Fabric development in the mantle section of a paleo-transform fault and its effect on ophiolite obduction, New Caledonia

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ABSTRACT

The Bogota Peninsula shear zone has been interpreted as a paleo-transform fault in the mantle section of the New Caledonia ophiolite. New, detailed field measurements document the rotation of foliation, lineation, and pyroxenite dikes across a 50-km-wide region. Deformation intensity recorded by folding and boudinage of dikes increases toward a central, 3-km-wide mylonitic zone. We used several additional methods to characterize fabric patterns across the shear zone. The shape-preferred orientation of orthopyroxene grains, computed from outcrop tracings, closely parallels field fabrics, with increased alignment and flattening near the center of the shear zone. The lattice-preferred orientations of olivine are consistent with high-temperature fabrics; the a axes within the mylonitic core were used to constrain the orientation of shear zone boundaries. Seismic anisotropy calculations, based on the lattice-preferred orientation of olivine, indicate 5%–11% shear-wave anisotropies, with increased values in the center of the shear zone. The magnetic silicate fabric in the rocks, determined from anisotropy of magnetic susceptibility techniques, broadly matches field fabrics but provides less consistent information across the shear zone than other fabric methods.

This suite of field and laboratory data provides a unique and detailed view of strain and fabric patterns across a shear zone in oceanic mantle lithosphere. Because the primary mantle fabrics seem to be related to the present distribution of ophiolitic rocks in New Caledonia, we propose that ophiolite obduction and Neogene extension may have been controlled by preexisting fabrics and structures in the oceanic lithosphere.

INTRODUCTION

Studies of modern oceanic transform faults and fracture zones have provided an increasingly detailed picture of the topography, geometry, and evolution of these important plate boundary structures (e.g., Menard and Atwater, 1969; Bonatti, 1978; Choukroune et al., 1978; Fox and Gallo, 1984; Lawson et al., 1996). However, actual in situ observations are difficult, and, although steep fracture zone walls provide some insight into deeper rocks and structures (e.g., Fox et al., 1976; Prinz et al., 1976), it is impractical to study structures developed across the entire transform system at anything but the shallowest lithospheric levels.

Where oceanic transform faults are preserved on land, such as within ophiolites, they can be investigated by more direct methods and provide invaluable insight about deformation patterns preserved in deeper lithospheric levels. Well-exposed transform faults, not just shear zones, are relatively rare in ophiolites. The best-known examples include the Arakapas fault zone in the crustal section of the Troodos ophiolite in Cyprus, where sheeted dikes are rotated 90° into alignment with the fault zone (e.g., Moores and Vine, 1971; Simonian and Gass, 1978; MacLeod et al., 1990), and the Coastal complex in the Bay of Islands complex, Newfoundland, which has been interpreted as a fracture zone or transform fault that may have facilitated ophiolite obduction (Karson and Dewey, 1978; Casey et al., 1983; Suhr and Cawood, 2001). In both of these examples, most of the ophiolitic material is located on one side of the inferred plate boundary, providing a glimpse of half of the transform system deformation.

The Bogota Peninsula shear zone in the New Caledonia ophiolite also has been interpreted as a paleo-transform fault (Prinzhofer and Nicolas, 1980; Nicolas, 1989). Unlike the Arakapas fault and the Coastal complex, however, this shear zone is exposed entirely within peridotites from the mantle section of the ophiolite. We used several field and petrofabric techniques to characterize strain patterns across the shear zone and to develop a conceptual model for its evolution during progressive deformation. We then linked these strain patterns to ophiolite obduction and subsequent extension, suggesting that primary mantle fabrics and structures may have controlled, at least in part, these major tectonic events.

TECTONIC HISTORY

The island of New Caledonia represents the exposed portion of the Norfolk Ridge (Fig. 1), a microcontinental ribbon that rifted from the eastern Gondwanaland margin during the Late Cretaceous (e.g., Dubois et al., 1974; Crawford et al., 2003). This rifted Mesozoic-age material forms the basement in New Caledonia (Aitchison et al., 1995; Cluzel et al., 2001). Younger geologic structures on the island developed in two phases of postrifting activity, which are summarized next.
Eocene Convergence

Based on dike ages in New Caledonia and plate reconstructions of the southwest Pacific, a phase of convergence affecting New Caledonia began at 53–55 Ma (Crawford et al., 2003; Cluzel et al., 2006). Most tectonic models suggest that the South Loyalty Basin was subducted beneath the Loyalty arc during this time (Eissen et al., 1998; Cluzel et al., 2001; Crawford et al., 2003; Spandler et al., 2005). This northeast-facing convergence eventually ceased when the Norfolk Ridge reached the subduction zone at ca. 37 Ma (Cluzel et al., 2006).

During convergence, several thrust sheets were emplaced onto New Caledonia, including the basaltic Poya terrane along the southern coast, the high-pressure Pouebo terrane in the northeast, and the ophiolite nappe (Fig. 1). The Poya terrane is interpreted as allochthonous material, such as seamounts, scraped off the downgoing slab during subduction and accreted to the forearc before emplacement onto the Norfolk Ridge (Cluzel et al., 2001). Radiolarians within the Poya terrane are typically Late Cretaceous to Paleocene in age (Cluzel et al., 2001). The Pouebo terrane represents the subduction and metamorphism of oceanic crust and associated sediments (Spandler et al., 2004, 2005), making it a likely metamorphic equivalent of the Poya terrane (Cluzel et al., 2001; Whattam et al., 2008). It is difficult to determine the age of the Pouebo terrane protolith, but peak metamorphism in these rocks occurred at ca. 44 Ma, followed by rapid cooling and exhumation of a coherent block from 40 to 34 Ma (Baldwin et al., 2007). The ophiolite nappe is predominantly composed of mantle material and presently drapes into the South Loyalty Basin from the Belep Islands to the southern part of New Caledonia, based on gravity data (Collot et al., 1987, 1988). The
ophiolite is considered either Late Cretaceous (e.g., Prinzhofer, 1981) or Eocene (What- tam et al., 2008) in age, and was the last to be emplaced onto New Caledonia between 37 and 34 Ma (Cluzel et al., 2006).

### Neogene Extension

Regional extension, which modified both onshore (Lagabrielle et al., 2005; Chardon and Chevillotte, 2006; Lagabrielle and Chauvet, 2008) and offshore (Dupont et al., 1995; Lafoy et al., 1996; Auzende et al., 2000; Chardon et al., 2008) structures, followed Eocene convergence and has lasted until the present day. Based on a study of small-scale faults throughout the island, Lagabrielle et al. (2005) suggested that there have been two phases in this extensional history.

The first phase, beginning in the Oligocene, had extension directions from 90° to 140° and primarily affected peridotites in the ophiolite nappe. The western Caledonian fault zone (known as the Sillon in older literature), which forms the abrupt southwestern boundary of the Massif du Sud, is interpreted as a large detachment fault that facilitated extension during this phase (Fig. 1; Lagabrielle and Chauvet, 2008). This phase of postorogenic collapse likely facilitated unroofing of the high-pressure Pouebo terrane (Rawling and Lister, 2002) and has been linked to possible slab break-off (Cluzel et al., 2005).

