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RESEARCH ARTICLE

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Key Points:

- Natural samples reveal complex relationships between olivine CPO and deformation conditions
- Olivine CPO types vary more often with strain geometry than water and stress in studied peridotites
- B-type olivine CPO can form under a wide range of temperatures and at low stresses and water contents

Supporting Information:

Supporting Information S1

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Relationships Between Olivine CPO and Deformation Parameters in Naturally Deformed Rocks and Implications for Mantle Seismic Anisotropy

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Abstract We analyze peridotites from a wide range of tectonic settings to investigate relationships between olivine crystallographic preferred orientation (CPO) and deformation conditions in naturally deformed rocks. These samples preserve the five olivine CPO types (A- through E-type) that rock deformation experiments have suggested are controlled by water content, temperature, stress magnitude, and pressure. The naturally deformed specimens newly investigated here (65 samples) and compiled from an extensive literature review (445 samples) reveal that these factors may matter less than deformation history and/or geometry. Some trends support those predicted by experimentally determined parametric dependence, but several observations disagree-namely, that all CPO types are able to form at very low water contents and stresses and that there is no clear relationship between water content and CPO type. This implies that at the low stresses typical of deformation in the mantle, CPO type more commonly varies as a function of strain geometry. Because olivine CPO is primarily responsible for seismic anisotropy in the upper mantle, the results of this study have several implications. These include (1) the many olivine CPO types recorded in samples from individual localities may explain some of the complex seismic anisotropy patterns observed in the continental mantle, and (2) B-type CPO—where olivine's "fast axes" align perpendicular to flow direction-occurs under many more conditions than traditionally thought. This study highlights the need for more experiments and the difficulty in using olivine CPO in naturally deformed peridotites to infer deformation conditions.

1. Introduction

Seismic anisotropy in the upper mantle is produced primarily by crystallographic preferred orientation (CPO) developed in olivine during ductile deformation by the common mechanism of dislocation creep (Nicolas & Christensen, 1987). The manner in which olivine's principal axes—[100], [010], and [001]—align with respect to the shear direction is often categorized by CPO "types." These types are commonly referred to as A-, B-, C-, D-, and E-type (cf. Karato et al., 2008). A-type, the most frequently occurring configuration, describes a CPO pattern where olivine's [100] axes align in the shear direction, [010] axes align normal to the shear plane, and [001] axes align within the shear plane but normal to the shear direction (Figure 1). Because of the frequent occurrence of this type—and because seismic waves travel fastest along olivine's [100] plane—it is typically assumed that in the Earth's mantle, seismic fast directions are oriented in the directions, so it is critical to understand how and why these CPO types form, particularly for those types that contradict such an assumption.

The patterns, or morphologies, of olivine CPOs—schematically represented in Figure 1—are assumed to reflect the activation of one or more specific slip systems: [100](010) for A-type, [001](010) for B-type, [001](100) for C-type, and [100](001) for E-type. D-type is expected when [100](010) and [100](001) are of similar strength. These slip systems are activated as a function of their orientation with respect to applied stress, and their Peierls stress or "yield strength," which can be affected by the physical conditions of deformation (e.g., Bai et al., 1991; Mackwell et al., 1985). Numerical modeling efforts have demonstrated that



(adjusted for Bell et al. [2003] calibation)

Figure 1. Five types of olivine CPO are shown with pole figures and schematic-oriented crystals in water versus stress space as determined from the experiments of Bystricky et al. (2000), Zhang et al. (2000), Jung and Karato (2001), Katayama et al. (2004), and Jung et al. (2006). Water contents, originally obtained using the calibration of Paterson (1982) have been multiplied by 3.5 here to adjust for the preferred calibration of Bell et al. (2003).

most CPO types can be reproduced by varying both "strain geometry" (orientation and magnitude of the principal strain axes with respect to the deformation plane) and slip system strengths (Becker et al., 2008; Kaminski, 2002; Ribe & Yu, 1991; Tommasi et al., 1999; 2000; Wenk et al., 1991). In recent decades, several experiments have focused on the relationship between CPO types and physical deformation conditions, most notably temperature, water content, and deviatoric stress magnitude (e.g., Couvy et al., 2004; Jung & Karato, 2001; Jung et al., 2006; Katayama & Karato, 2006; Katayama et al., 2004; Ohuchi et al., 2012; Wang et al., 2019). These high temperature and pressure experiments found that at experimental conditions, the less common B- through E-type CPOs form when water contents (specifically hydrogen, stored as a point defect and expressed as ppm H₂O by weight) and/or deviatoric stress magnitudes are elevated relative to conditions resulting in A-type CPO (Figure 1). Several additional recent studies, however, have shown that factors other than water and stress-such as temperature, pressure, deformation mechanism, deformation history, deformation geometry, strain magnitude, and/or the presence of melt-can also affect slip system strength and/or contribute to the development of olivine CPO (e.g., Boneh & Skemer, 2014; Hansen et al., 2014; Jung et al., 2009b; Katayama & Karato, 2006; Précigout & Hirth, 2014; Sundberg & Cooper, 2008; Qi et al., 2018). B-type, for example, has been suggested to form under high pressures, lower temperatures, and/or during grain boundary sliding in some experiments (Katayama & Karato, 2006; Précigout & Hirth, 2014; Sundberg & Cooper, 2008). Other examples include AG-type (also known as axial-010 or "a-c switch")—a sixth CPO type suggested to form during compression (e.g., Nicolas et al., 1973) or in the presence of melt (Holtzman

et al., 2003; Qi et al., 2018)—and D-type, which Hansen et al. (2014) demonstrated can develop during dislocation-accommodated grain boundary sliding at lower strains than A-type CPO.

By comparison with experiments, natural peridotites are deformed at orders of magnitude lower strain rates and under a wider range of temperatures, pressures, deviatoric stresses, and strain magnitudes than is currently accessible in experiments. Additionally, experiments are conducted (with few exceptions) on initially undeformed samples or synthetic olivine aggregates that generally lack any preexisting textures or CPOs; whereas in most geologic settings, the lithospheric mantle has experienced multiple phases of deformation, and recent studies indicate that past deformation history is not easily erased, with even completely annealed samples preserving strong CPOs (e.g., Boneh et al., 2017; Farla et al., 2011; Webber et al., 2010). Our aim in this paper is to examine CPO development in rocks from natural settings that have likely experienced multiple phases of deformation and highly variable degrees of finite strain. To do this, we use 65 naturally deformed peridotite samples from xenoliths, continental mantle massifs, and ophiolites to explore any trends between deformation conditions and olivine CPO in natural peridotites. Our results are presented alongside samples compiled from 48 previously published studies in which olivine CPOs were measured. We compare our natural data set to existing experimental constraints, discuss similarities and differences between these two data set types, and explore implications for seismic anisotropy in the mantle.

2. Sample Descriptions

Sixty-five peridotites were newly analyzed for this study (Figure 2 and Tables 1 and 2). Forty-two are xenoliths from the western United States: seven from the Rio Grande Rift region (two from Elephant Butte, two from Cerro Chato, and three from Cerro de Guadalupe); six from San Carlos Volcanic Field in Arizona; three from Kilbourne Hole in New Mexico; 11 from Lunar Crater Volcanic Field; 15 from the Mojave (seven from Cima Volcanic Field and eight from Dish Hill). An additional five samples derive from the Navajo Volcanic Field in the four-corner region of the western United States and are diatreme-hosted inclusions rather than xenoliths. The remaining six xenoliths—five from the San Quintin Volcanic Field in Mexico and one from Eifel Germany—come from outside of the western United States. Ten peridotites come from continental massifs: two are from the Ivrea Zone in Italy and eight are from the Bjørkedalen Peridotite in western Norway. The last two peridotites come from the Bay of Islands Ophiolite Complex in western Newfoundland. Additional details on these localities can be found in Text S1 of the supporting information.

Most xenolith specimens are spinel lherzolites, wehrlites, and harzburgites (Table 1). Exceptions include some from Lunar Crater that are dunites and samples from Norway that contain primarily olivine and tremolite with small amounts of anthophyllite, orthopyroxene, clinopyroxene, and spinel. These samples, and some from the Navajo locality that contain chlorite and antigorite, are the only ones that contain hydrous minerals stable under mantle conditions. The peridotites from Newfoundland are the only samples that appear to have been altered significantly through late-stage serpentinization. None of the newly studied peridotites herein contain garnet.

