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The shallow $P$-velocity structure of the southern Dead Sea basin derived from near-vertical incidence reflection seismic data in project DESIRE

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SUMMARY

As a part of the DEad Sea Integrated REsearch (DESIRE) project a near-vertical incidence reflection (NVR) experiment with a profile length of 122 km was completed in spring 2006. The profile crossed the southern Dead Sea basin (DSB), a pull-apart basin due to the strike-slip motion along the Dead Sea Transform (DST). The DST with a total displacement of 107 km since about 18 Ma is part of a left-lateral fault system which connects the spreading centre in the Red Sea with the Taurus collision zone in Turkey over a distance of about 1100 km. The seismic experiment comprises 972 source locations and 1045 receiver locations. Each source was recorded by $\sim$180 active receivers and a field data set with 175 000 traces was created. From this data set, 124 444 $P$-wave first-break traveltimes have been picked. With these traveltimes a tomographic inversion was carried out, resulting in a 2-D $P$-wave velocity model with a rms error of 20.9 ms. This model is dominated by a low-velocity region associated with the DSB. Within the DSB, the model shows clearly the position of the Lisan salt diapir, identified by a high-velocity zone. A further feature is an unexpected laterally low-velocity zone with $P$-velocities of 3 km s$^{-1}$ embedded in regions with 4 km s$^{-1}$ in the shallow part on the west side of the DSB. Another observation is an anticlinal structure west of the DSB interpretated to the related Syrian arc fold belt.

Key words: Tomography; Controlled source seismology; Transform faults.

INTRODUCTION

In the area west and south of the Dead Sea several regional seismic studies has been carried out to describe the deep velocity distribution (Ginzburg et al. 1979a,b; El-Isa et al. 1987a,b; Ginzburg & Ben-Avraham 1997; ten Brink et al. 2006; Mechie et al. 2009) with one exception of Ryberg et al. 2007 focussing to the shallower structure. The main goal of the field experiment was to investigate the crustal structures across the Dead Sea Transform (DST) in the upper few kilometres with seismic reflection data. The complete crustal cross-section and interpretation can be found in Weber et al. 2011. With the existing reflection seismic data set only the first arrivals of the $P$-wave were used to derive a velocity model using traveltime tomography with the aim to resolve shallow small-scale velocity anomalies which are not resolvable in wide-angle refraction models.

The DST marks the plate boundary between the Arabian Plate and the African Plate. It trends generally NNE and extends from the spreading centre in the Red Sea to the subduction zone at the Taurus mountains in Turkey (Fig. 1a, inset) and consists of a complex fault system with well-developed deep basins. The Dead Sea basin (DSB), with a length of about 130 km, a width of 7–18 km and a depth of $\sim$10 km (ten Brink et al. 1993; Garfunkel 1997; Hall 1997) is the largest pull-apart basin in this fault system and one of the deepest pull-apart basins in the world.

The crust east and west of the DSB is 30–35 km thick (Garfunkel & Ben-Avraham 1996; Garfunkel 1997; Mechie et al. 2009) and is characterised by a surface elevation difference of $\sim$300 m between the top of the western block (650 m) and the top of the eastern block (950 m). Both the west and the east blocks are gently tilted away from the DSB. In the area of the Dead Sea Integrated Research (DESIRE) near-vertical incidence reflection (NVR) experiment (Fig. 1a) a maximum altitude difference of 1350 m between the DSB and the mountain tops of the eastern block is observed.

GEOLOGICAL SETTING

At the borders of the DSB geological formations of the Upper Cretaceous (West) and the Precambrian through Lower Cretaceous (East) are exposed (Fig. 1b). On both sides of the DSB, the age of the rocks decreases with increasing distance to the basin.

