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SLAB MELTING AS A BARRIER TO DEEP CARBON SUBDUCTION

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Interactions between crustal and mantle reservoirs dominate the surface inventory of volatile elements over geological time, moderating atmospheric composition and maintaining a life-supporting planet¹. Whilst volcanoes expel volatile components into surface reservoirs, subduction of oceanic crust is responsible for replenishment of mantle reservoirs^{2,3}. Many natural, ‘superdeep’ diamonds originating in the deep upper mantle and transition zone host mineral inclusions indicating an affinity to subducted oceanic crust⁴⁻⁷. Our experiments show that the majority of slab geotherms will intersect a deep depression along the melting curve of carbonated oceanic crust at depths of ~ 300 to 700 km, creating a barrier to direct carbonate recycling into the deep mantle. Low-degree partial melts are alkaline carbonatites that are highly reactive with reduced ambient mantle, producing diamond. Many inclusions in superdeep diamonds are best explained by carbonate melt – peridotite reaction. A deep carbon barrier may dominate the recycling of carbon in the mantle and contribute to chemical and isotopic heterogeneity of the mantle reservoir.

Altered oceanic crust incorporates appreciable carbon, which is added by magmatic and hydrothermal processes⁸, and by addition of CO₂ during interaction of basalt with seawater⁹. Together, these alteration processes result in subducting lithosphere that contains an average of ~ 2 wt.% CO₂ in the uppermost volcanic section and 100 - 5000 ppm CO₂ throughout the remaining 7 km of crust⁸. Crustal carbon initially contains a mixture of reduced hydrocarbons⁸ and oxidised carbonates⁹. However, metamorphic re-equilibration of slab carbon with ferric iron and/or oxidizing fluids produced during serpentine dehydration at sub-arc conditions, likely converts most slab carbon to carbonate¹⁰. Some of this carbon is returned to the exosphere in volcanic arcs, but both theoretical¹¹ and experimental¹² studies suggest that a significant quantity of carbon may survive beyond slab dehydration, and be subducted into the mantle.

Carbon is insoluble in mantle silicate minerals¹³ and is stored either as carbonate, carbide or diamond depending on the oxidation state. Under oxidising conditions carbonate lowers the melting point (solidus) of mantle peridotite by some five hundred degrees compared with volatile-free mantle¹⁴. However, at the more reducing conditions prevailing deeper in the upper mantle and transition zone, carbon will be stored as diamond or carbide minerals¹⁵, where it does not appreciably influence melting.

35 Superdeep diamonds originate from depths beneath the lithospheric mantle (≥ 200 km) and are the
36 only direct samples of the deep mantle carbon reservoir. Inclusions in these diamonds are
37 dominated by upper mantle and transition zone minerals, which are mostly associated with
38 subducted mafic lithologies rather than peridotite^{4,7,16}. Many superdeep diamonds are made of
39 isotopically light carbon^{6,7} and, where measured, their inclusions contain isotopically heavy
40 oxygen¹⁷, unambiguously indicating an origin from recycled surface material^{6,7,17}. The elevated
41 trace element abundances of many silicate inclusions suggest crystallization from a low-degree
42 melt, thought to be generated from melting of subducted oceanic crust^{7,18}. Here we examine the fate
43 of subducting carbonated MORB (mid-ocean ridge basalt) as it reaches the transition zone, and the
44 potential for melt-mantle reactions to reproduce superdeep diamonds and their distinctive inclusion
45 assemblages.

46
47 Previous experimental studies have investigated the melting behaviour of carbonated basalt at
48 elevated pressures, but only one extends beyond 10 GPa¹⁹. These studies show a remarkable
49 diversity in melting behaviour making extrapolation to higher pressures difficult. In addition, the
50 bulk compositions employed in previous studies often contain considerably more CO₂ than mean
51 oceanic crust, and fall outside the compositional field of natural MORB rocks (see Methods, EDF1
52 and EDT1). To better understand the melting behaviour of deeply subducted oceanic crust we have
53 determined the melting phase relations of a synthetic MORB composition containing 2.5 wt.% CO₂
54 between 3 and 21 GPa (Methods). Our starting composition replicates the major element
55 composition of basaltic rocks from IODP hole 1256D²⁰ and falls within the range of natural crust
56 compositions²¹ (EDF1).

57
58 We observe subsolidus phase assemblages containing garnet, clinopyroxene, an SiO₂ polymorph,
59 and Ti-rich oxide at all pressures. The carbon component was either CO₂, dolomite, magnesite or
60 magnesite plus Na-carbonate depending on pressure, and the positions of solid carbonate phase
61 boundaries are consistent with previous studies^{22,23}. Near-solidus partial melts are CO₂ bearing
62 silicate melts below 7 GPa, and silica-poor calcic carbonatites above 7 GPa. The alkali component
63 of carbonatite melts increases with pressure (EDF4), and all melts have high TiO₂/SiO₂ (see
64 Methods and extended data items for detailed results).

65
66 The melting temperature of carbonated oceanic crust is tightly bracketed from ~ 3 to 21 GPa (figure
67 1). Melting temperatures increase steadily with increasing pressure until about 13 GPa, when the
68 solidus dramatically drops over a narrow pressure interval by ~ 200 °C. This drop in solidus
69 temperature is caused by a change in clinopyroxene composition towards a more Na-rich

70 composition above 13 GPa due to dissolution of Na-poor pyroxene components into coexisting
71 garnet. Eventually, clinopyroxene becomes so sodium-rich that a coexisting Na-carbonate mineral
72 ($[\text{Na}_{0.97}\text{K}_{0.03}]_{0.33}[\text{Ca}_{0.86}\text{Mg}_{0.11}\text{Fe}_{0.03}]_{0.67}\text{CO}_3$) stabilizes in the subsolidus assemblage, causing the
73 depression along the solidus. The loss of Na-poor clinopyroxene component, and the extended
74 stability of sodic clinopyroxene in the absence of an alternative Na-bearing silicate phase, is
75 consistent with previous studies²⁴. Above 16 GPa the solidus changes little with pressure, remaining
76 at ~ 1150 °C, consistent with the solidus observed in a sodium-rich simplified system where sodic
77 carbonate ($[\text{Na},\text{K}]_{0.33}\text{Ca}_{0.67}\text{CO}_3$) controls melting temperatures²⁵. The major difference between this
78 work and the previous study of carbonated MORB above 8 GPa¹⁹ is the different phase assemblage
79 resulting from the lower and more realistic CO_2 and CaO contents of our bulk composition.
80 Previous bulk compositions with higher CaO contents (EDF1 and 5) are located on the Ca-rich side
81 of the majorite-clinopyroxene tie-line and stabilise aragonite as the carbon-hosting phase, which can
82 incorporate considerable Na_2O . The lower CO_2 content in our bulk composition results in a smaller
83 proportion of carbonate, of which the dominant species is Na-poor magnesite. Thus, sodic
84 clinopyroxene remains stable as an alkali-host, coexisting with stoichiometric Na-carbonate to high
85 pressures.

86

87 The deep solidus depression in carbonated oceanic crust at uppermost transition zone conditions
88 creates a key control on the recycling of mantle carbon. Extrapolation of the range of modern-day
89 oceanic crustal geotherms into the transition zone²⁶ reveals that the majority of slabs will intersect
90 our solidus for carbonated recycled MORB (figure 1), producing carbonatite melt. Given the
91 expected temperature profile in the average subducted slab²⁶ we estimate that melting would occur
92 to depths of at least 7 km into the crustal section. Only the coldest modern day slabs escape the
93 solidus depression and are able to carry their carbonate cargo beyond the transition zone. If ancient
94 slabs were hotter³, it appears likely that carbonate subduction through the transition zone and into
95 the lower mantle has been limited throughout Earth's history. Whilst the natural variability of
96 subducting slabs (e.g. composition, age, temperature) will have created some range in melting
97 behaviour, the depression of the carbonated eclogite solidus will remain an efficient barrier. Thus,
98 direct recycling of carbon into the lower mantle may have been highly restricted throughout most of
99 Earth history, instead being redistributed throughout the upper mantle.

100

101 Carbonatitic melts are predicted to be mobile at mantle conditions due to their low viscosity and
102 ability to wet silicate minerals²⁷, so should percolate out of the slab and infiltrate the overlying
103 peridotitic mantle²⁵. Experiments suggest that below ~ 250 km, ambient mantle oxygen fugacity is

104 reducing, and a free metal phase may be present in the mantle²⁸. Under such conditions carbonate
105 melt is unstable and will reduce to diamond plus oxygen by a ‘redox-freezing’ reaction²⁸ such as:
106



108

109 Thus, the expulsion of carbonatite melts due to melting of oceanic crust along the solidus
110 depression provides an ideal environment for diamond growth across a depth interval of ~ 300 –
111 700 km. We predict that the interaction between MORB-derived carbonatite melt and ambient
112 peridotite is capable of reproducing many of the characteristics of superdeep diamonds and the
113 mineral inclusions that they capture from this depth interval^{4,5}. The most common silicate minerals
114 identified in superdeep diamonds are majorite garnet, and a titanium-bearing, calcium-silicate phase
115 commonly interpreted as retrogressed ‘calcium perovskite’^{4,6,7,18}. Barometric estimates of the
116 crystallization pressures for these majorite inclusions indicate they crystallised between 10 and 16
117 GPa⁵, and inclusions of calcium perovskite are constrained by their chemistry to have formed
118 between ~ 10 and 20 GPa^{6,18}. These pressures are remarkably consistent with the range of pressures
119 at which slab crustal geotherms are predicted to intersect the carbonated solidus depression (figure
120 1).

121

122 Redox reactions in the mantle are complex and involve silicates, many containing iron that exists in
123 both ferrous (Fe^{2+}) and ferric form (Fe^{3+}). To test the melt-mantle interaction model we recreated
124 the infiltration process in a second set of experiments by partially equilibrating a model slab melt
125 with an iron-metal-bearing transition zone peridotite assemblage at 20 GPa (see Methods for
126 details). We observe a reaction zone between the alkaline carbonatite melt and the initial peridotitic
127 assemblage of majorite, wadsleyite, calcium-silicate perovskite and iron metal that consists of
128 sodium-rich majoritic garnet, $\text{Ca}[\text{Si}, \text{Ti}]\text{O}_3$ perovskite, ferrous ringwoodite ($\text{Mg}\# \sim 75$),
129 ferropericlasite ($\text{Mg}\# \sim 0.4$) and diamond (EDF6, 7 and EDT3). We compare the resulting mineral
130 compositions with previous experimental data for peridotite and MORB systems to investigate
131 whether natural inclusion assemblages might preserve a record of mineral-melt reactions.

132

133 The compositions of the majority of superdeep majoritic garnet inclusions are not typical of those
134 expected in either peridotitic or eclogitic bulk compositions (figure 2) and instead lie between these
135 two end-members. Kiseeva et al.¹⁶ described these intermediate compositions as pyroxenitic, and
136 suggested that the transition zone may harbour a large component of this rock type. Our results
137 suggest an alternate explanation. In figure 2 the majoritic garnets produced during the experimental
138 melt-mantle interaction are intermediate between peridotitic and eclogitic compositions, and cover

139 much of the range seen in the diamond inclusions. The chemical imprint imparted by the MORB-
140 carbonatite on the peridotitic mantle is recorded in the inclusions as elevated Ca#, Na and Ti
141 contents alongside depleted Mg#. Our experiments only demonstrate the composition of garnets
142 produced near the beginning of melt-mantle interaction sequence, and we suggest that the
143 intermediate character of the natural inclusions records a snap shot of the infiltration and reaction of
144 slab-derived carbonatite melt with peridotite.

