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Project Title: Analysis of Los Angeles Region Seismic Experiment (LARSE) II teleseismic data for 3D lithospheric structures in the western Transverse Ranges, San Fernando Valley, and Santa Clarita Valley

Monica D. Kohler, UCLA

During 2000 I focused my SCEC-related efforts on two projects: analysis of LARSE II passive teleseismic data and development of a new raypath density-based inversion parameterization. The LARSE II teleseismic data analysis included 1) travel-time inversion for modifications to my upper mantle P-wave velocity model to be included in the SCEC 3D Seismic Velocity Model, Version 3, 2) calculation of receiver functions for Moho depth variations below the western Transverse Ranges, San Fernando Valley, and Santa Monica Mountains (collaborative work with L. Zhu), and 3) SKS splitting measurements for anisotropy below the LARSE II array (collaborative work with P. Davis).

The complete LARSE II passive dataset is archived at the IRIS DMC and data analysis is now well on its way to completion. Teleseismic travel times have been measured by cross correlation for approximately 50 teleseisms recorded by the majority of the array with good signal-to-noise. The travel times were combined with a previously compiled file of SCSN travel times for inversions to update my uppermost mantle velocity model (Kohler, JGR, 1999). The new model parameterization consists of grids that are 10 by 10 km where the data are dense, and 20 by 20 km where they are sparser. A related task included removing crustal velocity and Moho depth variation effects from the teleseismic travel times by raytracing through the SCEC 3D Seismic Velocity Model, Version 2. Thus, the effects of crustal heterogeneity are stripped off the travel times before the mantle velocities are determined.

In collaboration with L. Zhu, receiver functions have been computed for 4527 P waveforms from 87 teleseismic events.[1] The San Fernando Basin and the Santa Clarita Basin are well imaged with the basin bottoming at 6 to 8 km depth. In addition, a low-velocity patch exists near the surface under the Antelope Valley which might be an old sedimentary basin or low-velocity rocks. The Moho is seen clearly as a continuous flat feature at a depth of 34 km under the Mojave Desert. It terminates near the downward extension of the San Andreas fault. There are no detectable P-to-S conversions at a depth range of 20 to 40 km between the Santa Monica Mountains and the western Transverse Ranges. Instead, several offset conversion bands can be seen between depths of 40 to 50 km suggesting thickened crust. In a related effort, M. Bostock and I have just begun a collaborative project to compute crustal and upper mantle structure by applying his inverse-scattering formalism (Bostock, AGU abstract, 2000) to LARSE II teleseisms. I have also contributed to an SKS splitting analysis[2] in which splitting parameters calculated for LARSE II data show approximately east-west fast direction polarization with values consistent with previous studies.

The final task included development of a new inversion parameterization technique that accounts for variations in raypath density using continuous basis functions to represent velocity perturbation. The development of the wavelet-type inverse theory formulation is finished and testing on synthetic datasets has begun.[3] In this approach, velocity perturbations are represented as 2D Fourier series within layers of constant thickness. The raypath density for each layer is calculated at the minimum station spacing and converted into a Fourier series representation

(Fig. 1a,b). The raypath density map is assumed to represent how well fine-scale structure can be resolved and is used as a spatially-variable representation of the spatial Nyquist for velocity structure at any point. As each teleseismic ray is traced back from each station, the cutoff frequency for the travel-time perturbation representation of the ray at the layer midpoint is determined by the raypath density at that point. I have begun testing this scheme on synthetic data sets and velocity models (Fig. 1c,d) and found that it works well when the appropriate damping parameter is chosen (Fig. 2). Where the data are dense and fine-scale structure can be resolved, the inversion reproduces the small-scale structure from the initial model. Where the data are sparse, the inversion produces only smoothed, large-wavelength structure. Use of a 2D Gaussian filter prevents spurious edge effects, streaks, and bulls-eye effects not related to velocity structure or station coverage.

[1]Zhu, L., and **M. D. Kohler**, Preliminary results of crustal structure from the LARSE II passive recording experiment using teleseismic P-to-S converted waves, Eos, Trans., AGU (abstract), *81(48)*, F850, 2000.

[2]Davis, P. M., S. Baher, and **M. D. Kohler**, Seismic anisotropy of the Southern California uppermost mantle across the San Andreas fault, Eos, Trans., AGU (abstract), *81(48)*, F862, 2000.

