The Internal Structure of an Oceanic Core Complex: An Integrated Analysis of Oriented Borehole Imagery from IODP Hole U1309D (Atlantis Massif)

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[1] Oceanic core complexes at slow-spreading ridges represent the uplifted footwalls of large-offset ‘detachment’ faults that initiate at steep dips, and rotate to flatten via a ‘rolling hinge’ mechanism in response to flexural unloading. A key question is whether oceanic core complex development is accommodated entirely by displacement on the detachment fault zone, or if significant internal deformation of the footwall occurs during flexure and rotation. We investigate this by constraining the internal architecture of the Atlantis Massif oceanic core complex (Mid-Atlantic Ridge, 30°N) using Formation MicroScanner borehole wall images and cores from the 1416 m-deep Integrated Ocean Drilling Program Hole U1309D. Two distinct sets of structures are observed. N-S-striking, E-dipping structures dominating the upper 385 m are interpreted as a brittle to semi-brittle zone of fracturing in the footwall. Structures with this geometry occur down to 750 m below seafloor, suggesting that the detachment damage zone extends deep into the footwall. The nature of this deformation is, however, enigmatic: several cataclastic shear zones with reverse geometry in their current orientations may be rotated extensional faults or relate to shortening at the base of the flexing beam of a very weak footwall. By contrast, E-W-striking, N- and S-dipping structures dominate the lowermost kilometer of the borehole. They likely represent conjugate fractures formed in the hanging wall of a late, E-W normal fault zone, separating gabbroic rocks of the central dome of Atlantis Massif from serpentinized peridotite to the south, responsible for post-detachment uplift of the southern margin of the massif.

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1. Introduction

[2] Understanding the mechanisms of denudation of oceanic core complexes (OCCs) at intermediate to ultra-slow spreading mid-ocean ridge segments is important because they are responsible for creating significant regions (locally up to 60%) of new seafloor [Smith et al., 2006, 2008; Schroeder et al., 2007]. OCCs develop when plate divergence is accommodated by large-slip normal faults during protracted periods (1–4 Myr) of reduced magma supply at the ridge axis. The lower oceanic crust and in some cases the upper mantle is denuded to the seafloor via large-offset (km-scale) normal faults [e.g., John and Cheadle, 2010]. These so-called oceanic detachment faults create OCCs characterized by broad, domal bathymetric highs with spreading-parallel corrugations and smaller-scale striations on the detachment fault slip-surfaces. Similar structures, known as metamorphic core complexes, are found associated with extreme extension in continental settings [John, 1987; Davis and Lister, 1988], including rifted continental margins [Froitzheim and Manatschal, 1996; John and Cheadle, 2010]. Sea-floor mapping using bathymetric, side scan sonar and backscatter techniques has now documented numerous OCCs on both the Mid-Atlantic Ridge (MAR) and South West Indian Ridge (SWIR) [Cann et al., 1997; Blackman et al., 1998, 2002, 2006, 2011; Tucholke et al., 1998; Karson, 1999; MacLeod et al., 2002, 2009; Escartin et al., 2003, 2008; Cannat et al., 2006], as well as the Philippine Sea [Ohara et al., 2001; Tani et al., 2011]. When active, oceanic detachment faults may constitute the plate boundary [deMartin et al., 2007; Schroeder and John, 2004; Smith et al., 2006; Tucholke et al., 1998], representing a new class of seafloor spreading [Escartin and Canales, 2011].

[3] Atlantis Massif is an OCC exposed on the western flank of the MAR at the intersection with the Atlantis Transform Fault at 30°N (Figure 1a). The central dome of the massif was drilled and logged during Integrated Ocean Drilling Program (IODP) Expeditions 304 and 305, which ran back-to-back from November 2004 to March 2005 [Blackman et al., 2006]. The main goal of the two expeditions was to document the structural and lithologic characteristics of the OCC footwall, leading to an improved understanding of the processes that control OCC formation and subsequent evolution.

[4] In this paper we present an analysis of wireline logging data from IODP Hole U1309D sited on the central dome of Atlantis Massif, focusing specifically on data from the Formation MicroScanner (FMS) tool. We integrate logging data sets with those derived from observations on recovered drill core to provide insights into the structural architecture of the OCC footwall. Specifically, we address the question of the extent to which OCC development is controlled entirely by displacement on a discrete detachment surface/fault zone (exposed at the seafloor), or whether important detachment-related structures are present at depth that partition strain within the footwall: i.e., how does the footwall of an OCC respond internally to the stresses imparted upon it by flexure due to unloading of the hanging wall? Using logging data to investigate these issues has two distinct benefits: (i) these data uniquely provide the true spatial orientation of structural features intersecting the borehole, complementing azimuthally unoriented observations made on core pieces; and (ii) logging data are continuous, whereas hard-rock ocean drilling produces a discontinuous record due to variations in core recovery. We combine our structural data with direct seafloor observations, and discuss these results in terms of the interplay between detachment-related structures and those likely to relate to post-detachment tectonic uplift of the southern margin of Atlantis Massif.

