

Originally published as:

Weber, M., Abu-Ayyash, K., Abueladas, A., Agnon, A., Alasonati Tasárová, Z., Al-Zubi, H., Babeyko, A., Bartov, Y., Bauer, K., Becken, M., Bedrosia, P. A., Ben-Avraham, Z., Bock, G., Bohnhoff, M., Bribach, J., Dulski, P., Ebbing, J., El-Kelani, R., Förster, A., Förster, H.-J., Frieslander, U., Garfunkel, Z., Goetze, H. J., Haak, V., Haberland, C., Hassouneh, M., Helwig, S., Hofstetter, A., Hoffmann-Rothe, A., Jäckel, K. H., Janssen, C., Jaser, D., Kesten, D., Khatib, M., Kind, R., Koch, O., Koulakov, I., Laske, G., Maercklin, N., Masarweh, R., Masri, A., Matar, A., Mechie, J., Meqbel, N., Plessen, B., Möller, P., Mohsen, A., Oberhänsli, R., Oreshin, S., Petrunin, A., Qabbani, I., Rabba, I., Ritter, O., Romer, R. L., Rumpker, G., Rybakov, M., Ryberg, T., Saul, J., Scherbaum, F., Schmidt, S., Schulze, A., Sobolev, S. V., Stiller, M., Stromeyer, D., Tarawneh, K., Trela, C., Weckmann, U., Wetzel, U., Wylegalla, K. (2009): Anatomy of the Dead Sea transform from lithospheric to microscopic scale. - *Reviews of Geophysics*, 47, RG2002, 10.1029/2008RG000264.

ANATOMY OF THE DEAD SEA TRANSFORM FROM LITHOSPHERIC TO MICROSCOPIC SCALE

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Received 12 February 2008; revised 12 August 2008; accepted 24 October 2008; published XX Month 2009.

[1] Fault zones are the locations where motion of tectonic plates, often associated with earthquakes, is accommodated. Despite a rapid increase in the understanding of faults in the last decades, our knowledge of their geometry, petrophysical properties, and controlling processes remains incomplete. The central questions addressed here in our study of the Dead Sea Transform (DST) in the Middle East are as follows: (1) What are the structure and kinematics of a large fault zone? (2) What controls its structure and kinematics? (3) How does the DST compare to other plate boundary fault zones? The DST has accommodated a total of 105 km of left-lateral transform motion between the African and Arabian plates since early Miocene (~20 Ma). The DST segment between the Dead Sea and the Red Sea, called the Arava/Araba Fault (AF), is studied here using a multidisciplinary and multiscale approach from the μm to the plate tectonic scale. We observe that under the DST a

narrow, subvertical zone cuts through crust and lithosphere. First, from west to east the crustal thickness increases smoothly from 26 to 39 km, and a subhorizontal lower crustal reflector is detected east of the AF. Second, several faults exist in the upper crust in a 40 km wide zone centered on the AF, but none have kilometer-size zones of decreased seismic velocities or zones of high electrical conductivities in the upper crust expected for large damage zones. Third, the AF is the main branch of the DST system, even though it has accommodated only a part (up to 60 km) of the overall 105 km of sinistral plate motion. Fourth, the AF acts as a barrier to fluids to a depth of 4 km, and the lithology changes abruptly across it. Fifth, in the top few hundred meters of the AF a locally transpressional regime is observed in a 100–300 m wide zone of deformed and displaced material, bordered by subparallel faults forming a positive flower structure. Other segments of the AF have a

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52 transtensional character with small pull-aparts along them.
 53 The damage zones of the individual faults are only 5–20 m
 54 wide at this depth range. Sixth, two areas on the AF show
 55 mesoscale to microscale faulting and veining in limestone
 56 sequences with faulting depths between 2 and 5 km.
 57 Seventh, fluids in the AF are carried downward into the
 58 fault zone. Only a minor fraction of fluids is derived from
 59 ascending hydrothermal fluids. However, we found that on

69 **Citation:** Weber, M., et al. (2009), Anatomy of the Dead Sea Transform from lithospheric to microscopic scale, *Rev. Geophys.*, 47,
 70 XXXXXX, doi:10.1029/2008RG000264.

71 1. INTRODUCTION

73 [2] Large faults are the most prominent surface expres-
 74 sions of crustal and lithospheric processes driven by
 75 motions within the interior of the Earth. Key sites to study
 76 large strike-slip faults include the San Andreas Fault (SAF)
 77 in California, USA; the Alpine Fault in New Zealand; the
 78 North Anatolian Fault (NAF) in Turkey; and the Dead Sea
 79 Transform (DST) in the Middle East (Figure 1). These and
 80 other fault zones constitute some of the most visible
 81 expressions of plate tectonics and can be traced at the
 82 surface for hundreds of kilometers. Active faults are also
 83 the locations of large earthquakes responsible for the risk
 84 associated with seismic hazard. In recent years it became
 85 widely accepted that transform faults represent pathways for
 86 the movement of fluids, especially water/brines, and sites of
 87 enrichments in minerals. Despite their social and economic
 88 relevance and much progress in their understanding in the
 89 last decades, one of the most challenging problems and still
 90 one of the key questions of plate tectonics remains: to
 91 understand in more detail the initiation and the spatial and
 92 temporal evolution of active large fault zones.

93 [3] Recently, several authors have addressed this question
 94 summarizing the present state of knowledge. One focus has
 95 been the internal structure of crustal fault zones and the
 96 mechanism of faulting at different depth ranges [Chester *et al.*,
 97 1993; Holdsworth *et al.*, 2001; Ben-Zion and Sammis,
 98 2003; Faulkner *et al.*, 2003]. Other studies have used
 99 geophysical methods [e.g., Ritter *et al.*, 2005a] to show
 100 similarities and differences between geophysical images of
 101 fault zones. The most recent and comprehensive review,
 102 including an exhaustive list of recent references, is given by
 103 Handy *et al.* [2007] and, from a geophysical perspective, by
 104 Mooney *et al.* [2007]. One of their key recommendations
 105 and challenges for the future is to study fault zones in a
 106 multidisciplinary approach focused on natural laboratories
 107 and looking at interacting processes within faults. Because
 108 of the wide range of parameters and forces controlling the
 109 shape and kinematics of faults the comparison of the results
 110 derived at several natural laboratories is an absolute neces-
 111 sity to gain a better insight into these processes. In that
 112 context, Mooney *et al.* [2007] list several outstanding key
 113 questions concerning faults, for example, (1) How far down
 114 into the crust do fault damage zones extend? (2) Does
 115 crustal deformation become narrower or wider in the lower
 116 crust? (3) Is there a physical relationship between pro-

the kilometer scale the AF does not act as an important fluid 60
 conduit. Most of these findings are corroborated using 61
 thermomechanical modeling where shear deformation in the 62
 upper crust is localized in one or two major faults; at larger 63
 depth, shear deformation occurs in a 20–40 km wide zone 64
 with a mechanically weak decoupling zone extending 65
 subvertically through the entire lithosphere. 66

nounced seismic low-velocity zones and high-conductivity 117
 zones? 118

[4] Here we address some of the key questions and give 119
 an overview of our multidisciplinary investigation (Figure 2) 120
 that aimed to quantify the physical processes responsible 121
 for forming the Arava/Araba Fault (AF) segment of the 122
 DST. This is done by combining findings from seismology, 123
 electromagnetics, gravity, geothermics, petrology, geochem- 124
 istry, field mapping based mainly on surface geology as well 125
 as satellite image interpretation and remote sensing, and 126
 thermomechanical numerical simulations. The spatial scales 127
 analyzed range from 10 + 5 to 10 – 5 m with the goal of 128
 understanding the evolution of the DST through time. These 129
 findings are compared to results known from the most 130
 intensively studied fault worldwide, the SAF. For details, 131
 see Li *et al.* [1990], Brocher *et al.* [1994], Holbrook *et al.* 132
 [1996], Henstock *et al.* [1997], Unsworth *et al.* [1997, 133
 2000], Ryberg and Fuis [1998], Fuis *et al.* [2001], Hole *et al.* 134
 [2001], and Becken *et al.* [2008]. The most recent 135
 overviews are given by Mooney *et al.* [2007], Wilson *et al.* 136
 [2005], and Fuis *et al.* [2008], who compared the 137
 transpressional plate boundaries of the SAF and the Alpine 138
 Fault in New Zealand, and by Jiracek *et al.* [2008], who 139
 studied the SAF, the Alpine Fault, and the Yarlung-Tsangpo 140
 Suture in southern Tibet. 141

[5] In section 2 we summarize the results of our multi- 142
 disciplinary geophysical approach to imaging and mapping 143
 the AF, from lithospheric to crustal scale, then from the 144
 kilometer to meter scale, and finally down to the μm scale. 145
 These results, summarized in Figure 3, are then compared to 146
 petrological and geochemical studies together with findings 147
 from surface geology and thermomechanical modeling. In 148
 section 3 we compare our results for the AF with those from 149
 specific locations of the SAF, finding some similarities but 150
 also significant differences. Section 4 contains the discus- 151
 sion and conclusions and an outlook on future challenges. 152

153 2. DEAD SEA TRANSFORM

[6] The DST is a ~ 1000 km long left-lateral fault zone 154
 that extends from the Red Sea spreading center to the 155
 Zagros zone of plate convergence (Figure 1). It cuts through 156
 a continental area whose crust was shaped by the end- 157
 Proterozoic Pan-African Orogeny. Later, from Cambrian to 158
 Paleogene times, it behaved as a stable platform and was 159
 covered by extensive sediment cover. The platformal history 160
 was interrupted by a period of faulting and magmatism 161

