Origin of continental margin morphology: Submarine-slide or downslope current-controlled bedforms, a rock magnetic approach

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Abstract

Morphological features observed in both swath bathymetry and seismic reflection data are not unique, which introduces uncertainty as to their origin. The origin of features observed in the Humboldt Slide has generated much controversy because the same features have been interpreted as a submarine failure deposit versus current-controlled sediment waves. It is important to resolve this controversy because similar structures are observed on many continental margins and the origin of these features needs to be understood. Anisotropy of magnetic susceptibility (AMS) measurements on sediment samples acquired from the Humboldt Slide reveal that the top ∼ 8 m have not experienced post-depositional deformation. This suggests that these features are formed by primary deposition associated with downslope currents. Using the same AMS technique on a core acquired north of the Humboldt Slide in a region with no geophysical evidence for post-depositional deformation, we were able to identify a ∼ 1 m thick deposit that appears to be a small slump.

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1. Introduction

Geologists have long appreciated the importance of submarine landslides and failures in the development of unconformities (Embley and Jacobi, 1986; Booth et al., 1993; Evans et al., 1996). Recently there has been much debate concerning the identification of submarine landslides and rotational slumps in seismic reflection data (Dillon et al., 1993; Gardner et al., 1999; Holbrook, 2001; Holbrook et al., 2002; Lee et al., 2002; Trincardi et al., 2004). The controversy arises, in large part, because the stratigraphic geometry of many deposits, previously identified as retrogressive slumps, is not unique and could equally be generated from down- or along-slope currents (e.g. Blake–Bahama collapse structure, Holbrook et al., 2002; Humboldt Slide in the Eel River Basin, Fig. 1). Deposition and erosion of the slope may be caused by a number of different processes (e.g., slope failure, incision and overbank deposits, bottom currents, shelf-edge deltas). Understanding how these processes sculpt the continental slope is critical to generating quantitative geologic models of continental slope evolution (Pratson and Coakley, 1996; Driscoll and Diebold, 1999). Developing a test to discriminate between these alternative scenarios, retrogressive slumps versus current-controlled deposits, would provide valuable new insights into the origin of these deposits and their relative importance in the construction and evolution of continental margins.
Gardner et al. (1999) interpret the Humboldt Slide as having been formed by retrogressive failure and rotation of blocks above a shear zone. Lee et al. (2002) reinterpreted the Humboldt Slide complex as sediment waves emplaced by downslope gravity flows and argued that sediment waves are infilling a slide scar. The differing interpretations of the structure by Gardner et al. (1999) and Lee et al. (2002) highlight the ongoing controversy regarding the origin of this type of morphological feature around the globe (of which the Humboldt Slide complex is only one such example). These two interpretations (primary depositional features versus retrogressive faulting and internal deformation) are based on the same Huntec (Dodds, 1980) seismic data set. The difficulties of interpreting such structures are illustrated by the Blake–Bahama outer ridge. Holbrook (2001) reinterpreted existing USGS data from the Blake–Bahama outer ridge and refined the previous interpretation of Dillon et al. (1993) that the observed features were normal faults associated with a gas-hydrate collapse structure. By examining the stratal geometry, Holbrook (2001) suggested the structures were growth faults recording several events, not just one event as suggested by Dillon et al. (1993). Subsequently, Holbrook et al. (2002) conducted an expensive 3D seismic survey to define the nature of the collapse structures. After acquiring the seismic data set, Holbrook et al. (2002) concluded that the features were in fact not growth faults, but were actually sediment waves.

In this paper, we apply a test based on sedimentary fabric as characterized by anisotropy of magnetic susceptibility (AMS) for assessing the origin of these ambiguous features of the Humboldt Slide. The test is based on magnetic fabrics to determine the extent of post-depositional deformation, a prediction of the retrogressive failure hypothesis. We will outline and describe the results.

Fig. 1. The Eel River Basin is located in northern California, just north of the Mendocino Triple Junction. The Eel River enters the ocean south of Humboldt Bay. The slide is bounded by the Little Salmon Fault Zone to the north, the Eel Canyon to the south and up dip is delineated by the shelf break. The bathymetry is a compilation of the STRATAFORM EM1000, MBARI EM300 and the NOAA coastal relief 3 second data. The red inset shows the location of the study area (Fig. 2) with the core locations marked with orange circles.
2. Geologic setting

The ongoing deformation, uplift, and erosion of the Californian hinterland provide vast amounts of sediment to the U.S. Pacific continental margin (Clarke, 1987; Field and Barber, 1993). The relatively high rate of sedimentation and recurrence of earthquakes (Couch, 1980) makes the Eel River Basin an ideal locale to examine slope failure and consequent slide deposits. In addition to strong forcing functions, vast amounts of data have been acquired in the Eel Basin as part of the ONR STRATAFORM project (Nittrouer, 1999). This background allows us to place our results into a well-defined geological framework.

