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**Key Points:**
- High-sampling rate is important for pressure gauges to record seismic body waves, Rayleigh waves, and tsunamis and their dispersive features.
- Theoretical relation between pressure and vertical acceleration \((p = \rho_pH\rho_s)\) is valid for a long time \((\sim 3\text{ hr for the 2010 Chile earthquake})\).
- A relationship between pressure and vertical velocity \((p = \rho_p c_v v_z)\) holds only at the first \(P\) wave arrival, but not for later phases.

**Supporting Information:**
- Supporting Information S1

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**Abstract**
Recent developments of ocean-bottom pressure gauges (PG) have enabled us to observe various waves including seismic and tsunami waves covering periods of \(T \sim 10^2\)–\(10^3\) s. To investigate the quality for broadband observation, this study examined the broadband PG records (sampling rate of 1 Hz) around Japan associated with the 2010 Chile earthquake. We identified three distinct wave trains, attributed to seismic body waves, Rayleigh waves, and tsunamis. Clear dispersive features in the Rayleigh waves and tsunamis were explained by theories of elastic waves and gravity waves. Quantitative comparison between pressure and vertical velocity \((p = \rho_p c_v v_z)\) holds only at the first \(P\) wave arrival, but not for later arrivals. Similar results were confirmed for various earthquakes with different source-station distances and magnitudes, suggesting the robustness of these relations. The results demonstrate that the high-sampling rate \((\geq 1\text{ Hz})\) is necessary to observe seismic-wave dispersion and PG can record both seismic waves and tsunamis with reasonable quality for waveform analyses, whereas conventional onshore and offshore seismometers or tide gauges can observe either of seismic waves and tsunamis. Utilizing the high-sampling PG in combination with the seismic and tsunami propagation theory for estimating earthquake source process or analyzing wave propagation processes in the ocean will deepen our geophysical understanding of the solid-fluid coupled system in the Earth and contribute toward disaster mitigation.

**Plain Language Summary**
Recent developments of offshore ocean-bottom observation networks have enabled us to use high-sampling (one or more samples per second) seafloor pressure gauge (PG) data. This study investigated PG records with broadband period range (seconds to hours) around Japan during the 2010 Chile earthquake. We identified a seismic \(P\) wave train arriving \(\sim 20–30\text{ min after the focal time}.\) Another seismic wave train due to the surface Rayleigh wave during \(\sim 70–110\text{ min and tsunami during }\sim 24–72\text{ hr was also confirmed},\) which showed a dispersive feature; long-period waves arrive earlier than short-period waves. The dispersion theories obtained from the elastic and fluid dynamics thoroughly explained these features. We also compared wave amplitudes between the PG and nearby ocean-bottom seismometer and confirmed the validity of the relationship between pressure and vertical acceleration and between pressure and vertical velocity. This study demonstrates PG can record both seismic and tsunami signals clearly with reasonable quality, while conventional seismometers or tide gauges can either. Ultrabroadband observation of PG plays an important role in deepening our understanding of geophysical wave propagation processes in the solid-fluid coupled system in the ocean and enables delivery of essential information for earthquake early warning and disaster mitigation.

**1. Introduction**
Ocean-bottom pressure gauges (PGs) have recently been developed worldwide for tsunami observations (e.g., Mungov et al., 2013; Rabinovich & Eblé, 2015; Titov et al., 2005; Tsushima & Ohta, 2014). Compared with tsunami observation using the conventional tide gauges, PGs are advantageous as they are less affected by complex coastal site effects (e.g., Kubota et al., 2018). Owing to this advantage, PGs have significantly contributed to our understanding of tsunami generation and propagation processes derived from solid and fluid dynamics (e.g., Allgeyer & Cummins, 2014; Inazu & Saito, 2013; Kubota et al., 2020; Lay et al., 2016; Maeda et al., 2013; Poupardin et al., 2018; Rabinovich et al., 2013; Saito et al., 2010; Sandanbata et al., 2018; Satake et al., 2005; Watada et al., 2014; and references therein). PGs have further been utilized for observing much
longer time scale phenomena, such as infragravity waves (e.g., Tonegawa et al., 2018) and oceanographic and geodetic phenomena (e.g., Baba et al., 2006; Fukao et al., 2019; Inazu et al., 2012; Wallace et al., 2016).

In addition, recent studies have reported that the PG observations can record other geophysical waves, including ocean-acoustic and seismic waves (e.g., Bolshakova et al., 2011; Levin & Nosov, 2009; Matsumoto et al., 2012; Mizutani et al., 2020; Nosov & Kolesov, 2007; Webb & Nooner, 2016). Dynamic pressure changes associated with the coseismic seafloor vertical accelerations, recognized as a reaction force to the seafloor lifting up the water column, can also be observed (e.g., An et al., 2017; Filloux, 1982; Ito et al., 2020; Nosov & Kolesov, 2007; Saito, 2019). This indicates the ability for PGs to be utilized as vertical accelerometers for the sea-bottom motions (An et al., 2017; Kubota et al., 2017). One strong advantage for using PGs in seismic wave observation is that the signal never saturates, unlike high-sensitivity ocean-bottom seismometers (OBSs). The recent development of seafloor pressure observations with a sampling rate exceeding 1 Hz (hereafter, high-sampling rate) in comparison with the low sampling rate of the conventional PG (e.g., the Deep-ocean Assessment and Reporting of Tsunamis [DART] system, 1/15 Hz; e.g., see Rabinovich & Eblé, 2015) has contributed extensively to our understanding of ocean-acoustic waves and seismic waves. Given the PG observations of various geophysical waves, PG could be utilized for ultrabroadband geophysical observation, including both seismic and tsunami waves, which will be useful to extract earthquake source information and the ocean’s ultrabroadband wave propagation process. However, PG records have primarily been utilized for analyzing tsunamis and not extensively for seismic wave signals. Recent studies have discussed PG quality and performance in tsunami observation, which featured a period range of $10^2$–$10^3$ s (e.g., Rabinovich & Eblé, 2015; Saito et al., 2010; Tsushima & Ohta, 2014), whereas those in seismic waves have not been examined in detail.