The second phase, affecting Upper Pliocene—through Quaternary-age rocks, has variable extension directions of 90°, 40°, and 0°–10° (Lagabrielle et al., 2005). The obliquity between these directions and the trend of the Norfolk Ridge has resulted in sinistral transtension (Chardon and Chevillotte, 2006). Lagabrielle et al. (2005) suggested that this extension is due to the modern plate-tectonic setting, perhaps related to the flexure of the oceanic lithosphere currently subducting under the Vanuatu island arc.

### NEW CALEDONIA OPHIOLITE

The ophiolite sheet has two primary exposures on New Caledonia. The large continuous sheet in the south is known as the Massif du Sud and includes the Bogota Peninsula shear zone, which is the focus of this study. The series of klippen along the northwestern coast are in fault contact with the Poya terrane; several klippen are deformed in the Belep shear zone (Fig. 1). In this section, we compare the compositional and fabric patterns between these two exposures as well as the previous tectonic interpretations of the two shear zones.

#### Massif du Sud

The Massif du Sud has a well-developed mantle section that is ~1–3 km thick, is composed primarily of harzburgite and dunite, and has a few layered gabbros and mafic lavas (Avias, 1967; Prinzhofer et al., 1980). The contact between the mantle and crustal sections of the ophiolite is observed at a few locations, where it is subhorizontal (Prinzhofer et al., 1980). Regional mapping within the mantle section (Prinzhofer and Nicolas, 1980; Prinzhofer et al., 1980) demonstrates generally E-W–to NW-SE–striking foliations with <30° southerly dips and subhorizontal N-S–trending lineations (Fig. 1).

On the Bogota Peninsula along the northeast coast, foliation becomes subvertical and N–S–striking, while lineation remains subhorizontal and N–S–trending (Fig. 2). Fabrics are rotated over a 50-km-wide region around a central 3-km-wide high-strain zone. Fabric strength, as defined by both the lattice-preferred orientation (LPO) of olivine and stretched orthopyroxene grains with aspect ratios up to 25:1, increases toward the center of the shear zone (Prinzhofer and Nicolas, 1980). Shear sense indicators are consistently dextral across the Bogota Peninsula and surrounding coastline. We refer to the entire region of rotated fabrics as the Bogota Peninsula shear zone.

#### Ophiolite Klippen

The ophiolite klippen have a different compositional and fabric character than the Massif du Sud. Plagioclase lherzolites are exposed in many of the klippen, with local diopside harzburgites and spinel lherzolites (Moutte, 1982). The fabric patterns and orientations are generally less consistent than those in the Massif du Sud, both within each klippe and between klippen. This inconsistency has been attributed to late deformation and rotation of blocks (Leblanc et al., 1980; Nicolas, 1989). Lagabrielle and Chauvet (2008) interpreted these klippen as isolated allochthonous remnants of the ophiolite nappe displaced along the western Caledonia fault zone in response to regional extension.

Several of the northernmost klippen are also deformed by the 120-km-long Belep shear zone (Fig. 2). Outside the shear zone, fabrics are similar to those in the Massif du Sud with E–W–to NW-SE–striking foliations and subhorizontal N–S–trending lineations (Fig. 1; Nicolas, 1989). Inside the shear zone, the NW-striking foliation becomes vertical, while lineation remains horizontal. In the Tiebaghi Massif, the subhorizontal foliations in harzburgites outside the shear zone can be traced to vertical mylonitic fabrics in lherzolites (Moutte, 1982).

#### Previous Tectonic Interpretation of Structures

Both the Bogota Peninsula and Belep shear zones have been interpreted as paleotransform faults by previous workers and therefore have been used to infer ridge-transform geometries for rocks in the Massif du Sud and ophiolite klippen. We describe the data and reasoning for each interpretation next and attempt to place each shear zone into its original tectonic context (Fig. 2).

For the Bogota Peninsula shear zone, several field observations are important: (1) A strain gradient is observed on both sides of the Bogota Peninsula toward a central mylonitic zone (Fig. 2B), showing consistent dextral shear sense everywhere, (2) foliation in the shear zone is perpendicular to that in the Massif du Sud (Fig. 1), and (3) lineation remains consistently N–S–trending and subhorizontal inside and outside the shear zone (Fig. 1). Based on these data, Prinzhofer and Nicolas (1980) interpreted the Bogota Peninsula shear zone as a N–S–striking, dextral paleotransform fault.

Nicolas (1989) used these same field data to suggest that the Massif du Sud formed along the south flank of an E–W–striking ridge (relative to present geographic coordinates). The E–W–strike is based on (1) the assumption that the N–S–trending lineations record relative plate motion and (2) that a ridge should be perpendicular to plate motion. The south flank interpretation is based on a shear sense inversion within the Massif du Sud (Nicolas, 1989), a pattern that has been used to infer flanking directions in the Oman ophiolite (see Fig. 2.2 in Nicolas, 1989).

These interpretations of ridge-transform geometry imply that the Massif du Sud formed in an inner corner environment bound on the west by the Bogota Peninsula shear zone (Figs. 2D and 2E). From the available data, it is difficult to determine whether the shear zone records deformation only on the Massif du Sud plate, or whether it records deformation across the entire transform portion of the system. In the former case, both ridge segments would be located in the South Loyalty Basin, as illustrated in Figure 2E. In the latter case, the western ridge segment should be south of the present-day Bogota Peninsula and therefore onshore.

Similar field fabric patterns are used to suggest that the Belep shear zone represents a transform fault (Sécher, 1981). However, the noncontinuous nature of exposures and conflicting shear sense indicators complicate this argument. The shear zone affects the western sides of Poum and Tiebaghi, all of Yandé, and the eastern sides of Art and Pott, requiring a change from NW-striking on New Caledonia to more N-striking near the Belep islands.
Figure 2. Maps of the (A) Belep and (B) Bogota Peninsula shear zones, showing foliation trajectories and shear sense indicators. (C) Interpretation of the Belep shear zone from Nicolas (1989), where the diffuse fabric patterns and opposite shear sense in the Belep Islands are linked to asthenospheric flow patterns. (D) Interpretation of fabric patterns from Bogota Peninsula shear zone and Massif du Sud, where the shear zone records dextral motion along a transform fault and the Massif du Sud records spreading-related fabrics on the south flank of an E-W–trending ridge. Nicolas (1989) suggested that the Bogota Peninsula shear zone may represent only one half of the transform fault system, meaning that the present exposure on New Caledonia reflects rocks presently separated by a fracture zone (and not rocks between two ridge segments). Diagrams A–D were modified from Nicolas (1989). (E) Using these interpretations of the Belep and Bogota Peninsula shear zones, we show possible ridge-transform geometries if both reflect transform faults. For the Bogota Peninsula shear zone, the distance between ridge segments is unknown and used solely to illustrate the expected sense of shear.
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Our field measurements from the Bogota Peninsula shear zone are primarily from coastal outcrops, where wave-cut terraces provide excellent three-dimensional exposures and relatively fresh rocks (Fig. 3). Most of these coastal sites are in the northern structural domain of Bogota Peninsula, where steeply dipping foliations are observed. In contrast, rocks in the southern domain (the highlands) are often deeply weathered and have moderately inclined foliations. Prinzhofer and Nicolas (1980) first noted these two structural domains and interpreted the contact between them as a low-angle, E-W–striking thrust fault.