The Humboldt Slide deposit mantles a bowl-shaped depression that extends from the outer shelf to the middle slope on the Eel Margin (Fig. 2). On the basis of the internal geometry and surficial morphology of the Humboldt Slide deposit, two competing hypotheses have emerged: (1) The Humboldt Slide deposit and internal geometry were formed by retrogressive failure and rotation above a shear zone (detachment) with minimal lateral translation of the deformed sediment carapace (Gardner et al., 1999), and (2) The deposit records primary deposition by density currents (hyperpycnal flows) cascading down a pre-existing slide scar (Lee et al., 2002). In the primary deposition scenario, previous slope failure and evacuation of the failed material over-steepened the local slope and created the Humboldt Slide scar. Acceleration of the density flows in response to the locally over-steepened slope gives rise to current-controlled bedforms in this region.

The Eel River Basin has a narrow shelf (22 km from the Eel River to the shelf break above the Humboldt Slide) such that sediment can potentially escape over the shelf break onto the shelf slope and beyond (Alexander and Simoneau, 1999). The typical winter swell can resuspend sand in 50–80 m of water, whereas large storms can rework sand on the middle to outer shelf, and perhaps down to the upper slope (Alexander and Simoneau, 1999). Deposition rates derived from $^{210}$Pb and $^{137}$Cs reveal high sediment accumulation rates (SAR) on the shelf, and on the slope in the area of the Humboldt Slide (Alexander and Simoneau, 1999).

Fig. 2. Location of XStar CHIRP (Fig. 4), Huntec (Fig. 5), and multi-channel seismic (MCS; Fig. 19) lines on the Humboldt Slide amphitheater. The core locations are marked with orange dots. Cores 1, 2 and 7 are coincident with the Chirp, Huntec, and MCS E–W lines. Core 5, which was acquired outside of the Humboldt Slide, was used as a control core for the study. Image courtesy United States Geological Survey.
3. Methods

3.1. Sedimentary AMS fabrics

Early workers such as Ising (1942), Rees (1961), and Marino and Ellwood (1978) suggested the use of AMS to test the reliability of natural remanent magnetism (NRM) measurements from sediments. From these measurements, they concluded that “normal” (oblate) AMS fabrics should generally yield robust results for paleomagnetic field studies. They found poor or incorrect results from samples that showed distorted magnetic fabrics and suggested a wide range of possible causes, including slumping.

Kanamatsu et al. (2001) summarized the potential ways that AMS fabric could be altered by internal (mineralogical) changes and physical reorientation of magnetic grains. Internal changes may be from magnetostriction, growth or dissolution, and brittle or plastic deformation of individual grains. Physical reorientation can be non-coaxial (simple) shear or compaction, which does not alter the fabric within each grain; however, the strain will still alter the overall magnetic fabric.

A range of laboratory experiments have been conducted to examine the depositional controls on magnetic fabrics. For example, Rees and Woodall (1975) investigated a variety of materials using both running-water deposition and deposition from slurries slumping. The experimental results suggest a systematic variation in the AMS fabric with changes in the critical shear stress in the bottom boundary layer (i.e., increasing water current velocity). Additional experiments with plaster mixtures...

Fig. 3. Schematic diagram illustrating development of sedimentary magnetic anisotropy fabric. Eigenvector directions are plotted in the equal area projections whereby circles are the directions $V_3$ associated with the minimum eigenvalue $\tau_3$, triangles are $V_2$ associated with $\tau_2$ and squares are the directions $V_1$ associated with the maximum eigenvalue $\tau_1$. The histograms to the right are bootstrapped eigenvalues for the specimens showing the 95% confidence intervals for $\tau_1$, $\tau_2$, $\tau_3$. a) Quiet water: the $V_3$ directions are typically vertical and $\tau_1$ and $\tau_2$ are not significantly different (i.e., the fabric is oblate). b) Moderate flow: the $V_3$ directions are deflected from the vertical by the current, but the fabric is still typically oblate. The bootstrap mean declination for $V_3 (D_{\text{sub} 3})$ is the inferred direction of flow, which closely approximates the paleocurrent direction measured from ripples (Schwehr and Tauxe (2003)). c) Uniaxial horizontal deformation: the eigenvalues will be significantly different (i.e., the fabric is triaxial). In extreme cases, the $V_3$ directions will ultimately become horizontal. Arrows show the direction of compression that is orthogonal to the $V_1$ orientation. Note that scales vary for the histograms.

found that the eigenvector associated with the maximum eigenvalue can align either parallel or perpendicular to flow direction depending on flow conditions (Rees, 1983).

Several authors have recognized the effects of sediment deformation on NRM (see Tarling and Hrouda, 1993 and Tauxe, 1998 for summaries). For example, Rosenbaum et al. (2000) examined a core (OL-92) from Owens Lake, CA that contains sediments ranging from 800 ka to the present. The original interpretation of the OL-92 magnetic record was that there were a number of geomagnetic excursions in the Brunhes Chron (Glen and Coe, 1997). Rosenbaum et al. (2000) found that sediment deformation was associated with a number of these “excursions” and that the eigenvector associated with the minimum magnetic susceptibility (here called $V_3$) could be used as an indicator for deformation. If the direction of $V_3$ is significantly deflected from vertical, that portion of the core might have been deformed, and therefore should not be used for field direction or field intensity. Rosenbaum et al. (2000) arrived at the conclusion that the deformation observed in the OL-92 core was a result of fluidization. A seismic survey conducted after coring shows that the OL-92 core is located in the Owens Valley Fault Zone (Brooks and Johnson, 1997) and the deformation could have resulted from a combination of drilling and faulting.