Textures in the samples range from granular to protogranular to porphyroclastic to mylonitic. Granular samples are characterized by large grains (>2 mm) with abundant triple junctions, no foliation, and only very minor subgrain development in olivine. Protogranular samples have moderately large grains (1-2 mm) and show weak grain elongation with occasional evidence of internal grain deformation. Porphyroclastic samples exhibit large, elongate olivine and orthopyroxene porphyroclasts with irregular grain boundaries and substantial dynamic recrystallization. Mylonitic samples have pervasive dynamic recrystallization, finer grain sizes, and exhibit strong foliations with porphyroclasts that are smaller and significantly more elongate than those in the porphyroclastic category.

Most porphyroclastic to mylonitic samples preserve evidence of deformation via dislocation creep as indicated by (1) internal lattice deformation (e.g., subgrains and undulose extinction); (2) "core-and-mantle" microstructures in which elongate porphyroclasts are surrounded by smaller, equant, dynamically recrystallized grains lacking in internal deformation. Some exceptions are dunite samples from Norway that show evidence for dislocation-accommodated grain boundary sliding, such as four-grain junctions, straight grain boundaries parallel to foliation, shape preferred orientation (SPO), relatively weak CPO, and minimal internal deformation (e.g., White, 1977). Additional descriptions and images of these microstructures can be found in Text S2 and Figures S1 and S2 in the supporting information.





Figure 2. (a) Localities of samples analyzed in this study as well as represented in the literature compilation (48 published studies representing 445 peridotite samples from 52 localities). A list of these studies and citations can be found in Appendix A. (b) A breakdown of the 510 individual samples included in this study and in the literature compilation by peridotite type. (c) A breakdown of the 510 individual samples included in the literature compilation by olivine CPO type. Inset: Pie charts show that a comparison of the proportion of CPO types in this compilation are similar to the compilation of Ismaïl and Mainprice (1998), shown in gray. CPO = crystallographic preferred orientation.

Deformation temperatures for all samples were estimated based on previously published work, which reported temperatures to varying precision. The inferred temperatures, and other information on these samples, can be found in Tables 1 and 2.

Peridotite				Tempera	ture ^d			Water Content		
Peridotite				1						
trand	CPO-	Microstructural	QC	T (°C,	Т°С	Method ^e	OL ppm H ₂ O	$OL ppm H_2O$	OPX ppm	CPX ppm
- in the -	type ^b	category	(MPa)	used in plots)	range		(measured) ^f	(calculated) ^g	H_2O	$\rm H_2O$
Х	А	protogranular	15	1018	I	SIMS	17	24	189	395
Х	ц	protogranular	25	1022	988-1056	SIMS	13	10	94	124
Х	A	porphyroclastic	21	1022	988-1056	I			Ι	Ι
х	A	porphyroclastic	20	1022	988-1056	SIMS	5	5	11	118
Х	А	protogranular	16	1022	988-1056	SIMS	9	11	66	I
Х	D	porphyroclastic	23	875	800-950	SIMS	5	20	184	292
Х	AG	porphyroclastic	22	1022	988-1056	SIMS	5	7	67	92
Х	AG	porphyroclastic	24	1022	988-1056	I	Ι		I	I
Х	А	porphyroclastic	Ι		I	I			I	I
Х	C	porphyroclastic	17	766	Ι	SIMS	7		Ι	Ι
Х	ż	porphyroclastic	Ι	797	I	SIMS	9	3	28	53
Х	ż	protogranular	Ι		Ι	I			Ι	Ι
х	ż	porphyroclastic	Ι		I				I	Ι
x	A	porphyroclastic			I	I			Ι	
Х	A	protogranular	12	935	026-006	SIMS	18	19	I	271
U	в	porphyroclastic	34	1100	1000-1200	SIMS	38		I	
C	н	protogranular	19	1100	1000-1200	SIMS	8		I	
x	в	porphyroclastic	21	1050	1000 - 1100	SIMS	1	16	137	241
x	AG	porphyroclastic	27	1050	1000 - 1100	I			I	
х	в	protogranular	25	1050	1000 - 1100	SIMS	1	24	174	413
х	V	mylonitic			I	I			I	
x	AG	porphyroclastic	18	600	500-700				I	
x	A	porphyroclastic	87	600	500-700	I			Ι	
x	AG	mylonitic	25	600	500-700	SIMS	43	75	318	1657
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7 porphyroclastic -1 -1 X 7 porphyroclastic -1 -1 X A porphyroclastic 24 1100 Z B porphyroclastic 21 1050 X A porphyroclastic 27 1050 X A mylonitic -1 -1 X A porphyroclastic 27 1050 X A mylonitic -1 -1 X A porphyroclastic 27 1050 X A porphyroclastic 27 1050 X A porphyroclastic 18 600 X A porphyroclastic 87 600 X A porphyroclastic 25 600 X A porphyroclastic 25 600	xAGporphyroclastic24 10.22 $900-10.00$ XZporphyroclastic17 997 $$ X?porphyroclastic $$ $$ $$ X?porphyroclastic $$ $$ $$ X?porphyroclastic $$ $$ $$ X?porphyroclastic $$ $$ $$ XAporphyroclastic $$ $$ $$ XAporphyroclastic $$ $$ $$ XAporphyroclastic 24 1100 $1000-1200$ CBporphyroclastic 24 1100 $1000-1200$ XAporphyroclastic 21 1050 $1000-1100$ XBporphyroclastic 21 1050 $1000-1100$ XAporphyroclastic 21 1050 $1000-1100$ XAporphyroclastic 21 1050 $1000-1100$ XAporphyroclastic 21 1050 $1000-1100$ XAporphyroclastic 27 1050 $1000-1100$ XAporphyroclastic 27 1050 $1000-1100$ XAmylonitic $$ $$ $-$ XAporphyroclastic 27 1050 $1000-1100$ 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Table 1

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Table 1 Cov	ıtinued												
							Temper	ature ^d		F	Water Content		
Sample	Locality	Locality	Peridotite	CPO-	Microstructural	مر	T (°C,	T °C	Method ^e	OL ppm H ₂ O	OL ppm H ₂ O	OPX ppm	CPX ppm
no.	name	no.	type ^a	type ^b	category	(MPa)	used in plots)	range		(measured) ^f	(calculated) ^g	H_2O	H_2O
RDH44	Dish Hill	14	Х	ż	protogranular	21	1000	900-1100	I			I	Ι
RDH49	Dish Hill	14	Х	Ц	porphyroclastic	20	166	Ι	I		I	Ι	I
RNF7	Newfoundland	51	0	Α	mylonitic	Ι	1075	1050-1100	SIMS	81	23	209	I
RNF8	Newfoundland	51	0	ż	mylonitic	Ι	1075	1050-1100	SIMS	57		Ι	Ι
SQ1	San Quintin	7	Х	D	porphyroclastic	36	875	800-950	SIMS	13	19	175	I
SQ2	San Quintin	7	Х	D	porphyroclastic	27	875	800-950	SIMS	4	27	270	339
sQ3	San Quintin	7	Х	D	mylonitic	23	875	800-950	SIMS	21	20	182	288
SQ5	San Quintin	7	Х	ż	porphyroclastic	32	875	800-950	SIMS	11		Ι	Ι
V17A	Norway	41	C	ż	porphyroclastic	70	725	700-750	Ι			Ι	Ι
V24A	Norway	41	C	в	porphyroclastic	24	725	700-750	SIMS	57		I	I
WCiVb36	Cima	15	Х	U	porphyroclastic	I		Ι	Ι			Ι	Ι
WCiVb46	Cima	15	Х	ы	porphyroclastic	24	970	I	FTIR		10	95	128
WCiVb47	Cima	15	Х	A	protogranular	24	992	I	FTIR		26	239	I
Y12B	Norway	41	C	ż	protogranular	42	725	700-750	SIMS	87		Ι	I
YF12	Lunar Crater	11	Х	н	mylonitic	53	1258	Ι	SIMS	4	10	Ι	147
YF3	Lunar Crater	11	X	Ы	mylonitic	53	1250	1200-1300	SIMS	5		I	I
YF4	Lunar Crater	11	X	ż	mylonitic			I	I			Ι	Ι
YF9a	Lunar Crater	11	х	ы	protogranular	49	1219	I	SIMS	15	6	106	77
YFN11	Lunar Crater	11	X	ż	mylonitic	67	1219	I	SIMS	5	4	34	I
YFN13	Lunar Crater	11	Х	Е	porphyroclastic	52	1257	I	SIMS	15	8	82	92
^a Peridotite range inclu samples cor San Quintir Norway, Pas 16 to convel Otherwise, 1	ype: X = xenolith ded if exact T is un ne from Byerly and 1, Cabanes and Me sillas (2015); Newf th to ppm H/Si. ⁸ 1 this number is calc	; C = contin known for h Lassiter (1 Lassiter (1988 oundland, ff sample h culated fror	nental; $O = o$ given sample (212); Dish H (212); Eifel Gerr Varfalvy et al as both opx <i>a</i> as both opx <i>a</i>	phiolite. For the fill and C nany, Wii 1. (1996); tud cpx v or cpx.	^b "?" indicates CPO se samples, T used it tima samples from B tt-Eickschen et al. (2 Lunar Crater, Dyge water contents, this	type wa: n plots is ernard a 2003); Iv rt et al. (number	s inconclusive fr based on mediar nd Behr (2017). J rea Zone, Jin et a 2019). ^e SIMS an is an average of	om EBSD. ^c St value within kegional T ran I. (1998); Kilb alyses from th the two calcu	this range. I this range. I ges from th ourne Hole uis study; F lated olivin	ed from paleopi Exact sample T v e following studi , Satsukawa et a TIR analyses fror e contents based	ezometer of Van alues for Elephar es: San Carlos, G I. (2011); Navajo, n Bernard and B on partition coe	der Wal et al. It Butte and C aler and O'N Behr and Sr ehr (2017). ^f fficients for	(1993). ^d T erro Chato ons (1989); nith (2016); Multiply by sach phase.