Field Fabrics

The alignment of spinel and/or pyroxene grains defines foliation and lineation in the peridotites along the Bogota Peninsula (Fig. 4). Careful analysis of foliation and lineation orientations across the 50-km-wide study region in Figure 5 demonstrates a more complex pattern than is evident when fabric data are simply combined into shear zone and non–shear zone fabrics (Fig. 1). Instead of observing N–S–striking vertical foliations in the shear zone, the

(continued)
strike of foliation rotates from NW-SE outside Bogota Peninsula to NNE-SSW in the center of the shear zone; this rotation is accompanied by increasing dips, from ~60° on the margins of the study area to 75°–90° in the center. Instead of observing N-S–trending lineations everywhere, the subhorizontal lineation directions rotate clockwise with foliation from NW-SE– to NNE-SSW–trending in the center of the shear zone.

Based on these patterns of rotation, we divide the shear zone into five domains: a center with NNE-striking foliations, two near-field regions with N-striking foliations, and two far-field regions with NW-striking foliations (stereographic projections in Figs. 3 and 6). The stippling in Figure 5 highlights these five domains. This interpretation is noted here because the remaining field data are either presented from these five domains or, if fewer data are available, in terms of only three domains (center, near field, and far field), given the overall symmetry across the shear zone. 

Dikes

Several generations of dikes intrude the ultramafic rocks along Bogota Peninsula. Crosscutting relationships indicate the relative timing from earliest to latest as pyroxenite, feldspathic pyroxenite, hornblende gabbro, and diabase dikes (Prinzhofer and Nicolas, 1980). While Nicolas (1989) reported that all dikes on the Bogota Peninsula shear zone were vertical with an average strike of 30°, we demonstrate that dike orientations depend on position across the shear zone, as well as composition. Here, we focus on pyroxenite dikes and diabase dikes, which are present in great enough quantity to determine robust patterns across the shear zone.

Pyroxenite Dikes

Two sets of pyroxenite dikes can be identified in each of the five shear zone domains based on their deformation behavior. Poles illustrated as solid circles in Figure 6 represent dikes that are necked or boudinaged, whereas open circles represent dikes that display constant thickness. Like foliation and lineation, both sets rotate clockwise from the far field toward the center of the shear zone.

The stretched dikes rotate from NW-striking with a variety of dips in the far field to NNW-striking and steeper in the near field; dikes from the center have similar orientations to near-
field dikes, but this set includes dikes with true N-S strikes. Except for within the center, these stretched dikes are subparallel to foliation. Qualitatively, the stretching recorded by this dike set increases toward the center of the shear zone, as illustrated by the greater separations between boudins in the shear zone interior. Further, boudinage is observed on both horizontal and vertical surfaces in the center, as opposed to primarily on horizontal surfaces in far-field regions (Fig. 7).

The dikes with constant width also change orientation across the shear zone. In the far field, these dikes include subhorizontal orientations, WNW-striking surfaces with a variety of dips, and N-S-striking steep surfaces. In the near field, the dikes either parallel the boudinaged set or are steeply dipping and NNE-striking (in the NW) or gently dipping and E-W-striking (in the SE). In the center, there are few dikes with constant thickness; those present strike NE with a variety of dips.

Folded pyroxenite dikes were occasionally observed with axial planar foliations and isoclinal fold shapes (Fig. 7E). Except for three folds within the center of the shear zone, folded pyroxenite dikes were exclusively found in the far-field regions. We plot the orientations of folded dikes of any composition, including pyroxenite dikes, in Figure 6 because so few folds were observed across the region. Because the axial planes were often difficult to measure accurately in the field, we plot the poles to fold limbs. Since fold shapes are typically isoclinal, these demonstrate that fold axial planes are subparallel to foliation for each of the three shear zone domains. These data also

Figure 5. Graphs showing the orientation of (A) foliation and (B) lineation across the shear zone, where distance is measured relative to the westernmost station studied (KA04-06). Foliation strikes and lineation trends were also added from Prinzhofer and Nicolas (1980). For comparison, the shape-preferred orientation (SPO) foliation ($S_1$-$S_2$ plane) and lineation ($S_1$) have been added to A and B, respectively. (C) The magnitude and shape of the SPO ellipsoid. The stippled regions in each chart show the shear zone center and near-field regions based on field fabric orientations. The gray horizontal bar denotes the average value calculated from four stations within the main Massif du Sud.
Figure 6. Lower-hemisphere, equal-area projections from the five (or three) shear zone regions highlighted in Figure 3. These regions were defined empirically based on consistent foliation and lineation orientations. Many of the other types of data (dike orientations, joint orientations, shape-preferred orientation [SPO]) support these divisions as they also rotate with the changing foliation orientation. For smaller data sets, we combined data from the two near-field and two far-field regions to better illustrate the patterns across the shear zone. For pyroxenite dikes, filled circles represent dikes showing boudinage or necking while open circles represent dikes with constant thickness.
Figure 7. Photographs of pyroxenite (A–E) and diabase (F) dikes from the ophiolite. (A) Subhorizontal, weakly boudinaged dike from the Massif du Sud can be compared with boudinage for dikes in the far-field (B), near-field (C), and central (D) regions of the shear zone. Note the near total transposition into the foliation plane and vertical boudins in D from the center of the shear zone. (E) Folded dikes are common in the far-field regions but were rarely observed elsewhere. (F) An undeformed diabase dike.
demonstrate that folds are most common in the far field.

To summarize, both the boudinaged and folded dikes support our division of the shear zone into five domains. The low-strain far-field domains preserve pyroxenite dikes with a variety of orientations, including folded dikes. The high-strain center domain preserves a narrow range of dike strikes, and most are quite boudinaged. We also note here that orthopyroxene blebs, significantly larger than the average orthopyroxene grain size, were observed (Fig. 4D). We interpret these as relic boudins from pyroxenite dikes, which are so extended that it is no longer possible to trace them as continuous planar structures. These blebs may help explain why boudinaged dikes from the center of the shear zone are not parallel to foliation, unlike the pattern from the other four domains. These dikes would have become so extended that they could no longer be recognized.

**Diabase Dikes**

Diabase dikes are mostly restricted to the center of the shear zone, with consistent ~30° strikes and steep dips to the NW (Fig. 6). These dikes commonly crosscut field fabrics in harzburgites. The dikes may show slight changes of orientation relative to position in the shear zone—those from near-field regions have more northerly strikes than the main NE-striking set from the center of the shear zone. Boudinage of diabase dikes was not observed (Fig. 7), although several dikes showed minor warping and bending.

**Joints**

Joint orientations were measured on marine terraces at select stations along the coastline. We collected data about the representative orientations of nonhorizontal, systematic joint sets, but we did not record information about joint spacing or density.

The poles to joints for the five shear zone regions are plotted in Figure 6. Two distinct joint sets are present in each region. These sets are mutually perpendicular and become steeper in the center of the shear zone. The poles to joints mimic the rotation of foliation, so that the two sets are always parallel and perpendicular to foliation.

