

1 **The Effect of Seamount Subduction on the Formation of Holocene Marine Terraces:**  
2 **A Comparison of Kinematic and Mechanical Plate Subduction Models**

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11 **Key Points:**

- 12 • Traditional plate subduction models fall short in explaining Holocene marine terrace  
13 formation.
- 14 • A novel mechanical model addresses stress changes and deformations near a subducted  
15 seamount.
- 16 • Assessing megathrust earthquakes using Holocene marine terraces must account for plate  
17 interface irregularities.  
18

## 19 **Abstract**

20 Marine terraces have long been a subject of paleoseismology to reveal the rupture history of  
21 megathrust earthquakes. However, the mechanisms underlying their formation, in relation to  
22 crustal deformation, have not been adequately explained by kinematic models. A key challenge  
23 has been that the uplifted shoreline resulting from a megathrust earthquake tends to subside back  
24 to sea level during subsequent interseismic periods. This study focuses on the remaining permanent  
25 vertical deformation resulting from steady plate subduction and examines it quantitatively using  
26 three plate subduction models. Specifically, we pay attention to the effects of irregular geometries  
27 in the plate interface, such as subducted seamounts. Besides a simplified model examination, this  
28 study employs the plate geometry around the Sagami trough, central Japan, to compare with  
29 surface deformation observation. The subduction models employed are the kinematic subducting  
30 plate model, the elastic/viscoelastic fault model, and the mechanical subducting plate model  
31 (MSPM). The MSPM, introduced in this study, allows for more realistic simulations of crustal  
32 displacements by imposing net zero shear stress change on the plate boundary. Notably, the  
33 presence of a subducted seamount exerts a significant influence on surface deformation, resulting  
34 in a concentrated permanent uplift above it. The simulation of earthquake sequence demonstrates  
35 that coseismic uplifts can persist over time and contribute to the formation of marine terraces. The  
36 results demonstrated that the geological observations of coseismic and long-term deformations can  
37 be explained by the influence of a subducted seamount, previously identified in seismic surveys.

## 38 **Plain Language Summary**

39 This study explores how marine terraces are created resulting from plate subduction. Existing  
40 models struggle to explain why these terraces persist. In traditional models, the ground lifted  
41 during earthquakes sink back by the same amount after the earthquake, but this doesn't match real  
42 observations. In this study, we used a simulation to understand the crustal deformation around the  
43 plate subduction zone. Specifically, we looked at how uplift happens when there is an irregularity  
44 on the plate boundary. Because previous models did not consider the effects of such irregularity,  
45 we also made a new subduction model. As a result, we found irregularities on plate boundary can  
46 lead to permanent deformation that is more significant than in the previous simulation. testing our  
47 model on the Boso Peninsula in central Japan, the simulated deformation matched real observation  
48 of marine terraces. This research highlights the importance of considering plate geometry when  
49 studying the crustal deformation and earthquake history using marine terraces.

## 50 **1 Introduction**

51 Accurate assessment of the seismic hazard of a particular area requires a thorough  
52 understanding of the past earthquakes that have occurred on the relevant fault. However, the  
53 intervals between great earthquakes can span hundreds or even thousands of years, exceeding the  
54 range of modern instrumental observations which are typically limited to around one hundred years  
55 at best. Consequently, we must rely on historical records and geological data to reveal earthquake  
56 occurrence histories.

57 Holocene marine terraces are widely recognized as an important geological record of past large  
58 earthquakes, especially around subduction zones. When megathrust earthquakes occur along  
59 subduction zones, they can generate intense uplifts and subsidence in the surrounding areas. Such  
60 uplifts may create a stair-case coastal landform by emerging a beach and wave-cut bench. This  
61 phenomenon has been observed in recent earthquakes such as the 1923 Taisho Kanto earthquake  
62 (Shishikura, 2014), the 2004 and 2005 Sunda megathrust earthquakes (Briggs et al., 2006), and

63 the 2016 Kaikoura earthquake (Clark et al., 2017). While some of these uplift events include  
64 movements on upper plate faults branching from the plate interface (e.g., Clark et al., 2017), others  
65 are attributed to slips on the plate interfaces. Recurrence of such uplifts over time can lead to the  
66 development of Holocene marine terraces, which have been observed on various coasts around  
67 subduction zones and studied extensively seeking to understand earthquake recurrence (Shimazaki  
68 and Nakata, 1980; Ramos and Tsutsumi, 2010; Wang et al., 2013; Litchfield et al., 2020), including  
69 those attributed to upper plate faulting. Therefore, the Holocene marine terraces are highly  
70 valuable records for investigating past megathrust earthquakes.

71 However, the approach of using marine terraces to investigate past earthquake recurrence has  
72 been subject to questions. While the uplift accumulation can be explained by slip recurrence on a  
73 fault when coseismic deformations are attributed to intraplate faulting (e.g., Ninis et al., 2023),  
74 understanding the formation of marine terraces uplifted due to interplate slip is not straightforward.  
75 Specifically, the back-slip model, which is a well-known kinematic model for crustal deformation  
76 resulting from earthquake recurrences along subduction zones developed by Savage (1983),  
77 assumes that the coseismic slip and interseismic back-slip are equal in magnitude but opposite in  
78 direction, resulting in a net zero amount of slip on the fault after an earthquake sequence. This  
79 assumption suggests that the total amount of crustal deformation will also be net zero, when elastic  
80 deformation is assumed. As a result, the formation of marine terraces is deemed unlikely under  
81 this model.

82 The existence of marine terraces along various coasts without significant upper plate faulting  
83 raises questions about the permanent uplift resulting from plate subduction. One explanation for  
84 this uplift was proposed by Sato and Matsu'ura (1988), who suggested that steady subduction can  
85 generate permanent vertical deformation through fault slip in an elastic-viscoelastic stratum.  
86 Fukahata and Matsu'ura (2006; 2016) indicated that this permanent deformation is caused by the  
87 interaction between the curvature of plate interface and gravitational compensation. In addition,  
88 Kanda and Simons (2010; 2012) proposed that an elastic model can account for permanent vertical  
89 deformation by assuming steady slips on the upper and lower interfaces of the subducting slab,  
90 resulting from the effect of plate bending. While these models focused on long-term deformations  
91 and did not distinguish into individual earthquake sequences, the accumulation of deformations  
92 resulting from earthquake sequences should ultimately yield the same distribution.

93 However, these models focus on longer spatial and temporal scales of over 100 km and 10–  
94 100 thousand years, which are more relevant to great-scale terrains such as island arcs. To study  
95 the deformation later than the Holocene glacial retreat (<10k years BP) and its impact on marine  
96 terraces, the paleo-seismological investigations must focus on smaller-scale spatiotemporal  
97 deformation histories. Recent geological investigations of marine terraces have yielded essential  
98 observations. For example, the Holocene marine terraces on the southernmost tip of the Boso  
99 Peninsula in central Japan (Figure 1) indicated that their elevation distribution abruptly decreases  
100 within a short distance and its typical wavelength ranges 5–10 km (Komori et al., 2021).  
101 Furthermore, these marine terraces showed elevation changes along the strike direction of the  
102 subduction zone, indicating inhomogeneity in the subduction geometry.

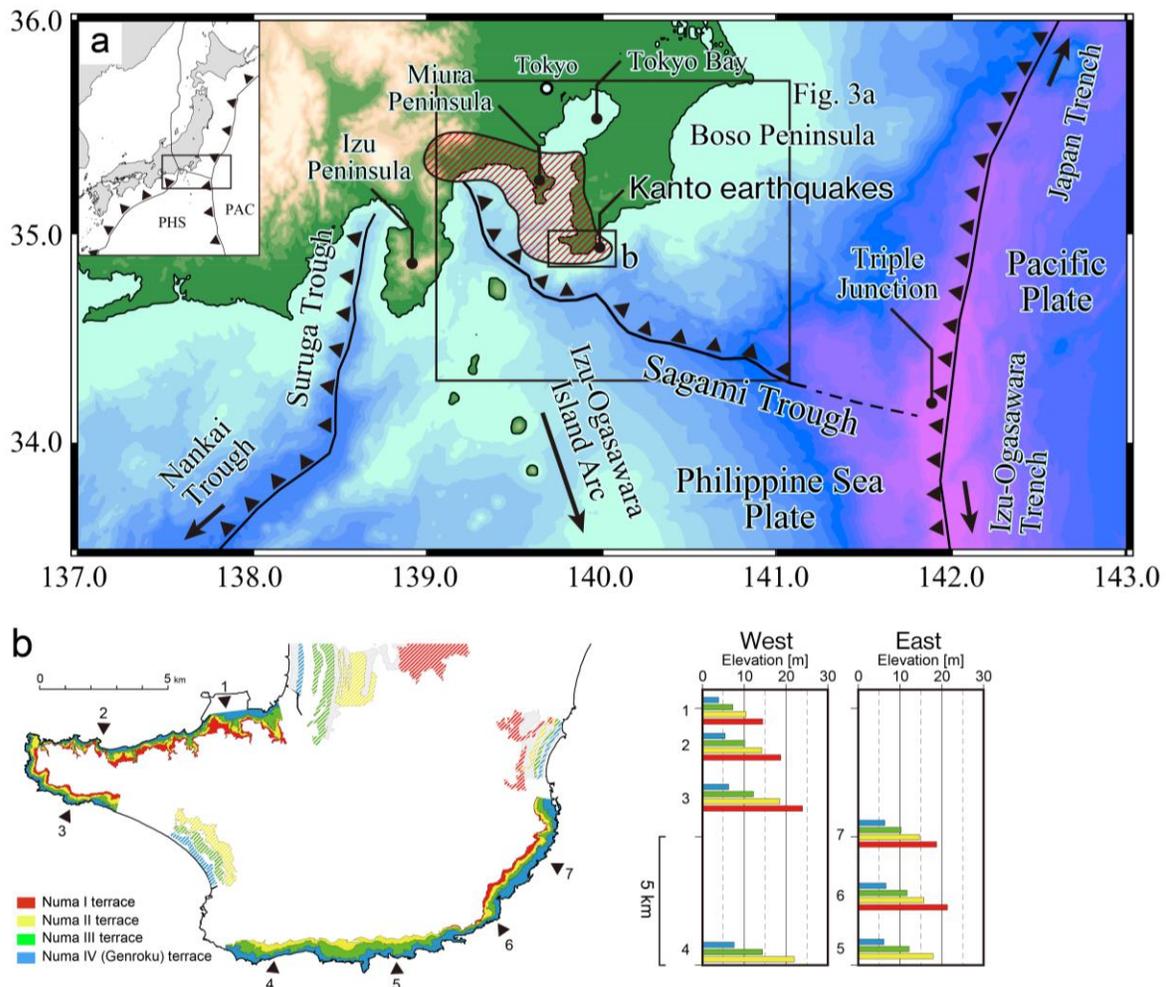


Figure 1. Survey region of this study and geological observations. (a) Tectonic setting of the Sagami Trough. The red meshed area indicates the estimated source region of the historical Kanto earthquakes (Sato et al., 2005; Sato et al., 2016). (b) Distribution of the Numa terraces after Komori et al. (2020). Right panels show the elevation distribution of the Numa terraces at each reference point, indicated by triangles in the left map.

103

104 Many previous studies have focused on the influence of plate interface irregularities on various  
 105 tectonic and seismological phenomena (e.g., Wang and Bilek, 2011). Specifically, the subduction  
 106 of seamounts has been explored through various model simulations, including its impact on the  
 107 tectonic formation of accretionary prisms (Miyakawa et al., 2020) and its effects on earthquakes,  
 108 including slow slip events, resulting from the irregular geometry and fluid intrusion (van Rijnsingen  
 109 et al., 2018; Sun et al., 2020). However, the investigation of crustal deformation due to subduction  
 110 over timeframes ranging from 1,000 to 10,000 years, which is the focus of this study, has not  
 111 received sufficient attention through quantitative model simulations. This timescale falls in a  
 112 middle ground between the previous objectives, where the deformation is transient and negligible  
 113 compared to that in active tectonic structures, while the sequence of coupling and rupture can be  
 114 approximately averaged. Consequently, modeling of plate subduction involving multiple

115 earthquake sequences and mechanically consistent crustal deformation model over a millennium  
116 scale is required.

117 Regarding crustal deformation, the impact of subducting seamounts on the surface and seafloor  
118 geometry has been suggested by analogue experiments (Dominguez et al., 1998; 2000) and  
119 geological observations (Kodaira et al., 2000; Gardner et al., 2001). The long-term deformation  
120 can be broken down into an accumulation of deformations resulting from individual earthquake  
121 sequences. Therefore, such long-term deformation patterns are presumed to reflect the asymmetry  
122 of deformation between inter- and coseismic periods. Given the significance of understanding the  
123 deformation sources for interpreting Holocene marine terraces, it is imperative to conduct a  
124 quantitative investigation of the effects of subducted seamounts over timescales spanning  
125 thousands of years.

126 In the Sagami Trough subduction zone, the target region of this study, a seismic reflection  
127 survey detected the bump geometry of a subducted seamount (Tsumura et al., 2009), and its effect  
128 on crustal deformation has been discussed (Sato et al., 2016). However, previous modeling  
129 investigations have encountered the difficulty in simulating the formation of the Holocene marine  
130 terraces, which was possibly resulted from the assumption of smooth plate interface geometry and  
131 underestimation of the effect due to interface irregularities.

