Estimation of the Slip Rate Along the Unruptured Fault Segment of the M7.2 1896 Rikuu Earthquake, Northeast Japan

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Abstract The Ou Backbone Range (OBR) is one of the most seismically active intraplate regions in northeast Japan. Over the past 200 years, it has been the locus of five large magnitude earthquakes, including the M7.2 1896 Rikuu earthquake, which produced coseismic surface deformation along the northern segment of the Eastern Margin Fault Zone of Yokote Basin (EFZYB), while the southern segment remained unruptured. Despite the extensive paleoseismic investigations along the ruptured segment, the slip rate estimates, and the recurrence interval remained largely unresolved. In this study, advances have been made in the field of estimating the slip rate and recurrence interval for the unruptured segment. This study reports the long-term slip rate, and recurrence interval along the Higashi Chokai san Fault and the signatures of active faulting along the Kanazawa and Omoriyama Faults. Based on the seismic reflection and borehole survey, we have estimated a vertical displacement of 29.3–36.1 m across the Higashi Chokai san Fault, a long-term slip rate of 1.6–1.9 mm/yr over the past 35 ka, and a recurrence interval of about 3,800–4,600 years.

Combining our results with the already published data along the OBR, revealed two seismic gaps, first along the western margin of the OBR, along the partial rupture zone of the 1970 Akita-ken earthquake, and second, along the eastern margin of the OBR, north of the 2008 Iwate-Miyagi Nairiku earthquake (IMEQ) rupture zone. Given most of the slip is released elastically, we propose that these seismic gaps have a potential to produce earthquakes with magnitudes Mw 7.1 and Mw 7.3, respectively if they were to rupture in a single event.

1. Introduction

The northeast Japan arc is one of the most seismically active plate boundary systems in the world recording a convergence rate of ~9 cm/yr between the Pacific and the Eurasian plates (Seno et al., 1996) (Figure 1a). The plate boundary has hosted several large magnitude earthquakes, of which the 2011 Tohoku mega-subduction earthquake (Mw 9.0–9.1), claimed 15,641 lives and caused around 300-billion-dollar worth of damage, making it one of the five deadliest earthquakes of the modern era (Ammon et al., 2011; Kagan & Jackson, 2013; Mori et al., 2011). Apart from the plate boundary earthquakes, several intraplate earthquakes for example, the 1896 Rikuu earthquake (M7.2) (Matsuda et al., 1980), the 1995 Northern Miyagi earthquake (M7.0), the 1970 Akita-ken Nantobu earthquake (M6.2) (Hasegawa et al., 1974), the 1998 Iwate volcano earthquake (M6.1) (Miura et al., 2000), and the 2008 Iwate-Miyagi Nairiku earthquake (M7.2) (Matsuura & Kase, 2010) have occurred on the faults in the over-riding plate in the Tohoku region (Figure 1a). The 1896 Rikuu earthquake is one the largest on-land reverse fault earthquakes in the history of Japan (Matsuda et al., 1980), which ruptured beneath the Ou Backbone Range (OBR) and reactivated the preexisting northern segment of the Eastern Margin Fault Zone of Yokote Basin (EFZYB), while leaving the southern segment unruptured (Kagohara et al., 2009; Matsuda et al., 1980; Miura et al., 2002) (Figures 1a and 1b). Since then, extensive work has been conducted along the ruptured segment of the EFZYB to document the fault slip rate, the timing of the penultimate and the earlier events, and establishing the recurrence interval for large earthquakes (Kagohara et al., 2009; Matsuda et al., 1980; Miura et al., 2002; Research Group for Senya Fault, 1986; Imaizumi et al., 1997; Sato et al., 2002). Along the southern unruptured segment of EFZYB, some of the pioneering work was conducted by the Active Faults Research Group of Japan (1991), Sawa et al., (2011, 2013), Sugito (2013) to document the probable trace of the active fault, however, the estimation of slip rates and recurrence interval remained largely unresolved. In this study, advances have been made through lidar data, seismic reflection, and borehole survey to calculate the fault slip rate and recurrence...
interval of the south unruptured fault segment of the EFZYB, located along the western margin of the OBR in northeast Japan (Figure 1a).

For decades, northeast Japan has been the locus of several shallow-focus earthquakes, with its distribution concentrated mainly along two prominent regions (Hasegawa et al., 1994; Miura et al., 2000). The first is beneath the Pacific Ocean along the plate boundary subduction zone, while the other is within the intraplate
region along the OBR (Hasegawa et al., 2005). Over the last 200 years, OBR has witnessed five large magnitude earthquakes, with their focal depths ranging between 10 and 15 km (Miura et al., 2004). These shallow focus depths are attributed to the shallow depth of the brittle-ductile transition zone, and steep geothermal gradient beneath the OBR (Hasegawa, et al., 2000). Among the five large magnitude earthquakes, few have completely ruptured the seismogenic thickness, while others did not reach the surface. This complex earthquake rupture behaviors’ of the recent earthquakes along the OBR makes it important to understand the rupture pattern and directivity to assess the future implications of the seismic hazard in the region.

In this study, we document the evidence of active faulting at three sites along the unruptured segment of the EFZYB, (a) along the Kanazawa Fault, (b) along the Omoriyama Fault, and (c) the Higashi Chokai san Fault (Active Fault Research Group of Japan, 1991; AIST, 2010; Awata, 1999; Kagohara et al., 2009; Sawa et al., 2013; Sugito, 2013; Taniguchi et al., 2007) and estimate the slip rate and recurrence interval along the Higashi Chokai san Fault (Figures 2a and 2b). Toward this goal, we have extensively used aerial photo-in-
terpretation, digital elevation model (DEM) obtained from the lidar data, and field investigations to identify the tectonic geomorphic features associated with the recent faulting. We have assessed the subsurface geometry of the Higashi Chokai san Fault through the seismic reflection survey, and measured the vertical displacement through borehole drilling investigations. Further, we have combined our results with the already published data from the past earthquake ruptures, to discuss the tectonic implications, slip potential, earthquake rupture directivity, and the associated seismic gaps along the OBR.

2. Regional Setting

2.1. Tectonic and Geologic Settings

The Tohoku region forms a part of an island arc, bordering to its west by the Japan Sea and to its east by the Japan trench (Seno et al., 1996; Yoshii, 1979) (Figure 1). Along the trench, the Pacific Plate is subducting beneath the northern Honshu, and this ongoing subduction is expressed in the form of intraplate compression, which is accommodated by several thrust and fold belts, bounded by the sedimentary basins (Nakata & Imaizumi, 2002). From east to west, at least three mountain ranges and two lowlands have been identified, namely the Kitakami Mountain, the Ou Backbone Range, and the Dewa Range bounded by the Kitakami lowland and the Yokote Basin, respectively (Figure 1a). The Ou Backbone Range (OBR) is one of the prominent ranges that runs through the center of the Tohoku Island arc (Kagohara et al., 2009; Miura et al., 2002). It forms a part of a large back-arc spreading rift basin (Yamaji, 1990), which during early to middle Pliocene experienced a vast regional east-west compression (Sato & Amano, 1991). The continued compression elevated the OBR to ∼1000 m, thereby reversing the sense from normal faulting to reverse faulting, through tectonic inversion (Sato et al., 2002).

In the study area, OBR can be classified into three geomorphic units, from west to east, (a) the Mahiru Mountain bounded to the west by the Yokote Basin, (b) N-S trending Yuda Basin in the center, and (c) the Ou Backbone Range to the east (Figure 2a). The elevation of the OBR varies from 200 m in the Yuda Basin to 750 m along the Mahiru Mountain, which is further classified into the Senya Hill, Misaka Hill, and the Yuzawa Hill (Figures 2a and 2b). The OBR is separated from the Yokote Basin by an east-dipping reverse fault, viz, Eastern Margin Fault Zone of Yokote Basin (EFZYB) and from the Kitakami lowland by a west-dipping Kitakami Teichi Seien Fault Zone (KTSFZ) (Figures 1a and 2a) (Sato et al., 2002). These faults converge at a depth of ∼13–15 km to form a pop-up structure (Sato & Hirata, 1998; Sato et al., 2002) (Figure 1b). As reported, the M\text{\textsubscript{\text{JMA}}} 7.2 Rikuu earthquake rupture propagated to the west and produced surface deformation along the northern segment of the EFZYB, whereas the 2008 M\text{\textsubscript{\text{JMA}}} 7.2/M\text{\textsubscript{\text{w}}} 6.9 Iwate-Miyagi Nairiku earthquake propagated to the east, and ruptured the southern segment of the KTSFZ, producing a uplift along the pop-up structure (AIST, 2008; Ishiyama et al., 2008; Matsuda et al., 1980; Matsu’ura & Kase, 2010).

