Broad-band $Lg$ attenuation modelling in the Middle East

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SUMMARY
We present a broad-band tomographic model of $Lg$ attenuation in the Middle East derived from source- and site-corrected amplitudes. Absolute amplitude measurements are made on hand-selected and carefully windowed seismograms for tens of stations and thousands of crustal earthquakes resulting in excellent coverage of the region. A conjugate gradient method is used to tomographically invert the amplitude data set of over 8000 paths over a $45° \times 40°$ region of the Middle East. We solve for $Q$ variation, as well as site and source terms, for a wide range of frequencies ranging from 0.5 to 10 Hz. We have modified the standard attenuation tomography technique to more explicitly define the earthquake source expression in terms of the seismic moment. This facilitates the use of the model to predict the expected amplitudes of new events, an important consideration for earthquake hazard or explosion monitoring applications. The attenuation results have a strong correlation to tectonics. Shields have low attenuation, whereas tectonic regions have high attenuation, with the highest attenuation at 1 Hz found in eastern Turkey. The results also compare favourably to other studies in the region made using $Lg$ propagation efficiency, $Lg/Pg$ amplitude ratios and two-station methods. We tomographically invert the amplitude measurements for each frequency independently. In doing so, it appears the frequency dependence of attenuation in all regions is not compatible with the power-law representation of $Q(f)$, an assumption that is often made.

Key words: Body waves; Seismic attenuation; Seismic tomography; Wave scattering and diffraction; Wave propagation; Asia.

1 INTRODUCTION
$Lg$ is a crustally guided shear wave and is a very significant phase at regional distances over continental paths. It generally propagates at velocities ranging from 3.6 to 3.1 km s$^{-1}$ and is recorded over a relatively wide frequency band ranging from 0.2 to 10 Hz. The amplitude of this phase, in relation to the amplitude of other regional phases, such as $Pn$ and $Pg$, is important in discriminating earthquakes from explosions (e.g. Kim et al. 1993; Walter et al. 1995; Hartse et al. 1997). The amplitude of $Lg$ has also been used to estimate the yield of nuclear tests (e.g. Nuttli 1986; Hansen et al. 1990). At local distances, the amplitude of shear waves are important for characterizing the earthquake hazard. For all these applications, characterizing the attenuation of $Lg$ over a wide frequency band is necessary to account for the observed large variations in amplitude.

In this paper, we present a broad-band $Lg$ attenuation model of the Middle East. The complicated tectonics of this region, as expressed by a topographic map showing tectonic provinces (Fig. 1), are well known and lead to large variations in factors that affect crustally propagating $Lg$, such as sediment thickness, crustal velocity and crustal thickness. Briefly, active continental collision between Arabia and Eurasia had lead to the formation of mountain ranges ( Zagros) and high plateaus ( Turkish Plateau, Iranian Plateau) on one side, whereas shields and platforms (Arabian Shield and Arabian Platform) have been relatively unperturbed on the other. Older collisions resulting from the closing of the Tethys Ocean are expressed in other orogenic zones ( Caucasus, Kopet Dagh) and remnants of oceanic crust ( Black Sea, South Caspian, Meditreanean) at the fore-region of the stable continent (Ukrainian Shield, Kazakh Platform).
Furthermore, in active tectonic regions, the aforementioned crustal properties (e.g. crustal thickness, etc.) are heterogeneous and can change significantly over very small lateral ranges. As a result, we expect large attenuation variations from region to region in this area. It is therefore advantageous to make amplitude measurements for a large number of paths, so as best to cover the region with redundant data and crossing paths.

