Kinematic modeling of deformation in the Troodos ophiolite, Cyprus

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The Troodos ophiolite in Cyprus is an ideal location to study deformation at a ridge–transform intersection. The ophiolite exposes both the fossil Arakapas transform fault and the Solea graben, an extinct spreading ridge. I analyzed paleomagnetic data from gabbros, which show vertical axis rotations that vary with position relative to the Arakapas fault belt and the Solea graben. This study characterizes the deformation between the Arakapas fault and the Solea graben based on my paleomagnetic rotation data and sheeted dike orientations by constructing numerical models. I test seven two–dimensional kinematic models, each of which represents heterogenous simple shear, to better constrain the spatial distribution of deformation due to motion along the Arakapas fault belt. I estimate that least 40 km of displacement has been accommodated along the transform fault and that 90% of this displacement is accommodated within 1.5 km of the fault. These types of quantitative estimates have not been made before for the Troodos ophiolite.

*Keywords:* Troodos Ophiolite; numerical models; mid–ocean ridges; shear strain; paleomagnetism; magnetic susceptibility anisotropy
1 Introduction

Transform faults are important components of divergent plate boundaries, allowing lateral motion between offset spreading ridges. However, it is difficult to study the deformation recorded by active faults because modern spreading ridges are located at great depths. Ophiolites, which are pieces of oceanic lithosphere thrust onto continents (Moores and Vine, 1971) provide an ideal opportunity to study the deformation recorded in fossilized oceanic lithosphere.

The Troodos ophiolite in Cyprus was important for the development of plate tectonics, as it contributed to the understanding of sea-floor spreading processes. The ophiolite contains a well-preserved transform fault known as the Arakapas fault belt (Simonian and Gass, 1978). Extensive debate in the literature focused on the sense of motion along the Arakapas fault belt with arguments for left-lateral motion, right-lateral motion or both senses of motion due to a reversal in slip direction (e.g. Simonian and Gass, 1978; Bonhomme et al., 1988; Macleod and Murton, 1995). Paleomagnetic studies eventually settled this debate by demonstrating vertical axis clockwise rotation consistent with dextral motion (e.g. Bonhomme et al., 1988; Macleod and Murton, 1995; Granot et al., 2006).

In this study, I construct numerical models of deformation due to the dextral motion on the Arakapas fault belt. Instead of simply stating that motion along the fault belt was dextral, my numerical models allow me to place quantitative constraints on the magnitude of deformation accommodated on the transform fault and where this deformation is concentrated. I use sheeted dike orientations (Carr and Bear, 1976) and paleomagnetic data from my own analysis and others (Abelson et al., 2002; Granot et al., 2006) to constrain my kinematic modeling.
2 Geologic background

Ophiolites were first described by Moores and Vine (1971) as oceanic material that was uplifted and emplaced onto land, and the Troodos ophiolite served as the case example. The complete ophiolite stratigraphy was characterized as tectonized peridotite, gabbro, sheeted dike complex, pillow lavas and deep–sea sediments as shown in Figure 1 (Moores and Vine, 1971). Since this initial description, consensus in the literature on the origin of ophiolites as developed and it is currently believed that the Troodos ophiolite and other Mid east ophiolites were likely formed 85 Ma (Moores and Vine, 1971) in a supra–subduction zone equivalent to a modern fore–arc environment (Clube et al., 1985). The Troodos ophiolite was then uplifted following a collision between a subduction zone south of Cyprus and Eratosthenes seamount (Ben–Avrahem, 1982) or the Egyptian continental margin (Moores et al., 1984) 5- 10 Ma after its formation (Clube et al., 1985).

Figure 1: Geologic map of the Troodos ophiolite (modified from Moores and Vine, 1971).
Figure 2: Dike strike and dips north of the Arakapas fault belt from the geologic map by Carr and Bear (1976) superimposed on map modified from Moores and Vine (1971).

Spreading in the Troodos ophiolite is related to three grabens— the Solea, the Minterso, and the Larnaca grabens (Varga and Moores, 1985) illustrated in Figure 1. The grabens were identified because the sheeted dikes dip towards each graben, and the regions between the grabens are less faulted and tilted (Varga and Moores, 1985). Most spreading in the
Figure 3: Two opposite models for explaining the observed dike orientations in Cyprus. In a), dikes are injected into a sigmoidal stress field and remain fixed in that orientation. In b), dikes are injected at a NS orientations but are rotated by deformation due to the transform fault (modified from Morris et al., 1990).
Troodos ophiolite occurred at the Solea graben, both by magmatic accretion and by normal faults that accommodate extension (Varga and Moores, 1985). The Mitsero graben, in contrast, formed by amagmatic spreading, interpreted as occurring simultaneously to the final extension event at the Solea graben (Allerton and Vine, 1991). Allerton and Vine (1991) suggest that a change in spreading direction terminated extension at the Solea and Mitsero grabens. This affected the orientation of the Arakapas transform fault, changing it from EW (090) to EWE (075). Spreading then to jumped to the Larnaca graben and magma was intruded into the now leaky transform fault (Allerton and Vine, 1991). No evidence for a spreading ridge is observed in the Anti–Troodos plate south of the Arakapas fault belt in the Limassol forest complex (Morris et al., 1990).

