S32C-08 1525h

Anisotropic Parameter Estimation From PP and PS Waves in 4-Component Data

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In recent years we have seen that in many cases anisotropy cannot be neglected in seismic processing routines. Furthermore we know that sea-bed properties play an important role in marine seismics, particularly for sea-bed multicomponent seismic, because they significantly influence the amplitudes, phases and travel times of seismic waves.
In this work we investigate how reliably we can extract anisotropy parameters from P- and PS-wave data using near-surface layer multicomponent data from the Alba Field, North Sea. Near-surface data is affected by apparent anisotropy due to a vertical velocity compaction gradient. Furthermore, we can see phase reversals in the pressure wave data set which will lead to erroneous anisotropy parameter values. Thus, the imaging of the target area is affected. We extract the P-wave anisotropy parameter η (Alkhalifah and Tsvankin, 1995) which can be related to Thomsen's ϵ and δ formulation (Thomsen, 1986) by $\eta = (\epsilon - \delta)/(1 + 2\delta)$ from P-wave data and compare the result with reference values from a model, built according to well, VSP and surface data. As the parameters show differences up to an order of magnitude we use PS-wave data to estimate the effect of anisotropy. The effect of anisotropy on the converted waves in a single layer can be described by the parameter

$$\chi = (\gamma_0 - 1)\gamma_{eff}^2 \eta, \quad with \ \gamma_{eff} = \frac{\gamma_2^2}{\gamma_0},$$

where γ_0 is the vertical velocity ratio, γ_2 the rms velocity ratio and γ_{eff} the effective velocity ratio. This equation is true for a single TIV layer. In this case η can be calculated from χ and vice versa. Thus, estimating χ from converted-wave data we can compute η for the near-surface events. The resulting parameters fit the reference values much better and show values of the same order of magnitude.

S32D MCC: 3009 Wednesday 1340h

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

Presiding: **I G Wong,** Seismic Hazards Group, URS Corporation; **G Anderson,** U.S. Geological Survey

S32D-01 1340h

Is earthquake triggering driven by small earthquakes?

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Using a catalog of seismicity for Southern California, we show that the number of triggered earthquakes increases exponentially with the mainshock magnitude. The trade-off between this scaling and the distribution of earthquake magnitudes controls the relative role of small compared to large earthquakes. We show that seismicity triggering is driven by the smallest earthquakes, which trigger fewer aftershocks than larger earthquakes, but which are much more numerous. The magnitude of triggered earthquakes obeys the Gutenberg-Richter distribution, with the same b-value as other earthquakes, independently of the size of the triggering earthquake. We propose that the non-trivial scaling of the number of aftershocks with the mainshock magnitude emerges from the fractal spatial distribution of aftershocks. These observations highlight the importance of taking into account the effect of small earthquakes to model earthquake triggering.

S32D-02 1355h

Importance of direct and indirect triggered seismicity and derivation of Bath's law for the largest aftershock

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Using the simple ETAS branching model of seismicity, which assumes that each earthquake can trigger other earthquakes, we quantify the role played by the cascade of triggered seismicity in controlling the rate of aftershock decay as well as the overall level of seismicity in the presence of a constant external seismicity source. We show that, in this model, the fraction of earthquakes in the population that are aftershocks is equal to the fraction of aftershocks that are indirectly triggered and is given by the average number of triggered events per earthquake. Previous observations that a significant fraction of earthquakes are triggered earthquakes therefore imply that most aftershocks are indirectly triggered by the mainshock. For some values of the parameters, the ETAS model also reproduces Bath's law, which states that the average magnitude difference $\langle \Delta m \rangle$ between a mainshock and its largest aftershock is 1.2, regardless of the mainshock magnitude. We first point out that the standard interpretation of Bath's law [Vere-Jones, 1969] in terms of the two largest events of a self-similar set of independent events is incorrect, because it neglects the selection procedure entering the definition of aftershocks. We reconcile Bath's law with (i) the existence of a universal Gutenberg-Richter (GR) law for all earthquakes and (ii) with the empirical observation (productivity law) that each earthquake of magnitude m triggers other earthquakes at a rate $\sim 10^{\alpha m}$ with $\alpha \approx 0.8$.

S32D-03 1410h

Testing of hypotheses associated earthquake clustering features by using the stochastic reconstruction method

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Based on Ogata's ETAS model and the thinning procedure, we firt give an iterative approach to estimate the background occurrence rate, which is a function of locations but constant of time, and the parameters of the clustering struction, simultaneously. We then discuss the methods for sperating the seismicity in a given catalogue into a background component and a clustering component and for classifying the earthquakes into different clusters stochastically. The keys of these methods are the probabilities of one event being a direct offspring of another event or being a background event. Making use of these probabilities, we can reconstruct the functions associated with the characteristics of earthquake clusters, which enable us to test a number of important hypotheses about the earthquake clustering phenomena. In this paper, using the JMA catalogue for illustruction, the results show the following assertions: (a) The function forms provided on each clustering component in the formulation of Ogata's ETAS model are good enough as a first approximation in describing earthquake clusters; (b) the background events trigger offspring in a different ways from the triggered events; (c) The diffusion of the aftershock sequence is mainly caused by the cascades of individual trigger processes, while each individual process for triggering direct offspring is not diffusive; (d) the scales of the area of triggered events are mainly associated with the triggering ability of their direct ancestor.

