

1 **Changes in non-dipolar field structure over the**
2 **Plio-Pleistocene: New paleointensity results from**
3 **Hawai‘i compared to global datasets**

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12 **Key Points:**

- 13 • Global compilations of paleointensity do not conform with the hypothesis that Earth’s
14 time-averaged magnetic field is a dipole (GAD field)
- 15 • We present 31 new paleointensity results from 0-4 Ma from the Hawaiian Islands
- 16 • Results for 0-1.5 Ma are inconsistent with a GAD field when compared to global
17 datasets reanalyzed the same way, but are consistent for 1.5 to 2.5 Ma

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18 Abstract

19 A foundational assumption in paleomagnetism is that the Earth’s magnetic field
 20 behaves as a geocentric axial dipole (GAD) when averaged over sufficient timescales. Com-
 21 pilations of directional data averaged over the past 5 Ma yield a distribution largely com-
 22 patible with GAD, but the distribution of paleointensity data over this timescale is in-
 23 compatible. Reasons for the failure of GAD include: 1) Arbitrary “selection criteria” to
 24 eliminate “unreliable” data vary among studies, so the paleointensity database may in-
 25 clude biased results. 2) The age distribution of existing paleointensity data varies with
 26 latitude, so different latitudinal averages represent different time periods. 3) The time-
 27 averaged field could be truly non-dipolar.

28 Here, we present a consistent methodology for analyzing paleointensity results and
 29 comparing time-averaged paleointensities from different studies. We apply it to data from
 30 Plio/Pleistocene Hawai’ian igneous rocks, sampled from fine-grained, quickly cooled ma-
 31 terial (lava flow tops, dike margins and scoria cones) and subjected to the IZZI-Thellier
 32 technique; the data were analyzed using the BiCEP method of Cych et al (2021, doi:10.1029/2021GC009755),
 33 which produces accurate paleointensity estimates without arbitrarily excluding specimens
 34 from the analysis. We constructed a paleointensity curve for Hawai’i over the Plio/Pleistocene
 35 using the method of Livermore et al (2018, doi:10.1093/gji/ggy383), which accounts for
 36 the age distribution of data. We demonstrate that even with the large uncertainties as-
 37 sociated with obtaining a mean field from temporally sparse data, our average paleoin-
 38 tensities obtained from Hawai’i and Antarctica (reanalyzed from Asefaw et al., 2021, doi:10.1029/2020JB020834)
 39 are not GAD-like from 0 - 1.5 Ma but may be prior to that.

40 Plain Language Summary

41 Paleomagnetists make the assumption that the Earth’s magnetic field behaves like
 42 a bar magnet centered at the spin axis, known as a Geocentric Axial dipole or GAD. Com-
 43 pilations of the magnetic field’s direction are largely consistent with this assumption, but
 44 compilations of its strength (paleointensity) are not. A number of causes for this could
 45 be: 1) The different experimental methods and the criteria used to pass or exclude pa-
 46 leointensity data might cause differences in records. 2) The ages of records differ between
 47 locations. 3) The field really doesn’t behave like a bar magnet. To test this, we performed
 48 paleointensity experiments on rocks collected in Hawai’i and compared our results to re-
 49 sults of similar age from other locations analyzed using the same methodology. The three
 50 locations analyzed in this study do not produce time-averaged paleointensities consis-
 51 tent with a GAD field for the most recent 1.5 million years, but a GAD field cannot be
 52 ruled out before this time. This indicates that differences in time-averaged field strength
 53 in global records can be unrelated to differences in methodology or age between stud-
 54 ies.

55 1 Introduction

56 Paleomagnetists use the direction of the magnetization acquired in the Earth’s an-
 57 cient magnetic field to obtain estimates of the ancient latitude at which the rock formed.
 58 Calculation of a latitude relies on the assumption that the Earth’s magnetic field is struc-
 59 tured like a bar magnet when averaged over sufficiently long timescales, so that the mag-
 60 netic field is vertical at the poles, and horizontal at the equator, also termed a Geocen-
 61 tric Axial Dipole (GAD). Estimates of the Earth’s magnetic field direction, taken from
 62 different latitudes over the past 10 Ma conform relatively well to a GAD field, with a
 63 small hemispheric asymmetry (Cromwell et al., 2018). On the other hand, estimates of
 64 the Earth’s magnetic field strength (the paleointensity) averaged over the last 5 Ma con-
 65 sistent show a behaviour incompatible with a strongly dipolar field. A seemingly per-
 66 sistent feature in paleointensity data is the presence of weak paleofields at high south-

ern latitudes (Lawrence et al., 2009; Asefaw et al., 2021; Tauxe et al., 2022), which causes a hemispheric asymmetry in the paleointensity data. This is seen in paleointensities from the MagIC database over the last 5 Ma (plotted in Figure 1a) where the mean paleointensity at 80°S would be produced by a centered magnetic dipole with a moment of around 40 ZAm², whereas the mean paleointensity at 20°N would require a dipole moment with a magnitude closer to 80 ZAm². Attempts to fit Giant Gaussian Process (GGP) models to paleointensity data to determine the structure of the time-averaged field have found that the field consistently requires a strong quadrupole term 15-30% the strength of the dipole field (Muxworthy, 2017; Shcherbakov et al., 2019), producing this asymmetry. However, such a large quadrupole is completely incompatible with the directional data.

Three different hypotheses could explain the non-dipole like behaviour of global time-averaged paleointensity records: bias in paleointensity estimation, comparison of temporally distinct data in a time-varying field, and genuine non-dipole field behavior. Regarding the issue of bias, paleointensity estimation involves normalizing the observed natural remanent magnetization (NRM) to a magnetization acquired in a known laboratory field. The accurate determination of a paleointensity therefore requires that the acquisition of a magnetization be reproducible in the laboratory. However, it has been shown (e.g., Levi, 1977; Dunlop & Özdemir, 2001; Krása et al., 2003; Tauxe et al., 2021) that some rocks have non-reproducible magnetizations, which can lead to biased paleointensity estimates. Global paleointensity records may be confounded by these biased estimates, leading to an apparent non-dipole signature. Alternatively, geomagnetic intensity variations through time may not be well averaged. The majority of paleointensity determinations are made with volcanic rocks, which record an instantaneous snapshot of the magnetic field at the time they cool. Archeomagnetic data indicate that the Earth’s magnetic field strength can vary strongly over decades to centuries (e.g., Shaar et al., 2020), so numerous paleointensity estimates are necessary for a good average. If the field strength varies over long timescales (e.g., millions of years), then comparing the “average” of two studies may not be meaningful if the units sampled are of different ages. And finally, it is also possible that the geomagnetic field is not in fact GAD-like but has long-term non-axial dipole contributions (as suggested by Wilson, 1970; Cromwell et al., 2013; Tauxe et al., 2022).

Paleomagnetists have identified behaviors in a paleointensity experiment that deviate from theoretical expectations and may lead to bias and recent studies have made a greater effort to eliminate such biased results. In most paleointensity studies, results from paleomagnetic specimens are excluded from the analysis if they fail a set of “selection criteria” which are phenomenological descriptions of these behaviors. Alternatively, the BiCEP method (Cych et al., 2021) attempts to find a relationship between the apparent paleointensity and one of these commonly used selection criteria (curvature, Paterson, 2011), and attempts to correct for the bias induced by the non-ideal behavior, obtaining accurate results without excluding data from the analysis based on arbitrary criteria. Recently, a study (Tauxe et al., 2022) which used the strict CCRIT criteria (Cromwell et al., 2015) and the BiCEP method on paleointensity studies from several latitudes found that there is still a discrepancy between these time-averaged paleointensities and those expected for a GAD field, making our first hypothesis (apparent non-dipole behavior is caused by bias in paleointensity estimation) unlikely to be the cause of inaccurate paleointensities.