In the region of the NVR experiment, the information from five boreholes has been used to constrain the possible geological
structure at depth along the NVR profile (Fig. 2). The locations of four boreholes are shown in Fig. 1(b). The fifth borehole (Swaqua 1) is outside the area covered by this map. In the Hazerim 1 borehole (Gvirtzman 2003), which is located about 10 km west of the westernmost vibroseis source point, thin Tertiary units and Upper and Lower Cretaceous units which, in turn, overlie Jurassic units. In the Zohar 8 Deep borehole (Gilboa et al. 1993), which lies about 1 km south of the processing line at profile km 41, the depths of these units are all shallower. Linear interpolation between these two boreholes yields a westward dip of approximately 0.8 to 2.6° for the various geological units. From the Zohar 8 Deep borehole to the Lot 1 borehole (Kashai & Croker 1986), the geological units dip downward with an angle between 0.9 to 3.1°. The basin is mainly filled with Neogene sediments and here in the El Lisan 1 borehole, the top of the Lisan salt diapir has been penetrated at shallow depth (Kashai & Croker 1986). A few kilometres east of the DSB, exposures of Lower Cretaceous rocks overlying units of the Cambrian can be observed in the wadi ‘Ibn Hammad’ a few kilometres south of the profile. At greater depths, Lower Cretaceous units overlying units of the Silurian in the borehole Swaqua 1, which lies approximately 100 km east of the easternmost shot point. Between the wadi exposure and the Swaqua 1 borehole, an eastward downdip of 0.3° of this boundary was estimated. The decreasing age of the geological units from Cambrian (at profile km 80) across the Ordovician (at profile km 87, Förster et al. 2006) to the Silurian leads to the assumption of an eastward dip of these units. At larger depths, Precambrian units with an eastward downdip of 0.8° underlie the units of the Cambrian. This boundary is the result of a reconstruction of the tectonic processes in this region using additionally the DESERT P-velocity model (Ryberg et al. 2007; Weber et al. 2011).

DATA

Along a seismic line were 972 seismic source points distributed, of which 864 of them were vibrator source points, located west of the Dead Sea and 108 were explosive source points located east of the Dead Sea. The vibrator source points spaced ~50 m apart consisted
Figure 2. Lithostratigraphic section along the profile based on information from five boreholes and an outcrop in Wadi ‘Ibn Hammad’. The solid red line indicates the topography of the NVR profile and yellow highlighted numbers indicate the dip in degrees of the linearly interpolated contacts between geologic units. The depth and dip of the geologic units east of the DSB are estimated by the correlation of the DESERT profile with the DESIRE profile (Weber et al. 2011).
of 4–7 vibrators with 276 kN peak force each using a linear 24-s-long up-sweep of 10–50 Hz. Each vibrator point was repeated 7–16 times and the resulting data were vertically stacked to increase the signal/noise ratio. Using a roll-along acquisition technique, each spread with a length of 18 km consisted of 90 sources. In contrast to the vibrator sources west of the DSB explosive sources were used east of the DSB. These source points were charged with 20–50 kg and were detonated in boreholes at 12–30 m depth. Here the sources were located ∼500 m apart and each spread consisted of nine sources. On both sides of the DSB each source was recorded by a spread of 180 three-component geophones with a spacing of 100 m, a natural frequency of 4.5 Hz and a sample interval of 5 ms. With a daily spread movement of 4.5 km, 20 days were needed to finish this experiment.

The different source types require a marginally different data processing and lead to seismograms with different characteristics. In contrast to the standard data processing for traveltime tomography (bandpass filter, linear moveout correction, etc.) the vibroseis data had been correlated with the vibroseis pilot sweep and a zero to minimum phase conversion filter was applied for a better identification of the first-arrival onsets of the P wave (Fig. 3a). The vibroseismograms are characterised by a higher dominant frequency band with resultant higher resolution in contrast to the lower-frequency explosive data (Fig. 3b). Nevertheless, in both record sections a decreasing signal/noise ratio with increasing distance to the source is visible.

