

Super El Niño can raise coastal extreme sea-level risk tenfold

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Abstract

El Niño-Southern Oscillation is a dominant mode of global climate variability with a range of societally relevant impacts. However, its effect on global coastal sea-level extremes remains poorly quantified, especially for very strong (“super”) El Niños, which are rare. Here we analyse global reanalyses, hindcasts, and reconstructions to quantify how El Niño alters extreme sea level frequencies through mean sea level, tide, surge, and wave processes with distinct geographic fingerprints. We find low-frequency mean sea-level anomalies are the primary source of increased risk in the tropical Pacific during El Niño events, but that wave, tide, and surge play important secondary roles. Cross-basin thermocline tilt produces sea-surface height anomalies that correspond to an order-of-magnitude increase in exceedance risk along the tropical east Pacific, with the response steepening for strong events. The coincidence of the projected 2026–2027 Super El Niño with local maxima of long-period astronomical tidal cycles may produce 99th-percentile extremes in up to half of all high waters at some sites with potential major flooding and ecological impacts—the background shift alone is equivalent to 40–50 years of projected sea-level rise under SSP1-2.6.

As the dominant mode of nonseasonal variation in global climate, El Niño–Southern Oscillation (ENSO) is associated with a wide range of well-documented, societally relevant impacts on marine ecosystems and fishery productivity [1], rainfall and drought [2, 3], monsoon suppression and amplification in the Indian Ocean [4], and inland flooding and landslides [5]. The year 2026–27 is projected to become the strongest El Niño on record [6], with a central estimate of $+2.74^{\circ}\text{C}$ on the Relative Oceanic Niño Index (RONI) [7, 8], which is 0.9 standard deviations (0.34°C) above the next largest historical event (1982–83). As such, the projected strength of this very-strong (“super”) El Niño has raised global concern over its impacts [9].

Less is understood about how ENSO affects coastal sea-level extremes, which arise from the combined effects of mean sea level, tides, surges, and waves. Mean sea level, surges, and waves all vary under ENSO [10–16], but whether total water level exceeds a local flood threshold depends on all components jointly. A background anomaly of a given size changes risk very differently depending on the local tide–surge–wave climate, and can be swamped where tides are large [13, 17]; a surge or wave event matters most when it coincides with high spring tides or an elevated background [18]. In California, for example, the record surges of 1998 fell on neap tides, whereas the January 1983 storms met the highest spring tides in four years [19]. Similarly, in 2015–16 the seasonal high-water peak arrived before the largest waves [15].

Prior work has nonetheless largely examined these components separately. To our knowledge, the contributions of mean sea level, tide–surge, and waves to the change in extreme-water-level risk have not been separated at the event level, and the coincidence of large waves with extreme still-water levels has been examined only in case studies [15, 19]. Most analyses also composite over events or regress on an ENSO index [11, 14, 17, 20], implicitly assuming that responses scale linearly with event magnitude (defined by sea surface temperature), an assumption least tested for the rare, very strong events of greatest concern. Finally, most evidence is local or regional, concentrated on the US West Coast [11, 12, 14, 15, 17–20] and Australia [21], where long records exist; elsewhere, observations of extreme coastal water levels and their relation to ENSO are sparse [10]. The few global assessments were conducted at annual resolution and found that tide-dominated total water levels are largely insensitive to ENSO [13], focus on interactions between climate modes [22], or rely on satellite shoreline proxies [23–25]. It therefore remains unclear how much ENSO amplifies or reduces coastal risk, which components drive the change and how they combine, and how this varies in space and with event magnitude: questions made particularly pressing by the developing 2026–27 super El Niño.

To address these crucial gaps, we quantify ENSO impacts on sea-level extremes globally, extending the global approach of Muis et al. [13] with an event-level decomposition that jointly considers mean sea level, tide, surge, and wave effects. To do this, we use a low-frequency sea level reconstruction, a global high-frequency (tide and surge) hindcast and associated wave reanalysis. We combine monthly background sea level M from a hybrid tide–gauge reconstruction (1950–2015) [26], hourly tide-plus-surge water levels from a global tide and surge reanalysis (1950–2015) [27] and hourly ERA5 waves (1979–2015) [28] at 3,197 coastal sites around the world; with ENSO state from the Relative Oceanic Niño Index (RONI) [7, 8] and the equatorial ocean

state from satellite altimetry (Methods). Each high water is treated as a trial, and an extreme is a high water above z_{99} , the site’s 99th-percentile of high waters in neutral months. The risk ratio RR_{EE} is the probability that a high water exceeds z_{99} during El Niño relative to the same probability in neutral conditions, compared within the same calendar months, so that values above 1 indicate increased risk. To attribute the change, we form counterfactual states that pair El Niño tides and surges with a neutral background (RR_{EN}) and neutral tides and surges with the El Niño background (RR_{NE}), and split the change in risk, $RR_{EE} - RR_{NN}$, into background and tide-surge contributions, ϕ_M and ϕ_X , using Shapley values [29]; $f_M = \phi_M / (RR_{EE} - RR_{NN})$ is the share due to the background. Wave effects are measured by RR_{HW} , the El Niño change in the probability that an extreme high water is accompanied by large wave forcing. Finally, we project the background shift for 2026–27 from its site-wise relationship with RONI and convert it, together with the astronomical tide (nodal and perigean cycles), into a conditional risk ratio. Uncertainties are 95% intervals from bootstrap resampling of whole ENSO events. We focus here on El Niño, but the methods employed are similarly applicable to La Niña and other climate variability.

El Niño reshapes coastal extreme risk

We find that the typical El Niño raises the risk of sea-level extremes along the eastern boundary of the Pacific Ocean by increasing background sea level by ~ 5 cm. When coupled to tides and surges, the background M increases the risk ratio RR_{EE} of high-water levels above the 99th-percentile of neutral-condition water levels (z_{99}) more than fivefold at some sites in the tropical east Pacific (regional median 2.33 along Mexico–Ecuador), with a corresponding reduction in risk around the Maritime Continent. Across the coastal sites, RR_{EE} is significantly greater than 1 at 1,264 of 3,197 sites and significantly less than 1 at 316 (Fig. 1b). The increase spans the Pacific coast of the Americas: north of Mexico–Ecuador, Alaska–California has a regional median of 1.55 (significant at 90% of sites), about 1.7 between 20 and 55°N and falling to 1.1 poleward of 62°N; to the south, the Peru–Chile median of 1.70 falls from 3.3 north of 20°S to 1.4 south of 35°S. Smaller increases occur in the Caribbean (1.22), the Gulf of Mexico (1.17) and eastern North America (1.20), where fewer sites are individually resolved (27%, 8% and 18%). Decreases are concentrated in the western Pacific and Maritime Continent (median 0.53; significant at 59% of sites), the north Indian Ocean (0.61) and southern and eastern Australia (0.84), mirroring the east–west dipole of the background anomaly (Fig. 1a). Elsewhere, including the northeast Atlantic, the Mediterranean and Africa, regional medians lie between 0.9 and 1.2 and few sites are resolved.

Background sea level drives most risk

Background sea level changes are the primary means through which El Niño changes the risk of sea-level extremes. We quantify the fractional contributions of background sea-level anomalies M and the sum of tides plus surges X to the change in exceedance risk of still water level H with the associated Shapley values ϕ_M and ϕ_X (Methods). Mexico–Ecuador has the largest contribution from background sea level variability

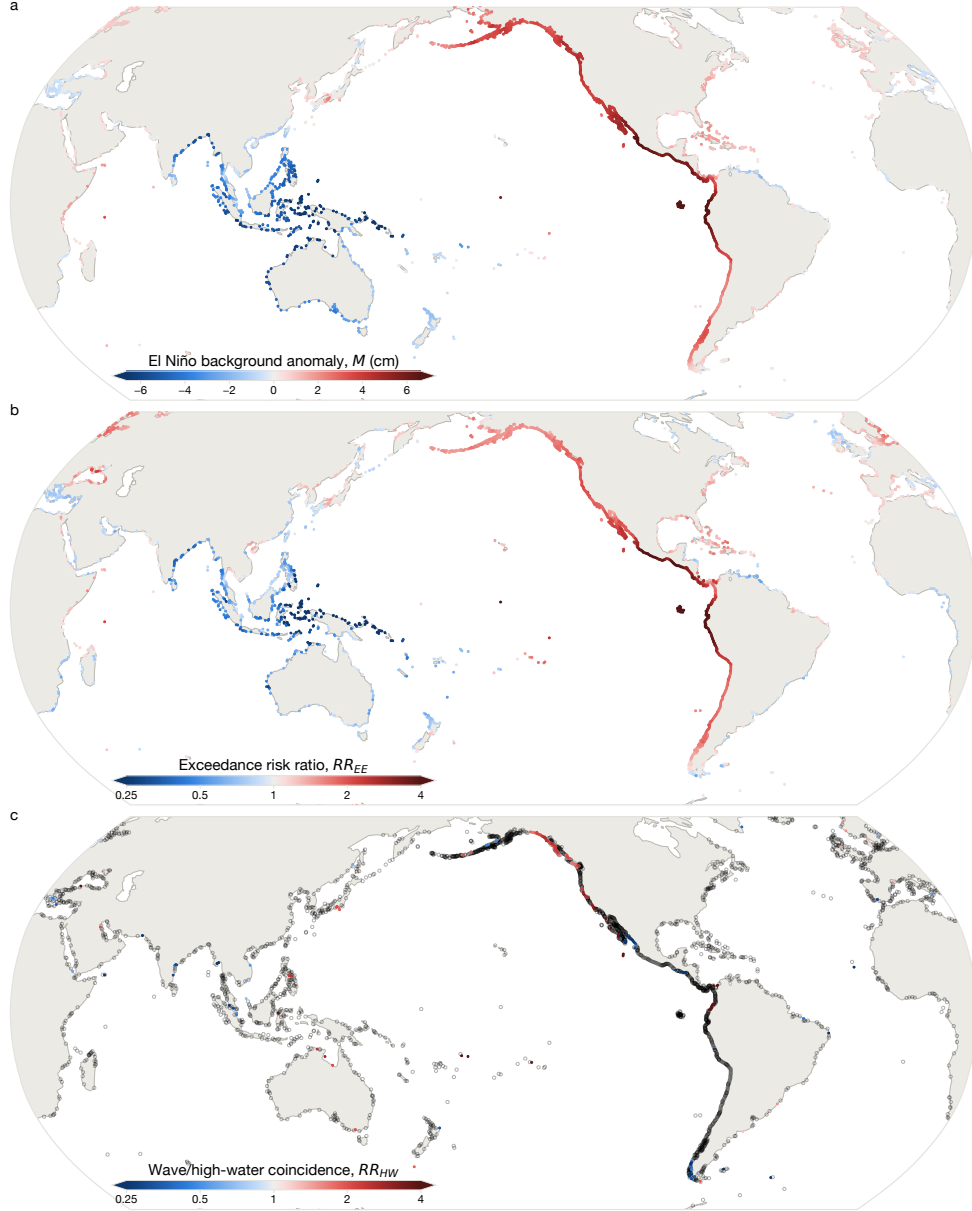


Fig. 1 ENSO leaves multiple, geographically distinct fingerprints on coastal hazards. a) Composite anomaly of the slowly varying background sea level M during El Niño months relative to calendar-month-matched neutral months, 3,197 coastal sites. **b)** Risk ratio RR_{EE} for exceedance of the site-specific 99th-percentile high water, El Niño against month-matched neutral, on the neutral full-state threshold $z_{99} = P_{99}(X_N + M_N)$. **c)** Wave/high-water coincidence RR_{HW} ; the El Niño-to-neutral ratio of the probability that an extreme high water ($H > z_{99}$, the threshold of b) is accompanied by a large wave state $W > W_{95}$ within ± 6 h, at 2,943 matched GTSM and ERA5 wave sites, 1979–2015. Filled points are sites whose event-block bootstrap interval excludes 1, and at least one El Niño extreme high water is accompanied by large waves; open grey points are unresolved sites.

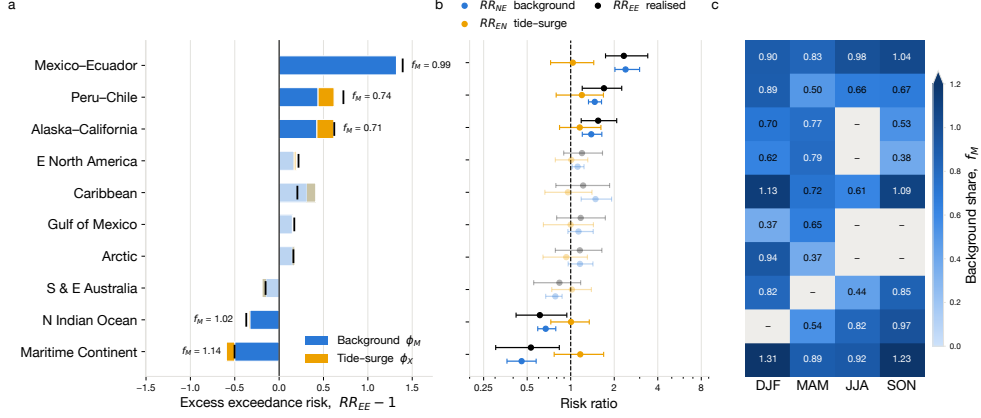


Fig. 2 Attribution of background and tide-surge sea level variability on coastal extreme risk. **a)** Exact two-player Shapley decomposition of the excess exceedance risk into a background share ϕ_M and a tide-surge share ϕ_X . Bar length is the total effect and the internal division is the attribution; black ticks mark the total. The Shapley decomposition closes exactly at each site. Small differences between the summed regional-median contributions ($\phi_M + \phi_X$) and the regional-median total (black ticks) arise because medians are taken separately for each quantity. Regions are medians over their sites, ordered by total effect. Regions are shown where the median $|RR_{EE} - RR_{NN}| \geq 0.15$; muted regions are where RR_{EE} interval includes 1. **b)**, The four counterfactual states behind that split: RR_{NE} pairs El Niño background with neutral weather, RR_{EN} pairs neutral background with El Niño weather, and RR_{EE} is the realised state. Points are regional medians; bars are 95% event-block bootstrap intervals over whole ENSO events and single neutral years. Donor-consistent pairing is used throughout, so the four states share one estimator and the residual interaction term is -0.005. **c)**, The same background share $f_M = \phi_M / (RR_{EE} - RR_{NN})$ computed season by season. Blank cells are seasons in which the regional median $|RR_{EE} - RR_{NN}| < 0.15$.

with a resolved increase, $f_M = 0.99$. Here, background variability raises the risk ratio to 2.39 for typical El Niño events. This El Niño sea level fingerprint (Fig. 1a) is well established and well understood [30] (Supplementary Note 2). Weakened trade winds trigger eastward-propagating Kelvin waves and westward-propagating Rossby waves, which together relax the mean equatorial thermocline and sea-surface height gradient [30, 31]. After the Kelvin waves reach the eastern boundaries, associated anomalies progress poleward as coastally trapped waves, effecting sustained coastal sea-level anomalies along the west coast of the Americas [32, 33].

While mean sea-level anomalies M make primary contributions to increased risk during El Niño events, the effects of tides plus surges X remain important in some areas. At higher latitudes along Pacific coastlines (e.g., Alaska-California), tides and surges account for about 30% of the change ($f_M = 0.71$). This is likely a consequence of an El Niño-deepened Aleutian Low [34] that drives storm surges and wave events (Fig. 1c), which is reflected by sustained $f_M = 0.69$ north of 55°N , consistent with prior work [12, 35]. A different spatial structure characterizes Peru-Chile: its northern coastline mirrors the M -dominated tropics, while its mid-latitude southern sites lie in a westerly belt largely insulated from El Niño, diluting the regional median.

El Niño effects on risks of sea-level extremes extend beyond the Pacific. The Caribbean shows a modest increase in risk ratio ($RR_{EE} = 1.22$) for a small change

in background anomaly, although, as in the Gulf of Mexico, the regional total is not resolved by the bootstrapped intervals. More generally, the response in the Caribbean and Gulf of Mexico is presumably mediated by atmospheric teleconnections [36, 37], with signals lagging the RONI signal by 2 to 3 months when estimated by cross-correlation. Despite this common mechanism, their risk profiles differ based on local surge climates and seasonality (Fig. 2c). The Gulf of Mexico is modulated by extratropical winter storms, yielding a low background share ($f_M = 0.61$) strongly influenced by atmospheric forcing. By contrast, the Caribbean appears background-dominated ($f_M = 1.15$, the tide-surge contribution slightly offsetting the background). This high f_M is produced by a quiet local surge climate, likely reflecting a suppression of tropical cyclone formation [38], an underrepresentation of such variability in the tide-surge reanalysis (Methods), or both. With little competing storm variability, even modest mean sea-level anomalies dictate the extreme tail.

Waves add a distinct hazard

Superimposed on still water levels, wave effects also make distinct contributions to heightened risk of extremes during El Niño. The coincidence risk ratio RR_{HW} (Methods) measures whether El Niño changes the share of extreme high waters that are accompanied by large wave forcing, separately from the change in how often extremes occur, and it shows a different geographical footprint from M (Fig. 1c). El Niño raises it along the northeast Pacific from California to British Columbia (US West Coast median 1.85, with resolved increases at 24% of sites). It lowers it across the north Indian Ocean (0.54), around the Maritime Continent (0.89) and along the coasts of Africa, where many sites record no El Niño extreme high water accompanied by large waves. Where El Niño raises the background strongly, as along Mexico–Ecuador (0.94), extreme high waters need less help from waves, so the share accompanied by large waves need not rise even though wave intensity does (Supplementary Note 10). Thus, an unchanged wave–high-water coincidence does not imply unchanged wave forcing: elevated background sea level allows extreme high waters to occur across a broader range of wave conditions. No statistically significant change is observed for the east coast of Australia.

Small Island Developing States in the Pacific are particularly vulnerable to ENSO-related risk [39, 40]. At Kiritimati in the Line Islands, El Niño raises extreme-water risk almost sixfold ($RR_{EE} = 5.7$), but through the background alone: the share of extreme high waters accompanied by large wave forcing is unchanged ($RR_{HW} = 0.98$ [0.43, 2.35]; 9.1% in neutral conditions). Among islands, resolved increases lie instead at the edge of the tropics and in the storm belt: in the Revillagigedo Islands off Mexico the share rises from 4–5% to 16% ($RR_{HW} = 3.2$ – 3.8), and in the Aleutians from 18–26% to 34–45% (1.7–2.0). Beyond the Pacific, few increases are resolved, and many sites in the north Indian Ocean and along African coasts record no El Niño extreme accompanied by large waves, suggesting teleconnections associated with El Niño [41, 42]. Similar results are obtained with significant wave-height H_s in place of wave run-up forcing (Supplementary Fig. S12), indicating that our conclusions are robust to the quantification of wave effects.

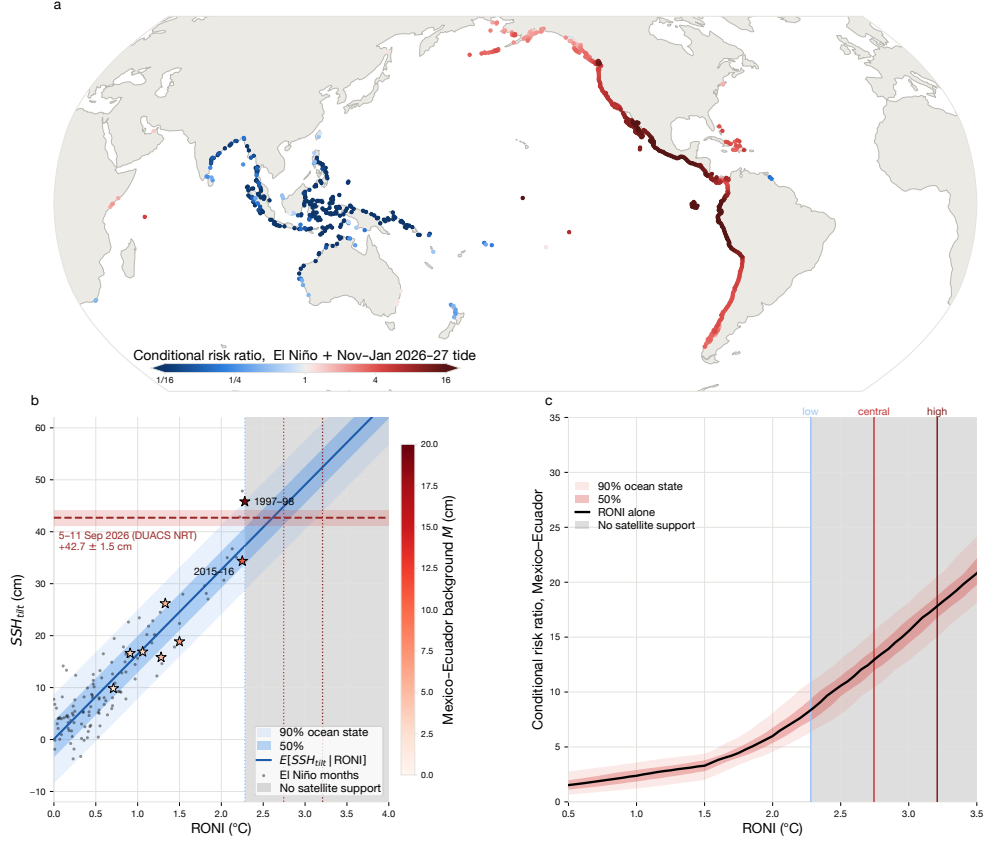


Fig. 3 Conditional risk outlook for projected 2026–27 Super El Niño. **a)** Conditional risk ratio of 99th-percentile extremes, factoring in both ENSO-driven background sea level and the astronomical tide for November 2026–January 2027 (nodal and perigean cycles) with mean ocean state (issuable sites only). **b)** Fitted and extrapolated linear relationship between RONI and SSH_{tilt} with key events overlaid; stars mark the eight satellite-era El Niños from 1993–2015 at their RONI maximum, coloured by the Mexico–Ecuador background M . **c)** Conditional risk ratio for Mexico–Ecuador conditioned on RONI and SSH_{tilt} . Shaded bands in **b)** and **c)** are the 50% and 90% ranges of SSH_{tilt} at fixed RONI, estimated at the event scale from the 1993–2015 record; they describe ocean-state variability, not confidence intervals. Grey shading marks amplitudes beyond satellite support, vertical lines the low, central and high 2026–27 scenarios, and the dashed line in **b)** the 7-day mean of near-real-time tilt for 5–11 September 2026. The band, ± 1.5 cm, is the standard deviation of 7-day means about their own monthly mean over 1993–2025, which scales a weekly value against the monthly points plotted.

