Earthquake Recurrence intervals in Complex Seismogenetic Systems

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ABSTRACT

We examine the association of recurrence intervals and dynamic (entropic) states of shallow (crustal) and deep (sub-crustal) seismogenetic systems, simultaneously testing if earthquakes are generated by Poisson processes and are independent (uncorrelated), or by Complex processes and are dependent (correlated). To this effect, we apply the $q$-exponential distribution to the statistical description of interevent times, focusing on the temporal entropic index (measure of dynamic state), in connexion to the $q$-relaxation interval that constitutes a characteristic recurrence interval intrinsically dependent on the dynamic state. We examine systems in different geodynamic settings of the northern Circum-Pacific Belt: transformational plate boundaries and inland seismic regions of California, Alaska and Japan, convergent boundaries and Wadati-Benioff zones of the Aleutian, Ryukyu, Izu-Bonin and Honshū arcs and the divergent boundary of the Okinawa Trough.

Our results indicate that the $q$-exponential distribution is universal descriptor of interevent time statistics. The duration of $q$-relaxation intervals is reciprocal to the level of correlation and both may change with time and across boundaries so that neighbouring systems may co-exist in drastically different states. Crustal systems in transformational boundaries are generally correlated through short and long range interaction; very strong correlation is quasi-stationary and $q$-relaxation intervals very short and extremely slowly increasing with magnitude: this means that on occurrence of any event, such systems respond swiftly by generating any magnitude anywhere within their boundaries. These are attributes expected of SOC. Crustal systems in convergent and divergent margins are no more than moderately correlated and sub-crustal seismicity is definitely uncorrelated (quasi-Poissonian). In these cases $q$-relaxation intervals increase exponentially, but in Poissonian or weakly correlated systems their escalation is much faster than in moderately to strongly correlated ones. In consequence, moderate to strong correlation is interpreted to indicate Complexity that could be sub-critical or non-critical without a means of telling (for now). The blending of earthquake populations from dynamically different fault networks randomizes the statistics of the mixed catalogue.

A possible partial explanation of the observations is based on simulations of small-world fault networks and posits that free boundary conditions at the surface allow for self-organization and possibly criticality to develop, while fixed boundary conditions at depth do not; this applies particularly to crustal transformational systems. The information introduced by $q$-relaxation may help in improving the analysis of earthquake hazards but its utility remains to be clarified.
1. Introduction

A recurrence interval is a statistical estimate of the likelihood of an earthquake to occur, typically based on historical data and used in the analysis of earthquake hazard and risk, so as to design structures that will withstand events of given severity and return period. The standard definition of recurrence interval (number of years on record plus one, divided by the number of events) assumes that the events are generated by point (Poissonian) processes, meaning that events of similar size have a stationary probability of occurrence and are independent of each other, as well as of their predecessors and successors. In calculating recurrence intervals, the number of events is typically taken from, or estimated on the basis of the standard Frequency–Magnitude (F-M) distribution of Gutenberg and Richter, or modifications/extensions of the F-M distribution (e.g. Molnar, 1979; Kagan, 1997). Although indisputable, the F-M distribution is static and says nothing about the dynamics of the fault network, or about correlation (dependency) in the energy released by successive earthquakes. Accordingly, standard and “improved” estimators of earthquake recurrence are based on the distribution of magnitudes over a given seismic region and time period, thus comprising approximations to the long-term average of the true recurrence interval. This might lead to misestimation if the dynamics of the seismogenetic system is not Poissonian.

The obvious and unique parameter directly associated with the recurrence interval is the lapse between consecutive earthquakes over a given area and above a magnitude threshold: this is referred to as interevent time, waiting time, calm time etc. Understanding the statistics of earthquake frequency vs. interevent time is apparently essential for understanding the dynamics of the active fault network. However, interevent times have generally not been used in the estimation of earthquake recurrence intervals. In this Authors’ interpretation, this is due to a majority endorsement of the idea that seismogenesis is fundamentally a point process in time. If so, seismogenesis should obey additive Boltzmann-Gibbs thermodynamics and be memoryless. The most influential realization of this idea is the Epidemic-Type Aftershock Sequence (ETAS) model and its modifications (e.g. Ogata, 1988, 1998; Console and Murru, 2001; Helmstetter and Sornette, 2003; Ogata and Zhuang, 2006; Marzocchi and Lombardi, 2008; many others). Because interevent times are strictly positive, their statistics should observe the exponential distribution. However, contrary to “expectation” the empirical frequency–interevent time (F-T) distributions are generally found be power laws. For this reason, in the context of statistical seismology they have been analysed with tailed standard statistical models reducible to power laws in some way or another. Examples of this approach are the gamma distribution and the Weibull distribution (e.g. Bak et al., 2002; Davidsen and Gold, 2004; Corral, 2004; Martinez et al, 2005; Talbi and Yamazaki, 2010). Nevertheless, Molchan (2005) has shown that for a stationary point process, if there is a universal distribution of interevent times, then it must be an exponential one. Saichev and Sornette (2007) rebutted by showing that an approximate unified law compatible with empirical observations could be found by incorporating the Omori-Utsu law of aftershocks; the same
Authors went on to develop a theory of the statistics of interevent times in the framework of the ETAS model and to argue that empirical observations can be explained in this context. Working from a statistical physics perspective, some researchers proposed \textit{ad hoc} mechanisms for the generation of power laws by a combination of correlated aftershock and uncorrelated background processes (e.g. Saichev and Sornette, 2013; Hainzl et al, 2006; Touati et al, 2009).

We argue that it is difficult to conceptually reconcile the expected from point processes exponential distributions, with experimentally observed power-laws that imply altogether different dynamics. As to why, consider that in order to provide realistic descriptions of seismicity, Poissonian theories must rely on irrefutable but obviously non-Poissonian empirical laws: the Gutenberg-Richter distribution is a power-law that \textit{cannot} be derived on the basis of Boltzmann-Gibbs thermodynamics and the Omori-Utsu aftershock distribution is a Zipf-Mandelbrot power-law and also inconsistent with the Boltzmann-Gibbs formalism. Accordingly, Poissonian theories posit that seismicity comprises some mixture of Poissonian and non-Poissonian processes and that its dynamic expression is at the same time Poissonian and non-Poissonian (where aftershocks are concerned). This is a contradiction in terms with no apparent resolution because the fundamental empirical laws are introduced axiomatically and do not emerge from the theories: Poissonian models effectively are \textit{ad hoc} constructs that albeit well formulated and generally elegant, are probably akin to grand unification constructs such as string or M-theories: necessarily multi-parametric, unnecessarily complicated and possibly challenging the principle of maximum parsimony.

An alternative approach is Complexity. In this view seismicity expresses a fractal fault network (system) that may be sustainably non-equilibrating, or may even evolve/transit between equilibrating (Poissonian) and non-equilibrating (Complex) states. Non-equilibrating states require a significant proportion of successive earthquakes to be \textit{dependent} through short and long range interaction that introduces delayed feedback. The dependence is known as \textit{correlation} and confers memory manifested by power-law distributions of dynamic parameters such as energy release rates and interevent times in particular. Non-equilibrating states (Complexity) can be critical, subcritical and non-critical. The critical extreme is occupied by Self-Organized Criticality (SOC), an internal bottom-up process postulating that seismicity continuously evolves toward a stationary critical state with no characteristic spatiotemporal scale, so that earthquakes develop spontaneously and have a chance of cascading into large events (e.g. Bak and Tang, 1989; Sornette and Sornette, 1989; Olami et al., 1992; Bak et al, 2002; Sornette, 2004; Bakar and Tirmakli, 2009; many others). The allure of SOC is that it is self-consistent and also predicts several observed properties of earthquake occurrence: the Gutenberg-Richter and Omori-Utsu laws emerge \textit{naturally} in simulated fault networks. The non-critical part is occupied by several and frequently top-down mechanisms, (e.g. blended dynamics, external forcing etc.), able to maintain a fault network in states of non-equilibrium; a list can be found in Sornette (2004) and Sornette and Werner (2009). Notable among these is the Coherent Noise Model (Newman,
shown by Celikoglu et al. (2010) to generate power-law distribution in the time dependence of successive events.

The statistical properties of Complex systems can be evaluated on the basis of Non-Extensive Statistical Physics (NESP), which was introduced by Constantino Tsallis as a generalization of the Boltzmann-Gibbs formalism of thermodynamics (Tsallis, 1988, 2001, 2009; Gell’mann and Tsallis; 2004; Tsallis and Tirnakli, 2010). As such, it is appropriate for the analysis of Complexity evolving in a fractal-like spacetime and exhibiting scale invariance, long-range interaction and memory; the formalism is summarized in Section 2 and, more extensively, in the supplementary material. NESP comprises a physically intuitive description of seismicity with a minimal set of parameters; suffice to say that it provides a theoretical platform on which to construct the Gutenberg-Richter law from first principles (Section 2.1). During the past several years, NESP applications to seismology have attracted considerable attention and several researchers studied the properties of F-T and F-M distributions. A long but non-exhaustive list is provided with the supplementary material, while extensive collections of review and research papers can be found in Vallianatos and Telesca, (2012), in Chelidze et al., (2018), and in the references therein.

Studies conducted by Efstathiou et al., (2015, 2016, 2017), Efstathiou and Tzanis, (2018); Tzanis et al., (2013, 2018) and Tzanis and Tripoliti, (2019), have appraised the dynamics of seismogenesis on the basis of correlation, as this is specified by the so-called entropic indices (Section 2) and principally by the index associated with the distribution of interevent times. Their analysis considered both full (whole) and background seismic processes (in which aftershocks were removed by stochastic declustering) so as to assess the dynamic states of the seismogenic continuum. Furthermore, the dependence of spatial and temporal correlation on magnitude and separation between successive events (interevent distance) was scrutinized. It was found that seismogenic systems may have very different dynamics, from SOC to Poissonian, that the state/level of correlation may be closely associated with their geodynamic setting and that it can change with time. Transformative systems generally appeared to be more correlated or even SOC, while systems of convergent/divergent plate margins appeared to be Complex but only moderately correlated. Finally, systems of Wadati-Benioff zones appeared to be generally Poissonian.

Herein we revisit the temporal entropic index, but this time in direct association with another important dynamic parameter: the $q$-relaxation interval which is the NESP analogue of the relaxation time and comprises an alternative definition of the recurrence interval (Section 2.1). In the analysis of interevent time distributions, the $q$-relaxation interval is the characteristic time required by a seismogenic system to produce an earthquake above a given magnitude and is expected to depend on, and convey information about the dynamic state of the system. For instance, due to long-range correlation and bottom-up organization, a critical system should be able to generate earthquakes of any magnitude within short intervals after the occurrence of any event. Conversely, a Poissonian system
would generate earthquakes within intervals dependent on its productivity, but monotonically/ non-linearly increasing with event size.

We apply the analysis described in Section 2.2 to many different, (single-fault and composite) seismogenetic systems of the northern Circum-Pacific Belt: these include the transformational plate boundaries and inland seismic regions of California, Alaska and Southwest Japan, the convergent plate boundaries and Wadati-Benioff zones of the Aleutian, Ryukyu, Izu-Bonin and Honshū arcs, and the divergent plate boundary of the Okinawa Trough. We adopt this approach for several reasons: a) Confirm and clarify the results of previous work with new information; b) Explore how and why the geodynamic setting may affect the dynamic state of seismogenetic systems; c) Assess the statistical effects of blending the seismicity of different systems in a single catalogue; d) Conduct a preliminary investigation of whether $q$-relaxation intervals can help in improving the analysis of earthquake hazard and risk. The large volume of data and results prohibits presentation of full and background processes in a single piece of work. Herein we report on the analysis of full processes, and reserve the presentation of background processes for follow-up work.

2. Non Extensive Approach to the Statistical Physics of Earthquakes

2.1. Overview

In statistical mechanics, an $N$-component dynamic system may have $W=N!/\Pi_i N_i!$ microscopic states, where $i$ ranges over all possible conditions (states). In classical statistical mechanics, the entropy of that system $S$ is related to the totality of these microscopic states by the Gibbs formula $S=-k \sum p_i \ln(p_i)$, where $k$ is the Boltzmann constant and $p_i$ is the probability of each microstate. If the components of the system do not interact and are statistically independent of each other, its entropy factorises into the product of $N$ identical terms, one for each component; this is the Boltzmann entropy $S_B=-Nk \sum p_i \ln(p_i)$. A basic property of this formalism is additivity (extensivity): the entropy of the system equals the sum of the entropy of their components. However, it is now widely appreciated that a broad spectrum of non-equilibrating natural and physical systems does not conform to this requirement. Such non-additive systems, which are also commonly referred to as non-extensive after Tsallis (1988), include statistically dependent (interacting) components, in consequence of which they acquire memory (feedback) and can no longer be described with Boltzmann-Gibbs (BG) statistical physics.

Tsallis (1988, 2009) formulated an appropriate description of non-extensive systems by introducing the concept of Non Extensive Statistical Physics (NESP) as a direct generalization of Boltzmann-Gibbs statistical physics. If $x$ is some dynamic parameter, the non-equilibrium states of non-extensive systems can be described by the entropic functional
(1)

\[ S_q(p) = \frac{k}{q-1} \left[ 1 - \int_0^x p^q(x) dx \right], \]

where \( p(x) dx \) is the probability of finding the value of \( x \) in \( [x, x+dx] \) so that \( \int_0^1 p(x) dx = 1 \), and \( q \) is the entropic index. In the limiting case \( q \to 1 \), Eq. (1) converges to the Boltzmann–Gibbs functional \( S_{BG} = -k \int p(x) \ln(p(x)) dX \). Like the Boltzmann-Gibbs, the Tsallis entropy is concave and fulfils the H-theorem but is not additive when \( q \neq 1 \). For a mixture of two statistically independent systems \( A \) and \( B \), the Tsallis entropy satisfies \( S_q(A, B) = S_q(A) + (1-q) S_q(A) S_q(B) \). This is known as pseudo-additivity and is further distinguished into super-additivity (super-extensivity) if \( q < 1 \), additivity when \( q \to 1 \) (i.e. Boltzmann-Gibbs statistics) and sub-additivity (sub-extensivity) if \( q > 1 \). Accordingly, the entropic index is a measure of non-extensivity.

By maximizing \( S_q \), it can be shown that when \( q > 0 \) and \( x \in [0, \infty] \), the cumulative probability function (CDF) of \( x \) is the \( q \)-exponential distribution (Tsallis, 1988, 2009; Abe and Suzuki, 2005)

\[ P(>x) = \exp_q \left( -\frac{x}{x_0} \right) = \left[ 1 - (1-q) \left( \frac{x}{x_0} \right) \right]^{\frac{1}{1-q}} \]

(2)

where \( x_0 \) is a characteristic value (\( q \)-relaxation value) of \( x \) and

\[ \exp_q(x) = \begin{cases} 
1 + (1-q)x, & x > 0 \\
0, & x \leq 0
\end{cases} \]

is the \( q \)-exponential function, such that for \( q = 1 \), \( \exp_q(x) = e^x \). Eq. (2) is a \( q \)-exponential distribution and for \( q > 1 \) defines a CDF of the Zipf-Mandelbrot kind. For sub-extensive systems with \( q > 1 \), \( P(>x) \) is a power-law with a tail. For extensive (random) systems with \( q = 1 \), \( P(>x) \) is an exponential distribution. Finally, for super-extensive systems with \( 0 < q < 1 \), \( P(>x) \) is a power-law with a cut-off so that \( P(>x) = 0 \) whenever the argument becomes negative; such systems are characterized by bounded correlation radii.

Eq. (2) can be used to derive the NESP equivalent of the Gutenberg-Richter Law (F-M distribution) in a manner relating energy and magnitude. This was pioneered by Sotolongo-Costa and Posadas (2004) and refined by Silva et al., (2006) and Telesca (2011, 2012); it comprises a first principles approach based on NSEP-compatible “fragment-asperity models” that consider the interaction of asperities and the fragments filling space between them, (which is supposed to modulate earthquake triggering). These models differ only in their assumption of how the energy stored in the asperities and fragments scales with their characteristic linear dimension. The model proposed by Telesca (2011, 2012), which assumes that the energy scales with the area of the fragments and asperities \( (E \propto r^2) \) so that \( M \propto \sqrt[3]{\log(E)} \); this is consistent with the empirical laws of energy–moment and moment–magnitude scaling and compatible with the well-studied rate-and-state friction laws of rock failure. Accordingly, the F-M distribution used herein is
\[ P(> M) = \frac{N(> M)}{N_0} = \left(1 - \frac{1 - q_m}{2 - q_M} \frac{10^{2M}}{\alpha^{2/3}} \right)^{2-q_M} \]

with the constant \( \alpha \) expressing the proportionality between the released energy \( E \) and the fragment size \( r \) and \( q_m \) is the magnitude entropic index.

2.2. Bivariate earthquake frequency distributions: Construction and NESP-based modelling

As stated in the Introduction, standard and “improved” definitions of the recurrence interval are based on the static cumulative Frequency – Magnitude (F-M) distribution. Accordingly they comprise approximations by proxy to the true long-term average recurrence interval over a given seismic region. Herein, the evaluation of earthquake recurrence will be attempted on the basis of a physically relevant parameter: interevent time. However, earthquake magnitudes and interevent times are not exactly unrelated. It is understood that the larger the magnitude scale, the longer the recurrence interval and interevent time. Accordingly, and in order to ensure the rigour of the analysis, the frequency distribution of interevent times (F-T) will be evaluated conditionally on the frequency distribution of magnitudes (F-M). Instead of considering only one-dimensional (univariate) F-T distributions, the analysis will be based on bivariate Frequency – Magnitude – Interevent Time (F-M-T) distributions thereby introducing additional constraints on the permissible variation of parameters, in line with the approach adopted in previous work (e.g. Efstathiou et al., 2015, 2017; Efstathiou and Tzanis, 2018; Tzanis et al., 2013, 2018; Tzanis and Tripoliti, 2019).