A total of 124 444 P-wave arrivals were picked. To obtain information about the pick quality, a reciprocity test was carried out (Zelt 1999), by checking the difference between the traveltime from shot A to the station at shot B and the traveltime from shot B to the station at shot A. The exact traveltime at the position of the source was determined by linear interpolation of the picked traveltime of the neighbouring stations. With the picked traveltimes 86 765 reciprocal time differences could be calculated with a rms error of 19.8 ms, a number which acceptable for our tomographic inversion result.

**TOMOGRAPHIC INVERSION**

For the tomographic traveltime inversion, the approach of Zelt & Barton (1998) was used.

To create an initial model, a common mid point (CMP) gather was constructed for each successive 10-km interval of the profile,
Figure 4. (a) Initial model for the P-velocity tomographic traveltime inversion (vertical exaggeration 5:1). (b) The result of the tomographic inversion of 124,444 traveltimes, characterized by a rms error of 20.9 ms (vertical exaggeration 5:1). Only regions from the deepest node crossed by at least 1 ray at each node column are shown. The model shows a high-velocity zone in the centre of the basin and a subtle low-velocity zone at shallow depth (800 m) on the western shoulder from profile km 10 to 40. Numbers 430 and 929 indicate the positions of the sources shown on the map (Fig. 1a) and the corresponding record sections in Fig. 3. WBF, Western Boundary Fault; EBF, Eastern Boundary Fault; B, Border between Israel and Jordan. (c) The hit count distribution shows that the rays focus in the high-velocity areas, which leads to a higher coverage and resolution in the high-velocity shallow parts of the model (vertical exaggeration 5:1).

from which a 1-D P-velocity model was created by trial-and-error traveltime matching. The resultant 1-D P-velocity models were linearly interpolated to create a 2-D P-velocity initial model. Finally each model node column was vertically shifted in accordance to the topography (Fig. 4a).

This 2-D initial model was used in our traveltime tomography to derive the final P-velocity model displayed in Fig. 4(b). For the forward modelling a constant horizontal and vertical node spacing of 50 m was used which resulted in a model with 2481 horizontal and 161 vertical nodes. In contrast to the forward modelling, the inversion was carried out for different cell sizes following the iterative approach of Ryberg et al. (2007). The cell size was thus decreased from initial values of 4 km by 1.6 km to 100 m by 50 m. The inversion result calculated with a larger cell size was used as
Figure 5. Results of the stability test. (a, c, e, g and i) show in black the velocity depth function of a complete initial model (vertical exaggeration 5:1). Note that the initial models in (a), (c) and (e) are laterally inhomogeneous. (b, d, f, h and j) are the corresponding result of the traveltime tomography. (b) Only 10 per cent randomly selected traveltimes were used for the inversion with the same initial model as shown in Fig. 4(a). (d) P-velocity model for an initial model with 10 per cent slower velocities than in Fig. 4(a). (f) P-velocity model for an initial model with 10 per cent faster velocities than in Fig. 4(a). (h) P-velocity model for a 1-D single gradient initial model. (j) P-velocity model for an initial model which simplified the 1-D velocity depth function of the final velocity model (Fig. 4b) into a combination of piecewise continuous linear gradients with depth. These fairly similar models indicate a very robust inversion, for example, all models show the low-velocity layer on the west at shallow depth and the high-velocity zone in the DSB.
the initial model for the inversion with a smaller cell size (for details of this approach see Ryberg et al. (2007)).

Special effort was spent to evaluate the pick uncertainty, which is, amongst others, used as a stop criterion for the inversion. Due to the large number of manually detected P-wave first arrivals, an increasing pick uncertainty with increasing source–receiver offset was assumed. The pick uncertainty was set to a quarter of the period of the first arrival signal, which was measured at several constant source–receiver offsets. Due to the large differences in the period between a vibroseis source (west) and an explosive source (east), two different pick uncertainties were calculated and applied according to the type of source.

