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A Paleomagnetic Paleolatitude Determination from the Upper Cretaceous Units of the Gold Beach Terrane, Southwest Coastal Oregon

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A PALEOMAGNETIC PALEOLATITUDE DETERMINATION FROM THE
UPPER CRETACEOUS UNITS OF THE GOLD BEACH TERRANE,
SOUTHWEST COASTAL OREGON

BY
NOEL LINER

Accepted in Partial Completion
Of the Requirements for the Degree
Master of Science

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ABSTRACT

The paleolatitude for the Gold Beach terrane has been under debate from several sources (Bourgeois, 1980b; Bourgeois, 1985; Jayko and Blake, 1985; Seiders and Blome, 1987; Aalto, 1989) since the 1980 publication of Joanne Bourgeois’ PhD thesis connecting the Upper Cretaceous formations of the Gold Beach terrane to the Nacimiento Block, currently in the vicinity of Southern California. During the Cretaceous, the Nacimiento Block was farther south than its present position by about 200 km (Blake and Jayko, 1985). The opposing hypothesis of Seiders and Blome (1987) connects the Gold Beach terrane to the neighboring Klamath block, implying at most 300 km of northwards transport for the Gold Beach terrane (Seiders and Blome, 1987). Paleomagnetic study of the Upper Cretaceous units of the Gold Beach terrane has determined that the units originated south of their present position by 1550 +/- 500 km, at a latitude of 34° during the Campanian, placing the origin of the Upper Cretaceous units at the Cretaceous latitude of northern Baja, Mexico. A compaction correction was applied using the results of AMS-determined rock magnetic fabric, showing a small amount of post-Cretaceous inclination shallowing. After correction for tectonic rotation, AMS fabric implies an eastward derived sediment source, possibly from the Nacimiento Block. These results substantiate the hypothesis of Bourgeois (1980), Blake and Jayko (1985) for a southerly latitude of origin of the Gold Beach terrane and rule out the hypothesis of Seiders and Blome (1987) for a local origin connected to the Klamath block. The units are overlain by the Mio-Pliocene Empire Formation at Blacklock Point, however several neighboring late Cretaceous terranes are overlain by the mid-Eocene Lookingglass Formation, capping the emplacement age of the Gold Beach terrane to about 50 Ma. The displacement estimated by this study compares favorably to plate motion rates derived by Engebretson et al., (1984) for translation of this terrane on the Kula plate to its present position by 50 Ma.
Acknowledgements

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INTRODUCTION

The North American Cordillera is composed of a collage of terranes accreted during the Mesozoic (Jones et al., 1983). Paleomagnetic evidence from Cordilleran terranes indicates that the western edge of North America has consistent paleomagnetic discordance when compared to North America cratonic paleopoles (Beck, 1980). This pervasive discordance of paleopoles between the North American craton and Cordilleran terranes indicates significant poleward translation accompanying terrane accretion.

Paleomagnetic and geologic evidence suggests that many of the Mesozoic rocks of Southwest Oregon are exotic to their present location (Blake et al., 1985). Geologic and paleomagnetic studies of the Gold Beach terrane of southwest Oregon (Figure 1) indicate northward latitudinal translation of at least 1200 km (Blake et al., 1985; Jayko and Blake, 1993). Bourgeois (1980) hypothesized that the upper Cretaceous units of the Gold Beach terrane originated at least as far south as present-day southern California based on sediment provenance of the Upper Cretaceous units of the Gold Beach terrane. Seiders and Blome (1987) opposed this view on the basis of conglomerate clast compositions, arguing the Gold Beach terrane originated near the neighboring Klamath block (Figure 2) (Seiders and Blome, 1987).

The paleogeography of the North American craton from the middle Jurassic to the Cenozoic is uncertain, including the nature of the early Jurassic margin along western North America, and the amount and timing of translation of terranes from their places of origin to their location of accretion (Umhoefer, 2003). Further, the amount of post-accretion translation is unknown for many terranes, particularly during the Mesozoic. There is a need to document the timing and location of terrane accretion to the North American Margin. Quantification of post-accretion latitudinal displacement of Cordilleran terranes can offer a better understanding of continental and offshore plate interactions. This study documents the timing and total latitudinal translation of a Late Cretaceous marine terrane on the outboard margin of the Cordillera, helping to establish a more thorough understanding of North America’s paleogeography and tectonic evolution.

It is the aim of this study to quantify the amount of translation of the Gold Beach terrane by use of paleomagnetic methods. A successful paleomagnetic study constrains timing and amount of displacement for these units and adds to the growing body of evidence that the western Cordillera has grown by accretion of exotic terranes. These results will be
used to test the conflicting models of Bourgeois (1980) and Seiders and Blome (1987), which place the Gold Beach terrane at vastly different latitudes during the Cretaceous. This study relies on two hypotheses – 1) the geocentric axial dipole model is representative of the behavior of the ancient magnetic field (Tauxe, 1998), and 2) paleolatitude can be accurately estimated from paleomagnetic inclinations using AMS fabric to correct for possible inclination shallowing in these rocks (Hodychet al., 1999).
GEOLOGY AND SETTING OF THE GOLD BEACH TERRANE

LOCATION and GEOLOGIC SETTING

The geology of the Southern Oregon coast was first mapped by Diller in 1907, then by Howard and Dott in 1968, who divided the Upper Cretaceous units into the Hunter’s Cove Formation and the Cape Sebastian Sandstone. Bourgeois (1980b), broke the Upper Cretaceous Houstenaden Creek Formation out of the Hunter’s Cove Formation based on its differing sedimentology.

The Gold Beach terrane is an assemblage of deformed marine sediments located in southwest coastal Oregon (Figure 3). The Gold Beach terrane is located south of the town of Gold Beach. Gold Beach outcrops are accessible in three places along highway 101. The northernmost exposure is north of Cape Sebastian at Blacklock Point. The middle exposure, 0.5 km south from Crook Point, is approximately 1.0 km in length. The southernmost exposure is the most extensive, covering approximately 4.0^ km, and is bordered by Houstenaden Creek to the south. All outcrops are bounded to the west by the Pacific Ocean. The Gold Beach terrane is fault bounded to the east by the older Jurassic Colebrook Schist and Eastern Franciscan Belt, which in turn are overlain by mildly deformed Eocene sedimentary rocks.

The Gold Beach terrane consists of the Upper Jurassic Otter Point Formation, Campanian (83 Ma) Houstenaden Creek Formation and Campanian – Maastrichtian (83 Ma to 74.5 Ma) Cape Sebastian Sandstone and Hunter’s Cove Formation (Bourgeois and Dott, 1985) (Figure 4). The Otter Point Formation is the basement of the Gold Beach terrane, being unconformably overlain by the Upper Cretaceous strata (Dott, 1971). Based on the differing sedimentology, Bourgeois developed a depositional and tectonic model for placement of the Upper Cretaceous formations over the Otter Point Formation in an environment of vertical faulting accompanied by fluctuations in relative sea levels (Figure 5). Upper Cretaceous Gold Beach units are dated from Campanian to Maastrichtian using dinoflagellates, angiosperm pollen, and peleceypods (Bourgeois and Dott, 1985). Overlapping the Gold Beach Terrane are Mio-Pliocene deposits of the Empire Formation (Diller, 1907).
**GEOLOGY of the OTTER POINT FORMATION**

The basal Jurassic Otter Point Formation of the Gold Beach terrane is composed primarily of interbedded volcanic sandstone, mudstone, conglomerate and andesitic breccia and tuff. The Otter Point Formation is a pervasively sheared and folded assemblage of oceanic origin (Dott, 1970). Aalto (1989, 1990) interpreted the Otter Point Formation as derived from a relatively undissected allochthonous magmatic arc on the basis of sandstone petrology, in contrast to neighboring Franciscan terrane rocks, which represent a transitional continental arc-continental margin environment (Blake et al., 1985).

**GEOLOGY of the HOUSTENADEN CREEK FORMATION**

The Cape Sebastian Sandstone and Hunter's Cove Formation were originally named by Howard (1961), and later described by Dott (1971). Dott's original description of the Cape Sebastian Sandstone indicates the formation is approximately 250 m thick from the lower conglomerate facies to the laminated and burrowed silty sands at the top of the section (Dott, 1971). These 250 m of strata were named the “Lower sequence” or “Lower Cape Sebastian Sandstone” by Hunter (1970), who recognized an unconformity beneath the Cape Sebastian Sandstone. Outcrops of Houstenaden Creek Formation originally were mapped as Hunter's Cove Formation. Bourgeois (1980b) recognized a section of the “Lower Sequence” as distinct and named it the Houstenaden Creek Formation. The Houstenaden Creek Formation is the oldest of the Cretaceous units and unconformably overlies the older Jurassic (Tithonian to Valanginian, ~150 to 136 Ma) Otter Point Formation.

Houstenaden Creek Formation a coarsening-upward series of turbidites representing basin plain to inner fan facies. The lowermost parts of the formation are interbedded shales and fine sandstones of possibly Albian age (Bourgeois and Dott, 1985). Going up section, the stratigraphy abruptly changes to coarse sandstones and laminated mudstones (Figure 6) overlain by a coarsening-upward sequence of amalgamated sandstones (Figure 7), channelized deposits, and conglomerate (Figure 8). Conglomerate increases to the top of the section. Conglomerate is typically composed of well-rounded, resistant clasts of intermediate to silicic volcanic material and some redeposited intraformational concretions (Bourgeois, 1980). Clasts from the Jurassic Otter Point Formation are not found in the Houstenaden Creek Formation. Using the facies association and deep sea fan model of
Mutti and Ricci Lucci (1972), Bourgeois interpreted the entire Houstenaden Creek Formation as a rapid progradation of a submarine fan.

The Houstenaden Creek Formation crops out north of Cape Sebastian and 10 km to the south of Cape Sebastian at Burnt Hill Cove. The bottom of the formation is not exposed, and pervasive faulting between outcrops makes 500 m a minimum estimate of thickness, which possibly reaches up to 700 m (Bourgeois, 1980b).

The upper sandstones of the Houstenaden Creek Formation were dated as Campanian based on the presence of well preserved dinoflagellates and angiosperm pollen (Bourgeois and Dott, 1985). Poorly preserved fossil palynomorphs and foraminifera from the lower shales indicate a possible age of Albian for the base of the Houstenaden Creek Formation (Bourgeois, 1985), however those fossils were found in a tectonic zone, and may not represent the true age of this formation (Blake, 2004, pers. Comm.).

**GEOLOGY of the CAPE SEBASTIAN SANDSTONE**

The Cape Sebastian Sandstone is a 200 m thick fining-upward sequence that was deposited in shoreface to outer shelf (foreshore to basin plain) environments. The section grades from a coarse conglomerate to a hummocky cross-bedded coarse sandstone to alternating burrowed sandy-siltstones and laminated, fine sandstone layers at the top of the section. The Cape Sebastian Sandstone is Campanian in age on the basis of *Inoceramus* fossils. The unit encompasses 1 to 2 million years of depositional history (Bourgeois, 1980a). The Cape Sebastian Sandstone appears conformably overlain by the Hunter's Cove Formation, although the exact nature of the contact is obscured by faulting (Dott, 1971).