Figure 1b shows the age distribution of latitudinally binned absolute paleointensity data in the MagIC database (without selection). It is apparent that different latitude bins have different age distributions. Because of this, the average paleointensity from each bin is representative of a different time period, and is not an average paleointensity for the whole of the last 5 Ma. High quality paleointensity data, analyzed in a consistent manner, are needed to determine whether temporal sampling is the cause of

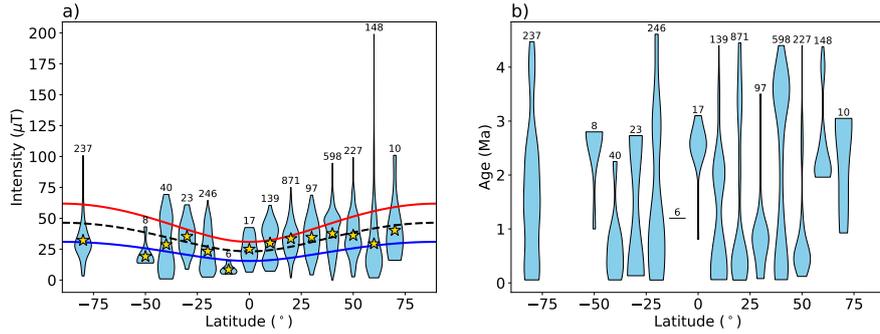


Figure 1. Violin plots showing latitudinal binned distributions of a) paleointensity and b) age for reported paleointensity results from the MagIC database aged between 50 ka and 5 Ma. In a violin plot, the width of the violin represents the frequency of intensities in that latitude bin, with the widest point in the violin representing the modal value. The number of data points in each bin are noted above the violins. The yellow stars in a) are the mean paleointensity value at each latitude bin and the solid blue, dashed black and solid red lines represent the expected mean values for a dipole field with a strength of 40, 60 and 80 ZAm^2 respectively.

119 apparent non-dipolar behavior, or if the time-averaged field is truly non-dipolar, as out-
 120 lined in our third hypothesis.

121 In this paper, we present paleointensity estimates from rapidly cooled volcanic mater-
 122 ial from lava flows, dikes and vent deposits (scoria and spatter cones) aged 0-4 Ma
 123 from the Hawai‘ian islands. In Section 2, we describe how we collect samples in the field
 124 (2.1), how we conduct paleointensity experiments (2.2) on specimens therefrom, how we
 125 analyze our results using the BiCEP method which produces accurate estimates for spec-
 126 imens magnetized in known fields (2.3), and how we obtain ages for our samples using
 127 $^{40}Ar/^{39}Ar$ dating (2.4). In Section 3, we show the results of our paleointensity study in
 128 Hawai‘i. We provide a discussion of the disadvantages of traditional methods of paleoin-
 129 tensity selection in Section 4.1. Section 4.2 discusses how our results suggest that scoria
 130 may be a useful lithology for obtaining high-quality paleointensity estimates, and are
 131 in agreement with estimates from other lithologies. In Section 4.3 we fit a model to our
 132 results in an attempt to derive a time average that accounts for uneven temporal sam-
 133 pling. We then apply the same methodology to studies from Northern Israel and Antarc-
 134 tica which targeted similar materials. This allows us to test whether poor temporal sam-
 135 pling or non-dipole behavior is responsible for the weaker paleointensity at high latitudes.
 136 Our results indicate that there is a persistent non-dipole component in the Earth’s mag-
 137 netic field over at least the past 1.5 Ma with older data being much more consistent with
 138 a GAD field.

139 2 Methods

140 2.1 Field Methods

141 Our results come from samples collected over three field seasons from outcrops on
 142 the Hawai‘ian islands. Samples were collected from the islands of Hawai‘i, Maui, Moloka‘i,
 143 and O‘ahu in an attempt to get a representative average paleointensity over the past 4
 144 Ma. This study targeted predominantly glassy and fine grained igneous material from
 145 lava flow tops and bottoms, scoria cones and dike margins. Néel theory (Néel, 1949) pre-
 146 dictes the physics of “uniaxial single domain” grains which should behave ideally in a pa-

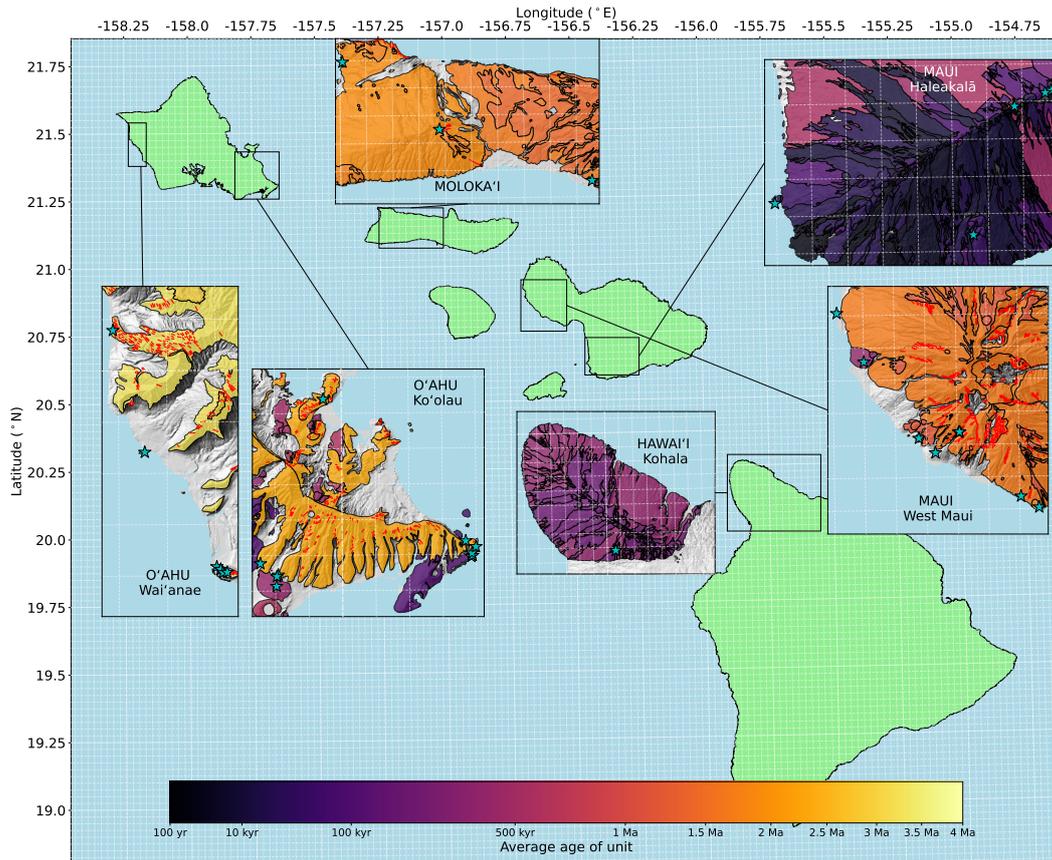


Figure 2. Maps showing sampling localities for successful sites used in this study (blue stars). Insets are labeled with the name of each island in capital letters and the name of the volcano (if applicable) in lowercase. Each map shows samples from a different Volcano/Island. Colors represent ages of units (Sherrod et al., 2007), with darker colors indicating younger flows (see colorbar), and dike locations indicated by red lines. Topographic data: U.S. Geological Survey (USGS). 2015. USGS 10-m Digital Elevation Model (DEM): Hawai'i. Coastline data: Hawai'i Statewide GIS Program.

147 leointensity experiment. Only very small magnetic particles exhibit single-domain be-
 148 havior, and so we sampled rapidly cooled materials most likely to contain these fine grains.

149 In the field, we collected small unoriented hand samples using a hammer and chisel;
 150 this allowed us to obtain smaller pieces of material and was less destructive than obtain-
 151 ing oriented specimens with a drill. Maps of our sampling localities are shown in Fig-
 152 ure 2 and details regarding location, age and material are given in Table 1.

153 2.2 Laboratory Work

154 Each sample was crushed with a mortar and pestle to produce multiple paleomag-
 155 netic specimens with masses on the order of 0.1 g. Specimens were weighed and glued
 156 into 1 cm wide borosilicate glass tubes using a high temperature, low magnetic moment
 157 glue (KaSil). We subjected each specimen to the IZZI-Thellier method (Tauxe & Staudigel,
 158 2004; Yu et al., 2004). This is a step-wise double heating experiment in which the
 159 NRM is replaced by a Thermal Remanent Magnetization (TRM) acquired in a known

Table 1. Ages and locations for sites from this study that passed CCRIT or BiCEP. Locations for all sites, including those that did not pass CCRIT or BiCEP are listed in the supporting information. Latitudes and Longitudes are referenced to the WGS84 standard. Codes in the methods column represent the method or citation used. AP: Ar-Ar age plateau, II: Ar-Ar inverse isochron, MP: Ar-Ar mini-plateau, O05: K-Ar age (Ozawa et al., 2005), S03: K-Ar age/stratigraphic relationship (Sherrod et al., 2007), SR: Stratigraphic relationships with other dated units (Sherrod et al., 2007), T03: K-Ar age (Tagami et al., 2003), TF: Ar-Ar Total Fusion Age. For more detailed explanation, see Section 2.4