2. Geology of Atlantis Massif and IODP Hole U1309D

[5] Atlantis Massif is located near 30°N at the intersection of the slow-spreading MAR (full-spreading rate ∼23.6 mm/yr, NUVEL-1 model [DeMets et al., 1990]) and the Atlantis Transform Fault (Figure 1a). The massif is inferred to represent an OCC exposed via long-lived detachment faulting [Cann et al., 1997; Blackman et al., 1998, 2002, 2006]. Corrugated striations are seen on the central portion of the massif in the bathymetric data.
Blackman et al., 1998, 2002] and the detachment fault itself is exposed over an 8–10 km wide by 15 km long area that forms the central, elongate doubly plunging domal seafloor topography (Figure 1a). The seafloor dips ~5°W away from the ridge-axis on the western side of the dome, and then curves to dip gently ~12°E toward the ridge-axis before disappearing beneath an eastern block composed of basalt [Blackman et al., 2006]. The Southern Ridge of Atlantis Massif (Figure 1a) is almost 1 km higher than the central dome region and deformation zones observed at the top of the exposed southern flank during submersible surveys provide evidence of a long-lived normal fault [Schroeder and John, 2004; Karson et al., 2006]. Dredged samples from the Southern Ridge include gabbroic rocks and serpentinized peridotite, and on its southern wall just below the apex is the active, serpentinite-hosted, Lost City hydrothermal vent field [Kelley et al., 2001; Kelley, 2005; Früh-Green et al., 2003].

[6] Eight holes were drilled at Atlantis Massif during IODP Expedition 304/305 [Blackman et al., 2006]. IODP Hole U1309D, located ~15 km west of the axial valley on the central dome of Atlantis Massif where the seafloor coincides with the detachment fault surface (Figure 1a), reached a maximum depth of 1415.5 m below sea floor (mbsf). In addition to the remarkable depth of penetration, an exceptionally high average core recovery of 74% was also achieved. The recovered section, shown graphically in Figure 1b, is dominated by gabbroic rocks (91%) that span a wide range of modal compositions (gabbro, gabbronorite and olivine gabbro, plus minor Fe-Ti rich oxide gabbro) and are complexly interlayered on scales of centimeters to hundreds of meters [Blackman et al., 2006; Ildefonse et al., 2007; John et al., 2009]. Minor ultramafic rocks comprise 5% of the remaining recovered core, basalt/diabase 3% and felsic dykes 1%.

[7] Based on sea-surface magnetic anomalies, Atlantis Massif was initially interpreted as having formed via detachment faulting over the past ~2 Ma [Zervas et al., 1995]. This was further constrained by near-bottom magnetic measurements that recorded a small positive magnetic anomaly 1 km to the west of the massif, interpreted as Anomaly C2n (1.77–1.95 Ma).
of the GeoFrame software package produced by Schlumberger Ltd. The fully processed data are available for download from the IODP-USIO log database (http://brg.ldeo.columbia.edu/logdb/).

[11] Processed data were uploaded into GeoFrame at the Borehole Research Facility, University of Leicester, UK, and two image normalizations applied within the BorNor module of GeoFrame. For the static normalization, the resistivity range of the entire depth interval is calculated and translated into 256 color levels. This highlights absolute variations in conductivity and permits distinct areas hundreds of meters apart to be compared and characterized. In contrast, the dynamic normalization assigns the full color range within 2 m sliding windows, thus enhancing fine-scale, subtle resistivity contrasts. The results are fully oriented, unwrapped images of the borehole wall. On a single pass uphole, the FMS pads image 22% of the formation wall for a nominal borehole diameter of 26 cm or 10.25 inches (individual FMS pads are 45 mm wide). If subsequent passes do not follow the same uphole path, a larger surface area is imaged; however, this was only the case for 10% of IODP Hole U1309D.