The conglomerate at the base of the Cape Sebastian Sandstone ranges from 10 to 20 m thick, and originated in a shoreface environment (Bourgeois, 1980a). Clasts are composed of rounded detritus from the Otter Point Formation, and include calcareous mudstones, volcanics, quartz, diorites, concretions of intraformational scour, and redeposition blocks (Dott, 1971). The conglomerate grades to a hummocky cross-stratified coarse-grained sandstone up to 50 m thick. This hummocky sandstone decreases in abundance going up-section, alternating with fine sandstone and laminated, burrowed siltstones. Hummocky sandstone is interpreted as a nearshore to wavebase facies along an open platform subject to episodic, high-energy wave activity (Figure 9) (Bourgeois, 1980a).
The upper hummocky bedded and burrowed facies to laminated facies extend from the lower shoreface zone to outer shelf. The environment on which the Cape Sebastian Sandstone was deposited would have appeared very similar to the present day northern California/Oregon coast where seasonal, high-energy storms produce waves that scour sediments deposited during fair weather to depth of wave base. Fluvial systems input a generous volume of material eroded from neighboring coastal mountains of steep topographic relief into a deepening basin (Bourgeois, 1980a). The Eel River basin in northern Humboldt County is a modern analogue to the type of basin that probably existed during deposition of the Cape Sebastian Sandstone (Bourgeois, 1984).

The Cape Sebastian Sandstone is estimated to encompass at most 2 million years of depositional history (Bourgeois, 1980a). In order to accommodate this high sediment input and produce the 250 m of fining-upwards stratigraphy, the Cape Sebastian Sandstone is interpreted as having undergone at least 250 m of relative sea level rise. This could have occurred contemporaneously with active vertical faulting and true sea level rise during the Cretaceous worldwide high sea-level stand (Bourgeois 1980a). By the time of deposition of the Cape Sebastian Sandstone, the Houstenaden Creek Formation was juxtaposed in fault contact with the Otter Point Formation, as the Cape Sebastian Sandstone was deposited on both.

**GEOLOGY of the HUNTER’S COVE FORMATION**

The Hunter’s Cove Formation was named by Howard (1961). The Hunter’s Cove Formation conformably overlies the Cape Sebastian Sandstone with a gradational contact. Hunter’s Cove Formation is a 300 m thick, fining upwards series of turbidite sequences, ranging from nearshore to basin plain environments. Hunter’s Cove Formation is dated from late Campanian to early Maastrichtian on the basis of diagnostic fossils Inoceramus, and the ammonoid Anapachydiscus (Dott, 1971).

The Hunter’s Cove Formation has four distinct lithofacies. The most common rock types are interbedded shales and sandstones. The basal lithofacies is thick sandstone interbedded with occasional shales. Bouma sequence b and c are common in these coarser grained sandstones, with distinctive convolute lamination (Figure 10) (Aalto, pers. comm. 2003). These laminations are interpreted as fluid escape structures generated by
liquefaction (Dott, 1971). A coarse channel fill conglomerate is prominent as a sea stack exposed on Hunter's Cove beach north of Pistol River (Figure 11). This channel scour and fill body contains redeposited clasts of Cape Sebastian Sandstone in Hunter's Cove Formation (Figure 12) (Bourgeois, 1980b), and was sampled for a conglomerate test. This section was described by Bourgeois as part of the “chaotic zone” and is correlated to a zone of complex slumps and possible fault-scarp breccias. Going upsection the shale-to-sandstone ratio increases, with thick sandstone beds decreasing in abundance and interbedded shales and fine sandstones increasing (Figure 13). Complexly deformed and overturned bedding is the most striking feature of these thin-bedded sandstones and shales. Deformation of these beds is interpreted to be fault-influenced during or shortly after deposition (Bourgeois, 1980b). Scour marks and trace fossil burrows are common on the undersides of fine-grained sandstone beds (Figure 14). Recently, a new species of foraminiferid diagnostic of environment was described from Hunter's Cove Formation. Bathysiphon Aaltoi was demonstrated on the basis of paleo-ecological reconstruction to inhabit a trench slope or floor, an environment similar to where modern analogue Bathysiphon Filiformis exists today (Miller, 1995). The discovery of this fossil supports the interpretation of this facies as a distal turbidite in a basin plain environment.

CONFLICTING PALEOGEOGRAPHIC MODELS for the GOLD BEACH TERRANE

The location of origin of the coastal Gold Beach terrane is still unresolved (Aalto, 1990). Bourgeois (1980b) inferred the Gold Beach terrane to have a paleolatitude as far south as southern California, based on the affinity of the Gold Beach terrane sediments to rocks of the Nacimiento block (Figure 15).

Sieders and Blome (1987) used quartzite conglomerate clasts to tie the Gold Beach terrane to the Klamath block, inferring approximately 300 km of transport. However the study was flawed for a number of reasons. Ages of radiolarian cherts used by Seiders and Blome (1987) to tie the Otter Point Formation to the Klamath mountains are found in both units, however the Klamaths also have radiolarian cherts of much older (Permian) ages that are not found in the Otter Point Formation (Bourgeois, 1987). Conglomerate clasts derived from the Otter Point Formation could have sources other than the Klamath mountains. A paleomagnetic study conducted by Mankinen et al. (1984) showed no latitudinal translation of the Klamath block. In light of the discordant paleomagnetic evidence from the Otter Point
Formation (Blake et al., 1985), this furthers the case that the Gold Beach terrane has no provenance association with the Klamaths. When the evidence used by Seiders and Blome (1987) was reinterpreted by Jayko (1993), she correlated the Gold Beach terrane to rocks at least as far as 1200 km south, adjacent to the Sierra Nevada batholith (Figure 16).

A previous paleomagnetic study by Blake et al. (1985) of the Jurassic Otter Point Formation placed the Gold Beach terrane a similar distance of 1200 km to the south of its present day location. However, the study could have been biased by the high uncertainty of the Late Jurassic cratonic reference pole used at the time, and unaccounted effects of inclination shallowing. For my thesis, cratonic poles from the Cretaceous and from the Jurassic with a high degree of certainty are used as a reference for paleolatitude reconstruction using the results of this study and the results of Blake et al.'s (1985) study of the Jurassic member. In addition, the Upper Cretaceous units were tested for bias by inclination shallowing.
SAMPLING AND METHODS

FIELD SAMPLING

Samples were collected in the field using a modified chainsaw motor equipped with a water-cooled, diamond-tipped drill bit. The azimuth and hade of the drilled cores were measured using a non-magnetic slotted tube and a Brunton compass adjusted for local declination. A mark was made down the slot with a brass wire, and the core was then extracted from the outcrop face, and labeled. Cores were extracted from each site from all three formations.

Care was taken to ensure a wide enough stratigraphic range was sampled to cover the 10,000 years of depositional history necessary to account for paleosecular variation (Tauxe, 1998). Because the 700 m of stratigraphic section that make up the Upper Cretaceous formations cover 18 million years of depositional history, a spacing of just 1 m between sites is equivalent to nearly 20,000 years. Inferring a linear rate of deposition for multiple turbidite sequences (which occur as discrete events) is incorrect, however, sites were stratigraphically spaced at least 10 m apart, ensuring the effects of paleosecular variation would be averaged in the rock record.

Cores were transported to Western Washington University where they were cut into 2.2 cm long specimens. End pieces from each core were stored as chips for later testing to determine magnetic mineralogy. On average, two specimens were retrieved from each core sample. After cutting and labeling, samples were stored in a Lodestar Magnetics magnetically shielded room at Western Washington University's Northwest Paleomagnetism Laboratory.

A total of 48 cores from 3 sites was collected from the Cape Sebastian Sandstone. All cores are from the hummocky-bedded units, visible on the wind and wave swept cape north of the Hunter's Cove Formation beds.

The Hunter's Cove Formation was sampled at 8 sites with 115 cores collected. Sampling was done on amalgamated, thick bedded sandstones, conglomerate and intensely deformed distal turbidites.

A total of 110 cores comprising 11 sites was collected from the Houstenaden Creek Formation. Samples were gathered from Salal cove, north of Cape Sebastian, and from outcrops at Burnt Hill Cove and Landslide Bay located in Samuel P. Boardman state park.
Lithology of samples ranged from medium-grained, bioturbated sandstones to intraformational conglomerate and concretions (Table 4).

**MAGNETIC FABRIC MEASUREMENTS**

All minerals in a rock have low-field magnetic susceptibility, and for many minerals magnetic susceptibility is anisotropic, with differing values of susceptibility related to crystal axes. For magnetite, grain shape is the main determining factor for differing susceptibility values along each axis of the magnetic ellipsoid shape (Nye, 1985; Tarling and Hrouda, 1993). All of these magnetic minerals contribute to the bulk susceptibility of a rock specimen (Tarling and Hrouda, 1993). By inducing a magnetism in the specimen, the bulk susceptibility of that specimen may be measured.

The anisotropy of magnetic susceptibility (AMS) was measured for each specimen on an Agico KLY3-S Kappabridge susceptibility meter. AMS data were calculated and reduced using SUSAR, the software provided with the instrument. Site means and bootstrap confidence ellipses were plotted using Lisa Tauxe’s (1998) plotams and bootams software.

AMS provides a proxy for mineral alignment in rocks, which can then be related to geologic processes such as strain or paleocurrent alignment. The AMS fabric of these rocks was investigated to establish whether a positive correlation exists between the fabric and regional folding. AMS can provide a reliable method to interpret fabric affected by strain (Borradaile, 1988; Richter et al., 1993). Use of fold axis and maximum AMS axis alignment has been successfully demonstrated to establish strain as the dominant process by which mineral alignment occurred (Hirt et al., 1987). In the absence of visible cleavage, incipient cleavage may develop that controls AMS fabric. In order to establish that the fabric is depositional in nature and does not reflect a tectonic overprint, AMS fabric was compared to a regional fold axis determined by bedding plane intersection from bedding measured at different locations in the field. If the AMS fabrics record depositional processes then rock fabric can be used to interpret likely paleoflow/downslope direction (Tarling and Hrouda, 1993). If there is post-deposition compaction, then these data can be used to evaluate possible effects of compaction on paleomagnetic inclinations.
LOW TEMPERATURE TREATMENT

The initial natural remanent magnetization (NRM) was measured on a 2-G Enterprises model 755 Superconducting Rock Magnetometer. The NRM is the total magnetization acquired by a rock as a result of various mechanisms.

For the first step in demagnetization specimens were immersed in liquid nitrogen for two periods of twenty minutes each. Immersion of specimens in liquid nitrogen (77 K) was used to attain a more clearly defined paleomagnetic result. This immersion cycles magnetite through the Verwey transition at 120 Kelvin (Verwey, 1939). The Verwey transition is the temperature at which magnetite grains change from cubic to monoclinic symmetry (Dunlop and Ozdemir, 1997). Cycling magnetite through the Verwey transition reduces the contribution to magnetization by large multi-domain grains while preserving the harder fraction of remanence (Ozdemir et al., 2001). The process relies on magnetic memory of single domain and pinned-wall multi-domain grained magnetite to faithfully reproduce the more reliable magnetism when reheated to room temperature (Dunlop and Ozdemir, 1997). The NRM was measured after each cycle of this low temperature demagnetization.

PALEOMAGNETIC MEASUREMENTS

After low temperature treatment specimens underwent either stepwise thermal demagnetization or stepwise alternating field demagnetization. Thermal demagnetization occurs when magnetic minerals carrying remanence in a specimen are heated past their respective unblocking temperatures, the temperature at which they experience loss of magnetism (Tauxe, 1998). Different magnetic minerals have different unblocking temperatures, making their presence easy to identify. All thermal demagnetization occurred in a stepwise fashion, to reveal as much of the rock’s characteristic magnetization as possible (Zijderveld, 1967).

Stepwise thermal demagnetization was performed using an ASC-TD 48 magnetically shielded oven. Susceptibility was measured and recorded between thermal steps using a Bartington MS-2 susceptibility meter. Specimens were demagnetized until magnetization was 5% of original NRM or magnetization became so erratic as to warrant removal of specimen from circulation for the study.