The 35 km–long Arakapas fault belt trends EW, ranges from 0.5 to 1.5 km in width, and is characterized by significant valleys filled with brecciated rocks (Simonian and Gass, 1978). There are few structural kinematic indicators within the fault belt, so other structures in the ophiolite are used to determine the sense of motion. The most commonly used dataset is sheeted dike orientations north of the Arakapas fault belt (e.g. Carr and Bear, 1976; Fig. 2). Near the Solea graben, the dikes are oriented approximately WNW and NS. East of the graben, dikes change orientation and strike approximately NE to near due E near the Arakapas fault belt. Two models have been proposed to explain the changing dike orientations: sinistral motion or dextral motion along the fault.

In the sinistral offset model, spreading ridges had a right–stepping geometry and may or may not have been joined by a fault (Fig. 3a). This spreading geometry induced a sigmoidal stress field where dikes were injected in an “S” shape and have remained in this orientation since their intrusion (Simonian and Gass, 1978). The field evidence supporting this model are mylonites which outcrop on the cm–to–meter scale and cannot be traced more than 100m along strike (MacLeod and Murton, 1995).

In the dextral motion model, the spreading ridges had a left–stepping geometry with a
transform fault between them. Sheeted dikes were injected at a NS orientation but experienced clockwise rotation due to drag imposed by the Arakapas fault belt (Fig. 3b; Simonian and Gass, 1978). This model is supported by paleomagnetic data (Bonhomme et al., 1988), which demonstrate the clockwise vertical axis rotations, expected in dextral systems. Mylonites have been inferred to help accommodate space problems from block rotations in a right-lateral shear zone (MacLeod and Murton, 1995). Because the dextral motion model is clearly preferred in the literature (e.g. Bonhomme et al., 1988; MacLeod and Murton, 1995; Granot et al., 2006), I assume dextral motion along the Arakapas fault belt for the remainder of this paper.
3 Paleomagnetism

Paleomagnetism is the study of the Earth’s ancient magnetic field recorded in rocks containing ferro–magnetic minerals. For the gabbros and sheeted dikes in this study, I rely on thermal remanent magnetism (TRM), where rocks preserve the direction of the Earth’s magnetic field when cooled below their Curie temperature. The direction of the recorded magnetic field is described by declination, which is the azimuth between the horizontal component of the total magnetic vector and geographic north, and inclination, which is the angle between horizontal and the vertical component of the total magnetic vector.

Figure 4: Simplified geologic map (modified from Moores and Vine, 1971) showing paleomagnetic rotations in gabbros based on data from this study and from Abelson et al. (2002) and Granot et al. (2006). Magnetic vectors oriented north represent stations where the gabbros have experienced no rotation and vectors not oriented north, represent stations where the azimuthal difference between north and the vector denotes the magnitude of vertical axis rotation.
For tectonic problems, declination and inclination are useful for determining whether rocks have changed their orientation or position since their TRM was acquired. If the declination from a specific rock does not coincide with the expected declination for a rock of that age and location, the rock may have experienced a net rotation. Similarly, if the inclination from a specific rock does not coincide with the expected inclination for a rock of that age and location, the rock may have experienced a net latitudinal translation.

At each of 16 stations that were sampled in the ophiolite, two gabbro samples were collected (Fig. 4). The hand–samples were drilled and cores were oriented in the laboratory. Paleomagnetic measurements were made on 2–5 specimens from each station using a superconducting quantum inference device (SQUID) and an alternating field (AF) demagnetizer at the University of Minnesota’s Institute of Rock Magnetism. Remanent field measurements using the SQUID were made after applying AF peak values of 0 mT, 5 mT, 10 mT, 15mT, 20mT, 30mT, 40mT, 60mT, 80mT, 100 mT, 125mT, 150mT and 180mT until the specimens retained less than 10\% of their initial magnetization. At each of these AF peak values, magnetic grains with coercivities less than the field strength are randomized resulting in a decrease of the magnetic intensity and a change in the direction of the remaining field.
Figure 5: Diagrams in a) and b) are representative Zijderveld plots of the alternating field demagnetization experiments. In b), the secondary magnetization can be seen in the 0–5 mT demagnetization step. c) is an equal–area hemisphere projection of the magnetization vector from each station and the Troodos mean vector (Clube and Robertson, 1986).
### Table 1: SUMMARY OF PALEOMAGNETIC RESULTS