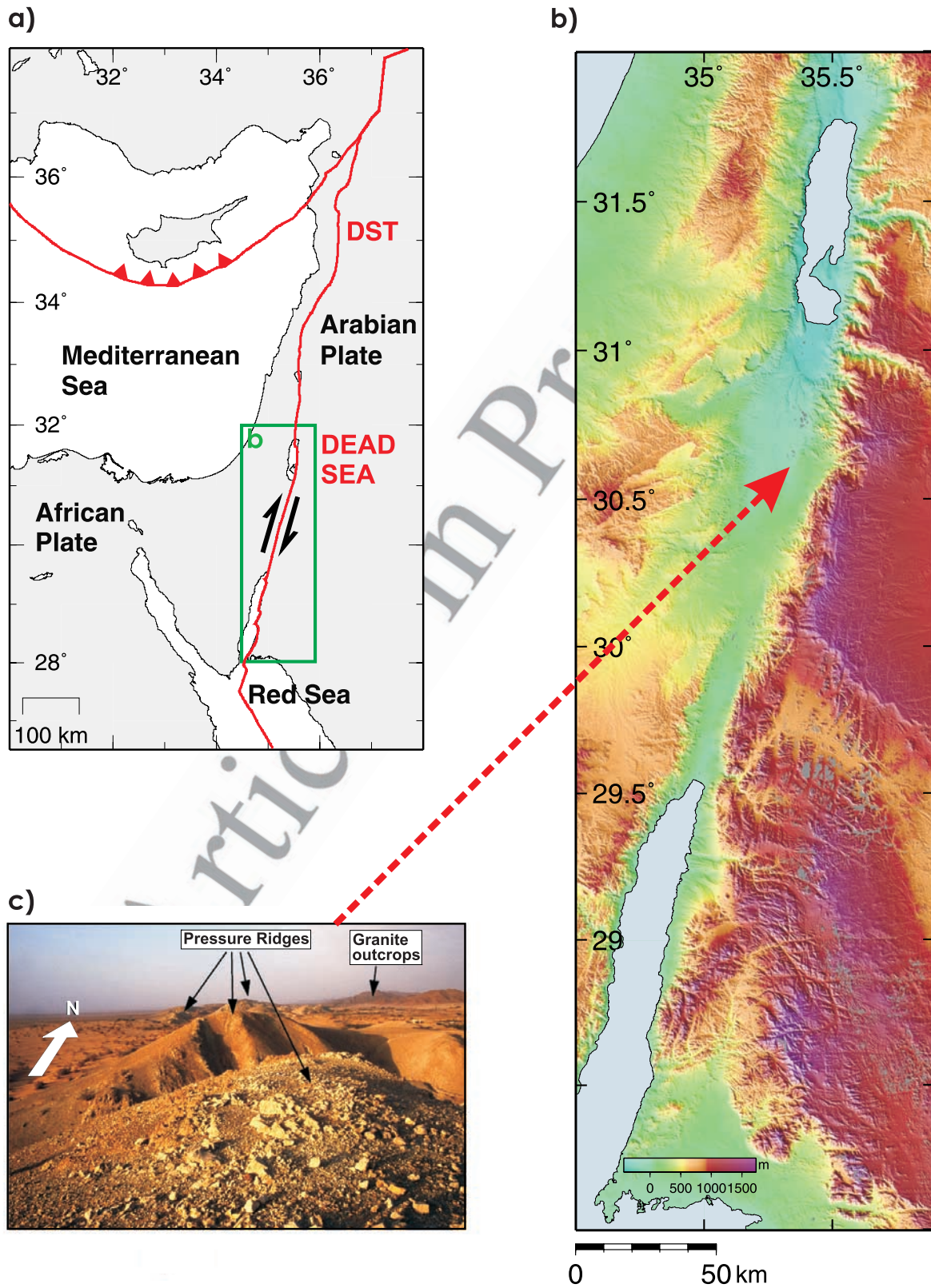


Figure 1. (a) Tectonic setting of the Levant. DST, Dead Sea Transform. (b) Relief map of the region between the Dead Sea and the Red Sea (green box in Figure 1a). (c) Pressure ridges in the Arava/Araba Valley between lines P9 and P10 in Figure 2f and at line 5 in Figure 2g.

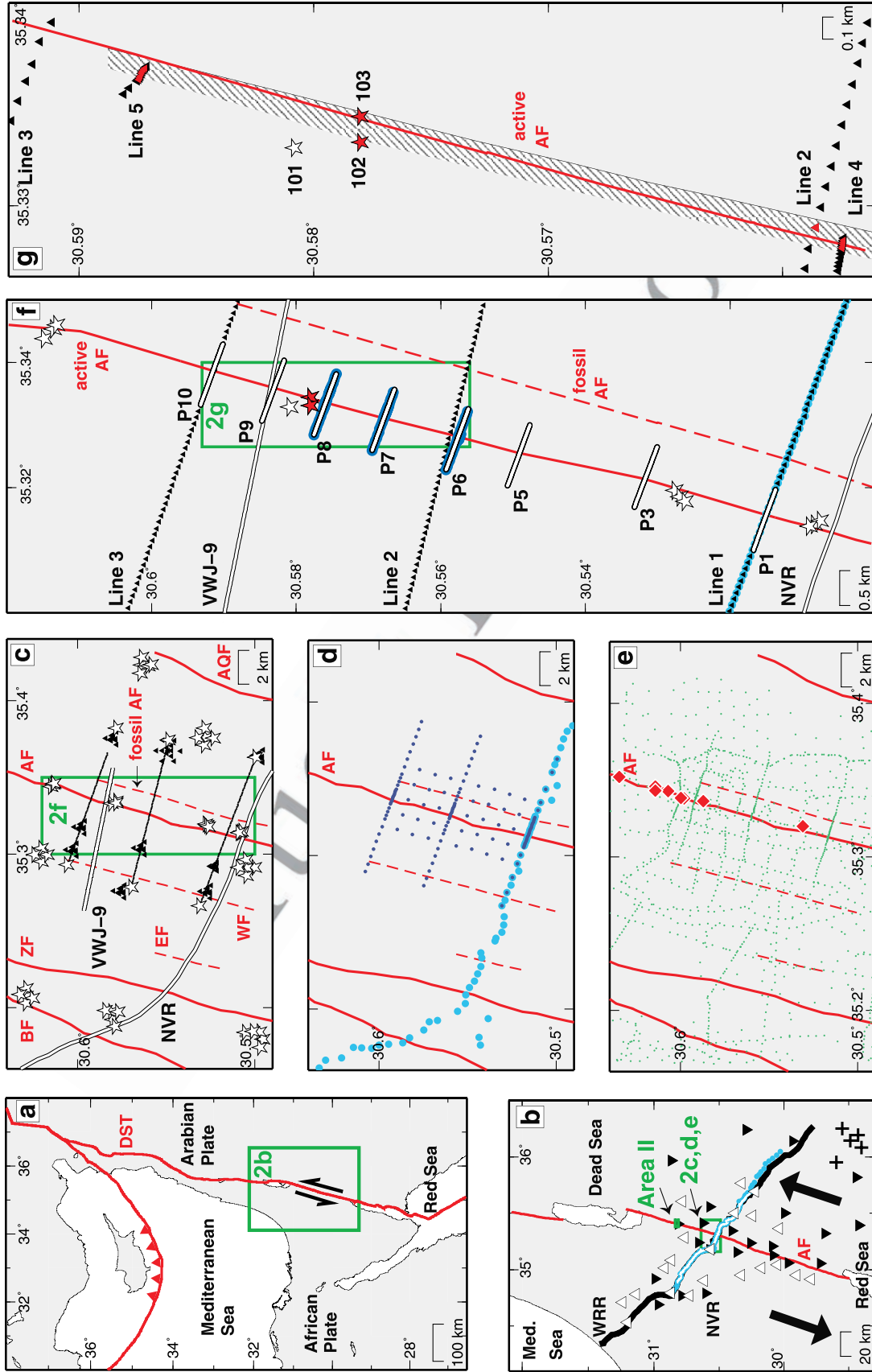


Figure 2

162 related to shaping the Levant continental margin in the early
163 Mesozoic, by intraplate magmatism in the Early Cretaceous,
164 and by mild folding and shearing in Late Cretaceous to
165 Miocene times [Garfunkel, 1988, and references therein].
166 The DST is one of the new plate boundaries that formed
167 when this continental area broke up in mid-Cenozoic times
168 [Freund, 1965; Gass, 1970; Garfunkel, 1981; Bosworth *et*
169 *al.*, 2005, and references therein].

170 [7] The ~600 km long part of the DST south of Lebanon
171 has a broad arcuate shape in map view and is mostly marked
172 by a conspicuous valley, while its more northern part is less
173 regular. Our study deals with the southern segment, focus-
174 ing on an area that was little affected by previous Phanero-
175 zoic deformation. Matching of the rock bodies and
176 structures across the transform along this segment indicates
177 a left lateral offset of ~105 km (uncertainty is a couple of
178 kilometers) and that the motion postdated ~20 Ma old dikes
179 [Quennell, 1958; Freund *et al.*, 1970; Bartov, 1974; Bandel
180 *and Khouri*, 1981; Eyal *et al.*, 1981]. Regional plate
181 kinematics give an independent similar estimate of the
182 offset, ~100 km (uncertainty < 20 km), and also show that
183 much of the offset occurred already in Miocene times [Joffe
184 *and Garfunkel*, 1987; Garfunkel *and Beyth*, 2006]. The
185 structure along the southern half of the DST [Garfunkel,
186 1981; Garfunkel *and Ben-Avraham*, 2001; Ben-Avraham *et*
187 *al.*, 2008] is dominated by longitudinal strike-slip faults
188 arranged en echelon that produce a series of deep pull-apart
189 basins (also called rhomb grabens) alternating with much

190 shallower fault-controlled structural saddles. These struc-
191 tures are embedded in a 3–20 km wide depressed zone that
192 forms a valley delimited on both sides by normal faults so
193 that the southern segment of DST superficially resembles
194 extensional rifts. These faults displace very young sedi-
195 ments and have a direct physiographic expression, indicat-
196 ing their continuing activity [Garfunkel *et al.*, 1981]. The
197 strike slip faults trend at a small angle to the overall
198 elongation of this part of the DST, which indicates a small
199 component of transtension that is augmented by the longi-
200 tudinal normal faults. In addition, near our transect and
201 farther south a part of the lateral motion took place on
202 several faults crossing the margins of the topographic valley
203 [Eyal *et al.*, 1981; Sneh *et al.*, 1998], but farther north it was
204 concentrated in the topographic valley. The development of
205 the DST was accompanied by uplifting of the flanking
206 regions. Also, igneous activity took place in a wide region,
207 mainly east of it and along its northern half, but it was very
208 limited near our transect [Garfunkel, 1989; Sneh *et al.*,
209 1998].

210 [8] Since the advent of plate tectonics the DST has been
211 considered a prime site to examine large shear zones
212 [Quennell, 1958; Wilson, 1965; Freund *et al.*, 1970]. Large
213 historical earthquakes on the DST with magnitudes up to 7
214 [Garfunkel *et al.*, 1981; Ambraseys *et al.*, 1994; Amiran *et*
215 *al.*, 1994; Klinger *et al.*, 2000b; Ken-Tor *et al.*, 2001;
216 Migowski *et al.*, 2004; Agnon *et al.*, 2006] and the recent
217 1995 Nueiba *M*7.2 event [Hofstetter *et al.*, 2003] as well as

