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► **To cite this version:**

Blandine Gardonio, David Marsan, Thomas Bodin, Anne Socquet, Stephanie Durand, et al.. Change of deep subduction seismicity after a large megathrust earthquake. 2023. hal-04260791

HAL Id: hal-04260791

<https://hal.science/hal-04260791v1>

Preprint submitted on 26 Oct 2023

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Article

Keywords:

Posted Date: June 27th, 2023

DOI: <https://doi.org/10.21203/rs.3.rs-3071456/v1>

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Additional Declarations: There is **NO** Competing Interest.

Change of deep subduction seismicity after a large megathrust earthquake

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Subduction zones are home to the world's largest and deepest earthquakes. Recently, large-scale interactions between shallow (0-60 km) and intermediate (80-150km) seismicity have been evidenced before megathrust earthquakes along with large-scale changes in surface motion. Large-scale deformation transients following major earthquakes have also been observed possibly due to a change in slab pull or to a bending/unbending of the plates, which suggests the existence of interactions between the deep and shallow parts of the slab. In this study, we analyze the spatio-temporal variations of the Japanese seismicity before and after the 2011/3/11 M_w 9.0 Tohoku-oki earthquake. We observe that the rate of the intermediate to deep (150-450km) seismicity underwent a deceleration of 55% south of the rupture zone and an acceleration of 30% north of it, coherent with the GPS surface displacements. This shows for the very first time how a megathrust earthquake can affect the stress state of the slab over a 2500 km lateral-range and a large depth range, demonstrating that earthquakes interact at a much greater scale than the surrounding rupture zone usually considered. Our vision of subduction zones must consequently be renewed as a whole, from shallow to large depths.

Recent studies suggest that large subduction earthquakes can affect the plate stress regime by showing either a synchronicity of the shallow and intermediate seismicity before large megathrust earthquakes^{1,2,3,4}, a slab wide deformation transient concomitant with deep and shallow earthquake swarms before a large deep-focus earthquake⁵ or a gravity field change^{6,7}. The observation of large-scale surface displacements reversals, several months before megathrust earthquakes also suggests the existence of deep precursors⁸. However, the observations of such large-scale interactions are rare and their mechanisms, and spatio-temporal extent remain poorly understood. To tackle this riddle, one solution is to analyze of the seismic productivity as a proxy for the slab stress conditions after a megathrust earthquake^{9,10}. In that respect, the 2011/3/11 M_w 9.0 Tohoku-oki earthquake is an ideal case study, since it is the best recorded megathrust earthquake of the instrumental period of seismology¹¹⁻¹⁶ and it occurred within one of the best-monitored regions of the world, Japan. Several studies conducted after the Tohoku-oki earthquake evidenced a preseismic phase^{1,6,17-22}. After the earthquake, there was a shift from predominant compression to tension in the forearc region off the coast²³⁻²⁷. The offshore GPS-Acoustic stations located above the main coseismic slip area have shown large landward post-seismic displacement due to an early viscoelastic relaxation of the oceanic mantle below the subducting plate^{28,29}. Furthermore, segments adjacent to the Tohoku-oki earthquake experienced a landward increase of surface velocity that could either be due to the acceleration of the subducting plate^{30,31} or to a viscoelastic response³². The Tohoku-oki earthquake is thus characterized by both a large-scale preparatory phase and a large-scale post-seismic response. Here, we investigate the intermediate and deep seismicity (from 150 to 450km depth and even 680km in the Izu-Bonin area) of the entire Pacific slab, over the 2,500 km long Japanese island arc. The spatio-temporal evolution of the seismicity rate λ (the number of earthquakes per unit of time) is analyzed using a probabilistic approach. Our study provides new observations on the consequences a large mega-thrust earthquake on intermediate and deep seismicity, and hence on the stress state of a large portion of the Pacific slab.

Observed variations in rate of intermediate and deep seismicity

Our aim is to investigate the temporal and spatial evolution of the background seismicity, as a proxy for changes in loading rates. We work on 21 years of earthquake activity located in the subduction zone of the Pacific slab, recorded by the JMA (Japanese Meteorological Agency) catalogue from 2000 to 2021. To avoid a signal mostly affected by aftershocks and foreshocks sequences, we removed dependent earthquakes following the method of Marsan et al. (2017)³³ (see Methods, Fig. S1 to S4). Keeping only the independent background seismicity³⁴ evidences temporal and spatial changes in loading rates (Fig. S5). 144 reference points are defined over a depth-range of 80km to 640km (colored dots in Fig. 1). For each point, we select the 200 closest earthquakes with magnitude ≥ 3.5 which corresponds to the magnitude of completeness at these depths. Note that there are fewer earthquakes in the Japan Sea so the distance between the reference points and the related earthquakes reaches 600km in this region whereas it is 300km for the other points (Fig. S6a). For every set of earthquakes, we investigate the temporal evolution of the seismicity rate given by $\lambda = N/\Delta t$, where N is the number of events occurring in the time period Δt (see Methods). We compute the relative change of seismicity rate λ at the time of the Tohoku-oki earthquake as $(\lambda_a - \lambda_b)/\lambda_b$, where λ_a and λ_b are the seismicity rates after and before (over 11 years) the Tohoku-oki earthquake. The effect of the Tohoku-oki earthquake on the seismicity rate at depth is far from being homogeneous over the Japanese islands and we observe three main areas of opposite behaviors (Fig. 1a). Underneath the island of Hokkaido, the seismic rate increases, up to 30% between 150 and 450km depth. This is confirmed by the stack of all the time series at latitude $> 42^\circ\text{N}$ (Fig. S6b). The maximum change is found at 200km at 45°N latitude (Fig. 2a and b, red curve). We detect no significant effect of the Tohoku-oki earthquake on the deep seismicity (Fig. 1a). The stack of the time series of reference points lying between 35°N and 42°N shows no change of the seismic rate at the time of the Tohoku-oki earthquake (Fig. S6c). The southern part of the Japanese island arc, from 26° to 35°N latitude, shows a decrease of the seismicity rate from -55 to -

10% with an average of -30% (Fig. 1a and S6d). The maximum change of -55% is found at 34°N latitude and 280km depth (Fig. 2a and b, blue curve). By increasing the length of the analyzed period, the maximum change underneath Hokkaido and the negative change in the south attenuate with time until 2017 (Fig. 1b). The attenuation in the south continues until 2021, while there is a down-dip migration of the acceleration pattern in the north (Fig. 1c).

Statistical significance of the variation in the seismicity rate

In order to test the robustness of the observations, we perform various statistical tests on the onset of the seismicity rate change, focusing on the 2000/3/11-2013/3/11 period when we see the most significant changes due to the Tohoku-oki earthquake. We use different statistical approaches (see Methods) assuming that the rate of seismicity follows a non-homogeneous Poisson process. First, we analyze the probability for a change in seismicity rate, assuming that it occurs at the time of the Tohoku-oki earthquake. To that aim, we follow the formalism of Marsan and Wyss (2011)³⁵. The probability density function (pdf) of the rate change r is shown in Fig. 2c. A rate change r less than or greater than 1 means a decrease or increase in seismicity rate, respectively. At 34°N latitude, the probability density for a rate change after the Tohoku-oki earthquake is maximum at $r_{\max}=0.45$ meaning that it is most likely to have undergone a decrease of seismicity rate in that region at that time (Fig. 2c, blue curve). Conversely, an increase of the seismicity rate under the island of Hokkaido is confirmed with $r_{\max}=1.27$ (Fig. 2c, red curve). It is possible to track this evolution by examining the pdf of the seismicity rate, λ before and after the Tohoku-oki earthquake, separately (Fig. S7). We compute the standard deviation of r , σ , for a change at every grid point (Fig. S8). Plotting r_{\max} , $r_{\max}-\sigma$ (Fig. S8 left) or $r_{\max}+\sigma$ (Fig. S8 right) leads to the same conclusions as we still see values close to 1 under Hokkaido for $r_{\max}-\sigma$ and a decrease of the earthquake rate around 34°N latitude for $r_{\max}+\sigma$. The high probability of having a change in seismicity rate at the time of the Tohoku-oki earthquake is thus confirmed whether we see an increase or a decrease of λ . We then focus on the probability of this change occurring at any other time. We again compute the pdf of r

using the same dataset but now exploring different times of change (Fig. S9 and S10). While r_{\max} , i.e. the value that maximizes the pdf, is stable between 2000 and Tohoku-oki, there is a significant change of slope of the maximum at the time of the Tohoku-oki earthquake (black lines in Fig. S10) implying that the regime changes at this precise moment only, both at 34°N and 44°N latitude. More specifically, $r_{\max} > 1$ between 2000 and the Tohoku-oki earthquake at 44°N latitude and $r_{\max} < 1$ at 34°N, showing the increase/decrease, respectively, of the seismicity rate at the time of the megathrust earthquake. Finally, we estimate the probability of having a change in the seismicity rate at any given time for the whole region by stacking the pdf of seismicity rate change calculated at each reference point (see Methods and Fig. 3). This probability shows a significant change at the time of the Tohoku-oki earthquake (Fig. 3a) indicating that a change in seismicity rate is most likely at the time of the Tohoku-oki earthquake, although its likelihood varies with latitude (Fig. 3b). Seeing no effect of the Tohoku-oki earthquake on the seismicity rate at the northern (above 45°N latitude) and southern (below 28°N latitude) tips confirms the spatial delimitation of our observations. Thus, the Tohoku-oki earthquake affected the stress state of the slab from the Izu-Bonin to the Hokkaido islands with a dichotomy between north (acceleration) and south (deceleration). We further clustered the 144 reference points according to their r_{\max} values at the time of Tohoku-oki earthquake applying the k-means clustering method (an iterative, data-partitioning algorithm that assigns n observations to exactly one of k clusters) considering 4 clusters (Fig. S11a). For each cluster thus formed, we display in Fig. S11b-e the mean of the pdf computed at each time step for all the points in the cluster. The first cluster is located underneath Hokkaido (Fig. S11b) and its r_{\max} value is maximum at the time of the mega-thrust (Fig. S11b). It corresponds to the area where the seismicity rate increases after the Tohoku-oki earthquake. Cluster 2 corresponds to locations where the seismicity rate is constant ($r_{\max}=1$, Fig. S11c). Cluster 3 gathers the southern points where there is a decrease of the seismicity rate after the megathrust earthquake. Its r_{\max} value is indeed minimal at the time of the Tohoku-oki earthquake. The same applies to

cluster 4 with a smaller decrease, r_{\max} being close to 1. Note that all the grid points in the northern and southern tips belong to either cluster 2 or 4 which are the ones showing the mildest effect of the Tohoku-oki earthquake, hence outlining indicating the spatial limit of the Tohoku-oki effect. This confirms the dichotomy between the northern and southern parts of the Pacific slab and the extent of the Tohoku-oki effect on the intermediate and deep seismicity.