132 We conducted a modeling study on crustal deformation concerning marine terrace formation,  
133 which is resulting from plate subduction. Recognizing the inadequacy of previous models to  
134 explain residual uplift following earthquake sequences, we started with a simple modeling  
135 examination to establish the asymmetry of crustal deformation distribution between interseismic  
136 coupling and coseismic rupture, rather than relying on individual case studies. The suspected factor  
137 contributing to this asymmetry is irregular geometry at the plate interface. However, since  
138 traditional subduction models often implicitly assume a smooth interface geometry, introducing  
139 irregularities into such models may lead to mechanically inconsistent assumptions and potential  
140 misinterpretation of crustal deformation.

141 To address this concern, this study proposed a mechanically consistent subduction model  
142 designed to accommodate complex plate interface geometries, including irregularities such as  
143 subducted seamounts, and evaluated its impact on simulated deformation with a simple modeled  
144 subduction geometry. Finally, we compared the long-term vertical deformation distribution  
145 observed in the Boso Peninsula with the model simulation results, discussing the significance of  
146 plate interface geometry in assessing crustal deformation histories around subduction zones.

## 147 **2 Sagami Trough Subduction Zone**

148 The Sagami Trough is a convergent plate boundary where the Philippine Sea Plate (PHS)  
149 subducts in a northwestward direction beneath the continental plate of northeast Japan at a rate of  
150 approximately 30–40 mm per year (Seno et al., 1993; DeMets et al., 1994) (Figure 1a). This  
151 subduction zone exhibits a highly complex geometry, with the eastern and western ends marked  
152 by the triple junction, where the Pacific Plate subducts below the PHS, and the Izu Peninsula, a  
153 collided volcanic island, respectively. Historical documents record the occurrence of two interplate  
154 earthquakes along this plate boundary: the 1703 M8.2 Genroku Kanto earthquake and the 1923  
155 M7.9 Taisho Kanto earthquake (hereafter, the 1703 Genroku earthquake and the 1923 Taisho  
156 earthquake, respectively) (Usami et al., 2013). The 1923 Taisho earthquake caused an uplift of  
157 approximately 2 m in the coastal area around Sagami Bay. Additionally, geological evidence  
158 shows that the southernmost tip of the Boso Peninsula experienced an uplift of approximately 6 m  
159 during the 1703 Genroku earthquake. While we have built less consensus about the interval times

160 of the Kanto earthquakes, the elastic recovery of these earthquakes probably has not been fully  
161 completed. The source fault of the 1923 Taisho earthquake is broadly acknowledged to be on the  
162 upper boundary of PHS from geodetic and teleseismic inversions (Sato et al., 2005; Nyst et al.,  
163 2006). While some geodetic inversion argued a possibility of activation of an inland fault (Pollitz  
164 et al., 1996), subsequent geological studies have proved no recent activity on the corresponding  
165 faults.

166 In addition to the marine terraces formed in historical era, older uplifted coasts are also  
167 recognized at the southernmost part of the Boso peninsula. These terraces, known as the Numa  
168 terraces, have been the subject of numerous geological and geomorphological studies. (Watanabe,  
169 1929; Matsuda et al., 1978; Nakata et al., 1980; Kawakami and Shishikura, 2006; Komori et al.,  
170 2020; 2021) (Figure 1b). The Numa terraces are classified into four levels, namely Numa I, II, III,  
171 and IV in descending order (Nakata et al., 1980). The lowest one, Numa IV, is the uplifted coast  
172 caused by the 1703 Genroku earthquake. The distribution pattern of these terrace platforms  
173 suggests that the Numa terraces likely represent records of similar type >M8 class megathrust  
174 earthquakes, referred to as Genroku-type earthquakes. Besides the Genroku-type earthquakes,  
175 there are also earthquakes that occur more frequently but cause minor uplift up to 1–2 m. Beach  
176 ridges distributed along the western coast of the Boso peninsula imply the recurrence of  
177 earthquakes similar to the 1923 Taisho earthquake (Shishikura, 2014).

178 Previous studies have extensively discussed the formation scenario of the Numa terraces and  
179 the occurrence history of the Kanto earthquakes. Some of these earlier studies, such as Matsuda et  
180 al. (1978), attempted to correlate the distribution of Numa terraces with the pattern of coseismic  
181 uplift and interseismic subsidence associated with historical earthquakes. However, the similarity  
182 in the spatial distribution of marine terraces does not necessarily provide straightforward evidence  
183 for the recurrence of characteristic earthquakes because the influence of interseismic deformation  
184 is much greater than the variation in coseismic deformations. Sato et al. (2016) explored the  
185 permanent uplifts caused by the plate subduction at the southernmost Boso peninsula. They used  
186 the kinematic formula within an elastic/viscoelastic half-space (Sato and Matsu'ura, 1988) to  
187 demonstrate this permanent surface deformation. However, their findings suggested that long-term  
188 deformation around the subduction zone could be approximated as steady motion, and they  
189 concluded that the formation of the Numa terraces was not directly related to the Kanto  
190 earthquakes, except for Numa IV. Noda et al. (2018) proposed an explanatory model for the current  
191 elevation distributions of the Numa terraces by combining steady uplift and sea level fluctuations,  
192 a concept often applied to late-Pleistocene marine terraces. This model hypothesized that the Numa  
193 terraces might have a reversal formation age (i.e., a higher terrace is younger than a lower terrace)  
194 at certain locations. However, subsequent geological studies (Komori et al., 2020; 2021) did not  
195 find evidence to support such a feature in the Numa terraces.

196 In addition, previous studies have highlighted several discrepancies between existing models  
197 and geological observations of the Numa terraces. One notable inconsistency lies in the  
198 concentrated distribution of permanent uplift caused by plate subduction. In conventional crustal  
199 deformation models introduced later, the characteristic wavelength of deformation is typically  
200 comparable to plate thickness, extending broadly up to 100 km from the trench axis. However, our  
201 previous geomorphological study revealed a steep decrease in elevation within 10 to 20 km  
202 distance (Komori et al., 2020). Such feature is possibly seen in a place where upper plate faulting  
203 occurs (e.g., Clark et al., 2017), but no evidence of active inland fault is confirmed around this  
204 area. Furthermore, the feature of Numa terraces where the relative elevations do not correspond  
205 proportionally to their formation intervals (Komori et al., 2021) serves as another example of how

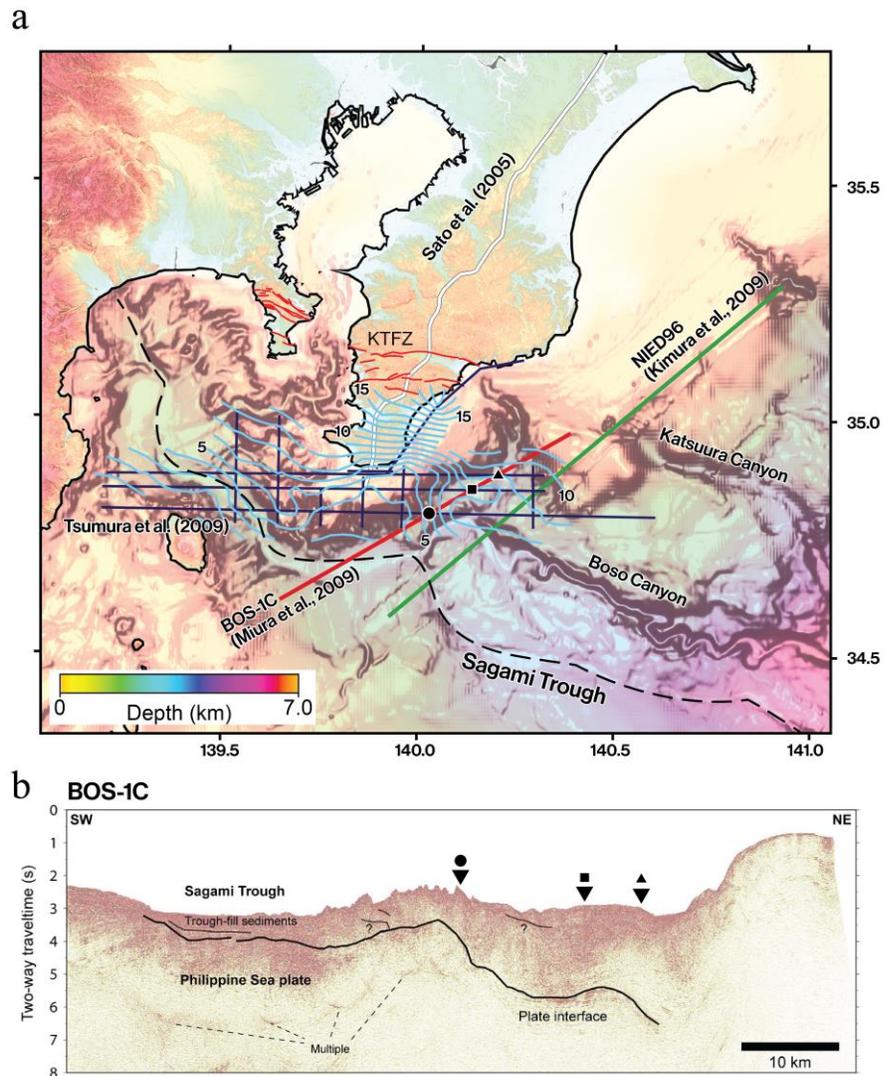


Figure 2. (a) Bathymetry map around the survey region and the profile lines of the previous reflection surveys (Sato et al., 2005; Kimura et al., 2009; Miura et al., 2009; Tsumura et al., 2009). The blue contour lines indicate the estimated depth of upper PHS by Tsumura et al. (2009), where the dark-blue straight lines are the survey profiles. The red lines indicate the inland active faults, where KTFZ stands for Kamogawa-teichi fault zone. (b) Post stack time migrated reflection image of the BOS-1C profile (Miura et al., 2009). Solid black line is our interpretation of the plate interface. Triangles indicate the positions of intersection with the survey lines of Tsumura et al. (2009).

206 conventional crustal deformation models fail explain the formation history, although this issue is  
 207 not examined in this study. These contradictions suggest that a more fundamental understanding  
 208 of crustal deformations is necessary for evaluating past earthquake histories.

209 This study aims to explore the relationship between permanent uplift, namely the accumulated  
 210 deformation resulting from multiple earthquake sequences, and plate interface geometry. Previous  
 211 reflection surveys have extensively investigated the tectonic structure around the Sagami Trough  
 212 subduction zone and the upper interface geometry of the PHS. Figure 2a illustrates the profiles

213 from these earlier surveys (Sato et al., 2005; Kimura et al., 2009; Miura et al., 2009; Tsumura et  
214 al., 2009). Tsumura et al. (2009) conducted surveys in the nearest shore region to our study area  
215 (Figure 2a) and reported the presence of a subducted seamount. Furthermore, Miura et al. (2009)  
216 obtained a cross-section in the southeast offshore Boso (Figure 2b) that intersects multiple survey  
217 lines from Tsumura et al. (2009). A comparison of these cross-sections in the migrated time  
218 sections (see Figure 7 in Tsumura et al. (2009)) reveals comparable positions of the reflectors at  
219 the cross points. As a result, these two independent surveys strongly suggest the existence of an  
220 irregular geometry, possibly a subducted seamount, beneath the southernmost part of the Boso  
221 Peninsula. The tectonics in this subduction zone (Figure 1a) suggests that this subducted seamount  
222 is possibly a part of the Izu-Ogasawara Island Arc.

223

224 In geological studies conducted in other regions, upper plate faults branching from the main  
225 thrust have been identified as potential causes of permanent deformations around subduction zones  
226 (Plafker et al., 1969; Litchfield et al., 2020). In the surrounding region of this study area, there is  
227 no clear evidence of significant activity of intraplate faults in the upper plate. Approximately 20  
228 km north of the study area, in the central part of the Boso Peninsula, an active fault zone  
229 (Kamogawa-teichi fault zone) is recognized (Nakajima et al., 1981). However, geological records  
230 of recent activities in the late Quaternary are not evident in this fault zone (Komatsubara, 2017).  
231 In the offshore region, Kimura et al. (2009) identified several splay faults branching from the main  
232 thrust. However, the branching faults in the shallower part, which likely form Boso Canyon at the  
233 seafloor, do not connect to other reflection survey results in the nearshore (Miura et al., 2009;  
234 Tsumura et al., 2009). It is possible that this branch fault has merged with the main thrust, where  
235 Boso Canyon meets the Sagami Trough. Another branching fault in the northeast, potentially  
236 exposed as Katsuura Canyon, appears to connect to the Kamogawa-teichi fault zone based on the  
237 seafloor topography (Kimura et al., 2009). Consequently, for the purpose of our modeling work,  
238 we assume that upper plate faulting does not significantly contribute to the crustal deformation in  
239 the region and that coseismic deformation is due to subduction interface earthquakes.

### 240 **3 Subducting Plate Models**

241 The crustal deformation models accompanying plate subduction have been proposed by  
242 various modeling studies. However, it is challenging to find an ideal model that explains all  
243 phenomena around subduction zones. Instead, these models have been developed with different  
244 scales focusing on specific phenomena. Because the target phenomena range from momentary  
245 earthquake events to long-term deformation leading to island-arc formation, we must choose the  
246 most appropriate model depending on the purpose.

247 This study aims to investigate the deformation resulting from a repetition of interseismic  
248 coupling and coseismic ruptures. Each co- and interseismic deformation depends on the range of  
249 coupling patches and slip amount. Hence, even with highly simplified first-order approximated  
250 models, such as back-slip models, the expected errors due to fault geometry may result within a  
251 negligible range. However, when considering the cumulative effect of these deformations over  
252 time, the differences in assumptions regarding how a plate subducts become significant. Therefore,  
253 we compared four subduction models, including three existing models and one newly developed  
254 model, while paying attention to the irregularity on the plate interface (Figure 3). In this section,  
255 we first review the settings and characteristics of the subduction models used in previous studies.

