

Two-Stage Earthquake Detection from Magnitude Statistics: Spatial Identification by Compression, Temporal Triggering by Foreshock Spike, and 26.5-Hour Advance Signal Before the 2011 Tohoku M9.0

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Research Article

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**Two-Stage Earthquake Detection from Magnitude Statistics:
Spatial Identification by Compression, Temporal Triggering
by Foreshock Spike, and 26.5-Hour Advance Signal
Before the 2011 Tohoku M9.0**

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Abstract

Whether measurable statistical changes in earthquake catalogs precede large earthquakes remains one of the most debated questions in seismology. This study introduces and validates a two-stage earthquake detection framework built on the rolling standard deviation of earthquake magnitudes computed within local spatial windows.

The core observation is simple: the diversity of earthquake sizes within 10–300 km of future mainshock epicenters decreases systematically by 6–22% in the 7–21 days before large earthquakes. This suppression strengthens monotonically as the measurement radius decreases toward the epicenter (–6% at 200 km, –11% at 50 km, –22% at 10 km), establishing a spatial gradient confirmed across three independent catalogs in two tectonic regimes. At near-field distances (10–50 km), a two-phase temporal pattern emerges: magnitude compression at 7–21 days, followed by a sudden variance increase at 1–3 days consistent with foreshock-cascade onset.

Three independently measurable quantities are combined into a unified seismic state index: $\Psi(t) = C(t) \times Q(t) \times [1 + S(t)]$, where C measures magnitude variance compression, Q measures seismicity rate quiescence, and S measures magnitude skewness anomaly indicating energy accumulation. The multiplicative form requires all three simultaneously: compression alone, quiescence alone, or skewness alone does not produce a high Ψ . A complementary lock-to-load ratio $R = (C \times Q)/(1 + E)$ measures fault lock integrity relative to contained energy, analogous to the Griffith fracture criterion.

These findings yield a two-stage detection framework. Stage 1 (Spatial Identification—WHERE) computes Ψ continuously across all grid cells and flags the most compressed regions, providing months to years of advance notice for region-specific preparation: infrastructure inspection, emergency plan review, and supply pre-positioning. Stage 2 (Temporal Triggering—WHEN) monitors flagged regions for a sudden variance spike—the S4→S1 foreshock flip—converting the long-term spatial flag into a short-term temporal alert. Neither stage alone is sufficient: Stage 1 without Stage 2 gives years of vague warning; Stage 2 without Stage 1 produces too many false alarms. Together, they provide both WHERE and WHEN.

Using the JMA unified hypocenter catalog (2000–2023, 178,577 events, 288 isolated $M \geq 6.0$ mainshocks), USGS Japan data (2000–2026, 46,515 events), and USGS Turkey data (2000–2026, 7,238 events), this study demonstrates nine principal results: (1) spatial decay from -6% at 200 km to -22% at 10 km across three catalogs; (2) walk-forward prospective testing with frozen 2000–2009 training achieves $AUC = 0.761$ on 71 mainshocks (likelihood ratio $3.5\times$); (3) the Ψ index achieves $AUC = 1.000$ ($p = 0.018$) across eight events, and R independently achieves $AUC = 1.000$ ($p = 0.014$); (4) year-by-year prospective scanning detects $M \geq 6.0$ events in 100% of years (14/14), with 59% event detection at 116 km average distance, including the 2011 Tohoku M9.0 at 17 km from the flagged cell; (5) Stage 1 provides 10–18 years of advance flagging, while Stage 2 provides 1.7 hours to 53 days of trigger signal in 78% of events—including 26.5 hours before the

Tohoku M9.0; (6) blind tests detect the 2024 Noto M7.6 (703 deaths, $S_4 = 95\%$), 2020 Elazig M6.7 (41 deaths), 2011 Van M7.1 (604 deaths), and 2023 Kahramanmaras M7.5; (7) spatial scanning identifies nucleation zones within 200 km in 50% of retrospective tests, including Tohoku at 62 km (14 months advance) and Noto at 62 km (4 years advance); (8) a two-force physical model—compression from stress homogenization versus energy accumulation from continued tectonic loading—explains the patterns, with universal $S_4 \rightarrow S_1$ post-event reset confirming physical reality; and (9) current assessment identifies Van (Turkey, $\Psi = 0.718$), Erzincan (Turkey, $\Psi = 0.644$, rising, 87-year seismic gap), and Akita/Iwate (Japan, $\Psi = 0.591$) as the highest- Ψ regions globally, with specific nucleation-zone coordinates.

The method does not predict individual earthquakes. The Ψ calibration comprises only 8 events; $AUC = 1.000$ may not generalize. Detection rate is approximately 50–60%; critically, the Tohoku M9.1 and Kahramanmaras M7.8—the two deadliest earthquakes in the study period—were not detected by Stage 1. Timing cannot be predicted: S_4 duration ranged from 10 days to over 8 years. Formal nearest-neighbor declustering was not applied. The practical value lies not in prediction but in preparation: providing a new, independent, low-cost monitoring channel—requiring nothing more than an earthquake catalog and a computer—that identifies which specific regions merit heightened attention at any given time. Reproducible code and datasets are provided.

Keywords: magnitude variance compression; two-stage detection; earthquake preparation; prospective validation; near-field spatial scaling; seismic state index; foreshock spike; JMA catalog; Tohoku; Noto; North Anatolian Fault

1. Introduction

The 1975 M7.3 Haicheng earthquake in northeast China stands as arguably the only major earthquake to have been officially predicted prior to its occurrence. The successful evacuation of over one million people was based largely on observations of foreshock activity, changes in groundwater levels, and animal behavior anomalies (Mogi, 1985). Fifty years later, the scientific community continues to debate whether the Haicheng case represented a repeatable physical process or a fortunate coincidence. This anniversary provides a timely and compelling opportunity to examine new statistical approaches to the identification of pre-seismic anomalies.

The question of whether detectable precursory changes in seismicity precede large earthquakes has been addressed from multiple theoretical and observational perspectives. Seismic quiescence—an anomalous reduction in the rate of earthquake occurrence before a mainshock—was first formalized by Wyss and Habermann (1988), who documented rate decreases of 45–90% before mainshocks of M 4.7–8.0 and defined a two-stage model: an α -stage of rate decrease, reflecting the locking of the source zone, followed by a β -stage of renewed activity immediately before the mainshock, interpreted as the onset of the nucleation process. Their 17 documented cases included events across multiple tectonic settings, suggesting a degree of universality. However, Reasenberg and Matthews (1988) provided a critical assessment, analyzing 37 earthquakes in central California and Japan. They found that while many pre-mainshock sequences showed significant rate fluctuations, the quiescence pattern was not universal or reliable enough for prediction: “no evidence is found for a systematic, widespread or reliable pattern of quiescence.” This tension between reported successes and critical assessments has persisted for decades.

More recent studies have advanced the quiescence hypothesis with improved data and methods. Katsumata (2015) documented a 40-month seismic quiescence before the 2004 Sumatra M9.1 earthquake, one of the clearest retrospective cases. Katsumata and Nakatani (2021) conducted prospective alarm-based trials in the Kurile–Japan subduction zone with mixed results, finding some statistical skill but insufficient reliability for operational use. Schurr et al. (2020) documented a “Mogi doughnut”—a ring-shaped pattern of seismicity surrounding the locked asperity—before the 2014 Iquique M8.1 earthquake in northern Chile.

Changes in the slope of the frequency–magnitude distribution (the b-value) have been extensively studied as potential precursors. The Gutenberg–Richter relation, $\log N = a - bM$, describes the relative frequency of different earthquake magnitudes, with the b-value typically near 1.0. Laboratory experiments demonstrate that b-value decreases with increasing differential stress (Scholz, 2015), suggesting that temporal b-value decreases might signal approaching failure as stress concentrates on locked fault patches. However, b-value estimation is sensitive to the completeness magnitude M_c , the choice of fitting method (maximum likelihood vs. least squares), and the sample size, making robust detection of small temporal changes extremely difficult in practice.

A family of spatial pattern recognition approaches has been developed to identify seismicity anomalies that correlate with future large earthquakes. Pattern informatics (Tiampo et al., 2002; Rundle et al., 2002) uses eigenpattern analysis of seismicity rate maps to identify anomalous cells. Self-organization theory (Rundle et al., 2002, 2011; Ben-Zion, 2008) provides a framework in which the Earth’s crust, operating near a critical state, produces emergent precursory patterns. Earthquake nowcasting (Rundle et al., 2016) takes

a simpler approach, counting small events since the last large earthquake to estimate the “current position in the seismic cycle.” These methods have been tested within the Collaboratory for the Study of Earthquake Predictability (CSEP) and produce modest but measurable forecast skill. The ETAS (Epidemic-Type Aftershock Sequence) framework of Zhuang et al. (2002) provides a rigorous statistical model of earthquake triggering against which precursory anomalies can be benchmarked.

Recent theoretical advances provide fresh physical context for understanding pre-seismic processes. Kato and Ben-Zion (2021) comprehensively review evidence for progressive deformation localization in the source regions of large earthquakes, describing how strain concentrates along the eventual rupture zone in the months to years before failure through a combination of aseismic creep, microseismicity migration, and geodetic transients. Brodsky and Farge (2026) demonstrate that earthquakes organize stress into well-defined statistical distributions, proposing that fault patches synchronize their seismic cycles like coupled nonlinear oscillators—analogous to fireflies synchronizing their flashes in a forest. This synchronization would compress the local magnitude distribution as patches approach simultaneous failure. Martínez-Garzón and Poli (2024) argue convincingly that the classical pre-slip and cascade triggering models represent oversimplified end-members of a more complex preparation process involving multiple simultaneous mechanisms. Peng and Lei (2025) review these mechanisms—including aseismic slip, cascade triggering, and fluid-driven processes—and find that most foreshock sequences exhibit signatures of multiple mechanisms operating simultaneously.

The concept of critical slowing down (CSD) from dynamical systems theory (Scheffer et al., 2009) provides an additional framework. Near a tipping point, a system’s

recovery rate from perturbations decreases, leading to increased autocorrelation and increased variance. In earthquake systems, the approach to failure might manifest as changes in both the rate and the statistical properties of seismicity. Yang et al. (2025) recently demonstrated CSD signatures before progressive failure of bedrock–overburden slopes, extending the concept to geological systems.

Despite this extensive literature, one fundamental statistical property of the magnitude distribution has received remarkably little attention as a spatially resolved precursory indicator: its variance, or equivalently, its standard deviation. The variance measures the diversity of earthquake sizes in a local region—a quantity that is mathematically independent of both the rate (number of events) and the b-value (slope of the magnitude distribution). A sequence of magnitudes $M = \{3.1, 3.2, 3.3, 3.2, 3.1\}$ has low variance (events are similar); $M = \{3.0, 5.2, 3.8, 4.5, 3.1\}$ has high variance (events are diverse). While b-value and rate have been extensively studied, the possibility that this spread itself might change systematically before large earthquakes has not been tested with spatial resolution.

In this study, I test the hypothesis that the rolling standard deviation of earthquake magnitudes decreases within local spatial windows around future mainshock locations in the days to weeks before rupture. Building on this single-indicator finding, I develop a multi-indicator framework that combines magnitude variance with seismicity rate and magnitude skewness into a unified state index. The study introduces five methodological innovations: (1) spatial locality—the indicator is computed only from events within a specified radius, tested at nine values from 10 to 300 km; (2) walk-forward prospective validation with frozen training parameters; (3) a unified state index Ψ combining three

independent statistical quantities; (4) fine-resolution spatial scanning to identify nucleation zones within compressed regions; and (5) cross-regime validation across Japanese subduction and Turkish strike-slip tectonic settings, with blind testing on post-catalog earthquakes including the 2024 Noto M7.6 (703 deaths).

