

1 **A Mineral Scale Geochemical Investigation of Ultramafic Rocks**
2 **From the San Carlos and Kilbourne Hole Xenolith Localities,**
3 **Southwestern U.S.A.; Insights Into the Origin of the Regional**
4 **Shallow Mantle**

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11 **Abstract**

12 Twenty-one ultramafic xenoliths from the Quaternary San Carlos, Arizona and Kilbourne Hole,
13 New Mexico localities were investigated petrographically. We conducted a series of mineral scale
14 analyses of major and trace elemental chemistry via electron microprobe and laser ablation ICP-
15 MS techniques. Samples comprise various peridotites and pyroxenites derived from the shallowest
16 mantle beneath the modern western U.S.A. region. Peridotites are interpreted to represent the
17 primary depleted shallowest mantle lithospheric rocks of the region, whereas clinopyroxenites,
18 websterites and wehrlites are former melts or cumulate of mafic melts that enriched and/or
19 metasomatized the mantle at some unknown time. Both suites equilibrated at around 1000-1110⁰
20 C and 0.9-1.5 GPa, corresponding to depths of 30-50 km beneath the surface, immediately beneath
21 the modern Moho. Trace elemental concentrations, Mg# and other depletion indicators in
22 peridotites are inconsistent with an (accreted) oceanic plate (Farallon) origin for these rock suites.
23 Instead, they represent a melt-modified native North American depleted mantle lithosphere. We
24 show that to a first order, differences in chemistry between cores and rims of various lithologies
25 are minuscule, lacking diffusional profiles. The implication is that most metasomatic processes
26 that enriched the mantle there are recent, Mesozoic or younger.

27

28 **Keywords:** ultramafic, western U.S.A., mantle lithosphere, thermobarometry

29 **Resume**

30 Nous avons étudié pétrographiquement vingt et un xénolites ultramafiques issus des gisements
31 quaternaires de San Carlos (en Arizona) et de Kilbourne Hole (au Nouveau Mexique). Des
32 analyses minérales in situ en éléments majeurs et traces ont été réalisées par microsonde
33 électronique et LA-ICP-MS, respectivement. Ces échantillons consistent en une large variété de
34 péridotites et de pyroxénites issues de la partie supérieure du manteau lithosphérique sous l'actuel
35 région Ouest des États-Unis. Les péridotites sont interprétées comme des reliques originellement
36 appauvries issues de la partie supérieure de la lithosphère, alors que les clinopyroxénites,
37 webstérites et les wherlites sont interprétées comme d'anciens liquides magmatiques ou des unités
38 cumulatives formées à partir de ces mêmes liquides mafiques, ayant enrichi et/ou métasomatisé
39 cette portion du manteau. Les deux assemblages lithologiques se sont équilibrés à des conditions
40 de température autour de 1000-1110°C et de pression avoisinant les 0.9 à 1.5 GPa. Ces conditions
41 correspondent à une profondeur de 30 à 50 km, juste en dessous du Moho moderne. La signature
42 géochimique des péridotites en éléments en trace, en Mg# et d'autres indicateurs
43 d'appauvrissement suggèrent que ces unités ne sont pas liées génétiquement à l'accrétion d'une
44 plaque océanique (Farallon). A contrario, ces roches représentent des vestiges appauvris du
45 manteau lithosphérique nord-américain ayant partiellement été modifiés par la présence de liquides
46 magmatiques. Notre étude démontre que les différences géochimiques entre les cœurs et les
47 bordures des différents échantillons lithologiques sont minimales et exemptes de profils de diffusion.
48 Cela implique que la plupart des processus métasomatiques ayant enrichi le manteau sont
49 relativement récents, probablement mésozoïques.

50 **Mots-clés :** ultramafique, ouest-américain, manteau lithosphérique, thermobarométrie

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54 1. Introduction

55 Mineral scale geochemical and isotopic investigations represent the gold standard of unraveling
56 continental upper mantle origin and evolution, as revealed through xenoliths found in basaltic
57 rocks. The great majority of such xenoliths are fragments of the mantle lithosphere. Tools based
58 on major and trace elemental chemistry include thermobarometry, the ability to estimate how
59 depleted a mantle fragment is and therefore how likely it is to belong to a former oceanic domain
60 (Luffi et al., 2009, Quinn et al., 2018) or the continental mantle. The ability to actually date (i.e.,
61 obtain a radiometric age of) the extraction of a mantle fragment from the convective mantle is
62 more restricted and is based on the opportunistic application of certain isotopic systems, such as
63 Re-Os depletion ages. Still, a tremendous amount of information can be obtained based on
64 chemical information alone, especially if performed *in-situ*, within grain. Electron microprobe
65 techniques have allowed the determination of major elemental concentrations on ultramafic
66 minerals for decades. The advent of laser ablation single collector ICP-MS techniques in recent
67 years allowed for complementary studies of trace (as well as major) elemental concentrations to
68 be carried out in such materials.

69 Here we present new mineral chemistry data on various ultramafic xenoliths from Quaternary
70 basaltic rocks from San Carlos (Arizona) and Kilbourne Hole (New Mexico). Both are classic
71 locations (Frey and Prinz, 1978; Wilshire et al., 1988) in the geologic literature. Materials from
72 these locations have been used as standards for various geochemical materials over the past five
73 decades and the xenolith occurrences are famous for richness in material and size of xenoliths.
74 Despite these features, petrologic studies of both localities are somewhat antiquated and numerous
75 petrologic and tectonic questions still linger about the origin of these rocks. For example, it is still
76 a matter of debate as to whether they represent native North American lithospheric fragments or
77 subducted-accreted Farallon materials. Another issue of debate is whether the abundant signs of
78 metasomatic re-fertilization (via passage of melts) seen in these rocks are young extensional
79 features (syn- basaltic magmatism), relatively young subduction-related Cordilleran features (35-
80 200 Ma) or inherited from a previous Wilson cycle.