A typical demagnetization schedule started from 80° C and continued in 30° C steps, until reaching 300° C then proceeded in 10° C steps to 350° C, back to 30° C- 40° C steps
up to 540° C, then 10° C steps to 600° C. This pattern is tailored to demagnetization of goethite in the low temperature ranges, pyrrhotite in the 300° to 350° range, and magnetite in the 550° C - 590° C range (McElhinny and McFadden, 2001). Justification for these temperatures is based on the assigned Curie temperatures of ~160° C for goethite, ~320° C for pyrrhotite (Tauxe, 1998) and 580° C for magnetite (Dunlop and Ozdemir, 1997). There is the possibility of various types of titano-magnetite being part of a specimen's magnetic mineralogy, and hence, the interim temperature steps between the above-mentioned thermal unblocking temperatures are valuable in assessing the true demagnetization path of any specimen subjected to thermal demagnetization.

At least one specimen per site undergoing thermal demagnetization was selected for alternating field demagnetization (AFD). AFD was performed using a D-Tech D-2 Alternating Field Demagnetizer at 5 to 10 mT steps until fluctuations in NRM intensity and direction displayed by the specimen became consistently erratic. For well-behaved specimens, intensity decay was smooth until intensity reached a level of 5% of its original value. Most specimens were fully demagnetized by ~120 mT.

Alternating field demagnetization works by applying an oscillating magnetic field to a specimen. Grains with magnetic moments having coercivities below the peak alternating field will track the field and become randomized as the field decays. Theoretically, the net contribution of the affected grains to the magnetic moment is zero, owing to half of them being aligned one direction, and the other half being aligned in the opposite direction (Tauxe, 1998). In practice, specimens undergoing alternating field demagnetization can potentially acquire an anomalous magnetization perpendicular to the demagnetization field, skewing the measured resultant vector (Dankers and Zijderveld, 1981). The resulting spurious magnetization is named gyroremanent magnetization (GRM). In order to overcome the potential effect of GRM, specimens were sequentially demagnetized along three axes at the same level, with measurements taken between each step. The resultant vectors are later combined to produce a smooth demagnetization path. The entire process is known as AGRM – anti-gyroremanent magnetization.

**PALEOMAGNETIC ANALYSIS**

Directional components of the magnetic vectors were determined using principal component analysis (PCA) (Kirschvink, 1980). Zijderveld diagrams were used to identify
different components along a specimen's demagnetization path (Dunlop, 1979). First-removed components are defined by a straight-line path starting from the first demagnetization step of the specimen until either a change in direction occurs or the magnetization degenerates into random noise. Second removed components were defined as straight-line paths of different direction than the first removed components occurring in a higher temperature or mT range. In order to be considered for determining site means, straight-line segments had to cover at minimum four demagnetization steps and maintain an acceptable maximum angular deviation value (MAD). For specimens whose second-removed components had been determined to be affected by magnetic overprinting, plane analysis was used in an attempt to determine an acceptable direction.

Two types of line fits are available for use during PCA. Free line fits use only the points selected along a specimen's demagnetization path when calculating direction and MAD value. The overall site mean quality can be greatly improved by use of anchored line analysis during PCA, particularly for sites whose free line site mean directions are somewhat scattered. Anchored lines are the directions from the origin along the demagnetization path. This type of analysis can be deemed appropriate if the overall result is a higher quality line fit for demagnetization paths trending to the origin so long as the direction remains relatively unchanged. Anchored line analysis was used for one site in this study for a higher quality final site mean.

Declination, inclination and magnetic moment can be converted into Cartesian coordinates $X_1$, $X_2$, and $X_3$. These values are used in the orientation tensor $T$, a 3x3 matrix based on the transformed Cartesian coordinates. The largest eigenvalue $\tau_1$ corresponds to the principal eigenvector, $V_1$; $\tau_2$ corresponds to the mid-value eigenvector $V_2$; and the least eigenvalue $\tau_3$ corresponds to the least eigenvector $V_3$. The standard deviation $\sigma$, is defined by $\sigma = \sqrt{\tau}$ and the transformation from magnetic directions to Cartesian coordinates is given by

$$B = \left(\frac{X_1 + X_2 + X_3}{3}\right)$$
$$D = \tan^{-1}\left(\frac{X_2}{X_1}\right)$$
$$I = \sin^{-1}\left(\frac{X_3}{B}\right)$$
Quality of line and plane fits were assessed using MAD values. The MAD value for a line fit component is defined by:

\[ MAD_{\text{line}} = \tan^{-1}\left( \sqrt{\left(\sigma_2\right)^2 + \left(\sigma_3\right)^2} / \sigma_1 \right) \]

The eigenvector \( V_3 \) is associated with the least eigenvalue \( \tau_3 \), and can be taken as the pole to the best-fit plane in which the unknown component of magnetization must lie. The MAD angle for a plane is defined by:

\[ MAD = \tan^{-1}\left( \frac{\tau_3}{\tau_2} + \frac{\tau_3}{\tau_1} \right) \]

Magnetization components were categorized based on three parameters – 1) visual clarity of line segment trending to origin on Z-plot, 2) MAD values and 3) angle by which the line fit of the second-removed component deviates from a path to the origin on Z-plot. Only components with MAD values less than 20 and an angle to the origin less than 20° were considered for determining site means.

Site means were calculated using Fisher (1953) methods in the computer program IAPD. Non-Fisher distributions were quantified using Tauxe’s (1998) bootstrap methods.

A Fisher distribution approximates a normal distribution dispersed on a sphere. The Fisher distribution is given by:

\[ F = \left( \frac{K}{4\pi \sinh K} \right) \exp(K \cos \alpha) \]

where \( \kappa \) is the precision parameter; as \( \kappa \to \infty \), dispersion of directions \( \to \) zero. A high value for \( \kappa \) indicates tight clustering of data about a true mean direction while a \( \kappa \) value of 1 indicates a uniform distribution of data points about a sphere.

\[ k = \left( \frac{N-1}{N-R} \right) \]

where \( R \) is the resultant vector given by

\[ R = \left( \sum X_{1i} \right)^2 + \left( \sum X_{2i} \right)^2 + \left( \sum X_{3i} \right)^2 \]
and $X_{1i}$, $X_{2i}$, $X_{3i}$ are the Cartesian coordinates of the declination ($D_i$), inclination ($I_i$) of the magnetization and total field magnitude ($B$).

Quality of the site mean is reflected by the size of $\alpha_{95}$. The $\alpha_{95}$ value is an estimate of the cone of 95% confidence about the mean. High $\alpha_{95}$ values show a high uncertainty of mean. The $\alpha_{95}$ value is defined as

$$\alpha_{95} = \cos^{-1}\left[1 - \frac{N - R}{R} \left(\frac{1}{p} \left(\frac{1}{(N-1)}\right) - 1\right)\right]$$

or,

$$\alpha_{95} = \frac{140}{\sqrt{(K+N)}}$$

Final site groupings were characterized as either Fisher (1953) or non-Fisher distributed using Lisa Tauxe's (1998) $bootdi$ program. The $bootdi$ program uses bootstrap methods that generate a “paradataset” suitable for quantifying non-Fisher distributed data sets. The bootstrap software operates by randomly selecting a list of points from the existing data set, then these para-data points are used to calculate a mean direction. A total of 500 data points is drawn for the para data set. Using the default settings, the program generates 1000 para-data sets for bootstrap calculations. Variation that include 95% of the resulting directions in the generated para-dataset are reflective of the distribution of directions within the original dataset’s 95% confidence limits (Tauxe, 1998). When examining non-Fisher distributed data, the bootstrap-derived error limits are more accurate than $\alpha_{95}$ (Tauxe, 1998).

Plane analysis was used in an attempt to determine whether there was an unresolved magnetization skewing the final site mean distribution for non-Fisher data sets. Planes were anchored to the origin along the second removed component’s demagnetization path. Poles to the planes were plotted on an equal area plot. The poles to the planes have a polarity determined by the order of vector multiplication (Tauxe, 1998; R. Burmester pers. comm., 2003). Vectors are calculated from the $D$, $I$ and $M$ values (declination, inclination and magnetization) at each point along the demagnetization path. These are then transformed into three arithmetic means by user-determined endpoints. By inserting the three arithmetic means into a 3x3 matrix, the data points yield eigenvalues and eigenvectors relative to data points which are most, midways and least distributed.
The principal eigenvector corresponds to the best-fit line through the data (greatest scatter). The eigenvector associated with the least eigenvalue is taken as the pole to the best-fit plane. Planes are perpendicular to the pole defined by the least eigenvector. Polarity is thus determined by the difference between the first and last points along the demagnetization path. Using right hand rule, the thumb of the right hand points towards the pole (the least squares eigenvector), the fingers of the right hand curl in the direction of the unresolved or partially unresolved demagnetization path.

**FIELD TESTS METHODS**

**CONGLOMERATE TEST**

The conglomerate test works by taking oriented samples from conglomerate clasts in outcrop, and comparing the ancient magnetic direction of the clasts to the characteristic magnetic direction of the parent material. DRM is acquired during deposition of the conglomerate; the age of conglomerate should be contemporaneous with age of primary magnetization (McElhinny and McFadden, 2000). The conglomerate test works for conglomerate composed of the same rock whose direction is being tested. A random scattering of directions from the clasts versus a well clustered set of directions from beds of the same formation is evidence that the formation retains its primary magnetic direction.

Randomness of data for the conglomerate test should be demonstrated numerically. Watson (1956) defined the parameter $R_0 = \sqrt{7.8\left(\frac{N}{3}\right)}$.

If the value of $R$ is less than $R_0$ randomness cannot be disproved. This is beneficial for testing whether the directions isolated from conglomerate clasts are truly random.

Outcrops were sampled for a conglomerate test from intraformational brecciated clasts in the Houstenaden Creek Formation (Figure 11) and rounded conglomerate derived from the Cape Sebastian Sandstone in an olistostrome deposit of Hunter's Cove Formation (Figure 12).

**TILT TEST**

The fold test is based on the concept that when layered strata are unfolded, their magnetic directions will be similar if the magnetization pre-dates folding (Enkin and Watson, 1996, Tauxe 1998). A positive fold test is a good indication that the magnetization was acquired before folding and is possibly an original DRM. However, because folding can be a multi-phase process, a simple unfolding of beds about their strike may not reflect the true
complexity of the folds. There may be plunging folds, or multiple episodes of folding and rotation in a rock’s history. The bootstrap approach can also be applied to the fold test (Tauxe, 1998) to determine what constitutes a statistically well clustered data-set as well as the degree of unfolding at which directions most tightly cluster. The bootstrap fold test relies on the magnitude of the primary eigenvalue $\tau_1$. The tightness of the grouping is reflected in the relative magnitude of $\tau_1$. The behavior of $\tau_1$ during unfolding reveals the correction at which the directions are most closely spaced. Para-data sets drawn from the observed paleomagnetic directions are used to evaluate the behavior of $\tau_1$ as a function of tilt-correction. The data are graphically displayed showing the tightest grouping of major eigenvalues and 95% confidence intervals for representative para-data sets against percent unfolding. A fold test result whose 95% confidence interval overlaps 100% and not 0% is considered positive.

Without extensive field work it may be impossible to distinguish multiple folding or tilting and rotation events. Many of the coastal exposures of rocks are very complicated, commonly having overturned beds, different ages of folding, and poor outcrop exposure. It would be more accurate in this case to refer to the fold test for the Houstenaden Creek Formation as a “tilt test”. For this “tilt test” possible multiple phases of folding or rotations on faults cannot be accounted for, and the beds are untilted about the axis of their strike. A positive tilt test result is better clustering of directions when the layers are untilted rather than in-situ. Samples were collected from folded beds for a fold test from the Hunter’s Cove Formation and tilted beds from the Houstenaden Creek Formation.