Stratigraphically, the OBR consists of an alternating sequence of Miocene to Pliocene marine, and terrestrial sediments thrust over a 400-m thick Pleistocene terrestrial sequence of poorly sorted, and unconsolidated alluvial fan debris deposited within the modern Yokote Basin (Kagohara et al., 2009; Sato & Amano, 1991) (Table 1). Within the OBR, the base of the Senya and the Misaka Hills consists of the Yuda
and the Mahirugawa Formation, which are composed of lava, and the pyroclastic rocks, and overlain by the Miocene Yoshizawagawa Formation, which is composed of the deep marine siliceous mudstone (Kagohara et al., 2009) (Figure 2b) (Table 1). Further to the south, the Yuzawa Hill is composed of the cretaceous granites, which is overlain by the Late Miocene acid pyroclastic rock, which are, in turn, overlain by the Late Miocene to Early Pliocene dark gray mudstone, Late Pliocene to Holocene sandstone, and the recent terrace deposits (Figure 2b) (Table 2).

2.2. Active Fault Distribution Along EFZYB

The study focusses on the EFZYB, which extends for ~60 km, and separates the Yokote Basin from the OBR (Active Faults Research Group of Japan, 1991). The EFZYB is classified into at least five fault segments, from north to south, the Shiraiwa Fault, Ota Fault, Senya Fault, Kanazawa Fault, and the Omoriyama Fault (Matsuda et al., 1980; Research Group for Senya Fault, 1986) (Figure 2a). The detailed geomorphic and seismic reflection survey along the Shiraiwa Fault revealed an east-dipping thrust fault extending for ~14 km in length with a dip angle of 25°, while the Ota Fault segment extends for ~5 km, and shows a steep thrust ramp, and a shallow frontal ramp subsurface geometry (Kagohara et al., 2009; Sawa et al., 2013). The Senya Fault is the most studied N-S trending east-dipping thrust fault in this region, extending for at least 24 km in length with a dip angle of 35° (Kagohara et al., 2009). The Kanazawa and the Omoriyama Faults are N-S trending reverse faults that extend for at least 17 and 23 km, respectively, with a dip angle of 45° (AIST, 2010). The Kanazawa Fault is a range-bounding fault, whereas the Omoriyama Fault is located within the Yokote Basin, ~3.5 km westward of the Kanazawa Fault along the low-relief Akasaka Hills (Figures 2a and 2b). The deformation along both the faults has resulted in west-facing discontinuous fault scarps (Sawa et al., 2011, 2013; Taniguchi et al., 2007; Watanabe et al., 2011).

The geomorphic studies along the Shiraiwa and Senya Faults have suggested a late Pleistocene dip-slip rate of 1.2 ± 0.2 mm/yr and 0.9 ± 0.3 mm/yr, respectively along the shallow thrust ramp, while a millennial dip-slip rate of 1.1 ± 0.3 mm/yr along the deeper thrust ramp (Imaizumi et al., 1997; Kagohara et al., 2009; Matsu’ura & Sugaya, 2017). A geomorphic investigation along the Senya Fault revealed an average vertical displacement rate of 0.8 mm/yr over the past 23 ka, a coseismic vertical slip of at least 3.5 m and a recurrence interval of 3,000–4,000 years (Matsuda et al., 1980).

The Shiraiwa, Ota, Senya, and some parts of the Kanazawa Fault were reactivated during the 1896 Rikuu earthquake (Mw 7.2) but, due to the lack of seismological and geodetic measurement, the earthquake parameters were not constrained (Thatcher et al., 1980). However, the extensive field investigations along the EFZYB rupture zone revealed a surface rupture extending of at least 36 km along the Shiraiwa Fault, Ota Fault, Senya Fault, and along the northern section of the Kanazawa Fault, before terminating along the rest of the Kanazawa Fault (Active Fault Research Group of Japan, 1991; Matsuda et al., 1980) (Figures 2a and 2b).

| Period  | Epoch  | Name                      | Composition                      |
|---------|--------|---------------------------|----------------------------------|
| Neogene | Pliocene| Ainono Fr.                | Dark gray mudstone               |
|         | Miocene| Onnagawa and Kusanagi Fr. | Rhyolite and dacite lava          |
|         |        | Sannai Fr.                | Hard mudstone                    |
|         |        | Kaneyama, Nagao, Oanazawa Fr. | Acid pyroclastic rock and mudstone |
|         | Early  | Hosogoshizawa, Semi and Oinosawa Fr. | Pyroxene andesite lava |
| Mesozoic| Cretaceous| Abukuma Granites         | Hornblende-biotite tonalite, granodiorite |
3. Methods

3.1. Geomorphological Analysis

We targeted three locations, Yokote Site (along the Kanazawa Fault), Hiraka Site (along the Omoriyama Fault), and the Yuzawa Site (along the Higashi Chokai san Fault) in the unruptured fault segment of the EFZYB (Figures 2a and 2b). To characterize the fault trace, extensive geomorphological analysis was performed along the deformed Quaternary landforms, using aerial photo-interpretation, lidar DEM analysis, and the field survey. We used two sets of Geospatial Information Authority of Japan (GSI) aerial photographs, shot in the years 1946 and 1976 with a scale of 1:12,000 and 1:8,000, respectively. From the GSI stereo-pair images, several anaglyphs were generated (Goto & Sugito, 2012), which helped in the identification of the fault-related geomorphic indicators, such as, tectonic scarps, deformed fluvial terraces, tilted surfaces, and the uplifted fan surfaces. However, at some locations it was difficult to identify these features due to the dense vegetation cover, there we complemented our analysis with the 2-m lidar DEM to quantify the elevation of the fault scarps and differentiating different landforms. The lidar DEM was constructed from the .las pointcloud/.shp file using IWD interpolation on ArcMap v10.7 (Arrowsmith & Zielke, 2009). The different levels of fluvial terraces were delineated based on the topographic profiles, and the varying slope angles, extracted from the hillshade DEM, generated from the 3D Analyst and Spatial Analyst Tools in the ArcToolbox (Davis, 2012). The application of the aerial photo-interpretation and lidar data set is standardized, and most widely used methodology to identify the fault related geomorphic indicators (Arrowsmith & Zielke, 2009; Kondo et al., 2008; McCalpin, 2009). Based on the geomorphic analysis, we have measured 37 m lateral extent of the south unruptured segment of the EFZYB.

3.2. Seismic Survey

To reveal the subsurface geometry of the faults, three high-resolution S-wave seismic profiles A-A’, B-B’, and C-C’ were acquired at the Hiraka, and Yuzawa sites, respectively (Figures 4h and 7). A 460-m long A-A’ seismic profile was acquired across the Omoriyama Fault along the low-relief Akasaka Hill (Figure 4h). The B-B’ and C-C’ profiles were acquired along 988 and 194 m long Maemori and Omotemachi seismic line across the Higashi Chokai san Fault on the alluvial fan surface of Shirako River (Figure 7). For each cross-section profile, the vertical axis shows the subsurface depth (m) with positive amplitudes in black color. To acquire a high-resolution seismic profile, significant signal at 20–90 Hz was considered through a band-pass filter in SeisSpace® ProMAX®. The amplitude of the signal was adjusted by applying amplitude gain control and deconvolution. Further, through prepolymerization oscillation recording, the noise was suppressed, and S/N ratio was improved compared to the original oscillation recording. After the above treatment, NMO (Normal Moveout) correction was performed based on the velocity structure constructed by the velocity analysis, and further, CDP (Common Depth Point) polymerization was performed. Depth conversion was carried on the time sections before and after migration using the smoothed velocity structure, and each depth section was obtained.

3.3. Borehole and Dating

Based on the surrounding topography, and relation to the fault geometry, three boreholes were drilled (BH1–BH3) on the alluvial fan surface of the Shirako River near the township of Yuzawa. The boreholes BH1–BH3 were drilled up to a depth of 30, 38, and 35 m, respectively, to measure the vertical displacement of the marker litho-unit and further, to calculate the long-term slip rate. Two boreholes BH1 and BH2 were drilled on the hanging-wall side, whereas BH3 was drilled on the footwall side. The boreholes were drilled via TOHO-DO-D drilling machine with a core recovery rate of more than 90%. The collected cores were cleaned and sampled for radiocarbon dating. The lithology in the boreholes was matched based on composition, color, clast type, and fabric (the details are discussed in the results section).

For radiocarbon dating, we have collected the detrital charcoal, bulk organic sediments, and the wood samples, which were then sent to Beta Analytic Inc. (Laboratory No. 551773, Miami, USA). The samples containing wood and plant material were pretreated with the acid/alkali/acid washes, whereas the samples containing organic sediments were treated with acid washes (Broek et al., 2010). The roots and microfossils were removed from the samples using <180-micron sieves, followed by an acid wash to remove the
The conventional radiocarbon ages were calibrated to the calendar years using the IntCal13 or Marine13 radiocarbon database in OxCal v 4.4 (Reimer et al., 2013). Further, the high-probability density (HPD) range method was used, to statistically refine the likelihood of the age range within 95.4% probabilities (Ramsay, 1995).