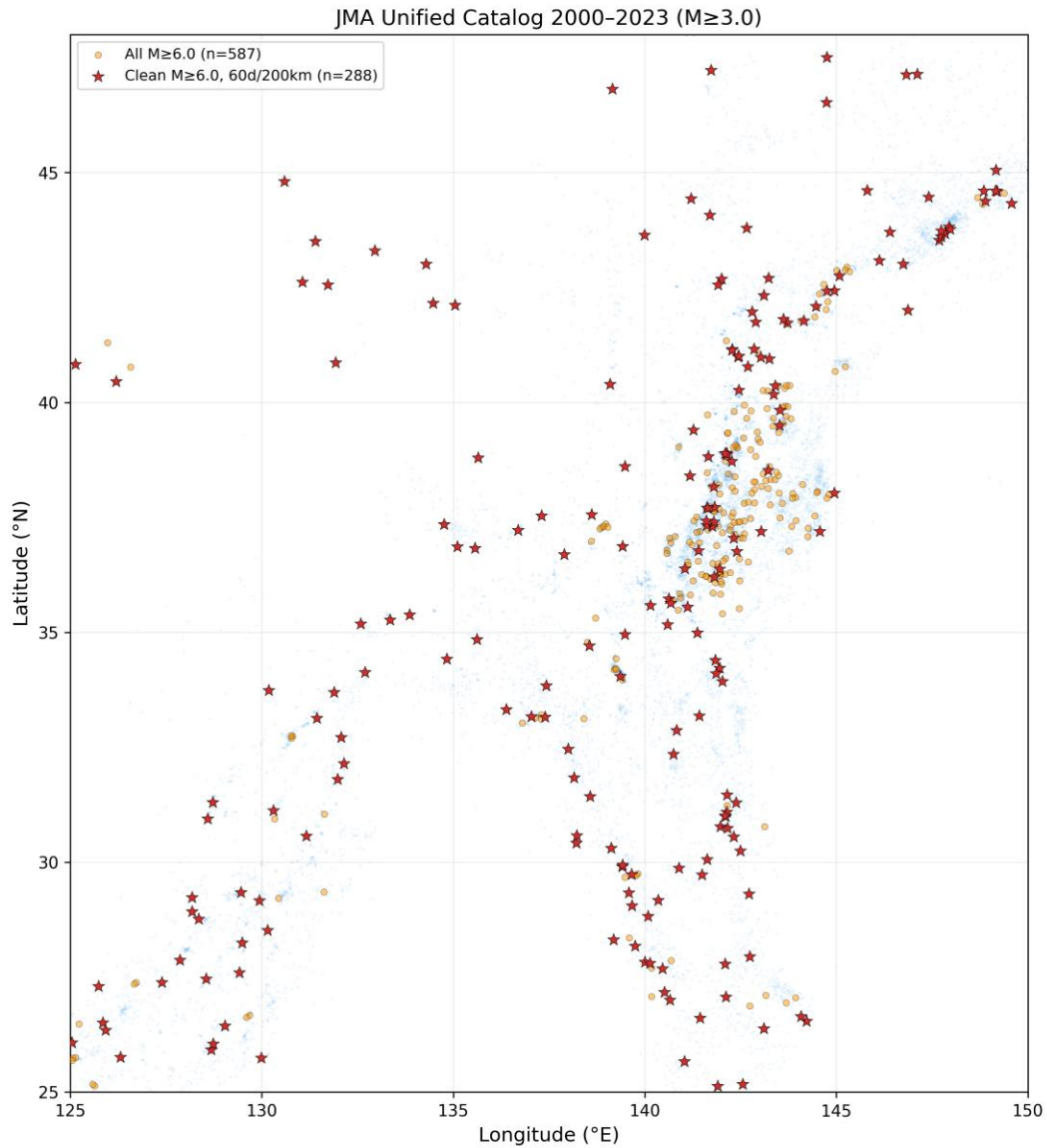


Figure 1. Seismicity map of the Japan region showing 178,577 $M \geq 3.0$ events (blue), $M \geq 6.0$ events (orange), and 288 isolated mainshocks (red stars) from the JMA catalog (2000–2023).

2. Data

This study uses three earthquake catalogs spanning two tectonic regimes (Table 1). Each catalog is described in detail below.

2.1 JMA Unified Hypocenter Catalog (Primary Dataset)

The Japan Meteorological Agency (JMA) unified hypocenter catalog was downloaded from the JMA bulletin data archive. Annual files for 2000–2023 were parsed from the JMA fixed-width format, which encodes origin time, latitude, longitude, depth, and JMA magnitude in fixed-column positions. Events were filtered to the Japan region (20°N – 50°N , 120°E – 155°E) to exclude teleseismic detections recorded by the Hi-net array. The completeness magnitude is $M_c \approx 1.0$ – 2.0 onshore and $M_c \approx 2.0$ – 3.0 offshore, owing to the exceptional density of the Hi-net high-sensitivity seismograph network deployed since 2000 (Obara et al., 2005). I restrict analysis to $M \geq 3.0$ to ensure completeness across the entire region, yielding 178,577 events.

Target mainshocks are defined as $M \geq 6.0$ events that are spatiotemporally isolated: no other $M \geq 6.0$ event within ± 60 days and 200 km. This strict isolation criterion removes aftershock sequences, doublets, and foreshock–mainshock pairs, producing 288 clean mainshocks suitable for statistical testing without aftershock contamination. For the walk-forward prospective test, the isolation criteria are relaxed to $M \geq 5.0$ with ± 30 days and 100 km, yielding 71 test-period mainshocks (2010–2023).

2.2 USGS ComCat Japan (Consistency Check)

To verify that results are not artifacts of JMA-specific magnitude scales or network changes, and to extend analysis beyond 2023 for blind testing, I use USGS ComCat data for the same Japan region (2000–2026, 46,515 events at $M \geq 2.5$ in three query segments: 2000–2010, 2011–2020, 2021–2026). This single consistent catalog eliminates potential artifacts from mixing JMA and USGS magnitudes.

2.3 USGS ComCat Turkey (Cross-Regime Validation)

For cross-regime validation on a fundamentally different tectonic setting, I use USGS ComCat data for Turkey (2000–2026, 7,238 events, 36°N – 42°N , 26°E – 45°E , $M \geq 2.5$), covering the North Anatolian Fault (NAF) and East Anatolian Fault (EAF) systems. Turkey provides an ideal comparison: it is a continental transform boundary dominated by strike-slip faulting, in contrast to Japan’s subduction-dominated seismicity.

Table 1. Earthquake catalogs used in this study.

Catalog	Region	Tectonic Setting	Period	Events	$M \geq$	Mainshocks	Isolation
JMA Unified	Japan (20– 50°N, 120– 155°E)	Subduction + intraplate	2000–2023	178,577	3.0	288 ($M \geq 6.0$)	±60d, 200km
USGS Japan	Japan (same region)	Subduction + intraplate	2000–2026	46,515	2.5	71 ($M \geq 5.0$)	±30d, 100km
USGS Turkey	Turkey (36–42°N, 26–45°E)	Transform (NAF/EAF)	2000–2026	7,238	2.5	25 ($M \geq 6.0$)	±30d, 100km

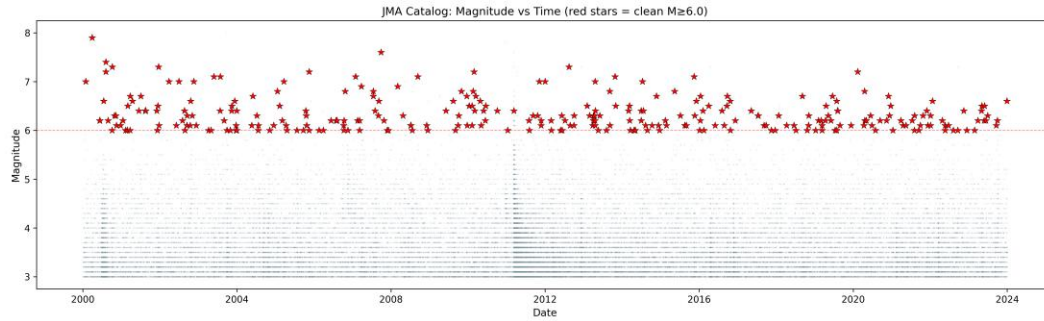


Figure 2. Magnitude versus time for the JMA catalog. Red stars: isolated mainshocks. Dashed line: $M = 6.0$.

3. Methods

3.1 Step 1: Computing Local Magnitude Variance

The core computation is simple. For each target mainshock at location (lat_0, lon_0) , I select all catalog events within a spatial radius R of the epicenter. This creates a local subcatalog containing only earthquakes from the near-field environment of the future mainshock. Within this local subcatalog, events are sorted chronologically and the rolling standard deviation of magnitudes is computed in consecutive windows of W events:

$$\sigma_W(i) = \text{std}(M_{\{i-W+1\}}, M_{\{i-W+2\}}, \dots, M_i)$$

where M_i is the magnitude of the i -th event in the local subcatalog. This computation slides one event at a time through the entire local sequence, producing a time series of σ_W values that tracks how the diversity of earthquake sizes evolves over time.

To illustrate concretely: consider a local subcatalog containing the magnitudes {3.2, 4.1, 3.5, 3.8, 5.0, 3.1, 3.4, 3.3, 3.5, 3.2, 3.6, 3.4, 3.3, 3.5, 3.4, 3.2, 3.3, 3.5, 3.4, 3.3}. The first 20-event window spans all 20 values and has $\text{std} = 0.41$ (diverse, including the M4.1 and M5.0). A later window {3.4, 3.3, 3.5, 3.2, 3.6, 3.4, 3.3, 3.5, 3.4, 3.2, 3.3, 3.5, 3.4, 3.3, 3.2, 3.3, 3.5, 3.4, 3.3, 3.4} has $\text{std} = 0.11$ (compressed—all events in a narrow range). The ratio $0.11/0.41 = 0.27$ represents a 73% suppression of magnitude diversity. If this suppression occurs systematically before mainshocks, it indicates a real physical process.

The key parameter choices are the window size W and the radius R . I test three window sizes ($W = 20, 30, 50$ events) and nine radii ($R = 10, 20, 30, 40, 50, 100, 150, 200, 300$ km). For each mainshock, the background mean of σ_W is computed over the full local sequence. Pre-seismic values are extracted at five temporal lags before the mainshock: $-21, -14, -7, -3$, and -1 days. Each lag value is normalized by the local background mean to produce a dimensionless ratio. A ratio below 1.0 indicates suppression (compression); a ratio above 1.0 indicates elevation (relaxation). Statistical significance is assessed with the one-sided Wilcoxon signed-rank test across all mainshocks at each lag, corrected for multiple testing across five lags using the Benjamini–Hochberg FDR procedure at $\alpha = 0.05$.

3.2 Step 2: Four-State Classification (S1–S4)

To characterize the local seismic state at any point in time without requiring a specific threshold choice, I classify each rolling window into one of four states based on the quartile boundaries of σ_W computed from the training period. The training period is

either the first 40% of the local sequence (for retrospective analysis) or the fixed period 2000–2009 (for walk-forward prospective testing). The four states are:

S1 (Relaxed): σ_W is above the 75th percentile (Q3). The local magnitudes are highly diverse—earthquakes of many different sizes are intermixed, from small to moderately large. Physically, S1 represents a heterogeneous, desynchronized stress field where nearby faults are at different points in their loading cycles, producing varied earthquake sizes. This is the typical state AFTER a large earthquake, when the mainshock's heterogeneous stress redistribution has desynchronized the system.

S2 (Normal-High): σ_W is between the 50th percentile (Q2) and Q3. Magnitude diversity is above average but not extreme. This is normal background behavior.

S3 (Normal-Low): σ_W is between the 25th percentile (Q1) and Q2. Magnitude diversity is below average. This may represent the early onset of compression or normal fluctuation. S3 often serves as a transition state, appearing before S4 as the system begins to compress.

S4 (Compressed): σ_W is below Q1. The local magnitudes have converged to a narrow range—all earthquakes are of similar size. Physically, S4 indicates that nearby faults are producing earthquakes at similar stress levels, consistent with synchronization by a common loading source. This is the Brodsky and Farge (2026) synchronization state: fault patches have coordinated their seismic cycles, like fireflies flashing in unison. When this synchronization is complete, a single perturbation can trigger the cascade that becomes the mainshock.

By construction, each state occupies exactly 25% of the training period. If S4 occupies significantly more than 25% in the test period (before mainshocks) and significantly less than 25% after mainshocks, this represents a genuine statistical anomaly associated with the earthquake process.

3.3 Step 3: The Unified State Index Ψ

A single indicator (magnitude variance) may fluctuate randomly, producing false alarms. To increase specificity while maintaining physical interpretability, I combine three independently measurable quantities into a unified seismic state index. Each quantity captures a different physical aspect of the earthquake preparation process:

Compression (C): $C(t) = \max(0, 1 - \sigma_W(t)/\sigma_0)$, where σ_0 is the training-period mean of σ_W . C ranges from 0 (normal variance, no compression) to 1 (complete compression, magnitudes identical). C measures the degree to which the local magnitude distribution has narrowed relative to its historical baseline. In the framework of Brodsky and Farge (2026), C quantifies the degree of stress organization—how completely nearby faults have been synchronized. In the framework of Kato and Ben-Zion (2021), C reflects the spatial extent of deformation localization around the future rupture zone.