**REVERSALS TEST**

The Earth’s magnetic field is known to reverse polarity. Rocks magnetized over a sufficiently long time can record both polarities and thus one can test for ancient magnetization by comparing directions from rocks of the same formation that have different polarities. By flipping the direction of one data-set to its antipode, we can see if it shares a common mean with the data-set having the opposite polarity. When these two data-sets come from rocks of the same formation, we have what is known as a reversals test. In a positive reversals test the two means will be indistinguishable when compared to each other in this manner.
The bootstrap can be applied to the reversals test for determining if the two data sets are distinct. When intervals of the cartesian coordinates of each data set are plotted with their respective 95% intervals, overlap of the intervals indicates the data cannot be distinguished at the 95% confidence level, and hence share a common mean. This is especially useful in cases where the mean direction of each data set lies outside the $\alpha_{95}$ confidence region of the other, yet the confidence regions overlap when viewed on an equal-area plot.

A bootstrap reversals test was done for samples taken from the Houstenaden Creek Formation.

**INCLINATION SHALLOWING METHODS**

Bedding planes of sedimentary rocks supply paleohorizontal so that inclination of magnetization relative to bedding can be used to calculate paleolatitude. However, using the paleomagnetic inclination obtained from sedimentary rocks is complicated by possible inclination flattening (Deamer and Kodama, 1990). Inclination flattening occurs naturally in some sedimentary rocks resulting from the electrostatic attraction between a boundary layer of ions surrounding polar clay particles and the minute magnetite grains aligning with the earth's magnetic field at time of deposition (Anson and Kodama, 1987). The result of this attraction is a mechanical rotation of acicular magnetite grains into the plane of the clay particle, thus shallowing inclination and producing a magnetic fabric mimicking platy clay mineral fabrics (Kodama and Sun, 1990). This process becomes more pronounced as burial depth (strain from compaction) increases and/or clay content increases (Kodama and Sun, 1990). To test for inclination shallowing a comparison of rock fabric (in this case, using the KLY-3 Kappabridge to measure the rock's AMS) to the inclination of the magnetic vector characteristic of the rock's ancient magnetization is made. If a positive correlation is established, a compaction correction can be used following the methods Hodych et al., (1999).

Comparison of minimum susceptibility axis over the maximum susceptibility axis ($K_{min}/K_{max}$ – garnered from AMS data) to tangent of inclination from second-removed components was used to estimate inclination shallowing. The comparison works best for specimens dominated by multi-domain or pseudo-single domain grain size of magnetic minerals (Hodych et al., 1999). This comparison is only valid for second removed
components reflective of the magnetic field at the time of deposition (depositional remanent magnetization, or DRM) or second removed components reflective of the magnetic field shortly after deposition (post-depositional remanent magnetization, or pDRM) and if the AMS fabric is determined to be unaffected by tectonic forces.

**ROCK MAGNETIC PROPERTIES METHODS**

Two methods were used in an attempt to determine magnetic mineralogy in several specimens. The thermomagnetic properties of samples from different formations were measured on the KLY-3 Kappabridge. This low-field susceptibility experiment is intended to indicate the presence of magnetic minerals based on Curie temperature. The other method used to determine magnetic mineralogy is thermal demagnetization of multi-component isothermal remanent magnetization (IRM), otherwise known as the Lowrie method (Lowrie, 1990).

High temperature susceptibility experiments performed with the furnace attached to the KLY-3 Kappabridge susceptometer were used to help determine ferromagnetic mineralogy. Due to the low field that the specimens are exposed to with this machine, only magnetic minerals with a low coercivity, such as multi-domain magnetite are revealed by this method. This provides an incomplete picture, so another method that can be used to detect high-coercivity magnetic minerals such as pyrrhotite is needed. This requirement is met with the Lowrie (1990) method.

The Lowrie test requires that the sample’s mineralogy has not been altered by thermal unblocking experiments. Due to the nature of thermal demagnetization – i.e., the sample being exposed to high temperatures possibly in excess of 600° C – it is probable that secondary (magnetic) mineral growth will occur during the heating/cooling process (R. Burmester, pers. comm., 2003). Occurrence of secondary mineral growth renders thermally demagnetized specimens useless for tests that detect original magnetic mineralogy in the sample. However a specimen fully demagnetized by application of alternating field technique is suitable for testing with the Lowrie method. The demagnetized specimen is subjected to an applied magnetic field along the core’s z, y, and x axes. Each axis is sequentially magnetized in progressively lower magnetic fields (Lowrie, 1990). The specimen then undergoes stepwise thermal demagnetization and measurement (Lowrie,
The data, when plotted, show thermal unblocking of magnetization held by grains with different coercivities (Lowrie, 1990).
RESULTS

ROCK MAGNETIC PROPERTIES

Thermomagnetic testing on the KLY-3 Kappabridge detected magnetite in all samples tested (Figure 17). Due to the low strength of the induced magnetic field by the instrument, the higher coercivity mineral pyrrhotite could not be detected by this test.

Samples from all three formations were tested for magnetic mineralogy by use of the Lowrie method. The majority of the magnetization was carried in the intermediate coercivity range along the y-axis (saturated in a 0.4 T magnetic field), consistent with magnetite. A drop in magnetization was very pronounced at 320° C for all specimens tested except those originating from sites 00KH1 and 00KH3 (Figure 18). All sites showed a noticeable drop in magnetization beginning about 560° C. These results are consistent for the presence of pyrrhotite (320° C Curie temperature) (Dekkers, 1983) and magnetite (580° C Curie temperature) (Dunlop and Ozdemir, 1997).

LOW TEMPERATURE TREATMENT

Most specimens cycled through LTD show a significant drop in magnetization upon their first immersion in liquid nitrogen, and markedly less drop upon a second immersion. Magnetization losses varied, ranging from 20% to 80% of original magnetization. Specimens with upward inclinations exhibited an increase in magnetization upon cycling through LTD, presumably due to partial unblocking of the first removed, downward polarity component.

Specimens from this study experiencing a loss of greater than 50% of NRM intensity upon cycling through LTD were generally poor carriers of remanent magnetization. Most specimens experiencing this significant drop did not resolve a vector at demagnetization steps higher than ~240° C. This result is probably due to a significant proportion of multidomain-grained magnetite present in the specimens, as evidenced by the large drop in intensity in these specimens after LTD (Dunlop and Ozdemir, 1997). LTD appears a good indicator for probable magnetic behavior upon thermal demagnetization of specimens.

PALEOMAGNETIC RESULTS

Thermal demagnetization behavior varied. Some specimens were fully demagnetized by temperatures as low as 350° C. In general, specimens that had a weak initial magnetization demagnetized at lower temperatures.
Most specimens heated past 450° C exhibited fluctuations in magnetization and upward spikes in susceptibility. Increases in susceptibility indicate chemical changes in the specimen (Schmidt, 1992). Increases in magnetization at high temperatures show acquisition of a new magnetism by the specimens: possibly conversion of pyrrhotite to magnetite in the 450° temperature range (Van Velzen and Zijderveld, 1992), or oxidation of iron-bearing silicates upon heating.

All three formations displayed first-removed components, defined by temperature steps up to about 240° C. Most directions cluster well when viewed in geographic coordinates. The first components appear to reflect the recent magnetic field at their location. Scatter of these directions significantly increases upon tilt correction, which indicates the first-removed components are a recent overprint, and not characteristic of the rock's DRM (Figure 19) (Zijderveld, 1967).

**CAPE SEBASTIAN and HUNTER'S COVE FORMATION**

Specimens demagnetized from the Cape Sebastian Sandstone and Hunter's Cove Formation could not be used in determining a paleolatitude for this study. Specimens from these formations overwhelmingly produced poor to unresolvable components past about 240° C (Figures 20, 21).

**HOUSTENADEN CREEK FORMATION**

Sites used in final analysis for this study all come from the Houstenaden Creek Formation. Specimens from Houstenaden Creek Formation tended to have stronger initial magnetization than the Cape Sebastian Sandstone and Hunter's Cove Formations, and thermal demagnetization generally produced high-quality two component demagnetization paths. Generally, specimens from the Burnt Hill Cove locality (00KHU sites) had steady decreases in magnetization during stepwise thermal demagnetization (Figure 22). Sites 00KH1 and 00KH3, from the Landslide Bay locality, had reverse-polarity second-removed components, and displayed initial increases in magnetization to ~240° C, corresponding to loss of the first removed, normal polarity component. A drop in magnetization was observed to coincide with unblocking of the reverse polarity, second-removed component (Figure 23).

Results of principal component analysis for second removed components produced two categories of data; sites 00KH1 and 00KH3 that produced inclination up (Figure 24), Fisher-distributed data sets, and sites 00KHU1 through 00KHU5 that have an inclination
down, non-Fisher distributed data set (Figure 25). The presence of both normal (downward) and reverse (upward) components of magnetization ensures that enough stratigraphic section was sampled to average out the effects of paleosecular variation (Van der Voo, 1989).

The final streaked distribution of second-removed components from the sites of the Burnt Hill Cove locality (00KHU) necessitated the use of plane analysis to infer a demagnetization endpoint. The results of plane analysis on site 00KHU4 indicate a clearly preferential demagnetization path towards a declination \( \sim 105° \), and an inclination \( \sim 62° \) (Figure 26). Plane analysis for sites 00KHU1, 2, and 5 were less certain for a preferential direction, but overall a trend towards more easterly and steep inclinations could be resolved. Therefore, specimens at the high end of the streaked pattern with steeper inclinations are interpreted to yield the most likely demagnetization endpoint.

From the Landslide Bay locality two sites (00KH1 and 00KH3) produced second-removed components that were useful in attaining a final site mean. The final Landslide Bay site mean has a declination of 248°, and an inclination of -52°. Results from these sites were used to establish displacement and rotation of the entire terrane with respect to the North American craton following the techniques of Beck (1989) and Butler (1990). The Late Cretaceous reference pole was taken from Diehl (1991) and Gunderson and Sherriff (1991). Results after tilt-correction indicate translation of 1650 +/- 500 km coordinates, and a clockwise rotation of 122° (Table 1).

**FIELD TESTS RESULTS**

**CONGLOMERATE TEST**

The conglomerate test for the Cape Sebastian Sandstone was positive. Magnetic components removed from Cape Sebastian conglomerate in Hunter's Cove Formation (site 00KHC3) scatter widely, indicating the clasts retain their original magnetization. Two conglomerate clasts sampled had visible bedding. Specimens from both samples have very well defined directions, and have an inclination of 50°. Unfortunately, the rest of the samples tested from the Cape Sebastian Sandstone did not yield a usable magnetic vector for a comparison, so the conglomerate test was not useful for this study.

Samples taken from a breccia deposit at site 00KHU3 of the Houstenaden Creek Formation were also used in a conglomerate test. The second-removed components from
clasts at this site are loosely clustered about a shallowly down and easterly direction. The directions at this site fail Watson's (1956) test for randomness and hence fail the conglomerate test (R value of 8.26 which is greater than the Ro value of 4.84). The low scatter shows either a type of overprint dominates the rock's NRM or low transport of clasts. AMS minimum axes for this site are not scattered, and show a northwest trending girdle. One would expect a conglomerate to have randomly distributed AMS fabric as a proxy for the clast distribution, instead the girdle distribution of mineral fabrics measured by AMS could reflect folding of the clasts about a NW axis – consistent with the dominant folding structures throughout coastal northern California and southern Oregon. An attempt at uncovering a magnetic direction was made using pole to plane analysis for this site. The results of plane analysis show randomly oriented magnetic vectors from the conglomerate clasts hinting at the potential for a positive conglomerate test that has been obscured, either by the nature of the clast transport or by a spurious magnetic overprint (Figure 27).