Given that AMS is extremely sensitive to strain (Housen et al., 1996; Kanamatsu et al., 2001), magnetic fabric has recently been used to detect subtle deformation of sediments and to distinguish geomagnetic features from deformational artifacts (e.g., Rosenbaum et al., 2000; Cronin et al., 2001). Cronin et al. (2001) suggested that AMS could be used to detect slumps not otherwise obvious from the geologic field evidence (so-called “crypto-slumps”, Schwehr and Tauxe, 2003). These studies show that AMS can distinguish between post-depositional deformation and primary depositional features. Although AMS has long been used to detect deformation in a variety of geological applications, our aim is to determine the geological origin of a continental margin deposit. AMS may provide a powerful method that can be used in a number of geological settings to detect post-depositional deformation where existing data are equivocal.

There is always the possibility that there might be complications in the AMS results caused by diagenetic effects on the magnetic mineralogy. Such changes can be detected using rock magnetic methods such as anhysteretic remanent magnetization (ARM), isothermal remanent magnetization (IRM), low field bulk susceptibility ($\chi_{lf}$ or sometimes written just $\chi$) and high field susceptibility $\chi_{hf}$ (e.g. Banerjee et al., 1981; King et al., 2001).
3.2. Development of magnetic fabric in sediments

Here we summarize decades of research on AMS fabric in sediments in a variety of current regimes (see Fig. 3; Tarling and Hrouda, 1993). In quiet water conditions (Fig. 3a), there is a tendency for elongate particles to lie sub-parallel to the bedding plane. As the magnetic susceptibility is usually at a maximum parallel to the long axis of particles, the direction of maximum magnetic susceptibility ($V_1$) will tend to lie close to the plane of the bedding. However, there is no preferred direction within the bedding plane, therefore the direction of intermediate magnetic susceptibility ($V_2$) and $V_1$ will be indistinguishable as will the associated eigenvalues ($\tau_2$ and $\tau_1$). Hence, the magnetic fabric will be oblate with a vertical $V_3$ direction.

In moderate water currents (Fig. 3b), especially on inclined bedding planes, particles may be slightly deflected, resulting in off-vertical $V_3$ directions. Here too, we expect the fabric to be characterized by an oblate AMS ellipsoid, but the $V_3$ direction will be deflected in the direction of the paleocurrent.

What happens to the magnetic fabric during post-depositional deformation is more complex. Initial theoretical work on the relationship between magnetic fabrics and actual grain fabrics with respect to strain was conducted by Owens (1974), Hrouda and Hruskova (1990), and Housen et al. (1993). Most studies using AMS fabric to determine strain have been applied on tectonic scales and have examined weakly metamorphosed rocks (e.g. Pares et al., 1999; Kanamatsu et al., 2001). Studies on such low-grade metamorphic rocks are complicated because chemical changes during metamorphism may affect the magnetic minerals to a significant degree.
some degree. It is difficult to separate physical from chemical effects on magnetic fabrics. Nonetheless, research shows that strain alters magnetic fabric and can have effects such as deflecting the minimum susceptibility vector from vertical, and aligning the maximum eigenvector perpendicular to the axis of compression (see Fig. 3c). By working with sediments that have not experienced diagenesis from deep burial and heating, chemical changes should be minimal.

3.3. AMS and slumping

Cronin et al. (2001) investigated a section of limestone in Italy to define the paleomagnetic field in the Cretaceous Normal Superchron. The paleomagnetic data displayed several intervals in which the direction deviated significantly from the expected normal direction. Such data are often interpreted as excursions of the geomagnetic field. In order to rule out slumping as a possible cause, they used AMS fabrics to characterize the “ordinary” and “deviant” intervals. The deviant intervals were triaxial, while the ordinary intervals were oblate. These results strongly suggest that the deviant directions were the result of crypto-slumping, which is soft sedimentary slumping sub-parallel to bedding that leaves little to no visible record in the outcrop.

As suggested by Rosenbaum et al. (2000) and Cronin et al. (2001), it appears that even minor amounts of soft sediment deformation can have a profound effect on the paleomagnetic record. However, such deformation can be extremely difficult to detect based on visual observations alone, hence the term crypto-slump (Schwehr and Tauxe, 2003).

Schwehr and Tauxe (2003) pursued the idea that soft sediment deformation can be detected through the use of AMS by investigating both crypto-slumped sediments...
from a marine environment, along with the sediments from within and above the slump to confirm observations of Cronin et al. (2001). They found a crypto-slump in a shale that can be traced laterally to a slumping event observed in the outcrop. Without the excellent exposure and lateral continuity along the outcrop, its slumped nature would not be easily detected. Schwehr and Tauxe (2003) then developed a test for post-depositional deformation based on the AMS characteristics of slumped versus undeformed sediments.

In essence, the test assesses whether the $V_3$ directions are vertical and distinct from $V_1$ and $V_2$, and whether the fabric is oblate (as expected for undisturbed sediment), or triaxial or prolate (as expected for disturbed fabrics). They used a statistical bootstrap approach to perform this test (see Constable and Tauxe, 1990; Tauxe, 1998 for more details).