4. Results

4.1. Site 1: Yokote

4.1.1. Geomorphology

The site is located 2-km north of the Yokote Railway Station at 140°33.782′E, 39°19.826′N, at the confluence of the Yokote and the Yoshizawa Rivers. These rivers drain the Mahiru Mountain (in the hanging wall of the Kanazawa Fault), and flows westward toward the Yokote Basin (Figure 3a). At least five levels of displaced fluvial terraces are identified in the hanging wall of the Kanazawa Fault along the Yoshizawa River, and classified as T0–T4, in order of increasing elevation (Figure 3a). The elevation of these terraces are determined from the river perpendicular topographic profiles extracted from the 2-m lidar DEM (Figure 3a inset). Terrace T4 tread is the most widespread fluvial terrace, identified along the banks of the Yoshizawa River, at an elevation of 12–15 m above the current riverbed. Based on the flat morphology, and wide lateral extent, we expect T4 to have deposited post Last Glacial Maximum (LGM). Similar widespread surfaces posidating LGM have been reported from several other regions in central Tohoku (Toyoshima, 1984).
Terrace T3 is identified on the right bank of the Yoshizawa River, at an elevation of 10–11 m above the river bed, T2 at 7–9 m, T1 at 5–6 m, T0 at 3–4 m, and Floodplain (Fp) at 1–2 m above the current river bed.

At the site, where the Yoshizawa River meets the Yokote River, the Kanazawa Fault has deformed all the four levels of terraces (Figure 3a). In order to measure the vertical displacement across the deformed terraces, several elevation profiles were extracted, which revealed 18 ± 1 m vertical displacement in terrace T4, 14 ± 1 m in terrace T3, 10 ± 1 m in terrace T1, and 7 ± 1 m along T0 (Figure 3b). In the hanging-wall, a gentle back-tilt is observed in the terraces T0, T1, and T3, which suggests bulldozing of the sediment successions along the fault trace, and a gentle flattening of the fault with depth (Figure 3b). Similar observations of back tilting in the fluvial terraces, as a result of tectonic uplift has been reported from the sedimentary basins in the Himalaya and New Zealand (Litchfield et al., 2010; Malik & Nakata, 2003).

The observed deformation along the terraces, with T4 (oldest terrace) showing the largest vertical displacement, and the terraces T0–T3 showing back tilting, provides a key evidence, suggesting that the Kanazawa Fault has been tectonically active since the Quaternary period. However, due to the absence of age control on these terraces, it was not possible to calculate the uplift rate along the Kanazawa Fault. Nevertheless, the observation supports the idea that the terraces have been deformed by multiple earthquake events.
4.2. Site 2: Hiraka

4.2.1. Geomorphology

The Hiraka site is characterized by a NNW-SSE striking low-relief landform extending for almost 10 km within the Yokote Basin, herein named as Akasaka Hill (Figure 4). The landform is located west of the Mahiru Mountain within the Yokote Basin, and unlike the Mahiru Mountain, it is incised by wide, alluvium-rich valleys (Figure 4a). A west to east trending long topographic profile (A-B) across the Yokote Basin (Figure 2c), indicates an abrupt rise in elevation from ~60 to ~110 m, followed by a drop in elevation to ~80 m, and again rising to ~250 m. Therefore, based on the topographic profile, the landforms within the Yokote Basin can be classified into a foreland alluvium, followed by an isolated low-relief Akasaka Hill, a widespread alluvial plain, and the Mahiru Mountain (Figure 2c). The width of the alluvial plain decreases southeastward, where it merges with the Mahiru Mountain.

The Akasaka Hill is bounded to the west by an east-dipping Omoriyama Fault, which has displaced the Quaternary alluvium, resulting in the south-west facing fault scarps, showing varying height of 1.2–2.7 m (Figure 4). The Akasaka Hill is surrounded by the Quaternary alluvium, and shows almost same lithology (Late Miocene to Pliocene dark gray mudstone) as the Mahiru Mountain, may suggest foreland propagation of the Mahiru Mountain along the Omoriyama Fault.

4.2.2. Seismic Survey

To determine the subsurface geometry, and deformation pattern of the Omoriyama Fault, a seismic reflection survey was carried out across the 1–1.5 m high fault scarp near the Aratokoro village at ~140°31.67′E, 39°16.972′N (Figures 4g and 4h). The depth section across the X-Y profile revealed continuous reflections up to a depth of 45 m, after which the reflections become discontinuous and weak. However, based on the reflection pattern, there is a clear indication of tectonic bulge, and back tilting between CDP 300 and CDP 150, which could be due to the sub-surface dragging of the lithounits along the fault. Further, a systematic change in the reflection can be identified at CDP 100, which marks an east-dipping line of discontinuity along the warped reflections (Figure 4h). The line of discontinuity can be traced from the surface to a depth of at least 60 m beneath the Akasaka Hill. The thrust plane shows a variable dip of ~20° near the surface and ~26° at the depth (Figure 4h). Further, from CDP 20 to CDP 60, the reflections are marked by a west-dipping synclinal fold, which may indicate presence of another fault trace west of the Omoriyama Fault (Figure 4h). However, a detailed paleoseismic investigation is required to understand the near surface geometry, and past seismic history along the Omoriyama Fault.

4.3. Site 3: Yuzawa

4.3.1. Geomorphology

The Yuzawa site is located near the township of Yuzawa, at the piedmont fan surface of the Shirako River, which drains the southern extent of the Mahiru Mountain (Figure 5). Based on the elevation difference and the slope angle, two generations of alluvial fan surfaces were identified (Figures 5d and 5e). The widespread Shirako River fan surface (AF-1) is sloping westward at an angle of 0.6°, is overlain by another fan surface (AF-2) from the nearby stream, sloping southeastward at an angle of 0.9° (Figures 5d and 5e, P3, P4). However, none of them have been dated so far, nonetheless, based on their elevation difference, and varying slope angles, we suspect AF-2 to be younger than the AF-1. The entire area of AF-1 is occupied by the town of Yuzawa, and has been subjected to anthropogenic disturbances; therefore, we used lidar DEM to identify the fault trace (Figure 5). Several long and short topographic profiles (P1–P5) were extracted across the fan surface. Particular caution was exercised to ensure that all the measurements were made on the same geomorphic surface, to ensure the tectonic origin of the observed elevation difference.

Based on the topographic slope break, two thrust fault scarps of the Higashi Chokai san Fault were identified, and herein referred as YF1 and YF2 (Figure 5). Across the study area, the scarp shows a total vertical displacement of 3.2–3.9 m (Figures 5b, 5c, and 5f; P1, P2, P5) which is a manifestation of active deformation along the Higashi Chokai san Fault. Further, to observe the lateral continuation of the fault trace, several profiles were extracted along the entire Yuzawa Hill (Figure 5). At the township of Iwasaki, 5-km northeast of the Yuzawa site, a flexure scarp was identified both in the anaglyph/lidar DEM, and in the field, showing
Figure 5.
a topographic break, and an elevation of 3 m (Figures 5g and 5h). At 3.5 km southeastward of the Yuzawa city, near the township of Sekiguchi, the scarp further degrades, and disappears, and reappears near the township of Yakushizawa, 8 km southeast of the Yuzawa city (Figures 5i–5l). The presence of fault scarps at the rangefront of the Yuzawa Hill may indicate that the entire hill has been tectonically deformed.

4.3.2. Borehole Survey

Three boreholes were drilled on the alluvial fan surface (AF-1) of the Shirako River (Figures 6a and 6b). Borehole 1 (BH1) was drilled in the hanging-wall of YF2, whereas BH2 and BH3 were drilled on the hanging-wall and footwall of YF1, respectively. The lithology in the BH1 is composed of typical alluvial fan debris deposit with subangular pebble and cobble gravel layers separated by small lenses of fine sand and silt (Figure 6a). Based on the varying lithology, size, color, and fabric of the clastic layers, the BH1 is separated into five lithounits. The top 1.7 m in BH1 (Unit-A) is considered as topsoil: A horizon based on its color, texture, and composition. Unit-B ranges in thickness from 1.7 to 2.4 m and consists of loosely bound, poorly
sorted, subrounded clast-supported gravel clasts ranging in size from 5 to 6 mm. Unit-C ranges from 2.4 to 5.3 m in thickness, and consists of brown, poorly sorted, loosely bound gravel composed of pebble clasts with an average size ranging from 10 to 40 mm. The gravel is predominantly composed of rhyolite, rhyolite tuff, and a small amount of andesite (Figure 6a). Unit-C’ ranges in thickness from 5.3 to 6 m and consists of subrounded, poorly sorted, grayish to bluish clast-supported gravel clasts. Unit-M ranges in thickness from 6 to 30 m and consists of suite of pale brown to gray gravels with dispersed black angular fragments. The gravel is predominantly composed of dacite, rhyolite, andesite, with dispersed sandstone and basalt fragments. Unit-M is distinguished from the overlying units based on its density, predominance of dacite, and the appearance of basalt fragments (Figure 6a). Small lenses of fine and medium-grained organic-rich sand and silt were encountered at various depths from which the sediment for radiocarbon dating was collected.