S32D-04 1425h

Detection of Anomalous Seismicity as a Stress Change Sensor

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Seismic quiescence has been attracting the attention as one of the precursors to a large earthquake. Also, it has been getting more attention than ever that the stress changes due to a slip can explain the mechanism of triggering another event. Likewise, seismic quiescence due to stress lowering transferred from a rupture or silent slip elsewhere can be discussed as a general phenomenon. Specifically, the quiescence in an active focal area could be found in the shadow zones where Coulomb stress change would be negative from a seismic or aseismic slip nearby. The seismicity rate lowering could be well seen in the regions where the activity has been high, including aftershock activity itself. To see the seismicity change explicitly, however, we need to fit a statistical model of either the modified Omori formula or its extension (the ETAS model [Ogata, 1992]), to the seismic activity before the suspected triggering time. Then, the activation and lowering of the seismicity relative to the predicted one by the model are of great concern to explore matching the pattern of Coulomb's stress changes due to a rupture or silent slip. A number of such examples from the recent activities in Japan will be shown. These lead us to a summarized observation that even a small size of negative ΔCFF down to the order of millibars can trigger such lowering, which is also supported by the Dieterich's seismicity rate equation [Dieterich, 1994; Toda and Stein, 2002]. Thus, we expect that aftershock and general seismic activity relative to the modeled rates is sensitive enough to measure a slight negative stress change. An earthquake prediction scenario from the implication of the relative quiescence is based on the asperity hypothesis in the sense that precursory slip around asperities applies more shear stress to the asperities, which in turn promotes the rupture of the main fault. However, on the other hand, aseismic slips in some region is not necessarily a precursor to a large event. Indeed, we have observed a number of aseismic slips (silent earthquakes; sometimes repeated in the same region) with no subsequent large events in the last decades. This is consistent with some empirical results that (relative) quiescence was not always followed by a large event. Therefore, distinguishing whether or not an aseismic slip leads to the rupture of asperity remains a further difficult research theme in earthquake prediction. At present, this issue can only be described in terms of probabilistic prediction [Ogata, 2001]. References Dieterich, J., H., J. Geophys. Res., 99, 2601-2618, 1994. Ogata, Y., J. Geophys. Res. 97, 19845-19871,1992. Ogata, Y., J. Geophys. Res. 106, 8729-8744, 2001. Ogata, Y., Jones, L.M. and Toda, S., J. Geophys. Res., 108, 10.1029/2002JB002009, 2003. Toda, S. and Stein R.S., J. Geophys. Res., 107, 10.1029/2001JB000172, 2002

URL: http://www.ism.ac.jp/~ogata/

S32D-05 1440h

Bayesian Analysis of Aftershock Sequences

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The Reasenberg-Jones (1989) or RJ model describes the time-magnitude hazard of an aftershock sequence by combining the modified Omoris law (Utsu, 1961) and the Gutenberg-Richter (1954) rule. The parameters estimated separately from the power-decaying law and magnitude-frequency relation are then integrated for analyzing aftershock sequences in California. Wiemer (2000) further applied the RJ model to spatial grids over a possible aftershock area for evaluating the hazard or probability of a future aftershock. In this study, we propose a Bayesian analysis of the RJ model for a aftershock sequence with a non-informative prior or prior distributions based on previous aftershock sequences. Note that, under a non-informative prior, the Bayesian version of the RJ model based on posterior modes of the related parameters provides with an approximation to the estimated RJ model which maximizes the associated likelihood. The three versions of the RJ model, including the Bayesain, maximum likelihood and the integrated ones, will be applied to the Chi-Chi (M=7.6, 1999/9/20 UT) aftershocks for comparative purpose.