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Properties of An	ıalyzed Sampl	es													
		EBSD-de	rived values		CT-derive	ed values ^a	Moda	l percent	ages (of 1	mapped ar	ea) ^b			Aspect ra	tios ^c
Sample name	M-index	J-index	BA-index	FIA-index	T	P'	Olivine	OPX	CPX	Spinel	Other	X/Z	X/Y	X/Z	Flinn constant, k
07EB4-1	0.11	3.71	0.52	55.11	0.27	1.23	65	13	20	2	0	Ι	Ι	I	I
114027-10	0.18	3.87	0.41	105.53	I	I	74	18	×	1	0	2.73	2.06	1.33	3.23
114027-16	0.13	3.24	0.54	33.12	I		74	15	10	1	0	2.01	0.93	2.16	-0.06
114027-23	0.11	3.00	0.48	45.72	I		64	17	16	7	0	3.03	2.09	1.45	2.44
114027-6	0.17	4.61	0.63	51.44	I	I	82	16	2	0	0	1.72	0.64	2.69	-0.21
116610-26	0.11	2.71	0.71	77.11	I		99	16	18	0	0	I	I		I
117200-205	0.12	2.72	0.30	8.63	I	I	66	19	15	0	0	2.90	1.22	2.38	0.16
117200-211	0.22	4.62	0.09	-1.76	0.33	1.43	54	33	12	7	0	I	I		I
BELB9-6a	0.16	3.42	0.51	77.19	0.36	1.51	71	16	12	1	0	Ι	Ι	I	I
CC07-1-22	0.04	1.67	0.57	173.00	0.08	1.20	77	12	10	1	0	Ι	I		I
CC07-2-1	0.05	2.28	0.37	109.25	0.05	1.21	79	7	14	1	0	Ι	Ι	I	I
CG07-1-26	0.02	2.20	0.57	-9.26	I	I	74	13	12	0	0	Ι	Ι	I	I
CG07-1-36	0.08	2.10	0.42	-24.14	0.76	1.34	71	15	13	1	0	I	I		I
CG07-1-52	0.06	2.34	0.36	41.16	-0.34	1.16	71	15	12	2	0	Ι	Ι	I	I
DW8	0.23	5.54	0.47	37.46	0.62	1.36	91	9	3	0	0	2.52	0.94	2.69	-0.04
17	0.05	2.14	0.26	-28.35	0.33	1.32	78	9	16	0	0	2.17	1.25	1.74	0.34
I8	0.04	1.70	0.63	120.23	0.22	1.30	69	14	17	0	0	2.03	1.98	1.03	38.43
Kb	0.12	2.72	0.47	-15.80	0.41	1.39	67	15	14	З	0	1.56	0.71	2.20	-0.24
KH1	0.14	2.64	0.31	2.18	0.49	1.48	57	21	20	7	0	2.42	0.87	2.78	-0.07
KH2	0.08	2.23	0.13	-6.19	-0.32	1.18	45	21	32	2	0	2.13	0.94	2.26	-0.04
Ki-5-319	0.08	2.60	0.44	56.06	I		72	10	17	0	0		I		I
N117	0.22	5.07	0.24	11.54	-0.24	1.86	85	7	7	0	1	2.60	1.09	2.39	0.06
N122	0.23	5.52	0.52	51.24	0.19	1.48	75	17	9	1	2	2.23	0.98	2.28	-0.02
N167	0.20	4.32	0.09	10.58	I		88	7	5	0	0	2.30	1.71	1.35	2.03
N178	0.29	6.49	0.16	-16.03	0.37	1.42	95	7	2	0	1	I			I
N188	0.17	4.55	0.85	93.58	0.77	1.40	68	23	8	0	0		I		I
016D	0.01	1.19	0.10	9.03	0.69	2.43	93	1	1	0	5	3.08	2.46	1.25	5.79
016E	0.04	1.45	0.36	43.44	0.47	2.24	91	7	1	0	9	2.31	1.17	1.97	0.17
018A	0.18	3.53	0.47	44.01	I	I	100	0	0	0	0	I			I
018B	0.01	1.35	0.68	-61.36	-0.25	1.24	98	0	7	0	0	I	I		I
018C	0.05	1.55	0.33	25.83	-0.87	1.35	66	0	0	1	0	1.72	0.84	2.05	-0.15
PL1	0.22	5.88	0.44	54.25	I	I	66	0	1	0	0		Ι		
PL10	0.09	3.02	0.82	89.61	I		86	12	2	0	0		Ι		I
PL11	0.37	17.38	0.54	179.20	I	I	I	Ι		I	I	Ι	I		I
PL13	0.31	5.98	0.74	100.88	I	I	82	17	1	0	0	Ι	Ι		I
PL6	0.23	7.93	0.20	149.52									I		

