Geomorphic imprints of repeated tsunami waves in a coastal valley in northeastern Japan

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ABSTRACT

Devastating tsunami waves induced by The 2011 off the Pacific coast of Tohoku Earthquake severely inundated valleys along the ria-type (notched rocky) coast in northeastern Japan, causing erosion of bedrock on valley-side slopes. Using high-resolution topographic data obtained by terrestrial and airborne laser scanning, we investigated detailed morphology of the eroded valley in a typical site at Aneyoshi. We found several characteristic features likely affected by tsunami wave erosion: 1) a series of small cliffs on the valley-side slopes related to the tsunami flow depth, 2) an inverse asymmetry of side slopes at ingrown meander bends, 3) an exceptionally wide valley bottom near the coast, and 4) a knickpoint located around the maximum tsunami run-up point far back from the coast. These topographic features are ascribable to repeated tsunami wave erosion since the mid Holocene.

Keywords: tsunami, bedrock erosion, terrestrial laser scanning, cliff, meander, knickpoint

1. Introduction

The 2011 off the Pacific coast of Tohoku Earthquake (hereafter Tohoku-oki earthquake) (Mw 9.0) on March 11, 2011 caused extreme tsunami waves, which inundated extensive coastal areas of northeastern Japan. Patterns in the tsunami wave invasion are different between lowland plains and narrow valleys. The tsunami waves spread over several kilometers from coastlines in coastal plains leaving a certain amount of sediment deposition (Goto et al., 2012; Richmond et al., 2012), while the waves ran up tens of meters in height to cause erosion rather than deposition in narrow valleys along the ria-type, notched rocky Sanriku Coast (Goto et al., 2011; Mori et al., 2011, 2012; Komatsu et al., 2014). Such different responses between coastal lowlands and rocky coasts have also been observed elsewhere (Dawson, 1994; Umitsu et al., 2007; MacInnes et al., 2009). Even bedrock can be, if not uncontroversial, eroded by a single tsunami, if the wave height and velocity are large enough (e.g., Young and Bryant, 1992). However, due to the limited number of reports or ambiguous evidence on bedrock erosion by tsunami along rocky coasts, the impact of tsunami erosion in coastal morphology remains unclear in many places around the world where tsunami can be effective (e.g., Courtney et al., 2012).

Valleys facing the Pacific Ocean along the rocky coast in northeastern Japan are suitable for investigating morphological signatures due to tsunami wave erosion after the

Tohoku-oki earthquake (Komatsu et al., 2014). Hayakawa et al. (in press) performed topographic measurement of several slopes near the coasts of Sanriku and found small cliffs in bedrock near the height of maximum tsunami surface. Among them, a valley at Aneyoshi, facing the Pacific Ocean near Miyako City, is a typical site affected by the 2011 tsunami wave erosion (Fig. 1), where not only its valley bottom but also valley-side slopes were eroded. We performed detailed morphological surveys in the valley to examine topographic features affected by tsunami wave erosion.

Fig. 1

2. Study area

The southern Sanriku Coast is ria type rocky coast, whose lithology is generally composed of dacitic pyroclastic rocks with thin layers of mudstone of early Cretaceous age (Yoshida and Katada, 1984). The local lithology of the study site of Aneyoshi is homogeneous, composed of the dacitic pyroclastic rocks with complex orientations of beddings and numerous joints. Although decadal-scale subsiding movement has been measured with a mean rate of 5–10 mm/yr (Kato, 1983), and the 2011 earthquake caused about 0.5 m of subsidence around the study site (Geospatial Information Authority of Japan,

2011), marine terraces of Marine isotope stage (MIS) 5e along the nearby coast at 18–29 m a.s.l. indicate ca. 0.2 mm/yr of a long-term uplift rate (Koike and Machida, 2001).

In the site of Aneyoshi, two major rivers drain into the same coast: one forms the east-facing valley, and the other one comes from the north (Fig. 1B). Analyses were performed in the east-facing main valley (MV) because it was directly affected by the invasion of tsunami waves with a run-up height of ca. 40 m. The other northern valley (NV) has fewer erosional features and lower maximum tsunami height (ca. 24 m) (Haraguchi and Iwamatsu, 2011). The MV was used as a camping site until the 2011 tsunami, and human modification has been generally less in the entire valley except for the roads and concreted small river banks along the valley bottom.