The BH2 and BH3 cores were drilled on AF-1 to a depth of 38 and 35 m, respectively, with an aim to recover the same lithologic units as exposed in the BH1 core from which to measure the vertical displacement and, in turn, calculate a slip rate. Based on the various distinctive lithology, color, clast size, and fabric, the stratigraphy in the BH2 and BH3 cores are classified into 15 and 8 lithostratigraphic units, respectively (Figure 6b). Unit-A based on its color and texture is considered as topsoil-A horizon, at the depth of 0.8 and 2.3 m in BH2 and BH3, respectively. In BH2, Unit-B ranges in thickness from 0.8 to 1.5 m and consists of loosely bound, clast-supported, poorly sorted gravel to pebbles clasts (Figure 6a). Unit-B’ in BH2, consists of 1.5–3 m thick, medium to coarse-grained sand with dispersed gravel clast. It is followed by Unit-C which ranges in thickness from 3 to 5.3 m and consists of poorly sorted, clast-supported, gravel clasts. The average size of the clasts ranges from 10 to 20 mm and is predominantly composed of andesite, rhyolite, and mudstone fragments. Unit-D ranges in thickness from 5.3 to 6.6 m and is composed of organic-rich yellowish-gray sandy silt (Figure 6a). Unit-E consists of grayish-brown poorly sorted, clast-supported subrounded gravel clasts and ranges from 6.6 to 9 m depth. The average size of the clasts ranges from 6 to 10 mm and are typically composed of rhyolite, granite, and a small amount of basalt. Unit-F is composed of yellow to bluish-gray organic silt with a mixture of fine to medium-grained sand with interspersed lenses of coarse sand mixed with gravel, and ranges from a depth of 9–11.1 m. Unit-G consists of poorly sorted brown gravel clast with matrix-supported, coarse sand and ranges up to 11.9 m. It is followed by Unit-H, which consists of clast-supported bluish-gray, subangular to subrounded gravel clast, ranging in size from 2 to 70 mm and ranges up to 13.4 m. Unit-I is composed of coarse-grained silty sand with interspersed brownish-gray gravel clast and ranges up to 15.5 m. Unit-J consists of subrounded, poorly sorted, matrix-supported gravel clasts.
with interspersed subangular grayish pebbles, and ranges in thickness from 15.5 to 28.7 m. The average size of the gravel ranges from 6 to 10 mm and is predominantly composed of granodiorite, andesite, basalt, and rhyolite. The matrix is coarse-grained sand. Unit-K ranges in thickness from 28.7 to 30.7 m and is composed of yellowish-brown clay. Unit-L consists of subrounded gravel clast dispersed in sandy unit and ranges in thickness from 30.7 to 34.2 m. Unit-M consists of clast supported subangular to subrounded greenish-gray pebbles and gravel clasts. The average size of the clasts ranges from 6 to 20 mm and is predominantly composed of dacite, andesite, and rhyolite. The unit is differentiated from the overlying units based on the dominance of greenish dacite, and the absence of granite (Figure 6a).

We were able to recover most of the lithologic units of BH2 in BH-3, except few units such as Unit-B and Unit-C, which are missing in BH3, and due to the limitations in the drilling till 33 m we were unable to penetrate the units K-M in BH3.

4.3.3. Age Constraint and Correlation

A total of 37 samples were collected from BH1, 2, and 3 for radiocarbon analysis (Figure 6, Table S1). In BH1, two organic sediments were collected from the Unit-A and Unit-C which gave an age of 670 ± 30 years and 3,750 ± 30 years, respectively. From the Unit-M, seven samples were collected, of which three samples (charcoal, wood, and plant material) were older than >43,500 years, while the rest of the four samples yielded an age of 35,170 ± 240 years, 36,620 ± 310 years, 29,440 ± 150 years, and 28,640 ± 120 years from the top to bottom (all from the organic sediment). We suspect the two younger ages (29,440 ± 150 years and 28,640 ± 120 years) in the cluster of older ages could be due to the contamination of the sediments with the younger charcoal or rootlet. Hence, we have considered 35,170 ± 240 years to represent the lower age bound of the Unit-M (Figures 6a and 6b).

In BH2, a total of 10 samples were dated, from Unit-D, three samples were collected YM-2 5.84–5.86 (organic sediment), YM-2 6.25–6.28 (peat) and YM-2 6.58–6.59 (plant material) which yielded almost consistent ages ranging from 8,990 ± 30 years to 9,820 ± 30 years. From Unit-F, two organic sediments were collected YM-2 10.45-10.48 (organic sediment) and YM-2 11.04-11.04 (organic sediment) which yielded an age of 25,460 ± 100 years and 24,580 ± 90 years, respectively, while the subsequent underlying units (Units-H and Unit-J) gave younger ages. The older ages from Unit-F and Unit-J' in BH2 can be explained by its fine-grained organic silt lithology, which deposits over a longer time span and are more susceptible to derived older wood or charcoal contaminations. From Unit-M only one sample is dated (YM-2 37.24–37.27) which yielded an age of 21,940 ± 80 years, which is relatively younger than that of Unit-M in BH1. We interpret younger age of Unit-M in the BH2 to represent the younger stratigraphy which got eroded from the hanging-wall scarp (BH1) and deposited on the footwall (BH2). The ages of lithounits in BH2 ranged from 8,990 ± 30 years to 21,940 ± 80 years (Figures 6a and 6b). Similarly, from BH3 a total of 18 samples were collected of which nine samples were collected from the Unit-F which yielded an older age with respect to the underlying units. We suspect older ages from the Unit-F is due to its fine-grained lithology, which takes more time to deposit and makes it more susceptible to the contamination from the older organic sediments or to the reworking. The ages of lithounits in BH3 ranged from 3,210 ± 30 years to 17,670 ± 50 years (Figures 6a and 6b).

Units-A can be correlated in all the three boreholes, whereas Unit-B and Unit-C is observed only in the BH1 and BH2. Unit-C in BH1 unconformably overlies the lower Unit-M, which is recognized based on the density contrast, the appearance of angular fragments of andesite, predominance of dacite, and a large age gap between the units above and below the unconformity. Based on the similarity in the lithology and age, Unit-M can be traced in BH1 and BH2 at the depth of 6 and 34.2 m, respectively, whereas due to the drilling limitations, the unit was not observed in BH3.

4.3.4. Seismic Survey

Two seismic reflection profiles were acquired across the Higashi Chokai san Fault to image its subsurface geometry (Figure 7). The profiles B-B’ and C-C’ extends for 988 and 194 m, respectively, crossing the alluvial fan surface. The high amplitude reflectors are visible in both of the profiles up to a depth of 80 m. Profile B-B’ reveals flat and continuous reflectors from CDP 200 to CDP 600, which becomes folded and discontinuous from CDP 610 to CDP 800 across an east-dipping trace of discontinuity, interpreted as a fault (Figures 7b and 7c). The truncation of reflectors along the east-dipping discontinuity can be correlated with the
identified surface trace of the YF2 fault scarps. Hence, the reflections in the B-B' seismic profiles revealed a northeast dipping thrust fault with a measured dip angle of ~30° (Figure 7). Similarly, the C-C' seismic profile is characterized by flat and continuous reflectors from CDP 20 to CDP 40, which then becomes folded from CDP 80 to CDP 140; the reflectors are truncated at the east-dipping discontinuity (Figure 7b). The folding of reflectors observed in both of the profiles can be interpreted as fault-related fold structures, which at the surface coincide with the YF1 and YF2 fault scarps.

The reflectors imaged in the profiles B-B' were correlated with the stratigraphy in the boreholes (BH1 and BH2) drilled on the hanging and footwall of the YF2. The correlation is based on the lateral extent and continuity of the reflectors. In profile B-B', from CDP 200 to CDP 350, a continuous and strong reflector is observed to a depth of 84 m below the surface (98 m). The reflector merges at a depth of ~9–10 m in BH1 and is interpreted to mark the top of Unit-M (Figure 7c). The reflector can be traced in the entire profile, almost continuous but folded at CDP 650 and CDP 750, beneath the surface trace of YF2 (Figure 7c). The western limit of the reflector is observed at CDP 950, to a depth of 54.7 m below the surface (86.3 m), and ~32 m in the BH2. This prominent reflector is interpreted to mark the top of Unit-M in the BH2. The correlation of the reflector, with the stratigraphy in boreholes, drilled in the hanging and footwall of YF2, indicates a vertical displacement of 29.8 m.