After the Tohoku-oki earthquake, many small magnitude earthquakes went undetected, especially at great depths, masked by the high productivity of larger, shallower, aftershocks. To address this issue, we analyze again the JMA catalog taking a magnitude of completeness of 4.5 and considering the 70 closest earthquakes to each reference points (Fig. S12). We still see the positive and negative changes underneath Hokkaido and at 34°N latitude, respectively. Note that the seismicity rate change in the Izu-Bonin at depth is still negative but less statistically significant. We finally carry out the same exercise (taking $M_c = 4.5$ and the 70 closest events) using the USGS catalog and observe a strong increase underneath Hokkaido and a less visible decrease at 34°N latitude (Fig. S13).

Possible causes of changes in the seismicity rate

We have shown that the changes of seismicity rate are statistically significant on the entire Japanese Pacific subduction zone between 150 and 450 km depth, over a spatial extent spanning 2500km from 26° to 48° of latitude. Interestingly, the maximum and the minimum rate changes are both at major bends of the Pacific subducting slab³⁶ (Fig. 1). Furthermore, the dip is larger in the Southern bend, around 34°N latitude, where the maximum negative variation was recorded, than underneath Hokkaido, where the maximum positive variation was recorded. The analysis of the Coulomb stress transfer on the area of the rupture zone of the Tohoku-oki earthquake, i.e. at a smaller depth than this study (0-150km), highlights a cessation of the thrust earthquakes in the rupture zone that might last for centuries^{9,10}. To get a first order estimate of the elastic loading/unloading on the subduction interface, we compute the Coulomb stress change for two receiver points located on the slip

interface at depth, where we observe the maximum acceleration and deceleration of the seismicity rate (orange and yellow stars in Fig. 4a). To do so, we use COULOMB3.3^{37,38} and impose a 50m of slip on one dislocation located on the rupture zone of the Tohoku-oki earthquake. The receiver points are located according to the interface geometry³⁶ and we vary the rake angle from -180 and 180° for comparison with the rake values of faults in the surrounding area (Fig. 4b). A positive Coulomb stress change, i.e. a loading of the fault, is found for most dominant rake values at 44°N latitude, while a negative Coulomb stress change, an unloading, is found for most of the characteristics rakes at 34°N latitude coherent with an increase/decrease of the seismicity rate at these latitudes. Our GPS data analysis (Fig. 1) confirms the enhanced landward motion that follows the Tohoku-oki earthquake observed in the Hokkaido region^{28,29}, that induces an anticlockwise rotation of the velocity change and an enhanced trenchward motion in Honshu. This landward GPS velocity change in Hokkaido is located right above the maximum positive change of the seismicity rate observed at 44°N latitude. Furthermore, we see a possible second, larger-scale, clockwise rotation of the displacement in the south, at 34°N latitude, where we observe the maximum negative change. Therefore, there is a clear spatial correlation between the surface displacements and the changes in the seismicity rate which places constraints on the location of the changes at great depth that GPS alone cannot resolve. Note that a M_w 7.6 aftershock of the Tohoku-oki earthquake, which occurred offshore the Boso peninsula, and its subsequent postseismic slip could be responsible for part of this velocity change³⁹. Such rotation patterns have different characteristics according to their timing in the seismic cycle⁴⁰ and different mechanical models could explain the enhanced landward motion seen in Hokkaido : an apparent increase in coupling in regions adjacent to the megathrust^{30,31}; or a viscoelastic relaxation of both oceanic and continental plates caused by coseismic and postseismic fault slip^{28,29,41} that results in gear-train rotation patterns³².

Because the deep seismic response after the Tohoku-oki earthquake is both extremely fast and extensive, a rapid and large scale stress transfer along the slab is required and could be facilitated by

a low-viscosity channel, similar to the one modelled down to 135km to explain the post-seismic deformation following the Maule earthquake⁴². As the variation in seismicity rate disappears slowly with time, this low viscosity is likely a transient feature, that can be modeled as a Burgers rheology combining Maxwell and Kelvin moduli commonly invoked in visco-elastic post-seismic relaxation models, including after Tohoku-oki earthquake^{26,43,44}.

The Tohoku-oki earthquake has large-scale consequences on the very-deep (150-450km and deeper in the Izu-Bonin area) seismicity of the Pacific plate. We observe significant changes in deep seismicity rate together with changes in surface displacement rates over a 2500 km lateral-range and up to 450km depth, demonstrating that interactions between shallow and deeper parts of the slab exist at every stage of the seismic cycle. This questions our understanding and our vision of subduction zones that must be considered as a whole. In the future, taking greater depths into account in 3D rheological models and seismicity studies will bring more insights to the stress state of the slab and its continuity as a geodynamical object.

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Acknowledgments: Earthquake data were kindly provided by the Japan Meteorological Agency in cooperation with the Ministry of Education, Culture, Sports, Science and Technology. We acknowledge the help of the National research Institute for Earth science and Disaster Prevention, Tsukuba, for making available their waveform data. The authors would like to thank Cécile Lasserre, Pascal Bernard, Claudio Satriano, H el ene Lyon-Caen and Marianne M etois for insightful discussions. **Funding:** We acknowledge support from the European Research Council (grant number 681346-REALISM and 865963 DEEP-trigger). **Author contributions:** BG conceived the study. BG, DM, TB, SD, YR developed the statistical methods, AS and MR analysed the GPS data. All authors participated to the writing of the manuscript. **Competing interests:** Authors declare no competing interests. **Data and materials availability:** All data is available in the main text or the supplementary materials.

Supplementary Materials:

Materials and Methods

Figures S1-S13

Methods:

1) Declustering

We perform declustering of the earthquake dataset so to remove aftershocks that would otherwise contaminate the analyzed changes in seismicity rates. This is done based on the method already described in Marsan et al. (2017), which uses an epidemic-type model of seismicity with a spatially heterogeneous background rate. We here provide the details of the method. Earthquake occurrences are modeled as a number of earthquakes per unit time and unit area $\lambda(x, y, t)$, defined as the sum of two contributions :

$$\lambda(x, y, t) = \mu(x, y) + v(x, y, t)$$

in which v accounts for triggering by previous earthquakes, and μ is the activity that would occur in the absence of any such interactions, i.e., the background rate, which we here assume to depend on position (x, y) but not on time. Any earthquake with index i occurring at $\{x_i, y_i, t_i\}$, with magnitude m_i , triggers aftershocks with rate :

$$v(x, y, t) = \sum_{i|t_i < t} \frac{K e^{\alpha m_i}}{(t + c - t_i)^p} \cdot \frac{\gamma - 1}{2\pi} \cdot \frac{L_i^{\gamma-1}}{\left((x - x_i)^2 + (y - y_i)^2 + L_i^2 \right)^{\frac{\gamma+1}{2}}}$$

hence the product of the Omori-Utsu law with a power-law spatial density. Parameter values are fixed for all parameters based on either previous analyses or independent knowledge, except for K which depends on completeness magnitude and the regional level of seismicity, and that therefore cannot be fixed a priori (we describe below how the value of K is determined). We set α to 2.0, which is a typical median value when considering previous studies, $p = 1$ which corresponds to the simple Omori's law, $c = 10^{-4}$ days (the exact c value has no impact on our study as long as it is sufficiently less than several minutes), $\gamma = 2.0$ which lies in the interval already proposed by Felzer and Brodsky (2006) and Marsan and Lengliné (2010). The aftershock zone radius is defined as $L_i = 0.2 \times 10^{0.5(m_i - 2.5)}$ (in km), which has the same $10^{0.5m}$ scaling as in Utsu and Seki (1955) and Van der Elst and Shaw (2015). This radius is 4 times larger than the rupture radius of a 30 bar stress drop earthquake (Eshelby 1957).

The values of $v(x_i, y_i, t_i) / K$ are computed once and for all, making the rest of the computation fast with an Expectation-Maximization method. The probability that earthquake i is a background

earthquake is $\omega_i = \mu(x_i, y_i) / \lambda(x_i, y_i, t_i)$. We start by taking arbitrary (but non-zero) values of ω_i , e.g., $\omega_i = 0.5$, and smooth these values to obtain the a priori background rate:

$$\mu(x, y) = \sum_i \omega_i e^{-\sqrt{(x-x_i)^2 + (y-y_i)^2} / \ell} \cdot \frac{1}{2\pi\ell^2}$$

where ℓ is a smoothing parameter taken equal to 50 km. Parameter K is then computed as:

$$K = \frac{\sum_i (1 - \omega_i)}{\sum_i F_i}$$

where $F_i = e^{ami} (\ln(t_e + c - t_i) - \ln c)$, with t_e the ending date of the studied period. This corresponds to the Maximum Likelihood Estimate of K knowing ω_i for $p=1$ as assumed here. Given $\mu(x, y)$ and K , the a posteriori probabilities ω_i can then be computed, and the procedure is thus iterated until all values eventually converge to their final estimates. The latter do not depend on the initial, arbitrary choice of ω_i .

To demonstrate how the declustering affect the data, we show in Figures S1, S2 and S3 the earthquake catalog for all magnitude ≥ 2.5 earthquakes before and after declustering, for two selected periods and (extended) zones. The declustering method yields probabilities of being a background earthquake for each event as the final product. We here randomly draw background earthquakes based on these probabilities – for the purpose of simplifying the visual aspect of the plots. Figure S1 and S3 demonstrate that visually evident aftershock clusters are indeed removed efficiently, which is the case for all aftershock sequences except the remarkable case of the 2011 Tohoku-oki mainshock itself (Figure S2). Here the aftershock sequence is only attenuated but is still very visible in the declustered catalog. This is mainly due to our assumption that aftershock triggering is controlled by the epicentral distance between the mainshock and its aftershocks ; this assumption is clearly not valid for an extended source like a M9 mainshock when performing an analysis at the regional scale as we do, hence the poor behavior of the method for this particular mainshock.