The example from Schwehr and Tauxe (2003) shows that AMS is able to distinguish between sedimentary structures and deformation in situations where field observations are ambiguous. However, the AMS bootstrapping technique developed by Cronin et al. (2001) and Schwehr and Tauxe (2003) may not be easily applicable to cores such as those collected in the Eel River Basin because of different compaction and deformation states. Bootstrapping conducted by Cronin et al. (2001) and Schwehr and Tauxe (2003) grouped samples into stratigraphic layers; however, such sampling is not currently possible with today’s coring technology. For each stratigraphic layer, there may be different magnetic grain distributions and concentrations in addition to the possibility of different flow regimes and directions. We employ bootstrap statistics on sediment zones in this study, but the results should be used with caution.

### 3.4. AMS applied to the Humboldt Slide

The two alternative hypotheses for the formation of the Humboldt Slide deposit (slope failure and sediment waves) predict very different fabrics that can be measured using AMS. The deformational hypothesis of Gardner et al. (1999) predicts a triaxial fabric with the

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maximum axes of susceptibility (the $V_1$ eigenvectors) being either poorly grouped, perpendicular to the most compressive stress (see Fig. 3c), or approximately north-south. The density current hypothesis of Lee et al. (2002) predicts oblate fabrics with the minimum axis of susceptibility (projected into the lower hemisphere) deflected in the direction of paleocurrent flow (see Fig. 3b), or approximately westward. The center region of the core should display oblate AMS fabric if the features are depositional, whereas broad deformation will show dominantly triaxial fabric throughout.

3.5. Seismic data

Seismic lines covering the top half of the Humboldt Slide were acquired during August 1999 as part of the ONR STRATAFORM project (cruise TTN-096). The CHIRP seismic system (e.g. Schock et al., 1994; Quinn et al., 1998; Gutowski et al., 2002) is a modified EdgeTech XStar system with an ADSL link from the fish to the topside computers. The data were collected with a 50 ms sweep from 1 to 6 kHz. The XStar SEG-Y records were processed with seismic-py and SIOSEIS (Henkart, 2006), and were plotted with pltsegy. Fig. 4 shows the section of the cruise data relevant to this study.

The Huntec data presented in Fig. 5 were collected during August 1995 on cruise W-2-95-NC. The data were processed with a combination of Sonarweb and seismic-py. A hydrophone (channel 2) is mounted on a tail behind the fish, which experiences a large amount of motion, so we processed only channel 1 (Galway, 2000). Huntec data from cruises W-2-95-NC and W-1-96-NC have been presented in Gardner et al. (1999) and Lee et al. (2002).

3.6. Coring

In November 2001, large-diameter piston cores were acquired using the Oregon State University Coring Facility on board the R/V Thompson. The core sites were selected based on CHIRP seismic data (Fig. 4) and on EM-1000 swath bathymetry (Fig. 2). P-code differential GPS was used to locate the core sites and yielded 10 m, or better, accuracy of the core location. Core locations and lengths are summarized in Table 1.
The lengths of the piston cores range from 5.9 to 7.8 m. The piston cores have an inner diameter of 10.2 cm which minimizes deformation associated with coring allowing undisturbed samples for AMS analysis to be acquired away from the liner effects.

Core 5 (Figs. 2 and 6), collected north of the Humboldt Slide, serves as a control because it is located in an area with minimal deformation based on the seismic and bathymetric data. Core 5 was collected at a depth similar to that of cores 1, 2, and 7 and is located in a portion of the slope characterized by gullies (referred to as rills) described by Spinelli and Field (2001). We predicted that the AMS results for the control core would show a normal sedimentary fabric, perhaps with a signature of current flow down or across slope (i.e., gravity sheet flows or slope-parallel contourites).

Cores 1, 2, and 7 were acquired in the primary study site within the Humboldt Slide (Figs. 4 and 7–9). These cores sampled across the crest of one sedimentary structure located in the center of the Humboldt Slide at a water depth of 460 m. This feature has a wavelength of about 150 m and an amplitude of approximately 6 m. Given fish layback uncertainty (as marked in Fig. 4), it is not possible to determine exactly where on these structures each core was acquired; however, it is clear that the three cores have sampled both the upslope and downslope components of the slide on feature c. Note that the estimated core penetration is shown on Fig. 4.

3.7. Paleomagnetics

From the cores, we collected 8 cm³ paleomagnetic sample cubes with a typical sampling interval of 10 cm. The down core AMS measurements provide the key data for interpreting the Humboldt Slide as either a retrogressive failure or a downslope current-controlled deposit (i.e., sediment waves) infilling a slide scar. Magnetic measurements were performed at the Scripps Paleomagnetic Laboratory. NRM measurements were conducted on 3-axis CTF and 2-G cryogenic magnetometers (designated Bubba and flo respectively), located in a magnetically shielded room. Alternating field (AF) demagnetizations were accomplished using an SI-4. After best fit directions for each sample were found, the Fisher statistics (Fisher, 1953) were applied to each core section to get a best fit declination ($D^\parallel$, $K$. Schwehr et al. / Marine Geology xx (2007) xxx–xxx

A best fit tensor is derived from the 15 measurements with a field of 1 T.