Nodal and perigean tidal cycles and sea-level rise

Superimposed on ENSO effects are other long-period signals, including sea-level rise and interannual astronomical tidal cycles, that can further exacerbate risk. Sea-level rise across the east Pacific rim has increased exceedance risk by a factor of 3.07 between 1950 and 2015. Astronomical tides contribute through the 18.6-year lunar nodal cycle and the 8.85-year perigean cycle [43], whose 4.4-year half-period modulates high tides

along the east Pacific rim at least as strongly as the nodal cycle (median amplitudes of the November–January 99th-percentile high water, 3.0 and 3.6 cm). Together they raise risk over the same coastline by a factor of 1.47 (median over E Pacific rim) when averaged over November 2026–January 2027 relative to a cycle-mean tide.

The coincidence of El Niño during different phases of the nodal and perigean cycle can amplify or offset the effect of the background anomaly M on risk [19]. Although it was one of the three strongest El Niños on record (RONI +2.25 °C), the 2015–16 El Niño peaked near a minimum of the astronomical tide along much of the east Pacific rim, which on its own lowered risk along the US West Coast by about 40% (tidal risk ratio 0.57; Supplementary Fig. S4). In contrast, 2026–27 peaks near the maximum, which on its own roughly doubles risk over the same coast (1.95), a 3.4-fold swing between the two events. This varies along the rim: at Mexico–Ecuador the November–January 2026–27 astronomical anomaly is only +0.5 cm, so the change in risk there is almost entirely due to El Niño.

Outlook for the 2026–27 event

The risk of sea-level extremes is projected to increase dramatically along the eastern boundary of the Pacific Ocean (Fig. 3a). We determine this by combining projections of the super El Niño of 2026–27 with predictions of the astronomical tide for November 2026–January 2027, when the event is expected to peak (Methods). The El Niño projection is conditioned on the central estimate for RONI, with the spread reflecting the variability from historical ocean states (Methods). The central scenario (+2.74 °C) and the low and high scenarios (+2.28 and +3.21 °C) are the 50th, 10th and 90th percentiles of the September 2026 NOAA Climate Prediction Center outlook for October–December 2026 [6, 44], which is issued on RONI and so enters the projection without conversion (Methods). El Niño background variability is projected by regressing M on RONI using a piecewise-linear (hinged) fit where steepening is resolved, and defaulting to a linear relationship where not. Across Mexico–Ecuador, the response steepens: the background gains 4.2 cm per °C of RONI below the +1.5 °C strong-event threshold and 11.2 cm per °C above it; a 2.6-fold increase in sensitivity (95% event-block interval 1.6–3.7). These projections are qualitatively robust to reasonable alternative modelling assumptions (Methods; Supplementary Fig. S10, Table S6), but show that extrapolating from moderate events underestimates the background response to strong events. Surge and wave effects are omitted here: wave intensity and coincidence rise with amplitude, but both carry predictive skill only along Mexico–Ecuador (Supplementary Note 10, Table S7). As such, risk is likely underestimated where tides and surges carry more of the change, such as Alaska–California ($f_M = 0.71$).

Coastal M anomalies are strongly tied to the cross-basin sea-surface height tilt (SSH_{tilt}) that reflects changes in the depth of the thermocline: over 1993–2015 the Mexico–Ecuador background correlates with SSH_{tilt} at $r = 0.92$, against $r = 0.87$ for RONI (Supplementary Note 5; Supplementary Figs. S5 and S13). While SSH_{tilt} is correlated with El Niño indices, such as RONI, similar temperature anomalies can yield drastically different cross-basin tilts, and thus different shifts in M (Fig. 3b): 1997–98

and 2015–16 reached almost the same RONI (2.28 and 2.25°C) but very different tilts (45.8 against 34.4 cm), and the Mexico–Ecuador background reached 18.5 cm in 1997–98 against 10.7 cm in 2015–16. Because operational forecasts are issued for RONI, our outlook is conditioned on RONI, and SSH_{tilt} enters as the range of ocean states observed at a given RONI (Fig. 3b-c; Methods). Near-real-time altimetry gives a 7-day mean SSH_{tilt} of 42.7 cm for 5–11 September 2026, the highest for that calendar week in the 34-year record (next: 34.1 cm in 1997). The August 2026 monthly mean, 35.7 cm, is likewise the highest August on record (2015: 28.3 cm). In all eight El Niños of that record (1993–2015) the tilt continued to rise after August and peaked with RONI, so this anomaly is expected to intensify further (Supplementary Note 5).

The central estimate of the conditional risk ratio for Mexico–Ecuador is the highest of any coarse region (defined in Fig. A1), reaching 12.9 (90% ocean-state range: 10.4–15.6), peaking at 17.8 (15.1–20.5) under the high scenario (Fig. 3c). Practically, this shifts the historical 99th-percentile high water from a rare occurrence (once every 52 days) to a recurrent hazard, exceeded every 4.0 days and, at the most affected sites, exceeded by up to half of all high waters at the anticipated peak. Conditioning the projection on the tilt observed in September 2026, rather than on its expected value at a given RONI, raises the central Mexico–Ecuador risk ratio from 12.9 to 15.2 (90% range 12.9–17.6; Supplementary Note 13, Table S9). Taken individually, Peru has the highest risk increase by country, with low, central, and high estimates of 13.3, 21.2, and 30.5. Analogous curves for all regions and along-shore subregions are provided in Supplementary Figs. S6–S7 and Table S2. Central estimates elsewhere on the rim are smaller but still elevated: 5.7 along the US West Coast, 3.3 in British Columbia, 1.9 in Alaska and 3.9 in Chile. Beyond the Pacific, the Caribbean reaches 4.1 (21 issuable sites), because its small tidal range lets a background rise of only 4 cm, together with the tidal term, raise risk severalfold, whereas around the Maritime Continent the neutral 99th-percentile high water would almost never be exceeded (central $RR \approx 0.01$ at the median site). Because extrapolation of this high-amplitude steepening contributes materially to the projected magnitude, we additionally repeat the outlook assuming a purely linear ENSO response: the central Mexico–Ecuador background anomaly decreases from 19.3 to 13.6 cm and the corresponding risk ratio from 12.9 to 7.9 (Supplementary Fig. S11).

While we focus on sea-level extremes, defined statistically in terms of high-water percentiles, El Niño-driven changes in mean sea levels, tides, surges, and waves are also apparent in flood reports (Supplementary Note 3). Using a recent flood news report database [45], we find the same east–west dipole in 13,528 coastal flood events from El Niño months: background sea level improves prediction of the flood day along Mexico–Ecuador and Alaska–California, but degrades it around the Maritime Continent and in the north Indian Ocean, where El Niño lowers sea level (Supplementary Note 3, Fig. S2–3). Waves and tide–surge effects show a relatively larger contribution to reported floods but with more varied fingerprints. Shifts in background M can also have large ecological impacts by changing the duration of tidal inundation [46], which alters intertidal ecosystem exposure [10, 11, 47]. Under the central projection, the mean duration of a high-water excursion above mean higher high water (MHHW) at Mexico–Ecuador increases by 36% (+0.76 h), and that of a low-water excursion below

mean lower low water (MLLW) decreases by 31% (-0.74 h). The western Pacific and Maritime Continent show the opposite effect: a 19% decrease above MHHW (-0.50 h) and a 19% increase below MLLW ($+0.50$ h) (Supplementary Fig. S8, Table S3).

Finally, it is instructive to relate ENSO-driven sea-level extremes to future conditions expected under sea-level rise [11, 15, 19, 20]. Translating the projected 2026–27 event background anomaly into an equivalent interval of local sea-level rise using IPCC AR6 projections [48, 49], we illustrate this compounding threat (Table A1). For Mexico–Ecuador under SSP1-2.6, and omitting the effects of vertical land motion, we find that the low, central, or high projections for the 2026–27 event are equivalent to 30, 40, or 51 years of baseline (linear) sea-level rise, respectively, which is 1.2, 1.6 or 2.1 times larger than the amount of sea-level rise that occurred along that region since 1950. Ultimately, while an El Niño event persists for only a single year, its background anomaly compresses decades of projected sea-level rise into a single season, offering a vivid, real-world preview of our coastal future.

These results highlight El Niño as a compound coastal sea level hazard whose magnitude cannot be fully inferred from surface warming alone. Across the tropical Pacific, the dominant control is the evolving equatorial ocean state, which can raise coastal background sea level enough to transform rare extremes into recurrent high waters. Tides, surges, and waves amplify or offset this response regionally. In 2026–27, an exceptional Pacific tilt could temporarily raise coastal sea level by as much as decades of future sea-level rise, while a favourable phase of the astronomical tide further compounds this risk, dramatically increasing the potential for coastal flooding. Incorporating ocean state alongside ENSO forecasts therefore offers a direct route from seasonal climate prediction to advance warning of coastal extremes.

1 Methods

1.1 Datasets

Three datasets are combined to capture background sea level variability M , tide–surge effects X , and waves W . These datasets are described below.

1.1.1 Background sea level M .

The background sea level M is taken from the hybrid sea level reconstruction of [26], updated in April 2022 and supplied on a coastal grid, and we refer interested readers to the original text for details. The monthly mean sea level reconstruction combines tide-gauge records with a probabilistic treatment of the large-scale field, using wind stress, sea level pressure, and sea surface temperature in gap filling. We take the variable with glacial isostatic adjustment removed but the *seasonal cycle retained*, monthly at 18,719 coastal points from 1900–2015, and map it to nearest point if within 1 km; otherwise inverse distance weighting over the four nearest, provided these lie within 100 km (383 of the 3,197 sites exceed 1 km and use the weighted fallback). The local long-term trend is then removed by Locally Weighted Scatterplot Smoothing (LOWESS) with a 30-year span so monthly through decadal variability is preserved, and M is centred on the 1950–2015 baseline mean. Retaining the seasonal cycle is what allows the composition in (4) to supply the steric and mass seasonal signal that the barotropic model does not carry; the seasonal-free variant of the same series is retained in parallel and is the one used by every regression on an ENSO index.

Note that M is reconstructed large-scale climatic sea level variability, combining sterodynamic, mass, fingerprint, and circulation contributions. No pressure correction is applied to M , and the amount of static pressure response it retains varies. Sensitivity testing (not included) found this variability to have negligible impacts on the results presented. The reconstruction ends in December 2015, during the 2015–16 El Niño, so that event is excluded from event-consistency and leave-one-out statistics by default. Its onset, to December 2015, is retained for the amplitude fits of Eq. 15 and in sensitivity tests.

1.1.2 Tide and surge data

Tide and surge data come from the GTSMv3.0 global tide-and-surge reanalysis configuration developed and made available by Aleksandrova et al. [27]. We here provide a condensed summary, but interested readers are referred to the original text for comprehensive implementation and benchmarking details.

The model is implemented in Delft3D Flexible Mesh [50] as a depth-averaged hydrodynamic model on an unstructured grid with spatial resolution reaching about 2.5 km along most global coasts and 1.25 km along the European coast. The model is forced with hourly ERA5 10 m winds and mean sea level pressure fields, includes a spatially varying annual mean sea level field, has no open boundaries, and represents tides through 60 tidal constituents, while storm surges arise from barometric pressure

and wind stress forcing using a Charnock drag formulation. Bathymetry is assembled from GEBCO globally, EMODnet in European waters, and Bedmap2 beneath Antarctic ice shelves. Here, we only use the model-estimated still water levels.

The reanalysis spans the period 1950–2024, and we make use of hourly water levels. While global coastal coverage is available, the 16,517 native coastal stations are thinned by requiring a minimum separation of 25 km and discarding sites further than 100 km from the nearest background sea level reconstruction M point (described above), which leaves 1,984 sites globally. Because the eastern Pacific carries the signal this study is about, and a 25 km thinning there discards the alongshore structure the coastal waveguide imposes, that basin alone is then restored to the model’s native station spacing. This yields 3,206 candidate sites, of which nine have no reconstruction point within 100 km and are dropped, leaving **3,197 sites**: the network used everywhere throughout this manuscript. The densification is confined to the eastern Pacific, so pooled global medians are weighted more heavily toward that basin than they would be on a uniformly thinned network, and regional values are reported throughout for that reason. A separate quality screen flags 53 sites (1.7%) whose decadal standard deviation is anomalous relative to their neighbours; these are retained but carry a flag, and the conditional outlook is provided only at the 1,209 sites that additionally pass three tests: (i) leave-one-El-Niño-out skill in predicting each held-out event’s exceedance rate is positive relative to neutral climatology ($1 - SSE_{ENSO}/SSE_{clim} > 0$; 2015–16 excluded); (ii) the 95% event-block interval on the background increment $\Delta RR_M = RR_H - RR_X$ excludes 0; (iii) the 95% interval on RR_H excludes 1. Of the 1,217 sites passing all three, 8 are dropped because the nearest GESLA gauge within 50 km shows an observed El Niño increment of opposite sign [51]; 275 of the remainder are corroborated by a gauge of the same sign and 934 have no gauge within 50 km.

The GTSM mean sea level trend is removed through the secular term T_j of Eq. 4. A known issue with the reanalysis is the underestimation of extremes due to tropical cyclones (TCs). This may mean that the risk ratios estimated from GTSM in regions prone to TCs do not fully capture this variability [52]. El Niño events tend to produce greater wind shear in the Atlantic, which hinders hurricane formation. As such, changes in hurricane-induced extremes may be underestimated in the Atlantic basin in the associated risk ratios. Note: neither the Gulf of Mexico nor the US East Coast were found to have significant excess exceedance risk – i.e., their bootstrap intervals on the excess risk contained zero (Fig. 2b). As such, no such risk was resolved in this analysis. In contrast, hurricane activity generally increases in the Northern tropical east Pacific which could lead the Mexico tide–surge share to be underestimated.

1.1.3 Wave data

Wave data are taken from the ERA5 reanalysis [28]: specifically, hourly significant wave height H_s and mean wave period T , for the period 1979–2015. ERA5 extends to 1940, but before 1979 it assimilates no satellite observations and its wind and wave fields are less constrained and less homogeneous. We therefore use 1979–2015, which ends with the background reconstruction and contains ten complete El Niño events and the onset of 2015–16. Each analysis site is matched to its nearest *valid* wave cell

within 100 km (median separation 23.5 km) rather than to its own rounded cell, which would fall on land or on cells the wave model does not resolve; the 254 sites with no valid cell inside that radius are excluded. Panel c of Fig. 1 therefore uses 2,943 sites drawn from 2,066 distinct cells; RR_{HW} is defined at 2,876 of them and its interval excludes 1 at 254. At a further 177 sites no El Nino extreme high water is accompanied by large waves ($RR_{HW} = 0$) – these sites are treated as unresolved. Wave forcing is the deep-water runup-forcing proxy

$$W = \sqrt{H_s L}, \quad L = \frac{gT^2}{2\pi}, \quad (1)$$

where H_s is the deep-water significant wave height and L is the deep-water wavelength, which is related to the mean wave period T [53]. Note that Eq. 1 is used without its foreshore-slope terms. Equation (1) is therefore a proxy for the forcing that drives runup, *not* an estimate of runup, as no beach slope is assumed anywhere. T is the mean rather than the peak period, so L is systematically underestimated by a factor that varies with spectral shape. ERA5 wave fields are undefined under sea ice, and those hours are excluded from the counts rather than scored as small waves — a distinction that matters only in Hudson Bay and Labrador, where coverage falls to about 69%. An ablation which only considers significant wave-height H_s and not the run-up proxy is provided in Supplementary Note 9 and Fig. S12 and shows similar behaviour.

1.1.4 Tropical-Pacific ocean state

The reconstruction above ends in 2015 and is coastal, so the equatorial ocean state is taken instead from DUACS L4 gridded altimetry [54], product `SEALEVEL_GLO_PHY_L4_MY_008_047` (doi.org/10.48670/moi-00148). This has a 0.125° spatial resolution and covers the period from 1993 to 2025. Three boxes are defined: SSH_E (5°S – 5°N , 160° – 80°W), SSH_W (5°S – 5°N , 120° – 160°E), and the index

$$SSH_{\text{tilt}} = SSH_E - SSH_W, \quad (2)$$

is positive when the east stands high relative to the west, i.e., a flattened thermocline. Processing matches the *seasonal-free* background exactly: the calendar-month climatology is removed, then the secular component is removed under the same LOWESS convention, so that an index and its target remain comparable and no apparent skill can be a trend artefact. This is the reason the two background series of the previous section are not interchangeable. The tilt carries no seasonal cycle, so any quantity regressed on it must use the seasonal-free background, while H itself takes the seasonal-retaining one. When estimating the joint distribution of RONI and SSH_{tilt} in Fig. 3, the conditional spread is quoted at the *event* scale rather than the month scale, because months within one event share an ocean state.

The present state is taken from the near-real-time DUACS stream (`cmems_obs-sl_glo_phy_ssh_nrt_allsat-l4-duacs-0.125deg_P1D`), which is the operational twin of the reprocessed product rather than the same data. Over their 565 overlapping days the two tilts correlate at $r = 0.993$ with an rmsd of 0.81 cm and

a standard deviation ratio of 1.002; the +0.52 cm mean offset is removed before the recent value is ranked against the reprocessed record. The frozen seasonal climatology is reused and the LOWESS secular term is *held* at its December 2025 value rather than extrapolated, an assumption worth -0.15 cm had it instead continued at its last five-year slope. The near-real-time months are provisional and are drawn as such.

1.2 Definition of El Niño and indices

ENSO state is defined from the Relative Oceanic Niño Index (RONI) of the NOAA Climate Prediction Center [7, 8], the index on which CPC’s operational ENSO outlooks are issued. RONI is the sea surface temperature anomaly over the Niño-3.4 box (5°S – 5°N , 170° – 120°W) minus the anomaly averaged over the tropics (20°S – 20°N), rescaled so that its variance matches that of Niño-3.4, and is published as overlapping three-month means. Because it is measured relative to the tropical mean, it needs no detrending: secular warming of the tropics does not progressively reclassify late years as warm. The published values are used as issued, each three-month season being assigned to its centre month (e.g., December–February 1983 to January 1983). A month is considered El Niño when the index is at or above $+0.5^{\circ}\text{C}$, La Niña at or below -0.5°C , and neutral in between, with a five-consecutive-month persistence requirement before an event is declared; $+1.5^{\circ}\text{C}$ marks a strong event, and $+2.0^{\circ}\text{C}$ a very strong or “super” event. An *event* is a maximal run of qualifying months and is the unit of both cross-validation and resampling throughout. Over 1950–2015 this gives 20 El Niño events, eight of them strong. The CPC outlook for 2026–27 is issued on the same index, so the scenario amplitudes enter the projection without conversion. The whole analysis was repeated with a conventional Niño-3.4 index (ERSSTv5, linearly detrended over 1950–2015); the results are compared in Supplementary Note 12 and Supplementary Table S8.

1.3 Risk ratio estimation and Shapley attribution.

Composing the response variable.

The analysis requires a water level that carries both the high-frequency chronology of tide and surge and the slowly varying background, with no band counted twice and none discarded. Two facts about these products constrain how they may be combined. GTSM is barotropic: forced by ERA5 winds and sea level pressure, it resolves the wind- and pressure-driven monthly variability and the astronomical long-period tides, but it contains no steric, barystatic, or baroclinic-circulation seasonal cycle. The reconstruction contains all of them. The two therefore *overlap* in the monthly band — both carry the wind- and pressure-driven part — and each also holds something the other cannot.