Figure S4 shows the selected deep earthquake dataset analyzed in this study, before and after declustering. This subset is not processed separately from the rest of the catalog: we decluster the catalog as a whole, and afterward extract the selected (deep) subset out of the overall declustered

catalog. While 57 % of all earthquakes are found to be aftershocks, this reduces to 38 % when considering only the deep selected earthquakes. This drop is expected as intermediate and deep activity is less prone to large aftershock triggering (Hainzl et al., 2019 ; Sippl et al., 2019). Figure E offers a complementary view to Figure S5 by plotting the inter-event time vs date before and after declustering, showing that the aftershock sequences have efficiently been removed.

2) Statistical analysis of the earthquake catalogue

In this study, we make the assumption that the period of interest is divided into 2 periods with durations [$\equiv t_b, \equiv t_a$], representing the period of time before and after the time of change, respectively, and characterized by constant seismicity rates [λ_b, λ_a]. In each of these periods, the rate of seismicity is assumed to follow a Poisson process, i.e, the probability to observe N events in a period given the rate λ is

$$p(N|\lambda) = \frac{(\lambda \Delta t)^N \exp(-\lambda \Delta t)}{N!}$$

First statistical analysis: we impose the time of change between the two periods at the Tohoku-oki earthquake occurrence and given observed N_a and N_b , and given $\equiv t_a$ and $\equiv t_b$, we look for the values of [λ_b, λ_a] that maximize $p(N_b|\lambda_b)$ and $p(N_a|\lambda_a)$, which yields : $\lambda_b = N_b/\equiv t_b$ and $\lambda_a = N_a/\equiv t_a$.

Second statistical analysis: for a given time of change between the two periods, the probability to observe a rate of $\lambda_b = N_b/\equiv t_b$ is :

$$p(\lambda_b) = \frac{\Delta t_b (\lambda_b \Delta t_b)^{N_b} \exp(-\lambda_b \Delta t_b)}{N_b!}$$

Following Marsan and Wyss (2011), we compute the probability that the rate is increased by more than a given ratio r:

$$P\left(\frac{\lambda_a}{\lambda_b} > r\right) = \int_0^{\infty} d\lambda_b p(\lambda_b) \int_{r\lambda_b}^{\infty} d\lambda_a p(\lambda_a)$$

The pdf of r is then $p(r) = -\frac{dP}{dr}$. Together with equation (1) this leads to :

$$p(r) = \frac{\Delta t_a^{1+N_a} \Delta t_b^{1+N_b}}{N_a! N_b!} r^{N_a} \int_0^{\infty} d\lambda \lambda^{1+N_a+N_b} e^{-\lambda(\Delta t_b + r \Delta t_a)}$$

Under the integral, we recognize the Gamma function $\Gamma (Na+Nb+2)$ which is the continuous prolongation of the factorial (i.e., $n! = \Gamma (n+1)$), therefore

$$p(r) = \frac{(N_a + N_b + 1)!}{N_a! N_b!} \Delta t_a^{1+N_a} \Delta t_b^{1+N_b} \frac{r^{N_a}}{(\Delta t_b + r \Delta t_a)^{2+N_a+N_b}}$$

Third statistical analysis: We want to compute the probability for the position of the change point to be at a given time t_0 . If N events occur at times $d = \{t_1, t_2, \dots, t_N\}$ and the rate $\lambda(t)$ is variable with time, following [Rasmussen, 2018], the probability to observe those events at these times is

$$p(d \vee \lambda(t)) = \prod_{i=1}^N \lambda(t_i) \exp\left(-\int \lambda(t) dt\right)$$

where the integral is computed from the start to the end of the time interval under consideration.

Our model $\lambda(t) = \lambda_b$ if $t < t_0$ and $\lambda(t) = \lambda_a$ if $t > t_0$ with t_0 the change point. The probability to have a change in the earthquake rate at t_0 given the observations is

$$p(t_0 \vee d) = \int \int p(t_0, \lambda_b, \lambda_a \vee d) d\lambda_b d\lambda_a$$

Using Bayes' Theorem, this becomes :

$$p(t_0 \vee d) = \int \int p(t_0, \lambda_b, \lambda_a) p(d \vee \lambda_b, \lambda_a, t_0) d\lambda_b d\lambda_a = p(t_0) p(\lambda_b) p(\lambda_a) \int \int p(d \vee \lambda_b, \lambda_a, t_0) d\lambda_b d\lambda_a$$

where $p(t_0) p(\lambda_b) p(\lambda_a)$ is called k in the following hence

$$p(t_0 \vee d) = k \int \int \lambda_b^{N_b} \lambda_a^{N_a} \exp(-\lambda_b \Delta t_b - \lambda_a \Delta t_a) d\lambda_b d\lambda_a$$

We now perform a change of variable where $u = \lambda \Delta t$

$$p(t_0 \vee d) = k \frac{1}{\Delta t_b^{N_b+1}} \int u^{N_b} \exp(-u) du \cdot \frac{1}{\Delta t_a^{N_a+1}} \int u^{N_a} \exp(-u) du$$

Under the two integrals, we recognize again the Gamma function. Using Stirling's approximation :

$$N_b! = \sqrt{2\pi N_b} \left(\frac{N_b}{e}\right)^{N_b}$$

we get

$$p(t_0 \vee d) = k \frac{1}{\Delta t_b^{N_b+1} \Delta t_a^{N_a+1}} \sqrt{2\pi N_b} \left(\frac{N_b}{e}\right)^{N_b} \sqrt{2\pi N_a} \left(\frac{N_a}{e}\right)^{N_a}$$

Since $\left(\frac{1}{e}\right)^{N_a+N_b}$ is a constant, it does not depend on t_o , and lumping together all the constants in a new k factor, we finally obtain

$$p(t_o \vee d) = \frac{k \sqrt{N_b N_a} N_b^{N_b} N_a^{N_a}}{\Delta t_b^{N_b+1} \Delta t_a^{N_a+1}}$$

used to compute Fig. 3.

3) GPS data processing

The data of the 1439 GEONET stations have been processed in double difference using GAMIT/GLOBK software [Herring et al., 2018]. For each day, the GEONET data were split into sub-networks of about 40 sites chosen to minimize the baseline between stations and improve the resolution of phase ambiguity. Subnetworks share 2 common sites with nearby other sub-networks to “tie” the solutions together, following a similar approach than the one presented in the framework of the PBO project Herring et al. [2016]. A reference tie network containing at least 1 station from every other sub-network is also constructed to provide additional stability to the entire network combination. For each sub-network, we reduce 24-hr measurement sessions to daily estimates of station position, choosing the ionosphere-free combination and fixing the ambiguities to integer values. We use IGS final products for the satellite orbits, satellites clocks and Earth orientation parameters (<https://www.igs.org/products>). Following [Herring et al., 2016], the orbit parameters are fixed to the IGS values. Ocean loading corrections are applied at each stations, using FES2004 (Finite Element Solution) [Lyard et al., 2021] ocean tidal loading. The Vienna Mapping Function (VMF1) [Bohm et al., 2006] is used to map the tropospheric delay in zenithal direction. The zenith delay is estimated every 2-hours and 1 gradient parameter is estimated per day. We apply atmospheric tidal and non-tidal loading correction, following Tregoning and van Dam [2005]

recommendation. The different sub-networks are then combined together into a single daily solution, in a regional stabilization approach using a Kalman filter with GLOBK software to obtain daily loosely constrained solutions. Then daily solutions are combined into a multiyear solution to derive the time series, and to express the solutions in the ITRF2014 [Altamimi et al., 2017] with a 7-parameter transformation using regional IGS sites. The daily GNSS position time series [gnss products, 2019] are available at https://doi.osug.fr/staging/GNSS_products/GNSS.products.Japan.html. These GNSS position time series are then corrected from any jumps associated with documented material changes. The change of velocity before and after the Tohoku-oki earthquake are simply calculated as the difference of slope in each time series, between a linear trend before the earthquake from 2008/01/01 to 2011/03/08 and a linear trend after the earthquake from 2011/03/13 to 2013/03/10.

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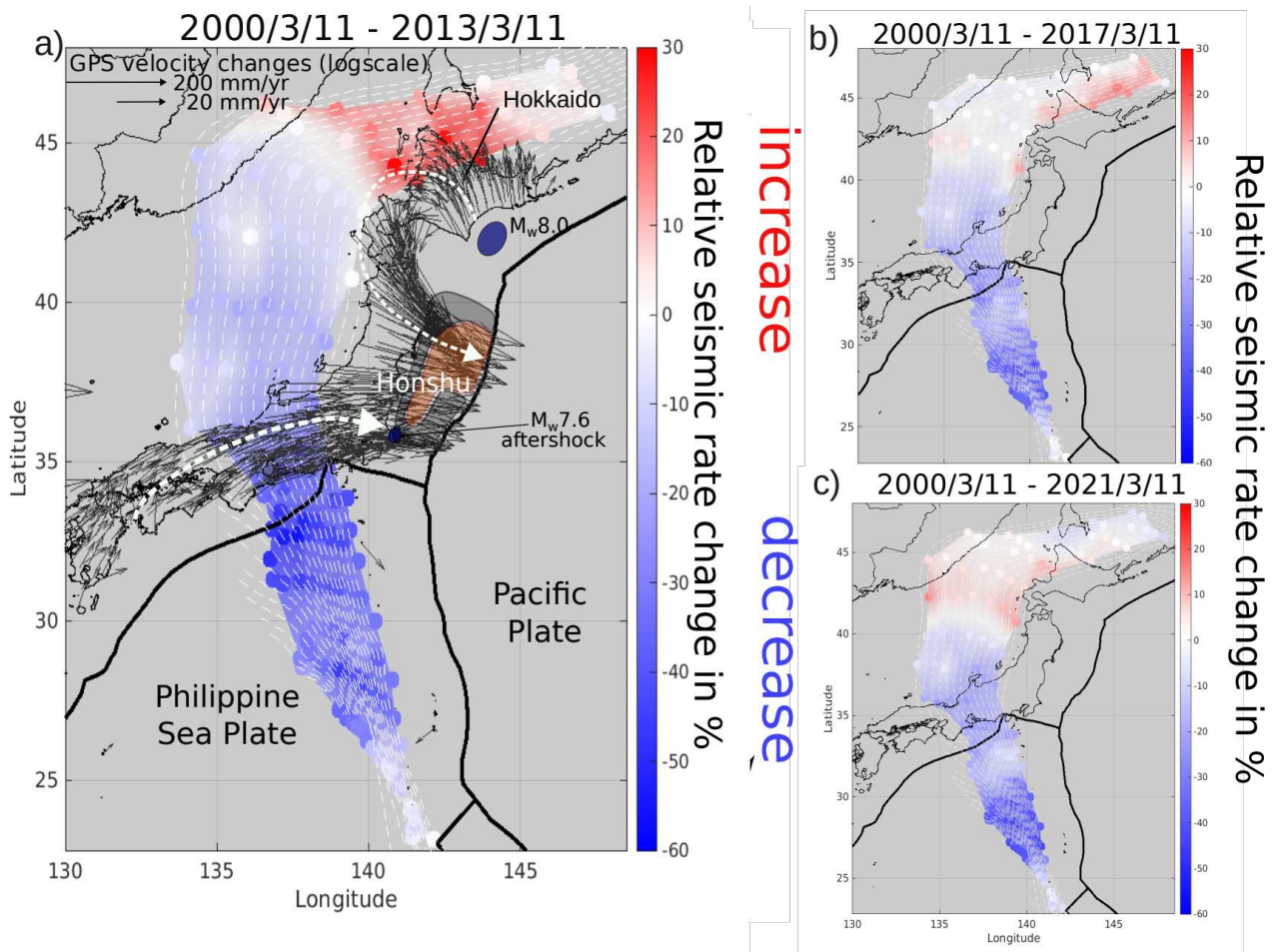


