Title: Weak faults at megathrust plate boundary respond to tidal stress

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Author #1: Takashi Tonegawa, Japan Agency for Marine-Earth Science and Technology (JAMSTEC), 2-15, Natsushima-cho, Yokosuka, 237-0061, Japan,
tonegawa@jamstec.go.jp

Author #2: Toshinori Kimura, Japan Agency for Marine-Earth Science and Technology (JAMSTEC), 2-15, Natsushima-cho, Yokosuka, 237-0061, Japan,
kimurat@jamstec.go.jp

Author #3: Kazuya Shiraishi, Japan Agency for Marine-Earth Science and Technology (JAMSTEC), 2-15, Natsushima-cho, Yokosuka, 237-0061, Japan,
kshiraishi@jamstec.go.jp

Author #4: Suguru Yabe, National Institute of Advanced Industrial Science and Technology, 1-1-1, Umezono, Tsukuba, Ibaraki, 305-8560, Japan, s.yabe@aist.go.jp

Author #5: Yoshio Fukao, Japan Agency for Marine-Earth Science and Technology (JAMSTEC), 2-15, Natsushima-cho, Yokosuka, 237-0061, Japan,
Author #6: Eiichiro Araki, Japan Agency for Marine-Earth Science and Technology (JAMSTEC), 2-15, Natsushima-cho, Yokosuka, 237-0061, Japan,
araki@jamstec.go.jp

Author #7: Masataka Kinoshita, The University of Tokyo, 1-1-1, Yayoi, Bunkyo, 113-0032, Japan, masa@eri.u-tokyo.ac.jp

Author #8: Sanada Yoshinori, Japan Agency for Marine-Earth Science and Technology (JAMSTEC), 2-15, Natsushima-cho, Yokosuka, 237-0061, Japan,
sanada@jamstec.go.jp

Author #9: Seiichi Miura, Japan Agency for Marine-Earth Science and Technology (JAMSTEC), 2-15, Natsushima-cho, Yokosuka, 237-0061, Japan,
miuras@jamstec.go.jp

Author #10: Yasuyuki Nakamura, Japan Agency for Marine-Earth Science and Technology (JAMSTEC), 2-15, Natsushima-cho, Yokosuka, 237-0061, Japan,
yasu@jamstec.go.jp

Author #11: Shuichi Kodaira, Japan Agency for Marine-Earth Science and Technology
(JAMSTEC), 2-15, Natsushima-cho, Yokosuka, 237-0061, Japan,

kodaira@jamstec.go.jp

Indicate the corresponding author

T. Tonegawa (tonegawa@jamstec.go.jp)
Abstract

Lateral spatial variations of weak portions at the plate boundary in subduction zones have been estimated primarily by the distribution of slow earthquakes mainly occurring around seismogenic zones. However, the detailed depth profile of weak faults remains elusive. Here, we deployed 6 ocean bottom seismometers in the Nankai subduction zone, Japan, to observe reflections originated from drilling vessel Chikyu ship noise (hydroacoustic P wave) that was persistently radiated from a fixed position at the sea surface, and retrieved P-to-s (Ps) reflections from multiple dipping faults near the plate boundary. The Ps amplitudes were stacked and compared according to the degrees of tidal stresses, and they were large at high tide (compression). A migration technique shows that the locations where velocity contrasts fluctuate were estimated at both the megasplay fault and another fault between the megasplay fault and the top of the oceanic crust. This indicates that the physical properties of those faults are changed by tidal stress. The physical-property changes are attributed to fluid connections and isolations within the faults due to tidal stress fluctuations, inducing the variation of seismic
anisotropy. Such a variation was confirmed by a three-dimensional numerical simulation for wave propagation with anisotropic medium. Our observation suggests that multiple weak faults are present around the plate boundary, and the obtained changes of fault physical properties may have implications for our understanding of tidal triggering of earthquakes.

Keywords

Ambient noise, Seafloor observation, Ps reflection, Ship noise, Tidal response, Megasplay fault

Main Text

Introduction

Tidal stresses result from the deformation of the Earth and ocean loading due to the gravitational pull of the Sun and Moon, and these can trigger ordinary (Cochran et al. 2004; Tanaka et al. 2006; Ide et al. 2016) and slow earthquakes (Rubinstein et al. 2008; Nakata et al. 2008; Lambert et al. 2009; Ide 2010) by stress fluctuations on plate
boundaries in subduction zones. The tidal sensitivity of slow earthquakes is affected by
the fluids within faults (Hawthorne and Rubin 2010; Houston 2015; Yabe et al. 2015;
Nakamura and Kakazu 2017). Under wetter conditions, the physical properties of the
plate boundary and thereby its frictional properties are presumably affected by external
periodic forces. However, there has been little observation on the tidal response of
plate-boundary properties because it requires continuous natural or artificial seismic
sources as well as persistent retrieval of seismic waves sampling the boundary. The
discovery of such phenomena could allow for the identification of mechanically weak
faults that are closely linked to rupture propagations of megathrust earthquakes and
tsunami generation, and this could be the key to understanding tidal triggering of
earthquakes.

A scientific drilling project was carried out between October 2018 and March
2019 by the International Ocean Discovery Program (IODP) with the drilling vessel
D/V Chikyu (hereafter Chikyu) (IODP expedition 358) (Fig. 1) at the C0002 site located
off the Kii Peninsula in the Nankai subduction zone, Japan, beneath which recurring
slow slip events have occurred around the plate boundary (Araki et al. 2017). Because
the *Chikyu* has a potential to excite incessant seismic sources during the project, we installed 6 ocean bottom seismometers (OBSs) at distances ranging between 100–800 m from the drilling site in the period from November 2018 to May 2019 to retrieve seismic reflections beneath the seafloor originated from the *Chikyu*.