4.3.5. Slip Rate

We relate change in the elevation of the marker unit Unit-M, measured in the boreholes, and seismic reflection profiles, to the recent fault movements, which has uplifted the unit in the hanging-wall of the Higashi Chokai san Fault. The elevation difference between the Unit-M in BH1 and BH2 when projected onto the topographic profile across YF2, resulted in a vertical displacement of 36.1 m (Figure 6). Similarly, from B-B' seismic reflection profile, we have estimated a vertical displacement of 29.3 m (Figure 7c). The estimated vertical displacement (Vs) of 29.3 and 36.1 m yielded a vertical slip rate (U) of 0.83–1.02 mm/yr over the last 35,170 ± 240 years. To calculate the slip rate along the Higashi Chokai san Fault, the dip angle (θ) is derived from the seismic reflection data, which shows a mean dip angle of 30° (Figure 7c). Following the Caskey (1995) relationship, \( S = V_s \times \cos(\phi)/\sin(\phi+\theta) \), where \( S \) is the slip dip, \( V_s \) is the vertical displacement, and \( \theta \) is the dip angle of the fault, and \( \phi \) is the slope of the surface (AF-1) (0.6°), we have calculated the dip slip (\( S \)) of 57.5 and 70.0 m, and a long-term slip rate of 1.6 mm/yr to 1.9 mm/yr over the last 35,170 ± 240 years. Note that the estimated slip rate on the Higashi Chokai san Fault, represent the minimum slip rate because, in order to calculate the cumulative slip rate, we need to measure the vertical offset of Unit-M in BH1 and BH3. However, due to the limitation of drilling in BH3 to 33 m depth, it was not possible to penetrate Unit-M. Therefore, the calculated slip rate 1.6 mm/yr to 1.9 mm/yr represents the minimum long-term slip rate measured over a time duration of 35 ka along the Higashi Chokai san Fault.

Without the knowledge of the timing of the last large earthquake along the southern segment of the EFZYB, the estimation of the recurrence interval becomes critically important. Therefore, we used Wallace's, (1970) standard relation (\( R = D/U \)) to determine the average recurrence interval, where, \( R \) is the recurrence interval, \( D \) is the vertical slip per event, and \( U \) is the vertical slip rate. From this relationship, we have estimated a probable recurrence interval of about 3,800–4,600 years using the vertical slip rate of 0.83–1.02 mm/yr and vertical slip (\( D \)) of 3.9 m measured from the topographic profile across the fault scarp at the Yuzawa site (Figure 5f). It is assumed that the elevation of fault scarp used represents a single-event vertical slip, similar to the 3.5 m coseismic vertical slip reported by Matsuda et al. (1980) (along the northern ruptured segment of EFZYB during the 1896 Rikuu earthquake). The recurrence interval estimated from our studies are in coherence with the 3,000–4,000 years estimated from the northern ruptured segment of EFZYB (Research Group for Senya Fault, 1986). Further, to calculate the possible magnitude of the earthquake, we used, \( M_w = 2/3 \log_{10} (M_0) − 10.73 \) (Hanks & Kanamori, 1979) relationship, where \( M_0 = \mu \times \text{slip} \times L \times W \); the shear modulus (\( \mu \)) \( 3 \times 10^{11} \text{N/m}^2 \), \( L \) is the length of the active fault, and \( W \) is the downdip locking width. Following this relation, we have estimated \( M_w \) 7.1 for a 37 km long south unruptured segment of the EFZYB and 15 km deep brittle-ductile interface beneath the OBR (Hasegawa, et al., 2000) and \( M_{\text{MMA}} \) 7.6, following the Matsuda (1977) relationship (\( \log(D) = 0.6M − 4 \)), where \( D \) is slip per event and \( M \) is the magnitude, it is one of the most common methods applied to determine the magnitude of the earthquakes along the Japanese inland faults). Note that the estimated slip rates are considered minimum due to the limitation of data acquisitions, so the corresponding magnitude of the earthquake would be slightly underestimated.
5. Discussion

5.1. Structure of Higashi Chokai San Fault

The signatures of active faulting and slip rate estimates are very well documented along the historically active northern ruptured segment of the EFZYB (Imaiizumi et al., 1997; Kagohara et al., 2009; Matsuda et al., 1980). However, in our study, which is based on the geomorphic, seismic, and borehole survey data, we provide evidence of Quaternary faulting along the unruptured segment of the EFZYB, along the western margin of the Mahiru Mountain. We document the evidence of westward-facing thrust fault scarps, deformed alluvial fan surfaces, and back tilted fluvial terraces suggesting tectonic uplift along the Kanazawa Fault, Omoriyama Fault, and the Higashi Chokai san Fault (Figures 3–5). We further support our results with the subsurface seismic reflection data which indicates strong folding and dragging of the subsurface litho-stratigraphic units along the Omoriyama Fault and Higashi Chokai san Fault due to the ongoing rapid east-west shortening (Figures 4h and 7).

The Higashi Chokai san Fault is the frontal east-dipping thrust fault located ~8 km west of the Omoriyama Fault. It is bounded by an anticline in its’ hanging wall (Yuzawa Hill) which thrusts over the Yokote Basin sediments (Figures 2b and 2c). The presence of folded subsurface stratigraphic units, along with the observed unconformity, and large vertical displacement in the borehole stratigraphy drilled in the Quaternary alluvial fan deposits, favor the evidence, suggesting westward propagation of the Higashi Chokai san Fault from the Omoriyama Fault, with the two joining at the depth. However, more subsurface reflection data is required to confirm the inference that they join at the depth. Similar thrust-front migration has been observed along the ruptured segment of the EFZYB with the Kawaguchi Fault as the master fault and the Senya Fault as the propagated fault (Fujiwara, 1954). As reported during the 1896 Rikuu earthquake, the main rupture zone of the earthquake migrated along the frontal Senya Fault instead of the Kawaguchi Fault (Ike-da, 1983), making it important to understand the geometry of the frontal-most Higashi Chokai san Fault in the southern unruptured segment of the EFZYB. The evidence of foreland migration of the thrust faults and their capability of hosting large magnitude earthquake has been reported from the other parts of the world, such as the Himalaya and the Western Transverse Range, California (Levy et al., 2019; Yeats, 1986).

5.2. Tectonic Implication

In the last 200 years, five M ≥ 6 earthquakes have occurred along the ~100 km long OBR, nucleating at the depth of 12–15 km (Hasegawa et al., 1974; Komatsubara & Awata, 2001; Matsu'ura & Kase, 2010; Suzuki et al., 2010) (Figure 8a). In view of the high seismicity, various studies/models have been proposed to understand the generation of large magnitude inland earthquakes (Hasegawa et al., 2005; Shibazaki et al., 2008; Umino & Hasegawa, 2002). Based on the dense GPS network and seismic tomography studies, the OBR has been characterized as a zone of high strain concentration marked by shallow seismicity cutoff depths, with respect to the adjacent Kitakami Mountain and Dewa Range (Miura et al., 2004; Tanaka & Ishikawa, 2002). The studies further revealed spatial variation in the distribution of the low seismic velocity zone or high Vp/Vs ratio beneath the OBR, which coincides with the location of the two active volcanoes that is, Mt Akitakomagatake to the north, and Mt Naruko to the south (Hasegawa et al., 2005). Interestingly, the region of high Vp/Vs ratio zone also coincides with the hypocenters of the 1998 Iwate volcano earthquake (M6.1) and the 1996 Onikobe earthquake (M6) (Nishimura et al., 2005; Takada & Furuya, 2009). The plausible explanation for the observed seismicity is due to the partial melting and upwelling of the fluids, which may intrude and weaken the overlying crust, facilitating large earthquakes (Umino & Hasegawa, 2002; Zhao et al., 2000). However, it was further proposed that the mechanical weakening of crust beneath these volcanoes results in concentration of stress away from it, causing elastic deformation perhaps in between the region of Mt Akitakomagatake and Mt Naruko (Hasegawa et al., 2005) (Figure 8a).

Given that the frequency of the earthquakes is also controlled by the slip rate and width of the interseismic decoupling zone (Stevens & Avouac, 2015), a very shallow width is observed (∼15 km) along the OBR, which may suggest a reduced capacity along the OBR to store the elastic strain, which results in frequent failures, as illustrated by over five large magnitude earthquakes in the region. For example, 1896 Rikuu earthquake (M7.2) occurred in the zone of high-stress concentration, and propagated westward along the east-dipping reverse fault (Matsuda et al., 1980). The event produced a total of 36 km and 9–14 km of
surface ruptures along the strike of the Senya Fault and the Kawafune Fault, respectively, and ∼3.5 m of vertical displacement (Matsuda et al., 1980) (Figures 8a and 8b). The depth of the event is not known with certainty because the event predated the installed seismic network, but it was suggested not deeper than 15 km (Takagi et al., 1977). Given its magnitude and evidence of surface ruptures, we can infer that the event has ruptured the entire seismogenic thickness.