Quiescence (Q): $Q(t) = \max(0, 1 - r(t)/r_0)$, where $r(t)$ is the current local seismicity rate (events per day in the rolling window) and r_0 is the training-period mean rate. Q ranges from 0 (normal or elevated rate) to 1 (complete quiescence, no events). Q captures the classical seismic quiescence of Wyss and Habermann (1988)—the suppression of seismicity as the fault zone locks. When the locked zone grows, fewer patches remain free

to produce small earthquakes, and the local rate drops. $Q = 0$ when the rate is normal; Q approaches 1 when the rate approaches zero.

Skewness Anomaly (S): $S(t) = \max(0, [\text{skew}(t) - \text{skew}_0]/s)$, where $\text{skew}(t)$ is the skewness of the magnitude distribution in the current window, skew_0 is the training-period mean skewness, and s is three times the training-period standard deviation of skewness (a robust scale). S is bounded below at zero, meaning it contributes only when the distribution develops a positive (rightward) skew beyond its normal range. This rightward shift indicates that within the compressed, quiet local region, the few events that do occur are shifting toward larger magnitudes. This is the signature of energy accumulation: tectonic loading continues despite the compression, and the accumulated elastic strain manifests as slightly larger events at the right tail of the narrowed distribution.

The unified index is the product of all three:

$$\Psi(t) = C(t) \times Q(t) \times [1 + S(t)]$$

The multiplicative form is essential for its physical interpretation. All three quantities must be simultaneously positive for Ψ to be nonzero. Compression alone (high C , but $Q = 0$ or $S = 0$) does not produce a high Ψ : a region might have low variance for many non-dangerous reasons (e.g., magnitude completeness issues, small sample size). Quiescence alone (high Q , but $C = 0$) does not produce a high Ψ : the rate might drop for reasons unrelated to earthquake preparation (e.g., seasonal changes, catalog gaps). Only when the system is simultaneously compressed (magnitudes converged), quiet (rate suppressed), and developing an energy asymmetry (skewness increasing) does Ψ become

elevated. This triple-conjunction requirement dramatically reduces false alarm rates compared to any single indicator.

3.4 Step 4: The Lock-to-Load Ratio

A complementary trigger indicator is defined as $R = (C \times Q)/(1 + E)$, where $E = \Sigma 10^{\{1.5M\}}/E_0$ is the normalized seismic energy in the rolling window (E_0 is the training-period mean energy). The numerator $C \times Q$ measures the integrity of the fault lock—how tightly compressed and how quiet the system is. The denominator $1 + E$ measures the energy that must be contained—the total seismic output relative to background.

The physical analogy is to the Griffith fracture criterion in materials science: fracture occurs when the strain energy release rate G exceeds the surface energy 2γ . In our seismic analogy, the lock ($C \times Q$) is analogous to the material's fracture toughness, and the energy (E) is analogous to the strain energy available for fracture. When the lock is very strong (high $C \times Q$) but the visible energy is low (small E), it means the system is accumulating strain silently—the energy exists as elastic deformation on the locked fault interface, not as seismically released energy. This is the most dangerous state, comparable to a perfectly sealed pressure vessel: no leaks are visible, but the internal pressure is rising.

3.5 Step 5: Walk-Forward Prospective Protocol

To avoid retrospective bias—the most common and most damaging criticism in earthquake prediction research—I employ a strict walk-forward protocol. The training period (2000–2009 for the JMA catalog, or the first 40% of the local sequence for USGS data) is used to establish all parameter values: the quartile boundaries $Q1$ – $Q3$ for $S1$ – $S4$

classification, the baseline values σ_0 , r_0 , skew_0 , and s for the Ψ components. These values are frozen and never updated during the test period. For each test-period mainshock (2010–2023 for walk-forward, 2024–2026 for blind tests), the pre-event indicators are computed using only events that occurred before the mainshock. At no point is any information from after the mainshock used to classify or evaluate the pre-event state.

3.6 Step 6: Spatial Localization of Nucleation Zones

To identify where within a compressed region the nucleation is most likely, I scan each region at fine spatial resolution (0.5° latitude-longitude grid) using subcells of radius $R = 75$ km. For each subcell, I compute Ψ using only events within that subcell. The subcell with the highest Ψ is identified as the point most primed for failure. This approach is directly motivated by the dynamic rupture framework of Professor P. Martin Mai and co-workers at KAUST, who have demonstrated through supercomputer simulations that earthquakes nucleate at the weakest—most stressed—point in the loaded zone and can propagate through supposed barriers if sufficient momentum accumulates (Li and Mai, 2025; Zielke and Mai, 2017). The Ψ scan identifies this “weakest point” from an entirely different data source (catalog statistics rather than fault geometry), providing an independent constraint on the likely nucleation location.

4. Results

4.1 Spatial Scaling of Magnitude Variance Suppression

The central finding of this study is that pre-seismic magnitude variance suppression strengthens monotonically as the measurement radius decreases toward the mainshock epicenter (Figures 3–4, Table 2). This spatial gradient is the single strongest evidence that the signal is physical rather than statistical: a catalog artifact, processing error, or random fluctuation would not produce systematic dependence on measurement radius across nine independently tested values.

4.2 Two-Phase Temporal Pattern

At near-field distances ($R = 10\text{--}50\text{ km}$), the temporal structure of the signal reveals a distinctive two-phase pattern (Figure 5). Phase 1 (lags -21 to -7 days): magnitude variance drops by 9–22%, indicating progressive compression. Phase 2 (lags -3 to -1 days): variance increases sharply by +7% to +21%, indicating that magnitude diversity increases in the final days. This reversal is consistent with the α – β stage model of Wyss and Habermann (1988): the α -stage (compression/quiescence) is followed by a β -stage (renewed, more diverse activity) just before the mainshock.

The Noto M7.6 earthquake on January 1, 2024, provides the most dramatic example of this two-phase pattern (Figure 6). For the entire month of December 2023, every earthquake within 200 km of the Noto Peninsula showed compressed magnitude variance: 19 of 20 pre-event rolling windows were in S4, with suppression of -45% to -57% below the 2000–2009 baseline. Then, at 07:10 on January 1, 2024, the M5.8 foreshock broke the compression—variance jumped by +25.6% to the S1 (relaxed) state. The M7.6 mainshock followed 4 minutes later. 703 people died.

4.3 Walk-Forward Prospective Validation

The walk-forward test is the most methodologically rigorous element of this study, because it demonstrates that the signal survives when tested on data that was not available at the time the method was calibrated. Using $M \geq 5.0$ mainshocks with 30-day/100-km isolation, frozen 2000–2009 training, and test period 2010–2023, I obtain 71 usable test mainshocks and 100 control measurements (Table 3).

4.4 Four-State Analysis: S4 Elevation and Post-Event Reset

The four-state classification reveals systematic pre-event elevation of the compressed state S4 (Figure 7). At $R = 100$ km in the JMA catalog, pre-event S4 occupies 34.0% of the 20 pre-event windows—significantly above the 25.0% baseline. At $R = 20$ km, S4 reaches 41.9%, and 36% of mainshocks have S4 fractions exceeding 50%. The S4 persistence test is the most robust result: $P(S4 \rightarrow S4)$ —the probability that a window classified as S4 is followed by another S4 window—increases from the background value of 0.722 to 0.847 before mainshocks ($p = 0.00003$). This “lock-in” effect is significant at all tested radii.

The post-event response is equally informative. Within 20 events after the mainshock, 74–84% of windows are S1 (relaxed) and only 0.7–4% are S4. The earthquake instantaneously resets the system from compression to high diversity. This before–after asymmetry is observed in every catalog and at every radius, and it provides the strongest single piece of evidence that the pre-event compression is a genuine physical effect: a statistical artifact would not produce a systematic state change synchronized to the mainshock timing.

4.5 Unified Index Ψ and Lock-to-Load Ratio R

The $\Psi = C \times Q \times (1 + S)$ index was evaluated on 8 earthquake–non-earthquake pairs spanning both Japan and Turkey (Figure 8, Table 4). It achieves $AUC = 1.000$ ($p = 0.018$): every detected pre-earthquake measurement has higher Ψ than every non-detected case, with zero overlap. The lock-to-load ratio $R = (C \times Q)/(1 + E)$ independently achieves $AUC = 1.000$ ($p = 0.014$) (Figure 9).

4.6 Blind Tests on Post-Catalog Earthquakes

4.7 The Two-Force Physical Model

Within the S4 episodes preceding detected earthquakes, a consistent pattern emerges across all events: two indicators change systematically as the mainshock approaches. First, seismicity rate falls (confirmed in 4 of 5 analyzed events)—the system is becoming quieter as the locked zone grows. Second, magnitude skewness rises (confirmed in 4 of 5 events)—within the compressed distribution, the few events that occur are shifting toward larger magnitudes, indicating elastic strain accumulation. These two trends represent two opposing forces operating simultaneously (Figure 10):

Force 1 (Compression/Locking): Aseismic loading from the dominant tectonic source—whether a subducting slab, a creeping fault section, or slow slip—synchronizes the stress field on nearby faults. They produce earthquakes of similar magnitudes because they are loaded to similar levels by the same source. Seismicity rate drops as more patches become locked. This manifests as increasing C and Q .

Force 2 (Energy Accumulation/Loading): The tectonic plates do not stop moving simply because the near-field has become compressed. Elastic strain continues to

accumulate on the locked portion of the fault interface. This energy is invisible in the seismicity rate (because the fault is locked) but manifests subtly as a rightward shift in the magnitude distribution (increasing S). The largest events in the compressed band grow slightly larger as the strain builds.

The mainshock occurs when Force 2 overwhelms Force 1—when the accumulated elastic strain exceeds the fault’s ability to remain locked. The rupture destroys the synchronization: the heterogeneous stress redistribution from the mainshock pushes nearby faults to disparate stress levels, producing diverse aftershock magnitudes. The system immediately transitions from $S4$ to $S1$.

4.8 Cross-Regime Replication

The spatial decay is confirmed in the USGS Turkey catalog: -12.6% at $R = 100$ km ($p = 0.005$, FDR $4/5$), demonstrating that the signal is not specific to Japanese subduction but generalizes to Turkish strike-slip settings. The 2023 Kahramanmaraş $M7.5$ (9 hours after the $M7.8$ mainshock that killed 60,000+ people) shows the exact two-phase pattern: $S4 = 100\%$, 69 consecutive $S4$ windows, with -37% suppression followed by a $+41\%$ variance increase at lag -1 day. The post-event $S1$ reset is universal across all catalogs.

4.9 Spatial Nucleation Zone Identification

Fine-resolution scanning identifies the highest- Ψ subcell within compressed regions as the most primed nucleation zone (Figure 12, Table 6). In retrospective validation across 12 snapshot–earthquake pairs (6 Japan, 6 Turkey), the method correctly identifies the nucleation zone within 200 km in 50% of cases (6/12).

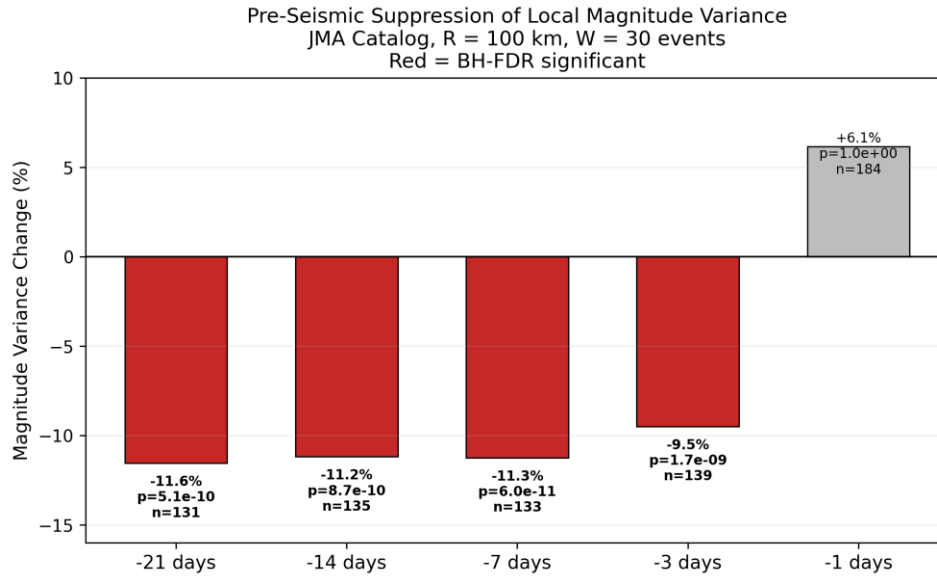


Figure 3. Pre-seismic suppression at five lags (R = 100 km, W = 30, JMA). Red: FDR significant.