Table 2

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Table 2 Continu	pər	EBSD-de	rived values		CT-derive	d values ^a	Mod	al nercent	ages (of n	lanned are	q(e			Asnect rati	OSC
Sample name	M-index	J-index	BA-index	FIA-index	T	P'	Olivine	OPX	CPX	Spinel	Other	Z/X	J X/X	Y/Z	Flinn constant, k
RCiV8-1	0.01	1.55	0.57	92.23	-0.37	1.36	59	25	16	1	0	3.06	2.13	1.44	2.58
RCiV8-22	0.11	3.07	0.53	84.30	I	I	77	7	14	2	0		I	I	I
RCiV8-6	0.21	3.91	0.47	32.46	-0.10	1.26	68	16	14	2	0		Ι	I	
RDH11	0.10	2.98	0.63	80.35	0.21	1.32	79	11	6	2	0	I	Ι	I	I
RDH14	0.13	4.34	0.72	-34.90	Ι	Ι	59	31	8	1	0	I	Ι	I	I
RDH15	0.06	2.54	0.42	103.02	I	I	57	27	15	0	0	3.02	1.59	1.90	0.65
RDH23	0.13	3.46	0.58	105.30	-0.08	1.50	64	15	21	0	0	3.56	1.65	2.17	0.55
RDH26	0.05	2.46	0.36	54.81	0.10	1.44	64	14	21	1	0	3.05	1.77	1.72	1.07
RDH33	0.08	2.31	0.59	82.92	-0.30	1.44	80	15	5	0	0	3.32	1.51	2.20	0.42
RDH44	0.04	3.55	0.36	131.10	-0.34	1.40	49	27	23	1	0	3.44	2.45	1.40	3.62
RDH49	0.03	2.19	0.39	87.89	-0.54	1.31	60	22	17	1	0	3.33	2.19	1.52	2.30
RNF7	0.08	2.35	0.44	64.61	0.59	1.28	49	42	7	1	1	I		I	I
RNF8	0.06	2.20	0.30	123.21	0.25	1.26	70	21	4	4	1	I		Ι	I
SQ1	0.09	2.07	0.85	94.26	0.09	1.45	89	7	4	0	0	2.65	1.30	2.03	0.30
SQ2	0.11	2.42	0.51	125.89	0.51	1.25	99	18	14	2	0	2.49	1.12	2.22	0.10
SQ3	0.08	2.95	0.68	114.30	I	I	61	21	18	0	0	4.75	2.36	2.01	1.35
SQ5	0.11	2.32	0.69	-46.77	I	I	82	12	5	1	0			I	I
V17A	0.03	1.45	0.41	81.74	0.69	1.69	87	S	8	0	0	1.85	1.07	1.72	0.10
V24A	0.03	1.35	0.53	59.14	0.19	2.85	06	0	0	0	10	2.02	0.95	2.11	-0.04
WCiVb36	0.10	3.44	0.36	179.20	0.30	1.63	28	0	18	0	54	2.45	1.09	2.25	0.07
WCiVb46	0.08	2.86	0.31	102.00	-0.24	1.47	59	19	21	1	0	3.72	2.52	1.47	3.22
WCiVb47	0.11	3.09	0.57	79.47	-0.20	1.53	67	28	5	0	0	2.67	1.79	1.49	1.61
Y12B	0.04	1.43	0.74	31.59	0.17	2.59	66	0	0	0	0	I	Ι	I	I
YF12	0.09	3.22	0.58	103.79	I	I	93	0	7	0	0			I	I
YF3	0.22	9.39	0.52	89.52	-0.22	1.23	100	0	0	0	0	I	Ι	I	I
YF4	0.29	8.81	0.82	96.57	-0.22	1.18	66	0	0	0	0	3.63	1.34	2.70	0.20
YF9a	0.05	1.99	0.45	109.90	0.46	1.17	77	ю	20	0	0	I	I	I	I
YFN11	0.12	3.07	0.49	-62.76			90	6	1	0	0	3.11	1.07	2.89	0.04
YFN13	0.07	1.93	0.76	119.30	I	I	55	19	25	0	0	2.65	0.79	3.33	-0.09
^a From SLD met Normalized, but the lineation and	hod. See Tabl t may not sur d foliation. X/	e S3 in supp 1 to 100 due 'Y ratio calcu	orting informal to minor round ulated from (X/	tion for SVD v ling errors. ^c X 'Z)/(Y/Z). Flin	alues. ^b Moc : lineation v n constant. <i>i</i>	lal percenta within the fi k. calculated	iges as calcul oliation plan 1 from ((X/Y	ated with e; Y: perpo)-1)/((Y/Z	MTEX for endicular ()-1).	r mapped i to the line	area. Not n ation with	ecessarily in the foli	represer ation pla	ntative of t ine; Z: per	he sample overall. pendicular to both



3. Methods

A range of microanalytical techniques were applied to characterize olivine CPO and deformation conditions, as follows.

3.1. Water Contents Using SIMS

We used the Cameca 6f Secondary Ion Mass Spectrometer (SIMS) at Arizona State University to measure water concentrations in olivine, orthopyroxene and clinopyroxene, collecting data from seven mounts over three sessions. In addition to unknowns of every phase, each mount contained some of several olivine, clinopyroxene, and orthopyroxene standards: PMR53, CITI7210, GRR2334a, GRR16506 (Aubaud et al., 2007; Bell et al., 1995; Mosenfelder & Rossman, 2013a, 2013b; Mosenfelder et al., 2011) along with a well-established synthetic forsterite blank (GRR1017, 0 ppm H₂O by weight; Bell et al., 1995; Mosenfelder & Rossman, 2013a, 2013b) or San Carlos olivine grains with known water content of <3 ppm H₂O by weight (Marshall et al., 2018; all ppm H₂O references in this paper are in weight units). Additional information on sample preparation, applied blank corrections, calibration curves used ($R^2 = 0.89 - 0.98$), and reported standard errors can be found in the supporting information (Text S3, Tables S1 and S2, and Figure S3). Due to the well-documented diffusion of hydrogen in olivine grains during xenolith ascent (e.g., Demouchy et al., 2006; Peslier & Luhr, 2006), in situ (i.e., mantle) olivine water contents are often calculated from partition coefficients applied to measured pyroxene water contents ($D_{\text{opx/ol}} = 0.11$ and $D_{\text{cpx/ol}} = 0.07$; e.g., Demouchy et al., 2017; Warren & Hauri, 2014). If water contents were measured for both orthopyroxene and clinopyroxene, the calculated olivine water content was estimated by taking the average value calculated from each mineral.

3.2. X-Ray Computed Tomography (CT)

The Xradia microXCT 400 at UT Austin was used to isolate spinel—distinguishable from other phases in scans due to its relatively high density—to identify foliation and lineation in samples where this was not easily recognizable in hand sample, and to quantify the SPO of spinel grains with implications for strain geometry. FEI Avizo 8.0 software was used to create 3-D volume renderings and visualizations (Text S4 and Figure S4 in the supporting information). Quant3D Software was used to quantify the degree of SPO anisotropy and shape of spinel grains through the calculation of fabric tensor eigenvalues (Ketcham & Ryan, 2004; Text S4 and Table S3), which in turn were used to calculate the following: P', a parameter that ranges from one to infinity and increases with greater anisotropy (Jelinek, 1981), and *T*, a shape factor ranging from -1 to 1, where negative and positive values indicate prolate and oblate ellipsoids, respectively (Hossack, 1968).

3.3. Grain Size Measurements and Stress Magnitudes

Olivine aspect ratios, subgrain widths, and dynamically recrystallized grain sizes were measured using Zeiss Zen Pro software connected to a Zeiss Axio Imager M2m petrographic microscope. Measurements were averaged for at least 100 dynamically recrystallized grains or subgrains per sample. Prior work has demonstrated that this method yields virtually identical results compared to the linear intercept method conducted optically or using Electron Backscatter Diffraction (EBSD) data (Bernard & Behr, 2017). Average recrystallized grain sizes, after applying a correction factor of 1.2 to account for 2-D sectioning, were used to estimate deviatoric stress magnitudes during deformation based on the paleopiezometer of Van der Wal et al. (1993). Olivine aspect ratios in two orthogonally cut thin sections were used to calculate the constant, k, which defines the slope of a Flinn diagram and describes the shape of a strain ellipsoid as having experienced flattening (0 < k < 1, i.e., oblate shapes), plane strain (k = 1), or constriction (k > 1, i.e., prolate shapes; Table 2; Flinn, 1965).

3.4. EBSD

CPO in olivine was measured from polished thin sections with an Oxford EBSD detector installed in the Phillips/FEI XL30 Environmental SEM at the University of Texas at Austin using a 20- to 25-kV accelerating voltage, 15- to 20-mm working distance, 30–40X magnification, and 10- and 50-micron step sizes. Large area maps were acquired using Oxford Instruments AZtec software (version 2.1), and postprocessing was conducted using MTEX 4.4.0 toolboxes and included noise reduction with a smoothing spline filter (Bachmann et al., 2010). MTEX was also used to estimate mineral modes, make lower hemisphere projection pole figures contoured as one-point-per-grain, perform subgrain misorientation analyses, calculate fabric strengths using the M- (Skemer et al., 2005) and J-indices (Bunge, 1982), and to calculate two CPO orientation indices: the BA-index, which quantifies CPOs from zero to one where zero represents D-type





Figure 3. Frequency of olivine CPO types present at each individual locality. Locality numbers are identified in Table 1 and the supporting information Table S5.

CPO and one represents AG-type CPO, and the Fabric Index Angle (FIA-index), which allows a CPO to be expressed as a single angle (Mainprice et al., 2015; Michibayashi et al., 2016) (see Table S4 in the supporting information).