Valley-side slopes under the tsunami flow depth exhibit bare bedrock due to the removal of surface materials such as soils and vegetation by the tsunami (Fig. 2A). Some thin patches of soil and fallen trees remain in the bedrock areas along the valley-side slopes. The surface water of the river draining into the MV seeps in the ground in the downstream part of MV (ca. 500 m from the coastline) because of thick gravel deposition on the valley floor and low amount of discharge. Bedrock is hardly observed in the riverbed. The valley exhibits a meander form (Figs. 1C, 2C), and the reach downstream of the major meander (within ca. 500 m from the coast) has a wider valley bottom (ca. 30–100 m), while the upstream reach has a narrower valley bottom (<20 m).

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Fig. 2

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A prominent knickpoint is formed at ca. 150 m upstream of the wide/narrow

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transition point (Fig. 2D). Although the riverbed at the present knickpoint is covered with

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artificial concrete and its bedrock is only partly exposed, the location of the knickpoint is not

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related to any specific anomalies in lithology, such as a particularly resistant rock layer

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judging from the bedrock outcrops in the surrounding slopes. No such prominent knickpoint

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is found in the downstream reach of MV nor in NV.

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The offshore wave height of the tsunami around the coast was about 20 m, and due to

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the narrowing of the valley width, the maximum run-up height of the tsunami was ca. 40 m in

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the inland part of the Aneyoshi valley (Konno et al., 2011; Tsuji et al., 2011; Shimozone et

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al., 2012). In the past, similarly high run-ups at this site have also been recorded historically.

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At the time of the Meiji-Sanriku earthquake in June 1896 (Mw 8.2–8.5), the associated

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tsunami wave came into the valley to run up to ca. 20–35 m, which is similar to that observed

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for the 2011 earthquake (Iki, 1896; Yamana, 1896). A tsunami run-up height of 20.8 m was

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also recorded for the Showa-Sanriku earthquake in March 1933 (Mw 8.4) (Otuka, 1934).

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111 **3. Methods**

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113 To examine detailed morphological features, we employed high-resolution
114 topographic datasets obtained from terrestrial and airborne laser scanning after the
115 earthquake. Although similar high-resolution terrain data are not available from before the
116 earthquake, the analysis on the topographic data enabled identification of characteristic
117 features likely affected by tsunamis in the study site. Here we describe the details of data
118 acquisition and analysis, particularly focusing on steep segments in the valley-side slopes and
119 knickpoint recession rate in the mainstream of the valley.

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121 *3.1. Terrestrial and airborne laser scanning data*

122 Terrestrial laser scanning was carried out in the field in November 2011 to obtain
123 high-resolution topographic data of tsunami-affected valley-side slopes. We used a mid-range
124 scanner of GLS-1500 by Topcon Inc. Distance and angle accuracies of the laser range
125 measurement are 4 mm at 150 m and 6", respectively. The obtained point cloud data with
126 XYZ coordinates, intensity of reflected laser pulse and RGB color information are managed
127 in Topcon ScanMaster and ESRI ArcGIS software.

128 To capture the entire shape of the tsunami-affected side slopes in the valley, point
129 clouds from 13 scan positions were obtained. The point clouds were registered using
130 reference targets by distance resection and tie-pint methods. Geographic coordinates of

reference points in UTM Zone 54N obtained by a carrier-phase GNSS receiver (Nikon-Trimble GeoXH Explorer 6000) and the base-point records by GEONET (GNSS Earth Observation Network System operated by Geospatial Information Authority of Japan) were used for georeferencing the merged point cloud with horizontal and vertical accuracies of less than 10 cm.

As an initial filtering of the raw point cloud data (7,342,548 points), artificial objects such as electric poles that were constructed after the tsunami, and vegetation cover including fallen trees were manually identified and removed from the point cloud. The filtered data (6,507,913 points; Fig. 3A) were then converted into a gridded digital elevation model (DEM) using a simple interpolation method of local minimum filtering, which gives minimum value in a certain search circle for each grid cell, showing bare land surface beneath objects (DTM). The resolution of the TLS-DTM was determined to be 10 cm based on the filtered point cloud density (75.4 points/m²) and location accuracy (< 10 cm).