In the present study, we examine the high-sampling PG record featuring a period range of $10^5$–$10^7$ s, with particular focus on the seismic wave signals. Section 2 describes the PG and additional data sets used and the method conducted for the analyses. Section 3 outlines the results. Section 4 interprets the results by examining the origins of the identified PG signals based on both the propagation theories of seismic waves and tsunamis and the comparison with other nearby instruments. Section 5 compares the PG waveform with nearby OBSs to discuss the quantitative relationship between pressure change and seismic waves. Section 6 examines the quality of the PG observation by comparing the spectral amplitudes and the background noises to discuss the performance and limitations of the high-sampling PG observation. Finally, section 7 discusses future potential of the high-sampling PG for furthering our geophysical understanding and for contribution to practical disaster mitigation. The conclusion is summarized in section 8.

2. Data and Method

2.1. Data

We examined the records of onshore and offshore instruments in northern Japan (Figure 1) during some moderate local to regional earthquakes and major regional to global earthquakes. We here primarily focused on the 2010 Chile earthquake (Mw 8.8; Duputel et al., 2012) because of its large magnitude and ideal source-station distance (Figure 1; the other examples are discussed in section 5). The approximate distance along the great circle path from Chile to Japan is 17,000 km (angular distance of 150°). Station information is listed in supporting information Table S1. We use PGs at KPG1 and KPG2 (dark blue inverted triangles in Figure 1b) and three-component acceleration records from a OBS at KOBS1 (red circle) of the Off-Kushiro cabled-observatory, operated by the Japan Agency for Marine-Earth Science and Technology (Hirata et al., 2002; Kawaguchi et al., 2000). A coastal tide gauge at Hanasaki Port operated by the Japan Meteorological Agency (blue diamond) and a Streckensen STS-2 onshore broadband seismometer at Kushiro (KSRF; green triangle) and a nearshore tiltmeter at Samani (orange square) operated by National Research Institute for Earth Science and Disaster Resilience (NIED, 2019; Okada et al., 2004) were also utilized. The detailed specification of the instruments is described in Text S1. The velocity records from the onshore seismometer are converted to accelerograms. The horizontal components of the original seisograms and tilt records are along the north-south and east-west directions, which are rotated to the radial (R) and transverse (T) directions along the great circle path (Figure 1). We resampled all of these data sets with 1 Hz. Note that the instrument responses at the onshore and offshore seismometers were not removed. We also examine the PGs of the DART system deployed by the
National Oceanic and Atmospheric Administration in the Pacific Ocean (Titov et al., 2005; Mungov et al., 2013; black inverted triangles in Figure 1). Refer to Table S2 for information on the DART stations.

Figure 2a shows the time series after the Chile earthquake from the PGs (dark blue), tide gauge (blue), onshore seismometer (green), OBS (red), and tiltmeter (orange). The waveforms from DART records are shown in Figure S1. Tidal variations are observed by the PGs and tide gauge. We also confirm the small tidal fluctuations in the tiltmeter. Figure 2b shows the high-pass filtered records with a cutoff period of 10,800 s (3 hr, ~0.1 mHz). Seismic waves are clearly observed, except with the tide gauge. Figure 2c depicts the band-pass filtered records (passbands of 30–10,800 s, ~0.1–33 mHz). Tsunamis are clearly detected by the PGs and the coastal tide gauge. A pressure change of \( \Delta p = 1 \text{ hPa} \) corresponds to a sea-surface height change of \( \Delta \eta = 1 \text{ cm} \) (\( \Delta p = \rho_w g_0 \Delta \eta, \rho_w \sim 1.03 \text{ g/cm}^3 \): seawater density, \( g_0 = 9.8 \text{ m/s}^2 \): gravitational acceleration); therefore, the tsunami amplitudes in PGs and tide gauges are comparable. Tsunami-related tilt changes are also recorded with the coastal tiltmeter (e.g., Kimura et al., 2013; Nishida et al., 2019). The amplitudes of seismic

Figure 1. Location map of this study. (a) The 2010 Chile earthquake (Duputel et al., 2012) and great circle paths to the stations (yellow lines). DART tsunami stations are shown by inverted triangles. (b) Enlarged view around northern Japan. Dark blue inverted triangles are the PGs, the blue diamond is the coastal tide gauge, the green triangle is the onshore broadband seismometer, the red circle is the OBS, and the orange rectangle is the nearshore tiltmeter.

Figure 2. Time series at onshore and offshore sensors associated with the 2010 Chile earthquake. (a) Raw data. (b) High-pass filtered data (cutoff period of 10,800 s). The waveform around the first P wave arrival is enlarged in (b'). (c) Band-pass filtered data (30–10,800 s). Note that the scales are different in each panel and U, R, and T denote the vertical, radial, and transverse component, respectively.
waves and tsunamis in the PGs are almost comparable, whereas tsunamis recorded in the tiltmeter (orange trace in Figure 2c) have much smaller amplitudes than the seismic waves, by magnitudes of $10^2$.

2.2. Spectrogram Analysis

We examine the features of the high-sampling PG data based on the spectrogram analysis which calculates the Fourier transform with a moving time window. In this study, we applied Aki and Richards’s (1980) definition of the Fourier transform, as

$$X(\omega) = \int_{-T_L/2}^{T_L/2} x(t)e^{i\omega t}dt,$$

where $T_L$ is the time window length of each bin and $\omega$ is the angular frequency ($=2\pi f$, $f$ is frequency). Here, the spectrogram $A(t, \omega)$ is defined as

$$A(t, \omega) = \left| \int_{-T_L/2}^{T_L/2} x(t+\tau)e^{i\omega \tau}d\tau \right|.$$

For the calculation of the spectrogram, we adopt the Fourier amplitude spectra, but not power spectra. In order to examine the spectral features of the seismic waves, we set the time window length of each bin ($T_L$) at 512 s and the time shift ($\Delta T$) at 60 s (i.e., $t = n\Delta T$, $n$ is an integer number). We also investigate tsunamis based on the spectrogram with $T_L = 8,192$ s and $\Delta T = 1,800$ s. We further examined the spectral features with broadband period ranges including both seismic waves and tsunamis, using $T_L = 1,024$ s with $\Delta T = 60$ s. Prior to calculating the Fourier transform of each bin, we removed the mean value and applied a Hanning taper (Blackman & Turkey, 1958) with a length of 0.05$T_L$ to both edges of the window.