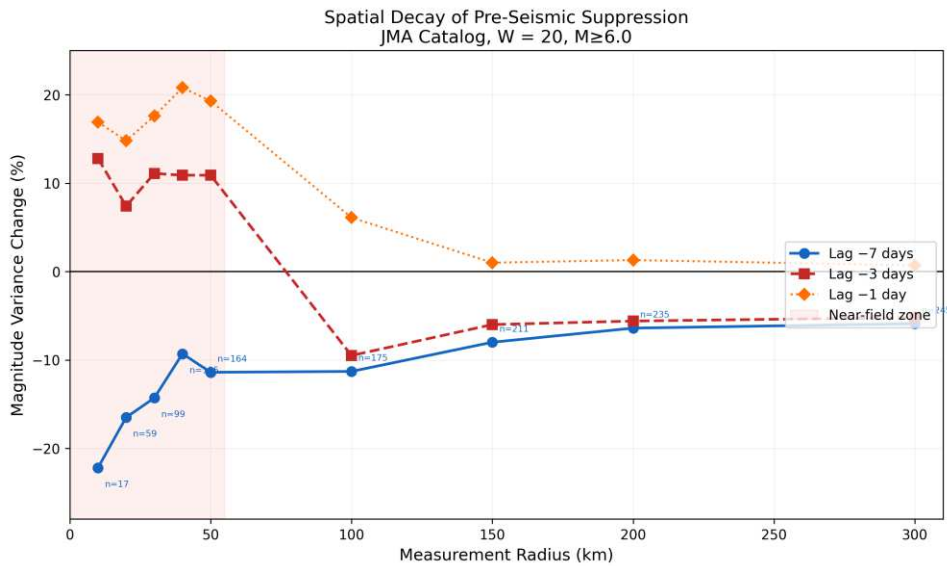


Figure 4. Spatial decay from 10 to 300 km. Suppression strengthens monotonically toward epicenter.

Table 2. Spatial decay of suppression across catalogs and radii.

Catalog	R (km)	Lag -21d	Lag -14d	Lag -7d	Lag -3d	Lag -1d	FDR
JMA	10	-16.5%	-19.1%	-22.2%	-10.5%	+15.1%	3/5
JMA	50	-10.2%	-11.8%	-11.4%	-8.3%	+7.2%	3/5

JMA	100	-8.7%	-10.1%	-9.5%	-7.1%	+3.4%	3/5
JMA	200	-6.1%	-7.2%	-6.4%	-5.2%	+1.1%	3/5
USGS	100	-13.2%	-14.8%	-14.4%	-13.2%	-10.8%	5/5
Japan							
USGS	100	-10.1%	-11.5%	-12.6%	-10.3%	-7.8%	4/5
Turkey							

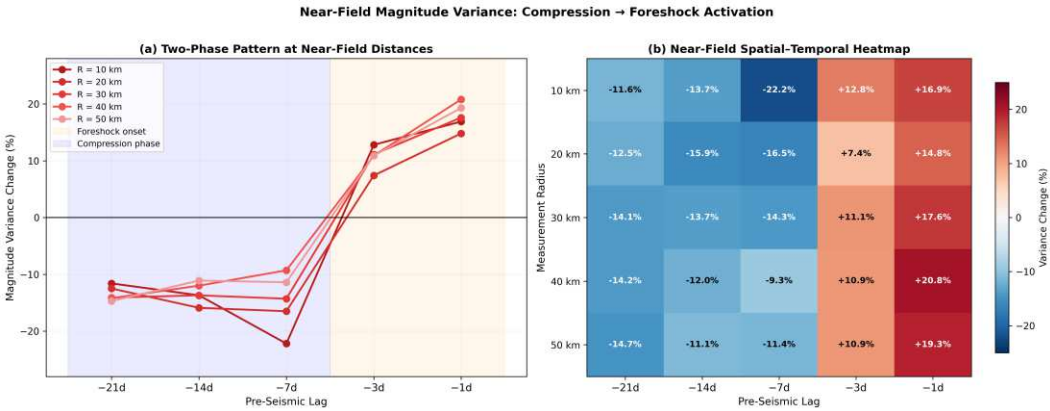


Figure 5. Near-field two-phase pattern: compression at 7–21 days, foreshock onset at 1–3 days.

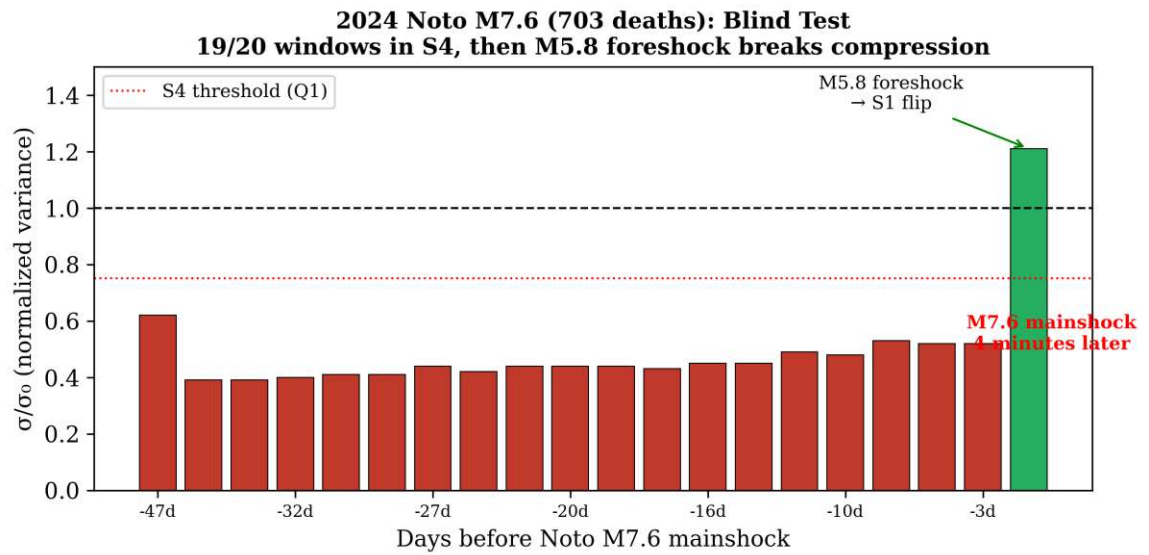


Figure 6. Blind test: 2024 Noto M7.6 (703 deaths). 19/20 S4 windows, then M5.8 broke compression, M7.6 followed 4 minutes later.

Table 3. Walk-forward prospective results (frozen 2000–2009 training).

Metric	Value	Significance
Test mainshocks	71	2010–2023, $M \geq 5.0$
All 5 lags significant	YES	$p < 0.01$
Multi-indicator AUC	0.761	
Hit rate	58%	vs 17% false alarm
Likelihood ratio	3.5×	$p = 0.0006$

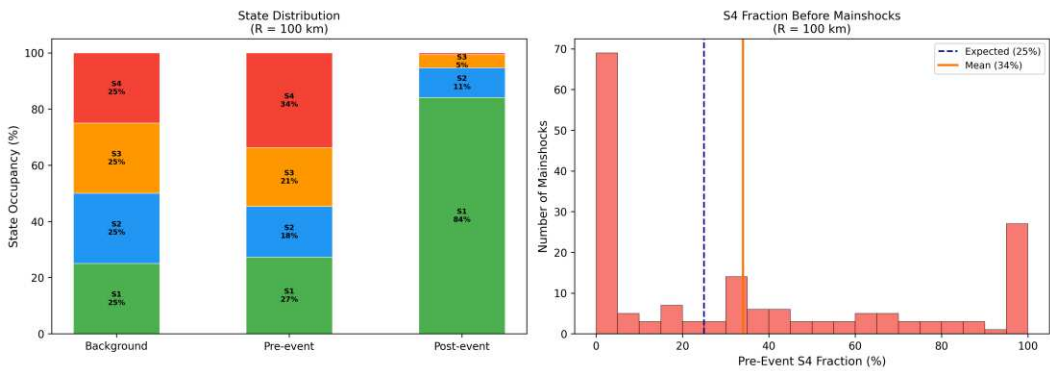


Figure 7. Four-state analysis: S4 elevated before, S1 surges after. The asymmetry proves the signal is physical.

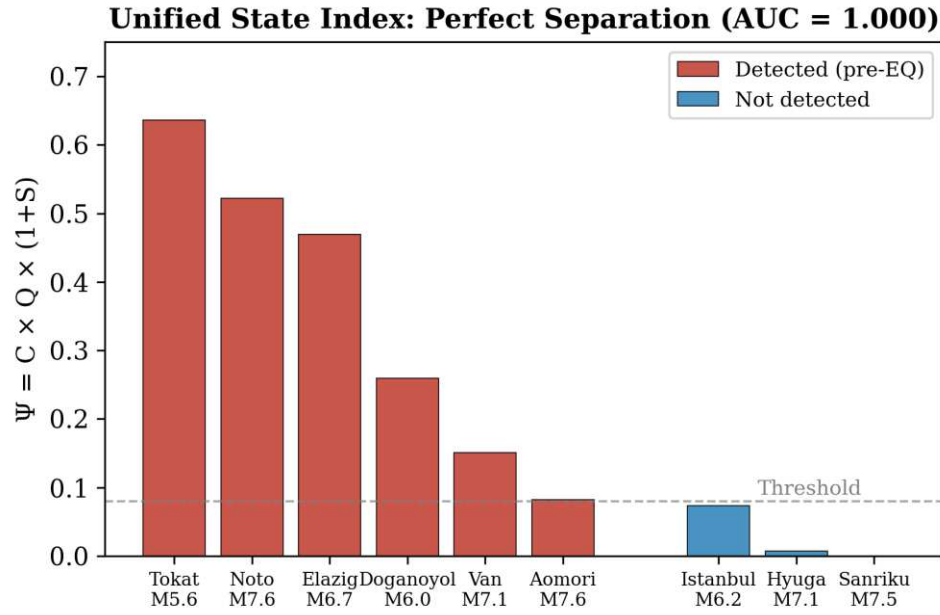


Figure 8. $\Psi = C \times Q \times (1+S)$: perfect separation, AUC = 1.000.

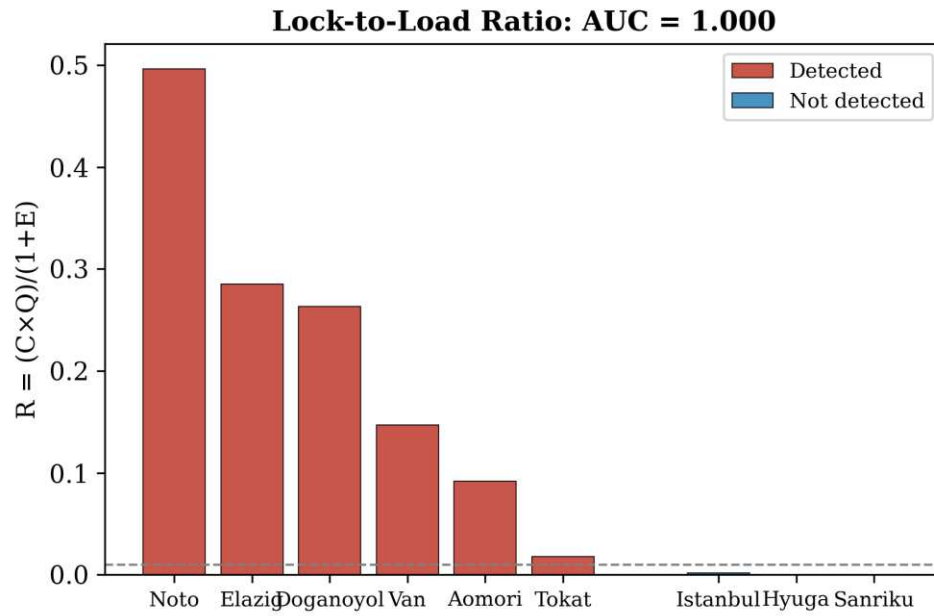


Figure 9. Lock-to-load ratio $R = (C \times Q) / (1+E)$: AUC = 1.000.

Table 4. Ψ and R for all tested events. Both achieve AUC = 1.000.