145

146 Experimental Ca-perovskites have high titanium (~ 40 - 60 mol% CaTiO₃) and are essentially
147 magnesium free, features observed throughout the global range of 'Ca-perovskite' inclusions
148 (EDF8). Thus, our reaction experiments reproduce the unique characteristics of diamond-hosted 'Ca-
149 perovskite' inclusions. Crystallisation by reaction between a low-degree carbonated melt and
150 peridotite is also consistent with the extremely elevated trace element contents of diamond-hosted
151 'Ca-perovskites' inclusions²⁴.

152

153 Probably the most abundant inclusions in superdeep diamonds are magnesium-iron oxide
154 ([Mg,Fe]O), which are often interpreted to indicate diamond growth in the lower mantle⁴. However,
155 our experiments demonstrate that ferropericlase can be produced in reactions between carbonatitic
156 melt and reduced mantle peridotite at upper mantle pressures rather than requiring a lower mantle
157 origin²⁹. Figure 3 demonstrates that natural ferropericlase inclusions are almost all iron-rich relative
158 to ferropericlase expected in mantle peridotite, and their compositions form arrays toward higher
159 NiO and lower Na₂O with increasing magnesium number. Our experimental ferropericlase
160 compositions lie at the end of the arrays and are iron-rich because the peridotite starting material
161 was initially iron-saturated. We suggest that, like the majorite inclusions, the array of intermediate
162 ferropericlase compositions record the progressive reaction of carbonatite melt and ambient mantle.

163

164 The melting phase relations of recycled oceanic crust suggest that slabs should undergo melting and
165 loss of carbonate components in the transition zone (figure 4), a process that has considerable
166 implications for the deep carbon cycle. The compositions of diamond-hosted inclusions provide
167 strong evidence of this process and confirm that carbon must survive subduction beyond sub-arc
168 dehydration reactions. We predict that carbon is rarely transported beyond the transition zone and
169 instead refertilises the upper mantle as diamond. Oxidation of diamond-bearing mantle upon
170 upwelling can lead to redox melting¹⁵ beneath the lithosphere and contribute significantly to the
171 generation and geochemical signature of surface lavas. This process also likely contributes to the
172 formation of distinctive chemical and isotopic reservoirs in the mantle³⁰. Superdeep diamonds
173 provide a physical record of carbon recycling above subducting slabs, which can be used to infer

174 the residence time of carbon in the mantle. This residence time is regulated by rates of subduction,
175 convective mantle upwelling and melting beneath the lithosphere, and could occur over a range of
176 timescales, perhaps as short as tens to hundreds of millions of years, suggesting the mantle carbon
177 cycle can be significantly more vigorous than previously estimated^{2,3}.

178

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239

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244

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246 from the literature and wrote the manuscript as part of his Ph.D. studies. M.J.W. and S.C.K.
247 provided training in experimental techniques, assisted during interpretation of results, provided
248 advice and assisted with manuscript preparation in their roles as A.R.T.'s Ph.D. supervisors. R.A.B.
249 provided training and assistance with experimental techniques and sample preparation alongside
250 contributing to the scientific content and preparation of the manuscript.

251

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253 www.nature.com/reprints. The authors declare no competing financial interests. Correspondence
254 and requests for materials should be addressed to A.R.T. (a.r.thomson@ucl.ac.uk).

255 **Main text figure legends**

256 **Figure 1: The melting curve of carbonated MORB (this study) compared to hot and cold**
257 **subduction geotherms²⁶.** The stability fields of carbon-bearing phases are identified in different
258 colours. Experiments performed marked by filled triangles indicating their relationship to the
259 solidus, larger symbols mark solidus brackets. The solidus ledge creates a narrow depth interval
260 where slab temperatures intersect the melting curve, producing a focussed region of melt generation
261 at the top of the transition zone.

262

263 **Figure 2: Composition of majoritic garnet minerals from previous experimental studies,**
264 **inclusions in diamonds and reaction experiments (this study).** The red field outlines the
265 approximate range of peridotitic majorite compositions, the blue field outlines the range of MORB
266 majorites from pressures above the carbonated MORB solidus ledge ($> \sim 9$ GPa). Data and
267 corresponding references for this figure are provided in the online source data file.

268

269 **Figure 3: Composition of ferropericlase minerals from previous experimental studies,**
270 **inclusions in diamonds and reaction experiments (this study).** Blue arrows indicate the
271 compositional evolution expected as melt-mantle interactions progress. Data and corresponding
272 references for this figure are provided in the online source data file.

273

274 **Figure 4: Schematic of the deep mantle carbon cycle as described in the text.** Arrows represent
275 paths and estimates of the relative magnitudes of carbon fluxes. Downwelling slabs dehydrate at
276 sub-arc depths but retain the majority of their carbon cargo. Upon reaching the transition zone they
277 produce carbonatite melts (this study) along the solidus ledge that infiltrate [28] and react with the
278 overlying mantle (this study). This causes diamond production, refertilisation and associated
279 metasomatism of the surrounding mantle. The melting of recycled crust in the transition zone
280 essentially prevents carbon transport into the lower mantle.

281

282

283 **METHODS**

284 **Starting materials.** The starting material for experiments to determine the melting phase relations
285 of carbonated MORB (ATCM1) replicates basalts from the IODP 1256D from the Eastern Pacific
286 Rise²⁰ (the reported composition of IODP 1256D basalts is the average of all analyses presented in
287 table T17 of the cited issue) with an added 2.5 wt.% CO₂ (EDT1). This material was formed by
288 mixing high purity SiO₂, TiO₂, Al₂O₃, FeO, MnO, MgO, Ca₃(PO₄)₂ and CaCO₃, that were fired
289 overnight at temperatures of 400 – 1000 °C, of appropriate weights in an agate mortar under
290 ethanol. This mixture was decarbonated and fused into a crystal-free glass in a one-atmosphere tube
291 furnace by incrementally increasing temperature from 400 to 1500 °C prior to drop quenching into
292 water. Subsequently weighed amounts of CaCO₃, Na₂CO₃ and K₂CO₃ were ground into the glass,
293 introducing the alkali and CO₂ components. After creation the starting material was stored at 120
294 °C to avoid absorption of atmospheric water. Starting material ATCM2 replicates the near-solidus
295 melt composition measured in melting experiments at 20.7 GPa and 1400/1480 °C. This was
296 created by grinding natural magnesite and synthetic siderite with high purity CaCO₃, Na₂CO₃,
297 K₂CO₃, SiO₂, TiO₂, Al₂O₃ and Ca₃(PO₄)₂. Synthetic siderite was created in a cold-seal pressure
298 vessel experiment run at 2 kbar and 375 °C for 7 days. A double Au capsule design containing iron
299 (II) oxalate dehydrate in the inner and a 1:1 mixture of CaCO₃ and SiO₂ in the outer capsule
300 produced a pale beige powder confirmed as siderite using Raman spectroscopy. The material for a
301 sandwich experiment, to ensure near-solidus melt compositions were accurately determined at 20.7
302 GPa, was formed of a 3:1 mixture of ATCM1:ATCM2.

303

304 The transition zone peridotite mineral assemblage in reaction experiments was synthesised at 20.7
305 GPa and 1600 °C for 8 hours from a mixture of KR4003 natural peridotite³¹ with an added 2.5 wt.%
306 Fe metal. In reaction runs the recovered synthetic peridotite was loaded in a second capsule,
307 surrounded by the ATCM2 near-solidus melt composition. Additional reaction-type experiments
308 were performed on ground mixtures of peridotite and melt compositions. In these experiments PM1
309 pyrolite³² was used as the peridotite component and mixed with ATCM2 melt in 9:1, 7:3 and 1:1
310 weight ratios in Fe capsules. A single mixed experiment was performed in a Au capsule and used a
311 starting mix of PM1:Fe:ATCM2 in 16:1:4 molar ratio.

312

313 **Experimental techniques.** High-pressure experiments were performed using a combination of end-
314 loaded piston cylinder (3 GPa) and Walker-type multi anvil (5 - 21 GPa) experiments at the
315 University of Bristol. Piston cylinder experiments employed a NaCl-pyrex assembly with a straight
316 graphite furnace and Al₂O₃ inner parts. Temperature was measured using type D thermocouple
317 wires contained in an alumina sleeve and positioned immediately adjacent to the Au₈₀Pd₂₀ sample

318 capsule that contained the powdered starting material. We assume that the temperature gradient
319 across the entire capsule (< 2 mm) was smaller than $20\text{ }^{\circ}\text{C}^{33,34}$. The hot piston-in technique was
320 used with a friction correction of 3% applied to the theoretical oil pressure to achieve the desired
321 run conditions³⁵.

322

323 Multi-anvil experiments were performed using Toshiba F-grade tungsten carbide cubes bearing 11,
324 8 or 4 mm truncated corners in combination with a pre-fabricated Cr-doped MgO octahedron of 18,
325 14 or 10 mm edge length respectively. The relationship between oil-reservoir and sample pressure
326 for each cell was calibrated at room and high temperature ($1200\text{ }^{\circ}\text{C}$) by detecting appropriate room
327 temperature phase transitions of Bi, ZnTe and GaAs and bracketing transformations of SiO_2
328 (quartz-coesite and coesite-stishovite), Mg_2SiO_4 (α - β and β - γ) and CaGeO_3 (garnet-perovskite).
329 Calibrations are estimated to be accurate within ± 1 GPa. In all experiments desired run pressure
330 was achieved using a slow, Eurotherm controlled, pressure ramp of ≤ 50 tonnes/hour. Experiments
331 were heated after high pressure was reached with high-temperatures generated using stepped
332 graphite (18/11 cell) or straight LaCrO_3 furnaces (14/8 and 10/4 cells) and monitored with type C
333 thermocouple wires. Two 10/4 experiments, performed during a period of repeated LaCrO_3 heater
334 failures, used rolled $40\text{ }\mu\text{m}$ thick Re furnaces. Temperature was quenched by turning off the furnace
335 power prior to a slow decompression ramp (half the rate of experiment compression) to ambient
336 conditions. Samples were contained in Au capsules unless temperatures exceeded its thermal
337 stability, in which case $\text{Au}_{80}\text{Pd}_{20}$ or $\text{Au}_{75}\text{Pd}_{25}$ capsules were used. Run durations all exceeded 600
338 minutes and are reported in extended data tables 2 and 3. Temperature uncertainties were believed
339 to be less than ± 20 , 30 or $50\text{ }^{\circ}\text{C}$ for 18/11, 14/8 and 10/4 cells respectively^{36,37}.

340

341 Recovered samples were mounted longitudinally in epoxy, polished under oil and repeatedly re-
342 impregnated with a low viscosity epoxy (Buelher EpoHeat) in order to preserve soft and water-
343 soluble alkali carbonate components present in run products.

