The 2011 Magnitude 9.0 Tohoku-Oki Earthquake: Mosaicking the Megathrust from Seconds to Centuries

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Geophysical observations from the 2011 Mw 9.0 Tohoku-Oki, Japan earthquake allow exploration of a rare large event along a subduction megathrust. Models for this event indicate that the distribution of coseismic fault slip exceeded 50 m in places. Sources of high-frequency seismic waves delineate the edges of the deepest portions of coseismic slip and do not simply correlate with the locations of peak slip. Relative to the Mw 8.8 2010 Maule, Chile earthquake, the Tohoku-Oki earthquake was deficient in high-frequency seismic radiation—a difference that we attribute to its relatively shallow depth. Estimates of total fault slip and surface secular strain accumulation on millennial time scales suggest the need to consider the potential for a future large earthquake just south of this event.

The 2011 Tohoku-Oki earthquake occurred on the megathrust where the Pacific Plate subducts below Japan at an average rate of about 8 to 8.5 cm/yr (Fig. 1) (7). Historically, many Mw 7 to Mw 8 earthquakes have occurred on the Japan Trench megathrust (2). Geodetic observations of crustal strain during the interseismic period have been used to infer spatial variations in the degree of plate coupling (i.e., regions of the megathrust expected to produce large earthquakes) for this section of the Japan Trench (3). Generally, these models infer high coupling in regions where earthquakes were known to have already occurred (Fig. 1 and fig. S1) with only partial or even no coupling from the trench to a point approximately midway between the trench and the coastline—precisely the region where the 2011 Tohoku-Oki earthquake occurred. It is fundamentally difficult to use land-based data to assess the state of coupling on distant portions of a megathrust. Historically, the Jogan earthquake of July 13, 869 AD may be the only documented event to have occurred with a possible magnitude and location similar to that of the 2011 earthquake (4).

Observations of the 2011 Tohoku-Oki earthquake from a dense regional geodetic network and globally distributed broadband seismographic networks, as well as open ocean tsunami data, allow the construction of a family of models that describe the distribution and evolution of subsurface fault slip. Surface displacements due to the Tohoku-Oki earthquake were observed by over 1,200 continuously recording GPS sites installed and operated by the Geodetic Survey of Japan (GSI). Here, we use data sampled at 5 min intervals to produce individual three-component positional time series from which we isolate coseismic displacements (Fig. 1) (5). Significant quasi-permanent displacements due to the mainshock occurred over the entire northern half of Honshu, with peak GPS-measured offsets exceeding 4.3 m horizontally and 66 cm of subsidence (Figs. 1 and 2). We also isolate surface displacements associated with an Mw 7.9 aftershock that occurred about 30 min after the mainshock (Fig. 1). The spatial extent and the azimuth of the horizontal displacement vectors, indicate that the aftershock was located to the south of the mainshock in the Ibaraki segment. Peak horizontal GPS-measured displacements for this aftershock are approximately 44 cm (Fig. 1). We constrain the distribution of coseismic slip on the shallowest portions of the megathrust using observations of open ocean tsunami wave heights measured by deep sea-bottom pressure gauges (Fig. 1). Based on their spatial and azimuthal distribution, we selected 12 sensors in the Pacific ocean east of the Japan trench. The closest of these pressure gauges detected a maximum tsunami wave height of more than 1.9 m (Fig. 2A).

We first describe static coseismic slip models based on the GPS observations of coseismic offsets and the seafloor pressure gauge data (Fig. 3 and fig. S2). Static models constrain the final distribution of slip for the event but not its temporal evolution. We adopt a novel fully Bayesian probabilistic formalism requiring no a priori spatial regularization (5, 6). We conservatively define the section of
the megathrust directly involved with the earthquake by considering only the areas where inferred slip exceeds 8 m (approximately 15% of the maximum slip value depending on the model). The model predicts maximum seafloor subsidence of about 2 m located 50 km offshore Sendai and Kamaishi, and maximum seafloor uplift just under 9 m about 50 km from the trench (Fig. 2). This model fits the GPS and tsunami data with variance reductions of 99.7% and 90.1%, respectively. Residual GPS displacements after removal of the model predictions are shown in fig. S3.

