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Hypersolidus deformation in the lower crust of the Josephine ophiolite: evidence for kinematic decoupling between the upper and lower oceanic crust

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Abstract

New mapping and structural observations in the lower crust of the Josephine ophiolite provide insights into the geometry and kinematics of hypersolidus flow beneath an oceanic spreading center. The lower crust, defined here as the sequence of rocks overlying mantle peridotite and beneath exposure of 100% sheeted dikes, can be divided into lower wehrlite–dunite and upper gabbroic sections. The contact between the two is mutually intrusive where exposed. Hypersolidus fabrics are the dominant structures observed. No pervasive crystal-plastic deformation is observed. Restoration of the hypersolidus foliations and igneous modal layers to their on-axis orientation indicates that they strike approximately perpendicular to the strike of the inferred paleo-spreading center as defined by the orientation of sheeted dikes and on-axis, oceanic faults. Hypersolidus lineations define a dispersed 3-D flow pattern in the lower crust, whereas extension directions in the upper crust (i.e. poles to sheeted dikes, oceanic normal faults) are unidirectional and perpendicular to the paleo-ridge axis. Collectively, these observations are consistent with partitioning of deformation between the upper and lower crust, and local ridge-parallel extension in the partially-molten lower crust due to possible subsidence of the thickened, axial upper crust. However, some component of kinematic coupling between the lower crust and mantle peridotite driven by asthenospheric flow cannot be ruled out.

Keywords: Ophiolite; Suprasubduction zone; Lower crust; Hypersolidus foliations; Magma chamber processes; Klamath Mountains

1. Introduction

One of the principal questions regarding the understanding of oceanic spreading centers is the nature of lower crustal deformation and its affects on the formation and evolution of magma chambers beneath the ridge axis. Early work in the lower crustal sections of the Oman, Troodos, and Bay of Islands ophiolites suggested that oceanic magma chambers were steady-state features in which magma input matched extension, and axial magma chambers were large bodies in which crystallizing phases settled onto an inward dipping floor (i.e. towards the spreading ridge; Sleep, 1975; Dewey and Kidd, 1977; Pallister and Hopson, 1981; Casey et al., 1983 and others). However, recent seismic studies over active intermediate- to fast-spreading ridges have demonstrated that the volume of material beneath the ridge axis that has seismic velocities appropriate for a melt body is much smaller (< 100 m thick) than the magma chamber interpreted from ophiolite studies (Detrick et al., 1987; Kent et al., 1994; Dunn and Toomey, 1997; Dunn et al., 2000). The remaining lower crust is defined by anomalously low velocities that are interpreted to result from the presence of 0–20% melt (Sinton and Detrick, 1992; Dunn et al., 2000), and perhaps up to 50% depending on the melt topology (Mainprice, 1997). Seismic studies beneath slow spreading ridges (< 50 mm yr^{-1}) display low velocity zones consistent with a magma chamber only beneath a few seamounts located within the axial valley (e.g. Barclay et al., 1997).

One way to explain the accretion of the entire lower crust from a few small melt bodies is by crystallization and translation of the magma (melt + crystals) down and away from the melt lenses as spreading continues (e.g. 'gabbro glacier' from Quick and Denlinger, 1993; Nicolas et al., 1993; Boudier et al., 1996; Buck, 2000; Fig. 1a). Kinematic

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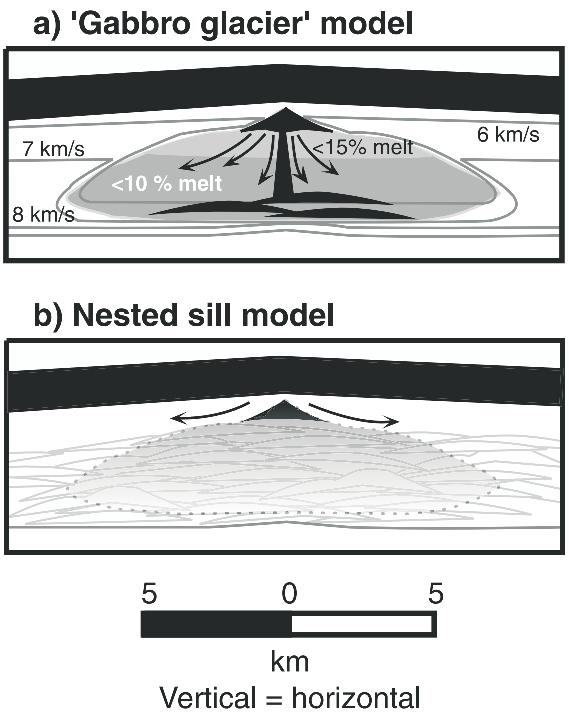


Fig. 1. Endmember models for magma chamber formation and evolution beneath oceanic spreading centers where magma supply is high enough to produce a 'complete' ophiolitic pseudo-stratigraphy. Both models portray a view looking parallel to the ridge axis (perpendicular to the spreading direction). (a) 'Gabbro glacier' model (e.g. Nicolas et al., 1993; Quick and Denlinger, 1993) in which melts crystallize in the shallow, seismically-observed melt lens beneath the sheeted dikes (shown in black). Magma is then translated down and outward from the melt lens via hypersolidus flow to generate the layered, lower crust. Also shown are seismic velocities that define the low velocity zone beneath the ridge axis (after Nicolas et al., 1993). (b) Nested sill model with seismic low velocity zone outlined with dots and gradient shade pattern (e.g. Boudier et al., 1996; Kelemen and Aharanov, 1998). White layer beneath the sheeted dikes (black) represents the upper-most gabbros. Note that in both cases the flow direction in the lower crust as defined by hypersolidus structural elements is parallel to the spreading direction.

and microstructural arguments in this model require that foliations and lineations reflect large strains in the *hypersolidus state* associated with transfer of magma from the melt lens to the lower crust (Nicolas et al., 1993; Nicolas and Ildefonse, 1996). If the seismic constraints on melt fraction are correct, then hypersolidus flow must occur at low melt fractions (<20%) near the base of the crust (Nicolas et al., 1993), yet still preserve microstructures formed in a hypersolidus state (e.g. Pallister and Hopson, 1981; Yoshinobu and Hirth, 2002). An alternative model suggests that the majority of the lower crust is formed by emplacement of numerous small sills (e.g. Kelemen and Aharanov, 1998; Dick et al., 2000; Fig. 1b). Formation of hypersolidus structures in this model may be due to plate separation (extension) (e.g. Quick and Denlinger, 1993), processes associated with magma emplacement, and/or coupling between the lower crust and flowing upper mantle (Nicolas and Boudier, 1995). In order to evaluate these hypotheses, it is necessary to establish the geometry and conditions of formation of structures within the lower crust because each model makes testable predictions regarding the deformation mechanisms, geometry, and kinematics of flow in the lower crust.