3. Results

The spectrograms for the PGs at KPG1 and KPG2 are shown in Figure 3. The spectrograms during the first 3 hr from the origin time are illustrated in Figures 3a and 3b ($T_L = 512$ s). The spectrogram confirms two distinct seismic wave trains during the seismic wave arrival at $\sim$20 min (periods of $T \sim 3$–20 s, $\sim$50–300 mHz) and $\sim$70 min ($T \sim 10$–200 s, $\sim$5–100 mHz) (black and red arrows in Figures 3a and 3b, respectively). The second wave train at $\sim$70 min displays a clear dispersive feature (i.e., long-wavelength waves arrive earlier than short-wavelength waves). Figures 3c and 3d depict the spectrogram for 120 hr after the origin time ($T_L = 8,192$ s). A tsunami-attributed wave train, showing a clear dispersion ($T \sim 60$–1,000 s, $\sim$1–20 mHz), can be confirmed in the spectrograms at $\sim$24 hr (blue arrow in Figures 3c and 3d). The arrival timing and duration of seismic waves and tsunamis are significantly different; hence, it is difficult to display both in one figure. In order to display both wave signals, we show the spectrograms using the logarithmic scale for the elapsed time in Figures 3e and 3f ($T_L = 1,024$ s).

We also calculate the spectrograms for other onshore and offshore instruments to compare with the PG spectrograms (Figure 4; see Figures S2 and S3 for the spectrograms with linear scale for the elapsed time). The seismic wave trains at $\sim$20 and $\sim$70 min are detected by the tiltmeter, onshore seismometer, and OBS, but not by the coastal tide gauge, which is similar to the PG spectrogram. Another distinct wave train can also be confirmed at $\sim$40–45 min in the horizontal components of the seismometers (e.g., black arrow in Figure 4e). We also identify an additional seismic wave train in the transverse component at the Station KSRF at $\sim$70 min (e.g., green arrow in Figure 4f), but this is not observed in the PG spectrogram. We recognize small amplitude increases associated with tsunamis in the tiltmeter spectrogram at $\sim$24 hr (blue arrows in Figures 4a–4c), but the dispersive feature cannot be observed.

4. Interpretation on Results of Spectrogram Analysis

4.1. Seismic Wave Trains

We interpret the wave trains in the PG and the other instruments identified by the spectrogram analysis based on the theory of seismic waves and tsunamis. The onset timing of the first waves in the pressure records ($\sim$20 min) is identical to the onshore seismometer (Figure 2). We calculate the theoretical travel time of the core phases such as the PKP phase, the $P$ wave penetrating into the outer core, at KSRF using the Global Earth Structure Model AK135 (Kennett et al., 1995; Figure S4). We obtain a travel time of PKP...
phase of 1,193.74 s (~19.8 min), which accurately reproduced the observed arrival time (Figure S4). Thus, we conclude that this wave train is the seismic body P waves including PKP and PKIKP, PKiKP, PKPab, and PKPbc phases (e.g., Blom et al., 2015). Furthermore, the theoretical travel time of the SS phases, free-surface-reflected S waves leaving a source downward, at KSRF is 2,570.55 s (~42.7 min; Figure S4). This also corresponds with the onset of the waves confirmed in the horizontal seismograms at ~40–45 min.

Figure 3. Spectrograms for PGs at (a, c, e) KPG1 and (b, d, f) KPG2. (a, b) Spectrograms for seismic waves. (c, d) Spectrograms for tsunamis. (e, f) Spectrograms for both seismic and tsunami waves with a logarithmic scale on the horizontal axes. The high-pass (a, b, e, and f) and band-pass (c and d) filtered waveforms are also shown. Distinct wave train arrivals are denoted by the colored arrows (black = seismic P waves, red Rayleigh waves; pink = Rayleigh waves from the opposite direction to the first Rayleigh wave; blue = tsunamis).
Figure 4. Spectrograms for (a) tide gauge at Hanasaki, (b, c) tiltmeter at SAMH, (d, e, f) onshore seismometer at KSRF, and (g, h, i) ocean-bottom seismometer at KOBS1. The horizontal axes are displayed on a logarithmic scale. The Love wave train arrival is denoted by the green arrow. See the caption in Figure 3 for the other description.

(Figure 4e) and also indicates the body wave. The direct S waves do not arrive at the stations off Hokkaido because they do not propagate into the fluid outer core (e.g., Shearer, 2009).

The dispersion is detected in the second seismic wave train at ~70 min. To identify the origin of this wave, we examine the particle motion of the seismometers at KSRF and KOBS1 (Figure 5). The pressure waveform at KPG1 is depicted in Figure 5 (red trace). The particle motions in both OBS and onshore seismometer records demonstrate retrograde motion during the arrival of this wave at 4,200–4,800 s (black arrow in Figure 5). This indicates a seismic surface wave (Rayleigh wave). For the time window during 100–110 min (6,000–6,600 s) from the origin time, we confirm another retrograde particle motion with the rotation direction opposite to the first Rayleigh wave (gray arrow in Figure 5). A corresponding spectral amplitude increase is also confirmed (pink arrow in Figure 3). These comparisons suggest another Rayleigh wave train from the opposite direction to the first Rayleigh wave arrival and delayed by ~30 min. It is also worth pointing out that the Rayleigh wave arrivals at 5–10 mHz is slightly delayed compared to those at 20–50 mHz (Figures 3a and 3b). This observation has an opposite sense to typical understanding of surface dispersion that the longer-period waves arrive earlier (e.g., Shearer, 2009). A similar dispersion pattern also appears in the KOBS1 record, in particular, the vertical component (Figure 4g).
Examining the particle motion of the seismometers suggests that the wave train arriving to the PG at \(\sim 70\) min is the Rayleigh wave. We further examine the dispersive feature of the Rayleigh wave train recorded by the PGs based on the elastic wave propagation theory. We calculate the theoretical group velocity dispersion of the Rayleigh wave based on the AK135 global structure model (Kennett et al., 1995) incorporating a seawater layer with a thickness of 4.5 km. The velocity structure and the dispersion relationship is depicted in Figure S5. Figure 6 shows the comparison between the theoretical arrival time of the Rayleigh wave and the spectrograms. The theoretical Rayleigh wave arrivals at each period are consistent with the spectral amplitude peak in the spectrograms (red lines in Figure 6). In addition, the onset of the delayed Rayleigh wave train is also explained (pink lines) when considering a path along the opposite direction of the great circle path on the Earth (\(\sim 23,000\) km). This suggests that the delayed wave train is also due to the Rayleigh wave but radiated to the opposite direction of the first Rayleigh wave train.