Let TWL be the GTSM total water level and M the background reconstruction. A naive composition $H = (\text{TWL} - \text{MSL}_{\text{annual}}) + M$, with M taken from the reconstruction variable whose seasonal cycle has been removed upstream, is wrong in both directions at once. Removing only an *annual* mean sea level leaves GTSM’s own monthly variability in the fast term, so the wind- and pressure-driven band is added twice; and taking the seasonal-free background discards the steric and mass

seasonal cycle, which nothing else in the composition supplies. We quantified both defects directly. The double count is regionally split and follows the shelf: the fraction of the background’s monthly variance already linearly present in the GTSM series is 71% in the NE Atlantic, 52% in the Arctic and 47% at Alaska–California, against 7.9% at Mexico–Ecuador and 2.6% at Peru–Chile. The omission runs the other way and is largest in the tropics and semi-enclosed seas: the median peak-to-peak seasonal cycle is 7.4 cm in GTSM against 15.5 cm in the reconstruction, a gap of 12.3 cm in the western Pacific, 11.8 cm in the Gulf of Mexico and 11.5 cm at Mexico–Ecuador.

The adopted composition.

We therefore take the background *with* its seasonal cycle, and remove from the fast term only the part of its monthly mean that the background already explains. For site j , let $\bar{X}_j(m)$ be the mean of TWL over calendar month m , let $T_j(m)$ be its secular trend, and let S_j be the climatology of $\bar{X}_j - T_j$ by calendar month. Let $\tilde{M}_j$ denote the background with its own calendar-month climatology removed, and $\tilde{X}_j = \bar{X}_j - T_j - S_j$. The projection coefficient

$$b_j = \frac{\sum_m \tilde{X}_j(m) \tilde{M}_j(m)}{\sum_m \tilde{M}_j(m)^2} \quad (3)$$

is the amount of the background that the model’s own monthly mean already carries, measured per site. The fast term and the response are then

$$X_j(t) = \text{TWL}_j(t) - [T_j(m) + S_j(m) + b_j \tilde{M}_j(m)], \quad H_j = X_j + M_j, \quad (4)$$

where m is the month containing hour t and M is now the background *retaining* its seasonal cycle. Removal is exact: the monthly mean is formed over contiguous whole months, and no value is interpolated. In the non-seasonal monthly band (4) reduces to $H = M + X_\perp$ with $X_\perp = \tilde{X} - b \tilde{M}$ orthogonal to $\tilde{M}$ by construction, so the shared band is counted exactly once, and the model’s independent barotropic variability survives. Everything sub-monthly, including the tide–surge interaction that motivated using native total water levels, is untouched.

The coefficient b_j is strongly regional, which is the reason a single global rule will not do: across all 3,197 sites the median is 0.29 with an interquartile range of 0.12 to 0.76, and it is far higher on the broad extratropical shelves than on the narrow, thermocline-driven eastern Pacific margin — the same geography as the shared-variance split above. Long-period tides are not lost by this operation. Harmonic analysis of the reconstruction against 82 gauges recovers S_a at an amplitude ratio of 0.91 ($r = 0.93$) and S_{sa} at 0.77 ($r = 0.88$), so the Solar annual and Solar semi-annual tides are carried by M and removing them from the fast term costs nothing. Sensitivity testing of the adopted composition against alternatives on GESLA tide gauge data [51] is provided in the Supplementary Note 1 (Supplementary Table S1).

Two background series.

Because the background now retains its seasonal cycle, two versions of M are employed. The seasonal-retaining series is the one that enters H through (4). This is used in Figs. 1-2. The seasonal-free series is employed when estimating quantities that are regressed

onto RONI: the amplitude and hinge fits, the conditional risk ratios in Fig. 3. This is essential as those indices are themselves deseasonalised against a frozen calendar-month climatology.

High waters and threshold.

Risk is estimated at the level of individual high waters, defined as each local maximum of the hourly series separated by at least 8 hours. Over 1950–2015 this yields 46,573 high waters at the median site, $\approx 706 \text{ yr}^{-1}$. Each high water is treated as a trial, so exceedance probabilities are fractions of high waters rather than rates per unit time. The site-specific threshold z_{99} is the 99th-percentile of high waters of H drawn from neutral months only, so that by construction $\Pr(H > z_{99} \mid \text{neutral}) = 0.01$, corresponding to ≈ 7.1 exceedances per year. The threshold is re-estimated on H as defined in (4); it is not carried over from any other composition.

Month-matched exceedance probability.

El Niño events are strongly phase-locked with seasons [55]. To account for these seasonal effects, El Niño high waters are compared only against neutral high waters occurring in the same calendar month. For each site j , we aggregate high waters into counts n_{bmj} and exceedances e_{bmj} indexed by block b , and calendar month $m \in \{1, 2, \dots, 12\}$. Blocks are defined as either complete contiguous El Niño events (20 El Niño events over 1950–2015 – see above definition), or individual years for the neutral states (64 years over the same period). Counts are pooled across blocks for each calendar month *prior* to any ratio being formed such that

$$\hat{p}_{mj}^S = \frac{\sum_{b \in S} e_{bmj}}{\sum_{b \in S} n_{bmj}}, \quad S \in \{E, N\}, \quad (5)$$

where S refers to the condition (E = El Niño, N = neutral). Pooling in this manner ensures events of unequal lengths contribute proportionally to the number of high waters they contain rather than a mean of per-event rates. The calendar-month mixture of the conditioned states defines a single weight vector,

$$w_m = \frac{\sum_j \sum_{b \in E} n_{bmj}}{\sum_{m'} \sum_j \sum_{b \in E} n_{bm'j}}, \quad \sum_m w_m = 1, \quad (6)$$

corresponding to the probability of being in a given month m given state E . This is applied to *both* states,

$$\hat{p}_j^S = \sum_m w_m \hat{p}_{mj}^S, \quad (7)$$

to ensure the neutral reference is re-weighted onto the El Niño seasonal occupancy. Weights are summed over sites, which yields a single seasonal reweighting across the network. Testing showed retaining per-site weights had negligible effects on the risk ratio estimates described below. Therefore, the pooled vector is used because a single weighting makes differences between sites attributable to the sites rather than being relative to their individual weights.

Risk Ratio.

Weighting each probability $\hat{p}_j^S$ by the total number of high waters $n_j^S = \sum_m \sum_{b \in S} n_{bmj}$ at site j and in state S , we define the risk ratio as

$$\text{RR}_j = \frac{(\hat{p}_j^E n_j^E + \frac{1}{2}) / (n_j^E + 1)}{(\hat{p}_j^N n_j^N + \frac{1}{2}) / (n_j^N + 1)}. \quad (8)$$

Note that the risk ratio has a Haldane–Anscombe continuity correction, which can accommodate the possibility of zero occurrences of a given state. Its effect on the point estimates is negligible (median 0.08%), and where a site has fewer than five neutral exceedances ($\hat{p}_j^N n_j^N < 5$) the RR is treated as undefined. The correction is nevertheless retained for estimating confidence intervals: the event-block bootstrap (described below) resamples 20 whole El Niño events with replacement, and at six sites a replicate can draw no El Niño exceedance at all (at most 1.2% of replicates), which without the correction would make that replicate’s log risk ratio undefined. The correction removes that failure mode. Uncertainty estimates are described in the dedicated uncertainty section 1.4.

Given the month-matched exceedance probabilities, the risk ratio is a multiplier *on the exceedance probability per high water, within the months El Niño occupies*. It is not the change in the expected annual number of exceedances, and the two differ whenever exceedances are seasonally concentrated.

Counterfactual states.

We construct four states from realised distributions of X and M such that

$$p_{AB} = \Pr(X_A + M_B > z_{99}), \quad A, B \in \{E, N\}, \quad (9)$$

where the first subscript A denotes the state of the high-frequency tide–surge chronology X and the second B the state of the background M . Again, E denotes El Niño, and N denotes neutral. p_{EE} is the realised El Niño state, p_{NN} the neutral baseline, p_{EN} pairs the El Niño high-frequency chronology with a neutral background, and p_{NE} a neutral high-frequency chronology with the El Niño background state. All risk ratios are then referenced against the neutral state as a common baseline, carrying the *same* continuity correction and the same counting convention as (8), so that the four-state quantities and the two-state risk ratio are computed identically and remain comparable:

$$\text{RR}_{AB} = \frac{(p_{AB} n_j^A + \frac{1}{2}) / (n_j^A + 1)}{(p_{NN} n_j^N + \frac{1}{2}) / (n_j^N + 1)}, \quad \text{RR}_{NN} \equiv 1, \quad (10)$$

where n_j^A is the number of high waters contributing to state A at site j . The $p_{NN} n_j^N < 5$ censor of (8) applies here unchanged. The Haldane–Anscombe continuity correction is once again employed with a median effect below 0.1%, but is necessary for keeping bootstrapped log risk ratios defined.

For the two realised states p_{EE} and p_{NN} , each high-water retains the background of the month in which it occurred. Counterfactual states instead substitute a *donor* background drawn from the opposite state at the same calendar month. The donor background is averaged over all admissible donors (read: backgrounds from matched months). This is necessary to produce an unbiased estimate of the counterfactual scenario as the realised pairing carries a covariance between X and M .

If one neglects this and instead mixes two months directly (e.g. leaves two of the four states carrying covariance and two without it), the apparent additive interaction $I = \text{RR}_{EE} - \text{RR}_{EN} - \text{RR}_{NE} + \text{RR}_{NN}$ at the median site is -0.017 , against -0.005 with donor-consistent states; the difference is an artefact of the pairing rather than a property of the ocean, and donor averaging removes it. Donor averaging removes this effect. Note, however, that while these indicators are averaged over donors, the background values themselves are never averaged, so the variance of M is preserved. We therefore use donor-consistent quantities for all four states in the decomposition described below.

Shapley Decomposition.

Using the assembled counterfactual states, one can precisely attribute the contribution of both the background M and the fast chronology X to the total risk. We treat the background M and fast chronology X as two players with a characteristic function $v(\emptyset) = 0$, $v(\{M\}) = \text{RR}_{NE} - \text{RR}_{NN}$, $v(\{X\}) = \text{RR}_{EN} - \text{RR}_{NN}$ and $v(\{M, X\}) = \text{RR}_{EE} - \text{RR}_{NN}$. The exact two-player Shapley values are thus

$$\phi_M = \frac{1}{2}[(\text{RR}_{NE} - \text{RR}_{NN}) + (\text{RR}_{EE} - \text{RR}_{EN})], \quad (11)$$

$$\phi_X = \frac{1}{2}[(\text{RR}_{EN} - \text{RR}_{NN}) + (\text{RR}_{EE} - \text{RR}_{NE})]. \quad (12)$$

The two values ϕ_M and ϕ_X are the mean marginal contribution of each player over both orders of addition. This is the unique allocation of the interaction described above in the sense that it satisfies efficiency, symmetry, additivity, and the null-player property as described in [29]. For two players, the interaction described above is split evenly between players. The decomposition is exact and closes with

$$\phi_M + \phi_X = \text{RR}_{EE} - \text{RR}_{NN}. \quad (13)$$

Because the decomposition closes, the split can be read as a share. We define the *background share*

$$f_M = \frac{\phi_M}{\text{RR}_{EE} - \text{RR}_{NN}}, \quad (14)$$

as the fraction of the total change in exceedance risk attributable to the background M , with $1 - f_M$ the fraction attributable to the fast chronology X . Two properties follow from (14) being a ratio of signed quantities and govern how it is reported. First, it is undefined where the total is small, so f_M is defined only where $|\text{RR}_{EE} - \text{RR}_{NN}| \geq 0.15$; 871 of the 3,197 sites return no value, and a share of nothing is never reported. Second, it is *not* bounded to $[0, 1]$: where the two players act in opposite directions, $\phi_X < 0$ gives $f_M > 1$ (the background more than accounts for the total, and the weather partly

offsets it), which occurs at 651 sites. Where $RR_{EE} < RR_{NN}$ — the El Niño *lowers* extreme-water-level risk, as across the Maritime Continent — f_M is correspondingly the share of a decrease. Regional values are the median of per-site shares rather than the ratio of regional medians, since the two differ, and they are shown only where at least eight sites produce a value, and the regional median $|RR_{EE} - RR_{NN}|$ exceeds 0.15.

A note on exceedance rates.

Exceedance rates are estimated per high-water peak. A 1% exceedance rate, corresponding to z_{99} , occurs in expectation 7.1 times per year. This count is essentially invariant to the composition in Eq. (4): removing a monthly mean shifts the series by a constant within each month, so a local maximum can appear or vanish only at a month boundary; relative to the naive composition, the number of high waters changes by a median of 0.003% (at most 5.8%).

Wave and tide–surge coincidence

For each high water, large-wave forcing is scored when the maximum of W within ± 6 h exceeds W_{95} , the site’s 95th percentile of W over neutral-month high waters (ice-covered hours excluded). H is the composition of (4), and an extreme high water is one above the z_{99} of the risk ratio. The coincidence probability is the share of extreme high waters accompanied by large-wave forcing, $p = \Pr(W > W_{95} \mid H > z_{99})$. To keep the month matching of the risk ratio, the joint probability $\Pr(H > z_{99}, W > W_{95})$ and $\Pr(H > z_{99})$ are each month-matched with the site’s El Niño monthly occupancy as weights (months with ≥ 5 El Niño and ≥ 20 neutral high waters), and p is their ratio, for El Niño and neutral months separately. Using this, we obtain $RR_{HW} = p^E/p^N$. It is undefined at sites with fewer than five El Niño or five neutral extreme high waters with wave data. Because it conditions on an extreme high water, RR_{HW} describes the wave state that accompanies extremes, separately from how often extremes occur, which is RR_{EE} . Intervals use 400 replicates of the event-block scheme described below.

1.4 Uncertainty Estimates

All confidence intervals in this study are computed using a bootstrap, and no approach used resamples observations independently. This is essential to handle the spatiotemporal autocorrelation in coastal water levels. In our context, the 792-month record contains only twenty El Niño events, eight of them strong, and neighbouring high waters within an event are not independent. As such, resampling throughout is intended to be done at the level in which the data are exchangeable, and this differs by statistic. One approach is employed in the main text, and all reported confidence intervals are at the 95% level unless explicitly stated otherwise. Two additional approaches, a 36-month moving-block bootstrap (Supplementary Note 2) and a site-clustered bootstrap (Supplementary Note 3), are described in the Supplementary Information.

Event-block resampling for risk ratios.

All risk ratios and counterfactual decompositions make use of event-block resampling. The record is partitioned into blocks that are resampled whole: a warm block is one

complete, contiguous El Niño event (20 of them), and a neutral block is one calendar year (64). A replicate draws 20 warm blocks and 64 neutral blocks with replacement and then recomputes the entire estimator from those blocks — the calendar-month weights, the pooled exceedance counts, the Haldane–Anscombe continuity correction and the minimum-neutral-exceedance mask are all re-derived, never held fixed at their observed values. Crucially, the donor months used to build the counterfactual states are resampled in the same replicate, so uncertainty in the donor sets propagates into the counterfactual risk ratios rather than being treated as known. We use differing numbers of samples depending on the data:

1. 200 for the four-state risk ratios and their Shapley decomposition (Figs. 1b and 2
2. 400 for the wave-coincidence statistic
3. 1,000 for the regional knot rule
4. 2,000 for the conditional sea level models and the scenario projection

1.5 Projecting to an El Niño event of a given amplitude.

Several results are expressed for an El Niño of a stated amplitude rather than for the observed events, and all of them share one fit, which is defined here once. At each site, the background is regressed on RONI over the *warm limb* only — the months with a non-negative index — because the response is asymmetric and a fit through both limbs would average two different regimes.

Two forms are carried,

$$\Delta M_j(A) = \alpha_j + \beta_j A \quad \text{and} \quad \Delta M_j(A) = \alpha_j + \beta_j A + \gamma_j \max(A - A_0, 0), \quad (15)$$

with the second adding a hinge at $A_0 = 1.5^\circ\text{C}$. The breakpoint $A_0 = 1.5^\circ\text{C}$ is fixed *a priori* at the conventional strong-El-Niño threshold and is not estimated from the coastal response. This preserves a one-parameter test of steepening through γ and avoids selecting a breakpoint from the same observations used to evaluate it. We separately test sensitivity to this choice by profiling the breakpoint over $0.6\text{--}2.2^\circ\text{C}$ and by fitting a free-breakpoint model using whole-event resampling (Supplementary Fig. S10). These analyses are used to assess robustness to breakpoint placement rather than to redefine the estimator. The regression uses the seasonal-free background, for the reason given above.

Sites are grouped along the eastern Pacific rim: Alaska, British Columbia, US West Coast, Mexico, west Central America, Colombia–Ecuador, Peru, Chile. Elsewhere, coarse regions are used (shown in Fig. A1), giving 23 regions, each with at least 35 sites. The eastern Pacific is resolved finely because that is the one basin where the response varies materially over a few hundred kilometres of coast. For region r define the composite background as the cross-site median

$$\bar{M}_r(t) = \text{median}_{j \in r} M_j(t), \quad (16)$$

and fit (15) to $\bar{M}_r$ over the warm limb, giving a single γ_r that is used only to decide whether region r adopts the hinge. The region adopts the hinge when a 95% event-block bootstrap interval on γ_r excludes zero, using 1,000 replicates and the same block scheme as every other interval in this study (§1.4). The test is *two-sided*, so a region may adopt the hinge because its response flattens above A_0 as well as because it steepens. Sites in adopting regions are then projected with their own hinged fit (α_j , β_j , γ_j) and all other sites with the linear fit. Nine regions are identified as significant, covering 1,530 of the 3,197 sites. For the 2026–27 outlook, we additionally evaluate structural sensitivity by repeating the projection with the purely linear warm-limb model at all sites. The linear model is refitted independently rather than obtained by setting $\gamma = 0$ in the hinged fit, because the corresponding estimates of α and β differ. All other ocean-state conditioning, astronomical tide contributions, exceedance curves and site-screening procedures are held unchanged (Supplementary Fig. S11). The case for a regional rather than a per-site decision, its calibration against a circular-shift null, the adopting regions and the sensitivity tests are set out in Supplementary Information (Supplementary Figs. S9–S11, Tables S4–S6).

Adding astronomical tidal effects

The outlook adds a deterministic astronomical tidal term η_j for the season in which the event is expected to peak, November 2026–January 2027, so the risk ratio is read at $\Delta M_j(A) + \eta_j$. GTSM carries only about a sixth of the 18.6-year nodal modulation of M_2 that tide gauges record (1.1% against 7.4% peak-to-peak), so the term is synthesised rather than read from the model. Eight constituents (M_2 , S_2 , N_2 , K_2 , K_1 , O_1 , P_1 , Q_1) are fitted to the model tide at each site over 2010–2012, and the tide of each November–January season is rebuilt in continuous time with the nodal amplitude factors and phase corrections [56] evaluated at the season’s midpoint. The synthesis therefore carries the 18.6-year nodal cycle and also the 8.85-year perigean cycle, whose 4.4-year half-period arises from the changing alignment of lunar perigee with the syzygies and solstices [43]. η_j is the anomaly of the season’s 99th-percentile high water from its 1950–2014 November–January mean. The choice of season matters at Mexico–Ecuador, where the central risk ratio is 10.9 for October–December, 12.9 for November–January and 13.6 for December–February (Supplementary Note 4).

Converting to a risk ratio

A projected shift is converted to a risk ratio directly from the site’s empirical exceedance curve, without fitting a tail distribution. No extreme-value model is assumed anywhere in this study. The curve is built from all neutral-month high waters of H , pooled over 1950–2015 and all calendar months: the risk ratio is the fraction of these high waters that exceed z_{99} , the threshold of the risk ratio, after a uniform shift by $\Delta M_j(A) + \eta_j$, divided by the unshifted fraction (0.01). It is therefore a ratio per high water for months in which the background stands $\Delta M_j(A)$ above normal, i.e. near the event peak, not an event or annual average.

Ocean-state range

The central projection depends on RONI alone. The ocean state enters only through the spread around it. We take $p(SSH_{tilt} | A)$ to be Gaussian, with a mean given by a linear fit to the warm-limb months of 1993–2015 (16.3 cm per °C) and a standard deviation equal to the residual spread of the eight satellite-era events at their RONI maximum (5.2 cm; 95% CI 3.4–11.5). Each draw of the tilt anomaly $\delta = SSH_{tilt} - E[SSH_{tilt} | A]$ is applied basin-wide, shifting site j by an additional $c_j \delta$, where c_j is the partial regression coefficient of M on SSH_{tilt} at fixed RONI (warm limb, 1993–2015). Regional bands are the 5–95% range, across draws, of the median risk ratio over issuable sites (20,000 draws at the scenario amplitudes; 2,000 for the continuous curves in Fig. 3c and Supplementary Figs. S6–S7).

1.6 Sea-level rise equivalence

To give the projected El Niño background shift a familiar scale, we express it as the number of years of projected sea-level rise that would accumulate the same displacement (Table A1).

The numerator is the site-wise shift $\Delta M_j(A)$ of (15), with the astronomical tidal term *excluded*, because the question asks about the shift from the El Niño and the tidal cycles are a separate, independently varying contribution.

The denominator is the IPCC AR6 regional sea level projection [48, 49], taken from the FACTS medium-confidence output at the median quantile, on the AR6 one-degree grid; each analysis site is matched to its nearest grid cell (median separation 38 km, maximum 76 km) and every trajectory is referenced to its own 2026 value, so the quantity is cumulative rise since 2026. We use the `nov1m` variant, which excludes the AR6 background vertical land motion. Land motion therefore enters neither side, and the comparison is of the oceanic contribution alone. It is consequently *not* an interval of relative sea-level rise, and at coasts undergoing rapid uplift or subsidence the relative figure would differ substantially. Three scenarios are reported: SSP1-2.6, SSP2-4.5 and SSP5-8.5.