Site	Island	Lithology	Lat. ($^{\circ}$ N)	Lon. ($^{\circ}$ E)	Age (Ma)	$\pm 2\sigma$	Method
HW306	Hawai'i	Vent Deposit	20.04470	-155.73437	0.1900	0.0700	SR
ML001	Moloka'i	Dike	21.13719	-157.15547	2.0700	0.0200	TF
ML012	Moloka'i	Vent Deposit	21.08955	-157.01053	1.6100	0.0300	AP
ML015	Moloka'i	Vent Deposit	21.19876	-157.24734	1.7700	0.0200	AP
MU004	Maui	Vent Deposit	20.77605	-156.53433	1.4300	0.0200	AP
MU009	Maui	Vent Deposit	20.81885	-156.61782	0.6100	0.0120	T03
MU011	Maui	Vent Deposit	20.83016	-156.63110	1.2300	0.0690	II
MU012	Maui	Vent Deposit	20.88931	-156.67484	0.3000	0.0216	AP
MU013	Maui	Vent Deposit	20.92685	-156.69633	0.5840	0.0100	T03
MU023	Maui	Vent Deposit	20.61085	-156.31100	0.0765	0.0635	S03
MU025	Maui	Vent Deposit	20.70692	-156.25424	0.0950	0.0450	S03
MU027	Maui	Vent Deposit	20.70551	-156.25857	0.0950	0.0450	S03
MU031	Maui	Vent Deposit	20.69669	-156.28040	0.0670	0.0404	AP
MU036	Maui	Vent Deposit	20.63397	-156.45102	0.0106	0.0085	II
MU106	Maui	Dike	20.83446	-156.59879	1.4900	0.0500	AP
MU109	Maui	Dike	20.83440	-156.59798	1.5500	0.0500	AP
MU111	Maui	Dike	20.83471	-156.59808	1.4500	0.0600	AP
MU113	Maui	Lava Flow	20.78467	-156.54893	1.1000	0.0600	AP
OA003	O'ahu	Flow	21.29434	-157.81123	2.5500	0.0800	AP
OA008	O'ahu	Flow	21.40440	-158.17461	3.7100	0.0600	AP
OA014	O'ahu	Dike	21.51972	-158.22772	3.4900	0.1700	AP
OA015	O'ahu	Flow	21.46033	-158.21154	3.1000	0.0300	AP
OA019	O'ahu	Flow	21.30938	-157.65783	2.8400	0.0600	MP
OA026	O'ahu	Flow	21.29836	-157.65380	2.7700	0.1300	SR
OA028	O'ahu	Flow	21.29907	-157.65273	2.7200	0.0800	AP
OA030	O'ahu	Vent Deposit	21.27831	-157.79929	0.3800	0.1100	O05
OA100	O'ahu	Vent Deposit	21.28628	-157.79791	0.4800	0.0400	O05
OA101	O'ahu	Vent Deposit	21.28521	-157.79900	0.4800	0.0400	O05
OA104	O'ahu	Flow	21.30080	-157.65320	2.1800	0.3500	AP
OA108	O'ahu	Dike	21.30527	-157.65027	2.2500	0.1700	AP
OA114	O'ahu	Dike	21.41002	-157.76354	2.8700	0.0600	AP
OA116	O'ahu	Dike	21.40308	-158.17264	3.7200	0.0500	MP
OA117	O'ahu	Dike	21.40308	-158.17264	3.7200	0.0500	MP
OA123	O'ahu	Sill Margin?	21.40149	-158.17141	2.5900	0.0900	AP
OA124	O'ahu	Dike	21.40168	-158.16927	3.2500	0.0100	MP

laboratory field. Under the IZZI protocol, the order of the in-field and zero-field steps alternates at each temperature step. Under ideal conditions, the ratio of the magnetization lost in a zero-field step to the magnetization gained in an in-field step is the ratio of the ancient field (B_{anc}) to the laboratory field (B_{lab}). For this study, multiple lab fields were used for different specimens, as we observed that the choice of B_{lab} affected whether our specimens passed or failed some of our criteria (see Section 2.3). After each heating, specimens were measured in four positions using the 2G Cryogenic Magnetometer at the Paleomagnetic Laboratory at Scripps Institution of Oceanography.

For sample characterization, a PMC MicroMag 3900 Alternating Gradient Magnetometer was used to measure First Order Reversal Curves (FORCs, Pike et al., 1999) using the xFORC protocol and software of Zhao et al. (2017) on sister specimens from sites used in this study. Sample material from several sites was used to produce doubly polished thin sections. Thermo Fisher Scientific Phenom Desktop Scanning Electron Microscope (SEM) was used to produce Back Scattered Electron (BSE) images, as well as Electron Dispersive X-Ray Spectroscopy (EDS) point observations and maps for identifying the elemental compositions of minerals. These analyses were undertaken at the Scripps Institution of Oceanography Paleomagnetic Laboratory.

2.3 Analysis of Data

To make sure that we have unbiased results, we used two different analysis methods on our data to obtain an estimate of the ancient field. Primarily, we used the Bias Corrected Estimation of Paleointensity (BiCEP) method (Cych et al., 2021) of estimating paleointensities, but we also looked at results using the criteria of Cromwell et al. (2015) (CCRIT). CCRIT are a strict set of selection criteria which exclude many specimens to obtain only results in which we have high confidence. The BiCEP method uses all specimens (without evident alteration), but uses the curvature criterion of Paterson (2011) as a predictor for the bias of the paleointensity yielded by each specimen. This was shown by Tauxe et al. (2022) to produce an unbiased site-level paleointensity estimate while excluding fewer data than the strict CCRIT criteria.

BiCEP assumes that the magnetization records a single field, and thermochemical alteration of the specimen has not occurred. To make certain of this, we used the minimal selection criteria (see Paterson et al., 2014 for definitions and references), $DANG < 10$, $DRAT < 10$. In addition, we use a new parameter, $MAD_{Coe} < 5$ which just uses the zero-field first steps. The set of temperature steps on the Arai plot which maximize the FRAC criterion while passing the MAD_{Coe} , $DANG$ and $DRAT$ criteria. The vast majority of our specimens pass these criteria with ease, and the ones that do not would unambiguously be rejected by almost any other set of criteria.

Site results from BiCEP have a 95% credible interval which is equivalent to the full width of the 2σ interval from traditional selection criteria methods (e.g., CCRIT). We considered a site level result from BiCEP acceptable if it has a credible interval with a full width less than 40% of the median value, or $16 \mu\text{T}$, whichever is greater (the original BiCEP A or B criteria of Cych et al., 2021 only include the former criterion). This is equivalent to criteria of $\pm 10\%$ or $4 \mu\text{T}$ used for the CCRIT at a site level (for a Gaussian distribution, the full width of the 95% credible interval is equal to four standard deviations). An example of BiCEP being used to estimate B_{anc} and its uncertainty for a site is shown in Figure 3.

To compute a time-averaged paleointensity, we used the “Age Hyperparameter Reversible Jump Markov Chain Monte Carlo” (AH-RJMCMC) method (Livermore et al., 2018). This model fits piecewise linear curves to paleointensity data in a probabilistic fashion. The output of the model is a distribution of paleointensity curves, 95% of which lie within a “95% credible” envelope (displayed in Figures 4 and 8). The uncertainties in the paleointensity curves become large during time periods where there are few or no