We also calculate the theoretical dispersion relationship using different structure models to compare with the observed seismograms. We reveal that a model excluding the seawater layer from the AK135 structure model (Figure S6) cannot explain the short-period Rayleigh waves (greater than \(\sim 50\) mHz; Figure S7). We also compare the theoretical arrival time based on the Preliminary reference Earth model [PREM] (Dziewonski & Anderson, 1981), which includes a seawater layer (Figure S8). The observed dispersive features in the Rayleigh waves are reasonably reproduced by the Preliminary reference Earth model, although the theoretical arrival time is slightly delayed by \(\sim 10–20\) min in periods at \(\sim 10–50\) s (\(\sim 10–100\) mHz; Figures S9 and S10). This result demonstrates that the Earth structure model is important for reproducing
the dispersive feature, which suggests Rayleigh wave dispersion recorded by the PG record can be useful to constrain the subsea floor structure, as well as the ocean-bottom seismometers.

In the particle motion of the onshore seismometer, a particle motion along the transverse direction is also confirmed during 3,600–4,200 s (Figure 5). This indicates the Love wave is also observed in the horizontal components of the seismometers. In order to examine the Love wave train in the PGs, we calculate the group velocity and theoretical arrival times of the Love wave (green line in Figure 6). The spectral amplitude increase due to the Love wave cannot be confirmed in the KPG1 spectrograms, while the corresponding signal is recognized in the transverse component of the seismometer at KSRF (green arrow in Figure 4f). This result is expected given the transverse motion of Love waves (e.g., Shearer, 2009) while the pressure changes due to the seafloor motion are primarily caused by the seafloor vertical accelerations (e.g., Saito, 2019).

4.2. Tsunami Wave Trains

The wave train arriving at ~24 hr, which was attributed to a tsunami, also demonstrated the dispersion. Based on the gravity wave theory (e.g., Pedlosky, 2013; Saito, 2019), the tsunami phase velocity \(c\) and group velocity \(U\) are given as

\[
c = \frac{2\pi}{\omega} = \frac{\omega}{k},
\]

\[
U = \frac{c}{2} \left[ 1 + \frac{2kH_0}{\sinh(2kH_0)} \right],
\]
where \( k \) is the wave number (=\( 2\pi/\lambda \); \( \lambda \) is wavelength), \( T \) is the period, and \( \omega \) is the angular frequency. The angular frequency \( \omega \) follows the dispersion relation:

\[
\omega^2 = g_0 k \tanh(kH_0),
\]

where \( g_0 \) is the gravitational acceleration and \( H_0 \) is the water depth. We calculate the theoretical arrival time of the tsunami energy using the group velocity at each period (blue line in Figures S5b and S5c). The theoretical arrival time explains the spectral peak in the PG (Figure 6a). The tide gauge spectrogram, meanwhile, did not show clear tsunami dispersion (Figure 6b). The main cause for this is its low sampling rate (1/60 Hz). However, the complex coastal site effect associated with local bathymetry might be another possible cause (e.g., Geist, 2018; Tanioka et al., 2019). The dispersive feature was also not clearly recognized in the tiltmeter spectrogram (Figures 4b and 4c).

We investigated the spectral feature of the PGs of the DART stations (Mungov et al., 2013; Titov et al., 2005; black inverted triangles in Figure 1). The information of the DART stations is listed in Table S2, the waveforms are depicted in Figure S1, and the spectrogram of the DART data is in Figure 7. The dispersive feature can also be confirmed, even in the spectrogram of the DART Station 32412, closest to the epicenter (~2,400 km from the source; Figure 7a), although the temporal delay of the shorter-period components is too large for other distant DART stations (Figures 7b–7e) and KPG stations (Figure 4). We can also recognize the dispersive Rayleigh wave signals in these DART records. However, because the sampling rate of these DART systems is low (\( \leq 1/15 \) Hz), it is difficult to recognize the dispersive feature from the spectrograms.

In the KPG1 and KPG2 spectrograms (Figures 3c and 3d), after the tsunami arrival at 24 hr, the amplitudes at periods of 50–500 s (20–200 mHz) increase between 24 and 36 hr. This feature is also confirmed in the DART spectrograms (Figure 7). We also verify amplitude variations above 20 mHz following tsunami arrival in the onshore seismometer at KSRF (Figures S3f and S3g) and nearshore tiltmeter at Samani (Figures S3k and S3i). No local earthquakes or aftershocks occurred during the tsunami arrivals, so this may suggest that these shorter-period tsunamis are generated by long-period ones with the effects of reflections and scattering by local bathymetry.

In addition, the delayed shorter-period tsunamis \( (T < \sim 100 \text{ s}) \) at greater than \( \sim 30–40 \) hr are not recognized in the DART spectrograms. The static pressure change at seafloor \( p \) due to sea-surface height change (i.e., tsunami) can be expressed as follows (e.g., Saito, 2019):
5. Quantitative Relation Between Pressure and Vertical Seismogram

In section 4.1, we verify that the PG clearly observed the Rayleigh wave signal. The current section compares the quantitative relationship between the pressure change and the seafloor vertical seismic motion in detail. When the period is longer than the resonant period of the acoustic fundamental mode, we focus on the period bands of 10–100 s, in which Rayleigh waves are the most dominant signal in the records. The amplitudes and phases of the pressure change at KPG1 and the vertical acceleration at KOBS1 are very similar (Figure 8a). The vertical displacement expected at KPG1 obtained from Equation 7 is similar to the vertical displacement at KOBS1 as well (pink trace in Figure 5). We assess the similarity of the pressure change and vertical acceleration based on the correlation coefficients (CCs), which is defined as

\[
CC(t) = \frac{\int_{-T_L/2}^{T_L/2} p(t+\tau)a_z(t+\tau)d\tau}{\sqrt{\int_{-T_L/2}^{T_L/2} p(t+\tau)^2d\tau \int_{-T_L/2}^{T_L/2} a_z(t+\tau)^2d\tau}}
\]  