The spatial distribution of slip (Fig. 3) can be divided into several sections. The central section contains the highest estimated slip values with peak displacement of around 60 m. The up-dip limit of the forearc is an active accretionary prism that extends about 50 km landwards of the trench. Generally, the majority of the fault slip does not extend below this zone, with the exception being just up-dip of the region of maximum fault slip where estimated slip values near the trench range from about 5 to 15 m. A tendril of slip extends over 100 km north from the central slip zone and just down-dip from the inferred source of the 1896 Mw 8.0 Sanriku earthquake. Average slips in this region are approximately 5 to 10 m—similar to those inferred for the 1896 earthquake (7). A lobe of about 10 m of fault slip extends down-dip toward the Oshika Peninsula east-northeast of Sendai. This lobe overlaps a couple of the inferred historical Miyagi-Oki rupture areas. Slip in the up-dip portion of these rupture areas extends 20 m. Another tendril of significant slip (5 to 10 m) extends southwards of the main high slip asperity. This tendril clearly overlaps the inferred locations of the 1938 Fukushima earthquake sequence.

We estimated probability distributions for derived scalar rupture quantities including rupture area, potency, scalar seismic moment, and static stress drop (fig. S4). Estimates of moment magnitude range from 8.8 to 9.2. We note that static slip models are relatively insensitive to absolute scaling of the elastic moduli, thus estimates of moment are less certain than estimates of potency. Estimates of static stress drop vary between 2 and 10 MPa depending on the area of fault considered. These values are high relative to previous estimates for megathrust events, which typically lie in the 1 to 5 MPa range (8) and reflect the relatively small area over which there is high slip. As a point of comparison, models of the 2010 Mw 8.8 Maule, Chile earthquake typically find twice the along-strike extent of slip and half the peak slip as our model for the Tohoku-Oki event (9, 10).

We also developed two kinematic finite fault models incorporating one and two fault planes, respectively, broadband seismic data, and GPS observations, but no tsunami data (5, 11). Examples of the displacement and velocity waveform fits are shown in fig. S5. The inferred moment rate function suggests that most of the rupture occurred in a little over 3 min (fig. S6). We find a low average rupture velocity of about 1.2 km/s (fig. S6). This model explains 99% of the variance of the GPS data. The peak estimated fault slip in this model is about 45 m, a little less than found in the static model. This difference is due to the imposed smoothing in the kinematic model, which is absent in the static model. The extension of slip to the south (offshore Fukushima) is evident in the kinematic model; however, it is located up-dip relative to the static model, presumably due to the absence of the tsunami constraints in the kinematic model. The Mw 7.9 aftershock occurred just beyond the southernmost extent of the mainshock slip area with an estimated maximum slip of about 4 m for this event (Fig. 3) (5).

Observations at the high-frequency (HF) end of the seismic spectrum (2 to 4 Hz) can constrain rupture direction and duration (12). 90% of the energy release in this frequency band occurs within about 3 min (fig. S7). Unilateral rupture propagation would result in an azimuthally dependent duration. This earthquake displays uniform durations at most azimuths with slightly shortened durations in the down-dip direction, suggesting bilateral along-strike rupture with some down-dip propagation (fig. S7).

We developed an image of the rupture process at high frequencies (between 0.5 and 1 Hz) using back-projection of teleseismic array waveforms (13) based on high-resolution array processing techniques (5, 14, 15). The most energetic of the high-frequency sources as a function of time during the rupture are systematically down-dip of the regions of largest fault slip (Fig. 3). The robustness of the locations of the HF radiators relative to the assumed hypocenter is supported by the consistency between the results obtained with USArray data and those from the European array (Fig. 3 and fig. S8).

HF radiation is usually assumed to be spatially correlated to seismic slip (16) or uniformly distributed over the fault (17). This assumption contrasts with the spatial complementarity between HF and low frequency (LF) source properties observed for this earthquake. Such a relationship has been inferred for other earthquakes, although not systematically (18), and might reflect the general lack of correlation between HF and LF in ground motions (19, 20). Dynamic rupture models generate HF radiation mostly during sudden changes of rupture speed along sharp contrasts of fault rheology or geometry, or along remnant stress concentrations from previous earthquakes (21), which can define the boundaries of the slip area.

Instead of relying on characteristics of the rupture dynamics to explain the predominance of HF radiators down-dip of the region of significant coseismic slip, an alternative explanation is that they occur at the transition between brittle and ductile regions. Recent observations of medium-sized earthquakes down-dip from aseismic sections as well as of
slow-slip and non-volcanic tremor phenomena suggest the presence of frictional heterogeneities in the brittle-ductile transition regions of subduction megathrusts (22). Coseismic triggering of compact brittle asperities embedded in the ductile fault matrix could explain the relative locations of HF and lower frequency slip and the apparent lack of HF radiators at shallow depths.