IRM's were imparted with an ASC impulse magnetizer using a 100 mT alternating field and a 40

Table 2). D was then applied to each core section such that the AMS eigenvectors are geographically oriented.

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The V1 declination (D_{V1}) shows the direction of the eigenvector associated with the maximum eigenvalue. This direction is only meaningful if the τ1 eigenvalue is statistically distinguishable from τ2. Therefore, the V1 directions marked as prolate (cyan) and triaxial (red) in the Shapes column are significant. The V1 declination is meaningful when the fabric shape is either oblate or triaxial (τ3 distinct from τ2). The inclination of V3 (I_{V3}) is often used as a proxy for detection of bed rotation (e.g. Rosenbaum et al., 2000; Kanamatsu et al., 2001; Housen and Kanamatsu, 2003).

After magnetic measurements were made, the samples were wet sieved to remove the clay to medium silt fraction using a 47 μm screen. Finally, a range of grain size separates were sieved to determine silt, sand, and organic debris fractions; the organic debris included branches, twigs, seeds, etc.

4. Results

Cores 1, 2, and 7 were collocated with CHIRP seismic data to constrain the geometry and stratigraphy of the three prominent highs being studied (marked a–c in Fig. 4). These three cores were collected across high c. The seismic reflection data were used to determine where to sample the features because in certain areas the deformed features are mantled by a pelagic drape. The thickness of the drape varies systematically from the top of the slide complex to the base (Fig. 5). At the top of the slide structure, there is little to no detectable pelagic drape overlying the deformed features at the locations of cores 1, 2, and 7 as observed in the CHIRP and Huntec seismic lines (Figs. 4 and 5a). Examination of our core locations (Fig. 4) and co-registered seismic data reveal that all three cores penetrated through any pelagic drape into the underlying sedimentary features.

The CHIRP system imaged faint seaward dipping reflectors with high-amplitude landward dipping reflectors. There is a marked asymmetry with the landward dipping sequences being much thicker than the seaward dipping units. In fact, across some features only the landward dipping sequences are observed with individual horizons cropping out at the sea floor on the seaward slope (Fig. 4a).

The first test to determine if post-depositional deformation occurred is to examine the NRM directions for evidence of rotation. The seaward and landward limbs of these features (Fig. 4a–c) exhibit a range of dips from 2.1° to 4.1° (Fig. 4; inset table). Cores 1, 2 and 7 penetrate structure c which has a seafloor landward slope of 2.2° and a seaward slope of 4.1°. If there is E–W compression or rotation with a northerly fold axis, there should be a shallowing of approximately 0.5° in α zones occur at the top of each of the piston cores. This zone is characterized by high $\chi_{lf}$ anisotropy, ARM, and IRM. The individual shape fabrics are predominately triaxial. Bootstrap eigenvectors have a tight cluster near vertical. Blue squares are $V_1$; yellow triangles are $V_2$; and red circles are $V_3$. The cyan dots are the bootstrap eigenvectors; $V_1$, $V_2$, $V_3$ eigenvectors are not distinguished. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
the NRM inclination compared to the expected geocentric axial dipole (GAD) inclination of 60.0°. Given that the \( \alpha_{95} \) confidence values range from 8 to 26° (Table 2: cores 1, 2, and 7), such small rotations are below the resolution of this approach, and therefore, we must rely on the AMS technique for detecting post-depositional deformation.

The AMS results are shown in Figs. 6–9. Core 5 (Fig. 6) was collected as a control for this study and as such, it is located outside of the Humboldt Slide. Cores 1, 2, and 7 were acquired within the slide feature to assess whether the sediment carapace has experienced rotation and deformation. Based on the observation of the AMS fabric observed in the cores, four distinct zones termed \( \alpha, \beta, \beta', \) and \( \gamma \) were identified. Core 5 is used as the “type section” because it exhibits all four zones. Here we will describe the characteristics that define the four zones.

### 4.1. \( \alpha \) — alpha

The first zone, \( \alpha \), is characterized by high susceptibility and high anisotropy observed in the AMS data. Individual samples exhibit a predominantly triaxial AMS fabric type with near-vertical orientation of the minor eigenvector (see Figs. 6–9: \( I_{V3} \)). The Hext (Hext, 1963) average inclination of \( V_3 \) for a group of samples is recorded in Table 3 as \( I \). The \( V_1 \) eigenvectors, as observed in the equal area projection (Fig. 10), have no preferred orientation (Figs. 6–9: \( D_{V1} \)). The confidence ellipses and best fits derived by bootstrap statistics for the different zones are reported in Table 3. The \( \alpha \) zone is also characterized by large ARM, IRM and \( \chi_{lf} \) (we plot only \( \chi_{lf} \) in Figs. 6–9). An increase in the coarse silt fraction correlates with the boundary between \( \alpha \) and \( \beta' \) in core 2 (Fig. 11: grain size). When plotting IRM versus \( \chi_{lf} \), the different zones can be delineated as shown in Fig. 12. \( \alpha \) exhibits high and transitional \( \chi_{lf} \), ARM, and IRM, whereas low \( \chi_{lf} \), ARM, and IRM are characteristic of the other zones (Fig. 12).