where \(T_L\) is the length of the time window. We use the two time window lengths for the CC calculation; \(T_L = 60\) s (gray line in the middle panel in Figure 8a) and \(T_L = 300\) s (black line). We obtain CC \(\sim 1\) during the dominant Rayleigh wave arrival (\(\sim 70\) min) and also other seismic waves (\(\sim 1–3\) hr). Recent studies have reported the validity of Equation 6 in the time domain in the coseismic data set of a few minutes from the origin time (An et al., 2017; Kubota et al., 2017). Since the signal-to-noise ratios in the KPGs are so high, they had good agreement with seismograms for a much longer duration than the results from previous studies. We further compare the pressure change at KPG1 and the vertical acceleration at KOBS1 for other local to regional moderate earthquakes and regional or global major earthquakes, the 2006 Kuril earthquake (Mw 8.3, epicentral distance \(\sim 890\) km; Figure 8b), the 2007 Kuril earthquake (Mw 8.1, 950 km; Figure 8c), the largest foreshock of the 2011 Tohoku earthquake (Mw 7.3, \(\sim 390\) km; Figure 8d), the 2011 Tohoku earthquake (Mw 9.0, 420 km; Figure 8e), the 2012 Sumatra earthquake (Mw 8.6, \(\sim 6,700\) km; Figure 8f), and the 2018 Hokkaido Iburi Eastern earthquake (Mw 6.6, 230 km; Figure 8g). We obtain similar results regardless of their very different source-station distances and magnitude, which suggests the robustness of the observed relation.

We calculate the temporal variation of the coherency and phase difference between two waveforms in the frequency domain. The procedure to calculate the coherency and phase difference is summarized in Text S2. Our results confirm the high coherency (greater than \(\sim 0.9\)) and zero-phase-difference at periods of \(\sim 10–50\) s (\(\sim 20–100\) mHz; Figures 8d and 8e) during the Rayleigh wave arrival. We further assess the
The correlation between the pressure and vertical acceleration with different dominant periods (Figure 9). The correlation between the pressure and acceleration is low in the shortest period bands ($T < 10$ s). This is due to the dynamic relation being derived under the assumption that the wave period is longer than the resonant period of the acoustic fundamental mode $T_0 = \frac{4H_0}{c_0} \sim 6$ s. We also find that the correlation between the pressure and acceleration is very high in periods of $T \sim 10$–50 s (Figures 9c and 9d), which infers that the dynamic relation holds well at the period bands of $\sim 10$–50 s. Conversely, at longer periods ($T > 50$ s), the correlation is not as high and unstable (Figures 9e–9h). The low correlation at the longer period is caused by the noises attributed to the seafloor vertical displacement due to the long-period ambient oceanographic water waves. This is referred to as the compliance noise (e.g., An et al., 2020; Crawford et al., 1991, 1998; Webb, 1998; Webb & Crawford, 1999). However, we detected relatively high correlation during the Rayleigh wave arrival ($\sim 70$–100 min) even at the longer periods up to $T \sim 400$ s ($\sim 2.5$ mHz; bottom panels in Figures 9e–9g). This indicates that the PGs can observe the Rayleigh wave signals up to the periods of $\sim 400$ s where the Rayleigh wave signals are larger than these compliance noises.

In contrast to the dynamic pressure relation (Equation 7), a relationship between the pressure change and vertical velocity $v_z(t)$ is proposed in several previous studies (e.g., Bolshakova et al., 2011; Matsumoto et al., 2012; Nosov et al., 2007):
We also compare the pressure change at KPG1 and the vertical velocity at KOBS1, with periods shorter than the acoustic resonant period $T_0 \sim 6$ s. As a result, we can recognize high correlation ($\sim 1$) during the theoretical $P$ wave arrival in this period band (Figure 10a). The comparisons between pressure change at KPG1 and the vertical velocity at KOBS1 for other moderate to major earthquakes confirm that this tendency is robust (Table S3 and Figures 10b–10g). If we carefully review the $P$ wave arrivals, we find that the amplitude of the pressure change $p(t)$ does not agree with $\rho_w c_0 v_z(t)$, although the phases agree with each other. The amplitude ratio between $p$ and $\rho_w c_0 v_z^2$ in the first arrival of the seismic waves differs for each event (see Figure 10). This tendency is most evident in the 2011 Tohoku-Oki earthquake (Figure 10e). The local site effect due to the station location difference might be the possible cause. However, this seems implausible because the station difference between KPG1 and KOBS is only $\sim 4$ km, and any systematic relation in the amplitude ratio among these events cannot be identified.
We also point out that the high correlation is confirmed only in the P wave arrivals, but neither in the entire wave trace nor in the later arrivals such as S waves (Figures 10b–10g). The pressure \( p \) tends to be larger than \( \rho_w c_0 v_z \) around the P arrivals. We should note that a relation \( p = \rho_w c_0 v_z \) can be derived assuming a P wave propagation through fluid medium (e.g., Saito, 2019). However, the wavefields are more complicated than a P wave propagation through fluid medium, in particular, in the later wave arrivals where the waves are trapped by the reflections between the sea surface and the seafloor. The previous results of numerical simulations suggest this feature; the pressure changes at the seafloor due to trapped P waves or ocean-acoustic waves exceed \( \rho_w c_0 v_z \) (Bolshakova et al., 2011; Saito, 2017). Although Matsumoto et al. (2012) compared the Fourier amplitudes of the seismograms and pressure records and found the consistency with the relation \( p = \rho_w c_0 v_z \) in the frequency domain, they discussed only the Fourier amplitudes and not the phases. Our results show that the validity of the theoretical relationship between the pressure and velocity is very limited and holds only at the first arrival of the seismic waves, not at the latter arrivals.

Notably, the time window at \( \sim 6,300 \) s demonstrates relatively high correlation (Figure 10a; see Figure S12a for more detail), which might be due to the P wave associated with the Mw 7.4 aftershock (GCMT) which occurred \( \sim 1.4 \) hr after the Chile earthquake. We also examined the correlation between the pressure and vertical velocity with several different period bands (Figure S12), although it seems the correlation is low in longer period ranges (\( T > T_0 \sim 6 \) s).
6. Quantitative Comparison in Amplitudes of Tsunami and Seismic Waves

We finally investigate the quantitative relationship between the spectral amplitude and the background noise, using the 10 Hz sampled original data. The time window of 58.25 hr (∼0.1 × 222 s) is used for the Fourier transform. Figure 11 shows the spectral amplitudes of the onshore and offshore records for background (gray) and coseismic (black) signals. In order to quantitatively compare the amplitudes between each record, the units are converted to apparent pressure change ([Pa·s] = [(m⁻¹, kg·s⁻²·s)]. We assume that a sea height change of 1 cm equals a pressure change of 1 hPa (10² Pa) to convert to apparent pressure change for the tide gauge at Hanasaki. The vertical accelerations in the OBS at KOBS1 and onshore seismometer at KSRF are converted to the apparent pressure change by using Equation 6, with \( \rho_w = 1.03 \text{ g/cm}^3 \) and \( H_0 = 2,218 \text{ m} \) (water depth at KOBS1).

