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Lessons on the calculation of static stress loading from the 2003 Bingol, Turkey earthquake

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Abstract

The 2003 Bingol earthquake ($M_w = 6.4$) occurred very close to a region along the east Anatolian fault zone which was identified in 2002 as posing a particularly high seismic risk. This damaging earthquake occurred on a conjugate right-lateral blind fault that was inconsistent with the stress-change field calculated for preceding large earthquakes in the region. In this paper, four reasons which might be responsible for this stress discrepancy are identified and investigated individually. Firstly, co-seismic stress changes are considered. The time frame of the previous stress calculations is extended to include the large earthquakes in the 1780s which were not included in the earlier study. A sensitivity analysis is then conducted on the more recent events since 1822 to examine the effect of errors in their location and sizes. The possibility of the occurrence of a small magnitude event close to the 2003 epicentre is considered. We argue that, barring the occurrence of a low-probability, unmodelled local event, the Bingol earthquake was unlikely to have been triggered by co-seismic stress transfer from any known sequence of previous earthquakes. Finally we examine and modify the secular loading model used in the 2002 study and show that loading which is properly constrained by regional GPS data produces a positive stress change on the 2003 rupture. As a result of our examination of the stressing history of the Bingol hypocentre we argue that it is through a combination of historical seismology, guided and constrained by structural geology, directed paleoseismology in which the locations and extent of historical events are confirmed, and stress modelling which has been informed by detailed GPS data, that an integrated seismic hazard program might have the best chance of success.

Keywords: seismic hazard; coulomb stress modelling; historical seismology

1. Introduction

There is a growing body of literature suggesting that neighbouring earthquakes interact elastically by influencing the occurrence time and location of subsequent earthquakes [1,2]. Most of the studies report a positive correlation between the stress field caused by preceding events and the location of the subsequent events [3–10].

Earthquake triggering has been extensively studied in Turkey and in many cases a correlation between positive Coulomb stress and the location of subse-

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quent events has been observed. Stein et al. [6], for example, studied large earthquakes that occurred along the NAF between 1939 and 1992 and showed that 9 out of 10 consecutive earthquakes were set up by the preceding events. After studying 29 earthquakes mostly occurring on the western extensions of the North Anatolian Fault (NAF) in the Marmara and North Aegean Sea regions in the last century, Nalbant et al. [8] concluded that the occurrence of 24 of them could be explained by stress loading due to their predecessors. They also highlighted that certain areas having active faults and positive stress loads should be regarded as posing particularly high seismic risk. The deadly 1999 Izmit earthquake occurred in one of these areas showing the importance of incorporating the stress calculations in seismic hazard studies.

This success prompted an investigation of seismic risk along the East Anatolian Fault Zone (EAFZ) by taking a similar but more refined approach (i.e. [10], here referred to as EAFZ02). The Coulomb stress perturbation due to the most reliable historical earthquakes and instrumentally recorded events since 1822 was combined with tectonic loading to produce a 180-yr stress change model for the region; this model was then resolved onto known active structures to produce a seismic hazard map. The tectonic load was calculated using the deep dislocation technique in which dislocation surfaces are introduced below the seismogenic layer and moved proportional to tectonic loading rates. The authors of that study found that the occurrence of 9 out of 10 events in the historical catalogue could be explained by advance-to-failure due to the stress interactions with previous events. The results also clearly indicated two highly loaded areas which were of particular concern: the first extended along the EAFZ between the cities of Kahraman Maras and Malatya and the second was