81 We address some of these questions in this preliminary petrologic study using major and trace
82 elemental chemistry on all principal ultramafic lithologies from these iconic locations. A

83 subsequent study focusing on radiogenic and stable isotopic systems will complement the findings
84 presented here.

85 **2. Petrography**

86 We studied 21 peridotite and pyroxenite xenoliths from the San Carlos and Kilbourne Hole
87 localities (Fig. 1). The peridotite suite of samples investigated includes 14 spinel lherzolites and 2
88 spinel wehrlites, while the pyroxenite suite is represented by 5 olivine websterites. The sampling
89 at both locations was carried out so that the petrographic and textural variety of the samples
90 recovered (estimated visually in the field) is the most representative for each location, according
91 to previous investigations (Frey and Prinz, 1978; Dromgoole and Pasteris, 1987; Wilshire et al.,
92 1988).

93 San Carlos samples used for this study (n=10) are olivine-rich spinel lherzolites (n=6), spinel-
94 bearing olivine websterites (n=3), and spinel wehrlite (n=1) (Fig. 2). The majority of samples
95 collected from this location can be classified as Group I, according to the terminology of Frey and
96 Prinz (1978). They have Cr-rich clinopyroxenes and spinels and high forsterite content in olivines
97 ($Fo > 86$). The olivine wehrlite sample (SC07) is the only Group II xenolith used in this study. It
98 has Fe-rich olivines (Fo_{71}), Ti-rich clinopyroxenes, and Al- and Fe-rich spinels.

99 San Carlos xenoliths show a wide variation of textures, which we classified using the systematics
100 of Mercier and Nicolas (1975). The predominant textural type is protogranular (SC02, SC03,
101 SC04, SC05, SC06, SC13), with medium to coarse grained, dominantly equant crystals of olivine
102 and orthopyroxene (1-5 mm) and much smaller clinopyroxenes and spinels. The grain boundaries
103 tend to be curvilinear although, locally, polygonal aggregates of same mineral forming triple
104 junctions were observed. Spinels occur mainly as interstitial, vermicular shaped grains, or rounded
105 blebs enclosed in peripheral parts of orthopyroxenes. Rare polygonal spinel crystals have been
106 detected. The samples display no evident mineral elongation, which could define any lineation or
107 foliation to the rocks.

108 Porphyroclastic texture is present in one sample (SC10). The sample hosts large, elongated
109 porphyroclasts of orthopyroxene (up to 8 mm) within a matrix of fine to medium grained,
110 polygonal neoblasts of olivine and clinopyroxene (representing approximately 50% of the sample).

111 The orthopyroxene porphyroclasts have irregular shapes with curved boundaries, and are slightly
112 aligned, imprinting a weak foliation to the sample. Spinel occurs as a subordinate phase, with
113 irregular grain shape, either fully enclosed or peripheral to olivine crystals. Several larger
114 clinopyroxene crystals contain annealed cracks, forming trails of minute sulfide and fluid
115 inclusions without any preferred orientation.

116 Two samples have mosaic equigranular textures (SC09 and SC11). They are characterized by fine-
117 grained, equant crystals, with straight-lined grain boundaries, often converging at triple point
118 junctions. Spinels mainly occur as small, disseminated grains, with curvilinear boundaries. Some
119 spherical spinel inclusions were observed within olivine crystals. The equigranular samples show
120 no textural anisotropy.

121 Poikilitic texture has been observed in the wehrlite sample (SC07). The sample is coarse grained
122 (2-3 mm average), and it contains large poikilitic clinopyroxene crystals that enclose rounded
123 grains of olivine. Spinels are scattered throughout the sample as small blebs fully enclosed in
124 clinopyroxenes. Orthopyroxenes were absent in this sample. Interstitial glass is present, and it
125 occurs as a discontinuous network of melt pockets surrounding the silicate minerals. Locally, tiny
126 quench crystals of apatite and plagioclase were found in association with melt pockets. No other
127 discrete metasomatic phases were identified, even though amphiboles (pargasite, kaersutite) and
128 phlogopite have been reported previously at this location, especially in Group II xenoliths (Frey &
129 Prinz, 1978).

130 Most samples lack any visible reaction products (pyrometamorphic textures) commonly developed
131 through decompression-induced partial melting and melt infiltration during xenolith transport.
132 Only three samples (SC04, SC07 and SC09) have clinopyroxenes with porous (spongy-textured)
133 outer margin, indicative of an incipient partial melting reaction. The spongy border zones are thin
134 (20-30 μm) and well delimited from the inner, homogeneous cores.

135 Sulfides are scarce (average 3 grains per polished section) or absent in San Carlos samples. They
136 are small (20-40 μm) and occur mostly as isolated spheroidal droplets enclosed in clinopyroxenes.
137 Some sulfide inclusions display irregular margins, with apophyses radiating from the rims,
138 considered to be decrepitation features resulted during the entrainment of the xenoliths in the
139 basaltic magma. Similar features in San Carlos xenoliths have been first described by Andersen et

140 al., (1987). Linear or curved healed fractures in clinopyroxenes may have associated trails of
141 minute sulfide and fluid inclusions. Interstitial sulfides have been also observed; they have either
142 irregular, elongated shapes, and occur along curvilinear silicate grain boundaries or rounded blebs
143 located at triple junctions, commonly in close association with glass pockets.