PG spectra (Figure 11a) demonstrate an amplitude increase for the period range of ∼60–5,000 s, which are associated with tsunamis. Conversely, in the tide gauge spectra, the significant increase in the tsunami amplitude is limited within the period range of ∼2,000–4,000 s (∼0.25–0.5 mHz; Figure 11b), which is due to the coastal site effect (e.g., Geist, 2018; Tanioka et al., 2019). The broadband amplitude increase in the PG spectra demonstrates that PGs are capable of detecting the wider period range of tsunamis and are much less affected by the coastal site effect than coastal tide gauges.

Our results further confirm that amplitude increases are related to the seismic waves in the spectra of the PG and seismometers (Figures 11a, 11c, and 11d). Small spectral amplitude increases at periods of ∼5 s correspond to the body waves (marked by green arrow). In the periods of ∼10–50 s (∼20–100 mHz), where the Rayleigh wave is dominant, we verify the signal amplitude increases in the PG (orange text). This Rayleigh wave-related amplitude increase can also be observed in the spectra of the OBS and onshore seismometer. The background noise level in the frequency band of Rayleigh waves in the KPG1 is small compared to the KOBS1. This suggests that the KPG1 can detect seismic Rayleigh wave signals with qualities similar to or better than the KOBS1 during this event. One challenge in the seafloor seismic observation is that the installation environment (i.e., we cannot fix the sensor to the ground) cannot be controlled. This can easily be achieved in the onshore seismic observation. The lower noise level in the PG in this frequency band may suggest that the PG will provide good supplementary information for seafloor seismic observation.

7. Future Applicability of the High-Sampling PG

We summarize the key results of the spectral analyses in Figure 12. Five key findings can be drawn from our analyses:

1. High-sampling-rate (≥1 Hz) is necessary for PGs to detect the dispersive Rayleigh waves in broadband periods (∼10–400 s, ∼2.5–100 mHz) and dispersive tsunamis (∼50–1,000 s, ∼1–20 mHz) very clearly.
2. Sampling rate of the typical DART system (∼1/15 Hz) is not sufficient to fully analyze the wave propagation processes of the seismic waves and tsunamis.
3. Seismometers (≥100 Hz sampling) can record seismic wave signals clearly but do not record tsunamis, even when installed at the ocean floor.
4. Coastal tiltmeters observe both seismic wave and tsunami signals but do not clearly record the dispersion.
5. Tide gauges cannot clearly detect tsunami dispersion or seismic waves.

In particular, Finding (1) cannot be explicitly indicated if the sampling rate of the PG is low (e.g., the third-generation DART system, 1/15 Hz; e.g., see Rabinovich & Eblé, 2015). Our result suggests that the high-sampling PG observation would be a good candidate for the alternative and backup tools of seismic wave observations. This corresponds with An et al. (2017) and Kubota et al. (2017) who also demonstrated that the seismic waves retrieved from the PG records greatly contribute to accurately determining the centroid moment tensor solution of moderate-sized offshore earthquakes. New, wide, and dense offshore observation networks incorporating OBSs and PGs were recently installed in the deep-sea region (e.g., Howe et al., 2019; Kanazawa et al., 2016; Kaneda et al., 2015; Kawaguchi et al., 2015; Mochizuki et al., 2016; Rabinovich & Eblé, 2015; Uehira et al., 2016). These new PG networks feature a higher sampling rate (≥1 Hz). Furthermore, the sampling rate of the next generation DART system (DART4G) will possibly be higher than the third-generation system (e.g., An et al., 2017; Angove et al., 2019; Rabinovich & Eblé, 2015). Utilizing the array of high-sampling PGs will enable us to analyze the ultrabroadband geophysical wave propagation,
including tsunamis and seismic waves. Developments of the offshore PG networks will facilitate the analysis of ultrabroadband pressure signals covering periods of $10^0$–$10^3$ s. PGs can further be employed for observing much longer time scale phenomena (e.g., Baba et al., 2006; Fukao et al., 2019; Inazu et al., 2012; Tonegawa et al., 2018; Wallace et al., 2016; Figure 12). Using these new PG networks will create new possibilities for understanding the geophysical wave propagation processes in the solid-fluid coupled system and the generation processes of earthquakes and tsunamis. Several studies have already begun to use an array of the high-sampling PGs for the broadband geophysical wave propagation analyses (Fukao et al., 2018, 2019; Mizutani & Yomogida, 2019; Sandanbata et al., 2018).

Real-time earthquake early warnings are especially important for developing nations exposed to a high tsunami risk (e.g., Mulia et al., 2019). For more reliable and early earthquake warnings, it is desirable to install

![Figure 11. Fourier amplitude for (a) KPG1 and (b) HANA and vertical components for (c) KOBS1 and (d) KSRF. The gray and black lines depict the power spectra with a time window of ~2.5 days before and after the 2010 Chile earthquake, respectively. Characteristic spectral bands of tsunami and seismic waves are shown by blue and red arrows, respectively. Note that the instrumental responses are not removed.](image)

![Figure 12. Schematic illustration of geophysical phenomena in the ocean. The geophysical phenomena which can be observed by the high-sampling-rate ocean-bottom PGs and other instruments are shown by colored arrows. Note that the application of PG to geodetic phenomena is shown by the dashed arrow, because this time scale range is not covered in this study.](image)
OBSs in these networks, despite the higher economic costs associated with construction. Nevertheless, this present study demonstrates that seismic wave signals can still be observed by high-sampling-rate PGs, even if they do not incorporate OBSs. Recently, Nakamura et al. (2019) estimated earthquake body-wave magnitudes based on high-frequency seismic wave signals in PGs for real-time data analyses. The present study demonstrates the significant potential of longer and wider period seismic signals in near-field high-sampling PG records for real-time estimation of the moment magnitude and centroid moment tensor (e.g., Kubota et al., 2017). The broadband observations of high-sampling PG are significantly valuable for real-time earthquake warnings and disaster mitigation.