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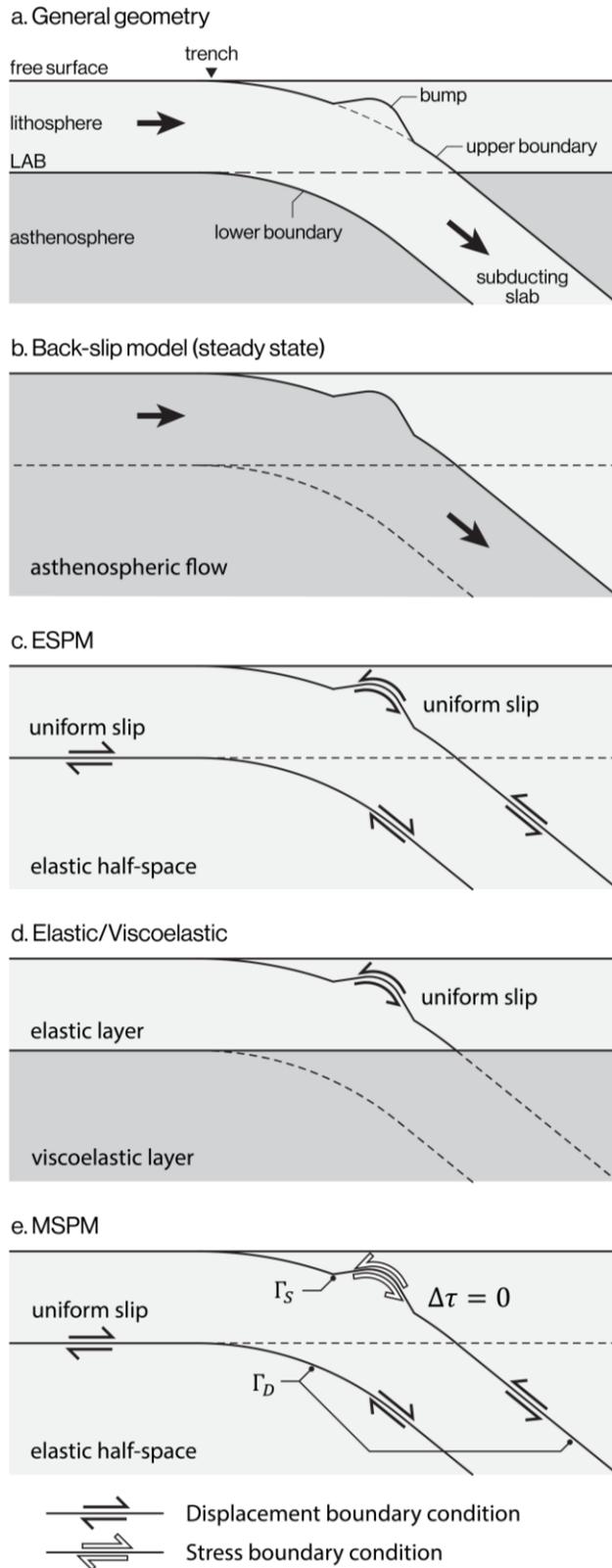


Figure 3. Schematic illustration of subduction models. (a) General geometrical setting of plate subduction. (b) Schematic representation of the imposed steady state assessed from the back-slip model, following the interpretation by Kanda and Simons (2010). (c) Slip configuration for the steady state of ESPM (Kanda and Simons, 2010). Uniform slip is imposed on the entire plate interfaces (double arrows). (d) Schematic structure illustration of the elastic/viscoelastic model. Uniform slip is imposed on the plate interface above the LAB. (e) Boundary conditions of MSPM. Black solid arrows and white arrows indicate the interfaces where uniform slip is imposed ( $\Gamma_D$ : area of displacement boundary condition) and no shear stress change occurs ( $\Gamma_S$ : area of stress boundary condition), respectively.

### 258 3.1 Back-slip Model

259 The back-slip model is the simplest approximation for subduction zones, first proposed by  
260 Savage (1983), that explains the strain accumulation during the interseismic period by assuming a  
261 back-slip solution due to the plate interface coupling that is given to entirely compensate the steady  
262 motion of the plate subduction (forward slip on the plate interface). This model is widely used  
263 because it only requires information on the slip deficit, the amount of back-slip, on the coupling  
264 zone, rather than constraining the slips on decoupled zone across a large part of plate interface.  
265

266 However, the back-slip model exhibits a critical weakness. It assumes that steady slip along  
267 the plate boundary does not induce deformation within the upper crust, which is actually a very  
268 strong assumption. According to Kanda and Simons (2010), the back-slip model corresponds to  
269 assuming a viscously deformable subducting plate, and therefore no permanent deformation of the  
270 upper plate is produced under the steady state beyond the perturbation of the earthquake cycles,  
271 meaning the recurrence of the stress build-up and seismic slip. Figure 3b schematically illustrates  
272 this imposed steady state, where no deformation occurs within upper plate, and ignored the elastic  
273 property or complex asthenospheric flow beneath the plate interface. This model is a first-order  
274 approximation suitable only when the influence of forward slip on the entire plate interface is  
275 negligibly small. Theoretical analyses demonstrate that such a condition is attained only at a limit  
276 of flat interface geometry (or zero curvature) and uniform slip distribution (or zero slip gradient)  
277 (Romanet et al., 2020). Because the subduction interface is inevitably curved, the effect of the  
278 steady forward slip is accumulated in nature, contradicting the assumption of the back-slip model  
279 as pointed out by Matsu'ura and Sato (1989). Therefore, while it is applicable for problems like  
280 kinematic inversion of interseismic coupling, it cannot be used to model the long-term permanent  
281 deformations over multiple earthquake cycles.

### 282 3.2 Elastic Subducting Plate Model

283 Kanda and Simons (2010; 2012) proposed a subduction model in an elastic half space to  
284 overcome the problems with the back-slip model in treating the long-term deformation (Figure  
285 3c). This model, elastic subducting plate model (ESPM), assumes steady slips on the upper and  
286 lower interfaces of the subducting plate. ESPM considers the long-term effect of the steady  
287 forward slip, which was neglected in the back-slip model. Further, the imposed lower interface  
288 introduced the elasticity of the plate and the asthenospheric viscoelasticity with 2D and 3D  
289 structures. As a result of bending of the subducting plate, ESPM produces the long-term permanent  
290 vertical deformations on the surface with steady subduction (forward slip). Kanda and Simons  
291 (2010) explained that this deformation is caused by the strain accumulation within the subducting  
292 plate and will remain unless the flexural stresses are released by inelastic behavior.

293 ESPM is an advanced subduction model that accounts for long-term permanent deformation  
294 resulting from steady forward slip, a factor overlooked in the back-slip model. However, one of  
295 the boundary conditions employed in ESPM, namely the uniform slip amount on the entire plate  
296 boundary, might become a strong assumption depending on the geometry of subduction zones. In  
297 other words, there is no mechanical validation for the assumption that slip amount becomes  
298 uniform over time. For example, recent modeling studies of earthquake events have frequently  
299 utilized dynamic rupture simulations driven by stress drops rather than kinematic slips. These  
300 simulations have revealed that the resulting slip amount is markedly sensitive to fault alignment,  
301 even under the same initial stress conditions (e.g., Ando and Kaneko, 2018). Consequently, in  
302 long-term deformation scenarios, the slip amount is also likely influenced by local irregularities in

303 interface geometry, even when the same stress conditions are applied due to the large-scale  
304 configuration of subduction zones. The back-slip model targets a snapshot behavior during  
305 earthquake cycles and thus can disregard the inhomogeneity accumulated over a long period. In  
306 contrast, if the model considers a longer timescale involving multiple earthquake cycles, it should  
307 account for the non-uniform distribution of accumulated slip on the plate interface.

### 308 3.3 Multilayered Elastic/Viscoelastic Half-space Model

309 Besides these kinematic models that assumes an elastic half-space, crustal deformation  
310 accompanying plate subduction has also been modeled using elastic/viscoelastic layered models  
311 (Matsu'ura and Sato, 1989) (Figure 3d). This model has an advantage over ESPM in the treatment  
312 of the transient behavior of the bulk viscoelasticity due to the direct Maxwellian modeling of the  
313 asthenospheric viscoelasticity. Since the stress in the viscoelastic asthenosphere is relaxed after  
314 the Maxwell time of the viscoelastic relaxation (Fukahata and Matsu'ura, 2016), the lower  
315 boundary of the elastic lithosphere behaves like the free surface in the steady state without the  
316 transient behavior. This property engages for the validity of the slipping lower surface imposed in  
317 ESPM to model the asthenospheric behavior. Fukahata and Matsu'ura (2016) explored the  
318 mechanism of permanent deformation resulting from steady subduction in this elastic/viscoelastic  
319 model, confirming that vertical deformation arises from the interaction between lithosphere  
320 bending due to the curvature of the plate interface and gravitational compensation. However, due  
321 to the theoretical limitation, their viscoelastic structure is horizontally layered, unable to account  
322 for the 2-D or 3-D structure of the subducting plate that can be important to model the case of the  
323 Sagami Trough with significant geometrical irregularity.

### 324 3.4 Limitations of Previous Models and needs for Updating Models

325 As stated above, the previous studies of ESPM and the multilayered viscoelastic models  
326 revealed the steady forward slip or the steady plate subduction with the curved plate geometry is  
327 important to generate the permanent uplift. However, these models only considered the first-order  
328 scale of the subduction interface geometry with assuming the uniform slip rate. Their major  
329 limitations arise from that they did not account for stress changes induced by local irregularities  
330 along the plate boundary like a subducting seamount seen in the Sagami Trough (Tsumura et al.,  
331 2019). The local geometrical structures can generate shorter wavelength patterns of permanent  
332 uplift and stress changes along the plate interface. Such a local stress can modify the slip  
333 distribution on the plate interface, and the non-uniform slip can further contribute to form the uplift  
334 patterns, where the uniform slip distribution cannot be premised. In this study, we aim at exploring  
335 the underlying mechanism of the permanent uplift in the Sagami Trough subduction zone by  
336 focusing on the irregular geometry of the plate interface. We keep our model simple as possible  
337 but the previously introduced assumption of the uniform slip is not presupposed. Moreover, we  
338 test whether the inferred subducting seamount can quantitatively explain the spatial distribution of  
339 the long-term vertical displacement rate recorded in the Numa terraces.

## 340 4 Model Setting

### 341 4.1 Mechanical Subducting Plate Model

342 Both ESPM and the elastic/viscoelastic model, described in the previous section, demonstrated  
343 permanent deformations resulting from the curvature of the plate interface. However, these models  
344 assume uniform slip distribution on the plate interfaces for steady state and neglect the other source

345 of elastic deformation, such as slip gradient (Romanet et al., 2020). As demonstrated in the  
346 following investigation, their assumption is approximately valid with a sufficiently smooth plate  
347 interface geometry but is not when it has an irregular geometry with large curvatures. Therefore,  
348 this study proposes a new subduction model that can simulate the spatial changes in slip  
349 distribution due to the irregular geometry on the plate interface, extended from the previous  
350 subduction models.

351 The new subduction model, MSPM, first considers the average movement over a long time  
352 period and applies boundary conditions as slips and stress changes on the plate interfaces. For  
353 example, subduction models focusing on extended time periods, such as thermomechanical models  
354 utilizing finite elements, often assume the plate interface as a thin, plastically weak layer (Bessat  
355 et al., 2020). This layer is qualitatively a boundary unable to sustain shear stress. Therefore, we  
356 can employ a boundary condition that the accumulated shear stress on the plate interface is  
357 negligible compared with the total slip amount.

358 Subsequently, we simplify earthquake sequences for convenience. Namely, by assuming a  
359 constant recurrence interval and persistent rupture regions, the stress accumulation per one co- and  
360 interseismic sequence aligns with the average value of long-term accumulation, which is negligibly  
361 small. Of course, it is widely acknowledged that actual individual earthquake ruptures exhibit  
362 wide-ranging variations, and it should be noted that this assumption is relatively strong. There is  
363 room for discussion regarding how the interseismically accumulated stress is allocated to each  
364 individual rupture. However, at this moment, we aim to evaluate the average behavior of recurrent  
365 earthquakes.

366 Consequently, the subduction model proposed in this study uses shear stress as the boundary  
367 condition instead of slip deficit, which is accumulated during interseismic periods and reduced to  
368 the level of the sliding frictional strength at the coseismic timing. To compute the interseismic  
369 stress accumulation, we developed the mechanical subduction model, MSPM, based on the  
370 configuration of ESPM (Figure 3e), by replacing the displacement boundary condition of the upper  
371 interface to the stress boundary condition to consider the nonuniform distribution of slip rates. The  
372 lower interface of the subducting slab remains the same with that of ESPM, applying the uniform  
373 displacement rate. In other words, this model operates as a stress drop model reproduces coseismic  
374 slips that release an equivalent amount of shear stress accumulated during interseismic periods due  
375 to external force. The advantage of this model is that the effects of the irregular plate interface  
376 geometry is introduced to determine the spatial variation of the slip rate in a physically consistent  
377 manner.

