Impact of crustal deformation detection by the DSI (difference of split-band interferograms) method with PALSAR-2 data: A case study on the 2016 Kumamoto Earthquake

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2 **interferograms) method with PALSAR-2 data: A case study on the 2016**
3 **Kumamoto Earthquake**

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11

12 **Abstract**

13 Interferometric Synthetic Aperture Radar (InSAR) is a useful tool for detecting
14 surface deformations at high spatial resolutions. When InSAR is applied to huge surface
15 deformations, clear fringes with complicated phase gaps often appear in the
16 interferograms. Although the surface deformations in such areas are important for
17 understanding their mechanisms and for investigating disasters, it is difficult to convert
18 the data on such fringes to surface deformation information because of difficulties
19 associated with phase unwrapping. To resolve these difficulties, we created multiple
20 SAR pairs with different frequencies using a band-pass filter and derived the difference
21 of interferograms which are generated from these SAR pairs. Generally, its result
22 corresponds to the result of SAR observations made with long-wavelength radar.
23 Therefore, a phase wrap was less likely to occur, and phase unwrapping was easy to
24 accomplish. We applied this method to the PALSAR-2 data pairs for the 2016
25 Kumamoto Earthquake and succeeded in identifying huge crustal deformations with
26 complicated phase gaps in the vicinity of surface ruptures. Comparing these results with
27 the crustal deformations observed from GNSS measurements, the root-mean-squares of

the differences were found to be approximately 4 cm. Although this accuracy was poorer than that of conventional InSAR, it was nearly equivalent to that of the offset-tracking method. Furthermore, its spatial resolution was significantly better than that of the offset-tracking method. However, the disadvantage of this method is that its detection accuracy is significantly degraded in zones with low coherence, due to noise amplification. The standard deviation of the noise component was approximately 2 cm for pixels with coherences above 0.7. However, for pixels with a coherence lower than 0.2, the standard deviation was greater than 10 cm, and the noise component occasionally exceeded 1 m. Despite the disadvantages of this method, it is effective for the detection of huge crustal deformations with high spatial resolution in areas where phase unwrapping methods for conventional InSAR are inappropriate.

Keywords

DSI, difference of split-band interferograms, InSAR, crustal deformation, the 2016 Kumamoto Earthquake

Introduction

Interferometric Synthetic Aperture Radar (InSAR) is a useful tool for detecting the surface deformations associated with earthquakes, volcanic activity, landslides, and so on. In particular, the L-band SAR which is installed on the ALOS-2 (SAR sensor name: PALSAR-2) and upcoming ALOS-4 and NISAR has been utilized as an effective tool for detecting surface deformations in heavily vegetated areas. This is due to its capacity to penetrate thick vegetation and its high stability of coherence. Another advantage of the L-band SAR in InSAR applications is its robust capacity to detect huge deformations with large spatial gradient. Generally, such huge deformation produces a dense fringe, which makes it difficult to solve modulus 2π ambiguities. Considering simple 1-D phase unwrapping as an example, it is impossible to solve its ambiguities in case where the phase gaps of neighboring pixels exceed a half cycle. Because the wavelength of the L-band is longer than that of the C- and X-bands, such a phase gap is unlikely to occur. Thus, it is relatively easy to detect huge deformations using L-band SAR. However, when large earthquake events cause surface ruptures, even if a clear fringe with high coherence is obtained, many phase gaps exceeding half a wavelength

often appear in interferograms. In such cases, the phase information around the surface rupture is not converted to the corresponding amount of crustal deformation. Figure 1 illustrates the interferogram depicting crustal deformation around the surface rupture associated with the 2016 Kumamoto Earthquake (April 16, 2016, $M_{JMA}7.3$) derived from PALSAR-2 data. Clear fringes are also seen near the fault, but their pattern is extremely complicated, and many phase gaps exceeding a half cycle are apparent. It is difficult to correctly unwrap the phases in this region. However, information about the crustal deformations in such areas is important for improving the understanding earthquake occurrence mechanisms, investigating damaged areas, and so on. Therefore, it is meaningful to obtain surface deformation information for such areas. Himematsu and Furuya (2016) applied the offset-tracking method to PALSAR-2 data and obtained information on crustal deformation near the fault. Thus, the offset-tracking method is an effective tool; however, a large correlation window is necessary for high accuracy, and thus, there is a trade-off between detection accuracy and spatial resolution in this method.

