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The origin of Earth's first continents and the onset of plate tectonics

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Methods

High pressure-temperature experiments. Experimental runs were carried out on a ½ inch (1.27 cm) end-loaded piston cylinder press at the NERC Recognised Experimental Geoscience Laboratories, School of GeoSciences, University of Edinburgh, U.K using talc-Pyrex®-graphite assemblies. Capsules were centred in assemblies using alumina spacers, within 1mm of the thermocouple junction. Experiments were performed using the hot-piston-out technique in which runs were pressurised to 110% of the reported final pressure, heated, and pressure allowed to bleed off to the desired run pressure. Pressure was monitored and maintained during runs within 0.25 kbar of the reported pressure, and a 15% correction to nominal pressure applied to compensate for pressure loss due to internal friction, previously calibrated for the assembly based on the quartz-coesite transition, kyanite-sillimanite transition, jadeite-albite-quartz reaction and the melting point of diopside. Temperature was measured with a Pt-Pt13%Rh thermocouple. No correction for the effects of pressure on thermocouple EMF was applied. Temperature gradients across the sample are small over the temperature range used here, in the order of 10°C, and temperature during runs did not deviate more than 5°C from the reported values. Experiments were quenched by shutting off power to the heating circuits, and fell to below 300°C after 5 seconds, and to below 100°C after 12 seconds. The starting material was a finely-powdered natural glass sample [primitive, depleted and anhydrous Kroenke-basalt sample: 1187-8 ([Fitton and Godard, 2004](#))], with up to ~3 vol.% of crystallised olivine, from the Ontong Java Plateau. Approximately 0.035-0.05 grams of the sample powder and ~2-3 wt.% 18.2MΩm deionised water were loaded in

welded Ag₇₀Pd₃₀ capsules and run for 24 or 48 hours. Experiments are designed to simulate partial melting of oceanic plateau material in a possible Eoarchaean subduction zone thus requiring an oxygen fugacity (fO_2) close to nickel-nickel oxide (NNO) mineralogical equilibrium. However, experiments cannot be buffered using an NNO assemblage due to Ni reaction with the Pd-bearing capsule material. Also, several studies show that fO_2 cannot be buffered using solid buffers in water undersaturated experiments ([Skjerlie & Patino Douce, 2002](#)). Nevertheless, although not explicitly buffered, the talc-Pyrex® experimental assembly used in this study is believed to impart a NNO oxygen fugacity on the experimental runs. Many previous fluid-absent experiments in the literature do not approach equilibrium, but the use of our natural glass enables us to do so. Equilibrium conditions in our experiments are shown by (1) all mineral phases being homogeneous and un-zoned and (2) similar phase proportions and compositions being derived from experiments run for 24 and 48 hours at the same pressures and temperatures.

Major element analyses. Experimental charges were encased in epoxy resin, ground down and polished for imaging and major and trace element analysis. Samples were carbon-coated and imaged on an FEI XL30 Field Emission Gun Environmental SEM (ESEM) at the School of Earth and Ocean Sciences at Cardiff University, U.K. Major element and high concentration trace element analyses of run products were determined at the School of GeoScience, University of Edinburgh using a five spectrometer Cameca SX100 electron microprobe instrument with 15kV acceleration voltage. Mineral products were measured using a fully focussed beam and two conditions of 4 nA (major elements) and 100 nA (minor and trace elements). For glass analyses, a defocused beam of 2 nA and 5 μm was used for major elements to avoid Na loss and to target the relatively small melt pools, and 80 nA at 5 μm was used for minor and trace elements.

Trace element analyses. Trace element analyses on the glasses were determined on a secondary ion mass spectrometer (SIMS) at the University of Edinburgh and a laser ablation inductively-coupled plasma mass spectrometer (LA-ICP-MS) at Cardiff University. SIMS analysis was performed on a Cameca IMS-4f ion microprobe. Analyses were made using a 160° primary beam of 15 keV impact energy. A 1nA beam was focussed to a 5-8 micron spot and 4.5 keV positive secondary ions were measured. Molecular ion overlaps were significantly reduced by use of energy filtering (ion energies between 55 and 95 eV measured). Corrections were made for overlap of the light rare earth elements (REE) on the heavy REE and HoO on Ta. Concentrations were determined using a combination of STHS-1, GSD-1G and SRM610 glass standards. LA-ICP-MS analysis was carried out using an ESI P213 laser ablation system and a thermo X Series 2 ICP-MS. Reference standards BIR, BHVO, BCR and NIST612 were used to calibrate the instrument, correct instrument drift and ensure the accuracy and precision of the analyses. In the unknown reference standard (BCR) Ti, Y, Zr, Pr, Nd, Dy, Lu, Th and U did not deviate more than 5% from published values whereas Sr, Nb, Ba, Ce and Gd did not deviate by more than 5-10% and La, Tb, Ho, Yb and Ta did not deviate more than 10-18%.

Modelling the addition of an aqueous slab-derived fluid. We model the addition of aqueous fluid-mobile elements to our tonalite melts by using the methodology of ([Kogiso et al., 1997](#)) who studied dehydration processes experimentally on a natural amphibolite under open system conditions. According to [Kogiso et al. \(1997\)](#) the mobility of an element can be calculated by:

$$E_m = \frac{C_0 - C_l}{C_0}$$

Where E_m is element mobility, C_0 is the element concentration in a starting amphibolite and C_1 is the concentration of an element in the amphibolite after fluid loss. The concentration of an element in a slab-derived fluid is then calculated based on the extracted water content in the altered slab by:

$$C_{sf} = \frac{\sum(C_i X_i) E_m}{H_2O}$$

Where C_{sf} is the element concentration in the slab-derived fluid, C_i is the concentration of the element in the starting rock unit(s), X_i is the proportion that the rock unit(s) contribute to the slab fluid concentration, and H_2O is the mass fraction of water. Note that [Kogiso et al. \(1997\)](#) do not supply E_m data for potassium, but because of the similar ionic charge and radii we use the Rb E_m figure for potassium. We also assume a H_2O content of 1% for our calculations. It is estimated that the slab-fluid would be derived from the upper sections of the OJP; however, estimating the relative proportions of the different rock types that contribute to the slab fluid (X) is extremely difficult. Therefore, we assume that there is a 1% contribution from high-Nb basalts, 2.5% from volcanoclastic rocks on the shear surface and then equal contributions (~32.2%) from the Kroenke, Kwaimbaita and Singgalo-type rocks from the OJP ([Fitton and Godard, 2004](#)). It is important to note that a variety of different proportions can produce similar results to ours, and the interactive Tables DR4 and DR5 (Supplementary Information) allow the reader to modify the proportions to explore all available options. The final composition of our combined tonalite melts and slab-fluids can then be determined using mass balance:

$$C_{T+sf} = C_T(1 - f) + C_{sf}f$$

Where C_{T+sf} is the concentration of an element in a mixture of our tonalites with an added fraction (f) of the slab-derived component, and C_T is the concentration in our tonalite melts (Tables DR4 & DR5 Supplementary Information). We find that most incompatible and slab-fluid mobile elements can be modelled by mixing 4% C_{sf} with 96% of our tonalites.