Seismic interferometry is capable of detecting waves propagating between two receivers by cross-correlating the ambient noise recorded at both the receivers (Shapiro 2005; Brenguier et al. 2007). Applying this technique to continuous seismic records observed at a near-vertical array of OBS04 and a borehole sensor (Kopf et al. 2011; Kopf et al. 2016) (Fig. 1b), we found that the downgoing *P* waves generated by the *Chikyu* ship noise were dominant during the non-drilling period, while upgoing *S* waves from the drill-bit torque at depth could be observed in the drilling period (See details in Materials and Methods section): *Chikyu* produces hydroacoustic *P* waves in the sea water by, e.g., internal instrumental and external thruster noises, from the fixed drilling position at the sea surface, and they are converted to downgoing *P* waves at the seafloor. Therefore, we examined the *Chikyu* ship-noise records acquired by the 6 OBSs during the non-drilling periods of the project, in order to persistently retrieve the
seismic waves reflected from downgoing $P$ waves at the deep structure around the plate boundary.

In the Materials and Methods section, we confirm the wavefield near the seafloor during non-drilling period with seafloor and borehole sensors, correct the amplitudes of the 6 OBSs with site amplifications, and extract reflections through a deconvolution technique from both airgun shot experiments conducted by two vessels and Chikyu ship noise. In the Results section, we compare the amplitude variations of reflections in the deconvolved waveforms using the calculated tidal stresses, and image the locations, where the effects of tidal stress can be observed, through a migration technique. We finally discuss a possible model that explains reflection amplitudes induced by tidal stresses.

Materials and Methods

Ambient noise wavefield during the drilling period

Six ocean-bottom seismometers (OBSs) were installed at a range of distances between 100 m and 800 m from C0002. Four OBSs were placed at 100, 100, 200 and
250 m by the remotely operated vehicle (ROV) installed on D/V Chikyu and the two remaining OBSs were dropped in free-fall from R/V Yokosuka at 700 and 800 m from the drilling site; their location errors are ~2 m and 10–20 m, respectively.

To investigate the ambient noise wavefield during the drilling project, we calculated the cross-correlation function (CCF) of continuous records observed by OBS04 and a borehole sensor that is deployed at a depth of 900 m beneath the seafloor (Figs. 1b and 2a) (Kopf et al. 2011; Kopf et al. 2016). The OBS contains a 4.5 Hz short-period sensor with three components. The length of time window was 80 s. Spectral whitening was performed during the calculation of the CCF (Brenguier et al. 2007). Because we have the rotation speed information of the drill-bit (RPM: rotation per minute) in the drilling vessel (D/V) Chikyu (hereafter Chikyu) (Fig. 2b), the CCFs averaged over a one-hour period using ambient noise records of the vertical component (ZZ) between the two sensors were calculated for two conditions: RPMs of 0 (non-drilling period) and 32 (drilling period). This processing was also performed using the transverse component (TT), in which the transverse direction at the two sensors is perpendicular to the directions from the Chikyu to the two sensors.
Figure 2d shows four CCFs (ZZ and RPM = 0, ZZ and RPM = 32, TT and RPM = 0, and TT and RPM = 32). Because the reference site is OBS04 (seafloor) for the CCF calculation, peaks with large amplitudes at the positive/negative lag time in the CCFs represent the downgoing/upgoing wave. The ZZ-CCF for RPM = 0 shows the downgoing P wave at a lag time of 0.5 s, and the TT- and ZZ-CCFs for RPM = 32 show the upgoing S wave at a lag time of 1.8 s. This means that the upgoing S wave and downgoing P wave are dominant in the ambient noise wavefield during the drilling and non-drilling period, respectively.

**Correction on the amplification of seafloor records**

Seismic waves observed at the seafloor are amplified by soft materials in the marine sediments at shallow depths (Yabe et al. 2015). The amplified amplitudes at the OBS are corrected by zero or weak amplification of those observed at the borehole sensor. We used four deep earthquakes associated with the Pacific Plate subducting beneath the Japanese Island (Table 1). The epicenters of these events are located within 200 km from the drilling site, which ensures near-vertical incidence of P wave into the
OBSs. $W_{ij}$ is the root mean squared amplitude (RMS) of the vertical component for the $i$-th event and the $j$-th station, and is given by

$$W_{ij} = \sqrt{\frac{w_{ij}^2(t)}{w_{io}^2(t)}}$$  \hspace{1cm} (1)

where $w_{ij}$ and $w_{io}$ are the vertical components at an OBS and borehole sensor, respectively, for a time window from $-1$ to 9 s from the P arrival time. The RMS of the horizontal component, $H_{ij}$, is defined as

$$H_{ij} = \sqrt{\frac{u_{ij}^2(t) + v_{ij}^2(t)}{u_{io}^2(t) + v_{io}^2(t)}}$$  \hspace{1cm} (2)

where $u_{ij}$, $v_{ij}$ and $u_{io}$, $v_{io}$ are the horizontal components at an OBS and borehole sensor, respectively. Using these at a frequency band of 5–15 Hz, the site amplification factors for the vertical and horizontal components at the $j$-th station can be written as

$$W_j = \frac{\sum_{i=1}^{4} W_{ij}}{4}, \quad H_j = \frac{\sum_{i=1}^{4} H_{ij}}{4}.$$  \hspace{1cm} (3)

The estimated values for each of these variables are summarized in Table 2. The estimated values for different events were stable, which led to small errors. Although the amplification factors in the horizontal component are larger than those in the vertical component, such large values were also obtained previously in this region (Kubo et al.)
Moreover, based on this correction, in the Discussion section, we do not incorporate sediment layers in numerical simulations when comparing the observed and calculated amplitudes.