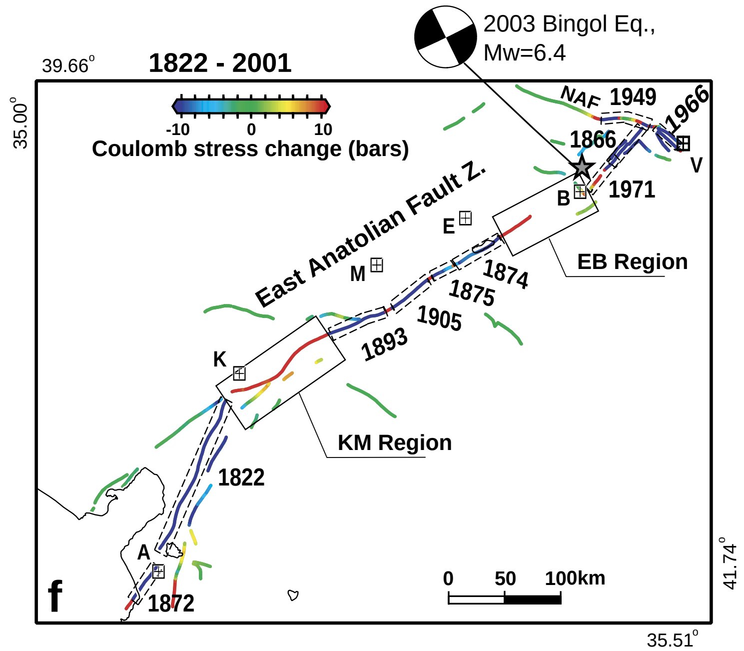

Fig. 1. Coulomb stress change along the East Anatolian fault zone (EAFZ) due to the large earthquakes and tectonic loading since 1822. This figure is modified from [10] and shows their main conclusions. The calculations were carried out on the digitised active faults. The rupture lengths of events are indicated by dashed rectangles labelled with their years of occurrence. The stress changes are mapped mid-crustal depth, 6.5 km. An average value, 0.4, of the effective coefficient of friction is used in the calculations. Hot colours, representing positive stress change, indicate tendency toward failure while cold colours show less likelihood to failure. Two areas were identified in 2002 as posing particularly high seismic risk; the KM and EB regions. The 2003 Bingol earthquake (solid star, epicentral location 39.01° N, 40.51° E) occurred very close to the EB region with a pure strike–slip focal mechanism solution (USGS-NEIC). The important cities along the EAFZ are A, Antakya; K, Kahraman Maras; M, Malatya; E, Elazig; B, Bingol and V, Varto. NAF stands for the North Anatolian fault.

located on an inferred extension of the EAFZ between the cities of Elazig and Bingol (Fig. 1).

The 1 May, 2003 Bingol earthquake (M_w=6.4) occurred within this second area with a focal mechanism which was consistent with left lateral rupture of a buried segment of the EAFZ, prompting suggestions that this represented a success for the idea of using Coulomb stress modelling to identify fault segments posing short term seismic risk (Fig. 1). This "success", however, depended on confirmation of the orientation of the causitive fault. One of the authors visited Turkey within 2 days of the event to attempt to confirm the mechanism by measuring the orientation of any surface ruptures. None existed. Furthermore, the first 2 days of the aftershock distribution which were poorly located did not clearly identify the orientation of actual rupture plane (pers. comm., Ömer Emre), [11]. The aftershock distribution of the first several days recorded by a temporary local network was published on the internet 10 days after the mainshock [12] and unambiguously showed that the event was a right lateral failure on an unmapped structure conjugate to the EAFZ [see also 13]. Resolution of the EAFZ02 stress-change model onto the identified Bingol fault showed a negative stress load (Fig. 2); the 2003 Bingol earthquake was not consistent with the stress field modelled in the 2002 study.

Here we revisit the 2002 calculations and attempt to identify the reasons for the discrepancy between the EAFZ02 stress model and the occurrence of the Bingol earthquake. Specifically we investigate the effect of four elements of EAFZ02 which could be problematic with a view to developing more effective protocols for future hazard assessments. Firstly, older historical earthquakes which were not included in the previous study may have significantly affected the stress field, these are included and discussed. Secondly, the historical earthquakes used in the 2002 study may have been incorrectly modelled due to uncertainties in magnitude and location. Thirdly, smaller earthquakes ($4.0 \leq M \leq 6.0$) located very close to the 2003 hypocentre, and whose stress perturbations were not modelled in 2002, may have had a significant, local effect on the stress field; we examine instrumental catalogues in the region to elucidate this possibility. Finally, we examine the method of calculating the secular stress used to load the faults and suggest that the use of regional

