Tectonic Tremors in the Northern Mexican Subduction Zone Remotely Triggered by the 2017 Mw8.2 Tehuantepec Earthquake

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Full paper

Keywords: 2017 Tehuantepec earthquake, tectonic tremor, remote triggering, wavefield simulation

DOI: https://doi.org/10.21203/rs.3.rs-73471/v2

Abstract

Surface waves from the 2017 Mw8.2 Tehuantepec earthquake remotely triggered tectonic tremors in the Jalisco region, approximately 1000 km WNW in the northern Mexican subduction zone. This is the first observation of tremor triggering in this region and one of the largest known examples of a triggered tremor in the world. Although prior studies have found tectonic tremors triggered by teleseismic waves in subduction zones and plate boundaries, further investigation of tremor triggering is crucially important for understanding the causative mechanism. We calculate the stress and strain changes across the three-dimensional plate interface attributable to seismic waves from the earthquake by full wavefield simulation. The maximum magnitude of the dynamic strain tensor eigenvalues on the plate interface, where tremors likely occur, is approximately 10 -6 . The subducting slab geometry effectively amplifies triggering waves. The triggering Coulomb failure stress changes resolved for a thrust fault plane consistent with the geometry are estimated to be approximately 10-40 kPa. The relationship between the triggering stress and triggered tremor amplitude may indicate that the [[EQUATION]] of the rate-state-dependent friction law is 10 to 100 kPa.

Introduction

Large earthquakes can remotely trigger tectonic tremors as well as earthquakes at great distances during the passage of seismic waves through regions close to failure. Previous studies have found tectonic tremors immediately triggered by teleseismic waves in subduction zones and at plate boundaries worldwide, including the Nankai subduction zone (e.g., Miyazawa and Mori 2005, 2006), the Cascadia subduction zone (e.g., Rubinstein et al. 2007), Taiwan (e.g., Peng and Chao 2008), the Hikurangi subduction zone (Fry et al. 2011), the Mexican subduction zone (Zigone et al. 2012), and the San Andreas Fault (e.g., Ghosh et al. 2009; Hill et al. 2013; Peng et al. 2009), where ambient tremors have also been reported. Moreover, there are observations of tremors triggered in different tectonic settings where ambient tremors have not yet been detected (Aiken et al., 2013, 2015, 2016; Obara 2012; Peng et al, 2013). The amplitudes of triggered tremors are highly correlated with the peaks of passing waves (Miyazawa and Brodsky 2008; Gomberg 2010). These findings are not only intriguing because transient stress changes can stimulate a fault to slip but also remarkably important for understanding the mechanism that causes tectonic tremors, which are still enigmatic phenomena at plate interfaces. For example, Rubinstein et al. (2007) and Miyazawa and Brodsky (2008) showed that movements enhancing the plate movement played an important role in triggering tremors, indicating the mechanism of slip on the plate interface.

We show that tectonic tremors in the northern Mexican subduction zone, i.e., Rivera and Cocos subduction zones beneath the Jalisco region, where tremor triggering has not been reported previously, were remotely triggered by the 2017 Mw8.2 Tehuantepec earthquake approximately 1000 km to the southeast. The 2017 Tehuantepec earthquake occurred at 04:49:46.7 on September 8, 2017 (UTC) (global centroid moment tensor (CMT) catalog), in the southern Mexican subduction zone with a mechanism of normal faulting within the Cocos Plate and ruptured the whole thickness of the Cocos Plate down to the mantle (e.g., Melgar et al. 2018; Suárez et al. 2019) (Figure 1). Notably, the 2017 Tehuantepec earthquake is also known to have triggered tremors along the western coast of North America (Chao et al. 2017) and activated tremors in the southern Mexican subduction zone (Husker et al. 2019). In the southern Mexican subduction zone, tremors were triggered by the 2010 Mw8.1 Maule earthquake accompanied by slow slip (Zigone et al. 2012). Since triggered tremors previously found around the world are located in different tectonic settings, a comparison of the triggering process is required to help us comprehensively clarify the source mechanism. This study investigates the characteristic triggering process by calculating strain and stress changes at depth using a forward technique of elastic wave simulation for a three-dimensional structure and by estimating the parameter of a friction law from the observed waves.