**ORTHOPYROXENE MACROSCOPIC FABRICS**

In the field, tracings of orthopyroxene grains were made on sheets of clear plastic at thirty stations along the coast. Three oriented tracings, each with 80–300 individual grains, were collected at each station on outcrop surfaces that were approximately mutually perpendicular. Data were also collected from four stations in the Massif du Sud to facilitate comparison with fabrics developed outside of the shear zone.

**Determining the Shape-Preferred Orientation Ellipsoid**

Tracings were digitized using a flatbed scanner; the resulting images were processed using ImageJ (Rasband, 2009). For each tracing, a two-dimensional fabric ellipse was calculated using the Intercept method (Laueune and Robin, 1996) from the software package Ellipsoid 2003 (Launeau et al., 1994; Launeau and Robin, 1996, 2005; Robin, 2002). For each station, these two-dimensional ellipses were combined mathematically into a three-dimensional fabric ellipsoid using Ellipsoid 2003. This ellipsoid represents the shape-preferred orientation (SPO) of the population of pyroxene grains.

A significant advantage of SPO analysis is that, in addition to orientation information, the method provides quantitative constraints on strain magnitude. Using the principal axes (S, S, S) of each SPO ellipsoid, we can determine the natural shear ε and shape parameter ν. When ε = 0, the SPO ellipsoid is spherical; ellipticity increases with increasing values of ε (Nadai, 1963; Hsu, 1966; Hossack, 1968). The values of the shape parameter have the range −1 < ν < 1, where ν < 0 corresponds to constrictional fabrics, ν = 0 is plane strain, and ν > 0 represents flattening fabrics (Ramsay and Huber, 1983).

**SPO Ellipsoid Results**

The results of our SPO analysis are reported in Table 1. To examine spatial changes across the shear zone, the values of ε and ν are plotted graphically in Figure 5C. The biggest difference between values from the Massif du Sud (in gray) and those from the shear zone occurs in the center, where ε is elevated, between 0.2 and 0.3, and fabrics change from prolate/plane (in the far field) to plane strain/oblate (in the center). A secondary shear zone originally noted by Prinzhofer and Nicolas (1980) is also evident northwest of the central zone, as indicated by higher ε and ν values, steeper foliations, and S-plunging lineations.

The orientations of the fabric ellipsoids are also plotted in Figure 5 as an SPO foliation (S, S, S plane) and lineation (S) to facilitate direct comparison with field fabrics on a station-by-station basis. The consistency between these calculated SPO fabrics and the measured field fabrics is quite clear, although the plunges of the SPO lineations are often steeper than those of field lineations in the center of the shear zone (Fig. 5). The orientations of S, S, and S are also plotted from the three shear zone domains in Figure 6. These data illustrate the clockwise rotation of fabrics from the far field to the center; their better clustering in the center highlights the increased fabric strength in the high-strain portion of the shear zone.

**Microstructural Fabrics**

We measured olivine lattice-preferred orientations (LPO) across the shear zone to better characterize the temperature and style of deformation in peridotites (Nicolas and Christensen, 1987; Drury and Fritz Gerald, 1998). We also computed predicted seismic anisotropy values based on the fabric strength and orientation of olivine LPO. Both are described here.

**Olivine LPO**

Olivine LPO was measured at 12 sites across the shear zone (A–L in Fig. 3). In each sample, 102 grains were oriented using a universal stage mounted on a petrographic microscope. Because of the moderate degree of serpentinization in these rocks, this technique was invaluable for identifying and orienting individual grains. The LPO patterns are presented in their in situ geographic orientation (and not the more standard rock fabric reference frame) in Figure 8. When possible, we also show the orientations of field foliation and lineation, which were independently measured at the time of sample collection. To test the repeatability of these measurements, LPO values were measured from two different samples at the same site (site D).

All sites show point distributions for the three crystallographic axes consistent with the high-temperature fabrics expected in mantle conditions. In detail, sites from the center of the shear zone (A–H) have subhorizontal a axes that are ~10°–15° clockwise of the field lineation, consistent with dextral shear sense (Nicolas, 1989). The b and c axes form more weakly defined point distributions, where b axes often lie near the pole to field foliation. Sites from the near field (I and J) and far field (K and L) have slightly weaker fabric patterns than those from the center. The a axes in these regions rotate with the changing foliation strike.

**Seismic Anisotropy**

Using the LPO data, we computed seismic velocities and shear-wave anisotropy values with the Ani2K software (based on Mainprice, 1990). In these calculations, seismic velocities
are determined using the LPO, density (Crosson and Lin, 1971), and elastic stiffness coefficient of olivine (Abramson et al., 1997). The Voight-Reuss-Hill averaging technique was used for these calculations; the choice of averaging technique affects the absolute velocities but not the anisotropy parameters (Christensen and Lundquist, 1982). Further, calculated anisotropy values are typically higher than laboratory measurements by ~1% (Christensen, 2002). Given these caveats, our seismic anisotropy results likely overestimate the true values, but they are still useful for comparison with shear-wave splitting results from other lithospheric-scale shear zones.

The maximum, minimum, and average compressional and shear-wave velocities ($V_p$, $V_s$) and anisotropies ($AV_p$, $AV_s$) for each LPO measurement are reported in Table 2 (see definitions of these values in Mainprice and Silver, 1993). Select seismic properties are illustrated graphically in Figure 8. There is no statistical difference between velocities or anisotropies for sites in the center of the shear zone (A–H) and those flanking the shear zone (I–L). The compressional wave velocities are typically of those calculated from LPO patterns in mantle xenoliths, but most other parameters (especially $AV_p$ and $AV_s$) are slightly higher than values calculated from xenoliths (see table 4 in Titus et al., 2007), which is not surprising given the aforementioned caveats.

### MAGNETIC FABRICS

For the same subset of sites across the shear zone (A–L), we applied a combination of high-field and low-field anisotropy of magnetic susceptibility (AMS) techniques. These two types of AMS measurements are used in tandem to isolate the primary magnetic silicate fabric from olivine and pyroxene. The methods and results of this analysis are presented here as a test case for the applicability of this technique to mantle rocks.

### AMS Methodology

For mantle rocks, secondary minerals like magnetite typically dominate the low-field AMS signal (MacDonald et al., 1988; Bina and Henry, 1990; Richter et al., 1996; Borradaile et al., 1995).
Figure 8. Results of olivine fabric analysis. Each column represents data from a single station labeled A–L on Figure 3. The stations are arranged from west to east, and the portion of the shear zone (center, near field, far field) from which samples were collected is also indicated. The first three rows show lower-hemisphere, equal-area projections for the olivine lattice-preferred orientation (LPO) data. Data have been contoured as percent area, and $N = 102$ for each site. The bottom three rows show calculated seismic anisotropy parameters from these data, including compressional wave velocity ($V_p$), shear-wave anisotropy ($AV_s$), and polarization directions. Contours for $V_p$ are in 0.2 km/s, and the contours for $AV_s$ are in 1% delay time; the maximum and minimum values are denoted next to each plot (and also in Table 3). (Continued on following page).
Figure 8 (continued).
and Lagroix, 2001). These minerals form during serpentinization at temperatures <500 °C (e.g., Saad, 1969), so that their shape anisotropy is often unrelated to mantle flow fabrics (MacDonald et al., 1988). We wanted to separate the primary, mantle fabric due to paramagnetic components (olivine, orthopyroxene, chromian spinel) from the secondary, serpentine-related fabric. This separation is possible in high magnetic fields (>0.7 T), where the saturation magnetization of ferromagnetic minerals is reached (at least in the absence of hematite or maghemite), because the slope of the hysteresis loop becomes constant and represents the contribution from paramagnetic minerals only (Kelso et al., 2002; Martín-Hernández and Ferré, 2007). This technique has been applied successfully to other mantle rocks (Ferré et al., 2005).