### 4.2. \( \beta' \) — beta prime

The \( \beta' \) zone is observed in cores 2 and 5 and is characterized by rotation of the \( V_3 \) eigenvectors away from vertical determined for the individual sample...
measurements (Figs. 6 and 8: $I_{V3})$. The deflection of $V_3$ away from the vertical is also observed in the equal area projections (Fig. 13). A marked decrease in total anisotropy as evidenced by the eigenvalues is also characteristic of the $\beta'$ zone. (Figs. 6 and 8). $V_1$ exhibits slightly more grouping in $\beta'$ than in the overlying $\alpha$ zone,
nevertheless a strong preferred orientation is not observed (Fig. 13). The AMS fabrics for individual samples are predominately triaxial for the $\beta'$ zone.

4.3. $\beta$ — beta

Underlying the $\beta'$ zone in cores 2 and 5, there is a pronounced shift to more oblate fabric for individual samples in the $\beta$ zone. However, the transition from $\beta'$ to $\beta$ is not well defined by either the eigenvalues or $\chi_{lf}$. In cores 2 and 5, the $\beta$ zone is defined by near-vertical $V_3$ orientation with tight clustering (Figs. 8 and 6: $I_{V3}$). Note that in core 7, even though the eigenvalues by sample show some scatter, the bootstrap vectors show a tight cluster near vertical (Fig. 14).

In cores 1 and 7, $\beta'$ is not observed and the $\alpha$ zone mantles $\beta$. When $\alpha$ directly overlies $\beta$, the zones are delineated by a marked shift in $\chi_{lf}$ (Figs. 7 and 9). In core 1, a marked decrease in overall anisotropy appears to correlate with the boundary between $\alpha$ and $\beta$. In core 1, the decrease in overall anisotropy is more subdued for the transition between $\alpha$ and $\beta$ (Fig. 7) than in the other cores. In the $\beta$ zone, $V_1$ for all cores shows weak grouping; however, there does not appear to be a preferred orientation for $\beta$. As mentioned before, $\beta$ is characterized by low $\chi_{lf}$, ARM, and IRM (Figs. 11 and 12).

In all four cores, only small zones of bioturbation are observed. In general, the layering is clearly visible, and undisturbed as revealed by the core photos. The boundaries between layers are sharp, and mottling and smearing of layer boundaries is not commonly observed. A large number of organic-rich layers are clearly visible in the cores with a maximum thickness of 30 cm (core 2 from 568–589 cm for the vascular plant material deposit; Fig. 6).

The core photos show a reciprocal relationship for the $\beta$ zone and the occurrence of dark, organic-rich layers depending on whether the cores were acquired within or outside the slide region. Cores 1, 2, and 7 show an increase in occurrence and thickness of dark organic-rich layers in the $\beta$ zone compared to the $\alpha$ zone. The dark organic-rich layers are comprised mostly of
vascular plant matter (Fig. 15). In core 5, the \( \alpha \), \( \beta' \), upper \( \beta \) zones are characterized by dark organic-rich layers, while the lower beta zone is largely devoid of such layers. In core 5, the thickest dark, organic-rich layers occur in the \( \gamma \) zone (Figs. 6 and 16).

The \( \gamma \) zone is only observed in core 5, which was acquired to the north of the Humboldt Slide. \( \gamma \) is characterized by a marked deflection of \( V_3 \) from vertical in both the individual and group samples (Figs. 6 and 17).

In core 5, the \( \beta \) zone above and below the \( \gamma \) zone is characterized by oblate fabric, with \( \gamma \) being predominately triaxial. The \( \gamma \) zone is indistinguishable from \( \beta \) based on \( \chi_{16} \) eigenvalues, ARM and IRM. Within the \( \gamma \) zone, the sediments appear to show signs of post-depositional deformation as a dark, organic-rich horizon may have been folded (Fig. 16 inset).

### 5. Discussion

In order to fully understand a marine slump, it is helpful to review the features that are typically expected. Comparison with a small slide exhibiting minimal run out shows markedly different features than those observed in the Humboldt Slide. Fig. 18 is a CHIRP seismic image of the Gaviota Slide from the Santa Barbara Basin that exhibits features typical for many slides and provides valuable insights for the expected sedimentary structures and morphology (Schwehr et al., in press). For the Gaviota slide, the failed material has not moved far downslope with minimal translation from the evacuation to accumulation zone. If the failure is retrogressive in origin, then after initiation, the failure propagates upslope from the point of initial failure and terminates at a headwall scarp. Along the seaward extent of the slide complex, the toe often exhibits signs of compressional deformation (Fig. 18). The \( V_1 \) eigenvalues showed a preferred orientation as a result of downslope compression (Fig. 18: inset). Between the upper headwall scarp and the toe, there is an evacuation zone, where the material has vacated, or a zone of thinned and extended material.

According to Gardner et al. (1999), the Humboldt Slide is thin skinned; however, little or no accommodation zone for this slide is observed. This does not match the model for other slides where either a catastrophic failure mobilizes the sediment into a turbidity current, or sediment is removed from an evacuated zone to an accumulation zone downslope (Fig. 18).