[12] Inclined planar features, such as faults, fractures or veins that intersect the borehole are represented as sinusoidal traces on the unwrapped FMS images (Figure 2a). The orientation of an individual feature is found by generating a best fitting sinusoid through GeoFrame. The dip of a feature is equal to the arctangent of the sinusoid amplitude \( h \) divided by the borehole diameter \( d \), which is calculated from the FMS calipers. Since the images are oriented with respect to magnetic north using the magnetometer on the GPIT instrument, the true dip direction of a feature is the orientation of the sinusoid minimum.

4. Results and Discussion

4.1. Data Quality and Potential Sources of Error

[13] From our systematic investigation of FMS images of the borehole wall of IODP Hole U1309D between 97 and 1415 mbsf we have identified 4324 individual planar conductive features, and made quantitative measurements of their orientations using GeoFrame. It is important to recognize that there is an element of subjectivity to the identification of features, for two principal reasons: (1) good borehole wall contact is not always made by all four pads, and (2) the toolstring is not always in contact with the borehole wall. The dip of a feature is equal to the arctangent of the sinusoid amplitude \( h \) divided by the borehole diameter \( d \), which is calculated from the FMS calipers. Since the images are oriented with respect to magnetic north using the magnetometer on the GPIT instrument, the true dip direction of a feature is the orientation of the sinusoid minimum.
pads, and therefore features may not necessarily be visible on all four resistivity image traces; and (2) because the borehole wall is not completely imaged, in intervals of the borehole with multiple or intersecting features there may be a number of different ways of fitting individual planes to the observed conductive traces. In order to provide a measure of confidence we have therefore subdivided the data set into three sub-groups based on the quality and robustness of each sinusoidal fit, following the definition of MacLeod et al. [1995] (Figure 2b): Quality Group 1 = trace visible across all pads, with plane defined unambiguously ($N = 1330$); Quality Group 2 = trace visible on 3 pads, with orientation relatively well constrained ($N = 1681$); Quality Group 3 = trace visible on 1–2 pads with some potential for ambiguity in the orientation of the plane ($N = 1313$). Additional better quality categories for features that could be traced across 5 to 8 pads were not employed as there

Figure 2. (a) An inclined planar feature (such as a fault or vein) intersecting the borehole can be represented by a sinusoidal trace on an unwrapped FMS image. The orientation of the feature is the direction of the sinusoid minimum, and the dip is calculated as the arctangent of the sinusoid amplitude, $h$, divided by the borehole diameter, $d$. (b) Unwrapped FMS image over an interval illustrating sinusoidal traces with different qualities of fit following criteria in MacLeod et al. [1995]. Quality Group 1 is shown in blue and has the most robust fit, Quality Group 2 is represented in green and Quality Group 3, with the least confident fit, is shown in red. (c) Equal area stereographic projections showing the poles-to-planes of features categorized as Quality 1, 2 and 3 (same colors as above). Kamb contours have been used to highlight trends in the data.
are only 37 narrow (3.6 m average vertical thickness) intervals where repeat FMS passes followed different uphole tracks. Of the 433 features (10% of the total data set) identified in these intervals, those with traces visible on 5 to 8 pads were included in Quality Group 1.

[14] Figure 2c shows the poles-to-planes of all features seen in the FMS imagery plotted on equal area stereographic projections according to their data quality. Although all three sub-groups display similar overall distributions, the lowest quality category exhibits a larger degree of scatter, evidently because features traced across only 1–2 FMS tracks have the least precise spatial control. For this reason we do not utilize these data (Quality Group 3) further in this paper.

[15] The orientations of the remaining data are plotted in Figure 3a. It is evident that significantly smaller numbers of features are sub-horizontal or sub-vertical than with intermediate dips. Martel [1999] demonstrated that a bias is inherent in borehole fracture data sets due to the lower probability of intersecting a fracture that is dipping parallel to the borehole axis than one that is perpendicular. Similarly, for identical fracture populations with uniform spacing, a larger depth interval is required to intersect the same representative number of fractures if they dip at an angle closer to the borehole axis. To illustrate these effects, we generated a synthetic, randomly distributed set of poles-to-planes (resulting in a near uniform distribution of dips) and applied a probability density function to replicate the effects of borehole bias using the methodology in Martel [1999]. In the resulting synthetic data (Figure 3b), steeply dipping features are under-represented as a result of borehole bias. The decrease in frequency with increasing dip of planar features determined from FMS imagery of IODP Hole U1309D (histogram of Figure 3a) may at least partly result from the same effect.