144 Kilbourne Hole samples examined for this study (n=11) are Cr-diopside bearing spinel lherzolites
145 (n=8), spinel-bearing olivine websterites (n=2) and spinel wehrlite (n=1) (Fig. 2). All samples
146 belong to Group I of Frey and Prinz (1978) and display predominantly protogranular and
147 subordinately porphyroclastic – tabular equigranular textures, according to the texture
148 classification of Mercier and Nicolas (1975).

149 Samples with protogranular texture (KH21, KH25, KH26, KH27, KH28, KH30, and KH31) have
150 coarse olivine and orthopyroxene grain size (2-4 mm), with mostly curvilinear grain boundaries.
151 Locally, olivine crystals form aggregates with polygonal boundaries. Clinopyroxenes are
152 significantly smaller and occur in direct contact with larger orthopyroxene crystals. Spinels have
153 amoeboid shapes, occurring commonly as interstitial blebs around orthopyroxene grains. Textural
154 anisotropy or deformation features are absent in this sample group.

155 Samples with porphyroclastic transitioning into tabular equigranular texture (KH22, KH32, KH33,
156 and KH35) have finer grain size (< 1mm) with few larger, elongated, porphyroclastic olivine and
157 orthopyroxene crystals. Grain boundaries are straight-lined, commonly forming triple point
158 junctions. Clinopyroxenes form small, polygonal grains scattered in the samples. Spinels occur
159 both as rounded to sub-rounded inclusions in olivine and as interstitial, xenomorphic shaped
160 grains. Abundant intergranular spinels display the characteristic holly-leaf shape, but some well-
161 developed, subhedral crystals are also present. The samples show weak lineation, characterized
162 mainly by the alignment of elongated spinel grains.

163 With the exception of sample KH22, Kilbourne Hole samples studied here show no evidence of
164 reaction or alteration of their primary mineralogy during transport. Interstitial spinel grains in
165 sample KH22 show thin (approximately 20 μm) reaction coronas comprising small grains of
166 partially interconnected secondary, high-Cr spinels surrounded by glass. Spinels display sharp
167 boundaries between the core and the rim, and their cores show no evident zoning pattern.

168 No hydrous mineral phases were identified in the Kilbourne Hole samples used in this study.
169 Sulfides are a very minor phase, with maximum five grains per polished section. Some samples
170 are completely devoid of sulfides. Where present, sulfides are small (< 50 μm across), round blebs
171 fully enclosed within silicate phases (mainly clinopyroxenes) or discrete, interstitial grains located
172 at grain boundaries, commonly in association with melt pockets.

173 3. Analytical methods

174 3.1. Electron-probe micro analysis

175 Major element, along with selected trace element compositions for the major, rock-forming
176 silicates (olivines, clinopyroxenes, and orthopyroxenes) and spinels were analyzed in polished
177 thick sections of approximately 1 mm thickness. We used a Cameca SX-100 electron-probe
178 microanalyzer at the Lunar and Planetary Laboratory, University of Arizona. Five spectrometers
179 were employed using the crystal arrangements TAP, LPET, LPET, TAP, LLIF. The instrument
180 was calibrated using well-characterized natural and synthetic standards. During each analysis,
181 major elements (Na $K\alpha$, Mg $K\alpha$, Al $K\alpha$, Si $K\alpha$, K $K\alpha$, Ca $K\alpha$, Mn $K\alpha$, Fe $K\alpha$, Ti $K\alpha$, and Ba $L\alpha$
182 for silicate minerals; plus Cr $K\alpha$ for oxides) were measured at 20 kV accelerating voltage and 20
183 nA beam current, counting for 10 seconds on peak. Beam conditions were then changed to high-
184 current (299nA, at 20 kV accelerating voltage) and longer counting time for the measurement of
185 selected trace elements (F $K\alpha$, P $K\alpha$, S $K\alpha$, Cl $K\alpha$, Cr $K\alpha$, Sc $K\alpha$, V $K\alpha$, Ni $K\alpha$, Co $K\alpha$, Zn $L\alpha$, Y
186 $L\alpha$, Ga $L\alpha$, Zr $L\alpha$, Sr $L\alpha$, Sn $L\alpha$, As $L\alpha$, Pb $M\alpha$). Representative mineral compositions are reported
187 in Appendix 1. Further details regarding the analytical routines used in this study are available at
188 <https://www.rockptx.com/wp-content/uploads/2017/02/microprobe-analytical-routines-2016.pdf>.

189 3.2. Laser ablation inductively coupled plasma mass spectrometry

190 Major and trace element concentrations in olivines, clinopyroxenes, orthopyroxenes, and spinels
191 were obtained *in situ* by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-
192 MS) on the same thick sections used for electron probing. The analyses were conducted on an
193 Agilent 7500a ICP-MS instrument, coupled to a GeoLas 2005, 193 nm wavelength laser ablation
194 system, at the State Key Laboratory of Geological Processes and Mineral Resources, China
195 University of Geosciences. Laser sampling was performed in He carrier gas, single spot ablation

196 style, with a laser beam diameter of 40 μm . Data acquisition time for each analysis was set to 50
197 seconds on sample, followed by approximately 20 seconds of background acquisition. Well-
198 characterized reference glasses of natural and synthetic compositions such as NIST: SRM 610
199 (Pearce et al., 1997), USGS: BCR-2G, BHVO-2G, BIR-1G, GSD-1G, GSE-1G (Jochum et al.,
200 2005) and MPI-DING: ATHO-G, ML3B-G, T1-G (Jochum et al., 2006) were used after every 8
201 sample analyses. An in-house software (ICPMSDataCal) was used to perform time-dependent drift
202 of sensitivity and mass discrimination correction and quantitative data calibration. More detailed
203 operating conditions, data calibration strategy and software, as well as method uncertainties are
204 found in Liu et al. (2008). Analyses are reported in Appendix 2.