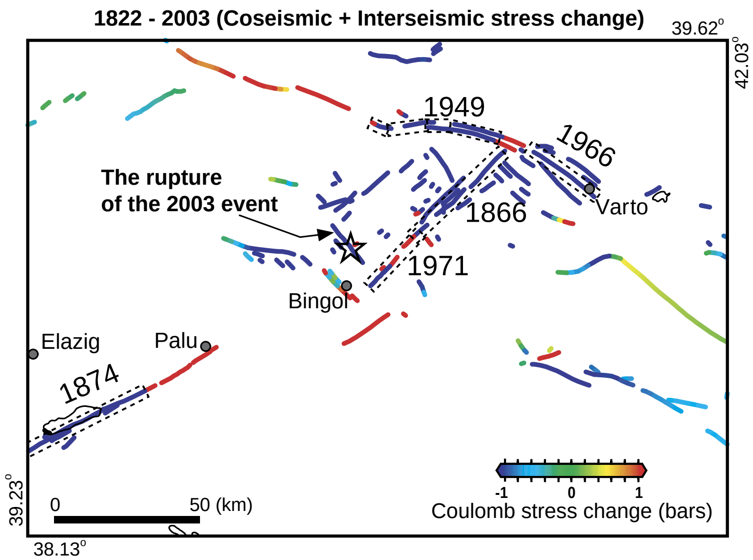

Fig. 2. A closer look at all the active faults in the vicinity of the 2003 event. Even though the 2003 fault was not mapped before and did not break the surface, we have inferred the rupture length of the event from magnitude-fault length empiric relations [14], and digitised it to show stress change on it. The total Coulomb stress change is plotted at a time just before the 2003 Bingol event resulting from all earthquakes and tectonic loading since 1822, using the same methodology as in [10]. The stress ranges along the 2003 fault from −2.1 bars at the NW to −13.6 at the SE tip.

GPS data might be important in the modelling of secular tectonic loading.

2. Possible causes for explanation of the occurrence of the 2003 event

Stress modelling studies inherently, and necessarily, assume a zero-stress baseline at some arbitrary time in the past and that stress changes since that time will dominate the total stress field. To be able to get the most correct stress picture on a fault, one has to start the calculations as far back in time as possible. In practice this starting time is controlled by the data quality requirements of the modelling in terms of location, magnitude and rupture geometry. As discussed in EAFZ02, the quality of the historical earthquake data along the EAFZ becomes suspect prior to the 1822 event. For this reason, two large earthquakes that occurred in 1784 and 1789 [14] were excluded from that study. Here we test whether including our best understanding of these two earthquakes in our calculations significantly changes the stress changed modelled on the 2003 hypocentre. As the events are poorly located, we assign them to the closest mapped active faults in the area that can support them. To do this take into account the damage to known habitation centres [14,15]. The town of Palu (Fig. 2), for example, was heavily damaged during the 1789 earthquake and aftershocks were strongly felt in that area [15]. The damage distribution due to the 1784 event was very close to the North Anatolian fault, so it is most likely that it occurred on it. Fig. 3 shows the result of including these events, beginning the tectonic loading in 1784, and employing the same computer code and methodology (homogeneous elastic half space, $\mu = 0.4$) as in the EAFZ02 paper. As the figure clearly shows, most of the 2003 fault plane, including the hypocentral area, experiences negative stress with positive stress (0.9 bars) only at the SE tip. The inclusion of these events hence cannot explain the occurrence of the Bingol earthquake and since there are no other significant large earthquakes in the area in recent historical times it is unlikely that missing large earthquakes are important. Note that these results are insensitive to the coefficient of friction. Since the high angle between conjugate faults of about 90° implies that the effective coefficient of friction in the area is very low, we check our results by computing only shear stress changes on the hypocentre (i.e. $\mu = 0$.)