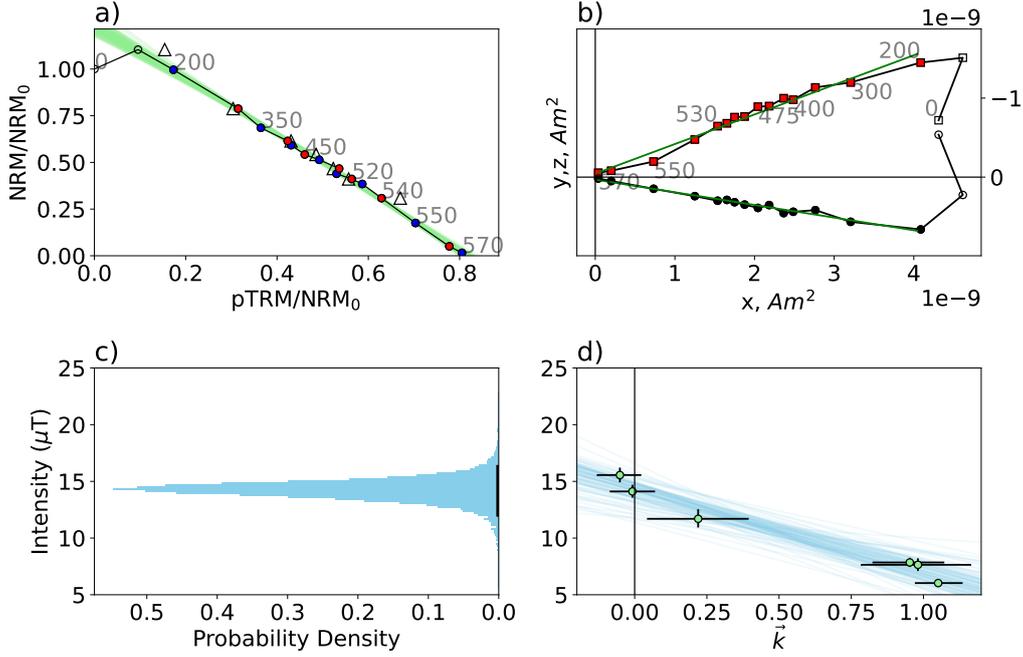


Figure 3. Example of BiCEP being used to obtain a paleointensity for site MU111. a) Arai plot (Nagata et al., 1963) for specimen MU111A05, red dots represent steps where the zero-field measurement was made first, and blue dots represent in-field first steps. Open circles represent temperature steps not used for this analysis. Triangles are pTRM checks and green curves are BiCEP’s circular fits to the data. b) Zijdeveld plot (Zijdeveld, 1967) showing magnetic direction data. Open symbols are steps where the temperature steps were not used. Green line is a principal component analysis fit to the directional data. c) Histogram of possible site mean intensities from BiCEP. d) BiCEP fit showing the predicted relationship (blue lines) between intensity (y axis) and the curvature criterion (\vec{k} , x axis) for all six measured specimens from this site.

211 data. Taking the temporal average of each paleointensity curve produces a distribution
 212 of possible time-averaged paleointensities. We discuss the rationale for using this approach
 213 in Section 4.3. A similar approach using curve fitting to calculate time-averaged pale-
 214 intensities was recently used in Bono et al. (2022).

215 2.4 Age Constraints

216 We obtained a range of radiometric ages for our samples that span the past 4 Ma.
 217 Rocks from 23 of our successful sites were analyzed at the Argon Geochronology lab at
 218 Oregon State University (OSU) for age determination. 200-300 μm pieces from each sam-
 219 ple were prepared by acid leaching in an ultrasonic bath according to the procedure of
 220 Koppers et al. (2000). This was followed by irradiation of the samples in the OSU TRIGA
 221 CLICIT nuclear reactor. Samples were then incrementally heated using a defocused CO₂
 222 laser, and the isotopic composition of the released argon was measured using an ARGUS-
 223 VI multi-collector mass spectrometer. Eighteen of our ages were calculated using argon-
 224 argon (Ar-Ar) plateaus. Sites MU011 and MU036 were calculated using inverse isochron
 225 ages, which are typically used when the initial sample contains excess ⁴⁰Ar. Three ages
 226 from sites OA019, OA116 and OA124 were calculated using mini-plateau ages, using less
 227 than 50% of the Ar released (Heaton & Koppers, 2019) and site ML001 was calculated

228 using a total fusion age because neither a reliable plateau or isochron age could be cal-
 229 culated. The mini-plateau and total fusion ages should be considered less reliable than
 230 the inverse isochron or plateau ages, as these samples could have been affected by ar-
 231 gon recoil or loss (Schaen et al., 2021). For sites OA030, OA100 and OA101, we used ex-
 232 isting potassium-argon (K-Ar) ages (Ozawa et al., 2005), and on West Maui, existing K-
 233 Ar ages (Tagami et al., 2003) were similarly used for sites MU009 and MU013. Mapped
 234 scoria cones at sites MU023, MU025 and MU027 have good age constraints over the timescale
 235 we are interested in from K-Ar dating and stratigraphic relationships outlined in Sherrod
 236 et al. (2003). Finally, sites HW306 and OA026 have their age constrained by stratigraphic
 237 relationship with other Ar-Ar dated flows. Ages obtained using the inverse isochron and
 238 plateau methods generally yielded consistent results when analysis with both methods
 239 was possible. Additionally, our inverse isochron age for site MU011 was highly consis-
 240 tent with the previous K-Ar age obtained by Tagami et al., 2003. A full table of ages
 241 is given in Table 1, and plots of the age vs cumulative argon released can be found in
 242 Supplementary Figure S1.

243 3 Results

244 Results are listed in Table 2. We obtain passing results from 35 sites (Table 2): 31
 245 passed BiCEP and 21 passed CCRIT. Some of the results that pass CCRIT do not pass
 246 BiCEP, but those sites that pass both methods exhibit good agreement between one an-
 247 other. Because BiCEP gives a more objective analysis, and because we obtain more pass-
 248 ing results with this method, we use only the results that pass BiCEP for the rest of our
 249 analyses.

250 We plot our results versus age in Figure 4. It is apparent that our results support
 251 the hypothesis that the more recent field (over the past ~ 1.5 Ma) is considerably higher
 252 than that from 1.5-4 Ma (e.g., Tauxe, 2006), supporting the hypothesis of a potential
 253 long period variation in the field strength (Selkin & Tauxe, 2000; Tauxe, 2006; Ziegler
 254 et al., 2011). It is also worth noting that in Figure 1, latitudes which have age distribu-
 255 tions skewing towards ages older than 1 Ma (e.g., 80°S , 60°N , 0°) tend to have averages
 256 that agree with a ~ 40 ZAm^2 dipole, whereas the majority of latitudes with mostly younger
 257 results tend to agree with a 60-70 ZAm^2 dipole moment, so qualitatively our hypoth-
 258 esis that the missing dipole may be caused by temporal sampling seems plausible. How-
 259 ever, the data from Antarctica (Asefaw et al., 2021) span the entire last 4 Ma but also
 260 have an average field consistent with a 40 ZAm^2 axial dipole strength, so temporal sam-
 261 pling alone does not explain all of the deviation from a GAD field.

262 The high paleointensity results over the past 1.5 Ma come predominantly from vent
 263 deposits (scoria and spatter cones), whereas older results come predominantly from dikes
 264 and lava flows. The dikes and lava flows are associated with the early shield building stages
 265 of Hawai'ian volcanoes, whereas the vent deposits are predominantly from the later stages
 266 of volcanic construction. The difference in lithology being coupled with a difference in
 267 field strength may be concerning, however our young, high field strength results agree
 268 well with the average paleointensity from lava flows in the HSDP2 core (Cai et al., 2017;
 269 Tauxe & Love, 2003, reanalyzed in Tauxe et al., 2022), shown as grey triangles in Fig-
 270 ure 4, although the variance of the HSDP2 data is larger. Additionally, results from sev-
 271 eral scoria cones yielded fields weaker than $30 \mu\text{T}$, including two cones on Moloka'i older
 272 than 1.5 Ma. This leads us to believe that our results from scoria are accurate.

273 4 Discussion

274 4.1 Pitfalls of selection criteria

275 We used the BiCEP method to obtain site level paleointensity estimates, and pre-
 276 fer this over the CCRIT method (and all other sets of selection criteria in use by var-

Table 2. Paleointensity results from specimens in this study which passed BiCEP and CCRIT. n_{pass} : Number of passing specimens. n_{tot} : Total number of specimens. For CCRIT results B_{min} and B_{max} represent the bounds of the 2σ interval, and so a full width of 40% or 16 μT is considered to have passed. The method column represents the preferred paleointensity result (BiCEP) when a site passed both BiCEP and CCRIT