Two equivalent intervals are given for each. The *linear* figure holds the AR6 near-term rate — the mean 2026–2050 rate, 4.5 to 5.7 mm yr^{−1} over the eastern rim depending on scenario — constant thereafter, and is what neglecting the acceleration would give. The *accelerating* figure follows the AR6 trajectory itself, $\Delta S_j(t) - \Delta S_j(2026) = \Delta M_j$, by interpolation between the decadal steps. Where the shift is large enough to be reached late in the century, the acceleration shortens the interval, as expected; where it is small and reached early, the accelerating figure is slightly *longer*, because the rate in the first decades is below the 2026–2050 mean that the linear column uses. Both are reported as medians of per-site ratios rather than ratios of medians, and regionally rather than pooled. The comparison is one of magnitude alone: a single El Niño delivers this displacement within a season and removes it again, and sea-level rise does not.

Supplementary information. Supplementary Information accompanies this paper. Supplementary Notes 1–13 cover: the tide-gauge evidence for the composition of H (Note 1, Table S1); the wind impulse-response analysis of the mechanisms

setting M (Note 2, Fig. S1); the reported-flood (Groundsource) analysis (Note 3, Figs. S2–S3); the astronomical tidal state of 2026–27 against 2015–16 (Note 4, Fig. S4); the co-evolution of RONI and SSH_{tilt} (Note 5, Fig. S5); conditional risk against amplitude by region and along the eastern Pacific (Note 6, Figs. S6–S7, Table S2); tidal-excursion duration (Note 7, Fig. S8, Table S3); the regional knot rule and its sensitivity to knot placement and functional form, including Fig. 3 recomputed with a linear response (Note 8, Figs. S9–S11, Tables S4–S6); the H_s -only wave ablation (Note 9, Fig. S12); the dependence of wave intensity and wave–high-water coincidence on El Niño amplitude (Note 10, Table S7); the sensitivity of the background to the equatorial tilt at fixed RONI (Note 11, Fig. S13); the sensitivity of every result to the choice of ENSO index (Note 12, Table S8); and the 2026–27 outlook after conditioning on the measured ocean-state as of September 2026 (Note 13, Table S9). Every supplementary figure and table is generated reproducibly in the accompanying notebook, `ENSO_coastal_supp.ipynb`.

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Declarations

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Appendix A Extended Data

Table A1 Equivalent sea-level rise due to El Niño background shift (cm). Regional background sea level shift in *cm* is converted to number of years of sea-level rise from 2026 based on both linear and accelerating IPCC AR6 trends [48].

Region	Sites	(cm)	SSP1–2.6		SSP2–4.5		SSP5–8.5	
			Linear	Accel.	Linear	Accel.	Linear	Accel.
Low ENSO scenario (RONI = 2.28°C)								
Mexico–Ecuador	491	14.1	29.5	30.1	26.8	26.6	22.4	22.7
Peru–Chile	304	5.8	16.5	17.6	15.2	16.9	12.7	14.6
Alaska–California	575	7.6	19.7	19.9	18.2	19.0	16.0	17.1
Eastern rim (pooled)	1370	9.0	21.9	22.0	20.8	21.1	18.3	19.0
Central ENSO scenario (RONI = 2.74°C)								
Mexico–Ecuador	491	19.3	40.4	40.2	36.7	35.6	30.6	29.6
Peru–Chile	304	7.1	20.1	20.7	18.5	19.6	15.4	16.9
Alaska–California	575	9.4	24.3	23.8	22.5	22.5	19.7	20.2
Eastern rim (pooled)	1370	11.1	27.7	27.2	26.5	25.8	23.2	23.0
High ENSO scenario (RONI = 3.21°C)								
Mexico–Ecuador	491	24.5	51.4	49.8	46.7	43.3	38.8	36.1
Peru–Chile	304	8.4	23.8	23.8	21.8	22.2	18.0	19.0
Alaska–California	575	11.3	28.8	28.4	26.8	26.1	23.5	23.3
Eastern rim (pooled)	1370	13.2	33.4	32.5	32.0	30.2	28.3	27.0

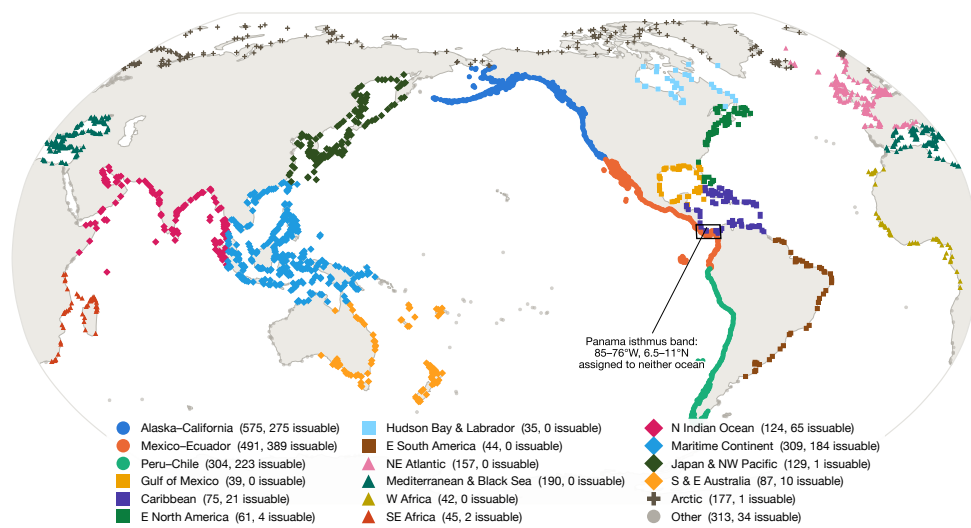


Fig. A1 Labeled map of coarse regions considered in this manuscript.

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1 Supplementary Information for Super El Niño can
2 raise coastal extreme sea-level risk tenfold

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Overview

This document contains the Supplementary Notes, Figures and Tables that support the main text. Sections 1–13 are cited in the main text as Supplementary Notes 1–13. Every figure and table is generated by the accompanying notebook, `ENSO_coastal_supp.ipynb`, which writes the data behind each panel alongside it. Regions are those of Extended Data Fig. 1 unless stated otherwise. Notation follows the Methods: H is the total water level, M the background sea level, X the tide-surge level, W the waves (either H_s significant wave-height, or run-up proxy (Methods)), A the RONI amplitude, and SSH_{tilt} the cross-basin sea-surface-height difference.

1 Composition of H and the seasonal cycle of M

The background reconstruction M and the barotropic tide-surge model X both carry variability at monthly periods, so their naive sum $M + X$ counts the shared band twice. We compared four compositions of the monthly background: M alone; the model’s monthly mean $\bar{X}$ alone; the naive $M + X$; and the adopted $M + X_{\perp}$, in which the part of $\bar{X}$ that projects onto M is removed (Methods). Deleting $\bar{X}$ entirely makes the monthly mean of H equal to M and so removes the double count, but it also discards the barotropic variability that the model resolves and the reconstruction does not.

The four were scored against 702 GESLA4.1 [1] tide-gauge records over 1950–2015 in the non-seasonal monthly band, with the calendar-month climatology and a linear trend removed from every series identically (Supplementary Table S1a). $M + X$ has the highest correlation and the largest error (standard deviation 1.075 times the observed, RMSE 4.65 cm), the signature of a band counted twice. M alone errs the other way, retaining 64% of the observed monthly spread. $M + X_{\perp}$ has the lowest error (3.67 cm).

The choice was then tested on the quantity the paper reports: over the matched gauges, the ordinary-least-squares slope of observed on modelled El Niño RR_H with $H = X + M$ must move toward 1 and the correlation must not fall (Supplementary Table S1b). With the RONI classification used throughout this study, the correlation criterion is met but the slope criterion is not: for the adopted composition r rises from 0.862 to 0.868, but the slope overshoots 1 (1.148 against 0.904), so $|1 - \text{slope}|$ grows from 0.096 to 0.148, against M alone: $r = 0.874$, $|1 - \text{slope}| = 0.178$. When the same test was run with Niño-3.4 (Supplementary Note 12), both criteria were met (r from 0.844 to 0.870, $|1 - \text{slope}|$ from 0.157 to 0.099). The gauge evidence for the composition therefore rests on the monthly-band comparison above, which does not depend on the ENSO index; on the reported risk ratio the adopted composition matches the gauges about as well as the naive one, erring high rather than low. The correction matters little where M and X share little variance and a great deal where they share much: RMSE between $M + X_{\perp}$ and naive $M + X$ moves from 4.08 to 4.15 cm at Mexico–Ecuador and from 5.85 to 5.93 cm at Peru–Chile, but from 6.00 to 3.63 cm in the NE Atlantic and from 7.48 to 3.99 cm in the Arctic.

Supplementary Table S1 Tide-gauge evidence for the composition of H . **a**, Each candidate monthly background scored against 702 gauge records, 1950–2015, in the non-seasonal monthly band (calendar-month climatology and linear trend removed from every series identically): standard deviation relative to the gauge, correlation, and RMSE. **b**, correlation and ordinary-least-squares slope of observed on modelled El Niño RR_H over 708 matched gauges. The criterion, fixed before either variant was computed, was that the slope move toward 1 without the correlation falling.

a Monthly band, 702 gauges			
Composition	sd ratio	r	RMSE (cm)
M only	0.638	0.81	4.00
$\bar{X}$ only	0.596	0.76	4.37
$M + X$ (naïve)	1.075	0.86	4.65
$M + X_{\perp}$ (adopted)	0.754	0.82	3.67
b El Niño RR_H , 708 gauges			
Composition	r	slope	$ 1 - \text{slope} $
$M + X$ (naïve)	0.862	0.904	0.096
M only	0.874	1.178	0.178
$M + X_{\perp}$ (adopted)	0.868	1.148	0.148

2 Mechanisms of the background anomaly

El Niño indices are composite metrics: they correlate with coastal sea level but are not the mechanism that drives it. This note examines that mechanism empirically (Supplementary Fig. S1).

Wind impulse-response function

We fit a causal multi-input finite impulse-response filter, here called the impulse-response function (IRF), from equatorial zonal wind stress τ_x to the background $M(t)$ at each site. Causal means that only present and past forcing enters, so the fitted response is an accumulation of wind stress rather than a contemporaneous association:

$$M(t) = c + \sum_{k=0}^2 \sum_{j=0}^L h_j^k \tau_x^k(t-j) + e_t, \quad (\text{S1})$$

where h_j^k is the sea-level response j months after a one-month wind pulse in box $k \in \{0, 1, 2\} \equiv \{\text{W}, \text{C}, \text{E}\}$. With $L = 18$ (19 lags) there are 57 predictors. Monthly zonal wind stress is taken from ERA5[2]; the monthly climatology and linear trend are removed before $\cos \varphi$ -weighted averaging over three boxes (5°S – 5°N ; 140 – 170°E , 170 – 200°E , 200 – 270°E). Wind stress is not smoothed, since smoothing would inflate the apparent memory of the system. After the leading L months are consumed by the lags, 774 months (1951-07 to 2015-12) remain.

Regression. Let $\mathbf{y} \in \mathbb{R}^{n_{\text{tr}}}$ be M at one site over the n_{tr} training months and $\mathbf{X} \in \mathbb{R}^{n_{\text{tr}} \times 57}$ the lagged wind-stress design matrix, ordered lag-major,

$$\mathbf{X}_{t, 3j+k} = \tau_{x, t-j}^k, \quad k \in \{0, 1, 2\}, \quad j = 0, \dots, L, \quad (\text{S2})$$

so box k occupies columns $k, k+3, \dots, k+54$. $\mathbf{h}$ has units of cm per Nm^{-2} ; in Supplementary Fig. S1a it is rescaled to cm per one-standard-deviation wind pulse. Because training months are autocorrelated and the lag columns are nearly collinear, unregularised least squares is dominated by noise. We therefore penalise curvature in lag, encoding the physical assumption that the ocean’s response to a wind pulse varies smoothly with lag; an ordinary ridge penalty would instead shrink each lag independently.

Smoothness penalty. For one box with kernel $(h_0, \dots, h_L)$ the second difference at lag j is $h_j - 2h_{j+1} + h_{j+2}$. Collecting these in $\mathbf{D} \in \mathbb{R}^{(L+1) \times (L+1)}$, whose i th row holds $(1, -2, 1)$ from column i , the penalty is $\|\mathbf{D}\mathbf{h}\|^2$. The last rows are zero-padded rather than truncated, which imposes $h_{L+1} = h_{L+2} = 0$: without it $\mathbf{D}^\top \mathbf{D}$ has a two-dimensional null space (constant and linear kernels); with it $\mathbf{D}^\top \mathbf{D}$ has full rank and the kernel must decay to zero by the end of the window, a damped-response prior. The full penalty applies this box by box,

$$\mathbf{P} = 10^{-3} \mathbf{I}_{57} + \sum_{k=0}^2 \mathbf{S}_k^\top \mathbf{D}^\top \mathbf{D} \mathbf{S}_k, \quad (\text{S3})$$

102 where $\mathbf{S}_k$ selects box k 's 19 columns. The factor 10^{-3} fixes the relative weight of a small
 103 isotropic ridge, which makes $\mathbf{P}$ strictly positive definite while leaving it dominated by
 104 smoothness; the penalty strength is the separate scalar λ .

105 **Estimation.** Predictors and target are centred but not standardised, so that
 106 the IRF keeps physical units and each box contributes according to its physical
 107 importance; c is then the training-fold mean of M . The kernel is

$$\hat{\mathbf{h}}(\lambda) = (\mathbf{X}^\top \mathbf{X} + \lambda n_{\text{tr}} \mathbf{P})^{-1} \mathbf{X}^\top \mathbf{y}, \quad (\text{S4})$$

108 with $\lambda \geq 0$ chosen by cross-validation ($\lambda = 0$, unregularised least squares, is
 109 admissible). Confidence bands on h_j^k are from a 36-month block bootstrap with 500
 110 resamples.

111 **Cross-validation.** Skill is always estimated out of sample by contiguous-block
 112 cross-validation with a lag embargo, because sea level is autocorrelated and a test
 113 month t shares wind inputs back to $t - L$ with its neighbours. In the outer loop the
 114 774 months are divided into $K_{\text{out}} = 8$ contiguous blocks of 96 or 97 months, each
 115 held out in turn, and every training month within $L = 18$ months of the test block is
 116 discarded, so no month is predicted by a model that saw wind it shares. In the inner
 117 loop the remaining training months are divided into $K_{\text{in}} = 4$ blocks with the same
 118 embargo, and λ is chosen from $\{0\} \cup \{10^{-3}, \dots, 10^4\}$ (21 log-spaced values plus zero)
 119 by minimising summed inner-fold squared error, independently for every site and outer
 120 fold. The selected model is refitted on the full training set and applied once to the test
 121 block.

122 **Skill.** Out-of-fold predictions are concatenated into one series per site and scored
 123 as

$$R^2 = 1 - \frac{\sum_{t \in \mathcal{M}} (\hat{M}_t - M_t)^2}{\sum_{t \in \mathcal{M}} (M_t - \bar{M}_{\mathcal{M}})^2}, \quad (\text{S5})$$

124 where $\mathcal{M}$ is the set of scored months. Eq. (S5) is a pooled out-of-sample statistic,
 125 unbounded below. Where $\mathcal{M}$ is restricted (for example to strong El Niño months), the
 126 model is still fitted on all months and only the scoring is restricted. Contemporaneous
 127 wind alone scores -25% to -144% . This is what wind-forced ocean dynamics predicts
 128 when an instantaneous forcing is regressed on its own time integral; it reflects the
 129 estimator, not evidence against the mechanism.

130 **Identifiability.** The western and central wind boxes correlate at $r = 0.79$. Their
 131 combined gain is well determined, but its partition between the two boxes is not, and
 132 we make no claim about it.

133 Results

134 El Niño weakens, and at times reverses, the easterly equatorial trade winds. The
 135 wind anomaly excites eastward-propagating equatorial Kelvin waves, which cross the
 136 basin in about two months and deepen the thermocline in the east. The fitted IRFs
 137 reproduce the expected basin-scale dipole (Supplementary Fig. S1a): at zero lag a west-
 138 erly anomaly in the central box raises the background in the east (Mexico–Ecuador,
 139 Peru–Chile, Alaska–California) and lowers it on the Maritime Continent. The eastern

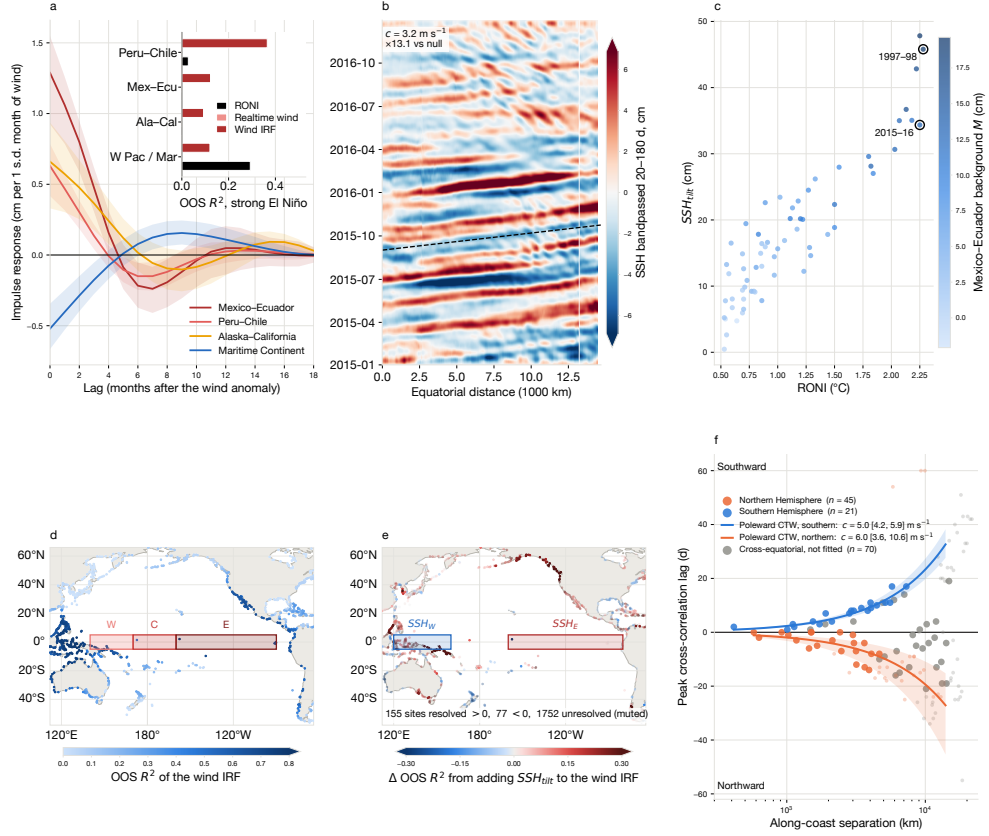
responses change sign at 4–6 months and decay over roughly a year. Across the tropical Pacific the IRF explains a median of 61% of the out-of-sample variance of M , and up to 79% (Supplementary Fig. S1d).

Wind history matters most where contemporaneous indices fail. In strong El Niño months (inset of Supplementary Fig. S1a), RONI alone has essentially no out-of-sample skill on the eastern rim, whereas the IRF explains roughly 0.1–0.4 of the variance. On the Maritime Continent the order reverses: RONI outperforms the IRF by a factor of more than two, consistent with the central-box wind being only one part of the western Pacific response. Daily GLORYS12V1 sea-surface height [3], product GLOBAL_MULTIYEAR_PHY_001_030 (doi.org/10.48670/moi-00021) along the equator in 2015–16 shows an eastward-propagating signal at $c = 3.2 \text{ ms}^{-1}$, characteristic of equatorial Kelvin waves (Supplementary Fig. S1b).

A second determinant of M is the state of the upper tropical Pacific, in particular thermocline depth, for which the cross-basin height difference SSH_{tilt} has been proposed as a proxy [4]. In a 1.5-layer model of the equatorial Pacific, the surface displacement is $\eta = h \Delta\rho/\rho$, where h is thermocline depth and $\Delta\rho$ the density contrast across it [5]. SSH_{tilt} and RONI rise together across satellite-era El Niño months (Supplementary Fig. S1c), but they separate at the largest amplitudes. In 1997–98 (RONI 2.28°C , SSH_{tilt} 45.8 cm) the Mexico–Ecuador background reached 18.5 cm; in 2015–16, at almost the same index (2.25°C) but with a weaker tilt (34.4 cm), it reached 10.7 cm. The coastal response therefore depends on the ocean state an event reaches, not only on its SST amplitude.

