S31G MCC: 3009 Wednesday 1020h

Stress Transfer, Triggered Earthquakes, and Time-Dependent Seismic Hazard II *(joint with G, T, NG)*

Presiding: **J Gomberg,** U.S. Geological Survey**; M W Stirling,** Institute of Geological and Nuclear Sciences

S31G-01 1020h

Testing hypotheses of earthquake occurrence

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We present a relatively straightforward likelihood method for testing those earthquake hypotheses that can be stated as vectors of earthquake rate density in defined bins of area, magnitude, and time. We illustrate the method as it will be applied to the Regional Earthquake Likelihood Models (RELM) project of the Southern California Earthquake Center (SCEC). Several earthquake forecast models are being developed as part of this project, and additional contributed forecasts are welcome. Various models are based on fault geometry and slip rates, seismicity, geodetic strain, and stress interactions. We would test models in pairs, requiring that both forecasts in a pair be defined over the same set of bins. Thus we offer a standard "menu" of bins and ground rules to encourage standardization. One menu category includes five-year forecasts of magnitude 5.0 and larger. Forecasts would be in the form of a vector of yearly earthquake rates on a 0.05 degree grid at the beginning of the test. Focal mechanism forecasts, when available, would be also be archived and used in the tests. The five-year forecast category may be appropriate for testing hypotheses of stress shadows from large earthquakes. Interim progress will be evaluated yearly, but final conclusions would be made on the basis of cumulative five-year performance. The second category includes forecasts of earthquakes above magnitude 4.0 on a 0.05 degree grid, evaluated and renewed daily. Final evaluation would be based on cumulative performance over five years. Other types of forecasts with different magnitude, space, and time sampling are welcome and will be tested against other models with shared characteristics. All earthquakes would be counted, and no attempt made to separate foreshocks, main shocks, and aftershocks. Earthquakes would be considered as point sources located at the hypocenter. For each pair of forecasts, we plan to compute alpha, the probability that the first would be wrongly rejected in favor of the second, and beta, the probability that the second would be wrongly rejected in favor of the first. Computing alpha and beta requires knowing the theoretical distribution of likelihood scores under each hypothesis, which we will estimate by simulations. Each forecast is given equal status; there is no "null hypothesis" which would be accepted by default. Forecasts and test results would be archived and posted on the RELM web site. The same methods can be applied to any region with adequate monitoring and sufficient earthquakes. If fewer than ten events are forecasted, the likelihood tests may not give definitive results. The tests do force certain requirements on the forecast models. Because the tests are based on absolute rates, stress models must be explicit about how stress increments affect past seismicity rates. Aftershocks of triggered events must be accounted for. Furthermore, the tests are sensitive to magnitude, so forecast models must specify the magnitude distribution of triggered events. Models should account for probable errors in magnitude and location by appropriate smoothing of the probabilities, as the tests will be "cold hearted:" near misses won't count.

S31G-02 1035h INVITED

Conditional Probabilities for Earthquake Triggering in Global and Regional Datasets

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We define earthquake triggering as the existence of a pair of similar sized events that are close in position and time, above chance occurrence in time, but allowing for the spatial correlation of epicentres. The global CMT catalogue contains a finite triggering signal with a similar correlation function to that predicted in a critical point system. No strong directional dependence is found, but the Coulomb triggering hypothesis cannot be tested quantititively due to the fault plane ambiguity in many of the focal mechanisms. The correlation length and mean triggering distance are on the order 10-20km, similar to the seismogeneic thickness, and increase only gradually with time after a given event, in the form of a power law of exponent $H \sim 0.06$. This compares with $H \sim 0.5$ expected for Fickian stress diffusion, but is quantitatively consistent with delayed stress corrosion failure in a random heterogeneous medium. The regional study is based on ISC data with a smaller magnitude threshold. We do not detect a strong bias in the results based on the choice of region for concentric circles centred on the Gulf of Corinth area as a test case. The results of an analysis of the most active Flinn-Engdahl-defined regions are consistent with the model formed from the global study, but show significant regional variation in the model parameters that depend on the local tectonic style. The parameters of the model may be used directly to calculate a conditional probability of further similar-sized events, once one has occurred.