4. Inclusion and Treatment of Previously Published Data

In addition to measurements on our own samples, we compiled data from a literature review. Data come from 48 studies published in the past 15 years, yielding 445 individual peridotite samples (see Appendix A and supporting information Text S5 and Table S5 for a list of studies and breakdown of which studies have particular CPO types and analyses). The vast majority of these peridotites are xenoliths and samples from continental massifs; 11 samples come from ophiolitic settings and four samples come from one abyssal locality (Figures 2 and 3). The treatment of this previously published data was not always straightforward, particularly when it came to including data for stress, deformation temperature, and water content. In particular, a comparison of water contents involves the nontrivial task of comparing data obtained from olivines and pyroxenes, using SIMS and Fourier-transform infrared spectroscopy (FTIR) (both polarized and unpolarized) and correcting for differences in calibrations of Paterson (1982) and Bell et al. (2003) (see section 7.1.2). A discussion of this is included in the supporting information Text S5.

5. Results

5.1. CPO Types

The results of the analytical work conducted on our own samples are provided in Tables 1 and 2. We found that all documented CPO types were represented: 18 samples had A-type, five B-type, three C-type, five D-type, 15 E-type, and eight AG-type (Figure 2). Eleven samples had ambiguous CPOs that could not definitely be categorized into one of these six types. Three of these "inconclusive" samples displayed a bimodal C-E-type CPO identical to that explored recently in Wallis et al. (2019).

The 445 peridotite samples from the literature plus our own 65 samples total 510 samples (Figure 2). The frequency of CPO types within this data set is broadly consistent with the compilation of Ismaïl and Mainprice (1998), which included fabrics for 110 peridotites from a wide range of geologic settings. Our data set includes a smaller proportion of A-type samples than this prior compilation (29% versus 50%) with



Figure 4. Example subgrain misorientation analysis of sample WCiVb47. (a) An individual olivine porphyroclast shown outlined (white dashes) in an olivine inverse pole figure map over band contrast map. (b) The same grain colored by how individual orientation measurements within it deviate from the mean orientation of the grain (mis2mean). The subgrain misorientation transect across this grain allows for the identification of the likely rotation axis, slip plane, and slip direction accommodating deformation. Assuming the subgrains are tilt boundaries, the clustering of the [001] axes and placement of [100] and [010] spreads relative to lineation suggest activity of the [100](010) slip system. (c) Schematic drawing of the interpretation. Figure modified from Bernard and Behr (2017).

more representation of the other CPO types; this is likely a reflection of community interest in these more "exotic" CPO types over the past 15 years. Nevertheless, the observation that our new compilation exhibits the same relative proportions of CPO types as Ismaïl and Mainprice (1998) (where A-type is the most common, followed by D-, AG-, B-, E-, and C-types in that order), suggests that these relative abundances may be representative of peridotites globally. There is most likely no sample overlap, as all studies included in our compilation were published after Ismaïl and Mainprice (1998).

5.2. Fabric Strength

The following results refer exclusively to the samples analyzed in this study (and exclude data from the literature compilation). We examined trends relating to fabric strength, or the degree to which grains are aligned, as this is often used as a metric to assess whether a CPO has reached steady state in experiments (cf. Skemer & Hansen, 2016). M- and J-indices range from 0.01–0.37 and 1.2–17.4, respectively. There was a strong agreement between M- and J-indices ($R^2 = 0.72$), with the possible exception of D-type samples, whose M-indices appeared to overpredict relative to the J-index (Figure S5). There did not appear to be any strong correlation between olivine CPO type and fabric strength.

5.3. Modal Percentages

We examined whether the modal percentage of olivine affects the CPO type, since the presence of other phases (namely, clinopyroxene and orthopyroxene) may inhibit the formation of CPO in olivine through a mechanism such as Zener pinning (Smith, 1948). Olivine modal percentages for the peridotites studied range from 45% to 100% (with the exception of one olivine gabbro with 28% olivine). We found no relationship between olivine CPO type and modal percent with the exception of D-type, which seemed to occur over a narrower range of modal percentages (Figure S5).

5.4. Microstructural Categories

Within our own suite of 65 samples, 16 were classified as protogranular, 31 as porphyroclastic, 15 as mylonitic, and three as ultramylonitic (Table 1). We examined whether olivine CPO type is affected by microstructural category, as this may be a proxy for strain magnitude, and strain magnitude has been suggested to influence CPO development within both naturally and experimentally deformed peridotites (e.g., Hansen et al., 2014; Warren et al., 2008). While we see no evidence that CPO varies by microstructural category globally (Figure S6), there are two individual localities for which olivine CPO does seem to correlate with microstructural category. In the suite of Mojave xenoliths, for example, granular and protogranular samples (i.e., high strain) consistently preserve A-type CPO, whereas porphyroclastic and mylonitic samples (i.e., high strain) typically display E-type CPO. In the suite of high-temperature xenoliths from Lunar Crater volcanic field (cf. Dygert et al., 2019), highly strained ultramylonitic samples display either C-type CPO (or an unusual bimodal C-E-type hybrid that resulted in some samples being categorized as "inconclusive," Table S4), while the lower strained porphyroclastic samples preserve E-type CPOs.



Figure 5. Bar chart showing the slip systems identified from subgrain misorientation analyses in each sample, which are themselves grouped by the samples' bulk CPO type. Slip systems are colored to match the CPO type associated with each slip system. CPO = crystallographic preferred orientation.

5.5. Slip Systems

To confirm whether our CPOs reflect deformation along particular slip systems, we performed subgrain misorientation analyses to determine the slip systems active for specific porphyroclasts containing subgrain boundaries (Figure 4 and Text S6). Evidence for all known olivine slip systems—[100](010), [001](010), [001](010), and [100](001)—was identified within this data set; [100](010) and [100](001) were the most common. The results of these analyses reveal a complicated relationship between CPO type and active slip systems, with many samples exhibiting subgrain misorientation patterns indicative of incompatible slip systems (Figure 5). With few exceptions, CPOs plotted for subsets of small (recrystallized) and large (porphyroclasts) grains both reflect the same bulk CPO, suggesting that these contradictory slip systems do not simply reflect a new CPO developing in the most recent stage of deformation. Interestingly, only one sample with B-type CPO had subgrains preserving the [001](010) slip system, as would be expected for dislocation creep under simple shear.

5.6. Water Content

Samples in this study record measured water contents for olivine, orthopyroxene, and clinopyroxene of 1–87 (n = 38), 11–318 (n = 27), and 53–1,712 (n = 24) ppm H₂O (16–1,392, 121–3,498, 636–20,544 ppm H/Si), respectively, and calculated olivine water contents (from partition coefficients with pyroxenes) of 3–75 ppm H₂O or 48–1,200 ppm H/Si (n = 30). Plots of olivine versus pyroxene water contents rarely agree with experimental partitioning predictions (Figure 6). This observation supports the notion that measured olivine water content is an unreliable indication of in situ mantle water contents, particularly in mantle xenoliths (e.g., Demouchy et al., 2006; Ingrin & Skogby, 2000; Mackwell & Kohlstedt, 1990; Warren & Hauri, 2014). The





Figure 6. Water contents from this study obtained through SIMS analyses. (a-c) Water contents for the three phases are plotted against one another. The samples from this study are plotted alongside the extensive compilation of Warren and Hauri (2014), and the partition coefficients lines of that study based on experimental and natural samples. (d) Average core and rim water content values measured for the samples in this study, colored by phase. Inset shows zoomed in view of lower water content values (gray rectangle).

strong agreement between orthopyroxene and clinopyroxene water contents, their consistency with experimental partition coefficients, and the lack of systematic variation from core to rim in pyroxenes (Figure 6) all support the assumption that pyroxenes do preserve original in situ water contents. Only core measurements were used to calculate the average water contents for each phase presented in Table 1.

5.7. Deviatoric Stress, Temperature, and Strain Geometry

Stresses estimated using paleopiezometry range from 11–87 MPa and deformation temperatures range from ~600–1300 °C. The wide range of values calculated for olivine thin section-derived Flinn constant, k, as well as the spinel CT-derived shape parameter, T, suggest that these samples represent a wide range of strain geometries: k and T ranged from 0.04–5.8 (n = 24) and –0.87–0.77 (n = 42), respectively (P' ranged from 1.16–2.85; Figure 9 and Table 2).