Detection Of Triggered Tremors

The triggered tremors with dominant periods ≤1 s during the passage of surface waves with dominant periods of approximately 20 s can be detectable by simple signal processing with high-pass filtering. Figure 2 shows the velocity waveforms for the Mw8.2 Tehuantepec earthquake at station CJIG of the National Seismological Service (SSN) of Mexico (SSN 2017) (Figure 1). The high-pass filtered waveforms at 8 Hz in Figure 1a show the tremor recurrently and periodically triggered during the passage of surface waves (Figure 2b). Three tremors were triggered in phase with passing surface waves from 310 s to 380 s after the earthquake initiation, while they seemed to be followed by further tremor activity with smaller amplitudes, including two minor tremors. These characteristics are very similar to the dynamically triggered tectonic tremors previously found. Comparing the high- and low-frequency components, we can identify that the first-arrival high-frequency waves from 140 s to 240 s represent the P-wave arrival and its coda waves, and the second ones from 250 s to 300 s correspond to the arrival of the S-wave. These high-frequency signals at the arrivals of P- and S-waves were also observed at other stations. Both P- and S-waves are known to be capable of triggering tectonic tremors (Miyazawa 2012; Hill et al. 2013); however, we could not distinguish tremors from teleseismic body waves in the present study. The amplitude of large triggered tremors here observed exceeds 1 μm/s, which is roughly one order of magnitude larger than that of the background tremors.

We could not identify any transiently triggered tremors at other tremor regions in the Mexican subduction zone, e.g., Oaxaca and Guerrero, close to the 2017 Tehuantepec earthquake, where Zigone et al. (2012) found triggered tremor, using both broadband and strong-motion records, because the codas of body waves arrive together with the surface waves and hinder the detection of tremors by high-pass filtering. Other techniques, such as the reference spectrum method (Miyazawa 2012), might be useful for detection; however, in this case, a dense seismic network for a statistical test and a reference event are required. This situation does not mean that no transiently triggered tremors occurred in other areas because Husker et al. (2019) found clear signs of tremor burst, particularly in Oaxaca subsequent to the 2017 Tehuantepec earthquake, and the activity might have started during the passage of the seismic waves.

The locations of the triggered tremors were not obtained because of the relatively low spatial coverage of the seismic network. The two closest stations to CJIG are located within approximately 180 km to 220 km, which are too far to observe these triggered tremors. Further, because CJIG is the only station that recorded the triggered tremors, they likely occurred in the region of background tremor activity, similar to the results of previous studies (Brudzinski et al. 2016; Idehara et al. 2014, Maury et al. 2018). The magnitudes are also not available, although the tremors that exceed 1.5 μm/s in amplitude could be the largest triggered tremors ever recorded in the world.

Full Wavefield Simulation For The 2017 Tehuantepec Earthquake

We examine the strain and stress changes on the plate interface, where tremors likely occur, associated with passing seismic waves from the 2017 Tehuantepec earthquake. For this purpose, the inverse approach using observed waveforms at a single station near the source via transport kernels in a one-dimensional structure (Miyazawa and Brodsky 2008; Miyazawa 2019) is applicable. However, the exact location of the triggered tremors is unknown, and the triggering wave propagates along the trench and is significantly affected by the structure of the subducting slab. Therefore, we first obtain an overview of the spatial distribution of these changes from the full waveform simulation in the three-dimensional structure. We use an open-source seismic wave propagation code (OpenSWPC) by Maeda et al. (2017) for elastic wave simulation. This approach solves the equations of motion in three-dimensional Cartesian coordinates with viscoelastic constitutive equations using a finite difference method. The spatial grid size is 1 km in the horizontal and vertical directions, and the time step is 0.05 s. The computational region has an area of 1000 km × 2000 km and extends from the topographic surface to a depth of 1000 km (Figure 1). The distance is measured based on the Gauss-Krüger coordinate system, where the source location of the Tehuantepec earthquake is set as the origin. The geometrical model is generated by combining ETOPO1 for topography (Amante and Eakins 2009), CRUST 1.0 (Laske et al. 2013) for the crust including the Moho, and Slab2 (Hayes et al. 2018) for the subducting plate. The subduction zone geometry model of Slab2 for Central America is extrapolated to cover a wider area for the numerical simulation using the surface command in the Generic Mapping Tool (GMT) (Wessel and Smith 1998) with a tension factor of 0.25. Seawater fills from the ocean floor to the surface. We do not employ spherical structures of the earth. Table 1 shows the parameters used for the structure, where we refer to and modify the standard values in subduction zones from CRUST 1.0 and OpenSWPC. They are in agreement with the tomographic structures below central Mexico (Iglesias et al., 2001, 2010). We assume a point source obtained by the W-phase inversion (Ye et al. 2017) with a moment rate function of the Herrmann function. Table 2 shows the source parameters of the Tehuantepec earthquake, where the rise time is determined through trial and error.