344

345 **Analytical techniques.** Polished and carbon coated run products were imaged in backscatter
346 electron mode (BSE) using a Hitachi S-3500N SEM microscope with an EDAX Genesis energy
347 dispersive spectrometer to identify stable phases and observe product textures. Subsequently,
348 wavelength dispersive spectroscopy (WDS) was performed using the Cameca SX100 Electron
349 Microprobe or the Field Emission Gun Jeol JXA8530F Hyperprobe at the University of Bristol to
350 achieve high precision chemical analyses of run products. Analyses were performed using an
351 accelerating voltage of 15 or 12 kV on the respective instruments, with a beam current of 10 nA.
352 Calibrations were performed during each session using a range of natural mineral and metal

353 standards and were verified by analysing secondary standards (as in [6]). Silicate phases were
354 measured using a focused electron beam whereas carbonates and melts were analysed using an
355 incident beam defocussed up to a maximum size of 10 μm . Count times for Na and K were limited
356 to 10 seconds on peak and 5 seconds on positive and negative background positions. Peak count
357 times for other elements were 20 - 40 seconds. Additional analyses of the calcium perovskite phases
358 grown during reaction experiments, measuring only SiO_2 and MgO content, were made using the
359 Jeol instrument at 5 kV and 10 nA to ensure reported MgO contents were not influenced by
360 secondary fluorescence from surrounding material.

361

362 The identity of experimental-produced minerals was determined using Raman spectroscopy as a
363 fingerprint technique. Spectra were collected using a Thermo Scientific DXRxi Raman microscope
364 equipped with an excitation laser of either 455 or 532 nm.

365

366 **Choice of bulk composition and comparison with previous studies.** Studies that investigate the
367 alteration of oceanic crust have demonstrated that carbon incorporation does not simply occur by
368 the addition of a single carbonate species to MORB⁹. It instead appears to occur by a complex
369 amalgamation of hydrocarbon and graphite deposition related to hydrothermal fluxing above
370 magma chambers at the mid-ocean ridge⁸ and underwater weathering^{9,38-40} where seawater-derived
371 CO_2 reacts with leached crustal cations, often in veins. It is believed that the quantity of biotic
372 organic carbon in the crustal assemblage is negligible compared with abiotic organic compounds
373 and inorganic carbonates⁸. These processes result in a layered crustal assemblage that, in the
374 uppermost few hundred metres can contain up to a maximum of 4 wt.% CO_2 in rare cases^{9,39} but
375 more commonly < 2 wt.% CO_2 [8, 9, 39]. Beneath 500 m depth the carbon content drops to between
376 100 and 5000 ppm CO_2 throughout the remainder of the 7 km thick basaltic section⁸, and is mostly
377 organic hydrocarbon species. The upper 300 m are regularly altered and can be generally thought to
378 have compositions similar to the altered MORB rocks analysed by Kelley et al.⁴¹. Deeper portions
379 of the MORB crust retain their pristine MORB compositions. It is therefore apparent that
380 carbonated eclogite bulk compositions used in previous studies, where at least 4.4 wt.% CO_2 was
381 added to an eclogite by addition of ~ 10 wt.% carbonate minerals, may not be good analogues of
382 naturally subducting crustal sections. The compositions of these starting materials from previous
383 studies^{19,42-46} can be found in EDT1. We do not include the composition of the starting material
384 used by [47] or [48] as these studies were conducted in simplified chemical systems so are not
385 directly comparable with these natural system compositions.

386

387 However, as some of the previous studies rightly identify and discuss, the composition of deeply
388 subducted MORB is not likely to be the same as that entering the subduction system. One process
389 widely believed to alter the composition of downwelling MORB is sub-arc slab dehydration. PT
390 paths of subducted slabs²⁶ can be compared with experimental studies of hydrous, carbonated and
391 H₂O-CO₂-bearing eclogite compositions^{12,24,42,43,49} and thermodynamic models^{11,50} to conclude that
392 slabs experience dehydration at sub-arc conditions (i.e. 1-5 GPa) but will generally not reach high
393 enough temperatures to undergo melting. Therefore, they will by-in-large retain their carbon
394 components although some fraction may be lost by dissolution into aqueous fluids^{51,52}. It is believed
395 that sub-arc dehydration is capable of removing SiO₂ from the subducting assemblage, and previous
396 carbonated MORB compositions were therefore designed to be significantly silica undersaturated
397 (relative to fresh/altered MORB)^{19,43-45}. Whilst studies^{e.g. 53-56} do indicate that SiO₂ can become
398 soluble in H₂O at high-pressures, they infer that the solubility of silica in hydrous fluids only
399 exceeds ~ 1 wt.% at T > 900 °C at 1 GPa (higher T at higher P). In contrast, slab dehydration occurs
400 on all prograde slab paths at T < 850 °C. Additionally, Kessel et al.⁵⁷ measured the composition of
401 quenched hydrous fluids coexisting with MORB at 4 GPa and 800 °C; their data indicates a
402 maximum of ~ 12 wt.% SiO₂ can dissolve in the fluid. Given that there should be considerably less
403 than 10 wt.% H₂O (more likely << 5 wt.% H₂O) in subducting assemblages this suggests a
404 maximum SiO₂ loss in subducting MORB lithologies of ~ 0.6 – 1.2 wt.%. The compositions used in
405 previous studies have SiO₂ depletions ranging from 3 wt.% up to, more commonly, 6 – 10 wt.%
406 SiO₂ relative to MORB.

407

408 We further investigated the effect of oceanic crust alteration and sub-arc dehydration on the
409 composition of subducted MORB rocks by compiling a dataset of altered MORB⁴¹ and exhumed
410 blueschist, greenschist and eclogite facies rocks from exhumed terrains worldwide to compare them
411 with fresh MORB²¹, our starting material and previous starting materials. We then assess the
412 relevance of our starting material based on the composition of natural MORB rocks, rather than
413 using models of the subduction process that contain few observable constraints. Results of this
414 comparison are plotted in EDF1. This analysis confirms that relative to fresh MORB, altered
415 MORB and exhumed crustal rocks are somewhat depleted in SiO₂, up to a maximum of 6 wt.%
416 SiO₂ in the most extreme case, but more commonly 0 – 3 wt.% SiO₂. Thus, many previous starting
417 materials are too silica undersaturated to be good analogues of subducting MORB. Furthermore,
418 this analysis reveals that altered and exhumed MORB are not enriched in CaO compared with fresh
419 MORB, if anything they actually contain lower CaO on average. In contrast, all previous starting
420 materials are enriched in CaO compared with fresh MORB. This is because most previous studies
421 introduced the carbon component to their experiment by adding ~ 10 wt.% calcite to an eclogite-

422 base composition. We note that SLEC1⁴³ was not created in this manner, but instead this
423 composition falls far from the MORB field as the authors used an eclogite xenolith erupted by a
424 Hawaiian volcano as a base material. By plotting the position of the maj-cpx join, defined by the
425 composition of our experimental phases plotted in EDF5, onto EDF1a we demonstrate that our bulk
426 composition (ATCM1), ALL-MORB²¹, the vast majority of the fresh MORB field, altered⁴¹ and
427 exhumed MORB samples fall on the CaO-poor side of this join, i.e. on the Mg+Fe rich side.
428 Therefore, magnesite will be the stable carbonate phase in these compositions at high pressure
429 (above dolomite breakdown). In contrast, all previous bulk compositions plot on the Ca-rich side of
430 this join, and therefore in a different phase field to the overwhelming majority of subducted MORB.
431 This difference causes a significant difference in the phase relations of our starting material relative
432 to those used in previous studies.

433

434 We acknowledge that no single bulk composition can be a perfect analogue for the entire range of
435 subducting MORB compositions, however ATCM1 is a good proxy for sections of the MORB crust
436 between ~ 300 m and 7 km depth that have unaltered major element compositions and low CO₂
437 contents. Additionally, ATCM1 remains a better analogue for the uppermost portions of the MORB
438 crust than starting materials employed in previous studies because its CO₂ content is within the
439 range of natural rocks whilst it is also not oversaturated in CaO or over depleted in SiO₂. This is
440 despite it falling towards the SiO₂ rich end of the compositional spectrum of subducting MORB
441 rocks.

442

443 **Slab fO_2 and carbonate survival to transition zone conditions.**

444 Recent experiments have suggested that carbonate in eclogitic assemblages may be reduced to
445 elemental carbon, either graphite or diamond, at depths shallower than 250 km⁵⁸. However,
446 subducting slab geotherms are much colder than the experimental conditions investigated by this
447 study, and additionally they are believed contain significant ferric iron that is further increased
448 during de-serpentinisation¹⁰. Indeed, several observations of carbonate inclusions in sub-
449 lithospheric diamonds^{e.g. 6,7,59} require that slab carbon remains oxidised and mobile until diamond
450 formation, far deeper than 250 km. Given the numerous observations from natural diamond
451 samples, the general uncertainty in the mantle's fO_2 structure and the lack of any conclusive
452 experimental evidence that subducting carbon becomes reduced prior to reaching the transition zone
453 we posit that nearly all subducting carbon is stable as carbonate throughout the upper mantle in
454 subducting MORB assemblages.

455

456 **Experimental Results – Carbonated MORB melting.** EDT2 presents the run conditions,
457 durations and phase proportions in all carbonated MORB melting experiments, which are also
458 summarised in EDF2. Phase and melt compositions are presented in the supplementary tables.
459 Phase proportions are calculated by mass balance calculations that use the mean composition of
460 each phase as well as the reported 1σ uncertainty in this mean as inputs. We note that the 1σ
461 uncertainty for some oxides in garnet and clinopyroxene minerals occasionally exceeds 1 wt.%,
462 although it is normally much smaller than this. These large uncertainties are a function of the small
463 crystal sizes present in some runs, and not a function of sluggish reaction kinetics. Phase proportion
464 calculations were run in a Monte Carlo loop of 10,000 calculation cycles where a varying random
465 error was added to each oxide in each mineral phase during each iteration. Overall the distribution
466 of varying random errors for each oxide form a Gaussian distribution with standard deviation equal
467 to the reported 1σ uncertainty of measurements. The reported proportions are the numerical mean of
468 all calculation cycles and the r^2 value reports the average squared sum of residuals. Low r^2 values
469 indicate that chemical equilibrium is likely to have been achieved and that mineral and melt
470 compositions have been accurately determined.

471

472 Representative BSE images of the polished experiments are shown in EDF3. Garnets in
473 experiments at all pressures contain abundant SiO_2 inclusions. In subsolidus experiments the
474 number of inclusions increases and the definition of mineral boundaries deteriorates, which makes
475 accurate analysis of garnet compositions increasingly challenging. In supersolidus runs, garnet
476 minerals adjacent, or near to, carbonatite melt pools have well defined edges and contain fewer
477 inclusions. However, far from quenched melts the textures of garnets remain small and pervasively
478 filled with inclusions, indicating the influence of melt fluxing on mineral growth. With increasing
479 pressure, garnets become increasingly majoritic, with increasing quantities of octahedral silicon.

480

481 Clinopyroxene was observed in all subsolidus experiments, as euhedral crystals that are often
482 spatially associated with the carbon-bearing phase. Cpx abundance falls with increasing pressure
483 and their compositions becoming increasingly dominated by sodic components (jadeite, aegerine
484 and $\text{NaMg}_{0.5}\text{Si}_{2.5}\text{O}_6$) at high pressure (EDF5). Cpx only disappears from the stable phase
485 assemblage in supersolidus experiments at 20.7 GPa. SiO_2 is observed in all runs and are small,
486 often elongated tabular-shaped crystals. An oxide, either TiO_2 at low pressure or an Fe-Ti oxide
487 above 13 GPa (as in [24]) are observed in all subsolidus runs.