With all picked first-arrival traveltimes a P-velocity model with a rms error of 20.9 ms was obtained. The final P-velocity model (Fig. 4b) shows a high-velocity body in the centre of the DSB from profile km 60–70. Further, on the west side of the DSB at 800 m depth from the surface, a horizontal low-velocity zone (LVZ) with P-velocities of about 3 km s\(^{-1}\) is embedded in regions with P-velocities of 4 km s\(^{-1}\). This velocity distribution can also be seen in the seismograms—the apparent velocity increases up to an absolute source–receiver offset of 3 km and decreases then up to an offset of 6 km with increasing apparent velocities again (Fig. 3a). The higher velocities above the LVZ lead to a concentration of rays with affiliated higher resolution there (Fig. 4c).

The rms error is only one characteristic value for the quality of velocity model fit. However, this value does not explain in detail how good a model represents the data. To further evaluate the velocity model and analyse the inversion quality, it is necessary to carry out some additional tests. All test inversions have been carried out with the same inversion parameters that are also used for the inversion of the final P-velocity model.

At first, a stability test (Ryberg et al. 2007) was carried out, which comprised inversions with different initial models or number of traveltime picks (Fig. 5). This test shows the extent to which the final P-velocity model (Fig. 4b) depends on the initial model and also on the ray coverage. First, an inversion was carried out with only 10 per cent of the data, randomly selected from the full data set. The resulting model is characterised by a rms error of 20.8 ms (Fig. 5b). To study the influence of the initial model, an initial model was tested with 10 per cent higher velocities than used for Fig. 4(a) (see Fig. 5c). In this case the resulting model (Fig. 5d) has a rms error of 21.5 ms. Next, an initial model with 10 per cent lower velocities than the initial model for Fig. 4(a) (see Fig. 5e) was applied for the inversion and then the resulting model is characterised by a rms error of 25.7 ms (Fig. 5f). To study the LVZ on the west side of the model, an initial model with a velocity of 4 km s\(^{-1}\) at the surface and a constant gradient of 0.012 km s\(^{-1}\) per km was used, which corresponds to a difference of 1 m s\(^{-1}\) per node using a node spacing of 50 m (Fig. 5g). This resulted in a model with a rms error of 23.7 ms (Fig. 5h). Finally, a more complicated 1-D initial model was used, which approximates the velocity distribution of the western part of the final model constructed of successively different constant gradients with depth (Fig. 5i). This initial model leads to a tomographic inversion result with a rms error of 26.8 ms (Fig. 5j). All five test inversions show only small differences in their model properties and in rms error compared to the preferred model in Fig. 4(b). The main features of the final velocity model are similar, for example, the high-velocity areas in the DSB and the LVZ in the shallow regions of the western blocks. Due to the smaller number of data, the inversion result using 10 per cent randomly selected data is characterised by a reduced depth of ray penetration and a rms error lower than the final P-velocity model.

Our next test was a recovery test based on the final velocity model. From the final velocity model (Fig. 4b), synthetic traveltimes were calculated and after this a Gaussian-distributed random noise jitter with the same magnitude as the final rms error for the real data inversion was added to the synthetic traveltimes. These noisy synthetic traveltimes were used as observed data for a tomographic inversion with the same initial model (Fig. 4a) and inversion parameters as were used for the inversion of the final velocity model (Fig. 4b). Artefacts of the inversion should removed in the resulting model and main features should be ‘boosted’. The resulting model (Fig. 6) shows the main features of Fig. 4(b): for example, the high-velocity area in the DSB and the LVZ at shallow depths in the western block of the model.