Distinguishing whether structures in the partially molten lower crust represent processes of magma emplacement, lithospheric plate separation, drag from the underlying upwelling asthenosphere, or some other process requires geometric and kinematic linkages to be made between upper crustal, lower crustal, and upper mantle structures. This paper presents new field and structural data from the crustal section of the Josephine ophiolite that adds insight into the geometric and kinematic evolution of the lower oceanic crust. The Josephine ophiolite is an excellent field area to evaluate geometric relations between the upper and lower oceanic crust because, unlike other ophiolites such as Oman and Troodos, the entire crustal sequence can be restored to its on-axis orientation (e.g. Harper, 1982). Restoration of hypersolidus structures and microstructural observations indicate that flow in the axial lower crust was locally 3-D and included a component of along-axis flow, in contrast to the ridge-perpendicular unidirectional extension in the upper crust.

2. Geologic setting of the Josephine ophiolite

The 162 Ma Josephine ophiolite, located in the Klamath Mountains of northern California and southern Oregon, represents a complete slice of oceanic lithosphere (Fig. 2; Harper, 1984; Harper et al., 1994). Emplacement of the ophiolite occurred by more than 100 km of underthrusting along a low angle fault beneath the western edge of North America between ~155 and 150 Ma (Harper et al., 1994). Emplacement-related deformation and metamorphism is minor in the northern half of the ophiolite (prehnite–pumpellyite facies), whereas in the southern half sheeted dikes and pillow lavas are weakly to strongly foliated (Harper et al., 1988).

In ascending order the ophiolite consists of >1000 km^2 of (1) tectonized harzburgite and other ultramafic rocks, (2) an up to ~1.5-km-thick section of mixed 'cumulate' ultramafic and minor mafic rocks in gradational/intrusive contact with (3) weakly layered 'cumulate' (lower) gabbroic rocks, (4) ~200–600 m of variable-textured (upper) gabbro, (5) up to >1 km of sheeted dikes, and (6) 150–>400 m of pillow lavas (Fig. 2b; Harper, 1984; Yoshinobu, 1999). The lower crust, defined here as the sequence of rocks overlying tectonized peridotite and beneath exposure of 100% sheeted dikes, can be divided into lower wehrlite–dunite and upper gabbroic sections (Fig. 3; see below).

Formation by spreading in a suprasubduction zone environment is indicated by the temporal relationship between magmatism in the Josephine spreading center and a coeval arc to the northwest (Rogue Formation and Chetco

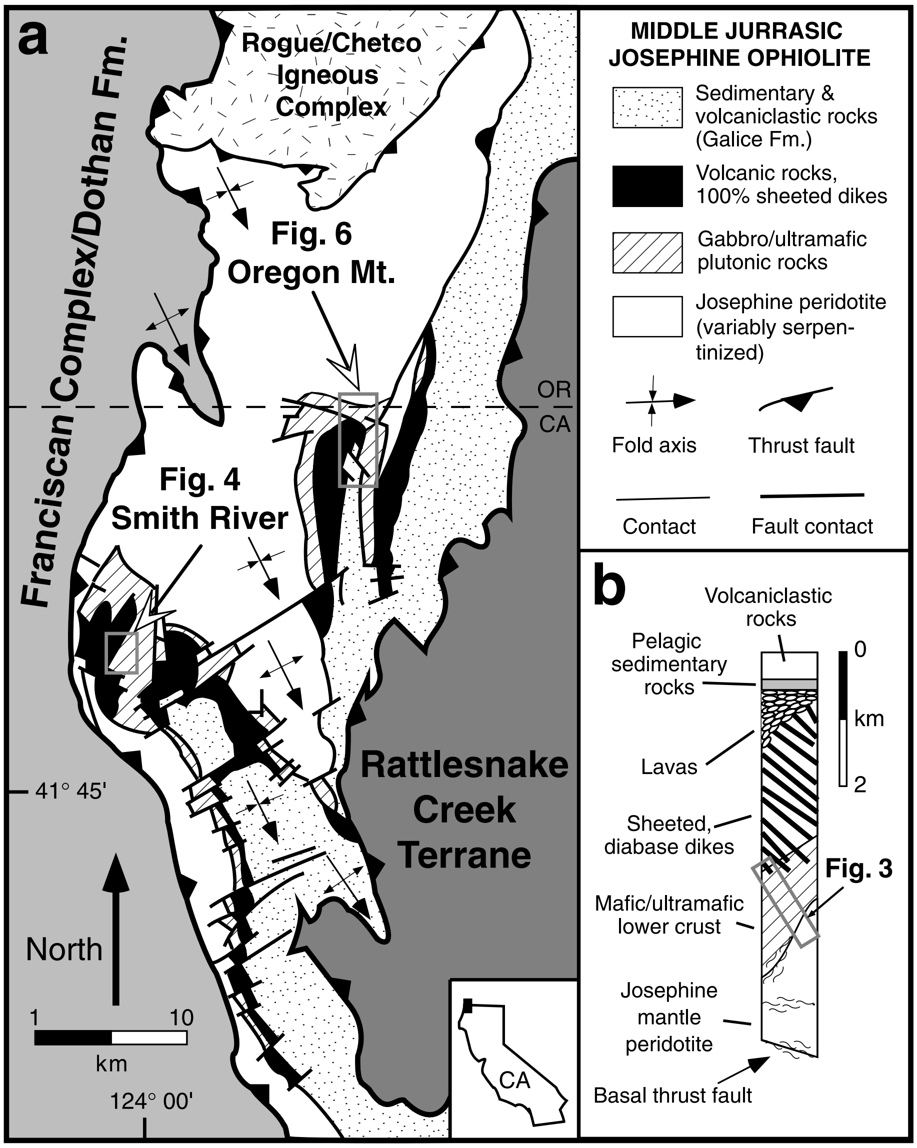


Fig. 2. (a) Simplified geologic map of the Josephine ophiolite (modified from Alexander and Harper, 1992); (b) generalized crustal column for the Josephine ophiolite; note that pillow lavas and flows have steeper dips down section, implying eruption during growth faulting.

Igneous Complex, Fig. 2a), the trace element composition of eruptive lavas and sheeted dikes, the crystallization sequence of lavas, dikes, and plutonic rocks (i.e. olivine–clinopyroxene–plagioclase), and the presence of tuffaceous cherts and volcaniclastic rocks overlying and locally interbedded with the pillow lavas (Harper, 1984). New trace element geochemistry is consistent with a more complex setting in which the Josephine spreading center propagated from a back arc setting into the Jurassic arc (Harper, 2003).