A bivariate F-M-T distribution can be constructed as follows: A threshold (cut-off) magnitude \( M_{th} \) is set and a bivariate frequency table (histogram) representing the empirical incremental distribution is first compiled. The empirical cumulative distribution is then obtained by backward bivariate summation as

\[ N_{\tau m} = \sum_{j-D_j}^{\tau} \sum_{i-D_i}^{m} \{ H_{ij} \Leftrightarrow H_{ij} \neq 0 \} \], \( \tau = 1, \ldots, D_T \), \( m = 1, \ldots, D_M \)

(4)

where \( H \) is the incremental distribution, \( D_M \) is the dimension of \( H \) along the magnitude axis and \( D_T \) is the dimension of \( H \) along the \( \Delta t \) axis. In this construct, the cumulative frequency (earthquake count) is \( N(\{M \geq M_{th}, \Delta t : M \geq M_{th}\}) \), so that the empirical probability \( P(\{M \geq M_{th}, \Delta t : M \geq M_{th}\}) \) is simply

\[ \frac{N(\{M \geq M_{th}, \Delta t : M \geq M_{th}\})}{N_0}, \quad N_0 = N(M = M_{th}, 0) = \parallel N \parallel_0 \].

(5)

A thus constructed empirical cumulative F-M-T distribution is shown in Fig. 1a (solid circles): It is based on a set of 6,358 events with \( M_{th} \geq 3.5 \), which occurred in the seismic region of northern California during 1968-2017.5 (see Section 3.1 for details). The distribution is shown in logarithmic frequency scale and comprises a well-defined surface in which the end-member \((M \geq M_{th}, \Delta t = 0)\) is the one-dimensional empirical Gutenberg – Richter law and the end-member \((M = M_{th}, \Delta t)\) is the one-
dimensional F-T distribution.

Assuming that magnitudes and interevent times are statistically independent, i.e. that the sequence of events does not depend on fault hierarchy, the joint probability \( P(M \cup \Delta t) \) factorizes into the probabilities of \( M \) and \( \Delta t \) in the sense \( P(M \cup \Delta t) = P(M) \ P(\Delta t) \). Then, by considering the empirical and escort probabilities to be identical,

\[
\frac{N(\{M \geq M_{th}, \ \Delta t : M \geq M_{th}\})}{N_0} = \left(1 - \frac{1 - q_M}{2 - q_M} \cdot \frac{10^H}{\alpha^{\Delta t}}\right) \cdot \left(1 - (1 - q_r) \cdot \Delta t \frac{\Delta t}{\Delta t_0} \right)^{\frac{1}{1 - q_r}},
\]

where \( q_M \) and \( q_r \) are the entropic indices for the magnitude and interevent times respectively and \( \Delta t_0 \) is the \( q \)-relaxation interval, analogous to the relaxation time often encountered in the analysis of physical systems. \( \Delta t_0 \) it is the characteristic time required by an active fault network to generate an earthquake of magnitude \( M(\geq M_{th}) \) and as such it is by definition a type of recurrence interval enjoying the cardinal advantage of being based on direct observations of its dynamic expression.

On taking the logarithm and setting \( a = \log(N_0) \), Eq. (6) becomes

\[
\log N(\{M \geq M_{th}, \ \Delta t : M \geq M_{th}\}) = a + \left(2 - q_M\right) \cdot \log \left(1 - \frac{1 - q_M}{2 - q_M} \cdot \frac{10^H}{\alpha^{\Delta t}}\right) + \frac{1}{1 - q_r} \log \left(1 - \Delta t_0^{\Delta t} (1 - q_r) \Delta t\right)
\]

Eq. (7) is a generalized bivariate law of the Gutenberg – Richter kind in which \( b_q = \left(2 - q_M\right) \cdot (q_M - 1)^{-1} \) is the NESP generalization of the \( b \) value (also see Telesca, 2012).

The parameters of Eq. (7) can be approximated with non-linear least-squares. Because these are all positive (bounded from below), and the entropic indices are also bounded from above, they are solved with the trust-region reflective algorithm (e.g. Moré and Sorensen, 1983; Steihaug, 1983), together with least absolute residual (LAR) minimization to suppress outliers. Fig. 1a illustrates the fitted model (continuous surface). The solution is associated with 154 degrees of freedom and the approximation is excellent (correlation coefficient \( R^2 \approx 0.994 \)). The magnitude entropic index \( q_M = 1.534 \pm 0.002 \) so that \( b_q = 0.87 \), which compares very well to \( b \) values of the order of 0.87-0.91 computed for this data set with conventional techniques. The temporal entropic index \( q_r \) is approximately 1.342 \pm 0.002 and indicates moderate sub-extensivity. The \( q \)-relaxation interval \( \Delta t_0(M \geq 3.5) \), i.e. the characteristic time for the recurrence of events with \( M \geq 3.5 \) over the entire seismic region of northern California is \( 4.38 \times 10^3 \pm 5.1 \times 10^5 \) years or 1.599 days. Finally, the energy scaling constant \( \alpha = 75.59 \pm 26.7 \).

Fig. 1b presents a succinct statistical appraisal of the result, performed by fitting a normal location-scale distribution (dashed line) and a Student-t location-scale distribution (solid line) to the cumulative probability of the sorted residuals (r). Approximately 90% of the residual population, for which -
0.2 \leq r \leq 0.101, is normally distributed. The tail forming at \( r > 0.1 \) is fairly well fitted with the \( t \)-location-scale distribution and thus represent statistically expected outliers. It is interesting to note that the properties of the distribution are determined by the populous small-moderate magnitude scales and interevent times, and that outliers are mainly observed at moderate-large magnitudes and longer interevent times. Outliers frequently arise from flaws of the catalogue, (e.g. omitted events, glitches in magnitude reporting etc.), but in some cases could be genuine exceptions to the norm: for instance, they may correspond to rare, externally triggered events. Such details will not be examined herein.

When this type of analysis is carried out for different magnitude thresholds one obtains tables and graphs of the variation of the entropic indices, \( q \)-relaxation interval, energy scaling constant and other parameters pertaining to the numerical solution of Eq. (7). Fig. 2 illustrates the analysis of a catalogue of 19281 events with \( M \geq 3.0 \), which occurred in the seismic region of northern California during 1968-2017.5 (see Section 3.1 for details). Fig 2a illustrates the variation of the entropic indices \( q_M \) and \( q_T \) with magnitude; Fig. 2b the increase of the \( q \)-relaxation, where it is interesting to observe that it can be described by an exponential function; Fig. 2c illustrates the variation of the energy scaling constant \( \alpha \), Fig. 2d the variation of the goodness of fit (R\(^2\)) and, finally, Fig. 3e the variation of the degrees of freedom associated with the solution of Eq. (7).

### 3. EARTHQUAKE DATA AND ANALYSIS

The present study focuses on analysis of the \( q \)-relaxation time \( \Delta t_0 \) in association with the temporal entropic index \( q_T \) using data from major seismic regions of the northern half of the Circum Pacific Belt: California, Alaska, the Alaskan-Aleutian Arc, the Ryukyu Arc and Okinawa plate, the Izu-Bonin Arc, and the Honshū Arc and Okhotsk plate. Basic information about the tectonic setting and properties of the earthquake catalogues is given in Section 4 and Table 1 of the supplementary material. As explained in Section 2.2, the magnitude entropic index \( q_M \) and energy scaling constant \( \alpha \) are computed so as to constrain the estimation of \( \Delta t_0 \) and \( q_T \), but are not otherwise considered; the properties of \( q_M \) have thoroughly been investigated in previous work (Efstathiou et al, 2015, 2017; Efstathiou and Tzanis, 2018; Tzanis et al., 2018, Tzanis and Tripoliti, 2019). The results are summarized in Tables 2-4 of the supplementary material. In order to maintain experimental rigour, parameter estimation was not performed for catalogue subsets containing less than 500 events and results were not considered unless associated with a goodness of fit \( R^2 \geq 0.97 \).

The joint examination of \( \Delta t_0 \) and \( q_T \) is useful because the temporal entropic index designates dynamic state of a seismogenetic system, which can be extensive (Poissonian) or non-extensive, with the later having the possibility to be Non-Critical, Sub-Critical or Critical. If the system is non-extensive, it generates a sequence of correlated events which depend on their predecessors and influence their successors. The degree of correlation is evaluated by the entropic index so that if \( q \neq 1 \), the system is...
non-extensive, whereas if \( q \to 1 \) the system is extensive (uncorrelated and memoryless). It follows that the \( q \)-relaxation interval is indivisibly associated with the dynamic state of the system and the joint evaluation of \( \Delta t_0 \) and \( q_T \) may provide information by which to characterize the state of a seismogenetic region. Previous work (Efstathiou et al., 2015, 2017; Tzanis et al., 2018) has demonstrated that the systematic observation of experimental values \( q_T \geq 1.15 \) would be compelling evidence of sub-extensive dynamics and has classified the degree of sub-extensivity (correlation) as: insignificant when \( q_T < 1.15 \), weak when \( 1.15 \leq q_T < 1.3 \), moderate when \( 1.3 \leq q_T < 1.4 \), significant when \( 1.4 \leq q_T < 1.5 \), strong when \( 1.5 \leq q_T < 1.6 \) and very strong when \( 1.6 \leq q_T \).

3.1 California, USA

The prominent seismogenetic feature of California is the San Andreas Fault (SAF). It comprises a NW to NNW oriented, 1300 km long, right-lateral transformational boundary between the Pacific plate to the west and the North American plate to the east, and has generated several large earthquakes during the past two centuries. The SAF system (main, “sibling” and co-lateral faults) is generally thought to comprise three major segments: The Mojave segment in South California, between Salton Sea (approx. 33.36°N, -115.7°E at SE California) and Parkfield, Monterey County (approx. 35.9°N, -120.4°E), the central segment between Parkfield and Hollister (approx. 36.85°N, -121.4°E), and the northern segment from Hollister and through the San Francisco bay area to Mendocino Fracture Zone (offshore, approx. 40.36°N, -124.5°E).

The SAF accommodates about 75% of the total motion between the North American and Pacific plates. The remaining 25% is accommodated by NNW-SSE right-lateral shear deformation concentrated in a zone east of the Sierra Nevada mountain range, called the Walker Lane or Eastern California Shear Zone (Wesnousky, 2005; Guest et al., 2007). The northern terminus of the Walker Lane is located at approximately (40.3°N, -120.6°E), between the Pyramid Lake, Nevada and Lassen Peak, California where the Honey Lake Fault Zone meets the transverse tectonic zone forming the southern boundary of the Modoc and Columbia plateaus with the Great Basin. The Walker Lane extends southward of the intersection of the Death Valley with the Garlock Fault, crosses the Mojave Desert and terminates on the San Andreas Fault between Salton Lake and the San Bernardino Mts.

To complicate things, California is geologically divided into northern and southern by the SW-NE left-lateral Garlock fault which extends for almost 250 km between its junction with the East California Shear Zone (ECSZ) at the north-eastern edge of the Mojave Desert (35.6°N, -116.4°E) and its junction with the SAF at Tejon Pass (34.8°N, -118.9°E). This major tectonic boundary developed to accommodate the strain differential between the E-W extension of the Great Basin eastwards of the ECSZ (e.g. Wernicke et al., 1988), and the NW-SE right lateral transformation of the ECSZ and SAF. Thus, the right-lateral motion on the SAF and ECSZ locks up in the area of the Garlock, where local
variations in the mode of deformation and earthquake focal mechanisms are observed (e.g. Jones, 1988; Hardebeck and Hauksson, 2001; Fialko, 2006; Becker et al, 2005). Between 37.7°N and 35.1°N, the left-lateral motion of the Garlock fault generates a restraining bend and a broad S-shaped westward displacement of the SAF, known as the “Big Bend”.

North of the Garlock Fault and bounded by the SAF to the east and the Walker Lane to the west, lies the west-tilting semi-rigid Sierra Nevada (or Sierran) microplate, whose interior (Central Valley) is characterized by the absence of significant faults and large earthquakes (Hammond et al., 2012; Saleeby et al., 2009; McCaffrey 2005; Dixon et al., 2000; Goter et al, 1994). South of the Garlock extends the South California Seismic Region or SCSR. The northern boundary of the SCSR is the WNW-ESE Santa Ynez and Garlock Fault zones from the Pacific coast (34.5°N, -120.5°E) and through Tejon Pass to approximately (35.5°W, -116.4°E) in the Mojave Desert. Then, it turns to the south and eastward of the South Bristol Mts. Fault (34.6°N, -115.6°E) runs to approximately (32.0°N, -114.5°E) in the Gran Desierto de Altar (Sonora, Mexico), north of the head of the Gulf of California; it turns westwards to approx. (32°N, -117°E) south of Tijuana, Mexico, and then to (32°N, -119°E) off the coast of Mexico. Finally, it turns north and runs parallel to the coastline and west of the San Clemente and Santa Cruz Islands up to 34.5°N. The SCSR is characterized by several major faults and numerous branches that create a complex seismic landscape.

The seismicity of Northern California is monitored by North California Seismic Network (NCSN) and the respective regional catalogue is published by North California Earthquake Data Centre (http://www.ncedc.org). The seismicity of Southern California is monitored by the Southern California Seismic Network (SCSN) and the regional catalogue is published by the South California Earthquake Data Centre (http://www.data.scec.org). Prior to analysis, both catalogues were pre-processed (homogenized) and thoroughly examined for completeness; details can be found in Efthathiou et al. (2017) and Efthathiou and Tzanis (2018), as well as in the supplementary material. As a result, both catalogues were reduced to the $M_L$ (local magnitude) scale. The NCSN catalogue was found to be complete for $M_L \geq 3.0$ as of 1968. As for the SCSN catalogue, the sustainable magnitude of completeness ($M_c$) was found to be approximately 3.0 during the period 1968-1975 and to decrease to 2.5 as of the early 1980’s. As demonstrated by Efthathiou and Tzanis (2018), if $M_c \geq 3.0$ the SCSN catalogues yield almost identical results for the periods 1968.0-2017.5 and 1980.0-2017.5. Accordingly and in order to examine small magnitude scales, the SCSN catalogue is considered only for the period 1980.0-2017.5 in which $M_c \geq 2.5$.

In addition to the division into northern and southern regions, the broader Pacific – North-American boundary can further be sub-divided into six earthquake source sub-areas as illustrated in Figs. 3 and 4. These are the north and south segments of the San Andreas Fault, the north and south segments of the Walker Lane (Sierra Nevada Range and Eastern California Shear Zone), the Mendocino Fracture
Zone in north California and the Inner Continental Borderland region in southern California.

3.1.1 San Andreas Fault (SAF) System

The source area of the north segment of San Andreas Fault (henceforth nSAF), extends north of the Garlock Fault between Parkfield and the Mendocino Fracture Zone (Fig. 3). In this study, its outline is defined to the north by the line joining the northern terminus of the SAF/Shelter Cove section (40.2°N, -124.3°E), the northern terminus of the Bartlett Springs Fault System (Lake Mountain fault) and the Battle Creek Fault (40.5°N, -121.9°E); to the east by the Battle Creek Fault, the Foothills Fault system (roughly 39.3°N, -118.8°E) and the Kern Gorge fault and White Wolf fault zone (35.3°N, -118.6°E) and to the West by an offshore imaginary line parallel to the Pacific Coast.

nSAF exhibits different dynamics with reference to the M7.2 Loma Prieta event of 1989.797 (Fig. 5): prior to that event, the temporal entropic index $q_T$ indicates moderate to strong correlation (sub-extensivity) over all magnitude scales (Fig. 5a); after the event, only indicated insignificant to weak correlation is observed (Fig. 5c). Analogous observations were made by Efstathiou et al. (2017), albeit with small differences due to different earthquake populations in the post-1990.0 period. Similar behaviour can be observed in the $q$-relaxation intervals ($\Delta t_0$). Prior to 1989.0, $\Delta t_0$ is $2.6 \times 10^3$ years (roughly one day) at $M_{th}=3$ and increases exponentially to $2.77 \times 10^2$ at $M_{th}=3.9$ (10 days), and to $4.35 \times 10^2$ years (16 days) at $M_{th}=4.1$ (Fig. 5b). Moreover, it is comparable to the standard recurrence interval for $3 \leq M_{th} \leq 3.9$, although the two appear to deviate at larger magnitude scales. Following 1990.0, $\Delta t_0$ varies exponentially from $9.65 \times 10^3$ years (3.5 days) at $M_{th}=3$, to $8.1 \times 10^3$ years (approx. 29.6 days) at $M_{th}=3.9$ (Fig. 5d). With respect to the strongly correlated period 1968.0-1989.0, this represents an almost threefold increase; $\Delta t_0$ also appears comparable to the standard recurrence interval in the sense that the latter varies exclusively within the 95% simultaneous prediction intervals associated with the former. When the entire 49.5 year period since 1968.0 is considered, the results are intermediate to those of Fig. 5, apparently due to the mixing of different dynamics prior to, and after the Loma Prieta earthquake. Thus, $q_T$ indicates weak to moderate correlation (Fig. 6a), while $\Delta t_0$ increases from $5.8 \times 10^3$ years (2.1 days) at $M_{th}=3.0$, to $6.4 \times 10^2$ years (23.4 days) at $M_{th}=3.9$ and $1.25 \times 10^2$ years (45.5 days) at $M_{th}=4.2$ (Fig. 6b). The differences between the $q$-relaxation and standard recurrence intervals are very small for $M_{th} \leq 3.4$ but systematically diverge thereafter, with the standard recurrence interval being as low as $9.45 \times 10^2$ years (34.5 days) at $M_{th}=4.2$.