S31G-03 1050h INVITED

The Importance of "Time-Varying" Input for Probabilistic Seismic Hazard Assessment

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The standard methods of modern probabilistic seismic hazard assessment (PSHA) typically assume a Poissonian (time-independent) model of earthquake occurrence. Input to Poissonian-based hazard estimates are simply the magnitudes and recurrence intervals of earthquakes on earthquake sources, allowing the PSHA to be computationally simple, and probably adequate for quantifying the long-term distribution of hazard in a region (e.g. peak and spectral accelerations expected with a 10% probability of exceedance in 50 years, equivalent to a 475 year return time). However, a serious deficiency in these standard methods of PSHA is that they ignore "time-varying" phenomena, such as the time elapsed since the last earthquake on an earthquake source, the association of foreshocks and aftershocks with a given mainshock, and interactions between the different earthquake sources in a region. Quantification of the influence of these factors on the Poissonian hazard estimates is especially important for estimating shorter-term hazard (i.e. time periods of tens of years or less), and significant efforts in recent years have focused on developing "time-varying" PSHAs that take these factors into account. We will show examples of PSHAs from New Zealand, USA and elsewhere to illustrate how "time-varying" hazard estimates differ from standard Poissonian estimates. For example, we observe a 30% increase in the peak ground acceleration expected along the Alpine Fault of New Zealand in a "time-varying" model over the standard Poissonian model, at the 10% probability of exceedance in 50 years level.

S31G-04 1105h

Viscoelasticity, Postseismic Slip, Fault Interactions, and the Recurrence of Large Earthquakes

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Since Reid formulated the elastic rebound hypothesis, our view of earthquake occurrence has been based on the idea of uniform loading leading to recurrent failures. This view was reinforced by the discovery of plate tectonics but, recently, many studies have demonstrated the existence of transient, high deformation rates after large earthquakes due to either viscoelastic processes or post-seismic slip. Viscoelastic response of the lower crust and upper mantle to large earthquakes results in temporarily higher deformation rates in the region surrounding the mainshock. These higher deformation rates will result in faster than average reloading of strain energy onto the mainshock fault. If post-seismic slip is a planar downward extension of the mainshock, then the post-seismic slip will increase the stress stored on the mainshock fault plane. Again, this will result in faster than average loading of strain energy onto the mainshock fault plane for some time immediately following the mainshock. Thus, the loading of strain energy onto seismogenic faults is not temporally uniform and, in this study, I consider the effects of transient deformation on the estimation of earthquake probabilities by modifying the Brownian Passage Time (BPT) model of earthquake recurrence. The BPT inter-event time distribution is derived from a process where a state variable starts at 0, evolves by a superposition of a linear trend and Gaussian white noise until reaching a value of 1 when failure occurs and state is reset to 0. For earthquake recurrence the linear trend represents the uniform deformation due to plate motions and the noise represents fault interactions and other unknown perturbations to the process. To approximate the effects of viscoelasticity and post-seismic slip; I add a decaying exponential term to the BPT model's uniform loading term. The resulting inter-event time distributions remain approximately lognormal but the balance between the level of "noise" and the coefficient of variability of the inter-event time distribution changes depending on the shape of the loading function. For a given level of noise in the loading process, transient deformation generally has the effect of increasing the coefficient of variability of earthquake inter-event times. Conversely, the level of noise needed to achieve a given level of variability is reduced when transient deformation is included. This would then increase the effect of known fault-interactions modeled as steps in state due to changes in stress or strain because the steps in state due to the known fault-interactions would be larger with respect to the "noise." If earthquake probabilities are estimated solely by empirically fitting inter-event times, the transient loading effect need not be modeled because it produces a similar-shaped distribution as the standard BPT model. However, if the goal is to estimate earthquake probabilities based on the physics of seismogenic processes, then transient deformation must be included. For example, a transient-loading BPT model for the 1906 San Francisco earthquake produces a 60% greater variability in earthquake recurrence when compared to a uniform loading model and the same level of noise. Thus, if transient deformation is included in a BPT model for the faults of the San Francisco Bay region and the noise level is reduced to maintain the same variability in earthquake recurrence, then the effect of the stress change of the 1906 earthquake is greater than for a constant-loading BPT model. Prior work may therefore underestimate the effect of the 1906 stress shadow.