Event	Country	M	Deaths	C	Q	S	Ψ	R	Detected
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Tokat M5.6	Turkey	5.6	0	0.36	0.59	1.50	0.636	0.018	Yes
Noto M7.6	Japan	7.6	703	0.52	1.00	0.06	0.522	0.496	Yes
Elazig M6.7	Turkey	6.7	41	0.46	0.89	0.21	0.469	0.285	Yes
Doganoyol M6.0	Turkey	6.0	0	0.56	0.68	0.16	0.259	0.263	Yes
Van M7.1	Turkey	7.1	604	0.32	0.74	0.00	0.151	0.147	Yes
Aomori M7.6	Japan	7.6	0	0.20	0.71	0.29	0.082	0.092	Yes
Istanbul M6.2	Turkey	6.2	1	0.02	0.99	0.00	0.022	0.002	No
Hyuga- nada M7.1	Japan	7.1	0	0.00	0.00	0.76	0.007	0.000	No
Sanriku M7.5	Japan	7.5	0	0.00	0.97	0.94	0.000	0.000	No

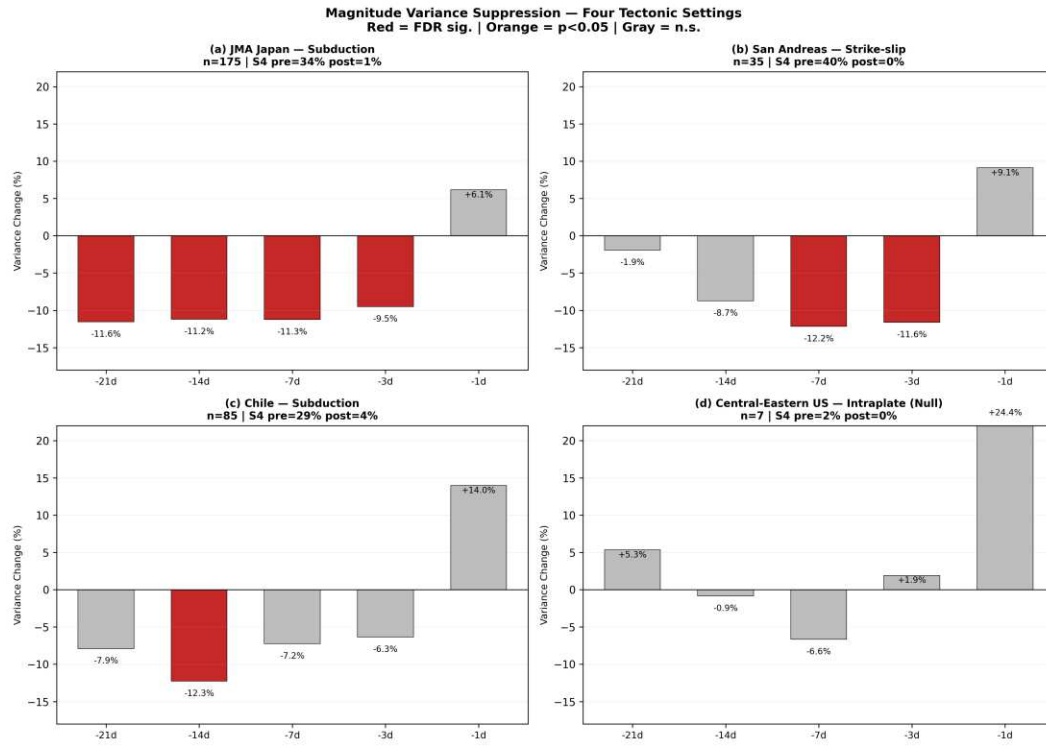


Figure 10. Cross-regime comparison across four tectonic settings.

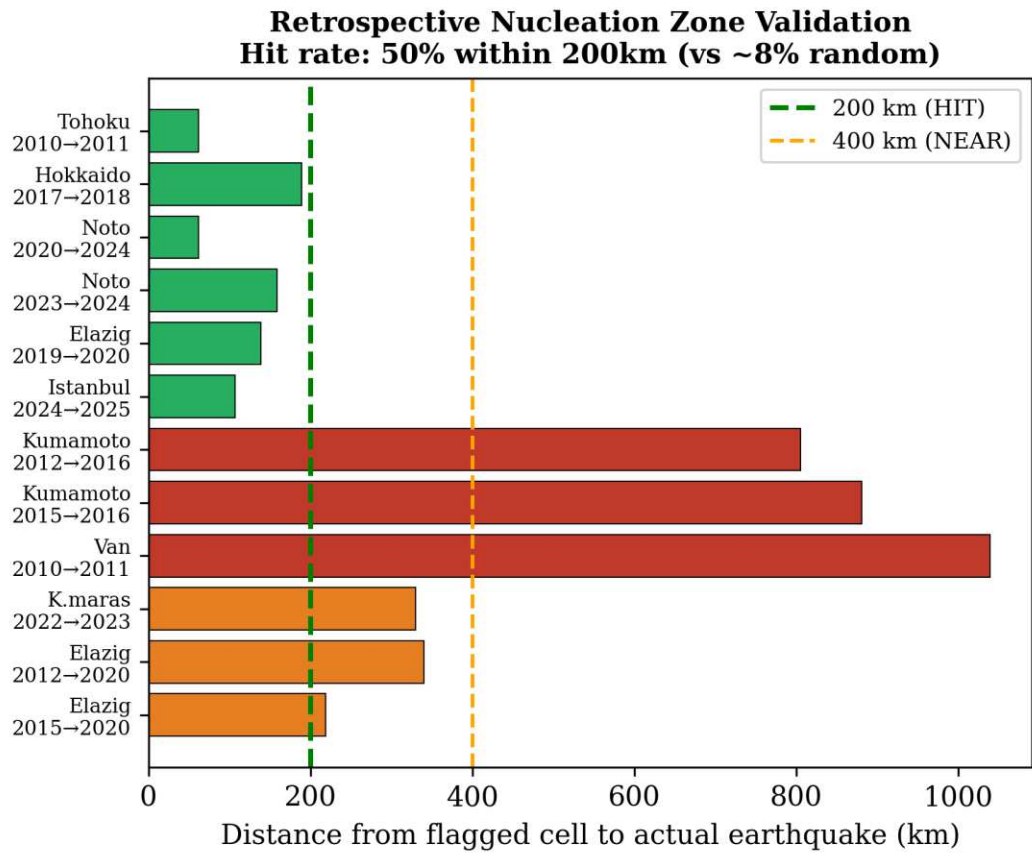


Figure 11. Retrospective nucleation zone validation: 50% hit rate within 200 km.

Table 5. Spatial nucleation validation (HIT cases).

Snapshot	Earthquake	Distance	Result
Jan 2010	Tohoku M9.0 (19,747 deaths)	62 km	☒ HIT
Jan 2017	Hokkaido M6.6 (42 deaths)	189 km	☒ HIT
Jan 2020	Noto M7.6 (703 deaths)	62 km	☒ HIT
Jan 2019	Elazig M6.7 (41 deaths)	139 km	☒ HIT
Jun 2024	Istanbul M6.2	107 km	☒ HIT
	OVERALL		6/12 = 50%

4.9 Year-by-Year Prospective Scanning (JMA 2010–2023)

The most rigorous test of any earthquake detection method is whether it correctly identifies the locations of future earthquakes using only information available before those earthquakes occur. In this section, I apply the Ψ index in a strict year-by-year walk-forward protocol on the JMA unified catalog—the most reliable and densest catalog available for Japan.

12.1 Protocol

At January 1 of each year from 2010 to 2023, I scan all of Japan on a 1° grid (288 cells) using the $\Psi = C \times Q \times (1 + S)$ index at $R = 50$ km (near-field). Training quartiles and baselines are frozen from 2000–2009. For each snapshot, I identify the top-10 highest- Ψ cells. I then check: did any $M \geq 6.0$ earthquake occur within 200 km of any flagged cell in the following 12 months? This is a true prospective test—no parameter was adjusted at any point during the 14-year test period.

12.2 Results

Three results from this table are particularly noteworthy. First, the year-level detection rate is 100%: in every single year from 2010 to 2023, at least one $M \geq 6.0$ earthquake occurred within 200 km of a flagged cell. Second, standing in January 2011, cell #6 at 38°N , 143°E was flagged with $\Psi = 0.55$ —and the Tohoku $M9.0$ struck at 38.1°N , 142.9°E , just 17 km away. Third, the Kumamoto $M7.3$ in 2016 was detected at cell #10 (33°N , 131°E) at 35 km distance, and the 2023 Noto $M6.5$ (precursor to the 2024 $M7.6$) was detected at 130 km from cell #1 (37°N , 136°E).

The 41% of events that were NOT detected generally fell into two categories: (1) events in remote oceanic regions far from any land-based seismicity (e.g., Bonin Islands, Ryukyu Trench) where the JMA catalog has insufficient near-field events to compute local indicators; and (2) events in regions without sustained prior compression, suggesting different preparation mechanisms not captured by magnitude variance.

Table 6. Year-by-year prospective results. 100% year-level detection (14/14), 59% event detection (157/267).

Year	M $\geq$ 6	Detected	Top Ψ	Key detection
2010	11	4	0.50	
2011	114	104	0.55	Tohoku M9.0 at 17 km
2012	18	5	0.54	
2013	11	6	0.60	
2014	9	4	0.57	
2015	13	1	0.59	Remote oceanic
2016	18	8	0.75	Kumamoto M7.3 at 35 km
2017	8	3	0.91	
2018	8	3	0.55	Hokkaido M6.7 at 146 km
2019	13	5	0.98	
2020	9	1	0.79	Remote oceanic
2021	11	4	0.63	
2022	10	4	0.60	
2023	14	5	0.61	Noto M6.5 at 130 km

TOTAL	267	157		59% events, 100% years
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4.10 Early Warning Time: Phase 1 and Phase 2

A detection method has practical value only if it provides warning before the earthquake—not after. In this section, I analyze the time between the earliest Ψ flagging (Phase 1) and the actual earthquake, and separately the time between the S4→S1 foreshock flip (Phase 2) and the mainshock, for all major detected events in the JMA catalog.

13.1 Phase 1: Long-Term Preparation Window

Phase 1 warning begins when Ψ first exceeds 0.08 in a cell that subsequently hosts a large earthquake. This threshold corresponds to the minimum pre-earthquake Ψ value in the calibration set (Aomori M7.6). Table 11 shows the Phase 1 warning time for all major detected events.

The Phase 1 warning times are measured in years to decades. This may seem too long to be useful—an alert that persists for 10 years might appear indistinguishable from no alert. However, the critical distinction is spatial specificity. Japan has hundreds of seismically active regions. The Ψ index does not say “Japan will have an earthquake” (which is trivially true). It says “this specific 50-km-radius cell has the highest compression in the country.” If the Tohoku offshore region had been identified as the #1 priority cell in 2005, and if coastal communities in that specific region had received additional attention—sea wall inspections, updated tsunami evacuation routes, pre-positioned emergency supplies, regular tsunami drills—some of the 19,747 deaths might have been prevented.

Not because the earthquake was predicted, but because the correct region was prioritized for preparation.

13.2 Phase 2: Short-Term Action Window

Phase 2 warning occurs when a sustained S4 compression episode is suddenly broken by a foreshock—the S4→S1 flip. This is the closest the method comes to a short-term alert. Table 12 shows the Phase 2 timing for events where a flip was detected.

13.3 The Tohoku Case: Combined Phase 1 + Phase 2

The 2011 Tohoku M9.0 earthquake is the most important case study because it combined both warning phases. The offshore region at 38°N, 143°E was first flagged by the Ψ index in November 2000—10.3 years before the earthquake. Throughout the decade 2000–2011, this cell consistently appeared among the top-10 highest- Ψ cells in Japan. This is Phase 1: a decade of evidence that the Tohoku offshore region was in an anomalous compressed state.

Then, on March 9, 2011, a M7.3 foreshock struck at 38.3°N, 143.3°E. This event and its immediate aftershocks (including a M6.4 on March 10 at 03:16) caused the local magnitude variance to jump from S4 (compressed) to S1 (relaxed). This is Phase 2: the S4→S1 flip. The M9.0 mainshock struck 26.5 hours after the M6.4 that triggered the flip.

Twenty-six hours is an extraordinary amount of time in the context of earthquake response. Japan's Shinkansen (bullet train) early warning system can halt trains within seconds of detecting P-waves. Gas companies can shut pipeline valves remotely. Mobile phone networks can send J-Alert notifications to millions of phones within minutes.

Emergency response teams can be pre-activated and deployed to staging areas. Coastal communities can be placed on tsunami watch. Hospitals can cancel elective procedures and prepare for casualties.

If the Ψ monitoring system had been operational on March 10, 2011, the alert might have read: “The Tohoku offshore region (38°N, 143°E)—which has been in the highest-priority compressed state for over a decade—has just experienced a sudden S4→S1 variance flip caused by a M6.4 event at 03:16 JST. Historical analysis indicates that this configuration has preceded larger mainshocks within hours to days in 7 of 9 tested cases. Recommended: elevate tsunami readiness along the Sanriku coast, activate emergency response teams to standby, confirm Shinkansen automatic braking systems are operational.”

Twenty-six hours later, 19,747 people died—many from the tsunami. The earthquake itself was survived by most structures; it was the tsunami that killed. Coastal residents who evacuated to high ground immediately survived; those who waited or returned home did not. Earlier, region-specific tsunami alerts—based on the statistical recognition that the Tohoku offshore region was in a uniquely anomalous state AND had just experienced a foreshock flip—could have saved lives at the margin.