The south segment of the San Andreas Fault (sSAF) is sandwiched between the Eastern California Shear Zone and the Inner Continental Borderland Region (Fig. 4); it accounts for 75% of the local slip rate between the Pacific and North American plates and comprises a tripartite system of large sub-parallel faults: the eponymous San Andreas Fault in the east, the San Jacinto Fault (SJF) in the centre and the Elsinore Fault (EF) in the west. To the north, the sSAF terminates on the left-lateral Garlock
fault. The southern boundary of the sSAF, if any, is not clearly defined. The eponymous fault terminates at the southeast corner of the Salton Sea but is thought to connect with the Imperial Fault (IF) though the extensional Brawley Seismic Zone (BSZ). The San Jacinto and Elsinore faults also extend to the SE, with San Jacinto also terminating against the Imperial Fault and Eslinore continuing into Mexico as the Laguna Salada Fault (LSF) where the M7.2 Baja California event of 2010 has occurred.

Fig. 6c illustrates the variation of the temporal entropic index with threshold magnitude. $q_t(M_{th})$ is lower than 1.2 at small magnitude scales (insignificant-weak correlation), but for $M_{th}>3.6$ increases steeply and for $M_{th}=3.9$ exceeds 1.6 (very strong). The shape of the $q_t(M_{th})$ curve will be explained below. Efstathiou and Tzanis (2018) attributed the increase of correlation to corresponding increase in the interaction radii associated with the increase in the size of events and also found that correlation is very strong up to interevent distances (ranges) of 150 km, dropping to moderate thereafter. The $q$-relaxation interval $\Delta t_0$ increases exponentially from $4.22\times10^3$ years (1.5 days) at $M_{th}=2.5$, to $1.58\times10^2$ years (5.81 days) at $M_{th}=3.0$, and 0.158 years (57.7 days) at $M_{th}=4.2$ (Fig. 6d). The $q$-relaxation and standard recurrence intervals are comparable for $M_{th}=2.9$ but the former diverges thereafter, becoming significantly longer at larger magnitudes.

Inasmuch as the accuracy of epicentre locations allows it, it is interesting to study two major components of sSAF separately, in order to examine their individual dynamics and study the effect of blending their statistical properties. Accordingly, Figs. 7a-b present results for the Elsinore – Laguna Salada faults (EF/LSF) and Fig. 7c-d the same for the San Jacinto/San Andreas – Brawley Seismic Zone – Imperial faults (SJF/SAF). In Fig. 7a, $q_t(M_{th})$ can be seen to increase steeply from 1.2-1.4 at small magnitude scales (insignificant-moderate correlation), to 1.54 at $M_{th}=3$ (strong correlation) and to 1.7-1.9 at $M_{th}=3.2$ (very strong correlation). In Fig. 7b, the $q$-relaxation intervals are practically the same for all threshold magnitudes and oscillate about a mean value of $1.1\times10^2 \pm 5.3\times10^3$ years (3.9±1.95 days). This implies that soon after any event, EF/LSF responds by generating practically any magnitude (up to the observed $M_l=4$), anywhere within its boundaries! The standard recurrence interval is vastly different from the $q$-relaxation interval! In Fig. 7c, $q_t(M_{th})$ is seen to increase quasi-linearly from 1.09 at $M_{th}=2.5$ (insignificant correlation) to 1.35 at $M_{th}=3.9$ (moderate correlation). In Fig. 7d, the $q$-relaxation appears to increase exponentially from $8.3\times10^3$ years (3 days) at $M_{th}=2.5$ to 0.1 years (38.5 days) at $M_{th}=3.9$; $q$-relaxation and standard recurrence intervals are comparable within the range of observations, although the latter appears to diverge and escalate faster than the former at larger magnitudes.

It is apparent that the two sSAF sub-systems exhibit very different dynamics, the nature of which will be discussed in Section 4. Their analysis, however, helps in understanding the combined properties of the tripartite sSAF system. Thus, the insignificant to weak correlation observed in Fig. 6c up to
$M_\text{th}=3.6$ can be explained by the quasi-Poissonian nature of the SAF/SJF and the randomization effected by blending statistically different earthquake populations (Fig. 6c). Conversely, the steep increase and strong correlation observed at magnitudes larger than 3.6 can be explained by the very strong correlation of EF/LSF seismicity in combination with the moderate to strong correlation of SJF/SAF seismicity: randomization is still present but not as severe! It is also worth noting that in both Fig. 7a and 7c, the correlation increases with $M_\text{th}$.

3.1.2 The Walker Lane

The northern segment of the Walker Lane defines the eastern boundary of the Sierra Nevada microplate and will henceforth be referred to as SNR (Fig. 3). In this study, its source area is bounded to the north by the line joining the Battle Creek Fault and the northern termini of the Butt Creek and Almanor fault zones (roughly 44.5°N, -121.2°E) up to -116°E; to the east by the -116°E meridian; to the south by the Garlock Fault and to the west by the White Wolf and Kern Gorge fault zones, the Foothills Fault system and the Battle Creek Fault. As shown by Efstathiou et al., (2017), SNR was not affected by the Loma Prieta event and its dynamics did not change during the period 1968.0-2012.0 considered therein. The present analysis (1968.0-2017.42) is shown in Fig. 8a and 8b and confirms that of Efstathiou et al., (2017). The temporal entropic index $q_T(M_\text{th})$ varies from strong (1.53) at $M_\text{th}=3$ to very strong (1.88) at $M_\text{th}=4.2$, increasing in a quasi-linear fashion (Fig. 8a). This would again signify progressive increase of interaction radii, namely longer-range interaction. In addition, Efstathiou et al. (2017) showed that $q_T$ indicates moderate to high correlation (1.36-1.66) over all ranges (interevent distances), which is very significant for interpretation. The $q$-relaxation intervals are very short (Fig. 8b): they vary from $2.5\times10^3$ years (1 day) at $M_\text{th}=3.0$, to $1.44\times10^2$ years (5.3 days) at $M_\text{th}=4.2$. This is analogous to the case of EF/LSF and implies that SNR will swiftly generate any magnitude (at least up to $M_L=4.2$). In Fig. 8b, the standard recurrence interval is substantially longer than the $q$-relaxation interval and that the difference escalates non-linearly.

The south segment of the Walker Lane, also known as Eastern California Shear Zone (ECSZ) extends southward of the intersection between the Garlock Fault and the Death Valley and runs across the Mojave Desert terminating on the San Andreas Fault between Salton Lake and the San Bernardino Mts (Fig. 4). This is a zone of dextral strike-slip faults that accommodates approximately 25% of the total motion between the North American and Pacific plates (Dixon et al., 2000; Miller et al., 2001) and has generated large earthquakes (e.g. Landers, 1992, $M_w=7.3$ and Hector Mine, 1999, $M_w=7.1$). The eastern expanse of the ECSZ is delimitied by the diffuse extensional deformation of the Basin and Range province. Although its origin is still open to debate, it has been suggested that it formed by northward propagation of the plate boundary in the Gulf of California due to the northward motion of the Baja California microplate (Faulds et al., 2005a,b; Harry, 2005; McCrory et al., 2009).
Fig. 8c illustrates the variation of the temporal entropic index with magnitude: $q_T(M_{th})$ exhibits quasi linear increase of generally very strong correlation from about 1.7 at $M_{th} = 2.5$ to higher above 1.9 for $M_{th} \geq 3.7$; it also averages to 1.8±0.03 for $M_{th} \leq 3.0$ and 1.95±0.03 for $M_{th} \geq 3.1$. The results are compatible with those of Efstathiou and Tzanis (2018), who also determined that correlation is persistently and uniformly strong across ECSZ. As in EF/LSF and SNR, the $q$-relaxation intervals are very short (Fig. 8d); they vary from $9.56 \times 10^{-6}$ years (a few minutes) at $M_{th} = 2.5$, to $1.14 \times 10^{-3}$ years (0.42 days) at $M_{th} = 3.0$ and to $1.05 \times 10^{-2}$ years (3.8 days) at $M_{th} = 3.8$. The standard recurrence intervals are much longer than the $q$-relaxation interval and the difference escalates non-linearly. As it turns out, the entire Walker Lane (SNR and ECSZ) is persistently and very intensely correlated over all magnitudes and ranges, and also exhibits very short $q$-relaxation intervals.

3.1.3 Mendocino Fracture Zone (MFZ) and Inner Continental Borderland region (ICB)

The Mendocino Fracture Zone is bounded by the coordinates 40°N to 43°N and -123°E to -128°E (Fig. 3) and comprises a E-W right-lateral transformational plate boundary between the Pacific and Gorda plates, off the coast of Cape Mendocino in northern California (e.g. Dickinson and Snyder, 1979; Furlong and Schwartz, 2004). It extends westward from its transform–transform–trench junction with the San Andreas Fault and the Cascadia subduction zone (Mendocino Triple Junction), to the southern end of the Gorda Ridge at approx. (40.4°N, -128.7°E); it then continues on as an inactive segment for several hundred kilometres. The MFZ includes the most active part of northern California (Yeats, 2012) and according to Dengler et al. (1995), the north coastal region accounted for about 25% of the seismic energy released in California in a 50 year period.

Fig. 8a shows that $q_T(M_{th})$ exhibits a peculiar oscillation but always remains lower than 1.3 (insignificant to weak correlation), averaging to 1.2±0.07. This is compatible with the results of Tzanis et al., (2018), who also found that correlation is no higher than moderate ($q_T \leq 1.4$) over all ranges. The $q$-relaxation intervals increase exponentially from $1.2 \times 10^{-2}$ years (4.4 days) at $M_{th} = 3.0$, to $9.6 \times 10^{-2}$ years (35 days) at $M_{th} = 4.2$ (Fig. 8b). For $M_{th} \geq 3.7$ they are comparable to the standard recurrence intervals but escalate at a lower rate so that by $M_{th} = 4.2$ they are approximately 10 days shorter; the trend appears to persist at larger magnitude scales.

The Inner Continental Borderland region (ICB, see Fig. 4), contains several faults and extends offshore and to the west of the southern California mainland, from Point Conception to the Vizcaino Peninsula in Baja California. ICB is a tectonically complex system in which seismicity appears more diffuse than in the mainland, although this may be an artefact of lopsided network geometry and structural heterogeneity (Astiz and Shearer, 2000; references therein). The area can be divided into four major sub-parallel groups of dextral faults which, from east to west are: i) the Newport–Inglewood (NIF) and Rose Canyon (RCF) faults that make landfall at San Diego and perhaps connect with the Vallecitos and San Miguel faults in Baja California; ii) the Palos Verdes (PVF) – Coronado
Bank (CBF) fault that makes landfall near Ensenada, Mexico; iii) the Santa Cruz – Santa Catalina – San Diego Trough – Bahia Soledad (SDTF) fault that makes landfall south of Punta, Mexico; iv) the Santa Cruz – San Clemente – San Isidro fault zone (SCF). During the past 50 years, several moderate ($M_L 5$ to $6$) earthquakes have occurred in the region, consistent with the right-lateral deformation of the local Pacific–North American plate boundary and the regional tectonics of the San Andreas Fault system (e.g., Weldon and Humphreys 1986).

The variation of the temporal entropic index with threshold magnitude in ICB, is shown in Fig. 8c: for $M_{th} \leq 2.8$ correlation is insignificant and for $M_{th} \geq 2.9$ increases quasi-linearly to strong ($q_T = 1.52$ at $M_{th} = 3.4$); beyond this, earthquake populations are not sufficient to guarantee statistical rigour. The variation of $q_T$ is similar to that in sSAF and should admit the same interpretation. Again, this is compatible with Efstathiou and Tzanis (2018) who also determined that significant to strong correlation exists only at ranges shorter than 70km and is explainable by aftershock sequences, while it drops to weak-moderate at long ranges. The $q$-relaxation interval definitely increases between $M_{th} = 2.5$ (2.45×10$^{-2}$ years/ 9 days) and $M_{th} = 3.4$ (5.6×10$^{-2}$ years/ 20.4 days) but the short magnitude bandwidth and unenviable stability of $q_T$ estimators do not allow a definitive model of their growth (Fig. 8d). The durations of $q$-relaxation intervals are comparable to those of sSAF and appear to significantly differ from standard recurrence intervals (which escalate faster beyond $M_{th} = 3$).

### 3.1.4 Northern and Southern Seismic Regions

The results presented above show that the major seismogenetic systems of California are composites of sub-systems with very different dynamics. Given that many other broad seismic regions are composite, it is interesting to study the effect of mixing statistically different sub-systems in regional-scale earthquake catalogues. We, therefore, proceed to study the regional catalogues of NCSR (Fig. 3) and SCSR (Fig. 4).

Let us begin with NCSR, i.e. the blended nSAF, MFZ and SNR systems. As apparent in Fig. 10a, the temporal entropic index is consistently around 1.29±0.03 for threshold magnitudes up to 4.2 (moderate correlation) and steadily declines to 1.14±0.005 at $M_{th} = 4.6$ (insignificant correlation). The $q$-relaxation interval increases exponentially from 1.44×10$^{-3}$ years (approx. 0.5 days) at $M_{th} = 3$ to 0.11 years (40 days) at $M_{th} = 4.6$ and is practically identical to the standard recurrence interval (Fig. 10b). In SCSR, (blended ICB, sSAF and ECSZ systems), $q_T(M_{th})$ is lower than 1.2 at small magnitudes but increases steadily for $M_{th} \geq 3.5$ and attains values higher than 1.6 when $M_{th} > 4$ indicating very strong correlation (Fig. 10c). In short, it behaves in a fashion similar to sSAF! The $q$-relaxation interval increases exponentially from 2.61×10$^{-3}$ years (0.95 days) at $M_{th} = 2.5$, to 8.98×10$^{-3}$ years (3.3 days) at $M_{th} = 3.0$ and 0.13 years (46 days) at $M_{th} = 4.5$ (Fig. 10d). The $q$-relaxation and standard recurrence intervals are comparable in spite of instabilities in the variation of the former. The durations and variation of $q$-
relaxation intervals are comparable between the NCSR and SCSR. Careful comparison of Fig. 10 and Figs. 5–9 reveals that the results for the regional catalogues are intermediate to those of the individual seismogenetic fault systems and are apparently biased towards sub-systems contributing with larger earthquake populations. This is understandable considering that the mixing and chronological sorting of earthquakes from different and possibly non-interacting fault networks is bound to randomize the (composite) regional catalogue.

3.2 Alaska and the Alaskan – Aleutian Arc

The Aleutian Arc and Continental (mainland) Alaska source areas are bounded by the coordinates 50°N to 70°N and -196°E to -126°E. The principal geodynamic feature of this area is the North American – Pacific plate boundary (Fig. 11). The eastern boundary is defined by the Queen Charlotte – Fairweather (QC-F) dextral transform fault system, parallel to which the Pacific plate moves N-NW relative to the North American plate at a rate of approx. 50 mm/year. The boundary transits from transformational to convergent along a zone extending between (57.5°N, -137°E) and (59°N, -145.5°E), in which the Yakutat Terrane accretes to the North American plate; it then continues westwards as the Aleutian Arc and Trench system. Landward of the QC-F lays the right-lateral Denali transform fault. This is an arcuate feature running for approx. 750km in a northwesterly direction from about (59°N, -135.3°E) to about (63.5°N -147°E); it then bends westwards and continues almost parallel to the plate boundary for an additional 500km, to approx. (63°N, -155,2°E). The Aleutian Arc and Trench extends for approx. 3400km, from the northern end of the QC-F in the east (near 58.5°N, -137°E), to a triple junction with the Ulakhan Fault and the northern end of the Kuril-Kamchatka Trench in the west (near 56°N, -196°E). Westward of the Alaska Peninsula (Unimak Pass, 55.7°N, -164°E) the convergence transits from continental in the east to intra-oceanic in the west. Subduction along generates the Aleutian Volcanic Arc that extends as far as -182°E. The motion of the Pacific plate is always to the N-NW but due to the arcuate geometry of the trench, the vector of convergence changes from almost trench-normal in the east (Gulf of Alaska) to almost trench-parallel in the west. Along the continental part of the subduction, the rate of convergence varies from 56mm/year in the east (Gulf of Alaska), to 63mm/year in the west (); along the oceanic part, the rate varies from 63mm/year near Unimak Pass to 74 cm/year in the far west (e.g. DeMets and Dixon, 1999).

For the most part, seismic energy is released by large events. Within the North American plate (Continental Alaska), the highest seismicity rates are observed in southern Alaska, parallel to the plate boundary and decrease northwards. The transformational plate boundary also involves several secondary faults both seaward and landward of the transform faults, which accommodate a smaller fraction of the relative plate motion. The Aleutian Arc and Trench generate large numbers of earthquakes in the crust and in the subducting and overriding plates. Additionally, many earthquakes are associated with the Aleutian Volcanic Arc. Most large earthquakes have thrust mechanisms,
although some shallow (<30km) events are either strike-slip or normal. Most of the normal faulting events in the Aleutian outer rise region are caused by the bending of the Pacific plate as it enters the trench, while most of the shallow strike-slip events are concentrated along the island axis.