The ophiolite is inferred to have formed at slow- to intermediate-spreading rates (Alexander and Harper, 1992) based on: (1) geochemical and field evidence for episodic magmatism (Harper, 1988), (2) the presence of oceanic normal faults including an oceanic, antigorite mylonite detachment fault (Norrell and Harper, 1988, 1990; Alexander and Harper, 1992; Alexander et al., 1993), (3) fault-controlled hydrothermal circulation (Alexander et al., 1993), (4) on-axis serpentinization and related faulting (Coulton et al., 1995), (5) a continuous sheeted dike complex, and (6) on-axis tilting of the entire crustal sequence by ~50° (Harper, 1982) overlapping with extrusion as indicated by increasing dips of lavas with depth (Alexander and Harper, 1992). Recent observations from slow spreading centers indicate that a significant portion of the crust consists of serpentinized ultramafic rocks and gabbros, sometimes unconformably overlain by pillow lavas (see references in Karson, 1998). Intermediate- to fast-spreading ridges appear to be characterized by more continuous and thicker sections of volcanic rocks, sheeted dikes, and plutonic sequences (e.g. Oman, Nicolas et al., 1996; East Pacific Rise, Kent et al., 1994). The Josephine ophiolite has a thick (>1 km) sheeted dike complex developed throughout the entire ophiolite, a continuous

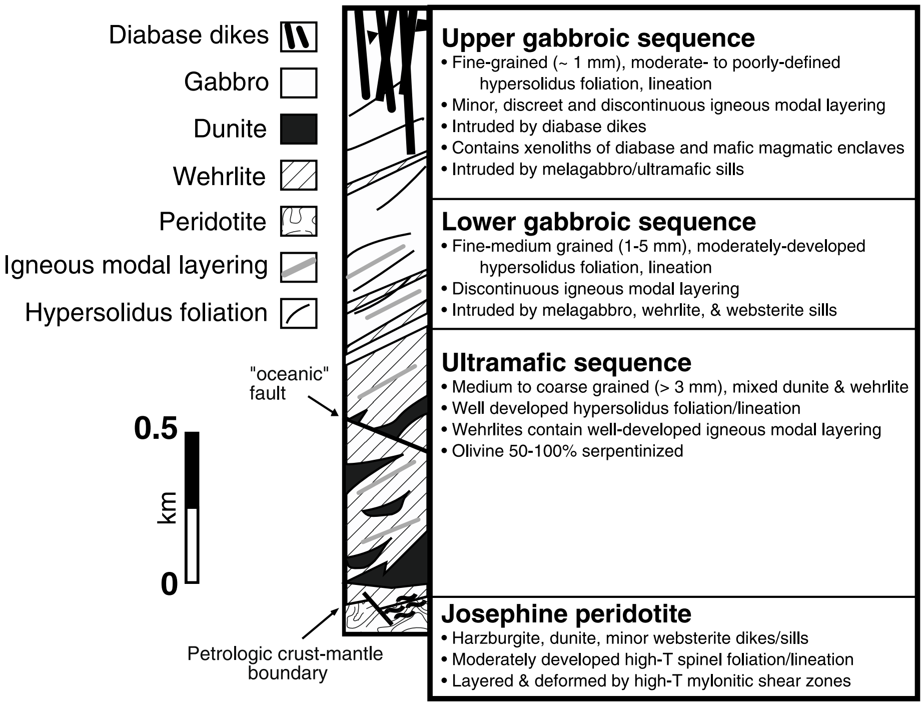


Fig. 3. Schematic lithologic column of the lower crust of the Josephine ophiolite displaying key lithologic and structural elements discussed in the text.

plutonic section, and no large-scale, low angle detachment faults. Therefore, the Josephine ophiolite likely represents either (a) the mid-segment of a slow-spreading ridge where crustal thickness and magma supply is highest (e.g. Detrick et al., 1995; Barclay et al., 1997), or (b) an intermediate spreading-rate ridge (e.g. Alexander and Harper, 1992).

3. Lower crustal section

3.1. Gabbroic sequence

The gabbroic portion of the lower crust in the Josephine ophiolite (up to 800 m thick; Fig. 3) is characterized by an assemblage of moderate sized intrusive bodies (> 100 m thick). Subhorizontal contacts are observed (e.g. sills), whereas subvertical or steeply dipping contacts are absent. Thus we infer length to thickness ratios in excess of 10:1. These larger gabbroic bodies are intruded by smaller sills and dikes ranging in thickness from 0.01 to 2 m. Compositionally, the larger intrusive bodies are gabbro with plagioclase:clinopyroxene ratios of about 65:35, whereas sills less than 2 m thick are of gabbroic and ultramafic composition. All of the rocks have undergone some degree of hydrothermal metamorphism with plagioclase and clinopyroxene variably altered to clinozoisite and actinolite, respectively. Grain sizes vary from 0.3 to 2.0 mm for both plagioclase and clinopyroxene. Felsic, quartz-bearing rocks (‘plagiogranites’) occur as dikes and small cross-cutting intrusions in the upper gabbros. Cross-cutting all of the above intrusions are fine-grained diabase dikes and sills with chilled margins; these diabase dikes are parallel to dikes in the overlying sheeted dike complex. Xenoliths of diabase locally occur in the top of the upper gabbros.

Sills form a conspicuous part of the upper gabbro along the Middle Fork of the Smith River (Fig. 4), and are divided into gabbroic and ultramafic suites. Excellent exposures of the sills occurs a few hundred meters below the structurally lowest exposure of 100% sheeted dikes and occupy between 10 and 70% of the outcrop (Figs. 4 and 5a). Sills in the upper gabbro are cut by diabase and gabbroic dikes. The sills can be distinguished from petrologic layers based on one or more of the following observations: (1) presence of chilled margins; (2) cross-cutting relationships; and (3) offset of markers on either side of the sill. Sill thicknesses range from centimeters to > 1.5 m and often taper along strike (Fig. 5b). The ultramafic sills in the upper gabbro are predominantly wehrlite, clinopyroxenite, and olivine websterite whereas the mafic sills are gabbroic in composition. A 135-cm-thick olivine websterite sill can be traced for at least 60 m in the upper gabbros. This sill contains well-developed modal layers and hypersolidus foliations that are discontinuously developed and surround thick (20–30 cm) layers of non-foliated, coarse olivine websterite (Fig. 5c). Within these non-foliated zones large (~ 8 mm) poikiloblasts of orthopyroxene occur and contain inclusions of clinopyroxene, olivine, and chrome-spinel (Fig. 5c).

3.2. Wehrlite–dunite sequence

Beneath the gabbroic section is a sequence of mixed wehrlite and dunite that ranges in thickness from 200 to

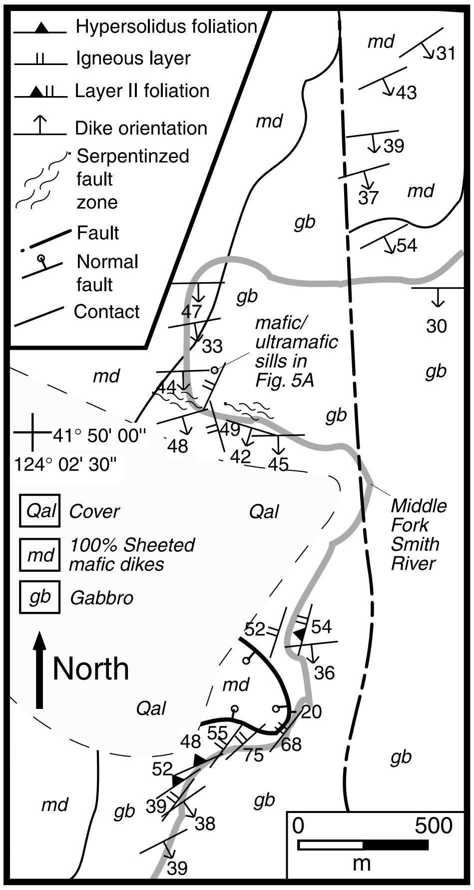


Fig. 4. Simplified geologic map of a portion of upper gabbros and lower sheeted dikes exposed along the Middle Fork of the Smith River. The contact between 100% sheeted dikes and underlying gabbros is gradational over 10 to a few tens of meters.