**Ps reflections from inline airgun shots within ±10 km**

Research Vessel (R/V) *Kaimei* (hereafter *Kaimei*) performed successive airgun shots along the NNW–SSE trending line within ±10 km from the drilling site (Fig. 3). Figure 3b shows the three-component seismic records, where two $P$-to-$s$ ($Ps$) reflected waves can clearly be seen in the two horizontal components. When a gently dipping interface, i.e., a megasplay fault, is present beneath the seafloor, a reflected $P$ wave emerges at a nearly constant two-way travel time within an offset of 10 km (Nakanishi et al. 2008): the megasplay fault is a major out-of-sequence thrust fault splayed off the plate interface. When the offset is small, a $Ps$ wave can also be observed in the same way. Based on $Ps$ – $P$ times of 6.1 and 7.6 s, a $Vp$ model (Shiraishi et al. 2019), and an empirical relation between $Vp$ and $Vs$ (Brocher 2005), two $Ps$ waves are reflected from the megasplay fault at a depth of 7 km and the top of the oceanic crust at
a depth of 9 km. For example, using a $V_p$ of 3.70 km s$^{-1}$ between the megasplay fault and the top of the oceanic crust (Shiraishi et al. 2019), we obtain a $V_s$ of 2.02 km s$^{-1}$ from the empirical relation (Brocher 2005). Using the differential travel time of 1.5 s, we obtained a thickness of 1.96 km between the two reflections.

**Ps reflections from airgun shot and ship noise by Chikyu**

Although the inline-shot records from Kaimei are useful for detecting $Ps$ signals, fixed-point airgun shots from Chikyu are more effective to compare with $Ps$ signals retrieved from Chikyu ship noise, because ship noise and airgun shots are generated from almost exactly the same location, whereas Kaimei was at least 400 m away from Chikyu.

To retrieve $Ps$ signals from airgun shot and ship noise, we applied a deconvolution technique in the frequency domain (e.g., Langston 1979; Ammon 1991) to the three-component records with a bandpass filter of 9–15 Hz: the horizontal component is divided by the vertical component in the frequency domain. Because we divided the deconvolved waveform by the maximum amplitude of the auto correlation
of vertical component (Ammon 1991), the absolute $Ps$ amplitudes are preserved irrespective of different OBSs and temporally-varying strengths of ship noise. Prior to the deconvolution, waveform amplitudes in the vertical and horizontal components were divided by the site amplification factors in Table 2. The time windows for airgun and ship noise (ambient noise) records were 20 s from the shot origin time and 80 s, respectively. For ship noise, hourly and daily waveforms in the radial and transverse components were obtained by stacking 80 s-deconvolved waveforms. Here, we only used 1-hour waveforms when the stacked number of the 80 s-deconvolved waveforms with RPM = 0 was > 40 among 45 segments ($\times 80$ s = 1 hour). This processing allows us to use the continuous 1-hour records in which non-drilling periods are dominant.

In Figs. 4 and S1, the deconvolved waveforms from 20 airgun shots were aligned at each OBS, while the daily waveforms from ambient noise on Julian days 1–120 of 2019 were aligned, in which Chikyu was located onsite on Julian days 1–58 and was away from the site after Julian day 59: this indicates that none of the signals after Julian day 59 were caused by ship noise. In these figures, we focused on lag times of 4–9 s by referring to $Ps$–$P$ times of 6.1 s and 7.6 s at the megasplay fault and the top of the
oceanic crust. Non-drilling periods (RPM = 0) were relatively secured during days 1–58 in 2019 compared with the project period in 2018 (Fig. 2b), and the location of the Chikyu was ranged within 10 m during the study period (Fig. 2c). Moreover, in order to evaluate the similarity of the deconvolved waveforms, we estimated cross-correlation coefficients (CCs) between the deconvolved waveforms using airgun shot and ship noise records, with a time window of 0.1 s and a time shift of 0.05 s (Figs. 4 and S1). Coherent signals between the two sources emerged at lag times of 5, 6, and 7.8 s in the radial-deconvolved waveforms (RDWs) at OBS03 (Fig. 4a), but signals were clearer in the transverse component and could be observed at all OBSs (Figs. 4b and S1). Here, the RDWs contain multiple peaks compared with the transverse-deconvolved waveforms (TDWs) (Fig. 4). Those peaks may have originated from the contamination of multiple reflections at shallow depth. In theory, the reflection coefficient of $P$-to-$S$ reflection from a horizontal boundary for a vertical incident angle is zero. However, when the incident angle of the $P$ wave and the boundary are slightly deviated from the vertical direction and horizontal plane, the reflection and its multiple reflections are possibly observed in the RDWs at smaller lag times and these are
deviated to other directions at larger lag times. Moreover, we estimated the CCs of the RDWs and TDWs between airgun shots and ship noises, and coherent signals were obtained with higher CCs in the TDWs (Fig. 4d). We therefore avoided the use of the RDWs in the subsequent processing. Although all of the signals in the TDWs obtained from the two sources were not completely matched, it is likely that the coherent signals originated from the same structural boundaries beneath the seafloor, including the megasplay fault, because peaks with high CC values could be observed in the TDWs between the airgun shots and ship noises (Fig. S1).