Many unwrapping algorithms for InSAR processing have been proposed. For

76 instance, the Brunch-cut method, which is one of the basic algorithms, solves for
77 modulus 2π ambiguities at each pixel while searching for an adequate path that connects
78 phases while considering residues and coherence (Goldstein *et al.*, 1988). The phase
79 connection is processed such that the phase difference between adjacent pixels does not
80 exceed a half cycle. However, as mentioned before, huge crustal deformations often
81 cause phase gaps exceeding a half cycle, so it is difficult to obtain unwrapped phases in
82 areas surrounded by such phase gaps. Although many other advanced phase unwrapping
83 algorithms such as minimum cost flow and statistical cost flow have been proposed
84 (Constantini, 1998; Chen and Zebker, 2000), it remains difficult to unwrap phases in
85 such areas.

86 Madsen and Zebker (1992) proposed a method for deriving the unwrapped phase
87 from the phase differences of interferograms with different frequencies. Such
88 interferograms are derived from SAR pairs created by dividing the chirp band using the
89 band-pass filter, which is called “split-spectrum method.” In the 1990s, there were
90 limited data sets with sufficiently high coherence for the application of the split-
91 spectrum method. Recently, L-band SAR with a wide bandwidth has become available,

and a chance which is obtained interferograms with sufficiently high coherence has drastically increased. Thus, the split-spectrum method has recently seen use as a general technique to estimate ionospheric delay (Bricic *et al.*, 2010; Rosen *et al.*, 2010; Gomba *et al.*, 2016; Furuya *et al.*, 2017; Wegmüller *et al.*, 2018). Therefore, recent L-band SAR data can also be useful for obtaining the unwrapped phase using the split-spectrum method. While this is an application of the split spectrum method, in this paper, we refer to it as the “DSI” (difference of split-band interferograms) to distinguish it from the split-spectrum method used for the estimation of the ionospheric delay.

We applied the DSI method to PALSAR-2 data for the 2016 Kumamoto Earthquake (Fig. 1), and succeeded in detecting huge crustal deformations near the fault. In this paper, we describe the fundamentals of crustal deformation detection using the DSI method with practical processing procedures, and discuss its detection accuracy, advantages, and disadvantages.

The basics of the DSI method

In this section, we describe the fundamentals of the DSI method used to obtain the

unwrapped phase. Generally, the interferometric phase is expressed as

$$\Delta\phi_{\text{obs}} = 4\pi f/c \Delta\rho_{\text{non-disp}} - 4\pi K/cf \Delta\text{TEC} + \text{bias}_1 \quad (1)$$

where $\Delta\rho_{\text{non-disp}}$ is the non-dispersive component of the slant-range change due to crustal deformation, orbit difference, atmospheric delay, etc. The second term on the right-hand side is called the dispersive component, which is mainly due to ionospheric delay. f is the radar frequency, c is the speed of light, ΔTEC is a total electron content along the radar propagation path, K is constant of $40.28 \text{ m}^3/\text{s}^2$, and bias_1 is the bias. First, it is necessary to estimate and remove the dispersive component. From Eq. (1), the response to frequency is different for dispersive and non-dispersive components. The split-spectrum method produces interferograms with different frequencies from SAR images that divide chirp bands, and divide the dispersive and nondispersive components based on the differences in responses (Bricic *et al.*, 2010; Rosen *et al.*, 2010; Gomba *et al.*, 2016; Wegmüller *et al.*, 2018). Next, the non-deformation components in the nondispersive component are removed. The phase component of the orbit difference can be estimated from the satellite ephemeris and a digital terrain model. In Japan,

125 atmospheric delay can be estimated with an accuracy of several centimeters, from the
126 results of the numerical weather model published by the Japan Meteorological Agency
127 (JMA) (Ozawa and Shimizu, 2010; Ozawa *et al.*, 2019). The deformation component
128 can be extracted by subtracting the non-deformation components from the initial
129 interferogram. However, it is a wrapped phase, and as described above, it is difficult to
130 obtain an appropriate unwrapped phase for huge crustal deformations with complicated
131 phase gaps. For example, in InSAR applications using the PALSAR-2 data (radar
132 wavelength: approximately 24 cm), a slant-range change of approximately 12 cm causes
133 a one cycle phase change; a deformation gap exceeding 6 cm makes it difficult to obtain
134 the unwrapped phase. Such huge deformations often appear due to earthquakes,
135 volcanic activities, landslides, glaciers, and so on. Conversely, if a microwave with a
136 long wavelength is used, this problem is unlikely to occur. Then, we used the DSI
137 method to obtain the unwrapped phase. The DSI method analyzes interferograms of the
138 different frequencies generated in the ionospheric delay estimation. Subtracting
139 dispersive and non-deformation components from their interferograms, the phase
140 difference ($\Delta\phi_{\text{diff}}$) is