Fig. 1. Relative change in the rate of intermediate and deep seismicity of the Pacific plate taking the Tohoku-oki earthquake as time of change. The data span a) from 2000/03/11 (yy/mm/day) to 2013/03/11; b) from 2000/03/11 to 2017/03/11 period, using the 260 closest events to sample the same area c) from 2000/03/11 to 2021/03/11 period using the 320 closest events to sample the same area. Negative values (blue) indicate a decrease of seismicity rate, positive values (red) indicate an increase of seismicity rate. Every 20km isodepth contours of the Pacific plate are shown from 80 to 400 km³⁴. Black arrows give the log scaled GPS post-seismic displacement. The orange and grey areas show the contours of the co and post-seismic slips¹².

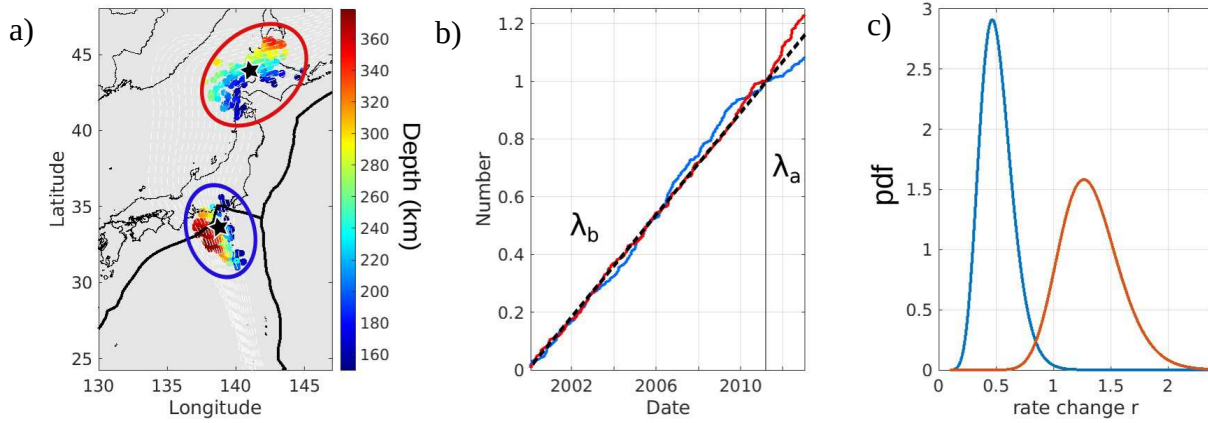


Fig. 2. Change of seismicity rate near latitudes 34°N and 44°N. a) Location of the earthquakes color coded with depth. They correspond to the 200 closest earthquakes of the two black stars. b) Cumulative number of earthquakes with time after declustering normalized at Tohoku-oki, for 34°N (blue) and 44°N latitude (red). Note the change of slope at the time of the Tohoku-oki earthquake, with λ_b the slope before and λ_a , the slope after the megathrust earthquake. c) probability density function (pdf) of the rate change (see Methods). A rate change $r < 1$, $=1$, >1 , means decrease, no change and increase, respectively. The probability of change at Tohoku-oki is maximum with $r_{\max}=0.47$ at 34°N (blue) and $r_{\max}=1.3$ at 44°N latitude (red).

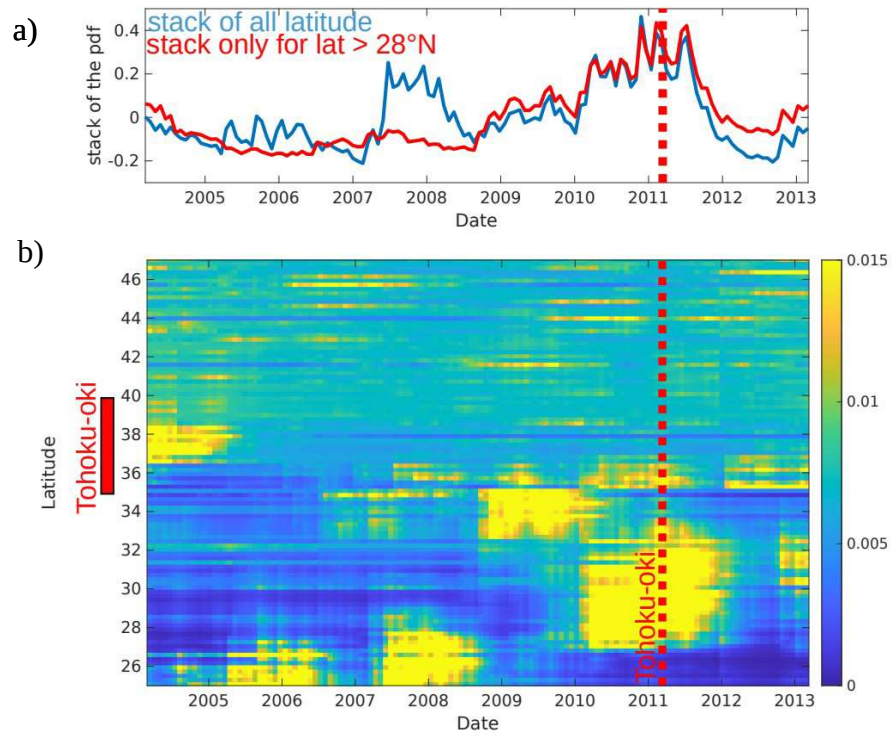


Fig. 3. Probability of having a seismicity rate change at any given time, stacked over all the reference points. a) Stack of all the data with or without points with latitude < 28°. b) Probability at each reference point plotted according to its latitude (see Methods).

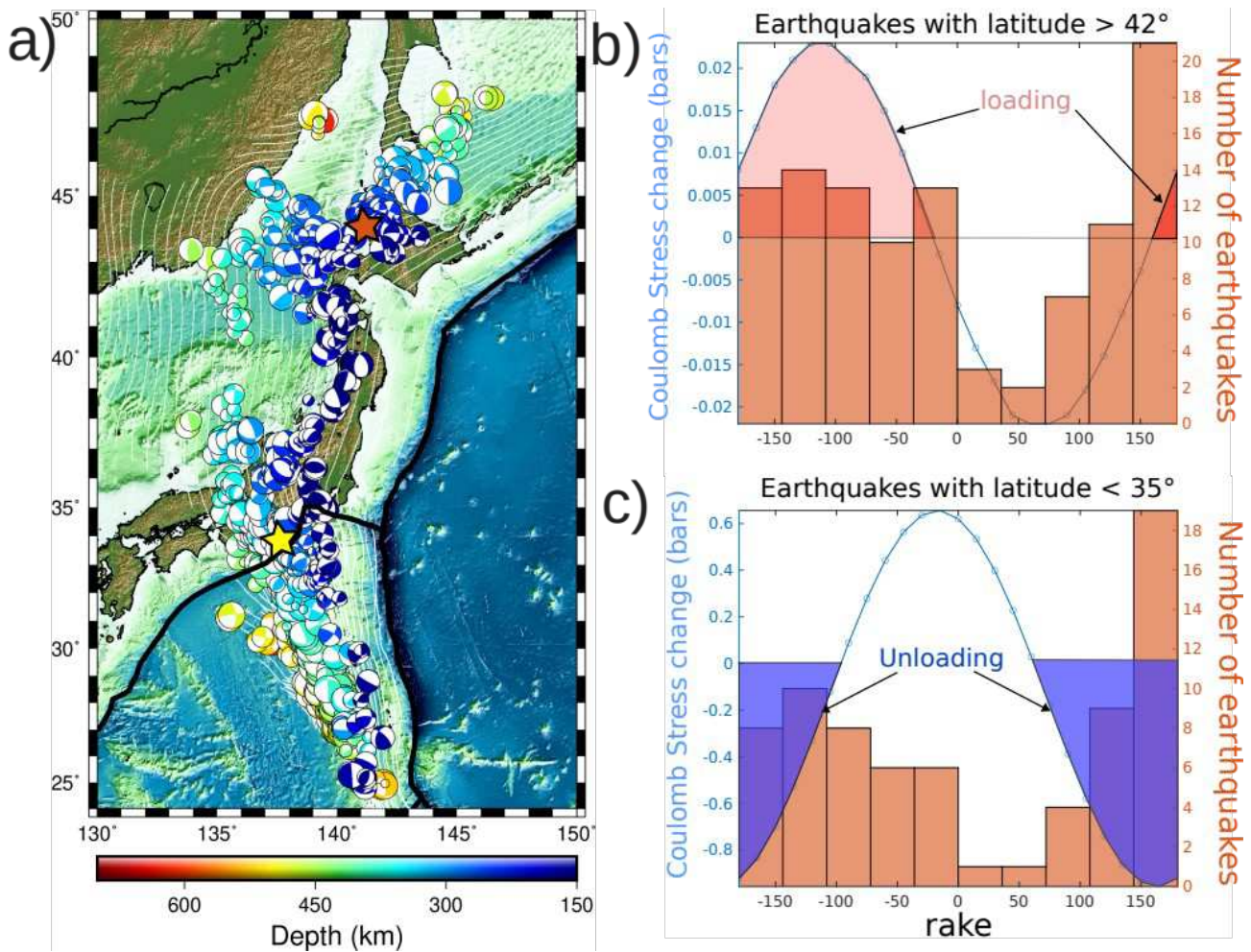


Fig. 4. a) Focal mechanisms of intermediate and deep earthquakes (source from NIED <https://www.fnet.bosai.go.jp/>). The orange and yellow stars locate the points where we compute the Coulomb stress change of b) and c). b) Blue: Coulomb stress change computed with COULOMB3 on a point located underneath Hokkaido, on the slab (lon=140.8°, lat=44.25°, depth=-204km, strike=250°, dip=26.7°, orange star in Fig. 4a) for different rake values, Orange: histogram of the rake values of all the earthquakes with latitude > 42°, c) Blue: same as b) for a point underneath 34°N latitude on the slab (lon=138.4°, lat=34.70°, depth=-181km, strike=161°, dip=39°, yellow star in Fig. 4a); Orange : same as b) for earthquakes with latitude < 35°. Here we assume a 0.4 friction coefficient.