Figure 2b shows the recorded and simulated waveforms at CJIG, which were bandpass filtered at 0.01 to 0.10 Hz. The triggered tremors shown in Figure 2a are in phase with the large-amplitude surface waves, but the consistency may not be seen with respect to amplitude. Note that the times of the simulated waves were shifted by -16 s considering the duration time. The simulated waveforms are generally consistent with recorded data for all the components except for the amplitudes of higher frequency surface waves after 350 s, but the reconstruction of the full wavefield is particularly difficult in a subduction zone (e.g., Takemura et al. 2019). Similar trends can be seen at the other stations along the trench. Considering the radiation pattern of surface waves, as the station is located near the extension of a nodal plane, the vertical and radial components are highly sensitive to the strike direction. Since the simulated P-waves are consistent with the recorded ones and the simulated Love waves cannot reconstruct the high amplitudes after 350 s, the differences are likely caused by the uncertainties in the three-dimensional structures, including heterogeneities, and by the assumption of a point source model to a certain extent. We examine the wavefield simulation using finite fault models with slip distributions, while there is no improvement of simulated waveforms at CJIG to fit the observed ones for this frequency range. The tomography studies on phase and group velocity inversion using surface wave observed in Mexico show strong horizontal heterogeneities in the crust between the 2017 event location and CJIG (e.g., Gaite et al. 2012; Castellanos et al. 2018), which are sensitive to the 10 s to 20 s surface wave features. Furthermore, the spherical structure, which is not modeled in the simulation, may not be negligible at very far fields and for very long periods, although this factor may have less influence than the reasons mentioned above. The frequencies analyzed in the present study are between 0.01 and 0.10 Hz, where dominant frequencies of waves are 0.05-0.10 Hz, which are more influenced by the crustal structure and slab geometry than those from the spherical effects. Therefore, the present results are sufficiently useful to investigate the characteristics of the wavefields and the spatial distribution pattern of strain changes on the plate interface in the following section, since our purpose is not to examine the exact structure or source model.

An additional movie file shows the simulated wavefields at a depth of 7 km [Additional file 1]. The seismic waves seem to propagate slightly slower onshore than offshore due to the slow velocity structure in the upper mantle. The waves propagating along the onshore toward the northwest direction, where the triggered tremors were observed, are accompanied by longer codas, which are caused by the wedge mantle structure between the crust and the subducted slab. This structure can also increase the amplitudes of surface waves, while the observed large amplitudes in the coda are not fully simulated.

Strain And Stress Estimation

Based on the full waveform simulation, we obtain the maximum magnitudes of eigenvalues of the dynamic strain tensor given as the representative maximum strain load at each position. Figure 3a shows the spatial distribution of maximum strain changes on the plate interface of Slab2. The value of strain changes generally decreases with distance from the epicenter. Anomalies of substantial strain changes are observed in the Guerrero and Jalisco regions to the northwest along the trench, where the plate interfaces are shallower than those in other regions. In the Jalisco region, the stress/strain changes have larger values within the tremor region than in the shallower or deeper regions (Figure 3b). The maximum strain changes in the Jalisco tremor region are estimated at approximately 1×10^{-6}, which is one to two orders of magnitude larger than the changes due to the earth tide. Note that these values may underestimate the actual ones because the simulated maximum amplitudes of surface waves underestimate the actual maximum amplitudes.