141

142
$$\Delta\phi_{\text{diff}} = \Delta\phi_{\text{high}} - \Delta\phi_{\text{low}} = 4\pi \Delta\rho_{\text{deform}} (f_{\text{high}} - f_{\text{low}})/c + \textit{bias}_2 \quad (2)$$

143

144 where f_{low} and f_{high} are the center frequencies of SAR images divided into two bands:

145 $\Delta\phi_{\text{high}}$ and $\Delta\phi_{\text{low}}$ are the phase differences of the interferograms of high and low bands,

146 $\Delta\rho_{\text{deform}}$ is the slant-range change due to crustal deformation, and $\textit{bias}_2$ is the bias. This

147 is consistent with the interferometric phase obtained from the SAR pair with a radar

148 frequency of $f_{\text{high}} - f_{\text{low}}$. The wavelength of this interferogram was $c/(f_{\text{high}} - f_{\text{low}})$. A phase

149 wrap does not occur when the slant-range change is less than a quarter of the

150 wavelength, and unwrapping is not necessary. The slant range change occasionally

151 exceeds its threshold, but even in such cases, the need to perform an unwrapping can be

152 eliminated by increasing the number of band divisions. This method is described in the

153 next section, along with an explanation of the actual procedure.

154

155 **Procedure of the DSI analysis: A case study on the 2016**

156 **Kumamoto Earthquake**

157 In this chapter, we describe the procedure for crustal deformation detection using the

DSI method, with a case study on the 2016 Kumamoto Earthquake using PALSAR-2 data. The sequence of the 2016 Kumamoto Earthquake began from a foreshock occurrence ($M_w6.1$) at 21:26 on April 14, 2016 (JST) in the Kumamoto Prefecture of the Kyushu district of western Japan. According to JMA, the depth of the hypocenter was 11.4 km. F-net, the broadband seismometer network of the National Research Institute for Earth Science and Disaster Resilience (NIED) (Okada *et al.*, 2004), derived a lateral slip mechanism with the north-south tension axis at a depth of 17 km. The mainshock ($M_w7.1$) occurred two days later, on April 16 at 1:25 JST. F-net determined the same lateral-slip mechanism as the foreshock, and the depth determined by the JMA was 11 km. In our previous study, we detected the crustal deformation associated with the earthquake by applying conventional InSAR to PALSAR-2 data, and showed that its result could be explained by the fault model, with right-lateral slips on three faults and one normal fault with a low angle; three right-lateral faults were identified along the Hinagu Fault, the Futagawa Fault, and its northeast extension, and the normal fault was parallel to the Futagawa Fault (Ozawa *et al.*, 2016). Dense fringes with complicated phase gaps appeared (Fig. 1(d)) in the vicinity of the Futagawa Fault, where surface

174 ruptures with dislocations exceeding 2 m were found (*e.g.*, Shirahama *et al.*, 2016, Toda
175 *et al.*, 2016). Complicated phase gaps were also observed in the western section of the
176 Aso caldera, where significant stress changes were induced by the ruptures associated
177 with the mainshock. In our previous study, crustal deformation was not obtained in
178 these areas because of difficulties associated with phase unwrapping (Fig. 1(c)). Using
179 the detection of such complicated deformations as an example, we describe the
180 procedure of the DSI analysis.

181 In this analysis, we analyzed the PALSAR-2 data obtained from Path 28 (descending
182 orbit, left-looking) and Path 126 (ascending orbit, left-looking). Most of the scenes were
183 mountainous areas covered by vegetation. The observation mode was SM1 for both
184 paths. Its center frequency is 1.2575 GHz, and the bandwidth of the SLC image is 80
185 MHz. The incidence angles were 32° and 24° at the scene center, respectively. The
186 observation dates were April 15, 2016, and April 29, 2016, for both paths, and
187 therefore, the results from their pairs did not include the crustal deformations associated
188 with the foreshock on April 14, 2016. In this analysis, we used RINC software, which is
189 an InSAR analysis tool developed by NIED (Ozawa *et al.*, 2016). The number of looks