Both low- and high-field AMS results can be represented geometrically by an ellipsoid with three mutually perpendicular principal axes: k1 ≥ k2 ≥ k3 (Jelinek, 1981). The bulk susceptibility k, is the average value of the principal axes and relates the induced magnetization in a sample to an applied field. By using the separation method, the low-field AMS ellipsoid characterizes the bulk rock fabric, whereas the high-field AMS ellipsoid characterizes the primary silicate fabric of the peridotites. Two other parameters are also useful for our analysis (Jelinek, 1981): (1) the degree of anisotropy P, which quantifies the strength of the magnetic fabric, where P = 1 for a perfect sphere, and P > 1 for increasing ellipticity; and (2) the shape parameter T characterizes the shape of the AMS ellipsoid, varying such that −1 ≤ T ≤ 1. When T = 1, the AMS ellipsoid is perfectly oblate; when T = 1, it is perfectly prolate.

In the laboratory, 20 mm cubic specimens were cut from oriented hand-samples with horizontal, vertical N-S, and vertical E-W faces. A Kappabridge KLY-3S was used to measure low-field AMS at a low alternating field of 300 A/m. A Vibrating Sample Magnetometer was used for high-field AMS measurements up to fields of ~1 T. High-field AMS was measured on a subset of low-field cubic specimens. Low-field and high-field AMS data are presented graphically in Figure 9, and site value averages are reported in Table 3.

### Low-Field AMS

The low-field AMS axes, representing 14–53 cubic specimens per site, are clustered tightly at the site level. Low-field k1 axes always form point distributions that plunge moderately to steeply toward many different azimuths across the shear zone. The k2 and k3 axes either form point distributions or together form a girdle along the plane perpendicular to k1. Like k1 orientations, k2 and k3 have inconsistent orientations across the shear zone. For most stations, a great circle connecting two of the low-field AMS axes is approximately parallel to field foliation, but the correlation is not perfect.

The bulk susceptibility k ranges from 350 to 2350 μSI, although 70% of specimens have k < 1000 μSI. The degree of anisotropy P ranges from 1.009 to 1.34, with ~90% specimens showing P < 1.1. The shape factor varies from −0.94 to 0.85, although 70% of specimens fall in the flattening field (T > 0). There are no consistent patterns for P or T relative to position within the shear zone.

### High-Field AMS

We might expect the long axis, k1, of the high-field AMS ellipsoid to parallel the lineation direction and the short axis to parallel the pole to foliation (Borradaile and Henry, 1997; Grégoire et al., 1998). However, our results, which represent 4–6 specimens per site, show no clear pattern in the high-field ellipsoid orientation. We attribute this scatter, in part, to the presence of hematite, which cannot be saturated even at high-field conditions. Further, fewer specimens were measured in high-field conditions because these measurements were more time-consuming (1 h per sample) than low-field measurements (5 min). Last, the low intrinsic degree of anisotropy for Fo0.75 (Belley et al., 2009) may contribute to the weakly anisotropic high-field AMS.

High-field susceptibilities, calculated from the high-field ellipsoid, vary from 350 to 615 μSI. The degree of anisotropy ranges from 1.02 to 1.13. The shape factor varies from −0.66 to 0.68. Similar to the low-field AMS results, there are no consistent patterns for P or T relative to position within the shear zone.

### COMPARISON OF FABRIC MEASUREMENTS

To facilitate comparisons between our shear zone measurements, we compiled field fabric and dike orientations, SPO from macroscopic orthopyroxene, LPO of olivine, low- and high-field AMS data, and joint measurements for stations A–L in Figure 9.

There is broad consistency between data sets that record high-temperature fabrics, including foliation, lineation, SPO, LPO, and high-field AMS. As discussed previously, the field and SPO lineations are typically parallel. Olivine a axes are clockwise from this orientation, which is expected for dextral systems (Tommasi et al., 1999). Although the high-field AMS patterns are less well defined, k1 is parallel to field lineation for sites with the most prolate ellipsoids (E and G). For several other sites (B, D, and F), k1 parallels lineation; these sites have less prolate ellipsoid shapes, meaning that the distinction between k1 and k3 is less important.

The records of lower-temperature fabrics from low-field AMS and joints are consistent on a station-by-station basis. The poles to joints, which are always gently plunging due to the subhorizontal coastal exposures, typically parallel low-field AMS axes that are also gently plunging. The orientation of joints (Figs. 6 and 9) also seems to mimic the rotation of foliation across the shear zone, suggesting that original ductile fabrics may control the orientation of later-stage brittle deformation and serpentinization.
Figure 9. Comparison of all data sets for select stations across the Bogota Peninsula. The lower-hemisphere, equal-area projections are organized by data type (in columns) and station location (rows A–L; see Fig. 3 for map locations). Columns show (1) field measurements; (2) orthopyroxene shape-preferred orientation (SPO); (3) olivine lattice-preferred orientation (LPO); (4) high- and low-field anisotropy of magnetic susceptibility (AMS) measurements; and (5) poles to joints measured in the field. The last row shows details about the different types of data shown on each projection. Occasionally, the different data sets were not collected at the same locations; this situation is indicated by offsets in the different rows (I–L).
<table>
<thead>
<tr>
<th>Field fabrics</th>
<th>Orthopyroxene SPO</th>
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<th>LPO b axes</th>
<th>LPO c axes</th>
<th>High-field AMS</th>
<th>Low-field AMS</th>
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<td><img src="image6" alt="High-field AMS L" /></td>
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- Field foliation
- Field lineation
- Poles to dikes
  $N = \# \text{ of poles to dikes}$

- Field foliation
- Field lineation
- Long axis
- Intermediate axis
- Short axis
- SPO foliation

- Field foliation
- Field lineation
- Contours of % area
- Dotted contour is 1% area
  $N = 102 \text{ for all samples}$

- Confidence ellipse
  - Mean $k_1$
  - Mean $k_2$
  - Mean $k_3$
  - Specimen $k_1$
  - Specimen $k_2$
  - Specimen $k_3$

  $N = \text{ total \# of specimens}$

- Poles to joints

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**Figure 9 (continued).**
MODEL OF SHEAR ZONE EVOLUTION

Using our extensive data set recording different components of strain and deformation across the shear zone, and existing fabric data from the Massif du Sud (Prinzhofer et al., 1980), we constructed a conceptual model for the development of the Bogota Peninsula shear zone over time (Fig. 10). This model is based on the interpretation of Nicolas (1989) illustrated in Figure 2D, where the Massif du Sud represents material formed at a spreading center and the Bogota Peninsula records dextral motion due to transform motion, either reflecting deformation recorded on one plate or across both plates.