This study benefits from improvements in attenuation tomography methodology as well as improved data coverage and processing. Our methodology uses carefully measured $Lg$ amplitudes and represents the amplitudes in several frequency bands in terms of source, path and site terms. We focused on making as many local to regional $Lg$ single station amplitude measurements as possible for the Middle East. Our goal is to maximize the number of crossing paths and avoid the severe restrictions that exist in two-station methods. We recognize that, in addition to attenuation, observed $Lg$ amplitudes are affected by the initial source amplitude, including corner frequency roll-off effects. In this study, we explicitly formulate the source term to account for this. Additionally, the effects of geometrical spreading are removed using standard corrections. We also recognize that $Lg$ is subject to blockage and attenuation to
amplitude levels below noise levels. We use both pre-event and pre-phase signal-to-noise thresholds to remove pure noise measurements from the inversion. In fact, mapping out strong attenuation and blockage regions in the Middle East provides useful information on the geophysical structure (e.g. Kadinsky-Cade et al. 1981; Rodgers et al. 1997a; Mellors et al. 1999; Gok et al. 2000; McNamara & Walter 2001; Baumgardt 2001; Sandvol et al. 2001). Finally, the source and geometrical spreading corrected amplitude measurements are then tomographically inverted for $Q(f)$ later in the paper. Yellow lines show the paths of waveforms shown in Fig. 2.

2 METHODOLOGY

In addition to along-path attenuation, the amplitude of $Lg$ waves is controlled by the initial source, the geometrical spreading with distance and site effects due to structure under the recording station. For a given frequency, the observed amplitude $A$ from event $i$ recorded at station $j$ is a product of a source term $S_i$, a geometrical spreading term $G_{ij}$, an attenuation term $B_{ij}$ and a site term $P_j$

$$A_{ij} = S_i G_{ij} B_{ij} P_j$$

(1)

Traditional amplitude tomographic techniques (e.g. Phillips & Stead 2008) then solve for each of the source, site and attenuation terms with trade-offs between them. However, this results in significant source terms for each event. In addition, for explosion monitoring or earthquake hazard purposes, we want to be able to apply the tomographically derived apparent attenuation to get a predicted amplitude, perhaps to use as a correction before forming a discriminant. In these cases, it would be unclear what source term is associated with the new event. Here we take a physics-based approach by explicitly defining the earthquake source in terms of the seismic moment. We begin by describing in more detail all of the terms of eq. (1), and how we will manage them in the inversion.

2.1 Geometrical spreading term

The geometrical spreading term $G_{ij}$ is that from Street et al. (1975) and is given as a function of epicentral distance $R$

$$G(R) = \begin{cases} \frac{1}{R} & \text{if } R < R_0 \\ \left(\frac{1}{R_0}\right)^n \frac{R}{R_0} & \text{if } R \geq R_0 \end{cases}$$

(2)

where for $Lg$, $n = 0.5$, and the critical distance $R_0$ is 100 km. This formula transitions from spherical spreading within the critical distance to decaying as a cylindrically spreading wave at larger distances.

2.2 Site term

The site term $P_j$ is simply a multiplicative factor that controls the near-station amplification due to local Earth structure.

2.3 Source term

The source term $S_i$ is formulated in terms of an earthquake model tied to a measurable parameter such as the seismic moment $M_0$ or moment magnitude $M_w$. Here we use the Magnitude Distance Amplitude Correction (MDAC) source model (Walter & Taylor 2001), where the earthquake source term can be expressed as a modified single corner frequency ($\omega_c$), with $\omega \rightarrow \omega_c$ high frequency fall-off (Aki 1967; Brune 1970, 1971):

$$S = \frac{FM_0}{\left(1 + (\omega/\omega_c)^2\right)^{1/3}}$$

(3)

where the corner frequency $\omega_c$ is specified as

$$\omega_c = \left(\frac{K\sigma}{M_0}\right)^{1/3}$$

(4)

$F$ and $K$ are constants that depend on medium properties, and the $FM_0$ product is the zero-frequency spectral level. $\sigma$ is the apparent stress (e.g. Wyss 1970) in the region. The source term in the tomography can be reformulated to solve for the change in moment from the assumed initial moment, where the initial value comes from a calibrated technique. In this case, to predict an $Lg$ amplitude for a new event, the moment is first determined and then the 2-D $Lg$ attenuation model and site effects are applied to get the predicted amplitude for an earthquake of that $M_0$ and location. This is very similar to the way 1-D path corrections are commonly used to correct $Lg$ amplitude measurements for event discrimination (e.g. Walter & Taylor 2001; Taylor et al. 2002) in the MDAC methodology.