[16] An additional factor that can influence the proportion of steeply dipping (lower quality sub-group) data acquired using FMS is that steeply dipping features are inherently more difficult to identify with confidence on FMS images. When features dip at angles greater than 80°, the associated sinusoidal traces have peak-to-peak amplitudes in excess of 2 m vertical height. This requires viewing the FMS images on a coarser scale to fit the appropriate depth interval into a single computer screen-shot. Furthermore, on dynamically normalized images, a trace corresponding to a steeply dipping feature can be represented by a variety of colors due to the 2 m sliding window dynamic normalization technique.

[17] Overall, although it is likely that the low numbers of steeply dipping planar features observed in IODP Hole U1309D (Figure 3a) is partly an artifact, it is impossible to quantitatively assess the extent of under-sampling of the true population [e.g., Terzaghi, 1965]. For this reason, we have not attempted to apply corrections to compensate for any borehole bias effect. We note from Figure 3a that in IODP Hole U1309D there are also relatively low numbers of sub-horizontal FMS features, which should be accurately sampled without significant bias (cf. frequency of features dipping between 0 and 10° in the real and synthetic data sets in Figure 3). This suggests an original non-uniform distribution of features.

[18] In addition to errors arising from the sinusoidal fitting process, errors in FMS image orientation might potentially arise from local magnetic anomalies where highly magnetic formations intersect the borehole. The most likely rock types encountered in IODP Hole U1309D that could give rise to such orientation errors are Fe-Ti-rich oxide gabbros and strongly serpentinized olivine-rich rocks, but the precise location of all occurrences of these more magnetic rock types cannot be determined from the core due to variations in recovery. As a precaution, therefore, we have used data from the GPIT magnetometer to locate regions in the borehole where the magnetic field varies by more than two standard deviations from the average in the hole. A total of 216 FMS features occur in such zones and could potentially be affected by orientation errors. These are eliminated from further analysis, leaving a total of 2795 planar features that we interpret and use in this synthesis.

4.2. Downhole Variation in the Orientation of FMS Features

[19] The overall distribution of orientations of features on electrical borehole wall images from IODP Hole U1309D (Figure 3a) in itself gives us little insight into the structural evolution of the Atlantis Massif OCC. To obtain greater insight we examine the downhole variation in orientation of FMS features and, where possible, attempt to correlate them with distinctive structural features identified in the cores.

[20] Our initial approach was to look for natural subdivisions of orientations downhole within the FMS data set without a priori reference to structural studies of the recovered core, because of the severe potential for bias in the latter due to the non-recovery of faulted material. The 2795 FMS features were
initially subdivided into 100 m depth bins and plotted on stereographic projections to highlight any clustering of FMS orientations. Where a bi-modal or scattered distribution was observed, the broad 100 m bin was split into finer 50 m, 25 m and 10 m bins as necessary. Six principal domains were subsequently defined by amalgamating depth intervals with characteristic orientations of FMS structures. Finally, we modified the true downhole depths (rather than curated core depths) of the upper and lower boundaries of each domain slightly in a couple of cases to coincide with what were clearly equivalent structural boundaries identified in the core. Some of our structural domains defined from the FMS features therefore coincide with those defined from the core [Blackman et al., 2006, 2011], but not all. Each of the six domains represents a substantial thickness (minimum = 167 m; maximum = 290 m), and is defined by a large number of FMS features ($N_{\text{min}} = 232; N_{\text{max}} = 682$).

The principal observations and characteristics associated with each domain are summarized in Table 1 and presented graphically in two formats: equal area stereographic projections that show the true geographic orientation of the poles to FMS features (Figure 4) and downhole plots of the density distribution of FMS features dipping toward specific quadrants (Figure 5). The most striking observations are the large number of E-W-striking (both N- and S-dipping) features present through much of the hole, and a paucity of W-dipping structures throughout the entire drilled section (Figures 4 and 5). There is also a distinct contrast between Domain I and Domains II–VI, with the uppermost domain being dominated by generally N-S striking, moderately E-dipping features. Planes of this orientation are also present deeper in the hole but are far less frequent (<1 per meter). Domains II–VI are instead dominated by generally E-W-striking features dipping either N or S.
4.3. Geologic Significance of FMS Features

[22] The planar features seen in the FMS images are narrow zones of high resistivity contrast most likely caused by fractures filled with conductive mineral veins, zones of cataclasism and/or seawater. Whereas crystal-plastic (oxide-) gabbro shear zones have been successfully identified on FMS images from other boreholes in oceanic gabbros [e.g., Haggas et al., 2005] such shear zones are uncommon in IODP Hole U1309D [Blackman et al., 2006, 2011]; hence such structures are unlikely to be significant in our data set. Here we explore whether comparing core-based observations of cataclasite, vein and fracture distributions with the FMS feature orientations across the six domains outlined above can allow us to correlate geometric information with core-based structural observations.