378 Besides, this mechanical model is similar to a concept of smoothing used in the recent geodetic  
379 inversion methods to evaluate the interseismic coupling that identify coupling patches instead of  
380 kinematic slip deficits (Johnson and Segall, 2004; Johnson and Fukuda, 2010; Herman et al., 2018;  
381 Herman and Govers, 2020; Lindsey et al., 2021). Conventional geodetic inversions employ  
382 smoothing parameters over the slip distribution to obtain steady results. However, such a constraint  
383 was not physically validated and might have overlooked the potentially seismogenic fault (Lindsey  
384 et al., 2021). The mechanical constraint inversion detects coupling patches on the plate interface  
385 and predicts physically reasonable slip distributions. The mechanical model employed in this study  
386 also can simulate each coseismic slip and interseismic deformation considering coupling patches,  
387 not only the steady state.

## 388 4.2 Model Geometry and Boundary Conditions

389 Using these subduction models, illustrated in Figure 3, the deformation patterns due to steady  
390 plate subduction and the recurrence of earthquakes are investigated. As previously mentioned, the  
391 behavior of each model would be influenced by irregularities at the plate interface. Hence, we  
392 explore the impacts of 3D model geometry and difference in the boundary conditions by  
393 considering several cases of plate geometries. Initially, we focused on a simple subduction  
394 geometry to compare the characteristics of the different subduction models introduced earlier.  
395 Figure 4 provides a visualization of the model geometry in this study. The geometry consists of a  
396 uniform cross-sectional profile along the trench axis, with the inclusion of a conical bump  
397 representing a subducted seamount. Additionally, the bottom interface of the slab is set parallel to  
398 the upper interface and has a thickness of  $H$ . In order to minimize the influence of model  
399 boundaries, we extended these surfaces with a sufficient length, although they are not depicted in  
400 this figure.

401 The geometry of the subducted seamount plays a crucial role in this investigation. To assess  
402 the model's sensitivity to stress changes and displacements, we explore the dependency on  
403 seamount geometries from an unusually tall bump with a height of 8 km and a radius of 15 km to  
404 the typical height of real seafloor seamounts not exceeding 4 km (Wessel et al., 2010). The  
405 seamount is adopted on the interface at a depth of approximately 10 km, as shown in Figures 4a  
406 and b.

407 In the ESPM and MSPM, we employed the elasto-static boundary element method with the  
408 triangle dislocation element (TDE) (Nikkhoo and Walter, 2015; Thompson et al., 2023) to  
409 implement the slip on the plate interfaces in a discretized manner. This method enables us to  
410 calculate displacements and stress changes within the elastic half-space based on linear  
411 convolutions of the Green's function with the slip amount assigned to each TDE. To impose the  
412 stress boundary condition, we calculated the shear stress change on the slip surface for MSPM by  
413 evaluating the stress at the center point of each TDE.

414 In the elastic/viscoelastic model, we employed the program developed by Hashima et al. (2008;  
415 2014), which is based on the formulation by Fukahata and Matsu'ura (2005; 2006). This model is  
416 capable of calculating displacement due to a point source or a line source. Consequently, we  
417 employed a different meshing geometry from the ESPM and MSPM. For the simple subduction  
418 geometry, a uniform flat geometry along the  $y$ -axis using line sources is initially modeled.  
419 Subsequently, the bump geometry is simulated by incorporating point sources through the addition  
420 and subtraction of the bump and flat surfaces, as illustrated in Figure 4c. This superposition is  
421 made possible due to the linear relationship between displacement and slip amount in Fukahata  
422 and Matsu'ura's (2005; 2006) formulation. Because the slip within the asthenosphere has no effect,  
423 considering complete viscoelastic relaxation, the slip is only assigned to the upper interface above  
424 the lithosphere asthenosphere boundary (LAB).

425 As described above, if each earthquake is assumed to be an average behavior of multiple  
426 earthquake sequences, the residual resulting from asymmetry between inter- and coseismic  
427 deformations coincides to the long-term deformation pattern due to steady subduction. Therefore,  
428 we first examined the steady subduction model. In ESPM, we adopted the displacement boundary  
429 condition proposed by Kanda and Simons (2010) for the slip rates of the  $i$ -th element on the upper

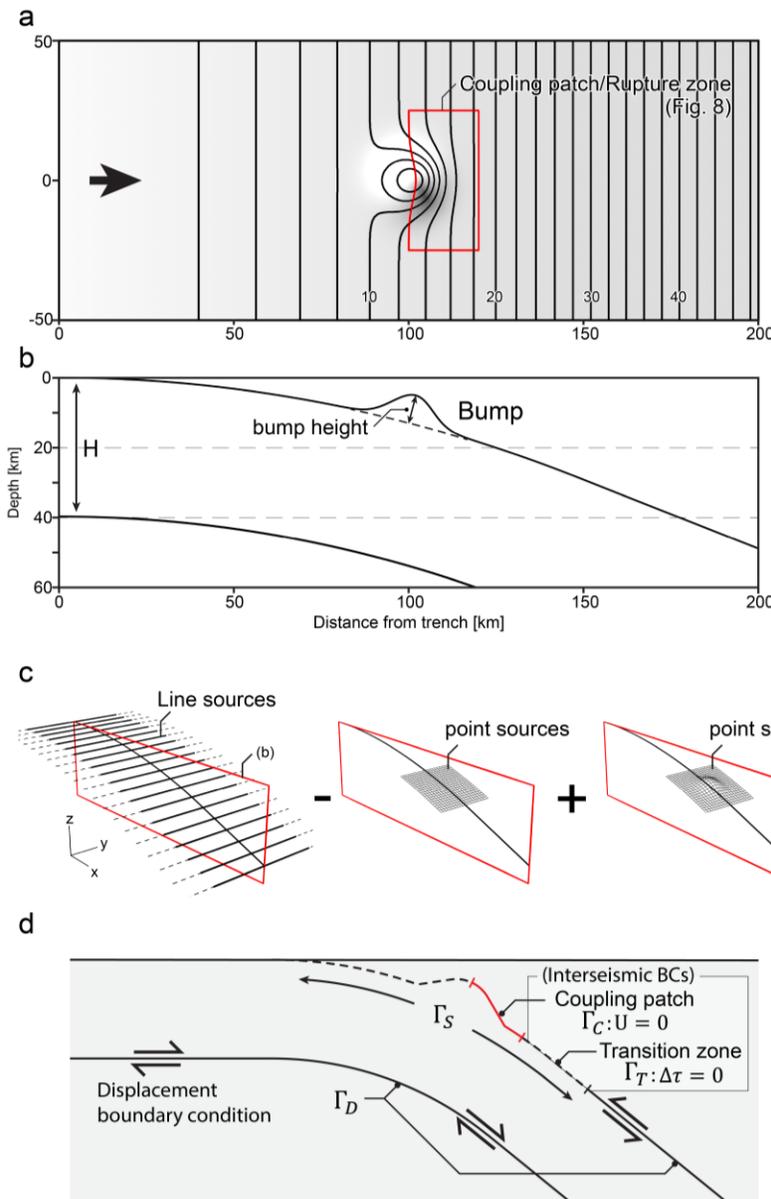


Figure 4. Geometry setting of the simple plate subduction model. (a) Plan view of the upper plate interface. The black lines indicate depth contours at 2 km interval. A conical-shaped bump with a height of 8 km is positioned at a depth of 10 km. The red rectangle indicates the rupture area and coupling patch in the earthquake sequence examination (Figure 8) (b) Cross-sectional view of the model geometry. The lower plate interface is set with a thickness  $H$  for ESPM and MSPM. (c) Schematic illustration of the superposition calculation used in the elastic/viscoelastic model. Refer to the main text for an explanation of this assumption. (d) Division of plate interfaces and boundary conditions in the earthquake sequence model using MSPM. The red and black broken lines correspond to the coupling patch and transition zone, respectively, applied during the interseismic period. The stress boundary condition is applied to the entire  $\Gamma_S$  during steady-state and coseismic events. The displacement boundary condition is applied to  $\Gamma_D$  during steady-state and the interseismic period.

430 and lower slab interfaces,  $V_{upper}^i$  and  $V_{lower}^i$ , respectively, where the uniform reverse and normal  
 431 slips with a rate of  $v_{pl}$  (mm/year) are imposed to be  $V_{upper}^i = v_{pl}$  and  $V_{lower}^i = -v_{pl}$  (Figure 3c).  
 432 Note that the lower interface mimics the LAB. The slip direction is parallel to the arrow depicted  
 433 in Figure 4, which is perpendicular to the trench axis.

434 For MSPM, we consider a mixed boundary condition given by displacement and stress rates  
 435 on the different areas of the plate interfaces. First, we define an area of interest on the upper  
 436 interface where the stress condition is calculated. This area of stress boundary condition (AOS) is  
 437 designated around the targeted geometry and coupling patches, denoted by  $\Gamma_S$  as depicted in Figure  
 438 4d. Outside of this AOS,  $\Gamma_D$ , a displacement boundary condition with a uniform slip rate is imposed  
 439 on the lower interface,  $V_{lower}^i = -v_{pl}$ , and on the upper interface,  $V_{upper}^i = v_{pl}$  when  $i \notin \Gamma_S$ . In

440 the AOS of the upper interface, we applied a stress boundary condition of the constant frictional  
 441 strength uniformly,  $\Delta\tau_S^i = 0$  when  $i \in \Gamma_S$  (Figure 3e); accordingly, the slip rate distribution in the  
 442 AOS can be linearly determined by the steady slip rate  $v_{pl}$ . The relationship between the vector  
 443 representing the shear stress change  $\Delta\tau_S$  in the AOS and the slip  $U_S (= U^i, i \in \Gamma_S)$  and  $U_D (=$   
 444  $U^i, i \notin \Gamma_S)$  on the inside and outside the AOS, respectively, are described as

$$\Delta\tau_S = G_{SS}U_S + G_{SD}U_D \quad (1)$$

445 Where  $G_{SS}$  and  $G_{SD}$  are the matrices representing the Green's functions calculated using  
 446 Thompson et al.'s (2023) code. The temporal differentiation of both side of the equations simply  
 447 gives the representation for the stress rate  $d\Delta\tau^i/dt$  and the slip rate  $V^i (= dU^i/dt)$  with the time-  
 448 independent Green's function,  $G$ . From the given boundary condition of the constant shear stress,  
 449 the stress boundary condition is reduced to  $d\Delta\tau_S^i/dt = 0$ . Thus, the slip rate distribution on the  
 450 shallower plate interface under the boundary conditions is linearly given by

$$V_S = -(G_{SS}^t G_{SS})^{-1} G_{SS}^t G_{SD} V_D, \quad (2)$$

451 where  $t$  denotes the transpose operation, and the indices denoting the number of elements are  
 452 omitted for a simple presentation. For calculation stability and reduction, we simplified the slip  
 453 direction and the calculation of shear stress change by considering only the direction parallel to  
 454 the subduction direction, regarding that the trench parallel component slip is negligible in a  
 455 relatively simpler geometry. For a more complex geometry, such as including branching fault, the  
 456 relaxation of this assumption would be needed.

457 In the elastic/viscoelastic model, we simulated the steady subduction by adopting the  
 458 configuration used in previous studies (Fukahata and Matsu'ura, 2016). The computation is  
 459 conducted using the viscoelastic boundary element method developed by Hashima et al. (2008;  
 460 2014). The steady state is approximated by considering the situation where viscoelastic relaxation  
 461 is completed. Consequently, we obtain the steady displacement and stress changes by applying  
 462 uniform slip to the entire upper plate interface to be  $V_{upper}^i = v_{pl}$  above the LAB at  $t = 0$ , after  
 463 enough time with zero rigidity in the asthenosphere, following the setting in Fukahata and  
 464 Matsu'ura (2016).

465 The structural parameters are given as shown in Table 1. In ESPM and MSPM, the structural  
 466 parameters in lithosphere are applied to the entire half-space.

#### 467 4.3 Earthquake Sequence Simulation

468 In addition to steady subduction, this study also explores an earthquake sequence using the  
 469 same subduction models. The geometry of the rupture region where uniform coseismic slip occurs  
 470 (in ESPM and the elastic/viscoelastic model) and the coupling patch (in MSPM) are defined  
 471 according to the configuration depicted in Figure 4a. We investigated how surface deformation  
 472 patterns change throughout the interseismic period depending on the subduction models.

473 The implementation of the earthquake sequence model using ESPM is straightforward. The  
 474 interseismic coupling zone, namely coseismic rupture zone, is set initially on the upper plate  
 475 interface, and uniform interseismic slip rate is assigned to the entire plate boundary, excluding this  
 476 coupling zone. An earthquake sequence is represented by a coseismic slip that releases the  
 477 accumulated slip deficit in the coupling zone.

478 In the elastic/viscoelastic model, coseismic slip is applied to designated rupture region at  $t =$   
 479  $0$ , and the post-seismic deformation or viscoelastic relaxation is taken into account. Displacements  
 480 caused by slip outside the rupture region can be treated as steady deformations with a fully relaxed  
 481 asthenosphere model, like the steady subduction model.

482 In the earthquake sequence model using MSPM, the AOS,  $\Gamma_S$ , is further divided into two parts;  
 483 the coupling patch,  $\Gamma_C$ , and the transition zone,  $\Gamma_T$  (Figure 4d). In the interseismic period, the slip  
 484 on the coupling patch is not allowed, i.e.,  $V_C^i = 0$  when  $i \in \Gamma_C$ , and the shear stress  $\Delta\tau_C$  is  
 485 accumulated there. The area surrounding the coupling patch steadily slip at a prescribed sliding  
 486 frictional strength,  $\Delta\tau_T^i = 0$  when  $i \in \Gamma_T$ , where the slip amount gradually increases without  
 487 accumulating shear stress there. The slip rate outside the AOS is uniform, same as the steady state.  
 488 Similarly in the case of steady state (equation 1), linear convolutions of the Green's function are  
 489 given by

$$\Delta\tau_C = G_{CC}U_C + G_{CT}U_T + G_{CD}U_D \quad (3)$$

$$\Delta\tau_T = G_{TC}U_C + G_{TT}U_T + G_{TD}U_D \quad (4).$$

490 Here, when the duration of the interseismic period is given by  $t_{cycle}$ ,  $U_C^i (= 0)$ ,  $U_D^i (=$   
 491  $\pm v_{pl}t_{cycle})$ , and  $\Delta\tau_T^i (= 0)$  are known, and therefore  $U_T$  can be linearly calculated using equation  
 492 4. Now, for equation 3, since we already know each slip distribution,  $U_C$ ,  $U_T$ , and  $U_D$ , the  
 493 accumulated shear stress on the coupling patch,  $\Delta\tau_C$ , is calculated straightforward.