[3]**Kohler, M. D.**, Using dense broadband USArray data to image lithospheric structures below the Pacific-North America plate boundary, Eos, Trans., AGU (invited abstract), *81(48)*, F893, 2000.

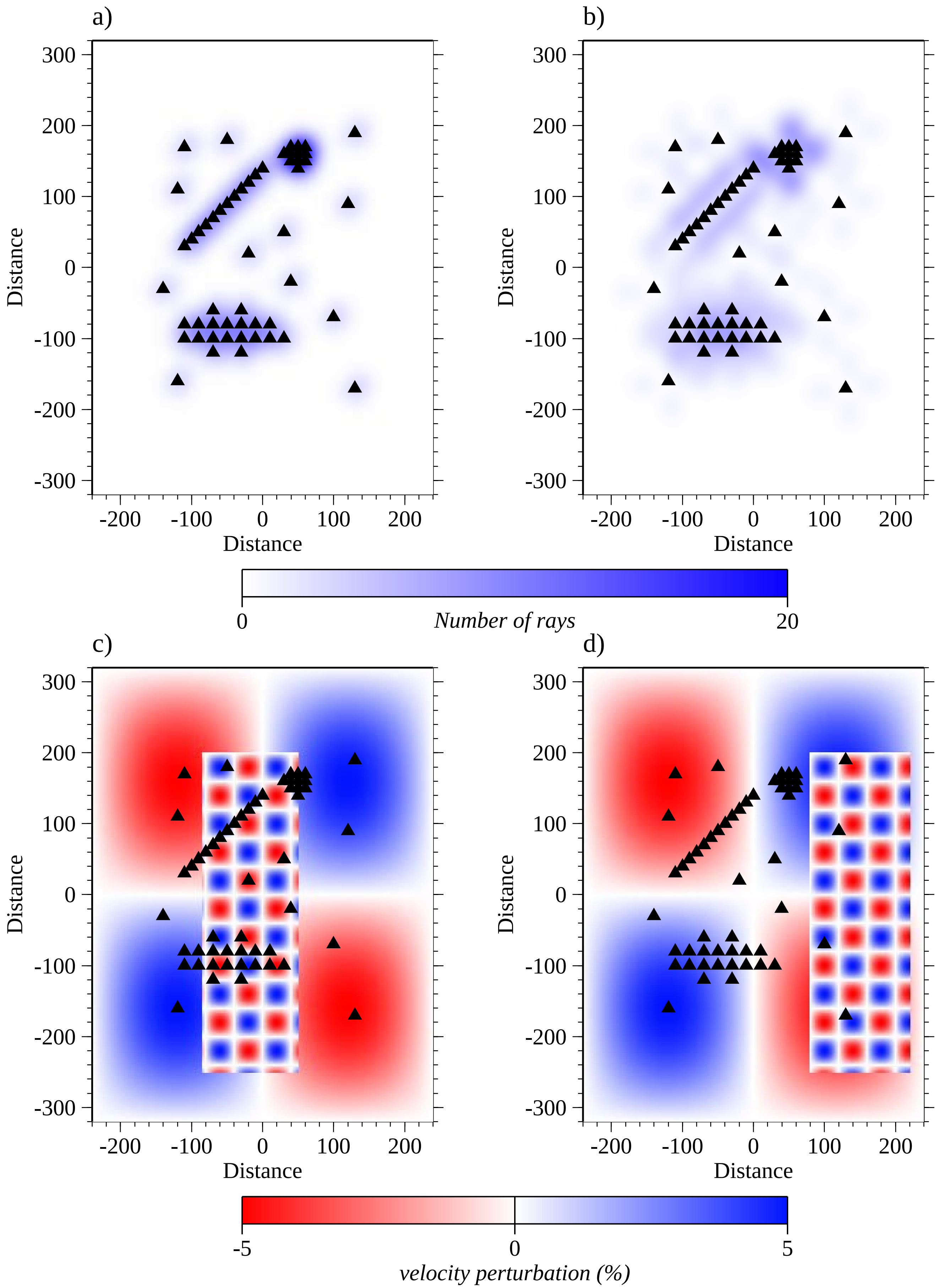

Figure 1. Plan view maps showing raypath coverage for a synthetic teleseismic travel-time dataset at depths of a) 30 km and b) 90 km. c,d) Input synthetic velocity models for tests of wavelet-type compared with standard, non-wavelet-type inversions. Triangles represent stations.