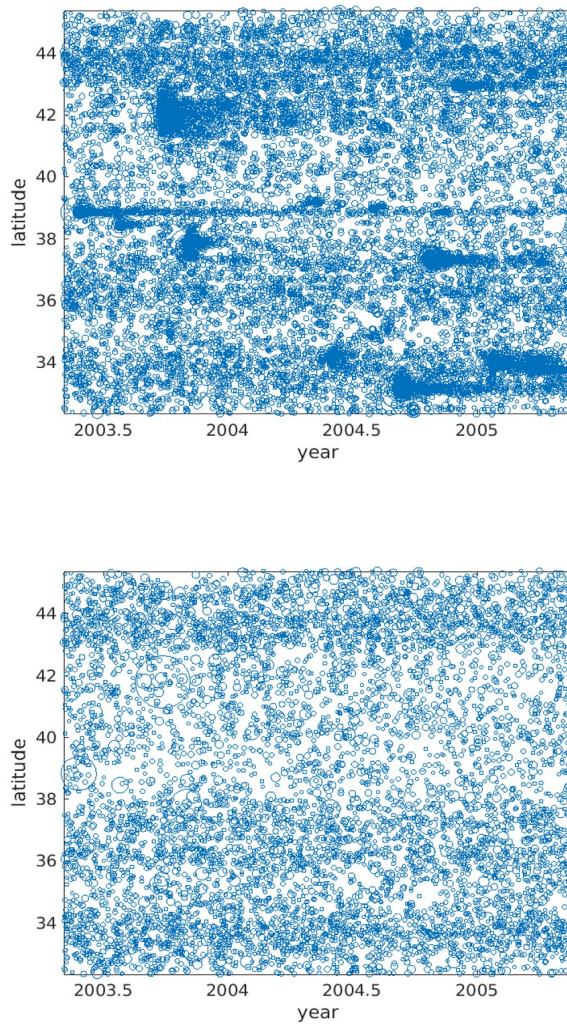


Fig. S1: (top) all $m \geq 2.5$ earthquakes in the JMA catalog, for the selected time period and latitude interval. The circle radius grows with the magnitude. Aftershock sequences are readily visible as clusters initiated by large shocks. (Bottom) same as top graph after declustering. The aftershock sequences are all removed.

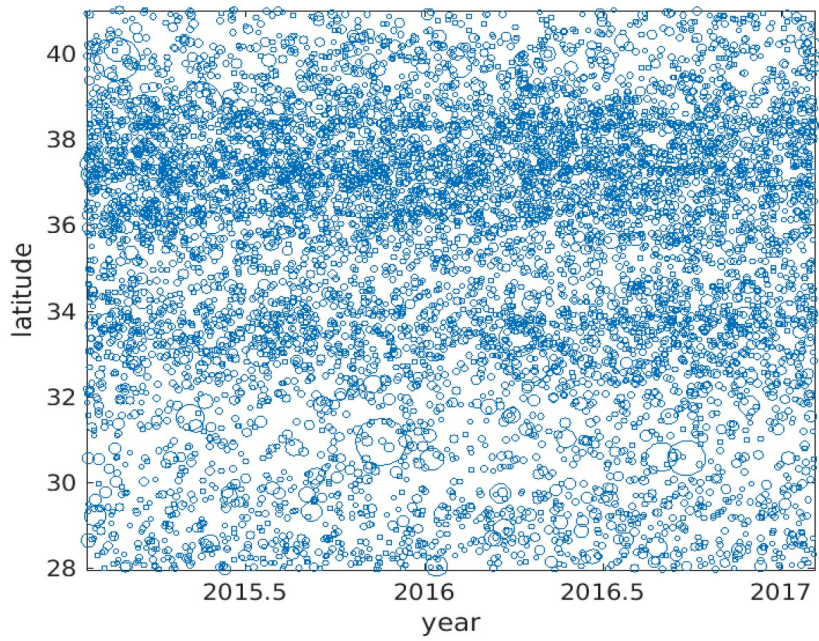
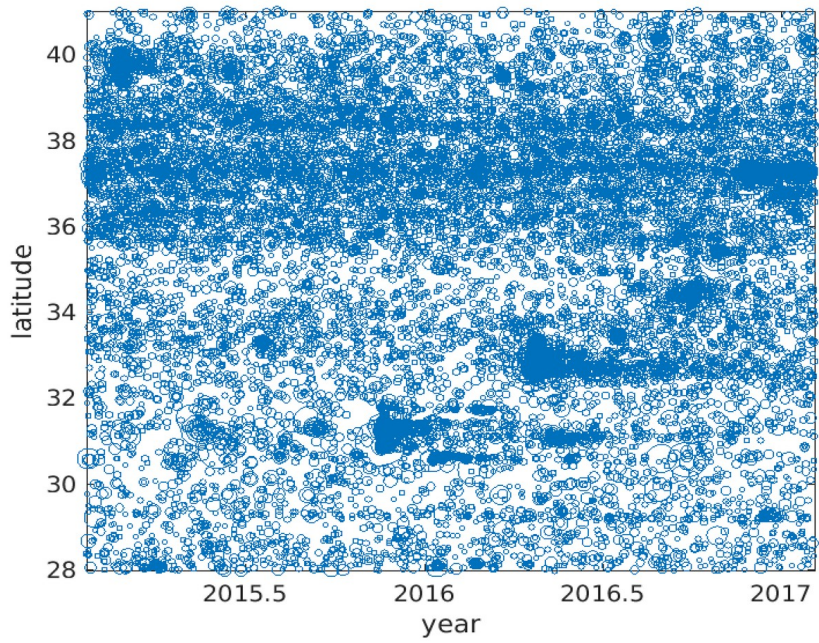


Fig. S2. Same as Fig. S1, for a time period after the 2011 Tohoku-oki-Oki megathrust earthquake.

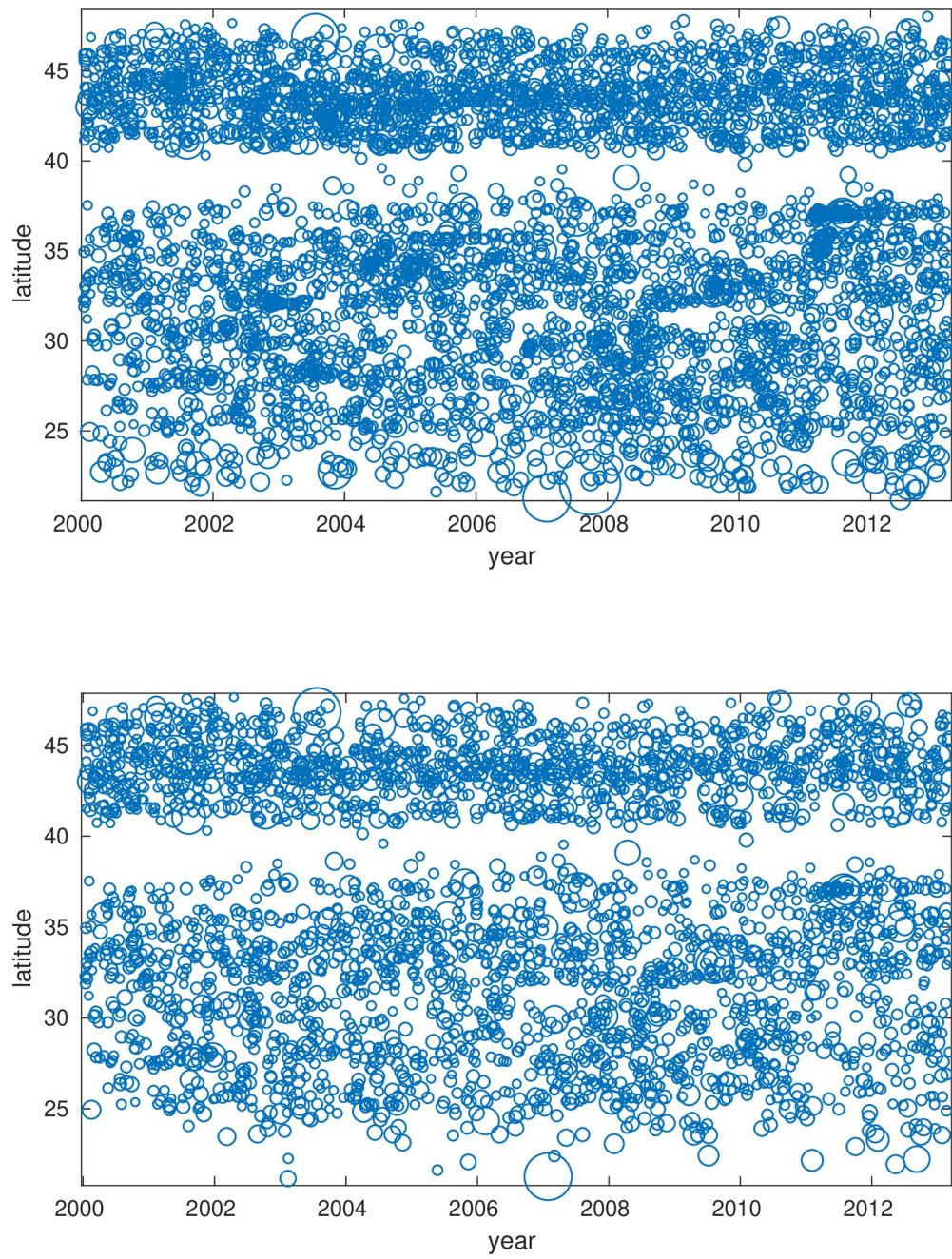


Fig. S3. Deep earthquakes selected in this study, before (top) / after (bottom) declustering.