494 At a seismic event, the coupling patch is allowed to slip to release the accumulated shear stress  
 495 during the interseismic period,  $\Delta\tau_C$ , while the shear stress change outside the coupling patch  
 496 persists zero,  $\Delta\tau_T^i = 0$  when  $i \in \Gamma_T$ . Combining the coupling patch and the transition zone into the  
 497 AOS again, the stress drop vector for a seismic event  $\Delta\tau_S$  is given by  $\Delta\tau_S^i = \Delta\tau_C^i$  when  $i \in \Gamma_C$  and  
 498  $\Delta\tau_S^i = 0$  when  $i \in \Gamma_T$ . Using the linear convolution of equation 1 and that the slip amount outside  
 499 the AOS at a seismic event is zero,  $U_D^i = 0$ , the coseismic stress distribution is calculated by  $U_S =$   
 500  $(G_{SS}^t G_{SS})^{-1} G_{SS}^t \Delta\tau_S$ .

#### 501 4.4 Crustal Deformation Simulation of the Sagami Trough

502 This study further investigates the crustal deformation distribution around the Sagami Trough,  
 503 simulating the observed plate interface geometry obtained from seismic surveys. The upper  
 504 interface geometry of PHS is created, as depicted in Figure 5, based on the observation results  
 505 presented in Figure 2. This simulation employs MSPM with the lower plate interface set to a  
 506 thickness of  $H = 40$  km. As the focus of this investigation is the effect of the subducted seamount  
 507 identified by Tsumura et al. (2009), the AOS is limited to the shallow part illustrated in Figure 5.  
 508 The displacement outside this region is constrained to be a uniform slip parallel to the subduction  
 509 direction, N30W, indicated by the arrow. Moreover, we simulate the coseismic and interseismic  
 510 deformations around the Sagami Trough by implementing a coupling patch, as depicted in Figure  
 511 5. The same method as in the previous section is applied to simulate earthquake sequences. This  
 512 allows us to evaluate the temporal deformation resulting from an earthquake sequence. The  
 513 structural parameters used in this simulation are the same as those used in the simple geometry  
 514 model (Table 1).

515

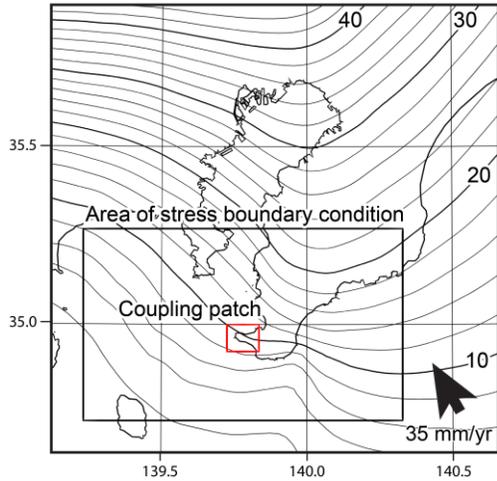


Figure 5. Geometry setting of the model simulation for the Sagami Trough subduction zone. The contour is the depth distribution of the upper plate interface of PHS, referring to Hashimoto et al. (2004), Hirose et al. (2008), and Tsumura et al. (2009). The black rectangle indicates the AOS, including a coupling patch for the earthquake sequence model, denoted by the red rectangle. Outside of the AOS is steady slip area, where uniform slip is imposed in the direction indicated by the arrow.

516

517 Table 1. Structural model

	h (km)	$\mu$ (GPa)	K (GPa)	$\eta$ (Pa·s)	$\rho$ (kg/m <sup>3</sup> )
Lithosphere	-	30	50	-	3000
Asthenosphere	40	50	90	$10^{19}$	3400

518 **5 Result**519 **5.1 Internal Stress Changes around the Interplate Bump**

520 Figure 6 presents the simulated distributions of deformation and stress change resulting from  
 521 steady subduction using different subduction models. The top, middle and bottom panels represent  
 522 the results obtained with ESPM (Kanda and Simons, 2010), the elastic/viscoelastic two layered  
 523 model (Fukahata and Matsu'ura, 2005; 2006), and MSPM (developed by this study), respectively.  
 524 In these figures, the displacement and the von-Mises stress change in the x-z plane are depicted  
 525 using arrows and color maps, respectively. In the elastic/viscoelastic model (Figures 6c and d), the  
 526 displacement is shown relative to the values obtained at a distant point from the subduction axis  
 527 in the hanging wall side. The arrows in the outer part of the subducting slab (bluish color) are  
 528 exaggerated by a factor of ten. Figures 6a, c, and e provide an overall view of the results, while  
 529 Figures 6b, d, and f offer closer views around the bump region. Figure 6g shows the slip amount  
 530 distribution on the fault using MSPM. In the case of ESPM and elastic/viscoelastic model, the slip  
 531 amounts are identical to the unit slip rate  $v_{pl}$  on the entire fault.

532 We can interpret the variations in the internal stress changes resulting from different employed  
 533 models. The stress changes resulting from steady subduction with a smoother plate interface, as  
 534 discussed by Kanda and Simons (2010) and Fukahata and Matsu'ura (2016), are insignificant  
 535 compared to the stress changes induced by the bump geometry introduced in this study. Figure 6  
 536 clearly demonstrates that noticeable stress changes occur around the bump geometry in all cases.  
 537 Note that in the elastic/viscoelastic model (Figures 6c and d), singularity values are observed  
 538 around the plate interface because this model employs point sources. The stress concentration  
 539 observed around the bump using ESPM is significantly larger than that using MSPM.

540

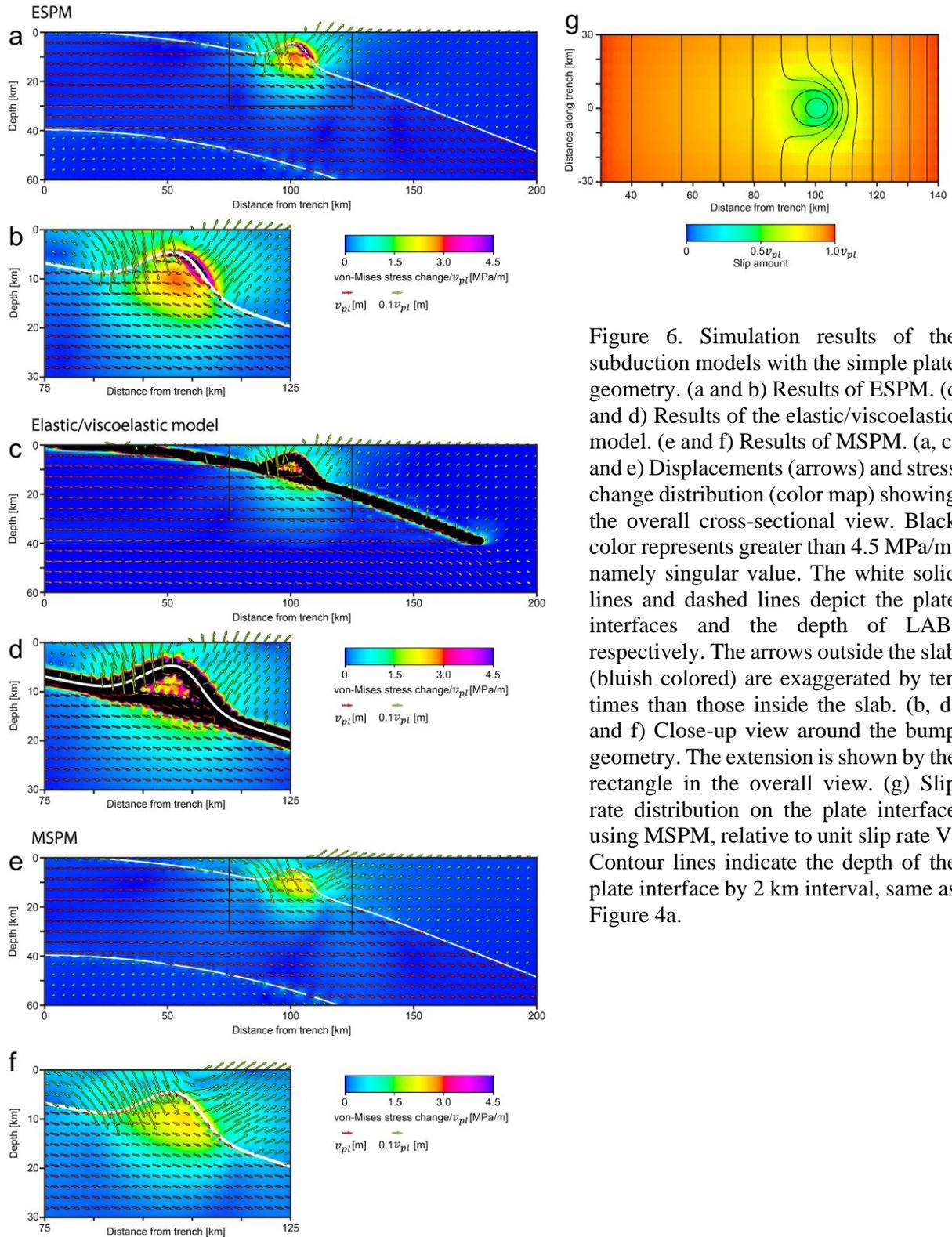


Figure 6. Simulation results of the subduction models with the simple plate geometry. (a and b) Results of ESPM. (c and d) Results of the elastic/viscoelastic model. (e and f) Results of MSPM. (a, c, and e) Displacements (arrows) and stress change distribution (color map) showing the overall cross-sectional view. Black color represents greater than 4.5 MPa/m, namely singular value. The white solid lines and dashed lines depict the plate interfaces and the depth of LAB, respectively. The arrows outside the slab (bluish colored) are exaggerated by ten times than those inside the slab. (b, d, and f) Close-up view around the bump geometry. The extension is shown by the rectangle in the overall view. (g) Slip rate distribution on the plate interface using MSPM, relative to slip rate  $V$ . Contour lines indicate the depth of the plate interface by 2 km interval, same as Figure 4a.

542 Figure 6g, showing the slip rate distribution around the subducted seamount calculated by  
 543 MSPM, indicates that the slip rate is lower in the vicinity of the seamount compared to the  
 544 surrounding areas. If a uniform slip rate  $V$  was applied in the entire area here, the result is identical  
 545 to ESPM. Hence, the essential difference between ESPM and MSPM is in this slip rate distribution.

## 546 5.2 Patterns of Surface Displacements

547 Figure 7a presents the permanent vertical surface displacements with each subduction model,  
 548 relative to the unit slip rate  $v_{pl}$ . If an averaged earthquake sequence is assumed, these patterns  
 549 coincide the residual resulting from asymmetry between inter- and coseismic deformations.  
 550 Specifically, permanent uplift is observed above the subduction side of the seamount (leading  
 551 flank), while subsidence occurs above its trench axis side (trailing flank). Although overall features  
 552 are comparable to each other, differences can be seen at the uplift peak. The peak uplift in MSPM  
 553 is more gradual and smaller than that in ESPM. The difference between ESPM and MSPM,  
 554 including displacements and stress changes, can be attributed to variations in slip distributions on

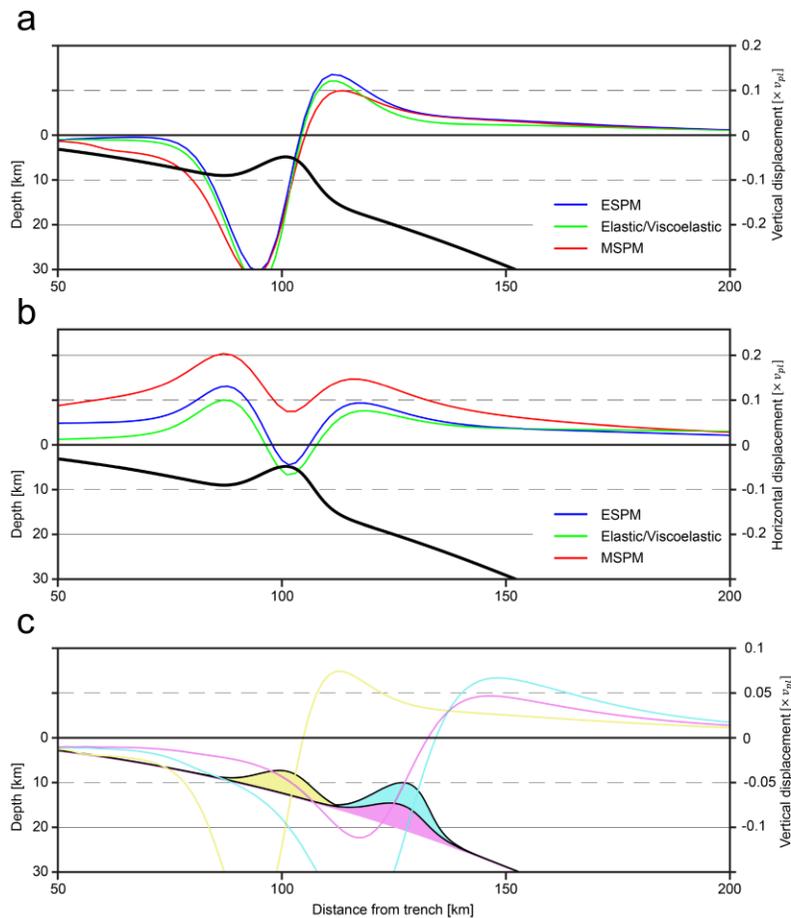


Figure 7. (a and b) Vertical and horizontal displacement distribution in each model. These results correspond to the arrows on the surface depicted in Figures 6 a, c, and e. Positive value indicates movements towards the subduction direction in (b). (c) Vertical displacement distributions with different bump geometries. The line colours correspond to the geometries of subducted seamounts. MSPM model was used for these simulations.