**TILT TEST**

A bootstrap tilt test was performed with specimens from the Burnt Hill Cove locality (00KHU) whose suitability was determined by pole to plane analysis and specimens from site 00KH1 of the Landslide Bay locality. When both sets of data are corrected for bedding, the greatest eigenvector and its associated paradata sets show minimum dispersion between 72% and 102% unfolding at the 95% confidence level. Because a portion of the tilt test overlaps at 100% unfolding, the tilt test is considered positive (Figure 28) (Enkin and Watson, 1996), and the rock's magnetization can be dated as at least prior to folding of the beds.

**REVERSALS TEST**

A simple reversals test was performed by flipping the directions from site 00KH1 to their antipode to see if they overlap the directions from the specimens of the Burnt Hill Cove locality used in the tilt test. The $\alpha_{95}$ cones of confidence for each data set used in this simple reversals test overlap, but the means of each site do not fall in the same $\alpha_{95}$ cone of confidence. In this case, one has to use a statistical test to tell if the data sets are distinct or whether they share a common mean. Data sets that can be shown to share a common mean can be said to have passed the reversals test. Using the bootstrap reversals test (Tauxe, 1989), the two data sets were shown to share a common mean at the 95%
confidence interval, and therefore these specimen show a positive reversals test (Figure 29).

**MAGNETIC FABRIC RESULTS**

Flinn (1962) plots were used to determine the characteristic ellipsoid shape of magnetic fabric for sites that produced useful two-component demagnetization paths. Most of those specimens have AMS ellipsoids that are oblate; the exceptions are from site 00KH1, whose specimens plot mostly in the prolate field (Figure 30).

AMS fabric for all sites in each formation was quantified using Tauxe’s *bootams* (1998) and plotted using Tauxe’s *plotams* (1998) software (Figure 31). AMS fabric from the Houstenaden Creek Formation was compared to a regional fold axis determined by bedding plane intersection from bedding measured in the field from both the Landslide Bay locality and the Burnt Hill Cove locality. AMS fabric seems to not correlate with regional folding in sites collected at Burnt Hill Cove (00KHU1-6) or at Landslide Bay (Figure 32). This suggests that the AMS fabric has a depositional origin in samples collected at Landslide Bay and Burnt Hill Cove.

Following successful isolation of paleomagnetic directions from the various Houstenaden Creek Formation specimens, correction of paleoflow directions for post-Cretaceous rotations was attempted. The total AMS declination correction of 102° counterclockwise rotation was arrived at by using a mean declination from second-removed components of all specimens used in the tilt and reversals tests, including the antipodal directions of the reverse polarity specimens.

Tilt and rotationally corrected AMS plots for sites 0KH1 and 00KH3 show a girdling of minimum axes to the west. This indicates a westerly directed paleoflow (Figure 33). In these sites I interpreted grain shape to be rollers either on a slope, or pushed by current flow. This interpretation is supported by a shape anisotropy dominated by oblate or prolate fabric at these sites and stratigraphic interpretation that a high-flow regime of coarse-grained Bouma A sequence sandstone acts as a traction carpet while being deposited (McElhinny and McFadden, 2000). Rotationally corrected paleoflow directions for sites from the Burnt Hill Cove locality are also westerly in direction, with a skewing of minimum axes in the direction of transport (Figure 33). Flinn plots for these sites show a mostly oblate fabric interpreted as a traction carpet deposit.
INCLINATION SHALLOWING RESULTS

The only sites that can be shown to have second-removed components free of extraneous magnetization and magnetic fabrics that are unaffected by folding are 00KH1 and 00KH3. Because of this, demagnetization paths and AMS data from these two sites were deemed appropriate for estimation of inclination shallowing following the methods of Hodych et al. (1999) using the second-removed magnetic vector from these rocks. Data from these sites show only a weak correlation of fabric to inclination. The $R^2$ value from the inclination correction of 0.0546 is indicative of only a very poor correlation between inclination and rock fabric. This result implies that there is minimal inclination shallowing in these rocks, with an inclination corrected to $-53.9^\circ$ from $-52.8^\circ$ (Figure 34).
DISCUSSION

Insights from rock magnetic testing allow some interesting speculations when applied to the demagnetization results. For sites 00KH1 and 00KH3 in which the Lowrie-Fuller test revealed only magnetite (Figure 18), the overall demagnetization quality of the specimens was good (Figure 23), the specimen directions within these sites were Fisher distributed (Figure 24), and the sites did not seem to contain or be influenced by an unresolved component of magnetization skewing their final distribution. For sites 00KHU1 to 00KHU5, which had clearly delineated demagnetization paths on a per specimen basis (Figure 22), but whose final site distributions were skewed by an unresolved component of magnetization (Figure 25), pyrrhotite was strongly detected along with magnetite (Figure 18). Similarly, for sites from the Hunter’s Cove Formation and the Cape Sebastian Sandstone, whose results were too poor to resolve a second-removed component (Figures 20, 21), pyrrhotite was indicated by the Lowrie (1990) test (Figure 18). Additionally, specimens taken from these sites had an initially weak magnetization prior to thermal cleaning. It therefore appears that the presence of pyrrhotite, commonly the result of secondary mineral growth, tends to obscure partially or completely the presumably ancient direction carried by magnetite in these rocks. Based on this reasoning, it seems likely that the second removed components from sites 00KH1 and 00KH3 are the most likely candidates for preserving an ancient direction in these rocks.

The multitude of specimens from the Cape Sebastian Sandstone and Hunter’s Cove Formation never resolved a definable ancient magnetic component. The wave-agitated zone from which the Cape Sebastian Sandstone derives does not seem suitable for the settling out from suspension of fine-grained magnetite necessary for preserving a stable record of remanence. The coarse-grained sandstones of this study and other studies seem to give consistently poor results (Donohoo, 2003; Heim, 2004). The fine-grained, outer shelf to abyssal plain deposits of the Hunter’s Cove Formation also gave very poor results, presumably because of its characteristically extremely weak magnetization. The Houstenaden Creek Formation of Bourgeois (1980) gave a low return of results for the effort, although it seems the ancient magnetic direction for the Campanian is preserved in at least some of these rocks. The Houstenaden Creek Formation samples are coarser grained than the Hunter’s Cove samples, and finer grained than the Cape Sebastian Sandstone.
There are two reasons to suspect that there is possibly some remagnetization of specimens from the Burnt Hill Cove locality (00KHU). The result of the conglomerate test for site 00KHU3 is inconclusive (Figure 27). Scattering of second-removed components at this site is too low to be called random, yet some visible scattering exists. This low scattering could be an artifact of deposition, as shown by AMS fabric. Plane analysis shows randomly oriented vectors trending from second-removed components at this site. Second-removed components from sites 00KHU1-5 demonstrate a streaked distribution on an equal area plot indicating a mix of overlapping magnetizations (Tauxe, 1998; Hagstrum and Jones, 1998). It appears there is a component of overlapping magnetization skewing the magnetizations of the normal polarity sites of the Houstenaden Creek Formation. Results of plane analysis on non-Fisher distributed sites were used for a successful tilt test and reversals test, indicating that an ancient magnetization is at least partially preserved at this locality.

The magnetic polarity of the 00KH and 00KHU sites is consistent with the biostratigraphic dating of the formation as Campanian. The magnetic field was reversed from 83 Ma to 79 Ma, and continued as normal polarity into the Maastrichtian. This result confirms Blake's (1985) assessment of the formation as being wholly Campanian in age, rather than the basal section being Albian, as the Albian is a time that spans the Cretaceous normal superchron, and has no observed reversed polarity during that time (Cande and Kent, 1995).

After investigating if the AMS fabric had been affected by folding, and determining that it had not, paleoinclination shallowing and paleoflow estimates could be obtained. Without testing for paleoinclination shallowing, the estimate of terrane transport has suspect validity (Kodama, 1997). The methods of Hodych et al. (1999) seemed most appropriate for this study, as the majority of second-removed components were arrived at by thermal demagnetization. The inclination shallowing methods developed by Kodama have relied on disaggregating samples and realigning the diasaggregated material in a strong magnetic field during compaction followed by alternating field demagnetization for attaining magnetic components used in his comparisons (Jackson et al., 1991; Kodama and Davi, 1995; Kodama, 1997). The use of AMS rock fabric for comparison to thermally demagnetized second-removed components is both simpler and less time consuming, and the results are shown to be effective (Hodych et al. 1999). The paleoinclination shallowing test for this
study showed only a weak correlation between mineral fabrics and inclination. When applied to the tilt-corrected paleolatitude determination for these sites, the AMS correction has a calculated paleolatitude of 34° +/- 3.8° (Table 2). This inclination-shallowing corrected latitude is virtually indistinguishable from the un-corrected paleolatitude, and places the Gold Beach terrane to the south by ~1550 +/- 500 km during the late Cretaceous.

Paleoflow directions were estimated for sites from the Houstenaden Creek Formation using final site means to rotate the tilt-corrected AMS data to their Cretaceous age orientations. When reconstructing the terrane to the south, the Gold Beach terrane lies outboard of the North American craton. If, as Bourgeois hypothesized, the upper Cretaceous units of the Gold Beach terrane had a continental source for their sediments (Bourgeois, 1980) then the terrane should have west directed paleocurrents when corrected for their Cretaceous age declination. The tilt and declination corrected AMS fabric shows a westerly paleoflow direction, supporting this model.

As an additional exercise, the paleomagnetic grand site mean used by Blake et al. (1985) from the Jurassic Otter Point Formation was recalculated with a new Jurassic paleopole from the North American craton (Beck and Housen, 2003). This result places the Gold Beach to the south of its present location by nearly 2100 km (Figure 35). Without the original data, it is impossible to account for inclination shallowing, so this estimate should be regarded with some caution. However, the results of this recalculation add internal consistency to this reconstruction.

Plate motion models derived by Engebretson et al. (1984) for the North American plate relative to the Kula/Farallon triple junction support a mechanism for this proposed northward translation during the late Cretaceous through the early Eocene (Figure 36) (Engebretson et al., 1984; Jurdy, 1984). When compared to the plate motion rates and expected displacements of terranes riding on the Kula and Farallon plates, the translation estimate of 1550 km between 83 Ma and 50 Ma supports the Kula plate as the most likely plate for northward translation of the Gold Beach terrane (Table 3).

Finally, restoring the Nacimiento block to its tentative place of origin 200 km to the south of its present position places it at just north of the Viscaino Peninsula, in Baja, Mexico (Jayko et al., 1993) (Figure 16). Bourgeois (1980) linked the upper Cretaceous units of the Gold Beach terrane to the Nacimiento block. The results of this study indicate that
during the late Cretaceous, the Gold Beach terrane was in the present-day vicinity of the Viscaínó peninsula, supporting both the hypotheses of Bourgeois (1980) and Jayko et al. (1993) for the Late Cretaceous latitude of the Gold Beach terrane. Using the results of this study, the evidence appears to rule out the Seiders and Blome (1987) model of an essentially in-situ location for the Gold Beach terrane, and supports the models of Bourgeois (1980), Blake et al., (1985), and Jayko et al. (1993) for a southerly latitude of origin.