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude</th>
<th>Longitude</th>
<th>N</th>
<th>Declination</th>
<th>Inclination</th>
<th>$B_{95\alpha}$ D</th>
<th>$B_{95\alpha}$ I</th>
<th>$F_{95\alpha}$</th>
<th>k</th>
<th>R</th>
</tr>
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<td>32° 52.144</td>
<td>2</td>
<td>230.6</td>
<td>30.2</td>
<td>.2</td>
<td>11.1</td>
<td>28.08</td>
<td>40.60</td>
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<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>35</td>
<td>34° 52.418</td>
<td>32° 52.890</td>
<td>3</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>36</td>
<td>34° 53.591</td>
<td>32° 53.906</td>
<td>3</td>
<td>265.8</td>
<td>-30.1</td>
<td>0.7</td>
<td>3.2</td>
<td>4.26</td>
<td>559.76</td>
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<td>32° 54.378</td>
<td>4</td>
<td>320.0</td>
<td>-8.4</td>
<td>1.5</td>
<td>7.3</td>
<td>7.92</td>
<td>101.1</td>
<td>4</td>
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<tr>
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<td>32° 55.613</td>
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<td>11.4</td>
<td>.6</td>
<td>12.5</td>
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<td>33° 04.174</td>
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<td>276.2</td>
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<td>9.19</td>
<td>120.59</td>
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<td>33° 00.355</td>
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<td>117.2</td>
<td>19.8</td>
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<td>75.01</td>
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<td>1.9</td>
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<td>305.4</td>
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<td>0.7</td>
<td>–</td>
<td>–</td>
<td>–</td>
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</tr>
<tr>
<td>55</td>
<td>34° 53.122</td>
<td>32° 57.005</td>
<td>5</td>
<td>331.6</td>
<td>24</td>
<td>11.5</td>
<td>21.7</td>
<td>14.55</td>
<td>22.88</td>
<td>4.9</td>
</tr>
</tbody>
</table>

$N$: Number of specimens measured  
$B_{95\alpha}$ D and $B_{95\alpha}$ I: 95% confidence interval for declination and inclination calculated using Bingham statistics  
$F_{95\alpha}$: 95% confidence interval calculated using Fisher statistics  
`: Paleomagnetic measurements were made but no conclusive magnetic vector was determined  
$k$: Dispersion measurement which increases as the remanent direction of each specimen is centered closer to the mean direction  
$R$: Magnitude of the summed direction cosines; the closer $R$ is to $N$, the tighter the vector cluster
Two representative demagnetization plots are shown in Figure 5a and 5b and the results from all stations are summarized in Table 1. Because my interest is in the deformation related to motion on the Arakapas fault belt, I am only concerned with primary remanent magnetizations and not the secondary overprint that can be seen in the lower demagnetization steps which are not connected to the origin by a single line as shown in Figure 5. The orientation of the primary signal was computed using Bingham statistics, and a heavier weight was applied to data further from the origin because these data have a greater signal–to–noise ratio. Ordinarily, for paleomagnetic analysis, one would apply a tilt correction to the rocks based on field evidence for the orientation of horizontal (e.g. bedding). However, this is not possible for gabbros, since they are gabbros are intrusive rocks.

Vertical axis rotations are calculated for each site by subtracting the magnetic vector from the well–documented Troodos mean vector where declination= 274° and inclination= 36° (α⁹⁵=12.3°) (Clube and Robertson, 1986). A dike whose paleomagnetic vector plots in the NE, SE or NW quadrants is inferred to have experienced clockwise rotation, while dikes whose paleomagnetic vectors plot in the SW quadrant are inferred to have experienced counterclockwise rotation. This assumption is based on both previous work suggesting clockwise rotation in a right–lateral system, as well as a higher probability of a small antithetic rotation (<90°) than a larger synthetic rotation (>270°).

The paleomagnetic results are shown in Figure 4. These data demonstrate that the direction of TRM vary across the region, suggesting that gabbros at different locations have undergone varying amounts of rotation. Within the Solea graben, there has been little to no rotation. East of the graben, rotations are mostly clockwise and increase with proximity to the Arakapas fault belt. My data are consistent with previous data particular well in regions in and near the Solea graben the station density is higher. My data is harder to compare to other’s data in regions greater than 8 km from both the ridge and the Arakapas fault belt belt because all datasets are less dense, but my data seem to show the same general trend
as other data. Inclination anomalies have not been interpreted.
4 Kinematic modeling

I constructed several kinematic models to quantify deformation in the ridge–transform intersection preserved in the Troodos ophiolite. To provide better constraints on the spatial distribution of deformation, my kinematic models attempt to match the orientations of sheeted dikes and rotations derived from paleomagnetic data from gabbros.