The $F$ constant is a function of the average $S$-wave radiation pattern $R_{ij}^{sp}$ and the source and receiver velocities and densities and is given (Walter & Taylor 2001) by

$$F = \frac{R_{ij}^{sp}}{4\pi \sqrt{\rho_s \rho_r \mu_s \mu_r}}$$

(5)
The constant $K$ that governs the corner frequency $\omega_c$ used in eq. (4) is determined by the radiated energy (Walter & Taylor 2001) and is given by

$$K = \frac{16\pi}{\beta_s^2 \left( R_{sv}^p \omega_c^3 / \alpha_c^2 \right) + \left( R_{sv}^s / \beta_s^2 \right)},$$

where $\zeta$ is the scale factor between the $P$-wave and $S$-wave corner frequencies. The parameter values we used for the average $S$ and $P$-wave radiation patterns are: $R_{sv}^p = 0.6$ and $R_{sv}^s = 0.44$ (Boore & Boatwright 1984). For the average source and receiver densities and velocities we used: $\rho_s = 2700$ kg/m$^3$, $\rho_t = 2500$ kg/m$^3$, $\beta_s = 3500$ m/s, $\beta_t = 2900$ m/s and $\alpha_s = 6000$ m/s. Here, we use $\zeta = 1$ indicating the source $P$ and $S$ corner frequencies are the same.

We generalize the apparent stress in eq. (4) to the following equation

$$\sigma = \sigma' (M_o / M_o')^{\psi}.$$  

(7)

For constant apparent stress, we would set $\psi$ to 0 and $\sigma'$ to the constant stress level. We have allowed the apparent stress to increase with increasing moment by specifying $M_o' = 1.0 \times 10^{15}$ N m at $\sigma' = 0.3$ MPa, and use $\psi = 0.25$ here to match the observations of Mayeda & Walter (1996). Later in the paper, source term assumptions, such as the choice of $\psi$ and the constancy of apparent stress, and its affect on attenuation results will be discussed in more detail.

Taylor et al. (2003) have previously incorporated aspects of MDAC methodology in $Lg$ tomography in Central Asia. Several important differences exist between our methodology and that employed in their study. For example, when possible we limit ourselves to events that have seismic moment estimates rather than deriving them directly from the data. In the present study, we are treating $Lg$ waves not sensitive to the high-frequency attenuation structure. As will be seen later, because of the large number of paths, there is a significant resolution improvement from the $Lg$-wave maps, but rather allows the source term to change with a preponderance of data. Second, it allows us to incorporate independent information into the problem, which can break the trade-off.

As in earthquake location and Earth velocity, even with the trade-offs between the two, we can do the same with the attenuation problem. In earthquake location, one can break the trade-off by incorporating independent information on the parameters into the problem. In location, we might use source locations derived from a local network, or a velocity model consistent with surface wave data. For the attenuation problem, we can incorporate seismic moments derived from long-period surface waves not sensitive to the high-frequency attenuation structure. Since the source term is explicitly defined in terms of moment, it is very easy to incorporate this independent information.

Further tests that we have performed show that there actually is not much trade-off between the source terms and $Q$, but there is some between the source terms and site terms. When we arbitrarily doubled all of the moments in the source term, the site terms dropped (by approximately $1/2$) to compensate. In an additional test, we randomized the starting magnitudes by perturbing the moment magnitude with a Gaussian distribution having a standard deviation of 0.15 magnitude units. In the case of both systematic and random perturbations to the initial source term, the attenuation maps were essentially unchanged in both variation and absolute magnitude. These tests indicate that, whereas there might be trade-offs between the site and source terms, by putting independent constraints on one of the terms, then we can resolve the other. This is even more so with values of attenuation.

2.4 Attenuation term

Lastly, the attenuation term $B_{ij}$ is expressed in terms of $Q$ and is given by

$$B_{ij} = \exp \left[ -\omega R_{ij} / (2 Q_{ij} v) \right].$$  

(8)

where $R$ is the event-station epicentral distance, $\omega$ is the angular frequency $(2 \pi f)$ and $v$ is the $Lg$ velocity. Based on observed travel times, we use a velocity of 3.4 km s$^{-1}$ for $Lg$. The $Q$ here refers to the quality factor for the total attenuation, including both intrinsic and scattering attenuation. There is some evidence that intrinsic and scattering $Q$ are about the same order of magnitude for $Lg$ waves (Lacombe et al. 2003).