488

489 The carbon-bearing phase in subsolidus experiments changes with increasing pressure. At 3 GPa
490 CO_2 , marked by the presence of voids in the polished sample, is stable. This converts to dolomite at

491 7.9 GPa, consistent with the position of the reaction $2cs + dol = cpx + CO_2$ [22]. Beyond ~ 9 GPa
492 dolomite becomes unstable and breaks down into magnesite + aragonite²³. Therefore, because the
493 ATCM1 bulk composition lies on the Mg+Fe²⁺-rich side of the garnet-cpx join (EDF5 and EDF1a),
494 magnesite replaces dolomite as the carbon host in the experimental phase assemblage. This differs
495 from experiments in previous studies, where aragonite was dominant because bulk compositions
496 fall on the opposite side of the garnet-cpx join. It is clear from the ternary diagrams (EDF5) that
497 while the tie-line between garnet and cpx remains magnesite and aragonite cannot coexist in a
498 MORB bulk composition. Finally, at pressures above 15 GPa, Na carbonate becomes stable in the
499 subsolidus phase assemblage. This is chemographically explained by the rotation of the garnet-cpx
500 tie-line with increasing pressure (EDF5). Its appearance can also be justified as a necessary host of
501 sodium at increasing pressure, since aside from clinopyroxene there is no other Na-rich phase stable
502 on the Mg+Fe side of the maj-cpx join.

503

504 The appearance of silicate melt, containing dissolved CO₂ (estimated by difference), defines the
505 solidus at 3 GPa. This may initially appear to contradict the results of some previous studies, which
506 find carbonatite melts are produced near the solidus of carbonated eclogite at pressures lower than 7
507 GPa^{e.g. 43,45,46}. However, this is easily explained by the differences in CO₂ and SiO₂ content used in
508 these studies. The higher CO₂ and lower SiO₂ contents of previous studies stabilise carbonate melt
509 to lower temperatures relative to silicate melts. Indeed, we note that our results are consistent with
510 those of Yaxley and Green⁴⁴ and Hammouda⁴² (the two previous studies with the least depleted
511 SiO₂) who also observed near solidus melts below 5 GPa were basaltic to dacitic silicate melts
512 containing dissolved CO₂. The results of Kiseeva et al.¹⁹ are not entirely self-consistent, in that at
513 some pressures between 3.5 and 5.5 GPa they observed silicate melts prior to carbonate melts (4.5
514 and 5 GPa), whereas this relationship is sometimes reversed (5 GPa in AuPd capsules) or both melts
515 were observed together (3.5 GPa). The observation of two immiscible melts in previous studies
516 likely reflects the maximum CO₂ solubility in silicate melts. Since our bulk composition has less
517 CO₂, akin to natural rocks, we do not observe liquid immiscibility.

518

519 In all experiments above 7 GPa near-solidus melt compositions are carbonatitic and essentially
520 silica-free. This result is notably different from [19] who reported that near solidus melts were a
521 mixture of silicate, carbonated silicate and carbonatite melts. We believe this contrast is caused by
522 the interpretation of experimental run textures. Whereas [19] identified regions of fine-grained
523 material consisting of mixtures of stable phases from elsewhere in the capsule as quenched melts,
524 we have not followed the same interpretation of these features. Although we do recognise similar
525 features in some run products we have interpreted these features as a consequence of poor crystal

526 growth in regions far from the influence of melt fluxing. In all supersolidus experiments we
527 observed regions of carbonatite material (typically < 1 wt.% SiO_2) that is fully segregated from
528 surrounding silicate minerals and possesses a typical carbonate-melt quench texture (EDF3).
529 Silicate minerals in close-proximity to these melt pools are larger than those elsewhere in the same
530 experiment, have well-defined crystal boundaries and contain few inclusions. Therefore, we
531 attribute the variable texture and regions of fine-grained material present in experiments to the
532 location of melt within experiments, which has a tendency to segregate to isolated regions of
533 capsules under influence of temperature gradients. Although melt segregation occurs in all
534 supersolidus experiments, the efficiency of segregation and size of melt pools significantly
535 increases with rising temperature above the solidus. EDF4 shows the highly systematic evolution of
536 the melt compositions reported from our study with increasing pressure, strongly supporting our
537 interpretations.

538

539 Carbonatite melts are calcic, $\text{Ca\#} > 0.5$ ($\text{Ca\#} = \text{Ca}/[\text{Ca}+\text{Mg}+\text{Fe}]$), despite subsolidus carbonates
540 being dominated by magnesite (EDF4). Melts have high concentrations of TiO_2 (typically 1 - 3.5
541 wt.%), P_2O_5 (0.4 - 1.5 wt.%) and K_2O (0.3 - 1.5 wt.%) and a variable Mg\# (0.33 - 0.7 defined as
542 $\text{Mg}/[\text{Mg}+\text{Fe}]$). The alkali content of melts, strongly dominated by Na_2O due to the bulk
543 composition, increases with pressure (from 1 to ~ 15 wt.% Na_2O at 7.9 and 20.7 GPa respectively;
544 EDF4). This increasing Na_2O content is driven by the decreasing compatibility of Na_2O in the
545 residual mantle phase assemblages as the abundance of stable clinopyroxene falls. At 20.7 GPa the
546 melt composition, as evidenced both by constant phase proportions and consistent melt/majorite
547 compositions, remains constant over a temperature interval of ~ 350 °C above the solidus. It is only
548 when temperature reaches 1530 - 1600 °C (runs #16 and #31) that the silica content of the melt
549 begins to increase (to 8.7 wt.%) and CO_2 content falls as melts start to become silica-carbonatites.

550

551 One experiment (#33) aimed to verify that measured low-degree melt compositions are accurate,
552 and are not affected by analytical problems related to the small size of melt pools, was conducted at
553 20.7 GPa. In this experiment the abundance of carbonate melt was increased by adding a mix
554 replicating the low degree melt composition ATCM2 to ATCM1 in a mass ratio of 1:3. If the
555 composition of low-degree melts has been accurately determined in 'normal' experiments then this
556 addition will have a negligible affect on phase relations or the compositions of the garnet, SiO_2 or
557 melt; it would simply increase the melt abundance. The result of this experiment has a similar
558 texture to all other experiments, where carbonatite melt segregates to one end of the capsule and is
559 adjacent to large, well-formed majoritic garnets. The far end of the capsule has a much smaller
560 crystal size, crystals have ragged edges, garnets are full of inclusions and SiO_2 is present along

561 grain-boundaries and triple junctions (EDF3h). Mineral and melt compositions, although not
562 exactly identical, are similar to those measured in ‘normal’ experiments (to achieve identical
563 compositions an iterative approach would be required that was not deemed to be necessary) thus
564 confirming that near-solidus melt compositions have been accurately determined. The presence of
565 fine-grained material away from segregated melt also acts to further confirm our hypothesis
566 regarding the vital importance of melt presence for growing large crystals during experiments.

567

568 **Subsolidus carbonate species at high pressure.**

569 Comparing our starting material and results with those of previous studies using ternary and
570 quaternary projections (EDF5) reveals that it is not possible for both magnesite and aragonite to
571 coexist alongside majorite and clinopyroxene due to stable mineral phase fields (see above). Thus,
572 in Mg-Fe dominated compositions, such as our starting material, magnesite is the stable carbonate
573 at high-pressure subsolidus conditions. Whereas, in Ca dominated compositions aragonite will be
574 the stable carbonate beyond the pressure of dolomite dissociation. Natural subducting MORB
575 compositions, which contain, at most, a similar quantity of CO₂ to our bulk composition¹¹, almost
576 all lie on the Ca-poor side of the majorite-clinopyroxene join (EDF1 and EDF5). In this situation, as
577 our experiments demonstrate, cpx remains an important Na-host in MORB assemblages to high
578 pressures alongside [Na,K]_{0.33}Ca_{0.67}CO₃ structured carbonate. Ca-rich compositions containing
579 subsolidus CaCO₃ experience different phase relations because aragonite can dissolve significant
580 Na₂O and so is the sole Na-host in these compositions. We conclude that because the majority of
581 natural MORB rocks fall on the Mg+Fe side of the maj-cpx join, like our bulk composition, that the
582 phase relations determined in this study are applicable to the case of natural subduction. Therefore,
583 the melting point depression we observe along the carbonated MORB solidus at uppermost
584 transition zone pressures is generally applicable to subducted oceanic crust.

585

586 **Experimental Results – Melt-mantle reactions.** Without the influence of slab-derived melts the
587 anhydrous transition zone peridotite assemblage at 20.7 GPa and 1600 °C (experiment G168 and
588 G176) is dominated by Na-poor majorite and wadsleyite (Mg# = 0.90) (EDF6, EDT3 and
589 supplementary tables). Upon reaction with the near solidus alkaline carbonatite defined during
590 melting experiments, ATCM2, a clearly defined reaction zone is observed between this ambient
591 peridotite assemblage and the infiltrating melt (EDF6). The products of this reaction are garnet
592 containing a notable Na₂X²⁺Si₅O₁₂ majorite component, Ca(Si,Ti)O₃ perovskite, ringwoodite,
593 ferropericlase and diamond. All of these phases were identified using Raman spectroscopy (EDF7)
594 and their compositions are presented in the supplementary tables. Raman spectroscopy alone, which
595 was performed prior to any sample polishing using diamond-based products, confirms the creation

596 of diamond during these reactions. We have not observed diamond using SEM techniques and
597 believe that it resides as sub-micron sized inclusions in the various reaction-product minerals where
598 it is seen by spectroscopic methods. The experiments performed on intimately mixed powders of
599 melt and pyrolite also form the same phase assemblages (EDT3) and mineral compositions from
600 those runs are also presented in the supplementary tables.

601

602 We observed the reaction products as new crystals floating in the residual carbonatite melt and/or
603 nucleated on the relics of the peridotite assemblage, thus creating zoned minerals. We have
604 demonstrated that the composition of majorite minerals crystallising during the reactions lie
605 between those expected for peridotitic and eclogitic minerals at a similar pressure and possibly
606 explain intermediate-composition diamond-hosted majorites (figure 2). We suggest that the full
607 range of intermediate inclusion compositions might be created by the gradual shift in phase
608 compositions, from those we observe towards more peridotitic minerals as the melt composition
609 reacts with increasing quantities of mantle material. Additionally we have shown that the
610 compositions of calcium perovskite (EDF8) and ferropericlasite (figure 3) formed during the
611 reactions are consistent with diamond-hosted minerals of those species. Further experiments, across
612 the solidus ledge and into the uppermost lower mantle pressure range are required to test whether
613 melt-mantle interactions account for all diamond-hosted inclusions.

614

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676 **Extended Data Captions**

677 **Extended Data Figure 1: Comparison of experimental compositions with natural rocks.**

678 ‘Fresh’ MORB rocks (red field), ALL-MORB ([21]; red circle), altered MORB rocks ([41]; pale
679 blue circles), exhumed blueschist, greenschist and/or eclogitic rocks (yellow circles) and starting
680 material from this (dark blue circle) and previous studies (green circles) of carbonated MORB
681 compositions. In **a** rocks altered MORB and exhumed rock compositions that fall on the Mg-Fe side
682 of the maj-cpx join from EDF5 plot below the dashed line, compositions that lie on the Ca side of
683 this join are plotted with as orange circles with yellow outlines or purple circles with blue outlines
684 and sit above the dashed curve. This confirms that magnesite will be the stable carbonate phase at
685 high pressure in vast majority of natural crustal rocks, as is the case for ATCM1. Data and
686 corresponding references for this figure are provided in the online source data file.