Fig. 3

The DEM after filtering trees and artificial structures is referred to as a TLS-derived digital terrain model (TLS-DTM). A filtered, 5-m resolution DEM derived from airborne laser scanning (ALS-DTM) has been provided by the Geospatial Information Authority of

Japan. The TLS-DTM covers the tsunami-affected valley-side slopes and the bottom of the main valley (Fig. 3B), while the ALS-DTM covers the whole catchment (Fig. 1B).

Using the color-mapped TLS point cloud, the upper limit of the tsunami wave inundation is traced as the boundary of vegetated and non-vegetated (exposed bedrock) domains in the valley-side slopes (Fig. 3A), being supported by airphoto images and the field observation data of Haraguchi and Iwamatsu (2011) (Fig. 1C). To look into small topographic features in the valley, transverse cross profiles are extracted from the TLS-TDM along the main valley at a spacing of 25 m (Fig. 3B). Also, a longitudinal profile along the valley was obtained from the ALS-DTM because it sufficiently covers the main valley to represent morphological features such as a knickpoint.

3.2. Identification of relatively steep portions by slope gradient analysis

Steep sections, including small cliffs, were quantitatively extracted from the cross profiles by an analysis of scalable gradients used to extract locally steepened reaches (Hayakawa and Oguchi, 2009; Hayakawa et al., in press). The method utilizes scale-dependent changes in topographic gradient to find relatively steep portions compared to an averaged gradient over longer measurement scales as described below.

In order to extract relatively steep portions along section lines, we use the method of topographic slope gradient analysis proposed by Hayakawa and Oguchi (2006, 2009),

which utilizes various lengths for measuring slope gradients. The method was originally applied to extract knickpoints or knickzones (relatively steep reaches) along stream courses, and it can also be applied to other cross section lines because the basic dataset for this analysis is elevation values along a section line sampled at a fixed interval. For each sampling point, scale-dependent slope gradient (G_d) is computed with various base lengths (d) as:

$$G_d = |e_1 - e_2| \cdot d^{-1} \quad (1)$$

where e_1 and e_2 are elevation values at $d/2$ upstream and downstream, respectively. A shorter d gives more local variations in gradient, while a longer d gives a trend of changes in gradient along the line. G_d with a certain range of d represents transition from local to trend gradient. When the range of d is set to cover various lengths from local to trend, sampling interval does not significantly affect the result if it is short enough compared to d . Then, within the range of d , the decrease rate of G_d with increasing d is obtained by linear regression for G_d and d as:

$$G_d = -R_d \cdot d + c \quad (2)$$

where c is an intercept of the regression line, and R_d is defined as the relative steepness, showing how much the point location is steeper than the adjacent points. By setting a threshold for R_d based on the statistical criteria (mean plus one standard deviation), relatively steep reaches can be extracted.

For the cross profiles obtained by the 0.1-m resolution TLS-DTM, sampling points with a regular interval (0.2 m) is set, and R_d is computed for the range of $d = 1.2 - 10$ m to extract meter-scale cliffs. Vertical positions of small cliffs (elevation at their middle point) in the cross profiles are then examined with regard to tsunami flow depth.

In addition, the method is also applied to the longitudinal profile to highlight the location of knickpoints. The longitudinal profile is obtained by the 5-m resolution ALS-DTM with a 7 m sampling interval, and R_d is computed for the range of $d = 12 - 300$ m to extract knickpoints having lengths of more than several meters.

3.3 Empirical equation to estimate knickpoint recession rates

Assuming that the prominent knickpoint is receding due to fluvial erosion (Ye et al., 2013), we estimate the recession rate of the knickpoint using an empirical equation with relevant physical factors including a stream discharge proxy, size of knickpoint and bedrock strength, which are obtained from the ALS-DTM, TLS point cloud and field measurements (Hayakawa and Matsukura, 2003).