We then estimate the stress changes resolved for the mechanism of the tectonic tremor. We assume that the tremor occurs on the plate interface with a mechanism having a strike of 319°, a dip of 29° and a rake of 90° for a fault plane consistent with the subducting slab geometry. Since the exact location of the triggered tremor is not available, we substitute the median epicenter of ambient tremors using the tremor catalog in Jalisco by Idehara et al. (2014), which is located approximately 50 km to the east of station CJIG (Figure 3b). We estimate the change in the Coulomb failure stress at the hypocenter on the plate interface, $\Delta\tau + \mu'\Delta\sigma_n$, where the effective friction coefficient μ' is assumed to be 0.2 and $\Delta\tau$ and $\Delta\sigma_n$ are the shear stress change in the slip direction and the normal stress change on the fault plane (negative for compression), respectively (Figure 4). The waveforms are bandpass filtered at 0.01 to 0.10 Hz. The Lamé parameters are set by the structure model in Table 1 for the synthetic case (Figure 4a). Since we cannot completely reconstruct the full wavefields as noted above, an inverse approach using observed waveforms is useful to make the results more robust. We use a linear kernel approach that computes continuous waveforms spanning the full spectrum at depth (Miyazawa and Brodsky 2008; Miyazawa 2019) to calculate the dynamic stress change resolved for the tremor mechanism at the depth of the plate interface. In this method, the stress changes associated with passing surface waves can be obtained from surface observations. For simplicity, we assume fundamental mode Rayleigh and Love waves propagating in an isotropic elastic medium. The Lamé parameters λ and μ are both 66 GPa at this depth, and the other parameters are the same as those described in Miyazawa (2019). Figure 4b shows the stress changes beneath station CJIG at a depth of 45 km. A systematic time difference in the arrival times of stress peaks between the simulation and estimation based on the observations occurs because of the difference in the location where we obtained the stress changes. As we cannot locate the triggered tremor source, we do not correct for travel time from the tremor source to station CJIG to directly compare the stress changes and corresponding tremor amplitudes. However, the triggering stresses for the three tremors in Figure 2 may correspond to the three peaks of 10 kPa to 40 kPa at times from 300 s to 370 s for the simulation (Figure 4a) and of 14.7 kPa, 28.4 kPa, and 22.6 kPa from 310 s to 380 s for the estimation from the observations (Figure 4b). The differences in these values appear primarily because the parameters and structures assumed in the simulation and estimation models are different and probably because station CJIG is located closer to the trench where the amplitudes of seismic waves decease more than those in the tremor region.

Due to the large ambiguous location of triggered tremors, we also examined the stress changes at the other location. Since the amplitude of triggered tremors is very large, they may occur as close as possible in the region of high background tremor activity, located approximately 30 km to the north-east of the station CJIG (Figure 3b). The depth of the plate interface is 3 km shallower than the previous one in this model. The stress changes for this location on the plate interface are shown in Figure 4a by dotted lines. The amplitudes are slightly larger than but comparable to those for the median location, primarily due to the difference in depths. These results indicate that there are no significant differences in the triggering stresses in the background seismicity area. Both the locations are reasonable in terms of travel time from tremor source to the station CJIG, if each tremor is triggered by the peak stress changes as mentioned above, although we cannot yet constrain the location only from a single station.

Large stress changes could likely trigger tremors; however, the observation of three tremors may not be enough to clarify the relationship between the amplitudes of triggered tremors and triggering stress, as shown by Miyazawa and Brodsky (2008). Since the tremor hypocenters are not obtained, we simply compare the peak values of tremor amplitude in the vertical component and the triggering stress (Figure 5), where the peak tremor amplitudes are measured from the envelope of high-pass filtered waveforms (Figure 2a) with 1 Hz lowpass filtering and the stress changes are estimated by the waveforms recorded at CJIG (Figure 4b). Two more small triggered tremors following three large ones are also used. Simulated values are not used here because the largest surface waves are not reconstructed. If we assume that the triggered tremors occurred at the same place, the relationship between the triggered tremor amplitude and the triggering stress derived by Miyazawa and Brodsky (2008) can be applicable to the peak values. The peak tremor amplitude, A_{tremor}, is represented by

$$A_{tremor} = Ca\sigma\left[\exp\left(\frac{\Delta CFF}{a\sigma}\right) - 1\right],$$

where a is an empirically derived constant, σ is the background stress (Dieterich 1994) and C is a constant. The parameters are obtained as $a\sigma$ = 56.84 kPa and C = 0.01965, which may not be robust due to the small number of datasets. The model curve and those when $a\sigma$ is 1, 10 and 100 kPa are shown in Figure 5, while there is no significant difference for $a\sigma >$ 100 kPa in this stress range. The relationship in our study may follow this model when $a\sigma$ ranges from 10 kPa to 100 kPa (Figure 5), which is consistent with the case of tremor triggering found in the Nankai region (Miyazawa and Brodsky 2008).