687

688 **Extended Data Figure 2: Experimental results/phase diagram and interpreted solidus**

689 **position.** The reactions $\text{cpx} + \text{CO}_2 = \text{dol} + 2\text{cs}$ and $\text{dol} = \text{mag} + \text{arag}$ are from [22] and [23] respectively.
690 The upper left curve is the anhydrous MORB solidus. N.B due to temperature gradients in
691 experiments at 8 GPa, a small quantity of dolomite is observed coexisting with melt in one
692 experiment above the solidus, present at the cold end of the capsule.

693

694 **Extended Data Figure 3: BSE images of experimental products. a** 7.9 GPa, 1250 °C, **b** 7.9 GPa,

695 1350 °C, **c** 13.1 GPa, 1350 °C, **d** 13.1 GPa, 1450 °C, **e** 20.7 GPa, 1100 °C, **f** 20.7 GPa, 1480 °C, **g**
696 20.7 GPa, 1600 °C, **h** sandwich experiment, 20.7 GPa, 1400 °C. The scale bar in each image is 10
697 μm .

698

699 **Extended Data Figure 4: Composition of experimental melts from this study.** Experimental
700 melts from selected previous studies marked with semi-transparent greyscale symbols. The effects
701 of increasing pressure, temperature and the effect of contamination due to partial analysis of silicate
702 minerals surrounding small melt pools are shown in **b**.

703

704 **Extended Data Figure 5: The composition of experimental phases from this study projected**

705 into two quaternary plots. **a** [Ca]-[Mg+Fe²⁺]-[Si+Ti]-[Na+K] and **b** [Mg+Fe²⁺]-[Ca]-[Al+Fe³⁺]-
706 [Na+K]. In both diagrams the grey fields are the compositional data projected onto the basal
707 ternary. The red field is the range of natural MORB compositions projected onto the basal ternary.

708 The yellow star plotted in the 4-component system and projected onto the basal ternary is ATCM1
709 (our bulk composition) whilst the black stars are bulk compositions from previous studies²⁵⁻²⁷.

710

711 **Extended Data Figure 6: BSE images of reaction experiments. a, b G169 and c, d G177.** In both
712 experiments a reaction zone and remaining carbonatite melt surrounds the unreacted peridotite
713 region. **a** An overview of G169. **b** A close up of the reaction in G169 containing newly crystallised
714 calcium perovskite, majorite, ferropericlasite and ringwoodite minerals. **c** A close up of the reaction
715 products in G177, which consist of small bright calcium perovskites, new majorite that is often
716 observed as a rim on relic peridotitic garnet and ringwoodite. **d** An overview of G177.

717

718 **Extended Data Figure 7: Raman spectra of minerals from reaction experiment G177**
719 **measured using a blue 455 cm⁻¹ excitation laser.** The position of the main peaks in each collected
720 spectrum have been labelled with their shift from the excitation laser in cm⁻¹.

721

722 **Extended Data Figure 8: Comparison of diamond-hosted calcium perovskite inclusions with**
723 **experimental mineral compositions in MgO vs Ti# space.** Data and corresponding references for
724 this figure are provided in the online source data file.

725

726 **Extended Data Table 1: Starting materials used in this and previous studies.** Ca# =
727 Ca/[Ca+Mg+Fe]. Mg# = Mg/[Mg+Fe].

728

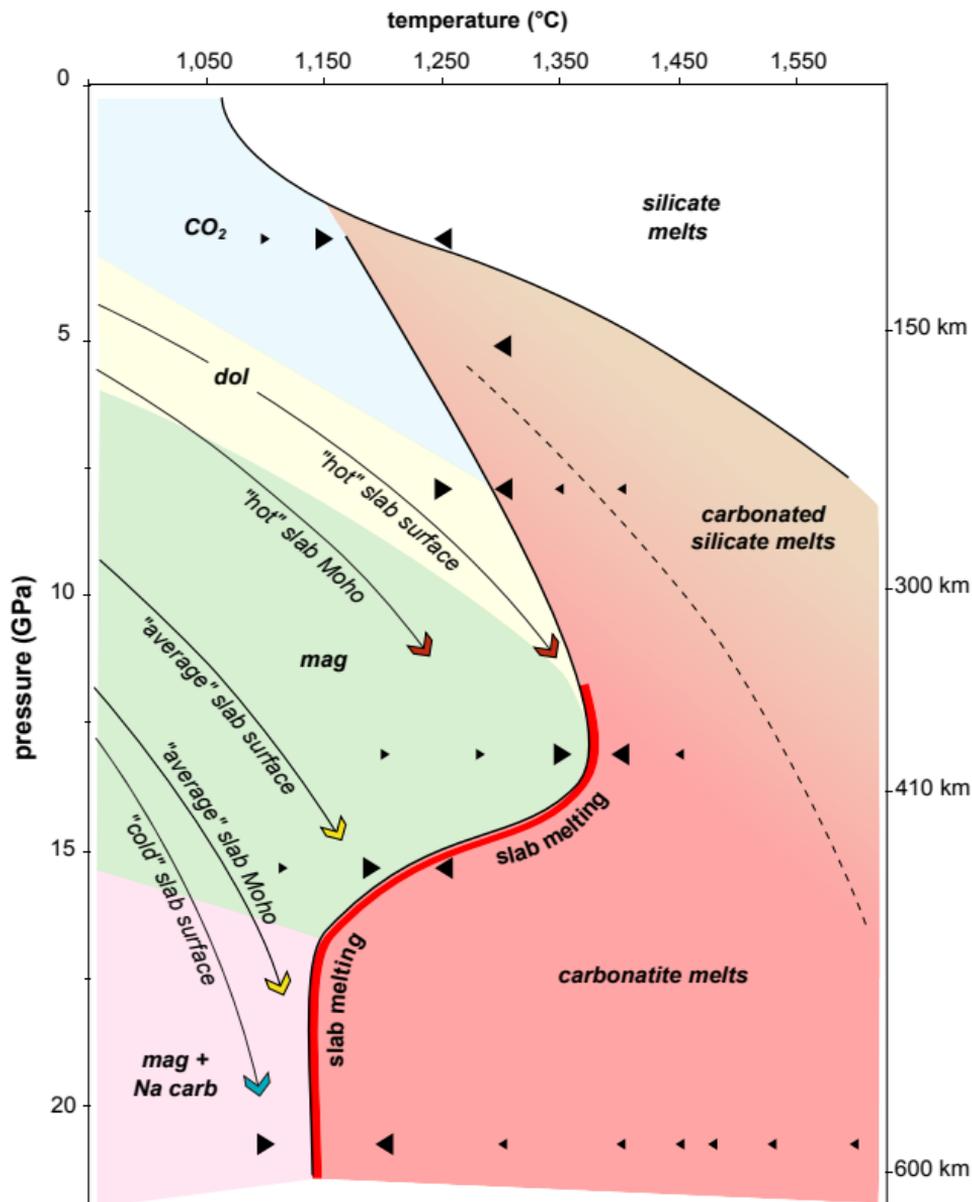
729 **Extended Data Table 2: Summary of run conditions and products for carbonated MORB**
730 **melting experiments.** Mass balance calculations were performed as described in the supplementary
731 information. Mineral abbreviations are as follows: gt = garnet; cpx = clinopyroxene; cs = coesite;
732 rut = rutile; maj = majoritic garnet; st = stishovite; FeTi oxide = iron-titanium-rich oxide phase; SM
733 = silicate melt; CM = carbonatite melt; dol = dolomite; mag = magnesite; Na carb = sodic
734 carbonate. Phase proportions are in wt. %.

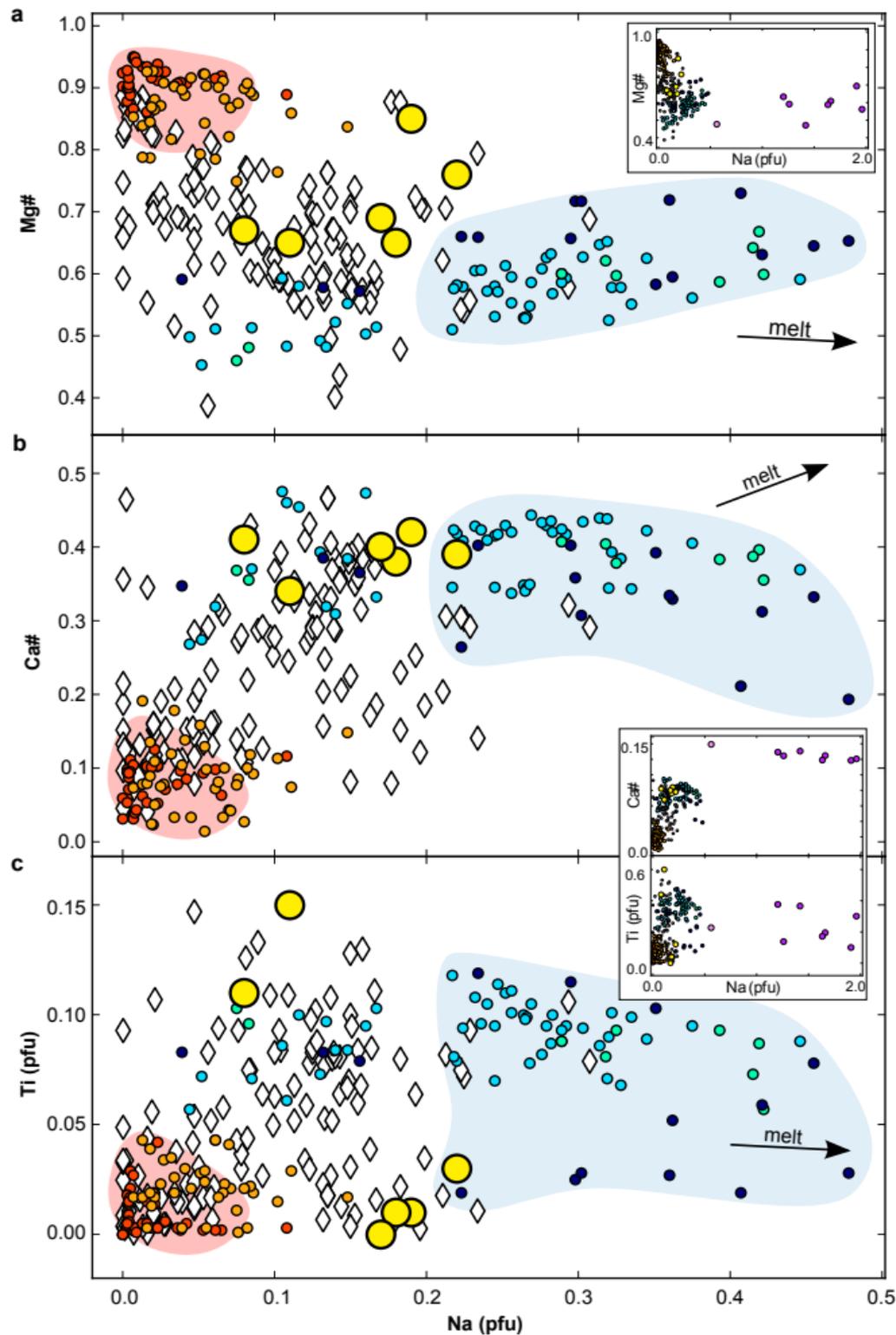
735

736 **Extended Data Table 3: Summary of reaction experiments run conditions and experimental**
737 **products**