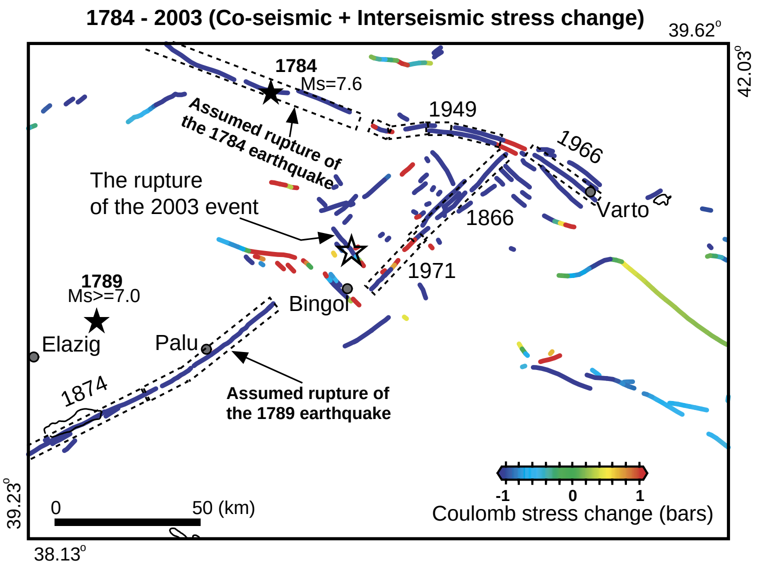

Fig. 3. Same as in Fig. 2 including older historical events and starting secular stress loading from 1784. Possible epicentres of the older historical events from [15] and [14] (solid stars) that are not modelled by Nalbant et al. [10] are also shown. Note that after inclusions of these events the stress change along the most of the 2003 rupture plane is negative.

and find that the shear stress at the hypocentre is negative with either the 1784 or 1822 baseline.

Significant uncertainties can arise in the estimation of location and magnitude of earthquakes using historical observations of intensity. These uncertainties, of course, imply corresponding uncertainties in the stress perturbations arising from the events. Additionally, intermediate sized earthquakes are rarely present in historical data yet such events may also influence the stress field.

Testing the importance of small ($4.0 \leq M \leq 6.0$) events is difficult as reliable event locations are only available from 1975. The closest event with respect to the epicentre of the 2003 earthquake is approximately 20 km distant which produces a very small stress change (10^{-4} bars) at the hypocentre of the Bingol earthquake. We cannot exclude the possibility, however, that an earlier intermediate size event may have occurred very close to the 2003 Bingol hypocentre and influenced its occurrence. This problem of our inability to model the stress due to small earthquakes near a future hypocentre is potentially important since it is known that large earthquakes can be triggered by small stress changes, and it is likely that dynamic stresses can propagate ruptures through regions of relatively low stress. The error in the hazard assessment is likely to be low, however, since the wavelength of the perturbation will be short and its propagation will not be common.

There are clearly some uncertainties involved in assessing the magnitude and faulting parameters of the historical earthquakes in the EAFZ02 study. Considering magnitudes 0.3 smaller and larger, which are the error limits involved in estimation of the magnitudes by Ambraseys [15], and adjusting the fault lengths and slip amounts accordingly, we find no significant changes around the hypocentral area (Fig. 4).

From the above, and not withstanding the small possibility of significant effect of an unmodelled and (very) local intermediate event, we can conclude that no reasonable change in the earthquakes in the 2002 study is likely to produce a significant change in the co-seismic part of the stress accumulated on the Bingol fault in the last 200 yrs. We therefore turn to the calculation of the secular component to explain the stress discrepancy.

The EAFZ02 study used the deep dislocation technique [16] to estimate the secular stressing increment for sections of the EAFZ. It is clear that this method, which seems to give good results for simple strike–slip geometries, has limitations, which might be severe, in regions of complex geometries such as the eastern terminus of the EAFZ against the NAF. In this tectonically complex corner, the left-lateral movement of the EAFZ is transferred to right-lateral movement mainly on the NAF and the area is cut by a complex network of conjugate faults which have been mapped