Adding contemporaneous SSH_{tilt} to the wind IRF (Supplementary Fig. S1e; satellite window 1993–2015, scored on its 56 El Niño months) raises out-of-sample R^2 most in the northeast Pacific, by a median of 0.27 between Oregon and Alaska. There, and only there, the tilt coefficient peaks at a lag of one month rather than zero, indicating that the coastal response is not instantaneous. This is consistent with coastally trapped waves (CTWs): on reaching the eastern boundary, equatorial Kelvin waves propagate poleward along the coast [6]. Using 20–180-day band-passed tide-gauge records along the eastern Pacific during 2015–16, peak cross-correlation lags increase poleward with along-coast separation in both hemispheres (Supplementary Fig. S1f): $c = 5.0$ [4.2, 5.9] ms^{-1} southward and 6.0 [3.6, 10.6] ms^{-1} northward, with the poleward sign in 58 of 61 non-zero within-hemisphere pairs. At these speeds an anomaly reaches the mid-point of the Alaska–California coast in 0.72 [0.41, 1.20] months, or 0.87 [0.74, 1.02] months using the better-determined southern speed, consistent with the one-month lag. The propagating signal rides on a broadly coherent adjustment, visible as pairs with near-zero lag.

These three pathways — accumulated wind forcing, the basin-scale ocean state, and poleward coastal propagation — together set the background anomaly M during El Niño, and they give it a degree of predictability beyond the SST index (Supplementary Note 5).



Supplementary Figure S1 Coastal response to ENSO-driven equatorial wind stress. a, Impulse response of the regional background M to a one-standard-deviation month of central-box zonal wind stress, with 95% intervals from a 36-month block bootstrap (500 resamples). Inset: out-of-sample R^2 in strong El Niño months ($A > 1.5^\circ\text{C}$) for contemporaneous RONI, contemporaneous wind and the wind IRF, on identical folds, embargo and penalty selection. **b,** Daily GLORYS12V1 sea-surface-height anomaly along the equator in 2015-16 (20–180-day band-pass). The dashed line is the beamformed slowness, $c = 3.2\text{ m s}^{-1}$, which exceeds a shuffled-column null by a factor of 13. **c,** El Niño months of the satellite era in the RONI- SSH_{tilt} plane, coloured by the Mexico–Ecuador background; 1997-98 and 2015–16 are ringed at their RONI maxima. **d,** Per-site out-of-sample R^2 of the wind IRF, 1951-07 to 2015-12, with the three forcing boxes (W, C, E). **e,** Per-site change in out-of-sample R^2 in El Niño months when SSH_{tilt} is added to the wind IRF (satellite window, 1993–2015; boxes mark the two limbs of SSH_{tilt}). Sites whose 95% event-block bootstrap interval (2,000 resamples) spans zero are muted; R^2 levels are not comparable with **d**. **f,** Lag of peak cross-correlation between eastern Pacific tide gauges against along-coast separation during 2015–16, by hemisphere, with a through-origin Theil–Sen fit per branch and its gauge-clustered 95% band. Cross-equatorial pairs are shown but not fitted. Note results are estimated on the thinned global coastal sites: $N = 1,984$.

3 Reported coastal floods

We test whether the drivers analysed in the main text leave a measurable signature in reported floods, using the Groundsource dataset of 2.6 million flood reports extracted from news articles in over 150 countries[7]. The dataset covers all flood types, predominantly rainfall-driven urban flooding, and does not attribute a driver. Coastal events were identified as follows:

1. records whose representative point lies within 25 km of the Natural Earth 10 m coastline are retained;
2. each is matched to the nearest analysis site within 50 km;
3. records at the same site starting within three days of each other are merged into one event (a case);
4. each case is matched to eight control days at the same site and in the same calendar month, drawn from the same year where possible and otherwise from the adjacent year, excluding any day within ± 7 days of any Groundsource record at that site. A case and its controls form a matched set, or stratum.

Each predictor is the maximum over a ± 1 -day window around the day. Because the background reconstruction ends in 2015, M is taken over 2000–2024 from the DUACS multi-mission altimetry product[8]. On the 2000–2015 overlap, replacing the reconstruction with DUACS retains about two thirds of the incremental background signal. The analysis comprises 69,643 strata at 1,985 sites. Regions with fewer than 150 strata (100 in the El Niño subset) are not reported.

Conditional logistic regression. Within stratum s , the probability that day i is the reported flood day is modelled as

$$\text{logit Pr}(\text{day } i \text{ of } s \text{ flooded}) = \alpha_s + \mathbf{x}_{si}^\top \boldsymbol{\beta}, \quad (\text{S6})$$

where α_s absorbs everything constant within the stratum (exposure, reporting intensity, tidal range) and $\mathbf{x}_{si}$ holds the drivers on that day. Conditioning on one flood day per stratum eliminates α_s :

$$\mathcal{L}_s(\boldsymbol{\beta}) = \frac{\exp(\mathbf{x}_{s,\text{case}}^\top \boldsymbol{\beta})}{\sum_{j \in s} \exp(\mathbf{x}_{sj}^\top \boldsymbol{\beta})}, \quad (\text{S7})$$

a softmax over the days of one stratum. The estimator therefore never compares coastlines with each other, only a flood day with non-flood days at the same site and season, so spatial variation in reporting cannot bias $\boldsymbol{\beta}$. A single global model is fitted, with drivers added to a rainfall baseline in a fixed order:

$$G_0 : \text{rain}; \quad G_1 : \text{rain} + X; \quad G_2 : \text{rain} + M; \quad G_3 : \text{rain} + H + W_R.$$

$G_1 \rightarrow G_2$ replaces X by $X + M$ and so isolates the background M . All rungs are evaluated on the identical complete-case set.

Added information. Models are scored by the held-out conditional log-likelihood of Eq. (S7) in nats per stratum, using eight folds grouped by site, with predictors

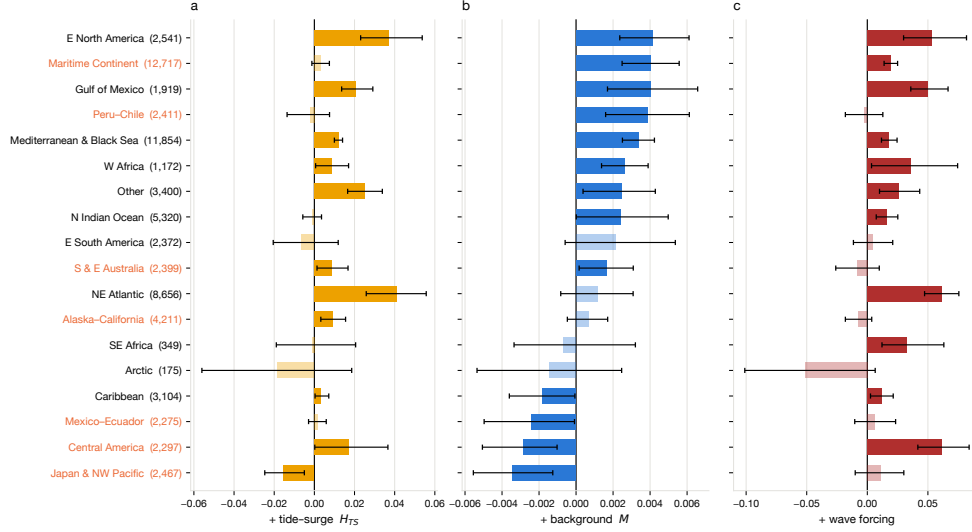
standardised on the training folds. A *nat* is a unit of information entropy, similar to a bit in binary but for natural logarithms. Here, models with no information will spread probability evenly over all nine days in a stratum with $\log 9 \simeq -2.2$ nats. Hence, by chance, a given matched set has a score of -2.185. An uninformative model spreads probability evenly over the nine days of a stratum and scores $-\ln 9 \approx -2.20$. Writing $\ell_s^{(G)}$ for the held-out log-likelihood of stratum s under model G , the added information of rung A over rung B is

$$\Delta_{A-B} = \frac{1}{S} \sum_{s=1}^S (\ell_s^{(A)} - \ell_s^{(B)}). \quad (\text{S8})$$

The log-likelihood is a strictly proper score and is evaluated out of sample, so $\Delta_{A-B} > 0$ only if the added driver sharpens the probability placed on the reported day given the drivers already present. An uninformative driver gives $\Delta \leq 0$ in expectation, and one whose sign is wrong for a coastline gives $\Delta < 0$ there. Increments are differenced within stratum and bootstrapped by site (2,000 resamples). Because a single global coefficient is fitted, regional values measure the information carried by the global relationship and are a lower bound on what a locally fitted model would recover.

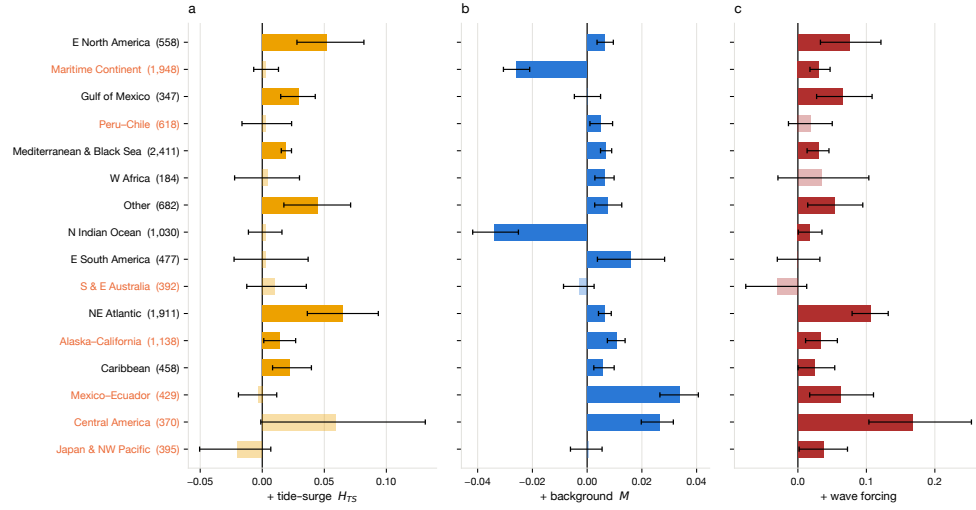
Uncertainty estimates: Site-clustered resampling. For the reported-flood analysis, coastlines are treated as independent rather than the individual flood reports. A replicate accordingly resamples *sites* with replacement and takes all of a drawn site's matched sets together. Resampling matched sets instead would treat that repetition as new information: measured directly on the same data, the stratum-level interval is 53% narrower in the all-months analysis and 59% narrower in the El Niño-restricted one — design effects on the interval width of $2.15\times$ and $2.46\times$ — which is enough to change which increments are reported as resolved. We use 2,000 resamples in both analyses.

Results. Pooled over all regions and years, each driver adds significant information: tide-surge $+0.0117$ [0.0091, 0.0143] nats per stratum, the background M $+0.0020$ [0.0014, 0.0025] and waves $+0.0232$ [0.0195, 0.0266]. By region (Supplementary Fig. S2), the background increment over all years is small everywhere ($|\Delta| \leq 0.004$) and of mixed sign. Restricting to strata whose case falls in an El Niño month (13,528 strata; Supplementary Fig. S3) resolves the ENSO sea-level dipole. On the eastern Pacific rim the background becomes strongly informative: $+0.034$ [0.027, 0.041] at Mexico–Ecuador, $+0.026$ in Central America and $+0.011$ at Alaska–California. In the west it carries information of the wrong sign under the global coefficient: -0.026 [-0.031 , -0.021] on the Maritime Continent and -0.034 [-0.042 , -0.025] in the north Indian Ocean, where El Niño depresses sea level and floods occur for other reasons. Peru–Chile is positive and only marginally resolved ($+0.005$ [$+0.001$, $+0.009$], 618 strata over a long coastline). The two eastern and western halves largely cancel in a pooled El Niño estimate, which is why results are reported by region. Because controls may come from the adjacent year and so from a different ENSO phase, the El Niño-subset contrast is largely El Niño against non-El Niño, and its magnitudes are not comparable with the all-years values. Supplementary Fig. S3 should be read for the pattern and Supplementary Fig. S2 for effect size.



Supplementary Figure S2 Information added by each flood driver, by region, 2000–2024. Held-out conditional log-likelihood gained per matched stratum (nats) on adding **a**, tide-surge X to rainfall; **b**, the background M ($X+M$); and **c**, wave forcing. Background from DUACS altimetry. Error bars are site-clustered 95% bootstrap intervals; faded bars span zero. Numbers beside region names are matched strata. Panels have separate x-axes because the wave increment is an order of magnitude larger than the background increment. Orange labels denote regions in the Pacific Basin.

In most regions waves carry the most information and tide-surge less; the background carries an order of magnitude less than waves outside El Niño. These relative magnitudes may partly reflect the dataset: larger, more destructive floods receive more coverage; the catalogue was designed for flash flooding rather than coastal flooding; and reporting density varies with digital access. With those caveats, all three drivers leave a measurable signature in reported floods, and the El Niño background leaves a regional fingerprint consistent with the extreme-water-level analysis.

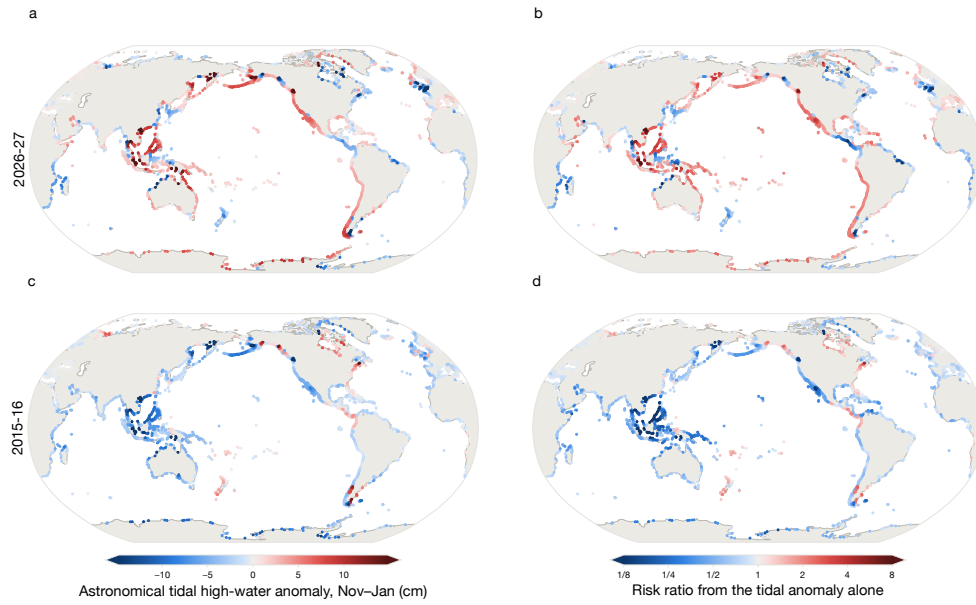


Supplementary Figure S3 As Supplementary Fig. S2, restricted to El Niño strata. Held-out conditional log-likelihood gained per matched stratum (nats) on adding **a**, tide-surge X to rainfall; **b**, the background M ($X + M$); and **c**, wave forcing. Only strata whose case falls in an El Niño month are kept; each stratum is kept whole. Rows are in the order of Supplementary Fig. S2, and regions with fewer than 100 strata are omitted. Because controls may fall in other ENSO phases, magnitudes are not comparable with Supplementary Fig. S2. Orange labels denote regions in the Pacific Basin.

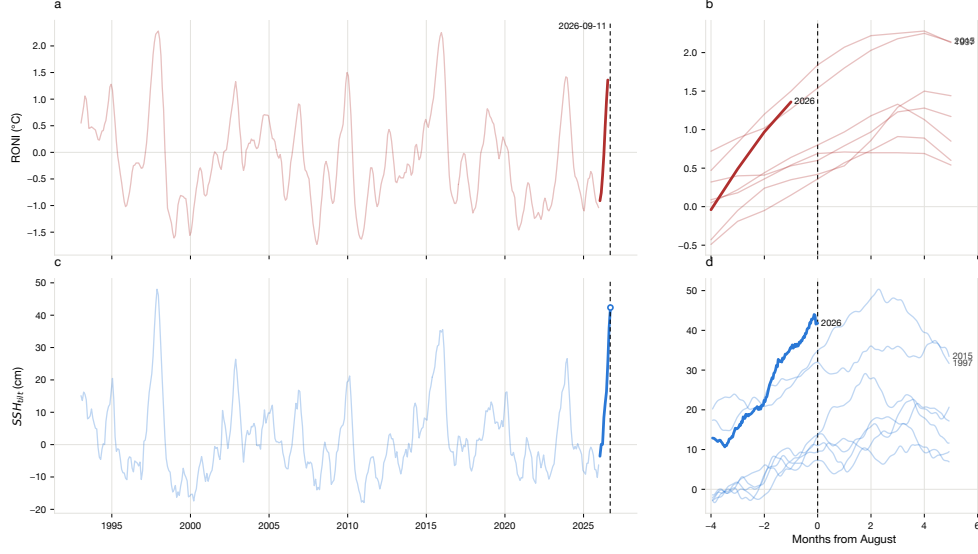
4 Astronomical tidal state in 2015–16 and 2026–27

Two astronomical cycles modulate high water on interannual time scales. The 18.6-year lunar nodal cycle acts in opposite senses on semidiurnal and diurnal coasts, so there is no single global nodal state. The 8.85-year cycle of lunar perigee, through its alignment with the syzygies and the solstices, gives a 4.4-year cycle in the high waters of any fixed season [9]. Both are carried by the tidal term of the outlook (Methods), which is evaluated for the November–January season in which El Niño peaks. Along the eastern Pacific rim the median November–January amplitudes are 3.0 cm for the nodal cycle, 3.6 cm for the 4.4-year perigean cycle and 0.2 cm for the 8.85-year cycle. GTSM’s own tide carries the same 4.4-year cycle (median amplitude 3.7 cm; site-by-site $r = 0.91$ with the synthesis).

Over November 2015–January 2016 the astronomical tide was near a minimum along most of the rim (median anomaly -1.9 cm), which on its own lowered exceedance risk at the 99th-percentile threshold along the US West Coast (median tidal risk ratio 0.57). In November 2026–January 2027 it is near a maximum ($+3.2$ cm; US West Coast 1.95; rim 1.47). The two fields are anticorrelated ($r = -0.71$; Supplementary Fig. S4), but not the near mirror images that the nodal cycle alone would give. Of the 2026–27 rim anomaly, about 1.3 cm is nodal and 1.7 cm perigean. At Mexico–Ecuador (issuable sites) the anomaly is small ($+0.5$ cm) and changes sign with the choice of season, from -2.8 cm for October–December to $+1.3$ cm for December–February, so the central risk ratio there is 10.9, 12.9, and 13.6 for the three seasons (12.2 with the calendar-year nodal cycle alone). Comparisons between the two events must carry the tide. The risk ratios shown are the tidal contribution in isolation, evaluated with no El Niño term, and should not be multiplied by the El Niño risk ratios of the main text.



Supplementary Figure S4 Astronomical tidal modulation of high water in 2026–27 and 2015–16. **a, b,** The astronomical (nodal and perigean) anomaly of the 99th-percentile high water for November 2026–January 2027 (**a**, cm) and the risk ratio it alone produces at the site-specific 99th-percentile threshold (**b**). **c, d,** The same for November 2015–January 2016. Each quantity shares one colour scale across both years; the risk-ratio scale is symmetric in $\log_2$ about 1.

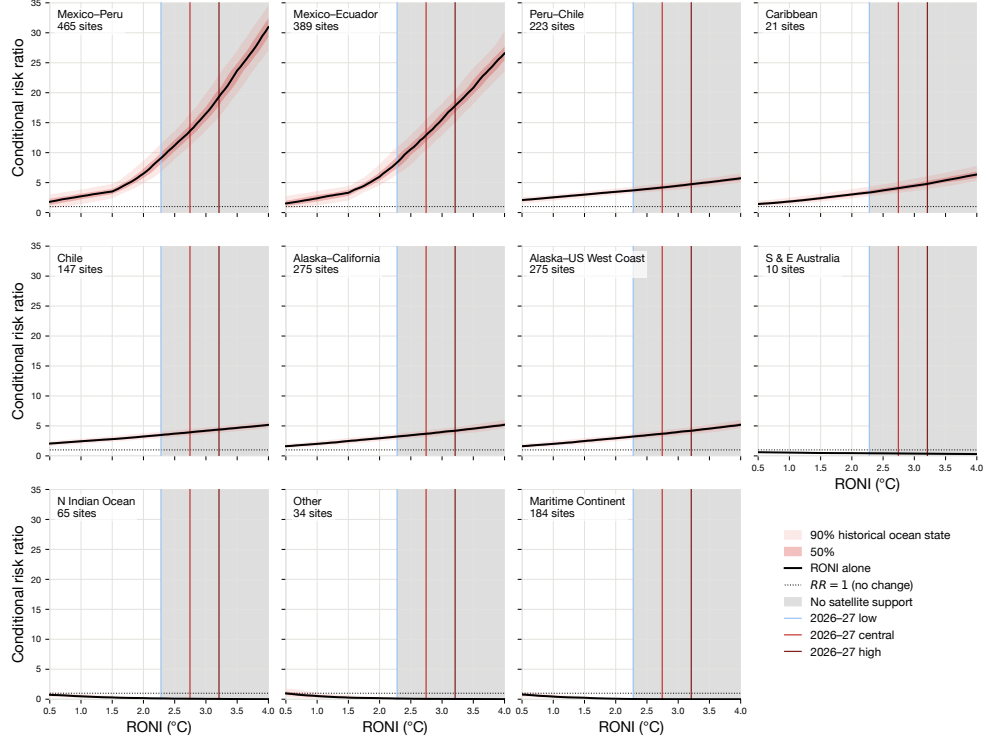


Supplementary Figure S5 RONI and SSH_{tilt} over the altimetry era. **a, c,** Monthly RONI (**a**; CPC official values, to June–August 2026, the latest published) and monthly mean SSH_{tilt} (**c**). Values before 2026 are faded. The dashed line marks the last day of the altimetry record, and the open circle in **c** the final, partial month. **b, d,** The same quantities for each satellite-era El Niño, aligned on August of the event year; 2026–27 is bold and 1997 and 2015 are named. SSH_{tilt} is the DUACS delayed-time product spliced to the near-real-time stream.