Figure 2. Setting of the DST and location of experiments within the Dead Sea Rift Transect (DESERT) project. Known faults are indicated in red. The area of subsequent blow-ups is indicated by a labeled green box. Note different scales. (a) Main faults of the Levant [see, e.g., Garfunkel, 1981; Ben-Avraham, 1985]. The left lateral displacement of 105 km at the DST is indicated by black arrows. The area of Figure 2b is given by the labeled green box. (b) The red lines indicate the main branches of the DST. Seismic/seismological experiments (transect) in black and white: black line, 260 km long wide-angle reflection/refraction (WRR) profile (13 shots and 99 receivers); white line, 100 km long near-vertical reflection (NVR) profile (roll along vibroseis source spacing 50 m, receiver spacing 100 m, and 90-fold coverage), coinciding with the inner part of the WRR. Passive long-term deployment (1.5 years) of seismological stations (21 short-period (open triangles) and 27 broadband stations (full triangles)). Electromagnetic/magnetotelluric (MT) experiments in blue: MT-long, light blue dots, 140 km long, coinciding mostly with NVR. Sites of thermal borehole measurements and surface heat flow determination are given by black crosses. The area of Figures 2c–2e is given by the labeled green box. Area II of the microstructural analysis (Figure 22, top right) is also indicated in green. (c) Central Arava/Araba Valley in Israel/Jordan. The red lines indicate the dominant active faults in the area (BF, Barak Fault; ZF, Zofar Fault; AF, Arava/Araba Fault; AQF, Al Quwayra Fault). The dashed red lines indicate the surface traces of three buried, now inactive faults (Eastern Fault (EF), Western Fault (WF), and fossil AF). Seismic experiments (for details, see section 2): NVR and VWJ-9, white lines; controlled source arrays (CSA) I, thin black lines, 3 × 10 km long; and CSA I miniarrays, black triangles, 9 station arrays (10 seismometers each). White stars indicate CSA I shot locations (7 × 5 shots, 4 × 3 shots on/near the AF, and 3 × 2 shots at end of the three CSA I lines). (d) As Figure 2c but for MT. MT regional profile, large, light blue dots; MT-pilot, dark blue dots, 10 km profile in the vicinity of the AF in the center of the MT regional profile; MT 3-D experiment, small, blue dots, 2 × 10 km and 7 × 4 km profiles. (e) As Figure 2c but for the local gravity survey. Green dots indicate the gravity sites. Locations of mesostructural and microstructural analysis on the AF (area I, Figure 22 (top left)) are given as red diamonds. (f) The red line shows the active AF, and the dashed red line indicates the surface trace of the buried fossil AF. Seismic experiments in black and white: NVR and VWJ-9, white lines; CSA I, black triangles, part of 3 × 10 km long lines (lines 1–3, station spacing 100 m). White stars indicate CSA I shot locations on/near the AF. Red stars indicate shots that produced fault zone guided waves (FZGW). CSA II, short white profiles, 8 × 1 km (P1–P10; station spacing 5 m on each line and shot spacing of 20 m). Electromagnetic experiments in blue: Long Offset Transient Electromagnetics, blue line, 8 km long along line 1 of CSA I (only partially shown); Short Offset Transient Electromagnetics, blue lines, 3 × 1 km long and coinciding with profiles P6, P7, and P8 from CSA II. (g) FZGW experiment on the active AF. Black triangles are stations of CSA I (lines 2 and 3, spacing 100 m (see Figure 2c); lines 4 and 5, spacing 10 m, with 20 stations per line). Stars indicate shot locations (shots 101–103). Red stars and triangles indicate shots and stations that produced or recorded FZGW, respectively. The shallow low-velocity segment of the active AF identified by FZGW is given by the shaded area.

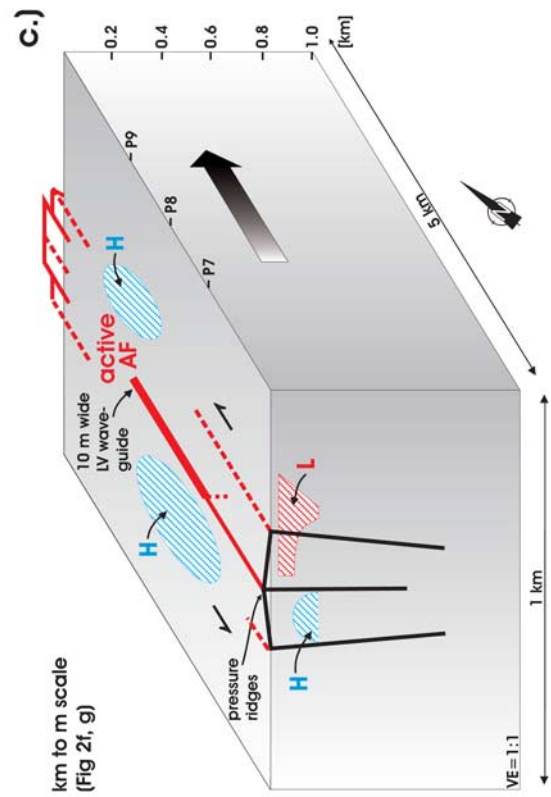
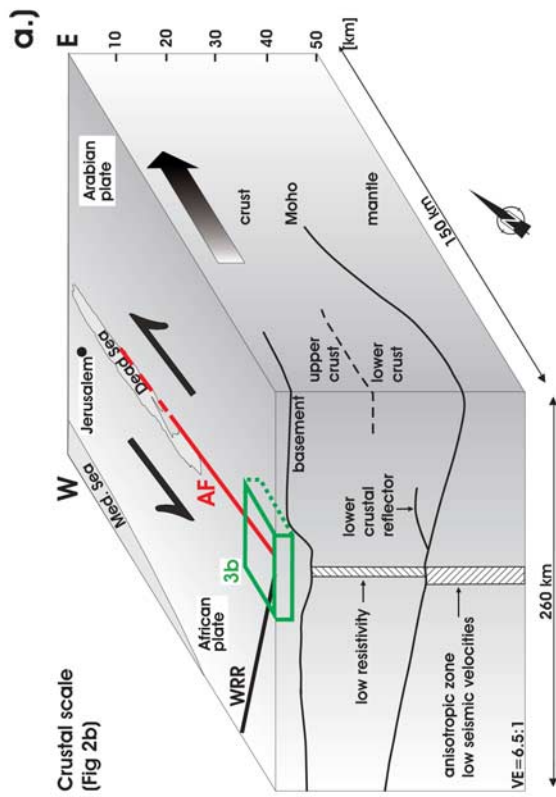
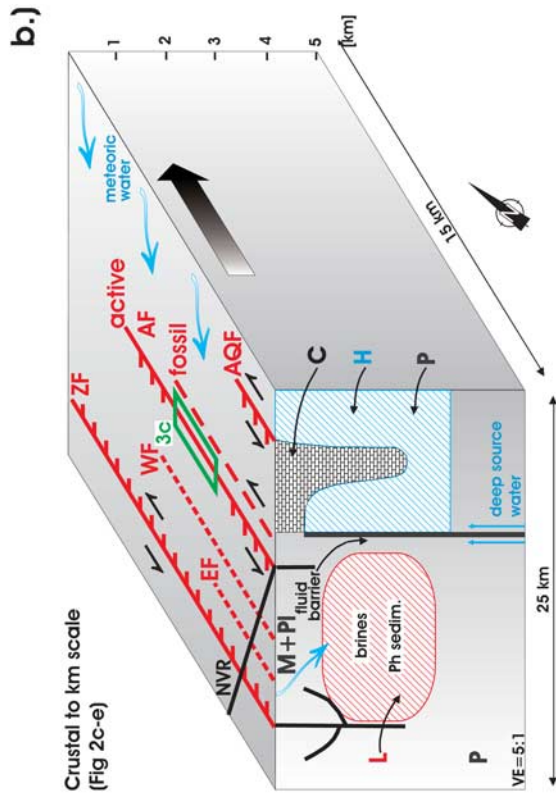


Figure 3

218 ongoing microseismic activity [e.g., *van Eck and Hofstetter*,
 219 1989; *Salamon et al.*, 2003; *Aldersons et al.*, 2003] show
 220 that the DST is a seismically active plate boundary
 221 (Figure 4). The historical catalog goes back for more than
 222 2000 years. It is based on documents by the local witnesses
 223 of devastating events, by pilgrims who came to the “Holy
 224 Land” shortly after the occurrence of the earthquakes, and
 225 by historians that provide detailed accounts of the history of
 226 the region including the felt earthquakes. Some of those
 227 events were catastrophic, causing casualties and widespread
 228 damage. Since no instrumental measurements were avail-
 229 able until the end of the 19th century and most of the literate
 230 population lived north of the latitude of the Dead Sea basin,
 231 the historic earthquakes are assumed to have location errors
 232 of about 100 km (or more) [*Salamon et al.*, 1996]. The
 233 establishment of a few regional seismological stations began
 234 in the early 20th century, followed by several WWSSN
 235 stations in the mid-20th century and at about 1980 by the
 236 seismic networks of Jordan and Israel. This enabled a more
 237 accurate localization of the seismicity along the Dead Sea
 238 Transform. A. Q. Amrat et al. (The Unified Earthquake
 239 Catalog of the Dead Sea Region, 2001, [http://www.relemer-](http://www.relemer-merc.org/)
 240 [merc.org/](http://www.relemer-merc.org/)) compiled and unified the historical and instru-
 241 mental catalogs of Jordan and Israel, now maintained on a
 242 regular basis, serving as a main source for seismotectonic
 243 studies and seismic hazard assessment. From Figure 4 it is
 244 obvious that the DST poses a considerable seismic hazard
 245 to Palestine, Israel, and Jordan. The present-day relative
 246 motion between Arabia and Africa is ~ 5 mm/a [*Klinger et*
 247 *al.*, 2000a; *Wdowinski et al.*, 2004; *Bartov and Sagy*, 2004;
 248 *Mahmoud et al.*, 2005; *LeBeon et al.*, 2006; *Reilinger et al.*,
 249 2006].

[9] Because of the political situation in this area and the
 fact that the DST is situated in the border region between
 Israel and Jordan, geophysical surveys in that area are a
 nontrivial logistical task. Here we present the results of an
 integrated geoscientific investigation along a 260 km long
 transect (Dead Sea Rift Transect (DESERT)) extending
 from the Mediterranean coast to the Arabian platform.
 The central part of the transect crosses the DST at a location
 in the Arava/Araba Valley which is as far away as possible
 from the center of the Dead Sea, a pull-apart basin, and the
 Red Sea, where active rifting occurs, and thus permits the
 study the DST and the tectonic processes controlling it
 without the complications of rifting. The fieldwork com-
 pleted by a team of scientists from Germany, Israel, Jordan,
 and Palestine started in spring 2000 and lasted until 2004.
 To facilitate identifying and locating the areas covered by
 these studies and the results of the analysis on the many
 scales employed, Figure 2 shows the spatially nested experi-
 ments. The areas covered are indicated sequentially by the
 labeled green boxes with mapped faults indicated in red.
 Figure 3 summarizes the results of these studies. A detailed
 discussion of the results leading to Figure 3 will now be
 given starting with the largest, the lithospheric scale.