Similarly, we find that large aftershocks with thrust mechanisms are located outside the region of coseismic slip (fig. S2), as has been documented for previous events (23, 24). Interestingly, for this event, aftershocks are dominantly down-dip of the regions of major coseismic slip—consistent with the idea that fault slip at shallow depths is relatively slow due to higher fracture energy and thus radiates energy less efficiently than at greater depths (8). This latter interpretation is supported by a comparison of seismic excitation between 0.5 and 4 Hz from the 2010 Maule and 2011 Tohoku-Oki events (fig. S9). While the value of peak slip inferred for the Tohoku-Oki earthquake was about 2 to 3 times larger than for the Maule earthquake and the area of appreciable slip for the Tohoku-Oki is approximately half that of the Maule earthquake, we find that the former produced much less HF radiation. We suggest that this difference is due to the fact that the Maule earthquake rupture is on average much deeper than the Tohoku-Oki rupture (fig. S10). Thus, it may be that shallow ruptures generally produce large displacements with relatively weak HF excitation. Such an interpretation is also broadly consistent with the general observations that tsunami earthquakes rupture very slowly with slip concentrated in the shallowest part of the megathrust (25).

Our inferred fault slip model suggests high static stress drop with large amounts of slip in a small region. An explanation for this behavior theorizes the existence of barriers that require much more stress accumulation than other regions before they rupture. Such barriers may pin the fault locally, limiting the amount of seismic slip occurring on neighboring areas that have lower thresholds for failure. When the strongest barrier finally ruptures, then surrounding areas can catch up. Subducted seamounts are the most obvious candidates for such barriers (26). Indeed, several seamounts are known to have subducted in this segment of the Japan Trench (27). The distribution of slip in the Tohoku-Oki earthquake suggests that small areas can have high effective yield stresses that serve to limit rupture propagation during some earthquakes but then eventually rupture with large slip during others.

The extent to which the 2011 earthquake was unexpected suggests that we should consider the potential for similar large events elsewhere on the Japan Trench megathrust. The secular interseismic velocity field for the Miyagi segment shows over 3 cm/yr of relative convergence across Honshu (Fig. 4). On average, faults in the interior and off the western coast of Honshu, are believed to account for between 1 and 2 cm/yr (28)—leaving 1 to 2 cm/yr associated with interseismic strain accumulation on the subduction interface. We adopt 1,100 years as a representative time period because this corresponds to the last large event that is inferred to have occurred in this region (4). Thus, over this time period, we must still account for 11 to 22 m of relative motion across Honshu. Similarly, at about 8.5 cm/yr of convergence, we must account for over 90 m of fault slip on the megathrust.

Earthquake activity offshore of Miyagi (Fig. 1) has been suggested to be dominated by M7+ earthquakes recurring every 30 to 40 years (2). However, it has already been established that the historical events are not exact repeats of one another (29). Further, such M7+ events only produce 3 to 4 m of fault slip and 5 to 20 cm of surface displacement per event. In the 2011 mainshock, fault slip in the region of the historical M7+ events ranged from 5 to 25 m, suggesting that the concept of a characteristic subduction earthquake with approximately the same slip per event at a given location may be of limited use (30). That the 2011 event produced approximately 50 m of slip up-dip of the historical Miyagi M7+ events is roughly consistent with a 500 to 1,000 year potential slip accumulation period. However, there is no basis on which to assume that the aforementioned interval of 1,100 years is representative of the recurrence interval of great earthquakes in this segment—it could be shorter by a factor of two and still be consistent with the surface displacement budget and the peak slip inferred in this recent earthquake.

The only previously recorded large events offshore Fukushima and Ibaraki occurred as a sequence in 1938, which taken together correspond to about an Mw 8.1 event (31). At about 8.0 cm/yr convergence, the 73 years since those earthquakes imply about 6 m of accumulated potential fault slip—surprisingly similar to the estimates of fault slip for this region during the 2011 event. However, this agreement is probably coincidental since there is no record of an equivalent set of earthquakes preceding the 1938 sequence. Similarly, the 2011 Mw 7.9 aftershock offshore of Ibaraki (Fig. 1) produced about 4 m of fault slip, implying a 50 year recurrence if these events are characteristic of this segment of the megathrust. There is no documentation of large (M8+) events prior to 1938 (31).