Localizing Deformation

For our conceptual model, we must decide the order in which strain was accommodated in the far field, near field, and center of the shear zone. Two models for shear zone evolution are common (Wojtal and Mitra, 1986; Means, 1995; Horzman and Tikoff, 2007): (1) localization, where deformation is first accommodated across a broad region, i.e., the far field in New Caledonia represents the first increment of deformation, but becomes focused in a narrower high-strain zone (Passchier, 1986; West and Hubbard, 1997); or (2) delocalization, where the shear zone width increases, i.e., from the center of the Bogota Peninsula shear zone to the far field, during progressive deformation (e.g., Aoya and Wallis, 2003).

Modern oceanic transform systems are instructive for choosing an appropriate model of shear zone development, and most data support a localizing model. Deformation at ridge-transform intersections, often inferred by the presence of active fault scarps, is typically expressed across a wider region near the intersection and becomes narrower within the transform valley away from the ridge (Fox and Gallo, 1984; Gallo et al., 1986; Kastens et al., 1986; Macdonald et al., 1986). The transform fault also marks a major thermal boundary between oceanic lithosphere of different ages (Parker and Oldenburg, 1973), and many thermal models (Phipps Morgan and Forsyth, 1988; Shen and Forsyth, 1992; Furlong et al., 2001) support the narrowing of the shear zone as the oceanic lithosphere moves away from the mid-ocean ridge (see, however, Behn et al., 2007, where the opposite path is predicted).

The diabase dikes from the Bogota Peninsula shear zone also support a localizing model. These dikes are undeformed and essentially restricted to the center of the shear zone. Thus, the last pulse of magmatism in the region occurred in the shear zone center, suggesting that this was the last active part of the system.

Shear Zone Evolution

Figure 10 shows our simplified, conceptual model for shear zone localization as four discrete time steps, although in reality, the deformation may have varied smoothly in space and time. The top row shows foliation trajectories that are active in a given step (solid lines) and those trajectories that have been frozen (dotted line). The bottom row shows dike orientations and their associated stretches exclusively within the actively deforming portion of the system. We purposely have not labeled this diagram with specific directions in space; instead, we discuss the orientation of the system in the following section.

In the first step, fabrics develop due to seafloor spreading. Based on mapping in the Massif du Sud (Prinzhofer, 1981), the orientations of pyroxenite dikes seem to vary widely, but their orientations were not sorted according to spatial location. Thus, we cannot determine if dike orientations vary systematically in space (e.g., with the foliation trajectories from the Massif du Sud). The few dikes that we observed in the Massif du Sud were typically subhorizontal and only weakly deformed (Fig. 7A), if at all.

In the second step, deformation across a wide shear zone causes steepening and rotation of foliation preserved in the two far-field regions. The fabrics developed in this phase may be slightly stronger than those in Massif du Sud based on the SPO patterns (Fig. 5). Two sets of dikes are commonly observed in the far-field regions. One set is steeply dipping and not quite parallel to foliation. This set shows evidence of boudinage on horizontal surfaces; these dikes also may form the limbs of isoclinal folds, the axial planes of which are subparallel to foliation. The second set of dikes is oblique to the foliation strike, gently dipping, and relatively undeformed. In reality, fabric development and dike injection in the Massif du Sud (first step) and far-field regions (second step) may have occurred at the same time. The steeper orientations in the far field may simply reflect the changing stress field near the ridge-transform intersection (e.g., Allerton, 1989; Gudmundsson, 1995).

In the third step, deformation becomes localized in the narrower shear zone preserved in the near-field regions. Continued deformation in this phase causes foliation to rotate and steepen. One set of pyroxenite dikes is now subparallel with foliation and quite stretched; the separation between boudins is observed on both horizontal and vertical faces. The second set of dikes
is relatively undeformed, steeply dipping, and oblique to foliation.

In the fourth and final step, deformation is concentrated in the 3-km-wide center of the shear zone. Foliation is subvertical, and only one set of subvertical dikes is still observed. These dikes are not parallel to foliation, but they are boudinaged. We infer that the set of dikes that was most parallel to foliation in the previous phase is now absent because the stretches were so large as to make these dikes unrecognizable in the field. The evidence for this comes from small, lozenge-shaped orthopyroxene blebs observed in the field, which were larger than normal orthopyroxene grains distributed in harzburgites (Fig. 4D).

**Orientation of the Shear Zone Boundaries**

The cartoon in Figure 10 implicitly assumes that the spreading ridge, which caused gently dipping foliation in the first step, is perpendicular to the shear zone boundaries, but this is actually difficult to determine from the available data. Recall that the previous interpretation of ophiolite fabrics (Nicolas, 1989) had an E-W–striking ridge responsible for the Massif du Sud and a N-S–striking transform fault, resulting in perpendicularity between the ridge and transform (Fig. 2). However, our more detailed account of fabrics across the shear zone (Fig. 6), and closer examination of the spatial patterns of foliation and lineation in the Massif du Sud suggest that this interpretation of the ridge-transform geometry requires revisions.

The first change we suggest is that the transform fault was striking NNE to NE instead of due N. As noted earlier, field foliation from the high-strain center of the shear zone (with its subhorizontal lineation) strikes 15°, while olivine a axes trend 30°. This magnitude and sense of angular discrepancy between field lineation and LPO are expected in dextral systems (e.g., Darot and Boudier, 1975; Nicolas, 1989). Given any reasonable model for fabric development in a shear zone, and assuming relatively high-strain values in a dominantly simple shear sense of angular discrepancy between field lineation and LPO are expected in dextral systems (e.g., Darot and Boudier, 1975; Nicolas, 1989). Given any reasonable model for fabric development in a shear zone, and assuming relatively high-strain values in a dominantly simple shear history, these data suggest that the shear zone boundary must be oriented 15°–35° east of N. Two independent data sets suggest that a NE-oriented value (30°–35°) is more appropriate. First, numerical modeling of LPO developed in transpression, transtension, and simple shear (Tommasi et al., 1999) shows that olivine a axes are closer to the shear direction than the field lineation (i.e., the long axis of the finite strain ellipsoid). Second, the consistent orientation of diabase dikes from the center of the shear zone (~35°) may be due to strong anisotropy developed in the transform system. Instead of reflecting the local extension direction at the time of intrusion, the dikes may have intruded parallel to the shear zone boundaries. Similar patterns have been observed in the Husavik-Flatey fault, a modern transform fault system in Iceland (Garcia et al., 2002). Therefore, 30°–35° is our best estimate of the shear zone boundaries.