in the interferogram generation were set to 11×15 for Path 28 and 8×12 for Path 126 so that the ground range was from 25 to 30 m. Although processing with a smaller number of looks is possible, we adopt a larger number of looks for reduction of decorrelation noise. The spectral shift filter described by Gatelli *et al.* (1994) was applied to the interferogram generation, and the spectral enhancement filter described by Goldstein and Werner (1998), with a filter coefficient of 0.8 and a window size of 32 pixels, was applied to the generated interferograms. For the simulation of the phase component due to topography, a numerical ellipsoidal height model created from a 10 m mesh digital elevation model published by the Geospatial Information Authority of Japan (GSI) and the EGM96 geoid model (Lemoine *et al.*, 1997) was used.

Next, we estimated the ionospheric delay using the split-spectrum method. The method by Brcic *et al.* (2010), Rosen *et al.* (2010), and Gomba *et al.* (2016) needs to unwrap the phase including the crustal deformation component; therefore, it was difficult to estimate the ionospheric delay near the fault, where unwrapping errors were likely to occur. Thus, we used the method described by Wegmüller *et al.* (2018) to extract the ionospheric delay component without unwrapping. This method creates two

206 SLC image pairs with different frequencies by dividing the chirp band using a band-
207 pass filter. In the case dividing to two bands, the center frequencies of their SLC images
208 were 1.2375 GHz (f_{low}) and 1.2775 GHz (f_{high}), and the bandwidth was 40 MHz. The
209 dispersive component ($\Delta\phi_{\text{disp}}$) was estimated as follows:

210

$$211 \quad 2 \Delta\phi_{\text{disp}} = \Delta\phi_{\text{obs}} - \Delta\phi_{\text{diff}} f_{\text{low}} f_{\text{high}} / f_0 (f_{\text{high}} - f_{\text{low}}) \quad (3).$$

212

213 where f_0 is the center frequency of the full-band SLC image, and $\Delta\phi_{\text{diff}}$ is the phase
214 difference between interferograms for high and low bands, corresponding to the
215 interferometric phase observed by a radar wave with a frequency of $f_{\text{high}} - f_{\text{low}}$. Its
216 frequency is 40 MHz, and the wavelength is approximately 7.5 m in this case study.

217 When this processing is applied to interferograms with the topographic and orbital
218 phase components removed, $\Delta\phi_{\text{diff}}$ usually falls between $-\pi$ and $+\pi$ radians, unless
219 crustal deformation exceeds a quarter of the wavelength which is 1.9 m in this case
220 study. However, crustal deformation in excess of 2 m was found during the 2016
221 Kumamoto Earthquake, and this is an extreme case that phase wraps of $\Delta\phi_{\text{diff}}$ appear.
222 Even in such a case, unwrapping is not necessary by increasing number of divisions.

223 We divided the chirp band into four bands (center frequency: $f_{1\sim 4}$), and connected the
224 phase differences of interferometric phases for (f_4-f_3) , (f_3-f_2) , and (f_2-f_1) . The phase
225 difference obtained for f_4-f_1 corresponded to $f_{\text{high}}-f_{\text{low}}$ in Eq. (3). Its frequency is 60
226 MHz. The frequency of $f_{i+1}-f_i$ was 20 MHz and corresponded to a wavelength of 15 m,
227 and a phase wrap did not appear when the slant-range change was less than
228 approximately 3.75 m; which was a quarter of the wavelength. Even if a larger
229 deformation occurred, it would have been possible to process the data without
230 unwrapping, by increasing the number of divisions further. The ionospheric delay
231 component was obtained by substituting the obtained phase difference ($\Delta\phi_{\text{diff}}$) into Eq.
232 (3). As $\Delta\phi_{\text{disp}}$ is a wrapped phase, it was necessary to apply unwrapping. However, its
233 spatial distribution was smooth, except in extreme conditions, and it was easy to obtain
234 an appropriate unwrapped phase. From its unwrapped phase, the ionospheric delay
235 component was estimated by masking pixels with low coherence and applying a
236 Gaussian filter. Next, we simulated the atmospheric delay from the numerical weather
237 model (Ozawa and Shimizu, 2010, Ozawa *et al.*, 2019) and removed it from the
238 interferograms of the four bands. Applying the same analysis used for the estimation of

239 $\Delta\phi_{\text{diff}}$ to these interferograms, we obtained the unwrapped phase of the crustal
240 deformation component for f_4-f_1 from Eq. (2), without any unwrapping.