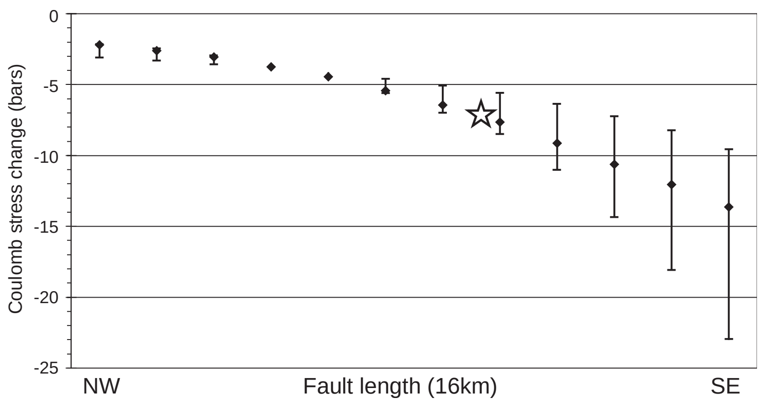

Fig. 4. Coulomb stress along the 2003 Bingol rupture at hypocentral (10 km) depth due to earthquakes and tectonic loading along the EAFZ since 1822. Diamonds show the values of stress computed using the methodology employed in the EAFZ02 study. Error bars indicate upper and lower limits of Coulomb stress changes due to perturbing fault length and slip of historical earthquakes within given magnitude uncertainty limits (± 0.3) [15]. Star indicates the location of the Bingol hypocentre.

at all scales. If, for example, the interaction between the NAF and EAFZ were to produce mechanical rotation thus complicating the strain distribution throughout the region that, the deep dislocation method modelling deterministic displacements on a few selected faults would be unlikely to model the stress accumulation accurately. We, therefore, investigate the effect of applying a different tectonic loading

