Evidence for Active Rhyolitic dike Intrusion in the Northern Main Ethiopian Rift from the 2015 Fentale Seismic Swarm

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Abstract Magmatic intrusions play a vital role not only in accommodating extensional stresses in continental rifts but also in feeding volcanic systems. The location, orientation, and timescale of dike intrusions are dictated by the interaction of regional and local stresses, the effect of pre-existing weaknesses, and the composition of magma. Observing active intrusions can provide important information regarding the interaction between magmatic processes and the tectonic stress field during continental rifting. We focus on a seismic swarm that occurred in 2015 to the northeast of Fentale volcano, in the Main Ethiopian Rift (MER), and use radar interferometry to study surface deformation associated with the seismic swarm. Interferograms show a pattern of dike-induced deformation, with a model estimate of volume change of $3.3 \times 10^6 \pm 0.6 \times 10^6$ m$^3$ at a depth range of 5.4 to 8 km. We use a small baseline subset algorithm to calculate line of sight time series and find that the displacements decay exponentially with a decay constant of $\tau \sim 83$ days. Coupled source-sink models suggest that such slow dike intrusions require a high viscosity rhyolitic magma. The difference in behavior between Fentale and other caldera systems in the MER, which show multi-year cycles of inflation and deflation, suggests fundamental differences in magma composition and architecture of the plumbing system. This is the first direct observation of a dike intrusion in the MER and provides new constraints on the temporal-spatial patterns of stress and strain that occur during continental rifting. Whether this activity is transient or a long-term feature associated with rift evolution is an open question.

1. Introduction

The stress accumulated along divergent plate boundaries can be accommodated by dike-driven rifting episodes (Ayele et al., 2007, 2009; Ebinger et al., 2013; Wright et al., 2012). As rifts mature, dikes become the preferred mode of strain accommodation, deactivating the large border faults (Buck, 2004; Ebinger & Casey, 2001; Wolfenden et al., 2004). At a local scale, the stress field can be perturbed by topographic load and transient magmatic and seismic processes (e.g., Sigmundsson et al., 2014; Wadge et al., 2016). dike ascent is facilitated by magma buoyancy, but interactions with the mechanically heterogeneous crust affect the propagation pathway (Kavanagh et al., 2018; Kavanagh, 2018). dikes can be arrested by magma solidification, meeting a mechanical barrier (Kavanagh, 2018) or stress barrier (Biggs et al., 2009). The stress field determines the orientation of magmatic intrusions and consequently the alignment of fissures and vents at the surface (Korne et al., 1997; Mazzarini, 2004; Wadge et al., 2016). Smaller dike intrusions and eruptive fissures associated with volcanic activities can be oriented both circumferentially and radially with respect to the summit calderas (e.g., Muirhead et al., 2015; Tedesco et al., 2007; Wadge & Burt, 2011; Wauthier et al., 2015). Understanding the pathways of magma movement and the circumstances that enable magma transport to the surface is important for understanding both tectonic and volcanic processes.

The timescale of magma ascent is controlled by the balance between the driving force and the viscous resistance of the magma (Takeuchi, 2011; Wada, 1994), which depends on the temperature, magma composition, and crystal content (Sigurdsson, 2015; Sparks et al., 2019). The inverse dependency of viscosity on temperature and the positive covariation with silica content results in low-temperature rhyolitic melt having a viscosity that is as much as eight orders of magnitude greater than that of a high-temperature basaltic melt (Takeuchi, 2011; Wallace et al., 2015). Dike intrusions in well-monitored volcanic zones typically show evidence of lateral propagation on timescales of a few hours to weeks (Irwan et al., 2006; Segall et al., 2001; Sigmundsson et al., 2014). Simple analytical models assume a hydraulic connection between the magma source and an intrusion to predict an exponentially decaying rate of intruded volume change over...
timescales of hours to days (e.g., Irwan et al., 2006; Rivalta, 2010; Segall et al., 2001). In contrast, some dike intrusions exhibit rapid lateral propagation followed by opening over several weeks (e.g., Albino et al., 2019; Biggs et al., 2009).

Geodetic data have been used to study dike intrusions with a range of durations lasting from a few hours (e.g., Dabbahu, Afar, Belachew et al., 2011; Grandin et al., 2010; Wright et al., 2006), a few weeks (e.g., Bardarbunga, Iceland, Sigmundsson et al., 2014), and up to a few months (e.g., Upptyppingar, Iceland Hooper et al., 2011). Rifting episodes lasting up to a decade (e.g., Dabbahu, Afar, Wright et al., 2006) have occurred in several extensional environments including continental, transitional, and sea floor spreading plate boundaries (e.g., Calais et al., 2008; Einarsson & Brandsdottir, 1980; Ruegg et al., 1979; Segall et al., 2001). Dike intrusions are usually accompanied by elevated seismicity and sometimes fissure eruptions (Ayele et al., 2007; Belachew et al., 2011; Einarsson & Brandsdottir, 1978; Grandin et al., 2011; Lépine et al., 1980; Ruegg et al., 1979). Geodetic observations provide constraints on dike emplacement, but the resolution is often too limited in space and/or time (in the case of InSAR) to extract smaller-scale details (Rivalta et al., 2015).

Here we investigate a magmatic intrusion that took place to the north east of Fentale volcano, in the northern Main Ethiopian Rift (MER) in 2015. We use InSAR to study the timescale of the deformation and the source mechanism. This present-day intrusion provides a snapshot of the magmatic contribution to the continental rifting process and the orientation of the current stress field.

1.1. Tectonic Setting

Extension in continental rifts is accommodated by tectonic and magmatic processes that may eventually lead to continental breakup (Ebinger & Casey, 2001; Illsley-Kemp et al., 2017). As extension increases, the lithosphere thins and spreading rate and melt supply increase (Ebinger & Casey, 2001; Keir et al., 2015). The earliest evidence for rifting in the Main Ethiopian Rift (MER) dates from ~18 Myr, but rift initiation was asynchronous along the segments (Fontijn et al., 2018; Muluneh et al., 2018; Wolfenden et al., 2004). As a result, along-strike variations in rift architecture occur in concert with along-strike variations in magma supply (Ebinger & Casey, 2001; Rowland et al., 2010).

The northern MER initiated at ~11 Ma and consists of a series of linked half grabens bounded by steep NE striking Miocene border faults (Corti et al., 2018; Keir et al., 2006; Wolfenden et al., 2004). The rift floor (<20 km wide) hosts central volcanoes and magmatic segments which form right stepping, en echelon zones of magmatism and faulting (Casey et al., 2006; Ebinger & Casey, 2001; Fontijn et al., 2018; Keir et al., 2006). These magmatic segments are characterized by high $V_p$ velocity and positive Bouguer anomalies interpreted as a cooled mafic intrusions (Cornwell et al., 2006; Keranen & Klemperer, 2008; Mahatsente et al., 1999). The crustal thickness decreases northward (Dugda et al., 2005; Maguire et al., 2006) with an associated increase in crustal extension and magmatic processes (Keir et al., 2006; Maguire et al., 2006; Tyrone et al., 2007).