The second revision involves the spreading direction versus ridge segment orientation responsible for fabric development in the Massif du Sud. If spreading were perpendicular to the ridge, the foliation strike and lineation trend

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**Figure 10. Schematic diagram illustrating how the Bogota Peninsula shear zone may have evolved as a localizing shear zone over time.** The data along the top row show foliation relationships (active and fossilized), whereas data along the bottom row show dike orientations and behaviors relative to foliation solely within the actively deforming zone. At t = 1, fabrics in the Massif du Sud were formed due to seafloor spreading, followed by shear zone localization at t = 2 far-field, t = 3 near-field, and t = 4 central domains. Late-stage diabase dikes were intruded in this last stage of deformation.
should also be mutually perpendicular. Even though the stereographic projections for the Massif du Sud, on average, show perpendicularity between these fabrics (Fig. 1), the fabric trajectories demonstrate “left-leaning” lineations relative to foliations (Fig. 11). This pattern suggests that spreading was oblique enough to the ridge segment to cause relative sinistral motion, which any model for the ridge-transform geometry must take into account.

The third revision involves the actual orientation of the ridge segments. This orientation is the most difficult to constrain with the available field data. Our best option is to assume that relative plate motion was parallel to the transform fault, a common observation in modern plate reconstructions. If spreading were oriented ~30°, the ridge segment would need to be clockwise from 120° to result in the sinistral sense of motion for the Massif du Sud. We suggest the most likely orientation is parallel to the 140° trend of New Caledonia and, locally, of the Norfolk Ridge.

This orientation is also consistent with the broad trends in magnetic (Bitoun and Recy, 1982; Lafoy et al., 1996) and gravity data (Collot et al., 1987) from the South Loyalty Basin.

Figure 11B illustrates our revised interpretation of the ridge-transform geometry, now oriented in space and accounting for the non-perpendicularity between ridge and transform segments. We show where the rocks currently exposed on the Bogota Peninsula may have originated, either preserving the true transform fault between two ridge segments, or half of a transform fault preserved on the Massif du Sud plate. These two possibilities for the history of the transform fault are discussed next.

**DISCUSSION**

Our rich and varied data set across the Bogota Peninsula shear zone provides detailed information about fabric evolution within a large-scale mantle shear zone. In this section, we discuss whether the shear zone represents a whole or a half of a transform fault. We then examine how mantle fabrics and their associated mechanical anisotropies may have controlled processes of ophiolite emplacement and later Neogene extension.

**Is the Bogota Peninsula Shear Zone a True Transform Fault?**

The high-temperature olivine LPO patterns support formation of the Bogota Peninsula shear zone within the oceanic lithosphere and not due to later reactivation of a preexisting structure. The stretched pyroxenite dikes, which rotate clockwise with foliation and show larger stretches in the center of the shear zone, suggest that the shear zone records progressive deformation. We used our detailed field data (Fig. 6) and fabric measurements (Figs. 5, 7, 8, and 9) from the Bogota Peninsula shear zone to develop a conceptual model for its evolution.
on 21 September 2019

serves the full transform fault. As noted by where the Bogota peninsula shear zone pre-

have been subducted and could therefore have original ridge-transform geometry is unlikely to

formation. Given this tectonic environment, the transform valley can be between 10 and 50 km wide (e.g., Choukroune et al., 1978; Garfunkel, 1981; Gallo et al., 1986; Parson and Searle, 1986; Lagabrielle et al., 1992). Within this valley, there may be multiple fault strands, or a single fault, that make up the principal transform displacement zone, which can be as narrow as 300 m and up to 1–5 km in width (e.g., Fox and Gallo, 1984; Parson and Searle, 1986; Mamaloukas Frangoulis et al., 1991). Thus, the 3-km-wide, high-strain zone on the Bogota Peninsula, as well as the smaller mylonitic zone in the northwestern far field (Fig. 3), can be thought of as the mantle continuations of the fault systems that would be observed at the surface within the transform valley.

Reconstructions of the tectonic development of the southwest Pacific may also provide insight on interpretation of the transform fault. Schellart et al. (2006) suggested that the South Loyalty Basin, where the ophiolite sheet likely formed, was once 750 km wide, compared to its present width of <100 km. The symmetry of seafloor spreading requires all ridge segments in the middle of the basin to be subducted with this amount of shortening. Thus, these features could not be preserved within the South Loyalty Basin and on New Caledonia because of the geometry required for a true transform fault (see Fig. 2). However, Whattam et al. (2008) suggested that the proto-ophiolite lithosphere formed in the forearc above a NE-dipping subducting slab east of New Caledonia. These authors argued that the ophiolite is much younger than its commonly assumed Late Cretaceous age, forming instead in the Eocene and emplaced not long after its formation. Given this tectonic environment, the original ridge-transform geometry is unlikely to have been subducted and could therefore have been emplaced directly onto New Caledonia.

Based on the available field evidence and tectonic reconstructions, we favor the model where the Bogota peninsula shear zone preserves the full transform fault. As noted by Whattam et al. (2008), ophiolite emplacement tends to occur shortly after formation instead of 30–50 m.y. later. When ophiolites are more than 10 m.y. old, they become difficult to obduct because their lithosphere is too thick (Dewey, 2003). Other suprasubduction zone ophiolites seem to share the pattern of obduction shortly after formation, including the nearby Papuan ophiolite (Lus et al., 2004) and the Semail ophiolite in Oman (Hacker et al., 1996). Further, the Troodos (Varga and Moores, 1985) and Oman ophiolites (Ceuleneer et al., 1988; MacLeod and Rothery, 1992; Le Mée et al., 2004) both preserve former spreading ridges within the sheet. Thus, the data and obduction models from other ophiolites suggest that the Bogota Peninsula shear zone represents a true transform fault.

Fabric Controls on Ophiolite Emplacement and Neogene Extension

Because joint orientations across the Bogota Peninsula shear zone rotate with the changing orientation of foliation (Figs. 6 and 9), we suggest that mechanical anisotropies from the original mantle fabrics controlled late-stage brittle deformation of these rocks. In this section, we propose that this reactivation may be true on a much larger scale as well, where inherited mantle fabrics controlled, at least in part, ophiolite obduction and much later Neogene extension. We also return to the interpretation of the Belep shear zone, which deforms several klippen in northwest New Caledonia.

Time Line of Deformation

Our proposed tectonic model is summarized in Figure 12, which shows a simplified map-view history of the ophiolite. This model is based primarily on fabrics and structures in New Caledonia and does not explicitly rely on a particular tectonic reconstruction for the region (e.g., Aitchison et al., 1995; Cluzel et al., 2001; Crawford et al., 2003; Sdrolias et al., 2003; Schellart et al., 2006), since these models differ with regard to the location where the ophiolite formed, the age of the ophiolite, the direction of subduction, and the original size of the South Loyalty Basin.

In the first panel (Fig. 12A), the material that would become the New Caledonia ophiolite formed in a ridge-ridge-transform environment. As discussed previously, we suggest that the spreading ridge segments were more or less parallel with the Norfolk Ridge (Fig. 11). The transform fault orientation forms a slightly acute orientation of foliation (Figs. 6 and 9), we suggest that mechanical anisotropies from the original mantle fabrics controlled late-stage brittle deformation of these rocks. In this section, we propose that this reactivation may be true on a much larger scale as well, where inherited mantle fabrics controlled, at least in part, ophiolite obduction and much later Neogene extension. We also return to the interpretation of the Belep shear zone, which deforms several klippen in northwest New Caledonia.