5 RONI and SSH_{tilt}

RONI and SSH_{tilt} rise and peak together across the altimetry era (Supplementary Fig. S5); across events, SSH_{tilt} peaks at or before RONI (median lead 0 months). The tilt integrates the wind forcing and so carries information about the ocean state that the contemporaneous SST index does not, and two events with the same RONI need not reach the same tilt (Supplementary Note 2). At the September anchor of the event year, the 2026 tilt already exceeds the mature peak of eight of the nine prior satellite-era El Niño events.

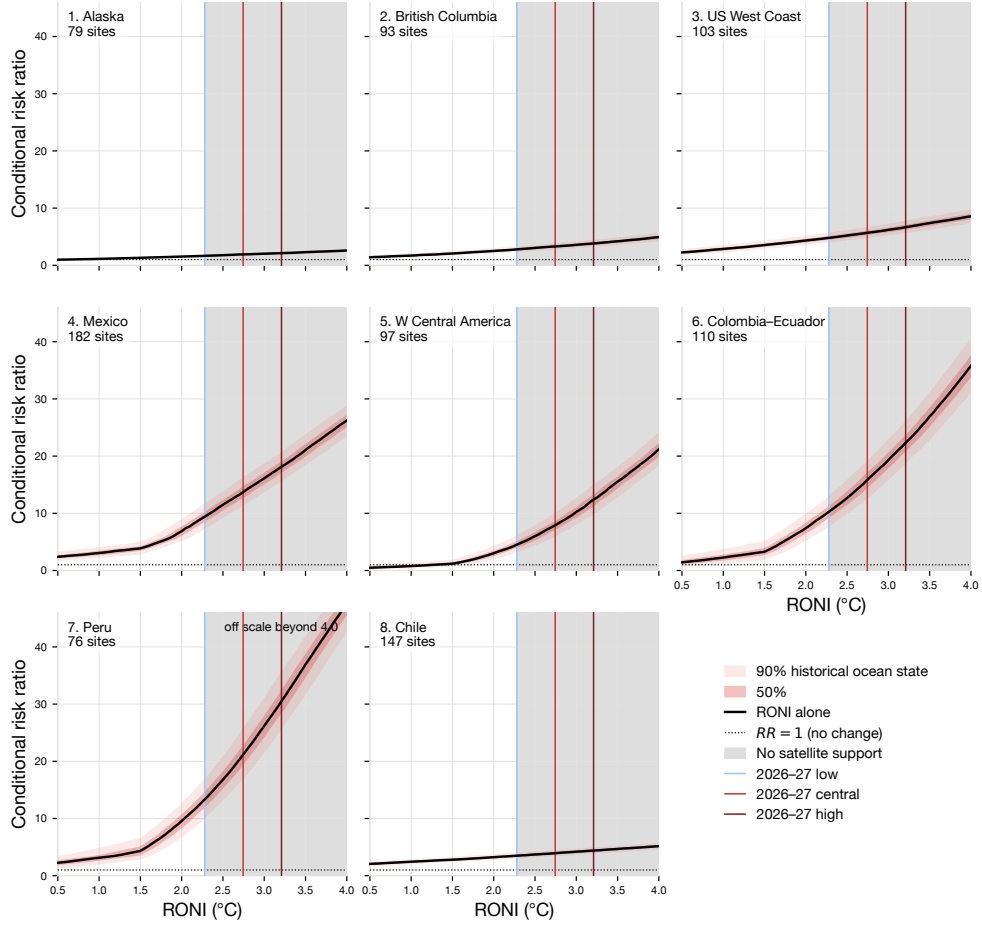
The tilt is the better predictor of the background. Over the satellite era (1993–2015, 276 months), the Mexico–Ecuador background correlates with SSH_{tilt} at $r = 0.92$ and with RONI at $r = 0.87$. In leave-one-event-out prediction over the eight El Niño events, a linear model in SSH_{tilt} has an RMSE of 2.24 cm, against 2.85 cm for a linear model in RONI and 2.67 cm for the hinged model (21% and 16% lower). Part of the apparent steepening of the response to strong events in RONI (Supplementary Note 8) may therefore reflect the ocean state that strong events tend to reach, rather than a threshold in SST amplitude. Because SSH_{tilt} is not operationally forecast, the projections of the main text are conditioned on RONI and carry the tilt as a source of spread (Supplementary Note 6).



Supplementary Figure S6 Conditional risk ratio against El Niño amplitude, by region. Each panel is Fig. 3c for one region. The black line is the amplitude-only projection; bands are the 5–95% and 25–75% ranges over ocean states drawn from $p(SSH_{tilt} | RONI)$. The dotted line marks $RR = 1$. Panels are ordered by the risk ratio at the central scenario and give each region’s site count. Vertical lines mark the three 2026–27 scenarios; grey shading marks amplitudes beyond the satellite-era maximum (2.28 °C). The blue bar in the western Pacific and north Indian Ocean panels marks where $RR < 0.01$ and has been clipped to the axis floor. Regions with fewer than ten sites are omitted.

6 Conditional risk projections for 2026–27

Supplementary Figs. S6 and S7 apply the construction of Fig. 3c region by region. The central line is the projection from RONI amplitude alone, including the November 2026–January 2027 astronomical tide; the bands are the spread induced by drawing the tropical Pacific ocean state from $p(SSH_{tilt} | A)$ at the event scale where A is the RONI amplitude. Because the conditional mean of SSH_{tilt} is linear in amplitude, the median of the bands tracks the central line by construction; only the spread is new. The spread is itself a result. It is wide on the eastern Pacific rim, where $\partial M / \partial SSH_{tilt}$ has a median of 0.23–0.34 cm per cm across the three rim regions of Supplementary Fig. S6 (issuable sites), and small in the Atlantic basins, where it is at most 0.10. The low scenario sits at the largest satellite-era amplitude (2.28 °C, 1997–98) and the central and high scenarios lie beyond it, so their values are extrapolations. Values at the three scenarios along the eastern Pacific rim are given in Supplementary Table S2.



Supplementary Figure S7 Conditional risk ratio against El Niño amplitude along the eastern Pacific. As Supplementary Fig. S6, with the eastern Pacific divided into eight along-shore regions, north to south. The regions nest the coarse scheme site for site: panels 1–3 pool to Alaska–California, 4–6 to Mexico–Ecuador and 7–8 to Peru–Chile. Bands are the 50% and 90% ranges over ocean states. Shading marks amplitudes beyond the satellite-era maximum.

Supplementary Table S2 Conditional risk ratio at the three 2026–27 scenarios along the eastern Pacific rim. Read from the curves of Supplementary Fig. S7 at the low, central and high scenario amplitudes (RONI of 2.28, 2.74 and 3.21 °C). RR is the amplitude-only projection; the band is the 90% range over historical ocean states conditional on that amplitude, not a sampling interval. The central and high amplitudes exceed the largest satellite-era value (2.28 °C). Regions run north to south and nest the three coarse eastern Pacific regions site for site.

Region	Sites	Low		Central		High	
		RR	90% range	RR	90% range	RR	90% range
Alaska	79	1.7	1.4–1.9	1.9	1.6–2.2	2.1	1.8–2.4
British Columbia	93	2.8	2.3–3.3	3.3	2.7–3.9	3.8	3.2–4.5
US West Coast	103	4.8	4.0–5.7	5.7	4.8–6.7	6.7	5.6–7.8
Mexico	182	9.4	7.2–11.8	13.7	11.4–16.3	18.1	15.6–20.7
W Central America	97	4.5	3.2–6.2	7.9	6.0–10.2	12.4	9.8–15.0
Colombia–Ecuador	110	10.3	7.7–13.1	15.8	12.6–19.2	22.3	18.6–26.2
Peru	76	13.3	10.0–17.1	21.2	16.9–25.8	30.5	25.7–35.7
Chile	147	3.5	3.0–4.0	3.9	3.4–4.5	4.4	3.8–4.9

7 Tidal-excursion durations

Supplementary Fig. S8 and Supplementary Table S3 report how the projected El Niño background shift changes the mean duration of a single tidal excursion at the 1,895 sites of the six Pacific-basin regions.

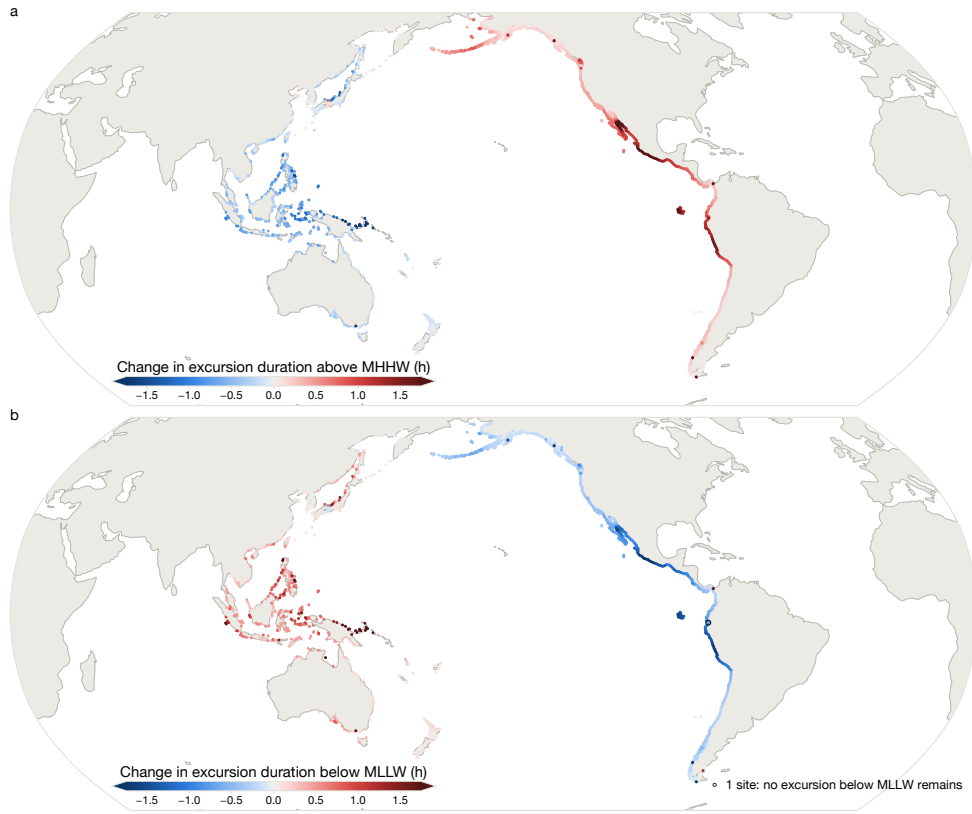
Series and datums. Durations are measured on the hourly total water level H over 1995–2014, approximately one full 18.6-year nodal cycle, so that nodal modulation averages out of the datums. Sites with fewer than 90% valid hours are excluded, and gaps of up to three hours are linearly interpolated. Mean higher high water (MHHW) is the mean, over tidal days of 24.84 h, of the highest local maximum in each day, with maxima at least six hours apart; mean lower low water (MLLW) is defined likewise from the minima. Datums and durations therefore share a vertical reference by construction. As an independent check, the datums agree with the GTSM annual tidal indicators for the same epoch at $r = 0.999$, with median offsets of -0.8 cm (MHHW) and $+2.8$ cm (MLLW).

Excursions. An excursion is a contiguous interval with H above MHHW or below MLLW. Entry and exit times are found by linear interpolation between the hourly samples either side of each crossing, which removes one-hour quantisation. Excursions touching the start or end of the record are discarded. A site’s duration is the mean over its excursions; a site is retained if it has at least ten excursions on each side, and all 1,895 qualify.

Projection. The shift at site j is $\delta_j = \Delta M_j(A)$ from Eq. (S9) at the central 2026–27 amplitude, $A = 2.74^\circ\text{C}$, under the regional knot rule (Supplementary Note 8). Raising the water level uniformly by δ_j is equivalent to lowering a fixed datum by δ_j , so projected durations are measured on the same series against $\text{MHHW} - \delta_j$ and $\text{MLLW} - \delta_j$, without resampling or refitting. At one Mexico–Ecuador site the shift lifts the series entirely above the lowered MLLW; it is counted as zero duration rather than treated as missing. Regional values are medians of per-site changes, which differ from the difference of regional median durations.

Supplementary Table S3 Change in the mean duration of a tidal excursion under the central 2026–27 background shift. Medians over each region’s sites (1995–2014 hourly water level). Shift is the per-site projected background change; base is the mean duration of a single excursion above MHHW or below MLLW. The median of the per-site changes is not the difference of the regional medians, so the baseline and the change are given rather than a projected duration.

Region	Sites	Shift (cm)	Range (m)	Above MHHW			Below MLLW		
				Base (h)	Δ (h)	Δ (%)	Base (h)	Δ (h)	Δ (%)
Mexico–Ecuador	491	+19.3	1.61	2.11	+0.76	+36	1.96	−0.74	−31
Peru–Chile	304	+7.1	1.25	2.39	+0.24	+10	2.02	−0.29	−14
Alaska–California	575	+9.4	2.45	2.31	+0.25	+9	2.57	−0.24	−8
Japan & NW Pacific	129	−1.1	1.36	2.42	−0.05	−2	2.80	+0.04	+2
S & E Australia	87	−3.4	1.61	2.08	−0.10	−5	2.05	+0.12	+6
Maritime Continent	309	−13.8	1.62	2.41	−0.50	−19	2.54	+0.50	+19



Supplementary Figure S8 Change in the mean duration of a single tidal excursion under the central 2026–27 background shift. a, Above mean higher high water (MHHW); **b,** below mean lower low water (MLLW), at 1,895 Pacific-basin sites, relative to the 1995–2014 hourly water level. The shift is the per-site warm-limb projection at $\text{RONI} = 2.74^\circ\text{C}$, hinged where the site’s region adopts the knot and linear otherwise. The open ring in **b** marks the one Mexico–Ecuador site where the shift (+25 cm) leaves no excursion below MLLW.

8 The knot in the warm-limb projection

The background response to El Niño in the tropical Pacific appears to steepen for strong events. The warm-limb response at site j is fitted in a linear and a hinged form,

$$\Delta M_j(A) = \alpha_j + \beta_j A \quad \text{and} \quad \Delta M_j(A) = \alpha_j + \beta_j A + \gamma_j \max(A - A_0, 0), \quad (\text{S9})$$

over the 408 months of 1950–2015 with $A \geq 0$, with the knot fixed a priori at the strong-event threshold $A_0 = 1.5^\circ\text{C}$. This note defines the rule that decides where the hinged form is used, and the tests of where the knot is placed (Supplementary Fig. S10), whether a knot is needed at all (Supplementary Table S6), and what removing it does to the projection (Supplementary Fig. S11).

γ is estimated from few months. Only 38 of the 792 months exceed A_0 (two more sit exactly on it). They come from seven El Niño events, and three of these — 1982–83, 1997–98 and 2015–16 — carry 73% of the weight on γ , measured as each event’s share of $\sum_t \max(A_t - A_0, 0)^2$. Every interval below is therefore an event-block bootstrap: whole El Niño events, whole La Niña events and single neutral years (69 blocks) are resampled with replacement, and the estimator is recomputed from scratch in each replicate.

Why the knot is used regionally

The obvious rule would adopt the knot at site j when a bootstrap interval on γ_j excludes zero. That rule is uncalibrated. The monthly background is spatially coherent far beyond the site spacing: the median nearest-neighbour distance is 40 km, and sites 100–200 km apart still correlate at $r = 0.98$ (Supplementary Table S4). The bootstrap resamples in time, one site at a time, so it represents the scarcity of strong events but not the redundancy between neighbours. One anomalous strong month moves γ_j along a whole coastline at once, and a per-site count records a single fluctuation as hundreds of detections.

Supplementary Table S4 Spatial coherence of the monthly background.
Median correlation between the monthly background series of site pairs, by separation, over 792 months at 3,197 sites. The median nearest-neighbour spacing is 40 km.

Separation (km)	Site pairs	Median r
0–25	358	1.000
25–50	984	0.998
50–100	2,293	0.994
100–200	5,595	0.980
200–500	11,068	0.953
500–1000	8,660	0.859

370 We measured this with a circular-shift null. For integer k the shifted index is

$$\tilde{A}_t^{(k)} = A_{(t-12k) \bmod T}, \quad (\text{S10})$$

371 where T is the record length in months. Shifting by whole years preserves both seasonal
 372 cycles, every spatial correlation among sites and the distribution of RONI, including
 373 the months above A_0 ; it destroys only the alignment between a month’s amplitude and
 374 its background. Across $k = 1, \dots, 12$ the shifted index correlates with the original at
 375 $|r| \leq 0.15$, so every detection on a shifted record is a false positive. Rerunning the full
 376 procedure (fit, 200-replicate bootstrap and decision rule) on each shifted record, the
 377 nominal 5% rule flagged a median of 7.2% of sites (range 2.8–17.2%); the worst shift
 378 declared 551 sites resolved, against 1,327 (41.5%) on the observed record. The sign
 379 of the detections also separates the observed record from the null: 92.2% of observed
 380 detections are positive, against a median of 40.9% and a maximum of 88.2% across
 381 shifts. A steepening is therefore supported by the agreement of many sites, not by any
 382 one of them, and a rule that chose the model site by site would be selecting on noise.

383 Regional grouping

384 Sites are grouped into 23 regions: eight along-shore regions on the eastern Pacific rim
 385 (Alaska, British Columbia, US West Coast, Mexico, West Central America, Colombia–
 386 Ecuador, Peru, Chile) and the coarse basin regions of Extended Data Fig. 1 elsewhere.
 387 The along-shore regions nest the coarse ones, so no site changes basin. For each region
 388 the composite $\bar{M}_r(t) = \text{median}_{j \in r} M_j(t)$ is formed and the hinged form of Eq. (S9) is
 389 fitted to it, giving one coefficient γ_r . The region adopts the knot when the 2.5th and
 390 97.5th percentiles of γ_r over 1,000 event-block replicates share a sign (i.e. excluding
 391 zero). Coefficients are still fitted per site; the regional fit determines only which form
 392 is used. Here, we use “adopt” to refer to the usage of the piecewise linear form over a
 393 given region.

394 The rule was calibrated before adoption on 24 circular shifts ($k = 1, \dots, 24$, 300
 395 replicates each). Under the null it adopts a median of one region and at most five;
 396 nine regions adopt on the observed record, more than under any shift (Supplementary
 397 Table S5, Supplementary Fig. S9). An alternative estimator, the median of the per-
 398 site γ_j within each region, was calibrated in the same way and rejected (it adopted five
 399 regions against a null maximum of five). Compositing before fitting removes site-level
 400 noise but not event-sampling uncertainty, which is the part that limits γ . The tropical
 401 eastern Pacific rim adopts coherently — Mexico, West Central America, Colombia–
 402 Ecuador and Peru — and reverts to a linear form poleward of it. Japan/NW Pacific
 403 and the Maritime Continent adopt a negative γ_r , a flattening above the threshold;
 404 the test is two-sided. The Arctic and Hudson Bay/Labrador are far noisier series and
 405 should be read with that in mind. Because the null adopts a median of one region,
 406 and no clear mechanism links the equatorial anomaly to these basins, the Arctic and
 407 Hudson Bay adoptions are probably spurious. They hold one issuable site between
 408 them, so they do not affect the outlook.