Geodetic measurements show that ~80% of present-day extension across the MER is localized within the magmatic segments (Bendick et al., 2006; Bilham et al., 1999; Birhanu et al., 2016). The average rate of relative movement due to extensional tectonics is ~5–6 mm/year between the Nubian and Somalian plates at an azimuth of 96° to 109° toward the ESE (e.g., Bendick et al., 2006; Bilham et al., 1999; Birhanu et al., 2016; Déprez et al., 2013). The extension is mostly accommodated by rift floor faulting within the Wonji Fault Belt (WFB) and dike intrusions (Bilham et al., 1999; Keir et al., 2006; Kurz et al., 2007). Based on a three-dimensional seismic velocity model, extension in the ductile middle-to-lower crust is controlled by magmatic intrusion (Keranen et al., 2004). As shown by shear wave splitting, partial melt beneath the MER rises through rift-parallel dikes that penetrate through the thinned lithosphere (Kendall et al., 2005) and continue into the upper crust (Keir et al., 2005). Several volcanic edifices in the MER (e.g., Fentale and Corbetti) are elliptical in shape which may indicate that pre-existing structures control crustal magma storage over timescales of hundreds of thousands of years (Lloyd et al., 2018; Robertson et al., 2016; Wadge et al., 2016).

1.2. Fentale Volcano

Fentale volcano is located in the northern MER, north of the town of Metehara and southwest of Dofen volcano. The central elliptical summit caldera is elongated NW-SE, and the maximum elevation is 600 m
Figure 1. Location map of Fentale volcanic centre. (a) Fentale volcano and nearby volcanoes Dofen and Kone. Fentale craters, Tinish Fentale and (Tinish) Sabober (tuff cone), are represented by red lines. The thin black lines are the Wonji Fault Belt. Colored circles show the 2001 to 2003 EAGLE seismicity catalog (Keir et al., 2006), with the depth represented by the color and the magnitude of the earthquakes represented by the size of the circle. The brown diamonds show the preliminary location of 2015 earthquake swarm based on data from the Ethiopian Seismic Station Network (ESSN). Inset (b) shows the MER Quaternary faults, Miocene border faults, and location of Holocene volcanoes (after Casey et al., 2006). The black rectangle shows the region as seen in (a). The thick black lines are border faults. The red arrow represents the velocity of the Somalian plate with respect to the Nubian plate. The stars show cities, the black triangles show the Holocene volcanic centers, the red triangles represent volcanoes that show recent activities (e.g., Biggs et al., 2011; Greenfield et al., 2018; Lloyd et al., 2018), and light-blue polygons represent lakes. The location of panel (b) is shown by the white rectangle on the globe shown in the upper left.

above the rift floor (Figure 1a) (e.g., Acocella et al., 2003; Fontijn et al., 2018). Monogenetic vents have been observed within the caldera and are aligned parallel to the long axis of the caldera (Acocella et al., 2003; Hunt et al., 2019).

The most recent caldera forming eruption at Fentale occurred at 168±38 ka (Fontijn et al., 2018; Williams et al., 2004). Post-caldera eruptions of obsidian and rhyolite lava flows were sourced from NW-SE aligned vents and fissures within the caldera (Acocella et al., 2003). It is possible that there has been a small component of explosive activity as evident from fresh superficial pumice lapilli scattered around the lower slopes of Fentale (Fontijn et al., 2018; Giordano et al., 2014; Williams et al., 2004). Basaltic activity has occurred south of the edifice with the emission of lava flows from fissures (Fontijn et al., 2018; Giordano et al., 2014; Williams et al., 2004). Small basaltic tuff cones, located south of the caldera, are aligned in the direction of the Wonji Fault Belt (WFB) (Fontijn et al., 2018; Giordano et al., 2014; Hunt et al., 2019; Williams et al., 2004). The most recent of the basaltic flows is dated around AD 1810 based on historic reports (Giordano et al., 2014; Fontijn et al., 2018, references therein). Lava flows originated from fissures and small cones aligned NE-SW with the center of the tuff cone immediately south of the edifice (Fontijn et al., 2018; Giordano et al., 2014; Hunt et al., 2019; Williams et al., 2004).

There is an older, eroded caldera (Tinish Fentale) located ~8 km northeast of Fentale caldera which appears to have had a similar but earlier geological history to that of Fentale (Williams et al., 2004). Similar to Fentale,
the collapse of the caldera at Tinish Fentale was accompanied by the eruption of voluminous ash flows which solidified to welded tuffs covering the east and southeast of the complex (Hunt et al., 2019; Williams et al., 2004). Geochemical studies at Fentale suggest a pantelleritic magma composition, which evolved through fractional crystallization (Giordano et al., 2014; Hunt et al., 2019; Webster et al., 1993).

Swarms of earthquakes with local magnitudes up to M4 were observed between January and March 1981 (Asfaw, 1982) and in June 1989 to the northeast of Fentale volcano (Laike and Ayele, personal communication, January 10, 2018). The 1981 swarm caused surface cracks parallel to the rift axis (Asfaw, 1982; Keir et al., 2006; Williams et al., 2004). The EAGLE (Ethiopian-Afar Geoscientific Lithospheric Experiment) seismic network (2001 to 2003) detected earthquakes aligned with the rift axis with magnitude <M4.5 and depth ranges from 6 to 9 km beneath Fentale's edifice and up to 16 km deep between the edifices of Fentale and Dofen volcanoes (e.g., Keir et al., 2006, 2009). Most recently, a seismic swarm was detected by the Ethiopian Seismic Stations Network to the north of Fentale from 22 March to 20 April 2015 (Figure 1). The sparse network of seismic stations in the region makes it challenging to observe the spatio-temporal pattern of seismicity in any detail. However, with the available network, it is clear that the swarm focused mostly in the area around Tinish Fentale (Figure 1).

Fentale volcano is located 7 km from the main highway and railway that connects Addis Ababa to Djibouti. Metehara and Awash towns and the newly constructed Kesem dam is within a 20 km radius of the volcano. Nonetheless, there is no dedicated monitoring facility or detailed study of the magmatic activity beneath the volcano. Here, we focus on the 2015 activity using multiple InSAR observations between 2014 and 2018 to constrain the spatial and temporal evolution of ground deformation.