205 **4. Results**

206 4.1. Mantle mineral chemistry: major and trace elements

207 Olivine, clinopyroxene, orthopyroxene and spinel grains were analyzed in detail, with 1 – 4 grains
208 per phase, cores and rims, for each sample (EMPA and LA-ICP-MS combined). Full mineral
209 datasets are available in Appendix 1 and 2. Minerals in the studied mantle xenoliths are generally
210 chemically homogeneous, showing no considerable inter- and intra-grain compositional variation.

211 4.1.1. Olivine

212 In San Carlos (SC) samples, olivine crystals have forsterite content ($\text{Fo}_i = 100 \times \text{Mg} / (\text{Mg} + \text{Fe})$,
213 normalized cation ratios) ranging between 87.8 and 89.7 for lherzolites and slightly more elevated
214 Fo contents for the olivine websterite samples (88.5 - 90.9) (Fig. 3), with comparable and limited
215 ranging NiO (0.35 – 0.42 wt.%) and MnO contents (0.10 – 0.18 wt.%). Calcium content shows
216 more variability (0.034 – 0.127 wt.% CaO), from sample to sample, but also from core to rim.
217 Many olivines, especially in the websterite sample group, have more elevated Ca contents towards
218 the rims (up to 60% increase of Ca at the rims of the grains). The olivines in the wehrlite sample
219 have the lowest Fo content observed (70.7), lower NiO (0.21 wt.%), and higher MnO (0.32 – 0.37
220 wt.%). Its calcium content from core to rim is 0.027 – 0.130 wt.% oxide, which is the most
221 significant rim enrichment for Ca in the San Carlos samples.

222 Olivines belonging to Kilbourne Hole (KH) lherzolite and websterite samples are largely similar
223 in their major element content. The lherzolite olivine Fo content ranges between 87.6 and 89.6,

224 and NiO between 0.37 and 0.42 wt.%, while the websterite olivines have 87.7 – 89.4% Fo and
225 0.36 – 0.40 wt.% NiO compositions (Fig. 3). Their Mn and Ca content overlaps as well, although
226 the lower and the higher ends of the range are characteristic to lherzolite (0.09 – 0.16 wt.% MnO;
227 0.025 – 0.084 wt.% CaO) and websterite samples (0.12 – 0.18 wt.% MnO; 0.035 – 0.093 wt.%
228 CaO), respectively. Few olivine crystals in the KH sample set show Ca compositional variation,
229 including normal (where Ca content increases at rims) and reverse (where Ca contents decreases
230 at rims) core-rim zoning patterns. However, Ca variation in KH samples is less pronounced than
231 in the SC samples investigated here (up to 25% Ca increase in KH26-lherzolite, KH31-websterite,
232 KH28-wehrlite). The KH wehrlite sample has olivines with low Fo (81.5), and NiO (0.26 wt.%),
233 and high MnO (0.21 wt.%). Calcium concentration is also higher than other KH samples (0.10 –
234 0.14 wt.%), but with much attenuated core-to-rim variation compared the SC wehrlite.

235 4.1.2. Clinopyroxene

236 San Carlos lherzolite samples have clinopyroxenes with an end-member composition of W_{042-}
237 $_{45}En_{49-52}Fs_{4-6}$ and Mg# ranging between 89.34 and 91.98. They are mostly homogeneous with
238 respect to Ti, Al and Na concentrations. Their TiO₂ content ranges between 0.11 and 0.54 wt.%,
239 Al₂O₃ between 3.27 and 7.35 wt.%, and Na₂O between 1.06 and 2.24 wt.%. Chromium displays
240 the largest amplitude of variation between SC lherzolites, with Cr# falling mainly into two distinct
241 compositional groups: a more fertile group with Cr# values of 5.74 – 7.58 (SC04 and SC05), and
242 a more refractory group with Cr# of 12.97 – 22.63 (SC02, SC03, SC06, SC13). The SC pyroxenites
243 have a clinopyroxene composition of $W_{021-46}En_{49-70}Fs_{4-8}$, with Mg# between 88.79 and 91.86,
244 more elevated Cr, Al and Ti contents (Cr#: 8.70 – 16.97; 4.38 – 13.55 wt.% Al₂O₃; and 0.23 – 2.40
245 wt.% TiO₂). Sodium is low (0.60 – 1.32 wt.% Na₂O) with the exception of sample SC09 where
246 Na₂O is slightly higher (2.04 – 3.51 wt.%) and variable from grain to grain, without any core-to-
247 rim heterogeneity. The wehrlite clinopyroxenes are Ti-augites characterized by an end-member
248 composition of $W_{037-47}En_{42-48}Fs_{11-14}$, and by Mg# and Cr# as low as 77.5 and 1.12, respectively.
249 Weak inter-grain compositional variation is observed, with TiO₂ content of 0.71 – 1.21 wt.%; 5.80
250 – 8.20 wt.% for Al₂O₃; and 1.05 – 1.19 wt.% for Na₂O.