Just after 74 years, in 1970, another earthquake struck the OBR with magnitude M6.2, 30 km south of the Rikuu earthquake rupture zone. The GPS receivers and focal mechanism indicated that the event occurred at a depth of 13 km on a 15 km wide and 15 km long rupture zone, with an average slip of 25 cm, which propagated westward along the east-dipping Omoriyama Fault (Hasegawa et al., 1974; Komatsubara & Awata, 2001) (Figures 8a and 8b). The rupture, however, did not reach the surface, instead partially ruptured the subsurface Omoriyama Fault (Komatsubara & Awata, 2001). After the 1970 earthquake, two earthquakes occurred along the OBR, one in 1996 and another in 1998; however, both of the earthquakes are related to the volcanic activities (Hasegawa et al., 2005; Umino & Hasegawa, 2002) (SR-Slip Rate, U-Vertical Slip Rate, RI-Recurrence Interval).

In 2008, another earthquake occurred, with magnitude Mw 6.9 (MJMA 7.2), south of the 1970 rupture area, along a west-dipping reverse fault (Matsu’ura & Kase, 2010) (Figures 8a and 8b). The geodetic observation showed that the earthquake initiated at a depth of 6 km, along a ∼40 km long and 18 km wide fault plane which ruptured in two distinct patches, initial southern patch recorded a larger slip of 6.2 m, followed by the northern patch which ruptured with a smaller slip (Suzuki et al., 2010). The earthquake produced
~6 km of surface rupture along the discontinuous fault scarps with a measured vertical displacement of 0.5–1.5 m (Aoi et al., 2008; Headquarters for Earthquake Research Promotion, 2008).

The compilation of the earthquake rupture data since 1896 along the ~100-km long OBR, revealed no characteristic time predictable event, instead a southward migrating rupture directivity (Figure 8b). The 1896 Rikuu earthquake ruptured the locked segment of the Senya Fault and propagated southward along the strike and ruptured the northern section of the Kanazawa Fault before terminating. What controlled the abrupt termination of the Rikuu earthquake is still poorly understood, but the residual stress at the southern end of the rupture zone, along with the ongoing E-W compression resulted in the failure of the downdip portion of the neighboring Omoriyama Fault in 1970 Akita-ken Nantobu earthquake. However, the rupture did not reach the surface, instead partially ruptured the locked portion of the fault. The residual stress from the incomplete rupture makes the area more prone to a much larger event which may strike sooner than would have anticipated from the recurrence interval. Similar observations of bimodal seismicity showing partial and complete ruptures have also been reported from the other fault systems such as Sumatra megathrust and Himalaya (Arora et al., 2019; Konca et al., 2008; Zilio et al., 2019).

5.3. Seismic Gaps and Seismic Potential

A well-developed earthquake data along the OBR provided an opportunity to estimate the seismic potential of the faults associated with the past earthquakes. The paleoseismic investigation along the northern segment of the EFZYB showed a clear evidence of at least two events prior to the 1896 Rikuu earthquake with penultimate event around 3,500 years ago (Research Group for Senya Fault, 1986). The most recent 1896 Rikuu earthquake produced a 36 km long surface rupture, and records a net slip rate of 1.1 ± 0.3 mm/yr along the Senya Fault (western margin of the OBR) (Kagohara et al., 2009; Matsuda et al., 1980; Matsu'ura & Sugaya, 2017; Research Group for Senya Fault, 1986). Based on the paleoseismic data, it is fairly well to infer that the substantial amount of accumulated strain has been released along the ruptured segment of the EFZYB associated with the 1896 Rikuu earthquake.

Along the southern unruptured segment of the EFZYB, based on the measured 37 km lateral extent and 3.9 m slip per event measured along the Higashi Chokai san Fault in this study, we have estimated the earthquake magnitude following the Hanks and Kanamori (1979) and Matsuda (1977), scaling relationship (Figures 8a and 8b). We suggest that if the southern unruptured segment of EFZYB were to rupture in a single event, then the estimated magnitude would be $M_w7.1/M_{MA}7.6$. During the 1970 Akita-ken Nantobu earthquake (M6.2), only a small fraction of the slip, around 25 cm was released during the partial rupture (Komatsubara & Awata, 2001), suggesting a potential of triggered earthquake along the southern un-ruptured segment of EFZYB.

Along the eastern margin of the OBR, the aftershock distribution of the 2008 Iwate-Miyagi Nairiku earthquake (M7.2) suggested ~40 km long rupture length at depth (Suzuki et al., 2010), and ~6 km of discontinuous surface ruptures through field investigation (Headquarters for Earthquake Research Promotion, 2008). During the earthquake, no associated deformation was observed north of the rupture zone, along the 62-km-long Kitakami Teichi Seien Fault Zone (KTSFZ), nor the region has witnessed any large magnitude earthquake in the past three centuries. The last known earthquake might have occurred around 4,500 years ago (Headquarters for Earthquake Research Promotion, 2012). Due to the lack of paleoseismic data, especially a well-founded recurrence interval, it is difficult to constraint the timing of last historical earthquake with reasonable certainty. But, based on the available data, we suggest that the region still holds potential for a future earthquake. From the Hanks and Kanamori (1979) and Matsuda (1977), scaling relationship, we suggest if the 62-km long KTSFZ were to rupture in a single event, the magnitude would be $M_w7.3/M_{MA}7.9$ (Figure 8b).

Our results highlight two seismic gaps along the OBR; the first section is between the 1896 Rikuu earthquake and the 1996 Onikobe earthquake with the partial rupture zone of the 1970 Akita-ken Nantobu earthquake (Figures 8a and 8b). And the second is north of the 2008 IMEQ rupture zone along the 62-km long KTSFZ along the eastern margin of OBR, where the records of the last historical large magnitude earthquake is not known, and the paleoseismic records indicate the most recent event around 4,500 years ago (Figure 8a).
6. Conclusion

Based on the extensive aerial photo-interpretation, and analysis of the lidar data, seismic reflection survey, and borehole data, we provide the evidence of late Quaternary faulting along the Kanazawa Fault, Omoriyama Fault, and the Higashi Chokai san Fault, along the unruptured segment of M7.2 1896 Rikuu earthquake in the Yokote Basin. The geomorphic observations include flexure fault scarps, deformed and back-tilted fluvial terraces, and alluvial fan surfaces, and folded and warped sub-surface strata imaged in the seismic reflection survey. The geomorphic signatures were observed all along the piedmont front of the Ma-hiru Mountain. Dating of the lithostratigraphic units in the boreholes, and the measurement of the vertical displacement from the hanging wall and foot wall of the Higashi Chokai san Fault allowed us to determine a vertical slip rate of 0.8–1.0 mm/yr and a long-term slip rate of 1.6–1.9 mm/yr for the past 35 ka. Further, from the measured single event vertical displacement, and vertical slip rate, we have estimated a probable recurrence interval of about 3,800–4,600 years.

After compilation of the past earthquake data along the 100-km long seismically active OBR, our result highlights southward migrating earthquake ruptures along the strike of the OBR, and bimodal seismicity with historical events showing both partial and complete ruptures. From the slip rate and recurrence interval estimates, we have determined the slip potential of the past earthquake ruptures to better access the seismic hazards. Our results highlight two seismic gaps along the OBR. The most alarming is the one between the 1896 Rikuu earthquake and the 1996 Onikobe earthquake along the partial rupture zone of the 1970 Akita-ken Nantobu earthquake. The findings suggest that the region along the Omoriyama Fault and its branching out Higashi Chokai san Fault has not completely ruptured during the 1970 earthquake event and holds a potential of $M_w \sim 7.1$ which may fail sooner than would have been anticipated from the recurrence interval.

Data Availability Statement

The active fault data set is available at the Active Fault Database of Japan (https://gbank.gsj.jp/activefault/index). Also available in English by clicking the on the tab, in upper right corner of the site. The aerial photographs used for preparing the geomorphic maps can be obtained freely from the Geospatial Information Authority of Japan (https://mapps.gsi.go.jp/maplibSearch.do#1). The radiocarbon age data are stored in the PANGAEA Data Archiving & Publishing/PDI-29003 repository (https://issues.pangaea.de/browse/ PDI-29003). The lidar data used in this study was obtained from Kokusai Kogyo Co., Ltd. The figures are prepared in freely available Generic Mapping Tools v 6.0.0 and licensed ArcGIS v10.7.

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References

Active Faults Research Group of Japan. (1991). Active faults in Japan: Sheet maps and inventories. (revised edition, p. 437). Tokyo: University of Tokyo Press.