**Stacking waveforms with tidal stress**

The tidal normal stress on the megasplay fault was calculated with a method described in a previous study (Yabe et al. 2015). The geometry of the megasplay fault was inferred with a strike = 270° and dip = 10°, which were set by referring to three-dimensional reflections and velocity models in this region (Shiraishi et al. 2019). The slip direction had a rake = 125°, which represents the thrust fault. The azimuth and angle were defined as clockwise from north and downward from the horizontal plane.
Figure 5a shows the resulting normal stress on the megasplay fault, in which negative values represent compression. We stacked 1-hour TDWs at each 5 days for the following three ranges in normal stress at every OBS: high tide < –2.4 kPa, middle tide between –2.4 kPa and 1.6 kPa, and low tide > 1.6 kPa (Fig. 5a). The envelope functions were calculated by stacking 5-day TDWs during 10–50 Julian days when the number of available TDWs were balanced among the three tide ranges (Figs. 6 and S2): during the 5 days, 120 hours (maximum) were divided by the three ranges of the tide, and we counted the numbers of 1-hour TDWs in the three ranges (Fig. 5b). The above thresholds for the three ranges of normal stress were determined so as to preserve the number of 1-hour TDWs to at least 5 during every 5-day period and for each range. Although the resulting counts on the 37th day were less due to the drilling periods (Fig. 2b) and the count for middle tide on the date was 5, those for high and low tides exceeded 9. We therefore compared TDWs between high and low tides in this study. The uncertainties of the envelope functions were the standard deviations obtained by bootstrapping the 5-day TDWs for each tide range for 5,000 times with repetition (Figs. 6 and S2).
To investigate the stability of the amplitudes in the stacked TDWs, we plotted 5-day moving averages of the TDWs for the normal stress ranges of high and low tides (Figs. 5c and d). Moreover, the delay time of each segment in the 5-day TDWs between high and low tides was also estimated by cross-correlating them with a time window of 0.2 s and a time shift of 0.1 s (Figs. 5c and d). If some signals in the TDWs were systematically delayed (or advanced) between high and low tides, the delay time should have showed the coherent patterns at similar lag times.

Migration P-to-s scattered amplitude and tidal-responded amplitude

Using the three-dimensional $V_p$ model (Shiraishi et al. 2019) and an empirical relation between $V_p$ and $V_s$ (Brocher 2005), we calculated differential travel times between direct $P$ waves from Chikyu to an OBS and $P$-to-$s$ ($Ps$) scattered waves in two-dimensional transects. The amplitudes of the TDWs were plotted onto locations where their lag times were matched with the differential times in the transects. Here, the TDWs were stacked over 10–50 days at every OBS, and the amplitudes in the stacked TDWs were assumed to have been due to $Ps$ waves. The TDW amplitudes at lag times
of 5–9 s were used for imaging, where \(P_s\) waves reflected around the plate boundary were expected. We only imaged below OBSs within a scattering angle of the \(P_s\) wave less than \(2^\circ\) from the vertical axis, because the number of OBSs was small for imaging the entire volume around the OBS array. The scattering angle was defined as the angle between the incident \(P\) wave and the \(P_s\) wave at scattering points.

In addition to the raw amplitudes, we also created transects for displaying the locations where amplitudes responded to tidal stress. To enhance the amplitude variations in response to tidal forces, we calculated the amplified differential \(P_s\) amplitudes (AD-\(P_s\)) between the envelope functions of the high-tide TDW, \(S_H(t)\), and low-tide TDW, \(S_L(t)\),

\[
F(t) = \frac{R(t)}{|R(t)|} S_H(t) \cdot R(t)^2 \cdot C, \quad (4)
\]

where

\[
R(t) = S_H(t) - S_L(t), \quad (5)
\]

and \(C\) is an arbitrary constant which only makes the average amplitude of \(F(t)\) comparable to that of \(S_H(t)\). We chose \(C\) as \(2.0 \times 10^6\). \(R(t)\) indicates the difference of \(P_s\) amplitudes in the envelope functions of \(S_H(t)\) and \(S_L(t)\) (eq. (5)). The positive/negative
polarity in $R(t)$ represents the velocity contrast increase/decrease at seismic discontinuities at high tide. When the large $Ps$ amplitude in $S_{th}(t)$ and large $Ps$ amplitude difference in $R(t)$ occur of the same lag times, the peak amplitudes in the $S_{th}(t)$ were amplified while preserving the polarities of $R(t)$. Thus, $F(t)$ shows $Ps$ reflections from seismic discontinuities with large impedance contrasts and large amplitude variations between high and low tides. The examples of $S_{th}(t)$ and $F(t)$ at OBS03 and OBS06 are displayed in Figs. 6b and S2b. Using these functions, we made ENE-WSW and NNW-SSE trending transects (red lines in Fig. 1b) and projected them onto the reflection images (Shiraishi et al. 2019) along nearby ENE-WSW and NNW-SSE trending lines (yellow lines in Figs. 1a and b). Because AD-$Ps$ at OBS01 was slightly large compared with that at other OBSs and the amplitudes in the TDWs may have been unstable during 1–58 Julian days (Fig. S1a), we removed this station from imaging.

Results

Tidal response of reflections

To investigate whether the amplitudes and travel times of $Ps$ waves respond
to tidal stress, we calculated the normal tidal stress, slip-parallel shear stress, and slip-orthogonal shear stress on the megasplay fault (Yabe et al. 2015) by assuming its geometry (Shiraishi et al. 2019) and thrust fault slip derived from the motion vector of the underlying ocean plate (DeMets et al. 2010) (Fig. 5a). Figure 6 presents the comparison of envelope TDWs using ship noise at high and low tide, mainly showing larger peak amplitudes of $P_s$ waves at high tide. This feature can be partially recognized at OBS02 and OBS06 (Fig. S2), although some peaks show smaller amplitudes at high tide. The stability of the larger amplitudes at high tide at OBS03 and OBS04 were also confirmed by 5-day moving average of the TDWs (Fig. 5). In addition, the time-shift estimation of waveform segments of the TDWs between high and low tide with a time window of 0.2 s and a time increment of 0.1 s did not vary systematically (Figs. 5c and d). These observations indicate that seismic velocity changes in the entire medium associated with tidal stresses are minor, whereas velocity contrasts at the boundaries within the accretionary prism fluctuated with tidal stress.

Imaging of tidal responding boundaries
Depth migrations of amplified differential $Ps$ amplitudes (AD-Ps) between high and low tide (eq. (4)) located where the velocity contrast fluctuates in the accretionary prism. The $Ps$ signals of the TDWs and the AD-Ps were strong at the megasplay fault at a depth of 7 km, whereas those signals were weak at the top of the oceanic crust at a depth of 9 km (Fig. 7). Here, for AD-Ps, a positive/negative amplitude represents a correspondingly positive/negative enhancement of $Ps$ amplitude at high tide, respectively, and AD-Ps at three stations close to the line are plotted. Additionally, the velocity contrast of a boundary at a depth of 8 km between the megasplay fault and the top of the oceanic crust fluctuates with tidal stress. More of the clear features of the two boundaries can be recognized in the NS-trending transect (Fig. 8).