S31G-05 1120h

Frequency-Dependent Dynamic Triggering

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Dynamic triggering has now been robustly documented for at least 4 M_w>7 earthquakes. As broadband seismic network coverage becomes denser, the continuing absence of mid-range triggering from M6-7 earthquakes motivates us to ask under what conditions do seismic waves trigger earthquakes. We provide observational evidence here that one important factor in triggering is the frequency of the seismic waves. The key evidence comes from the cases of remote triggering at Long Valley caldera. At least three regional $M_w > 6$ earthquakes since 1990 have produced shaking in Long Valley that exceeded the local peak particle velocity (and therefore transient elastic strain) of the 2002 Denali earthquake, yet the Denali earthquake triggered local seismicity and the regional ones did not. For instance, the Northridge earthquake (M_w=6.7, Δ=381 km) produced shaking with a peak-to-peak vertical amplitude of 1.4 cm/s in Long Valley and the Denali earthquake (M_w=7.9, Δ=3458 km) had peak-to-peak shaking of 1.0 cm/s in the same spot. However, the locally weaker Denali earthquake triggered local earthquakes and Northridge did not. The Denali-triggered events began within 10 s of the beginning of the Rayleigh wave particle motion, which is less than the duration of the Northridge wave train. Therefore, it is difficult to appeal to the difference in duration of the wavetrains. We turn instead to the alternative explanation that long-period waves ($\gtrsim$10–15 s) are more effective at triggering earthquakes than short-period ones. Under a low-pass filter with a corner at 15 s, Denali produced 2.5 times greater shaking at Long Valley than Northridge. The other examples are also consistent with this threshold. The sensitivity of triggering to long-period waves is consistent with a barrier-clearing model for triggering where local earthquakes are caused by a redistribution of pore pressure on a fault due to the unclogging of local flow paths by the seismically-induced flow. The long-period seismic waves may be more effective at clearing fractures than short-period ones because the longer the period of oscillation, the larger the total volume entrained. The result is that thicker barriers can be removed by long-period waves than short-period ones. The thicker barriers support larger head differences if all else is equal. Therefore, long-period waves may produce larger pressure changes on the fault than short-period ones of the same amplitude. Other potentially frequency-dependent mechanisms, such as rate and state friction, cannot explain the sustained triggering after the end of the seismic waves (Gomberg, JGR, 2001).

S31G-06 1135h

How Stress Transfer Between Volcanic and Seismic Zones can Suppress or Trigger Earthquakes