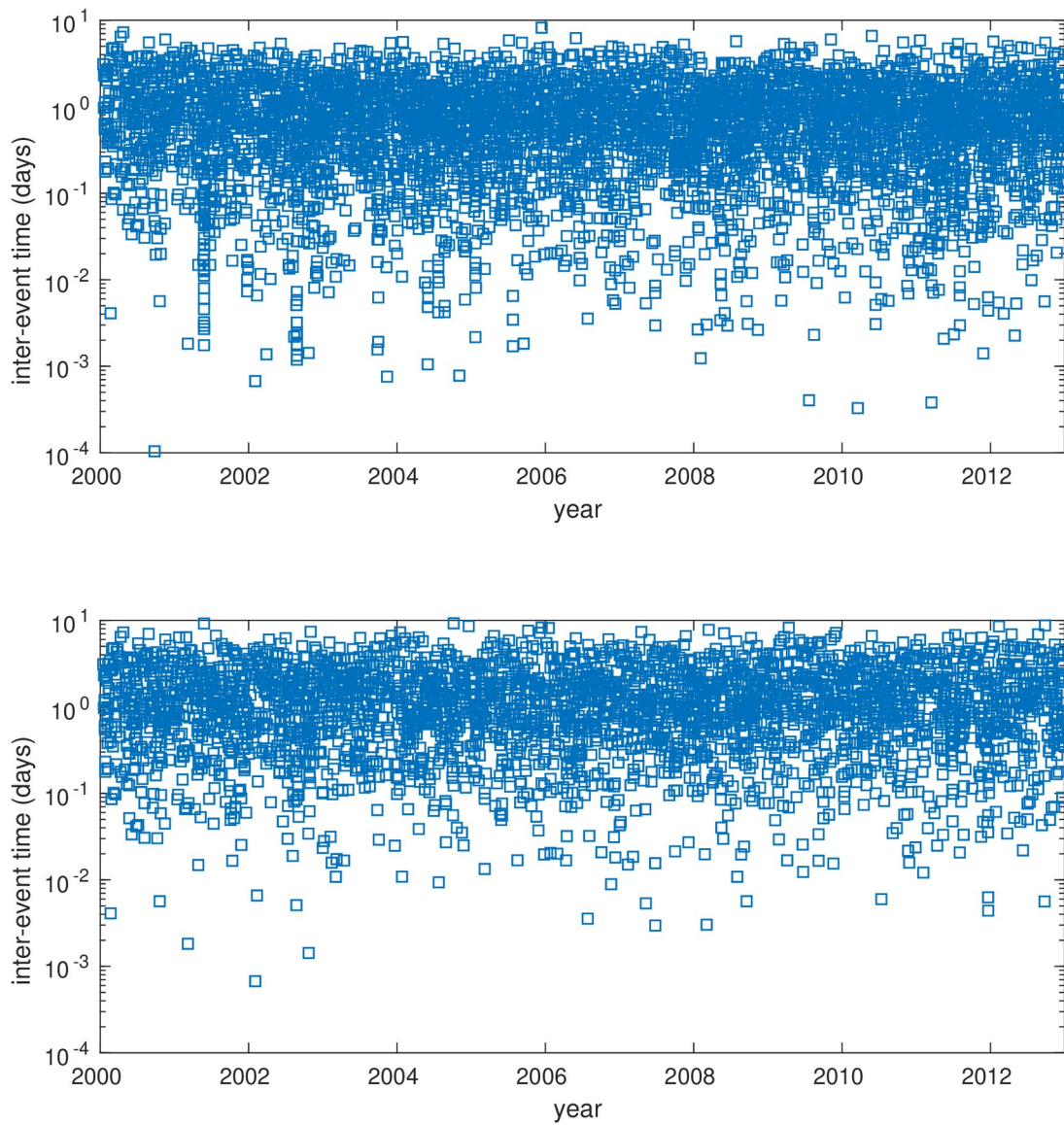


Fig. S4. Time between two successive earthquakes for the selected deep earthquakes, before (top) and after (bottom) declustering. The vertical ‘streaks’ corresponding to aftershock sequences (e.g., in 2001, 2002, 2004) are removed by the declustering.

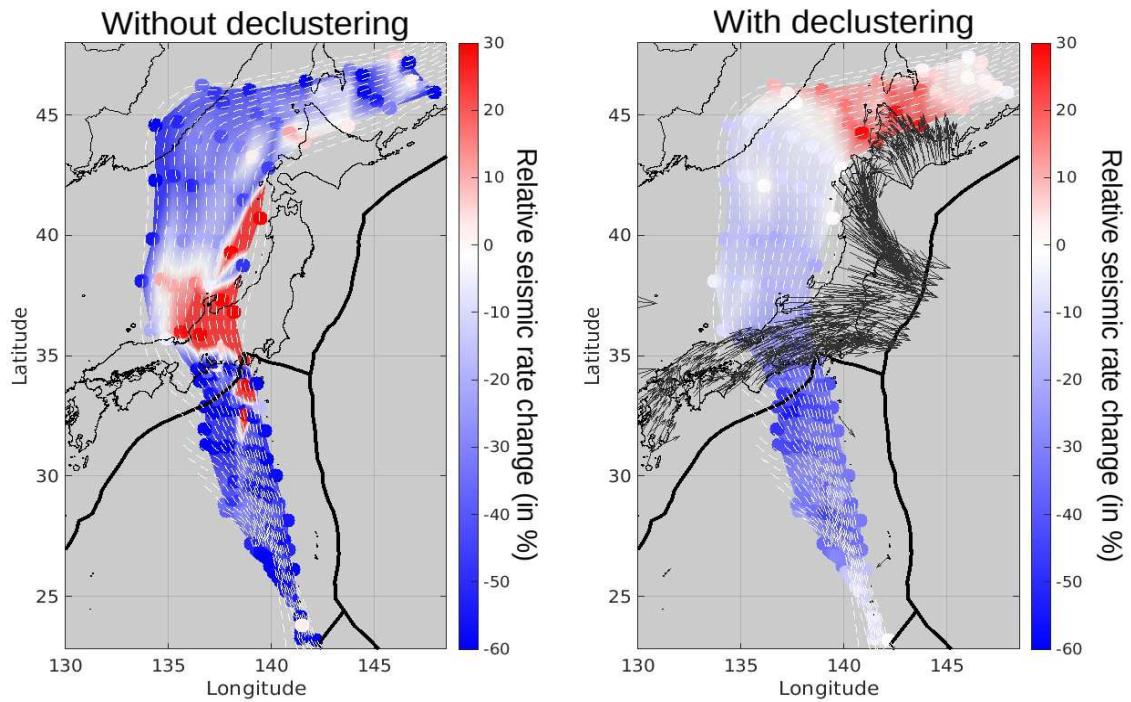


Fig. S5. Comparison of the relative seismicity rate change (in %) obtained before (left) and after (right) declustering. We see the dichotomy between North and South. After declustering, the decrease of the seismicity rate is less important in the South and the increase is more important in the North. The declustering method allows to unbiased the seismicity rate change, especially at latitudes 36 and 40°. Note that we remove the area close to the Tohoku-oki and the M_w 8.0 2003 Tokachi-oki rupture zones since the seismicity of the region is dominated by aftershocks even after declustering due to the very intense seismic activity (see Fig. S2).

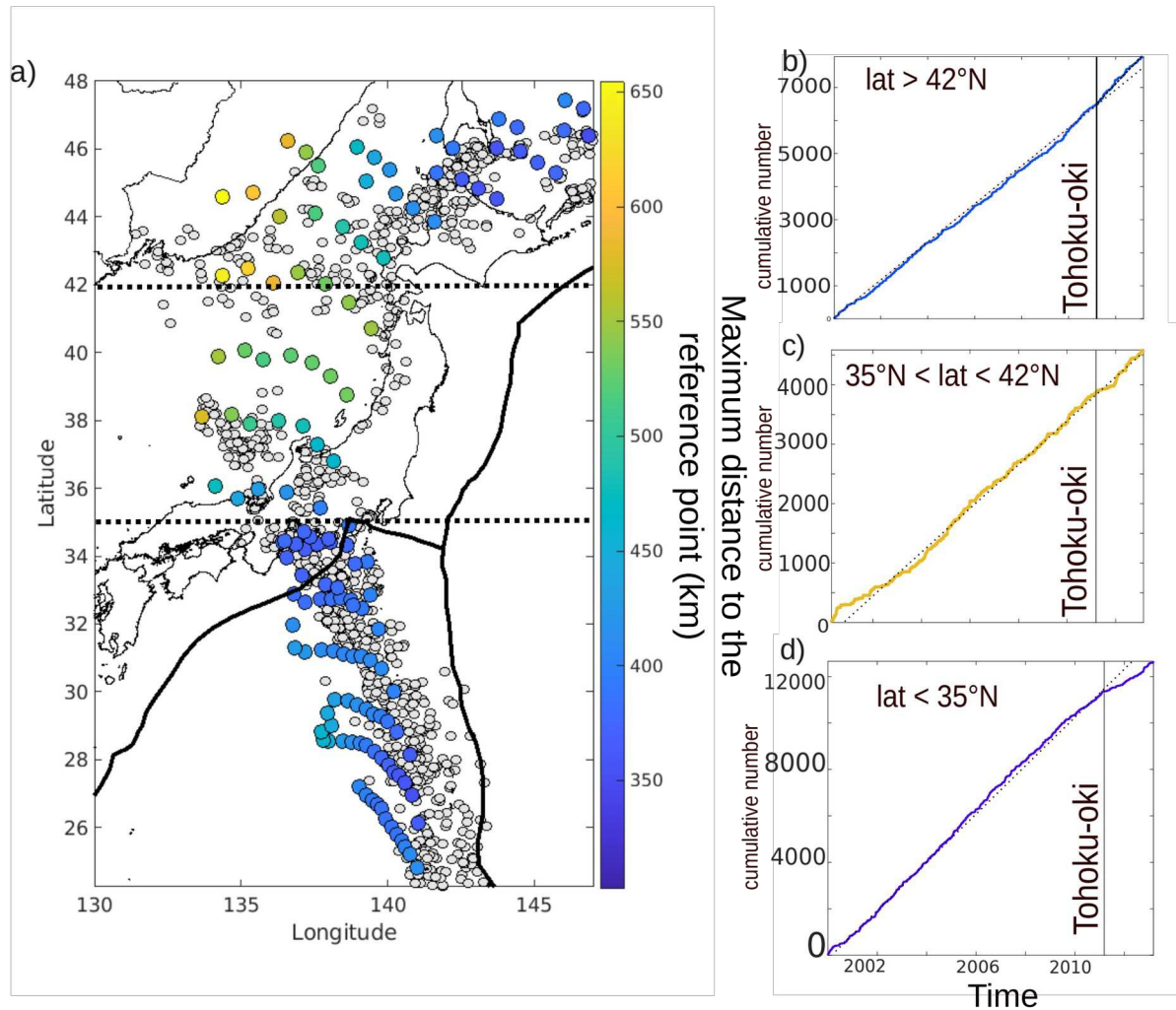


Fig. S6: Map of Japan showing the location of earthquakes studied in this paper selected with the method described in the main text (gray dots) and location of the reference points color coded with the maximum distance between the point and the farthest earthquake; b) Stack of the time series for all the reference points located at $\text{lat} > 42^\circ\text{N}$ earthquakes with time after selection and declustering, c) same as b for $35^\circ\text{N} < \text{lat} < 42^\circ\text{N}$ earthquakes. d) same as b for $\text{lat} < 35^\circ\text{N}$ earthquakes. Note that the total number of selected earthquakes is smaller for the Japan Sea than underneath Hokkaido or in the south. There is no significant change of seismic rate in this area while there is a clear increase and decrease in the North and South, respectively.