**Discussion**

**Velocity reduction in isotropic case**

A seismic exploration survey revealed the negative amplitude of $P$ waves reflected at the megasplay fault (Park et al. 2002), which is indicative of a velocity reduction within the fault. The larger positive amplitude observed at high tide in this
study indicates that the velocity reduction at the megasplay fault was further enhanced by vertical compression caused by tidal stress.

In order to confirm the velocity reduction, we conducted a three-dimensional finite-difference approach with rotated staggered grids and second-order calculation accuracies in time and space (Saenger et al. 2000). The model space ($x_1-x_2-x_3$) was $8 \times 8 \times 6$ km$^3$ and included four layers with a grid spacing of 0.01 km, and the $x_3$-axis was taken as vertically downward. The time step was 0.001 s. An absorbing boundary condition was assigned to each side except for the top (Clayton and Engquist 1977). The source location was set to $(x_1, x_2, x_3) = (4.0, 4.0, 0.0)$ at the model sea surface. A vertical force with a Ricker wavelet at a center frequency of 7.59 Hz (maximum frequency of 10.96 Hz) was applied to the source location. The stations were assigned at the seafloor locations (2 km depth) relative to the Chikyu in the observations (Fig. 9a). The $V_p$, $V_s$, and density of the homogeneous medium and fluid are 4.5 km s$^{-1}$, 2.25 km s$^{-1}$, and 2.5 g cm$^{-3}$, and 1.5 km s$^{-1}$, 0.0 km s$^{-1}$, and 1.0 g cm$^{-3}$, respectively (Table 4), in which $V_p$ is roughly referred to an averaged $V_p$ around the megasplay fault. In the isotropic case, a $V_s$ of 2.25 km s$^{-1}$ at layer 3 was reduced by 1, 3, 7, and 13 % (Table 4). The fault
thickness was set to be 0.05 km (Rowe et al. 2013), in which $V_p$ and $V_p/V_s$ were referenced to seismic exploration results (Shiraishi et al. 2019).

As a result, because the $P_s$ amplitude difference is linear with respect to $V_s$ reduction, the observed amplitude difference, $\sim 0.0002$ in Fig. 6a, can be linked to a $V_s$ reduction of 1.5–3.3 % (Fig. 9b). In this calculation, $P_s$ signals were not observed in the transverse component at OBS05 and OBS06. The $P_s$ amplitudes at these stations obtained in our observation may have been produced by a focusing effect of $P_s$ amplitudes at a gently curved interface of the megasplay fault that was imaged by a 3D seismic volume (Shiraishi et al. 2019), and so these were not used to estimate $V_s$ reductions. Because the obtained $V_s$ reduction of 1.5–3.3 % is large and elastic property perturbations linearly changed by the tidal stress are small, it is therefore necessary to consider other mechanisms capable of significantly changing the seismic velocity of the fault with small stress fluctuations and fluid concentrations.

**Fracture connection and isolation due to tidal stress**

A possible mechanism is fracture channeling with fluids in the fault by
external forces (e.g., (Hawthorne and Rubin 2010; Houston 2015)). Characteristic fractures are formed around a through-going fault (Y shear) (Bartlett et al. 1981) within the damage zone of a fault, with different direction groups including, e.g., Riedel (R) shears and antithetic R’ shears (Tchalenko and Ambraseys 1970). On land, these shear zones are found in the latest Cretaceous Mugi mélange in southwestern Japan, and this mélange is considered to be a fault rock along the plate boundary in the Nankai subduction zone (Kimura et al. 2012). These fractures around the deep plate boundary are potentially saturated with fluids, because the region between the fault and top of oceanic crust is characterized by high pore pressures (Tsuji et al. 2014), and the land mélange also shows evidence of fluid presence (Kimura et al. 2012). For a dip angle of the megasplay fault of 10° in this region (Shiraishi et al. 2019), the R shears are oriented close to horizontal (Fig. 10). When a small vertical-compressional stress is applied to R shears, their shapes are slightly compressed in the vertical direction, and internal fluids are expelled to Y shear and near-vertical R’ shears following the fluid pressure gradient. This results in fluid connections between shears, thereby enhancing seismic velocity anisotropy within the megasplay fault. By this mechanism, negative AD-Ps would be
caused only when near-vertical R’ shear expansion is prevalent and there is less connectivity of Y shears in the fault-parallel direction.

**Numerical simulation in anisotropic case**

To qualitatively investigate large $P_s$ amplitudes reflected at the megasplay fault, we also conducted numerical simulations for models in which the degree of anisotropy within the fault varied with the aspect ratio of cracks. Here, the assumptions for an effective medium in the damage zone along the megasplay fault are that highly/weakly connected fractures are described by inclusions of fault-oriented oblate spheroidal cracks with a small/large aspect ratio, with a constant porosity of the bulk medium at 0.07 by referring to the porosity $< 0.07$ of land outcrop samples that were considered to be a fossil seismogenic megasplay fault rock (Tsuji et al. 2008).