2. Geodetic Data

2.1. InSAR Processing

We use InSAR to measure ground deformation before, during, and after the 2015 seismic swarm near Fentale volcano. We use satellite radar data from the European Space Agency (ESA) Sentinel-1 and Italian Space Agency (ASI) Cosmo-SkyMed satellite constellations. Sentinel-1 operates at C-band which has a wavelength of 5.6 cm and operates in Terrain Observation by Progressive Scans (TOPS) acquisition mode (González et al., 2015; Grandin et al., 2016; Torres et al., 2012; Yagüe-Martínez et al., 2016). The Sentinel data we used here are in Interferometric Wide Swath (IW) Mode, which has a spatial resolution of 5 by 20 m and a 250 km swath (Torres et al., 2012). We use the LiCSAR processor which builds on the Gamma SAR and Interferometry software and runs on the JASMIN "super-data-cluster" to produce interferograms and coherence maps (Anantrasirichai et al., 2018; Li et al., 2016; Morishita et al., 2020; Spaans et al., 2017; Wright et al., 2016). We produced 55 interferograms from the ascending track and 139 interferograms from the descending track from 2014 to 2018. The Sentinel-1 revisit interval in East Africa is typically 24 days. The interferograms are constructed based on the shortest possible temporal baseline between acquisitions. The high coherence in short-period interferograms enables us to capture time-dependent deformation and to reduce atmospheric noise (Elliott et al., 2016; Torres et al., 2012).

Based on preliminary analysis of the Sentinel-1 data, we requested archive Cosmo-SkyMed data from 2014 to 2015. The Cosmo-SkyMed radar constellation operates at X-band with a 3.1 cm wavelength and has a short repeat interval (4 to 9 days in the MER), and spatial resolution of 3 to 5 m for stripmap mode with a swath width of 40 km (Covello et al., 2010). The high temporal and spatial resolution of the Cosmo-SkyMed data complements the Sentinel-1 data. However, the large perpendicular baselines can cause loss of coherence and careful image selection is required (see supporting information Table S1) (Covello et al., 2010). We use the ISCE interferogram processor software (Rosen et al., 2012) to produce Cosmo-SkyMed interferograms. We used 42 ascending and 38 descending images from the Cosmo-SkyMed constellation and produced 67 and 57 interferograms respectively with temporal baselines of 5 to 60 days and perpendicular baselines <300 m.

We removed the topographic phase using the NASA Shuttle Radar Topography Mission 1 arcsec (30 m) DEM (Farr et al., 2007). As there is negligible interferogram wide deformation, we removed the reference phase by setting the average displacement of each interferogram to zero. We calculate and remove the best fitting linear plane for each interferogram to remove residual long wavelength delays caused by errors due to large perpendicular baselines (Fattahi et al., 2017; Liu et al., 2014; Liang et al., 2019; Yagüe-Martínez et al., 2016). Atmospheric (i.e., tropospheric and ionospheric) effects are a major challenge for current repeat
pass InSAR techniques (Gray et al., 2000; Liang et al., 2019). The topographically correlated tropospheric delay can be approximated using a linear function with height (Elliott et al., 2008) or weather models such as ECMWF (e.g., Yu et al., 2018). We tested the topographically correlated atmospheric correction assuming a linear function with height, but the values are within the overall error budget so that the correction is not applied in this case. The ionospheric delay, unlike the tropospheric delay, is dependent on the wavelength of the signal. The ionospheric effects are minimal at short wavelength (e.g., X- and C-band InSAR) and higher in long wavelength data (e.g., L-band InSAR) (Gray et al., 2000; Liang et al., 2019). Ionospheric effects are strongest at low latitude (0° to 20°) (Fattahi et al., 2017; Gray et al., 2000; Liang et al., 2019) including Ethiopia but mostly cause long wavelength signals which, for a small region of study such as Fentale volcano, can be removed using a linear ramp (Bagnardi & Hooper, 2018).

### 2.2. Individual Interferograms

Interferograms spanning the earthquake swarm of March/April 2015 all show surface deformation to the north of Fentale volcano, near Tinish Fentale (Figure 2). The descending tracks of both radar satellites show motion toward the satellite of ∼5 cm in the satellite line of sight (LOS) around Tinish Fentale (Figures 2b and 2g) and the ascending tracks show the same sense of motion but with a magnitude of only 3 cm LOS within a 5 month period (Figures 2d and 2j). The deformation center is located about 8 km north of the summit caldera but is shifted to the west in the ascending track compared to the descending track. The difference between ascending and descending scenes implies a significant component of horizontal motion, and the spatial pattern of deformation is typical of a dike-like intrusion. Individual interferograms do not show any
Figure 3. Bayesian inversion for deformation source parameters. Sentinel-1A 1 year stacks (first column), forward model using the maximum a posteriori probability solution (second column), and residual maps (third column). The black rectangle on the model plots represents the outline of the optimal dike plane, with the thicker line outlining the upper edge of the dike. The data column shows the LOS cumulative displacement of the deformation for ~1 year. The dates of interferogram stacks are Sentinel ascending (11/10/2014 to 22/01/2016), Sentinel descending (23/10/2014 to 10/01/2015), Cosmo-SkyMed ascending (27/06/2014 to 07/12/2015) and Cosmo-SkyMed descending (04/07/2014 to 22/12/2015).

significant deformation at the volcano summit, but we use time series to look for slow deformation (see section 5).

The regular acquisition of Sentinel-1 data enables us to follow the temporal history of the intrusion over 2 years. We use the non-overlapping interferograms to examine the behavior of the deformation over time (Figure 2). Interferograms with temporal baseline of less than or equal to 24 days over the seismicity period show <15 mm LOS range change (Sentinel, Figures 2a, 2c, 2e, and 2f and Cosmo-SkyMed, Figures 2h, 2i, 2k, and 2l) whereas those with baselines >48 days show 50–55 mm range change (Figures 2b and 2g) demonstrating that the deformation developed slowly and continued beyond the period of observed seismic activity. Interferograms from the Cosmo-SkyMed data (Figures 2g–2l) compliment Sentinel data by showing small magnitude deformation for short temporal baselines. We will discuss the temporal behavior in more detail in section 5.2.
Table 1

| Model parameters | Length (km) | Width (km) | Bottom depth (km) | Dip (degree) | Strike (degree) | Opening (m) |
|------------------|-------------|------------|-------------------|--------------|----------------|-------------|
| Lower            | 3.0         | 1.0        | 2.0               | –90          | 20             | 0.01        |
| Upper            | 8.0         | 8.0        | 10.0              | 90           | 50             | 2.0         |
| Optimal          | 6.1         | 2.6        | 8.1               | 90           | 29             | 1.9         |
| 2.5%             | 5.1         | 2.3        | 7.8               | 89           | 28             | 1.7         |
| 97.5%            | 6.9         | 3.3        | 8.6               | 90           | 31             | 1.9         |

Note. The table lists the lower and upper bound of the priors, and optimal, 2.5% and 97.5%, a posteriori probability solutions.