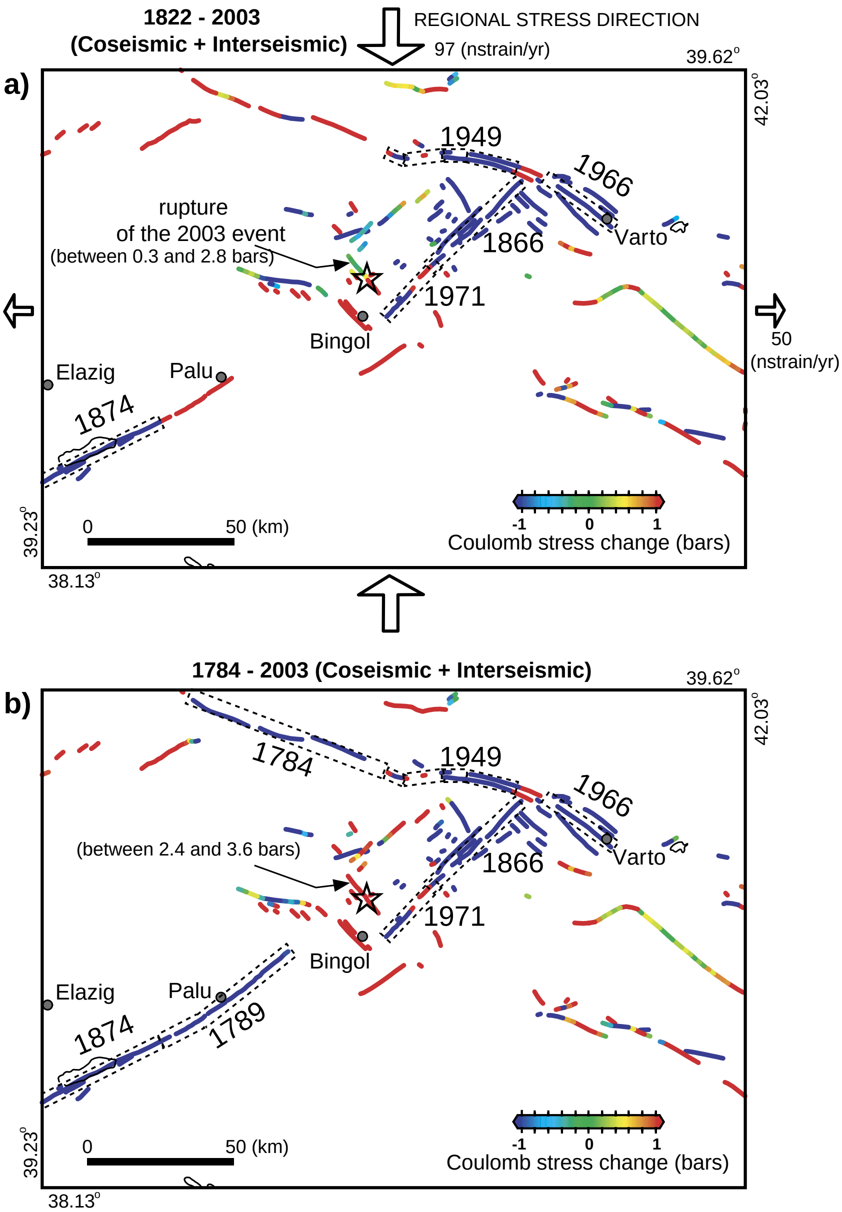

Fig. 5. The orientation of the far-field regional stress consistent with the principal strain direction and amplitudes obtained from GPS measurements [18,19]. Every fault in the diagram receives strain loading depending on its orientation. The end member stress values on the 2003 rupture plane from NW to SE are marked. Coulomb stress changes since 1822 (a), and since 1784 (b). Note that in both cases the stress change along the plane is positive in contrast to case shown in Figs. 2 and 3. Stress decreases due to the co-seismic effects are erased by the tectonic loading and bring the fault to failure state.

technique, the far field loading method, in which the entire region considered as a volume which is distorted in the directions of the principal strain axes at rates proportional to that obtained from regional GPS measurements. This loads all the faults in the study area based on their geometry and sense of slip. According to the Coulomb failure criteria, faults aligned near the critical angle to the principle loading axis will accumulate larger strain (and experiences larger strain rates) than those at unfavourable orientations. In earlier applications, this technique was used to find the geometry of the optimally oriented slip planes for failure [3,17]. Here, we employ the principal strain directions obtained from GPS data analysis [18,19]; the area is compressed in the direction of N–S and extended in E–W at average strain rates of 97 and 50 nstrain/yr respectively.

Note that InSAR modelling of deformation along the much simpler segment of the eastern North Anatolian Fault located 100 km NW of the Bingol area shows that the strain accumulates only along this single branch [20]. However dense GPS measurement results from California shows that deformation rates between the Pacific and North American plates are partioned among a number of parallel faults [21]. The total deformation rate in a complicated tectonic environment such as the Bingol area is also likely to be portioned along the small faults. To determine the actual deformation rate along these faults one has to measure the deformation from a GPS network in which the station separations are proportional to the fault lengths. In other words a denser GPS network would be necessary. Unfortunately, as is discussed in next section, the GPS network in our study area is sparse hence the strain rates calculated as above may slightly over estimate the loading on smaller faults (<30 km in length).

Fig. 5a shows the stress perturbation on the faults due to both co-seismic and tectonic loading since 1822. With this modified tectonic loading model the 2003 fault receives more pronounced strain/stress loading than from the deep dislocation method. The average tectonic load on the fault is approximately 0.0376 bars/yr. The stress change on the 2003 fault with this loading is positive and ranges from 0.3 bars at the NW tip to 2.8 bars at the SE. The stress at the 2003 hypocentre is approximately 0.26 bars at the time of the Bingol earthquake. During this time period, accumulation of secular stress loading is about 6.8 bars, but the fault receives negative stress changes due to the co-seismic slip which is about -6.54 bars. If we include the earlier historical events and start the tectonic loading in 1784, the stress change along the 2003 fault ranges from 2.4 at the NW tip and 3.6 bars at the SE. The stress change at the hypocentre is now about 1.75 bars, which consists of 8.23 bars tectonic loading during the 219 yrs and -6.48 bars co-seismic stress change. It thus appears that the stress accumulation on the 2003 rupture plane is due to tectonic loading.