Figure 6: a) Shows the Cartesian coordinate reference frame superimposed on the geometry of the Troodos ophiolite. The Arakapas fault belt parallels the positive $x$–axis, and the Solea graben parallels the positive $y$–axis. Thus the origin is located at the approximate position of the ridge–transform intersection. b) Shows a material line representing the initial orientation of a dike that forms an angle $\alpha$ with the positive $x$–axis and undergoes a simple shear deformation to a final orientation described by the angle $\theta$.

4.1 Model Description

Deformation is modeled as end–member simple shear on a Cartesian coordinate system. A plane strain deformation, such as simple shear, is appropriate for modeling the Arakapas fault belt. In fact, a common assumption in modern plate motion studies is that motion is always parallel to the transform fault. The model reference frame is defined such that the $x$–axis is parallel to the Arakapas fault belt; consequently the $y$–axis is approximately parallel
to the Solea graben. The origin is located at the ridge–transform intersection as illustrated in Figure 6a. Note that this coordinate system is defined using a mathematical Cartesian coordinate system and not the typical geologic coordinate system. In this reference frame, the two–dimensional simple shear deformation matrix $F$ is given by,

$$
F = \begin{pmatrix}
1 & \gamma \\
0 & 1
\end{pmatrix},
$$

(1)

where $\gamma$ is the shear strain. For $\gamma > 0$, Eq. (1) describes right–lateral motion.

A single simple shear deformation matrix represents a homogeneous deformation. However, it is clear from Figure 2 that deformation north of the Arakpas fault belt is inhomogeneous. To account for this heterogeneity, I model $\gamma$ in $F$ as a function of both $x$ and $y$ such that,

$$
\gamma(x, y) = \gamma(x)\gamma(y)
$$

(2)

For the $\gamma(x)$ component, I assume a constant spreading rate at the Solea graben, which implies that $\gamma$ increases linearly with distance from ridge. Therefore $\gamma(x) = ax$, where $a$ is a constant used as a proxy for time and is related to the rate of motion. For the $\gamma(y)$ component, I assume that dikes a greater distance from the Arakpas fault belt will experience less deformation due to fault–motion than dikes closer to the fault. Therefore $\gamma(y)$ is a monotonic function; the specific functions are described in further detail in Section 4.2.

To calibrate the kinematic models, I compare the results with two datasets: dike orientations from geologic maps (Carr and Bear, 1976) and paleomagnetic rotations from this study and others (Abelson et al., 2002; Granot et al., 2006) from gabbros. Note that the dike orientations and paleomagnetic rotation datasets are from different crustal levels in the ophiolite stratigraphy.
Figure 7: a) Average strike of dikes showing how the dike orientations (Fig. 2) have been smoothed to facilitate computational modeling. b) Average rotations computed from the gabbros showing how paleomagnetic directions (Fig. 4) have been smoothed to facilitate computational modeling. (Magnetic vectors which point in the direction of the Troodos Mean vector (dec= 274° (Clube et al., 1985)), exhibit no rotation. Azimuthal deviations from the Troodos Mean vector correspond to vertical axis rotations). Length of dike and magnetic vector correspond to the weight of each averaged data point used in modeling.

For the sheeted dike orientations, 202 dike orientations (Fig. 2) were visually grouped into ten regions of dikes with similar orientations (Fig. 7a). For each rectangular region, I computed the average dike strike for each rectangle using Bingham statistics. This smoothing of the data is necessary because the local variation within the dataset, such as a 90° strike difference of dikes in close proximity, is difficult to model computationally. In the modeling, dikes are treated as material lines that are deformed and rotated by the deformation matrix $F$. Their deformation path is tracked from their initial orientation, described by the angle $\alpha$ relative to the $x$–axis to their final orientation, described by the angle $\theta$ (Fig. 6b). Because the dike dataset consists of final orientations ($\theta$), I solve for the $\gamma(x, y)$ responsible for rotation and for the initial orientations ($\alpha$) assuming that all dikes have the
same initial orientation. To solve for $\alpha$, I use dikes in the NW quadrant, closest to the Solea graben and furthest from the Arakapas fault belt, which I interpret to have experienced the least deformation.