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In the South Iceland Seismic Zone (SISZ), major destructive earthquakes occur at intervals of 80-100 years, the largest events reaching M7.1. Although the zone is E-W trending, with an overall sinistral movement, most of the earthquakes are associated with NNE and ENE-trending strike-slip faults. Numerical models indicate that when the SISZ is loaded to failure, stresses are transferred to, and concentrate at, its ends: tensile in the northeast and southwest quadrants and compressive in the northwest and southeast quadrants. These model results agree with observations made when the SISZ was loaded to failure in June 2000, resulting in two M6.6 earthquakes. For several years prior to the June 2000 earthquakes, compression, uplift and intense earthquake swarms (some events exceeding M4) occurred in two active stratovolcanoes, located in the quadrants of compressive stresses. Most of this earthquake activity came to an end following the June 2000 earthquakes. Also, in the decades prior to the June 2000 earthquakes, the stratovolcano Hekla, located in one of the quadrants of tensile stresses, has erupted unusually frequently. These stratovolcanoes (with shallow magma chambers) act as “soft inclusions” to which stresses from the SISZ are transferred; they may be used as precursors to large earthquakes in the SISZ. Stress is also transferred from the volcanic zones to the seismic zones. For example, stress transfer from the overpressured 27-km-long feeder dike of the Laki 1783 eruption to the nearby SISZ triggered the 1784 earthquake sequence, the largest one in the SISZ. By contrast, dike injection in 1976 in the North Volcanic Zone during the Krafla Fires resulted in suppression of earthquake activity in the Husavik-Flatey Fault (HFF), an active transform fault partly exposed on land, for nearly two decades. For regional dikes such as these, the magmatic overpressure may reach tens of megapascals; part of this pressure is transferred as compressive stress to the nearby seismic zones, the SISZ and the HFF. The models presented here indicate that stress transfer from a dike injected in a volcanic zone either triggers or suppresses earthquake activity in a nearby seismic zone, depending on the location of the dike in relation to that zone. When the dike is so located that the dike-generated horizontal displacement is opposite to the normal displacement across the adjacent seismic zone, earthquakes are suppressed; when the dike-generated displacement is consistent with the normal displacement across the adjacent seismic zone, earthquakes are triggered.

S31G-07 1150h

Remotely Triggered Earthquakes in Intraplate Regions: Distributed Hazard, Dependent Events

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The central and eastern United States has experienced only 5 historic earthquakes with Mw above 7.0, the 1886 Charleston earthquake and four during the New Madrid sequence of 1811-1812 (three principal mainshocks and the so-called “dawn aftershock” following the first mainshock.) Careful consideration of historic accounts yields compelling evidence for a number of remotely triggered earthquakes in both 1812 and 1886, including several events large enough to be potentially damaging. We propose that one of the (alleged) New Madrid mainshocks, on 23 January 1812, may itself be a remotely triggered earthquake, with a location some 200 km north of the New Madrid Seismic Zone. Our proposed source location is near the location of the 1968 southern Illinois earthquake, which occurred on a blind thrust fault at 20-25 km depth. Intensity data for the 1812 event are consistent with expectations for a similarly deep event. Such triggered events presumably do not represent a wholly new source of hazard but rather a potential source of dependent hazard. That is, the common assumption is that the triggering will cause only a “clock advance,” rather than causing earthquakes that would not have otherwise occurred. However, in a low strain-rate region, a given dynamic stress change can potentially represent a much larger clock advance than the same change would cause in a high strain-rate region. Moreover, in regions with low seismicity and a short historic record, overlooked remotely triggered historic earthquakes may be important events. It is thus possible that significant events are currently missing from the historic catalogs. Such events–even if large–can be difficult to identify without instrumental data. The (interplate) 1905 Kangra, India earthquake, further illustrates this point. In this case, early seismic records provide corroboration of an early triggered event whose existence is suggested–but difficult to prove–based on detailed macroseismic data. In the central United States, where even moderate earthquakes are uncommon, our results suggest that the largest known historic earthquakes in three states (Kentucky, Illinois, and Mississippi) may have been remotely triggered earthquakes.

URL: http://pasadena.wr.usgs.gov/office/hough

S31H MCC: 3011 Wednesday 1020h

Upper Mantle and Lithosphere *(joint with V)*

Presiding: **Y Shen,** University of Rhode Island**; D L Schutt,** University of Wyoming

S31H-01 1020h

The Geometry of Recently Digested India: First Results From HIMNT Receiver Functions Spanning the Collision Zone in Nepal and Tibet