Site	n_{pass}/n_{tot}	B_{min} (μT)	B_{anc} (μT)	B_{max} (μT)	Method
HW306	8/8	30.8	36.8	42.9	BiCEP
ML001	7/7	23.2	31.2	39.2	BiCEP
ML012	6/6	28.1	29.0	30.2	BiCEP
ML015	5/5	5.5	12.0	16.7	BiCEP
MU004	11/11	39.3	42.3	45.5	BiCEP
MU009	6/6	31.1	36.6	42.4	BiCEP
MU011	5/9	19.2	26.5	33.8	CCRIT
MU012	6/6	31.8	34.6	37.6	BiCEP
MU013	8/8	14.8	19.2	23.8	BiCEP
MU023	8/8	26.1	31.0	35.6	BiCEP
MU025	7/7	33.9	42.1	50.2	BiCEP
MU027	6/6	19.7	24.7	30.7	CCRIT
MU031	10/10	34.6	40.4	46.0	BiCEP
MU036	9/9	10.4	10.9	11.4	BiCEP
MU106	10/12	22.1	28.8	35.0	BiCEP
MU109	7/7	15.9	18.8	21.9	BiCEP
MU111	6/6	12.1	14.3	16.2	BiCEP
MU113	8/8	38.1	43.7	49.7	BiCEP
OA003	11/11	26.9	29.2	31.3	BiCEP
OA008	4/4	14.9	20.2	26.2	BiCEP
OA014	10/12	10.3	13.0	15.6	BiCEP
OA015	8/8	35.3	39.7	44.5	BiCEP
OA019	15/15	20.5	22.9	25.3	BiCEP
OA026	8/8	12.5	15.0	17.4	BiCEP
OA028	8/8	29.4	33.1	36.8	BiCEP
OA030	16/16	45.6	48.9	52.2	BiCEP
OA100	6/12	50.0	51.0	52.0	CCRIT
OA101	9/9	37.3	43.0	48.3	BiCEP
OA104	3/8	15.8	17.6	19.3	CCRIT
OA108	8/8	13.2	19.5	25.5	BiCEP
OA114	6/6	21.8	25.3	30.2	BiCEP
OA116	8/8	21.7	24.9	28.2	BiCEP
OA117	5/5	19.2	23.7	28.1	BiCEP
OA123	6/8	10.3	13.8	19.0	BiCEP
OA124	7/7	33.8	36.8	40.2	BiCEP

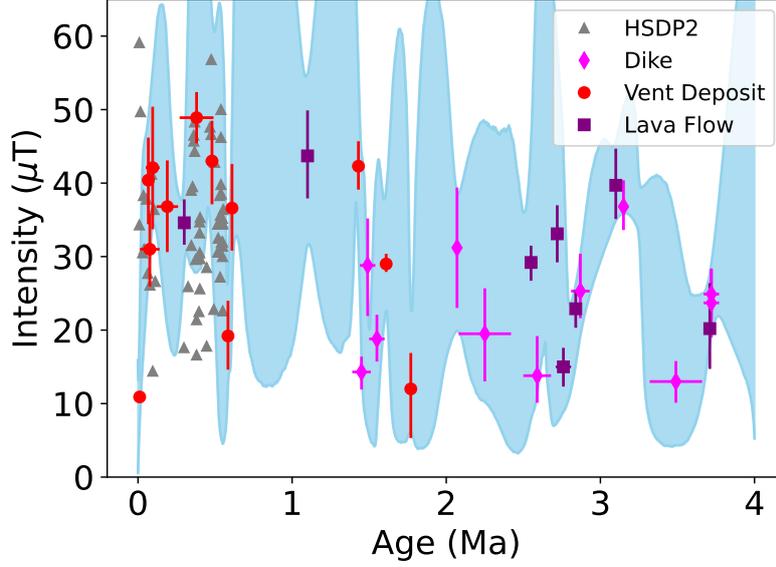


Figure 4. Paleointensity and age estimates from this study using the BiCEP method from lava flows (purple squares), scoria cones (red circles) and dike margins (pink diamonds). Error bars represent the 95% credible interval for intensity estimates, and the 2σ interval for age estimates. Grey triangles are other Hawai‘ian results from the HSDP2 core (Cai et al., 2017; Tauxe & Love, 2003, Tauxe et al., 2022), which have a similar distribution over this time period to our results. Blue envelope represents the 95% credible interval for the AH-RJMCMC model (Livermore et al., 2018) fit to the data (see Section 4.3).

277 various authors) as BiCEP produces many more site level results than CCRIT. Often, Bi-
 278 CEP passed sites where specimens failed the FRAC criterion of CCRIT, which speci-
 279 fies that a large proportion of the total magnetization of the specimen is needed to make
 280 a paleointensity estimate. BiCEP accounts for the uncertainty in curvature (and there-
 281 fore bias), introduced by using only part of a specimen’s Arai plot for a paleointensity
 282 estimate. This can be seen in Figure 3a, where specimen MU111A05 fails CCRIT due
 283 to low FRAC, but using a smaller part of the Arai plot translates to only a small increase
 284 in the uncertainty in curvature, shown by the green curves fit to the data.

285 In addition to the FRAC criterion in CCRIT, we identify cases in which criteria
 286 may reject a specimen if it has an ancient field much lower than the lab field. The MAD
 287 criterion may be exceeded if the laboratory magnetization acquired in an in-field step
 288 is not fully removed during a zero-field step, a consequence of a “high temperature pTRM
 289 tail” (Dunlop & Özdemir, 2000). This behavior is very noticeable in IZZI experiments
 290 (Figure 5), as the in-field first steps are more strongly affected by this effect. This leads
 291 to a zig-zag appearance in the Zijdeveld plot. The sizes of these tails are dependent on
 292 both the magnitude of the lab field, and the effect the tails have on MAD is dependent
 293 on the angle between lab and ancient field. If we call this angle θ , then the perpendic-
 294 ular part of the tails will be controlled by $B_{lab} \sin \theta$. If we assume no other sources of
 295 deflection to the MAD angle, the equation for the effect is:

$$\tan(\text{MAD}) \propto \frac{B_{lab}}{B_{anc}} \sin \theta. \quad (1)$$

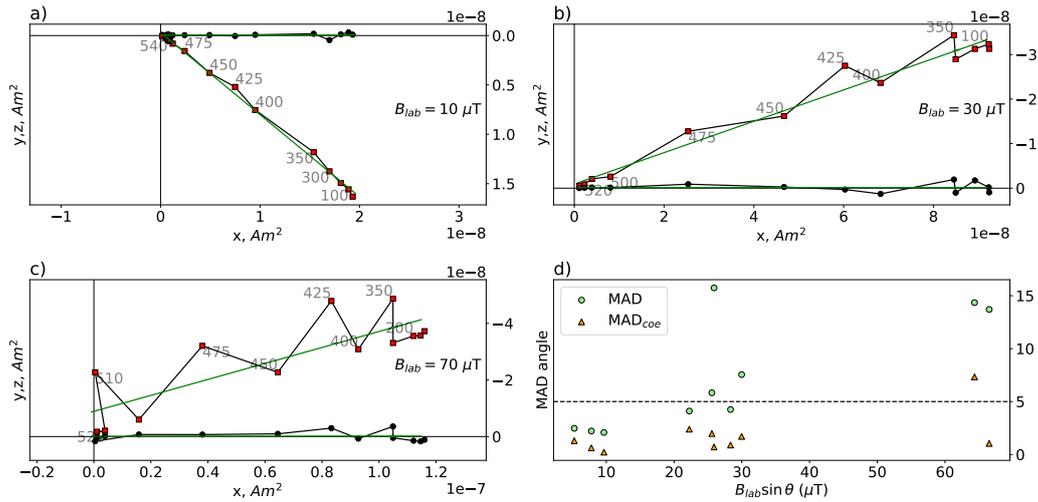


Figure 5. a)-c) Zijderveld plots of specimens from site OA014, showing zig-zagging behavior that progressively increases with lab field and d) Scatter plot showing the relationship between the MAD criterion, and the magnitude and angle of the lab field for all ten fully demagnetized specimens from this site. Paleointensity experiments were performed laboratory fields of a) a $10 \mu T$, b) $30 \mu T$ and c) $70 \mu T$. d) MAD (green circles) angle against the strength of the component of the lab field perpendicular to the ancient field direction (calculated by the PCA of the zero-field first steps). Orange triangles are the MAD of the zero-field first steps only (MAD_{Coe}). Horizontal dashed line represents the selection criterion (5) used in this study. Using MAD_{Coe} improves, though does not completely eliminate, the lab-field dependence of MAD. All MADs were calculated using temperature steps from $400-600^\circ C$ to avoid any potential viscous remanent magnetization (VRM).

296 This equation demonstrates that in the same lab field, sites with low ancient fields will
 297 be preferentially rejected with higher MAD, and sites with high ancient fields will be pref-
 298 erentially accepted.

299 To counteract the lab field-dependent effects, we used 10, 30 and 70 μT fields in
 300 our studies, which captures the range of the ancient field. At some sites with low esti-
 301 mated B_{anc} , there was an observably higher pass rate in lower fields. An example of this
 302 for site OA014 is illustrated in Figure 5. To treat specimens magnetized in different fields
 303 fairly, it is tempting to come up with a criterion for MAD which is dependent on Equa-
 304 tion 1. However, effects that we may be using MAD to look for (e.g., two component mag-
 305 netizations) will not be dependent on the lab field, and so we suggest calculating MAD
 306 for exclusively the zero-field first or “Coe” type steps (Coe, 1967). Although pTRM tails
 307 may still be present in these steps, they will be significantly reduced in in-field first steps.
 308 We call a MAD calculated using these steps MAD_{Coe} and how it compares to MAD for
 309 site OA014 is shown in Figure 5d. The use of MAD_{Coe} significantly reduces the lab field-
 310 dependent effects, but does not eliminate them entirely. Because pTRMs scale with the
 311 lab field used, there may be other unrecognized pTRM dependent effects. We recommend
 312 using a range of lab fields in paleointensity studies as the most robust way of compen-
 313 sating for these effects.