241 The slant-range change distributions obtained from the above analysis are shown in
242 Figs. 2(a) and 2(b), and the quasi-east-west (QEW) and quasi-up-down (QUD)
243 components converted using the method proposed by Fujiwara *et al.* (2000) are shown
244 in Fig. 3. Observation times for Path 28 and Path 126 differ by approximately half a
245 day, but it can be assumed that the crustal deformation that occurred during the time
246 difference in the observations is negligible because the maximum magnitude of the
247 earthquake that occurred during their time difference was $M_{\text{JMA}}4.2$. QEW is a direction
248 rotating approximately 3° clockwise from the east-west, and QUD is a direction
249 inclining approximately 8° to the north from the vertical. In general, the results show an
250 eastward displacement in the northern area of the Futagawa Fault and a westward
251 displacement in the southern area, which is consistent with the focal mechanism of the
252 main shock. Its boundary clearly coincides with the Futagawa fault. The eastern edge of
253 the QEW boundary becomes obscure around the western rim of the Aso caldera, and in
254 the western area, the boundary extends to the Hinagu Fault from the Futagawa fault. On

255 the other hand, the clear boundary of the QUD is observed in the western area of the
256 Futagawa Fault, while in the eastern area, the boundary is consistent with the
257 Idenokuchi Fault that is parallel to the Futagawa Fault (Figs. 3(f), (g)). Thus, the QEW
258 and QUD boundaries are different in this area, corresponding to the partitioning of a
259 strike-slip and a normal-slip, as described by Toda *et al.* (2016). Additionally, this study
260 clarified that subsidence gradually increased from the west, and that it exceeded 2 m in
261 the area between the Fugatawa and Idenokuchi faults. The maximum westward
262 displacement was also detected in this area, at 0.3–0.4 m.

263 Crustal deformation west of the Aso caldera, where a complicated fringe pattern
264 appeared (Fig. 1(c)), was also clearly obtained (Figs. 3(d)–(g)). As pointed out by
265 Fujiwara *et al.* (2016), crustal deformation shows that a north-down dislocation
266 dominated in the southern area, and a south-down dislocation dominated in the north
267 area, similar to a graben. In addition, this result shows that the deformation gap was
268 large in the southern area. Clear deformation gaps can be observed at points where
269 surface ruptures have been found by field surveys (Awata *et al.*, 2019; National Institute
270 of Advanced Industrial Science and Technology, 2019), and their extensions can also be

identified in these results (thin lines in Fig. 3). Although the locations of gaps can be also seen from the results of the conventional InSAR analysis, the DSI analysis is advantageous because it can obtain also its amount.

Discussions

As described in the previous chapter, the case study on the application of DSI analysis to the 2016 Kumamoto Earthquake yielded crustal deformation information that was consistent with previous researches. Although new seismological knowledge was not obtained in this case study, it was important to achieve reasonable results using DSI analysis. Since seismological interpretations of crustal deformation have been reported in many studies, we focus on the detection accuracy and limitations of the DSI method in this study.

First, we compared the crustal deformation information obtained by the DSI analysis with those from GEONET, which is the GNSS observation network operated by GSI (Sagiya *et al.*, 2000). Figure 4(a) shows the locations of the GNSS sites that were used in this comparison. Crustal deformations from GNSS were derived from the averaged

287 11-day difference coordinates of F3 solutions (Nakagawa *et al.*, 2009) for 8 Apr. 2016
288 and 29 Apr. 2016. Because crustal deformation from GNSS included the effects of the
289 foreshock on April 14, crustal deformation calculated from the fault model (the two-
290 fault model with uniform slip), by Kobayashi (2017), were subtracted from it. Figure 4
291 shows the comparison between the crustal deformations obtained from the DSI analysis
292 and GEONET (these values are shown in Tables S1 and S2). The reference site for their
293 deformation is the "960699" GEONET site (Open diamond in Fig. 4(a)). With the
294 exception of the "960701" GEONET site, they agree within 10 cm, and the root-mean-
295 squares of differences (rms) are 4.6 cm and 3.8 cm for Path 28 and Path 126,
296 respectively. Comparing the crustal deformations derived from conventional InSAR and
297 those from GEONET in the same way, the rms was 2.5 cm and 2.2 cm for Path 28 and
298 Path 126 (Tables S1 and S2), respectively, indicating that coincidence for the DSI
299 analysis is poorer than that for conventional InSAR. This is considered to arise from the
300 principal limitations of the DSI method. The SM1 mode of PALSAR-2 was observed
301 using a radar wave with a frequency of 1.2575 GHz, while the DSI analysis in this study
302 corresponds to an observation by a radar wave with a frequency of 60 MHz, indicating