2.5 Inversion

Taking the logarithm of eq. (1), combining it with eq. (8) and correcting for geometrical spreading results in

$$\log A_{ij} - \log G_{ij} = \log S_i + \log P_j - ((\omega \log v) / (2 Q_{ij} v)) R_{ij}. $$

(9)

On the left-hand side the raw amplitudes are corrected for geometrical spreading. After setting initial values on the right-hand side, we solve for an updated source term ($S$), a site effect term ($P$) and the attenuation term. It is then easy to back out the values for $M_o$ and the terms for the lateral attenuation quality factor $Q$.

In a recent study, Drouet et al. (2008) simultaneously inverted for source spectra, attenuation parameters and site responses in southern France using data from the accelerometric network, which contains a mixture of hard and soft rock sites. Although there are many similarities in the two methods, there are differences in the geometrical spreading (which does not vary with distance) and, most significantly, the $Q(f)$ is assumed to have a power-law frequency dependence, an assumption that we will be testing in our study. Furthermore, they analyse two regions where $Q$ and its frequency dependence are solved for, but do not vary within each region. Lastly, since our study focuses on weak motion from hard rock sites, we expect smaller site corrections.

A well-known problem in all tomographic inversion is the trade-off between various components of the problem. Menke et al. (2006) comment on the non-uniqueness of the coupled origin time–velocity tomography problem and suggest an analogous non-uniqueness between the source term and attenuation in amplitude tomography. The coupling of components is, in fact, one of the reasons why we have formulated the problem as we did. First, our methodology does not simply map errors of the source terms into bias in the attenuation maps, but rather allows the source term to change with a preponderance of data. Second, it allows us to incorporate independent information into the problem, which can break the trade-off.

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Further tests that we have performed show that there actually is not much trade-off between the source terms and $Q$, but there is some between the source terms and site terms. When we arbitrarily doubled all of the moments in the source term, the site terms dropped (by approximately $1/2$) to compensate. In an additional test, we randomized the starting magnitudes by perturbing the moment magnitude with a Gaussian distribution having a standard deviation of 0.15 magnitude units. In the case of both systematic and random perturbations to the initial source term, the attenuation maps were essentially unchanged in both variation and absolute magnitude. These tests indicate that, whereas there might be trade-offs between the site and source terms, by putting independent constraints on one of the terms, then we can resolve the other. This is even more so with values of attenuation.

3 DATA AND TOMOGRAPHY

We have made measurements at dozens of stations in the Middle East including GSN stations (i.e. AAK, ABKT, GAR, GNI, KBL, KIV, NIL, RAYN), PASSCAL deployments (Eastern Turkey Seismic Experiment, Saudi Arabia), GEONET stations (BGIO, CSS, EIL, ISP, MALTA), MEDNET (KEG), GEOSCOPE (ATD, HYB) and stations from local networks (i.e. Saudi Arabia, Jordan, Israel, Kazakhstan, United Arab Emirates, Azerbaijan, IIIES in Iran, KOERI in Turkey). Crustal seismicity levels in the region are...
high in particular areas. The majority of the events that we examine come from the very active Zagros Belt, although many events also originate along the Cyprian Arc, Red Sea and Gulf of Aden rifts, throughout Anatolia, along the Alborz and Kopet Dagh in Iran and in the Hindu Kush. We even have a large number of events in the mostly aseismic region of India, which are aftershocks of the Bhuj earthquake. In all cases, we limit the earthquakes to be within the crust (generally with depths less than 30 km).

In Fig. 2 we show examples of waveforms in the frequency bands we examined for two different events. The top plot is a central Iran event recorded at a station in southeast Iran. The top trace shows the unfiltered waveform, which shows the Lg phase coming in slightly faster than 3.4 km s\(^{-1}\). The rest of the traces show the same waveform where the signal has been filtered within the frequency band listed to the right. The Lg phase records very strongly along this path between 0.5 and 4 Hz, but starts to fall-off between 4 and 6 Hz and does not pass the pre-phase signal-to-noise test (explained later in this section) above 6 Hz. The bottom plot shows an event in the Gulf of Aqaba recorded at a station on the Arabian Peninsula. Here the Lg phase records well across all frequencies and dwarfs the other regional phases.