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S32D-06 1455h

The Prediction of Spatial Aftershock Probabilities (PRESAP)

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It is now widely accepted that the goal of deterministic earthquake prediction is unattainable in the short term and may even be forbidden by nonlinearity in the generating dynamics. This nonlinearity does not, however, preclude the estimation of earthquake probability and, in particular, how this probability might change in space and time; earthquake hazard estimation might be possible in the absence of earthquake prediction. Recently, there has been a major development in the understanding of stress triggering of earthquakes which allows accurate calculation of the spatial variation of aftershock probability following any large earthquake. Over the past few years this Coulomb stress technique (CST) has been the subject of intensive study in the geophysics literature and has been extremely successful in explaining the spatial distribution of aftershocks following several major earthquakes. The power of current micro-computers, the great number of local, telemeter seismic networks, the rapid acquisition of data from satellites coupled with the speed of modern telecommunications and data transfer all mean that it may be possible that these new techniques could be applied in a forward sense. In other words, it is theoretically possible today to make predictions of the likely spatial distribution of aftershocks in near-real-time following a large earthquake. Approximate versions of such predictions could be available within, say, 0.1 days after the mainshock and might be continually refined and updated over the next 100 days. The European Commission has recently provided funding for a project to assess the extent to which it is currently possible to move CST predictions into a practically useful time frame so that low-confidence estimates of aftershock probability might be made within a few hours of an event and improved in near-real-time, as data of better quality become available over the following day to tens of days. Specifically, the project aim is to assess the extent to which this is scientifically feasible in terms of our understanding of the physical phenomena which control the variation of seismicity following a large event due to stress redistribution and practically possible given present limitations on data availability, data quality and computational or data transfer speeds. The project is divided into a number of elements designed to reflect the temporal sequence of tasks that must be undertaken for the prediction of aftershock hazard. These tasks include determining a time-indexed sequence of slip distributions for both real and synthetic events, calculating a suite of time-indexed stress perturbations and quantitatively comprising predicted and observed aftershock distributions, and developing techniques for predicting likely strong ground motion from the predicted spatial distribution of aftershocks.

S32D-07 1510h

A Possible Triggering Mechanism for the Earthquake Swarms in Vogtland/NW-Bohemia: Data Analysis and Numerical Modeling

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Vogtland/NW-Bohemia (VB) at the German/Czech border region, central Europe, is characterized by recurring intraplate earthquake swarms, Quaternary volcanism, carbon-dioxide-rich mineral springs, a complex tectonic environment of intersecting fault systems, and an established connection between ascending magmatic fluids and seismicity. As the triggering mechanism of seismicity in VB is still unknown, this study aims to contribute by investigating the possible role of fluids in VB, based on the theory of linear poroelasticity. We assume that ascending fluids, of magmatic origin, trigger earthquakes by the mechanism of pore-pressure diffusion (i.e. relaxation). Two physical fields, the hydraulic diffusivity and the criticality (i.e. critical pore-pressure value leading to failure), both heterogeneously distributed in rocks, mainly control this triggering process. The results of the analysis of the year 2000 earthquake swarm data allowed the identification of diffusive characteristics, thus supporting this concept. It is further strengthened by a numerical model, with correlated high-diffusivity and high-criticality patches (corresponding to highly fractured and permeable compartments of the fault zone region in VB), which successfully simulates the general seismicity pattern of the earthquake swarms 2000, including the spatio-temporal clustering of the events, and the seismic activity migration.

S32D-08 1525h

Plate-Tectonic Analysis of Shallow Seismicity: Apparent Boundary Width, beta-Value, Corner Magnitude, Coupled Lithosphere Thickness, and Coupling in 7 Tectonic Settings

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A new plate model [*Bird*, 2003, G^3, 10.1029/2001GC000252] is used to analyze the mean seismicities of 7 types of plate boundary (CRB continental rift boundary, CTF continental transform fault, CCB continental convergent boundary, OSR oceanic spreading ridge, OTF oceanic transform fault, OCB oceanic convergent boundary, SUB subduction zone). We compare the plate-like (non-orogen) regions of model PB2002 with the CMT catalog to select apparent boundary half-widths, and then assign 95% of shallow earthquakes to one of these settings. A tapered Gutenberg-Richter model of the frequency/moment relation is fit to the subcatalog for each setting by maximum-likelihood. Best-fitting β values range from 0.53 to 0.92, but all 95%-confidence ranges are consistent with a common value of 0.61-0.66. To better determine some corner magnitudes we expand the subcatalogs by: (1) inclusion of orogens; and (2) inclusion of years 1900-1975 from the catalog of *Pacheco and Sykes* [1992]. Combining both earthquake statistics and the plate-tectonic constraint on moment rate, corner magnitudes include: CRB $7.64_{-.26}^{+.76}$, CTF $8.01_{-.21}^{+.45}$, CCB $8.46_{-.39}^{+.21}$, OCB $8.04_{-.22}^{+.52}$, and SUB $9.58_{-.46}^{+.48}$. Coupled lithosphere thicknesses are found to be: CRB $3.0_{-1.4}^{+7.0}$ km; CTF $8.6_{-4.1}^{+11}$ km; CCB $18_{-11}^{+?}$ km; OSR $0.13_{-0.09}^{+.13}$ km for normal-faulting and $0.40_{-.21}^{+?}$ km for strike-slip; OTF $12_{-7.1}^{+?}$, $1.6_{-0.5}^{+1.4}$, and $1.5_{-0.6}^{+1.2}$ km at low, medium, and high velocities; OCB $3.8_{-2.3}^{+13.7}$ km, and SUB $18.0_{-10.8}^{+?}$ km. Generally high coupling of subduction and continental plate boundaries suggests that here all seismic gaps are dangerous unless proven to be creeping. Generally low coupling within oceanic lithosphere suggests a different model of isolated seismic asperities surrounded by large seismic gaps which may be permanent.