303 that the sensitivity of crustal deformation detection is much lower. Phase noise due to
304 decorrelation at almost the same level as conventional InSAR appears on each
305 interferogram for the divided band, but they are not compensated within the
306 interferograms of different frequencies. Decorrelation noise in the DSI result is
307 amplified $f_0/(f_{\text{high}}-f_{\text{low}})$ times over that in conventional InSAR, corresponding to 21
308 times in this case study. Therefore, the accuracy of the DSI method is poorer than that
309 of conventional InSAR. In particular, for "960701" GEONET site, the slant-range
310 change results of the DSI analysis differ by more than 1 m from those from
311 conventional InSAR. This is much larger than the general accuracy of GNSS
312 observations, and therefore, a difference of 1 m is due to the error in the DSI analysis.
313 The coherence for this pixel was 0.04, which is extremely low. In such pixels, the phase
314 difference between dividing bands is not appropriately estimated due to noise
315 amplification, and the errors may have arisen from the misestimation of the phase
316 difference $\Delta\phi_{\text{diff}}$ in Eq. (3).

317 To investigate the relationship between the coherence and accuracy of the DSI
318 method, we analyzed the post-seismic pair of Path 126 (April 29, 2016, and October 14,

2016), where crustal deformation was much lower than that of the co-seismic pair. The observation mode was the same as that for the co-seismic pair. Figures 5(a) and (b) show slant-range change maps obtained using the DSI analysis and conventional InSAR, and Fig. 5(c) shows the coherence map. In the slant range change map from the DSI analysis, the area where high-frequency noise is dominant corresponds to the area with low coherence. In the differences between the slant-range change maps obtained from the DSI analysis and conventional InSAR, the deviations increased with decreasing coherence (Fig. 5(d)). The standard deviation of the pixels with a coherence of 0.7–0.8 or larger was 2 cm or less. It increased with the decreases in coherence, and for the pixels with a coherence of 0.5–0.6, was estimated to be 3.5 cm, corresponding to the rms of the comparison with GEONET. An rms of approximately 10 cm was obtained for a coherence of 0.2, but the difference occasionally exceeded 1 m (Figs. 5(d) and (e)). Noise increased significantly with decreases in coherence; thus, in DSI analysis, SAR pairs with high coherence are necessary for the acquisition of highly accurate of crustal deformation detections, and this method would be effectively applicable to L-band SAR data that can obtain high coherence. Additionally, the noise

due to the insufficient reduction of atmospheric delay was not amplified because it was included as a nondispersive component. Noises due to insufficient reduction of ionospheric delay are also not so different from that on the conventional InSAR.

Converting phase into the slant-range change, the ionospheric delay component is $K/f_{\text{high}}/f_{\text{low}}\Delta\text{TEC}$ for the DSI method and $K/f_0^2\Delta\text{TEC}$ for the conventional InSAR.

Next, the crustal deformation obtained from the DSI method was compared with that from the offset-tracking method. We used the GAMMA SAR processor for the offset-tracking analysis. In the co-registration of SAR images for the application of the offset-tracking analysis, we corrected for the parallax topography effect due to the satellite orbit difference and estimated the affine transformation coefficient by masking the areas with large deformations. In the offset-tracking analysis, we set a correlation window of 64×64 pixels for every 24 pixels and estimated the offset with the maximum correlation.

Because crustal deformation can be estimated with higher spatial resolution by the use of a smaller correlation window, we also attempted to perform an analysis using a correlation window of 32×32 pixels. However, spike noise significantly affected these results, and therefore, the results of the 64×64 correlation window was adopted for this

comparison.