738

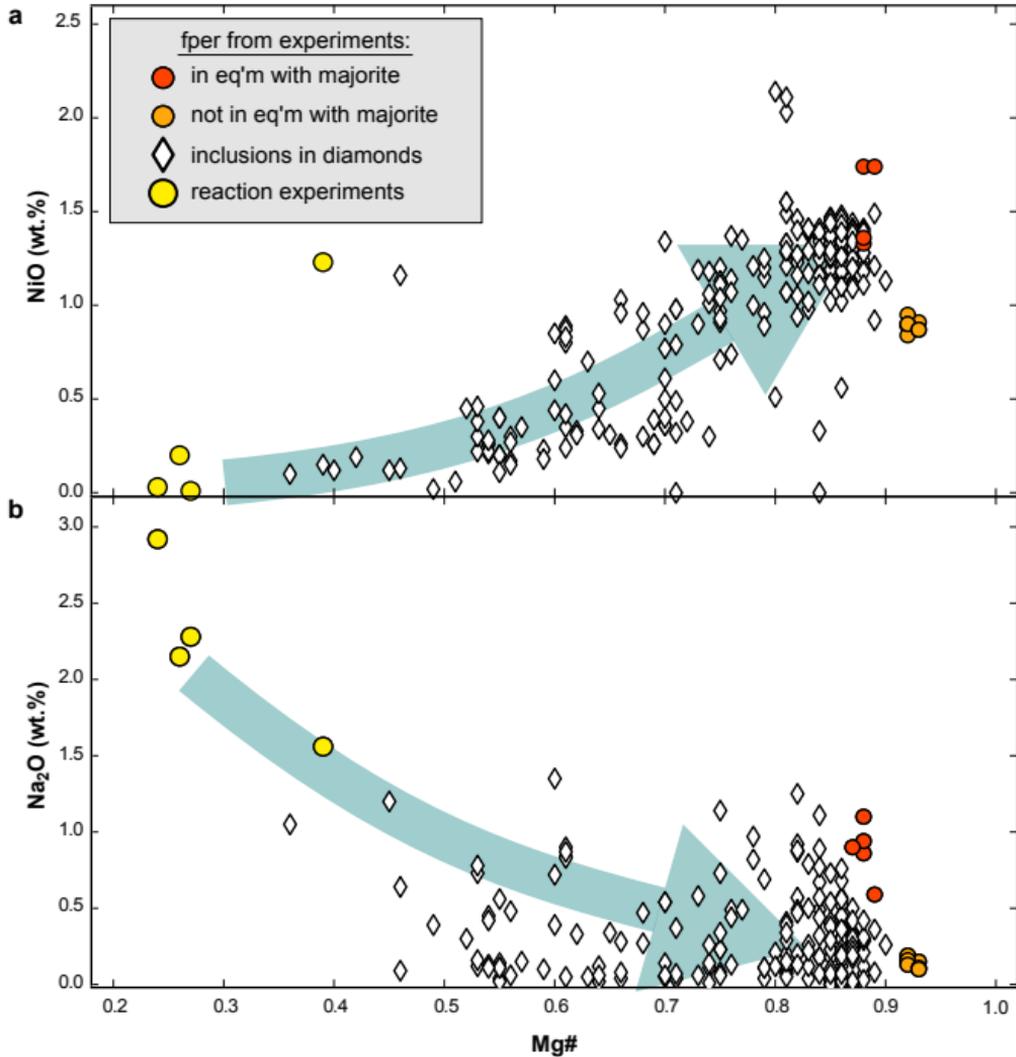
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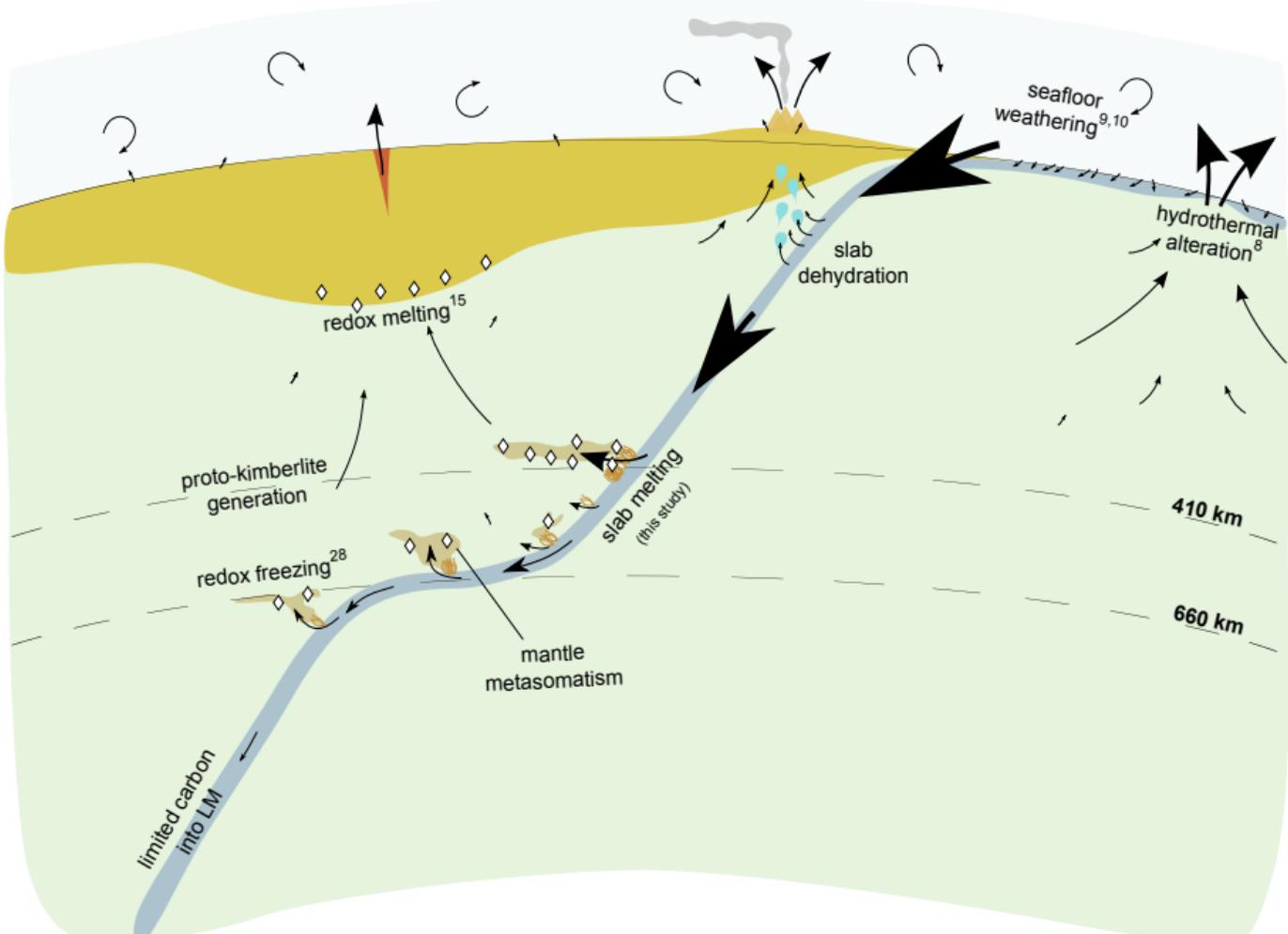


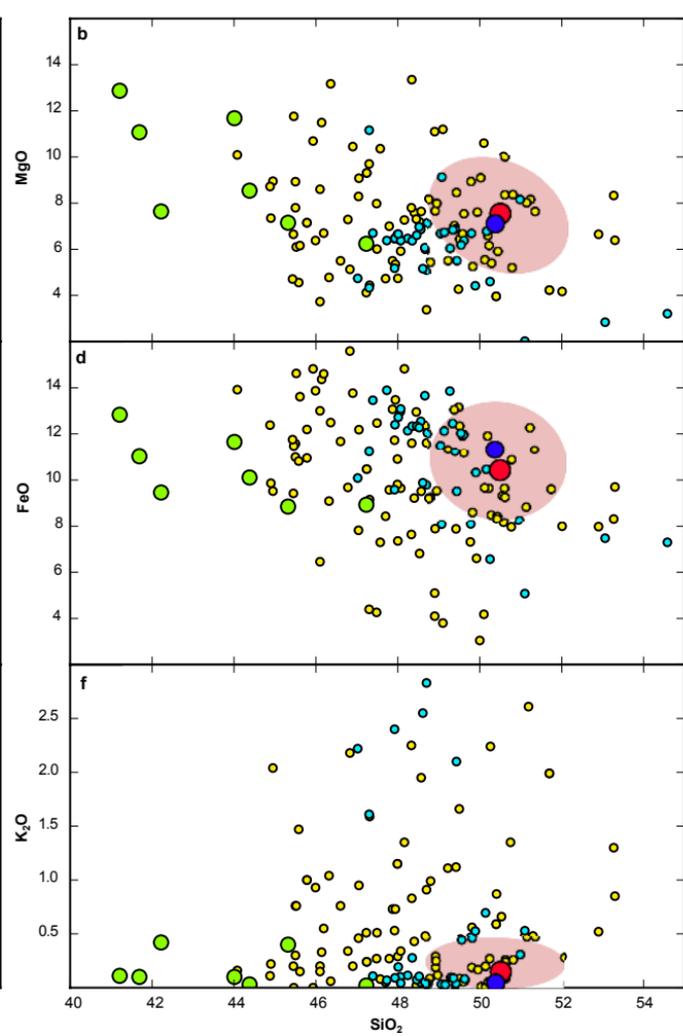
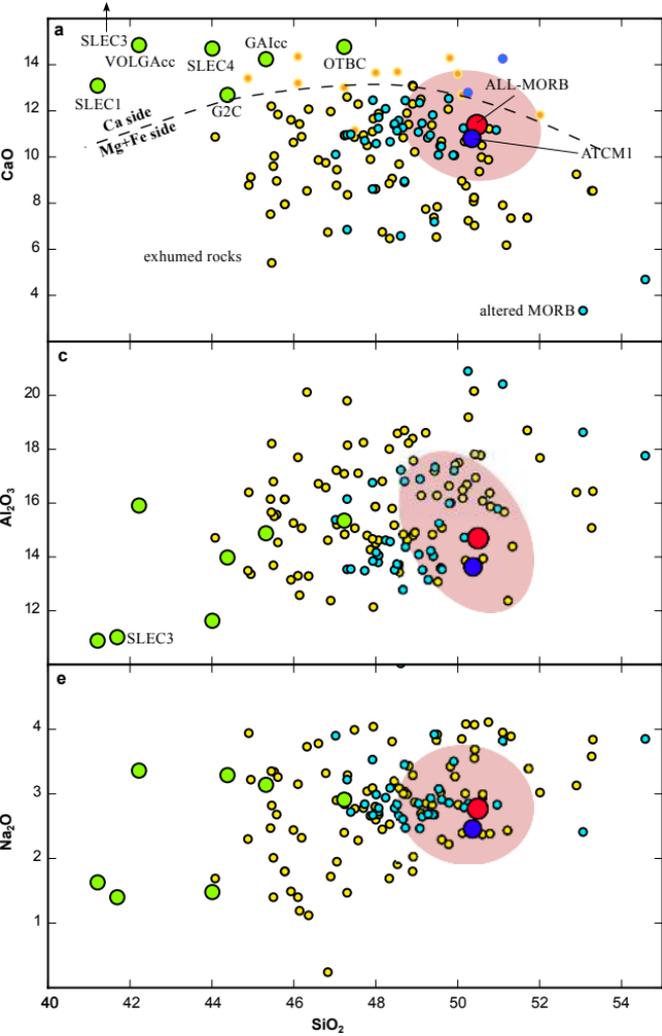