2.1. Lithospheric Scale

[10] Using fundamental mode Rayleigh waves at inter-
 mediate periods, *Laske et al.* [2008] showed that the sub-
 crustal S velocity under the region of the DST is, on average,
 5% lower than in the preliminary reference Earth model
 (PREM) [*Dziewonski and Anderson*, 1981] down to at least
 200 km. This is in agreement with the larger-scale study of
Pasyanos and Nyblade [2007], which shows that such an
 area of reduced velocities in the lithosphere stretches along

Figure 3. Structure and dynamics of the AF on different scales along the DESERT transect (Figure 2). Solid red lines indicate fault traces visible at the surface, and dashed lines are buried faults. Relative motions at faults are indicated by arrows (strike slip) and ticks (minor normal motion), and buried features (bodies) are outlined at the surface by dashed areas. Red bodies (“L”) are regions of low seismic velocity and low resistivity. Blue bodies (“H”) are regions of high seismic velocity and moderate to high resistivity. Geological units are also labeled by letters (P, Precambrian; Ph, Phanerozoic; C, Cambrian and Cretaceous; M, Miocene; Pl, Pliocene-Pleistocene). Subsequent blow-ups are indicated by a labeled green box. (a) Tectonic setting of the Levant [see, e.g., *Garfunkel*, 1981; *Ben-Avraham*, 1985]. The left lateral displacement of 105 km is indicated by arrows. The black line at the surface is the WRR profile; compare Figure 2b. The seismic basement is offset 3–5 km under the AF, whereas the Moho shows a gradual increase from 26 to 39 km from west to east with only a maximum of ± 1 km undulation under the AF. The Moho shallows to about 33 km toward the north on the eastern side. Prominent features in the crust are a lower crustal reflector under the eastern side and a subvertical zone with low resistivity located ~ 10 km west of the surface trace of the AF. Below the Moho a low seismic velocity and anisotropic zone with a few kilometers additional offset toward the west continues subvertically deeper into the upper mantle. Vertical exaggeration (VE) = 6.5:1. (b) Central Arava/Araba Valley in Israel/Jordan (green box in Figure 3a). The black line at the surface is the NVR profile; compare also to Figure 2c. Six left-lateral faults, three visible at the surface (solid red lines (ZF, AF, and AQF)) and three buried faults (dashed red lines (EF, WF, and fossil AF)), have been identified. A sedimentary body with low seismic velocities, low resistivity due to brines, and low densities (L) is juxtaposed opposite to a body of Precambrian age with high seismic velocities, moderate resistivity, and higher densities (H). The buried contact of these two bodies (fossil AF) is displaced 0.5–1 km toward the east with respect to the present-day surface trace of the AF (active AF). Note also geological units. The migration of fluids is indicated by blue arrows. VE = 5:1. (c) Central AF (green box in Figure 3b); compare also to Figures 2f and 2g. Several subparallel faults form the AF system on a hundred meter scale. Some of them are visible at the surface, and some are buried (dashed lines). Toward the north, pressure ridge structures dominate. Bodies with low seismic velocities and low resistivity (L) and high seismic velocities and high resistivity (H) can be identified west and/or east of the AF in the shallow subsurface. Note that L and H “bodies” are largely reversed in Figure 3c relative to Figure 3b. In the center of the sketch a ~ 3 km long and ~ 10 m wide shallow low-velocity zone detected by FZGW is indicated.

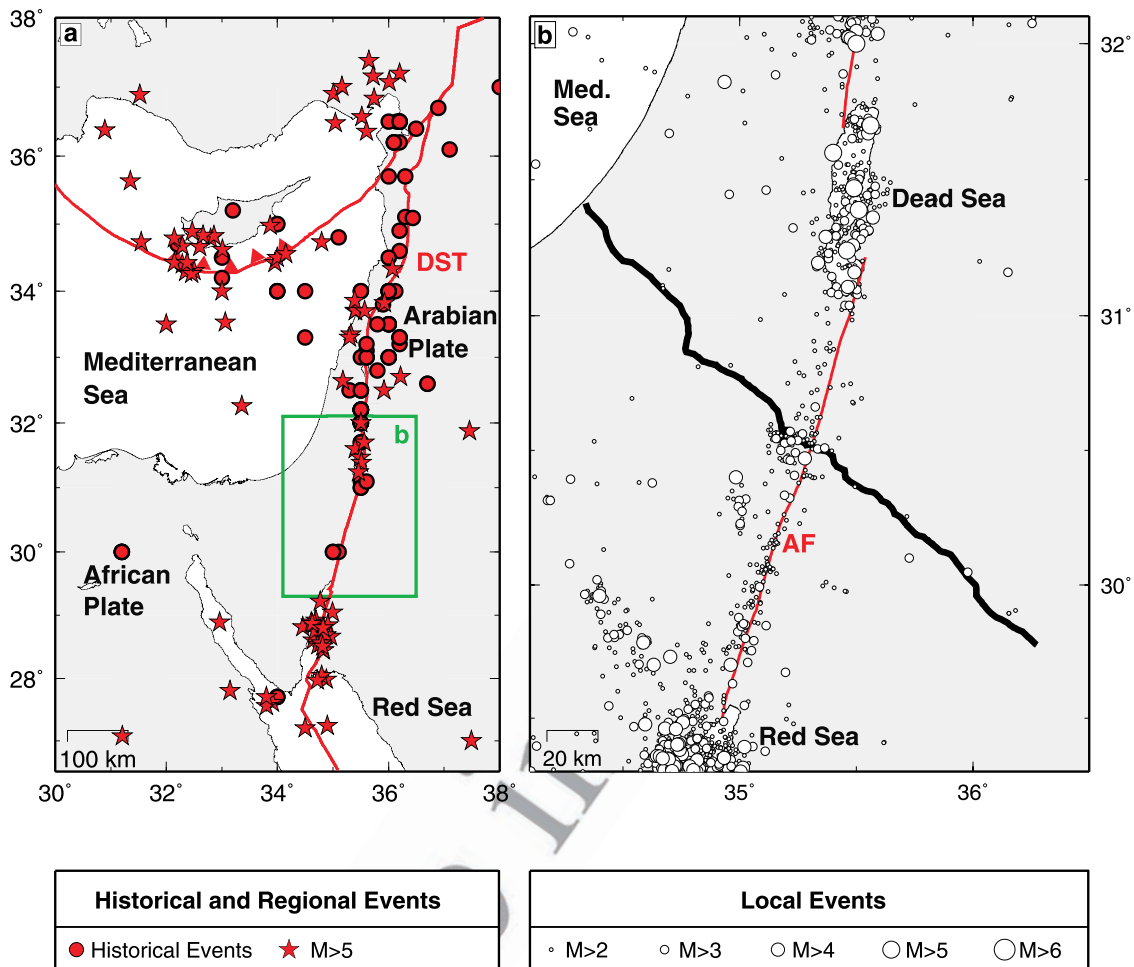


Figure 4. Seismicity of the Levant. Known faults are indicated in red. (a) Historic earthquakes of the last 2000 years are given as red dots, based on Migowski *et al.* [2004] and A. Hofstetter (personal communication, 2008). Earthquakes from 1900 to 2004 with magnitudes larger than 5 are shown as red stars (A. Hofstetter, personal communication, 2008). The area of Figure 4b is given by the labeled green box. (b) Local seismicity from 1983 to 2008 is shown as small white circles, scaled by magnitude (A. Hofstetter, personal communication, 2008). The DESERT profile is indicated in black.

282 the DST. This reduced value may be due to elevated
 283 temperatures connected to rifting processes in the Red
 284 Sea. The present-day conductive surface heat flow east
 285 and west of the DST, on the other hand, mimics the pre-
 286 Cenozoic thermal conditions not affected by thermal
 287 imprints. These imprints are associated with the young
 288 geodynamic processes that include the Red Sea rifting
 289 starting in the Oligocene [e.g., McGuire and Bohannon,
 290 1989; Bosworth *et al.*, 2005, and references therein],
 291 the eruption and emplacement of basaltic lava flows of Oligo-
 292 cene and younger age [e.g., Shaw *et al.*, 2003], and the
 293 formation of the Dead Sea Transform fault system starting
 294 in the Miocene [e.g., Garfunkel *et al.*, 1981; Garfunkel,
 295 1989; Ilani *et al.*, 2001; Shaw *et al.*, 2003; Bosworth *et al.*,
 296 2005; Sobolev *et al.*, 2005]. East of the DST (Figure 2b),
 297 a surface heat flow of 60.3 ± 3.4 mW/m² was determined at
 298 sites virtually unaffected at shallow depth by basalt extru-
 299 sions and high-heat-production granites [Förster *et al.*,
 300 2007]. Steady state geotherms calculated on the basis of
 301 60 mW/m² surface heat flow yield Moho temperatures of

~800°C at 37 km depth [Förster *et al.*, 2004]. West of the
 302 DST, geotherms based on surface heat flow of ~40 mW/m²
 303 [e.g., Ben-Avraham *et al.*, 1978; Eckstein and Simmons,
 304 1978] imply a considerably colder lithosphere, with a Moho
 305 temperature as low as 400°C [Stein *et al.*, 1993]. Both
 306 geotherms significantly underestimate the xenolith-derived
 307 lithospheric mantle temperatures [Stein *et al.*, 1993]. This
 308 discrepancy suggests that heating in close proximity to the
 309 DST, and the region east of it, may be due to lithospheric
 310 thinning. However, the thermal pulse at depth has not yet
 311 reached the surface owing to the time/length scale of
 312 thermal diffusion through the lithosphere [e.g., Turcotte
 313 and Schubert, 2002]. Laske *et al.* [2008] find an 80 km
 314 thick lid of normal *S* velocity, to the west of the DST,
 315 indicative of thermally unaffected crust and upper mantle.
 316 The whole region is furthermore underlain by an upper
 317 mantle, down to 410 km depth, with 3–4% reduced *S*
 318 velocities compared to PREM [Mohsen *et al.*, 2006].
 319

[11] Tomographic inversion of body waves (*P* and PKP
 320 phases) from 135 seismic events recorded at the 48 stations
 321