3. Conclusions and discussion

In this paper we have explored four possible explanations for the discrepancy between the stress field modelled by EAFZ02 and the occurrence of the Bingol 2003 earthquake. We have shown that, at least in this case, no reasonable change to the previous earthquakes used to calculate the co-seismic component of loading, which is consistent with present understanding of the seismic history of the region, makes a significant difference to the stress accumulation on the Bingol Fault. In particular we have shown that the locations of large historical earthquakes occurring in the region during the 1780s are constrained by the known active structure to occur in locations which cannot significantly stress the 2003 fault. Secondly, while the occurrence of a pre-instrumental intermediate sized very local event might change the co-seismic load, any reasonable change to the locations or magnitudes of the events modelled in the 2002 study are unlikely to have had any significant effect. In summary, our findings show that the 2003 Bingol earthquake is unlikely to have been triggered by previous earthquakes at least of a scale which can be modelled using currently available data. Note that a similar study [13] claimed that the 2003 earthquake was loaded towards failure by co-seismic stress due to the earlier events in the area. Their work differs from ours in that they started their calculation from the year 1874 and hence neglected a nearby event (Ms = 7.2) that occurred in 1866 on the Bingol–Karliova segment of the EAFZ. Most of the stress shadow on the 2003 rupture plane is caused by the

1866 and 1971 earthquakes, therefore the lack of inclusion of this event strongly affects their results.

In general, this work has underlined the importance of accurate modelling of the co-seismic stress perturbation field in regional seismic hazard studies based on historical seismicity. It is clear from, for example, Fig. 3 that while incorrect locations of previous events had little effect on the modelled stress on the Bingol fault, this is not the case in many important areas. We therefore suggest that some independent support for the predictions of historical seismology be employed better to constrain the predicted stress perturbations. Paleoseismology, as it is currently used, aims at the reconstruction of the event history of a particular fault segment; the questions it addresses relate to the timing of previous earthquakes on a known fault. The information required for seismic hazard assessment, however, might be obtained by directed paleoseismology in which the geometry of earthquakes whose timing is accurately known from historical studies is confirmed by mapping their surface ruptures. These geometries can then be used accurately to calculate resulting stress perturbations. Such investigations, together with accurate, high resolution structural geology to identify the active faults on which the resulting stress perturbations might be resolved will radically improve our computation of the hazard distribution.

Finally, we have shown that the technique used in EAFZ02 to compute the secular loading on the Bingol fault was probably inappropriate given the complexity of the structure in the Bingol area and that a more reasonable estimate of the incremental secular stressing which is consistent with regional GPS measurements results in significant positive loading on the Bingol hypocentre. This, we believe, is an important conclusion and it calls for detailed GPS surveying of seismically active areas so that the tectonic loading on identified active faults might better be estimated. We note with interest the current deployment of such a GPS network (pers. comm. Haluk Özener) in precisely the area of this study. A cautionary note, however, might be useful.

It is now widely accepted that many features of the space, time and energy distributions observed in the earth's crust are fractal exhibiting power–law probability functions. An important feature of power–law functions is the absence of a characteristic scale. Thus fractal fracture sets do not have a characteristic fracture length or spacing and fractal earthquake catalogues do not have any characteristic recurrence time. In such populations, up-scaling, the extrapolation of large scale properties from small scale observations, is not possible; equally true though less commonly discussed, these distributions also preclude down-scaling, the interpolation of small scale properties from large scale observations. Thus networks with apertures of 100 km or so might be appropriate to define secular loading increments for magnitude 7 earthquakes but could fail correctly to model strain heterogeneity which loads small faults to produce magnitude 6 earthquakes such as the 2003 Bingol event. More work is clearly required to investigate the scaling of deformation. In particular, models which predict fractal scaling in space also frequently predict fractal scaling in time. While sufficient resolution over sufficiently long periods is not yet available in GPS data, strain meters have measured long time series which might also be expected to show fractal scaling. Photogrammetry of sandbox models and examination of numerical simulations might also be useful.

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References

[1] R.A. Harris, Introduction to special section: stress triggers, stress shadows, and implications for seismic hazard, J. Geophys. Res. 103 (1998) 24347–24358.

[2] G.C.P. King, M. Cocco, Fault interaction by elastic stress changes: new clues from earthquake sequences, Adv. Geophys. 44 (2001) 1–38.