Rates of knickpoint recession, E [LT^{-1}], can be estimated by using an empirical law with relevant environmental factors (Hayakawa and Matsukura, 2003). Factors related to the erosion of a knickpoint can be summarized into a non-dimensional factor FR as:

$$FR = \frac{AP}{WH} \sqrt{\frac{\rho}{S_c}} \quad (3)$$

where $A [L^2]$ is catchment area above a knickpoint (measured with the ALS-DTM), $P [LT^{-1}]$ is mean annual precipitation in the catchment (assessed by the observation data of the nearest weather station at Miyako, $39^{\circ}38.8'N$, $141^{\circ}57.9'E$), $W [L]$ is the width of the knickpoint (measured in the TLS point cloud), $H [L]$ is the height of the knickpoint (measured in longitudinal profile derived from the ALS-DTM), $\rho [ML^{-3}]$ is water density (constant, 1), and $S_c [ML^{-1}T^{-2}]$ is unconfined compressive strength of bedrock composing knickpoint and surrounding riverbeds, estimated by an N-type Schmidt hammer rebound value measured in the field using the repeated impact method (Matsukura and Aoki, 2004; Viles et al., 2011). An empirical relationship derived from the knickpoint recession rates enables us to estimate the recession rate of a knickpoint E as:

$$E = a \cdot FR^b \quad (a = 99.7, b = 0.73) \quad (4)$$

The coefficient values were originally derived from a local study area in eastern Japan (Hayakawa and Matsukura, 2003), and the applicability of this equation has been validated for knickpoints across the world (e.g., Hayakawa and Matsukura, 2009).

4. Results

The cross profiles derived from the TLS-DTM frequently show stepped morphology in the valley-side slopes, which comprise small cliffs with a typical height of 0.8–3.8 m (Fig.

4A). Some cliffs are present at river banks located in the middle of the valley bottom (Fig. 4A), while others are more frequently found just below and slightly above the heights of tsunami flow, typically -8 m to +6 m from the flow depth (Fig. 4B).

Fig. 4

As noted, the valley bottom is relatively narrow (<20 m) on the upstream side, and the valley suddenly widens (>30 m) at ca. 500 m from the coast, just upstream of a major meander bend (Figs. 1C, 2C). The degraded outside slope of the meander bend (right bank) seems responsible for the formation of the widened valley, because only the outside slope is offset from the valley center. This type of valley asymmetry at a meander bend can be regarded as “inverse”, because fluvial erosion at meander bends generally leads to steeper outside slopes and gentler inner slopes. Although the weathered surfaces of only a few bedrock outcrops can be observed in the degraded outside slopes, the limited exposures of local rocks look similar to the surrounding outcrops without significant evidence of local weakness.

Along the longitudinal profile of the river draining MV, the prominent knickpoint observed in the field was successfully highlighted (Fig. 5A), and this corresponds with a clear break in the slope–area relation (Fig. 5B). The other knickpoints upstream are likely affected

by human modifications because the reach is located at the present village of Aneyoshi. The locality of the prominent knickpoint is, as noted, unrelated to lithology with absence of specifically hard rock layer. The upstream reach above the knickpoint follows a general trend of the slope–area relationship ($S = 0.03A^{-0.4}$), while the downstream reach shows a different trend. This indicates that different stages or formative processes have affected the upstream and downstream reaches. Also, the ca. 150-m long inner gorge immediately below the knickpoint suggests recent propagation of the knickpoint.

Fig. 5

Equation (4) with the measured parameters (catchment area of 1.0 km^2 , mean annual precipitation of $1,300 \text{ mm/y}$, knickpoint width and height of 4 m and 10 m , respectively, and bedrock strength of 111.2 MPa estimated from the mean Schmidt hammer rebound value of 53.8%) gives an estimate of the knickpoint recession rate to be 0.0095 m/y . This value is relatively low, but with respect to the range of knickpoint recession rates in Japanese hills and mountains (Hayakawa and Matsukura, 2003; Hayakawa, 2005; Hayakawa et al., 2008).

5. Discussion

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263 In the last one hundred years or so, at least two prominent tsunami events in 1896 and
264 1933 severely invaded the Aneyoshi valley prior to the 2011 event. Also, a tsunami event
265 similar to that in 2011 likely occurred in 869 (Jogan Tsunami) (Sugawara et al., 2013). The
266 subduction-zone earthquakes related to plate tectonics around the Japan Trench account for
267 the repeated tsunamis, and such seismic activities seem to have occurred through the Pliocene
268 and Quaternary (Regalla et al., 2010; Okada and Ikeda, 2012). A considerably lower sea level
269 during the Last Glacial period rose to reach its maximum stand in the mid Holocene around
270 6–7 ka, and has been relatively stable since then, with the present level being only 2–3 m
271 lower (Ota et al., 1990). Because the long-term uplift rate of 0.2 mm/yr in the region (Koike
272 and Machida, 2001) gives only ca. 1.4 m of uplift in 7,000 years, the relative sea level has
273 been almost stable since ca. 7 ka. Therefore, the inundations of past tsunamis into the valleys
274 at Aneyoshi seem to have occurred at a scale similar to that in 2011. If such tsunami
275 invasions have occurred every 50 years or so, as suggested from the three major recent
276 tsunami events (1896, 1933 and 2011), the number of tsunami invasions since 7 ka would be
277 roughly 140.