The earthquake data was extracted from the regional earthquake database of the Alaska Earthquake Centre (http://www.aeic.alaska.edu/html_docs/db2catalog.html) and comprises a total of 48,995 events recorded in the area 50°N to 70°N and -196°E to -126°E over the period 1968.0–2016.0 The catalogue was homogenized to the local magnitude scale ($M_L$) and thoroughly examined for completeness. Details can be found in the supplementary material and Tzanis et al., (2018). In the Aleutian Arc, the sustainable magnitude of completeness is $M_c \geq 4.4$ so that reliable analysis is limited to magnitudes above that level. Conversely, in Continental Alaska it is possible to consider all earthquakes with $M_L \geq 3$, for which the catalogue appears to be complete.

Seismogenesis in Alaska and the Aleutian Arc develops in a complex tectonic background, extends over a very large area and range of depths and exhibits regional variation. Along the broader North American – Pacific boundary (which is the focus of our analysis), it is possible to distinguish three classes of activity:

a) Crustal earthquakes in Continental Alaska primarily associated with the eastern transformational plate boundary, namely with the Queen Charlotte – Fairweather and Denali faults and the transitional zone spanned by the Yakutat and Wrangelian terranes. This source area will henceforth be referred to as Queen Charlotte – Denali Zone, or QCD (Fig. 11a).

b) Crustal earthquakes along the Alaskan – Aleutian Arc primarily associated with the convergent plate margin; these are crudely distinguished according to the depth of the Mohorovičić discontinuity, which is approx. 40 km beneath the Yakutat Terrane (Christeson et al., 2013) and 38.5 km along the Aleutian Arc (Janiszewski et al., 2013). The source area will henceforth be referred to as AT-C (Aleutian Trench – Crustal, see Fig. 11a).

c) Sub-crustal earthquakes in the Wadati-Benioff zone below the Mohorovičić discontinuity. This source volume will henceforth be referred to as AT-D (Aleutian Trench – Deep, see Fig. 11b).

This distinction provides an opportunity to compare the dynamics of earthquake populations generated in different seismotectonic settings and are subject to different environmental (crust vs. subducting slab) and boundary conditions (free in the crust vs. fixed in the slab).

Beginning with QCD, Fig. 12a shows that $q_q(M_{th})$ starts off as low as 1.1, but demonstrates steady linear increase, transcending the threshold of randomness at $M_{th}=3.2$ and climbing to 1.44 for $M_{th}=4.3$ (significant correlation). Analogous observations were made in the transformative seismogenetic systems of California, therefore, the same interpretation should apply. Tzanis et al., (2018) also found that long-range correlation is significant to strong at interevent distances of 300-600km and moderate thereafter. The $q$-relaxation interval increases with magnitude according to well-defined exponential law from $1.68 \times 10^{-2}$ years (6.1 days) at $M_{th}=3$ to 0.11 years (40 days) at $M_{th}=4.4$ (Fig.11b).
standard recurrence interval is very comparable to the $q$-relaxation interval for $M_{th}\leq 4$ but clearly diverges thereafter, escalating at a faster pace; very notably, the onset of divergence coincides with the transition from moderate to significant correlation (Fig. 10).

In AT-C, $q_T(M_{th})$ exhibits a slow upward trend from approx. 1.1 at $M_{th}=4.4$ to over 1.2 at $M_{th}\geq 5.1$, averaging to $\langle q_T(M_{th})\rangle=1.2\pm 0.054$ (Fig. 13a); it is evidently borderline between Poissonian and weakly Complex. Tzanis et al. (2018) have shown that correlation at ranges up to 200km is only moderate, dropping to weak-insignificant thereafter.

The $q$-relaxation interval increases exponentially from 1.4×10^{-2} years (5 days) at $M_{th}=4.4$ to 0.121 years (44.2 days) at $M_{th}=5.2$ (Fig. 13b); $q$-relaxation and standard recurrence intervals are generally comparable with the latter varying within the 95% prediction bounds of the exponential law fitted to the former.

Finally, $q_T$ is generally lower than 1.15 in AT-D (Fig. 13c). Tzanis et al., (2018) found that correlation is insignificant to weak for ranges up to 700km. The $q$-relaxation interval increases exponentially from 3.16×10^{-2} years (11.5 days) at $M_{th}=4.4$ to 0.121 years (44.2 days) at $M_{th}=5.2$ (Fig. 13d); $q$-relaxation and standard recurrence intervals are generally comparable/congruent with the latter varying within the 95% prediction bounds of the exponential law fitted to the former. On the basis of this evidence, the sub-crustal fault network appear to be Poissonian.

The analysis of seismicity along the Aleutian Arc a clear departure from hitherto observations: as opposed to the apparently sub-extensive dynamics of seismogenesis in the transformational plate boundaries in California and Alaska, this convergent plate boundary exhibits predominantly Poissonian characteristics. This is not the only case, as will be seen below.

### 3.3 North-West Segment of the Circum-Pacific Belt

This study area extends from 22°N to 46°N (Hokkaido, Japan) and from 122°E (east of Taiwan) to 146°E in the Pacific (Fig. 14); it includes several major convergent and one divergent plate boundaries, transformational plate boundaries and inland seismogenetic domains. Of these, the divergent and inland transformational systems are mainly crustal: earthquakes occur mostly in the brittle part of the upper lithosphere. The convergent systems are both crustal and sub-crustal. As with the Aleutian Arc, crustal and sub-crustal seismicity are examined by separating it according to the local depth of the Mohorovičić discontinuity.

The Mohorovičić discontinuity and upper mantle structures in and around the Japanese territories have been investigated with active and passive seismic studies (e.g. Iwasaki et al., 1990; Iwasaki et al., 2002; Yoshii, 1994; Nakamura et al., 2003; Nakamura and Umedu, 2009; Yoshimoto et al., 2004; Hasegawa et al., 2005; Shiomi et al., 2006; Chou et al., 2009 and Uchida et al., 2010). Analogous data exists in the 1°×1° global crustal model of Laske et al., (2013), available through [http://igppweb.ucsd.edu/~gabi/rem.html](http://igppweb.ucsd.edu/~gabi/rem.html). Information about the depth to the discontinuity was
assembled from all sources above and interpolated into the $0.1^\circ \times 0.1^\circ$ model illustrated in Fig. 15.

The earthquake data span the period 2002/1/1–2016/5/31 (2002.0–2016.42) and was extracted from the catalogue of the Japan Meteorological agency (JMA), available through the National Research Institute for Earth Science and Disaster Resilience (NIED) of Japan (http://www.hinet.bosai.go.jp); the agencies contributing data for this catalogue are listed in the “Acknowledgements” section. The JMA catalogue is homogeneous and complete for $M \geq 3.0$; information concerning its properties and the seismicity of the study area is given in the supplemental material. The epicentres of crustal earthquakes are shown in Fig. 15, on top of the Mohorovičić discontinuity model used for separating them. The hypocentres of sub-crustal earthquakes are illustrated in Fig. 16.

### 3.3.1 Ryukyu Arc and Subduction Zone (RKU).

This comprises the divergent Yangtze – Okinawa plate margin (Okinawa Trough) and the convergent Okinawa – Philippine Sea plate margin which forms the Ryukyu Trench and Arc. The two boundaries run parallel to each other roughly between (123°E, 23°N) and (132°E, 33°N) in Kyushu, Japan, forming an arcuate system that bulges to the southeast. The Ryukyu Trench marks the subduction of the (oceanic) Philippine Sea Plate beneath the (continental) Okinawa and Yangtze plates, which occurs at an average rate of 52 mm/yr. The Ryukyu Island Arc is a ridge comprising two parallel chains of more than 100 islands; those of the inner arc are Quaternary volcanoes created by the subduction of the Philippine Sea Plate and those of the outer arc are non-volcanic (Iwasaki et al., 1990). The Okinawa Trough is a rift structure comprising the back-arc basin of the Ryukyu Trench – Arc – Back Arc system (Lee et al., 1980; Kobayashi, 1985; Sibuet et al., 1987). The RKU catalogue is homogeneous by construction and complete for $M \geq 3.0$ (see supplement).

Crustal seismicity (RKU-C) is highly clustered and almost entirely confined to continental crust (Fig. 15); earthquakes are apparently aligned with the Okinawa Trough (presumably tectonic) and the Ryukyu Island Arc (presumably are tectonic and volcano-tectonic). Fig. 17a shows the temporal entropic index as a function of threshold magnitude; $q_f(M_{th})$ oscillates between 1.22 and 1.44 (weak to significant correlation) and has a mean value of 1.36±0.076 (moderate correlation). The oscillatory nature of the entropic index should be an effect of the data (e.g. magnitude reporting procedure) rather than a property of the seismogenetic system. Tzanis and Tripoliti (2019) detected weak correlation at intermediate-long ranges. The $q$-relaxation interval is not stably determined; it exhibits seemingly exponential increase from $1.53 \times 10^{3}$ years (½ day) at $M_{th}=3$ to $2.9 \times 10^{2}$ years (10.5 days) at $M_{th}=4.2$ (Fig.17b). Within the range of observations, the standard recurrence interval is relatively comparable to the $q$-relaxation interval and varies entirely within the 95% simultaneous prediction bounds of the latter (Fig. 17b); the instability associated with $\Delta t_0(M_{th})$, however, does not allow inference as to what would be the case at larger magnitudes.
Sub-crustal seismicity (RKU-D) is more or less evenly distributed in the subducting slab. Between the Ryukyu Trench and the Okinawa–Yangtze boundary focal depths are concentrated directly below the trench and confined to depths shallower than 100km; just behind the Okinawa Trough, they plunge abruptly into the mantle and reach depths no greater than 300km. In Fig. 17c, \( q(M_d) \) varies slightly between 1.1 and 1.2, averaging to 1.13±0.035 and indicating a dominantly Poissonian process. Tzanis and Tripoliti (2019) demonstrated absence of long-range correlation. The \( q \)-relaxation interval increases exponentially from \( 9.256 \times 10^4 \) years (8.1 hours) at \( M_d=3.1 \) to \( 4.8 \times 10^2 \) years (17.62 days) at \( M_d=4.9 \) (Fig.16d). Up to \( M_d=5 \) and quite possibly beyond, the standard recurrence interval is almost congruent with the \( q \)-relaxation interval and lies within the narrow 95% simultaneous prediction bounds of the model fitted to the latter (Fig. 17d). It is important to point out that these results are repeatable for any subset of the RKU-D catalogue, selected by different combinations of geographical boundaries and depth ranges.

3.3.2 Izu – Bonin Segment of the Philippine Sea – Pacific plate margin (PSP).

This intra-oceanic convergent margin forms the Izu-Bonin-Mariana Arc. Herein, only the 1400km Izu-Bonin segment will be considered, northward of 21°N in the northern Mariana Plate, and up to the interface of the Philippine Sea with the Okhotsk and Pacific plates at the Boso triple junction (roughly 141.9°E, 34.2°N). Crustal thickness along the Arc averages to 20-22 km. Subduction rates vary from 46mm/year in the north to ~34mm/year in the south (Bird, 2003; Stern et al., 2004). The Wadati-Benioff zone varies along strike, from dipping gently and failing to penetrate the 660 km discontinuity in the north, to plunging vertically into the mantle but failing to penetrate the 410km transition in the south (Fig. 16; also Stern et al., 2004). The north boundary of this system is the 340-kilometre long Sagami Trough (SAT), extending from the Boso triple junction in the east to Sagami Bay, Japan in the west and comprising the surface expression of the convergent boundary along which the Izu forearc of the Philippine Sea Plate is subducted under the Honshū forearc of the Okhotsk Plate (Nakamura et al., 1984; Ogawa et al., 1989). The (crustal and sub-crustal) PSP catalogue is homogeneous by construction and complete for \( M>3.0 \) (see supplement).

During 2002.0-2016.5, crustal seismicity (PSP-C) has taken place take in direct association with the Izu-Bonin trench and along the Bonin volcano island chain (Fig. 15). Significant (\( M \geq 6.5 \)) earthquakes are generally accompanied by low-intensity, short-lived aftershock sequences with only one exception (2010/12/22, M7.4). Otherwise, activity was distributed and continuous. Notable is an apparent decrease in earthquake production rates after approximately 2011 (see supplement). Fig. 18a illustrates the variation of the temporal entropic index with magnitude. Up to \( M_d=4 \), the \( q(M_d) \) varies between 1.2 and 1.3 and then drops to 1.15. Tzanis and Tripoliti (2019) have also detected weak to moderate correlation at long ranges. All these demonstrate approximately uniform weak correlation across PSP-C. The \( q \)-relaxation interval increases from \( 6.7 \times 10^3 \) years (2.4 days) at \( M_d=3 \) to \( 3.96 \times 10^2 \) years (14.5
days) at $M_{th}=4.3$, although the variation does not appear to observe a well-defined exponential law (Fig.17b). The standard recurrence interval is comparable to the $q$-relaxation interval, at least within the range of observations (Fig. 18b).

Sub-crustal seismicity (PSP-D) is rather evenly distributed in the subducting slab (Fig. 16). Eleven significant (6.5$\leq M \leq$7) and four major ($M>7$) events have taken place during 2002.0-2016.5, all followed by low-intensity/ short-lived aftershock sequences; earthquake activity is otherwise smooth and continuous. Fig. 18b shows that the sub-crustal temporal entropic index steadily decreases from a “high” of 1.15 at $M_{th}=3$ to under 1.1 at $M_{th}=3.5$ and has an average of 1.07±0.036. Tzanis and Tripoliti (2019) demonstrated total absence of long-range correlation. All these demonstrate practically random processes. The $q$-relaxation interval increases from $8.6\times10^4$ years (8 hours) at $M_{th}=3$ to $7.3\times10^3$ years (3.4 days) at $M_{th}=4.3$ and to $3.6\times10^2$ years (13.2 days) at $M_{th}=5$; the increase appears to follow a well-defined exponential law (Fig.17d). Up to $M_{th}=4.4$, the standard recurrence interval is congruent with the $q$-relaxation interval and does not appear to deviate significantly at larger magnitudes; rather, it appears to verge on the lower 95% simultaneous prediction bound associated with $\Delta t_0$, at least up to $M_{th}=5$ and possibly beyond (Fig. 18d). As with RKU-D the results are repeatable for any subset of the sub-crustal catalogue selected by different combinations of geographical boundaries and depth ranges.

### 3.3.3 South-West Japan (SJP).

The geological domain of south-western Japan comprises the Shikoku and southern half of Honshū islands, extending between Kyushu and the Itoigawa-Shizuoka Tectonic Line (ISTL). This area is part of the Amurian continental crust. Inland crustal deformation is dominated by the WSW-ENE right-lateral Median Tectonic Line (e.g. Tsutsumi and Okada, 1996), and the Niigata–Kobe Tectonic Zone (NKTZ) which, in SW Honshū, comprises a dense network of conjugate NW-SE and NE-SW strike-slip systems; the latter is bounded to the south by the MTL and can be explained by an E-W compressional stress regime (e.g. Taira, 2001; Sagiya et al., 2000). The westward extension of the MTL connects with a zone of north-south extension in central Kyushu (e.g. Okamura et al., 1992), which continues to the eastern end of the Okinawa trough. The MTL and NKTZ are part of the broad boundary between the Philippine Sea and Amurian plates, which converge along the Nankai Trough, off the coast of south-western Japan and generate significant intermediate depth seismicity. Several major earthquakes are known to have occurred along the Nankai mega-thrust, (actual interface between the two plates), with a recurrence period of one to two hundred years (Cummins et al., 2001 and references therein). The western boundary of SJP is the ISTL in central Japan, which is part of the slowly converging Amurian and Okhotsk plates and exhibits long recurrence intervals (e.g., Okumura, 2001). The ISTL transitions northward into the fold-thrust belt that defines the Amurian–Okhotsk plate boundary at the eastern margin of the Sea of Japan (Yeats, 2012). In the period 2002-2016.5,
earthquake activity has been intense, although not particularly prolific. The SJP catalogue is homogeneous by construction and complete for $M \geq 3.0$ (see supplement).

Crustal seismicity has mostly been concentrated in the NKTZ and scarcely along the MTL (Fig. 15). Only two large ($M \geq 6.5$) earthquakes took place, with one of them (2007/3/25, M6.9) accompanied by an extended aftershock sequence; otherwise, activity was limited to a series of distributed intermediate-sized events and their low-intensity, short-lived sequences (see supplement). As seen in Fig. 19a, the temporal entropic index increases quasi-linearly from 1.15 at $M_d = 3$ (marginally weak) to approximately 1.5 at $M_d \geq 3.7$ (marginally strong). Given that seismicity occurs in continental crust and in a predominantly strike-slip tectonic setting, this is fully analogous to other continental transformational systems considered herein. However, long-range correlation in no higher than moderate (Tzanis and Tripoliti, 2019). The $q$-relaxation interval increases from $9.17 \times 10^{-3}$ years (3.3 days) at $M_d = 3$ to $4 \times 10^{-2}$ years (14.7 days) at $M_d = 3.8$ (Fig. 19b); the small magnitude bandwidth does not allow for the variation to exhibit fully developed exponential growth (Fig. 18b). The standard recurrence interval is comparable to the $q$-relaxation interval up to $M_d = 3.4$ and clearly diverges afterward, escalating at an accelerating pace; this is also observed in other transformational systems.