In our study, an analyst has reviewed every seismogram. If identified, the Lg arrival is picked, otherwise a theoretical arrival of 3.45 km s\(^{-1}\) is used. The phase window starts at the pick and continues to a group velocity of 2.8 km s\(^{-1}\), with a minimum window length of 1 s. Amplitude measurements are then made in this window.

We measured Lg amplitudes for thousands of paths. For example, at 1 Hz, we have a total of 5889 paths at 53 stations from 2954 events. We tested the effect on the attenuation model of removing the paths from events recorded only at a single station. We lose significant coverage near the edge of the region, but find that the spatial variation of Q does not change over most of the coverage area. Therefore, to maximize coverage, we have included these events in our inversion.

At other frequencies, the number of paths remains high from 0.5 to 2.0 Hz, but falls to smaller numbers at higher frequencies (Fig. 3). At 6.0 Hz, we still have over 1000 paths, although the path coverage is not as even. Fig. 4 shows a path map for 1 Hz showing coverage of our study area. We have excellent coverage in the central portion of the region including eastern Turkey, Iran and the northern Arabian Peninsula. We have poorer coverage towards the periphery of our model, especially to the north and southeast where station coverage is sparse. We have no coverage of the broad oceanic region of the Arabian Sea where Lg does not propagate (e.g. McNamara & Walter 2001).

The input data consists of amplitude measurements for each path as well as moment estimates for each event. The amplitudes are time-domain rms amplitudes converted to pseudo-spectral amplitudes (e.g. Taylor et al. 2002) to be compatible with the frequency domain method we are using. We only select measurements where the signal-to-noise ratio (as determined from pre-event noise) exceeds 2.0 and the signal-to-noise ratio (as determined from pre-phase noise) exceeds 1.0. This is to ensure that all Lg amplitude measurements are actually from the Lg phase and not from Sn coda. The signal-to-noise criteria, along with the lower-frequency Nyquist for some stations with lower sampling rates, are the main reasons why we have fewer paths at higher frequencies.

Although doing our best to ensure that all Lg measurements are actually Lg amplitudes and not noise or Sn coda, it also has the additional benefit of removing blocked paths. Blockage regions, which do not propagate Lg for geometric reasons, are not treated as simply equivalent to high attenuation. Our resulting maps represent the attenuation that can be expected to affect Lg amplitudes along non-blocked paths. Blockage can be introduced later to eliminate paths that cross these regions (in our study area, the Red Sea, South Caspian Sea, Black Sea and Arabian Sea).

There is some concern that, by discarding (non-blocked) paths with high attenuation that have low signal-to-noise, the remaining paths will be biased to high Q. Whereas this is a legitimate concern, not using signal-to-noise criteria creates greater problems of including (and inverting) amplitudes of noise and Sn coda. A more sophisticated treatment might involve estimating an expected signal amplitude based on event size and path attenuation, coupled with a noise model. This would, of course, require an attenuation map. Another, or additional, approach would involve an inversion using inequalities. Both of these are sophisticated analyses that are beyond the scope of this paper, but data censoring will be the topic of a separate study by this research team.

Amplitude measurements are made in specific bands, notably 0.5–1, 1–2, 2–4, 4–6, 6–8, 8–10 and 10–12 Hz. In general, we will refer to each band by its low-frequency corner. The angular frequency used in eqs (3) and (8) fall within the band range. Since the low-frequency end of the band generally has higher amplitude than the high-frequency end, the band mid-point frequency can systematically underestimate the log averaged amplitude in the band (Rodgers et al. 1997b). We use a frequency that is one-fourth of the way up the scale, so for the 1–2 Hz band, a frequency of 1.25 Hz is used.

Values of seismic moment have been estimated either using moment tensor catalogs, such as the Global CMT (http://www.globalcmt.org/), regional waveform modelling or, in most cases, coda-derived estimates of this value (Mayeda et al. 2003). Ideally, all of our events would have such independent moment estimates. In about half of the cases, however, moments do not exist for events of interest. Therefore, where no moments are available, we have estimated a seismic moment from other magnitudes (\(M_0\), \(M_1\), etc.) using a regression relationship derived between the other moment and moments for those events where we have independent moments. While we fit our dataset empirically, the results compare favourably to the results of a study on the relationship between magnitude scales by Utsu (2002) over the magnitude range that we use. The catalog magnitudes are obtained from a variety of sources (e.g. PDE, ISC, regional networks) and vary somewhat in methods and estimated magnitudes.