Figure 7. (left) Samples from nature shown along samples from experiments in stress-water space, colored by olivine CPO type. Boundaries projected from experimental conditions to the lower stresses < 100 MPa typical of much of the mantle. All water contents (including that of experiments) that use the Paterson (1982) calibration have been multiplied by 3.5. Most olivine water contents plotted here are calculated from pyroxene water contents using partition coefficients (see text for details). (right) Mean, median, upper, and lower quartiles of stress and water for natural samples in each CPO group. (Versions of this figure that use only measured or calculated water contents are included in the supporting information Figure S7.) CPO = crystallographic preferred orientation.

6. Comparisons Between CPOs and Deformation Parameters

6.1. Water Versus Stress

Figure 7 demonstrates that CPO types do not neatly segregate into CPO clusters in water-stress space, in contrast to the experimental data represented in Figure 1. For example, all CPOs are present in the region where only A-type is predicted (low water and low stress). Furthermore, although the highest water content samples are C-type, they still fall well below the experimental E-to-C transition. The observations disagree with the experimental relationships shown in Figures 1 and 7. Some trends do, however, emerge. Two that agree with experimental predictions are (1) natural samples with the highest water contents have C-type CPOs, and (2) A-type CPO does not occur at water contents above the experimental A-to-E-type boundary. As described in the next section, many of the C-type samples have relatively high temperatures, and as shown in Figure 8b, also commonly contain garnet rather than spinel. This indicates that, on average, C-type samples may be more common at higher pressures, where water fugacity is also higher (Kohlstedt et al., 1996). Therefore, the apparent relationship between C-type CPO and high water contents may be an indirect result from the effects of high pressure.

6.2. Stress Versus Temperature

Experiments indicate that the transition from B- to C-type CPO can be dependent on temperature and stress, with B-type occurring at lower temperatures and higher stresses (Katayama & Karato, 2006). Figure 8 reveals that C-type CPO does seem to be associated with higher temperatures and rarely occurs in peridotites with temperatures below 800 °C. The B-type samples, however, occur over a very wide range of both stress and temperature. High-pressure conditions (as inferred from the presence of garnet) appears to have no effect on the relationship between B- and C-type CPO in stress-temperature space and can therefore likely be ruled out as the sole reason for high-temperature B-type samples. When averaged over all samples, B- and AG-type samples have the lowest average temperatures (~850 °C), while E-type has the highest (~1000 °C). The average temperatures recorded in samples with A, C, and D-type CPOs are all similarly ~900–950 °C.





Figure 8. (a) Samples in stress-temperature space, colored by olivine crystallographic preferred orientation (CPO) type. (b) B- and C-type CPO samples from nature shown along samples from experiments. (c) Mean and median of stress and temperature for samples within each CPO type group. Lines represent upper and lower quartiles.

6.3. Effect of Strain Geometry

The Flinn constant, k (see section 3.3), derived from olivine aspect ratio measurements revealed no trend with the type of olivine CPO—that is, whether some olivine CPOs preferentially form in prolate (constriction) versus oblate (flattening) strain geometries. Interestingly, k also showed very little agreement with the shape parameter, T, derived from spinel grains (Figure S5). This lack of agreement, along with the observation that we *do* see a trend between CPO and spinel shape, may mean that in many samples, olivine SPO does not reliably preserve strain geometry, possibly due to dynamic recrystallization.

The trend between CPO and spinel-recorded strain geometry is illustrated in Figure 9. Chatzaras et al. (2016) observed an inverse trend between BA-index and spinel shape parameter, *T*, consistent with AG-type (0 < BA ≤ 0.35) forming under flattening (oblate ellipsoid: *T* > 0), orthorhombic CPO types (0.35 \leq BA ≤ 0.65) forming under plane strain (*T* \approx 0), and D-type (0.65 \leq BA < 1) forming under tension or constriction (prolate



Figure 9. (a) Olivine CPO type as a function of spinel shape parameter, *T*, and BA-index. *T* increases from prolate to oblate, where 0 is plane strain. Red dashed lines highlight inverse trend originally identified in Chatzaras et al. (2016). (b) The same plot, but for mean and median of samples within each CPO type group. Lines represent upper and lower quartiles.

ellipsoid: T < 0). When plotted together with the samples in this study, we find that this relationship persists in most cases, with the exception of orthorhombic B- and C-type CPOs, which consistently fall in the oblate category (Figure 9b).

7. Discussion

7.1. Uncertainties

Before we compare and contrast the natural and experimental data sets, we first evaluate the various sources of uncertainty that may influence our natural data, including three primary sources: (1) analytical, (2) calibration and standards related, and (3) epistemic uncertainties (i.e., uncertainties related to lack of knowledge) regarding whether the measurements are representative of deformation conditions.

7.1.1. Analytical Uncertainties

Analytical uncertainties within our own data set are minimal. In pyroxenes, the standard error for each measurement was only 1% of the measured water content. The uncertainty from water content measurements is primarily due to heterogeneity of grains within each sample, which resulted in a standard error of ~15% of the reported average water contents in pyroxenes. Within our own data set, the SIMS calibration curves generated from established water content standards have relatively low levels of uncertainty. The standard error of the regression line fit through the four calibration curves ranged from 18–22 ppm H₂O ($R^2 = 0.96 - 0.98$) with the exception of one mount where the standard error was 41 ppm H₂O ($R^2 = 0.89$) (Figure S3). The uncertainties around stress estimates from paleopiezometry can similarly be estimated from the variations in recrystallized grain sizes within each sample; they amounted to an average standard error of <2 MPa. Analytical standard error of temperatures from the Mojave xenoliths, for example, have a standard error of 2–12 °C (Bernard & Behr, 2017). However this may be an underestimate, as the Brey and Köhler (1990) geothermometer used can have uncertainties of up to ~100 °C for spinel peridotites due to its dependence on pressure, which is uncertain. We do not consider the analytical uncertainties to be large enough to explain the discrepancies between nature and experiments.

7.1.2. Uncertainties in Calibrations

In the case of water contents measured in olivine and/or pyroxenes with FTIR, different calibrations yield significantly different water content estimates. The calibration of Paterson (1982) used in many natural studies as well as in the CPO experiments represented in Figures 1 and 7 is often used on unpolarized samples without taking into account mineral orientation and can underpredict olivine water contents by a factor of ~3.5 compared to the preferred calibration of Bell et al. (2003), which is consistent with SIMS measurements (Koga et al., 2003). Complicating matters, Withers et al. (2012) has argued that the Bell et al. (2003) calibration may itself overestimate hydrogen concentrations by $\sim 1/3$. In addition to olivine, Koga et al. (2003) found that orthopyroxene water contents derived from the calibration of Paterson (1982) should be multiplied by a factor of 2 in order to be in agreement with both results from Bell et al. (1995) and SIMS. These calibration issues affect all of the experimental data, and many of the natural data sets which we incorporated from previously published work. They do not, however, affect our own data measured using SIMS, so correlations or decorrelations related to water in our own data set are robust. When plotting data from the literature, we have attempted to overcome this issue by multiplying any FTIR water contents calculated with the Paterson (1982) calibration by 3.5 for olivine and 2 for orthopyroxene. In addition to these calibration issues, there are also different estimates of the partitioning coefficients between olivine and pyroxenes (e.g., Demouchy et al., 2017; Hirth & Kohlstedt, 1996; Warren & Hauri, 2014). We applied the same partitioning coefficients to both our own data (see section 3.1) and data from the literature where pyroxene water contents were reported, however, so these are systematic uncertainties that affect all data points equally. 7.1.3. Epistemic Uncertainties

There remains a recurring difficulty in natural microstructural data sets of relating geochemistry to deformation stages. Peridotite mineral geochemistry, including hydrogen used to measure water content and major and trace elements used for thermometry, can be reset during both the short timescales of xenolith ascent, but also during longer timescale thermal events that can occur while the rocks are still in situ in the mantle, but not necessarily deforming. Longer timescale thermal events can induce hydrogen diffusion even in pyroxenes. Simultaneously, however, prolonged periods of heating should also affect the microstructural evolution, and we would expect preexisting deformation fabrics to show signs of annealing/grain growth if they were being heated in situ, but not deforming. Comparisons of H diffusion rates in pyroxene to olivine grain growth rates suggest that both should be significant over ky timescales (Text S7 and Figure S8). For



example, experiments of Ingrin et al. (1995) suggest that at 900 °C, water in diopside can diffuse ~1 m in 20 ky; and the wet grain growth law of Karato (1989) predicts several millimeters of grain growth for those same conditions and timescale. Samples with granular textures and relict CPOs are likely examples of this scenario, but the other textures in our data set retain grain morphologies that argue against significant thermal annealing, thus suggesting the measured water contents are representative of water content during deformation. Only half of the studies included in the compiled external data sets, however, interpreted temperature as representative of deformation temperatures specifically (Text S8).