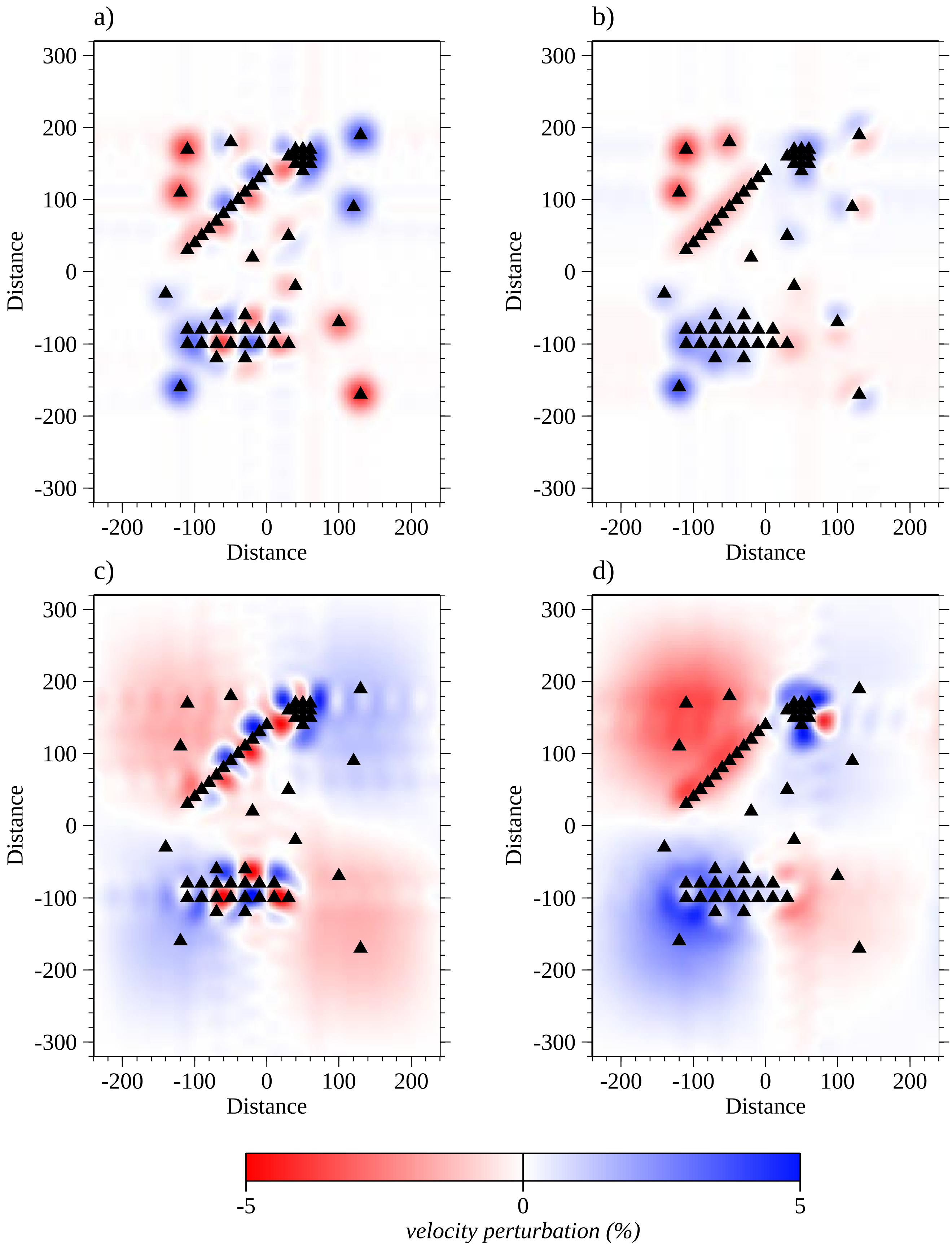

Figure 2. Results of inversions with a) synthetic input model 1, standard non-wavelet approach, b) synthetic input 2, standard non-wavelet approach, c) synthetic input 1, wavelet-type approach, and d) synthetic input 2, wavelet-type approach.