314 In addition to the lab field dependence of directional statistics, criteria which de-
 315 pend on pTRM checks (such as DRAT used in this study) have their own problems. Ther-
 316 mochemical alteration of magnetic material with blocking temperatures higher than the
 317 heating temperature can produce curved Arai plots with passing pTRM checks, as has
 318 been suggested by e.g. Wang and Kent (2021) and McClelland and Briden (1996). Ad-
 319 ditionally, although pTRM checks are used to detect alteration, they themselves may be
 320 caused by multi-domain carriers (Wang et al., 2013). Fortunately, the source of curva-
 321 ture is unlikely to matter for the BiCEP method, as it has been shown to yield accurate
 322 results when applied naively to a large test dataset, including passed and failed pTRM
 323 checks with no selection (Cych et al., 2021). However, the term “thermochemical alter-
 324 ation” describes a wide range of processes, and so in this study we cautiously excluded
 325 temperature steps where pTRM checks failed. More work is required to better under-
 326 stand thermochemical alteration processes and separate them from domain-state related
 327 behavior, for which the RESET method of Wang and Kent (2021) may be useful.

328 4.2 Sample Characterization

329 We have demonstrated our ability to obtain high quality paleointensity results from
 330 our samples using the BiCEP method. However, it is not clear what the primary car-
 331 riers of the magnetization are for these samples, particularly for samples from vent de-
 332 posits, which are relatively unstudied in the paleointensity literature. To attempt to char-
 333 acterize the domain state of our samples, we obtained First Order Reversal Curves (FORCs,
 334 Pike et al., 1999) for selected material from sites which passed BiCEP (and from some
 335 which failed). For this analysis we used sister specimens from the same samples for which
 336 the paleointensity results were acquired. FORCs are a qualitative way of assessing the
 337 domain state of a specimen using its hysteresis properties, and they can be decomposed
 338 into transient (tFORC), induced (iFORC) and remanent (remFORC) components us-
 339 ing to the protocol of Zhao et al. (2017). Specimens which contain “Single-Domain” (SD)
 340 grains which are ideal for the paleointensity experiment will have FORCs with a central
 341 ridge of positive values along the $H_a = -H_b$ axis (see e.g., Figure 6a). Specimens with higher
 342 numbers of non SD grains will have FORCs which have a spread along the $H_a = H_b$ axis.
 343 The iFORC which represents the induced part of the magnetization displays a pattern
 344 of three distinct “lobes” (e.g., Figure 6b,f) for a sample containing SD grains, whereas
 345 it may display four “lobes” or be extremely noisy for samples containing non-SD grains.
 346 The tFORC represents “transient hysteresis” which occurs in non-SD grains; specimens
 347 with just noise on the tFORC (e.g., Figure 6c) are most likely to be single domain.

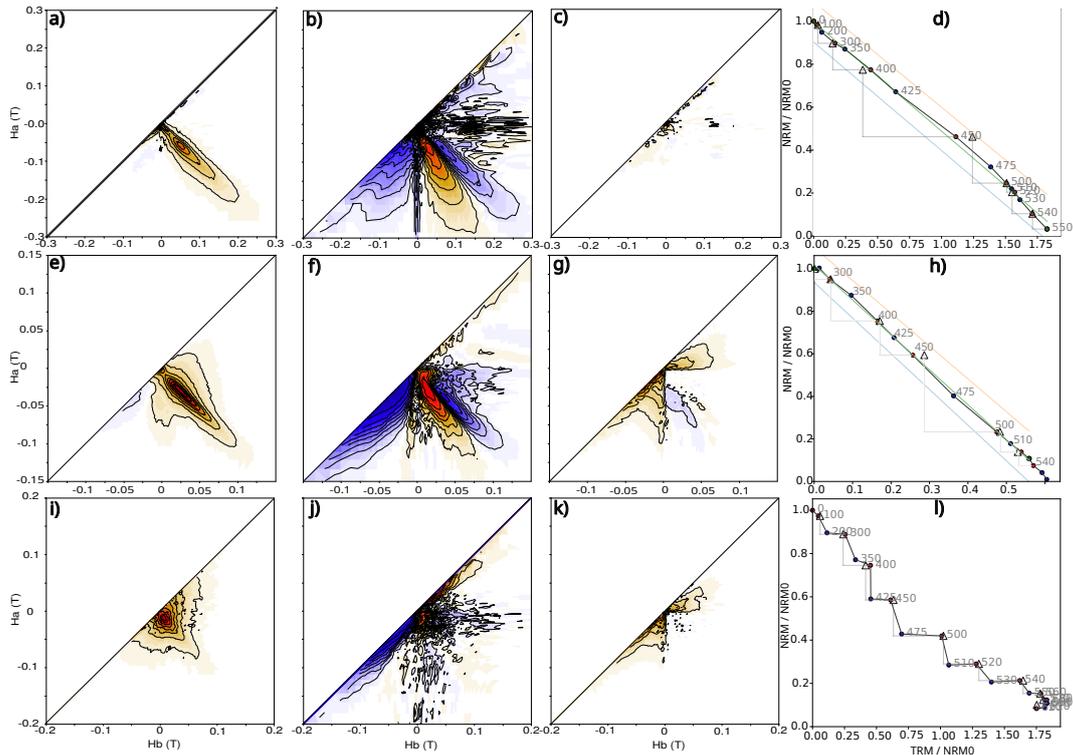


Figure 6. First Order Reversal Curves (FORCs) a),e),i), iFORCs b),f),j), and tFORCs c),g),k) calculated using the xFORC protocol (Zhao et al., 2017). All FORCs calculated using a smoothing factor of 2 and a non-linear color scale of 1, except for iFORCs which were calculated using a smoothing factor of 3 and a non-linear color scale of 10. Arai plots are plotted in d),h),l). FORCs use sister specimens from two sites that yielded passing results: OA030 (top row), OA014 (center row) and a site which did not pass CCRIT or BiCEP, HW305 (bottom row). Sites which yielded specimens with linear Arai plots tend to have an elongated central ridge and have 3 lobes in the iFORC (top and center rows), whereas sites with curved Arai plots tend to have more spread along the $H_a = H_b$ direction and have extremely noisy iFORCs with little information.

348 Examples of FORCs and Arai plots for different samples are displayed in Figure 6.
 349 The FORC interpretations generally agree with the paleointensity experimental results.
 350 FORCs obtained from dike samples have pronounced central ridges and three lobes in
 351 the iFORC if visible, and effectively no tFORC (Figure 6a-d). These samples generally
 352 had Arai plots which were straight lines, but sometimes underwent thermochemical al-
 353 teration at high temperatures. Samples from lava flows and vent deposits had central
 354 ridges, with small amounts of transient hysteresis and spreading along the $H_a = H_b$
 355 axis. These samples still have linear Arai plots, and often have three lobes present in the
 356 iFORC, which suggests that the majority of carriers in these specimens are single do-
 357 main (see Figure 6e-h). An example from a relatively coarse grained lava flow is given
 358 in Figure 6i-l. Samples like these had highly curved or zig-zagging Arai plots (Figure 6l)
 359 and generally had no central ridge and lots of spreading along the $H_a = H_b$ (Fig-
 360 ure 6i). These samples had pronounced tFORCs (Figure 6k), and only noise in the iFORCs
 361 away from the H_a axis (Figure 6j), observations which are consistent with the curved
 362 and zig-zagging Arai plots.