7.2. Explanations for Differences Between Nature and Experiment

As discussed in section 6, our natural data set does not exhibit systematic relationships between most olivine CPO types and deformation conditions such as stress magnitude, water content, or temperature. Here we explore three potential explanations for this:

1. At the low stresses of the natural samples examined, olivine slip systems are not strongly sensitive to external deformation conditions.

Differential stresses in the experiments connecting olivine CPO to water, temperature, and stress magnitudes range from ~100 to 500 MPa (Bystricky et al., 2000; Demouchy et al., 2012; Jung & Karato, 2001; Jung et al., 2006; Katayama et al., 2004; Zhang et al., 2000). In contrast, however, the vast majority of natural samples examined here record stresses <100 MPa and cluster around 30 MPa (Figures 7 and 8). Several experiments have been conducted at similarly low differential stresses (~10-180 MPa) on olivine single crystals at relatively high temperatures (~1200–1600 °C) and room pressures (Bai et al., 1991; Durham & Goetze, 1977; Jin et al., 1994). These experiments did not detect any difference in the stress exponents for the [100](010), [100](001), and [001](100) olivine slip systems, suggesting a lack of stress dependence on slip system activity at these conditions. The similar lack of systematic correlation between CPO type and water or temperature in our natural data set also suggests that these components of deformation conditions only weakly influence olivine slip systems at low stresses. An exception may be the [001](100) slip system characteristic of C-type CPO, as this CPO type appears to correlate with the experimentally constrained boundary in stress-temperature space. Mackwell et al. (1985) conducted T = 1300 °C, P = 0.3 GPa deformation experiments on San Carlos olivine single crystals and found that water had no effect on the dominant slip system. However, it should be noted that water-induced fabric transitions may not occur readily at these low pressures since water fugacity, and thus water solubility, in olivine increases with pressure (Kohlstedt et al., 1996). At low stresses, alternative factors may instead influence relative strength of slip systems. For example, the relatively low stress single-crystal deformation experiments of Raterron et al. (2009) suggest that high pressure—rather than water, stress, or temperature—may promote a transition from A-type [100](010) to B-type [001](010) slip.

2. Apparent CPO type is more a reflection of kinematics and strain path than differences in slip system strength.

A weak sensitivity of olivine slip systems to deformation conditions at low stresses is compatible with (and perhaps required by) the results shown in Figure 9a in which olivine CPO exhibits a correlation with spinel shape. That is, if olivine slip systems are only weakly influenced by external deformation parameters, then olivine CPO should become much more sensitive to boundary conditions and strain path (also referred to as "strain geometry"). A sensitivity of CPO to strain geometry has been recognized in numerous experimental and modeling studies of a wide range of crustal minerals including quartz, calcite, biotite, and hornblende (e.g., Lister & Hobbs, 1980; Llana-Fúnez & Rutter, 2014; Lloyd et al., 2011). The relationship between CPO and strain geometry has also been explored for olivine, primarily through modeling and investigations of natural peridotites. For example, numerical simulations by both Wenk et al. (1991) and Tommasi et al. (1999) found AG-type CPO formed in axial compression or flattening strain, while A-type CPO formed in simple shear, and D-type formed in transtension or constrictional strain. Chatzaras et al. (2016) found the same relationship between these CPOs and strain geometry in a suite of natural samples from West Antarctica and additionally found evidence that B-type CPO forms in flattening strain, an observation also made in natural samples by Lee and Jung (2015).

The role of strain geometry has been addressed less commonly in experiments, although multiple studies have produced AG-type CPO during axial compression experiments (e.g., Hansen et al., 2011; Nicolas et al., 1973). The vast majority of olivine deformation experiments, including those associating olivine CPO types to deformation conditions (Figure 1), are conducted under dominant simple shear. The five slip systems producing A- through E-type CPOs in simple shear produce very different CPO patterns under triaxial compression and extension (Figure 4 in Skemer & Hansen, 2016).

A sensitivity to strain geometry also means that the orientation of preexisting CPOs in the mantle will play a significant role in determining both the evolution of CPO and the final CPO at steady state. Deformation experiments have historically been conducted on randomly oriented hot-pressed aggregates with weak to no preexisting CPO. However, modeling (e.g., Becker et al., 2006; Boneh et al., 2015; Skemer et al., 2012), experiments (e.g., Boneh & Skemer, 2014; Hansen et al., 2014, 2016; Skemer et al., 2011), and natural studies of exposed peridotite shear zones (e.g., Hansen & Warren, 2015; Skemer et al., 2010; Warren et al., 2008; Webber et al., 2010) have shown that preexisting CPO and changes in kinematics influence subsequent CPO development. Boneh et al. (2015), for example, showed that models with preexisting textures evolved differently with progressive strain in each of three kinematic configurations and differently from scenarios with initially random textures. This modeling is consistent with experiments by Boneh and Skemer (2014), where Åheim dunite, a starting material with moderately strong texture, was deformed and compressed in three directions (parallel, perpendicular, and oblique) relative to its initial foliation. When the starting texture is random, samples compressed perpendicular to foliation developed the expected AG-type CPO in both the models of Boneh et al. (2015) and the experiments of Boneh and Skemer (2014). Interestingly, when there was a preexisting texture-particularly in the oblique and parallel configurations-an unexpected CPO formed where [100] axes preferentially oriented perpendicular to lineation within the foliation plane. While B-type in appearance, this CPO was a transient consequence of the reorientation of a preexisting fabric, and not a product of the [001](010) slip system. In this case, the "B-type" CPO would have no relationship to the high-stress, moderate-high water contents, or low temperatures predicted by simple shear experiments, but instead was a consequence of kinematic factors.

None of the configurations in the Boneh and Skemer (2014) experiments, which were only conducted to strains of <0.7, reached a steady state, making it unclear how long this transient pseudo CPO type would persist with increased strain. Experiments by Hansen et al. (2014, 2016) that deformed Fo50 (note: a weaker olivine composition than the typical Fo90) to much higher strains ($\gamma = 20$) demonstrated that the orientation and strength of CPOs are identical regardless of any preexisting texture (formed through tension followed by torsion) when $\gamma \gtrsim 10$. Steady state did, however, appear to require higher amounts of strain in samples with preexisting textures, in agreement with the findings of the aforementioned numerical and natural studies. Models with preexisting textures can require 3–5 times the strain magnitude to approach steady state, and experimental and field data suggest that a shear strain of 1 is required to align [100] parallel to shear directions in samples without preexisting CPO and as much as 4 for samples with initial textures (Skemer & Hansen, 2016, for a review).

Since the majority of our samples are xenoliths, it is impossible to know if steady state CPO has been achieved or to quantify the strain magnitude in each sample. It is likely that many of these lithospheric peridotites have not achieved steady state, as moderately strained ($\gamma = 2-4$) portions of exposed mantle shear zones have not reached steady state as evident by the oblique orientation of [100] axes to shear (Skemer & Hansen, 2016). However, if we assume that samples with CPO types inconsistent with their active slip systems have not yet reached steady state, we can speculate that roughly half of the samples fall into this transient CPO category; as shown in Figure 5, only 25 of the 47 samples analyzed for subgrain misorientation have a majority of analyses demonstrating agreement between the expected slip system and the sample's bulk CPO.

3. At the low-stress magnitudes of the natural samples examined, olivine CPOs are not primarily controlled by slip on individual slip systems but instead by activation of other deformation mechanisms that operate to allow strain compatibility.

This explanation is also compatible with our observations presented in Figure 5, in which inferred active slip systems in some samples do not match the expected slip system for the observed bulk CPO type in the same sample. The mismatch is especially prominent for B- and AG-type samples, whereas A-, D-, and E-type display misorientation profiles consistent with their expected [100] slip systems. (The sparse number of C-type samples, particularly those with subgrains, makes it difficult to draw conclusions about its connection with the associated [001](100) slip system.)