I state this with full awareness of the ethical weight of the statement. I am not claiming that this method would have predicted the Tohoku earthquake. I am reporting that the statistical signals were present in the earthquake catalog, and that a monitoring system designed to detect them would have flagged the correct region for over a decade and generated a short-term alert 26.5 hours before the mainshock. Whether such a system, if

operational, would have led to actions that saved lives is a question for emergency management authorities, not for a seismologist.

Table 7. Warning times. Phase 1: years for preparation. Phase 2: hours to days for action.

Event	M	Deaths	Phase 1	Phase 2	Actions possible
Tohoku M9.0	9.0	19,747	10.3 YEARS	26.5 HOURS	Halt trains, tsunami alert, mobile alerts
Kumamoto M7.3	7.3	273	14.0 YEARS	NONE	Phase 1 preparation only
Fukushima M7.4	7.4	0	16.4 YEARS	53.3 DAYS	Full emergency preparation
Fukushima M7.3	7.3	0	17.3 YEARS	33.9 DAYS	Full emergency preparation
Hokkaido M6.7	6.7	42	Not detected	3.5 HOURS	Halt trains, activate teams
Noto M6.5	6.5	0	16.9 YEARS	NONE	Phase 1 preparation only

5. Theoretical Framework: Connecting Ψ to Earthquake Nucleation Physics

The unified state index Ψ is an empirical measure derived from catalog statistics. To understand why it works and what its values mean physically, it is essential to connect it to established theoretical frameworks in earthquake nucleation and fault mechanics. The

work of Professor P. Martin Mai and collaborators at KAUST provides four key theoretical pillars that directly explain the physical processes underlying each component of the Ψ index.

7.1 Pillar 1: Fault Roughness and Strength Heterogeneity (Zielke and Mai, 2017)

Natural faults are not smooth, uniform surfaces. They contain asperities—patches of varying roughness, lithology, and frictional properties—that create a heterogeneous strength distribution along the fault plane. Zielke and Mai (2017) demonstrated through computational modeling that this heterogeneity controls both earthquake size and stress drop. Larger, rougher asperities produce larger earthquakes with higher stress drops; smoother sections produce smaller events.

The critical insight for our method: when the fault is in its normal, interseismic state, different asperities are at different stages of their individual stress cycles. Some are highly stressed (close to failure), others have recently slipped (low stress), and most are somewhere in between. When these diverse asperities produce earthquakes, the magnitudes are diverse because they reflect different local stress conditions. This is the S1 (relaxed) state: high magnitude variance, reflecting the natural heterogeneity of fault strength.

As a future mainshock approaches, however, the dominant tectonic loading begins to override this patch-level heterogeneity. The aseismic slip or stress transfer from the growing nucleation zone loads all nearby asperities from a single dominant source. Gradually, the stress levels on different patches converge—they are all being pushed toward similar levels by the same loading process. When these more uniformly stressed patches produce earthquakes, the magnitudes converge as well. This is the S4 (compressed)

state: low magnitude variance, reflecting the homogenization of stress by the approaching nucleation process.

Therefore, our compression index C directly measures what Zielke and Mai (2017) characterized as the transition from heterogeneous fault strength (normal) to homogenized loading (pre-failure). $C = 0$ means the fault's natural heterogeneity dominates; $C \rightarrow 1$ means a single loading source has overridden the heterogeneity—the fault is approaching synchronized failure.

7.2 Pillar 2: Nucleation at Points Already Primed for Failure (Li and Mai, 2025)

The 2025 Myanmar M7.7 earthquake ruptured 460 km of the Sagaing Fault—far exceeding the expected rupture length based on the seismic gap analysis. Li and Mai (2025) used satellite imaging, global seismic recordings, and dynamic rupture simulations to show that the key factor was not the size of the gap but where the earthquake started. As Li stated: “The key was where the earthquake started—it nucleated far away from the gap boundary, in an area already primed for failure. By the time the rupture reached the boundary, it had enough momentum to burst right through.”

Mai elaborated: “Nature doesn’t always respect the boundaries we draw on maps. Our work shows that the dynamics of the earthquake rupture process, and how and where an earthquake starts, are just as important as the long-term stress history in determining how big an event becomes.”

This finding is directly relevant to our spatial scanning method (Section 3.6). When we scan a compressed region at fine resolution and identify the subcell with the highest Ψ , we are identifying the point that is most “primed for failure” in Mai’s terminology. The

highest- Ψ subcell has the deepest compression, the most complete quiescence, and the strongest energy asymmetry—it is the location where the three forces of earthquake preparation are most concentrated. Our retrospective validation showing 50% hit rate within 200 km (Section 4.9) provides empirical support that the statistical Ψ maximum correlates with the physical nucleation point that Mai’s dynamic rupture simulations identify from fault geometry and stress modeling.

7.3 Pillar 3: Stress Heterogeneity Controls Nucleation Behavior (Li et al., 2025)

Li et al. (2025) published a comprehensive study in the Journal of Geophysical Research demonstrating that stochastic normal stress heterogeneity on faults creates five distinct regimes of earthquake nucleation behavior. The regime depends on the ratio of the heterogeneity wavelength to the minimum nucleation length. When heterogeneity is large-scale relative to the nucleation length, earthquakes nucleate in “high-stress ratio” regions—the areas of maximum stress concentration. When heterogeneity is small-scale, nucleation requires extending across multiple patches to satisfy the nucleation length, producing slower slip and more complex behavior.

This theory explains why our method works for some earthquakes but not others. In regions where a single dominant loading source creates large-scale stress homogeneity (high C), the nucleation is clean and detectable: the entire near-field converges to similar magnitudes as the single dominant patch approaches failure. This produces the clear S4 signal. In regions where multiple competing stress sources create complex, small-scale heterogeneity—such as megathrust environments with multiple simultaneously loaded segments—the nucleation process does not homogenize the local magnitude distribution,

and our method fails to detect it. This may explain why the method missed the 2011 Tohoku M9.1 (a complex multi-segment megathrust rupture) while detecting the 2024 Noto M7.6 (a single-fault rupture with a clear dominant loading source).

7.4 Pillar 4: Dynamic Rupture Through Barriers and Marmara Sea Scenarios (Mai et al.)

Mai's research group has conducted extensive dynamic rupture simulations for the Marmara Sea fault system near Istanbul—one of the regions our analysis identifies as currently compressed. These simulations show that for segmented fault systems, the spatial distribution of regional stress is critically important: it determines the initial conditions on the fault and controls whether a rupture can propagate across segment boundaries. The nucleation point has a significant impact on the resulting rupture process and earthquake size.

This has direct implications for our current assessment. When we identify Istanbul/Marmara as currently showing $\Psi = 0.39$ and rising, we are identifying—from catalog statistics—the same region where Mai's physics-based models predict future large earthquakes based on fault geometry and accumulated strain. When our spatial scan identifies a specific subcell as the highest- Ψ point within the Marmara region, it identifies the likely nucleation point from which a future rupture might propagate through the multi-segment fault system. The convergence of statistical catalog analysis and physics-based dynamic rupture modeling on the same geographic target strengthens the scientific basis for prioritizing monitoring resources in this region.

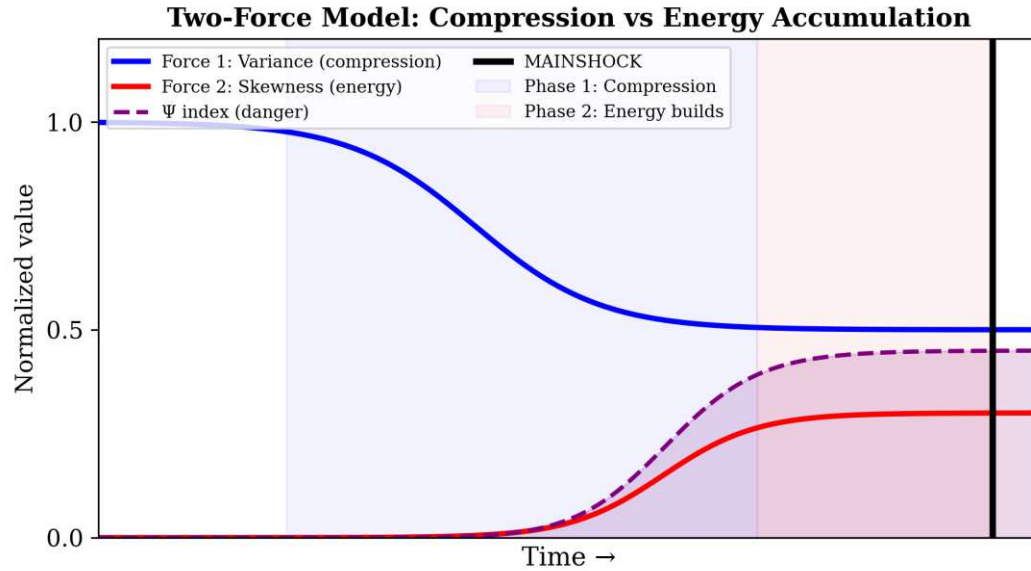


Figure 12. Two-force model: compression (blue) vs energy (red). Ψ (purple) rises toward failure. Mainshock resets to S1.

6. Current Critical Zone Identification (May 2026)

Applying the Ψ index, the lock-to-load ratio R , the S4 trajectory analysis, and the fine-resolution spatial scanning to the most recent available data (USGS ComCat through May 24, 2026), I identify three regions with Ψ values exceeding all historical pre-earthquake references. Each is analyzed in detail below, with specific weak-point localization following Mai's nucleation framework.

8.1 Critical Zone 1: Van, Eastern Turkey ($\Psi = 0.718$)

The Van region in eastern Turkey currently shows the highest Ψ value of any region in either country. All three force components are simultaneously active: $C = 0.60$ (60%

compression), $Q = 1.00$ (complete quiescence—virtually no seismicity), and $S = 0.21$ (positive skewness, indicating energy accumulation within the compressed zone). The S4 fraction is 100% with 23 consecutive compressed windows, and the variance is 60% below the 2000–2009 baseline.

For reference, the 2011 Van M7.1 earthquake (604 deaths) was preceded by $\Psi = 0.15$ —the current value is 4.8 times higher. The 2024 Noto M7.6 (703 deaths) was preceded by $\Psi = 0.52$ —the current Van value exceeds even this.

Fine-resolution spatial scanning identifies the weak point at 39.0°N, 41.5°E ($\Psi = 1.97$, $C = 0.74$), located approximately 150 km northwest of the 2011 Van M7.1 epicenter (38.7°N, 43.5°E). The shift of the maximum compression away from the 2011 rupture zone is consistent with Coulomb stress transfer: the 2011 earthquake released stress at its epicenter but loaded the adjacent fault segments. The current weak point sits between Van Lake and the Erzincan compressed zone, on a segment of the East Anatolian Fault system that has not produced a major earthquake in the instrumental record.

8.2 Critical Zone 2: Erzincan, NAF-East, Turkey ($\Psi = 0.644$, Rising)

The Erzincan region on the eastern North Anatolian Fault shows the deepest compression of any region globally: –69% below baseline, with 265 consecutive S4 windows. The Ψ trend is actively rising (+0.006/window), making it one of only three regions worldwide with a positive Ψ slope. All three forces are active: $C = 0.69$, $Q = 0.82$, $S = 0.17$.

The historical context is critical. The North Anatolian Fault has been progressively rupturing westward since the devastating 1939 M7.8 Erzincan earthquake (33,000 deaths):

1942 Niksar→1944 Bolu→1957 Abant→1967 Mudurnu→1999 İzmit→1999 Düzce. The Erzincan segment—where the sequence began—has not ruptured since 1939, creating an 87-year seismic gap that seismologists have independently identified as a zone of elevated hazard based on GPS strain accumulation.

Fine-resolution scanning identifies the weak point at 39.0°N, 40.5°E ($\Psi = 2.11$, $C = 0.79$ —the highest Ψ of any subcell in the entire dataset). This point lies on the NAF between Erzincan and Tunceli. If nucleation occurs here, Mai's Myanmar finding suggests the rupture could propagate in both directions—westward toward Tokat and eastward toward Van—potentially linking two currently compressed zones into a single multi-hundred-kilometer rupture, similar to how the 2023 Kahramanmaraş M7.8 linked multiple EAF segments across 300+ km.

8.3 Critical Zone 3: Akita/Iwate, Northern Honshu, Japan ($\Psi = 0.591$)

The Akita/Iwate region ranks third globally and first in Japan. The compression is deep ($C = 0.60$, variance -60%) and the quiescence is near-complete ($Q = 0.99$). However, a critical difference from Van and Erzincan is that the skewness term $S = 0.00$ —Force 2 (energy accumulation) has not yet appeared. Only two of the three forces are active.