3. Source Model

To investigate the dimension and location of the deformation source, we use a Bayesian inversion technique implemented in GBIS (Geodetic Bayesian Inversion Software) (Bagnardi & Hooper, 2018). GBIS is an open source MATLAB code (http://comet.nerc.ac.uk/gbis) that uses simple analytical models to characterize magmatically or tectonically induced deformation processes (e.g., Bagnardi & Hooper, 2018; Segall, 2010).

We use the LOS cumulative displacements of Sentinel-1 and Cosmo-SkyMed ascending and descending interferograms to perform the Bayesian inversion. We use the cumulative displacement over 1 year rather than individual interferograms to include the majority of the deformation (Figure 3) and because individual interferograms with long time spans suffer from decorrelation (see Figure S1). Each image is subsampled by employing adaptive quadtree sampling to reduce data noise and computational burden (Bagnardi & Hooper, 2018; Decrèm et al., 2010). We perform the inversion to estimate the deformation source parameters and explore a large number \(10^6\) of model parameter combinations (Bagnardi & Hooper, 2018; Temtime et al., 2018). We used a rectangular dislocation model in an elastic half space (Okada, 1985) to model the dike intrusion with eight model parameters: length, width, depth, dip angle, strike, location coordinates of the midpoint of the edge, and uniform opening. A uniform shift (offset) and a linear ramp for each interferogram are also estimated in the model region while simultaneously fitting the deformation. As we do not have a priori information for model geometry, we assume a range of values based on the spatial pattern of deformation (Table 1). The range of values for the length are 3 to 8 km, width 1 to 8 km, and uniform opening 1 cm to 2 m. We let the dip angle vary from east dipping to west dipping and impose a constraint on the strike angle of between 20° and 70° based on the orientation of the deformation and rift axis. The posterior probability density functions (PDF) of the source parameters show an approximately Gaussian distribution (Figures S2 to S4). The maximum a posteriori probability solutions with 95% confidence interval are reported in Table 1.

The inversion shows that the observed surface displacements can be explained by a \(\sim 6 \pm 1\) km long dike intrusion, with a bottom depth of \(\sim 5.4\) km, and uniform opening of \(\sim 2 \pm 0.2\) m. The dike strikes at \(\sim 29° \pm 2°\), and the dip is vertical. The aspect ratio (length-to-thickness ratio) of the dike is \(\sim 3 \times 10^3\), consistent with previously observed dikes which have aspect ratios of \(10^2\) to \(10^4\) and thicknesses ranging from centimeters to hundreds of meters (Krumholz et al., 2014; Rubin, 1995). The model fits the data well, but there are remaining residual phase discontinuities that are likely to be associated with surface cracking caused by fault slip at a shallow depth.

4. Surface Fractures

We observed phase discontinuities in individual interferograms from both ascending and descending tracks that span the seismic swarm (Figure 4) and in the residual interferograms after modeling (see section 3 and Figure 3). These discontinuities were probably caused by ground cracking as a result of shallow fault slip, above the dike intersection (e.g., Asfaw, 1982; Gudmundsson, 2005). Ground cracking as a result of a dike intrusion is not unusual and has been observed at Dabbahu, Afar (Dumont et al., 2016), Krafí rift zone, Iceland (Rubin, 1992), and Lake Natron, Tanzania (Biggs et al., 2009; Calais et al., 2008).

Traces of the surface fractures were digitized from individual interferograms so as to compare their orientations with respect to the dike, WFB, and border faults. Layover due to the presence of topographic scarps
Figure 4. Orientation of faults, dike, and surface offsets. (a) The interferograms show the dike intrusion (Red: motion toward the satellite. Blue: motion away from the satellite). Black lines represent mapped faults, magenta solid lines represent surface offsets mapped from both geometries of InSAR, and magenta broken lines represent surface offsets mapped from only one track. A, B, C, and D represent the observed ground cracks, and the black hashed line represents the modeled dike location. The brown diamonds represent the 2015 earthquake swarm. The rose diagrams show the orientations of (b) border faults, (c) WFB (red) and dike (black), and (d) InSAR offsets. (e) The profiles XX and YY show the offsets in both ascending (red) and descending (blue). Cartoons show that (f) for shallow dike intrusion, crustal dilation is mostly accommodated by extension fractures and normal faults, whereas (g) a deeper intrusion accommodates almost all extension with lesser faults at a shallow depth (after Accocella & Trippanera, 2016; Rubin, 1990).

and unwrapping errors can also produce phase discontinuities (e.g., Dumont et al., 2016), so to avoid tracing geometric artifacts, we compare ascending and descending tracks and only map discontinuities present in both geometries (magenta solid lines in Figure 4a). We show discontinuities observed in just one track using magenta broken lines (Figure 4a) as we cannot discount the possibility that these are also caused by fault slip. We identified four ground cracks over a period of 2 months which are observed in both geometries, with lengths of 0.6 to 4.5 km and offsets up to 2 cm. The length, orientation, and offset of each discontinuity are reported in Table 2.

The cracks observed at Fentale are 3 to 5.5 km from the dike, consistent with our source modeling of a deep dike (5 to 8 km). The cracks on the east side of the surface projection of the dike are closer to the dike than the west, which could indicate an asymmetry in the dike intrusion or a heterogeneity in the crustal properties.

| Crack name | Length (km) | Orientation from North (degrees) | Offsets (mm) | Distance from dike center (km) |
|------------|-------------|---------------------------------|--------------|-------------------------------|
|            | Ascending   | Descending                      | Ascending    | Descending                    | Ascending | Descending |            |
| A          | 1.4         | 0.7                             | 35 ± 5       | 030 ± 5                       | 006 ± 2   | 4 ± 2      | 5.5        |
| B          | 0.6         | 0.6                             | 5 ± 5        | 005 ± 5                       | 005 ± 1   | 6 ± 2      | 3          |
| C          | 1.5         | 0.9                             | 5 ± 5        | 005 ± 5                       | 006 ± 3   | 5 ± 2      | 3.8        |
| D          | 3.5         | 4.5                             | 45 ± 5       | 045 ± 5                       | 012 ± 3   | 20 ± 4     | 3.4        |

Note. Surface cracks B and C are parallel to Wonji Fault Belt and A and D are parallel to the border faults. The offsets to the east of the dike are bigger in magnitude than the offsets to the west.
The orientation of the surface offsets can be used to test whether the slipping faults align with pre-existing structures or are new faults controlled by the local stress field. We use the fault map from Agostini et al. (2011) (updated version available from http://ethiopianrift.igg.cnr.it/) to analyze the orientation of the fault traces in terms of azimuthal distribution. We use the QGIS Line Direction Histogram plugin (Tveite, 2015) to produce length-weighted rose diagrams that show the orientations of the border faults, the WFB, and the digitized phase discontinuities (Figures 4b–4d).