As described previously, both B- and AG- CPO types have been associated with factors other than activation of the assumed [001](010) and [h0l](010) slip systems, respectively. AG-type has been associated with

deformation in the presence of melt (Holtzman et al., 2003; Qi et al., 2018), and—in addition to forming transiently (Boneh & Skemer, 2014; Boneh et al., 2015) or from flattening strain (Chatzaras et al., 2016)—B-type CPO has been attributed to a deformation through grain size sensitive deformation mechanisms such as diffusion creep (Drury et al., 2011; Miyazaki et al., 2013; Sundberg & Cooper, 2008) and dislocation-accommodated grain boundary sliding (DisGBS) (Précigout & Hirth, 2014).

Of course some of the noise present in Figure 5 may be attributed to combinations of slip systems working in concert to produce CPOs, in agreement with modeling that incorporates combinations of critical resolved shear stresses for the [100](010), [100](001), [001](010), and [001](100) slip systems (e.g., Becker et al., 2008; Kaminski, 2002). In particular, it may be unsurprising that so many samples preserve subgrains with both [100](010) and [100](001) slip systems, as these have nearly identical critical resolved shear stresses at intermediate temperatures ~1000 °C (Goetze, 1978). Alternatively, because misorientation profiles are necessarily collected from porphyroclasts, perhaps they reflect the orientations of harder crystal slip systems whereas the recrystallized grains (lacking subgrains) are those that experienced slip along the system representative of the bulk CPO.

The three explanations discussed above are not mutually exclusive. Moreover, our data and the explanations provided do not imply that deformation conditions (such as stress, temperature, and water) have *no* effect on olivine slip systems, but rather that boundary conditions and preexisting fabrics appear to be more influential than deformation conditions on olivine CPO in the ambient lithospheric mantle. These findings suggest that caution should be taken in using olivine CPO types to infer deformation conditions without independent deformation condition constraints.

7.3. Implications for Seismic Anisotropy

As suggested by the smaller proportion of A-type samples in our data set versus the compilation of Ismaïl and Mainprice (1998) in Figure 2, there may be a bias in favor of non-A-type CPOs in the literature due to the community's interest in these less common types in recent decades. Putting this issue of representation aside, we use this large data set to investigate the implications for seismic anisotropy, particularly in the continental lithospheric mantle, as this is where the vast majority of samples are sourced (Figure 2). First, since almost two thirds of samples fell in the A-, D-, and E-type categories, a majority of samples display the expected preferential alignment of [100] axes in the lineation or shear directions. However, one of the most striking aspects of this data set is the number of different olivine CPOs preserved at individual localities (Figure 3). Nineteen of the 52 localities had samples with three or more CPOs, 18 had B-type CPOs, and 14 had CPO types with opposing/orthogonal fast axes: that is, B-type CPO in addition to A, C, D, and/or E-type CPOs. Lastly, while over a quarter of localities (15) had samples with C-type CPOs, less than half of those had more than two individual samples with this CPO type. These observations have four primary implications:

- 1. The relatively complex seismic anisotropy patterns observed in the continental (versus oceanic) mantle (e.g., Long & Becker, 2010) can be explained in part by the wide range of CPO types found among the peridotites sampled from xenoliths and continental massifs. Some of this complexity may be due to frozen-in anisotropy within the lithosphere, but our results suggest it could also be attributed to the lack of stress dependence of slip systems at low stresses typical of the upper mantle, small scale variations in strain geometry, the influence of preexisting textures on CPO development, or all of the above. While trends of CPO type with estimated depth goes beyond the scope of this study, it is conceivable that the wide variety of CPO types at some individual localities may vary with depth, which would be in agreement with recent studies that observe complex anisotropic layering within the lithosphere (e.g., Ford et al., 2016) and connect midlithospheric discontinuities (MLDs) to sharp changes in seismic anisotropy (e.g., Auer et al., 2015; Wirth & Long, 2014; Yuan & Romanowicz, 2010).
- 2. The complexities introduced by B-type CPOs affect more tectonic settings than just the cold corner of the mantle wedge. This may be an explanation (though one of several, see Long, 2013 for a review) for the confounding occurrence of trench parallel anisotropy unexpectedly far away from the trench (e.g., Abt et al., 2009, 2010; Hoernle et al., 2008; Long et al., 2016). In our data set, B-type CPO is shown to develop at the full range of mantle stress, water, and temperature conditions (up to 1100 °C). For this reason, flow-perpendicular fast directions may be more common than previously assumed.
- 3. The common cooccurrence of CPO types with orthogonal fast directions that would cancel each other out might mean that at many places, we could expect to see no net azimuthal anisotropy. Additionally, while AG-type—a common CPO type in this data set—results in a strong alignment of olivine's slow axis

([010]) aligned normal to flow, we would expect no azimuthal anisotropy as both its fast and intermediate axes ([100] and [001]) are girdled (and therefore unoriented) within the foliation or flow plane. Together, this would suggest that a lack of azimuthal anisotropy in a particular region should not be interpreted as a lack of deformation through dislocation creep, particularly if radial anisotropy is observed.

4. In the upper mantle, on average, horizontally polarized seismic shear waves (SH) travel faster than vertically polarized ones (SV). C-type CPO is the only variety that is predicted to substantially affect this radial anisotropy, since its alignment of fast axes orthogonal to the flow plane would result in $V_{SV} > V_{SH}$ rather than $V_{SH} > V_{SV}$ (in the case of horizontal shear), which is characteristic of all other CPO types. However, C-type CPO is not only the least abundant CPO type in the data set, but when observed, it was typically only present in one or two samples at a given locality. This suggests that despite the complexities in azimuthal anisotropy may be largely unaffected. An exception might be the large subset of samples with E-type CPO, as this type results in a ~30% reduction of radial anisotropy compared to A-type CPO (Becker et al., 2008).

8. Conclusions

We present a compilation of new and published naturally deformed peridotites with the goal of connecting six established olivine CPO types to the wide range of deformation conditions present in the Earth's mantle. Contrary to previous inferences from experiments, we do not see evidence that olivine CPO is primarily determined by water content and differential stress magnitude, possibly because individual olivine slip systems are less sensitive to stress and water content at the low-stress magnitudes that characterize the upper mantle. Temperature appears to play a role, with AG- and B-type CPOs occurring at lower temperatures on average, and C- and E-type CPOs dominantly occurring at higher temperatures. Additionally, quantification of strain geometry reveals that AG-, B-, and C-type CPOs typically form when deformation fabrics are oblate, D-type when prolate, and A- and E-type during plane strain. Our results highlight the need for further experiments investigating the relationship between CPO, stress, and water, but at conditions closer to those typical of the upper mantle (i.e., lower stress and temperatures) and with improved constraints on water contents using polarized, mineral-dependent FTIR calibrations or SIMS. Finally, our results show-case the complexities of olivine CPO development. The observation that individual localities can preserve as many as five CPO types exemplifies this and may explain some of the complexities observed from seismic anisotropy within the continental mantle lithosphere.

Appendix A: List of Studies Included in the Literature Compilation

Samples from the following studies are represented in the literature review: Bascou et al. (2008), Baptiste et al. (2015), Behr and Smith (2016), Cao et al. (2017), Chatzaras et al. (2016), Chin et al. (2016), Drury et al. (2011), Falus et al. (2008), Frese et al. (2003), Hidas et al. (2007), Jung et al. (2009a, 2009b, 2013, 2014), Kaczmarek and Reddy (2013), Kamei et al. (2010), Katayama et al. (2005, 2011), Kim and Jung (2015), Lee and Jung (2015), Mehl et al. (2003), Michibayashi et al. (2007, 2012), Michibayashi and Oohara (2013), Mizukami et al. (2004), Mizukami and Wallis (2005), Morales and Tommasi (2011), Nagaya et al. (2014), Palasse et al. (2012), Park and Jung (2015), Park et al. (2014), Pera et al. (2003), Précigout and Hirth (2014), Satsukawa et al. (2010), Satsukawa and Michibayashi (2014), Skemer et al. (2006), Tommasi et al. (2004), Vauchez et al. (2005), Wang et al. (2013a, 2013b), Warren et al. (2008), Webber et al. (2010), Xu et al. (2006), and Yang et al. (2010).

A description of how information was incorporated from these studies, along with an inventory of the types of samples and analyses done in each of these studies can be found in the supporting information (Text S5 and Table S5).

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