The calculation of the effective elastic constants for a medium containing fluid-filled fractures and their connections would be complicated. Instead, for the purpose of this study, we assumed a simple model in which fluid filled oblate spheroidal cracks, assuming Riedel shears, are horizontally developed within the
megasplay fault with a dip angle of 10°. When the cracks are partially connected by fluid-filled fractures, such as Y and R’ shears, due to an external force, individual fracture lengths are extended along the fault-parallel direction. This would be expressed by a slight decrease in the average aspect ratio of the cracks aligned to the fault-parallel direction with minor porosity changes. The $P$ wavelength used in this study is 300–500 m using a frequency of 9–15 Hz and a $V_p$ of 4.5 km s$^{-1}$ (Shiraishi et al. 2019), and the fault thickness at the Nankai accretionary prism toe was estimated to be 47.5 m from IODP Site C0007 (Rowe et al. 2013). Because the $P$ wavelength is sufficiently large compared to the fault thickness and therefore to the fracture sizes, the assumption that fracture connections can be described by the changes in the average aspect ratio of aligned cracks is acceptable.

We calculated effective elastic constants using differential effective medium (DEM) theory for hexagonal symmetry, in which aligned oblate spheroidal cracks with an aspect ratio are contained in a homogeneous medium of given porosity (0.07) (Nishizawa 1982; Tonegawa et al. 2013). The effective elastic constants transformed into those in which the axis was parallel to the normal direction of the megasplay fault
with a dip angle of 10° and strike = 270° by using the following equation (Nagaya et al. 2008).

\[ c_{i'j'k'l'} = U_{ii'}U_{jj'}U_{kk'}U_{ll'}C_{ijkl}, \]  

(6)

where the elastic constants \( C_{ijkl} \) \((i, j, k, l = 1, 2, 3)\) are transformed into those \( c_{i'j'k'l'} \) defined in the \((x_1'-x_2'-x_3')\) coordinate system. \( C_{ij} \) in Table 4 was converted into \( C_{ijkl} \) by

\[ C_{ijkl} = C_{jikl} = C_{ijkl} = C_{klij}. \]  

(7)

The matrix elements \( U_{ii'} \) can be written as

\[ U = \begin{pmatrix}
\cos \xi \cos \lambda & \cos \xi \sin \lambda & -\sin \xi \\
-\sin \lambda & \cos \lambda & 0 \\
\sin \xi \cos \lambda & \sin \xi \sin \lambda & \cos \xi
\end{pmatrix}. \]  

(8)

where \( \xi \) and \( \lambda \) are the tilt angle of the symmetry axis measured from the \( x_3 \)-axis and the azimuth of the axis measured clockwise from north (\( x_1 \)-axis), respectively. We set \( \xi = 10° \) and \( \lambda = 180° \) in the numerical simulations. The converted physical parameters were inserted in layer 3 in Table 5. In the isotropic case (aspect ratio = 1.0), \( V_p \) and \( V_s \) in the layer were reduced from 4.50 to 4.25 km s\(^{-1}\) and from 2.25 to 2.15 km s\(^{-1}\), respectively, because the porosity including fluids was set to 0.07. When the aspect ratio of cracks varied from 1.0, the fast and slow seismic velocities are slightly deviated
from those values. The aspect ratios are 0.2–1.0 with an increment of 0.2. The

calculated elastic constants are summarized in Table 4.

The results show that $P_s$ amplitudes tend to increase as the aspect ratio of

cracks decreases following a nonlinear curve (Fig. 10), except for the result at OBS05

as well as the isotropic case. Thus, our results indicate that fault-parallel connections of

fractures produce large $P_s$ amplitudes in our observation geometry, and that if fractures

are originally moderately connected (i.e. their aspect ratio is small), significant

amplitude variations can be induced by a small amount of additional connections

between fractures.

Moreover, we roughly estimated the preferred aspect ratio of cracks within

the megasplay fault in this region by using both our observation results and numerical

simulations. The observed $P_s$ amplitude and its difference between high and low tides

were $\sim 0.001$ and $\sim 0.0002$, respectively (Fig. 6). These values indicate that the aspect

ratio of cracks is close to 0.8–1.0, and that it varies within 0.6–1.0 between high and

low tides (Fig. 10). However, if the porosity within the megasplay fault was smaller

than 0.07, e.g., 0.05 in Fig. 11, the obtained $P_s$ amplitudes became small, thus
indicating that the aspect ratio of cracks and its variation between high and low tides would also be small. Because land outcrop samples within the fossil seismogenic megasplay fault shows a dominant aspect ratio of cracks of 0.1–1.0 (Tsuji et al. 2008), $Ps$ amplitudes obtained in this study can be explained by the 0.1–1.0 aspect ratio of cracks even if the porosity is slightly smaller than 0.05.

Moreover, using numerical simulations, we confirm whether the observed $Ps$ amplitudes are affected by (1) high velocity anomalies above the megasplay fault, detected by a seismic exploration study (Shiraishi et al. 2019), and (2) small velocity variations at shallow depths (Fig. 11). We examined the following two models: (1) a higher velocity zone above the megasplay fault than that at shallow depths without changing the velocity contrast at the megasplay fault, and (2) two layers above the megasplay fault with velocity reductions of 2.5 % and 1.5 % and thicknesses of 0.05 km and 0.03 km, respectively. The latter model assumes that other faults are present between the seafloor and the megasplay fault, and their velocities are slightly changed by tidal forces. As a result, the obtained $Ps$ amplitudes were slightly changed in model (1) from those in the original model but the characteristic in which $Ps$ amplitudes
increase with decreasing the aspect ratio was not changed, while $Ps$ amplitudes were
almost unchanged between models (1) and (2) (Fig. 11). This investigation indicates
that our results are not significantly affected by the velocity gradient and temporal
variations of velocity changes at shallow depth.

Conclusion

The physical properties of the plate boundary responding to tidal forces are
the important factor for understanding tidal triggering of earthquakes occurring at the
plate boundary. Such property changes are attributed to the presence and connection of
fluids within faults, and indicate the existence of mechanically weak interfaces.
Through the tidal response of reflection amplitudes, our study identified two such
parallel dipping faults around the plate boundary. Slow slip events in the shallow
Nankai subduction zone have been detected by pore pressure variations at the borehole
site (Araki et al. 2017) and also seafloor geodetic observations (Yokota and Ishikawa
2020), and they might have occurred at these multiple weak faults. Additionally, if the
physical properties of the weak portions within the megasplay fault have been
temporally varied with long-term fluid migration, and if the shear strength within the fault had been very weak, such slow slip events could be possibly triggered by tidal forces.