Referring back to eq. (9), we correct the amplitudes for geometrical spreading and solve for lateral \(Q\), the source terms \(S_i\), and the site terms \(P_j\). The region is parameterized into a grid of equal area cells. At frequencies of 0.5–4 Hz, we use 0.5° cells, whereas for higher frequencies we have fewer paths and use 0.75° cells. A single attenuation value is used as the starting model, which follows a power-law model

\[
Q(f) = Q_0 f^{\eta},
\]

where \(Q_0 = 275\) and \(\eta = 0.725\). During the inversion each frequency is solved for independently and, whereas the model quickly moves away from the starting model where there are data, the starting model persists in regions without any paths. The source terms are then converted to seismic moment or moment magnitude through the source term equations. We also solve for site terms at every station for each frequency band, which is representative of the local station site effects. A Laplacian function is used to smooth the variation in \(Q\) between adjacent cells and to regularize the tomography. A conjugate gradient method is used as the solver.
Figure 2. (a) An event in central Iran recorded at station ZHSF in southeast Iran. The top trace shows the unfiltered waveform over the same trace plotted in a filter comb from 0.5 to 10 Hz. (b) An event in the Gulf of Aqaba recorded at station AFIF in central Arabia.
For many reasons (e.g. focusing/defocusing, multi-pathing, blockage, source complications, etc.), seismic amplitude residual data is highly variable, much more so than seismic velocities. After the inversion, the rms error is reduced from a misfit of 0.60 log-units (\(\sim 4.0\)) with the starting model (single \(Q\), no site effects, initial source terms) to about 0.15 log-units (\(\sim 1.4\)) (Fig. 5). This is a variance reduction of about 94 per cent. Note that it is possible to fit the data to an even higher level by eliminating the smoothing and/or putting more of the amplitude variations into site and event corrections. This, however, is simply a function of overfitting the data with significantly more degrees of freedom.

4 RESULTS

The attenuation of 1 Hz \(Lg\) is shown by the map in Fig. 6. In each of the attenuation plots, warm colours represent low \(Q\) (high attenuation) and cool colours represent high \(Q\) (low attenuation). Note that \(Q\) is plotted on a logarithmic scale with values annotated every 0.25 in \(\log_{10}\) space.

The results show a significant variation in the \(Q\) values. Attenuation is very high over most of the Turkish–Iranian Plateau (\(Q < 200\), high along the Zagros Mts. and portions of the Red Sea and Gulf of Aden rifts (\(Q = 200–300\)), moderate in the northern Arabian Plate (\(Q = 300–400\)), low in the Arabian Platform and Kazakh Platform (\(Q = 400–600\)), and very low in the Arabian Shield and Indian Shield (\(Q > 800\)). The lowest \(Q\) is found in eastern Turkey (\(Q = 150\)) and the highest is found in the Indian Shield (\(Q = 2000\)). Although the results are very complicated, there appear to be significant correlations between \(Q\) values and age of most-recent tectonic activity, with areas of recent activity having low \(Q\) and areas of older thermo-tectonic events having high \(Q\).

One correlation (comparing Figs 1 and 6, for example) is the relationship between topography and crustal attenuation. Large topography variations, both high topography in mountain ranges and low topography in extended regions, are a surrogate for crustal deformation. It appears in our maps that tectonically-affected regions with significant crustal deformation, as indicated by topography variations, have lower \(Q\) (higher attenuation) than undeformed regions. No similar correlation exists between \(Q\) and sediment thickness. Another observation is that features that we would expect to produce large effects in the mantle (e.g. uplift of the Turkish and Iranian Plateaus) seem to also have large effects on \(Lg\) attenuation. This could be an effect of lithospheric thickness and heat flow on the crustal attenuation. Studies of \(Lg\) propagation have sought deterministic relationships between \(Lg\) amplitudes or amplitude ratios and path-specific crustal waveguide properties (e.g. Zhang & Lay...
Figure 5. Histograms of log10 amplitude residuals at 1.0 Hz for the initial model (light grey) and after the inversion (dark grey). The y-axis shows the frequency-percentage in percent.