The 2 cm/yr GPS-observed onshore convergence in this region (Fig. 4) implies that, in terms of the 1,100 year budget, there is between 0 and 10 m of surface convergence that cannot be plausibly associated with faults farther to the west (28). The combination of the 2011 mainshock and the M7.9 aftershock only produced about 2 m of surface displacement.

There is no record of a large event up-dip of the 1938 Fukushima and Ibaraki sequence. Thus, the slip budget on the megathrust and the surface velocity data suggest that an
earthquake similar to the 2011 event is possible offshore Ibaraki and Fukushima just south of the most recent event (Fig. 4). During such an event, the 1938 asperities and the M7.9 aftershock rupture area could experience much greater slip than has been documented for previous events, similar to what just occurred offshore Miyagi. However, if this region is in fact not strongly coupled, simple mechanical models (32) would predict high rates of post-seismic afterslip—as is inferred to have occurred after several large recent earthquakes (24). Thus, it is essential to monitor this region to quantify the extent of any post-seismic slip in order to further understand the long-term fault slip budget and associated seismic hazard.

References and Notes
5. See supporting material on Science Online.

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Supporting Online Material
www.sciencemag.org/cgi/content/full/science.1206731/DC1

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Fig. 1. Map of central and northern Honshu, Japan. Vectors indicate the horizontal component of the GPS displacements for the mainshock (yellow) and the $M_w$ 7.9 aftershock (orange). Approximate locations of historical megathrust earthquakes are indicated by closed curves colored by region—Pink: Tokachi (1968 M8.2), Green: Sanriku (1896 M8.5, 1901 M7.4, 1931 M7.6, 1933 M7.6), Purple: Miyagi (1897 M7.4, 1936 M7.4, 1978 M7.4, 2005 M7.2), Brown: Fukushima (1938 $M_w$ 7.4, 1938 $M_w$ 7.7, 1938 $M_w$ 7.8) (modified from (2, 3, 33)). Yellow and orange moment tensors indicate the W-phase centroid for the mainshock (34) and the GCMT location for the M7.9 aftershock. The closed yellow curve indicates the outline of the $M_w$ 9.0 mainshock (8 m slip contour). The region of inferred slip deficit or high plate coupling is indicated by dark blue nested contour lines for 35%, 70%, and 100% coupling (3). Barbed lines indicate subduction plate boundaries. The white arrow indicates the direction of convergence between the Pacific Plate and northeast Japan (1). T, S, and K indicate the cities of Tokyo, Sendai and Kamaishi. The yellow box in the inset reference map shows the region of this figure and the locations of deep-sea bottom pressure gauges used in this study, all superimposed on the peak tsunami wave heights predicted by our preferred earthquake source model.

Fig. 2. (Top) Observed (green) and predicted (white) deep ocean tsunami records of the Tohoku-Oki earthquake. The predicted records correspond to a model constructed using the mean of each fault slip parameter in the Bayesian inversion. These waveforms are superimposed on the map of maximum model-predicted tsunami height. (Bottom) GPS vertical coseismic surface displacements (circles colored and scaled with amplitude) as well as model predicted vertical seafloor displacements (filled contours). Other overlay features are as in Fig. 1.

Fig. 3. Inferred distribution of subsurface fault slip (color and black contours with a contour interval of 8 m). Fault slip associated with the $M_w$ 7.9 aftershock is indicated by nested 1 m orange contours. Historic earthquake ellipses are as in Fig. 1. Location of points of high-frequency radiation estimated using back projection methods with data from the European Union seismic array and the USArray are indicated by squares and circles, respectively, with color intensity indicating time of the activity relative to the beginning of the event and with size of the symbol proportional to amplitude of the HF radiation normalized to the peak value. The star indicates the location of the JMA epicenter. See fig. S2 for a plot of the slip model without overlays.

Fig. 4. Secular interseismic surface deformation (blue vectors) as observed by the GEONET continuous GPS network using the GSI F2 solution. The top-left inset shows the decrease of this deformation field at the latitude of the Miyagi (purple profile) and Ibaraki (orange profile) segments, with distance measured along the profile. The coseismic slip distribution is indicated by the yellow contours at 8 m intervals. The inset uses vectors within 100 km of the profile location. The question mark indicates a region of possible high seismic hazard. Other features are as in Fig. 1.
Supporting Online Material for

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GPS Processing

The 5-minute position estimates were calculated using Jet Propulsion Laboratory’s GIPSY-OASIS software (35). We use a single-station bias-fixed point positioning (36) and kinematic positioning strategy. For the kinematic positioning, the station position is a stochastic parameter in the model used to fit the GPS observations. Using the point-positioning technique, the GPS orbit and clock estimates, provided by JPL’s GPS Analysis Center, are held fixed. The resulting positions are therefore not relative to any site, which allows the results to be free of common mode errors introduced into double differenced high rate positions (37). Troposphere zenith delay and gradients were also estimated as stochastic parameters in our analysis.