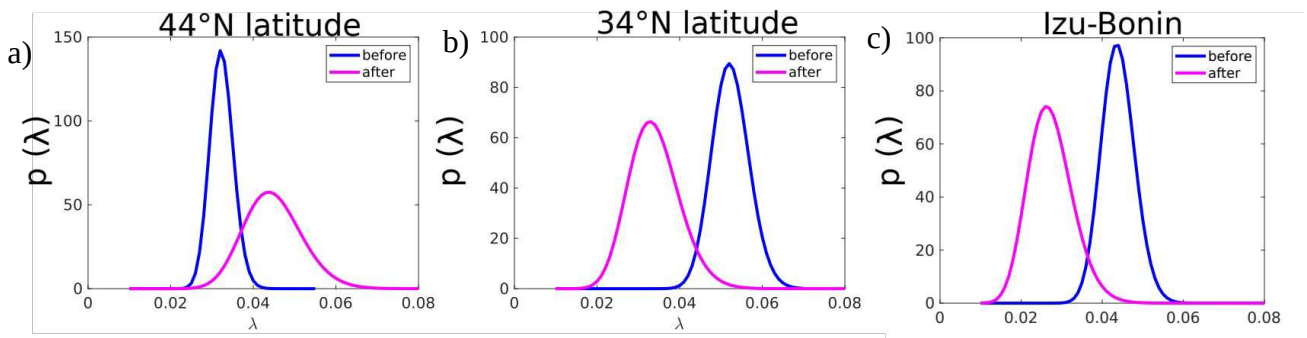


Fig. S7. Probability density function of the earthquake rate $p(\lambda)$, in 1/day, observed in a) Hokkaido (44°N latitude), b) 34°N latitude and c) Izu-Bonin area : before, i.e. λ_b (2000/3/11 – 2011/3/10, blue) and after, i.e. λ_a (2011/3/11 – 2013/3/11, pink) the Tohoku-oki earthquake. Note that with more earthquakes before the distribution of the λ is sharper before than after the Tohoku-oki earthquake.

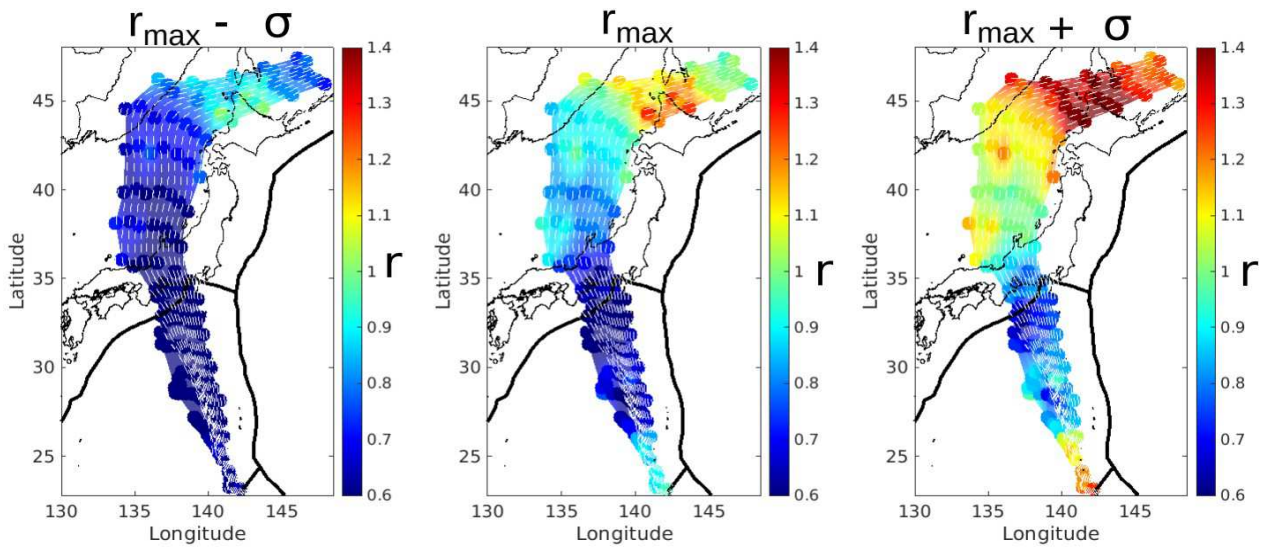


Fig. S8. Analysis of the uncertainty on the maximum of the pdf of r . To show the uncertainty on r_{\max} , i.e. the r value that maximizes $f(r)$ (middle), we show the r -values $r_{\max}-\sigma$ (left) and $r_{\max}+\sigma$ (right) with the same colorbar for comparison. We see that even after removing 1σ to the maximum, there is still an increase of the seismicity rate underneath Hokkaido. On the contrary, there is still a decrease of seismicity rate at latitude 34° when adding σ . Note that the data in the middle is the same as in Fig. 1 with the difference that the scale is here not in percentage.

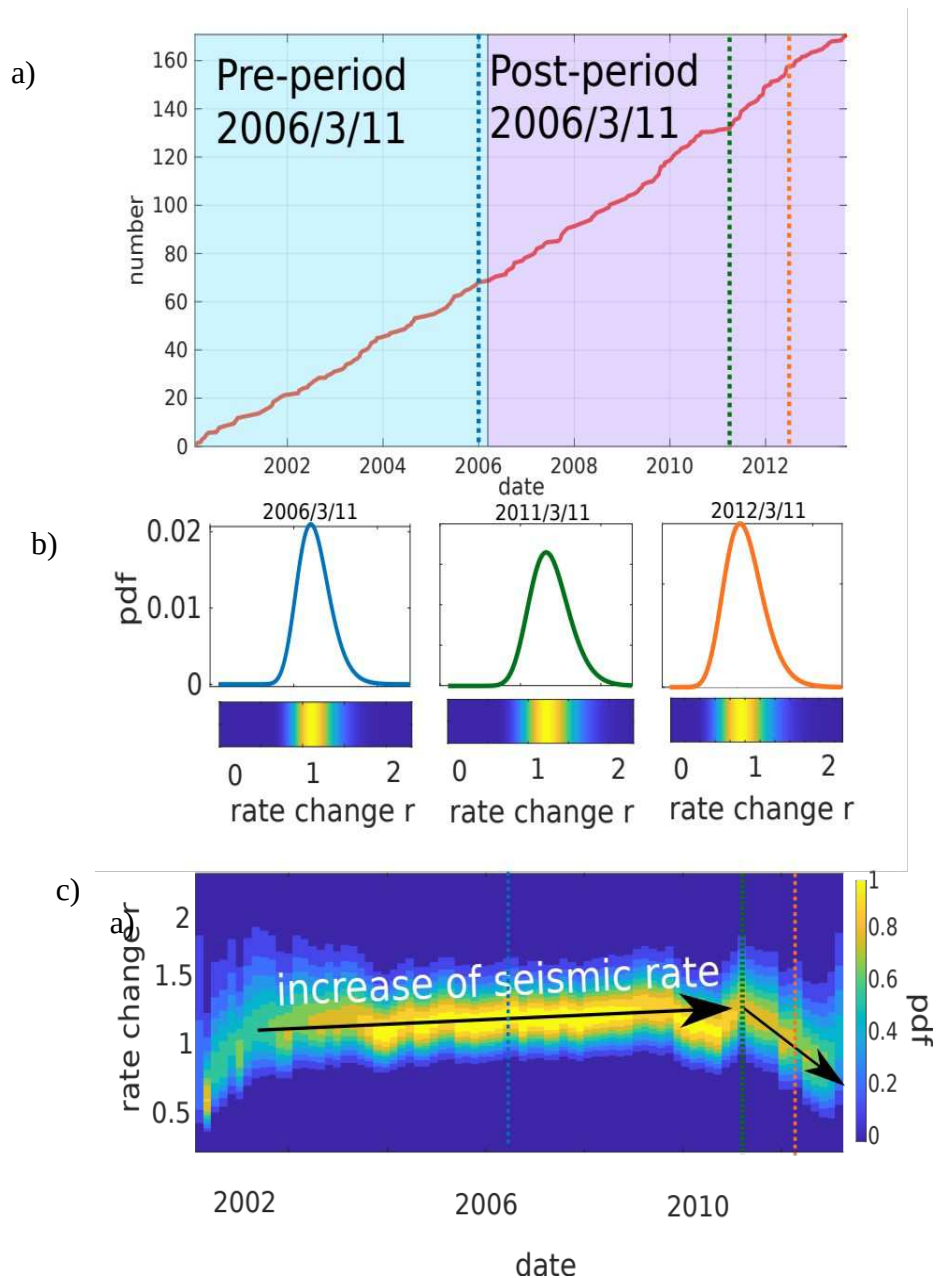


Fig. S9 Pdf of the seismicity rate change at each time plot computed for different times of change underneath Hokkaido. a) cumulative number of earthquakes with time, b) pdf of r , c) pdf of r for different times of change. The example shows that in 2006 (blue line), the increase of the seismicity rate due to the Tohoku-oki earthquake is in the post-period (between 2006/3/11 and 2013/3/11) so the r -value that maximizes the pdf (blue line) is slightly above 1. Choosing Tohoku-oki as the time of change, the pdf maximum is obtained for $r_{\max}=1.4$. Choosing a time of change in 2012 gives a $r < 1$ because a part of the increase is in the pre-period (between Tohoku-oki and 2012). We see that $r > 1$ before Tohoku-oki followed by a significant change of slope at the time of Tohoku-oki when r decreases after the megathrust earthquake.