800 m (ultramafic ‘cumulates’ of Harper, 1984; Figs. 3 and 6). Where exposed, the contact between the overlying gabbros and the ultramafic rocks in some localities display gabbroic sills intruding wehrlite. Wehrlitic xenoliths are sometimes observed within the gabbro sills. Elsewhere, clinopyroxene-rich dikes and sills, apparently derived from the underlying wehrlites appear to have intruded the base of the gabbro. These mutually intrusive cross-cutting relationships imply that both bodies were partially molten at the same time. Hypersolidus foliations are subparallel in both units, as discussed below.

With the exception of the area near Oregon Mountain, the base of the ultramafic unit is apparently defined by a thick, serpentinized shear zone that marks the tectonic contact with the underlying mantle sequence of the Josephine peridotite (Harper, 1984). At Oregon Mountain (Fig. 6), the contact is partially obscured by vegetation, but over an 8-m-wide interval a transition zone exists that juxtaposes coarse, clinopyroxene-rich wehrlite having a well developed hypersolidus foliation against harzburgite. Although olivine is partially to completely serpentinized, no thick serpentinized shear zone is observed.

In the vicinity of Oregon Mountain the ultramafic sequence approaches 800 m in thickness, is well-layered and is composed of numerous, interleaved bodies of wehrlite and dunite that range in thickness from ~1 to 400 m (Fig. 6). The intricate nature of the contacts between wehrlite and dunite are difficult to demonstrate at the regional map scale, although large bodies of dunite (>100 m thick) do occur (e.g. Fig. 6). Where exposed, the contact displays mutually interfingering relations between wehrlite and dunite and stranded rafts of wehrlite included in dunite. Based on cross-cutting relations and textural evidence, Yoshinobu (1999) interpreted the dunites to be replacive after clinopyroxene (e.g. Kelemen and Dick, 1995).

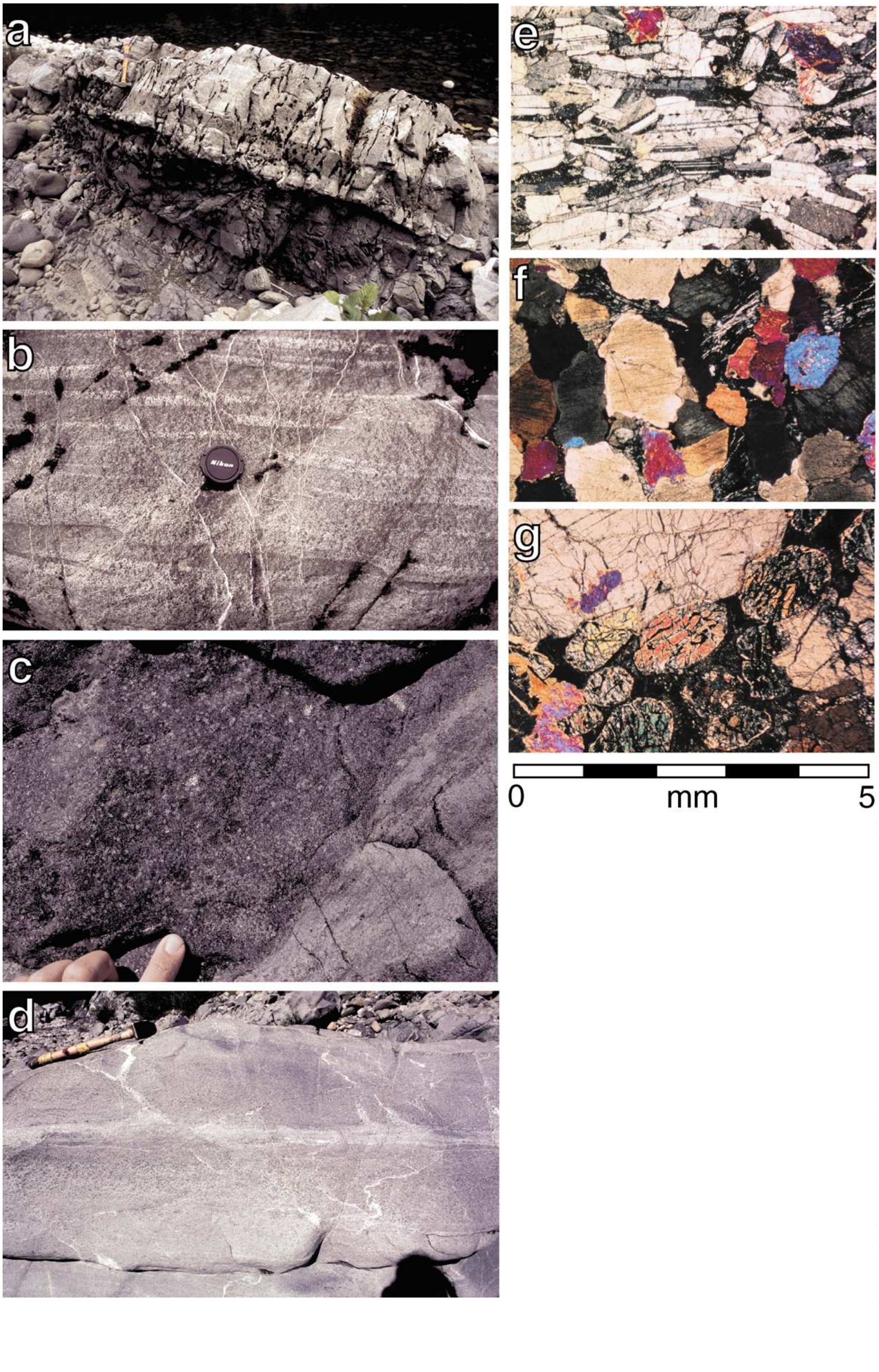
3.3. Hypersolidus structures

We use the modifier hypersolidus to describe any fabric element within the lower crust that formed while the rock was above its solidus, including modal layers and mineral alignments (foliations/lineations). Additional modifiers such as suspension-flow, crystal-plastic, etc., can then be added to denote the inter/intra-crystalline deformation mechanisms that accommodated fabric formation.

Discontinuous layering is observed in the gabbroic portion of the ophiolite but is not ubiquitous. Layering occurs as centimeter- to decimeter-scale intervals defined by changes in modal abundance and/or crystal size of pyroxene and/or plagioclase. Individual layers may contain hypidiomorphic granular textures or may be foliated. Hypersolidus foliations form parallel and locally obliquely to the igneous layering (Fig. 7; see next section). There is no systematic increase in the occurrence of layers in the gabbros with depth throughout the ophiolite, in contrast to other ophiolites (e.g. Pallister and Hopson, 1981). In fact, some of the most intricate and delicate centimeter-scale modal layers occur in the middle and upper gabbros.