[3] G.C.P. King, R.S. Stein, J. Lin, Static stress changes and the triggering of earthquakes, Bull. Seismol. Soc. Am. 84 (1994) 935–953.

[4] R.A. Harris, R.W. Simpson, P.A. Reasenberg, Influence of static stress changes on earthquake locations in southern California, Nature 375 (1995) 221–224.

[5] S.C. Jaumé, L.R. Sykes, Evolution of moderate seismicity in the San Francisco Bay region, 1850 to 1993: seismicity

changes related to the occurrence of large and great earthquakes, J. Geophys. Res. 101 (1996) 765–789.
[6] R.S. Stein, A.A. Barka, J.H. Dieterich, Progressive failure on the North Anatolian fault since 1939 by earthquake stress triggering, Geophys. J. Int. 128 (1997) 594–604.
[7] J. Deng, L.R. Sykes, Evolution of the stress field in southern California and triggering of moderate-size earthquakes: a 200-year perspective, J. Geophys. Res. 102 (1997) 9859–9886.
[8] S.S. Nalbant, A. Hubert, G.C.P. King, Stress coupling between earthquakes in northwest Turkey and the north Aegean Sea, J. Geophys. Res. 103 (1998) 24469–24486.
[9] C. Nostro, R.S. Stein, M. Cocco, M.E. Belardinelli, W. Marzocchi, Two-way coupling between Vesuvius eruption and southern Apennine earthquakes, Italy, by elastic stress transfer, J. Geophys. Res. 103 (1998) 24487–24504.
[10] S.S. Nalbant, J. McCloskey, S. Steacy, A.A. Barka, Stress accumulation and increased seismic risk in Eastern Turkey, Earth Planet. Sci. Lett. 195 (2002) 291–298.
[11] Ö. Emre, E. Herece, A. Doğan, O. Parlak, V. Özaksoy, R. Çıplak, S. Özalp, A report on the May 1, 2003 Bingol Earthquake (in Turkish), General Directorate of Mineral Research and Exploration, Report No. 10585, Ankara, 2003.
[12] S. Özalaybey, M. Ergin, 1 May 2003 (Mw = 6.4) Bingol earthquake aftershock monitoring, Preliminary report, 2003, http://www.ydbae.mam.gov.tr/english/index.html.
[13] C. Milkereit, et al., Implications of the 2003 Bingol earthquake for the interaction between the North and East Anatolian faults, Bull. Seismol. Soc. Am. 94 (2004) 2400–2406.
[14] N.N. Ambraseys, J.A. Jackson, Faulting associated with historical and recent earthquakes in the Eastern Mediterranean region, Geophys. J. Int. 133 (1998) 390–406.
[15] N.N. Ambraseys, Temporary seismic quiescence: SE Turkey, Geophys. J. Int. 96 (1989) 311–331.
[16] J.C. Savage, A dislocation model of strain accumulation and release at a subduction zone, J. Geophys. Res. 88 (1983) 4984–4996.
[17] R.S. Stein, G.C.P. King, J. Lin, Change in failure stress on the southern San Andreas fault system caused by the 1992 magnitude = 7.4 Landers earthquake, Science 258 (1992) 1328–1332.
[18] H.-G. Kahle, M. Cocard, Y. Peter, A. Geiger, R. Reilinger, A. Barka, G. Veis, GPS-derived strain field within the boundary zones of the Eurasian, African, and Arabian Plates, J. Geophys. Res. 105 (2000) 23353–23370.
[19] B. Oral, Global positioning system (GPS) measurements in Turkey (1988–2002): kinematics of the Africa–Arabia–Eurasia plate collision zone, Ph.D. thesis, Massachusetts Institute of Technology (1994) 344 p.
[20] T. Wright, B. Parson, E. Fielding, Measurement of interseismic strain accumulation across the North Anatolian Fault by satellite radar interferometry, Geophys. Res. Lett. 28 (2001) 2117–2120.
[21] S.J. Bourne, P.C. England, B. Parsons, The motion of crustal blocks driven by flow of the lower lithosphere and implications for slip rates of continental strike–slip faults, Nature 392 (1998) 655–659.