In a third set of tests, the size of resolvable structures in dependence of the position in the model was studied using a checker-board test with a perturbation range of –10 per cent to 10 per cent of the background velocities (Zelt 1998). Three different block sizes were used: 4 km by 2 km (Fig. 7a), 2 km by 1 km (Fig. 7b) and 1 km by 0.5 km (Fig. 7c). For those models, synthetic traveltimes were calculated with a smoothed version of the final P-velocity model as the

![Figure 6](image.png)

Figure 6. Result of the recovery test (vertical exaggeration 5:1). Synthetic traveltimes calculated from the final P-velocity model (Fig. 4b) have had random-noise added and the reinverted with the initial model shown in Fig. 4(a). Note the preservation of the low-velocity zone at shallow depth west of the DSB (profile km 10 to 40).
Figure 7. Results of the checker-board test. The test was carried out for three different block sizes (vertical exaggeration 5:1): (a) 4 km by 2 km, (b) 2 km by 1 km and (c) 1 km by 0.5 km. The depth of resolvable structures decreases with decreasing block size. Due to the smaller distances between the vibrator points (50 m) in the western part, this part of the model is better resolved than in the eastern part with larger distances between the explosive source points (500 m). 1 km by 0.5 km sized features can be resolved to a depth of $\sim$1.5 km. Larger features (4 km by 2 km) can be imaged down to a depth of $\sim$2 km. Further, a Gaussian-distributed random-noise jitter was added to the synthetic traveltimes as for the recovery test. As expected, the maximum depth of resolvable structure increases with increasing size of the structures. Structures with a size of 1 km by 0.5 km are resolvable to a depth of $\sim$1.5 km whereas structures with a size of 4 km by 2 km are resolvable down to a depth of $\sim$2 km. Furthermore, the checker-board tests show that the western part of the model is better resolved than the eastern part. The reason for this is the smaller distance between the sources caused by the higher number of sources on the west (the western part with 864 vibroseis sources and the eastern part with 108 explosive sources).

**DISCUSSION**

For correlation with the regional geology, the geological map (Fig. 8a), the information of five boreholes and the outcrops in the wadi ‘Ibn Hammad’ (Figs 2 and 8a) were used. Further for the area of the DSB a geological model based on gravity, magnetic and seismic refraction data (Garfunkel & Ben-Avraham 1996) was added to create a geological model.

Linear interpolation between the boreholes was used to provide an initial geologic model and this model was updated with the velocity distribution and the position of reflectors in the migrated depth section (Figs 8b and c).
Figure 8. Correlation of the $P$-velocity model with the geology and the migrated NVR depth section of Weber et al. (2011). (a) Geological map (enlarged part of Fig. 1b). Black dots indicate the positions of the vibroseis (west) and explosive (east) sources. HM, Hazerim 1; ZD, Zohar 8 Deep; LT, Lot 1; EL, El Lisan 1; SA, Swaqua 1. (b) The final $P$-velocity model from Fig. 5(b) overlain by a migrated depth section (Weber et al. 2011) and a geological model from Fig. 8(c). Note the good correlation between the transition from the Upper to the Lower Cretaceous, Lower Cretaceous to Jurassic and the position of the velocity shallowing in the western part of the velocity model (anticline). (c) Geological model for this area based on borehole information and Garfunkel & Ben-Avraham (1996) for the basin area. The flanks of the salt dome under the Lisan peninsula in the centre of the basin are not well constrained. WBF, Western Boundary Fault; EBF, Eastern Boundary Fault.

At the top of the model on the western and on the eastern block, a thin Tertiary unit could identified in the velocity model ($< 3 \text{ km s}^{-1}$) in accordance with the Hazerim 1 and Swaqua 1 boreholes and the surface geology.