363 We also obtained Back Scattered Electron (BSE) images using an Scanning Elec-
 364 tron Microscope (SEM), and Electron Dispersive X-Ray Spectroscopy (EDS) element

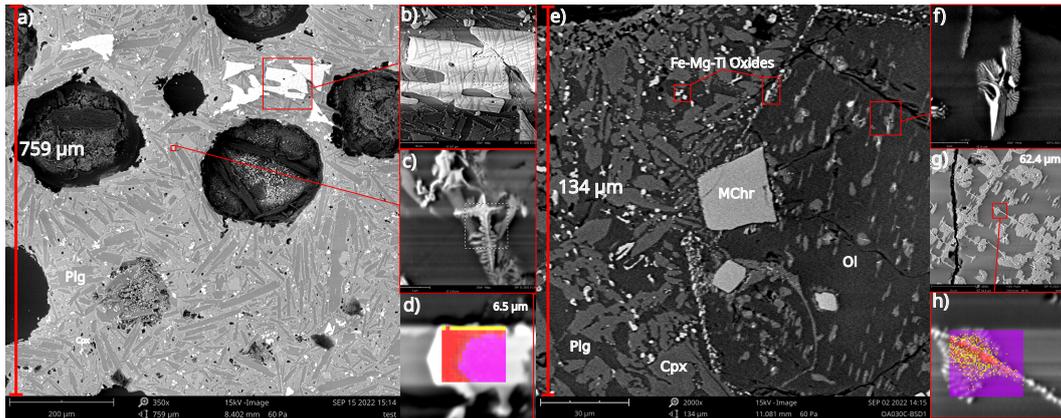


Figure 7. Back Scattered Electron (BSE) images and Electron Dispersive X-Ray Spectroscopy (EDS) maps of sister specimens from selected samples used in this study. Red text gives vertical field of view (FOV) for each image. a) BSE image of sample ML015A, a scoria vent deposit. b) Zoomed in image of large oxide in a), showing Fe-Ti exsolution textures. c) Zoomed in image of small oxide in a), showing elongate skeletal/cruciform structure. d) EDS element map of a typical oxide from another scoria vent deposit, ML012A, showing heterogeneous composition in the Iron-Titanium oxides. The atomic content of Si is shown in yellow, Fe in red, and Ti in pink. e) BSE image of sample OA030A, an agglutinated basanite vent deposit. f) Close up of high temperature alteration texture in olivine phenocryst. g) The same texture present in sample MU012A, a breccia from the bottom of a basanite lava flow. h) Close up of this texture with EDS element map. Colors are the same as d), with purple representing Mg. Note that the light colors in the BSE image represent an iron rich phase (interpreted as magnetite), which is surrounded by a phase richer in silicon than the surrounding olivine, interpreted as enstatite. Dominant mineral phases written on a) and e): Plg: plagioclase feldspar, Cpx: clinopyroxene, Ol: olivine, MChr: chrome spinel. Horizontal banding present in b),c),d),f),h) is an artifact of charging the sample that occurs in the SEM's EDS element mapping mode.

maps to identify iron oxides in several thin sections taken from our samples. Several pictures from these analyses are displayed in Figure 7 and further images of dike samples are displayed in supplementary Figure S2. Dike samples we analyzed contained no visible iron oxides in the glass, and almost no iron oxides in the groundmass. This is consistent with our FORCs and Arai plots (Figure 6a-d), which are indicative of this specimen containing a predominance of single domain grains, which are 10s of nm in scale and not resolvable by the SEM used in this analysis. By contrast, samples from vent deposits contained numerous micron-scale iron bearing oxides in the groundmass, and in some cases, larger iron oxides on the scales of 100s of microns (Figure 7a-d), size ranges where we would expect the grains to yield curved Arai plots. Many of these grains have elongated “cruciform” textures (Figure 7c) or have heterogeneous compositions (Figure 7a,d). One possibility is that these textures may persist to smaller scales, causing the larger grains to behave like assemblages of smaller, single domain, grains, due to their elongation or having smaller magnetic subregions separated by nonmagnetic lamellae. Another possibility is that these large grains do not contribute to the remanence. However, the lava flows and vent deposits have much higher NRM moments than the dikes, with mass normalized NRMs on the order of 10^{-2} to 10^{-3} Am²/kg, as opposed to the dikes which have moments on the order of 10^{-4} to 10^{-5} Am²/kg.

Two thin sections from sites MU012 and OA030 have numerous olivine grains which exhibit an unusual texture, as displayed in Figure 7e-h. This texture has been observed previously (Ejima et al., 2017; Blondes et al., 2012) and is interpreted as being caused by oxidation of olivine at temperatures above 800°C, which causes breakdown into an iron oxide (magnetite or hematite depending on formation conditions) and enstatite (see Figure 7h and figure caption). The temperature of the oxidation means that the samples were oxidized prior to gaining a magnetization, which means the NRM is a primary TRM acquired during cooling. Oxidation of this kind seems to typically occur in fire fountaining strombolian type eruptions (e.g., Del Moro et al., 2013) where the lavas remain at high temperatures in an oxidizing environment for a while (e.g., 950 °C for 24-48 hours as per Haggerty & Baker, 1967). OA030 is an agglutinated basanitic vent deposit, agreeing with this oxidative environment, whereas the MU012 sample was taken from breccia/clinkers in an a‘ā lava flow (rough fragmented pieces at the bottom of the flow), which may also undergo high temperature oxidation although the source is less clear.

Both sites with evidence for high temperature oxidation of olivines had highly linear Arai plots (see Figure 6h), with 16/16 specimens passing the strict CCRIT criteria for OA030, and 6/6 passing for MU012. Additionally a sample from OA030 has a FORC indicative of single-domain to single-vortex domain state, with a central ridge and three lobes in the iFORC (see Figure 6, middle row). This indicates that the oxides formed by this breakdown may have extremely desirable properties for paleointensity experiments. Similar to the smaller oxides found in our other vent deposits (Figure 7c), the elongation and finger-like structures present in these oxides could also explain their ideal behavior in the paleointensity experiment. These thin sections also contained numerous micron scale iron-titanium-magnesium oxides (interpreted as magnesioferrite) in the groundmass and around the outside of the olivine grains (Figure 7e), but because the majority of the remanence unblocks between 400 and 600°C (see Figure 6d), we believe that magnetite is the dominant remanence carrier in these specimens.

Despite the large iron oxides observed in vent deposits and lava flows from this study, we conclude that these lithologies provide a good source for paleointensity estimates, as they have a high success rate relative to our other lithologies owing to their strikingly linear Arai plots (see Figure 6, top row). Site MU113 provides further evidence for this, as material sampled from the inside of a lava tube gave an identical result to material sampled from a scoriaceous bomb entrained in the same flow. There are other reasons to favour these types of lithologies: The formation of these samples in an oxic environment at high temperature may help prevent thermochemical alteration during the pa-

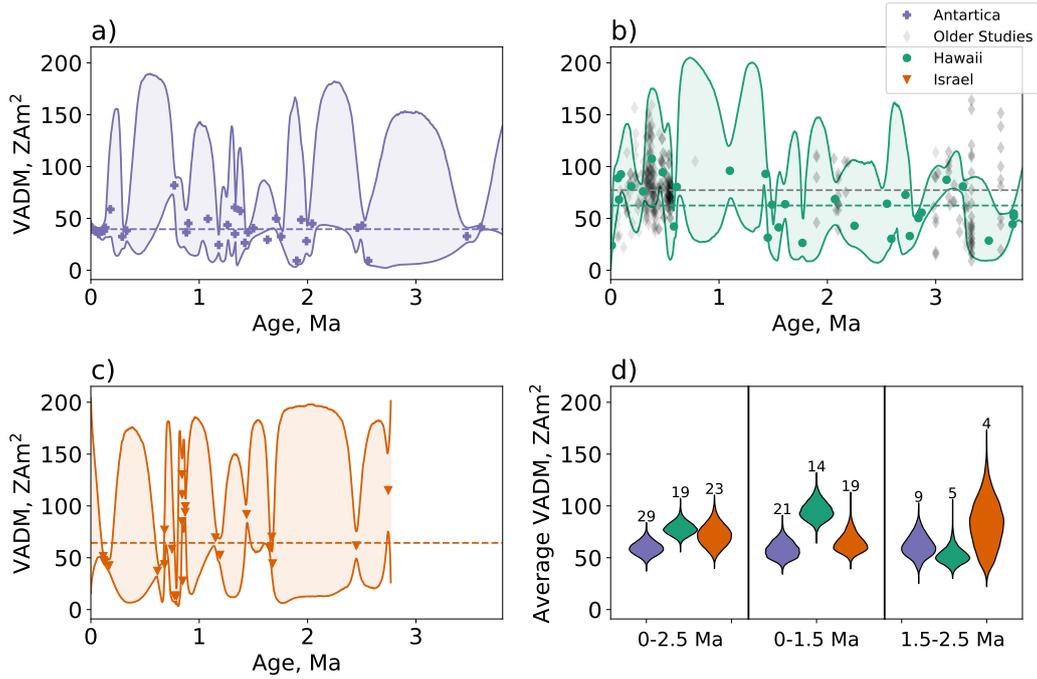


Figure 8. a) - c) Plots of VADM against age (symbols), and 95% credible envelopes for AH-RJMCMC models (Livermore et al., 2018) (shaded areas) for studies from a) Antarctica (purple plus symbols), b) Hawai'i (green dots), and c) Israel (orange triangles). Horizontal dashed lines are the average VADM of all paleointensity estimates (symbols) for each plot. In b), all unfiltered data in the MagIC database from Hawai'i aged between 50 ka and 3.8 Ma are plotted as grey diamonds, and the average VADM from these data are plotted as a grey horizontal line. d) Violin plots showing the distribution of averaged VADMs over different time periods, numbers refer to the number of paleointensity within these temporal ranges, although data outside these ranges may also contribute to these averages. Data from Hawai'i have a significantly higher average VADM than in Israel and Antarctica over the past 1.5 Ma, which is reflected in the averages from 0-2.5 Ma. Average VADMs for data older than 1.5 Ma appears to agree for all three locations.