To calculate the coseismic offsets, we use the time series sampled at 5-minute intervals in the time period of 2 hours before and after the time of the mainshock. For each component (north, east, up) of each time series, we fit the time series for a constant and two step functions – one for the mainshock and one for the aftershock, plus a linear post-seismic for the mainshock. To estimate errors on the coseismic jumps, we rescale the formal errors on the GPS time series using the RMS of the residuals after removing the constant and jumps. In the finite fault modeling, we then inflate the GPS errors further by a factor of 4 in order to account for systematic errors such as inaccuracies in the assumed elastic structure or the geometry of the fault.

Each component of the GPS positional data is modeled as follows:

\[ d(t) = C_0 + M_C \cdot H(t - t_M) + M_p \cdot (t - t_M) \cdot H(t - t_M) + A_C \cdot H(t - t_A) \]

where,

- \(d\) : Positional data from GPS time series.
- \(C_0\) : Constant offset
- \(M_C\) : Coseismic displacement caused by the mainshock at time \(t_M\)
- \(M_p\) : Amplitude modulating the post-seismic relaxation of the mainshock
- \(A_C\) : Coseismic displacement caused by the M 7.9 aftershock at time \(t_A\)
- \(H(\cdot)\): Heaviside function.

We perform a weighted least squares fit, solving for \(d_0, M_C, M_p, A_C\), with relative weights obtained from the formal errors on the observables. The formal errors, \(\sigma\), are rescaled as follows:

\[ \sigma^* = \frac{\sigma}{\sigma} \cdot \text{RMS}(d_{\text{obs}} - d_{\text{fit}}) \]

where \(d_{\text{obs}}\) and \(d_{\text{fit}}\) are the observed and predicted data values. We removed a three-sample window centered at the closest sample to the mainshock and aftershock since these are consistently observed to be outliers in the positional time series.
Preparation and Modeling of Tsunami Records

The distribution of slip as a function of depth along the fault, in particular the extent of slip found near the trench, is most strongly constrained by incorporating the tsunami waveform data. We use deep sea-bottom pressure sensor records retrieved from the NOAA website (http://www.ndbc.noaa.gov/dart/dart.shtml). We remove the ocean tidal signal using a polynomial filter. Because in deep water the amplitude of the tsunami wave is relatively small compared to the water depth, nonlinear propagation effects are negligible. Thus, we assume that the first oscillations of the tsunami signal are linearly related to the source (seafloor displacements) and that we can adopt a linear Green’s function approach. The tsunami Green's functions are estimated based on the linear shallow water approximation using the Cornell Multi-Grid Coupled Tsunami Model code (http://ceeserver.cee.cornell.edu/pll-group/comcot.htm). The bathymetry is extracted from the ETOPO2 bathymetry grid, with a resolution of 2 arcminutes.

Seafloor displacements are in turn related to slip on individual triangular subfault patches through elastic Green's functions. The Green’s function for each triangular patch is calculated by summing the contributions from a set of appropriately spaced and weighted point sources. For the purpose of modeling the tsunami wave heights, we treat slip during the earthquake as if it occurred simultaneously over the entire rupture. This approximation is valid because the earthquake rupture velocity is large compared to the tsunami propagation time.

We use 80 minute long records sampled at 1 min. We initially assume a 2.5 cm error for the tsunami records, but as with the GPS data, this error is inflated by a factor of 4 to account for systematic errors. After several tests, we found that we could not reconcile the timing of the first tsunami arrivals at four DART sensors closest to Japan with those observed at the distant stations offshore Alaska and Western USA. However, the waveforms at all 12 stations are well explained. Thus, we interpret these timing issues to be propagation effects from systematic errors in the bathymetry or the fact that dispersion has not been taken into account. Thus, in our inversions, we kept the records from the four nearest stations in absolute time, but allowed a time shift for the remaining 8 records. The applied time shift increases with propagation distance. With this time shift, we are able to fit all 12 full waveforms well.