The undifferentiated Mio-Pliocene Empire Formation deposits of Diller (1907) overlap the Gold Beach terrane constraining its time of arrival to at least Miocene. There are, however, several neighboring late Cretaceous terranes (such as the Sixes River, Yolla Bolly, Pickett Peak and Snow Camp terranes) that are overlain by the mildly deformed mid-Eocene Lookingglass Formation (Blake et al., 1985). Thus it seems likely that the Gold Beach terrane was emplaced at its present location with other coastal late Jurassic and Cretaceous terranes by about 50 Ma.
CONCLUSIONS

The results of this study bracket the Gold Beach terrane to a paleolatitude of 34° ±3.8° during the Campanian. This puts the upper Cretaceous units of the Gold Beach at a paleogeographic origin in present-day central Baja, Mexico, near the Viscaino Peninsula, supporting the models of Bourgeois (1980) and Jayko et al. (1993). Using plate motion models derived by Engebretson et al. (1984), the Gold Beach most likely translated northward on the Kula plate to its present location, and was emplaced by about 50 Ma (Blake et al., 1985). The entire package has subsequently been slightly offset and folded into its current lensoidal map pattern by faulting along the Pistol River shear zone during the Cenozoic (Dott, 1971; Bourgeois, 1980). This research indicates the Gold Beach terrane has undergone 1550 km+/- ~500 km of northward transport since the Late Cretaceous.
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Figure 1: Location and local terrane map of area surrounding Gold Beach terrane. Franciscan terranes are Jura-Cretaceous, Gold Beach is Upper Cretaceous overlying Upper Jurassic. Adapted from Aalto, 1990.
Figure 2: Conflicting origins for the Gold Beach terrane based on studies of Bourgoies (1980) connecting the Gold Beach to the Nacimiento block and Seiders and Blome (1987) connecting the Gold Beach to the Klamath block.
Figure 3: Geologic map of Cape Sebastian area adapted from Bourgeois (1980) and Dott (1971). Gold Beach and Blacklock Point are north of map area.
Quaternary

- QAL
  River alluvium

- QLS
  Major landslides

- PMT
  Pleistocene marine terrace deposits

Cretaceous

- KHC
  Hunter's Cove Formation
dominantly shale with thin sandstone beds; local lenses of coarse sandstone and conglomerate up to 30 feet thick

- KCS
  Cape Sebastian Sandstone
cross stratified, massive fine sandstone with local coarse conglomerate

- KHU
  Houstenaden Creek Formation
  interbedded shales and fine sandstones coarsening upwards with increasing volcanic and intraformational conglomerate

Jurassic

- JOP
  Otter Point Formation
  Black mudstone, graded sandstone, massive pebble conglomerates, bedded chert

- JOP
  Otter Point Formation
  Undifferentiated igneous rocks of Jurassic to Tertiary age

Faults

Dashed where approximately located or indefinite

Figure 3 (continued): Map key for geologic units
Figure 4: Stratigraphic section of rocks exposed in the Cape Sebastian area. Sampling sites approximately placed with figure numbers of respective photos. Location of 79 Ma approximate based on paleomagnetic results. Modified from Bourgeois (1980) and Bourgeois and Dott (1985).
Figure 5: Depositional and tectonic model for placement of the Upper Cretaceous formations over the Otter Point Formation in an environment of relative sea level changes and active vertical faulting. Adapted from Bourgeois (1980).
Figure 6: Sampling in Houstenaden Creek Formation, interbedded Bouma A and D sequence stratigraphy, Landslide Bay Locality, site 00KH3.
Figure 7: Outcrop at site 00KHU1. Lithic graywacke, Landslide Bay Locality. View looking Northwest.
Figure 8: Site 00KHU3. Intraformational conglomerate sampled for conglomerate test in Houstenaden Creek Formation. View to northwest.
Figure 9: Cape Sebastian Sandstone, sites 00KCS1,2, and 4. Hummocky cross-bedded sandstone records deposition of material to depth of storm wave base. Stratigraphically overlies the Houstenaden Creek Formation.
Figure 10: Outcrop at site 00KHC2. Close-up photo of Bouma C fluid escape structures and Bouma B laminated sands. Quarter for scale on bottom left.
Figure 11: Outcrop of site 00KHC3. Channel scour and fill in Hunter's Cove Formation. Clasts are Cape Sebastian Sandstone. Clasts sampled for conglomerate test of Cape Sebastian Sandstone. Staff for scale (1.5 meters).
Figure 12: Block model of deposition of Cape Sebastian Sandstone derived slump breccia onto Hunter's Cove Formation slope. Adapted from Bourgeois (1980).
Figure 13: Distal turbidite facies of the Hunter's Cove Formation. Locations of sites 00KHC4-8. Photo faces east.
Figure 14: Trace fossil burrows on underside of fine-grained sandstone beds from the Hunter's Cove Formation. Quarter for scale.
Figure 15: Quartz-Feldspar-Lithic diagram showing affinity of late Cretaceous Gold Beach terrane to Nacimiento Block. Diagram adapted from Bourgeois (1980).
Figure 16: Map showing proposed restored origins of Coast Range terranes. Blue arrow shows restored position of Nacimiento block, red arrow shows restored position of Gold Beach terrane (after Jayko and Blake, 1993). Green arrow shows restored origin of Gold Beach terrane from Seliders and Blome (1987).
Figure 17: Sample Kh 1-1, thermomagnetic curve and Curie temperature estimation by use of second derivative maximum. Curie temperature estimation made using Tauxe's *curie.exe* program (1998). Curie point indicates presence of magnetite.
A. Heating and cooling curves. Sample heated and cooled in Argon atmosphere to retard oxidation. Hopkinson's peak at 500 degrees typical of fine-grained magnetite. Magnetization versus temperature in degrees Celsius. Cooling curve shows increase in overall magnetization, indicating thermal alteration. B. First derivative of data from heating and cooling curves. C. Second derivative of data from heating and cooling curves. Spike represents estimated Curie temperature of 558 degrees, indicating magnetite.
Fig. 17 cont'd: Sample Kh 3-7, thermomagnetic curve and Curie temperature estimation by use of second derivative maximum. Curie temperature estimation made using Tauxe's *curie.exe* program (1993). Curie point indicates presence of magnetite.

A. Heating and cooling curves. Sample heated and cooled in Argon atmosphere to retard oxidation. Magnetization versus temperature in degrees Celsius. Cooling curve shows increase in overall magnetization, indicating presence of thermal alteration.

B. First derivative of data from heating and cooling curves.

C. Second derivative of data from heating and cooling curves. Spike represents estimated Curie temperature of 558 degrees, indicating magnetite.

A. Heating and cooling curves. Sample heated and cooled in Argon atmosphere to retard oxidation. Hopkinson's peak at 500 degrees typical of fine-grained magnetite. Magnetization versus temperature in degrees Celsius. Cooling curve shows increase in overall magnetization, indicating thermal alteration.

B. First derivative of data from heating and cooling curves.

C. Second derivative of data from heating and cooling curves. Spike represents estimated Curie temperature of 558 degrees, indicating magnetite.
Fig. 17 cont’d: Sample 00Khu2-10, thermomagnetic Curie temperature estimation, by use of second derivative maximum. Curie temperature estimation made using Tauxe’s curie.exe program (1998). Curie point indicates presence of magnetite.

A. Heating and cooling curves. Sample heated and cooled in Argon atmosphere to retard oxidation. Magnetization versus temperature in degrees Celsius. Cooling curve shows increase in overall magnetization, indicating presence of thermal alteration. Note strong initial intensity.

B. First derivative of data from heating and cooling curves.

C. Second derivative of data from heating and cooling curves. Spike represents estimated Curie temperature of 563 degrees, indicating magnetite.
Figure 17 cont'd: Sample 00Khu 4-6, thermomagnetic Curie temperature estimation, by use of second derivative maximum. Curie temperature estimation made using Tauxe's curie.exe program (1998). Curie point indicates presence of magnetite.

A. Heating and cooling curves. Sample heated and cooled in Argon atmosphere to retard oxidation. Magnetization versus temperature in degrees Celsius. Cooling curve shows increase in overall magnetization, indicating presence of thermal alteration.

B. First derivative of data from heating and cooling curves.

C. Second derivative of data from heating and cooling curves. Spike represents estimated Curie temperature of 555 degrees, indicating magnetite.
Figure 17 cont'd: Sample 00Khu 5-6, thermomagnetic Curie temperature estimation, by second derivative maximum. Curie temperature estimation made using Tauxe's curie.exe program (1998). Curie point indicates presence of magnetite.
A. Heating and cooling curves. Sample heated and cooled in Argon atmosphere to retard oxidation. Magnetization versus temperature in degrees Celsius. Cooling curve shows increase in overall magnetization, indicating presence of thermal alteration.
B. First derivative of data from heating and cooling curves.
C. Second derivative of data from heating and cooling curves. Spike represents estimated Curie temperature of 562 degrees, indicating magnetite.
A. Specimen from site 00KH1. Gradual steady decrease in magnetic intensity with increasing temperature, until 550°, where sudden drop in intensity indicates loss of magnetism within magnetite unblocking temperature range. Most of the magnetization is in the intermediate coercivity range, consistent with magnetite.

B. Specimen from site 00KH3. Gradual steady decrease in magnetic intensity with increasing temperature, until 550°, where sudden drop in intensity indicates loss of magnetism within magnetite unblocking temperature range. Most of the magnetization is in the intermediate coercivity range, consistent with magnetite.