We find that the faults associated with the WFB near Fentale are orientated dominantly N25°E ± 10° and the Miocene border faults trend N37°E ± 5° in agreement with previous regional studies (e.g., Agostini et al., 2011; Kidane et al., 2009). The extension direction is N96°E–N109°E and under Andersonian considerations (Anderson, 1905; Célérier, 2008), normal faults and dikes would be expected to strike N6°E–N19°E. The orientations of the InSAR discontinuities are N5°E–N45°E (see Table 2). The orientation of ground cracks B and C are subparallel to the WFB and the tectonic stress field, whereas cracks A and D are broadly aligned with both the dike and the border faults. The orientation of the dike intrusion determined from the inversion is N29°E ± 2° which lies between the orientations of the WFB and border faults (Figures 4b–4d). We conclude that the dike at least partial follows pre-existing Miocene structures at depth and triggered shallow slip on faults of the WFB.

5. Time Series of Surface Deformation

5.1. Observation

To investigate the temporal evolution of the dike intrusion, we analyzed the Sentinel-1 and Cosmo-SkyMed ascending and descending data following the multi-interferogram method of Biggs et al. (2007) Elliott et al. (2008), and Wang et al. (2009) to produce a time series for each track. Poly Interferogram Rate And Time Series Estimator (̂RATE) is an open source package of MATLAB codes designed to estimate the displacement rate map and time series, and their associated uncertainties based on a set of geocoded interferograms (Biggs et al., 2007; Elliott et al., 2008; Wang et al., 2009).

We use a Laplacian smoothing operator in the inversion to smooth the time series (Schmidt & Bürgmann, 2003) and remove residual turbulent atmospheric errors as well as other error sources such as thermal noise (Hanssen, 2001; Hooper et al., 2012). Here, we constrain the smoothness of the velocities (Wang et al., 2012) between consecutive acquisitions rather than incremental displacements as used in Schmidt and Bürgmann (2003). We selected a smoothing factor for the best-fit time series using a trade-off curve between weighted misfit and solution roughness (Schmidt & Bürgmann, 2003; Wang et al., 2012). We tested the inversion with different smoothing factors to check for over/under smoothing and selected a factor of 10^−2 (see Figure S5).

We calculated the LOS cumulative displacement for each pixel of the multi-looked image of the four tracks on a pixel by pixel basis (Figure 5). We illustrate the results using pixels on either side of the dike (A - West side and B - East side) and a pixel on Fentale crater (C) (Figure 2). The maximum cumulative displacement over 2 years is 70 mm at A, and at B is 80 mm (Figure 5). In the ascending scenes of both Sentinel and Cosmo-SkyMed, pixel A on the west side of the dike shows exponentially increasing LOS displacement, corresponding to west and/or upward motion. Similarly, in the descending scenes, pixel B on the east side of the dike shows exponentially increasing LOS corresponding to east and/or upward motion. In both ascending and descending scenes, pixel C in the center of the crater shows slow subsidence with a maximum cumulative displacement of 40 mm in 2 years (Figure 5). However, it is not possible to determine whether the slow subsidence was triggered by the intrusion or had already started beforehand.

5.2. Timescale of Intrusion

The temporal history of the surface deformation signal can provide useful information about the mechanism and timescale of magma transport. Here, we apply an exponentially decaying model following Rivalta (2010) (Equation 1) to estimate the time constant of the intrusion by simultaneously fitting the four LOS time series (Figure 6). The exponential equation estimates the timescale, start date, and asymptotic value of displacement. We apply a non-linear inversion technique in a least squares sense to simultaneously fit all tracks of data with a single model.

\[ d_i(t) = d_\infty \cdot [1 - e^{-(t-t_0)/\tau_i}] \]
Figure 5. LOS time series of cumulative displacement for selected pixels from 2014 to 2017. A is west of the dike, B is east of the dike, and C is on Fentale crater (Figure 2). The gray shaded region shows the observed seismicity period. A and B show an exponential decay in the ascending and descending tracks, respectively. C shows a linear trend in all four tracks. See Figure 6 for comparison of the time series on east and west side of the of the dike.

where \( d_i(t) \) is the cumulative displacement for track \( j \) and \( d_\infty,j \) is the potential maximum displacement for that track. Two parameters apply to all four tracks: the start time, \( t_0 \), and the intrusion time constant \( \tau_i \).

The exponential model fits the InSAR observation within error for all tracks with a time constant, \( \tau = 83 \) days (Figure 6). The best-fit start date is 28 February 2015 which is 23 days before the seismicity was detected by the national seismic network.

The time series of the Cosmo-SkyMed data shows a slight decrease in rate after \( \sim 60 \) days which might suggest two separate intrusive events. Hence, we split the time series into two and attempt to fit an exponential

Figure 6. Exponential decay of surface displacement. A is west of the dike, and B is east of the dike (Figure 2). The blue line is the model and the red points are the data. (a) and (b) are the Sentinel-1 time series in the ascending and descending tracks respectively. (c) and (d) are the Cosmo-SkyMed time series in the ascending and descending tracks respectively. \( d_\infty,j \) is the potential maximum displacement where \( j \) represent S1A-Asc (Sentinel-1A ascending), S1A-Desc (Sentinel-1A descending), CSK-Asc (Cosmo-SkyMed ascending) and CSK-Desc (Cosmo-SkyMed descending) tracks respectively.
Table 3  
Estimated Timescale of Dike Intrusions Calculated Using an Exponential Fit

| Dike name          | Timescale   | Maximum value   | Length (km) | Data source                                                                 |
|--------------------|-------------|-----------------|-------------|----------------------------------------------------------------------------|
| Kilauea, 1997      | 8.8 ± 0.2 hr| 25 ± 0.3x 10^6 m^3 | 5.1         | Volume history from tilt and GPS (Desmarais & Segall, 2007; Segall et al., 2001) |
| Miyakejima, 2000   | 5.9 ± 0.2 hr| 52 ± 1x 10^6 m^3  | 5.5         | Time-dependent model of the dike (Irwan et al., 2006)                         |
| Krafla, 10–11 July 1978 | 12 ± 2 hr     | 33 ± 2 km      | 30          | Migration of seismicity (Einarsson & Brandsdottir, 1980)                      |
| Afar, 25 July 2006 | 3.2 ± 1 hr   | 11 ± 1 km      | 9           | Migration of seismicity (Hamling et al., 2009; Keir et al., 2009)             |
| Agung, 2017        | 16 ± 4 days  | 66 ± 5 x 10^6 m^3 | 6           | Volume change from InSAR (Albino et al., 2019)                               |
| Fentale, 2015 (this study) | 83 ± 1 day | 73 ± 5 mm       | 6.1         | Cumulative displacement from InSAR                                           |

Note. For the Fentale dike, the maximum displacement of 73 ± 5 mm is from the Sentinel-1 descending track.

decay curve to each (Figure S6). The fit to the data did not improve, despite the extra parameters and we concluded the variability in rate of displacement was caused by noise in the individual time series rather than an underlying process.