Sub-crustal activity has generated two major events 37-45 km beneath the Nankai Trough, presumably associated with the Tonankai segment of the Nankai mega-thrust (2004/9/5, M7.1 and M7.4). These were accompanied by a prolific albeit short-lived sequence. Continuous distributed sub-crustal activity has otherwise occurred mostly beneath SW Honshū. The temporal entropic index has a mean of 1.11 ± 0.09 and indicates a practically Poissonian system, at least within the range of observations (Fig. 19c); $q_T$ is also seen to slowly increase from nearly 1 at $M_d = 3$ (“true” randomness) to an average of 1.2 at $M_d \geq 3.6$ (weak correlation). The increase may be due to weak short-intermediate range interaction in the voluminous aftershock sequence of the 2004/9/5 events. The $q$-relaxation interval increases from $2.18 \times 10^{-2}$ years (8 days) at $M_d = 3$ to $5 \times 10^{-2}$ years (18.4 days) at $M_d = 3.7$ (Fig. 19d). Here as well, the narrow magnitude bandwidth does not allow for its variation to exhibit fully developed exponential growth (Fig. 18d). The standard recurrence interval is generally shorter than the $q$-relaxation but comparable in the sense that it varies strictly within the 95% prediction bounds of the latter; it is difficult to infer whether the two quantities diverge at larger magnitude scales.

3.3.4 Honshū Arc and Okhotsk Plate Boundaries (OKH).

The Okhotsk Plate is bounded to the south by the Sagami Trough and to the west by the “slowly” (~10 mm/yr) converging fold-thrust boundary of the Amurian and Okhotsk plates that includes ISTL (e.g. Taira, 2001; Yeats, 2012). The eastern boundary is formed by rapid (~90 mm/yr) intra-oceanic convergence of the Okhotsk and Pacific plates in which the former overrides the latter along the Japan Trench, from the Boso triple junction (142°E, 34°N) to approximately (145°E, 41°N); it is also
responsible for the creation of the broad island arc of North-eastern Honshū. The Japan Trench is succeeded by the Kuril–Kamchatka Arc and Trench that extends up to the triple junction with the Ulakhan Fault and the terminus of the Aleutian Arc and Trench, near (164°E, 56°N).

The analysis presented herein will consider earthquake activity recorded up to the north of Hokkaido, Japan; northward of the line (146.8°E, 42.5°N) – (140.2°E 46.5°N) a very large part of the JMA catalogue does not contain reliable focal depth information and the crustal/sub-crustal parts of seismicity cannot be separated. The Amurian–Okhotsk boundary, although responsible for many strong earthquakes in the Sea of Japan and Sakhalin (Arefiev et al., 2006), has not been particularly active. Conversely, activity was significant along the Honshū Arc and prolific along the Japan Trench where many strong mega-thrust earthquakes have occurred, such as the 2003/M8.3 Hokkaido event (e.g. Watanabe at al., 2006) and the 2011/M9.0 Tōhoku mega-earthquake (e.g. Ozawa et al., 2011). A total of sixty four $M \geq 6.5$ earthquakes have been observed, twenty two of which prior to the 2001.19 Tōhoku mega-event and ten of which were major ($M \geq 7$); the remaining forty two mostly occurred as part of the Tōhoku aftershock sequence which included nine major events. At any rate, the analysis will consider seismicity from 2002.0 to 2011.19 (2011/3/10), just before the Tōhoku earthquake. After that time, the catalogue was overwhelmed by the high volume and very long-lasting aftershock sequence which extended both in and below the crust and obscured any other process. The OKH earthquake catalogue is homogeneous by construction and complete for $M \geq 3.0$ (see supplement).

Crustal seismicity (OKH-C) was mainly concentrated in the Pacific-Okhotsk forearc, although it was also significant along the Honshū Arc and backarc belts (Fig. 15). It included twenty four $M \geq 6.5$ earthquakes, of which seven occurred prior to 2001.19 and five were major. However, most of those were followed by relatively low-intensity and short-lived aftershock activity and only the 2004/10/23 M6.8 and the 2008/6/14 M7.2 events contributed with extended aftershock sequences. The temporal entropic index is shown in Fig. 20a; $q_f$ is of the order 1.3-1.4 for magnitudes up to 3.5 (moderate correlation), but increases steeply so as to attain to 1.7-1.9 at $M_d \geq 3.9$ (very strong correlation). Long-range correlation is generally moderate (Tzanis and Tripoliti, 2019). The $q$-relaxation interval is practically constant between $M_{th}=3$ ($\Delta t_q=3.6\times 10^{-3}$ years/ 1.32 days) and $M_{th}=3.9$ ($\Delta t_q=4.36\times 10^{-3}$ years/ 1.6 days); it cannot be fitted with a meaningful exponential law (Fig. 20b). The $q$-relaxation and standard recurrence intervals have nothing in common, with the latter exhibiting the “expected” non-linear increase. The steeply increasing $q_f$ and of the short, flat $q$-relaxation intervals strongly resemble the EF/LSF and Walker Lake in California (Section 3.1.2); it stands to reason that they are should be attributed to analogous causes. A possible explanation is that the very strong sub-extensivity is a result of Complexity and possibly criticality developing ahead of the 2011 Tōhoku mega-earthquake.

The Wadati-Benioff zone dips very gently and the sub-crustal seismicity is rather evenly distributed as far as west, as the eastern coast of Honshū and Hokkaido (Fig. 16); thereafter, it deeps steeply to the
north-west; the seismicity is highly clustered and reaches the depth of 500km at the southern part of the subduction zone, but is more dispersed and fails to penetrate the 410km discontinuity at the central and northern parts (Fig. 16). Sub-crustal seismicity (OKH-D) included thirty nine M≥6.5 earthquakes, fifteen of which occurred prior to the Tōhoku mega-event and seven were major, including the M8.3 Hokkaido earthquake of 2003/9/26. In general, they all have had low-intensity and short-lived aftershock signatures. The temporal entropic index is shown in Fig. 20c. In stark contrast to the crustal process, it indicates very weak correlation: \( q_T(M_{th}) \) is generally lower than 1.2 and has a mean value of 1.16±0.046. By removing aftershock sequences, Tzanis and Tripoliti (2019) found that background seismogenesis is quasi-Poissonian. The \( q \)-relaxation interval increases with magnitude from \( \times 10^4 \) years (7.7 hours) at \( M_{th}=3.2 \) to \( \times 10^2 \) years (7.4 days) at \( M_{th}=4.8 \), observing a well-defined exponential law (Fig.20d). In Fig. 20d, the standard recurrence interval is very comparable and always within the 95% simultaneous prediction interval associated with \( \Delta t_q(M_{th}) \). As with RKU-D/ PSP-D, the results are repeatable for any subset of the OKH-D catalogue, selected by different combinations of geographical boundaries and depth ranges.

4. DISCUSSION

The comparative analysis of the results presented in Section 3 demonstrates systematic patterns in the behaviour of the temporal entropic index, \( q \)-relaxation interval and standard recurrence interval, which develop in more than one intertwined ways and always in association with the tectonic framework. To begin with, our results clearly indicate that practically all variants of empirical interevent time distribution can be successfully described with the \( q \)-exponential distribution which, thus, emerges as a very potent universal descriptor of their statistical properties. To this end, although we did not provide formal proof, we believe that we did offer a set of compelling evidence.

Fig. 21 provides a colour-coded summary of all \( q_T(M_{th}) \) estimators shown in Fig. 5, 7, 8, 9, 12, 13, 17, 18, 19 and 20, in which the degree of correlation (non-extensivity) is classified according to the scheme specified in the prelude of Section 3 and Fig. 21. To facilitate comparisons, only estimators corresponding to \( M_{th} \geq 3.0 \) are included. The following general observations can be made:

**Crustal seismogenetic systems in transformational plate boundaries** are generally correlated (Fig. 21a); 56.3% of \( q_T \) estimators indicate above moderate correlation, 46.6% above significant and 37.9% above strong correlation. It was found that correlation may vary with time, as in the northern segment of the San Andreas Fault where it switched from strong to weak-moderate with respect to the 1989 Loma Prieta event. Correlation also varies across network boundaries so that neighbouring fault systems may co-exist in drastically different dynamic states. Finally, very strong correlation appears to be quasi-stationary, as in the case of the Walker Lane and the Elsinore – Laguna Salada faults in California. It is worth noting that these systems are either completely landlocked in the landward side
of the primary plate boundary (Walker Lane), or are principal elements of the primary boundary bounded by significant collateral seismogenetic sub-systems (EF/LSF). Such settings may be of significance in the development of correlation, as will be discussed below.

Another characteristic of transformative systems is the increase of correlation with magnitude; only exceptions are the north segment of the San Andreas Fault (nSAF) after the Loma Prieta earthquake and the Mendocino Fault Zone. Of the total 43.7% insignificant and weak \( q_T \) estimators shown in Fig. 21a, the 27.2% derive from magnitudes smaller than 3.5 and only the 16.5% from larger. This effect is thought to indicate long-range interaction in the sense that increasingly larger events (faults) are associated with increasingly larger interaction radii and respectively larger connectivity, thus increasing the range of interaction (correlation) within the fault network (e.g. Tzanis et al. 2018; Efstathiou and Tzanis, 2018). To this effect, explicit analysis has determined the presence of operational long range interaction in transformational plate boundaries (Efstathiou et al., 2017; Tzanis et al., 2018; Efstathiou and Tzanis, 2018).

**Crustal systems in convergent (divergent) plate margins** generally exhibit low levels of correlation. As seen in Fig. 21b, approximately 56.9% of \( q_T \) estimators are weak and 17.6% moderate. Only 19.6% indicate significant to very strong correlation, while the 9.8% strong and very strong correlation is exclusively observed in the Okhotsk plate and boundaries during the period leading to the 2011.19 Tōhoku mega-earthquake.

**Sub-crustal systems** are apparently Poissonian. In Fig. 21c, 75.34% \( q_T \) estimators are below the randomness threshold and the remaining 24.66% indicates only weak correlation. Because sub-crustal systems, especially those of Wadati-Benioff zones are very large, it is important to point out (and straightforward to demonstrate) that the results are independent of the size of the system and repeatable for any subset of their respective catalogues selected by different combinations of geographical boundaries and depth ranges.

Fig. 22 summarizes the characteristics of \( q \)-relaxation intervals for the three classes of seismogenetic systems studied herein and compares them to the average correlation determined for each system. For the sake of clarity, comparisons are based on the exponential models fitted to \( \Delta t_0(M_{th}) \), except for the cases where meaningful such models could not be computed. It is apparent that the duration of \( q \)-relaxation intervals is generally reciprocal to the level of correlation and that the law by which it escalates depends on the level of correlation. Thus, in systems exhibiting strong to very strong correlation, \( \Delta t_0(M_{th}) \) is generally short and relatively “flat”: these are the SNR and ECSZ segments of the Walker lane, (Fig. 22a), the EF/LSF (Fig. 22b), and the Okhotsk plate and boundaries (Fig. 22c). Notably, the ante-Loma Prieta period of nSAF may fall into this category, (see below). It would appear that upon occurrence of an event of any size, these networks respond within a short period by an event of any magnitude, anywhere within their boundaries! In almost every other crustal or sub-
crustal system with insignificant to significant correlation and insignificant to moderate long-range interaction, $\Delta t_0(M_{th})$ increases according to more or less well defined exponential laws. This includes the Inner Continental Borderland region (ICB) in California and Southwest Japan (SJP), although these models should be considered with caution as they are based on small magnitude bandwidths. The quasi-Poissonian systems of Wadati-Benioff zones also exhibit short $q_r$-relaxation intervals, especially at small magnitude scales; this, however, can easily be understood in terms of their large size and high productivity.

Comparisons between the $q_r$-relaxation ($qri$) and standard recurrence ($sri$) intervals allow for some general observations to be made:

a. In quasi-Poissonian sub-crustal systems, $qri$ and $sri$ are very comparable. In fact they are congruent in the Wadati-Benioff zones of RKU-D, PSP-D and OKH-D. They are also very comparable in the Wadati-Benioff zone of AT-D and beneath Southwest Japan (SJP-D) in the sense that throughout the range of observations and possibly beyond, $sri$ varies in the same manner and always within the prediction bounds of $qri$.

b. In insignificantly to weakly correlated crustal systems, (post-1990 nSAF, SJF/SAF, AT-C and PSP-C), the $q_r$-relaxation and standard recurrence intervals are generally congruent within the range of observations and possibly comparable beyond.

c. In moderately to strongly correlated crustal systems, namely MFZ, ICB, QCD and SJP-C, the $qri$ and $sri$ are generally comparable at small magnitudes (up to $M_{th} \approx 4$ in MFZ, 3 in QCD and 3.5 in SJP-C), but strongly diverge afterwards with the $sri$ generally escalating faster that $qri$ and indicating longer recurrences at larger magnitudes. This category possibly includes RKU-C although inference here is hampered by instabilities in the estimation of the $qri$.

d. In strongly – very strongly correlated crustal systems $qri$ and $sri$ are vastly incomparable. In EL/LSF, the entire Walker Lane (SNR/ECSZ) and the Okhotsk Plate, $qri$ is generally short and practically flat or very slowly increasing; beyond the range of observations it appears to increase, albeit by small increments and completely out of pace with $sri$. This category may include the ante-1989 nSAF which appears to be sui generis: in Fig. 6b the $qri$ and $sri$ are very comparable up to $M_{th} \approx 3.9$ but it is hard to assess how they behave at larger magnitudes because when correlation becomes very strong ($M_{th} \geq 4$), the $qri$ “shortens: and “flattens”, while the $sri$ continues to escalate.

e. The blending of earthquake populations from fault networks with different dynamics randomizes the statistics of the mixed catalogue. When correlation it thusly reduced to moderate levels, $qri$ and $sri$ are still incongruent. Such effects are observed in California (Sections 3.1.1 and 3.1.4).

To offer some explanation of the above observations, let us begin with the “flatness” of $q_r$-relaxation intervals in systems with persistent very strong correlation These systems exhibit persistent, strong to very strong long-range interaction which may amply explain why “upon excitation anywhere within its
boundaries, the system will respond by swiftly generating earthquakes of any magnitude, anywhere within its boundaries”. Given that this is exactly how Self-Organized Critical (SOC) systems are expected to behave, it is rather straightforward to propose that simultaneous observation of very strong correlation and short/ slowly increasing $q$-relaxation intervals is compelling evidence of SOC.

The above interpretation would imply by extension that networks with moderate to strong correlation and exponentially increasing but shorter than standard $q$-relaxation intervals are Complex sub-extensive but possibly not critical. Notably, such systems generally exhibit weak to significant long-range correlation (e.g. Tzanis et al., 2018; Efstathiou and Tzanis, 2018). Complexity and Criticality do not generally go hand in hand! As comprehensively discussed by Sornette and Werner (2009), there are non-critical mechanisms able to maintain a seismogenetic system in states of non-equilibrium. Their model-driven analysis suggests that quenched heterogeneity in the stress field and production rates may be of importance. In another instance, the Coherent Noise Model (Newman, 1996) describes a non-critical top-down process based on external stresses acting simultaneously and coherently onto all faults of a network without having direct interaction with them; Celikoglu et al, (2010) showed that this can generate $q$-exponential distributions of interevent times. Accordingly, seismogenetic systems with “moderate to strong correlation but shorter than standard $q$-relaxation intervals” may be Sub-Critical or Non-Critical and there is no obvious way of discriminating between them, at least for now. Their undeniably sub-extensive nature can amply explain the incongruence of $q$-relaxation and standard recurrence intervals! In a final note, nSAF may fall into this ambiguous category: the disappearance of strong correlation after the Loma Prieta earthquake implies that it was either due to a bottom-up self-organising process that completely dissolved when the fault relaxed, or it was altogether non-critical.

Let us, now, make a compelling question: What is that makes crustal systems sub-extensive and sub-crustal systems Poissonian? And why some sub-extensive systems may achieve criticality while others may not? A basic (and by no means conclusive) interpretation of the disparity can be formulated on the simple fact that crustal systems experience the first-order discontinuity of the Earth-Atmosphere interface while sub-crustal systems do not; it can also be based on the properties of non-conservative fault networks with small-world topologies (e.g. Abe and Suzuki, 2004, 2007; Caruso et al., 2005, 2007). We were pointed to this direction by the observed increase of correlation with magnitude, by the documented existence of some degree of long-range interaction in crustal seismogenetic systems (Efstathiou et al., 2017; Tzanis et al., 2018; Efstathiou and Tzanis, 2018; Tzanis and Tripoliti, 2019), by fruitful studies of non-conservative small-world Olami-Feder-Christensen models (e.g. Caruso et al., 2005; Caruso et al., 2007), and by strong evidence of small-worldliness in the seismicity of California and Japan (Abe and Suzuki, 2004, 2007).

In small-world networks, each fault is a node that belongs to the hierarchy of some local cluster of nodes and interacts with proximal or distal nodes according to the connectivity and range of its
hierarchical level: the higher the hierarchical level, the longer the range of connectivity. Upon excitation by some stress perturbation, a node accumulates energy in the form of strain and transmits it to connected nodes or releases it at various rates, thereby operating as a delayed feedback loop and inducing heterogeneity in the rates of stress transfer and release across the network; this appears to be important in the development of criticality (Yang, 2001; Caruso et al., 2007). Finally and very significantly, crustal networks are subject to free boundary conditions at the Earth-Atmosphere interface: near-surface faults comprise boundary elements of the network. In Olami-Feder-Christensen networks free boundary conditions compel the boundary elements to interact at different (delayed) frequencies with respect to deeper buried elements; this induces partial synchronization of the boundary elements, building long-range spatial correlations and facilitating the development of a critical state (e.g. Lise and Paszucki; Caruso et al., 2005; Hergarten and Krenn, 2011). In the particularly interesting study of Hergarten and Krenn (2011), the dynamics of the network are governed by two competing mechanisms: Synchronization, that pushes the system toward criticality and de-synchronization that prevents it from becoming overcritical and generates foreshocks and aftershocks. When the system reaches the critical state, synchronized failure transfers more stress to connected nodes and causes them to fail early, de-synchronizing with the rest of the system. When, however, the lag between de-synchronized failures becomes short again, the system can re-synchronize and repeat the cycle. This mechanism generates sequences of foreshocks, main shocks and aftershocks.