251 All clinopyroxenes in Kilbourne Hole mantle xenoliths are Cr-diopsides having relatively similar
252 end-member compositions of $W_{043-46}En_{48-50}Fs_{5-6}$ for lherzolites, $W_{044-47}En_{48-50}Fs_{5-6}$ for olivine

253 websterites, and $W_{0.42-4.4}En_{44-48}Fs_{9-12}$ for the wehrlite sample. KH lherzolites tend to be less
254 refractory than the SC equivalents, with slightly lower Mg# (88.7 - 90.57), Na_2O (1.56 – 1.98
255 wt.%), and Cr# (4.02 - 8.33), but richer in TiO_2 content (0.23 – 0.63 wt.%). The compositions of
256 clinopyroxenes in the websterite samples lie within the range of those analyzed in lherzolite with
257 respect to their end-member configuration and Mg# (89.28 - 90.52). They are characterized also
258 by lower Na_2O (0.41 – 1.93 wt.%), and moderate enrichment in Al and Ti contents (5.79 – 8.66
259 wt.% Al_2O_3 , 0.44 – 0.80 wt.% TiO_2). Chromium shows wider variation from sample to sample,
260 with Cr# as low as 1.10 – 3.24 for sample KH31, and 8.73 – 8.94 for KH35. In contrast, the wehrlite
261 sample has clinopyroxenes with remarkably low Mg#, between 78.35 and 83.59, and Cr#, between
262 0.16 and 6.21. On the other hand, clinopyroxenes are richer in Ti and Al (0.76 – 1.60 wt.% TiO_2 ,
263 and 7.50 – 9.25 wt.% Al_2O_3). In this sample, sodium has a limited range of variation (1.32 – 1.42
264 wt.% oxide).

265 4.1.3. Orthopyroxene

266 Orthopyroxenes in San Carlos samples are enstatite with similar end-member compositions ($W_{0.1-}$
267 $2En_{87-89}Fs_{9-10}$ in lherzolites, and $W_{0.1-2}En_{88-90}Fs_{9-11}$ in olivine websterites). No orthopyroxenes were
268 identified in the SC wehrlite sample. Lherzolite orthopyroxenes have Mg# ranging between 88.9
269 and 90.5, Cr# between 4 and 15.5, and are relatively Al-rich (2.3 – 5.4 wt.% Al_2O_3), and Na and
270 Ti poor (0.07 – 0.18 wt.% Na_2O , and 0.04 – 0.16 wt.% TiO_2). In comparison, orthopyroxenes
271 belonging to olivine websterite samples are slightly more magnesian (Mg# 89 – 91.2), and have a
272 more limited Cr# range (6.19 – 12.8) than the lherzolite-hosted orthopyroxenes, but otherwise,
273 their major element compositions overlap entirely. Most grains show moderate Al, Cr, Ca and Ti
274 zonation, with higher contents in cores than in rims.

275 Orthopyroxenes in Kilbourne Hole xenoliths are enstatite with end-member compositions of $W_{0.1-}$
276 $2En_{88-89}Fs_{10-11}$ for lherzolites, and $W_{0.1-2}En_{87-88}Fs_{10-11}$ in olivine websterites, and with slightly
277 higher ferrosilite component for the wehrlite sample ($W_{0.2}En_{84}Fs_{14}$). Lherzolite- and websterite-
278 hosted orthopyroxenes are characterized by a narrow range of Mg# (from 88.5 to 90.2) and Cr#
279 (between 2.8 and 6.4), and by moderate contents of Na and Ti (0.08 – 0.14 wt.% Na_2O ; 0.05 – 0.16
280 wt.% TiO_2). Aluminum content ranges between 3.7 and 5.2 wt.% Al_2O_3 and shows no significant
281 zonation. Orthopyroxenes in the wehrlite sample have Mg# of 85.5, Cr# of 5.3, are, in average

282 more aluminous (5.3 – 5.4 wt.% Al_2O_3), Ti and Na rich (0.17 wt.% TiO_2 , 0.14 wt.% Na_2O). The
283 analyzed orthopyroxene grains in the wehrlite sample preserve a weak variation in major elements,
284 expressed by core-to-rim enrichment in Ca, Ti and Al, compensated by depletion in Cr and Mg.

285 4.1.4. Spinel

286 Spinel from San Carlos lherzolites and websterites show a similar and broad compositional range,
287 characterized mainly by extensive substitution along the spinel (MgAl_2O_4) – magnesiochromite
288 (MgCr_2O_4) solid solution ($\text{Cr}_{0.16-0.78}\text{Al}_{1.15-1.83}$; atoms per formula unit – a.p.f.u.), and rather limited
289 substitution along the spinel – hercynite (FeAl_2O_4) end-member composition series ($\text{Mg}_{0.70-}$
290 $0.81\text{Fe}^{2+}_{0.19-0.30}$). As a result, lherzolite- and websterite-hosted spinels display a wide range of Cr#
291 (8.2 – 40.3) and consistently low values for Mg# (70 - 81). They also contain trace amounts of
292 Fe^{3+} (up to 0.10 a.p.f.u., calculated), and minor amounts of Ti and Mn (0.11 – 0.34 wt.% TiO_2 , and
293 0.08 – 0.19 wt.% MnO). In contrast, spinels belonging to the wehrlite sample are rich in the
294 hercynite component ($\text{Mg}_{0.50-0.52}\text{Fe}^{2+}_{0.48-0.50}$), and also contain higher amounts of Fe^{3+} (0.11
295 a.p.f.u.), TiO_2 (0.42 wt.%), and MnO (0.19 – 0.22 wt.%). The wehrlite-hosted spinels are
296 significantly less chromian (Cr# < 14) and less magnesian (Mg# < 52) than those in other SC
297 samples studied here.