555 the fault around the bump geometry (Figure 6g). The slip becomes smaller when the bump exists,  
556 leading to the smaller vertical displacement; in other words, ESPM means to impose unrealistically  
557 large slip along the bump. The horizontal component of the displacements in MSPM shows a bulk  
558 movement towards the subduction direction, which reflects the dragging due to the stacked bump  
559 (Figure 7b). Although the back-slip model results are not presented in this figure, it is worth  
560 mentioning that in the back-slip model, the displacements and stress change in the hanging wall  
561 consistently remain zero, regardless of the plate interface geometry, resulting from the subducting  
562 slab exhibits smooth deformation attributed to asthenospheric flow, as described in Kanda and  
563 Simons (2010).

564 As a result, the previous subduction models that assign uniform slip distribution along the  
565 entire plate interface for steady state may not accurately capture the displacement distribution and  
566 stress concentration around the bump geometry because of the enforced slip distribution ignoring  
567 the slip direction. In contrast, MSPM effectively represents the movement of the subducted bump  
568 stacking towards the hanging wall (Figure 7b), implying a dragging movement, and helps alleviate  
569 stress concentration around the bump.

570 The short-wavelength permanent vertical deformation, which was not effectively explained  
571 with the conventional model setting, can be qualitatively explained by all the models depicted in  
572 Figure 7a. While the deformation patterns are similar between the models, it is important to note  
573 that the differences among the models become larger for rougher and more irregular geometries,  
574 which could impact the analysis aimed to understand the fault geometry effect on the geodetic and  
575 geological observations.

576 Figure 7c depicts the surface deformation using MSPM with different subducted bump  
577 geometries. The yellow, magenta, and cyan lines represent the permanent vertical deformations  
578 associated with shallow short, deep short, and deep tall bumps, respectively, as indicated in the  
579 bottom part of the figure. It can be observed that shallower and larger bump geometries result in  
580 greater amounts of permanent displacement. Furthermore, in the case of the shallower bump, the  
581 short-wavelength deformation is more pronounced. It is important to note that the estimation of  
582 subducted seamount geometry is challenging and subject to uncertainties, with potential errors of  
583 a few kilometers. This analysis underscores the potential impact of different assumptions regarding  
584 the bump geometry, leading to different expectations for surface deformation.

585 Figure 8 displays the results of the earthquake sequence simulations. This earthquake sequence  
586 assumes that the rupture occurs over the same rupture pattern with a constant interval  $t_{cycle}$  in  
587 each model. The results in Figure 8 are the vertical displacements relative to the total subduction  
588 amount  $v_{pl}t_{cycle}$ . The red and blue lines represent the coseismic and interseismic vertical  
589 deformation patterns, respectively. The green line represents the total vertical deformation pattern,  
590 which is identical to the result shown in Figure 6g. The yellow lines are the snapshots at every  $1/5$   
591  $t_{cycle}$ . In the deformation pattern at  $t = t_{cycle}$ , namely at the completion of an earthquake  
592 sequence, the shaded portion corresponds to ‘residual uplift,’ where uplifts occur in both the  
593 coseismic and long-term average deformations. This residual uplift leads to the formation of  
594 marine terraces that remain above sea level. It should be noted that the specific patterns of  
595 coseismic and interseismic deformation are influenced by the position and size of the rupture area.  
596 Therefore, Figure 8 serves as an example illustrating possible deformation patterns that can arise  
597 from an earthquake sequence.

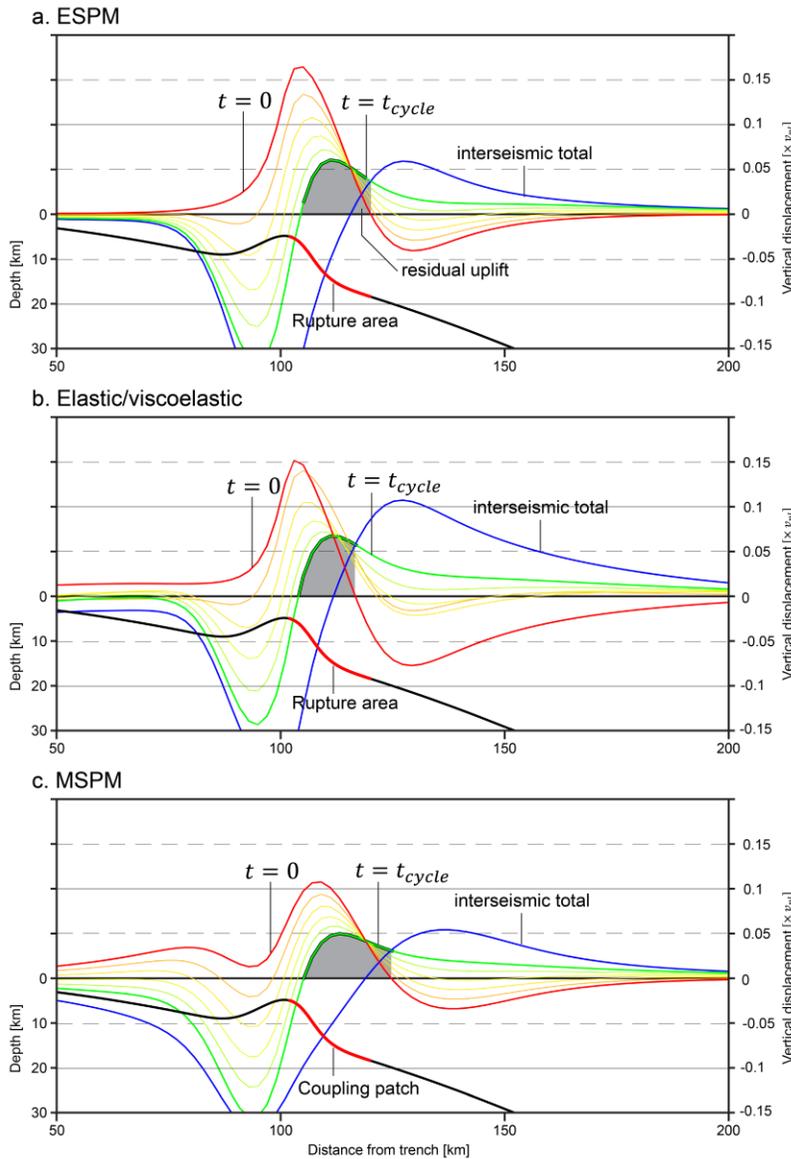


Figure 8. Transition of vertical displacements resulting from the earthquake sequence models. The red portion of the plate interface geometry indicates range of the rupture area (ESPM and Elastic/viscoelastic model) and coupling patch (MSPM), as shown in Figure 4. Red lines present the coseismic vertical deformation at  $t = 0$  and transits into the terminal deformation pattern at  $t = t_{cycle}$  depicted by the green lines. Yellow lines represent the snapshots of this transition at every  $1/5 t_{cycle}$ . The differences between red and green lines are interseismic total deformation, which is depicted by the blue lines. The shaded portions of the green lines indicate the residual uplift, where uplifts are observed both in coseismic and terminal deformation patterns.

598

599

### 5.3 Simulated Deformation Distribution of the Sagami Trough

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601

602

Figure 9 depicts the surface vertical deformation pattern using the geometry of the Sagami Trough subduction zone (Figure 5). In Figure 9a, the vertical deformation is shown for steady subduction (long-term average) using MSPM. Figure 9b illustrates the modeled coseismic vertical

603 deformation. The coseismic rupture is simulated by setting a coupling patch, shown as the red  
 604 rectangle in Figure 9b, that represents the southeastern coupling patch suggested by the results of  
 605 geodetic inversion (Sagiya, 2004; Noda et al., 2013), which is assumed as the main rupture portion  
 606 of the 1703 Genroku earthquake. The deformation amounts are expressed relative to the  
 607 convergence rate  $v_{pl}$  for the long-term deformation and total subduction amount  $v_{pl}t_{cycle}$  during  
 608 the interseismic period for the coseismic deformation, respectively. Figure 9c shows the  
 609 comparison between the simulated vertical displacement rate, as shown in Figure 9a, and the  
 610 observed elevation distribution of the highest paleo-shoreline, which indicates the sea level at the  
 611 Holocene highstand, compiled by Shishikura (2014). The observation points for the highest paleo-  
 612 shoreline are depicted in Figure 9a by red circles. For comparison, the amplitude of the simulated  
 613 vertical displacement rate is adjusted by assuming the convergence rate  $v_{pl}$  and the age of the  
 614 highest paleo-shoreline to be 35 mm/year (Seno et al., 1993) and 7,000 BP, respectively.  
 615 Moreover, considering the sea-level change after the Holocene highstand, the vertical  
 616 displacement is shifted by 5 meters.

617 As shown in Figure 9c, when considering the highest paleo-shoreline as indicative of long-  
 618 term deformation distribution, there is notable agreement between the observations and simulation  
 619 results on the eastern coast. In particular, the sharp decline in uplift rate from the southernmost tip

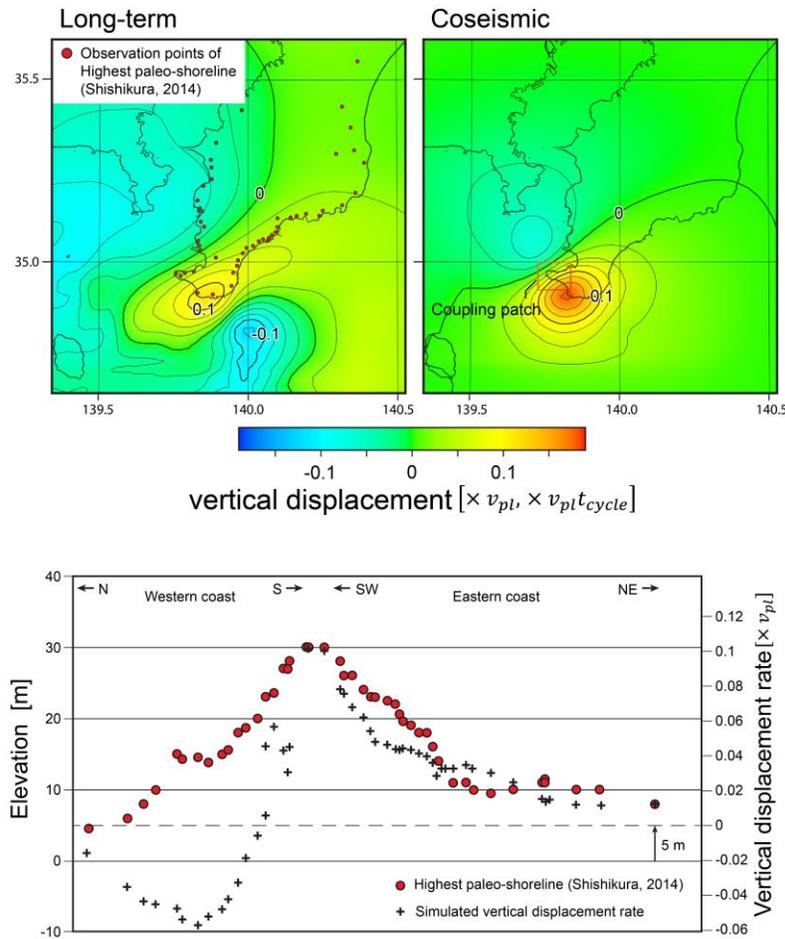


Figure 9. Simulated deformation distributions using MSPM with the model geometry of the Sagami Trough subduction zone. (a) Vertical displacement rate distribution with the steady-state assumption. (b) Coseismic vertical deformation distribution with the coupling patch representing the 1703 event, depicted by the red rectangle. (c) Comparison between the observed elevation distribution of the highest Holocene paleo-shoreline (Shishikura, 2014) and simulated vertical displacement rate. Observation points are displayed in (a). The amplitude of vertical displacement is calibrated assuming that the convergence rate and the age of the highest Holocene sea level are 35 mm/year and 7,000 BP. The vertical displacement is shifted by 5 meters reflecting the Holocene highstand.

620 towards the northeast, where the highest uplift rate closely corresponds to the observed data.  
621 However, on the western coast, while the deformation rate similarly decreases towards the north,  
622 it exhibits a long-term subsidence trend, contrary to the observed pattern. Figure 9b shows the  
623 coseismic vertical deformation amount at the southernmost tip of the Boso Peninsula, which is  
624 approximately  $0.18 v_{pl} t_{cycle}$ . Assuming that the rupture of this patch occurs every 2,000 years  
625 (Shishikura, 2014), the estimated coseismic uplift is approximately 5.4 m. This estimate is  
626 comparable to the observation of the marine terraces.