For the 27 sites with paleomagnetic data, rotations were visually grouped into ten rectangles of regions with similar rotations (Fig. 7b); I computed the average rotation for each rectangle using Bingham statistics. Five rotations with significant departure relative to the rotations recorded at nearby stations were not included in the dataset. Similar to the smoothing of the dike orientation dataset, smoothing of the paleomagnetic dataset is necessary because the local variation within this dataset, such as clockwise and counter clockwise rotation recorded within 2 km, is difficult to model computationally. The deformation path describing paleomagnetic rotations is different from the path for the dikes because paleomagnetic data reflect a bulk vertical axis rotation and not the rotation of a material line. The bulk rotation ($\omega$) is related to the simple shear component of deformation by,

$$\omega = -\gamma/2,$$

where a clockwise rotation is negative, and $\gamma$ is in radians (McKenzie and Jackson, 1983). Figure 8 graphically and mathematically illustrates how this rotation is related to the deformation matrix $\mathbf{F}$ through the velocity gradient matrix (McKenzie and Jackson, 1983).

My models for $\gamma(x, y)$ are compared to three datasets: the dike orientation dataset, the paleomagnetic rotation dataset, and both datasets where each dataset is given equal weight. For each function of $\gamma(x, y)$, the constants are determined by using non-linear least squares which minimizes the difference between the $\gamma$ predicted by the model and the $\gamma$ from the smoothed data for each rectangular region with data. The “goodness of fit” of each model is evaluated by summing the squared difference between the $\gamma$ predicted by the model and the $\gamma$ from the smoothed data: the smaller this sum, the better the fit.
Figure 8: Simple shear represented by its finite and infinitesimal components. The deformation path can be represented by the single finite deformation matrix given by $F$. Alternatively, the deformation path can be described by its infinitesimal stretching and vorticity components. The velocity gradient matrix ($L$) is the infinitesimal representation of $F$, and $L$ is decomposed into its stretching or symmetric component and its vorticity or skew-symmetric component. The vector fields describing these infinitesimal motions are shown. The finite transformation corresponding to each infinitesimal transformation is found by exponentiating each infinitesimal matrix. The exponentiated stretching matrix describes a pure shear deformation in a rotated coordinate system, and the exponentiated vorticity matrix describes a finite rotation.
4.2 Model results

I tested multiple expressions for $\gamma(y)$ including inverse functions, exponential functions and a superposition of two exponential functions to match the pattern of decreased rotation with increased distance from the Arakapas fault belt (Table 2). Four of the functions for $\gamma(x, y)$ converge to fit the smoothed data, while three do not converge.
Table 2: MODEL RESULTS

<table>
<thead>
<tr>
<th>Model</th>
<th>(\gamma(x, y))</th>
<th>Dike Model</th>
<th>Paleomagnetic Model</th>
<th>Combination Model</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>((a, b, c))</td>
<td>Prmag Error</td>
<td>Dike Error</td>
<td>Prmag Error</td>
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<tr>
<td>1</td>
<td>(ax^by)</td>
<td>0.66</td>
<td>0.02</td>
<td>(135,1.44,1.25)</td>
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<td>2</td>
<td>(ax \exp(-by))</td>
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<td>0.07</td>
<td>(134,52,19)</td>
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<tr>
<td>3</td>
<td>(a(x \exp(-by) + \exp(-cy)))</td>
<td>0.85</td>
<td>0.05</td>
<td>(134,26,19,19)</td>
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<tr>
<td>4</td>
<td>(ax^by)</td>
<td>0.67</td>
<td>0.15</td>
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<tr>
<td>5</td>
<td>(ax (\gamma-b)^c)</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>6</td>
<td>(ax (\csc(b y -1))^c)</td>
<td>-</td>
<td>-</td>
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</tr>
<tr>
<td>7</td>
<td>(ax \exp(-by^c))</td>
<td>-</td>
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\(\gamma(x, y)\): Equation used to the data

\(a, \alpha, b\) and \(c\): constants determined by the models

Dike model: Model based on dike orientation dataset

Paleomagnetic model: Model based on paleomagnetic rotation dataset

Combination model: Model based on dike orientation and paleomagnetic rotation dataset
The four converging models are shown in Figure 9 where the values of $\gamma$ throughout the modeled region are contoured in varying increments. The three primary differences between these models are 1) the magnitude of $\gamma$, 2) the curve spacing, and 3) the intersection of the curves with the $x$–axis. For the magnitude of $\gamma$, Model 1 predicts infinite $\gamma$ along the transform fault, while the three other models predict a finite values. Of the models that
Figure 10: Results from model 1: a) field showing smoothed dike orientations and dike orientations predicted from the dike dataset, b) contour of $\gamma(x, y)$ based on dike orientation dataset, c) field showing smoothed paleomagnetic data and the paleomagnetic rotations predicted by the paleomagnetic data (initial orientation of material lines was determined in part by using the dike orientation dataset), and d) contour of $\gamma(x, y)$ based on paleomagnetic rotation dataset.
predict finite values, the $\gamma$ values range from 17 (Model 2) to 7 (Model 4). Regarding the second difference between models which is the curve spacing, in Model 1 the contours of $\gamma$ are more tightly spaced than in any of the other models; this is related to the infinite $\gamma$ value at the transform fault because $\gamma$ must decrease from infinity at the transform fault to near 0 at a distance of 17 km from the transform fault. The third difference between models is where the contours of $\gamma$ intersect the $x$–axis. In Model 1, the contours intersect the $x$–axis at the origin. In contrast, contours in models 2, 3, and 4, intersect the $x$–axis at linearly increasing distances from the origin.

