DATA AND METHODS

Earthquake Location and Magnitude Procedures

Earthquakes were identified and located using data from seven three-component seismic stations deployed as part of the Ocean Observatories Initiative (OOI) Cabled Array (Kelley et al., 2014). An STA/LTA algorithm operating on a band-passed (7–50 Hz) vertical channel waveform was used to detect P-wave arrivals that were associated to form an initial set of hypocenters. S-waves were then identified using a kurtosis algorithm (e.g., Baillard et al., 2014) applied to the vector sum of horizontal components. Detections of whale calls and other in-water sources were excluded based on the coherence between the seismic waveforms and co-located hydrophone sensors. The remaining P- and S-wave detections were associated to form a catalog of 41,522 hypocenters, which were then located using a 1-D velocity model derived from multi-channel seismic studies (Arnulf et al., 2014) and the generalized earthquake-location (GENLOC) package (Pavlis et al., 2004). The earthquake hypocenters were relocated using the double-difference algorithm hypoDD (Waldhauser and Ellsworth, 2000). A final catalog of 19,049 earthquakes, each with at least nine defining phases, was derived from processing differential
arrival times using an 18-step iterative conjugate gradient method with the parameters listed in Table DR1. The median hypocenter location change after relocation is 230 m and the mean of the residuals is 0.029 s. Median relative location uncertainty was estimated to be 47 m in the horizontal and 103 m in the vertical direction.

Microearthquake moments ($M_o$) were determined in the frequency domain from P and S arrivals (Brune, 1970):

$$M_o = \frac{4\pi \rho v^3 r \Omega_o}{RK}$$

where \( \rho \) is the crustal density at the hypocenter (2700 kg/m\(^3\)), \( v \) is the average seismic velocity for P (5.09 km/s) or S (2.78 km/s) waves (Arnulf et al., 2014), \( r \) is the slant range from the earthquake hypocenter to the station, \( K \) is the free-surface correction taken to be 1.5 and 1.7 for P and S waves, and \( R \) is the radiation function assumed to be 0.42 and 0.59 for P or S waves, respectively (Toomey et al., 1985; Weekly et al., 2013). \( \Omega_o \) is the low frequency level of the amplitude spectrum calculated using a Boatwright (1980) model of the form:

$$\Omega(f) = \frac{\Omega_o e^{-\pi f / Q}}{[1 + (f / f_c)^4]^{1/2}}$$

The quality factor (\( Q \)), corner frequency (\( f_c \)) and \( \Omega_o \) are found simultaneously using a non-linear least squares procedure, and the spectral amplitudes \( \Omega(f) \) are obtained using a multi-taper (7) approach with a time-bandwidth product of 4. Moment magnitudes (\( M_w \)) were estimate from the seismic moments in Nm as (Hanks and Kanamori, 1979):

$$M_w = \frac{2}{3} \log_{10} (M_o) - 6$$
**Table DR1. HypoDD Parameters**

Parameters used in the hypoDD hypocenter relocation process. The max distance between linked pairs and residual threshold became progressively smaller as iterations progressed through the relocation routine in order to minimize and remove outliers.

<table>
<thead>
<tr>
<th>Iterations</th>
<th>P weight</th>
<th>S weight</th>
<th>Max link separation (km)</th>
<th>Factor multiplied by the standard deviation of the residuals</th>
<th>Damping</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-6</td>
<td>1</td>
<td>0.5</td>
<td>3</td>
<td>8</td>
<td>240</td>
</tr>
<tr>
<td>7-12</td>
<td>1</td>
<td>0.5</td>
<td>2</td>
<td>7</td>
<td>240</td>
</tr>
<tr>
<td>13-18</td>
<td>1</td>
<td>0.5</td>
<td>1</td>
<td>6</td>
<td>240</td>
</tr>
</tbody>
</table>

**Composite Focal Mechanisms**

Composite focal mechanisms are commonly used to decrease solution uncertainty when the number of recording stations is limited and/or the signal-to-noise ratio of the arrivals is low (e.g., Waldhauser and Tolstoy, 2011). Cross-correlating the vertical channel waveforms recorded on station AXEC1 identified groups of similar earthquakes. A correlation window of 1.4 s was used starting 0.1 sec before the P wave arrival in order to capture both the P- and S-wave energy.