4.3.1. Distribution of Vein Types

[23] Veins observed in the recovered core by Expedition 304/305 shipboard scientists are grouped according to their metamorphic grade (Figure 6), with corresponding significance in terms of their temperature of formation, relative timing, and rheology of their host rocks [Blackman et al., 2006, 2011]. Vein types vary in their distribution downhole, with the most prominent feature being the concentration of greenschist facies veins and those associated with alteration of ultramafic protoliths in Domain I (upper 400 m of the hole), which also has the highest overall number of veins. Amphibolite facies veins are sparse throughout the section and essentially absent below 800 mbsf; however, they are significant in core from Domain III upwards. Greenschist and sub-greenschist facies veins both reappear in Domains IV through VI, with tremolite veins in peridotite documented around the boundary between Domains VI and V.

[24] Because these variations downhole are broad in scale it is not possible to uniquely correlate particular vein types with specific domains (Figure 6). In addition, the vein groups cannot be distinguished in terms of their dip in the core reference frame [Blackman et al., 2006]. However, further insight into the geological nature and origin of vein facies observed in the FMS data potentially may be achieved by comparison with features throughout the section and essentially absent below 800 mbsf; however, they are significant in core from Domain III upwards. Greenschist and sub-greenschist facies veins both reappear in Domains IV through VI, with tremolite veins in peridotite documented around the boundary between Domains VI and V.

4.3.2. First-Order Reorientation of Core Features

[25] The direct correlation of observed vein types to any specific subset of FMS features on the basis of dip is clearly ambiguous. Similarly, observed brittle fractures have a broad range of dips in the core reference frame [Blackman et al., 2006]. However, further insight into the geological nature and origin of features observed in the FMS data potentially may be achieved by comparison with features...
observed in the recovered core and documented in their true geographic orientation.

Azimuthal orientations of core-based observations were initially unconstrained since individual core pieces are free to rotate inside the core barrel during drilling. Reorientation of core pieces to a geographic reference frame may potentially be achieved either by aligning magnetic remanence directions with an assumed reference direction, or by unambiguously matching individual core structures to their direct representations on oriented borehole wall images [MacLeod et al., 1992, 1994]. Morris et al. [2009] successfully applied the latter approach to reorient 34 core pieces distributed within the upper 400 m of IODP Hole U1309D to their original in situ orientations. Morris et al.’s [2009] analyses were restricted to this depth interval because of the more limited number of paleomagnetic and core structural data available below this depth. The reoriented pieces from the upper 400 m contain an insufficient number of veins and/or fractures to enable a statistical comparison to be made to the FMS data set. However, the restoration (Figure 7) did allow Morris et al. [2009] to calculate a well-defined mean in situ magnetization direction of the upper 400 m of IODP Hole U1309D (declination = 226.1°, inclination = −38.1°, α95 = 4.4°, N = 62). Tectonic rotation analysis based on the mean inclination of magnetic remanences below 400 mbsf [Morris et al., 2009] indicates an identical rotation angle and restored declination to that of the upper 400 m. Hence, the mean reoriented remanence direction reported by Morris et al. [2009] provides a

**Figure 4.** Six domains created by grouping intervals with similar orientation characteristics. Data are presented as poles-to-planes on equal area stereographic projections in the true geographical coordinate system. The true depth intervals for each domain are noted, along with the number of FMS features observed within the interval, N. Data from Quality Group 3 and data collected in the vicinity of highly magnetic rock types are not shown (see Section 4.1 for detailed explanation of data reduction). The strike and dip of the dominant structural orientation (and subsidiary structural orientation if appropriate) are determined from the peak cluster in shaded Kamb contours, where strike is specified 90° counter-clockwise from dip direction.
valid estimate of that of the entire hole, and a first-order reorientation of many more core pieces becomes possible by simply rotating piece-average magnetizations to the mean declination. Accordingly, we identified a subset of core features that occur in pieces with well-constrained magnetization directions (CSD ≤ 20°), and calculated reorientation angles required to align piece-average declinations in core coordinates to the hole-average declination (226.1°) in true geographic coordinates. This allows sufficient numbers of core features to be restored to their approximate in situ orientations (Figure 8) to facilitate a direct comparison with the FMS data. It must be noted, however, that this approach only provides a first-order reorientation as it assumes that individual piece-average magnetizations align precisely with the hole-average direction. In fact, the magnetization data of Morris et al. [2009] used to calculate the mean direction have considerable scatter (see lower left-hand stereographic projection of Figure 7), and this cannot be taken into account in any simple reorientation.