Fault Geometry

We construct a model of the surface of the subducted Pacific Plate using the GOCAD® commercial software package. Using interpretations of seismic reflection and refraction profiles, the plate interface is constrained to depths of about 30 km. The deeper slab geometry is constrained with published seismic tomographic and earthquake hypocenters. We interpolate a smooth interface between the seismic lines, and allow a misfit between the interface model and the seismic profiles up to a kilometer in depth in places where the seismic profiles conflict as to the depth to the top of the Pacific plate. The trench location is estimated from bathymetric features as well as from trench locations in the seismic profiles. For modeling purposes, we tessellate the fault geometry using triangular patches that conform to the complex 3-D geometry of the megathrust. We use a total of 419 triangular patches with a characteristic dimension that ranges between 11 and 27 km.
Static Finite Fault Modeling

For modeling the static slip distribution for the mainshock rupture, we adopt a Bayesian approach because it has several advantages over traditional inversion methodologies. Determining slip on a fault based on surface observations is, in general, an under-determined inverse problem, and this under-determinedness is only exacerbated when trying to constrain offshore slip with mainly land-based observations. Because of this ill-posedness, slip inversions have historically used \textit{a priori} constraints on the solution such as Laplacian smoothing and moment minimization. Since the inverse problem has no unique solution, any slip model produced in this manner is only one sample from a conceivably large space of possible models. Such a slip model is predestined to conform to some \textit{a priori} regularization, which may have been chosen without any physical basis.

In contrast, Bayesian analysis casts the solution to the inverse problem as an \textit{a posteriori} probability density function (PDF), \( P(M|D) \), the probability of a model \( M \) given observed data \( D \), which is proportional to \( P(M) \cdot P(D|M) \), the product of the \textit{a priori} probability of the model and the likelihood of the data. There are two main advantages to this approach. First, the inverse problem need never be evaluated because the \textit{a posteriori} PDF can be determined through Monte Carlo simulation, and thus \textit{a priori} regularization is not required to make the problem computationally stable. Second, computing the \textit{a posteriori} PDF yields not one solution to the under-determined inverse problem, but instead returns the ensemble of all possible source models that are consistent with the data and our \textit{a priori} assumptions about the physics of the earthquake source.

Our model for the earthquake rupture consists of two perpendicular components of slip on each cell of a complex triangulated mesh. We evaluate the data likelihood by calculating an L2 norm misfit function, \( P(D|M) = e^{-\frac{1}{2}(D-G \cdot M)^T C_D^{-1}(D-G \cdot M)} \), where \( G \) is a matrix of Green’s functions, \( C_D \) is the data covariance matrix, and \( M \) is the vector of slips on each patch. The Green’s functions are calculated using a 1-D velocity model modified from (40).

Our goal is to determine what parts of the rupture process are well constrained by the data and which are not. To this end, we use broad \textit{a priori} PDFs to describe our model so that the \textit{a posteriori} distribution is controlled by the data likelihood. The strike-slip component of slip is assigned a Gaussian \textit{a priori} probability with zero mean and a standard deviation of 10 m, while the dip-slip component is given a uniform probability between -10 m and infinity. (Negative slip denotes back-slip or normal faulting.) Drawing samples of the \textit{a posteriori} PDF is extremely costly for inverse problems with large numbers of free parameters. To make the sampling process computationally tractable we use the Cascading Adaptive Transitional Metropolis In Parallel (CATMIP) algorithm (6, 41) which combines an adaptive implementation of the Metropolis algorithm, resampling and simulated annealing in a parallel framework. We list the coordinates for each triangular patch and the estimates of slip for each patch in Tables 1 and 2. The ID number for each triangle is shown in Fig. S12.

Since the Bayesian approach applies no smoothing to the model, this leads to the question: to what extent are the details of the slip distribution to be believed? The answer to this question comes in two forms. The first is to look at the correlation between any two model parameters
(e.g., slip on neighboring patches) using the ensemble of all models that comprise the PDF. If the model is not resolved, we should find considerable anti-correlation between neighbors. As an example, we show the correlation between the dip-slip component of motion on the patch with the peak slip and all other patches (Fig. S11). Note that there is only slight anti-correlation with one neighbor, so the slip distribution appears to be well constrained. A second answer to the resolution question comes in terms of understanding the output of the Bayesian sampling process. Our solution for fault slip is not a single model, but rather represents derived statistical quantities (specifically, the mean or median) from the PDFs for each model parameter. If there are trade-offs between fault patches, the mean should not be affected, but the variance about the mean will increase.