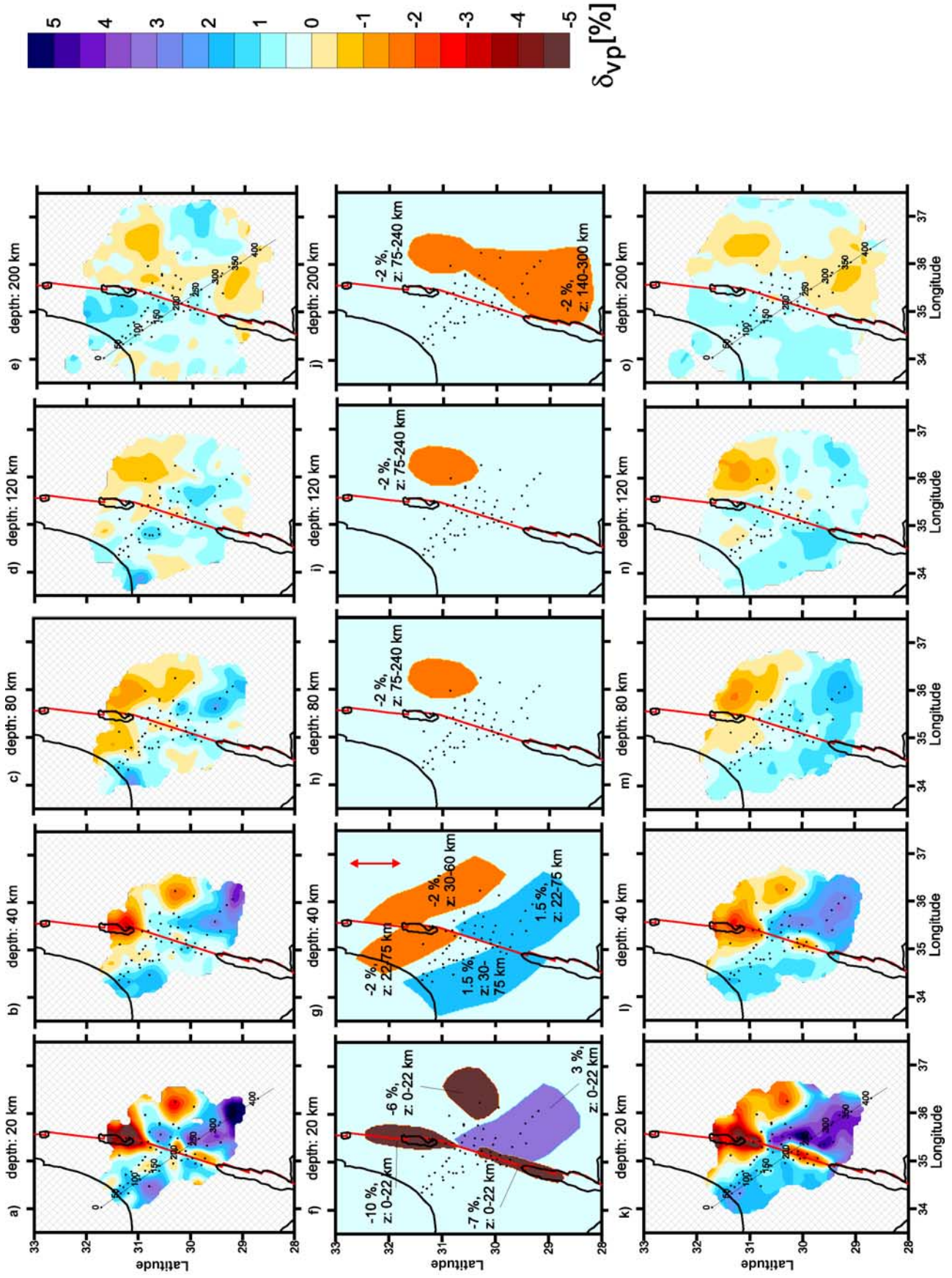


Figure 5

322 of the passive long-term deployment of DESERT (triangles in
 323 Figure 2b) is used to image the P velocity in the upper mantle
 324 under the DST [Koulakov et al., 2006]. After correction for
 325 surface topography and crustal thickness the amplitudes of the
 326 remaining subcrustal travel time anomalies are small. Figure 5
 327 shows the results of the inversion and corresponding resolution
 328 tests. One robust feature is the high-velocity P wave anomaly
 329 of about +1% from ~ 40 to 120 km depth in the SE
 330 (Figures 5a–5e), which could be due to a slightly lower
 331 temperature ($\Delta T < -100^\circ\text{C}$) and/or a preexisting lithospheric
 332 compositional anomaly in this region. Very little topography of
 333 the lithosphere-asthenosphere boundary is observed beneath
 334 the DST. This observation is confirmed by receiver function
 335 studies [Mohsen et al., 2006] showing a typical lithosphere-
 336 asthenosphere thickness of ~ 70 km east and south of the
 337 DST and an increase to 80 km in the north on the eastern side
 338 of the DST. Otherwise, no significant P velocity anomalies are
 339 observed in the upper mantle under the DST. The boundary
 340 between the African and the Arabian plate is detected under-
 341 neath the DST in the subcrustal lithosphere as a narrow,
 342 ~ 20 km wide zone of anisotropic velocity, most likely
 343 produced by fault-parallel mineral alignment in response to
 344 finite strain.

345 [12] Figure 6 shows the range of possible models (three
 346 clusters of models), based on the inversion of shear wave
 347 splitting observations (SKS waves) along the 100 km long
 348 near-vertical reflection (NVR) line (white line in Figure 2b)
 349 [Rümpker et al., 2003; Ryberg et al., 2005]. The preferred
 350 model is cluster 1. It shows that this zone, accommodating
 351 the transform motion between the African and the Arabian
 352 plate, is displaced ~ 10 km to the west relative to the surface
 353 trace of the AF. The geometry of this zone with an increased
 354 anisotropy of $\sim 1.8\%$ and a change in orientation of more
 355 than 10° with respect to the neighboring domains suggest
 356 subhorizontal, fault-parallel mantle flow within a vertical
 357 boundary layer that extends through the entire lithosphere.
 358 Interestingly, the width and position of this boundary layer
 359 in the upper mantle agree with the zone of relatively low
 360 resistivity obtained from the magnetotelluric studies (see
 361 section 2.2). The velocity anomaly below the crust, which is
 362 also slightly displaced toward the west relative to the
 363 surface trace of the AF (Figures 7d and 7i), agrees with
 364 the zone of anisotropic velocities detected in the SKS
 365 splitting studies (Figure 6). The apparent contradiction of
 366 substantial variation of seismic anisotropy beneath the DST
 367 and of a seismological homogeneous mantle under the DST
 368 found by Koulakov et al. [2006] is resolved by the obser-
 369 vation that even for high seismic anisotropy in such a

vertical zone the travel time residuals of subvertical P wave
 paths are close to zero [Sobolev et al., 1999]. Interestingly,
 all thermomechanical models of the DST have a similar
 20–30 km wide zone of shear flow located beneath the
 major fault that cut through the entire lithospheric mantle
 [Sobolev et al., 2005].

[13] Other estimates on the lateral extent of fault zone
 related deformation based on inferences of seismic anisotropy
 range from ~ 40 km for the Altyn Tagh fault in Tibet
 [Herquel et al., 1999] to 335 km for the Alpine fault in South
 Island, New Zealand [Baldock and Stern, 2005]. This
 indicates that the width of the decoupling zone increases
 with the total strain accumulated along the fault as suggested
 by the analysis of Baldock and Stern [2005]. However, for
 transtensional environments, e.g., within the India-Asian
 collision zone, some observations show a continuous rota-
 tion of SKS fast directions across the faults [McNamara et
 al., 1994], which may be indicative of neighboring regions
 of relatively abrupt changes in anisotropy [Rümpker and
 Ryberg, 2000]. In these cases the width of the vertical
 boundary layer may be reduced because of the more signifi-
 cant compressional component of deformation.

2.2. Crustal Scale

[14] Results of a simultaneous inversion of regional
 seismic phases for source location (not shown), Moho depth,
 and 3-D P and S velocity in crust and the uppermost mantle
 under the DST are given in Figures 7 and 8 [Koulakov and
 Sobolev, 2006]. The Moho depth increases strongly toward
 the east (Figure 8). Under Israel/Palestine the crust is similar
 to a thin continental margin, and under the Levant basin it is
 either the same or oceanic [Netzeband et al., 2006]. The
 narrow, NNE–SSW striking low-velocity P and S velocity
 anomaly identified in Figure 7 marks the position of the DST
 and is interpreted as sediments and a zone of fractured and
 deformed rocks in the middle and lower crust [Koulakov
 and Sobolev, 2006].

[15] In magnetotelluric (MT) images a 3–5 km wide,
 subvertical conductor (Figure 9, fault conductor 1 (FC1)) is
 detected in the crust and upper mantle west of the AF. It is
 spatially confined even in the ductile lower crust, but the
 bottom of the conductor is not resolved. Therefore, it might
 even penetrate the lithospheric upper mantle. The width and
 location of the conductor may correspond to the core of the
 lithospheric-scale shear zone which continues upper crustal
 faults down through the entire lithosphere [Sobolev et al.,
 2005] (see also section 2.5 for more discussion). This
 statement is in general agreement with the anisotropic
 vertical boundary layer in SKS splitting (Figure 6) and the

Figure 5. P velocity variation in the uppermost mantle under the DST based on tomographic inversion of 3366 P and PKP phases recorded at the 48 stations of the passive long-term deployment of DESERT (triangles in Figure 2b). The color bar gives the P velocity deviation in % (δV_p) from the reference model (IASP91 [Kennett and Engdahl, 1991]). (a–e) P velocity anomalies. (f–j) Synthetic model with description of anomaly parameters (amplitude and depth interval). Note the arrow indicating the offset of ~ 100 km in Figure 5g. (k–o) Synthetic P wave anomalies model using the synthetic model and the same ray configuration and parameters as in the data; noise with a root-mean-square (RMS) of 0.14 s is added. The synthetic model provides 35% of variance reduction and 0.22 s of data RMS, similar to those observed in the case of real data inversion. Modified from Koulakov et al. [2006], copyright 2006, Elsevier.

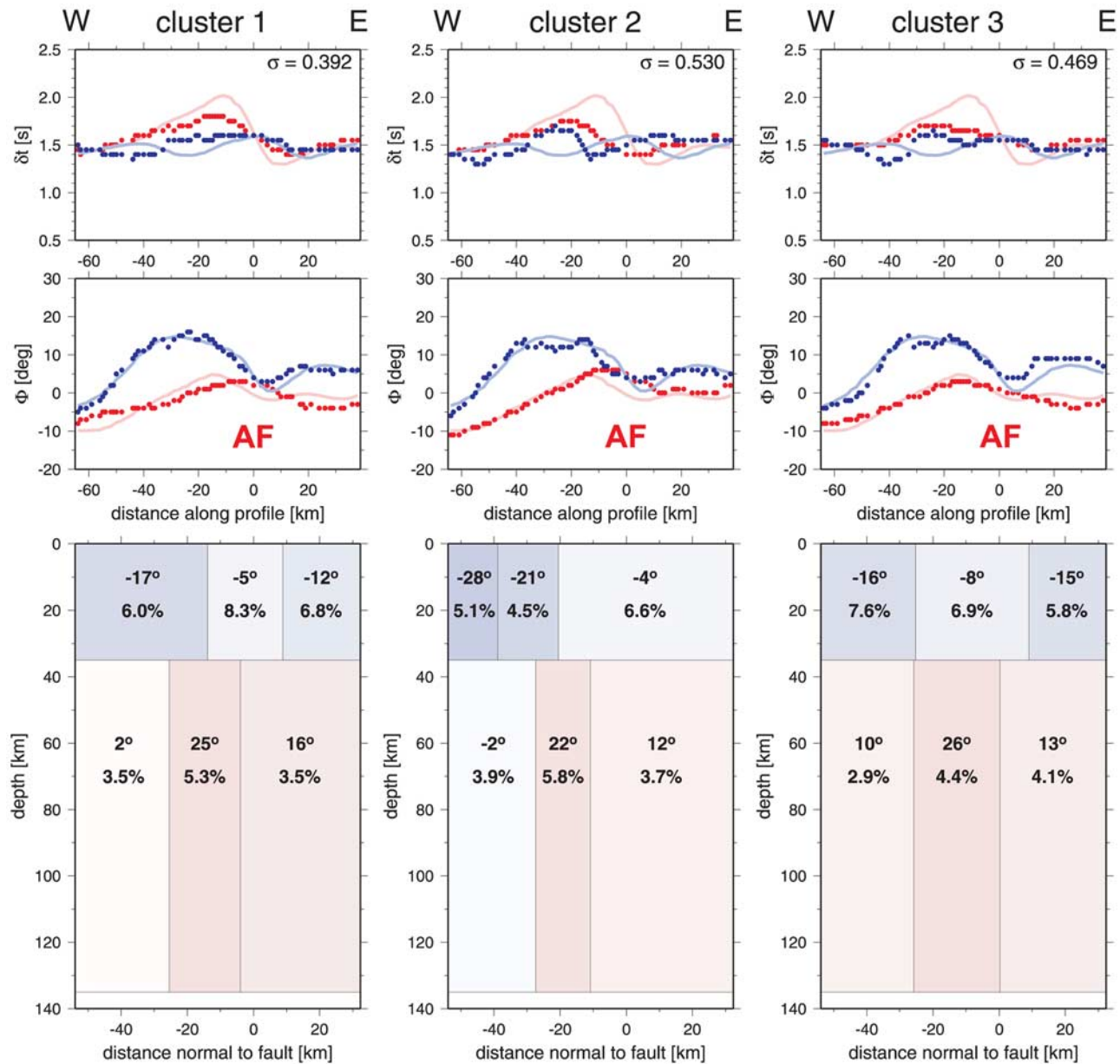


Figure 6. Model range of possible P wave velocity anisotropy in the crust and uppermost mantle under the DST on the 100 km long NVR profile (white line in Figure 2b) perpendicular to the AF (at 0 km). (bottom) Models for the three typical clusters of solution given by *Ryberg et al.* [2005]. The preferred model cluster is cluster 1 [see also *Rümpker et al.*, 2003]. The azimuth of the symmetry axis and the magnitude of the anisotropy are given in each block. (middle) Corresponding fast polarization direction Φ . Red corresponds to the period band 5–7 s, and blue corresponds to the results for the period band 2–5 s. Light red and light blue are the smoothed observed splitting parameters. (top) Corresponding delay time δt . All models are characterized by a central upper mantle zone of increased anisotropy, which is differently oriented with respect to its neighbors. Typical features of model cluster 1 are ≈ 20 km wide anisotropic zones in crust and upper mantle, model cluster 2 has typically narrower anisotropic zones, and model cluster 3 has broader crustal and upper mantle anisotropic zones. Modified from *Ryberg et al.* [2005].