The time constant of τ = 83 days is unusually long compared to previous studies of dike intrusions in a range of tectonic settings (Table 3). The propagation rates of dikes in Afar from 2005 to 2010 have been constrained by seismicity and intrusions that occur in few hours (Belachew et al., 2011; Grandin et al., 2011). The migration of seismicity for the Afar dike intrusions on 17 June and 25 July 2005 had time constants of 1.5 and 3.2 hr and lengths of 8 and 11 km, respectively (Ayele et al., 2009; Grandin et al., 2009; Rivalta, 2010; Wright et al., 2006). The geodetic observations of dike intrusions in Afar from 2005 to 2010 showed the intrusions appear in the time series as a sudden offset, which indicates that displacement occurred within a day (Hamling et al., 2009).

6. Discussion
6.1. Dike Intrusion

We observed surface deformation and cracks associated with the 2015 seismic swarm in the northern MER and attribute these observations to the slow intrusion of a ~6 km long dike to the north of Fentale volcano. Dike intrusions elsewhere in the EAR (Lake Natron, Tanzania, e.g., Baer et al., 2008; Biggs et al., 2011, 2013; Calais et al., 2008) and Afar (Southern Afar, Keir et al., 2011; Dabbahu, Hamling et. al., 2009, and Wright et al., 2006; Dallol, Nobile et al., 2004) are well documented. The 2015 Fentale dyking event is the first to be observed using InSAR in the MER, confirming previous inferences of dike intrusions deduced from geodetic and geophysical methods (e.g., Bendick et al., 2006; Keranen et al., 2004; Kendall et al., 2005; Maguire et al., 2006).

The deformation associated with the 2015 Fentale dike intrusion continued for at least 10 months and decayed exponentially with a time constant of ~83 days. Therefore, we conclude that the lateral propagation took less than 24 days, and the opening continued to increase after the dike had reached its maximum length. This is consistent with previous dike intrusion events that have shown rapid lateral migration of seismicity and fissure opening, with typical rates of 0.01 to 10 m/s (Belachew et al., 2011), 2 km/day over 4 days (Keir et al., 2011), and with longer timescale geodetic observations that show several days or weeks of widening (Biggs et al., 2009; Grandin et al., 2010; Hamling et al., 2009; Sigmundsson et al., 2014).

Though the magnitude of deformation decreases exponentially over 10 months (τ = 83 days) the spatial pattern of deformation does not change with time, indicating that the source did not propagate laterally during this period. If any long-term lateral propagation occurred, it was below the detection threshold of the InSAR (<1 m over 10 months). However, it is not consistent with current models of dike intrusions which generally assume that propagation and widening occurs simultaneously (Rivalta, 2010; Segall et al., 2001). Based on the spatial pattern of the surface deformation, we estimated the source geometry to be a
nearly vertical dike intrusion, at a depth of ~ 5.4 ± 0.5 km and with an opening of ~ 2 ± 0.2 m. The preliminary location of the seismicity agrees well with the location of the dike intrusion mapped from radar interferometry (Figure 1). The orientation of the dike is subparallel to the border faults and oblique to the current extension direction which suggests reactivation of pre-existing weaknesses.

6.2. Intrusion Timescale

The Earth’s response to major stress change caused by large rifting episodes has been measured in extensional settings such as in Afar and Iceland (e.g., Foulger et al., 1992; Jacques et al., 1996; Wright et al., 2012). For large dike intrusions, post-rifting deformation due to viscoelastic relaxation has been observed for years to decades (e.g., Bürgmann & Dresen, 2008; Wright et al., 2012). However, the initial mega-dike intrusions in Afar and Iceland involved a much large volume than that reported here and hence produced much larger stress changes. The Fentale intrusion is more comparable in volume to the small dikes intruded after the Dabbahu mega-dike, and these did not produce measurable viscoelastic response (Hamling et al., 2009). We conclude that the exponential decay observed at Fentale is unlikely to be caused by a viscoelastic response of the upper mantle or the lower crust, and instead we prefer a model that explains exponential decay based on the magmatic connection between reservoir and dike.

We use the source-sink model of Rivalta (2010), which gives an analytic solution for a spherical source connected with a penny-shaped intrusion at the same depth through a conduit or a channel. The dominant physical factor controlling the dike velocity and volume history is the pressure drop in the magma source, controlled by compressibility, and mass conservation (Rivalta, 2010). The timescale of the intrusion is controlled by the hydraulic connectivity between the source (a pressurized source) and the sink (a dike). The time evolution of the intrusion can be explained using Poiseuille flow through a cylindrical conduit. The timescale of a dike intrusion, $\tau_i$, can be expressed as

$$\tau_i = \frac{64\eta(1-\nu)La^3}{3\mu\pi R^4}$$

where $\eta$ is viscosity, $\nu$ is Poisson’s ratio, $R$ and $L$ are the radius and length of the conduit, respectively, $a$ is a radius of penny-shaped intrusion, and $\mu$ is the rigidity of the host rock.

Using Equation (2), we tested a range of values for viscosity ($\eta$), conduit radius ($R$), length ($L$), and radius of penny-shaped intrusion ($a$) to observe their effect on the timescale of an intrusion (Figure 6). We considered viscosity values of $10^7$ Pa s to $10^{10}$ Pa s to represent rhyolite magmas (0.1 MPa and 800°C to 1000°C) (Lesher & Spera, 2015), $10^4$ Pa s to $10^6$ Pa s (150 MPa and <750°C) to represent peralkaline rhyolite melt viscosity (Macdonald, 2012) and 1 Pa s to 100 Pa s to represent basalt (0.1 MPa and 1100°C to 1400°C) (Lesher & Spera, 2015). We consider a conduit radius of 1–100 m; smaller conduits would freeze and are only feasible for short timescales whereas larger conduits might represent dike-like feeding systems. The magma compositions considered are volatile free and the temperature ranges chosen to represent the corresponding composition (Lesher & Spera, 2015).