AIST. (2008). Advance reports on the 2008 Iwate-Miyagi Nairiku earthquake. National Institute of Advanced Industrial Science and Technology. Retrieved from http://unit.aist.go.jp/actfault/katsudo/jishin/IWATE_MIYAGI_INDEX.html

AIST. (2010). Active fault database of Japan, additional complementary survey of active faults result report No. I21-1 (in Japanese) activity history survey of the eastern margin fault zone (southern part) of the Yokote Basin. Retrieved from https://gbank.gsj.jp/activefault/ segment_param?SearchTYPE=&fval_type1=029-02&segment_id=029-02&topic_list=2&search_mode=2

Ammon, C. J., Lay, T., Kanamori, H., & Cleveland, M. (2011). A rupture model of the 2011 off the Pacific coast of Tohoku earthquake. Earth Planets and Space, 63, 693–696. https://doi.org/10.5047/eps.2011.05.015

Aoi, S., Morikawa, N., Kunugi, T., Suzuki, W., Pulido, N., & Fujihara, H. (2008). Characteristics of extreme found acceleration during the 2008 Iwate-Miyagi Nairiku earthquake. Abstract of 2008 fall meeting of seismological society of Japan.

Arora, S., Malik, J. N., & Sahoo, S. (2019). Paleoseismic evidence of a major earthquake event(s) along the hinterland faults: Pinjore Garden Fault (PGF) and Jhajra Fault (JF) in northwest Himalaya, India. Tectonophysics, 757, 108–122. https://doi.org/10.1016/j.tecto.2019.01.001

Arrowsmith, J. R., & Zielke, O. (2009). Tectonic geomorphology of the San Andreas Fault zone from high resolution topography: An example from the Cholame segment. Geomorphology, 113, 70–81.

Awata, T. (1999). Behavioral segments and their cascades for the earthquake occurrences in the Japanese surface ruptures. Proceedings of the paleoseismology workshop, 1999 GSI interim report.

Brock, F., Higham, T. F. G., Ditchfield, P., & Bronk-Ramsey, C. (2010). Current pretreatment methods for AMS radiocarbon dating at the Oxford Radiocarbon Accelerator Unit (ORAU). Radiocarbon, 52, 103–112. https://doi.org/10.1017/s0033822200045069

Caskey, S. J. (1995). Geometric relations of dip slip to a faulted ground surface: New nomograms for estimating components of fault displacement. Journal of Structural Geology, 17, 1197–1202. https://doi.org/10.1016/0191-8141(95)00023-7

Davis, O. (2012). Processing and working with LIDAR data in ArcGIS: A practical guide for archaeologists. (p. 23). Aberystwyth: The Royal Commission on the Ancient and Historical Monuments of Wales.
Fujisawa, K. (1954). On geomorphology in the northern part of the Yokote basin. *Annals of the Tohoku Geographical Association*, 7, 63–69. https://doi.org/10.5190/tga1948.7.63

Goto, H., & Sugito, N. (2012). Fault geomorphology interpreted using stereoscopic images produced from digital elevation models. *E-Journal GEO*, 7, 197–213. https://doi.org/10.4157/ejgeo.7.197

Hanks, T. C., & Kanamori, H. (1979). A moment magnitude scale” (PDF). *Journal of Geophysical Research*, 84(B5), 2348–2350. https://doi.org/10.1029/JB084ib05p02348

Hasegawa, A., Horuchi, S., & Umino, N. (1994). Seismic structure of the north-eastern Japan convergent margin: A synthesis. *Journal of Geophysical Research*, 99(22), 22295–22311. https://doi.org/10.1029/93jb02797

Hasegawa, A., Nakajima, J., Umino, N., & Miura, S. (2005). Deep structure of the northeastern Japan arc and its implications for crustal deformation and shallow seismic activity. *Tectonophysics*, 403, 59–75. https://doi.org/10.1016/j.tecto.2005.03.018

Hasegawa, A., Yamamoto, A., Umino, N., Horuchi, S., Zhao, D., & Sato, H. (2000). Seismic activity and deformation process of the overriding plate in the northeastern Japan subduction zone. *Tectonophysics*, 319, 225–239. https://doi.org/10.1016/s0040-1951(99)00296-6

Hasegawa, T., Hori, S., Hasegawa, T., Kasahara, K., Horuchi, S., & Koyama, J. (1974). On the focal mechanism of the Southeastern Akita Earthquake in 1970. *Journal of the Seismological Society of Japan*, 27, 302–312. https://doi.org/10.4249/izanishi1948.27.4_302

Headquarter for Earthquake Research Promotion. (2008). *Kitakami lowland west edge Fault Zone*. Headquarter for Earthquake Research Promotion. Retrieved from https://www.jishin.go.jp/regional_seismicity/rs_katsudansou/f013_kitakami/

Headquarter for Earthquake Research Promotion. (2012). *Evaluation of the western margin fault zone of the Kitakami lowland*. Headquarter for Earthquake Research Promotion. Retrieved from https://www.jishin.go.jp/main/choousa/katsudansou_pdf/13_kitakami-teichi.pdf

Ikeda, Y. (1983). Thrust-front migration and its mechanism; evolution of intraplate thrust fault system. *Bulletin of the Department of Geography, University of Tokyo*, 15, 125–159.

Imaizumi, T., Sato, H., Ikeda, Y., Ishimaru, T., Sakai, R., Yoneda, S., & Kubota, Y. (1997). Slip rate of the Senya fault, northeast Japan. *Programme and Abstracts of Japan Association for Quaternary Research*, 27, 84–85.

Ishiyama, T., Imaizumi, T., Otsuki, K., Kohsaka, S., & Nakamura, N. (2006). Field survey on surface rupture by the 2008 Iwate-Miyagi Nairiku earthquake.

Kagan, Y. Y., & Jackson, D. D. (2013). Tohoku earthquake: A surprise? *Bulletin of the Seismological Society of America*, 103, 1181–1194. https://doi.org/10.1785/0120120110

Kagohara, K., Imaizumi, T., Imaizumi, S., Miyachi, T., Sato, H., Matsuta, N., et al. (2009). Subsurface geometry and structural evolution of the eastern margin fault zone of the Yokote basin based on seismic reflection data, northeast Japan. *Tectonophysics*, 470, 319–328. https://doi.org/10.1016/j.tecto.2009.02.007

Konca, A. O., Avouac, J.-P., Sladen, A., Meltzer, A. J., Sieh, K., Fang, P., et al. (2008). Partial rupture of a locked patch of the Sumatra earthquake. *Journal of the Seismological Society of Japan*, 54, 33–44. https://doi.org/10.4249/izanishi1948.54.1_33

Kondo, H., Toda, S., Okumura, K., Takada, K., & Chiba, T. (2008). A fault scar in an urban area identified by LiDAR survey: A case study on the Toiagawa-Shizuoka Tectonic Line, central Japan. *Geomorphology*, 101, 731–739. https://doi.org/10.1016/j.geomorph.2008.02.012

Levy, Y., Rockwell, T. K., Shaw, J. H., Plesch, A., Driscoll, N. W., & Perea, H. (2019). Structural modeling of the Western Transverse Ranges: An imbricated thrust ramp architecture. *Lithosphere*, 11, 868–883. https://doi.org/10.1130/l1124.1

Litchfield, N., Wilson, K., Berryman, K., & Wallace, L. (2010). Coastal uplift mechanisms at Pakarue River mouth: Constraints from a combined Holocene fluvial and marine terrace dataset. *Marine Geology*, 270(1–4), 72–83. https://doi.org/10.1016/j.margeo.2009.10.003

Malik, J. N., & Nakata, T. (2003). Active faults and related late Quaternary deformation along the northwestern Himalayan Frontal Zone, India. *Annals of Geophysics*, 46(5), 917–936.

Matsuda, T. (1977). Estimation of future destructive earthquakes from active faults on land in Japan. *Journal of Physics of the Earth*, 25, S251–S260. https://doi.org/10.4249/pe1952.25.supplement_s251

Matsuda, T., Yamasaki, H., Nakata, T., & Imaizumi, T. (1980). The surface faults associated with the Rikuu earthquake of 1896. *Bulletin of the Earthquake Research Institute*, 55, 795–855.

Matsumura, T., & Kase, Y. (2010). Late Quaternary and coseismic crustal deformation across the focal area of the 2008 Iwate-Miyagi Nairiku earthquake. *Tectonophysics*, 487, 13–21.

Miyazaki, T., & Sugaya, K. (2017). Late Quaternary crustal shortening rates across thrust systems beneath the Ou Ranges in the NE Japan arc inferred from fluvial terrace deformation. *Journal of Asian Earth Sciences*, 140(1), 13–30.

McCalpin, J. P. (2009). *Paleoseismology: International Geophysics Series*. (second edition). Burlington, MA: Academic Press.