298 Kilbourne Hole lherzolites and websterites contain spinels that are highly magnesian (Mg# ranging
299 between 78.6 and 92.1, where the lower values represent mostly the lherzolites, while the higher
300 values, the websterites), and low chromium contents (Cr# ranging between 0.76 and 11.5). Their
301 end-member composition is dominated by spinel and show limited substitution along the spinel –
302 magnesiochromite and spinel – hercynite solid solution series ($\text{Cr}_{0.02-0.21}\text{Al}_{1.64-1.94}$; $\text{Mg}_{0.79-}$
303 $0.92\text{Fe}^{2+}_{0.08-0.21}$). In addition to these major elements, KH lherzolite- and websterite-hosted spinels
304 also contain small amounts of Fe^{3+} (up to 0.16 a.p.f.u.), Ti (0.02 – 0.13 wt.% oxide), and Mn (0.05
305 – 0.13 wt.% oxide). The KH wehrlite contains spinels that belong to the spinel – hercynite solid
306 solution series ($\text{Mg}_{0.67-0.76}\text{Fe}^{2+}_{0.24-0.34}$), with relatively high amounts of Ti (0.48 – 0.66 wt.% oxide),
307 and Fe^{3+} (0.08 – 0.2 a.p.f.u., calculated). The KH wehrlite spinels are also characterized by
308 extremely low Mg# (66.7 – 75.8), and Cr# (0.26 – 2.79).

309 4.2. Trace element compositions of clinopyroxenes

310 Trace element abundances in clinopyroxenes from San Carlos xenoliths show significant light rare
311 earth elements (LREEs) scattering, and nearly flat or smoothly descending slope for middle and
312 heavy rare earth elements (MREEs and HREEs, respectively) (Fig. 4.A). Most clinopyroxenes in
313 the San Carlos lherzolites display relatively flat MREE-HREE patterns, with wide range of CI
314 chondrite-normalized MREE and HREE concentrations, $(\text{MREE})_{\text{CI}} = 5.8 - 44.1$; $(\text{HREE})_{\text{CI}} = 1.74$
315 $- 24.2$; $(\text{Yb})_{\text{CI}} = 1.74 - 14.8$, and various degrees of LREE depletion, with $(\text{Ce}/\text{Yb})_{\text{CI}}$ between 0.12
316 and 0.98. As exceptions, lherzolites SC06 and SC02 contain clinopyroxenes with marked upward
317 convex REE patterns with apexes at cerium values, which show significant relative LREE
318 enrichments, $(\text{Ce}/\text{Yb})_{\text{CI}} = 4.56 - 3.05$. Websterite-hosted clinopyroxenes at San Carlos have LREE
319 abundances that vary from relatively depleted, $(\text{Ce}/\text{Yb})_{\text{CI}} = 0.64 - 0.99$ (SC10 and SC11), to highly
320 enriched $(\text{Ce}/\text{Yb})_{\text{CI}} = 5.6$ (SC09); while their MREE and HREE concentrations are confined to a
321 narrower range than in SC lherzolites, $(\text{MREE})_{\text{CI}} = 4.66 - 31.5$; $(\text{HREE})_{\text{CI}} = 2.45 - 12.8$; $(\text{Yb})_{\text{CI}} =$
322 $2.45 - 12.3$. The wehrlite sample contains clinopyroxenes with LREE-enriched profile $(\text{Ce}/\text{Yb})_{\text{CI}}$
323 $= 4.5 - 5.1$, and ample core to rim LREE and MREE enrichments ($\text{La}_{\text{core}} = 4.6$ ppm; $\text{La}_{\text{rim}} = 5.4$
324 ppm; $\text{Ce}_{\text{core}} = 12.1$ ppm; $\text{Ce}_{\text{rim}} = 18.3$ ppm; $\text{Nd}_{\text{core}} = 6.5$ ppm; $\text{Nd}_{\text{rim}} = 11.4$ ppm; $\text{Sm}_{\text{core}} = 1.6$ ppm;
325 $\text{Sm}_{\text{rim}} = 3.5$ ppm; $\text{Eu}_{\text{core}} = 0.8$ ppm; $\text{Eu}_{\text{rim}} = 1.3$ ppm;).

326 Primitive mantle-normalized extended trace element patterns of clinopyroxenes in the majority of
327 San Carlos samples (Fig. 4.B) are characterized by overall enrichments in the highly incompatible
328 elements (Th, U, LREE), with marked negative Nb anomalies relative to the LREE, and depletions
329 of Sr, Ti, Zr and Hf relative to elements with similar incompatibilities; $\text{Sr}/\text{Sr}^* = 0.42 - 0.88$; Ti/Ti^*
330 $= 0.16 - 0.78$; $\text{Zr}/\text{Zr}^* = 0.24 - 0.68$; and $\text{Hf}/\text{Hf}^* = 0.1 - 0.85$, (where: $\text{Sr}^* = (\text{Pr} + \text{Nd})/2$; $\text{Ti}^* = (\text{Eu}$
331 $+ \text{Gd})/2$; and Zr^* and $\text{Hf}^* = (\text{Nd} + \text{Sm})/2$). Few lherzolite-hosted clinopyroxenes show positive
332 deviations for $\text{Hf}/\text{Hf}^* = 1 - 1.3$, in the core of SC05 and in the rim of SC04, and $\text{Sr}/\text{Sr}^* = 1.03 -$
333 1.04 , in sample SC02, core and rim, and in the core of SC03.

