Supplementary Information

To illustrate the effect of variable mantle iron content on the elemental content of a putative planet, we used a similar accretionary model to that described previously in (1, 2). Planets were presumed to possess chondritic ratios of the involatile elements, with accretion occurring in 1% intervals. Metal-silicate equilibration occurred at a fixed fraction of the depth to the core-mantle boundary, with peak pressures of 20GPa reached at the tail-end of accretion, and equilibration temperature determined by the peridotite liquidus(3). Equilibration pressures imply that the planet's mass (M_P) is greater than that of Mars (M_M) but less than Earth's (M_E). i.e. $M_E > M_P > M_M$. The FeO contents of the accreting bodies were fixed throughout accretion, and each increment of arriving metal was equilibrated with the bulk mantle before being sequestered in the core. Elements were partitioned between the accreting metal and the bulk mantle at each step, using metal-silicate partition coefficients ($D_i = [i]_{metal} \big/ [i]_{silicate}$) from the sources detailed in (2). All planet's cores were presumed to contain a terrestrial-like S content of 2%(4), with metal phase compositions determined using the epsilon formalism of (5). Metal interactions were taken from (6, 7), and all metals except the major solutes (S, Si, O and Ni) were assumed to be at infinite dilution. For full details of the modelling approach and partition coefficients please see (2).

The modelled surface seawater compositions presented in figure 2 were generated using the PHREEQC modelling package, with putative Fe concentrations taken from (8, 9) and seawater compositions from (10-12). Many aspects of the ancient seawater composition are not precisely constrained, although there is a broad consensus on the general trends of many of the key parameters. Spatial and temporal variability in composition and redox level was surely present in the Archean (13) but is insignificant compared to the major changes that occur with rising oxygen levels moving forward in time. References for composition data are given in the figure 2 caption. With respect to changes in pH, we followed the approach of (14) and assumed pH increased from a low of 6.3 in the Archaean, to 7 in the interval between the GOE and the ~550Ma to the present value. The prevailing oxygen fugacity was set by assuming the initial atmosphere was in broad equilibrium with the mantle ($\log_{10}$ fO_2 8+/-2). The atmospheric oxygenation at the GOE boundary appears variable(15) , but post GOE is well established and relatively well constrained (GOE to pre-~550Ma $\log_{10}$ fO_2 2.7+/-0.3)(16). There is continued debate about the details of O_2 evolution between the GOE and the start of the Phanerozoic(17) but for the purposes presented here we simply use a single step at the Phanerozoic boundary to the present day value ($\log_{10}$ fO_2 -0.7+/-0.3).

Supplementary Information References

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