URL: http://moho.ess.ucla.edu/~kagan/kagan.html

S32E MCC: 3011 Wednesday 1600h

Global and Regional Tomography II *(joint with DI)*

Presiding: **D W Vasco,** Berkeley Laboratory, University of California, Berkeley**; R D van der Hilst,** Massachusetts Institute of Technology

S32E-01 1600h

Resolution Limit of Traveltime Tomography

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We derive an explicit formula relating a recovered 3-D seismic slowness image to the actual seismic slowness, under the assumption that a tomographically complete collection of finite-frequency traveltime shifts is inverted using linearized ray theory. The tomographically recovered image is blunted, or diminished in amplitude, and blurred, or spread laterally, as a result of wavefront healing and other finite-frequency diffraction effects. The 1-D spreading $\sim$ (amplitude reduction)$^{-1/3}$, so that total volume-integrated slowness is conserved.

S32E-02 1615h

3D mantle P-velocity model based on uniform tetrahedronization of the Earth and the natural-neighbour interpolation method

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Traditional parameterization of three dimensional (3D) Earth models consists of radial and horizontal basis functions with a constant lateral grid size, independent of depth. As a result, the horizontal basis functions become more densely distributed with increasing depth. The nominal resolution thus becomes finer. For example, in the lower mantle and D" region, current whole-mantle models appear to be characterized by particularly short-wavelength structures and high gradients, which might not be real features. In this study, we use instead a 3D parameterization generated by an "advancing front method" (NETGEN), given an initial triangulation on the surface of the Earth. The result is a number of tetrahedrons of optimal and roughly similar size making up of the Earth's volume. We then utilize the natural neighbour interpolation technique to set up the whole mantle 3D inverse problem. The functional value (in our case, perturbation to the P-velocity) at any point within the Earth can be found through interpolation of the values of its natural neighbours (on the order of 10). The number of nodes decreases with depth as the volume decreases. The basis function at each node is smooth, and dependent on the distances of other nodes around it. No separation of radial and horizontal basis functions is needed. Synthetic tests with 3D S-model S362 show that most of significant structures of S362 are recovered. Two major differences exist: (1) at 2,500 km and 2,800 km depth, the recovered model is smoother and has less short-wavelength signal than S362; (2) at the surface and CMB, the recovered S-velocity perturbation tends to be larger than S362. Currently we are deriving a whole mantle P-model. The preliminary result is promising. Compared to the traditional approach which employs both radial and horizontal basis functions, this parameterization allows a true 3D representation of the Earth's seismic velocity perturbations. It is also straightforward to build norm and roughness damping matrices and apply them to the target model. Because of the ease of the grid generation and the grid-dependent basis functions, it is also possible to introduce multigrid (finer resolution in a particular region) into the 3D seismic tomography to achieve multi-scale resolution.

S32E-03 1630h

Long-period mantle structure from Earth's free oscillation spectra

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Models of the Earth's long period three dimensional structure are obtained from the study of the splitting of Earth's free oscillation spectra. These data place important constraints on the aspherical compressional and shear wave velocity structure, and in addition has the capability to provide information on aspherical density structure. In general, *splitting functions* are measured from the normal mode spectra using a least squares inversion. These splitting functions are linearly dependent upon the aspherical velocity and density structure and are then used in a second least squares inversion to obtain tomographic velocity models. Many recent tomographic models, including S20RTS, have used splitting functions in addition to body wave and surface wave data. In principle, normal mode spectra could also be inverted directly to get tomographic models. In order to use splitting functions, it has been assumed that modes are isolated and therefore can be studied in the self-coupling approximation or that only small groups of modes need to be coupled. We will show that for many modes, this represents a poor approximation compared to full coupling, where large groups of modes are coupled. This implies that it is actually not accurate enough to use splitting functions. In order to access the influence of the splitting function approximation on the long period structure in tomographic models, we have performed tests for a data set of normal mode spectra of recent large earthquakes, including the Bolivia and Kuril Island events of 1994. We will show the different tomographic models obtained from a two-step inversions using splitting functions compared with a one-step direct inversion of the normal mode spectra. We will also show the difference between a direct inversion of normal spectra using the self coupling approximation versus the group coupling approximation.