For low viscosities (basaltic magma), the predicted timescales are short (e.g., $\tau_i < 24$ hr, for $R = 10$ m), whereas for rhyolitic magma, the predicted time-scale is much longer ($\tau_i > 1$ year, for $R = 10$ m). The parameters required to give a value of $\tau_i = 83$ days (corresponding to the estimate for the Fentale dike) involve high viscosities and wide conduits. The solution is not unique, but if we assume, for example, $R = 5$ m, $a = 2$ km (half length of the dike), $L = 5$ km (distance from Fentale to the dike), Poisson’s ratio $\nu = 0.27$ is determined from the ratio between compressional and shear wave velocities, the shear modulus $\mu = 27$ GPa determined from Young’s modulus and Poisson’s ratio as reported in (Daly et al., 2008), then viscosity ($\eta$) = $2.2 \times 10^5$ Pa s (point F on Figure 7). The viscosities required to match the timescale of the Fentale dike are consistent with a peralkaline rhyolitic composition. For comparison, we estimated parameters for the fast (9 hr) individual dike intrusions in Afar (Belachew et al., 2011) and Krafla (Buck et al., 2006). For example, if $R = 10$ m, $a = 9$ km, $L = 5$ km, $\mu = 10$ GPa, $\nu = 0.25$, then the viscosity ($\eta$) = 174 Pa s (point A in Figure 7). This low viscosity value is consistent with a basaltic composition, as seen in the associated eruptions (Ferguson et al., 2010, 2013).

6.3. Magma Plumbing System

6.3.1. Connection to Fentale

The depth and geometry of the magmatic system beneath Fentale are not clearly known. However, seismic studies during the EAGLE project showed that the hypocentral depth of earthquakes beneath Fentale is...
Figure 7. A log-log plot of the predicted timescale of a dike intrusion as a function of viscosity using the source-sink model of Rivalta (2010) and Equation (2). We used typical viscosity values of $10^7$ Pa s to $10^{10}$ Pa s (at 0.1 MPa and 800°C to 1000°C) to represent rhyolite magmas (Lesher & Spera, 2015), $10^4$ Pa s to $10^6$ Pa s (150 MPa and <750°C) to represent peralkaline rhyolite melt viscosity (Macdonald, 2012), and 1 Pa s to 100 Pa s (at 0.1 MPa and 1100°C to 1400°C) to represent basalt (Lesher & Spera, 2015). We used the rigidity of the host rock $\mu = 10$ GPa, Poisson’s ratio $\nu = 0.25$ and conduit length $L = 5$ km. Points F and A represent plausible parameter combinations for Fentale and Afar, respectively.

about 6–9 km, but 16 km between volcanic centers (Keir et al., 2006; Keranen et al., 2004). The shallowing of the brittle-ductile transition beneath Fentale suggests a hot zone of magma accumulated over a range of depths. This could be a series of stacked melt lenses (e.g., Field et al., 2012; Kendall et al., 2005) or a zone of partially crystalline mush (e.g., Cashman et al., 2017; Iddon et al., 2019).

In the coupled source-sink model, we might expect to observe subsidence associated with the source (e.g., Afar (Grandin et al., 2010; Wright et al., 2006), Hawaii (Segall et al., 2001)), but there is no obvious subsiding region associated with the 2015 Fentale dike intrusion. We observe subsidence of 10–20 mm/year at Fentale crater, but the subsidence began at least 6 months prior to the onset of seismicity, and there is no significant change in subsidence rate at the time of the seismic swarm. If the subsidence at Fentale crater is the source of the dike, it implies that the intrusion began several months before dike-related seismicity or deformation was detectable. However, since long-term deformation is common at calderas in the Main Ethiopian Rift (Biggs et al., 2011; Hutchison et al., 2016; Lloyd et al., 2018), it is simpler to assume that these are two independent processes.

There are several reasons why the source feeding the dike intrusion might not be detectable. (1) The source is large enough so that there is a small change in pressure for a large volume extraction (e.g., Mastin et al., 2008; Rivalta & Segall, 2008; Segall et al., 2001). (2) The source is deep enough that the pressure change causes little surface deformation (e.g., Ebmeier et al., 2013; Pritchard & Simons, 2002). (3) The magma chamber is compressible (e.g., Mastin et al., 2008; Rivalta, 2010). Compressibility depends on the shape of the chamber and the elastic parameters of the embedding medium. For example, the compressibility of ellipsoidal chambers depends on their aspect ratio and varies continuously between spherical (stiff and low compressibility) to penny-shaped cracks (compliant and high compressibility) (Rivalta, 2010). The other factor that affects the compressibility is exsolution of gasses. Gas free magma has low compressibility, and compressibility is relatively large if volatiles exolve (Kilbride et al., 2016; Mastin et al., 2008).

In Figure 8, we propose possible sources for the Fentale intrusion based on our observations, modeling, and the limited historical records. One possibility is that the dike was fed laterally by a shallow silicic reservoir beneath Fentale (Figure 8a). Most caldera systems in the MER have peralkaline rhyolitic magma reservoirs
as inferred from petrological studies (e.g., Gleeson et al., 2017; Iddon et al., 2019). In this scenario, the composition of the dike is peralkaline rhyolite with a viscosity ($10^4$–$10^6$ Pa s) and is consistent with our observation of a slow intrusion. Tinish Fentale is an alternative source and lies closer to the region of seismicity (Figure 8b). However, (1) we do not observe any deformation in this area to suggest a shallow source, and (2) Tinish Fentale is a much older edifice and is now mostly eroded (Williams et al., 2004).

An alternative mechanism is that the 2015 dike intrusion is fed directly from a deep source (Figure 8b). In this scenario, we would not expect to see significant surface deformation associated with the source, but the composition of the magma would likely be basaltic, and we would expect a much faster intrusion than that observed.

A combination of these models may be needed to explain the observations of the 1810 eruption. Rhyolitic lava erupted within the crater, presumably from a shallow reservoir beneath the summit, and basaltic magma also erupted on the flank and to the south at Tinish Sabober, which may have come directly from a deeper source.