Figure 18: Lowrie plots for specimens from sites 00KH1 and 00KH3.
Figure 18 cont’d: Lowrie plots for specimens from sites 00KHU1 and 00KHU2.
A. Specimen from site 00KHU1. Drop in intensity of intermediate coercivity component at 320 indicates loss of magnetism within pyrrhotite unblocking temperature range. Sharp loss of magnetism is apparent for magnetite starting at 550.
B. Specimen from site 00KHU2.
Figure 18 cont'd: Lowrie plots for specimens from sites 00KHU4 and 00KHU5.
A. Specimen from site 00KHU4. Drop in intensity of intermediate coercivity component at 320°C indicates loss of magnetism within pyrrhotite unblocking temperature range. Sharp loss of magnetism is apparent for magnetite starting at 550°C.
B. Specimen from site 00KHU5.
Figure 18 cont’d: Lowrie plots for specimens from sites 00KCS4 and 00KHC2.
A. Specimen from site 00KCS4. Drop in intensity of intermediate coercivity component at 320°C indicates loss of magnetism within pyrrhotite unblocking temperature range. Sharp loss of magnetism is apparent for magnetite starting at 550°C.
B. Specimen from site 00KHC2.
Figure 19: Equal area plots of first removed components for all sites.
A. First removed components for all sites in in-situ coordinates
B. Mean direction of first removed component
C. Scatter upon correction for tilt demonstrates nature of component as overprint.
Figure 20: Two examples of typical demagnetization paths from the Cape Sebastian Sandstone. First-removed component labeled as C1 in each plot. a) Zijderveld plot of specimen 00KCS1-1a. b) Equal area projection of demagnetization data of specimen 00KCS1-1a. c) NRM intensity decay during demagnetization for specimen 00KCS1-1a. Solid line is NRM decay, dashed line is decay of the vector difference sum. d) Zijderveld plot of specimen 00KCS2-14a. e) Equal area projection of demagnetization data of specimen 00KCS2-14a. f) NRM intensity decay during demagnetization for specimen 00KCS2-14a.
Figure 21: Two examples of typical demagnetization paths from the Hunter's Cove Formation. First-removed component labeled as C1 in each Zijderveld plot. a) Zijderveld plot of specimen 00KHC2-11b. This specimen was from coarse-grained sandstone. b) Equal area projection of demagnetization data of specimen 00KHC2-11b. c) NRM intensity decay during demagnetization for specimen 00KHC2-11b. Solid line is NRM decay, dashed line is decay of the vector difference sum. d) Zijderveld plot of specimen 00KHC5-3a. Specimen from fine-grained distal turbidite sandstone bed. e) Equal area projection of demagnetization data of specimen 00KHC5-3a. f) NRM intensity decay during demagnetization for specimen 00KHC5-3a.
Figure 22: Two examples of typical demagnetization paths from the Burnt Hill Cove locality of the Houstenaden Creek Formation. First-removed component is labeled as C1 and second removed component is labeled as C2 in each plot. a) Zijderveld plot of specimen 00KHU2-11b. b) Equal area projection of demagnetization data of specimen 00KHU2-11b. c) NRM intensity decay during demagnetization for specimen 00KHU2-11b. Solid line is NRM decay, dashed line is decay of the vector difference sum. d) Zijderveld plot of specimen 00KHU4-1a. e) Equal area projection of demagnetization data of specimen 00KHU4-1a. f) NRM intensity decay during demagnetization for specimen 00KHU4-1a.
Figure 23: Two examples of typical demagnetization paths from the Landslide Bay locality of the Houstenaden Creek Formation. a) Zijderveld plot of specimen 00KH1-10b. b) Equal area projection of demagnetization data of specimen 00KH1-10b. c) NRM intensity decay during demagnetization for specimen 00KH1-10b. Solid line is NRM decay, dashed line is decay of the vector difference sum. Notice increase in NRM intensity corresponds to removal of first-removed normal-polarity component. d) Zijderveld plot of specimen 00KH3-4a. e) Equal area projection of demagnetization data of specimen 00KH3-4a. f) NRM intensity decay during demagnetization for specimen 00KH3-4a. First-removed component is labeled as C1 and second removed component is labeled as C2 in each plot.
Figure 24: Site means of anchored line and free line analysis and associated alpha-95 cones of confidence on equal area plots for sites 00KH1 and 00KH3. a) 00KH1 anchored line site means in in-situ coordinates b) 00KH1 free line site means in in-situ coordinates c) 00KH1 anchored line site means in tilt-corrected coordinates d) 00KH1 free line site means in tilt-corrected coordinates. e) 00KH3 anchored line site means in in-situ coordinates f) 00KH3 free line site means in in-situ coordinates g) 00KH3 anchored line site means in tilt-corrected coordinates h) 00KH3 free line site means in tilt-corrected coordinates. See Table 1 for relevant statistical parameters and site mean Declination and Inclination.
Figure 24 (contd.)
Figure 25: In-situ and tilt-corrected second removed components from sites Khu1, Khu2, Khu4, and Khu5.
A. Distribution of second-removed components from sites Khu1, Khu2, Khu4 and Khu5 in in-situ coordinates.
B. Distribution of second-removed components from sites Khu1, Khu2, Khu4 and Khu5 in tilt-corrected coordinates.
Figure 26: Results of pole to plane analysis for site 00Khu4. Convergence of arrows show preferred demagnetization endpoint. Refer to text for details of analysis.
Figure 27: Equal area projection of second-removed components from breccia clasts sampled for conglomerate test at site 00KHU3 and associated alpha-95 circle of confidence. Site fails Watson's (1956) test for randomness. If the value of R is less than the parameter $R_o$ ($R_o = (7.813^*N/3)$), then randomness cannot be disproved. For this site, $R_o = 4.84$ and $R = 8.26$. Because $R > R_o$, this site fails Watson's (1956) test for randomness.

a) Dispersion of second-removed components on a per-specimen basis from site 00KHU3 in in-situ coordinates. b) Associated alpha-95 circle of confidence ($\alpha_{95}$ value of 16.4, $K$ value of 10.85). c) Results of pole to plane analysis from site 00KHU3. Arrows show randomly oriented vectors trending from conglomerate clasts. Random directions hint at the potential for a positive conglomerate test that has been overprinted by remagnetization.

d) Minimum AMS axes from this site, plotted for comparison. Minimum axes define a foliation at this site, which implies a non-random distribution of clasts. In this case, the conglomerate test could be construed as inconclusive in part due to the nature of the clasts not being transported enough to be randomized.
Figure 28: Bootstrap tilt test (Tauxe, 1998) for 00KHU specimens whose suitability was determined by plane analysis and 00KH1 specimens. a) Equal area plot of in-situ distributions of second-removed components used in tilt test. b) Equal area plot of tilt adjusted distribution of second-removed components used in fold test. c) Histogram of unfolded, tilt adjusted second-removed components used in tilt test. Dashed lines are representative of generated bootstrap data, solid line is unfolding curve of the data. Minimum dispersion of directions occurs between 72% and 100% unfolding.
Figure 29: The bootstrap test for a common mean a) Equal-area projection of means and alpha-95 cones of confidence for sites 00KH1 and 00KHU specimens used in Tauxe and Watson (1994) fold test. b) Equal-area projection of antipodal mean and alpha-95 cone of confidence for site 00KH1 and in-situ mean and cone of confidence for 00KHU specimens used in fold test. c) Histogram of $x_1$ Cartesian coordinates for in-situ KHU and KH1 bootstrapped means from 500 para-data sets and bounds that include 95% of components. d) same as c), for $x_2$. e) same as c) for $x_3$. f) Histogram of $x_1$ Cartesian coordinates for in-situ KHU and antipodal KH1 bootstrapped means from 500 para-data sets and bounds that include 95% of components. g) same as f) for $x_2$. h) same as g) for $x_3$. Because the confidence bounds overlap for all cartesian coordinates after reversals test, the bootstrapped means cannot be distinguished at the 95% confidence level. These data pass the bootstrap reversals test.
Figure 30: Flinn plot diagrams of sites 00KH1, 00KH3, and 00KHU1-00KHU5 (all sites used in this study to determine a final paleolatitude).
A. Site 00KH1. specimens range from triaxial to prolate; atypical.
B. Site 00KH3. specimens fall mostly in oblate field; typical of most sites.
C. 00KHU1 specimens lie mostly on triaxial line and in oblate field with a few prolate outliers.
D. 00KHU2 specimens lie mostly in oblate field.
E. 00KHU4 specimens lie mostly in oblate field.
F. 00KHU5 specimens lie mostly in oblate field.
Figure 31: Lower hemisphere equal-area projections of tilt-corrected AMS data and a
associated a-95 confidence ellipses from the Houstenaden Creek Formation. Site names
are labeled next to respective plots.
Figure 31 (continued)
Figure 31 cont'd.

- □ maximum
- ○ minimum
- △ intermediate
Figure 31 cont'd.
Figure 32: Comparison between in-situ AMS axes and fold axes as determined by bedding plane intersection for Houstenaden Creek Formation. Bedding plane orientations taken from the Landslide Bay locality and the Burnt Hill Cove locality in an attempt to get a regional fold axis. Fold axis intersection determined using the program GEOrient 9.1 by Rod Holcombe (2002). AMS axes and confidence ellipses plotted using Tauxes (1998) plotams program.
Figure 33: Equal area projection of tilt-corrected AMS and confidence ellipses. High flow regime causes rollers to acts as a "traction carpet", skewing minimum axes parallel to downslope/paleoflow direction.

a) Site 00Kh1 AMS data corrected for declination rotation. Girdling of min. axes from E to W with alignment of max axes perpendicular to streak. Inferred paleoflow direction to W, perpendicular to maximum axis.

b) Site 00Kh3 AMS data corrected for declination rotation. E to W streaking of minimum axes with perpendicular maximum axes shows weak paleoflow/downcurrent direction to W.
Figure 33 (cont'd): Equal area projection of tilt-corrected AMS and confidence ellipses. High flow regime causes rollers to acts as a "traction carpet", skewing minimum axes parallel to downslope/paleoflow direction.

a) Site 00Khu1 AMS corrected for declination rotation. Minimum axes skewed towards WSW. Paleoflow direction interpreted to be in direction of skew minimum axes. b) Site 00Khu2 AMS corrected for declination rotation. Girdling of minimum axes towards WSW. Paleoflow direction interpreted to be WSW.
Figure 33 (cont'd): Equal area projection of tilt-corrected AMS and confidence ellipses. High flow regime causes rollers to acts as a "traction carpet", skewing minimum axes parallel to downslope/paleoflow direction.

a) Site 00Khu4 AMS data and 95% confidence ellipses corrected for declination rotation. Minimum axes skewed towards SW. Paleoflow direction interpreted to be in direction of skew minimum axes.

b) Site 00Khu5 AMS data and 95% confidence ellipses corrected for declination rotation. Girdling of minimum axes towards WSW. Paleoflow direction interpreted to be towards WSW.
I. \( \frac{K_{\text{min}}}{K_{\text{max}}} \)

\[ y = 6.7783x - 5.4374 \]

\[ R^2 = 0.0546 \]

Figure 34: Indication shallowing test based on the methods of Hodych et al. 1999. Dataset derived from specimens used in final analysis from sites 00K1 and 00K3. Error bars on specimens are one standard deviation, solid error bars are +/- mean tan inclination.
Late Jurassic reference pole