Within the wehrlite–dunite sequence, igneous layering is commonly defined by changes in the modal abundance of olivine or clinopyroxene forming either dunite or clinopyroxenite layers. Where the effects of serpentinization are weakly developed, wehrlite contains subhedral olivine 0.5–1.0 mm in size, subhedral to anhedral clinopyroxene crystals 0.5–>10 mm, and trace amounts of euhedral chrome-spinel. Dunite is serpentinized; thus, only rarely observed are intact olivine crystals with chrome-spinel inclusions. Within both the gabbroic and wehrlite–dunite sequences layering is generally at an angle of ~20° to foliation (Fig. 7; see next section), although local, layer-parallel hypersolidus foliations are observed. Within the ultramafic sequence, some modal layers up to 10 or more centimeters thick are unfoliated yet sandwiched between foliated layers (Fig. 5c).

The preferred orientation of crystals with virtually no evidence of intracrystalline plastic deformation is the dominant microstructure observed in the mafic and ultramafic portions of the lower crust (Fig. 5e–g). Within the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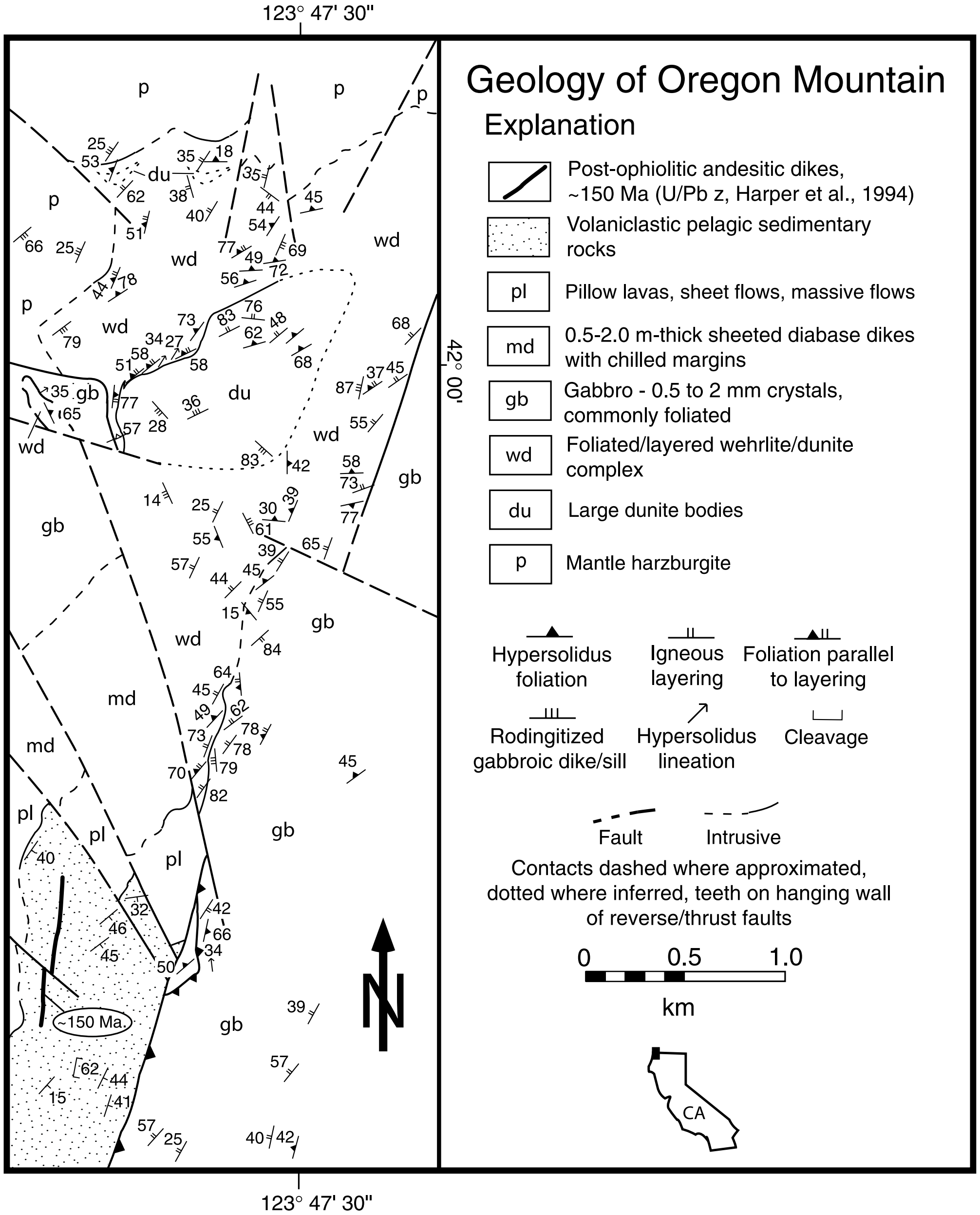

Fig. 6. Geologic map of the lower crustal section exposed at Oregon Mountain displaying structural and lithologic elements within the wehrlite–dunite and gabbro sequences. Only the largest dunite bodies are shown on the map; generally these bodies consist of mixed dunite and wehrlite at the 1–10 m scale.

Fig. 5. Photographs of various lithological and structural features in the lower crust of the Josephine ophiolite. (a) Sills in the gabbroic section along the Middle Fork of the Smith River (see location in Fig. 4; hammer in upper left for scale). (b) Fingering melagabbro sills pinched out along their length. Both the sills and igneous modal layers are displaced along discreet hypersolidus shear bands with normal-sense offset and are cut by late, hydrothermal veins. (c) Websteritic layer in ultramafic sill displaying coarse (~1 cm) orthopyroxene oikocrysts surrounded by a matrix of non-foliated fine- to medium-grained orthopyroxene, clinopyroxene, and olivine. The dark websteritic layer is sandwiched by a wehrlitic layer in which a well-developed hypersolidus foliation is preserved. (d) Igneous modal layer defined by accumulation of plagioclase with dikes that are interpreted to be 'escape' structures that accommodate upward migration of late crystallizing melt during compaction of the crystal mush. (e)–(g) Photomicrographs of hypersolidus microstructures from the lower crust of the Josephine ophiolite; all thin sections are cut parallel to lineation and perpendicular to foliation. (e) Aligned plagioclase crystals defining hypersolidus foliation. Note the straight grain boundaries and lack of any crystal-plastic deformation. (f) Aligned clinopyroxene crystals in a serpentinized olivine clinopyroxenite. Individual clinopyroxenes display no evidence for deformation but have serrated and lobate grain boundaries; moderately developed foliation is vertical. (g) Weakly serpentinized wehrlite. Note the lack of crystal-plastic deformation in olivine crystals.