627 Following the simplification assumption of earthquake sequences, as proposed in the  
628 examination show in Figure 8, the residual deformation pattern after an earthquake sequence, i.e.,  
629 relative height of terraces, matches the long-term deformation pattern, as shown in Figure 9a.  
630 However, as discussed in previous papers (Komori et al., 2020; 2021), the formation pattern of  
631 marine terraces in this region is complex and cannot be explained solely by a simple periodic  
632 rupture. The results presented in this study demonstrate the effects of subducted seamounts and  
633 the capability of MSPM to explain the rupture and deformation history along the subduction zone.

## 634 **6 Discussion**

### 635 **6.1 Optimal Subduction Model for Irregular Plate Interface**

636 The uplift accumulation observed around subduction zones has been the subject of extensive  
637 research due to geological observations, such as marine terraces. However, previous kinematic  
638 models primarily assumed a smooth plate boundary interface and attributed permanent uplift to  
639 deformation within the subducted slab or asthenospheric flow. It has become evident through  
640 observation, including the sand-box experiments (Dominguez et al. 1998; 2000) and the coastal  
641 landform (Kodaira et al., 2000; Gardner et al., 2001), that irregularities in the plate interface  
642 strongly influence permanent deformation. Additionally, the impact of fault curvature on the  
643 deformation was emphasized in previous mechanical investigations (Romanet et al., 2020). Thus,  
644 it is crucial to incorporate the effects of these irregularities when evaluating crustal deformation  
645 resulting from plate subduction. However, previous kinematic models often employed a boundary  
646 condition imposing uniform slip across the entire plate interface, which leads to unrealistic mass  
647 flow in the elastic medium along curved slip surfaces when local irregular curvature exists. To  
648 address this limitation, we have developed a new mechanical model capable of simulating plate  
649 subduction with irregular bump-shaped interfaces and compared its performance to the previous  
650 models.

651 The stress boundary condition in this study's model, MSPM, with the constant sliding friction  
652 is physically more reasonable than imposing a uniform slip on the entire fault. As depicted in  
653 Figure 6, the stress accumulations resulting from steady subduction around the subducted bump  
654 exhibit significant differences between MSPM and the previous models, ESPM and the  
655 elastic/viscoelastic model. The previous models show large stress accumulation in both the upper  
656 and lower sides of the subducted bump, which is caused by the bending of dislocation surface in  
657 the elastic medium. In contrast, MSPM simulates gradually decreasing displacements within the  
658 subducted bump (Figure 6f). Qualitatively, this deformation resembles the dragging of the stacked  
659 bump. Consequently, the displacement field, which was artificially simulated by the slips on the  
660 plate interface, is expressed by the gradual shear deformation within the bump. As a result of this  
661 deformation distribution, the artificial stress concentration in the upper side of the bump was  
662 eliminated.

663 Despite the significant difference within the internal stress condition, the vertical deformation  
664 demonstrated a qualitatively common pattern throughout those subduction models, where  
665 permanent uplift and subsidence concentrations occur above the leading and trailing flanks of the  
666 subducted seamount, respectively. As demonstrated in Figure 7, the difference in deformation  
667 amounts between each model are significant when employing identical plate interface geometry.  
668 However, when this is applied to actual plate geometries, the observation error of plate interface  
669 depth could exceed several kilometers. Furthermore, observation values are frequently derived  
670 from geological studies, further challenging to minimize observation errors. Consequently, the  
671 imperative to differentiate these models in practice may be overshadowed by the estimation errors  
672 stemming from observational inaccuracies.

673 However, analyses that attribute model approximations to observation errors can lead to  
674 misunderstandings and incorrect assumptions, as they may obscure mechanical inconsistencies  
675 and force overfitting between observations and simulation results. This is why the back-slip model  
676 has been overused inappropriately in problems related to subduction zones, disregarding its first-  
677 order approximation. Of course, it should be noted that MSPM is also an approximate model, but  
678 a step-by-step process to reduce mechanical inconsistency is essential. Additionally, as  
679 demonstrated in Figure 8, mechanical boundary conditions offer advantages in simulating more  
680 realistic behaviors of coseismic slip and interseismic coupling. Therefore, for the analysis of  
681 subduction zone deformations within a timescale of 10 to 100 thousand years, where the movement  
682 of subducting bump itself can be ignored, we would recommend utilizing the model with the  
683 mechanical boundary condition.

684 In this study, MSPM does not incorporate viscoelastic relaxation in the asthenosphere, like  
685 ESPM where steady slip on the bottom interface of the slab is assumed to simulate asthenospheric  
686 flow. As a result, the isostatic compensation resulting from the gravitational effect, which was  
687 focused in Fukahata and Matsu'ura (2016), is not accounted for in MSPM. While this effect may  
688 have a characteristic wavelength comparable to the lithosphere thickness and could be less  
689 significant in our current interest of the local surface deformation distribution, it is still worth  
690 discussing for precise estimations, particularly in cases of extreme uplift and subsidence. Figure  
691 6e illustrates that the elastic/viscoelastic model exhibits singular values on the slip surface, making  
692 it unsuitable for the stress boundary conditions. On the other hand, models in the elastic half-space  
693 can be used to simulate the viscoelastic effect. The condition where stress in the asthenosphere is  
694 fully relaxed after enough time can be approximated by assuming the rigidity of the asthenosphere  
695 is zero, which behaves like water, as discussed in Fukahata and Matsu'ura (2016). Thus, the  
696 permanent deformation can be modeled by incorporating boundary conditions that the stress  
697 accumulation on LAB is zero, where uniform slip is applied to in MSPM. Moreover, this  
698 configuration allows for the simulation of complex geometries in the subduction zone, unlike the  
699 elastic/viscoelastic model, which simulates a horizontally layered half-space.

## 700 6.2 Remaining Uplift after Earthquake Sequence

701 The relationship between marine terrace distribution and coseismic uplifts has long been paid  
702 attention. First, the primitive back-slip model fails to explain the mechanism of permanent  
703 deformation and marine terrace formation because the coseismic uplift is canceled out by  
704 interseismic subsidence. Studies conducted in other subduction zones have attempted to verify  
705 whether coseismic deformation would be completely recovered by interseismic deformation and  
706 matches the long-term deformation pattern, based on historical and geological records (Briggs et  
707 al., 2008; Wesson et al., 2015). Most of these studies have not produced observations indicating

708 that coseismic deformation corresponded to (reversed) interseismic deformations or the long-term  
709 deformation pattern. Consequently, the asymmetry between co- and interseismic deformations has  
710 been widely accepted from observational studies, while it is possibly attributed to upper plate  
711 faulting.

712 In a study by Sato et al. (2016), the recent deformation around the Sagami Trough was  
713 simulated using an elastic/viscoelastic model. It was concluded that coseismic uplifts are negligible  
714 due to subsequent interseismic subsidence, and they are considered as a ‘perturbation’ within the  
715 long-term steady uplift. In this study, the perturbation caused by coseismic uplifts was  
716 quantitatively evaluated using three subduction models. It was observed that if both long-term and  
717 coseismic uplifts are significant, the coseismic uplifts never return to sea level throughout  
718 earthquake sequences. This condition may occur above the leading flank of a subducted seamount.

719 This study examined deformations throughout each earthquake sequence. However, the  
720 earthquake sequences analyzed here (as shown in Figure 8) are based on idealized average-type  
721 earthquakes occurring over extended periods, akin to so-called characteristic earthquakes. The  
722 issue would arise when considering variations in individual earthquake ruptures and their resulting  
723 deformation patterns. Nonetheless, the consistent explanation of marine terrace formation was  
724 successfully demonstrated in at least the end-member earthquake sequences. As a result, the  
725 rebuttal to the previous argument seeking to attribute the causes of marine terrace formation to  
726 eustatic sea-level fluctuations (Noda et al., 2018) has been achieved.

727 Traditionally, paleoseismological studies have often estimated the magnitudes of past  
728 earthquakes and compared their similarity based on the elevation distribution of the marine  
729 terraces. However, the findings of this study suggest that the remaining terrace distribution does  
730 not directly indicate the coseismic uplift distribution. While it is possible to estimate the minimum  
731 magnitude of past earthquakes based on the extent of terrace formation, as it requires a sufficient  
732 initial uplift amount, the similarity of terrace distribution alone cannot identify the rupture region  
733 and characteristic earthquakes. Therefore, for a precise estimation of past rupture history along  
734 subduction zones, the correction of the interseismic deformation is essential, and must be based on  
735 other geological and geophysical data.

### 736 6.3 Simulation of Geological Observations

737 In this study, the long-term vertical deformation distribution around the Sagami Trough was  
738 evaluated using MSPM and the depth distribution of the PHS obtained from recent seismic  
739 surveys. The results showed that the long-term deformation is primarily influenced by the  
740 geometry of the plate interface because the influence of coupling patch will be declined over time.  
741 Thus, the permanent vertical deformation depicted in Figure 9a can be attributed to the subducting  
742 plate geometry shown in Figure 5. The southernmost tip of the Boso Peninsula exhibited the  
743 highest uplift rate, reaching  $0.12 v_{pl}$ . This location corresponds to the area above the leading flank  
744 of a subducted seamount, as observed by Tsumura et al. (2009). Although there are uncertainties  
745 associated with the convergence rate and seamount geometry, this uplift rate is comparable to the  
746 long-term uplift rate estimated from the height of the Holocene highest marine terrace observed in  
747 the region (Shishikura, 2014) (Figure 9c). Additionally, the elevation distribution, which peaks at  
748 the southernmost tip, is consistent with this uplift rate. Therefore, it can be concluded that there is  
749 a considerable possibility that the long-term deformation of the Boso Peninsula is influenced by  
750 the presence of the subducted seamount.

751 However, the overall distribution of the permanent vertical deformation does not necessarily  
752 align with the geological observations. For instance, although the model predicts long-term

753 subsidence around the Miura Peninsula, geological evidence such as Holocene and Pleistocene  
754 marine terraces suggests an uplift trend in this area (Figure 9c). This subsidence trend in this model  
755 possibly arises from the curvature of the model geometry in the northwestern part. The accuracy  
756 of the depth distribution, particularly in the deeper part of the subduction zone, is highly uncertain  
757 and may contribute to the discrepancies observed in the long-term deformation distribution.  
758 Additionally, the western end of the Sagami Trough exhibits a complex plate boundary due to the  
759 collision of the Izu Peninsula, which deviates from a simple steady subduction scenario  
760 (Hashimoto and Terakawa, 2018). This collision may introduce complexities that cannot be  
761 captured by the subduction models used in this study. Consequently, the crustal deformation  
762 around the Izu Peninsula may not be accurately simulated using the subduction models employed  
763 here. As a result, the coverage of this study is currently limited to the southern part of the Boso  
764 Peninsula, where the influence of the collision is smaller and the resolution of the depth distribution  
765 of the subducting plate is higher.

766 Figure 9b presents the distribution of coseismic vertical deformation when a coupling patch,  
767 represented by the red rectangle, is considered. The uplift observed at the southernmost tip of the  
768 Boso Peninsula is approximately  $0.18 v_{pl} t_{cycle}$ . Assuming a rupture interval of 2,000 years for  
769 this specific coupling patch, the estimated coseismic uplift is consistent with the observed  
770 elevation of the Genroku terrace, where the maximum elevation is approximately 7 m. It is  
771 important to note that the chosen rupture recurrence in this analysis is a subjective forward model  
772 and may not represent the actual recurrence pattern. However, this result suggests that the MSPM  
773 model has the potential to simulate realistic terrace formation, indicating its capability in capturing  
774 essential aspects of the process.

775 The formation history of the Numa terraces and the rupture history of the Sagami Trough  
776 require a more detailed and thorough discussion, taking into account the complexities observed in  
777 previous studies (Komori et al., 2020; 2021). These studies have shown that the formation intervals  
778 and relative heights of the Numa terraces are not consistent with each other, indicating a more  
779 complex pattern of terrace formation. Additionally, the rupture interval of 2,000 years, based on  
780 the terrace formation ages, is much longer than the typical recurrence interval for subduction  
781 earthquakes. This discrepancy strongly suggests that the rupture pattern of the Sagami Trough is  
782 not periodic and does not occur in the same region each time. To fully understand the rupture  
783 scenario of the Kanto earthquakes and provide a comprehensive explanation for the formation  
784 history, it is essential to employ a physically consistent model that considers coseismic,  
785 interseismic, and long-term deformations. The MSPM used in this study is well-suited for this  
786 purpose as it allows for the simulation of rupture recurrence that considers the accumulation and  
787 release of stress.

788 In addition to the effects of subducted seamounts, we cannot yet eliminate other potential  
789 sources of deformation within the hanging wall of subduction zones. Inelastic faulting, including  
790 splay faults and upper plate faults, can occur due to the compression stress field associated with  
791 plate subduction. These faulting events can contribute to the overall surface displacement field and  
792 result in complex deformation patterns (e.g., Hikurangi subduction margin, as discussed in Clark  
793 et al. (2017)). Analogue experiments conducted by Dominguez et al. (1998; 2000) have  
794 demonstrated that strain accumulation within the hanging wall caused by seamount subduction can  
795 be released through inelastic deformation.

796 Although major faults have not been identified in the Sagami Trough region based on seismic  
797 surveys, several studies have suggested the possibilities of such movements (Pollitz et al., 1996).  
798 In this context, model examinations and investigations into the possibility of inland faulting can

799 be valuable. The MSPM used in this study is a suitable tool for evaluating the stress conditions  
800 within the plates and can provide insights into the potential mechanisms and effects of inelastic  
801 faulting in the subduction zone. By considering multiple deformation sources and incorporating  
802 various geological and geophysical observations, a more comprehensive understanding of the  
803 deformation processes in the study region can be obtained.