Between profile km 20 and 30 the reflections indicate an anticlinal structure that can be associated with the Syrian Arc Fold Belt which extends from the west of the Sinai up to the north of the Dead Sea and contains many thrust faults (Quennel 1956; Sneh et al. 1998; Kesten et al. 2008). The western flank of this anticlinal structure can also be observed in the velocity model with shallowing high velocities. East of the anticlinal structure 2 strong reflectors correlate with the bottoms of the upper and Lower Cretaceous. These boundaries can also be verified in the velocity model. The bottom of the Upper Cretaceous consists of dolomites and limestones, which are characterized by higher velocities (Fig. 8b, yellow colour) in contrast to the underlying Lower Cretaceous sandstone and limestone (orange colour) (Gilboa et al. 1993). Deeper boundaries, for example, between the Jurassic and the Triassic cannot be confirmed.
due from the reflection data and the marginal ray coverage in the refraction data.

The position of the Western Boundary Fault (WBF) is clearly marked by a high-velocity area (green/blue colours) on the west side of the WBF and a low-velocity area (red and yellow colours) on the east side of the WBF. The velocity distribution in this part of the model supports the interpretation of a combination of a strike-slip and normal faulting (ten Brink et al. 1993; Garfunkel & Ben-Avraham 1996) which produce eastward-dipping reflectors between profile km 53 and 57 in the migrated depth section (Fig. 8b).

Only the width of the sinking blocks are smaller than published in Garfunkel & Ben-Avraham (1996, profile C). A possible reason for this is the more southward position of the DESIRE NVR profile where the distance between the WBF and the Lisan peninsula is shorter.

The horizontal position of the salt diapir in the DSB is well defined (profile km 63 to 70), but the deeper velocity structure in the salt diapir is not constrained due to the small number of rays covering this area caused by a gap of the sources between profile km 55 and 61. Furthermore, this area is characterized by a high-velocity–depth gradient, which leads to shallow penetration of rays. The expected complex inhomogeneous structure of the salt diapir is visible in the diffractions (hyperbolae) in the migrated depth section. The top of the salt diapir as seen in the Lisan 1 borehole is consistent with the velocity distribution in the tomographic model (red/yellow to green colours).

In the vicinity of the Eastern Boundary Fault (EBF) velocity structures cannot be easily correlated with a fault. The section east of the EBF looks like a large sagging structure hanging on the eastern flank, with more than 1 km drop-down. In the eastern part of the model, the transition between the Lower Cretaceous and the Cambrian/Ordovician/Silurian corresponds to a diffuse transition from the shallow, low P-velocities of about 3.5 km s\(^{-1}\) (orange colour) to higher velocities of about 4.0 km s\(^{-1}\) (yellow colour). The transition between the Cambrian and the Precambrian can be observed in the velocity model between profile km 71 and 110 with the transition from yellow to green and also in the migrated depth section between profile km 102 and 112 with strong reflections.

Some of the changes in the surface geology cannot be correlated with the velocity model. With respect to the basin flows east of the Dead Sea, we would expect an area of shallow, high seismic velocity in the eastern part of our model as well. However, this cannot be confirmed by the model. A possible explanation is that the flows are only a surface feature, that is fed by a small vertical mafic intrusions. A further example is the location of the WBF and EBF. The position of these faults are clearly visible in the geological map but not in the velocity model at very shallow depths (∼0 to 500 m).

**CONCLUSION**

With traveltime tomography a shallow P-velocity model of the southern DSB area was derived. This model with an rms error of 20.8 ms is characterised by two prominent features:

1. A horizontal low-velocity layer with 3 km s\(^{-1}\) embedded in regions with 4 km s\(^{-1}\) at shallow depth (∼800 m) in the western part of the model.
2. A high velocity zone in the middle of the DSB, which corresponds to the Lisan salt diapir.

Test inversions with different initial models and a different number of traveltime picks show that the inversion is very robust due to the large number of data and the high pick quality. A checker board test was carried out with different block sizes for an estimation of the size of resolvable structures in dependence of their positions in the model. A good correlation between the final P-velocity model, a migrated NVR depth section and a geological model was visible in several parts of these models.

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All figures were prepared using the Generic Mapping Tool (GMT; Wessel & Smith 1998).

**REFERENCES**