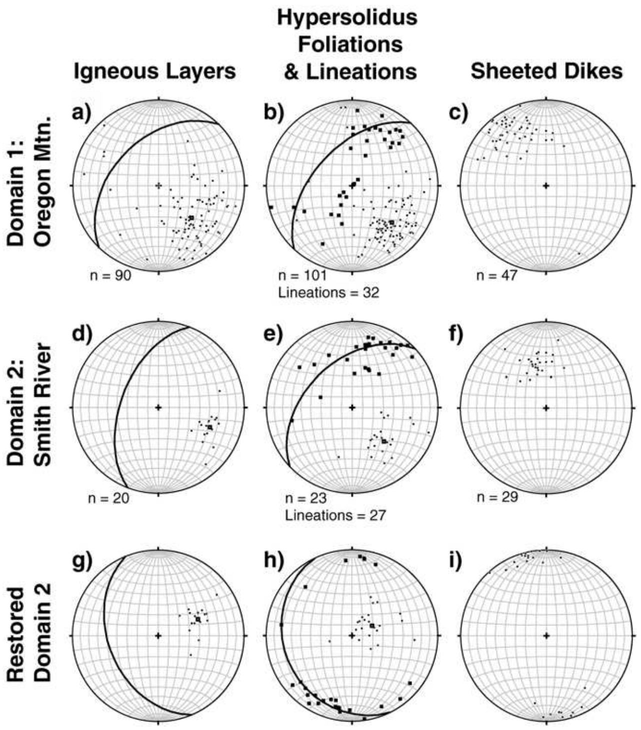

Fig. 7. (a)–(i) Equal area stereonets for various structural elements separated by domains in the Josephine ophiolite. Refer to Fig. 2a for location of domains. Dots = poles to planes; solid squares = lineations; plotted great circles = the average plane. See text for details of restoration.

gabbro, plagioclase laths define the foliation plane and have straight grain boundaries and aligned albite twins; triple-point junctures between feldspars are sometimes observed. Clinopyroxene also has a preferred mineral alignment with axial ratios greater than one and (100) twins that lay sub-parallel to the foliation plane. Both elongate clinopyroxene and plagioclase laths lie within the foliation and define a lineation. Sporadically developed asymmetric shear bands show a normal-sense deflection in the direction of the lineation (Fig. 5b). Minor 'tiling' of plagioclase feldspar laths is consistent with the lineation direction and shear band deflections. However, these structures are relatively uncommon.

The wehrlites contain a well-developed foliation defined by elongate, prismatic clinopyroxene crystals (1– > 20 mm in length) with subparallel cleavage planes and aggregates of partially- to wholly-serpentinized olivine. An irregularly developed mineral lineation is apparent in the foliation plane defined by aligned clinopyroxene laths. Clinopyroxene crystal boundaries are in some places cuspate suggestive of minor amounts of high-temperature crystal boundary migration (Fig. 5f). In general, however, these crystals show a distinct lack of any crystal-plastic deformation. Olivine in both the wehrlite and dunite is >80% serpentinized but maintain their original igneous crystal shapes. Where relict cores are preserved, olivine appears to be free of crystal-plastic deformation (e.g. Fig. 5g), although minor subgrains are observed in a few samples. Because of the lack of any appreciable crystal-plastic deformation in both the gabbroic and wehrlite–dunite sequences, we infer that the foliations, lineations, and minor shear bands formed in the presence of a melt phase during hypersolidus flow in the lower crust.

3.4. On-axis geometry of structures in gabbroic sequence

Structural data from the Smith River Domain (Fig. 2a) are in a favorable setting for reconstruction of the spreading-ridge geometry and for establishing geometrical relationships because conformable, overlying sedimentary rocks up-section provide a necessary paleo-horizontal datum and the entire ophiolite sequence is repeated on different limbs of folds and repeated by reverse faults (Harper, 1982). The gabbro section in the area of this study is situated in the hinge of a broad, gently southeast plunging (<10°) fold that was formed during post-thrust emplacement deformation of the ophiolite (Harper, 1982). We restore data from this locality (Fig. 7d–f) to their inferred on-axis orientation by first rotating the data 10° clockwise along an axis of 258° to restore the fold hinge and bedding in overlying sediments within the fold hinge to horizontal. This rotation results in sheeted dikes striking approximately east–west and dipping ~55° south, and igneous layering and hypersolidus foliations dipping moderately southeast, consistent with Harper (1982). A second rotation, inferred to have occurred at the ridge axis during normal faulting (Harper, 1982), is needed to restore the sheeted dikes and underlying cumulates to their orientation prior to this on-axis tilting. For the area of this study, sheeted dikes were rotated to vertical by rotating them 56° clockwise about an azimuth of 76° (Fig. 7i). This second rotation axis is the average strike of the sheeted dikes, which assumes the sheeted dikes were originally vertical and were rotated about their strike (= inferred ridge axis). This assumption made by Harper (1982) is supported by subsequent work giving independent estimates of the orientation of the ridge axis, including oceanic faults (strike and displacement vectors; e.g. Alexander et al., 1993) and the axis of tilting of lava flows whose dip increases with depth (Alexander and Harper, 1992).

After this restoration, hypersolidus foliations in the gabbro section of the study area have strikes that are nearly *perpendicular* to the strike of the overlying sheeted dikes, with an average dip of 22° to the west-southwest (Fig. 7h), similar to those for the ophiolite as a whole (Harper, 1982). The hypersolidus lineations scatter about a girdle defining the average foliation plane. Igneous layers also strike perpendicular to the dikes but dip on average more steeply (41° west-southwest average; Fig. 7g).

Restoration of hypersolidus fabrics in the wehrlite–dunite sequence is equivocal because a continuous transition from the crust–mantle boundary to the overlying sedimentary rocks is not preserved. However, foliations, lineation, and layers in the Oregon Mountain domain

(Fig. 7a–c) are subparallel with fabrics throughout the ophiolite (this study; Harper, 1982) suggesting that they may have undergone the same rotations as rocks in the Smith River domain.

4. Discussion

The recognition of oceanic fault zones, on-axis serpentinization, and growth faulting in the upper crust of the Josephine ophiolite (Alexander and Harper, 1992; Alexander et al., 1993; Coulton et al., 1995) and the presence of a well-developed sheeted dike complex attests to the continuous extensional deformation that occurred at the ridge-axis. In the lower crust, the lack of any pervasive crystal-plastic deformation implies that deformation must have been accommodated entirely by hypersolidus processes. In the following discussion we focus on (a) the origin of the structural geometry within the lower crust and the implications to interpreting hypersolidus structures in ophiolites, and (b) the nature of hypersolidus deformation mechanisms within the lower oceanic crust.

4.1. Structural geometry in the lower crust

A 3-D, schematic view of the geometrical relationships observed in the Josephine lower crustal section at the time of their formation at the ridge axis, but prior to ~50° on-axis tilting, is shown in Fig. 8a. Lineations define a dispersed 3-D pattern within foliation planes that dip in a direction parallel to the ridge axis. The dispersed pattern of lineations in the lower crust is in contrast to the upper crust in which poles to sheeted dikes form a well-defined, unidirectional extension direction (compare Fig. 7h and i). Igneous layering and hypersolidus foliations do not show a systematic increase in dip up section, as predicted by thermal models (e.g. Phipps-Morgan et al., 1994) and observed in the Oman and Bay of Islands ophiolites (e.g. Pallister and Hopson, 1981; Casey et al., 1983).

