial change in terrestrial carbon storage estimated by benthic δ^{13}C

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Terrestrial carbon storage is dramatically decreased during glacial periods due to cold temperatures, increased aridity, and the presence of large ice sheets on land. Most of the carbon released by the terrestrial biosphere is stored in the glacial ocean, where the isotopic signature of terrestrial carbon (δ^{13}C terrestrial carbon = -25‰) is observed as a 0.32-0.7‰ depletion in benthic foraminiferal δ^{13}C. The wide range in estimated δ^{13}C change is due to different subsets of benthic δ^{13}C data and different methods of weighting the mean δ^{13}C by volume. We estimate the glacial-interglacial δ^{13}C change of marine DIC using benthic *Cibicides* spp. δ^{13}C records from 486 core sites (seven to twelve times as many as previous studies). We divide the ocean into eight regions to generate linear regressions of regional δ^{13}C versus depth (0.5-5 km) for the late Holocene (0-6 ka) and Last Glacial Maximum (18-21 ka) and estimate a mean δ^{13}C decrease of 0.39 +/-0.05‰ (2σ). Assuming an isotope change of 0.04‰ in the surface ocean (0-0.5 km), we estimate a whole ocean δ^{13}C change of 0.33‰, equivalent to a terrestrial biosphere decrease of ~500 Pg C. If we account for uncertainty of +/-0.05‰ in our deep ocean estimate and a range of 0.34‰ for possible surface changes, we estimate a terrestrial biosphere decrease of 360-600 Pg C. Our estimate is smaller than a vegetation reconstruction estimate of 750-1050 Pg C [1] and within the uncertainty range of a recent dynamic global vegetation model estimate of 550-694 Pg C [2]. Additionally, our estimate falls in the middle of previous whole ocean δ^{13}C estimates of 0.32‰ [3] and 0.46‰ [4].

[1] Crowley, (1995), *Global Biogeochemical Cycle*, **9**, 377-389. [2] Prentice *et al* , (2011), *New Phytologist,* **189**, 988-998. [3] Duplessy *et al* , (1988), *Paleoceanography,* **3**, 343-360. [4] Curry *et al* , (1988), *Paleoceanography,* **3**, 317-341.

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