334 Clinopyroxene REE patterns from Kilbourne Hole xenoliths (Fig. 4.C) are similar to our San
335 Carlos patterns, with respect to LREE scattering, but display a smoother slope and narrower range
336 of values for the MREEs and HREEs. All analyzed Kilbourne Hole lherzolites contain
337 clinopyroxenes with pronounced LREE depletions, $(\text{Ce}/\text{Yb})_{\text{CI}} = 0.05 - 0.52$, and narrow range for
338 MREE and HREE concentrations, $(\text{MREE})_{\text{CI}} = 7.2 - 14.4$; $(\text{HREE})_{\text{CI}} = 6.5 - 15.1$; $(\text{Yb})_{\text{CI}} = 8 -$
339 13 . Core-rim variation is generally not remarkable, except for samples KH21 and KH22, which

340 display more significant rim enrichments in the most incompatible LREEs ($La_{core} = 0.02$ ppm;
341 $La_{rim} = 0.12$ ppm in lherzolite KH21, and $La_{core} = 0.42$ ppm; $La_{rim} = 1.2$ ppm; $Ce_{core} = 1.71$ ppm;
342 $Ce_{rim} = 2.66$ ppm in lherzolite KH22). The two websterite samples in the Kilbourne Hole sample
343 suite have contrasting trace element patterns. Websterite KH35 has a smoothly ascending LREE
344 slope, peaking at Eu values, with lightly depleted LREE, $(Ce/Yb)_{CI} = 0.7$, and relatively flat MREE
345 and HREE patterns, with values clustering closely around 10 x CI chondrite, resembling the
346 patterns observed in KH lherzolites. In contrast, websterite KH31 has LREE-enriched
347 clinopyroxenes with pronounced convex upward REE patterns with apex at neodymium $((Nd)_{CI} =$
348 $41)$, and steadily descending slope in the MREE and HREE ranges $((MREE)_{CI} = 20 - 35$; $(HREE)_{CI}$
349 $= 8.3 - 20.8)$, very similar to the wehrlite sample KH28. Websterite KH31 has a LREE/HREE
350 ratio of $(Ce/Yb)_{CI} = 3.2$, while the wehrlite KH28 has a slightly lower LREE/HREE value of
351 $(Ce/Yb)_{CI} = 2.3$.

352 Primitive mantle-normalized extended trace element patterns of clinopyroxenes in the Kilbourne
353 Hole samples (Fig. 4.D) are defined by generalized depletion in large ion lithophile elements
354 (LILEs) with the exception of a strong rim-enrichment in Ba, observed in lherzolite KH25, and by
355 relative enrichments in the more incompatible high strength field elements (HFSEs), U and Th.
356 Other, more compatible HFSE, such as Zr, Hf and Ti show variable degrees of depletion compared
357 to adjacent elements in the pattern: $Zr/Zr^* = 0.18 - 0.75$; and $Hf/Hf^* = 0.34 - 0.99$; and $Ti/Ti^* =$
358 $0.42 - 0.88$, where the lower end of the spectrum characterizes the websterite-hosted
359 clinopyroxenes, while the higher end of the spectrum characterizes the lherzolite and wehrlite-
360 hosted clinopyroxenes. Strontium content in KH lherzolites is limited to 18 – 42 ppm, and shows
361 moderate negative anomalies $Sr/Sr^* = 0.57 - 0.93$, and even small positive anomalies ($Sr/Sr^* = 1$
362 $- 1.1$) in sample KH30 and in the core of KH32. In wehrlite and websterite samples, Sr is distinctly
363 enriched (56 – 142 ppm) but, due to comparable enrichments in similarly compatible elements,
364 Sr/Sr^* ratios remain within the range of KH lherzolite samples.

365 4.3. Thermobarometry

366 Thermometry was performed on mineral cores and rims using Brey and Kohler (1990) as well as
367 the more recent calibration of Putirka (2008) (Fig. 5). Fig. 5 shows that there is very good
368 agreement between the two calibrations and hardly any differences between cores and rims of rock

369 forming minerals between the various lithologies. The temperature range is between 1000-1100
370 °C, which is clearly in the realm of lithospheric mantle temperatures, although they represent a hot
371 lithospheric mantle. These values are fully consistent with determinations made previously by
372 Galer and O’Nions (1989). Pressures of equilibration were calculated using Putirka (2008) and
373 indicate depth ranges between 0.9 -1.5 GPa, with the greater depths in that range characterizing
374 the Kilbourne Hole xenoliths. This is consistent with the slightly deeper Moho known independently
375 at that location (38 km) versus San Carlos where the Moho is at around 33 km. Overall our
376 pressure-temperature calculations are consistent with the high heat flow values reported from
377 Kilbourne Hole region (95 mW/m²). Clearly, these xenoliths at both localities span the shallowest
378 range of mantle (lithosphere) immediately below their respective Moho discontinuities but not
379 deeper than ~ 50 km.

380 4.4. Depletion of peridotites

381
382 Peridotites studies here have various degree of depletion based on modal mineralogy. However,
383 most of them are lherzolites and do not show extreme depletions as seen in the suboceanic mantle
384 (abyssal peridotites, etc.). One way of quantifying the amount of depletion is to use Y and Yb in
385 clinopyroxenes from peridotites, in order to determine how depleted these peridotites are relative
386 to a chondritic mantle (in %). This would broadly correspond to the percentage of melt lost from
387 a chondritic-like mantle. Results of this simple forward model are shown in Fig. 6 for both San
388 Carlos and Kilbourne Hole and show that the majority San Carlos and all of Kilbourne Hole
389 peridotites display small (<5%) amounts of depletion, with a few outliers at San Carlos near the
390 10% depletion mark. These results as well as the overall trace elemental patterns in spiderdiagrams
391 (Fig. 4) together argue against an oceanic mantle origin for these peridotites on melt depletion
392 arguments alone. Radiogenic isotopes (Sr, Nd, Pb) (Galer and O’Nions, 1989) are more ambiguous
393 at least for San Carlos, as they can be assigned to the MORB mantle field.