Any model that portrays the accretion of magma at the spreading axis and kinematics of hypersolidus flow in the Josephine ophiolite must take into account the episodicity of magmatism and faulting (Harper, 1988), the geometrical relations described above, and the lack of any high-temperature, sub-solidus, crystal-plastic deformation. Three factors point to an episodicity in magmatism and deformation in the Josephine ophiolite: (1) cross-cutting relationships between brittle faults, dikes, and pillow lavas indicate that soon after magmas were emplaced in the crust and/or extruded on the sea floor, brittle faulting resulted in ~50° tilting of the entire crustal section before the last lavas were erupted (Fig. 2b; Alexander and Harper, 1992); (2) serpentinized faults, which must have been active at temperatures appropriate for serpentine stability

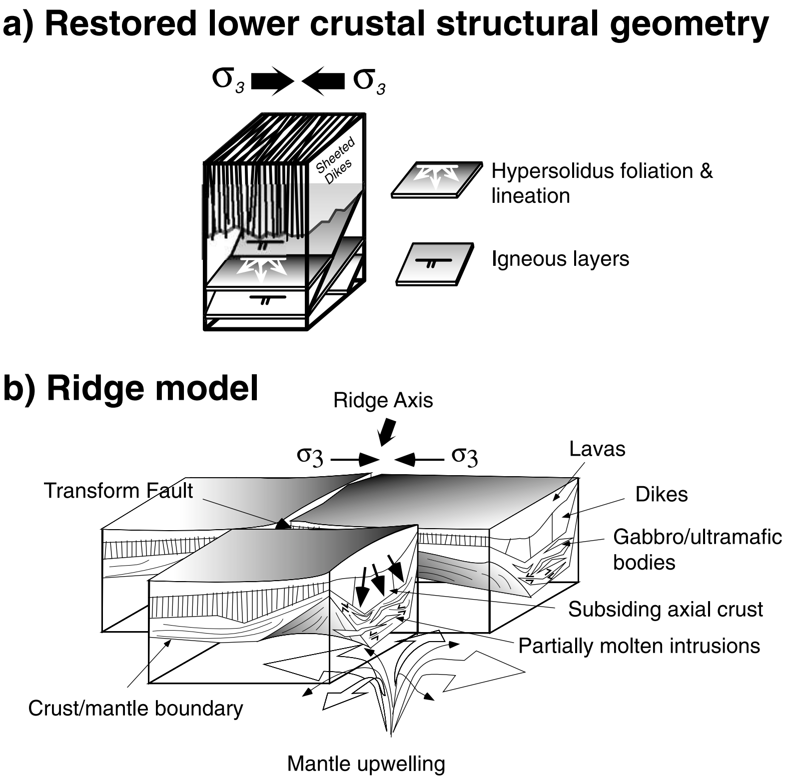

Fig. 8. (a) Schematic diagram illustrating the restored geometric relationships between hypersolidus foliations/lineations, igneous modal layers, and sheeted dikes at the axis of the Josephine spreading center based on structural relations exposed along the Middle Fork of the Smith River. Multiple arrows in the foliation plane reflect scatter of lineation data. (b) Schematic model of structural and lithologic evolution of the axial Josephine lower crust. Kinematic arrows in lower crust represent hypothesized hypersolidus shear zones. See Section 4 for details.

(<500 °C), are cut by diabase dikes chemically similar to the upper-most pillow lavas (Coulton et al., 1995); and (3) lavas, dikes, and cumulate phases within the lower crust display both highly fractionated and primitive trace element compositions (Harper, 1988; A.S. Yoshinobu, unpublished data). This is interpreted to reflect the presence of episodic magma chambers that allow fractionation to occur periodically and thus eruption of more evolved lavas (Harper, 1988). Lavas and dikes with primitive trace element compositions are interpreted to be mantle-derived melts that did not reside in a magma chamber and fractionate (Harper, 1988).

Within this context of episodic magma chamber construction, tectonism, and the lack of any sub-solidus crystal-plastic microstructures in the gabbros, we interpret the high angle between the strikes of the sheeted dikes (regional extension direction) and hypersolidus foliations /layering and the obliquity between poles to sheeted dikes and hypersolidus lineations to be the result of partitioning deformation between the upper and lower crust. The hypersolidus lineations are interpreted to reflect the local direction of flow of the magma, consistent with tiling directions in feldspars and minor hypersolidus shear bands in the gabbros (e.g. Fig. 7b). Because the lineations show a diffuse pattern we suggest that deformation in the lower crust was partitioned in a 3-D fashion that included a component of flow sub-parallel to the ridge axis.

We envision a complete magmatic and tectonic cycle that includes accretion of the lower crust via multiple sills and small intrusions with concomitant emplacement of fine-grained dikes and extrusion of volcanic rocks in the upper crust (Fig. 8b). As magmas are continually added to the middle and upper crust, the thickened crustal column isostatically subsides into the partially molten lower crustal cumulate pile (e.g. Karson, 1998). Subsidence events may also occur due to rapid draining of the shallow axial magma reservoir during an eruption and foundering of upper crustal rocks. Such subsidence may account for the generation of shallowly-dipping hypersolidus foliations, dispersed lineation patterns, compaction of the cumulate pile, and may promote further extraction of inter-cumulus melts (Fig. 7d).

Vertical mass transfer via subsidence may also be necessary in order to emplace numerous horizontally-extensive magma bodies in the lower crust. Emplacement of sills beneath the sheeted dikes by periodic (and rapid) shifts in the orientation of the far-field stresses may explain decoupling between the upper and lower crust (e.g. Gudmundsson, 1990; Parsons and Thompson, 1991). Both diabase and gabbroic/ultramafic sills have been identified in the upper gabbros and lower sheeted dikes, which is consistent with this interpretation (Harper, 1982; Yoshinobu, 1999). We speculate that the extension directions represented by the poles to the sheeted dikes likely reflect the far-field, plate-driven displacement field, whereas 3-D, hypersolidus flow in the lower crust represents the displacement field governed by axial subsidence and magma emplacement.

Kinematic coupling between the upper mantle and lower crust also may have contributed to the strain field in the lower crust (e.g. Nicolas and Boudier, 1995). However, existing structural work in the Josephine peridotite indicates that there is not a straightforward geometrical relationship between mantle foliations/lineations and structures in the lower crust (Harper, 1982; Yoshinobu, 1999). This may be due to the fact that the crust-mantle boundary is mostly a serpentinized fault. Future mapping studies should be focused around the Oregon Mountain area and regions to the north, where the crust-mantle transition is well-preserved.

4.2. Nature of hypersolidus deformation and the memory hypersolidus fabrics

If the lower crust is partially decoupled from the upper crustal extension direction, is there any evidence that is kinematically consistent with ridge-perpendicular extension in the lower crust? Although hypersolidus lineations are locally developed, asymmetric fabrics such as shear bands are rare. Therefore, if the lower crust was accommodating ridge-perpendicular extension via simple shear (e.g. Nicolas et al., 1993; Quick and Denlinger, 1993; Nicolas and Ildefonse, 1996; Buck, 2000), conventional evidence thought to represent such flow (e.g. S–C fabrics, shear bands, asymmetric porphyroclasts, drag folds, etc.) is not preserved in the crystallized rock.