The morphology of the Humboldt Slide suggests minimal translation down-slope because there is no downslope thickening or upslope thinning. Furthermore, the MCS data acquired across the region images individual layers that thicken and diverge toward the margin. The divergence of the horizons and the diminished dip up section may reflect long-term tectonic control in the region, suggesting fault-controlled accommodation (e.g., Driscoll and Hogg, 1995). Onlap and thinning are observed across the Little Salmon Anticline.

Lee et al. (2002) presented numerous lines of evidence based on stratigraphy and morphology that the Humboldt Slide features are current-controlled bedforms. Nevertheless, based on the same internal geometry and morphology Gardner et al. (1999) argued that these features are the consequence of post-depositional deformation. As previously mentioned, morphology is not unique and thus, the debate concerning the origin of these features continues. AMS measurements provide additional constraints on the origin of these features and are discussed below.

#### 5.1. \( \alpha \)— alpha

The high \( \chi_{16} \), ARM and IRM characteristic of the \( \alpha \) zone is different than the other underlying zones observed in the cores. Given that the high ARM and IRM values are only observed in the upper sections of the cores, Figs. 11 and 12 suggest that the base of \( \alpha \) is either (1) a diagenetic front delineating the top of the sulphate reduction zone, where biomediation consumes a fraction of the ferro magnetic grains; or (2) a mineralogical change reflecting a change in sediment...
discharge from the Eel River system. Diagenetic signatures in rock magnetic parameters (e.g. $\chi_{lf}$, ARM, IRM) have been reported by numerous authors from a wide range of environments around the world (e.g. Karlin, 1990b; Leslie et al., 1990b; Tarduno, 1994; Liu et al., 2004; Geiss et al., 2004; Kumar et al., 2005; Pan et al., 2005; Riedinger et al., 2005; Rowan and Roberts, 2006). These transitions are typically attributed to a sulfate reduction in sediments that preferentially consume magnetites with large surface area to volume ratios (e.g., smaller grain size). Karlin (1990b) concluded that magnetic mineral diagenesis is likely to occur in rapidly deposited, sulfidic sediments. The process may shutdown after the initial reactions, not because of the complete removal of magnetite, but from the magnetites becoming covered in a protective coating of pyrite (Egli, 2004).

The ARM versus $\chi_{lf}$ bi-plots show that the $\alpha$ zone is separate from the other zones that have moderate or low ARM and IRM values (Fig. 12). King et al. (1983) reported that for a line passing through the origin and through a group of measurements on an ARM versus $\chi_{lf}$ plot (they use $\chi_{ARM}$, which is a normalized form of ARM), the slope of the line is related to the size distribution of the magnetites in the samples. Steeper slopes are indicative of finer grained magnetites, whereas shallower slopes are evidence for coarser grained magnetites. If this relationship holds for the sediments in this study, then the most recently deposited sediments...
The marked shift to more oblate sample and group shape fabric defines the $\beta$ zone, which implies quiet water deposition. This conclusion is a bit surprising.

The results for the $\alpha$ zone show conflicting model interpretations based on that presented in Fig. 3 and are difficult to interpret. The $\alpha$ zone bootstrap inclination (Table 3: $I$) range from 83.1° to 88.1°. This tight vertical $V_3$ implies quiet water deposition as illustrated in Figs. 3a. Cores 2 and 5 show the strongest evidence for flow directions of 306° and 180° respectively based on $V_3$ $D$ (Table 3).

On the other hand, with the majority of samples being triaxial, one might be tempted to classify the sediments as deformed based on the triaxial histogram shown in Fig. 3c. It is important to distinguish the difference between sample level anisotropy (Figs. 6–9) and group level anisotropy (Fig. 3: bootstrapped eigenvectors). A group level bootstrap histogram tests for coherent deformation, whereas sample level anisotropy test for the statistical distinguishability of the eigenvectors for one specimen. A triaxial sample can imply a number of sediment histories including deformation or deposition under moderate flow conditions. Based on the observations, we interpret alpha to be indicative of deposition in moderate flow conditions.

### 5.2. $\beta'$ — beta prime

The $\beta'$ zone is only observed in cores 2 and 5 and is characterized by $V_3$ deviating from vertical. $\beta'$ has relatively stable $\chi_{16}$, ARM, and IRM. The individual samples are predominantly triaxial, but there are several samples that are prolate. Nevertheless, the $V_1$ orientation shows no preferred direction in the bootstrap. We interpret this to be moderate to strong flow conditions and core 2 may be on the apex of the feature which may have exposed this location to slightly greater currents and/or erosion. Conversely in core 5, the $V_3$ does not exhibit as much deflection, which may be indicative of more moderate flow. Core 2 sampled high flow conditions that appear to be centered on 325° (Table 3), which would be consistent with predicted flow directions. The observation that cores 1 and 7 do not exhibit $\beta'$ zones implies that the $\beta'$ zone observed in core 2 is a local feature with little lateral extent.