[27] The resulting reoriented core structures are subdivided based on type, including fractures, zones of cataclasite and veins (Figure 8). All three groups show broadly similar orientation distributions, and despite the uncertainties in reorientation discussed above, all three groups show a lack of N-S striking, W-dipping features (225°–315°) in purple and N-dipping (315°–045°) in green. The boundaries between the six principal domains (I–VI) illustrated in Figure 4 are shown by the solid black lines. No FMS data are available above the dashed line.

[28] The resulting reoriented core structures are subdivided based on type, including fractures, zones of cataclasite and veins (Figure 8). All three groups show broadly similar orientation distributions, and despite the uncertainties in reorientation discussed above, all three groups show a lack of N-S striking, W-dipping structures and overall distribution very closely comparable to that of the FMS data (Figure 3a). Noteworthy is the weak E-dipping preferred distribution of reoriented zones of cataclasite, found predominantly in the upper 400 m of the section, which is consistent with the N-S striking, E-dipping distribution of features seen in Domain I. However, as all three categories of reoriented core structures are broadly consistent with the orientation distributions determined by analysis of FMS images it is again not possible to uniquely assign FMS
features to any specific category of structure observed in the recovered core.

4.3.3. Direct Reorientation of Major Core Structures

[29] Structural logging of the recovered core from IODP Hole U1309D revealed several discrete intervals with significant thicknesses of cataclasite, interpreted as fault zones [Blackman et al., 2006, 2011; John et al., 2009]. The most significant of these, located between approximately 742 and 762 mbsf, is an amphibolite facies ultracataclasite studied in detail by Michibayashi et al. [2008]. A second interval between ~159 and 174 mbsf, hosts the greenschist facies cataclasite investigated by Hirose and Hayman [2008]. We have focused on these zones, plus two others present in the core at ~695 mbsf and ~785 mbsf [Blackman et al., 2006, Figures F189A and F189D]. In all four zones, we attempted to reorient individual core pieces containing fault rocks to geographical coordinates directly by matching the inclined core structures to their direct representations on oriented borehole wall images [MacLeod et al., 1992, 1994]. In each case there is typically 1–2 m of non-recovered material from each core [John et al., 2009] and hence a vertical window of uncertainty of that order in which to match each core feature to its most appropriate log representation. In all but the 785 mbsf case, however, we can relate cores and logs with reasonable confidence. Figure 9 highlights our best estimate of the geographic orientation of the high-strain cataclastic shear zones, together with all other FMS features from a window of approximately 5–10 m vertically either side of the fault zone. For the zone of cataclasite at 785 mbsf [Blackman et al., 2006, Figure F189D], the dip of the structure is very low (10° relative to the axis of the core) and FMS features of similar dip magnitude in the permissible depth interval have true geographic dip directions ranging from N through NE to E (Figure 9), and we have no additional criteria by which to favor any one over another. In this example, therefore, we cannot reliably reorient this particular structure.

[30] For the other three zones of thick cataclasite, we conclude they each strike roughly N-S and dip ~20°–50°E (Figure 9). Michibayashi et al. [2008] inferred the same orientation for the zone of fracture and cataclasite at 742–762 mbsf. Significantly, kinematic indicators for these shear zones all display clear, apparent reverse shear sense in their present-

Figure 6. Downhole plots of the distribution of vein type and dip magnitude (observations from the recovered core). Black = all vein types (including indeterminate veins, N = 230); red = amphibolite facies (hornblende/dark amphibole veins); purple = greenschist facies (green/brown/yellow actinolite veins often with secondary plagioclase halos, and late actinolite veins ± chlorite, albite and sphene); green = sub-greenschist facies (late veins/fractures with clay minerals, hydroxides or zeolites, carbonate veins and quartz veins); and blue = veins associated with the alteration of ultramafic protoliths (tremolite veins in peridotite, talc veins ± amphibole, chrysotile talc/serpentine veins and composite talc/tremolite veins). The boundaries between the six principal domains (I–VI) illustrated in Figure 4 are shown by the solid black lines, where no FMS data are available above the dashed line. Detailed vein group descriptions are provided by Blackman et al. [2006].
day orientations, i.e., the E sides have moved up [Blackman et al., 2006, Figure F189; Michibayashi et al., 2008, Figure 2]. Potential geologic implications of amphibolite to greenschist facies cataclastic shear zones with these kinematic indicators are discussed below.