Figures 2(b) and (e) show the slant-range change maps obtained from the offset-tracking analysis, and the differences with the results from the DSI analysis are shown in Figs. 2(c) and (f). Generally, the slant-range change maps obtained by offset-tracking analysis produced results very similar to those from the DSI analysis. Likewise, for the comparison of the DSI and GEONET results, we compared slant-range changes from the offset-tracking analysis with those from GEONET, and the rms of differences were 4.4 cm and 4.2 cm for Path 28 and Path 126 (Tables S1 and S2), respectively. These values almost parallel the coincidence of the DSI analysis. However, granular patterns with a wavelength of approximately several 100 m are superimposed over the entire area on slant-range change maps from the offset-tracking analysis. Such crustal deformation patterns are unnatural; and therefore, must be the result of estimation errors in the offset-tracking analysis. Additionally, comparing the spatial distributions of complicated deformations west of the Aso caldera, those from the DSI analysis are sharper than that those from the offset-tracking analysis. This indicates that the DSI analysis detected crustal deformations at a higher spatial resolution. However, in the

DSI analysis results, large amounts of random noise were obtained in the vicinity of the fault. Additionally, a significantly negative difference from the offset tracking method was also obtained near the fault. As coherence in this area was significantly low, it is possible that crustal deformation was not correctly estimated in the DSI analysis. Thus, it is possible that low coherence causes large misestimations in the slant-range changes derived from DSI analysis.

Summary

In this study, we described the use of the DSI method for detecting huge crustal deformations. In a case study, we applied the DSI method to PALSAR-2 data for the 2016 Kumamoto Earthquake, and detected crustal deformation in the areas where the unwrapped phase could not be obtained by conventional InSAR analysis, due to complicated phase gaps. Compared with conventional InSAR, the detection accuracy of the DSI method is poorer, but it can detect crustal deformation in areas where huge crustal deformation has occurred. In addition, the DSI method can detect crustal deformation with comparable accuracy and higher spatial resolution than the offset-

tracking method. However, the detection accuracy of the DSI method depends on coherence. An accuracy of approximately 2 cm was estimated for pixels with a coherence of more than 0.7–0.8, but an estimation error exceeding 1 m was occasionally obtained for pixels with a coherence of less than 0.2. The disadvantage of DSI analysis is that applications with high accuracy are limited to cases that use high-coherence data pairs. However, such a disadvantage will become insignificant in near future. New L-band SAR missions, PALSAR-3/ALOS-4 and NISAR, will commence in the near future, and observation frequency will be significantly improved from that of the current PALSAR-2/ALOS-2. Since they will stably provide InSAR results with high coherence, it is expected that the DSI method will become an effective tool for detection of huge crustal deformations.

Declarations

Ethics approval and consent to participate

Not applicable.

Consent for publication

400 Not applicable.

401

402 **List of abbreviations**

403 ALOS-2/4: Advanced Land Observing Satellite 2/4

404 DSI: Difference of Split-band Interferograms

405 EGM96: Earth Gravitational Model 1996

406 GEONET: GNSS Earth Observation Network System

407 GNSS: Global Navigation Satellite System

408 GSI: Geospatial Information Authority of Japan

409 InSAR: Interferometric SAR

410 JMA: Japan Meteorological Agency

411 JST: Japan standard time

412 NIED: National Research Institute for Earth Science and Disaster Resilience

413 NISAR: NASA-ISRO SAR

414 PALSAR-2: Phased Array type L-band SAR 2

415 QEW: Quasi-east-west

416 QUD: Quasi-vertical

417 SAR: Synthetic aperture radar

418 TEC: Total Electron Content

419

420 **Availability of data and materials**

ALOS-2/PALSAR-2 level 1.1 data can be purchased from RESTEC
(<https://www.restec.or.jp/en/>) or PASCO (<http://en.alos-pasco.com>).

Competing interests

The authors declare that they have no competing interests.

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Authors' contributions

TO developed the successful procedure of the DSI method and applied it to PALSAR-2 data on the 2016 Kumamoto Earthquake. YH analyzed the same data set by the offset-tracking method. All authors contributed to the preparation of the manuscript and discussion of the results.

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444 by the GSI were used in this study. We used Generic Mapping Tools (Wessel and
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Figure 1. PALSAR-2 interferograms depicting the crustal deformation associated with the 2016 Kumamoto Earthquake. Arrows in the figures indicate the along-track (A) and cross track (C) directions. (a) Interferogram for Path 28 (Descending, Left looking). (b) Interferogram for Path 126 (Ascending, Left looking). The two boxes correspond to the areas of (c) and (d). (c) Interferogram for Path 28, enlarged around the northeast of the Aso caldera. (d) Interferogram for Path 28, enlarged around the Futagawa Fault.