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Figure 35: a) Calculated paleolatitude for Upper Jurassic Otter Point Formation derived from b) Blake et al. (1985) grand site mean, with associated a-95 cone of confidence. Late Jurassic paleopole from Beck and Housen (2003). Expected declination and expected inclination refer to what the Declination and Inclination should be if the terrane were attached to North America at its present location during the Late Jurassic. The difference between the expected declination and inclination from the measured declination and inclination are the basis for calculating paleolatitude and tectonic rotation.
Figure 36: Plate motion reconstructions derived from fixed hotspot reference frame after Engebretson et al. (1984). Length of each arrow indicates 10 Myr of motion. Star indicates paleolatitude of 34° at ~80 MA derived from this study.
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<th>Tilt-corrected Site Mean Inclination</th>
<th>a95</th>
<th>K</th>
<th>Fisher type of fit</th>
<th>Plat</th>
<th>dist?</th>
</tr>
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<tbody>
<tr>
<td>Burnt Hill Cove</td>
<td>Khu1</td>
<td>2</td>
<td>19</td>
<td>126.1</td>
<td>33.1</td>
<td>28.4</td>
<td>79.3</td>
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<td>23.5</td>
<td>115.1 n/a</td>
<td>fl</td>
<td>42.3</td>
<td></td>
</tr>
<tr>
<td>168 24</td>
<td>Khu2</td>
<td>3</td>
<td>21</td>
<td>97.1</td>
<td>30.3</td>
<td>25.1</td>
<td>25.1</td>
<td>112.5</td>
<td>16.7</td>
<td>25.2</td>
<td>25.1 no</td>
<td>fl</td>
<td>35.0</td>
<td></td>
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<tr>
<td>Khu3/conglomerate</td>
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<td>17</td>
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<td>16.4</td>
<td>10.9</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>yes fl</td>
<td>n/a</td>
<td>154.7</td>
<td></td>
</tr>
<tr>
<td>153 34</td>
<td>Khu4</td>
<td>3</td>
<td>22</td>
<td>81.3</td>
<td>33.8</td>
<td>9.9</td>
<td>154.7</td>
<td>100.3</td>
<td>9.9</td>
<td>154.7</td>
<td>yes fl</td>
<td>46.3</td>
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<td>153 34</td>
<td>Khu4</td>
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<td>22</td>
<td>110.5</td>
<td>44.1</td>
<td>13.5</td>
<td>32.9</td>
<td>151.7</td>
<td>16.7</td>
<td>32.9</td>
<td>no fl</td>
<td>38.7</td>
<td></td>
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<tr>
<td>163 52</td>
<td>Khu5</td>
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<td>28</td>
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<td>21.1</td>
<td>16.6</td>
<td>31.7</td>
<td>161.9</td>
<td>16.7</td>
<td>31.2</td>
<td>no fl</td>
<td>32.7</td>
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<td></td>
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<tr>
<td>Khu1-5(w/o 3)</td>
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<td>16</td>
<td>102</td>
<td>103.6</td>
<td>33.3</td>
<td>9.1</td>
<td>18.6</td>
<td>135.8</td>
<td>8.7</td>
<td>20.5</td>
<td>no fl</td>
<td>40.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Landslide Bay</td>
<td>Kh1</td>
<td>7</td>
<td>13</td>
<td>79.1</td>
<td>-72.9</td>
<td>15.3</td>
<td>16.6</td>
<td>260.5</td>
<td>-52.7</td>
<td>15.1</td>
<td>16.9 yes fl</td>
<td>fl</td>
<td>-33.3</td>
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<tr>
<td>193 70</td>
<td>Kh3</td>
<td>4</td>
<td>12</td>
<td>163.2</td>
<td>-48.7</td>
<td>22.8</td>
<td>17.2</td>
<td>239.1</td>
<td>-34.4</td>
<td>22.8</td>
<td>17.2 yes fl</td>
<td>fl</td>
<td>-18.9</td>
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</tr>
<tr>
<td>146 62</td>
<td>Kh1</td>
<td>7</td>
<td>13</td>
<td>78.9</td>
<td>-73.4</td>
<td>12.4</td>
<td>24.8</td>
<td>260.5</td>
<td>-52.2</td>
<td>12.2</td>
<td>25.5 yes al</td>
<td>fl</td>
<td>-32.8</td>
<td></td>
</tr>
<tr>
<td>193 70</td>
<td>Kh3</td>
<td>4</td>
<td>12</td>
<td>145.5</td>
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<td>17.6</td>
<td>28.2</td>
<td>231.6</td>
<td>-48.6</td>
<td>17.6</td>
<td>28.2 yes al</td>
<td>fl</td>
<td>-27.2</td>
<td></td>
</tr>
<tr>
<td>00Khu &quot;b&quot;</td>
<td></td>
<td>6</td>
<td>16</td>
<td>85.5</td>
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<td>11.8</td>
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<td>105.6</td>
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<td>12.6 yes fl</td>
<td>fl</td>
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<td></td>
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<tr>
<td>00Khu1+00Kh3</td>
<td></td>
<td>11</td>
<td>25</td>
<td>122.0</td>
<td>-64.9</td>
<td>17.1</td>
<td>8.1</td>
<td>246.7</td>
<td>-52.8</td>
<td>10.0</td>
<td>17.1 yes al</td>
<td>fl</td>
<td>-33.4</td>
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<table>
<thead>
<tr>
<th>Formation</th>
<th>Site</th>
<th>Tilt-corrected Site Mean Declination</th>
<th>a95</th>
<th>Tilt-corrected Plat Inclination</th>
<th>AMS-corrected Site Mean Declination</th>
<th>AMS-corrected Plat Inclination</th>
</tr>
</thead>
<tbody>
<tr>
<td>Houstaden</td>
<td>00Khu1+00Kh3</td>
<td>-52.8</td>
<td>10</td>
<td>33.7</td>
<td>-53.9</td>
<td>34.3</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th>Present day</th>
<th>Magnetic field Declination</th>
<th>Inclination</th>
<th>a95</th>
</tr>
</thead>
<tbody>
<tr>
<td>Present day</td>
<td>Axial dipole field</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Late Cretaceous magnetic field</td>
<td>353.2</td>
<td>63.6</td>
<td>5.4</td>
</tr>
</tbody>
</table>

Table 1: Site-Mean directions for second-removed components from the Houstaden Creek Formation N/No, ratio of the number of samples demagnetized to those used in determining site-mean direction: a95 and k. Fisher (1953) statistical parameters; distribution attributes, appearance of directions when plotted on an equal area stereonet; type of fit, indicates either free line analysis (fl) or anchored line analysis (al) in which the origin of the Zijderveld plot is either not treated as part of the specimen's demagnetization path, or is treated as part of the specimen's demagnetization path respectively; Plat is calculated paleolatitude; 00Khu "b" are specimens whose directions were used in the fold-test and reversals test, after determining suitability using plane-fit analysis. Late Cretaceous direction from Diehl, 1991, and gundersonn and Sherriff, 1991.
<table>
<thead>
<tr>
<th>Late Cretaceous reference pole latitude</th>
<th>Late Cretaceous reference pole longitude</th>
<th>Late Cretaceous reference pole a95</th>
</tr>
</thead>
<tbody>
<tr>
<td>80.30°</td>
<td>189.50°</td>
<td>5.4</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>00Kh1, 00Kh3 site latitude</th>
<th>00Kh1, 00Kh3 site longitude</th>
<th>00/Kh1, 00Kh3 mean a95</th>
</tr>
</thead>
<tbody>
<tr>
<td>42.13°</td>
<td>-124.23°</td>
<td>3.9</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th>Expected declination</th>
<th>Expected inclination</th>
<th>Expected paleolatitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>66.05°</td>
<td>10.56°</td>
<td>48.38°</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th>00Kh1, 00Kh3 mean declination</th>
<th>00Kh1, 00Kh3 mean inclination</th>
<th>Calculated paleolatitude</th>
<th>Calculated paleolatitude error (+)</th>
<th>Calculated paleolatitude error (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>246.7°</td>
<td>52.8°</td>
<td>33.37°</td>
<td>3.90°</td>
<td>3.55°</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th>Total calculated transport</th>
<th>Calculated error in transport</th>
<th>Calculated latitude anomaly</th>
<th>Calculated latitude anomaly error</th>
</tr>
</thead>
<tbody>
<tr>
<td>1660 km.</td>
<td>+/- 500 km.</td>
<td>15.01°</td>
<td>4.53°</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>00Kh1, 00Kh3 mean declination</th>
<th>00Kh1, 00Kh3 AMS-corrected mean inclination</th>
<th>Calculated AMS-corrected paleolatitude</th>
<th>Calculated AMS-corrected paleolatitude error (+)</th>
<th>Calculated AMS-corrected paleolatitude error (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>246.7°</td>
<td>-53.9°</td>
<td>34.34°</td>
<td>4.00°</td>
<td>3.64°</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th>Total calculated AMS-corrected transport</th>
<th>Calculated error in AMS-corrected transport</th>
<th>AMS-corrected latitude anomaly</th>
<th>AMS-corrected latitude anomaly error</th>
</tr>
</thead>
<tbody>
<tr>
<td>1550 km.</td>
<td>+/- 500 km.</td>
<td>14.04°</td>
<td>4.64°</td>
</tr>
</tbody>
</table>

Table 2: Calculated paleolatitude for Upper Cretaceous Houstonian Formation derived from sites 00Kh1 and 00Kh3. Late Cretaceous paleopole from Diehl, 1991 and Gunderson and Sherriff, 1991. Expected declination and expected inclination refer to what the Declination and Inclination should be if the terrane were attached to North America at its present location during the Late Cretaceous. The difference between the expected declination and inclination form the measured declination and inclination are the basis for calculating paleolatitude and tectonic rotation.
<table>
<thead>
<tr>
<th>Farallon plate</th>
<th>Kula Plate</th>
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<tbody>
<tr>
<td><strong>time in Ma</strong></td>
<td>from - to</td>
<td>from - to</td>
</tr>
<tr>
<td>43-48</td>
<td>43-48</td>
<td>43-48</td>
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<tr>
<td>48-56</td>
<td>48-61</td>
<td>48-56</td>
</tr>
<tr>
<td>56-61</td>
<td>56-66</td>
<td>56-61</td>
</tr>
<tr>
<td>66-74</td>
<td>66-74</td>
<td>66-74</td>
</tr>
<tr>
<td>74-85</td>
<td>74-85</td>
<td>74-85</td>
</tr>
<tr>
<td><strong>speed km/mv</strong></td>
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<td>high</td>
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<tr>
<td>1.35</td>
<td>13</td>
<td>13</td>
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<tr>
<td>1.44</td>
<td>23</td>
<td>253</td>
</tr>
<tr>
<td>1.52</td>
<td>15</td>
<td>15</td>
</tr>
<tr>
<td>1.61</td>
<td>40</td>
<td>159</td>
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<td>1.75</td>
<td>40</td>
<td>159</td>
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<tr>
<td>1.86</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>2.56</td>
<td>256</td>
<td>618</td>
</tr>
<tr>
<td><strong>expected obliquity degrees</strong></td>
<td>low</td>
<td>high</td>
</tr>
<tr>
<td>17</td>
<td>13</td>
<td>13</td>
</tr>
<tr>
<td>23</td>
<td>63</td>
<td>63</td>
</tr>
<tr>
<td>19</td>
<td>15</td>
<td>15</td>
</tr>
<tr>
<td>21</td>
<td>40</td>
<td>40</td>
</tr>
<tr>
<td>21</td>
<td>40</td>
<td>40</td>
</tr>
<tr>
<td>18</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>47</td>
<td>256</td>
<td>618</td>
</tr>
<tr>
<td><strong>expected lateral displacement (km)</strong></td>
<td>low</td>
<td>total low</td>
</tr>
<tr>
<td>1563</td>
<td>3402</td>
<td>1078</td>
</tr>
<tr>
<td>1563</td>
<td>3402</td>
<td>1078</td>
</tr>
<tr>
<td>1563</td>
<td>3402</td>
<td>1078</td>
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Table 3: Obliquity and expected displacement within the Late Mesozoic and Cenozoic forearc region of northern California (after Jayko and Blake, 1993).
<table>
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<th>symbol</th>
<th>site</th>
<th>latitude</th>
<th>longitude</th>
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</thead>
<tbody>
<tr>
<td>Hunter's Cove Formation</td>
<td>Pistol River</td>
<td>00KHC</td>
<td>1</td>
<td>42° 30' N</td>
<td>124° 20' W</td>
</tr>
<tr>
<td></td>
<td>Pistol River</td>
<td>00KHC</td>
<td>2</td>
<td>42° 18.936' N</td>
<td>124° 25.053' W</td>
</tr>
<tr>
<td></td>
<td>Pistol River</td>
<td>00KHC</td>
<td>3</td>
<td>42° 18.963' N</td>
<td>124° 25.136' W</td>
</tr>
<tr>
<td></td>
<td>Hunter's Cove</td>
<td>00KHC</td>
<td>4 to 8</td>
<td>42° 19.218' N</td>
<td>124° 25.177' W</td>
</tr>
<tr>
<td>Cape Sebastian Sandstone</td>
<td>Cape Sebastian</td>
<td>00KCS</td>
<td>1 to 4</td>
<td>42° 19.155' N</td>
<td>124° 25.688' W</td>
</tr>
<tr>
<td>Houstenaden Creek Formation</td>
<td>Burnt Hill Cove</td>
<td>00KHU</td>
<td>1 to 5</td>
<td>42° 13.536' N</td>
<td>124° 22.940' W</td>
</tr>
<tr>
<td>Houstenaden Creek Formation</td>
<td>Landslide Bay</td>
<td>00KH</td>
<td>1</td>
<td>42° 13.106' N</td>
<td>124° 22.632' W</td>
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<tr>
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<td>Landslide Bay</td>
<td>00KH</td>
<td>2</td>
<td>42° 13.203' N</td>
<td>124° 22.735' W</td>
</tr>
<tr>
<td></td>
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<td>00KH</td>
<td>3</td>
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<td>124° 23.778' W</td>
</tr>
<tr>
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<td>Landslide Bay</td>
<td>00KH</td>
<td>4</td>
<td>42° 13.234' N</td>
<td>124° 22.785' W</td>
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</tbody>
</table>

Table 4: Rock formation, site locations and formation name symbols used in this study.