Other Teleseismic Events

We investigate the dynamic triggering of tremors by teleseismic waves from other large earthquakes from 2011 to 2019. From the ANSS earthquake catalog, we selected M ≥ 7.0 teleseisms, of which the peak ground velocities (PGVs) are expected to exceed 1.0 mm/s for surface waves of 20 s at CJIG. The PGV is estimated by using an amplitude-magnitude relationship (e.g., van der Elst and Brodsky 2010; Chao et al. 2013). Table 3 shows that the selected six large earthquakes and the PGV values are the measured maximum amplitudes of the vector velocity from the three components bandpass filtered from 0.01 Hz to 0.10 Hz recorded at CJIG. We cannot find any evidence of triggered tremors due to these earthquakes, by using the same signal processing as mentioned above for the 2017 Tehuantepec earthquake. The back azimuth is also an important factor affecting dynamic stresses (e.g., Miyazawa et al 2008; Hill 2012); however, the back azimuth of the Tehuantepec earthquake is 111 degrees, which is similar to the angle of the other five earthquakes. The observation that the triggered tremor only occurred for the 2017 Tehuantepec earthquake may be due to the large amplitude of the triggering waves. However, the amplitude of waves that triggered the first tremor in 2017 (Figure 2) is comparative to PGVs of the 2012 Mw7.4 and 2014 Mw7.2 earthquakes. There may be no clear threshold to trigger tremors if the amplitude of potential triggering waves at the surface is less than 10 mm/s.

Discussion

The main reasons why the tectonic tremors were triggered by the surface waves from the 2017 Tehuantepec earthquake include the large stress/strain changes in the tremor region. As shown in Figure 3, substantial strain changes of approximately 1×10^{-6} along the background tremor region in Jalisco indicate that the surface waves can effectively trigger tremors. The Coulomb failure stress changes effective for the triggering range from 10 to 40 kPa, which is consistent with previous results on tremor triggering in other regions (Aiken et al. 2013, 2015; Chao and Obara 2016; Miyazawa and Brodsky 2008; Peng et al. 2009). These large stress/strain changes are one to two orders of magnitude larger than those from earth tides and present abnormally large values. Note that these high values are also produced by the geometry of the subduction zone and the spatial relationship between the tremor region and the triggering earthquake, as seen in the full wavefield simulation. The results may indicate that the stress changes associated with passing P- and S-waves, which are comparable to the lowest triggering stress from the surface waves, were also capable of triggering the tremor. To date, we have not found clear evidence of triggering from body waves generated by the 2017 Tehuantepec earthquake.

One of the reasons why the tremors were found to be triggered in Jalisco and not in other closer regions due to the 2017 Tehuantepec earthquake is that the waves propagating along the trench are amplified due to the shallower plate interface when waves come closer to Jalisco. This should also occur in Guerrero, where the plate interface is shallower than those in other regions along the trench, while triggered tremors were not found due to poor signal-to-noise ratios. This also implies that the back azimuth from the source to the tremor region plays an important role in dynamic triggering. To date, there is no evidence to support this idea due to the small number of observations that can cause triggerable PGVs (Table 3), and the triggering stress level most likely matters.

The amplitude of triggered tremors of 1 μm/s is very large, roughly one order of magnitude larger than the amplitude of background ambient tremors, and one of the largest triggered tremor ever reported. The triggered tremors during the passage of surface waves are well known to sometimes have significant amplitude corresponding to large triggering stress because of a large slip area or alternatively because of the simultaneous occurrence of numerous events (Miyazawa and Brodsky 2008) and/or the triggering associated with slow slip (Miyazawa 2012). In the present case, we do not know whether there were slow slip events in Jalisco subsequent to the 2017 Tehuantepec earthquake (Husker et al. 2019; personal communication) because the tremor catalog used to detect slow slip events cannot be verified at only the single station CJIG. The transient stress changes either by body waves or surface waves indeed induced slip on the fault, thereby causing tremors, but the tremor activity might not have been followed by aftershock-like activity and might have returned to a low background level because the fault was not initially close to failure. This interpretation is similar to the characteristics of triggered seismicity with relatively large magnitude, as seen for the case of a dynamically triggered moderate earthquake, where the aftershocks abruptly returned to background levels within a month (Miyazawa 2016). If there was either triggered or background slow slip in Jalisco only when the 2017 Tehuantepec earthquake occurred, although such event has not been detected yet, this is consistent with no observation of triggered tremor by the similar amplitude of triggering waves from other large earthquakes. The tremor activity in Oaxaca subsequent to the 2017 Tehuantepec earthquake may have been caused by both dynamic and static stress changes (Husker et al. 2019).