Event clusters were defined based on similarity between waveforms (Fig. DR1) using a hierarchical approach with a correlation coefficient cutoff of 0.7. These data were subset to select clusters with earthquakes spanning a limited time range, resulting in a set of 100 clusters that each contain between 3 and 9 earthquakes. The clusters were then used to track the evolution of fault slip through time. The median time spanned by events within an individual composite is 29.3 days, however, this varied based on the event rate throughout the volcanic cycle (22.0 days pre-eruption, 1.2 days during the eruption, and 138.0 days post-eruption) (Fig. DR2). The average horizontal radius of the events within a composite cluster is 212 m.
P-wave polarities were reviewed by the author for the suite of 501 origins belonging to the composite groups (Levy, 2017). S/P amplitude data were measured in the 3-20 Hz frequency band using the vector sum of the peak amplitudes on each of the three channels, following the procedure of Yang et al. (2012). Focal mechanism solutions were found using HASH (Hardebeck & Shearer, 2002). The best solution was found at the average cluster location, but a set of acceptable focal mechanisms was calculated using an allowed polarity misfit of 10%, an epicenter uncertainty of 50 m, a vertical uncertainty of 100 m, and small variations in the velocity model. The preferred focal mechanism was then found by averaging the acceptable focal mechanisms after the removal of outliers.

Solutions run with and without S/P information returned similar nodal plane geometries; however, when the amplitude ratio data were included the average nodal plane uncertainty (NPU) decreased from 38° to 21° and the number of A and B quality (Hardebeck & Shearer, 2002) solutions increased from 25 to 81. To further assess the sensitivity of the solution to uncertainty in focal depth, focal mechanism parameters were analyzed for a range of depths between 0.7 and 2.0 km, in 0.1 km increments. The median nodal plane rotation is <24° across this depth range, with fault plane uncertainty typically minimized near the mean centroid depth of earthquakes within the cluster.

The mean strike and dip of the inferred fault planes (taken to be the outward dipping nodal plane returned for each solution) were estimated using the vector statistics package of Jones (2005); uncertainty in the mean was found using the 95th percentile of the chi-squared distribution. For the entire dataset the vector mean strike and dip is 345±13° and 70±2°, respectively. The strike and dip vector means were also calculated during the identifiable phases
of the volcanic cycle and were found to be $333\pm21^\circ/67\pm4^\circ$ (strike/dip for pre-eruption),

$338\pm15^\circ/69\pm5^\circ$ (syn-eruption), and $007\pm27^\circ/75\pm4^\circ$ (post-eruption).

**Analysis of Bottom Pressure Data**

Bottom-pressure recorder (BPR) data were averaged over 1-minute intervals and corrected for tides by subtracting the predicted pressure change associated with ocean tides, as estimated using the OSU tidal model (Egbert et al., 1994) implemented within SPOTL (Agnew, 2012). Pressure was converted to water column height assuming a density of $1030 \text{ kg/m}^3$. A least-squares linear regression was used to estimate the inflation rates during different phases of the eruption (Fig DR3).

**Seismic Moments and Cumulative Slip Estimates**

The seismic moment for each earthquake was found using the median of the individual P and S wave attenuation-corrected spectral amplitudes. The average slip ($d$) calculation along the eastern ring fault is estimated from the sum of the scalar seismic moments: $M_o=\mu Ad$. The shear modulus ($\mu$) was taken to be 1-4 GPa, appropriate for fractured basalt in the upper crust (Gudmundsson, 2016), and the area ($A$) is estimated based on the along-strike (3.0 km) and down-dip (1.75 km) dimensions of the seismically active eastern margin of the caldera.

Assuming a 1-4 GPa range of shear modulus, the cumulative slip along the eastern portion of the ring fault was estimated to be 8-30 cm in the three months prior to the eruption, 44-175 cm over the ~1-month period during the eruption, and <0.5 cm over the 19-month period post eruption. These estimates were then related to the differential vertical motion between BPR stations AXCC1 and AXEC2 to constrain the relative coupling between the eastern fault system and the geodetic uplift.
REFERENCES FOR DATA REPOSITORY


Gudmundsson et al., 2016, Gradual caldera collapse at Bárdarbunga volcano, Iceland, regulated by lateral magma outflow, Science 353, doi:10.1126/science.aaf8988

Hanks, T. C., and Kanamori, H., 1979, Moment magnitude scale, Journal of Geophysical Research, 84, 2348–2350


**FIGURE CAPTIONS:**

**Figure DR1.** a) Vertical channel waveforms for similar events in the cluster, used to generate a composite focal mechanism. b) Vertical and the horizontal waveforms for a single arrival shown with windows (boxes) used to estimate the noise, P-arrival, and the S-arrival amplitudes. c) Composite focal mechanism with red and blue indicating upward and downward polarity, respectively, and the size of the circle scaled to represent the S/P amplitude.

**Figure DR2.** Graph showing the length of time spanned by events in each composite focal mechanism cluster. The y-axis represents the cluster numbered from 0 to 100. Each horizontal line begins and ends at the times of the earliest and latest earthquake in a cluster. Background shading denotes phases of volcanic inflation and deflation.

**Figure DR3.** Time series of elevation data from BPR stations (a) AXCC1, (b) AXEC2 and (c) AXID1 processed to remove tidal variations. Slopes indicating inflation rates are shown by dashed black lines. Background shading denotes phases of volcanic inflation and deflation.
Fig. DR1
Fig. DR2
Slower Inflation Rate

Eruption

AXCC1

AXEC2

AXID1

Fig. DR3