The shear zone either recorded all of the transform deformation (left column) or half of the deformation preserved on one side of a fracture zone (right column). In the next panel (Fig. 12B), plate motion became convergent in the Eocene, and the ophiolite was obducted. We have shown the ophiolite sheet extending over all of New Caledonia based on exposures in the Massif du Sud, ophiolite klippen, and small remnants preserved in the Pouebo terrane. Some models suggest that the emplacement process was diachronous, beginning in the northern part of the island and continuing southward (e.g., Cluzel et al., 2001; Baldwin et al., 2007). We have no fabric or structural evidence to support or refute this idea, and as such we have simply shown the position of the ophiolite after obduction was complete. Because the ophiolite sheet drapes into the South Loyalty Basin (Collot et al., 1987), it probably did not experience significant rotation during emplacement.

After emplacement, we suggest that a feature from the non-Massif du Sud spreading ridge was reactivated as a dextral strike-slip or transpressional shear zone (Fig. 12C). This feature could have been the spreading ridge itself or a spreading ridge–parallel structure such as a normal fault or extensional core complex, which are both observed in other ophiolites (Varga and Moores, 1985; Cann et al., 2001) and in modern mid-ocean-ridge environments (e.g., Blackman et al., 1998; Idefonse et al., 2007). In fact, this structure may have been the original discontinuity that initiated ophiolite emplacement.

Reactivation of an inherited structure may explain two structural features on the island. First, the high-pressure Pouebo terrane experienced shortening before its exhumation (Rawling and Lister, 2002), causing folding that includes ophiolitic material. These folds hinges are asymptotic toward the present-day position of the western Caledonia fault zone (Fig. 12C). Second, reactivation might account for the formation of mylonites in the Belep shear zone, which is also along strike with the western Caledonia fault zone. Dextral motion is consistent with both features, except for the one sinistral shear sense indicator from the Belep Islands (Fig. 2A). Instead of using the antithetic shear sense to place the Belep Islands in the outer corner of a transform fault (Fig. 2B), perhaps the Belep shear zone reflects reactivation of material during or after ophiolite emplacement. In the top diagram in Figure 12C, note the alignment of the ophiolite klippen in the shear zone, which is different from their modern position (Fig. 12E). This arrangement allows the western
Fabric development in the mantle section of a paleotransform fault, New Caledonia

**Figure 12.** Cartoon illustrating how (A) fabric patterns formed within the ophiolite may have influenced both (B) ophiolite emplacement and (C–D) later deformation. (E) The modern setting. See text for details. The klippen deformed by the Belep shear zone in C are abbreviated Pott (Pt), Art (A), Yandé (Y), Poum (Pm), and Tiebaghi (T). Fold hinge orientations in C and normal faults in D within the Pouebo terrane are from Rawling and Lister (2002). HP—high pressure; WCFZ—western Caledonia fault zone.
across the island. This later phase of extension affected rocks not. This may account for the apparent bend in fault, relative to the Belep Islands, which were way (Titus et al., 2002). This motion accounts for the misalignment of the ophiolite klippen, which were displaced along the detachment fault, relative to the Belep Islands, which were not. This may account for the apparent bend in the Belep shear zone illustrated in Figure 2A. This later phase of extension affected rocks across the island.

Implications for the Tectonic Development of New Caledonia

This proposed reconstruction of ophiolite formation, obduction, and Neogene extension ties together four additional features about the ophiolite nappe, shear zones, and younger geologic structures on New Caledonia.

First, our model may explain compositional differences within the ophiolite sheet. The Massif du Sud is primarily harzburgitic, whereas the ophiolite klippen include spinel- and plagioclase-bearing lherzolites (Mouette, 1982; Ulrich et al., 2010). If the Bogota Peninsula shear zone represents a dextral transform fault, separating two island-parallel ridge segments (Fig. 12A), the rocks that become the Massif du Sud and klippen would have formed at two different spreading centers, thereby explaining their compositional differences.

Second, the style of ophiolite exposure across the island may be linked to the original oceanic plate-boundary geometry. If the Massif du Sud formed in an inner corner, the rocks within it would be older and presumably thicker than rocks northwest of the Bogota Peninsula shear zone, which became the klippen. Given its age and thickness, the Massif du Sud could have been obducted more easily as a continuous sheet.

Third, our reinterpretation of the Belep shear zone, not as an original transform fault but as a reactivated ridge-parallel structure, helps to explain its orientation relative to other ophiolitic structures. The current strike of the Belep shear zone is 60°–80° oblique to our revised estimate for the strike of the Bogota Peninsula shear zone. It is difficult to imagine how both shear zones could be transform faults from the same oceanic basin unless the klippen (or Massif du Sud) experienced major rotation during emplacement. However, fabrics in the klippen away from the Belep shear zone are parallel to those in the Massif du Sud (Fig. 1), making significant rotation during emplacement unlikely. Thus, reactivation of a spreading ridge-parallel feature may explain why the Belep shear zone is nearly perpendicular to the Bogota Peninsula shear zone and parallel to our inferred spreading ridge orientation.

Last, the ridge-transform geometry from our model (Fig. 12A) provides anisotropies within the ophiolite nappe that could have been exploited during obduction and Neogene extension. The Bogota Peninsula shear zone could have acted as a tear fault during obduction, allowing the continuous portion of the Massif du Sud to be emplaced with relative ease. The sharp western margin of the Massif du Sud, which is aligned with the southward continuation of the Bogota Peninsula shear zone, supports this suggestion (see Fig. 12E). Both the Belep shear zone and Neogene extensional structures, such as the western Caledonian fault zone, may have exploited the ridge-parallel anisotropy.

CONCLUSIONS

The Bogota Peninsula shear zone provides an unparalleled record of fabric development across the mantle section of a transform fault. Field foliations, lineations, and pyroxenite dikes show systematic rotation across a 50-km-wide region. Folded dikes are observed on the margins of the deforming zone, and boudinage of dikes increases toward a central 3-km-wide mylonite zone. Olivine LPO patterns demonstrate that a' axes are slightly clockwise of field lineations within the high-strain core, consistent with dextral motion along the shear zone. The shear zone is interpreted as recording the deformation on both sides of a transform fault. The Massif du Sud is attributed to rocks forming in an oceanic inner corner adjacent to a transform fault.

The strong material anisotropy from the inherited ridge-ridge-transform geometry may have played an important role in the obduction of the ophiolite: The initiation of obduction may have been facilitated by a former ridge segment (or ridge-parallel fabric), and the Bogota Peninsula shear zone may have acted as a tear fault during emplacement. The inherited material anisotropy also may have influenced subsequent Neogene extension, allowing a spreading ridge-parallel feature to be reactivated twice: first as a dextral shear zone linked to the formation of folds in the high-pressure Pouebo terrane and mylonitization in the Belep shear zone, and second as a detachment fault that facilitated the movement of ophiolite klippen into their present position along the northern portion of New Caledonia. Thus, inherited mantle fabrics may explain the present distribution of ophiolitic material across New Caledonia and provide an explanation for a secondary shear zone within the klippen, the orientation of which is nearly perpendicular to the Bogota Peninsula shear zone.

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REFERENCES CITED