The above discussion on resolution is only strictly true in the context of a perfect physical model, but no model can be perfect. Furthermore, the uncertainty in the present physical model is not considered as part of the observation error (i.e., the data covariance matrix) that goes into evaluating the data likelihood, and thus the a priori observation error and the a posteriori model uncertainties underestimate the true error. The model design used in this study ignores the effects of topography, simplifies the elastic structure of the Earth and the geometry of the fault, and assumes that the tsunami is caused by instantaneous and simultaneous slip on the fault. These limitations in the physical model can cause systematic biases that are not explored here. For instance, our current parameterization can only alter the timing of the tsunami waveforms by varying the location of slip. A more sophisticated model in which slip on the fault evolves as a function of time would cause time-dependent seafloor displacements. Such extra degrees of freedom would improve the fit to both the GPS and tsunami data while inevitably changing the a posteriori distribution of fault slip.

**The Mw 7.9 Aftershock**

For the Mw 7.9 aftershock, we adopted a small, tessellated rectangular fault patch with a position and fault orientation based on the GCMT mechanism. We also constrained the magnitude (Mw 7.9) to be consistent with the CMT mechanism. The distribution of fault slip for this event was calculated using a simulated annealing approach (42).

**Kinematic Finite Fault Modeling**

The assumed fault plane extends 280 km down-dip and 500 km along-strike. The fault is tessellated with a grid of 20 km × 25 km-sized patches. The slip rake on each patch is constrained to ±90°, with slip triggered by the rupture front. We constrain the total moment to the value given by the GCMT best double-couple moment. The rupture velocity is allowed to vary between 0.9 and 2.0 km/s. We adopt a 1-D layered velocity model extracted from CRUST2.0 (43). We considered both a two fault plane model and a single fault plane model. The two models show similar fits to the GPS and teleseismic data.

**High Frequency Seismic Wave Back Projection**

We used P-wave seismograms recorded by two large arrays at epicentral distances of 70°-90°, the USArray and the European network, which illuminate the source region from two almost orthogonal directions (Fig. S8). We band-pass filtered the waveforms between 0.5 and 1 Hz and
aligned them on their first P arrival by multi-channel cross-correlation (44). We then applied two different array processing techniques over 10-second-long sliding windows, Multiple Signal Classification (MUSIC) (14, 45) and coherent interferometry (CINT) (15, 46) with back-projection onto the source region at a reference depth of 15 km with travel times based on the IASP91 Earth model. Our results based on MUSIC and CINT are mutually consistent. Compared to classical beam forming, these high-resolution techniques resolve more closely spaced sources and are less sensitive to aliasing, yielding a sharper and more robust image of the rupture process. The results from both arrays are consistent with each other.

We use the JMA hypocenter to fix the absolute location. Use of a different hypocenter would result in a simple translation of the radiator locations. Expected errors in the epicenter are sufficiently small that the conclusion about the location of the high frequency radiators relative to the distribution of fault slip is robust.