Conclusions

The discovery of triggered tremors from the 2017 Mw8.2 Tehuantepec earthquake in the Jalisco region, as well as in the Oaxaca region as previously reported, indicates the possibility of the dynamic triggering of tremors in any other tremor region. Moreover, the relationship between the triggering stress and the triggered tremor amplitude seems to be similar to previous findings. This result suggests that more investigation of triggered tremors through continuous observations can systematically provide us with important information about parameters in a friction law on the plate interface, which might be useful for earthquake cycle simulation in the subduction zone.

Declarations

Ethics approval and consent to participate

Not applicable

Consent for publication

Not applicable

List of abbreviations

CMT: centroid moment tensor

GMT: Generic Mapping Tools

PGV: peak ground velocity

Availability of data and materials

The SSN data are available at https://doi.org/10.21766/SSNMX/SN/MX. The geometry models, CRUST 1.0 and Slab 2, are downloadable from https://igppweb.ucsd.edu/~gabi/crust1.html and https://doi.org/10.5066/F7PV6JNV, respectively. The OpenSWPC is an open source code available at http://github.com/takuto-maeda/OpenSWPC.

Competing interests

The authors declare that they have no competing interests.

Funding

This work was supported by the Japan Science and Technology Agency (JST) and the Japan International Cooperation Agency (JICA) under its Science and Technology Research Partnership for Sustainable Development (SATREPS) project (JPMJSA1510) (MM and MS) and partially supported by JSPS KAKENHI Grant Number JP16K05537 and JSPS Grant Number JPMXS05G2900001 (MM). This work was partially supported by CONACYT PN-2015-639 project (MS).

Authors' contributions

MM contributed to the simulation and modeling, and MS involved in signal processing. Both authors discussed the results and commended on the manuscript.

Acknowledgements

The authors appreciate the thoughtful comments by Chastity Aiken and an anonymous reviewer. Discussion with Allen Husker and Naofumi Aso helped develop this study. The SSN data were obtained by the Servicio Sismológico Nacional (México), and station maintenance, data acquisition and distribution were provided by its personnel. For data analysis, we used the computer systems of the Earthquake and Volcano Information Center of the Earthquake Research Institute (ERI), University of Tokyo, Japan. The figures were generated using GMT (Wessel and Smith 1998).

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Tables

Table 1. Structure model used for the full wavefield simulation.

Layer	Density (g/cm^3)	Vp (km/s)	Vs (km/s)	Qp	Qs
Topography (ETOPO1)	2.11	2.70	1.20	200	100
Crust 1	2.67	6.20	3.70	680	400
Crust 2	2.69	6.60	3.90	680	400
Crust 3 (Moho)	3.20	7.50	4.40	850	500
Slab Top (Slab2)	2.65	5.90	3.50	425	250
Slab (Oceanic Moho)	3.20	8.00	4.60	850	500

Table 2. Source parameters of the 2017 Tehuantepec earthquake.

Longitude	Latitude	Depth	Mw	Strike	Dip	Rake	Rise time
265.69°	15.34°	50 km	8.21	313°	77.7°	-95.5°	32 s

Table 3. PGV of other teleseismic earthquakes recorded at CJIG from 2011 to 2019.

Origin time (UTC)	Mw	Depth [km]	Distance [km]	PGV (mm/s)	Back-azimuth (deg)
2011-03-11	9.1	29	10,523	0.6147	313
2012-03-20	7.4	20	794	6.4976	114
2012-11-07	7.4	24	1,528	0.7079	112
2014-04-18	7.2	24	489	4.1535	118
2017-09-19	7.1	48	698	0.6566	98
2018-02-16	7.2	22	824	1.0696	114