However, it is conceivable that such deformation may have occurred earlier in the crystallization history of the lower crust via melt-enhanced deformation mechanisms (e.g. Nicolas et al., 1993; Park and Means, 1996). It is common to find non-foliated layers up to a meter thick that contain coarse-grained, hypidiomorphic granular textures sandwiched between zones containing well-developed hypersolidus foliations in both the gabbro and wehrlite–dunite sequences (Fig. 5c). Grain size in the foliated zones is commonly finer than in the adjacent isotropic layers. Where lineated, we speculate that these well-foliated zones may be the result of localized strain partitioning along discreet centimetric hypersolidus shear zones (e.g. arrows denoting low-angle shear planes in lower crust in Fig. 8b; Park and Means, 1996). These hypersolidus shear zones collectively may account for large total strains early in the crystallization history when interstitial melt could accommodate crystal sliding, rotation, and fabric development in the absence of pervasive intracrystalline plastic deformation (Nicolas, 1992; Park and Means, 1996).

One way to evaluate microstructural development and the conditions of deformation is to compare experimental rock deformation studies with naturally deformed samples. In comparing microstructures in olivine gabbro from ophiolite and oceanic spreading centers with flow laws derived from rock and mineral deformation studies, Yoshinobu and Hirth (2002) suggested that olivine crystals may display subgrain microstructures indicative of dislocation creep at melt fractions below 10%, while plagioclase deforms by melt-enhanced diffusion creep (based on experiments of Dimanov et al., 1998). Observation of such microstructures would imply that the melt fraction was low enough that tractions could be imposed on a framework of crystals by crystal-crystal interactions. Therefore, because of the lack of any appreciable crystal-plastic deformation within the lower crustal section of the Josephine ophiolite, we speculate that melt fractions may have been greater than 10% during formation of the hypersolidus fabrics. Such an interpretation is substantiated by the observation of fine, centimeter-scale modal and graded layering locally developed in the lower crustal section (Harper, 1984; Yoshinobu, 1999; see also Korenaga and Kelemen, 1998, for similar constraints from the Oman ophiolite). If large simple shear strains accommodated plate separation at low melt fractions, these structures should be disrupted, transposed and/or isoclinally folded (e.g. Bedard and Constantin, 1991) and evidence for dislocation creep should be observed in the crystallized rock (e.g. Boudier and Nicolas, 1995; Yoshinobu and Hirth, 2002). Although hypersolidus foliations are oblique to layers, only rarely do they disrupt layering (e.g. Fig. 5b). It is plausible that the enigmatic layer/foliation obliquity represents a form of hypersolidus S–C fabric such that the layers represent shear planes whereas the foliations represent flattening planes. In this scenario, displacement parallel to layers is entirely accommodated by inter-crystalline melt. However, if deformation continued during solidification, one would expect to see a transition from hypersolidus to subsolidus deformation mechanisms in the preserved microstructures regardless of whether the strain regime was simple or pure shear (Paterson et al., 1989; Dick et al., 1999; 2000). No such transition is observed in the lower crustal section of the Josephine ophiolite.

We suggest that the preservation of fabrics that formed in the hypersolidus state, and the lack of any appreciable crystal-plastic fabrics depends upon the ambient thermal regime and kinematic path during solidification of the lower crustal section. If the lower crust cooled slowly during continued extension, then one would predict that the rocks should display evidence for a transition from hypersolidus to subsolidus deformation, as has been observed in gabbros from the Southwest Indian and Mid-Atlantic Ridges (Karson, 1998; Dick et al., 2000). As noted above, crystal-plastic deformation in the lower crust of the Josephine ophiolite is restricted to narrow (centimeter-scale) shear bands locally developed in the upper gabbro (Alexander et al., 1993). Therefore, it is hypothesized that the gabbro and wehrlite–dunite sequences underwent rapid cooling following accretion in the lower crust and subsidence in the middle and upper crust. This conclusion is consistent with the fine-grained nature of the gabbros and stable isotopic studies that indicate cooling through 350 °C during hydrothermal circulation at the ridge axis before intrusion of the last diabase dikes (Alexander et al., 1993).

4.3. Implications to ridge magma chamber models

The results and interpretations presented here introduce additional complexities to models of spreading center magma chambers where magma supply is high enough to produce a continuous plutonic and volcanic crust. We suggest that the lower crust of the Josephine ophiolite was constructed by the emplacement of discreet batches of magma with sill-like geometries (e.g. Figs. 1b and 8b). Such an interpretation is consistent with geochemical data from the dikes and volcanic rocks in the Josephine upper crust (Harper, 1988) as well as drilling studies of modern slow-spreading oceanic crust from the Southwest Indian Ocean (Dick et al., 2000). However, we suggest that hypersolidus structures in the lower crust may be the result of magma emplacement dynamics (local upper crustal subsidence, internal flow) rather than due to plate spreading (Nicolas et al., 1993; Phipps-Morgan et al., 1994; Buck, 2000). Furthermore, hypersolidus structures within the Josephine ophiolite probably reflect localized deformation of the magma with melt fractions in excess of 10%. Such an interpretation is consistent with geophysical studies of contemporary oceanic spreading centers, but differs in the interpreted driving mechanism. Our model for the Josephine lower crust is most consistent with a ‘nested sill’ model (Fig. 1b; Boudier et al., 1996; Kelemen and Aharanov, 1998) and calls into question the notion that hypersolidus structures within the lower crust represent a continuous kinematic path from shallow melt lens through the formation of the entire lower crust (e.g. Fig. 1a; Nicolas et al., 1993).

5. Conclusions

Based on new geologic mapping, restoration of hypersolidus structures to their on-axis orientation, and microstructural observations, we draw the following conclusions. Hypersolidus fabrics (formation of mineral anisotropy, igneous modal layering, etc.) are the dominant structures observed. No pervasive crystal-plastic deformation is observed. Restoration of the hypersolidus foliations and igneous modal layers to their on-axis orientation indicates that they have strikes that are sub-perpendicular to the strike of the paleo-spreading center as defined by the orientation of sheeted dikes and on-axis, oceanic faults. Hypersolidus lineations define a dispersed 3-D flow pattern in the lower crust, whereas extension directions in the upper crust (i.e. poles to sheeted dikes, oceanic normal faults) are unidirectional and perpendicular to the inferred paleo-ridge axis. This geometry is attributed to hypersolidus flow in the lower crust that is both oblique and parallel to the ridge axis driven by subsidence of the axial upper and middle crust during continued accretion of basaltic magma. One implication is that hypersolidus fabrics may not be directly related to ridge perpendicular extension; thus, the lower crust is partially decoupled from the upper crustal strain field. If correct, this interpretation indicates that caution must be exercised when interpreting hypersolidus structural patterns in ophiolites where kinematic links between the upper and lower crust cannot be established.

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