278 Although the actual amount of erosion by the single tsunami event in the valley-side
279 slopes is difficult to quantify even for the 2011 case due to the lack of pre-event data, our
280 field observations suggest that erosion of exposed bedrock surfaces actually occurred at least

281 on the order of millimeters to decimeters (Hayakawa et al., in press). In fact, soils and
282 vegetation on the valley-side slopes, which had been present before the tsunami inundation,
283 as shown by old photographs of the camping site, were completely removed beneath the
284 maximum flow surface of the tsunami. This indicates that the tsunami flow was strong
285 enough to readily remove such unconsolidated materials, and such strong power of the
286 tsunami could also contribute to bedrock erosion beneath the soil. Furthermore, patchy
287 exposures of relatively fresh rock surfaces having yellow-ish colors, which are different from
288 weathered or mossy greyish colors, suggest the occurrences of erosion on bedrock surfaces.
289 The amount of possible erosion by the single tsunami event is small, but if similar events
290 occur more than one hundred times, the cumulative result may be erosion on the scale of tens
291 of meters. The characteristic erosional features found in the Aneyoshi valley could therefore
292 be formed by such repeated tsunami waves in the late Holocene.

293 The small cliffs found in the valley-side slopes are probably one of the characteristic
294 features of tsunami erosion (Fig. 4B). The concentration of small cliffs at the height of
295 tsunami flow is not likely due to lithological factors such as large joints nor weak interbedded
296 layers. Such features are not observed in the homogeneous pyroclastic rocks. Ocean storm
297 waves over the maximum (2–3 m) high stand of the sea level at 7 ka cannot be the cause of
298 their formation (mostly (83%) located >10 m a.s.l.). Also, flood erosion by the river draining
299 the MV is also improbable because of its small catchment area. It is plausible that the small

300 cliffs are affected by wave erosion from the tsunami because they tend to be located around
301 the 2011 tsunami flow surface. Tsunami wave pressure is high near the water surface
302 (Sunamura, 1975). Although the situation is different from erosion by ocean storm waves,
303 attacks by multiple tsunami waves caused by the 2011 earthquake could have higher impact
304 pressure at the run-up surface, as well as some high shear stress by run-up flow and backwash
305 accounting for the bedrock erosion beneath the maximum tsunami flow lines. Even if a single
306 tsunami event is insufficient to form all such cliffs, repeated attacks of tsunamis since 7 ka
307 may account for the numerous occurrences of the cliffs. Furthermore, the variation in the
308 elevation of the cliffs, some of which are positioned above the flow depth of the 2011
309 tsunami (Fig. 4B), can be explained by variations in the flow depth and run-up heights of
310 prehistoric tsunamis (Hayakawa et al., in press).

311 The inverse asymmetry observed at the meander bends and the associated very wide
312 valley bottoms could also be related to tsunami erosion. Incoming run-up and backwash of
313 tsunami waves could swirl at the incised meander bends, efficiently eroding the originally
314 steep outer-side slopes to degrade them. Degraded, less steep slopes, which resulted in
315 extraordinarily wide valley bottoms, may have been caused by a combination of slope
316 instability due to vegetation stripping and bedrock erosion by tsunami waves with water
317 velocities possibly exceeding 20 m/s (Komatsu et al., 2014). Komatsu et al. (2014) also

suggest that the effects of tsunami erosion on the slopes are limited to 500 m from the coast, where the valley direction considerably changes from east-facing to south-facing (Fig. 1C).

If the knickpoint was formed by coastal erosion during the sea-level highstand, the initiation of knickpoint recession would have been ca. 7 ka. However, given the estimated knickpoint recession rate, only 70 m of recession could have occurred in this time period. This distance is considerably less than the actual distance of ca. 500 m, from the probable former coastline to the modern knickpoint. Therefore, the significant tsunami erosion in the lower valley and much slower fluvial erosion in the upper valley could have led to the prominent knickpoint at their boundary.