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Receiver functions from the first broadband seismic deployment to cover the plains of south Nepal, the Lesser and Greater Himalaya, and southern Tibet, reveal the base of the Indian continental crust to be descending gently from 45 km depth in the plains through a steeper ramp in the transition through the high ranges, to 70-80 km depth underneath the Tibetan Plateau. Our data from a network of 29 stations span 20 months and provide roughly 3000 high signal to noise 3-component teleseismic P-SV seismograms, allowing the construction of 3D images of the Himalayan crustal and mantle structure. Receiver functions at most stations exhibit strong back azimuthal variations, and we calculate common conversion point stacks to correctly locate P to S converters. We see a clear Moho descending from the south to the north. The crustal thickness in Tibet is consistent with results from INDEPTH studies to the east of our deployment area. At the transition from the Nepalese foothills to the high Himalaya, we observe a midcrustal conversion at an average of 20 km depth, which coincides with a zone of shallow earthquakes relocated using our stations (Monsalve et al., this meeting). We observe along-arc variations in the depth of the Moho and of midcrustal structures. It is possible that some of these midcrustal features are associated with regions of partial melt observed in the INDEPTH data. Our new seismic results allow us to constrain previous inferences from gravity and structural studies for a clearer picture of the mechanisms of the Himalayan collision.

S31H-02 1035h

Probing the Southern Indian Shield with *P*-Wave Receiver-Functions

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We demonstrate a rare, successful case of using *P*-wavefield recorded on vertical-component seismograms alone to constrain crustal properties of the southern Indian shield. The data comprise of digital seismograms recorded by the short-period Gauribidanur Array (GBA) situated on Archean crust. The *P*-wavefield contains critical information, such as average speed of the *P*-wave in the crust (V_P), which is obscured in standard receiver-functions based on *P* to *SV* conversions. Moreover, conventional receiver-functions require all three components of seismograms, thus tripling the workload. In contrast to previous reports of *P*-wave receiver functions where the results have been equivocal, primary reflection off the Moho, the *PpPmp* phase, is visible and coherent in unprocessed data across the GBA array where the signal-to-noise ratio is excellent. We use eigen-image decomposition to estimate the source wavelet whose effect (including the instrument response) is then removed from seismic profiles by deconvolution. The resulting profiles show that the coda of *P*-wave is dominated by the *PpPmp* phase, whose uniform timing relative to the direct *P*-phase indicates a sharp, sub-horizontal Moho beneath GBA. Building upon previous reports of crustal thickness (35 $\pm$ 2 km) and average speed of the *S*-wave in the crust (V_S, 3.60 km/s), the *P*-wave receiver functions indicate a V_P of 6.4 $\pm$ 0.1 km/s in the crust. Low values of V_P, V_S, and the Poisson ratio (0.268 $\pm$ 0.011) are consistent with well-constrained results from Hyderabad, bolstering the interpretation that the entire thickness of Archean crust in southern India is felsic to intermediate in composition.

S31H-03 1050h

Surface-Wave Test of the Upper Mantle Low Velocity Zone

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Surface Waves hold the potential to detect the mantle low velocity zone (LVZ). We test the influence of a LVZ in the depth range of 100-180 km on the group and phase velocity dispersion curves of the fundamental and the first few higher modes as well as on the waveforms of both Rayleigh and Love waves. We invert calculated synthetic dispersion curves and real data from the Siberian Kraton area by the non-linear inversion Hedgehog method.
Our reference structure is the average 1D-model obtained from body wave tomography along profile Kraton. S-wave velocity is calculated for typical Vp/Vs ratios and density is estimated from the Nafe-Drake relation. We test several models: the reference model itself and the reference model with velocity reductions of 2% and 5% in the LVZ for different thicknesses of the LVZ. The group and phase velocity dispersion curves of the fundamental and higher modes are calculated for an elastic flat Earth model. Average errors ($\pm$ 0.06 and $\pm$ 0.04 km/s for group and phase velocity, respectively) are superimposed before the inversion. In order to check if the thickness of the LVZ can be resolved, synthetic seismograms are calculated for several depths of the sources, both for Rayleigh and Love waves, for the same source parameters and epicentral distances as for real events with travel paths across Siberia.
The differences between the seismograms are small, even when the source is located in or around the LVZ, which is the only situation where a LVZ can be detected. The LVZ cannot be identified in the ambient