8. Conclusions

We examined the performance of PGs in a wide frequency range by investigating the spectral features in the PG records in northern Japan associated with the 2010 Chile earthquake and comparing these with nearby onshore and offshore instruments. We calculated the spectrograms and revealed that the PGs clearly detected the wave trains due to body waves, Rayleigh waves, and tsunamis. The dispersive features were visibly recognized in the Rayleigh wave (periods covering ~10–400 s) and tsunami (~60–5,000 s) wave trains, which were explained by the propagation theories of seismic waves and tsunamis, respectively. The quantitative comparison between the pressure change and the vertical acceleration demonstrated that the dynamic relationship holds for ~3 hr from the origin time, whereas the relationship between the pressure change and the vertical velocity at higher period range (less than ~6 s) holds only at the first P wave arrival. This validity was confirmed in time domain from the real observation. Similar results seen for multiple earthquakes, regardless of their very different source-station distance and earthquake magnitude, suggest the robustness of the observed relation. This study demonstrated that high-sampling PGs can observe the broadband seismic wave and tsunami signals. In particular, seismic wave signals can be detected with similar quality as the OBS. The broadband geophysical wave observations of the high-sampling PG are important for furthering our understanding of the geophysical analyses and developing practical disaster mitigations.

Data Availability Statement

Coastal tide gauge data are available at the Intergovernmental Oceanographic Commission (IOC)’s website (http://www.ioc-sealevelmonitoring.org). PG and OBS data are acquired by Japan Agency for Marine-Earth Science and Technology (JAMSTEC, http://www.jamstec.go.jp/scdc/top_e.html). DART tsunami data are available at the website of the National Oceanic and Atmospheric Administration (NOAA, https://www.ngdc.noaa.gov/hazard/dart/2010chile_dart.html). Onshore broadband seismometer are available at the National Research Institute for Earth Science and Disaster Resilience (NIED)’s F-net (NIED, 2019; 10.17598/NIED.0005). Tiltmeter record used in this study, which is operated by NIED, is available online (10.17598/NIED.0018). We used TauP Toolkit (Crotwell et al., 1999) version 2.4.5 (https://www.seis.sc.edu/taupe) to calculate the theoretical arrival times. Figures in this paper were prepared using Generic Mapping Tools (GMT) software version 6.0.0 (Wessel et al., 2019).

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References

Aki, K., & Richards, P. G. (1980). Quantitative seismology: Mill Valley, CA: University Science Books.
Allgeyer, S., & Cummins, P. (2014). Numerical tsunami simulation including elastic loading and seawater density stratification. Geophysical Research Letters, 41, 2368–2375. https://doi.org/10.1002/2014GL059348
An, C., Cai, C., Zheng, Y., Meng, L., & Liu, P. (2017). Theoretical solution and applications of ocean bottom pressure induced by seismic seafloor motion. Geophysical Research Letters, 44, 10.272–10.281. https://doi.org/10.1002/2017GL075137
An, C., Shawn Wei, S., Cai, C., & Yue, H. (2020). Frequency limit for the pressure compliance correction of ocean-bottom seismic data. Seismological Research Letters, 91(2A), 967–976. https://doi.org/10.1785/0220190259
Angove, M., Arcas, D., Bailey, R., Carrasco, P., Coetzee, D., Fry, B., et al. (2019). Ocean observations required to minimize uncertainty in global tsunami forecasts, warnings, and emergency response. Frontiers in Marine Science, 6, 350. https://doi.org/10.3389/fmars.2019.00350
Baba, T., Hirata, K., Hori, T., & Sakaguchi, H. (2006). Offshore geodetic data conducive to the estimation of the afterslip distribution following the 2003 Tokachi-oki earthquake. Earth and Planetary Science Letters, 241(1–2), 281–292. https://doi.org/10.1016/j.epsl.2005.10.019
Blackman, R. B., & Turkey, J. W. (1958). The measurement of power spectra from the point of view of communications engineering. New York: Dover Publications.
Blom, N. A., Deuss, A., Paulsen, H., & Waszek, L. (2015). Inner core structure behind the PKP core phase triplication. Geophysical Journal International, 201(3), 1657–1665. https://doi.org/10.1093/gji/ggv103
Mochizuki, M., Kanazawa, T., Uehira, K., Shimbo, T., Shiomi, K., Kunugi, T., et al. (2016). *S-net project: Construction of large scale seafloor observatory network for tsunamis and earthquakes in Japan*. Paper presented at AGU Fall Meeting 2016, American Geophysical Union, San Francisco.

Mula, I. E., Gusman, A. R., Williamson, A. L., & Satake, K. (2019). An optimized array configuration of tsunami observation network off Southern Java, Indonesia. *Journal of Geophysical Research: Solid Earth*, 124, 9622–9637. https://doi.org/10.1029/2019JB017600

Mungov, G., Eblé, M., & Bouchard, R. (2013). DART® Tsunameter retrospective and real-time data: A reflection on 10 years of processing in support of tsunami research and operations. *Pure and Applied Geophysics*, 170(9–10), 1369–1384. https://doi.org/10.1007/s00024-012-0477-5

Nakamura, T., Araki, E., Takahashi, N., Suzuki, K., Yamamoto, S., Kofunaga, M., & Noda, S. (2019). Magnitude estimation by using ocean-bottom pressure data. Paper presented at JpGU Meeting 2019, Japan Geoscience Union, Makuhari, Japan.

National Research Institute for Earth Science and Disaster Resilience (NIED). (2019). *NIED F-net. National Research Institute for Earth Science and Disaster Resilience*. (NIED). (2019).

Nakamura, T., Araki, E., Takahashi, N., Suzuki, K., Yamamoto, S., Kofunaga, M., & Noda, S. (2019). Magnitude estimation by using ocean-bottom pressure data. Paper presented at JpGU Meeting 2019, Japan Geoscience Union, Makuhari, Japan.