5. Concluding remarks

To summarize, the coastal valley morphology at Aneyoshi appears to be affected by the 2011 and past tsunami wave erosion. The characteristic features, including small cliffs on inland valley-side slopes, inverse asymmetry of the side slopes at an incised meander bend, an exceptionally wide valley bottom of the downstream reach, and an over-receded knickpoint compared to its fluvial power of erosion, are not related to local lithological controls, and unlikely to be formed by fluvial and hillslope processes alone; whereas, impacts of repeated tsunami erosion since the early Holocene may well explain them.

Similar characteristics are also found at other sites along the southern Sanriku Coast such as step-like features comparable to the small cliffs at Aneyoshi, and shallow landslides induced by the tsunami that may correspond to erosion and widening of valleys at Aneyoshi (Hayakawa et al., in press). Such landforms may also be found in other areas subject to past large tsunamis, for example the coastal areas along the Nankai Trough in southwestern Japan. Geomorphological investigation in such places would potentially be effective to quickly or roughly estimate the possible magnitude of typical tsunamis from topography without depending on tsunami deposits, which are often unavailable in rocky coasts.

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FIGURE CAPTIONS

Figure 1. Plan view of the study area of Aneyoshi. (A) Overview of the eastern (Pacific Ocean) side of Honshu (Tohoku). (B) Overview of the main and north valley catchments at Aneyoshi. The background hillshade is derived from the 5-m ALS-DTM. (C) Downstream reach of the main valley. The aerial photographs are provided by Geographical Information Authority of Japan (GSI), taken just after the earthquake (March-April 2011). Positions and directions of photographs in Fig. 2, as well as approximate extents of narrow and wide portions of the valley are shown.

Figure 2. Photographs in the main valley of Aneyoshi. (A) Entire view of the downstream reach in the valley (modified after Hayakawa et al., in press). (B) View from the west to east at a meander bend. White dashed lines indicate tsunami flow depth. (C) Prominent meander bend at the upstream part of the widened reach, showing gentler outside and steeper inside slopes. (D) Downstream end of the knickpoint located near the maximum tsunami run-up point. Scratches on the cemented side walls near the knickpoint indicate effects of tsunami erosion.

Figure 3. (A) Perspective view of color-mapped point cloud obtained by TLS. White lines along the valley-side slopes indicate the traces of upper limit of bedrock exposures indicating

the tsunami erosion. (B) Plan view of the main valley (modified after Hayakawa et al., in press). The map extent is the same as Fig. 1C. Background is the ALS-DTM-derived hillshade and 1-m contour lines, superimposed by the TLS-DTM hillshade. Nine scan positions (SCN01–09) and GNSS benchmarks with ground elevation values are shown. White lines indicate upper and lower limits of the exposed bedrocks along the valley-side slopes, and the upper line indicates the elevation of the tsunami flow surface.

Figure 4. (A) Cross profiles of the main valley of Aneyoshi. Location of the sections is shown in Fig. 1c. Relatively steep portions (small cliffs) objectively extracted are highlighted with black solid lines. The dashed thicker gray line indicates the upper limit of the tsunami inundation, whereas the thinner gray line indicates the boundary between valley-side slopes and the valley bottom, defined as the breakline between the upper steep and lower flat zones. (B) Frequency distribution of relative elevation of cliffs to tsunami flow depth on the valley-side slopes. The tsunami depth (relative elevation = 0 m) corresponds to the thick gray lines in Fig. 4A. (modified after Hayakawa et al., in press)

Figure 5. Longitudinal view of the main valley of Aneyoshi. (A) Longitudinal profile of the downstream reach of the main valley. Locations of objectively extracted knickpoints are

512 highlighted as black solid lines. The longest and tallest one corresponds to the prominent
513 knickpoint observed in the field. The vertical black lines below the knickpoint indicate
514 location of small cliffs found in the cross profiles (Fig. 4). The tsunami flow depth of the
515 right bank is shown as the gray line. (B) Slope–area relationship along the entire main valley.
516 The dashed gray line indicates the location of knickpoint, which separates the regression lines
517 for upstream and downstream reaches.

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fig 1

fig 2

fig 3

fig 4

fig 5