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Declarations

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Ethics approval and consent to participate

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Not applicable

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Consent for publication

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Not applicable

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List of abbreviations

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OBS: ocean bottom seismometer, RDW: radial deconvolved waveform,

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TDW: transverse deconvolved waveform, AD-Ps: amplified differential Ps amplitude, ROV: remotely operated vehicle, D/V: drilling vessel,

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R/V: research vessel, CCF: cross correlation function, ZZ: vertical component correlation, TT: transverse component correlation, RPM:

500

rotation per minute, RMS: root mean square, $V_p$: P-wave velocity, $V_s$: S-
wave velocity, CC: cross correlation coefficient, $Ps$: $P$-to-$s$, R shear:

Riedel shear

**Availability of data and materials**

Seafloor seismometer data are available from JAMSTEC upon request.

The data that support the findings of this study are available from the corresponding author upon reasonable request.

**Competing interests**

The authors declare no competing interests.

**Funding**

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**Authors' contributions**

T.T processed the data, and drafted the manuscript. T.K. designed the observation. K.S. prepared the data for reflection profiles and drilling parameters of *Chikyu*. S.Y. calculated tidal stress. Y.F. and S.K. contributed to the interpretation. T.T., T.K., K.S., E.A., M.K., Y.S., S.M.,
Y. N., and S. K. acquired the data underlying this study. All authors contributed to the final manuscript.

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**Authors’ information**

*Japan Agency for Marine-Earth Science and Technology (JAMSTEC)*,
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Figure legends

**Fig. 1 Geometry of the Chikyu, drilling site, and OBSs.** a Location of the drilling site (C0002) and reflection profiles (Shiraishi et al. 2019) (yellow lines). b The red triangles show the locations of the 6 OBSs. The red lines show the locations of the migration images shown in Figs. 7 and 8. The yellow lines correspond to the reflection profiles used in Figs. 7 and 8 (Shiraishi et al. 2019). c Vertical cross-section showing the *Chikyu*, the *Kaimei*, the OBSs and the fault at depth. Solid and dashed lines indicate *P* and *S* waves, respectively.

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**Fig. 3 Airgun shot experiment from R/V Kaimei.** Map showing the locations of inline airgun shots from R/V *Kaimei* (black dots). All other symbols are as in Fig. 1. Three-component shot records observed at OBS02 at frequency bands 5–15 Hz. Left and right sides of the panels represent the NNW and SSE directions.

**Fig. 4 Radial- and transverse-deconvolved waveforms (RDWs and TDWs) for D/V Chikyu airgun and ship noise.** RDWs for D/V *Chikyu* airgun (left) with 20 shots and ship noise (right) during Julian days 1–58 at OBS03. The days without data (gray
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Fig. 7 Tidal response of TDWs. a Normal stress on the megasplay fault, calculated from the method (Yabe et al. 2015). Normal stress values within the light-blue rectangle were used to calculate the stacked envelope function in Fig. 6. Red lines represent the thresholds at high and low tides. b Number of stacked 1-hour TDWs within 5-day moving averages at (black line) high tide, (blue line) middle tide, and (red line) low tide. c The first two panels on the left represent 5-day stacked TDWs during Julian days 1–60 in 2019 for high and low tides at OBS03. The frequency band is 9–15 Hz. The right panel displays the time shift of each segment between high and low tides, and the time window of the segment and increment are 0.2 and 0.1 s, respectively. d Same as panel c, but for OBS04.

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image of amplified differential \( Ps \) amplitudes between high and low tide (eq. (4)) in the TDWs projected onto the reflection image along B–B’ (Fig. 1) (Shiraishi et al. 2019).

Zoom in on the box in panel a. The red arrows indicate multiple weak interfaces, including the megasplay fault.

**Fig. 9** Numerical simulations for \( Ps \) reflected waves. a The source location of D/V Chikyu was set at the sea surface. Bathymetry is flat. The red triangles represent the 6 OBSs, whose relative horizontal locations with respect to D/V Chikyu were preserved. The physical properties at each layer are summarized in Tables 3–5. b \( Ps \) amplitudes reflected at layer 3 as a function of \( Vs \) reductions in Table 3 for the 6 OBSs, yellow: OBS01, black: OBS02, light-blue: OBS03, orange: OBS04, red: OBS05, and magenta: OBS06. The two dashed arrows show the \( Vs \) reduction at OBS04 for an amplitude variation of 0.0002.

**Fig. 10** Numerical simulation for the \( Ps \) amplitude reflected at the megasplay fault containing different aspect ratios of cracks. a The absolute amplitudes of \( Ps \)
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Fig. 11 Numerical simulation results for Ps reflected waves in anisotropic media. a Velocity model (1) with a velocity gradient above the megasplay fault at $x_2 = 4$ km. Star indicates the source location. Solid and dashed lines in right panel represent 1D velocity model at $(x_1, x_2) = (4.0, 4.0)$ and the original model, respectively. b Velocity model (2) with two layers of velocity reductions and a velocity gradient above the megasplay fault. c $Ps$ amplitudes at OBS02 for (light-blue) model (1), (orange) model (2), and (black) the original model (Fig. 4) at aspect ratio of cracks of 0.2, 0.6, and 1.0. d $Ps$ amplitudes at OBS02 for (red) porosity of 0.05 and (black) porosity of 0.07 (original model; Fig. 10).
### Table 1: Deep earthquakes used for estimating site amplification.