I next discuss the results of Model 1 in more detail because this model had the smallest error (Table 2). However, a similar discussion could be made for the other kinematic
models that converge to fit the data. Figures 10 and 11 show the model predicted dike orientations and the smoothed data as well as the contours of $\gamma$ for the dike orientation dataset (Fig. 10a and 10b), b) the paleomagnetic rotation dataset (Fig. 10c and 10d), and c) both datasets combined (Fig. 11a and 11b). The dike–based model is able to predict the dike orientations near the Arakapas fault belt and 16 km from the Arakapas fault several km’s from the graben. At intermediate locations, the model both over and under predicts $\gamma$. The paleomagnetic–based model best matches data in zones of greatest clockwise rotation, and especially 4 km from the Arakapas fault and 7 km from the ridge. Discrepancies between the model and the paleomagnetic data occurs near the ridge as well as at 10 km from the Arakapas fault belt and 8 km from the Solea graben; in these zones, data exhibit both clockwise and counterclockwise rotations that cannot be interpreted by my kinematic model. The $\gamma$ contours are oblique to the Arakapas fault belt and the Solea graben. The dike orientation map and $\gamma$–contour map based on both datasets appear to be the average of the two individual models described above.

The models predict varying initial orientations ($\alpha$) of the sheeted dikes. Based on the dikes alone, the initial orientation is N22W- N27W compared to the N44W- N45W from the paleomagnetic data. Both models suggests that the dikes were not intruded parallel to the ridge, which is a typical first approximation for the initial dike orientation.
Figure 12: Accommodation of displacement near a shear zone. North of the Arakapas fault belt, $\gamma(x, y)$ is shown at three localities and increases with distance from the Solea graben. South of the Arakapas fault belt in the "Anti- Troodos" plate, $\gamma(x, y)$ is shown at three localities and increases with distance from the un–named ridge. Displacement accommodated by the transform is summed using integral calculus from displacement in each small horizontal linear zone shown in the colored curved by assuming that the shear strain decreases with distance away from the transform fault (Ramsey and Graham, 1970). In my model, the shear strain varies as a function of $x$ and for the displacement calculation, I substitute $\gamma(x)$ with the mean $\gamma(x)$ in my study area. Since their is no evidence for a ridge in the "Anti- Troodos" plate (Morris et al., 1990), the mean $\gamma(x)$ is the minimum average shear strain between spreading ridges and yields a conservative estimate of displacement.

The models based on sheeted dike dataset always predict greater average shear strain ($\gamma_{bulk}$) than models based on the paleomagnetic dataset reflecting the rotation of gabbros. For example, Model 2 predicts $\gamma_{bulk} = 1.31$ from dike orientation dataset and $\gamma_{bulk} = 1.08$ from the paleomagnetic rotation dataset. Thus, an additional 7° of bulk rotation is recorded in the dikes than in the gabbros. (This calculation requires integrating $\gamma(x, y)$ and cannot
be made for Model 1 because \( \gamma(x, y) \) for Model 1 is not integrable over the area of interest.)

Granot et al. (2006) find a constant rotational discrepancy of 5–20° between the dikes and the gabbros north of the Arakapas fault belt. These authors attribute this difference in rotation to a decoupling of the crustal layers at an early stage of magmatic accretion along the Solea graben.

The discrepancy between the two data sets could also be attributed to the variation of initial dike orientations, which I assume to be constant throughout the region. This assumption may not be appropriate for dikes near the Arakapas fault belt where the pre-existing anisotropy created by the Arakapas transform fault might cause dikes to intrude at orientations that are different from elsewhere in the ridge–transform intersection.

All kinematic models are more consistent with the dike orientation data than the paleomagnetic rotation data (see errors in Table 2). This may be due, in part, to the errors associated with each dataset: paleomagnetic data, which can be measured to \( \pm 10^\circ \), are noisier than dike orientation data, which can be measured to \( \pm 2^\circ \). Further, the paleomagnetic data near the ridge–transform intersection include both clockwise and counterclockwise rotations. Two different directions of rotation can not be accounted for in my simple shear model. Thus, the paleomagnetic model is least well–behaved in the ridge–transform intersection. There is a similar rotational discrepancy in the smoothed dike orientation data but it is smaller in magnitude, and therefore the dike orientations are easier to model.