Miura, S., Sato, T., Hasegawa, A., Suwa, Y., Tachibana, K., & Yui, S. (2004). Strain concentration zone along the volcanic front derived by GPS observations in NE Japan arc. *Earth and Planets Space*, 56, 1347–1355. https://doi.org/10.1029/2003epsl0353340

Mori, N., Takenhashi, T., Yasuda, T., & Yanagisawa, H. (2011). Survey of 2011 Tohoku earthquake tsunami inundation and run-up. *Geophysical Research Letters*, 38, L00G14. https://doi.org/10.1029/2011gl049210

Nakata, T., & Imaizumi, T. (2002). Digital active fault map of Japan. Tokyo: University of Tokyo Press.

Nishimura, T., Tanaka, S., Yamawaki, T., Yamamoto, H., Sano, T., Sato, M., et al. (2005). Temporal changes in seismic velocity of the crust around Iwate volcano, Japan, as inferred from analyses of repeated active seismic experiment data from 1998 to 2003. *Earth and Planets Space*, 57, 491–505. https://doi.org/10.1016/j.tecto.2005.03.018

Ozawa, A., Hiroshima, T., Komazawa, M., & Suda, Y. (1987). *Geological map of Japan 1:200,000, Akita and Oga*. Tsukuba: Geological Survey of Japan.

Ramsay, B. C. (1995). Radiocarbon calibration and analysis of stratigraphy: The OxCal Program. *Radiocarbon*, 37(2), 425–430. https://doi.org/10.2458/azu_js_rc.55.16947

Reimer, P. J., Bard, E., Bayliss, A., Beck, J. W., Blackwell, P. G., Bronk Ramsey, C., & van der Plicht, J. (2013). Int-Ca13 and Marine13 radiocarbon age calibration curves 0–50,000 years cal BP. *Radiocarbon*, 55(04), 1869–1887. https://doi.org/10.2458/azu_js_rc.55.16947
Research Group for Senya Fault. (1986). Holocene activity of the Senya Fault (Akita prefecture) and the tip of the fault from excavation survey in Senhata town. *Bulletin of the Earthquake Research, 61*, 339–402.

Sato, H., & Amano, K. (1991). Relationship between tectonics, volcanism, sedimentation and basin development, Late Cenozoic, central part of northern Honshu, Japan. *Sedimentary Geology, 74*, 323–343. [https://doi.org/10.1016/0037-0738(91)90071-k](https://doi.org/10.1016/0037-0738(91)90071-k)

Sato, H., & Hirata, N. (1998). Deep structure of active faults and evolution of Japan islands. *Kagaku*, 68, 63–71.

Sato, H., Hirata, N., Iwasaki, T., Matsuoka, M., & Ikawa, T. (2002). Deep seismic reflection profiling across the Oze backbone Range, northern Honshu Island, Japan. *Tectonophysics, 355*, 41–52. [https://doi.org/10.1016/s0040-1951(02)00133-6](https://doi.org/10.1016/s0040-1951(02)00133-6)

Sawa, S., Watanabe, M., & Suzuki, Y. (2011). Drilling survey across the active reverse fault zone along the eastern margin of the Yokote Basin, northern Japan. *Japan Geoscience Union Meeting, 2011*.

Sawa, S., Tsutsumi, H., Sugito, N., & Kagohara, K. (2013). i: 25,000 Active fault map of the metropolitan area Yokote basin eastern margin fault zone and its surroundings “Lake Tazawa” “Yokote” “Yuzawa” commentary. (p. 24). Geospatial Information Authority of Japan Technical Data.

Seno, T., Sakurai, T., & Stein, S. (1996). Can the Okhotsk plate be discriminated from the North American plate? *Journal of Geophysical Research, 101*(B5), 11305–11315. [https://doi.org/10.1029/96JB00532](https://doi.org/10.1029/96JB00532)

Shibazaki, B., Garatani, K., Iwasaki, T., Tanaka, A., & Iyo, Y. (2008). Faulting processes controlled by the nonuniform thermal structure of the crust and uppermost mantle beneath the north eastern Japanese island arc. *Journal of Geophysical Research, 113*. [https://doi.org/10.1029/2007jb005361](https://doi.org/10.1029/2007jb005361)

Stevens, V. L., & Avouac, J. P. (2015). Interseismic coupling on the main Himalayan thrust. *Geophysical Research Letters, 42*, 5826–5837. [https://doi.org/10.1002/2015gl064845](https://doi.org/10.1002/2015gl064845)

Sugito, N. (2013). GSI active fault map of Yuzawa. Retrieved from [https://maps.gsi.go.jp/#14/39.175/140.54166665/#base=std&ls=std%7Cafm](https://maps.gsi.go.jp/#14/39.175/140.54166665/#base=std&ls=std%7Cafm)

Suzuki, W., Aoi, S., & Sekiguchi, H. (2010). Rupture process of the 2008 Iwate-Miyagi Nairiku, Japan, earthquake derived from near source strong-motion records. *Bulletin of the Seismological Society of America, 100*(1), 256–266. [https://doi.org/10.1785/0120090043](https://doi.org/10.1785/0120090043)

Takada, Y., & Furuya, M. (2009). Aseismic slip during the 1996 earthquake swarm in and around the Kitakobe geothermal area, NE Japan. *Earth and Planetary Science Letters, 290*. [https://doi.org/10.1016/j.epsl.2009.12.024](https://doi.org/10.1016/j.epsl.2009.12.024)

Takagi, A., Hasegawa, A., & Umino, N. (1977). Seismic activity in the northeastern Japan arc. *Journal of Physics of the Earth, 25*. [https://doi.org/10.4294/jpe1952.25.supplement_s65](https://doi.org/10.4294/jpe1952.25.supplement_s65)

Tanaka, A., & Ishikawa, Y. (2002). Temperature distribution and focal depth in the crust of the northeastern Japan. *Earth Planets and Space, 54*, 1109–1113. [https://doi.org/10.1186/bf03353310](https://doi.org/10.1186/bf03353310)

Taniguchi, K., Nakata, T., Watanabe, M., Suzuki, Y., & Goto, H. (2007). Active fault position and shape study working group (2007): Study on the position and shape of major active faults for improving the long-term evaluation of active faults: East edge of Yokote Basin and the Kitakami lowland western margin active fault zone as an example. *Japan geoscience union meeting 2007 proceedings*. (pp. S149–P026).

Thatcher, W., Matsuda, T., Kato, T., & Rundle, J. B. (1980). Lithospheric loading by the 1896 Rikuu earthquake in northern Japan: Implications for plate flexure and asthenospheric rheology. *Journal of Geophysical Research, 85*, 6429–6435. [https://doi.org/10.1029/jb085ib11p06429](https://doi.org/10.1029/jb085ib11p06429)

Toyoshima, T. (1984). The sequence of river terrace developed in the last 20,000 years in the Ou Backbone Range, northeast Japan. Science reports of the Tōhoku University. Series 7. Geography. (pp. 88–105). Tohoku University.

Umino, N., & Hasegawa, A. (2002). Inhomogeneous structure of the crust and its relationship to earthquake occurrence. In Y. Fujinawa, & A. Yoshida (Eds.), *Seismotectonics in convergent plate boundary.* (pp. 225–235). Tokyo: Terrapub.

Wallace, R. E. (1970). Earthquake recurrence intervals on the San Andreas fault. *The Geological Society of America Bulletin, 81*, 2875–2890. [https://doi.org/10.1130/0016-7606(1970)81[2875:erirota]2.0.co;2](https://doi.org/10.1130/0016-7606(1970)81[2875:erirota]2.0.co;2)

Watanabe, M., Nakata, T., Suzuki, Y., Goto, H., Tsutsumi, H., Taniguchi, K., & Sawa, H. (2011). Tectonics landform and active structure of the Yokote Basin. Japan geoscience union meeting. *Tectonics, 9*, 365–378. [https://doi.org/10.1029/te090003p00365](https://doi.org/10.1029/te090003p00365)

Yeats, R. S. (1986). Active folds related to folding. In R. E. Wallace (Ed.), *Active tectonics*. (pp. 63–79). Washington, DC: National Academy Press.

Yoshii, T. (1979). A detailed cross-section of the deep seismic zone beneath north-eastern Honshu, Japan. *Tectonophysics, 55*, 349–360. [https://doi.org/10.1016/0040-1951(79)90183-5](https://doi.org/10.1016/0040-1951(79)90183-5)

Zhao, D., Ochi, F., Hasegawa, A., & Yamamoto, A. (2000). Evidence for the location and cause of large crustal earthquakes in Japan. *Journal of Geophysical Research, 105*, 13579–13594. [https://doi.org/10.1029/2000jb000026](https://doi.org/10.1029/2000jb000026)

Zilio, D., van Dinther, Y., Gerya, T., & Avouac, J. P. (2019). Bimodal seismicity in the Himalaya controlled by fault friction and geometry. *Nature Communications, 10*(48), 1–11. [https://doi.org/10.1038/s41467-018-07874-8](https://doi.org/10.1038/s41467-018-07874-8)