5. Discussion

5.1. Detachment-Related Structures

[31] Domain I in IODP Hole U1309D (95–385 mbsf) is distinct from all deeper domains in its high concentration of N-S striking, moderately E-dipping features (Figure 4). These structures are interpreted as detachment-related because of their ridge-parallel strike. They are, however, distinct in dip in comparison to the detachment fault itself: their mean dip of approximately 30°E is significantly steeper than the zero dip of the detachment surface at the location of IODP Hole U1309D.

[32] Although only limited data are available (N = 22), support for the detachment-related nature of the deformation in Domain I comes from the orientation of lineations on planar fracture and vein surfaces (Figure 10) in core pieces from the upper 300 m of the drilled section of IODP Hole U1309D. After reorientation of these lineation directions back to the true geographic reference frame, using the methodology described in section 4.3.2 above, an eastward plunging clustered distribution is resolved (mean azimuth/plunge = 112.0°/38.6°;
While shear sense indicators are not available for these lineation data, we note that this slip azimuth is consistent with the orientation of corrugations on the exposed detachment surface on the central dome of Atlantis Massif [Blackman et al., 1998, 2002].

Detachment-related structures comprising a damage zone of unknown thickness are to be anticipated in the footwall of OCCs [Schroeder and John, 2004; John and Cheadle, 2010]. The presence of N-S striking, E-dipping structures throughout Domain I implies that the damage zone associated with this detachment fault system may extend at least 385 m into the footwall. In addition, although not as pervasive as in the upper part of the hole, features with this orientation are present in significant numbers (>1 per meter) to depths as great as 762 mbsf, the boundary between Domains III and IV (Figures 4 and 5). Some of these clearly accommodated moderate strain under amphibolite and/or greenschist facies conditions. These observations therefore suggest that semi-brittle to brittle internal deformation of the Atlantis Massif OCC, albeit localized, persisted to more than 750 m into its footwall.

It is now well established that the footwalls of OCCs, including Atlantis Massif, are rotated substantially (45–65°) away from the median valley about sub-horizontal ridge-parallel axes as they are exhumed [Garcés and Gee, 2007; Morris et al., 2009; MacLeod et al., 2011]. A key prediction of flexure due to unloading of the hanging wall is extensional deformation of the footwall near its upper surface and shortening deformation near its base [Cannat et al., 2009]. Unfortunately the limited available kinematic data currently make it impossible to reliably differentiate extensional from contractional domains in the cored section, and kinematics cannot be determined directly from logging data.

Nevertheless, the N-S striking, moderately E-dipping zones of cataclasite that internally deform the Atlantis Massif footwall are very likely associated with these bending moments. Kinematic data that do exist from the fault zones we have restored (at 159–174 mbsf, 742–762 mbsf, 695 mbsf and 785 mbsf; Section 4.3.3 and Figure 9) indicate reverse shear sense (i.e., E-side upward) in their current orientation. Could these structures therefore be accommodating shortening at the base of the flexing beam of a very weak footwall? It is possible, though we urge caution for two reasons: (1) the intensity of shortening-related deformation should increase downward through the borehole, whereas it instead becomes much less important below 762 mbsf; (2) the E-side up shear sense observed in these N-S striking, E-dipping cataclastic fault zones (Figure 9) [Blackman et al., 2006, Figure F189; Michibayashi et al., 2008] is indicative of reverse faulting only if they were active in their current orientations. Rotating the 20–50° eastward dips by the 46° rotation found by Morris et al. [2009] restores these faults to a very steep easterly to vertical or slightly westerly
dip. If formed prior to rotation and passively tilted, these structures could therefore have initiated either as reverse faults or normal faults. Our only firm constraint is that their east sides moved upward. More careful consideration of the integrated thermo-chronologic, microstructural and rotation history of Atlantis Massif is clearly necessary, which is beyond the scope of this paper.

5.2. Post-Detachment Deformation

In contrast to the uppermost domain, FMS Domains II–VI recognized in IODP Hole U1309D are dominated by E-W-striking features (Figures 4 and 5). In particular, the boundary between Domains I and II is marked by a pronounced change from predominately NNE-striking to E–striking features (Figure 4). The presence of E-W-striking features is not easily

![Figure 9](image9.png)

Figure 9. Equal area stereographic projections of poles-to-planes of FMS features in the vicinity of cataclastic shear zones observed in core samples from IODP Hole U1309D. (a) 175 mbsf, (b) 695 mbsf, (c) 746 mbsf, and (d) 785 mbsf. Red symbols indicate structures associated with the shear zone observed in the core; black symbols indicate the orientation of FMS features in the specified interval surrounding the curated depth of the shear zone observed in the core (with associated 2 sigma Kamb contours).