### 5.3. $\beta$ — beta

The marked shift to more oblate sample and group shape fabric defines the $\beta$ zone, which implies quiet water deposition. This conclusion is a bit surprising.
considering the size and frequency of large organic-rich layers within the zone (e.g. Fig. 8: β and 15). This may be caused by how flows attach to and detach from the bottom as they travel over an undulating sea floor.

This is the zone where the best evidence should be found for the deformation predicted by the slope failure model (Gardner et al., 1999). The geometry observed in CHIRP and Huntec seismic data (Figs. 4 and 5) predict thickening of 2:1 or greater on the upslope limbs. This amount of compression is expected to create fabric like that illustrated in Fig. 3c. However, the eigenvectors plotted on stereonets with bootstrap eigenvectors for β most closely resemble Fig. 3a. Core 7 is the least like Fig. 3a, but the bootstrap eigenvectors are tightly clustered near vertical. The scatter in the eigenvectors can be traced to samples in four discrete regions located at ∼ 167, 261–282, 420–430, and 572 cm. These layers with V_3 deflected from the vertical might correspond to periods of higher flow. Because the V_1 eigenvalues shows no preferred orientation for β in Core 7, it does not appear to be recording post-depositional deformation.

5.4. γ — gamma

γ stands out as the largest region of deflected V_3. The I_{γ3} has a saw-toothed pattern indicating deformation that may not be coherent. This observation is supported by visual inspection of the core photograph, which shows a folded organic-rich layer (Fig. 16). Given the degree of deformation in γ, it is difficult to assess the direction of compression for this slump. The bootstrap D̄ for V_1 is 330°, but this value has an associated ζ of 90°, meaning that D̄ for V_1 is not significant. However, the lack of preferred orientation of V_1 in the bootstrap test is consistent with highly deformed and folded sediment (e.g., recumbent folding). Such a deformation pattern exhibits fabrics in the AMS that are similar to deposition under moderate to high flow conditions. Note that γ looks very different from the eigenvectors from the Gaviota Slide shown in Fig. 18 and much more like β′ in core 2. Nevertheless, visual examination of the core indicates that the fabric is associated with folding and deformation (Fig. 16).
5.5. The Humboldt Slide as a sediment wave field

We interpret the magnetic and sedimentological data presented here to indicate that there is little to no post-depositional deformation in the Humboldt Slide region.

On the basis of the AMS data, we have identified four types of sediment. There is evidence for moderate to strong flow events and small crypto-slump events (usually occurring as a number of events in a restricted region). These crypto-slump events may represent periods of rapid sea level change, high sediment accumulation rates and loading, or increased seismic activity, but currently there is insufficient evidence to test these hypotheses.

The rill area to the north of the Humboldt Slide shows greater evidence for deformation in comparison to the material in the Humboldt Slide amphitheater. The overlying several meter sediment package may be creeping over the underlying sediments.

We observe no evacuation zone or downslope thickening toward the toe of the slide. The units observed in the MCS data (Fig. 19a; described in Burger et al., 2003) thin seaward down the slide. Individual layers exhibit divergence and thickening towards the margin that may reflect long-term tectonic control in the region (Fig. 19b). The stratal geometry imaged in the MCS data is not consistent with the geometry predicted by the retrogressive failure model.

There is a geometric problem with this feature being interpreted as a slide deposit as it exhibits no thinning in the evacuation zone and no thickening in the accumulation zone. This implies that all deformation is accommodated by in-situ rotation and thickening of beds, with little to no translation. We are unable to identify any signature of such processes beyond occasional thin layers that appear to be deposited in moderate to high flow conditions. The orientation of the
6. Conclusions

The main results of our rock magnetic and seismic reflection study are summarized as follows:

1. The upper ~8 m of the Humboldt Slide sediments are not undergoing post-depositional deformation and folding.

2. The upper section sampled in this study appears to be the result of primary deposition, and thus, we interpret the features to be downslope current-controlled bedforms.

3. Based on MCS data, the thickness and dip of the subsurface sequences are not consistent with the features being a slide deposit. The sediment structures within the Humboldt Slide appear to be sediment waves that may mantle an older slide.

4. We identified a ~1 m thick slump layer located to the north of the Little Salmon Anticline in the region with extensive rills.

5. The change in how much and what fraction of material is delivered to the Humboldt Slide area (versus that north of the Little Salmon Anticline and down the Eel Canyon to the south) may have undergone a recent shift caused by the increasing frequency of combined large storm and flood events from 1955 to the present. This possibility requires further research.

Magnetic measurements allow us to test between the alternative hypotheses of slope failure, and sediment waves for the origin of the Humboldt Slide. Specifically, we are able to discern whether the morphology and internal geometry results from soft sediment deformation and retrogressive slumping, or downslope current-controlled deposition. The former predicts a triaxial AMS fabric with essentially north-south oriented maximum axes (Fig. 3c); the latter predicts an oblate fabric with a possible westward deflection of the minimum axes (Fig. 3a,b).

The morphology and internal architecture of the Humboldt Slide are not unique; there are numerous examples along other continental margins with similar morphology and ongoing debates regarding their origin (see examples in Lee et al., 2002). Magnetic methods for the detection of post-depositional deformation provides a new approach to determine the origin of such features on other continental margins.

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