409 To check that no single event drives an adoption, each of the eight events that
 410 reach A_0 was withheld in turn and γ_r refitted. None of the 8×9 refits changes the sign

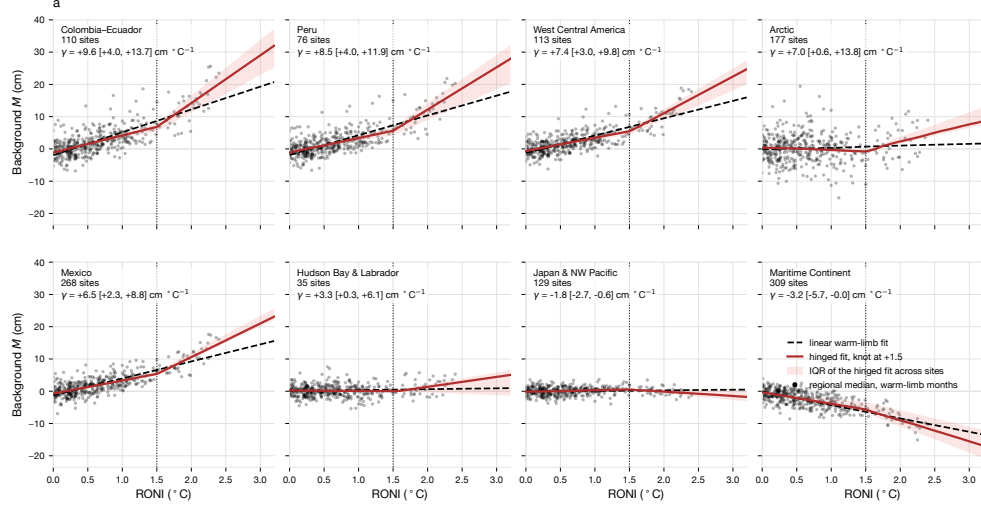
of an adopting region's coefficient; the largest change in magnitude is 37%, at Hudson Bay/Labrador when 1997–98 is withheld.

Supplementary Table S5 The regional knot rule. Knot coefficient γ_r of each region's composite background ($\text{cm } ^\circ\text{C}^{-1}$), fitted over the warm limb with $A_0 = 1.5^\circ\text{C}$, and its 95% event-block bootstrap interval (1,000 replicates). A region adopts the knot when the interval excludes zero. Scheme: along-shore (fine) regions on the eastern Pacific rim, basin (coarse) regions elsewhere. Under 24 circular-shift nulls the rule adopts a median of 1 region and at most 5.

Region	Scheme	Sites	γ_r	95% interval
<i>Adopt the knot</i>				
Colombia–Ecuador	fine	110	+9.6	[+4.0, +13.7]
Peru	fine	76	+8.5	[+4.0, +11.9]
West Central America	fine	113	+7.4	[+3.0, +9.8]
Arctic	coarse	177	+7.0	[+0.6, +13.8]
Mexico	fine	268	+6.5	[+2.3, +8.8]
Hudson Bay/Labrador	coarse	35	+3.3	[+0.3, +6.1]
Other (residual)	coarse	313	+1.4	[+0.2, +2.9]
Japan/NW Pacific	coarse	129	−1.8	[−2.7, −0.6]
W Pacific/Maritime	coarse	309	−3.2	[−5.7, −0.0]
<i>Do not adopt</i>				
US West Coast	fine	115	+6.0	[−0.1, +9.2]
NE Atlantic	coarse	157	+4.3	[−0.8, +7.9]
Alaska	fine	271	+3.8	[−4.3, +9.5]
British Columbia	fine	189	+3.4	[−1.9, +6.9]
Gulf of Mexico	coarse	39	+3.4	[−1.9, +9.5]
Chile	fine	228	+3.0	[−0.4, +4.8]
Caribbean	coarse	75	+1.2	[−0.2, +2.6]
W Africa	coarse	42	+1.0	[−0.5, +2.2]
SE Africa	coarse	45	+0.7	[−1.2, +2.3]
E North America	coarse	61	+0.0	[−3.0, +3.0]
E South America	coarse	44	−0.0	[−0.8, +0.9]
Australia S/E	coarse	87	−1.1	[−3.0, +0.9]
N Indian Ocean	coarse	124	−1.5	[−3.8, +1.3]
Mediterranean/Black Sea	coarse	190	−3.2	[−11.4, +3.0]

Sensitivity to knot placement

This and the next subsection use the Mexico–Ecuador composite (491 sites). The hinged form was refitted with A_0 on a grid from 0.60 to 2.20 $^\circ\text{C}$ in steps of 0.05 $^\circ\text{C}$. A candidate was admissible if at least eight months and three distinct El Niño events lay above it; 32 of the 33 grid points qualify. For each we record γ , the Akaike information criterion $\text{AIC} = n \ln(\text{SSE}/n) + 2p$ ($p = 4$ for the hinge, 3 for the linear fit), and the



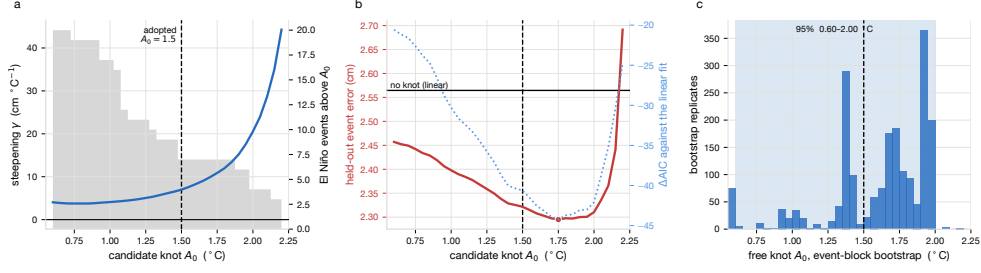
Supplementary Figure S9 The hinged and linear warm-limb fits in the regions that adopt the knot. Grey points are the regional composite background (median over all the region's sites), month by month, to which γ_r is fitted; γ_r and its 1,000-replicate event-block 95% interval are given in each panel. The dashed line is the median across sites of the linear fit, and the red line the median of the hinged fit ($A_0 = 1.5^\circ\text{C}$, marked), with its interquartile range across sites shaded. Eight of the nine adopting regions are shown, ordered by γ_r ; the residual class *Other* is omitted.

419 leave-one-event-out prediction error

$$\text{RMSE}_{\text{LOEO}} = \left[\frac{1}{n_E} \sum_e \sum_{t \in e} (\bar{M}_t - \hat{M}_t^{(-e)})^2 \right]^{1/2}, \quad (\text{S11})$$

420 where e runs over the 20 El Niño events, $\hat{M}^{(-e)}$ is fitted with every month of event
 421 e removed, and n_E is the number of warm-limb El Niño months. Holding out whole
 422 events prevents neighbouring months of one event from predicting each other. The
 423 knot was then left free: within each of 2,000 event-block replicates, A_0 was estimated
 424 by profile least squares over the admissible grid points, and the same free-knot model
 425 was scored by Eq. (S11) with A_0 re-estimated inside each training fold. Because AIC
 426 assumes independent residuals, the LOEO test provides an independent validation
 427 that is less affected by autocorrelation.

428 Over the Mexico–Ecuador region, γ is positive at every candidate, from +3.76 to
 429 +34.98 $\text{cm } ^\circ\text{C}^{-1}$, rising with A_0 as the months above the knot become fewer and more
 430 extreme (Supplementary Fig. S10a). In each of the nine adopting regions γ_r keeps
 431 its sign across the whole grid. On RONI the in-sample and out-of-sample criteria
 432 agree about location (Supplementary Fig. S10b): AIC and $\text{RMSE}_{\text{LOEO}}$ are both lowest
 433 at 1.75°C ($\Delta\text{AIC} = -44.3$ against the linear fit, -40.6 at 1.5°C ; $\text{RMSE}_{\text{LOEO}} =$
 434 2.295 cm). At the adopted knot $\text{RMSE}_{\text{LOEO}}$ is 2.322 cm , against 2.565 cm for the linear
 435 fit and 2.345 cm for the free knot. The free-knot bootstrap gives $A_0 = 1.70^\circ\text{C}$ with a
 436 95% interval of $0.60\text{--}2.00^\circ\text{C}$ (Supplementary Fig. S10c), $\gamma = +10.46 [+2.52, +27.02]$



Supplementary Figure S10 Sensitivity to the placement of the knot, Mexico–Ecuador composite. **a**, Knot coefficient γ at each of the 32 admissible candidate knots (blue); grey bars (right axis) are the number of El Niño events with months above the knot. **b**, Leave-one-event-out RMSE (red, Eq. S11; minimum marked) and in-sample ΔAIC relative to the linear fit (dotted blue, right axis) against knot location; the black line is the linear fit's RMSE. **c**, Knot location re-estimated freely within each of 2,000 event-block replicates; shading marks the 95% interval (0.60–2.00 °C). The dashed line marks the adopted $A_0 = 1.5^\circ\text{C}$.

437 and a slope ratio $(\beta + \gamma)/\beta$ of 3.37 [1.64, 7.14]. Note the lower-bound of the 95%
 438 interval (0.60 °C) is set by the grid edge. The data therefore support a steepening but
 439 do not locate it precisely; $A_0 = 1.5^\circ\text{C}$ is a convention the data are compatible with,
 440 not a threshold they resolve.

441 Sensitivity to the functional form

442 The hinged fit prescribes the shape of the steepening: two straight lines that meet at
 443 A_0 . A steepening found this way could be an artefact of that choice. We therefore refit-
 444 ted the relationship between RONI amplitude, A , and the Mexico–Ecuador composite
 445 background, $\bar{M}(A)$, with smooth curves that contain no breakpoint and are free to
 446 steepen, flatten or stay straight. We use this to evaluate whether steepening appears
 447 when the model is not told where, or whether, to bend; and whether it arises within
 448 individual events or from differences between events.

449 **Measuring steepening.** To compare curves of different shapes, each is sum-
 450 marised by how much steeper it is in strong months than in moderate ones. With
 451 $S(A) = d\bar{M}/dA$ the slope of the fitted curve,

$$\Delta S = \frac{1}{|\mathcal{T}_s|} \sum_{t \in \mathcal{T}_s} S(A_t) - \frac{1}{|\mathcal{T}_m|} \sum_{t \in \mathcal{T}_m} S(A_t), \quad \mathcal{T}_s = \{t : A_t > 1.5\}, \quad \mathcal{T}_m = \{t : 0.5 < A_t < 1.5\}, \quad (\text{S12})$$

452 the average slope over the observed strong months minus that over the observed mod-
 453 erate months, in $\text{cm } ^\circ\text{C}^{-1}$. A straight line gives $\Delta S = 0$; for the hinge, $\Delta S = \gamma$ exactly,
 454 so every model is on the scale of the hinge coefficient. Averaging over observed months
 455 evaluates each slope where there are data to support it.

456 **Models without a breakpoint.** We fitted restricted cubic splines [10] with 3, 4, 5
 457 and 6 spline knots, and a quadratic in A . The spline knots are placed at fixed quantiles
 458 of the warm-limb A_t (for four knots, the 5th, 35th, 65th and 95th percentiles: 0.04,
 459 0.33, 0.78 and 1.87°C), so their positions are not chosen by us. Beyond the outermost

460 knot, the curve is constrained to be straight, so it cannot invent curvature in the
 461 sparse strong-event tail. As such, the cubic spline will yield conservative estimates if
 462 steepening does exist beyond this point.

463 **Within or between events.** A steeper background in strong months can arise
 464 in two ways: within an event, the background rises faster as that event strengthens;
 465 or between events, strong events carry a larger background throughout. To separate
 466 them, each model was also fitted with one constant per event (69 blocks of events and
 467 neutral years). These event effects absorb each event’s average level, so the curve is
 468 estimated only from how the background changes as amplitude rises and falls within
 469 events. This is a strict test rather than a neutral one: the fitted event effects correlate
 470 with event peak amplitude ($r = +0.46$), so they remove part of the between-event
 471 contrast that the steepening describes. Both versions are reported, with 95% intervals
 472 from 2,000 event-block bootstrap replicates.

473 **Bayesian validation.** As an independent check with different assumptions, a
 474 hierarchical Bayesian version of the four-knot spline was fitted to the three along-shore
 475 subregions of Mexico–Ecuador (Mexico, West Central America, Colombia–Ecuador;
 476 1,224 subregion-months). Each subregion has its own curve, drawn towards a shared
 477 curve so that the three inform one another:

$$\bar{M}_{r,t} = \alpha_r + \mathbf{b}(A_t)^\top \boldsymbol{\theta}_r + u_{e(t)} + \varepsilon_{r,t}, \quad \varepsilon_{r,t} \sim \mathcal{N}(0, \sigma^2), \quad (\text{S13})$$

478 where $\mathbf{b}$ is the spline basis (columns scaled to unit variance); $\boldsymbol{\theta}_r = \boldsymbol{\theta}_0 + \tau_\theta \mathbf{z}_r$ and
 479 $\alpha_r = \mu_\alpha + \tau_\alpha z_r$ are the subregional curve and offset, where $z_r \sim \mathcal{N}(0, 1)$ and τ_α is the
 480 between-subregion standard deviation, so that $\alpha_r \sim \mathcal{N}(\mu_\alpha, \tau_\alpha^2)$; $\boldsymbol{\theta}_r$ is defined likewise,
 481 with a single scale τ_θ shared by all spline coefficients. $u_e = \sigma_u z_e$ is the optional event
 482 effect with $z_e \sim \mathcal{N}(0, 1)$.

483 Priors are $\mathcal{N}(0, 10^2)$ on μ_α and $\boldsymbol{\theta}_0$ and half-normal with scale 5 on τ_α , τ_θ , σ_u and
 484 σ . The model was sampled with the No-U-Turn Sampler[11] in NumPyro[12] (four
 485 chains, 1,500 warm-up and 2,000 retained draws each), and ΔS was computed from the
 486 shared curve on every draw. Sampling was well converged (Gelman–Rubin convergence
 487 diagnostic $\hat{R} \leq 1.009$, effective sample size ≥ 516 , two divergent transitions in 8,000
 488 draws in each full fit). Varying the prior scales 25-fold, from (2, 1) to (50, 25), moves ΔS
 489 between +3.23 and +3.49 without event effects, and between +2.06 and +2.32 with
 490 them. Because residuals are autocorrelated (lag-1 $r = +0.28$ in the Mexico subregion),
 491 the credible intervals are likely somewhat too narrow.

492 **Results.** Every model finds a steepening when fitted to the data as they stand
 493 (Supplementary Table S6). ΔS is +1.68 to +5.02 cm °C^{−1} for the splines and the
 494 quadratic, and +3.44 [+1.65, +5.19] for the Bayesian model, against +6.99 for the
 495 hinge. The steepening is therefore not an artefact of the breakpoint, although the
 496 hinge gives the largest estimate. It is not carried by one event: withholding each of
 497 the seven events with strong months in turn leaves the Bayesian ΔS between +2.80
 498 and +3.94, and +2.80 without 1997–98. With event effects, ΔS falls to between +1.0
 499 and +2.6 for the splines and the quadratic (+3.7 for the hinge), and the Bayesian
 500 estimate to +2.29 [+0.59, +4.01], still resolved. Part of the steepening is therefore
 501 between events — strong El Niños carry a disproportionately large background — but
 502 a substantial part persists within events, where the background also rises faster as an

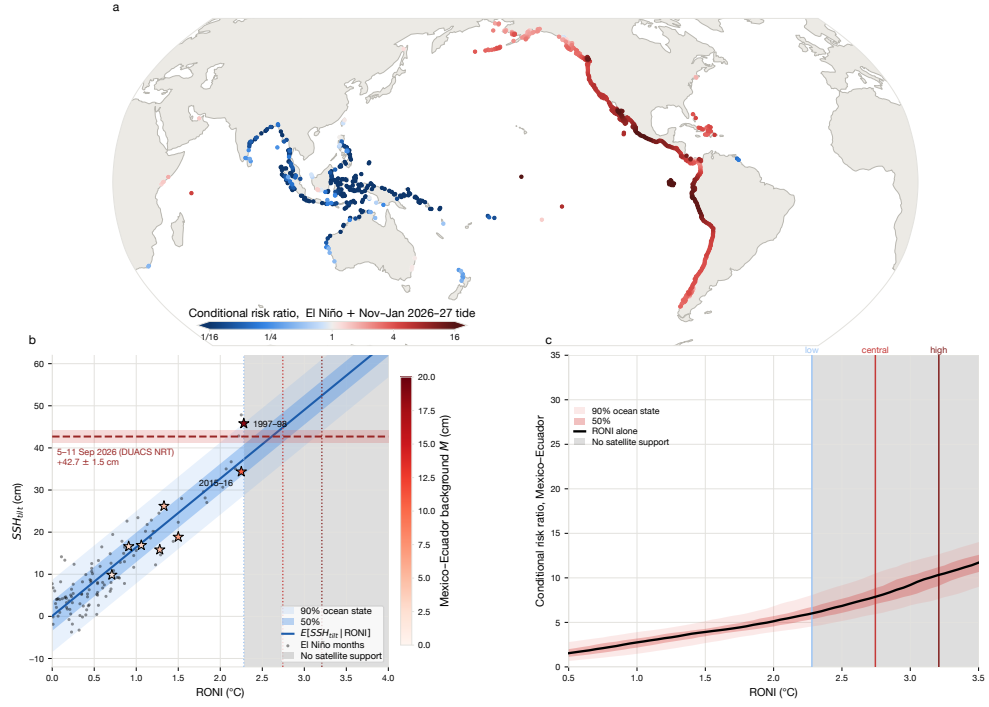
event strengthens. The between-event part is consistent with strong events reaching
a different ocean state (Supplementary Note 5), which RONI alone does not capture.
With Niño-3.4, the within-event part was not resolved (Supplementary Note 12).

Supplementary Table S6 Steepening in models without a breakpoint. ΔS ($\text{cm } ^\circ\text{C}^{-1}$) is the mean slope of the fitted curve, dM/dA , over observed months with $A > 1.5^\circ\text{C}$ minus the mean over $0.5 < A < 1.5^\circ\text{C}$ (Supplementary Eq. S12); for the hinge it equals γ . Mexico–Ecuador composite, 408 warm-limb months. Intervals and $P(\Delta S > 0)$ from 2,000 event-block bootstrap replicates. “With event effect” adds one constant per event block (69), so that the curve is estimated from within-event variation only. [†]Partially pooled over the three along-shore subregions; 95% credible interval and posterior probability (Supplementary Eq. S13).

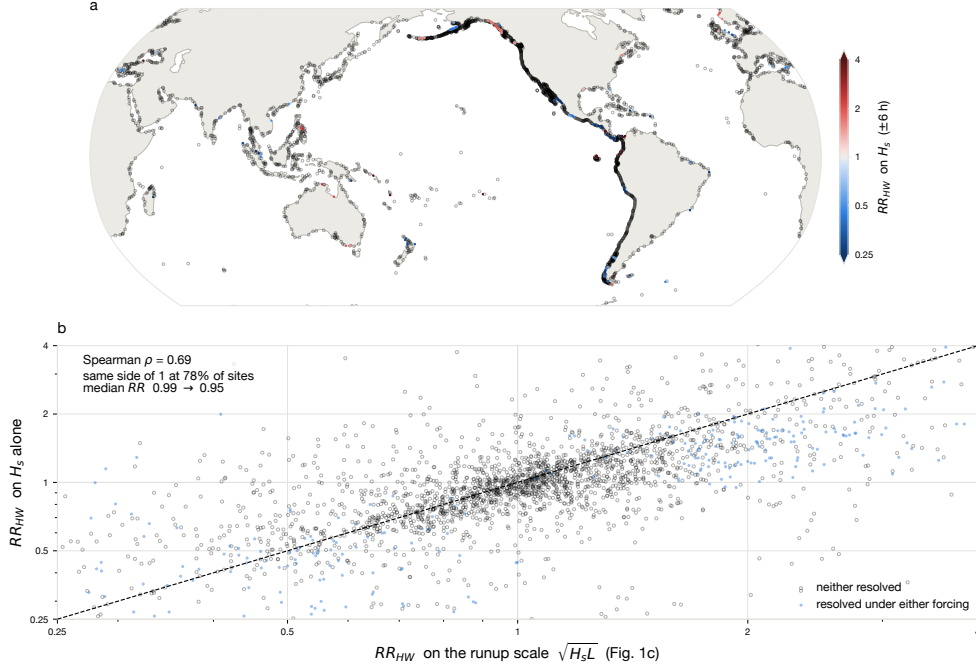
Model	No event effect			With event effect		
	ΔS	95% interval	P	ΔS	95% interval	P
Hinge, $A_0 = 1.5^\circ\text{C}$ (adopted)	+6.99	[+2.95, +9.43]	1.00	+3.71	[+0.37, +6.09]	0.98
Restricted cubic spline, 3 knots	+1.68	[+0.60, +2.47]	1.00	+1.01	[+0.26, +1.74]	1.00
Restricted cubic spline, 4 knots	+3.15	[+0.25, +5.10]	0.98	+2.05	[−0.22, +3.99]	0.96
Restricted cubic spline, 5 knots	+4.72	[+1.28, +6.80]	0.99	+2.33	[+0.16, +4.31]	0.98
Restricted cubic spline, 6 knots	+5.02	[+1.72, +7.08]	1.00	+2.43	[+0.01, +4.32]	0.98
Quadratic	+4.16	[+1.44, +5.94]	1.00	+2.60	[+0.71, +4.15]	1.00
Bayesian hierarchical spline, 4 knots [†]	+3.44	[+1.65, +5.19]	1.00	+2.29	[+0.59, +4.01]	0.99

Linear counterfactual

Supplementary Fig. S11 is Fig. 3 with the linear form of Eq. (S9) at every site. Because the linear fit has the larger slope at Mexico–Ecuador ($\beta = 5.55$ against $4.24 \text{ cm } ^\circ\text{C}^{-1}$ for the hinged fit), the two projections cross, the linear one being higher below about 1.7°C and lower above. At the low, central, and high 2026–27 scenarios, the knot raises the Mexico–Ecuador shift by 27%, 42% and 52%. Because the evidence above supports a steepening but does not fix its form, the main text also reports the linear projection, which is the more conservative of the two at the 2026–27 scenarios.