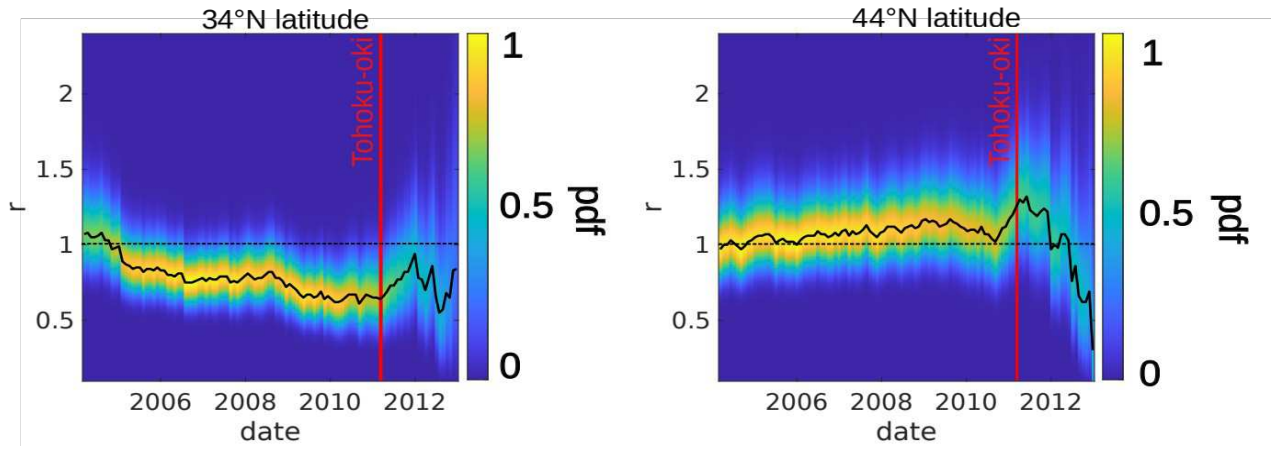


Fig. S10: pdf of the seismicity rate change at each time for 34°N latitude (left) and 44°N latitude (right) with the black line showing r_{\max} . A wider pdf means that the level of change is less constrained, as it is the case when one of the 2 periods contains a lower number of events.

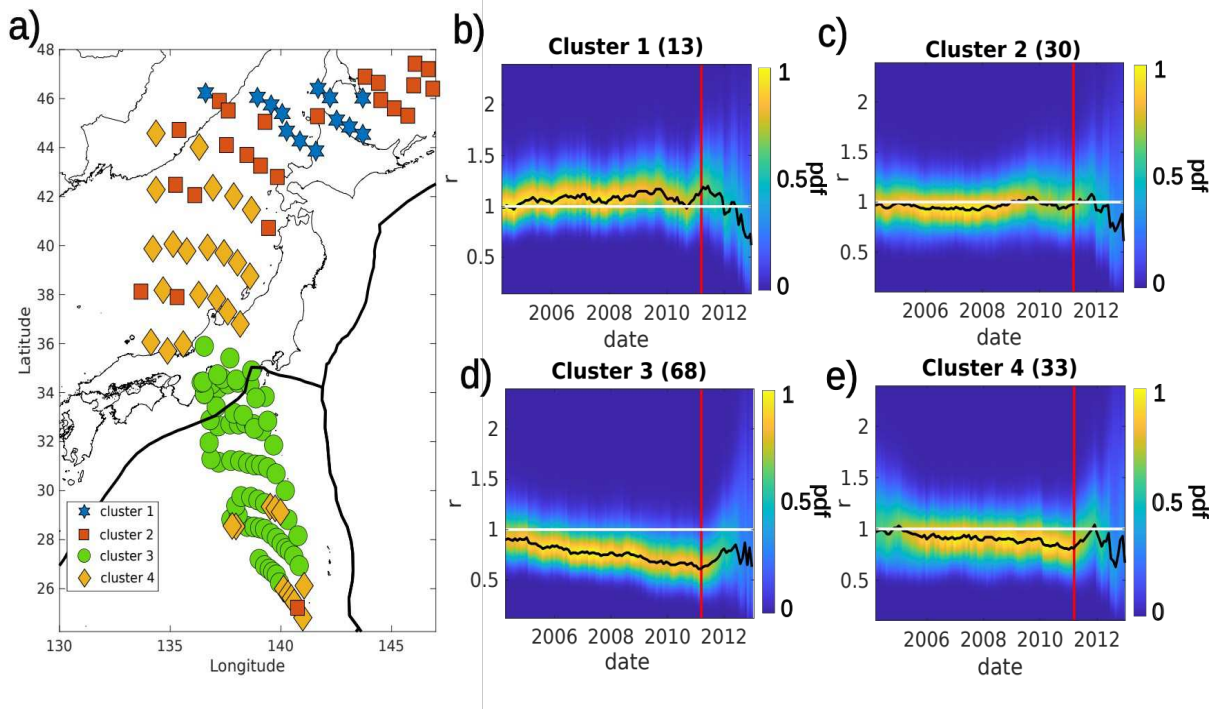


Fig. S11. Location and pdf of the seismicity rate change at each time of the clusters using the k-means clustering on the r_{\max} value at the time of the Tohoku-oki earthquake. The number of points present in each cluster is given in brackets. b-e) the black line shows r_{\max} and the red line shows the Tohoku-oki earthquake occurrence time. Note that points in the northern and southern tips belong to clusters 2 and 4 which undergo less effect of the Tohoku-oki earthquake indicating the spatial limit of the Tohoku-oki effect.

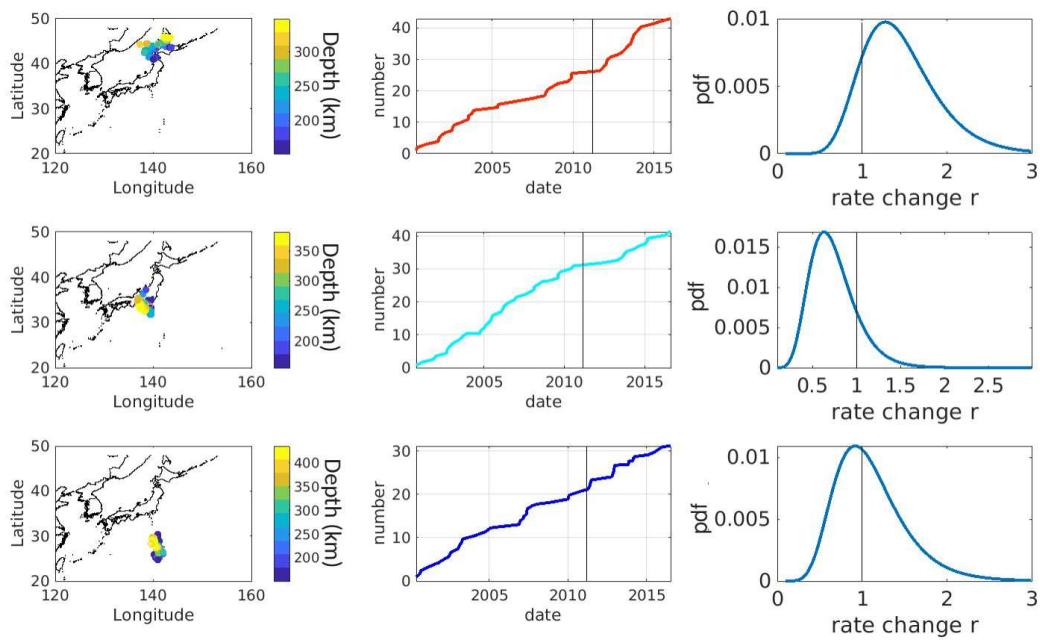


Fig. S12. Analysis of the seismicity from the JMA catalog at three different reference points : Hokkaido, Chubu and the Izu-Bonin. We use $M \geq 4.5$ earthquakes and the 70 closest events at each reference point and show the location (left), cumulative number (middle), probability of a change in seismicity rate (right). The acceleration and deceleration of the seismicity rate underneath Hokkaido and Chubu, respectively, is clear while the deceleration at the Izu-Bonin is less robust.

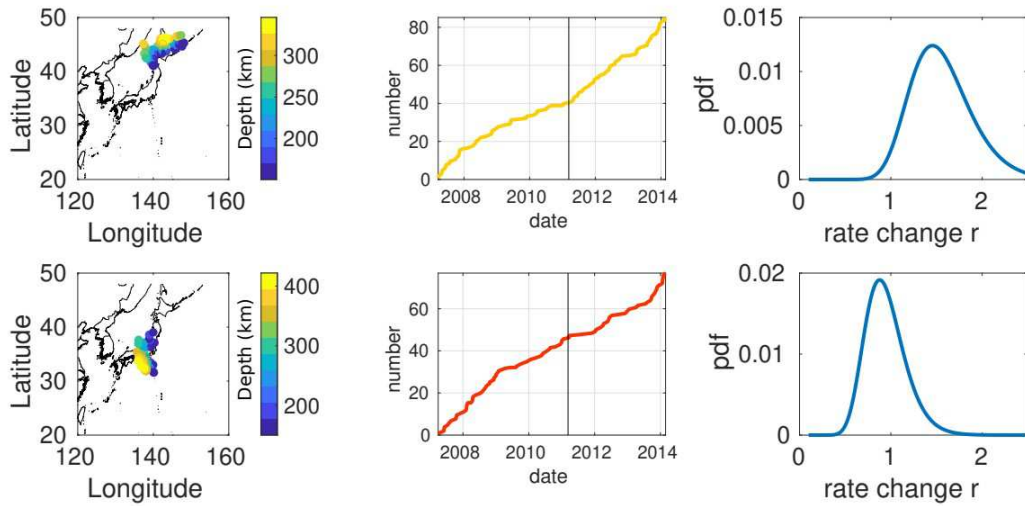


Fig. S13. Analysis of the seismicity from the USGS catalog at two different points : Hokkaido, Chubu and the Izu-Bonin. We select only $M \geq 4.5$ earthquakes and the 70 closest earthquakes at each reference point and show the location (left), cumulative number (middle), probability of a change in seismicity rate (right). The acceleration and deceleration of the seismicity rate underneath Hokkaido is clear while the deceleration underneath Chubu is less robust.