394

395 5. Interpretations

396 The results above have some important implications for the origin and age of the continental
397 mantle lithosphere currently underlying the American Southwest at 30-50 km. We summarize
398 them below:

- 399 a. The lack of any diffusional gradients from core to rim (cooling paths) suggest that
400 temperatures are recent; in other words, the heating of the lithospheric mantle recorded by
401 the xenoliths took place relatively recently, possibly during continental extension in the
402 Quaternary at both locations;
- 403 b. The same argument applies to metasomatism, which is probably also recent, and related to
404 the last Cordilleran orogenic cycle. It is unlikely that pyroxenite veins emplaced in a
405 previous Wilson cycle would show such uniform temperatures from core to rim;
- 406 c. The mild depletion displayed by peridotites from both locations in both Mg# as well as
407 trace elemental patterns is inconsistent with a Farallon (accreted oceanic slab) origin;
408 instead the mantle lithosphere at both locations is interpreted to be native North American;
- 409 d. Key markers such as Cr# in spinels and Mg# in olivines are also inconsistent with a cratonic
410 origin for the shallow mantle beneath the American Southwest; it is either a modified (re-
411 fertilized) cratonic mantle or more likely one formed as the root of the Mid-Proterozoic
412 terranes of the SW U.S.A and modified by Cordilleran-related metasomatism during the
413 Cenozoic.

414

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421 science data platform OSF at osf.io/rkv38. This contribution has the following document identifier
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423

424 **Figure captions**

425 **Fig. 1:** Simplified sketch map of the southwestern United States showing the distribution of
426 Cenozoic basaltic and basaltic andesite rocks, within the context of major physiographic provinces

427 of the region, modified after Kempton (1991). The locations of mantle xenoliths collected for this
428 study are shown in orange- San Carlos, AZ, and green- Kilbourne Hole, NM.

429 **Fig. 2.** A: Mineral modal composition for the investigated xenolith samples. B- I: Back-scattered
430 electron photomicrographs of representative textures observed in the studied samples: B -
431 protogranular (SC06); C - porphyroclastic (KH22); D – equigranular (SC11); E – poikilitic (SC09).

432 **Fig. 3:** Plot of spinel Cr-number vs. olivine Mg-number for peridotites (circles) and pyroxenites
433 (squares) from Kilbourne Hole (KH, red symbols) and San Carlos (SC, grey symbols), within the
434 olivine-spinel mantle array (OSMA), of Arai (1994). Domains for various tectonic settings (Arai,
435 1994) are shown for comparison. The melt depletion trend, marked by an arrow, is projected from
436 the Primitive Mantle (PM) composition of McDonough and Sun, (1995). $Cr\# = 100 \times Cr / (Cr +$
437 $Al)$; $Mg\# = 100 \times Mg / (Mg + Fe_{tot})$, normalized cation ratios; Fe_{tot} = total iron as ferrous. Wehrlite
438 samples were not plotted owing to their low (< 85) Mg#. Lher– lherzolite; web– websterite.

439 **Fig. 4:** CI chondrite-normalized rare earth element (REE) (panels A and C) and Primitive mantle
440 (PM)-normalized extended trace element abundances (panels B and D) of clinopyroxenes (cpx)
441 from San Carlos and Kilbourne Hole mantle xenoliths. Clinopyroxenes and whole-rock REE
442 compositions from various localities of mantle xenoliths along the Laramide corridor (diagonal
443 pattern field) (Geronimo, AZ: Menzies et al., 1985; Wikieup, AZ: McGuire & Mukasa, 1997; Dish
444 Hill and Cima, CA: Luffi et al., 2009; Kilbourne Hole, NM: Harvey et al., 2012; Cemetery Ridge,
445 AZ: Haxel et al., 2015; Crystal Knob, CA: Quinn et al., 2018), as well as for abyssal peridotites
446 (solid grey field) from Johnson et al., (1990) are shown for comparison. Normalizing values are
447 from McDonough and Sun (1995). Solid lines represent mineral core values, whereas dashed lines
448 represent mineral rim values. Lher– lherzolite; web– websterite; wher- wehrlite. The legend in
449 panel A applies to the entire figure.

450 **Fig. 5:** Geothermobarometry on San Carlos (grey symbols) and Kilbourne Hole (black symbols)
451 xenoliths: A: Correlation of core and rim temperatures calculated using major element-based two-
452 pyroxene geothermometer of Brey and Köhler, (1990), (T_{BKN}). Dashed lines represent $15^{\circ}C$
453 deviations from the 1:1 correlation line. B: Correlation of core and rim temperatures calculated
454 using major element-based two-pyroxene geothermometer of Putirka, (2008), ($T_{Putirka}$); C:
455 Correlation of calculated core temperatures between the two geothermometers used: T_{BKN} and

456 T_{Putirka}; D: P-T diagram, using calculated core temperatures (T_{Putirka}) and pressures (P_{Putirka}). Four-
457 phase peridotite stability fields are from Gasparik, (1984).

458 **Fig. 6:** Partial melting trends from Y and Yb in clinopyroxenes, following a generic model of
459 fractional melting of a primitive chondritic mantle. Y_{PM}= 3.34 ppm; Y_{bPM}= 0.346 ppm from
460 McDonough & Sun (1995).

461

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531 western United States.

Figure 1