Figure 2. Slant-range change maps obtained from the DSI and the offset-tracking methods and the differences between them. Arrows in the figures indicate the along-track (A) and cross track (C) directions. (a) Slant-range change map for Path 28, using the DSI method. (b) Slant-range change map for Path 28, using the offset-tracking method. (c) Differences between the slant-range change maps, using the DSI and offset-tracking methods for Path 28. (d) Slant-range change map for Path 126, using the DSI method. Dashed box corresponds to the area shown in (g) and (h). (e) Slant-range change map for Path 126, using the offset-tracking method. (f) Difference of slant-range change maps, using the DSI and offset-tracking methods for Path 126. (g) Enlarged slant-range change map, using the DSI method for Path 126. The area corresponds to dashed area in (d). (h) Enlarged slant-range change map by the offset-tracking method for Path 126. The area depicted corresponds to the dashed area in (e).

Figure 3. Two-dimensional crustal deformation calculated from the slant-range changes for Paths 28 and 126. Thin lines depict the surface ruptures associated with the 2016 Kumamoto Earthquake (Awata *et al.*, 2019; National Institute of Advanced Industrial Science and Technology, 2019). (a) Quasi-east-west (QEW) component of crustal deformation. It is a component rotated 3 degrees clockwise from East. The two boxes correspond to the areas of (d) and (f). Shaded areas show the caldera wall. (b) Quasi-up-down (QUD) component of crustal deformation. It is a component that inclines 8 degrees to the north direction from the vertical. The two boxes correspond to the areas in (e) and (g). (c) Topographic relief map. (d) Distribution of the QEW component enlarged about the northeast of the Aso caldera, corresponding to the Fig. 1(c). (e) Distribution of the QUD component, enlarged in the northeast of the Aso caldera, corresponding to Fig. 1(c). (f) Distribution of the QEW component, enlarged around the Futagawa Fault, corresponding to Fig. 1(d). (g) Distribution of the QUD component, enlarged around the Futagawa Fault, corresponding to Fig. 1(d).

Figure 4. (a) Distribution of the GEONET stations used in the comparison of slant-range changes. The background depicts the slant-range change maps for Path 126, using the DSI method. Open diamond indicates GEONET station “960699”, the reference point of crustal deformation. (b) Comparison of slant-range changes obtained from the DSI method and GEONET. Triangles and circles indicate the slant-range changes for Path 28 and Path 126, respectively.

Figure 5. (a) The slant-range change map about the post-seismic pair of Path 126, obtained using the DSI method. (b) The slant-range change map from conventional InSAR. (c) Coherence map. (d) Scatter plot of difference of slant-range changes obtained using the DSI (y-axis) versus coherences (x-axis). (e) The diagram depicts the relationship between the coherence and standard deviation of the differences in the slant-range changes.

619 Table S1. Comparison of the slant-range changes from GNSS, DSI, InSAR, and offset-

620 tracking for Path 28.

Site	GNSS [m]	DSI [m]	InSAR [m]	Off-track [m]	Coherence
021070	-0.129	-0.182	-0.144	-0.224	0.38
021071	0.288	0.246	0.254	0.307	0.40
950465	0.460	0.430	0.434	0.488	0.50
950466	-	-	-	-	-
960700	-0.037	-0.052	-0.056	-0.016	0.41
960701	-0.513	-0.442	-0.488	-0.555	0.11
960703	-	-	-	-	-
960704	-	-	-	-	-
960699	Reference site				0.45
rms*		0.046	0.025	0.044	

621

622 * The root-mean-square of differences in slant-range changes from GNSS.

623

624 Table S2. Comparison of the slant-range changes from GNSS, DSI, InSAR, and offset-
625 tracking for Path 126.

Site	GNSS [m]	DSI [m]	InSAR [m]	Off-track [m]	Coherence
021070	-0.067	-0.013	-0.071	-0.029	0.31
021071	0.083	0.107	0.041	0.073	0.36
950465	-0.120	-0.185	-0.154	-0.133	0.56
950466	0.057	0.038	0.044	0.003	0.44
960700	-0.005	-0.040	-0.017	-0.038	0.47
960701	0.137	-1.294	0.058	-0.019	0.04
960703	0.166	0.149	0.151	0.105	0.53
960704	0.125	0.102	0.121	0.062	0.63
960699	Reference site				0.55
rms*		0.038	0.022	0.042	

626
627 * The root-mean-square of the differences in slant-range changes from GNSS without
628 “960701”.

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