Based on the above considerations, it appears plausible that the level of sub-extensivity in crustal fault networks is associated with the degree of connectivity and synchronization of top-tier elements. In transformational plate boundaries, these may be contiguous segments of large transform faults that continuously “push” against each other and function as “hubs” that facilitate longitudinal transfer of stress and interaction between distant clusters. This may be of use in understanding why sub-networks of transformational boundaries experience different levels of sub-extensivity. For instance it stands to reason that if transfer of stress (interaction) can only take place lengthwise of a bounded and locked fault zone and cannot diffuse otherwise, the degree of correlation may rise to SOC; this might explain the Walker Lane and possibly the EF/LSF zones. Conversely, the Mendocino Fault Zone may not be able to achieve criticality because it is unbounded to the south and west so that stress can diffuse across the Pacific Plate. The other transform fault networks studied herein appear to fall “in between” these extremes.

In the crustal fault networks of convergent plate boundaries, top-tier faults are low-angle mega-thrusts, whose contiguous segments do not push against each other and are thus not highly connected as large transform faults, while stress may diffuse transversely across the convergent plates. Accordingly, the system is maintained in a state of weak-moderate sub-extensivity, mainly by to partial synchronization due to free boundary conditions at the Earth-Atmosphere interface: such systems are Complex but
arguably sub-critical. One significant exception is the very strong correlation of the entire Okhotsk Plate during the period leading to the exceptional 2011.19 Tōhoku mega-earthquake. Given the relatively short period of observations, it is impossible to infer whether this is persistent (SOC), evolutionary (Self-Organizing Critical) or altogether non-critical (e.g. externally driven). Moreover, the overwhelming and long (practically on-going) aftershock sequence does not allow confident inference as to the true state of correlation after 2011.19. Finally, our “interpretative explanation” posits that the particular conditions of rupture in sub-crustal fault networks inhibit connectivity and the fixed boundary conditions due to the lithospheric overburden prohibit synchronization, thus preventing the development of Complexity and only allowing earthquakes to occur as a series of quasi-independent events (quasi-Poissonian processes). At any rate, future research will show if all this holds water!

In concluding our presentation, there is no doubt that the information introduced by \( q \)-relaxation intervals, regarding the effect of the dynamic state on the response interval of a seismogenetic system, might be very useful in improving the techniques of earthquake hazard assessment. Exactly how it can be applied, however, is not of the present and requires significant additional research. It is certain that nothing new has to be done if a system turns out to be Poissonian but in the general case of Complexity things are far more complicated.

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Figure 1: (a) Solid circles represent the observed bivariate cumulative Frequency–Magnitude–Interevent Time distribution constructed according to Eq. 4 on the basis of 6,358 events with $M_L \geq 3.5$, spanning the period 1968-2017.42 in N. California Seismic Region. The continuous surface represents the model fitted using Eq. 7. (b) Probability analysis of the residuals (see Section 2.2 for details).
Figure 2: (a) Complete analysis of 19281 events with $M \geq 3.0$, observed in the seismic region of northern California during 1968-2017.42 (see Section 2.2 for details). The panels illustrate: a) the magnitude ($q_M$) and temporal ($q_T$) entropic indices, b) the $q$-relaxation interval $\Delta t_0$, c) the energy scaling constant $\alpha$, d) the goodness of fit ($R^2$), and e) the degrees of freedom associated with the numerical solution of Eq. 7.
Figure 3: Earthquake epicentres in the N. California Seismic Region during 1968-2017.42 ($M_c \geq 3$).

MFZ: Mendocino Fracture Zone; nSAF/ sSAF: respectively north and south segments of the San Andreas SNR: Walker Lane – Sierra Nevada Range; ECSZ: Eastern California Shear Zone.
**FIGURE 4**

**Figure 4**: Earthquake epicentres in the S. California Seismic Region during 1980-2017.5 ($M_c \geq 2.5$).  
**GF**: Garlock Fault; **ECSZ**: south segment of Eastern California Shear Zone; **nSAF**: north segment, San Andreas Fault; **SNR**: Walker Lane – Sierra Nevada Range; **sSAF**: south segment, San Andreas Fault; **SJR**: San Jacinto Fault; **EF**: Elsinore Fault; **BSZ**: Brawley Seismic Zone; **LSF**: Laguna Salada fault; **IF**: Imperial Fault; **PVF**: Palos Verdes Fault; **NIF**: Newport-Inglewood fault; **RCF**: Rose Canyon fault. Offshore faults include the Coronado Bank Fault (**CBF**), San Diego Trough Fault (**SDTF**) and San Clemente Fault (**SCF**). The Transverse Ranges include the San Gabriel Fault, the San Cayetano Fault, the Oak Ridge Fault, and Santa Ynez Fault.
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**Figure 22:** Summarization of the characteristics of $q$-relaxation intervals according to geodynamic setting – comparison to the average correlation. For clarity comparisons are based on exponential models fitted to $\Delta t_0(M_n)$, except for when meaningful models could not be obtained.
Earthquake Recurrence intervals in Complex Seismogenetic Systems

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1. Review of Non Extensive Statistical Physics and its applications in Seismology

1.1. Brief exposé of NESP.

In statistical mechanics, an $N$-component dynamic system may have $W=N!/\Pi N_i!$ microscopic states, where $i$ ranges over all possible conditions (states). In classical statistical mechanics, the entropy of that system $S$ is related to the totality of these microscopic states by the Gibbs formula $S=-k\sum p_i \ln(p_i)$, where $k$ is the Boltzmann constant and $p_i$ is the probability of each microstate. If the components of the system do not interact and are statistically independent of each other, its entropy factorises into the product of $N$ identical terms, one for each component; this is the Boltzmann entropy $S_B=-Nk\sum p_i \ln(p_i)$.

A basic property of this formalism is additivity (extensivity): the entropy of the system equals the sum of the entropy of their components. However, it is now widely appreciated that a broad spectrum of non-equilibrating natural and physical systems does not conform to this requirement. Such non-additive systems, which are also commonly referred to as non-extensive after Tsallis (1988), include statistically dependent (interacting) components, in consequence of which they acquire memory (feedback) and can no longer be described with Boltzmann-Gibbs (BG) statistical physics.

Tsallis (1988, 2009), introduced a thermodynamic description appropriate for non-extensive (non-additive) systems that comprises direct generalization of Boltzmann-Gibbs statistical physics and is known as Non Extensive Statistical Physics (NESP).

Letting $x$ be some dynamic parameter of a non-extensive system, the non-equilibrium states can be described by the entropic functional:

$$S_q(p) = \frac{k}{q-1} \left[ 1 - \int_x^\infty p^q(x)dx \right] ,$$

where $p(x)dx$ is the probability of finding the value of $x$ in $[x, x+dx]$ so that $\int_x^\infty p(x)dx = 1$, and $q$ is the entropic index. When $q \to 1$, Eq. (1) converges to the Boltzmann–Gibbs functional

$$S_{BG} = -k\int p(x) \ln(p(x))dx ,$$

The Tsallis entropy is concave and fulfils the $H$-theorem but when $q \neq 1$, it is not additive. For a mixture of two statistically independent systems $A$ and $B$, it satisfies

$$S_q(A, B) = S_q(A) + S_q(B) + (1-q) S_q(A) S_q(B).$$

This is known as pseudo-additivity and is distinguished into super-additivity (super-extensivity) if $q<1$, additivity when $q\to1$ (Boltzmann-Gibbs entropy) and sub-additivity (sub-extensivity) if $q>1$. It follows that the entropic index is a measure of non-extensivity in the system.

---

1 The designation “extensive” (complete/full according to Merriam-Webster’s definition), was introduced by Tsallis (1988) to label equilibrating systems, as opposed to those that are non-equilibrating and are “incomplete”, therefore non-extensive. The terms “additive” and “non-additive” are probably more appropriate but for consistency with the international literature, Tsallis’s terminology is adopted.
An additional characteristic of NESP is that it generalizes the definition of “expectation value” in accordance with the generalization of entropy. Thus, the \( q \)-expectation value of \( x \) is
\[
\langle x \rangle_q = \int_0^\infty x \cdot p_q(x) dx,
\]
where
\[
p_q(x) = \frac{[p(x)]^q}{\int_0^\infty [p(x')]^q dx'}.
\]
is an escort distribution. The concept of escort distributions was introduced by Beck and Schloegl (1993) as a means of exploring the structures of original distributions describing fractal and multi-fractal non-linear systems: the parameter \( q \) behaves as a microscope for exploring different regions of \( p(x) \) by amplifying the more singular regions for \( q > 1 \) and the less singular for \( q < 1 \).

By maximizing the Tsallis entropy, one obtains the probability density function (PDF):
\[
\hat{p}(x) = \frac{1}{Z_q} \exp_q \left[ -\frac{\lambda}{I_q} (x - \langle x \rangle_q) \right],
\]
\[
Z_q = \int_0^\infty \exp_q \left[ -\frac{\lambda}{I_q} (x - \langle x \rangle_q) \right] dx,
\]
\[
I_q = \int_0^\infty [\hat{p}(x)]^q dx
\]
where \( \lambda \) is an appropriate Lagrange multiplier associated with the constraint on the \( q \)-expectation value. The function
\[
\exp_q(z) = \begin{cases} 
(1 + (1-q)z)^{1/(1-q)} & 1 + (1-q)z > 0, \\
0 & 1 + (1-q)z \leq 0,
\end{cases}
\]
is the \( q \)-exponential function and comprises a direct generalization of the common exponential function such that \( q \to 1 \Rightarrow \exp_q(z) \to e^z \).

Eq. (5) is a \( q \)-exponential distribution. As evident in Eq. (6), when \( q > 1 \) it is a power-law PDF with a long tail indicating sub-extensivity (sub-additivity), if \( q = 1 \) is an exponential PDF indicating extensivity (additivity), and if \( 0 < q < 1 \) is a power-law PDF with cut-off such that \( \hat{p}(>x) = 0 \) for \( x < 0 \), indicating super-extensivity (super-additivity). In the last case, the cut-off appears at
\[
x_c = \frac{x_0}{1-q}, \quad x_0 = I_q \lambda^{-1} + (1-q)\langle x \rangle_q.
\]
Using the definitions of \( x_0 \) from Eq. (7) and the \( q \)-expectation value from Eq. (4), the PDF \( \hat{p}(x) \) can be expressed as
\[
\hat{p}(x) = \frac{\exp_q \left( x/x_0 \right)}{\int_0^\infty \exp_q \left( x'/x_0 \right) dx'}
\]
In the NESP formalism, the theoretical distribution to be fitted to the observed (empirical) distribution of $x$ is not the original stationary distribution $\hat{p}(x)$ but the escort probability $\hat{p}_q(x)$. Using this, the cumulative probability function (CDF) becomes:

$$\hat{P}(> x) = \int_x^\infty \hat{p}_q(x')dx'.$$

By substituting Eq. (8) in Eq. (4) and evaluating the integral, Eq. (9) reduces to:

$$\hat{P}(> x) = \exp\left(\frac{-x}{x_0}\right) = \left[1 - (1 - q)\left(\frac{x}{x_0}\right)^{1-q}\right].$$

which also is a $q$-exponential distribution that for $q>1$, defines a CDF of the Zipf-Mandelbrot kind.

Fig. S1 illustrates the $q$-exponential CDF (Eq. 10) for different values of $q$. For $q>1$ the CDF is a power-law with tail becoming increasingly longer (fatter) with increasing $q$: this translates to increasing correlation (interaction) and longer term memory. For $q\to1$, the power law converges to the common exponential distribution and system comprises a Poissonian (uncorrelated/memoriless) process. For $0<q<1$, the CDF is a power-law exhibiting a cut-off ($\hat{P}(> x)=0$) when the argument becomes negative and is characterized by a bounded correlation radius.

**Figure S1.** Four realizations of the $q$-exponential CDF plotted in linear (left) and double-logarithmic scale (right).

**1.2. NESP Applications to the analysis of seismicity: Overview**

During the past several years, NESP has attracted increasing attention with several researchers studying the properties of the F-T and F-M distributions. This includes studies of simulated $q$-exponential distributions emerging from critical seismicity models (e.g. Caruso et al, 2007; Bakar and Tirkakli, 2009), non-critical models, (e.g. Celikoglu et al, 2010), and rock fracture experiments (e.g. Vallianatos et al., 2012a). It also includes empirical studies of interevent time statistics based on the
one-dimensional \( q \)-exponential distribution specified by Eq. 10 (e.g. Abe and Suzuki, 2005; Carbone et al., 2005; Darooneh and Dadashnia, 2008; Vallianatos et al., 2012b; Vallianatos et al., 2013; Vallianatos and Sammonds, 2013; Papadakis et al., 2013, 2015; Michas et al., 2013, 2015; Antonopoulou et al., 2014). Finally, it includes studies of frequency vs. interevent distance distributions, related to the range of interaction (Abe and Suzuki, 2003; Batak and Kantz, 2014; Shoenball et al., 2015; Efstatiiou et al., 2017; Tzanis et al., 2018; Efstatiiou and Tzanis, 2018; Efstatiiou, 2019; Tzanis and Tripoliti, 2019). A review of NESP applications over a broad spectrum of spatial scales, from tectonic plates to country rock fractures and laboratory fragmentation experiments, is given by Vallianatos et al., (2016). Finally, extensive collections of review and research papers can be found in Vallianatos and Telesca, (2012), in Chelidze et al., (2018), and in the references therein.

Non-extensive analysis of the Frequency–Magnitude (F–M) distribution has been undertaken by Sotolongo-Costa and Posadas (2004), Silva et al. (2006) and Telesca (2011, 2012). These authors proposed NESP generalizations of the Gutenberg-Richter law based on physical models that consider the interaction between two rough fault walls (asperities) and the fragments filling the space between them (fragment-asperity model); this interaction is supposed to modulate earthquake triggering. In this model, the generalized Gutenberg-Richter law is approached by considering the size distribution of fragments and asperities and the scaling of size with energy. The transition from size to energy and magnitude distributions depends on how energy scales with size and with magnitude.

Sotolongo-Costa and Posadas (2004) assumed that the energy stored in the asperities and fragments scales with their linear characteristic dimension \( r \) \((E \propto r)\) or, equivalently, with the square root of their areas \( \sigma(E \propto \sigma^{1/2}) \); they also assumed that the magnitude scales with energy as \( M \propto \log(E) \). Darooneh and Mehri (2010) expand on the same model but assume that \( E \propto \exp(\sigma^{1/\alpha}) \) and \( M \propto \ln(E) \). We suggest that these assumptions are incompatible with the empirical laws of energy–moment and moment–magnitude scaling (e.g. Scholz, 2002; Lay and Wallace, 1995). Silva et al. (2006) revisited the fragment-asperity model and expressed Eq. (10) as

\[
\hat{p}(\sigma) = \left[ 1 - \frac{1 - q}{2 - q} (\sigma - \langle \sigma \rangle_q) \right]^{1 \over 1 - q}.
\]

Assuming that the energy scales with the characteristic volume of the fragments \((E \propto r^3)\), so that \( E \propto \sigma^{3/2} \) because \( \sigma \) scales with \( r^2 \), it is easy to see that \((\sigma - \langle \sigma \rangle_q) = (E/\alpha)^{2/3}\) with \( \alpha \) being a proportionality constant between \( E \) and \( r \). This yields the energy density function

\[
\hat{p}(E) = \left( \frac{2}{3} \frac{E^{-1/3}}{\alpha^{2/3}} \right) \left[ 1 - \frac{(1 - q)}{(2 - q)} \left( \frac{E^{2/3}}{\alpha^{2/3}} \right) \right]^{1 \over 1 - q}
\]

so that \( \hat{P}(> E) = N(> E)N_0^{-1} = \int_{E}^{\infty} \hat{p}(E)dE \), where \( N(> E) \) is the number of events with energy greater
than $E$ and $N_0$ is the total number of earthquakes. If the magnitude scales with energy as $M \propto \frac{1}{q} \log(E)$, for $q > 1$,

$$
\hat{P}(>M) = \frac{N(>M)}{N_0} = \left(1 - \frac{1 - q_M}{2 - q_M} \cdot \frac{10^{2M}}{\alpha^{2/3}} \right)^{\frac{2-q_M}{1-q_M}}.
$$

Eq. (12) has been used to investigate the seismicity of different tectonic regions (e.g. Telesca 2010a, 2010b; Telesca and Chen, 2010; Scherrer et al., 2015, Esquivel and Angulo, 2015).

Finally, assuming $E \propto r^3$ but that the magnitude scales with energy as $M \propto \frac{1}{q} \log(E)$, Telesca (2011, 2012) introduced a modified version of Eq. (12):

$$
\hat{P}(>M) = \frac{N(>M)}{N_0} = \left(1 - \frac{1 - q_M}{2 - q_M} \cdot \frac{10^M}{\alpha^{2/3}} \right)^{\frac{2-q_M}{1-q_M}}.
$$

We suggest that this model, by postulating that the energy released in the form of seismic waves scales with the effective area of the fault (fragments and asperities), is consistent with the empirical laws of energy–moment and moment–magnitude scaling and is also compatible with the well-studied rate- and-state friction laws of rock failure.
3. **On the construction of bivariate Frequency- Magnitude – Interevent Time (F-M-T) distributions.**

The bivariate F-M-T distribution is constructed as follows: A threshold (cut-off) magnitude $M_{th}$ is set and a bivariate frequency table (histogram) representing the empirical incremental distribution is first compiled. The cumulative distribution is obtained by backward bivariate summation, according to the scheme

$$N_{m\tau} = \sum_{j-D_T}^{r} \sum_{i-D_M}^{m} \{H_{ij} \Leftrightarrow H_{ij} \neq 0\}, \quad \tau = 1, \ldots, D_T, \quad m = 1, \ldots, D_M$$  (14)

where $H$ is the incremental distribution, $D_M$ is the dimension of $H$ along the magnitude axis and $D_T$ is the dimension of $H$ along the $\Delta t$ axis.