## 804 **7 Conclusion**

805 This study examined the formation of uplifted marine terraces around subduction zones,  
806 namely residuals resulting from asymmetry between inter- and coseismic deformations, focusing  
807 on the impact of plate interface irregularities. Because existing subduction models have implicitly  
808 assumed a smooth plate interface geometry, we first discussed the mechanical behavior around a  
809 bump on a plate interface and appropriate boundary conditions for such problems. The models  
810 utilized in this study differ in their approach to simulating plate subduction. ESPM and the  
811 elastic/viscoelastic model employ a uniform slip distribution on the plate interface, while MSPM  
812 imposes the constraint that the shear stress change should be net zero. Additionally, the  
813 elastic/viscoelastic model incorporates stress relaxation within the asthenosphere using a two-  
814 layered half-space model, whereas in ESPM and MSPM, the uniform slips on the bottom interface  
815 of the slab account for this movement. To examine the behavior of these models, a simple plate  
816 interface geometry with a bump shape was considered. The results showed that all three models  
817 were capable of producing localized uplift above the leading flank of the subducted seamount.  
818 However, there were notable differences in the displacement distribution within the crust. MSPM  
819 exhibited a more gradual displacement distribution compared to ESPM and the elastic/viscoelastic  
820 model. This difference arises from the extraordinary stress concentration that occurs when  
821 enforcing uniform slip on the bump in the models. In contrast, MSPM avoids such concentration  
822 by constraining the shear stress change to zero. Based on these findings, MSPM is considered to  
823 be a suitable model for simulating plate subduction when the plate interface exhibits local  
824 irregularities, such as a subducted seamount.

825 The analysis of vertical deformation around the subducted seamount revealed that it can play  
826 a crucial role in the formation of coastal landform, with larger vertical deformation than previously  
827 explained by the bending of the subducting plate. The patterns of permanent surface deformation  
828 slightly differ among the models used, but these differences are less significant compared to the  
829 variation caused by the size and geometry of the subducted seamount. Although it is currently  
830 challenging to directly validate the appropriateness of the models based on geological and seismic  
831 observations, it can be inferred that the significance of the subducted seamount in the deformation  
832 process is independent of the specific subduction model employed. In other words, regardless of  
833 the model used, the presence and characteristics of the subducted seamount have a substantial  
834 impact on the resulting deformation patterns and cannot be ignored.

835 The formation mechanism of marine terraces has been a subject of interest in understanding  
836 the recurrence of past earthquakes. Using the basic back-slip model, the coseismic uplifts are  
837 eliminated by subsequent interseismic coupling, making it difficult to explain the formation of  
838 marine terraces. Previous modeling studies using elastic/viscoelastic layered half-space models  
839 also suggested that individual earthquake sequences cannot generate sufficient remaining uplift to  
840 form marine terraces. However, this study demonstrates that the presence of subducted seamount  
841 can contribute to the coseismic and long-term uplifts, which provides a plausible mechanism for  
842 marine terrace formation through coseismic deformation. It should be noted that the correlation  
843 between the remaining deformation (i.e., relative heights of marine terraces) and the distribution

844 coseismic uplifts may not always be straightforward. Therefore, to accurately estimate the past  
845 rupture history from the present distribution of marine terraces, it is essential to carefully evaluate  
846 the interseismic deformation and employ a physically consistent model of rupture recurrence.

847 This study investigated the long-term deformation and coseismic uplift on the Boso Peninsula  
848 by using the observed geometry of PHS through seismic surveys. The long-term deformation  
849 pattern correlates the residual resulting from asymmetry between co- and interseismic  
850 deformations, namely the elevation distribution of Holocene marine terraces. The presence of a  
851 subducted seamount beneath the southern part of the Boso Peninsula, as indicated by the seismic  
852 survey conducted by Tsumura et al. (2009), was taken into account in the modeling. The  
853 employment of subducted seamount geometry led an uplift concentration at the southernmost tip  
854 of the Boso Peninsula. The simulated uplift rate was consistent with the estimated long-term uplift  
855 rate derived from the height of the Holocene highest terrace in this region. Furthermore, by  
856 incorporating coupling patches based on geodetic observations, the model also simulated a  
857 concentration of coseismic uplift at the southernmost tip of the Boso Peninsula, which corresponds  
858 to historical records.

859 The observation of the Numa terraces in the Boso Peninsula, with irregular formation intervals  
860 despite comparable relative heights, highlights the complexity of the rupture history along the  
861 Sagami Trough. It indicates that a more comprehensive rupture scenario is needed to explain the  
862 geological and geodetic observations, including marine terrace distribution, displacements of  
863 historical earthquakes, and present deformation observation from GNSS.

864 The verification of this study demonstrated significant differences in internal mechanical  
865 consistency between MSPM and conventional models. However, when compared to surface  
866 observations, the variations were negligible compared to the observational errors. Nevertheless,  
867 compared to the traditional first-order approximation approach, which unconditionally assigns  
868 uniform slip on the plate interface, the use of MSPM would reduce the potential for  
869 misunderstandings in interpreting deformations and movements in subduction zones. Furthermore,  
870 MSPM can reproduce more realistic behaviors in simulations of interseismic coupling and  
871 coseismic ruptures (Herman and Govers, 2020; Lindsey et al., 2021), without increasing the  
872 number of free parameters. Therefore, the utilization of the MSPM model would be recommended  
873 for interpreting future short-term deformations in subduction zones.

874 The distinction between MSPM and the elastic/viscoelastic model lies in their treatment of  
875 viscoelastic relaxation within the asthenosphere. MSPM assumes an elastic half-space and does  
876 not explicitly simulate viscoelastic relaxation, whereas the elastic/viscoelastic model incorporates  
877 viscoelastic behavior. While the elastic/viscoelastic model allows for a more realistic  
878 representation of the asthenosphere's viscoelastic relaxation, it faces limitations inaccurately  
879 simulating subduction with irregular plate interface geometries because it cannot directly calculate  
880 the stress change on the slip surface. In contrast, MSPM has an ability to approximate complete  
881 relaxation of stress in the asthenosphere over time by imposing a boundary condition that enforces  
882 zero stress accumulation on LAB. This modeling approach, which accommodates complex  
883 subduction geometries, offer as optimal combination of the models used in this study, unlike the  
884 horizontal two-layered model.

885 Despite the long efforts to understand the earthquake recurrence history through the analysis  
886 of vertical deformation recorded in coastal landforms, model explanations have faced challenges  
887 in encompassing observations at various scales. Specifically, the relationship between Holocene  
888 marine terraces and coseismic uplifts may have been overestimated due to their apparent  
889 correlation. The findings of this study have shed light on the significant influence of subducted

890 seamounts on permanent deformation around subduction zones, prompting a reevaluation of the  
891 interpretation of marine terrace distributions. It has become evident that marine terraces are  
892 influenced not only by coseismic uplift but also by interseismic and long-term deformations, which  
893 necessitates a proper assessment of the subduction mechanism and plate interface geometry in  
894 order to infer the past rupture history accurately.

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1115 **Figure 1.** Survey region of this study and geological observations. (a) Tectonic setting of the  
 1116 Sagami Trough. The red meshed area indicates the estimated source region of the historical  
 1117 Kanto earthquakes (Sato et al., 2005; Sato et al., 2016). (b) Distribution of the Numa terraces  
 1118 after Komori et al. (2020). Right panels show the elevation distribution of the Numa terraces at  
 1119 each reference point, indicated by triangles in the left map.

1120 **Figure 2.** (a) Bathymetry map around the survey region and the profile lines of the previous  
 1121 reflection surveys (Sato et al., 2005; Kimura et al., 2009; Miura et al., 2009; Tsumura et al.,  
 1122 2009). The blue contour lines indicate the estimated depth of upper PHS by Tsumura et al.  
 1123 (2009), where the dark-blue straight lines are the survey profiles. The red lines indicate the  
 1124 inland active faults, where KTFZ stands for Kamogawa-teichi fault zone. (b) Post stack time  
 1125 migrated reflection image of the BOS-1C profile (Miura et al., 2009). Solid black line is our  
 1126 interpretation of the plate interface. Triangles indicate the positions of intersection with the  
 1127 survey lines of Tsumura et al. (2009).

1128 **Figure 3.** Schematic illustration of subduction models. (a) General geometrical setting of plate  
 1129 subduction. (b) Schematic representation of the imposed steady state assessed from the back-slip  
 1130 model, following the interpretation by Kanda and Simons (2010). (c) Slip configuration for the  
 1131 steady state of ESPM (Kanda and Simons, 2010). Uniform slip is imposed on the entire plate  
 1132 interfaces (double arrows). (d) Schematic structure illustration of the elastic/viscoelastic model.  
 1133 Uniform slip is imposed on the plate interface above the LAB. (e) Boundary conditions of  
 1134 MSPM. Black solid arrows and white arrows indicate the interfaces where uniform slip is  
 1135 imposed ( $\Gamma_D$ : area of displacement boundary condition) and no shear stress change occurs  
 1136 ( $\Gamma_S$ : area of stress boundary condition), respectively.

1137 **Figure 4.** Geometry setting of the simple plate subduction model. (a) Plan view of the upper  
 1138 plate interface. The black lines indicate depth contours at 2 km interval. A conical-shaped bump

1139 with a height of 8 km is positioned at a depth of 10 km. The red rectangle indicates the rupture  
 1140 area and coupling patch in the earthquake sequence examination (Figure 8) (b) Cross-sectional  
 1141 view of the model geometry. The lower plate interface is set with a thickness  $H$  for ESPM and  
 1142 MSPM. (c) Schematic illustration of the superposition calculation used in the elastic/viscoelastic  
 1143 model. Refer to the main text for an explanation of this assumption. (d) Division of plate  
 1144 interfaces and boundary conditions in the earthquake sequence model using MSPM. The red and  
 1145 black broken lines correspond to the coupling patch and transition zone, respectively, applied  
 1146 during the interseismic period. The stress boundary condition is applied to the entire  $\Gamma_S$  during  
 1147 steady-state and coseismic events. The displacement boundary condition is applied to  $\Gamma_D$  during  
 1148 steady-state and the interseismic period.

1149 **Figure 5.** Geometry setting of the model simulation for the Sagami Trough subduction zone. The  
 1150 contour is the depth distribution of the upper plate interface of PHS, referring to Hashimoto et al.  
 1151 (2004), Hirose et al. (2008), and Tsumura et al. (2009). The black rectangle indicates the area of  
 1152 stress boundary condition, including a coupling patch for the earthquake sequence model,  
 1153 denoted by the red rectangle. Outside of the AOS is steady slip area, where uniform slip is  
 1154 imposed in the direction indicated by the arrow.

1155 **Figure 6.** Simulation results of the subduction models with the simple plate geometry. (a and b)  
 1156 Results of ESPM. (c and d) Results of the elastic/viscoelastic model. (e and f) Results of MSPM.  
 1157 (a, c, and e) Displacements (arrows) and stress change distribution (color map) showing the  
 1158 overall cross-sectional view. Black color represents greater than 4.5 MPa/m, namely singular  
 1159 value. The white solid lines and dashed lines depict the plate interfaces and the depth of LAB,  
 1160 respectively. The arrows outside the slab (bluish colored) are exaggerated by ten times than those  
 1161 inside the slab. (b, d, and f) Close-up view around the bump geometry. The extension is shown  
 1162 by the rectangle in the overall view. (g) Slip rate distribution on the plate interface using MSPM,  
 1163 relative to unit slip rate  $V$ . Contour lines indicate the depth of the plate interface by 2 km  
 1164 interval, same as Figure 4a.

1165 **Figure 7.** (a and b) Vertical and horizontal displacement distribution in each model. These  
 1166 results correspond to the arrows on the surface depicted in Figures 6 a, c, and e. Positive value  
 1167 indicates movements towards the subduction direction in (b). (c) Vertical displacement  
 1168 distributions with different bump geometries. The line colours correspond to the geometries of  
 1169 subducted seamounts. MSPM model was used for these simulations.

1170 **Figure 8.** Transition of vertical displacements resulting from the earthquake sequence models.  
 1171 The red portion of the plate interface geometry indicates range of the rupture area (ESPM and  
 1172 Elastic/viscoelastic model) and coupling patch (MSPM), as shown in Figure 4. Red lines present  
 1173 the coseismic vertical deformation at  $t = 0$  and transits into the terminal deformation pattern at  
 1174  $t = t_o$  depicted by the green lines. Yellow lines represent the snapshots of this transition at every  
 1175  $1/5 t_o$ . The differences between red and green lines are interseismic total deformation, which is  
 1176 depicted by the blue lines. The shaded portions of the green lines indicate the residual uplift,  
 1177 where uplifts are observed both in coseismic and terminal deformation patterns.

1178 **Figure 9.** Simulated deformation distributions using MSPM with the model geometry of the  
 1179 Sagami Trough subduction zone. (a) Vertical displacement rate distribution with the steady-state  
 1180 assumption. (b) Coseismic vertical deformation distribution with the coupling patch representing

1181 the 1703 event, depicted by the red rectangle. (c) Comparison between the observed elevation  
1182 distribution of the highest Holocene paleo-shoreline (Shishikura, 2014) and simulated vertical  
1183 displacement rate. Observation points are displayed in (a). The amplitude of vertical  
1184 displacement is calibrated assuming that the convergence rate and the age of the highest  
1185 Holocene sea level are 35 mm/year and 7,000 BP. The vertical displacement is shifted by 5  
1186 meters reflecting the Holocene highstand.