To improve future models, additional data should be collected near the Arakapas fault belt. No paleomagnetic data are found within 4 km of the Arakapas fault belt. Further, cross–cutting relationships between dikes in the field could be used to determine if there are relative age relationships between dikes. Using such age relationships, I could better determine which dikes were injected at some distance from the graben and therefore did not have the same initial orientation as other dikes. Dikes not intruded at the graben would not be included in the model. The enhanced constraints on deformation attainable from
data within 4 km of the Arakapas fault belt would permit both a more robust model of the data and constraint on the error in each model.

Figure 13: Maps showing crustal age contoured in millions of years for A) fast-spreading ridge (75 mm/yr), B) intermediate-spreading ridge (40 mm/yr), and C) slow-spreading ridge (5 mm/yr).

Despite some of the shortcomings in kinematic model, I can estimate the total displacement accommodated near Arakapas transform fault. To estimate this displacement, I do a back-of-the-envelope calculation using the method of Ramsey and Graham (1970) outlined in the Appendix 1. I construct a curve of $\gamma$ values relative to position perpendicular to
the Arakapas fault belt and find the area under this curve, which estimates displacement as shown in Figure 12. Because $\gamma$ varies as a function of both $x$ and $y$ and in my model, I calculate this curve for a particular $x$ value, which I chose to be 8 km from the Solea graben, which is the midpoint of my model area. Because this position is a minimum estimate of the midpoint between two ridge segments as illustrated in Figure 12, my estimate of total displacement is conservative.

This calculation suggests a minimum displacement of 40 km along the Arakapas fault belt. This 40 km of displacement was accommodated at varying distances from the Arakapas fault belt: 18% of the offset is accommodated within 0.1 km of the fault, 62% of the offset is accommodated within 0.5 km of the fault, and 94% of the offset is accommodated within 1.5 km of the fault.

To place the estimate of displacement in the context of a mid–ocean ridge environment, I use a reasonable range of spreading rates from fast, intermediate to slow (Keary and Vine, 1996). I test fast, intermediate, and slow spreading rates because geologic evidence from Cyprus has been used to support each (Allerton and Vine, 1987; Robinson et al. 2008; Varga and Moores, 1985). For a fast spreading rate of 75 mm/yr, crustal material 16 km from the ridge is 250 ka older than material at the ridge and spreading was active for a minimum of 530 ka. For an intermediate spreading rate of 40 mm/yr, crustal material 16 km from the ridge is 500 ka older than material at the ridge and spreading was active for a minimum of 1 ma. For an slow spreading ridge, crustal material 14 km from the ridge is 3.5 ma older that material at the ridge and spreading was active for a minimum of 8 ma.
6 Conclusion

I studied the deformation at a ridge–transform intersection in the Troodos ophiolite. Ophiolites expose a fossil sea–floor and as well as stratigraphically lower crustal zones, allowing interpretation of deformation at multiple crustal levels in a spreading–environment.

In this study, the paleomagnetic rotation dataset show vertical axis rotation consistent with previous studies that suggests dextral motion along the Arakapas fault belt. Kinematic modeling has better constrained the spatial distribution of deformation in the Troodos ophiolite and permits significantly more analysis of fault motion compared to previous qualitative models focused on determining slip–direction.

7 Acknowledgements

This research was supported in part by the Carleton College Kolenkow Reitz Fund for student research. I would like to all those who helped me put this project together– Sarah Titus showing me how to integrate my interest in mathematics with geology, as well as for initially getting me excited about this project, for countless hours of help discussing the science and for reading many drafts, Josh Davis for his mathematical insight, Mike Jackson and Julie Bowles at the Institute of Rock Magnetism for use of their magnetic laboratory equipment and help in interpreting results, Mark Zach for help with the drill–press, Bill Titus for invaluable advice, and my friends and family for friendship, laughs and encouragement over the past four years.
8 References

9 Appendix

9.1 Displacement and age calculations

Ramsey and Graham (1970) use calculus to calculate total displacement in a shear zone by finding the area under the curve of position perpendicular to the Arakapas fault belt versus $\gamma$ stated as,

$$\text{displacement} = \int_{-y}^{y} \gamma(y')dy',$$

(4)

where $y'$ is a dummy variable of integration, and all the deformation occurs within the bounds of $-y$ and $y$. Eq. (4) must be modified to estimate displacement accommodated along the Arakapas fault belt. I only model the deformation north of the Arakapas fault belt so I calculate the displacement north of the Arakapas fault belt and multiply this value by 2 to determine the total displacement. Because displacement in Eq. (4) is not a function of $x$, I calculate $\gamma(x)$ for a particular $x$ value and choose the mean $x$ (denoted by $\bar{x}$) of this study area because this represents the minimum distance between the two spreading ridges (Fig. 12). A conservative estimate of displacement is given by,

$$\text{displacement} = 2 \int_{0}^{y} a\bar{x}\gamma(y')dy',$$

(5)

where $a$ is one the parameters solved for in the kinematic modeling. Note that Eq.(5) cannot be used to evaluate the displacement predicted by Model 1, because $\gamma(x,y)$ for Model 1 does not converge on the interval [0,y].