An empirical such is presented in Fig. S2: It is based on a subset of 3,653 events extracted from the NCSN earthquake catalogue published by the North California Earthquake Data Center, over the period 1975-2012, excluding the Mendocino Fracture Zone and using a threshold magnitude $M_{th} = 3.4$ (for details see Section 3). The distribution is shown in linear (Fig. S2a) and logarithmic (Fig. S2b) frequency scales and comprises a well-defined surface in which the end-member ($M \geq M_{th}, \Delta t=0$) is the one-dimensional empirical Gutenberg – Richter law and the end-member ($M = M_{th}, \Delta t$) is the one-dimensional frequency – interevent time (F-T) distribution.

Outliers can be observed at the moderate- large magnitude scales and longer interevent times. They usually arise from minor flaws of the catalogue, such as omitted (sequences of) events, glitches in magnitude reporting etc. In some cases, however, they may comprise true exceptions to the continuum of the regional seismogenetic process; for example, they may represent rare, externally triggered events.

The existence of outliers has (in part) compelled the introduction of a significant constraint in the construction of the F-M-T distribution. According to Eq. (14), the summation in limited to the populated (non-zero) bins of the incremental distribution. Regardless of their origin, outliers have to be included in the summation. However, as illustrated in Fig. S2c and S2d, inclusion of unpopulated bins would have led to the generation of a stepwise function in which the unpopulated regions (of unknown probability densities) between the outliers and the (normal) populated bins would appear as patches of equal earthquake frequency (patches of uniform probability). In this case, the high probability zones of the empirical bivariate distribution would comply with well specified laws, but the lower probability zones would, for some unknown and unjustifiable reason, include uniform swathes. In one-dimensional distributions this effect may not influence parameter estimation by a significant factor and is usually neglected. In multivariate distributions however, in addition to the obvious absurdity, it would be numerically detrimental.
Figure S2. (a) Bivariate cumulative F-M-T distribution constructed according to Eq. (14) on the basis of 3,653 events with $M_l \geq 3.4$ extracted from the NCSR earthquake catalogue (Section 4 for details). (b) As per (a) but in logarithmic frequency scale. (c) As per (a) but including unpopulated bins in the summation, i.e. using the scheme $N_{mT} = \sum_{j=D_l}^{j-D_h} \sum_{i=D_{ij}}^{i-D_{ij}} H_{ij}$ instead of Eq. (14). (d) As per (c) but in logarithmic frequency scale.
4. Earthquake catalogues of Northern and Southern California.

The seismicity of Northern California is monitored by North California Seismic Network (NCSN) and the respective regional catalogue is published by North California Earthquake Data Center (http://www.ncedc.org). The seismicity of Southern California is monitored by the Southern California Seismic Network (SCSN) and the regional catalogue is published by the South California Earthquake Data Centre (http://www.data.scec.org).

In both catalogues, most earthquakes are reported in the $M_L$ and $M_w$ magnitude scales, while there is a considerable number of events in the duration ($M_d$) and amplitude ($M_x$) scales. The latter two have been exhaustively calibrated against the $M_L$ scale: Eaton (1992) has shown that they are within 5% of the $M_L$ scale for magnitudes in the range 0.5 to 5.5 and that they are virtually independent of the distance from the epicentre to at least 800 km. In consequence, $M_d$ and $M_x$ are practically equivalent to $M_L$. For the purpose of the present analysis, $M_w$ magnitudes were also converted to $M_L$ using the empirical formula of Uhrhammer et al., (1996): $M_w = M_L \times (0.997 \pm 0.020) - (0.050 \pm 0.131)$. Thus, both the catalogues were reduced to the local magnitude scale and are homogenous.

The basic characteristics and properties of the NCSN catalogue are summarized in Fig. S3 (adapted from Efstathiou et al., 2017). It is evident that as of 1968, the catalogue is complete for $M_L \geq 3.0$ (Fig. S3-b and S3-c).

**Figure S3.** a) The cumulative earthquake count of the NCSN catalogue for the period 1968-2012. b) Variation of the magnitude of completeness ($M_c$) with time in the NCSN catalogue with 95% confidence limits. c) Incremental (down triangles) and cumulative (open squares) Frequency-Magnitude distribution of the NCSN catalogue. Fig. S3-c was prepared with the ZMAP software (Wiemer, 2001; Woessner and Wiemer, 2005). The figure was adapted from Figs. 5 and 7 of Efstathiou et al., (2017).
Up to the early 1970’s the SCSN network was sparse and consisted of about 49 stations. As a result, the epicentral distribution maps compiled for the broader area of South California projected an image of somewhat diffuse seismicity (Hutton et al., 2010). During the 1980’s and early 1990’s the network improved qualitatively and quantitatively: more than 100 additional stations were installed, while past events were relocated and magnitudes re-determined. With denser network and modern data processing, it became clear that earthquakes activity was mainly clustered along and around the large active faults of the Late Quaternary (Fig. 4 of main article). As seen in Fig. S5-a, the sustainable magnitude of completeness (\(M_c\)) was approximately 3.0 during the early to mid-1970s and decreased after 1975, attaining a sustainable level of approximately 2.5 as of the early 1980’s. The spiky fluctuations observed in Fig. S5-a correspond to time-local instabilities in the estimation procedure caused by major aftershock sequences and should not be viewed as temporary changes in the detection threshold. Fig. S5 also shows that for the period 1968-2017 the catalogue is complete for magnitudes \(M_L \geq 3\) and comprises 10793 events (Fig. S5-b) while for the period 1980-2017 it is complete for \(M_L \geq 2.5\) and comprises 30117 events (Fig. S5-c). As demonstrated by Efstathiou and Tzanis (2018), for \(M_c \geq 3.0\), the NESP analysis of the SCSN catalogues yields almost identical results for both periods 1968-2017 and 1980-2017. Accordingly, and in order to study the dynamic expression of the small(er) magnitude scales, the SCSN catalogue is considered only for the period 1980-2017 in which \(M_c \geq 2.5\).

Figure S4: Cumulative event count of the SCSN catalogue for the period 1968 – 2017 (magnitude of completeness \(M_c \geq 3.0\)).
Figure S5. Attributes of SCSN earthquake catalogue: a) Magnitude of completeness ($M_c$) as a function of time for the period 1968-2017. b) Frequency–Magnitude distribution for the period 1968-2017; blue triangles denote the incremental distribution and red squares the cumulative. d) As per Fig. S5-b but for 1980-2017. Fig. S5-a was prepared with the ZMAP software (Wiemer, 2001; Woessner and Wiemer, 2005). The figure was adapted from Fig. 3 of Efthathiou and Tzanis, (2018)
5. **The earthquake catalogue of Continental Alaska and the Alaskan – Aleutian Arc.**

The earthquake data utilized for the source areas of Continental Alaska and the Aleutian Arc was extracted from the regional earthquake database of the Alaska Earthquake Centre (AEC, [http://www.aeic.alaska.edu/html_docs/db2catalog.html](http://www.aeic.alaska.edu/html_docs/db2catalog.html)) and comprises a total of 48,995 events recorded in the area 50°N to 70°N and -196°E to -126°E over the period 1968–2015. Most events are reported in the local magnitude ($M_L$) scale but a significant number is reported only in the surface ($M_S$) and body wave ($m_b$) scales. Fortunately, 1715 events are jointly reported in the $M_L$, $M_S$ and $m_b$ scales. It is, therefore, straightforward to generate calibration tables by which to convert $M_S$ and $m_b$ to $M_L$ (also see Tzanis et al., 2018). This exercise was carried out by robust re-weighted linear regression with a re-descending bisquare influence function. The $M_L$–$M_S$ relationship is shown in Fig. S6-a and the regression (calibration) formula is:

$$M_L = (1.074 \pm 0.018) \times m_b - (0.4099 \pm 0.0942), \quad 4 \leq m_b \leq 7.2.$$  

The $M_L$ – $m_b$ relationship is shown in Fig. S6-b and the corresponding regression formula is

$$M_L = (0.712 \pm 0.013) \times M_S + (1.651 \pm 0.066), \quad 3.5 \leq M_S \leq 7.5.$$  

The relationships between $M_L$ – $m_b$ and $M_L$ – $M_S$ are obviously linear which suggests that the regression coefficients are rather precisely determined. Thus, acknowledging the problems associated with the saturation of the local and body wave scales at the large magnitude end of the spectrum, and assuming that both relationships can be linearly extrapolated to smaller magnitude scales, it is possible to construct a homogeneous version of the AEC catalogue with all events reported in the local magnitude scale.

The AEC catalogue presents a conundrum: Fig. S6-c clearly shows that F–M distribution of seismicity recorded along the Aleutian Arc (Fig. 11 of the main paper) is bimodal, a feature not present in the seismicity of Continental Alaska (Fig. S6-d). For magnitudes between 3 and 4.3, the $b$ value is 0.47 and for $M_L \geq 4.4$ increases abruptly to 1.1. The origin of the bimodal distribution might be natural, (different physical mechanisms operating at small and intermediate-large magnitude scales), although $b$ values as low as 0.47 over so broad an area are not easy to explain. On the other hand, as can be seen in the incremental distribution (downward pointing triangles in Fig. S6-c) the escalation of frequency is faltering between $M_L = 3.9$ and 4.3 (events missing) and there is a rather suspicious leap of about 5500 events between $M_L = 3.0$ and 3.1 (event surplus), which is also difficult to explain naturally. Given also is the relatively sparse and nearly one-dimensional monitoring network along the Aleutians (see [https://earthquake.alaska.edu/network](https://earthquake.alaska.edu/network)) together with difficulties associated with the detection of small earthquakes. Finally, it is not difficult to verify that bimodality is definitely more pronounced in the western (oceanic) part of the convergence (west of Unimak Pass), where the network is sparsest. It is therefore not certain that the differences between the small and intermediate-large magnitude scales are natural. In consequence and as far as the Aleutian Arc and Trench is concerned, we only consider
earthquakes with $M_L \geq 4.4$, for which the F-M distribution, albeit imperfect, does not raise concerns as to its constitution. It is apparent that in that area, the homogenized AEC catalogue is complete for $M_L \geq 4.4$ (Fig. S6-c). Conversely, in Continental Alaska we shall consider all earthquakes with magnitudes $M_L \geq 3$, for which the catalogue appears to be complete (Fig. S6-d).

Figure S6. Relationship between (a) the local and surface wave magnitude scales and (b) between the local and body wave magnitude scales, for the area of Alaska and the Aleutian Arc. Analysis based on 1715 events jointly reported in the $M_L$, $M_S$ and $m_b$ magnitude scales, in the catalogue of the Alaska Earthquake Centre. The regression lines were fitted with robust linear least squares; broken lines mark the 95% prediction bounds. (c) The frequency – magnitude distribution of seismicity along the Aleutian Arc and Trench. (d) As per Fig. S6-c for continental Alaska. Down pointing triangles represent the incremental distribution; squares represent the cumulative distribution; broken lines are 99% prediction bounds. The figure was modified from Fig. 6 of Tzanis et al., (2018).

As evident in the foregoing, seismogenesis in Alaska and the Aleutian Arc develops in a rather complex tectonic background, extends over a very large area and range of depths and exhibits regional variation. However, the most significant earthquake source areas are located along the broader boundary between the North American and Pacific plates in which it is possible to distinguish three classes of earthquake activity:

d) Crustal earthquakes in Continental Alaska primarily associated with the eastern transformational plate margin; these include the Queen Charlotte – Fairweather and Denali faults and the
transitional zone spanned by the Yakutat Terrane and the Wrangelian Composite Terrane. This boundary will henceforth be referred to as Queen Charlotte – Denali Zone, or QCD; the cumulative earthquake count of the relevant catalogue is illustrated in Fig. S7-a.

e) Crustal earthquakes along the Alaskan – Aleutian Arc primarily associated with the convergent plate margin; these are crudely distinguished by the depth of the Mohorovičić discontinuity which is approx. 40 km beneath the Yakutat Terrane (Christeson et al., 2013) and 38.5 km along the Aleutian Arc (Janiszewski et al., 2013). This source area is referred to as AT-C (Aleutian Trench – crustal) and the relevant cumulative earthquake count is illustrated in Fig. S7-b.

f) Sub-crustal earthquakes along the Alaskan – Aleutian Wadati-Benioff zone, also distinguished by the depth of the Mohorovičić discontinuity. This source area is referred to as AT-D (Aleutian Trench - Deep) and the relevant cumulative earthquake count is also illustrated in Fig. S7-b. This provides an opportunity to study and compare the dynamics of earthquake populations generated in different seismotectonic settings, environmental conditions (crust vs. subducting slab) and boundary conditions (free in the crust vs. fixed in the subducting slab), and to inquire if these differences affect the dynamics of the fault network.

Figure S7. Cumulative event counts of the earthquake catalogues of: (a) the Queen Charlotte – Fairweather and Denali zone of transform faults (transformational North American – Pacific plate boundary); (b) the crustal (AT-C) and sub-crustal (AT-D) subsets of the Aleutian Arc and Trench (convergent North American – Pacific plate boundary).
6. The North-West Segment of the Circum-Pacific Belt and the JMA earthquake catalogue

The study area extends from approx. 22°N to 46°N (Hokkaido Island, Japan) and from 122°E (east of Taiwan) to 146°E in the Pacific Ocean (see Fig. 14 of main article). The earthquake data is produced by the Japan Meteorological Agency (JMA), in cooperation with the Ministry of Education, Culture, Sports, Science and Technology. The catalogue is based on data provided by the National Research Institute for Earth Science and Disaster Resilience (NIED, http://www.hinet.bosai.go.jp), the Japan Meteorological Agency, Hokkaido University, Hirosaki University, Tōhoku University, the University of Tokyo, Nagoya University, Kyoto University, Kochi University, Kyushu University, Kagoshima University, the National Institute of Advanced Industrial Science and Technology, the Geographical Survey Institute, Tokyo Metropolis, Shizuoka Prefecture, Hot Springs Research Institute of Kanagawa Prefecture, Yokohama City, and Japan Agency for Marine-Earth Science and Technology. The catalogue was made available by NIED, spans the period 01/01/2002–31/05/2016 and is homogeneous by construction and complete for $M \geq 3.0$ (Fig. S9).

The study area includes several major convergent and one divergent plate boundaries, transformational plate boundaries and inland seismogenic domains. Of these, the divergent, transformational and inland systems are mainly crustal: earthquakes occur mostly in the schizosphere (i.e. rigid, brittle part of the upper lithosphere). The convergent systems are both crustal and sub-crustal. In order to see if there are differences in the expression of crustal and sub-crustal fault networks, crustal and sub-crustal seismicity was examined separately by distinguishing it according to the local depth of the Mohorovičić discontinuity.

The depth to the Mohorovičić discontinuity was modelled by assimilating information from many different sources including active and passive seismic studies, (Iwasaki et al., 1990; Iwasaki et al., 2002; Yoshii, 1994; Nakamura et al., 2003; Nakamura and Umedu, 2009; Yoshimoto et al., 2004; Hasegawa et al., 2005; Shiomi et al., 2006; Chou et al., 2009 and Uchida et al., 2010), and the 1°×1° global crustal model of Laske et al., (2013). This information was assembled and interpolated into the 0.1°×0.1° grid illustrated in Fig. 15 of the main article.

The epicentres of crustal earthquakes are shown in Fig. 15 of the main article, on top of the Mohorovičić discontinuity model. The hypocentres of sub-crustal earthquakes are illustrated in Fig. 16 of the main article. Only earthquakes with magnitudes $M \geq 3$ are considered, as a compromise that ensures completeness, sufficiently large earthquake populations to guarantee statistical rigour and exclusion of very distant – very small events that are ostensibly uncorrelated and may randomize the catalogue. The seismogenetic systems and fault networks examined herein are as follows:
Figure S8: The Frequency – Magnitude distribution of (crustal and sub-crustal) seismicity in each of the four major earthquake source areas considered in the present work.

Figure S9: The cumulative event counts of the crustal (top) and sub-crustal (bottom) earthquake catalogues in each of the four major earthquake source areas considered in the present work.
6.1 Ryukyu Arc and Subduction Zone (RKU).

This source area comprises the divergent Yangtze – Okinawa plate margin (Okinawa Trough) and the convergent Okinawa – Philippine Sea plate margin, which forms the Ryukyu Trench and Arc. These run parallel to each other roughly between (123°E, 23°N) and (132°E, 33°N) in Kyushu Island, Japan, forming an arcuate system bulging to the southeast. The Ryukyu Trench marks the subduction of the (oceanic) Philippine Sea Plate beneath the (continental) Okinawa and Yangtze plates, which occurs at an average rate of 52 mm/yr. The Ryukyu Island Arc is a ridge comprising two parallel chains of more than 100 islands, with those of the inner arc being Quaternary island arc volcanoes created by the subduction of the Philippine Sea Plate, and those of the outer arc being non-volcanic (Iwasaki et al., 1990). The Okinawa Trough is a rift structure comprising the back-arc basin of the Ryukyu Trench – Arc – Back Arc system (Lee et al., 1980; Kobayashi, 1985; Sibuet et al., 1987).