Nishida, K., Maeda, T., & Fukao, Y. (2019). Seismic observation of tsunami at island broadband stations. *Journal of Geophysical Research: Solid Earth*, 124, 1910–1928. https://doi.org/10.1029/2018JB016833

Nosov, M. A., & Kolesov, S. V. (2007). Elastic oscillations of water column in the 2003 Tokachi-oki tsunami source: In-situ measurements and 3-D numerical modelling. *Natural Hazards and Earth System Sciences*, 7(2), 243–249. https://doi.org/10.5194/nhess-7-243-2007

Nosov, M. A., Kolesov, S. V., Denisova, A. V., Alekeev, A. B., & Levin, B. V. (2007). On the near-bottom pressure variations in the region of the 2003 Tokachi-oki tsunami source. *Oceanology*, 47(1), 22–37. https://doi.org/10.1134/S0001437007010005

Okada, Y., Kasahara, K., Hori, S., Obara, K., Sekiguchi, S., Fujiwara, H., & Yamamoto, A. (2004). Recent progress of seismic observation networks in Japan - hi-net, F-net, K-NET and KiK-net. *Earth, Planets and Space*, 56. https://doi.org/10.1186/BF03353676

Pedlosky, J. (2013). *Waves in the ocean and atmosphere: Introduction to wave dynamics*. Berlin: Springer Berlin. https://doi.org/10.1007/978-3-662-05131-3

Poujardain, A., Heinrich, P., Hébert, H., Schindelé, F., Jamelot, A., Reynond, D., & Sugioa, H. (2018). Traveltime delay relative to the maximum energy of the wave train for dispersive tsunami propagating across the Pacific Ocean: The case of 2010 and 2015 Chilean tsunamis. *Geophysical Journal International*, 214(3), 1538–1555. https://doi.org/10.1093/gji/ggy200

Rabinovich, A. B., & Eblé, M. C. (2015). Deep-ocean measurements of tsunami waves. *Pure and Applied Geophysics*, 172(12), 3281–3312. https://doi.org/10.1007/s00024-015-1059-1

Rabinovich, A. B., Thomson, R. E., & Fine, I. V. (2013). The 2010 Chilean tsunamis off the west coast of Canada and the northwest coast of the United States. *Pure and Applied Geophysics*, 170(9–10), 1529–1565. https://doi.org/10.1007/s00024-012-0541-1

Saito, T. (2017). Tsunami generation: Validity and limitations of conventional theories. *Geophysical and Applied Geophysics*, 170(9–10), 1529–1565. https://doi.org/10.1007/s00024-012-0541-1

Saito, T. (2019). *Tsunami generation and propagation*. Tokyo: Springer Japan. https://doi.org/10.1007/978-4-431-56850-6

Saito, T., Matsuoka, T., Obara, K., & Baba, T. (2010). Dispersive tsunami of the 2010 Chile earthquake recorded by the high-sampling-rate ocean-bottom pressure gauges. *Geophysical Research Letters*, 37, L23303. https://doi.org/10.1029/2010GL045290

Sandanbata, O., Watada, S., Satake, K., Fukao, Y., Sugioa, H., Ito, A., & Shiobara, H. (2018). Ray tracing for dispersive tsunami and source amplitude estimation based on Green’s law: Application to the 2015 volcanic tsunami earthquake near Torishima, south of Japan. *Pure and Applied Geophysics*, 175(4), 1371–1385. https://doi.org/10.1007/s00024-017-1746-0

Satake, K., Baba, T., Hirata, K., Iwasaki, S. I., Kato, T., Koshimura, S., et al. (2005). Tsunami source of the 2004 off the Kii Peninsula earthquakes inferred from offshore tsunami and coastal tide gauges. *Earth, Planets and Space*, 57(3), 173–178. https://doi.org/10.1186/BF03515111

Shearer, P. M. (2009). *Introduction to seismology* (2nd ed.). New York: Cambridge University Press. https://doi.org/10.1017/CBO9780511841552

Tanioka, Y., Shibata, M., Yamanaka, Y., Gusman, A. R., & Ioki, K. (2019). Generation mechanism of large later phases of the 2011 Tohoku-oki tsunami causing damages in Hakodate, Hokkaido, Japan. *Progress in Earth and Planetary Science*, 6(1), 30. https://doi.org/10.1186/s40645-019-0278-x

Tilov, V., Rabinovich, A. B., Mosjid, H. O., Thomson, R. E., & Gonzalez, F. I. (2005). The global reach of the 26 December 2004 Sumatra tsunami. *Science*, 309(5743), 2045–2048. https://doi.org/10.1126/science.1114576

Tonegawa, T., Fukao, Y., Shiobara, H., Sugioa, H., Ito, A., & Yamashita, M. (2018). Excitation location and seasonal variation of transoceanic infragravity waves observed at an absolute pressure gauge array. *Journal of Geophysical Research: Oceans*, 123, 40–52. https://doi.org/10.1002/2017JC013488

Tsuchiya, H., & Ohta, Y. (2014). Review on near-field tsunami forecasting from offshore tsunami data and onshore GNSS data for tsunami early warning. *Journal of Disaster Research*, 9(3), 339–357. https://doi.org/10.20965/jdr.2014.p0339

Uehira, K., Kanazawa, T., Mochizuki, M., Fujimoto, H., Noguchi, S., Shinbo, T., et al. (2016). *Outline of seafloor observation network for earthquakes and tsunamis along the Japan Trench (S-net)*. Paper presented at EGU General Assembly 2016, European Geosciences Union, Vienna, Austria.

Wallace, L. M., Webb, S. C., Ito, Y., Mochizuki, K., Hino, R., Henrys, S., et al. (2016). Slow slip near the trench at the Hikurangi subduction zone, New Zealand. *Science*, 352(6286), 701–704. https://doi.org/10.1126/science.aaf2349

Watada, S., Kusumoto, S., & Satake, K. (2014). Travel time delay and initial phase reversal of distant tsunamis coupled with the self-gravitating elastic Earth. *Journal of Geophysical Research: Solid Earth*, 119, 4287–4310. https://doi.org/10.1002/2013JB010841

Webb, S. C. (1998). Broadband seismology and noise under the ocean. *Reviews of Geophysics*, 36(1), 105–142. https://doi.org/10.1029/97RG02287

Webb, S. C., & Crawford, W. C. (1999). Long-period seafloor seismology and deformation under ocean waves. *Bulletin of the Seismological Society of America*, 89(6), 1535–1542.

Webb, S. C., & Nooner, S. L. (2016). High resolution seafloor absolute pressure gauge measurements using a better counting method. *Journal of Atmospheric and Oceanic Technology*, 33(9), 1859–1874. https://doi.org/10.1175/JTECH-D-15-0114.1

Wessel, P., Luis, J. F., Ujeda, L., Scharroo, R., Wobbe, F., Smith, W. H. F., & Tian, D. (2019). The Generic Mapping Tools Version 6. *Geochemistry, Geophysics, Geosystems*, 20, 5556–5564. https://doi.org/10.1029/2019GC008515