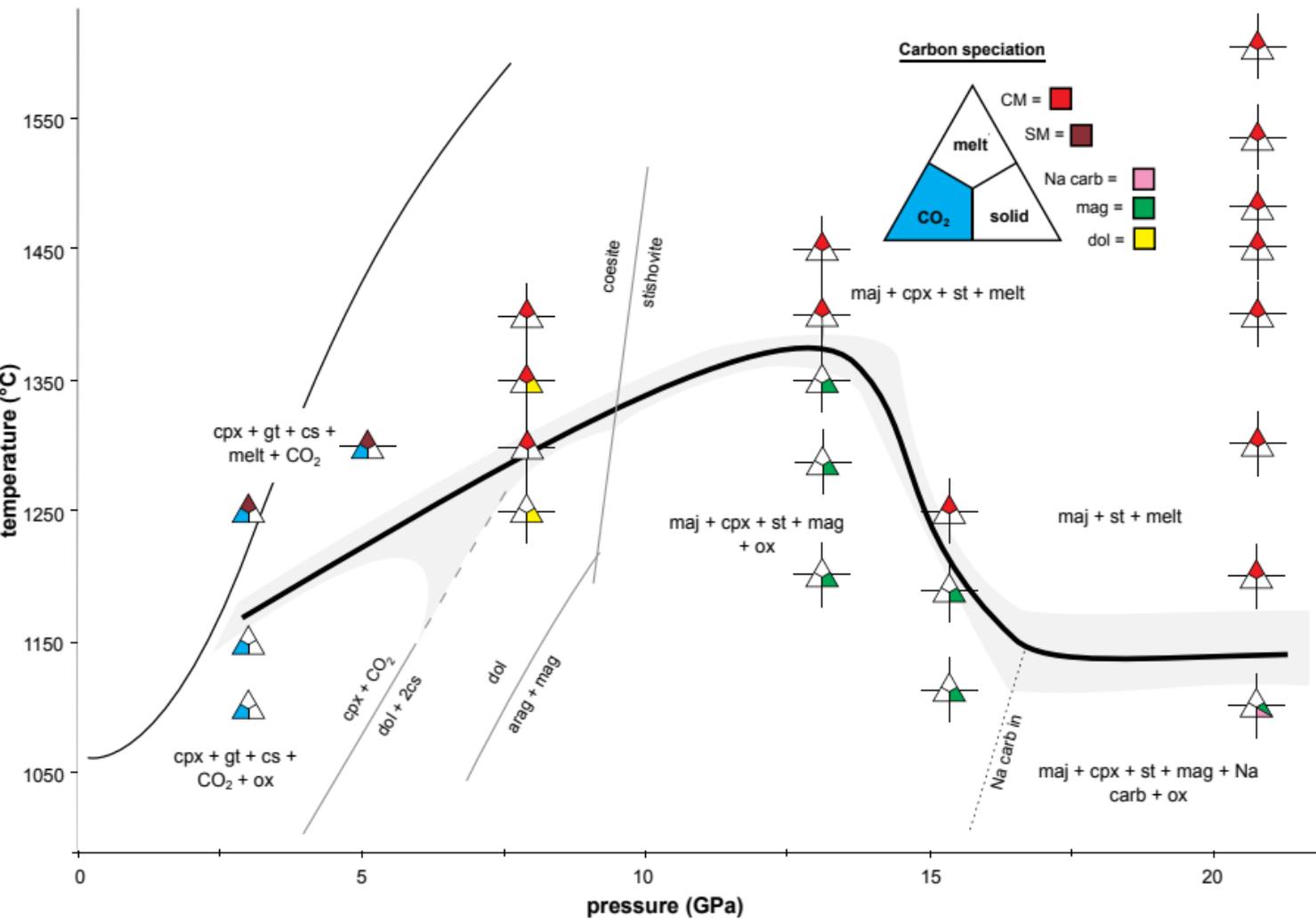
majorite from experiments on:

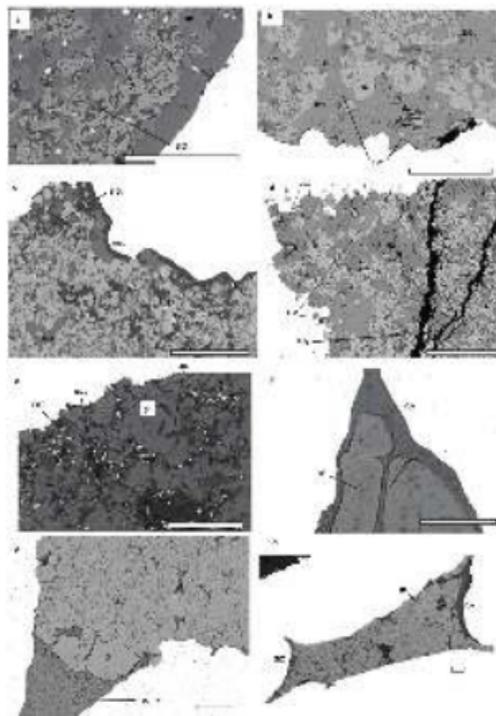
- | | |
|--|---|
| ● peridotite | inclusions in diamonds |
| ● peridotite + CO ₂ | reaction experiments |
| ● MORB | MORB melt 15.3 GPa |
| ● MORB + CO ₂ | MORB melt 20.7 GPa |
| ● MORB + H ₂ O | |