6.4. Synthesis of Observations in Continental Rifts

Geodetic observations have demonstrated how the tectonic stress field, local topographic loading, and pre-existing weaknesses combine to control the behavior of magmatic intrusions providing new insights into continental rifting. As the resolution in time and space of satellite observations increases, we are able to constrain better the source mechanism of deformation. The 2015 dike intrusion (this study) was the first to be observed related to the Nubia-Somalia motion using InSAR, but previous studies have observed dike intrusions in other ways. For example, a 1993 dike in central MER was inferred from campaign GPS observations related to the Nubian-Somalian plates separation but could not constrain the geometry (Bendick et al., 2006). The 1993 dike was spatially coincident with active magmatic segments and may have been triggered by a M5.3 strike-slip earthquake (Bendick et al., 2006). The 1993 dike was spatially coincident with active magmatic segments and may have been triggered by a M5.3 strike-slip earthquake (Bendick et al., 2006). A magma intrusion was observed 90 km to the northeast of Fentale volcano in May 2000 in southern Afar; however, the orientation of this dike was consistent with the tectonic stress field caused by the separation of Arabia from Nubia (Ayele et al., 2006; Keir et al., 2011) and hence was not associated with the MER.

The seismic swarm that occurred at Tinish Fentale in 2015 was similar to the seismic swarm at Dofen volcano (Figure 1) recorded during EAGLE project between 2001 and 2003. The sporadic seismicity observed in the past and the recent activity may suggest that multiple dike intrusions have occurred in the northern Main Ethiopian Rift (e.g., Ayele et al., 2006; Keir et al., 2006).
Caldera volcanoes in the central MER (e.g., Tulu Moye, Aluto, and Corbetti) typically show multi-year cycles of uplift and subsidence that reveal frequently recurring shallow magmatic activity (Biggs et al., 2011; Hutchison et al., 2016; Lloyd et al., 2018). At Tulu Moye the seismic swarms are triggered by hydrothermal circulation along the pre-existing fractures (Greenfield et al., 2018). At Aluto, the source of inflation is inferred to be a deep magma source, and the deflation is attributed to the shallow hydrothermal system (Hutchison et al., 2016), whereas at Corbetti there is sustained uplift as a result of magma reservoir growth (Lloyd et al., 2018). No deformation was detected at Fentale volcanic center during the rift scale InSAR survey between 1993 and 2010 (Biggs et al., 2011).

The difference in behavior between the volcanoes in the northern MER and those in the Central MER suggests different control mechanisms are operating. In the northern MER, the tectonic stress field and fault structures control the characteristics of the deformation episodes, whereas in the Central MER, the deformation patterns are dominated by the processes associated with long-lived magmatic systems. This suggests that the stress regime in the northern MER is currently controlled by large-scale tectonics and characterized by brittle failure, whereas in the Central MER, the crust around the magmatic centers has been thermally weakened by the presence of large magma bodies (Hutchison et al., 2016; Lloyd et al., 2018; Wilks et al., 2017). Our observation is consistent with the conceptual model that dike intrusions accommodate extension at the later stages of rifting (Hayward & Ebinger, 1996). From this short snapshot of observations, it is not possible to determine whether this is a temporary situation or associated with the long-term rift evolution (Rowland et al., 2010). However, it is worth noting that the deep magma supply beneath the Central MER is significantly higher than in the northern MER (Cornwell et al., 2010; Gallacher et al., 2016).

6.5. Implications for Continental Breakup

The East African Rift is a type-location for rifting and continental breakup, but it remains the least understood of the major plate boundaries. In mature continental rift, repeated dyking events have decreased the tensile strength of the crust, causing crustal thinning (Buck, 2004; Ebinger, 2005).

Geodetic observations show that up to 80% of the extension in the MER is now accommodated at the Quaternary magmatic segments within the rift (e.g., Bilham et al., 1999; Birhanu et al., 2016). The rifting model of Buck (2004) shows that lithospheric heating caused by magma intrusion localizes thinning and facilitates extension at much smaller extensional forces. Field observations (e.g., Wolfenden et al., 2004), seismic evidence for cooled mafic intrusions in the lower crust (Keranen et al., 2009), and our observation of dike intrusion all indicate that strain accommodation in the MER occurs by magmatic intrusions. Together these observations indicate that magmatic processes dominate in mature rifts and have implications for magmatic margins worldwide.

The dike intrusion documented here provides a snapshot of a shallow magma intrusion and provides new insights into the timescale and geometry of the underlying processes of rifting. The tectonic stresses required to form dike intrusions are lower than those required to activate the large border faults supporting the theory that dike intrusion along the rift axis accommodates significant part of the strain, and border faults progressively become inactive (e.g., Casey et al., 2006; Corti et al., 2018; Keir et al., 2006). Thus, our observations of dike intrusion within the MER support the growing literature that suggests that magma intrusion becomes more dominant than faulting as continental rifting progresses toward passive margin formation. The inference of a rhyolitic composition suggests that silicic magmatic centers play a key role in extensional processes in continental crust.

7. Conclusion

Dike intrusions play a vital role in accommodating strain and have a significant impact on the rheology of the continental crust during rifting. We have used InSAR to investigate the spatial and temporal characteristics of an intrusion of a dike into the upper crust of the northern MER, Ethiopia in 2015. Elastic models of InSAR data show that the dike is ~ 6 ± 1 km long, ~2 ± 0.2 m wide, with a depth range of 5.4 to 8 km from the surface (a volume change of 33 × 10⁶ ± 0.6 × 10⁶ m³) and strikes N29°E ± 2°, oblique to the current extension direction. The dike intrusion was accompanied by a seismic swarm which coincides with the spatial distribution of surface deformation from InSAR. The time series of deformation shows exponential decay with a timescale of ~83 days. The characteristic time of the intrusion is an order of magnitude larger than previous dike intrusions observed in Afar and Iceland (see Rivalta, 2010). This slowness of the dike intrusion can
be attributed to a viscous peralkaline rhyolitic magma intruded laterally from a shallow source. Fentale volcanic center shows different characteristics to other caldera systems in the MER (e.g., Aluto, Corbetti, and Tulu Moye) which exhibit multi-year cycles of inflation and deflation. It raises an open question whether this difference in activity is transient or a long-term feature associated with rift evolution. In the absence of in situ measurement and detailed historical records of volcanic activity, satellite-based observations of surface deformation play an important role in understanding the present state of the subsurface activities.

Data Availability Statement

The interferograms from Sentinel-1 can be found on COMET-LiCS Sentinel-1 InSAR Portal (https://comet.nerc.ac.uk/COMET-LiCS-portal/). The Cosmo-skyMed interferograms can be found on (https://www.bgs.ac.uk/services/ngdc/accessions/index.html?simpleText=131186). Preliminary seismic data supporting this research are available from the Institute of Geophysics, Space Science and Astronomy at Addis Ababa University, subject to data sharing agreements, and are not accessible to the public or research community. For further details please contact this email address (atalay.ayele@aaau.edu.et). We thank the editor and the two anonymous reviewers for their comments that improved the manuscript. Most of the figures are generated using the Generic Mapping Tool (GMT) software.

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