418 leointensity experiment, and fresh scoria is also easy to come by in Hawai'i, as many scoria
 419 cones are quarried. However, most preserved vent deposits are typically formed during
 420 the later stages of Hawai'ian volcanism, and consequently we have no results from
 421 scoria older than 2 Ma.

422 4.3 Temporal Distributions of Intensity

423 Mismatch between the observed distribution of paleointensities with latitude and
 424 the expected distribution for a GAD field (Figure 1a) could potentially be caused by in-
 425 consistencies in treatment of data among different paleointensity studies. To compare
 426 the time-averaged field from our model to data from different latitudes, we reanalyzed
 427 results from recent paleomagnetic studies in Northern Israel (Tauxe et al., 2022) and Antarc-
 428 tica (Asefaw et al., 2021) using the BiCEP method and the same criteria used for the
 429 Hawai'i samples. Tables of results from these re-analyses can be found in the Support-
 430 ing Information. Each of these studies yielded passing sites with results spanning the past
 431 2.5 Ma. For direct comparisons between locations, we convert each paleointensity result

432 to a Virtual Axial Dipole Moment (VADM) which is the moment of the geocentric dipole
 433 (measured in ZAm^2) that would yield the observed paleointensity at a given latitude.
 434 Our average VADM for Hawai'i is 62.4 ZAm^2 , which is similar to the 64.2 ZAm^2 value
 435 from Israel, but is significantly higher than the average in Antarctica (39.6 ZAm^2). Plots
 436 of VADMs with age for each location are shown in Figures 8a)-c), with average VADMs
 437 plotted as horizontal dashed lines. In Figure 8b we also plot all the data from Hawai'i
 438 in the MagIC database from this time interval in grey. The unfiltered data have a sig-
 439 nificantly higher variance than our data, and the weaker field seen prior to 1.5 Ma in our
 440 data is not apparent in the unfiltered Hawai'ian data, which have an average VADM of
 441 77.2 ZAm^2 . These differences could occur because more field variation is being captured
 442 by the larger dataset, or because the unfiltered data have more variance due to incon-
 443 sistency in their analysis (for example, preferentially taking the low temperature steps
 444 in a potentially sagging Arai plot). Despite the consistency in analysis of our data, the
 445 average VADM in Hawai'i and Israel is still very different to that found in Antarctica,
 446 indicating that inconsistency in analyses and biased paleointensities caused by Arai plot
 447 curvature are not the source of this mismatch.

448 Taking an average VADM of the entire age range of our data may not be represen-
 449 tative of the time-averaged field, because our data have different temporal distributions,
 450 with no data in Israel older than 2.75 Ma. In Hawai'i, this average does not capture the
 451 change in average field strength seen at 1.5 Ma, and in Israel, we have many paleointen-
 452 sity data which record a strong field and come from a small range of time around 850
 453 ka B.P. Because this time interval is oversampled, it will bias our average VADM towards
 454 these higher values. For this reason, we used the the AH-RJMCMC method (see Sec-
 455 tion 2.3), which produces a set of possible paleointensity curves for each locality. We took
 456 the average value of each curve over the past 2.5 Ma, and converted these averages to
 457 VADMs. At times where there are many paleointensity data, the curves produced by the
 458 AH-RJMCMC have high precision, but only locally, so the time period is not over-represented
 459 in the average. At times when there are few data, the model uncertainties become very
 460 large and revert to a uniform prior distribution(which we set as 0-220 ZAm^2), indicat-
 461 ing that we do not have enough data to resolve any inconsistencies in the VADM between
 462 localities at those times.

463 The models produced by the AH-RJMCMC analysis are shown in Figure 8a-c, and
 464 the distributions of the time-averaged VADMs for each locality are plotted on the vio-
 465 lin plots in Figure 8d. Using this methodology, it is apparent that the time-averaged VADMs
 466 over the last 1.5 Ma from Hawai'i and Antarctica are indeed not consistent with each
 467 other, but the time-averaged VADM in Israel could be compatible with either of the other
 468 latitudes. However, there is not enough evidence to confirm a difference in the tempo-
 469 ral average between Hawai'i and Antarctica from 1.5-2.5 Ma, with the average VADMs
 470 appearing consistent. This implies that poor temporal sampling is not the reason for in-
 471 consistent paleointensities at different latitudes, but that some form of genuine non-dipolar
 472 field behavior that causes higher fields in Hawai'i than Antarctica at least since 1.5 Ma.
 473 More paleointensity studies with high quality paleointensity data at different latitudes
 474 (especially from the southern hemisphere) are needed to better understand the sources
 475 of this non-dipolar behavior.

476 5 Conclusions

477 In this paper, we obtained 31 high quality paleointensity results from dikes, lava
 478 flow tops and vent deposits collected in the Hawai'ian islands, with ages ranging from
 479 0-4 Ma. We demonstrate a methodology for obtaining accurate time-averaged paleoin-
 480 tensities, with uncertainties which allow direct comparison between paleointensity stud-
 481 ies at different latitudes. The use of BiCEP allows for consistent comparison of results
 482 between different studies, and using the methodology of Livermore et al. (2018) allows
 483 us to obtain a time-averaged intensity, with uncertainty, which accounts for the tempo-

484 ral distribution of our paleointensity. Because these robust statistical approaches are used
 485 for calculating time-averaged paleointensities, we are able to exclude the hypotheses that
 486 inconsistency of our time-averaged VADMs is due to either biased paleointensity data,
 487 or inconsistent temporal sampling of paleointensities.

488 Applying the new methodology to data from the Hawai‘ian islands, we find that
 489 the time-averaged paleointensity in Hawai‘i over the past 1.5 Ma was higher than dur-
 490 ing the period from 1.5-4 Ma. Comparing results from paleointensity studies at three lat-
 491 itudes, we find that this period of high paleointensity is not recorded in rocks from Antarc-
 492 tica or Israel. We reiterate the conclusion of other recent papers (e.g., Tauxe et al., 2022)
 493 that the Earth’s magnetic field averaged over the past 1.5 Ma does not conform to a Geo-
 494 centric Axial Dipole. Further time averages at a greater range of latitudes and times will
 495 be needed to obtain better estimates of the structure of this time-averaged field.

496 Our results also indicate that vent deposits containing scoria, and olivine bearing
 497 rocks which are oxidized at high temperatures are potentially good lithologies for obtain-
 498 ing high quality paleointensity estimates, with higher success rates in the paleointensity
 499 experiment. Specimens from these lithologies have strong magnetizations and tend to
 500 alter less in paleointensity experiments. Additionally, these deposits are frequently quar-
 501 ried, allowing for easy access to fresh material in the field. Despite their useful proper-
 502 ties in paleointensity experiments, and their single-domain like FORCs, the size of iron
 503 oxides in these samples when viewed under a microscope is orders of magnitude larger
 504 than would be expected for single domain grains. Further study of the magnetic carri-
 505 ers in these samples should be undertaken to understand why they have such ideal rock
 506 magnetic properties.

507 Open Research

508 All data and interpretations are available at [https://earthref.org/MagIC/19614/
 509 9208acad-0f62-4d9e-b265-4c8907d40eb7](https://earthref.org/MagIC/19614/9208acad-0f62-4d9e-b265-4c8907d40eb7) and will be made available in the MagIC database
 510 at <http://earthref.org/MagIC/19614> on acceptance of this paper. Python notebooks
 511 for producing figures can be found at <https://github.com/bcych/hawaiian-paleointensity>
 512 and the release version associated with this paper can be found at the Zenodo reposi-
 513 tory <https://doi.org/10.5281/zenodo.7921097> (Cych, 2023).

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