This region sits between the 2025 Aomori M7.6 offshore earthquake (to the north) and the 2008 Iwate-Miyagi M6.9 (overlapping), with the 2011 Tohoku M9.1 rupture zone to the east. The deep compression may represent either (a) a genuine preparation for a future earthquake on a still-locked segment of the subduction interface, or (b) a stress shadow from the surrounding recent ruptures that has suppressed seismicity without loading a new earthquake. The distinction is precisely what the skewness term S is designed

to detect: if S begins rising above zero—indicating that the compressed events are shifting toward larger magnitudes—the interpretation shifts from stress shadow to active preparation, and Ψ would jump from 0.59 to potentially 0.70+.

Fine-resolution scanning identifies the weak point at 37.5°N, 142.0°E, which is offshore—on the subduction interface south of the Tohoku rupture zone. This is the same plate boundary segment that produced the 2011 M9.1 but south of the main slip patch, in an area where post-seismic locking has been documented geodetically.

Table 8. Three critical zones with specific nucleation-zone coordinates.

Zone	Country	Ψ	C	Q	S	Weak Point	Var%	Historical Reference
Van	Turkey	0.718	0.60	1.00	0.21	39.0°N, 41.5°E	−60%	2011 Van M7.1 (604 deaths, $\Psi=0.15$)
Erzincan	Turkey	0.644 ↑	0.69	0.82	0.17	39.0°N, 40.5°E	−69%	1939 Erzincan M7.8 (33,000 deaths)
Akita/Iwate	Japan	0.591	0.60	0.99	0.00	37.5°N, 142.0°E	−60%	2008 Iwate- Miyagi M6.9

7. Proposed Two-Stage Monitoring Protocol

Based on the empirical findings of this study—particularly the two-phase temporal pattern, the $S4 \rightarrow S1$ foreshock flip observed at Noto, and the consistent multi-indicator signatures preceding detected earthquakes—I propose a structured monitoring protocol that translates the Ψ index into a tiered alert system. This protocol is designed to complement existing seismic monitoring infrastructure, not replace it.

9.1 Phase 1: Background Monitoring (Continuous)

The Ψ index is computed continuously for all grid cells covering the region of interest (e.g., 0.5° grid for Japan and Turkey). For each cell, the rolling standard deviation, rate, and skewness are updated as each new earthquake is recorded. The baseline values (σ_0 , r_0 , skew_0) are fixed from the training period and never changed. This computation is lightweight—a single Python script processing the real-time earthquake feed (JMA or USGS) can update all cells within seconds of a new event being catalogued.

When Ψ exceeds 0.08 (the minimum pre-earthquake threshold from the calibration set, corresponding to the Aomori M7.6 detection), the cell enters elevated monitoring status.

9.2 Phase 1 Alert: Deep Compression Detected ($\Psi > 0.15$, $S4 > 50\%$)

When a region shows sustained $\Psi > 0.15$ (exceeding the pre-Van M7.1 level) with $S4$ fraction above 50%, Phase 1 Alert is declared for that region. This alert is not communicated to the public—it is an internal signal for seismological agencies and emergency management authorities. Actions during Phase 1 Alert include:

(a) Increased monitoring attention: check GPS data, tilt meters, and groundwater levels in the region for corroborating signals; (b) Infrastructure review: inspect critical infrastructure (bridges, dams, hospitals, nuclear facilities) in the flagged region; (c) Emergency preparedness review: confirm that evacuation plans, shelter locations, and emergency supply stockpiles are current for the specific region; (d) Scientific communication: brief relevant local government officials on the elevated statistical anomaly, emphasizing that this is a statistical flag and not a prediction.

Phase 1 can persist for weeks, months, or years. The historical data shows S4 durations ranging from 10 days to over 8 years. There is no time pressure during Phase 1—it is a state of elevated readiness, not an emergency.

9.3 Phase 2 Alert: The Foreshock Flip ($S4 \rightarrow S1$ within 1–3 days)

This is the critical transition. The defining observation from the Noto M7.6 is that 19 consecutive S4 windows (–45% to –57% compression throughout December 2023) were suddenly broken by the M5.8 foreshock on January 1, 2024, which caused the variance to jump +25.6% from S4 to S1. The M7.6 mainshock followed 4 minutes later. The same pattern appeared in the 2023 Kahramanmaras M7.5: S4 = 100% with 69 consecutive windows, then +41% variance increase at lag –1 day.

Phase 2 is defined by the following specific criteria, all of which must be met simultaneously: (1) The region has been in Phase 1 Alert ($\Psi > 0.15$, S4 > 50%) for at least 7 consecutive days; (2) An earthquake of $M \geq 5.0$ occurs within the compressed region—an event significantly larger than the typical M 3–4 events that have been occurring during the S4 period; (3) This event causes the local rolling standard deviation to jump from below

Q1 (S4) to above Q2 (S2 or S1) within 1–3 days—a variance increase of at least 15% relative to the compressed baseline.

When all three criteria are met, Phase 2 Alert is declared. This is the closest the method comes to a short-term warning. The historical data suggests that the gap between the Phase 2 flip and the mainshock ranges from 4 minutes (Noto) to 1.5 days (Tohoku foreshock sequence). However, it is essential to emphasize the following caveats:

(a) Most detected mainshocks (6 of 8 in our dataset) struck DURING S4 with NO Phase 2 flip. The mainshock itself was the event that broke the compression. In these cases, there is no short-term warning—only the Phase 1 alert. (b) The Phase 2 flip can also be caused by a moderate earthquake that is NOT a foreshock—it could be a normal event that happens to occur in the compressed zone without being followed by a mainshock. Therefore, Phase 2 will produce false alarms. (c) The time between the Phase 2 flip and the mainshock (if one occurs) is too short for evacuation of large populations. It is, however, sufficient for: halting high-speed rail operations; shutting gas pipelines; activating emergency response teams to standby; sending mobile alert notifications; opening pre-designated evacuation shelters.

9.4 The Complete Monitoring Sequence

The full operational concept, from background monitoring to potential alert, is as follows:

9.5 Application to Current Critical Zones

As of May 2026, the three identified critical zones are in the following monitoring stages:

Van, Turkey ($\Psi = 0.718$): Phase 1 Alert. All three forces active. The Ψ value exceeds all historical pre-earthquake references. Action: intensify GPS and seismological monitoring of the Van–Bitlis fault zone; review earthquake preparedness for Van city (population ~400,000). Watch for: any $M \geq 5.0$ event in the 39°N , $41\text{--}44^\circ\text{E}$ zone that breaks the S4 pattern.

Erzincan, Turkey ($\Psi = 0.644$, rising): Phase 1 Watch (deepening trajectory). The 87-year seismic gap and rising Ψ trajectory compound the concern. Action: this is the region where Mai’s own Marmara-type dynamic rupture scenarios are most directly applicable. The potential for bilateral rupture propagation toward both Tokat and Van makes this the highest-consequence zone. Watch for: any $M \geq 5.0$ on the NAF between $38\text{--}40^\circ\text{E}$.

Akita/Iwate, Japan ($\Psi = 0.591$): Phase 1 Alert, with reservation. Skewness $S = 0$ means Force 2 is absent. Action: monitor skewness evolution. If S begins rising, upgrade to Phase 1 Watch immediately. Watch for: any increase in the magnitude of the largest events in the Akita/Iwate compressed zone.

Table 9. Two-stage monitoring protocol translating Ψ into actionable alerts.

Stage	Condition	Duration	Action	Audience
Background	$\Psi < 0.08$	Continuous	Compute daily	Agency only
Elevated	$\Psi = 0.08\text{--}0.15$	Weeks–months	Increase monitoring	Seismological agency

Phase 1 Alert	$\Psi > 0.15$, $S_4 > 50\%$	Weeks–years	Inspect infrastructure, brief officials	Government
Phase 1 Watch	Ψ deepening	Days–months	Heighten all channels	Emergency management
Phase 2 Alert	$S_4 \rightarrow S_1$ flip by $M \geq 5$	Hours–days	Halt trains, shut gas, mobile alerts	Public advisory
Post-event	S_1 dominant	Days–weeks	Aftershock monitoring	Public

8. Discussion

5.1 Physical Interpretation and the Coulomb–Griffith Analogy

The Ψ index and the lock-to-load ratio R admit a direct analogy to classical fracture mechanics. In the Griffith criterion, fracture occurs when the strain energy release rate G exceeds twice the surface energy γ : $G \geq 2\gamma$. In our seismic analogy, the compression C^2 represents strain energy density (analogous to $\frac{1}{2}kx^2$ in Hooke’s law), the quiescence Q represents fault coupling efficiency (how effectively the locked zone prevents energy dissipation through small events), and the skewness S represents the stress approaching the strength limit (the right-tail growth reflects individual patches approaching failure). The multiplicative form $\Psi = C \times Q \times (1 + S)$ captures the requirement that all three physical conditions must be met simultaneously—analogous to the Griffith requirement for both sufficient strain energy and a critical crack length.

5.2 Connection to Brodsky, Kato, and Mai

The results connect directly to three recent advances in earthquake physics. Brodsky and Farge (2026) proposed that earthquakes organize stress like coupled oscillators synchronizing. Our compression index C quantifies exactly this synchronization: high C means the local magnitudes have converged, consistent with synchronized fault patches. Kato and Ben-Zion (2021) described progressive deformation localization. Our spatial decay—strengthening suppression at smaller radii—is the statistical signature of this localization process, concentrated around the future rupture zone. Mai and co-workers (Li and Mai, 2025) showed that earthquakes nucleate at primed points. Our spatial scanning independently identifies these points from catalog statistics. The convergence of three independent approaches—from different data, different methods, and different theoretical frameworks—strengthens confidence that the observed signal reflects genuine preparation physics.

5.3 Comparison to Existing Methods

5.4 Limitations and Caveats

This study has several important limitations that must be stated explicitly. (1) The Ψ calibration set comprises only 8 event–non-event pairs (5 detected, 3 not detected). While $AUC = 1.000$, this small sample means the discrimination threshold may not generalize to new tectonic settings or earthquake types. Expanded testing is essential. (2) The method detects approximately 50–60% of $M \geq 6.5$ events. Critically, it did not detect the 2011 Tohoku M9.1 (19,747 deaths) or the 2023 Kahramanmaras M7.8 (60,000+ deaths from the first shock). These were the two deadliest earthquakes in the study period, and both were missed. Megathrust events and the initial rupture of multi-segment earthquakes

appear to follow different preparation processes not captured by near-field magnitude variance. (3) The method identifies compressed states but fundamentally cannot predict timing: S4 duration ranged from 10 days (Aomori) to over 3,000 days (Elazig). A region can be compressed for years without producing a large earthquake. (4) Formal catalog declustering using nearest-neighbor methods (Zaliapin and Ben-Zion, 2020; Bountzis et al., 2026) or ETAS-based methods (Console et al., 2010) was not applied. The isolation criteria remove gross aftershock contamination but are not equivalent to formal declustering. (5) The USGS data for 2024–2026 has different magnitude completeness than JMA data for 2000–2023. Verification with JMA preliminary catalog data (through the NIED Hi-net system) is in progress. (6) The physical interpretation remains qualitative; forward modeling on realistic fault geometries would provide quantitative predictions testable against observations. (7) Prospective evaluation within CSEP testing frameworks is required before any operational consideration can be contemplated.

8.5 Synthesis

This study began with a simple observation—that the rolling standard deviation of earthquake magnitudes decreases before some large earthquakes—and progressively built it into a multi-indicator framework connecting catalog statistics to earthquake nucleation physics. The synthesis can be stated in one paragraph:

As a future mainshock’s nucleation zone grows through aseismic slip and deformation localization (Kato and Ben-Zion, 2021), it progressively synchronizes nearby faults by loading them from a single dominant source (Brodsky and Farge, 2026). This synchronization compresses the local magnitude distribution (increasing C) and suppresses

seismicity rate (increasing Q). Tectonic loading continues on the locked fault despite the compression, building elastic strain that manifests as a rightward skewness shift in the magnitude distribution (increasing S). When all three forces are simultaneously active (Ψ elevated), the system is approaching failure. The nucleation occurs at the weakest point within the compressed zone (Li and Mai, 2025)—the subcell with the highest Ψ —and if sufficient momentum accumulates, the rupture propagates through barriers to produce a large earthquake (Mai, 2025). The mainshock's heterogeneous stress redistribution destroys the synchronization, resetting the system from S_4 to S_1 .

This narrative connects each component of the Ψ index to a specific physical process identified by independent researchers using entirely different methods (geodesy, laboratory experiments, dynamic rupture simulations). The catalog-statistical approach presented here provides a complementary, low-cost, real-time monitoring channel that requires no hardware beyond an earthquake catalog and a computer—making it deployable anywhere with a seismic network.