![Figure 10](image10.png)

Figure 10. Structural lineation data from core pieces recovered from IODP Hole U1309D restored to the true geographic reference frame (see text for restoration methodology). Quantitative data are only available for 22 features over the interval 20–294 mbsf. The mean lineation direction (star, with associated 95% cone of confidence shown as a dashed line) is consistent with the slip azimuth inferred from the orientation of corrugations on the exposed detachment surface on the central dome of Atlantis Massif [Blackman et al., 1998, 2002]. Inset is a close-up photograph of lineations observed on a talc vein (interval 304-U1309D-29R-1, 70–74 cm).
reconciled with detachment-related deformation. Instead, we suggest here that they are related to the regional uplift of the massif.

[37] Bathymetric surveying and geological mapping and sampling of Atlantis Massif reveals a distinct contrast between the Southern Ridge, which is dominated by serpentinized peridotite, and the corrugated Central Dome, on which IODP Hole U1309D is located, which is instead characterized by gabbro [e.g., Blackman et al., 2002; Schroeder and John, 2004] (Figure 1). The two regions are separated by an E-W trending, N-dipping scarp at approximately 30°08′N, which appears to be a normal fault accommodating the uplift of the Southern Ridge relative to the Central Dome [e.g., Blackman et al., 2002; Karson et al., 2006].

[38] The FMS features that dominate below 385 mbsf have similar dip directions to widely spaced normal faults observed across the Southern Ridge that dip steeply to the N or NW, and relatively late faults that dip steeply toward the S or SW [Karson et al., 2006]. Although these faults are not traced northward beyond the crest of the massif, they are approximately parallel with the regional trend of the Atlantis Transform and are inferred to have formed in response to the uplift along the transform edge of the massif [Karson et al., 2006]. Such faults would provide preferential pathways for hydrothermal flow, and any density changes associated with resulting serpentinization would further contribute to uplift and reduce the frictional strength of the fault while temporarily decreasing permeability [Hirose and Hayman, 2008].

[39] It is then reasonable to suggest that the shallow-to-moderately dipping features observed in the lower half of IODP Hole U1309D, which would be in the hanging wall of such a major bounding fault, are likely to have formed in response to displacement on the principal structure (Figure 11). In this case, E-W-striking FMS features would represent synthetic/antithetic fractures associated with a major structure to the south of IODP Hole U1309D, in a damage zone projecting northward to intersect the hole below the main zone of detachment fault-related structures. This model would account for the observed change with depth from detachment-compatible N-S striking structures in Domain I to E-W-striking structures in lower domains.

6. Conclusions

[40] Over four thousand FMS features have been identified in continuous wireline logging data from IODP Hole U1309D over the interval 94–1415 mbsf. After eliminating the least robust data, the remaining 2795 FMS features are divided into six domains.

[41] Domain I is distinct from structurally deeper domains due to the dominance of N-S striking, E-dipping features. By comparison, Domains II–VI are characterized by a dominance of E-W-striking features (dipping to both N and S), although NNE-striking features are still present to the base of Domain III. We associate the N-S striking, E-dipping features with detachment-related deformation,
representing a semi-brittle to brittle damage zone associated with the detachment fault system that extends at least 385 m into the footwall and persists on more localized structures to more than 750 mbsf. Further evidence for detachment-related footwall deformation is provided by reoriented structural lineations associated with mineral veins in the upper 294 m of the hole that are consistent with the orientation of corrugations on the exposed detachment fault surface [Blackman et al., 1998, 2002].

[42] The E-W-striking, N- and S-dipping FMS features that dominate in the lower kilometer of the hole (Domains II–VI) are not compatible with detachment-related deformation. Instead, we suggest that these structures relate to later regional uplift of the massif, and represent conjugate structures in the hanging wall of a major E-W trending, N-dipping fault zone. This separates predominately gabbrorocks of the Central Dome from serpentinized peridotite of the Southern Ridge of Atlantis Massif. Structures intersecting Hole U1309D at depth are likely to have formed in response to displacement on this principal structure as it accommodated relative uplift of the Southern Ridge.

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