Figure 6. Attenuation map of $Lg$ $Q$ at 1 Hz in the Middle East. $Q$ values range from under 200 to over 1000. Note that the $Q$ scale is logarithmic. The map fades to white in regions with lower resolution.

Figure 7. Plot showing site amplification terms solved for at 1 Hz. Values range from 0.15 to 3.4. A value of 1.0 indicates normal site term values.

expected if the site terms reflect local effects. In fact, this is probably a very local effect like $Vs30$ (the average shear-wave velocity in the upper 30 m) and can vary over extremely short distances. We also performed a test where one station had the wrong gain, and found that it was directly absorbed into the site factor and had no impact on the attenuation maps.

The inversion also solves for source terms (which can be characterized as $M_w$ corrections) for each event. In general, changes to the event terms (shown in Fig. 8) are relatively small. Except for a few rare cases (which are probably poor magnitude estimates), the largest $M_w$ differences are 0.3–0.4 magnitude units and the overall average (rms) difference is only $\sim 0.2$ magnitude units. When we break it down by magnitude type, however, we find that the rms differences between the original and new $M_w$ is $\sim 0.23$ m.u. for converted magnitudes and only 0.17 m.u. for true moments. This supports our supposition that, when available, independent moments rather than magnitude regression derived moments should be used. It also demonstrates that the method is able to compensate for bad initial $M_w$s, rather than biasing the spatial attenuation. Small $M_w$ corrections are found across the frequency band except at 8.0 Hz where there are fewer paths and more events with amplitudes recorded at only a single station.

We compare our results to several other 1 Hz $Lg$ $Q$ models of the Middle East, those of Sandvol et al. (2001), Al-Damegh et al. (2004) and Zor et al. (2007), and one for the Indian Platform (Mitra et al. 2006). We have not compared our results to $Lg$ coda $Q$ maps of the region because they are based on the amplitudes from a different part of the waveform than the direct wave, have much longer scale lengths, and are more sensitive to the scattering component of $Q$.

Sandvol et al. (2001) estimated $Lg$/Pg relative attenuation by applying tomography to 4400 $Lg$/Pg amplitude ratios. Al-Damegh et al. (2004) added more data (for a total of 6200 paths) and improved the $Lg$/Pg efficiency maps. Because it is a relative attenuation
method, these studies do not obtain values of Lg Q. However, they find efficient Lg propagation within the Arabian Plate and some parts of the Red Sea and inefficient or blocked propagation in the Zagros and Turkish–Iranian Plateau. In general, this is consistent with the results that we find here.

Zor et al. (2007) used a two-station method to measure interstation Lg Q along 296 unblocked paths and 154 blocked paths. They then invert the measurements to obtain η, Q, and η models of a much smaller than the region we consider in this study. The results are more or less similar to the qualitative methods discussed above. They find low to normal Q values (250–350) in the northern Arabian Platform, high Q values (670–800) in the southern Arabian Plate, very low Q values in the Turkish Plateau (100–200). The lowest Q values are found in the East Anatolian Plateau (70–100). The authors also find a negative correlation between Q, η and η. In general, the Q values that they obtain are quite similar to those we find here.

Mitra et al. (2006) used the decay of spectral amplitudes with distance from three events to determine average Lg attenuation in the Indian Platform. Although their average values for the region Q = 655(±10) f 0.67(±0.03) are lower than the values that we found for the Indian Shield (Q > 800), they are more consistent with the higher values that they found for paths emanating from the shield—Q = 869(±45) f 0.64(±0.11).

Except as noted (e.g. grid size) we follow the same procedures for all other frequencies in our analysis. Results are shown in Fig. 9 and path density maps are provided in Supplementary Fig. 1. Ranges of Q values are recorded in Table 1. Some maps of Lg attenuation for different frequencies are quite similar, particularly those going from 0.5 to 1.0 to 2.0 Hz. In these three plots, the relative variations are generally similar, but the absolute Q values increases with frequency. This pattern seems to break down at higher frequencies, although there are several features in the 4.0 Hz map (eastern Turkey, Red Sea, central Iran) that seem persistent. Maps at higher frequencies are quite dissimilar, likely because the number of paths has fallen off rather significantly (Fig. 3).