418 region of low P and S velocity below the Moho (Figure 7),
 419 which also show anomalies displaced toward the west
 420 relative to the surface trace of the AF but do not have the
 421 lateral resolution of the MT experiment. The enhanced
 422 conductivity of FC1 could be explained by a pathway for
 423 fluids from the mantle or lower crust. Alternatively, this

region could have become permanently conductive by
 424 precipitation of conducting minerals such as graphite during
 425 the transform motion. The second large fault conductor
 426 appears at middle to lower crustal levels and could be
 427 related to the Al Quwayra Fault (AQF) (fault conductor 2
 428 (FC2) in Figure 9). Interestingly, FC2 appears to terminate
 429

Result of inversion of the ISC data in the Middle East region

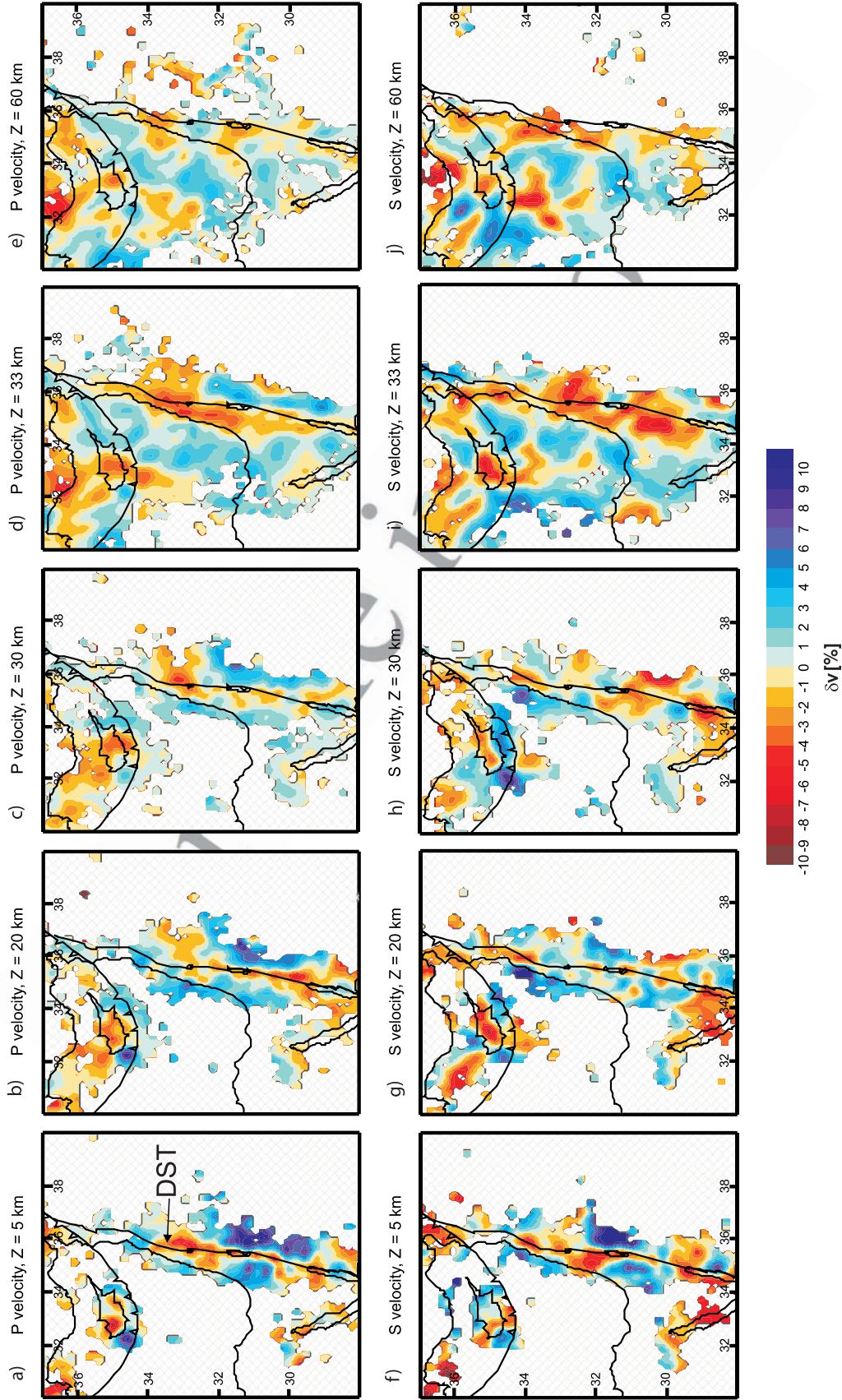


Figure 7. Crustal and uppermost mantle velocity anomalies (δv in % relative to AK135 [Kennett *et al.*, 1995]) in the Levant derived from regional earthquakes and seismic networks. (a–e) *P* velocities. (f–j) *S* velocities. The color bar gives the anomalies in percent. Modified from *Koulakov and Sobolev* [2006], copyright Blackwell Publishing.

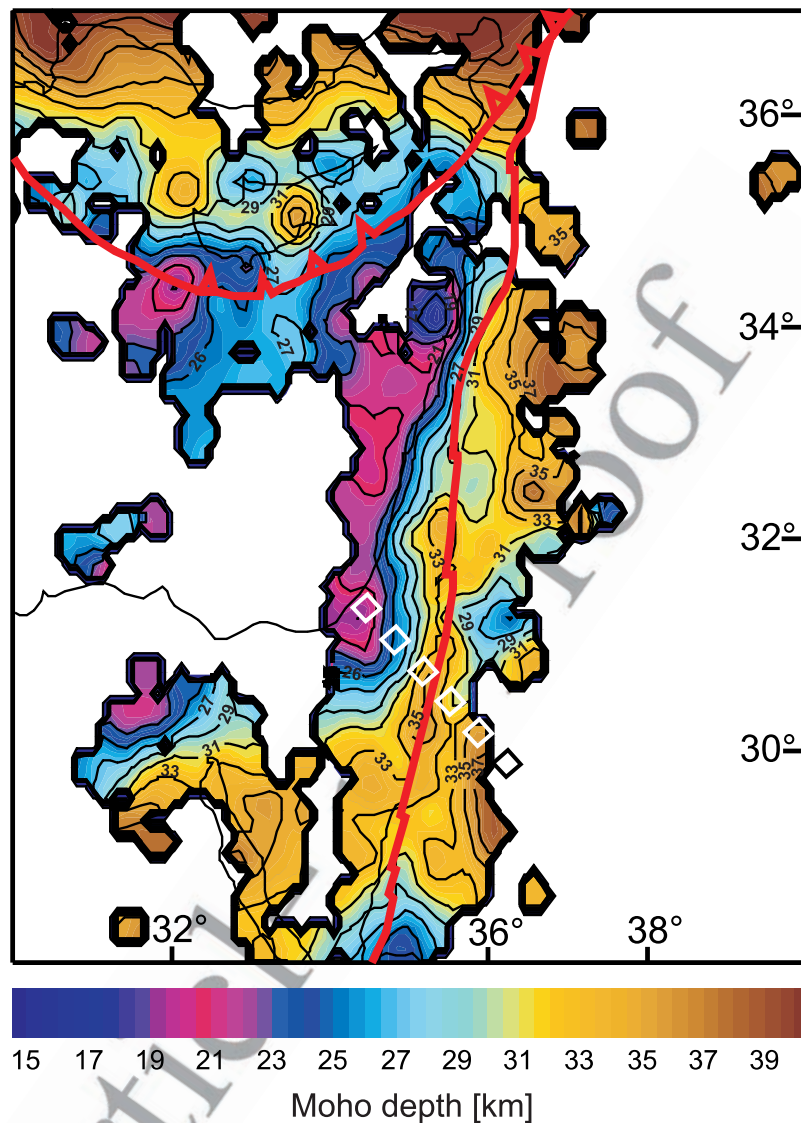


Figure 8. Crustal thickness in the eastern Mediterranean region derived from regional earthquakes and seismological networks in the Levant. The Moho depth was obtained simultaneously with the velocity model shown in Figure 7. The red lines are plate boundaries, the black lines are coast lines, and the white boxes indicate the DESERT profile. The color bar gives the depth of the Moho in kilometers. Modified from *Koulakov and Sobolev [2006]*, copyright Blackwell Publishing.

430 at the lower crustal reflector (LCR), discussed later in this
431 section in more detail.