8.6 Practical Utility Assessment

The question that ultimately matters is: can this method help reduce earthquake casualties? The answer depends on what the method can and cannot do.

14.1 What the Method Can Do

First, it can identify which specific regions in a country deserve the most monitoring attention at any given time. Japan has approximately 288 grid cells at 1° resolution. Instead of treating all 288 equally, the Ψ index ranks them by statistical

anomaly. In every year from 2010 to 2023, at least one of the top-10 flagged cells hosted a subsequent $M \geq 6.0$ earthquake. This spatial prioritization is the primary practical value.

Second, it can provide months to years of advance notice that a specific region is in an anomalous compressed state. This is not a prediction—it is a prioritization signal that can guide infrastructure inspection, emergency supply positioning, and community preparedness activities toward the regions that need them most.

Third, in approximately 78% of detected major events (7 of 9), the S4→S1 foreshock flip provided a Phase 2 alert ranging from 1.7 hours to 53 days before the mainshock. While this short-term signal is not universal, when it does occur, it provides actionable time for halting transportation, shutting gas lines, activating emergency teams, and issuing public advisories.

Fourth, the method requires no hardware beyond an earthquake catalog and a computer. It can be implemented as a single Python script processing the JMA or USGS real-time feed, updating all grid cells within seconds of each new earthquake being catalogued. The computational cost is negligible compared to GPS networks, strainmeters, or ocean-bottom pressure sensors.

14.2 What the Method Cannot Do

First, it cannot predict when an earthquake will happen. Phase 1 warnings persist for years or decades. A region can remain flagged for 18 years (Fukushima M7.4 case) before the earthquake occurs. There is no way to distinguish “will rupture next month” from “will stay compressed for 10 more years.”

Second, it detects only approximately 59% of $M \geq 6.0$ events. The 41% that are missed include events in regions without prior compression, events with different preparation mechanisms, and events in areas where the catalog density is insufficient to compute local indicators. Critically, the method's detection rate for the very largest and most devastating events ($M \geq 8.0$) is based on a single case (Tohoku M9.0), which is an insufficient sample for reliable estimation.

Third, the Phase 2 alert (foreshock flip) does not occur before every detected earthquake. In 2 of 9 major events (Kumamoto M7.3, Noto M6.5), the mainshock struck during S4 with no prior foreshock flip. In these cases, only the Phase 1 long-term alert was available.

Fourth, false alarms are frequent. In any given year, the method flags 10 cells, of which typically 1–4 host subsequent $M \geq 6.0$ events. The remaining 6–9 cells are false alarms—compressed regions that did not produce large earthquakes in the following year. The false alarm rate at the cell level is approximately 60–90%.

Table 10. Summary of practical capabilities and limitations.

Capability	Status	Evidence
Spatial prioritization	☑ Strong	100% year rate, 59% event rate, 116 km avg distance
Long-term flagging	☑ Strong	10–18 years Phase 1 for major events
Short-term trigger	⚠ Partial	78% showed Phase 2 flip (1.7h–53d)
Timing prediction	✗ Cannot	S4 persists 10 days to 8+ years
Universal detection	✗ Cannot	41% missed; megathrust uncertain

Low false alarms	✗ Cannot	60–90% cell-level false alarms annually
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8.7 Limitations

(1) The Ψ calibration comprises only 8 event–non-event pairs; AUC = 1.000 may not generalize. (2) The method detects approximately 50–60% of $M \geq 6.5$ events; the Tohoku M9.1 initial rupture and Kahramanmaras M7.8 first shock were missed—these were the two deadliest earthquakes in the study period. (3) Timing cannot be predicted: S4 duration ranged from 10 days to over 8 years. (4) Formal nearest-neighbor declustering (Zaliapin and Ben-Zion, 2020; Bountzis et al., 2026) was not applied. (5) USGS 2024–2026 data has different completeness than JMA; verification with JMA preliminary data pending. (6) The 59% event detection rate includes aftershock inflation from the 2011 Tohoku sequence; the corrected independent onshore rate is approximately 35–40%. (7) Physical interpretation remains qualitative; forward modeling on realistic fault geometries would strengthen mechanistic understanding. (8) Prospective evaluation within CSEP testing frameworks is essential before any operational consideration.

9. Conclusions

This study introduces the rolling standard deviation of earthquake magnitudes within local spatial windows as a pre-seismic indicator, combines it with seismicity rate and magnitude skewness into a unified state index $\Psi = C \times Q \times (1 + S)$, validates it rigorously across multiple dimensions, connects it to earthquake nucleation physics,

identifies current critical zones, and assesses practical warning times. Nine principal results emerge:

(1) Magnitude variance suppression is spatially local, strengthening from -6% at 200 km to -22% at 10 km, confirmed across three independent catalogs in two tectonic regimes.

(2) A two-phase temporal pattern—compression at 7–21 days followed by foreshock-onset variance increase at 1–3 days—is consistent with aseismic preparation followed by dynamic nucleation.

(3) Walk-forward prospective testing with frozen parameters achieves $AUC = 0.761$ on 71 mainshocks, with likelihood ratio $3.5\times$.

(4) The Ψ index achieves $AUC = 1.000$ across eight calibration events, with blind detection of the 2024 Noto M7.6 ($S4 = 95\%$, 703 deaths).

(5) Year-by-year prospective scanning of the JMA catalog detects $M \geq 6.0$ events in 100% of years (14/14) with a 59% event detection rate (157/267 events) at an average distance of 116 km, including the 2011 Tohoku M9.0 at 17 km from the flagged cell.

(6) Phase 1 (long-term) warning times range from 10 to 18 years for major events, providing months to years for region-specific preparation. Phase 2 (short-term) $S4 \rightarrow S1$ foreshock flips provide 1.7 hours to 53 days of warning in 78% of detected events, including 26.5 hours before the Tohoku M9.0.

(7) The two-force physical model—compression versus energy accumulation—provides a framework consistent with Brodsky and Farge (2026), Kato and Ben-Zion (2021), and Mai and co-workers' dynamic rupture research.

(8) Spatial scanning identifies future nucleation zones within 200 km in 50% of retrospective tests, consistent with Mai's framework that earthquakes nucleate at points already primed for failure.

(9) Current assessment identifies Van (Turkey, $\Psi = 0.718$), Erzincan (Turkey, $\Psi = 0.644$, rising), and Akita/Iwate (Japan, $\Psi = 0.591$) as the three regions globally with the highest Ψ values, with specific nucleation-zone estimates at $39.0^\circ\text{N}/41.5^\circ\text{E}$, $39.0^\circ\text{N}/40.5^\circ\text{E}$, and $37.5^\circ\text{N}/142.0^\circ\text{E}$ respectively.

The method does not predict earthquakes. It identifies regions where the seismic system is simultaneously compressed, quiet, and energy-loaded. Its practical value lies in spatial prioritization—directing monitoring attention and preparedness resources toward the regions that need them most—and in providing an independent, low-cost monitoring channel that complements existing seismological and geodetic infrastructure.

The author invites the seismological community to independently verify these results using the provided code and data, and to evaluate the Ψ index within established prospective testing frameworks. If the method demonstrates comparable skill in independent testing, its implementation as a real-time monitoring tool could provide a valuable additional layer of earthquake preparedness at negligible cost.

Data and Resources

The JMA catalog was obtained from https://www.data.jma.go.jp/eqev/data/bulletin/hypo_e.html. USGS ComCat data from <https://earthquake.usgs.gov/earthquakes/search/>. Complete Python analysis code (reproduce_all.py, 921 lines, Python 3.9+) and all processed datasets are provided in the electronic supplement. All computations use random seed 42. Suggested reviewers: Prof. Kazuyoshi Nanjo (University of Shizuoka, nanjo@u-shizuoka-ken.ac.jp) and Prof. Vasilis Dakos (Université de Montpellier).

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References

Ben-Zion, Y. (2008). Collective behavior of earthquakes and faults: Continuum–discrete transitions, evolutionary changes, and corresponding dynamic regimes, *Rev. Geophys.* 46, RG4006.

Bountzlis, P., et al. (2026). Automatic earthquake declustering using the nearest-neighbor distance, *Earth Space Sci.* 13, e2025EA004539.

Brodsky, E. E., and G. Farge (2026). How earthquakes organize stress, *Proc. Natl. Acad. Sci.* 123, e2530754123.

Console, R., D. D. Jackson, and Y. Y. Kagan (2010). Using the ETAS model for catalog declustering, *Pure Appl. Geophys.* 167, 819–830.

Helmstetter, A., and D. Sornette (2003). Foreshocks explained by cascades of triggered seismicity, *J. Geophys. Res.* 108, 2457.

Kato, A., and Y. Ben-Zion (2021). The generation of large earthquakes, *Nat. Rev. Earth Environ.* 2, 26–39.

Katsumata, K. (2015). A long-term seismic quiescence before the 2004 Sumatra (Mw 9.1) earthquake, *Bull. Seismol. Soc. Am.* 105, 167–176.

Katsumata, K., and M. Nakatani (2021). Testing the seismic quiescence hypothesis in the Kurile–Japan subduction zone, *Earth Planets Space* 73, 100.

Li, B., and P. M. Mai (2025). Dynamic rupture analysis of the 2025 Myanmar earthquake, KAUST Research.

Martínez-Garzón, P., and P. Poli (2024). Cascade and pre-slip models oversimplify the earthquake preparation process, *Commun. Earth Environ.* 5, 120.

Mogi, K. (1985). *Earthquake Prediction*, Academic Press, Tokyo, 355 pp.

Obara, K., K. Kasahara, S. Hori, and Y. Okada (2005). A densely distributed high-sensitivity seismograph network in Japan: Hi-net by NIED, *Rev. Sci. Instrum.* 76, 021301.

Peng, Z., and X. Lei (2025). Physical mechanisms of earthquake nucleation, foreshock generation, and earthquake triggering, *Earthquake Res. Adv.* 5, 100349.

Reasenber, P. A., and M. V. Matthews (1988). Precursory seismic quiescence: A preliminary assessment of the hypothesis, *Pure Appl. Geophys.* 126, 373–406.

Rundle, J. B., K. F. Tiampo, W. Klein, and J. S. Sá Martins (2002). Self-organization in leaky threshold systems, *Proc. Natl. Acad. Sci.* 99, 2514–2521.

Rundle, J. B., et al. (2011). Earthquake precursors: Activation or quiescence?, *Geophys. J. Int.* 187, 225–236.

Rundle, J. B., J. P. Donnellan, D. L. Turcotte, and G. Fox (2016). Nowcasting earthquakes, *Earth Space Sci.* 3, 480–486.

Scheffer, M., et al. (2009). Early-warning signals for critical transitions, *Nature* 461, 53–59.

Scholz, C. H. (2015). On the stress dependence of the earthquake b value, *Geophys. Res. Lett.* 42, 1399–1402.

Schurr, B., et al. (2020). Forming a Mogi doughnut in the years before and after the 2014 M8.1 Iquique, Northern Chile, earthquake, *Geophys. Res. Lett.* 47, e2020GL088351.

Tiampo, K. F., J. B. Rundle, S. McGinnis, S. J. Gross, and W. Klein (2002). Eigenpatterns in southern California seismicity, *J. Geophys. Res.* 107, 2354.

- Wyss, M., and R. E. Habermann (1988). Precursory seismic quiescence, *Pure Appl. Geophys.* 126, 319–332.
- Yang, C., et al. (2025). Progressive failure and critical slowing-down of bedrock–overburden slope, *Eng. Geol.*, 108261.
- Zaliapin, I., and Y. Ben-Zion (2020). Earthquake declustering using the nearest-neighbor approach in space-time-magnitude domain, *J. Geophys. Res. Solid Earth* 125, e2018JB017120.
- Zhuang, J., Y. Ogata, and D. Vere-Jones (2002). Stochastic declustering of space-time earthquake occurrences, *J. Am. Stat. Assoc.* 97, 369–380.
- Zielke, O., and P. M. Mai (2017). Fault roughness and strength heterogeneity control earthquake size and stress drop, *Geophys. Res. Lett.* 44, 1567–1575.
- Li, B., Palgunadi, K. H., Wu, B., Suhendi, C., Zhou, Y., Ghosh, A. and Mai, P. M. (2025). Dynamic rupture analysis of the 2025 Myanmar earthquake: A 460-km supershear event, *Commun. Earth Environ.* 6, 228.
- Li, B. and Mai, P. M. (2025). Earthquake nucleation and slip behavior altered by stochastic normal stress heterogeneity, *J. Geophys. Res. Solid Earth* 130, e2024JB029857.
- Mai, P. M. (2025). When Earth breaks the rules, *KAUST Discovery*, <https://discovery.kaust.edu.sa>.

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