With the set of inversions for several frequency bands, it is easy to produce frequency-dependent Q plots for any particular region. In Fig. 10, we assemble and compare of Q(f) for four well-covered regions within our study area: the Central Zagros, the Northern Arabian Platform, Indian Shield and Eastern Turkey. Locations are indicated by the white crosses in Fig. 1. For each point a 1° cell around the point was selected and averaged to provide mean Q with bars of standard deviation for each frequency. First, it appears that Eastern Turkey has the highest attenuation over the whole band. The Central Zagros is also a high attenuation region at low frequency, but has high Q at high frequencies. The Northern Arabian Platform has low attenuation at the low frequency end, but has moderate attenuation at high frequencies, whereas the Indian Shield has low attenuation across all frequencies. We also compare our results to other regions (Fig. 10) in the continental United States from Erickson et al. (2004). The variation in the Middle East is very large, with the highest Qs (from Arabia and India) being higher than the central United States and the lowest Qs (from Turkey and the Zagros) being lower than the Basin and Range.

In all cases, however, we find that Q(f) is not well represented by the power law (as represented by our starting model in eq. (10)), which is often assumed in many studies (e.g. Mitchell et al. 1997; Erickson et al. 2004; Mitra et al. 2006; Zor et al. 2007). We note that this divergence from the power law does not occur only at high-frequencies, where we have less confidence in the tomographic maps, but also at the lower-frequency end, where the inversion results are well-resolved. Whereas we might expect intrinsic Q to be frequency-dependent in a power-law sense, there is no reason for scattering Q (which depends on the correlation structure of scatterers) to be. For example, if there were a large number of scatterers at 1 Hz in the Indian Shield, then we would expect apparent Q to deviate significantly (to lower Q) from the power-law around this frequency.

One source parameter in the inversion that is not independently well constrained is the apparent stress. We experimented with both a constant apparent stress and an apparent stress that increases with moment as some recent studies have found (e.g. Mayeda & Walter 1996, Mayeda et al. 2007). We ran the models both for the case of ψ = 0 and ψ = 0.25. The change in the attenuation maps between the two cases was negligible (rms difference in log Q of 0.04), indicating that the attenuation maps are robustly determined for a range of source parameters. The source correction terms changed more significantly between the two cases. For the constant apparent stress case the source corrections at 1 Hz showed a less linear and stronger variation with moment than the case of ψ = 0.25, particularly for the largest events where there was large misfit between the original moment estimates (generally events having CMT solutions) and the new estimates. For this reason we prefer the ψ = 0.25, which is what is shown in Fig. 8.

5 CONCLUSIONS

We have developed a broad-band Lg attenuation model of the Middle East. We make single station absolute amplitude measurements and use a model-based formalism, which provides better path coverage and higher resolution than relative two-station measures. The amplitude measurements are then corrected for magnitude and distance effects to isolate the path attenuation. Tomographic inversion of the
Figure 9. Attenuation maps of $L_g Q$ at frequencies of 0.5, 1, 2, 4, 6 and 8 Hz. Note that the $Q$-scale varies from figure to figure to capture the variations in this parameter. The maps fade to white in regions with lower resolution.
The estimated region-specific $Lg\,Q$ model developed in this study can be used to constrain ground motion predictions for seismic hazard assessment. The wide range of inferred $Q$ values would indicate that strong ground motions for a given event would attenuate much more slowly in regions of high $Q$ and should be factored into probabilistic seismic hazard assessment in the Middle East.

Besides improving our attenuation models with more data, particularly at the highest frequencies, in the future we would like to investigate the possibility of inverting our frequency-dependent $Q(f)$ models for depth-dependent $Q(z)$ models. This will require more measurements and better tomographic maps at the highest frequencies, where the path coverage is poorest. This is complicated by several factors. The first is the fact that some of the stations considered have lower sampling rates. The second factor is that, even with higher $Q$, amplitude fall-off is more pronounced at higher frequencies due to the larger number of cycles. This requires larger events and shorter paths to make quality measurements.

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SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article:

Figure S1. Path density map at frequencies of 0.5, 1, 2, 4, 6 and 8 Hz.

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