432 [16] The results of a combined refraction/reflection sur-
433 vey from the Mediterranean to the Jordan highlands (black
434 and white line in Figure 2b) are shown in Figures 10 and
435 11c, respectively. The top two layers (Figure 10) consist
436 predominantly of sedimentary rocks with the seismic base-
437 ment below them offset by 3–5 km near the DST. Since the
438 wide-angle reflection/refraction (WRR) profile lies 30° off
439 the proposed symmetry axis derived from *S* wave (SKS)
440 splitting (Figure 6), the difference of the two *S* waves in the
441 controlled source WRR experiment (N–S and E–W) is too
442 small to be detected. The *P* and *S* wave velocity sections in
443 Figure 10 show a steady increase in the depth of the Moho
444 from ~26 km at the Mediterranean coast to ~39 km under
445 the Jordan highlands, with only a small but visible asym-

metric topography of the Moho under the DST. The lack of 446
447 significant uplift of the Moho under the DST, with the
448 Moho topography varying only ± 1 km, argues strongly
449 against a potential extensional rift structure in this area.
450 The *S* wave velocities (Figure 10b) and the corresponding
451 Poisson's ratios (Figure 10c) for the seismic basement can
452 be explained by felsic compositions typical for continental
453 upper crust, while those for the lower crust are representa-
454 tive of a mafic composition characteristic for continental
455 lower crust. These findings, of typical nonextended conti-
456 nental crust, a fact also corroborated by the gravity data
457 discussed later in this section, again speak strongly against
458 extensional processes in this area. The spatial coverage of
459 Moho topography is extended using the receiver function
460 method (RFM) for teleseismic data recorded at the stations
461 of the passive long-term deployment of DESERT (triangles

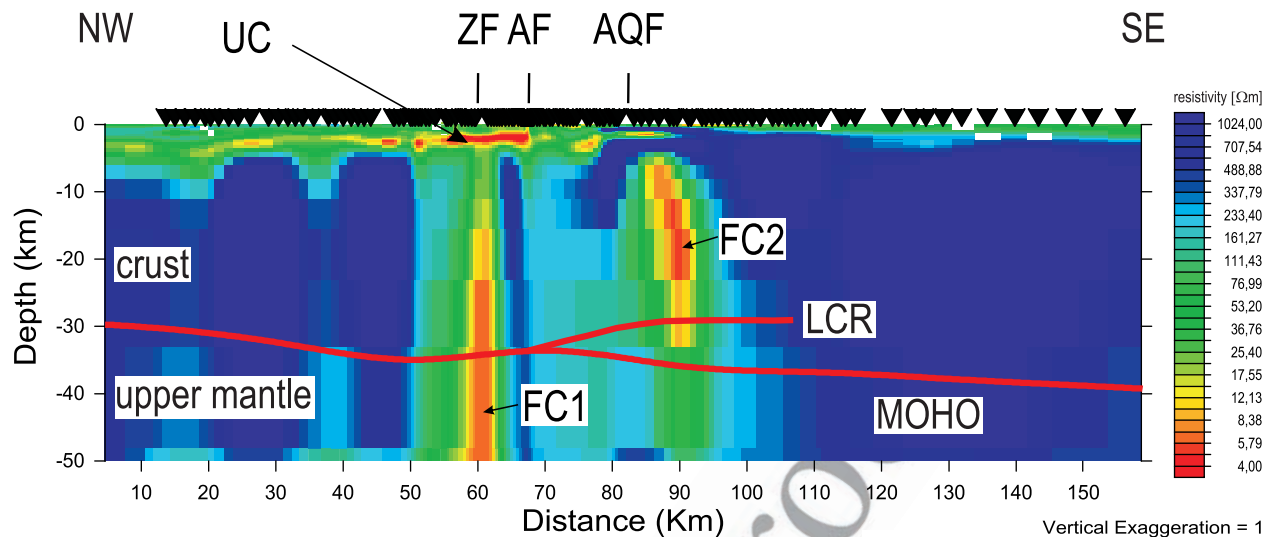


Figure 9. Regional-scale electrical resistivity cross section in Ω m along the magnetotelluric profile given as light blue dots in Figures 2b and 2d. Red and yellow indicate good conductors. The red lines are the Moho and the lower crustal reflector (LCR) determined from seismics and seismology; see Figures 10 and 11 and *DESERT Group* [2004], *Mechie et al.* [2005], and *Mohsen et al.* [2005]. UC is an upper crustal conductor between ~ 1.5 and 4 km depth, truncated toward the east at about the AF (for more details, see Figure 15 and discussion there). The locations of the ZF, the AF, and the AQF are indicated for orientation. These faults are associated with deep subvertical zones of high conductivity (fault conductor 1 (FC1) and fault conductor 2 (FC2)). Note also the corresponding seismic velocity anomalies in the lower crust and upper mantle west of the AF in Figure 6. The FC2 anomaly, associated with the AQF, appears to terminate at the LCR. No vertical exaggeration is applied. Modified from *Ritter et al.* [2005b].

462 in Figure 2b) and four permanent stations of the Israel and
 463 Jordan seismic networks [*Mohsen et al.*, 2005]. The Moho
 464 depth values under these stations are shown in Figure 11a.
 465 The depth of the Moho across the DST increases smoothly
 466 from about 30 km through 34–38 km toward the east, in
 467 agreement with the WRR data, but also shows significant
 468 north-south variations east of the DST, in agreement with
 469 gravity data discussed later in this section. The depth of the
 470 discontinuity associated with the LCR (see previous para-
 471 graph) and identified east of the DST (RFM study of
 472 *Mohsen et al.* [2005]), is shown in Figure 11b. Figure 11c
 473 shows the excellent agreement of the depth determination of
 474 the Moho and the LCR along the DESERT profile between
 475 the steep-angle (line drawing) and wide-angle (red band)
 476 controlled source techniques and the passive data RFM
 477 results (green and yellow symbols, respectively). The
 478 LCR is interpreted as horizontal north-south oriented shearing
 479 at this plate boundary, a process occurring also in the thermo-
 480 mechanical simulation of *Sobolev et al.* [2005] (see, e.g.,
 481 Figures 24d–24f). On the other hand, the LCR in the NVR
 482 profile (line drawings in Figure 11c) and the corresponding
 483 intermittent reflections in the WRR P wave data mainly east of
 484 the DST, from ~ 30 km depth, could also be from mafic
 485 intrusions associated with the nearby Cenozoic volcanism
 486 [*DESERT Group*, 2004]. The RFM data and the controlled
 487 source data can both be satisfied if the LCR comprises a small
 488 first-order discontinuity underlain by a ~ 2 km thick transition
 489 zone [*Mechie et al.*, 2005].

[17] Further constraints come from 3-D density modeling 490
 by *Götze et al.* [2006]. The 3-D density model is based on 491
 the newly compiled Bouguer anomaly for the area of the 492
 DST (Figure 12, top). Three cross sections across the DST 493
 are displayed in Figure 13. The density model along a profile 494
 coinciding with the WRR and NVR profiles (Figure 13b) 495
 matches well with the structure from the seismic data (black 496
 line in Figure 2b). The most negative Bouguer anomalies 497
 along the DST are mainly caused by the deep low-density 498
 sedimentary basins (most prominently the Dead Sea 499
 basin with sediment thicknesses of 8–12 km [*Garfunkel* 500
and Ben-Avraham, 1996, 2001; *Ginzburg and Ben-Avraham*, 501
 1997; *Ben-Avraham and Lazar*, 2006; *Ben-Avraham and* 502
Schubert, 2006; *Ben-Avraham et al.*, 2008]), whereas a 503
 shallow zone of high-density intrusion at 31°N , 36°E 504
 coincides with the local maximum gravity on the eastern 505
 flank of the DST. However, the Bouguer anomalies are, in 506
 general, characterized by higher values of 0 to -20 mGal in 507
 the NW and lower values of -60 to -80 mGal in the SE of 508
 the region (Figure 12, top). This trend reflects the gradual 509
 thickening of the crust from the Mediterranean toward the 510
 Jordan highlands found with the RFM (Figure 11). The 511
 gravity-derived Moho depth matches the RFM values. 512
 Therefore, the western region, near the Mediterranean, is 513
 characterized by the thinnest crust, whereas the Jordan 514
 highlands in the southeast are characterized by the thickest 515
 crust of 38–42 km (Figure 12, bottom). The thickness (up 516

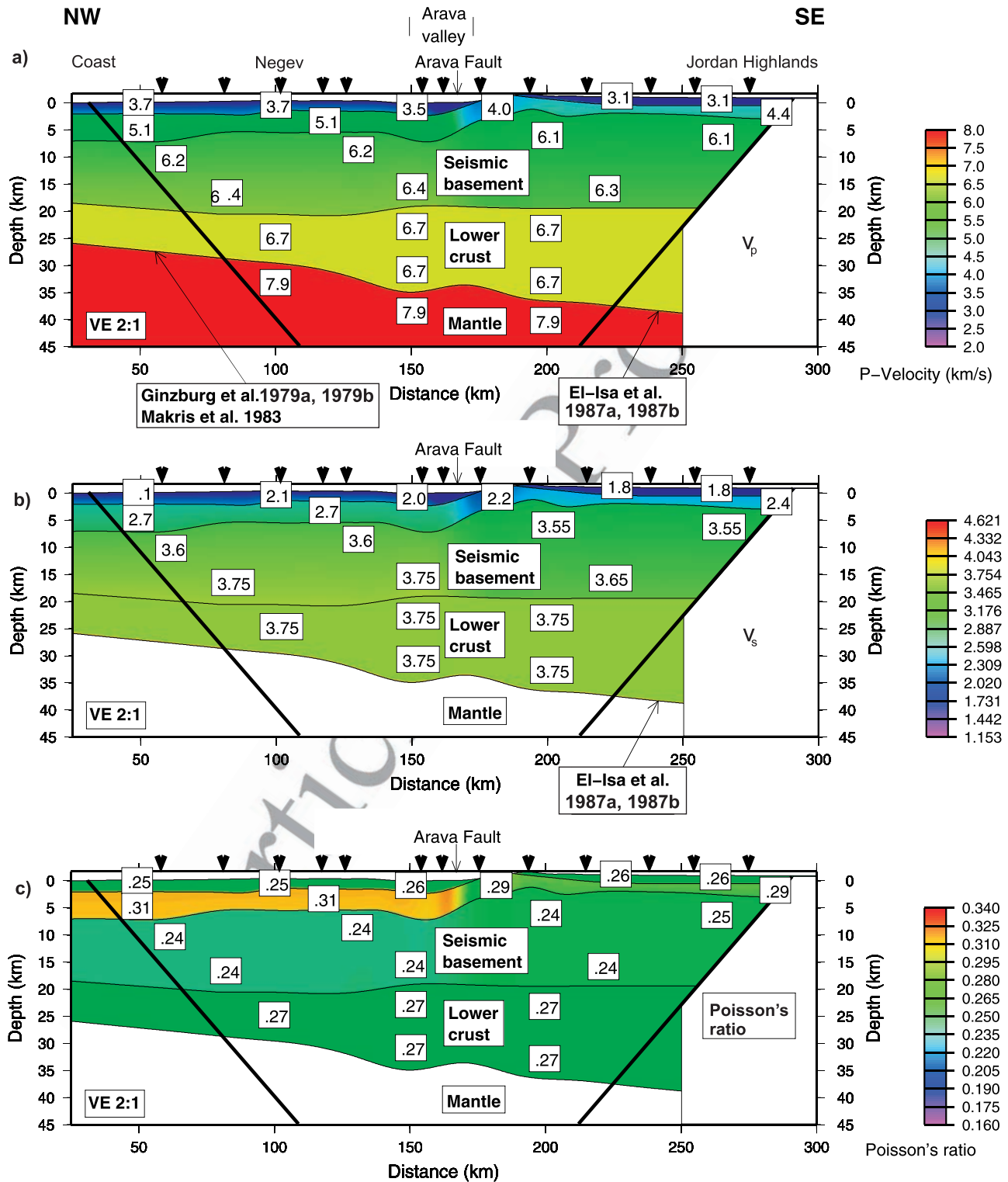


Figure 10. Crustal P and S velocity structure under the 260 km long WRR profile (black line in Figure 2b) from the Mediterranean coast (NW) to the Jordan highlands (SE). (a) P velocity model [after *DESERT Group*, 2004]. (b) S velocity model. (c) Corresponding Poisson's ratio model. Velocities are in km/s. Triangles at the top of each section represent the shot points. Only the region within the diagonal lines is resolved in this study. To the NW the boundaries and P wave velocities are based on previous work by Ginzburg et al. [1979a, 1979b] and Makris et al. [1983], while to the SE the boundaries and the P and S wave velocities are based on El-Isa et al. [1987a, 1987b]. Note vertical exaggeration of 2:1. Modified from Mechie et al. [2005], copyright Blackwell Publishing.