**Fig. S1**

Comparison of published fault coupling models (blue contours) constrained by GPS-based estimates of the secular interseismic surface velocity field. The contour intervals are equivalent to about (left) 50% and 95% coupling (47), (middle) 35%, 70% and 100% coupling (3), and (right) 30%, 60% and 90% coupling (28). Contours taken from original sources. Other features are as in Fig. 1. Differences between these models are mostly due to whether or not vertical deformation data is used, the amount of spatial smoothing that is applied, and the degree to which long term shortening on shallow on-land crustal faults is considered in each analysis.
Fig. S2
Inferred distribution of subsurface fault slip (color with superimposed contours at 8 m intervals) constrained by GPS and tsunami observations and derived from an unregularized Bayesian estimation method (6, 41). The black focal mechanisms are GCMT solutions for a period of 34 days following the mainshock. Only events with thrust mechanisms are shown.
Fig. S3
Comparison of Bayesian slip models to results from conventional damped least squares (LSQ). The CATMIP median and CATMIP mean models are the median and mean, respectively, of the Bayesian *a posteriori* probability density function calculated using the CATMIP sampling algorithm. The LSQ approach simultaneously minimizes data misfit, spatial roughness of the slip model, and the difference between the inferred moment and a target moment equivalent to Mw 9.0. Red arrows show the residual horizontal (top) and vertical (bottom) GPS coseismic offsets. All three models fit the observed GPS data with a variance reduction of about 99%. The median and mean of the Bayesian *a posteriori* distribution are nearly identical and have almost equal overall goodness of fit to both the GPS and tsunami data, but these two slip models differ in a low amplitude, long wavelength spatial mode. This small difference causes a trade-off in how well they fit the horizontal component of the GPS displacements relative to the vertical component. This trade-off may be due to limitations in our assumed Earth structure and fault geometry.
A posteriori PDFs for derived quantities including rupture area, potency, scalar seismic moment and static stress drop. We consider three different thresholds to define the extent of the earthquake rupture: the entire fault model (black), areas that slipped in excess of 10% of the maximum slip (red), and in excess of 20% of the maximum slip (blue).
(A) Sample waveform fits from the kinematic source model. Representative displacement (left column) and velocity (right column) waveform data (black) and fits (red) are shown. The number at the end of each trace is the peak amplitude of the observation in micro-meter or micro-meter per second. Station names are indicated to the left of the displacement seismogram with the numbers at the beginning of the trace indicating the source azimuth (top) and epicentral distance (bottom). Results from a larger sample (in velocity) are given in Fig. S5B for both the single and double plane solutions. (B) Comparison of an azimuthal sampling of P-waveforms against synthetics generated from the one-plane vs. two-plane solutions (see Fig. S6 for the corresponding slip models). The upper number indicates the azimuth with the lower number the distance in degrees. These two slip-models have been used to predict the waveforms globally. We consider the cross-correlations (cc) between data and synthetic waveforms (first 180s). Stations closer than 50 degrees contain the PP-wave in this time window, which lowers the fits. For example, see the recording at MAKZ near 150 s. Both the one and two plane models produce cc > .85 at over a 150 stations at ranges greater than 60 degrees.
Fig. S5 continued
Fig. S6
Kinematic fault slip models constrained by GPS measurements and teleseismic P-waveforms. Estimated fault slip (left) and predicted vertical seafloor displacements (right) are shown for the two-plane (top) and one-plane (bottom) kinematic models. Dip angles and depth are given in the northeast corner of each fault plane. White contours indicate temporal evolution of the rupture front, with time in seconds. The yellow star shows the epicenter used for each inversion. The respective moment rate functions are plotted in the insets.
Vertical component of smoothed teleseismic P-wave envelopes (band-passed at 2-4 Hz) for (A) the 2011 Tohoku-Oki earthquake and (B) the 2010 Maule earthquake as a function of azimuth. The envelopes are smoothed by convolving them with a triangle function of 2 s duration. Red dots represent the time by which 90% of the energy in the envelopes has been released. The slight relative decrease in this time at azimuths of 250 deg for the Tohoku event and 80 deg for the Maule event is caused by directivity.
Fig. S8
Azimuthal equidistant projection centered on the 2011 Tohoku-Oki earthquake epicenter showing the approximate ray paths to the European seismic array (blue) and USarray (red). Concentric circles indicate distance from the epicenter at 30° intervals.
Amplitude of the integrated teleseismic P-wave envelopes derived from the vertical component of seismograms in the (a) 2.0 to 4.0 Hz and (b) 0.5 to 2.0 Hz frequency bands, plotted as a function of distance from the centroid. Estimates for the 2011 Tohoku-Oki earthquake are indicated by black dots and estimates for the 2010 Maule with red dots. Black and red horizontal lines indicate the average of the log of the amplitude for each earthquake. Both the average and individual estimates for the Tohoku-Oki earthquake have been corrected for the difference in moment by a factor of $A/(10^{1.5\Delta M_w})$, where $A$ is observed amplitude and $\Delta M_w$ denotes the magnitude difference between the two earthquakes.
Fig. S10
Along-strike integrated potency (fault slip multiplied by area) as a function of depth (left) and distance from the trench (right). Also shown is the average bathymetric profile, $H$, versus distance from the trench (blue lines). Top row: 2011 Mw 9.0 Tohoku, Japan. Bottom row: 2010 Mw 8.8 Maule, Chile.
Correlation of the up-dip component of slip on the fault patch with maximum average slip with the dip-slip component of slip on all other patches. These values are calculated from the Bayesian \textit{a posteriori} PDF.
Fig. S12
Reference ID of each triangular patch used in the fault model. The geometry and estimated values of slip for each patch are given in Tables 1 and 2.
References and Notes


5. See supporting material on Science Online.