Supplementary Figure S11 Conditional risk projection for 2026–27 with a linear response to RONI. As Fig. 3, but with the linear warm-limb fit at every site in place of the regional knot rule. Ocean-state conditioning, astronomical tidal contributions, exceedance curves, thresholds and site screening are unchanged. Panel **b** does not depend on the response model and is identical to Fig. 3b. At the central scenario (2.74 °C) the Mexico–Ecuador background shift is 13.6 cm, against 19.3 cm under the knot rule, and the conditional risk ratio 7.9 against 12.9.



Supplementary Figure S12 Wave-high-water coincidence with large waves defined on H_s alone. **a**, As Fig. 1c, with a large wave defined as H_s above its neutral 95th percentile rather than the run-up scale $\sqrt{H_s L}$. Unresolved sites, including those at which no El Niño extreme high water is accompanied by large waves, are muted. **b**, RR_{HW} under the two definitions, site by site, with the 1:1 line; blue points are resolved under either definition.

9 Wave-high-water coincidence using H_s

Fig. 1c scores a large wave against the neutral 95th percentile of a deep-water run-up scale, $\sqrt{H_s L}$ with $L = gT^2/2\pi$, which includes mean wave period T as well as height. Supplementary Fig. S12 repeats the analysis with “large” defined on H_s alone, keeping the same sites, high-water trials, per-site neutral threshold rule, calendar-month matching and event-block bootstrap. Were T constant at a site, the two definitions would select the same trials, so any difference reflects wave period. The rank structure is largely preserved (Spearman $\rho = 0.69$), RR_{HW} falls on the same side of 1 at 78% of sites, and the median changes only from 0.99 to 0.95. Individual sites move by tens of percent, but Alaska–California stays above 1 (1.24; 1.05 with H_s) and the Maritime Continent and north Indian Ocean stay below it under both definitions. The run-up scale is retained in the main text because it is the more physically relevant forcing for run-up.

10 Dependence of wave risk on El Niño amplitude

The background shift is projected to 2026–27 through its dependence on RONI amplitude. We asked whether wave risk admits the same treatment, for two quantities: RR_W , the share of hours in which the run-up forcing exceeds its neutral 95th percentile, which describes the wave field alone; and RR_{HW} , the wave–high-water coincidence of Fig. 1c. Both were computed for each region and ENSO year (July–June) over November–March, which spans the ENSO peak and the boreal storm season and gives every year the same seasonal mix, so no calendar-month matching is needed. Each was regressed on the year’s peak monthly RONI, the amplitude scale of the 2026–27 scenarios, as $\ln RR = a + bA$ over the warm-limb years of 1979–2015. Six regions, fixed in advanced with the pooled eastern Pacific rim.

A significant slope shows that risk rises with amplitude on average, but not that amplitude is useful for projection, since the slope may only separate El Niño years from the rest. Each fit was therefore also scored by leave-one-event-out prediction against the mean of the other El Niño years where $\text{skill} = 1 - SSE_{\text{fit}}/SSE_{\text{null}}$, where the null predicts each El Niño year by the mean of the other El Niño years; positive skill means an event’s size adds information once the event is known to be underway.

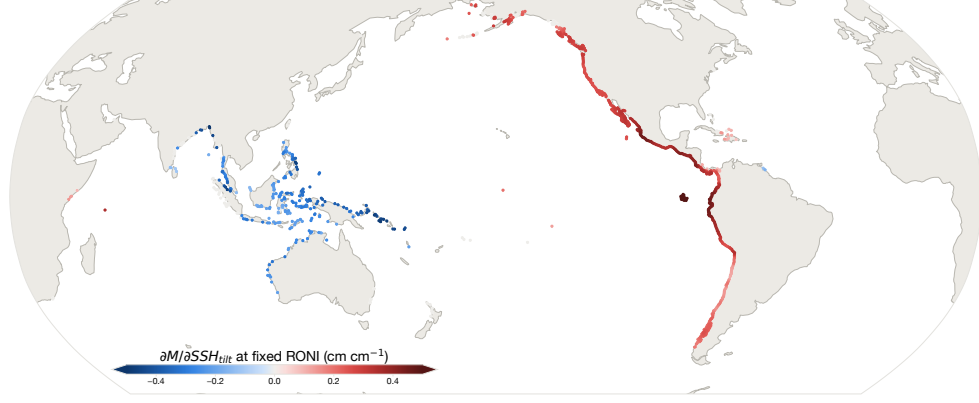
Wave intensity scales with amplitude on the eastern Pacific rim (Supplementary Table S7). At Mexico–Ecuador RR_W rises by a factor of about 1.5 per $^{\circ}\text{C}$ ($b = +0.38$ [$+0.19, +0.55$]), to 1.6 at $A = 2.0^{\circ}\text{C}$, and event size carries substantial skill ($+0.48$): the two largest events, 1982–83 and 1997–98, gave RR_W of 2.0 and 2.3, against a median of 1.05 for the five events below 1.4°C . At Alaska–California and in southern and eastern Australia RR_W also rises ($b = +0.21$ and $+0.19$), but event size adds nothing once an event is underway (skill -0.20 and -0.38). No slope is resolved in the western Pacific, the north Indian Ocean or eastern South America.

The coincidence scales with amplitude at Mexico–Ecuador ($b = +1.03$ [$+0.43, +1.59$], skill $+0.46$): 1982–83 and 1997–98 gave RR_{HW} of 5.2 and 4.3, against 0.6–1.2 in the four events below 1.4°C other than 1987–88. At Alaska–California the slope is barely resolved ($b = +0.14$ [$+0.01, +0.30$]) and event size adds almost nothing (skill $+0.04$): 1994–95, at 1.3°C , gave 1.8, whereas 1982–83, at 2.4°C , gave 0.9. Southern and eastern Australia also scales ($b = +0.47$ [$+0.06, +0.83$], skill $+0.29$), eastern South America declines ($b = -0.32$ [$-0.66, -0.09$]), and elsewhere, including the pooled eastern Pacific rim, the slope is indistinguishable from zero.

At Mexico–Ecuador both wave intensity and coincidence rise with amplitude and event size carries skill for both; elsewhere neither would support an amplitude-conditioned projection analogous to that of the background. Neither is projected to 2026–27. The central and high scenarios exceed the largest calibrating event (2.40°C , 1982–83), the wave record contains ten El Niño events and ends partway through 2015–16, and a log-linear fit grows rapidly beyond its data.

Supplementary Table S7 Dependence of wave and coincidence risk on El Niño amplitude. Log-linear fit, $\ln \text{RR} = a + bA$, of each region's risk ratio in November–March of each ENSO year on that year's peak monthly RONI, A , 1979–2015. RR_W : share of hours with run-up forcing above its neutral 95th percentile. RR_{HW} : the wave–high-water coincidence of Fig. 1c. Slope b (per $^{\circ}\text{C}$) with 95% event-block bootstrap interval (2,000 resamples). Skill: leave-one-event-out, against the mean of the other El Niño years (10 El Niño years); negative values mean that an event's size adds nothing once the event is known to be underway. RR at $A = 2.0^{\circ}\text{C}$, inside the calibration range (maximum 2.40°C). Regions are a fixed set, chosen in advance and not re-selected on the results shown; each region's share of resolved Fig. 1c sites is given in the accompanying data file.

Region	b	RR_W (wave intensity)			$\text{RR}_{2.0}$	b	RR_{HW} (coincidence)			$\text{RR}_{2.0}$
		95% interval	Skill				95% interval	Skill		
Alaska–California	+0.21	[+0.06, +0.36]	−0.20	1.45	+0.14	[+0.01, +0.30]	+0.04	1.28		
E South America	−0.00	[−0.19, +0.19]	−0.01	0.95	−0.32	[−0.66, −0.09]	−0.11	0.52		
Maritime Continent	−0.09	[−0.22, +0.10]	−0.06	0.81	+0.07	[−0.22, +0.41]	−0.07	1.23		
Mexico–Ecuador	+0.38	[+0.19, +0.55]	+0.48	1.58	+1.03	[+0.43, +1.59]	+0.46	2.87		
S & E Australia	+0.19	[+0.03, +0.37]	−0.38	1.27	+0.47	[+0.06, +0.83]	+0.29	1.34		
N Indian Ocean	+0.06	[−0.12, +0.27]	+0.02	1.07	+0.05	[−0.24, +0.37]	+0.01	0.99		
Eastern Pacific rim (pooled)	+0.21	[+0.09, +0.33]	+0.35	1.38	+0.06	[−0.11, +0.29]	−0.12	1.09		



Supplementary Figure S13 Sensitivity of the background to the equatorial tilt at fixed RONI. Site-wise partial slope $\partial M / \partial SSH_{tilt}$ (cm per cm) from the regression of the seasonal-free background on RONI and SSH_{tilt} over the warm-limb months of 1993–2015, at the 1,209 sites where the 2026–27 outlook is issued. Sites whose 95% event-block bootstrap interval spans zero are shown in pale grey.

11 Sensitivity of the background to the equatorial tilt

Supplementary Fig. S13 shows, for each site, the change in the seasonal-free background per centimetre of SSH_{tilt} when RONI is held fixed. It is the coefficient c_j of the regression $M_j(t) = a_j + b_j A_t + c_j SSH_{tilt}(t) + \varepsilon_j(t)$ over the warm-limb months ($A_t \geq 0$) of 1993–2015, and it is the sensitivity through which the ocean-state range of Fig. 3b,c enters the projection: each draw of SSH_{tilt} from its distribution at fixed RONI shifts the background at site j by c_j times the departure of that draw from the conditional mean. Of the 1,209 sites at which the 2026–27 outlook is issued, 1,084 have a 95% event-block bootstrap interval on c_j that excludes zero.

The slope is positive along the eastern Pacific boundary (median +0.34 cm per cm at Mexico–Ecuador, +0.26 at Alaska–California and +0.23 at Peru–Chile) and negative across the Maritime Continent and the north Indian Ocean (−0.28 and −0.28). It reproduces the El Niño background anomaly of Fig. 1a (pattern correlation $r = 0.98$ across the 1,209 sites), and its magnitude is largest near the equator and decreases poleward ($r = -0.46$ between $|c_j|$ and $|\text{latitude}|$). At a given El Niño amplitude, the equatorial ocean state therefore scales the existing background response rather than redistributing it.

12 Sensitivity to the choice of ENSO index

Every result in this study classifies and scales El Niño with the official Relative Oceanic Niño Index (RONI) of the NOAA Climate Prediction Center [13, 14]. Two properties motivate the choice. The operational outlook for 2026–27 is issued on RONI, so the scenario amplitudes enter the projection exactly as published; with any other index they must first be converted, which adds a step that is itself uncertain. And because RONI is measured relative to the tropical mean, it needs no detrending. A conventional Niño-3.4 index has to be detrended so that secular warming does not reclassify late years as warm, but a trend fitted over one period drifts once the index is used outside it: against RONI, Niño-3.4 detrended over 1950–2015 runs on average 0.17 °C low over 1950–1979 and 0.20 °C high over 2016–2024.

To measure how much the results depend on this choice, the entire analysis was repeated, by identical code from the ENSO classification onward, with the Niño-3.4 index from ERSSTv5 [15], with anomalies relative to 1950–2015, linearly detrended over the same period and smoothed with a three-month running mean. The thresholds, the persistence rule and the knot were unchanged. With Niño-3.4, the CPC outlook was first converted, by regressing Niño-3.4 on RONI over 1950–2015 ($\text{Niño-3.4} = -0.08 + 1.02 \text{ RONI}$, $r = 0.97$ – coefficients rounded), which places the three scenarios at 2.25, 2.73 and 3.20 °C.

The two indices correlate at $r = 0.97$ over 1950–2015 and assign the same state to 88% of months. They identify the same number of El Niño events (20), with boundaries that differ by a month or two. Most disagreements are weak La Niña months under Niño-3.4 that RONI classes as neutral, concentrated before 1980, where the detrended index runs low. Eight events are strong on RONI and nine on Niño-3.4 (1986–87 peaks at +1.46 °C on RONI), and the two rank the largest events differently: 1982–83 is the largest on RONI (+2.40 °C) and 2015–16 on Niño-3.4 (+2.54 °C), while 1997–98 and 2015–16 are almost tied on RONI (2.28 and 2.25 °C).

Supplementary Table S8 compares the key results throughout the manuscript. The historical risk ratios and their attribution are nearly unchanged: along the eastern Pacific rim the regional RR_{EE} differs by about 5% or less and f_M by at most 0.11, and the median background share over all sites is 0.84 against 0.74. The Caribbean and Gulf of Mexico responses are weaker with RONI and are no longer resolved at the regional level. The warm-limb steepening at Mexico–Ecuador is stronger with RONI ($\gamma = 7.0$ against 4.9 cm °C⁻¹), nine regions adopt the knot rather than eight, and part of the steepening persists within events, which it did not with Niño-3.4 (Supplementary Note 8). The 2026–27 outlook for Mexico–Ecuador is correspondingly higher, a central risk ratio of 12.9 (10.4–15.6) against 10.6 (8.1–14.1), almost entirely because of the steeper response: the scenario amplitudes differ by only 0.01–0.03 °C between the two routes, and with a purely linear response the two indices give 7.9 and 7.8.

Given this, none of the conclusions of the main text depends on the index. With either, El Niño raises the risk of extreme high water along the eastern Pacific rim and lowers it across the Maritime Continent, mostly through the background sea level; the background response steepens for strong events; and the 2026–27 outlook for Mexico–Ecuador is an increase in exceedance risk of about an order of magnitude.

Supplementary Table S8 Sensitivity of the main results to the ENSO index. Each row is computed twice by the identical pipeline: with CPC’s official Relative Oceanic Niño Index (RONI; used throughout this study) and with a conventional Niño-3.4 index (ERSSTv5, 1950–2015 base, linearly detrended over that period, three-month running mean). Both use the same thresholds ($\pm 0.5^\circ\text{C}$, five consecutive months; strong $\geq 1.5^\circ\text{C}$) and the same knot, $A_0 = 1.5^\circ\text{C}$. The two indices correlate at $r = 0.974$ over 1950–2015 and assign the same ENSO state to 88% of months. ^aWith Niño-3.4, the CPC outlook (issued on RONI) must first be converted, here by regressing Niño-3.4 on RONI over 1950–2015; with RONI it is used as issued. Intervals are as in the main text.

	RONI (this study)	Niño-3.4
<i>ENSO classification, 1950–2015</i>		
El Niño / neutral / La Niña months	199 / 433 / 160	196 / 394 / 202
El Niño events (strong, $\geq 1.5^\circ\text{C}$)	20 (8)	20 (9)
Largest event ($^\circ\text{C}$)	1982–83, +2.40	2015–16, +2.54
<i>Historical exceedance risk (Figs. 1–2)</i>		
Sites with RR_{EE} resolved above / below	1,264 / 316	1,313 / 298
1		
Mexico–Ecuador: $\text{RR}_{EE} (f_M)$	2.33 (0.99)	2.45 (0.97)
Alaska–California: $\text{RR}_{EE} (f_M)$	1.55 (0.71)	1.61 (0.67)
Peru–Chile: $\text{RR}_{EE} (f_M)$	1.70 (0.74)	1.78 (0.63)
Maritime Continent: $\text{RR}_{EE} (f_M)$	0.53 (1.14)	0.60 (1.22)
Median background share f_M , all sites	0.84	0.74
<i>Warm-limb response, Mexico–Ecuador</i>		
Slope below / above A_0 ($\text{cm } ^\circ\text{C}^{-1}$)	4.2 / 11.2	4.7 / 9.6
Knot coefficient γ [95%]	7.0 [2.7, 9.4]	4.9 [0.6, 9.2]
Steepening ΔS ($\text{cm } ^\circ\text{C}^{-1}$), Bayesian spline, without / with event effects [95%]	+3.44 / +2.29 [+0.59, +4.01]	+2.76 / +0.23 [−1.60, +2.04]
Regions adopting the knot (sites)	9 of 23 (1,530)	8 of 23 (1,343)
<i>2026–27 outlook, Mexico–Ecuador</i>		
Scenario amplitude, low / central / high ($^\circ\text{C}$)	2.28 / 2.74 / 3.21	2.25 / 2.73 / 3.20 ^a
Background shift, central (cm)	19.3	17.6
Conditional RR, central [90% ocean-state range]	12.9 [10.4, 15.6]	10.6 [8.1, 14.1]
Conditional RR, low / high	8.4 / 17.8	7.2 / 15.5
Same, linear response, central	7.9	7.8
Neutral 99th-percentile high water exceeded every (days)	4.0	4.9
Years of SSP1-2.6 sea-level rise, central / high	40 / 51	37 / 46
Multiple of sea-level rise since 1950, central / high	1.6 / 2.1	1.5 / 1.9
Excursion duration above MHHW / below MLLW	+36% / −31%	+33% / −28%
<i>Waves and reported floods</i>		
RR_{HW} , US West Coast (median)	1.85	2.22
RR_{HW} , Kiribati [95%]	0.98 [0.43, 2.35]	0.93 [0.41, 2.60]
Flood reports, El Niño strata: background gain, Mexico–Ecuador (nats)	+0.034 (429 strata)	+0.034 (463 strata)

13 Conditioning the outlook on the observed tilt

The outlook of Fig. 3 conditions on RONI alone: at each scenario amplitude A the ocean state is held at its expected value $E[SSH_{tilt} | A]$, and its uncertainty enters only as the spread of the bands. Because the risk ratio applies at the event peak, the tilt already observed in 2026 can inform the tilt at the peak. We test this for Mexico–Ecuador with a simple conditional model.

Model. For the nine satellite-era El Niño events with a pre-peak September, the tilt in the month of peak RONI, T_{pk} , is regressed on the peak amplitude A and the September tilt T_{Sep} , giving $T_{pk} = -0.6 + 10.2A + 0.58T_{Sep}$. Its leave-one-event-out error is 4.6 cm, against 5.8 cm for a fit on A alone to the same events; the September departure from $E[SSH_{tilt} | A]$ correlates with the peak departure at $r = 0.64$. The predicted peak departure, with the leave-one-out error as its standard deviation, replaces the unconditional spread, and is propagated through the same site sensitivities, astronomical tidal term and exceedance curves as Fig. 3c. With the ocean state at its expected value, this reproduces the published values exactly.

Result. The September 2026 tilt (42.8 cm, 1–21 September) is the highest for that month in the altimetry record. At the central scenario it implies a peak tilt of 52.3 cm, 7.5 cm above $E[SSH_{tilt} | A]$, and raises the Mexico–Ecuador risk ratio from 12.9 (90% range 10.4–15.6) to 15.2 (12.9–17.6; Supplementary Table S9). Conditioning on August instead, or assuming the tilt simply rises by its mean September-to-peak gain (7.2 cm), gives 14.6 and 14.4. The observed ocean state therefore places the outlook in the upper part of the published ocean-state range. It shifts the projection but does not narrow it, since the conditional spread (4.6 cm) is similar to the unconditional one (5.2 cm).

Caveats. The model rests on nine events over the extended altimetry record, and the 2026 September tilt lies 8 cm above any previous September, so the result is an extrapolation. The fitted weight on T_{Sep} (0.58) damps that anomaly relative to pure persistence. The amplitude is not updated from the tilt, because the CPC outlook already draws on the ocean state. The result is unchanged if 2004–05, which peaked in September itself, is excluded.

Supplementary Table S9 Mexico–Ecuador outlook conditioned on the observed equatorial tilt.
Risk ratio at the low, central and high 2026–27 scenarios (RONI 2.28, 2.74, 3.21 °C) and with the CPC outlook’s amplitude uncertainty folded in; median over the 389 issuable sites, with the 90% range over ocean states in parentheses (in the last column, over ocean states and amplitude). T_{pk} is the expected tilt at the event peak for the central scenario; “spread” is the standard deviation of the peak tilt about it (the event-scale spread for the published outlook, the leave-one-event-out error otherwise; an amplitude-only fit to the same events gives 5.8 cm). The 2026 tilt is 35.7 cm for August and 42.8 cm for 1–21 September. Fitted to the 9 satellite-era El Niño events with a pre-peak September (1994–95, 1997–98, 2002–03, 2004–05, 2006–07, 2009–10, 2015–16, 2018–19, 2023–24).

	T_{pk} (cm)	Spread (cm)	Low	Central	High	Outlook
Published outlook (RONI only)	44.9	5.2	8.4 (6.4–10.9)	12.9 (10.4–15.6)	17.8 (15.1–20.6)	12.9 (6.8–20.0)
Conditioned on September tilt (primary)	52.3	4.6	11.4 (9.2–13.6)	15.2 (12.9–17.6)	19.2 (17.0–21.9)	15.2 (10.0–21.0)
Conditioned on August tilt	50.4	5.9	10.7 (7.8–13.4)	14.6 (11.7–17.6)	18.9 (15.8–22.3)	14.5 (8.8–21.1)
September tilt plus mean growth	50.0	4.6	12.0 (10.1–14.3)	14.4 (12.2–16.8)	17.1 (14.6–19.4)	14.5 (10.7–18.5)

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