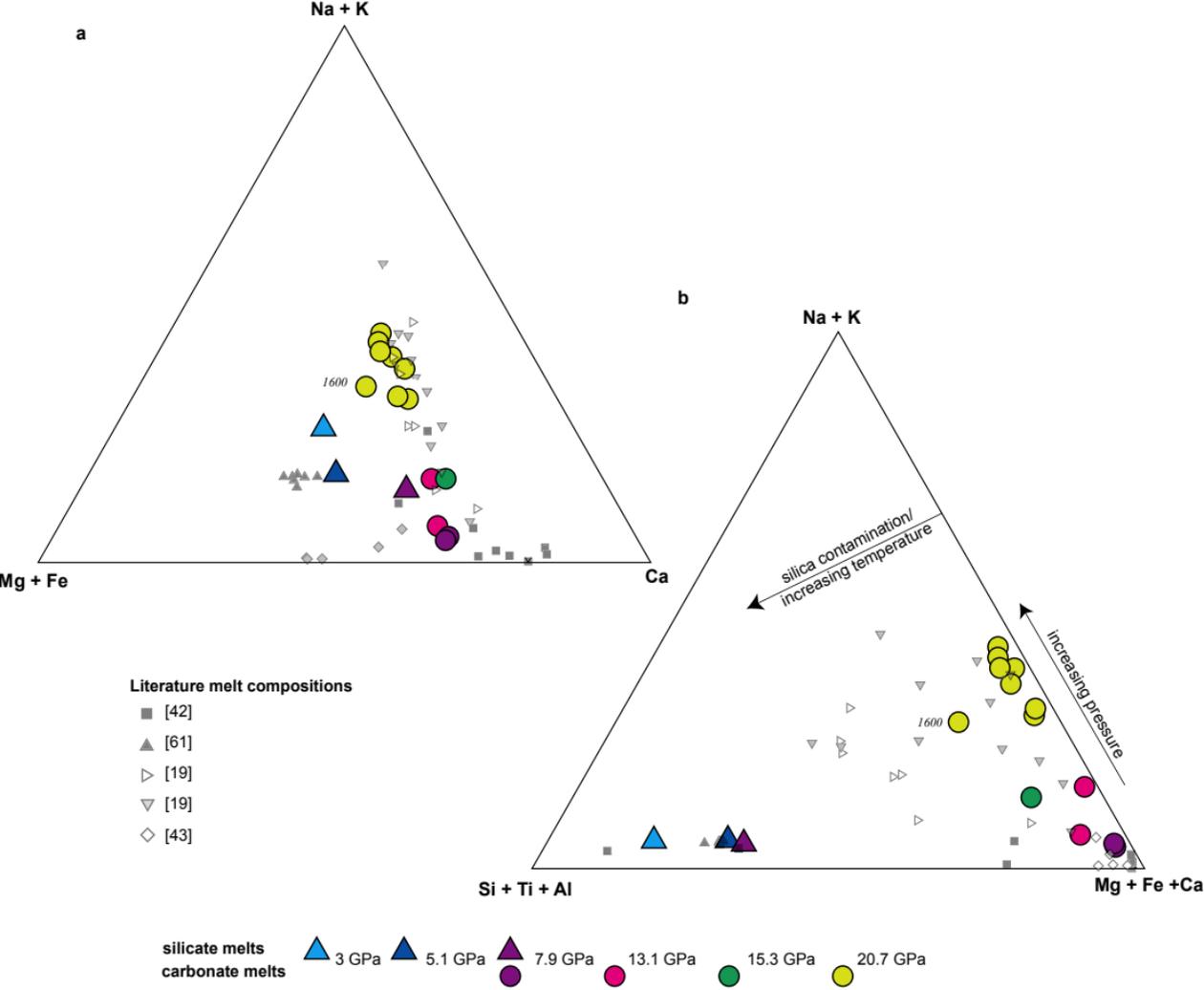


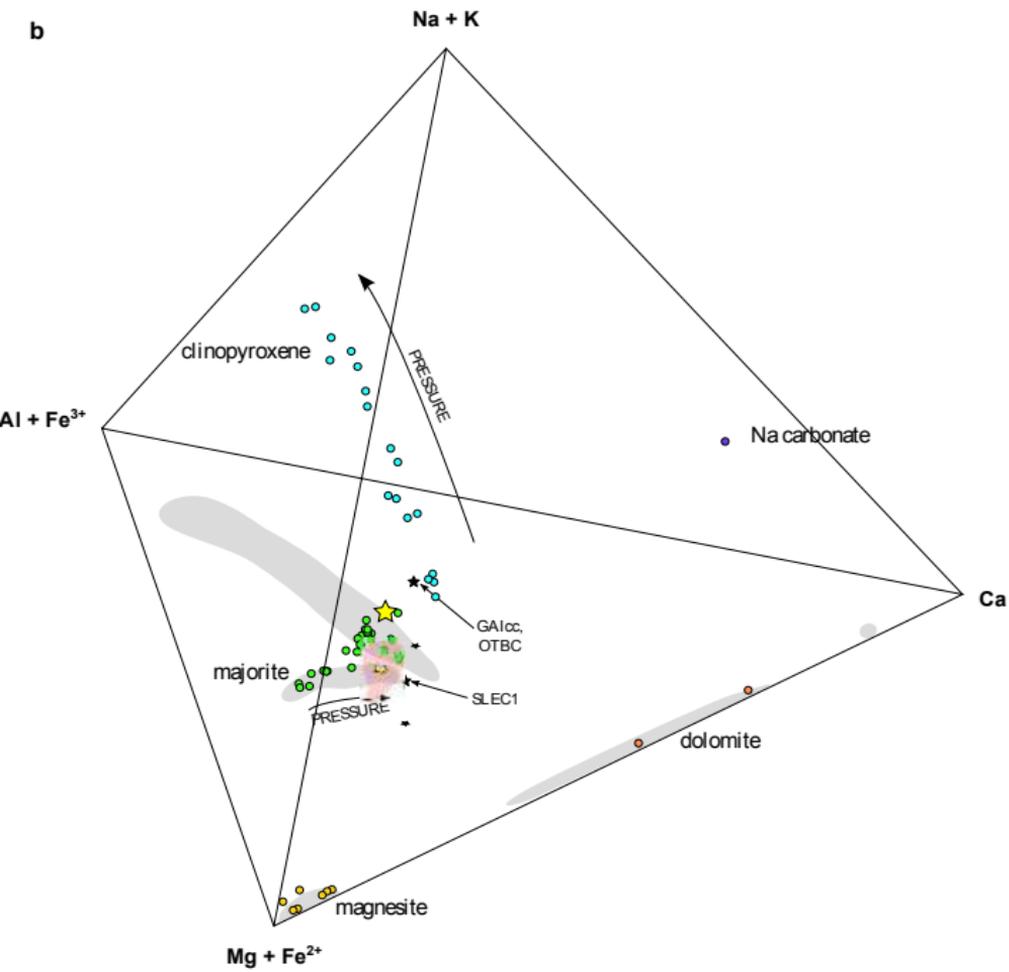
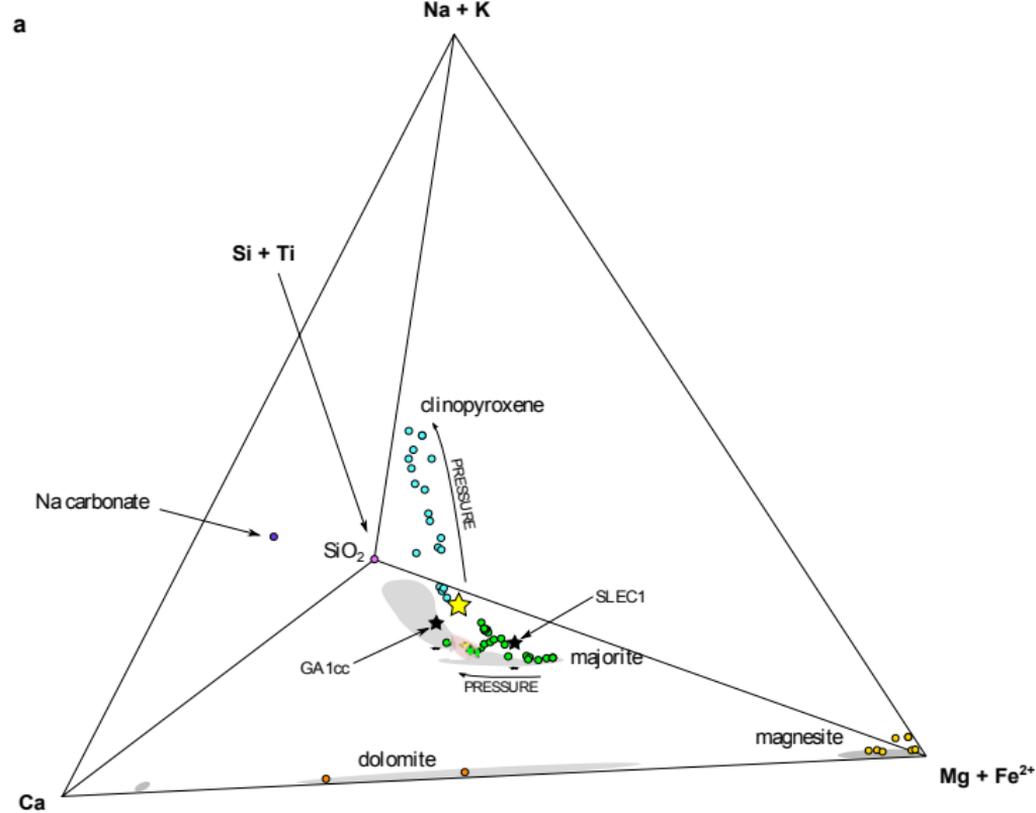


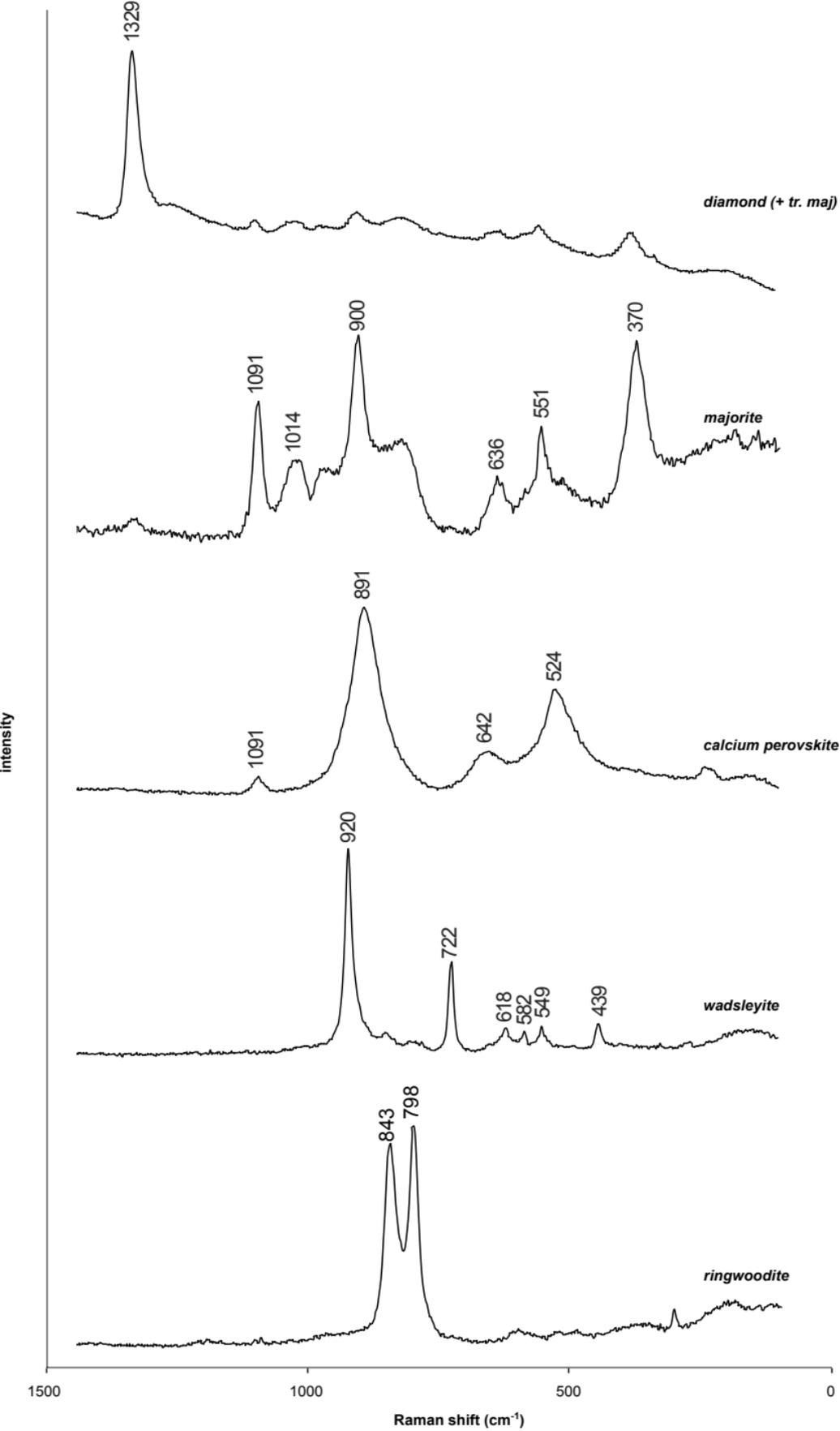












Year	Production	Value	Index	Production
1954	1,000,000	100	100	1,000,000
1955	1,100,000	110	110	1,100,000
1956	1,200,000	120	120	1,200,000
1957	1,300,000	130	130	1,300,000
1958	1,400,000	140	140	1,400,000
1959	1,500,000	150	150	1,500,000
1960	1,600,000	160	160	1,600,000
1961	1,700,000	170	170	1,700,000
1962	1,800,000	180	180	1,800,000
1963	1,900,000	190	190	1,900,000
1964	2,000,000	200	200	2,000,000
1965	2,100,000	210	210	2,100,000
1966	2,200,000	220	220	2,200,000
1967	2,300,000	230	230	2,300,000
1968	2,400,000	240	240	2,400,000
1969	2,500,000	250	250	2,500,000
1970	2,600,000	260	260	2,600,000
1971	2,700,000	270	270	2,700,000
1972	2,800,000	280	280	2,800,000
1973	2,900,000	290	290	2,900,000
1974	3,000,000	300	300	3,000,000
1975	3,100,000	310	310	3,100,000
1976	3,200,000	320	320	3,200,000
1977	3,300,000	330	330	3,300,000
1978	3,400,000	340	340	3,400,000
1979	3,500,000	350	350	3,500,000
1980	3,600,000	360	360	3,600,000
1981	3,700,000	370	370	3,700,000
1982	3,800,000	380	380	3,800,000
1983	3,900,000	390	390	3,900,000
1984	4,000,000	400	400	4,000,000
1985	4,100,000	410	410	4,100,000
1986	4,200,000	420	420	4,200,000
1987	4,300,000	430	430	4,300,000
1988	4,400,000	440	440	4,400,000
1989	4,500,000	450	450	4,500,000
1990	4,600,000	460	460	4,600,000
1991	4,700,000	470	470	4,700,000
1992	4,800,000	480	480	4,800,000
1993	4,900,000	490	490	4,900,000
1994	5,000,000	500	500	5,000,000
1995	5,100,000	510	510	5,100,000
1996	5,200,000	520	520	5,200,000
1997	5,300,000	530	530	5,300,000
1998	5,400,000	540	540	5,400,000
1999	5,500,000	550	550	5,500,000
2000	5,600,000	560	560	5,600,000
2001	5,700,000	570	570	5,700,000
2002	5,800,000	580	580	5,800,000
2003	5,900,000	590	590	5,900,000
2004	6,000,000	600	600	6,000,000
2005	6,100,000	610	610	6,100,000
2006	6,200,000	620	620	6,200,000
2007	6,300,000	630	630	6,300,000
2008	6,400,000	640	640	6,400,000
2009	6,500,000	650	650	6,500,000
2010	6,600,000	660	660	6,600,000
2011	6,700,000	670	670	6,700,000
2012	6,800,000	680	680	6,800,000
2013	6,900,000	690	690	6,900,000
2014	7,000,000	700	700	7,000,000
2015	7,100,000	710	710	7,100,000
2016	7,200,000	720	720	7,200,000
2017	7,300,000	730	730	7,300,000
2018	7,400,000	740	740	7,400,000
2019	7,500,000	750	750	7,500,000
2020	7,600,000	760	760	7,600,000
2021	7,700,000	770	770	7,700,000
2022	7,800,000	780	780	7,800,000
2023	7,900,000	790	790	7,900,000
2024	8,000,000	800	800	8,000,000

Source: U.S. Department of Commerce, Bureau of Economic Analysis, *Survey of Current Business*, Table 1.1.1, <https://www.bea.gov/data/gdp>.
 Note: All values are in constant 2017 dollars. The index is set to 100 in 1954.