The RKU catalogue is homogeneous by construction and complete for $M \geq 3.0$ (Fig. S8-a). As evident in Fig. 15 of the main article, shallow (crustal) earthquakes are highly clustered and almost entirely occur in areas with crustal thickness greater than 20km (continental crust); they are apparently aligned with the Okinawa Trough where they presumably are tectonic, as well as with the Ryukyu Island Arc where they presumably are tectonic and volcano-tectonic. During the period 2002-2016.5, crustal seismicity was dominated by a series of large to major earthquakes (2002/3/27, M7; 2005/3/20, M7; 2007/4/20, M6.7; 2015/11/14, M7.1; 2016/4/16, M7.3), and with the exception of the last major event, their low-intensity, short-lived aftershock sequences (Fig. S9-a).

As seen in Fig. 16 of the main article, sub-crustal seismicity is more or less evenly distributed over the subducting slab. Between the Ryukyu Trench and the Okinawa–Yangtze plate boundary, focal depths are mainly concentrated directly below the Trench and are confined to depths shallower than 100km; they plunge abruptly into the mantle just behind the Okinawa–Yangtze boundary and beneath the Okinawa Trough, to depths no greater than 300km. Sub-crustal seismicity has been characterized by numerous (15) large to major events, the largest of which (M7.2) occurred on 2010/2/27. The cumulative earthquake count curve indicates that all of these were associated with very low-intensity and very short-lived aftershock sequences (Fig. S9-b).

6.2 Izu – Bonin Segment of the Philippine Sea – Pacific plate margin (PSP).

The Philippine Sea – Pacific intra-oceanic convergent plate margin forms the Izu-Bonin-Mariana (or Izu-Ogasawara-Mariana) Arc and Trench. Herein, only the 1400km long Izu-Bonin segment will be considered, northward of latitude 21°N in the northern Mariana Plate and up to the interface of the Philippine Sea, Okhotsk and Pacific plates at the Boso Trench-Trench-Trench junction (roughly 141.9°E, 34.2°N). The Izu segment is punctuated by inter–arc rifts (Klaus et al., 1992; Taylor et al., 1991) and farther south also contains several submarine felsic calderas (Yuasa and Nohara, 1992). The
Bonin segment contains mostly submarine, and a few island volcanoes. Crustal thickness along the Izu-Bonin Arc averages to 20-22 km with a felsic middle crust. Subduction rates vary from 46 mm/yr in the north to ~34 mm/yr in the south (Bird, 2003; Stern et al., 2004). The Wadati-Benioff zone varies along strike, from dipping gently and failing to penetrate the 660 km discontinuity in the north, to plunging vertically into the mantle but failing to penetrate the 410 km transition in the south (Fig. 16 of main article; also Stern et al., 2004). The north boundary of this system is the 340-kilometre long Sagami Trough (SAT), extending from the Boso triple junction in the east, to Sagami Bay, Japan in the west and comprising the surface expression of the convergent boundary along which the Izu forearc of the Philippine Sea Plate is subducted under the Honshu forearc of the Okhotsk Plate (Nakamura et al., 1984; Ogawa et al., 1989). The SAT is known for major ($M \geq 8$) mega-thrust historical and instrumental era earthquakes, as for instance the M7.9 Great Kanto Earthquake of 1923/9/1 (e.g. Kobayashi and Koketsu, 2005).

The PSP catalogue is homogeneous by construction and complete for $M > 3.0$ (Fig. S8-b). During 2002-2016.5, crustal seismicity has taken place in direct association with the Izu-Bonin trench and along the Bonin Volcano Island chain where it is presumably volcano-tectonic (Fig. 15 of main article). Only three significant ($M \geq 6.5$) earthquakes have taken place, two of which (2005/2/10, M6.5 and 2006/10/24, M6.8) have low intensity, short-lived aftershock sequences and one (2010/12/22, M7.4) has a prolific sequence (Fig. S9-a). Otherwise, earthquake activity appears to be distributed and continuous. Notable also is an apparent decrease in production rates after approximately 2011 (post Tōhoku effect?).

As evident in Fig. 16 of main article, sub-crustal seismicity is rather evenly distributed in the subducting slab. Eleven significant ($6.5 \leq M \leq 7$) and four major ($M > 7$) events have taken place during 2002-2016.5, the most noteworthy of which occurred on 2007/9/28 (M7.6) and on 2015/5/30 (M8.1). The cumulative earthquake count curve shows that they all had very low-intensity and short-lived aftershock sequences, with earthquake activity appearing to be otherwise continuous. These characteristics are similar to RKU (Fig. S9-b).

6.3 South-West Japan (SJP).

The geological domain of south-western Japan consists of the Shikoku and southern half of Honshu islands and extends between Kyushu and the Itoigawa-Shizuoka Tectonic Line (ISTL). This area is part of the Amurian continental crust. Inland crustal seismicity is dominated by the WSW-ENE right-lateral Median Tectonic Line (e.g. Tsutsumi and Okada, 1996) and the Niigata–Kobe Tectonic Zone (NKTZ) which in SW Honshu comprises a dense network of conjugate NW-SE and NE-SW strike-slip systems; the latter is bounded to the south by the MTL and can be explained by an E-W compressional stress regime (e.g. Taira, 2001; Sagiya et al., 2000). The westward extension of the MTL connects with a zone of north-south extension in central Kyushu (e.g. Okamura et al., 1992), which continues to
the eastern end of the Okinawa trough. The MTL and NKTZ are part of the broad active boundary between the Philippine Sea and Amurian plates. These two plates converge along the Nankai Trough, off the coast of south-western Japan and generate significant intermediate depth seismicity. Convergence directions and rates are N–NW and 4.5cm/yr respectively (Seno et al., 1993). The tectonic coupling between the overriding and subducted plates has been evaluated to nearly 100% over the landward slope of the entire Nankai Trough (complete fault locking, Mazzotti et al., 2000). Several major earthquakes are known to have occurred along the Nankai mega-thrust, (interface between the two plates), with a recurrence period of one to two hundred years (e.g. Cummins et al., 2001 and references therein). The western boundary of SJP is the Itoigawa-Shizuoka Tectonic Line (ISTL) in central Japan, which is part of the convergent front of the Amurian and Okhotsk plates. The ISTL is about 250 km long, several km wide and exhibits long-term slip rates of around 8-10 mm/yr (e.g. Okumura et al., 1994; 1998) and long earthquake recurrence intervals (e.g., Okumura, 2001). Sagiya et al., (2002) determined E-W horizontal shortening of the order of 0.3μstrain/yr and Ikeda et al., (2004) indicated that ISTL comprises a low-angle, low-strength fault plane. Northward of Honshu, at the eastern margin of the Sea of Japan, the ISTL transitions into the fold-thrust belt of the Amurian–Okhotsk plate boundary (Yeats, 2012).

The SJP catalogue is homogeneous by construction and appears to be complete for $M \geq 3.0$, although the cumulative earthquake frequency exhibits a rather peculiar oscillation which will not be interpreted herein (Fig. S8-c). In the period 2002.0-2016.5, earthquake activity has been intense but not particularly prolific. Crustal seismicity has mostly been concentrated in the NKTZ and to a considerably lesser degree along the MTL (Fig. 15 of main article). Only two large ($M \geq 6.5$) earthquakes have occurred, with the most significant of them (2007/3/25, M6.9) accompanied by an extended aftershock sequence (Fig. S9-a); earthquake activity has otherwise been limited to a series of distributed intermediate-sized events accompanied by low-intensity, short-lived aftershock sequences. The sub-crustal activity has generated two major events at depths 37-45km beneath the Nankai Trough, presumably associated with the Tonankai segment of the Nankai mega-thrust (2004/9/5, M7.1 and M7.4). These were accompanied by a prolific albeit short-lived aftershock sequence that included two M6.5 events (on 7 and 2004/9/8). Sub-crustal activity has otherwise occurred mostly beneath south-western Honshu and has been continuous and distributed.

6.4 Honshu Arc and Okhotsk Plate Boundaries (OKH).

The Okhotsk Plate and plate boundary systems, are bounded to the south by the intra-oceanic convergent Philippine Sea – Okhotsk plate boundary (Sagami Trough, see above), to the west by the “slowly” (~10 mm/yr) converging fold-thrust boundary of the Amurian and Okhotsk plates which includes the ISTL, (e.g. Taira, 2001; Yeats, 2012), and to the east by the rapidly (~90 mm/yr) converging Pacific and Okhotsk plate boundary, in which the Pacific Plate is being subducted
underneath the Okhotsk Plate and forms the Japan Trench. The Japan Trench extends from the Boso triple junction near (142°E, 34°N) to approximately (145°E, 41°N) and is responsible for creating the broad island arc of North-eastern Honshu. The Japan Trench is succeeded by the Kuril–Kamchatka Arc and Trench system that extends up to the triple junction with the Ulakhan Fault and the terminus of the Aleutian Arc and Trench, near (164°E, 56°N).

The Okhotsk Plate and boundaries have experienced particularly intense seismic activity during 2002-2016.5. The analysis presented herein will consider seismicity up to the north of Hokkaido, Japan, because northward of the line (146.8°E, 42.5°N) – (140.2°E 46.5°N) a very large part of the JMA catalogue does not contain reliable focal depth information and is not possible to separate the crustal and sub-crustal parts of seismicity and calculate interevent distances. The Amurian – Okhotsk boundary, although responsible for many strong earthquakes in the Sea of Japan and Sakhalin (e.g. Arefiev et al., 2006), has not been particularly energetic. Instead, activity has been significant along the Honshu arc and prolific along the Pacific–Okhotsk subduction in which many strong mega-thrust earthquakes have taken place such as the 2003 M8.3, Hokkaido mega-thrust event (e.g. Watanabe at al., 2006) and the M9.0, 2011 Tōhoku mega-thrust mega-earthquake (e.g. Ozawa et al., 2011).

As before, the combined crustal and sub-crustal OKH catalogue is homogeneous by construction and complete for $M \geq 3.0$, although the shape of the cumulative frequency curve at the $M \geq 7$ magnitude range is unexpectedly oscillatory, a feature whose significance shall not be examined (Fig. S8-d). A remarkable sixty four $M \geq 6.5$ earthquakes have occurred within the Okhotsk plate system, twenty two of which have occurred prior to the 2001.19 Tōhoku mega-event and ten of which have been major ($M \geq 7$); the remaining forty two have mostly occurred as part of the Tōhoku aftershock sequence that included a total of nine major events. Crustal seismicity concentrated mainly along the Pacific-Okhotsk forearc and was dominated by the Tōhoku aftershock sequence although activity along the Honshu Arc and backarc belts has also been significant. This included twenty four $M \geq 6.5$ earthquakes, of which seven occurred prior to the Tōhoku mega-event and five were major (2003/9/26, M7.1; 2004/10/23, M6.8; 2008/6/14, M7.2; 2001/3/9, M7.3; 2001/3/10, M6.8).

With regard to sub-crustal seismicity, the Wadati-Benioff zone dips very gently and is rather evenly distributed as far as west, as the eastern coast of Honshu and Hokkaido; thereafter, it deeps steeply to the north-west and is rather unevenly distributed, being highly clustered and reaching the depth of 500km in the southern part of the zone, but more dispersed and failing to penetrate the 410km discontinuity in the central and northern parts. Sub-crustal activity included thirty nine $M \geq 6.5$ earthquakes, fifteen of which occurred prior to the Tōhoku mega-event and seven have been major, including the 2003/9/26, M8.3 Hokkaido event.

At any rate, the analysis presented herein will consider seismicity from 2002.0 to 2011.19 (2011/3/10), just before the Tōhoku earthquake. After that time, the catalogue was overwhelmed by the high
volume and very long-lasting aftershock sequence of the Tōhoku event, which occurred both in and below the crust and completely obscured any other seismogenetic process. Fig. S9-a and S9-b respectively show the cumulative earthquake counts of crustal and sub-crustal seismicity from 2002.0 to 2011.19, just before the Tōhoku event. As evident in Fig. S9-a, crustal seismicity was dominated by the extended aftershock sequences of the 2004/10/23 M6.8 and the 2008/6/14 M7.2 events, the rest contributing with relatively low-intensity and short-lived aftershock activity. The sub-crustal activity exhibits quasi-linear increase decorated with time-local jerks due to low-intensity and short-lived aftershock sequences of major events (Fig. S9-b). Interestingly enough, the 2003/9/26 M8.3 Hokkaido mega-thrust event had a relatively minor aftershock signature, with other significant contributions coming from the M7.1 2003/5/26, and M7.2 2005/11/15 events; all these have occurred at depths shallower than 45km.
TABLE 1: Summary of the earthquake catalogues used in the present analysis.

<table>
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<th>Source Area</th>
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<td>3.3</td>
<td>1.17</td>
<td>4.6</td>
<td>1.18</td>
<td>5.5</td>
<td>1.32</td>
<td>4.9</td>
<td>1.26</td>
<td>6.5</td>
<td>1.34</td>
<td>7.5</td>
<td>1.47</td>
<td>7.5</td>
<td>1.51</td>
<td>9.5</td>
<td>1.48</td>
<td>14.7</td>
<td></td>
<td></td>
<td></td>
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</tbody>
</table>

$q_T < 1.15$ | Insignificant | $1.15 \leq q_T < 1.30$ | Weak | $1.30 \leq q_T < 1.4$ | Moderate | $1.4 \leq q_T < 1.5$ | Significant | $1.5 \leq q_T < 1.6$ | Strong | $1.6 \leq q_T$ | Very Strong
**TABLE 4:** Temporal Entropic Indices and \( q \)-relaxation Intervals (in days) from Crustal and Sub-Crustal systems of Convergent Plate Margins.

<table>
<thead>
<tr>
<th>CRUSTAL SYSTEMS IN CONVERGENT PLATE BOUNDARIES</th>
<th>( q_T )</th>
<th>( \Delta t_0 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>AT-C</td>
<td>1.10</td>
<td>1.10</td>
</tr>
<tr>
<td>RKU-C</td>
<td>1.33 1.36 1.36 1.44 1.39 1.33 1.42</td>
<td>1.22 1.24 1.35 1.39 1.49</td>
</tr>
<tr>
<td>PSP-C</td>
<td>1.22 1.21 1.26 1.24 1.26 1.28 1.24 1.28 1.23 1.21 1.14 1.16</td>
<td>2.4 2.7 3.0 3.1 3.2 3.7 3.5 4.3 5.3 6.3 7.8 10.5 12.8 14.4</td>
</tr>
<tr>
<td>OKH-C</td>
<td>1.26 1.19 1.21 1.27 1.35 1.45 1.46 1.54 1.69 1.72 1.82 1.89</td>
<td>1.3 1.9 2.0 2.4 2.4 2.3 2.7 2.6 1.7 1.5 1.6 1.4</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>SUB-CRUSTAL SYSTEMS</th>
<th>( q_T )</th>
<th>( \Delta t_0 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>AT-D</td>
<td>1.07 1.06 1.01 1.00 1.11 1.00 1.00 1.15 1.15</td>
<td>11.5 13.7 22.0 23.9 27.6 35.7 41.5 43.5 44.2</td>
</tr>
<tr>
<td>RKU-D</td>
<td>1.19 1.14 1.11 1.09 1.08 1.12 1.13 1.12 1.10 1.13 1.17 1.12 1.12 1.15 1.10 1.18</td>
<td>0.3 0.5 0.6 0.7 0.9 1.1 1.2 1.7 2.1 2.7 2.7 3.1 4.1 5.1 5.8 7.9 10.4 9.7</td>
</tr>
<tr>
<td>PSP-D</td>
<td>1.15 1.14 1.11 1.12 1.10 1.06 1.07 1.06 1.04 1.03 1.04 1.03 1.06 1.03 1.08 1.09 1.09</td>
<td>0.3 0.4 0.5 0.5 0.6 0.7 0.9 1.1 1.3 1.6 1.8 1.9 2.3 2.7 3.3 4.3 4.8 6.7 7.9 10.9 13.3</td>
</tr>
<tr>
<td>SJP-D</td>
<td>1.00 1.06 1.07 1.04 1.16 1.19 1.08 1.27</td>
<td>8.1 8.5 9.4 11.2 14.4 13.8 18.4 14.1</td>
</tr>
<tr>
<td>OKH-D</td>
<td>1.15 1.13 1.15 1.13 1.16 1.13 1.16 1.19 1.20 1.19 1.22 1.16 1.08 1.07 1.13 1.24</td>
<td>0.3 0.4 0.5 0.6 0.7 0.9 1.0 1.1 1.3 1.6 2.3 2.9 3.8 5.2 6.1 6.5 7.4</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>( q_T ) Classification</th>
<th>( q_T &lt; 1.15 )</th>
<th>( 1.15 \leq q_T &lt; 1.30 )</th>
<th>( 1.30 \leq q_T &lt; 1.40 )</th>
<th>( 1.40 \leq q_T &lt; 1.5 )</th>
<th>( 1.50 \leq q_T &lt; 1.6 )</th>
<th>( 1.6 \leq q_T )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Insignificant</td>
<td>Weak</td>
<td>Moderate</td>
<td>Significant</td>
<td>Strong</td>
<td>Very Strong</td>
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7. Additional References


Telesca, L., 2010b. A nonextensive approach in investigating the seismicity of L’Aquila area (central Italy), struck by the 6 April 2009 earthquake (Mw 5.8), Terra Nova, 22, 87–93.


Vallianatos F., Michas G., Papadakis G. and Sammonds P., 2012b. A non-extensive statistical physics view to the spatiotemporal properties of the June 1995, Aigion earthquake (M6.2) aftershock sequence (West Corinth Rift, Greece), Acta Geophysica, 60 (3), 758-768.