Using reasonable spreading rates, I calculate the age of the crustal material relative to the geometry of the ridge–transform system by using the shear strain as a proxy for time. Using Model 2 along the Arakapas fault belt, $\gamma(x,0) = axe^{-b(0)} = ax$. From basic
kinematic equations, it follows that $x = art$ and therefore,

$$t(x) = \frac{x}{ar},$$

(6)

where $t$ is the age of crustal material and $r$ is a reasonable spreading rate.
9.2 Anisotropy of magnetic susceptibility

I measured the anisotropy of magnetic susceptibility in 144 gabbro specimens from 16 sites north of the Arakapas fault belt. Anisotropy of magnetic susceptibility is a measure of the spatial variation of magnetic susceptibility within magnetic grains. This tool has been used to examine flow patterns from spreading near the Solea graben (Abelson et al., 2001). I use this method to detect whether broad patterns in the anisotropy of magnetic susceptibility can be related to other data sets north of the Arakapas fault belt.

Because gabbros are ferromagnetic, they retain their internal magnetic field in the absence of an applied external magnetic field. The bulk susceptibility of a rock is the sum of the susceptibility of each mineral. For rocks with greater than 0.1 volume ferromagnetic minerals, the bulk susceptibility is dominated by the ferromagnetic minerals because of their high susceptibility (Tarling and Hrouda, 1993). For rocks with a lower percentage of ferromagnetic minerals, the susceptibility is dominated by paramagnetic minerals and in their absence, the susceptibility is dominated by diamagnetic minerals.

Figure 14: Kappabridge. In a), no specimen is placed in the Kappabridge so no change in voltage is produced in the upper coil, and therefore no voltage is induced in the lower coil. In b), a specimen produces a change in voltage in the upper coil, which according to Faraday’s law, produces a magnetic field which induces a voltage in the lower coil. This recorded voltage as well as the voltage recorded from 14 other non-unique specimen orientations are required to solve for the terms in the susceptibility matrix.
Low-field bulk susceptibility measurements were made on a KLY-35 Kappabridge (Fig. 14) at the Institute of Rock Magnetism. Because magnetic susceptibility is not constant over ranging temperature and varying applied fields, all measurements were made at 25° C and under magnetic fields less than 1 mT. Susceptibility measurements were made as shown in Figure 14 by measuring the susceptibility in 15 directions, six of which are independent. This repeat in measurement is required to determine the error.

There are several ways to describe anisotropy of magnetic susceptibility. Anisotropy of magnetic susceptibility is described by an ellipsoid where the major ($\kappa_1$) axis, is defined by the direction and magnitude of the greatest induced magnetism, the intermediate ($\kappa_2$) axis is defined by the direction and magnitude of the intermediate induced magnetism, and the minor ($\kappa_3$) axis is defined by the direction and magnitude of the least induced magnetism. Alternatively anisotropy of magnetic susceptibility can be described by degree of an anisotropy ($P_j$) or shape of ellipsoid ($T$), and bulk susceptibility ($\kappa_{bulk}$).
Table 3: SUMMARY OF ANISOTROPY OF MAGNETIC SUSCEPTIBILITY RESULTS

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"-" : Anisotropy measurements were made but no conclusive data.
Figure 15: Simplified geologic map modified from Moores and Vine (1971) showing the degree of magnetic anisotropy and magnetic foliations.

The anisotropy of magnetic susceptibility data were examined for spatial patterns in $\kappa_1$, $\kappa_2$, $\kappa_3$, $P_j$, T, and $\kappa_{\text{bulk}}$ and all results are summarized in Table 3. The only recognizable pattern was in the plane defined by the $\kappa_1$ and $\kappa_2$ axes which approximate magnetic foliation shown in Figure 15. The spatial pattern in the magnetic foliation mimics the strike of the sheeted dikes. In southern regions which are closer to the Arakapas fault belt, the magnetic foliation is oriented NE and ENE. At distances further from the Arakapas fault belt, the magnetic foliation is oriented NS. The similarity between the magnetic foliation and the
Sheeted dike orientations suggest that anisotropy of magnetic susceptibility data could be used instead of dike orientations or paleomagnetic data, which is useful because anisotropy data is much faster to collect than paleomagnetic data.