<table>
<thead>
<tr>
<th>Date</th>
<th>Latitude (°)</th>
<th>Longitude (°)</th>
<th>Depth (km)</th>
<th>Magnitude</th>
</tr>
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<tbody>
<tr>
<td>Eq. 1</td>
<td>2019/03/16 11:07:11</td>
<td>32.5795</td>
<td>136.9545</td>
<td>458.0</td>
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<tr>
<td>Eq. 2</td>
<td>2019/03/16 20:16:16</td>
<td>33.0767</td>
<td>138.3772</td>
<td>365.0</td>
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<td>Eq. 3</td>
<td>2019/03/20 04:17:30</td>
<td>34.1690</td>
<td>136.6558</td>
<td>359.9</td>
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<td>Eq. 4</td>
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<td>33.6270</td>
<td>137.2815</td>
<td>368.0</td>
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### Table 2: Estimated site amplifications at each OBSs.

<table>
<thead>
<tr>
<th></th>
<th>Amplification of the vertical component</th>
<th>Error for the vertical amplification</th>
<th>Amplification of the horizontal component</th>
<th>Error for the horizontal amplification</th>
</tr>
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<tr>
<td>OBS01</td>
<td>1.296</td>
<td>0.136</td>
<td>14.944</td>
<td>1.247</td>
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<td>OBS02</td>
<td>1.343</td>
<td>0.157</td>
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<td>0.881</td>
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<tr>
<td>OBS03</td>
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<td>0.118</td>
<td>19.446</td>
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<td>OBS04</td>
<td>1.406</td>
<td>0.157</td>
<td>19.595</td>
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<tr>
<td>OBS05</td>
<td>1.126</td>
<td>0.180</td>
<td>14.791</td>
<td>1.189</td>
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<tr>
<td>OBS06</td>
<td>1.300</td>
<td>0.195</td>
<td>16.214</td>
<td>1.046</td>
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</table>

### Table 3: Effective elastic constants as a function of aspect ratio of cracks.

<table>
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<tr>
<th>Aspect ratio</th>
<th>C11</th>
<th>C12</th>
<th>C13</th>
<th>C33</th>
<th>C44</th>
<th>Vp fast (km s⁻¹)</th>
<th>Vs fast (km s⁻¹)</th>
<th>Vp slow (km s⁻¹)</th>
<th>Vs slow (km s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.0</td>
<td>40.310</td>
<td>19.225</td>
<td>19.201</td>
<td>40.310</td>
<td>10.389</td>
<td>4.14</td>
<td>2.10</td>
<td></td>
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<tr>
<td>0.8</td>
<td>40.817</td>
<td>19.454</td>
<td>19.088</td>
<td>39.197</td>
<td>10.309</td>
<td>4.17</td>
<td>2.12</td>
<td>4.08</td>
<td>2.09</td>
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<tr>
<td>0.6</td>
<td>41.452</td>
<td>19.751</td>
<td>18.856</td>
<td>37.459</td>
<td>10.147</td>
<td>4.20</td>
<td>2.13</td>
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<td>2.08</td>
</tr>
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<td>0.4</td>
<td>42.249</td>
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<td>9.791</td>
<td>4.24</td>
<td>2.15</td>
<td>3.83</td>
<td>2.04</td>
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<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>4.32</td>
<td>2.17</td>
</tr>
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Table 4: Model space and physical parameters for numerical simulation.

<table>
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<th>Layer</th>
<th>Thickness (km)</th>
<th>$V_p$ (km s$^{-1}$)</th>
<th>$V_s$ (km s$^{-1}$)</th>
<th>Density (g cm$^{-3}$)</th>
</tr>
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<td>1.5</td>
<td>0</td>
<td>1</td>
</tr>
<tr>
<td>Layer 2</td>
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<td>4.5</td>
<td>2.25</td>
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<tr>
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<td>0.05</td>
<td>4.5</td>
<td>2.1375 (5%)</td>
<td>2.46</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2.0250 (10%)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1.9125 (15%)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1.8000 (20%)</td>
<td></td>
</tr>
<tr>
<td>Layer 4</td>
<td>1.95</td>
<td>4.5</td>
<td>2.25</td>
<td>2.46</td>
</tr>
</tbody>
</table>

Table 5: Model space and physical parameters for numerical simulation.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Thickness (km)</th>
<th>$V_p$ (km s$^{-1}$)</th>
<th>$V_s$ (km s$^{-1}$)</th>
<th>Density (g cm$^{-3}$)</th>
<th>Dip (°)</th>
<th>Strike (°)</th>
</tr>
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<tbody>
<tr>
<td>Layer 1</td>
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<td>1.5</td>
<td>0</td>
<td>1</td>
<td></td>
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</tr>
<tr>
<td>Layer 2</td>
<td>2</td>
<td>4.5</td>
<td>2.25</td>
<td>2.46</td>
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<td></td>
</tr>
<tr>
<td>Layer 3</td>
<td>0.05</td>
<td>*</td>
<td>*</td>
<td>2.46</td>
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<td>270</td>
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<td>Layer 4</td>
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<td>4.5</td>
<td>2.25</td>
<td>2.46</td>
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</tr>
</tbody>
</table>

*This value is listed in Table 4.
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**Fig. 9 Numerical simulations for Ps reflected waves.**

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Fig. 11 Numerical simulation results for Ps reflected waves in anisotropic media. a
Velocity model (1) with a velocity gradient above the megasplay fault at $x_2 = 4$ km. Star indicates the source location. Solid and dashed lines in right panel represent 1D velocity model at $(x_1, x_2) = (4.0, 4.0)$ and the original model, respectively. b Velocity model (2) with two layers of velocity reductions and a velocity gradient above the megasplay fault.

 c $Ps$ amplitudes at OBS02 for (light-blue) model (1), (orange) model (2), and (black) the original model (Fig. 4) at aspect ratio of cracks of 0.2, 0.6, and 1.0. d $Ps$ amplitudes at OBS02 for (red) porosity of 0.05 and (black) porosity of 0.07 (original model; Fig. 10).