517 to 38–42 km) and density of the crust confirm again that the
518 region of the DST is underlain by continental crust.
519 [18] In summary, integration of fundamental Rayleigh
520 waves, body wave tomography, RFM, and shear wave splitting

with *P* and *S* waves from controlled sources, combined with
521 MT studies and 3-D gravity data, allows us to derive an image
522 of the AF on the crustal scale (Figure 3a). The combination of
523 all the findings above suggests localized deformation and
524

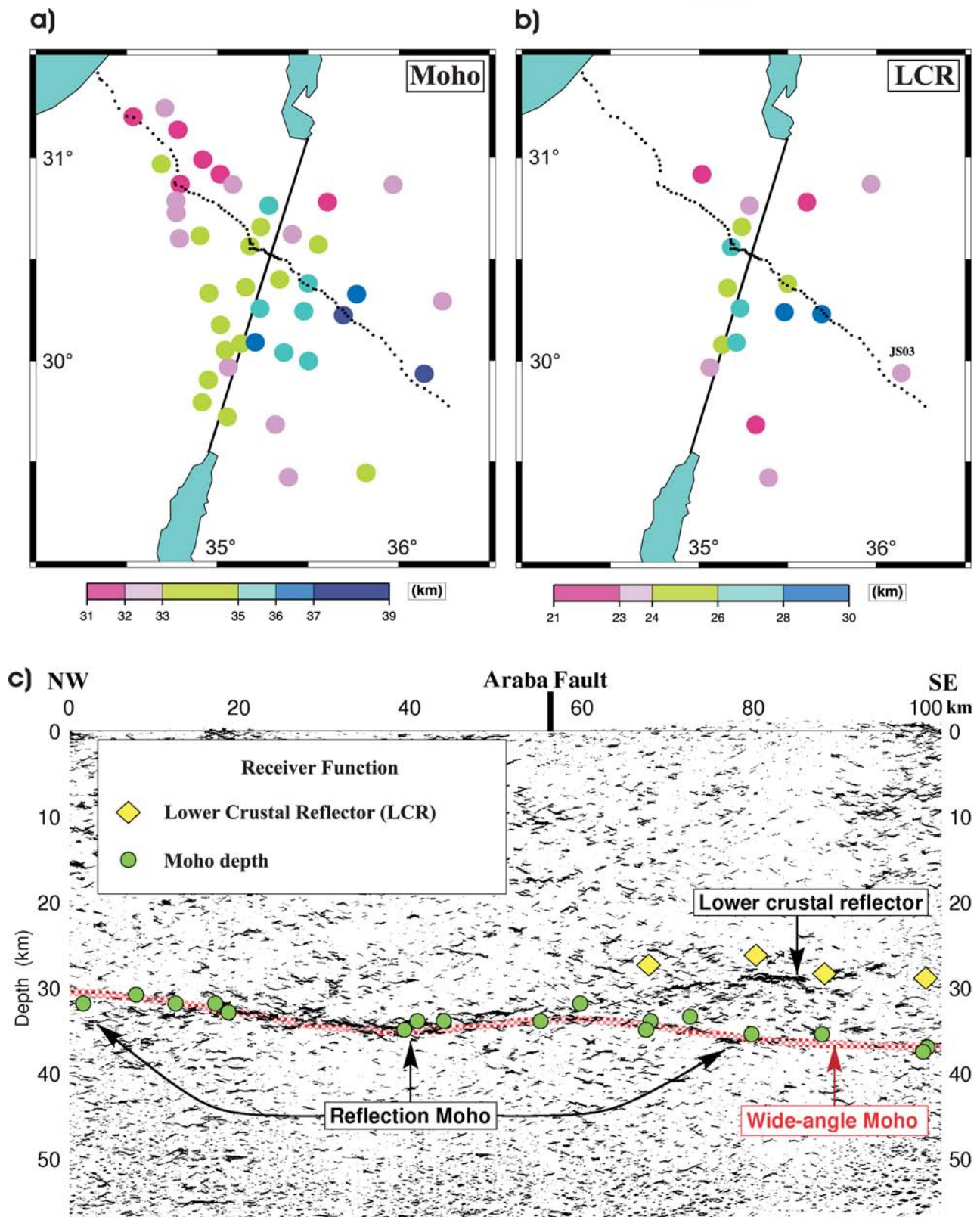


Figure 11

525 subhorizontal mantle flow within a narrow vertical boundary
 526 layer that extends through the entire lithosphere. We con-
 527 clude that rift-type deformation did not play a dominant role
 528 in the dynamics of the DST, a fact supported by the results
 529 of *Sobolev et al.* [2005]. All observations can be linked to
 530 the left-lateral movement of the two plates, accompanied by
 531 strong deformation within a narrow zone of minimum
 532 lithospheric strength cutting through the entire crust, con-
 533 sisting most likely of fault-parallel mineral alignment be-
 534 neath the AF that also gives rise to significant seismic
 535 anisotropy.

536 2.3. Kilometer to Meter Scale

537 [19] The elongated structure of the rift-like DST is
 538 composed of a series of large basins (the Dead Sea being
 539 the most prominent one), which are mainly attributed to
 540 dilatational jogs related to the left-lateral strike-slip motion.
 541 The formation of the Gulf of Aqaba/Elat had also been
 542 associated with pull-apart structures [*Garfunkel*, 1981; *Ben-*
 543 *Avraham*, 1985; *Garfunkel and Ben-Avraham*, 2001; *Ben-*
 544 *Avraham et al.*, 2008]. Along the AF the “rift” valley is
 545 further divided into a series of smaller subbasins [*ten Brink*
 546 *et al.*, 1999a]. Our study area is situated at the southern
 547 termination of the Dead Sea Basin. The fault-bounded
 548 structure of this depression had previously been revealed
 549 by commercial multichannel seismic reflection and borehole
 550 data, which suggested an asymmetric, graben geometry of
 551 flat subhorizontal strata of Miocene to recent sediment fill
 552 that becomes shallower toward the south [*ten Brink and*
 553 *Ben-Avraham*, 1989; *Garfunkel*, 1997; *Gardosh et al.*, 1997;
 554 *Zak and Freund*, 1981; *Neev and Hall*, 1979; *Rotstein et al.*,
 555 1991; *Ginzburg and Kashai*, 1981; *Kashai and Croker*,
 556 1987]. The Miocene Hazeva formation is considered the
 557 oldest basin fill unit, overlain by massive evaporite beds and
 558 sequences of marl, clay, sand, and gravel [*Gardosh et al.*,
 559 1997; *Zak*, 1967; *Garfunkel*, 1997]. A number of normal,
 560 often listric, faults dominate the structure in the southern
 561 Dead Sea Basin [*Gardosh et al.*, 1997]. The western side of
 562 our study area was previously analyzed by high-resolution
 563 seismic profiles [*Bartov et al.*, 1998; *Frieslander*, 2000].
 564 Several prominent faults cut the western half of the Arava/
 565 Araba Valley, with the subvertical Zofar Fault (ZF, Figure 2)
 566 being the most prominent. In the uppermost kilometer west
 567 of the ZF, Cretaceous and Permian rocks are found, while
 568 younger strata related to the Dead Sea Basin are exposed
 569 east of it [e.g., *Ritter et al.*, 2003]. Toward the center of the
 570 Arava/Araba Valley, additional north-south striking faults

were found at this longitude, often joining at depth forming
 negative flower structures [*Frieslander*, 2000].

[20] To study the uppermost crust in the vicinity of the
 AF in more detail, tomographic inversion techniques were
 applied to over 280,000 travel time picks of refracted *P*
 waves along the NVR line (red line in Figure 14a) [*Ryberg*
et al., 2007]. The *P* velocity structure is well resolved down
 to a depth of several kilometers (Figure 14b). It shows
 features that correlate well with surface geology, calibrated
 by boreholes in the vicinity of the DST [*Fleischer and*
Varshavsky, 2002; *Bartov et al.*, 1998; *Frieslander*, 2000].
 The tomographic *P* velocity model (Figure 14b) is character-
 ized by a thin layer (few 100 m) of low velocities (<2.6 km/s)
 in the depression containing the AF (range 40–60 km).
 Geology (Figure 14a) confirms that Quaternary alluvium
 (unconsolidated) is thin throughout this depression, which
 is underlain by consolidated sediments ranging in age from
 Jurassic to Miocene. Basement velocities (>5.0 km/s),
 interpreted as Precambrian igneous and metamorphic rocks
 (Figure 14c), are found at a depth of ~2 km below sea level
 west of the AF and 0–1 km below sea level east of the AF.
 Strong lateral velocity gradients (steps) are located at the
 following mapped faults, in order of step size: AF, Al
 Quwayra, Zofar, and Barak. The high-velocity block on
 the east side of the AF is interpreted as a high-standing
 sliver of the crystalline basement. The lithological contrast
 to the sediment basin west of it could create a hydrological
 barrier; see Figure 15 and *Ritter et al.* [2003], *Maercklin et*
al. [2005], and *Bedrosian et al.* [2007] for details.

[21] Magnetotelluric and seismic models [*Ritter et al.*,
 2003; *Maercklin et al.*, 2005; *Ryberg et al.*, 2007] provide
 complementary information about the resistivity and veloc-
 ity structure of the subsurface (Figures 15a and 15b),
 respectively. Using a new classification approach these
 independently derived models can be combined to map
 the subsurface structure where regions of high correlation
 (classes) provide structural and lithological information not
 always evident in the individual models [*Bedrosian et al.*,
 2007]. This method was applied to a 10 km long profile
 across the AF (seismics (central part of the NVR profile in
 Figure 2c) and coinciding magnetotellurics (Figure 2d)). Six
 prominent lithological classes are identified providing a
 clear delineation of stratigraphy in accordance with geologic
 results (Figure 15c and Table 1). All classes except 1 and 6
 are truncated at the AF. Classes 1 and 2 show typical values
 of igneous or metamorphic rocks (high-velocity sliver in
 Figure 14b), classes 5 and 6 can be attributed to sedimentary

Figure 11. Depth of Moho and the LCR in the vicinity of the DST combined with the 2-D reflection seismic image across the DST. (a) Map of Moho depth from *Mohsen et al.* [2005] derived using the receiver function method (RFM) at the stations of the passive long-term DESERT deployment (triangles in Figure 2b). The dotted line indicates the WRR profile. The Moho is deepest near the southeastern end of the DESERT profile and shallowest at the northwest. (b) Map of LCR (see also Figure 11c) occurring mainly east of the DST. (c) Automatic line drawing of the depth-migrated seismic common depth point section of the NVR experiment [after *DESERT Group*, 2004]. The black arrows mark the drop-off of reflectivity, generally interpreted as the Moho in seismic reflection data. The red band indicates the location of the Moho as derived from the WRR experiment in Figure 10a. The green and yellow symbols indicate the Moho and the LCR as determined with the RFM in Figures 11a and 11b, respectively, in a ~20 km wide corridor along the DESERT profile. Modified from *DESERT Group* [2004] and *Mohsen et al.* [2005] (copyright Blackwell Publishing), respectively.

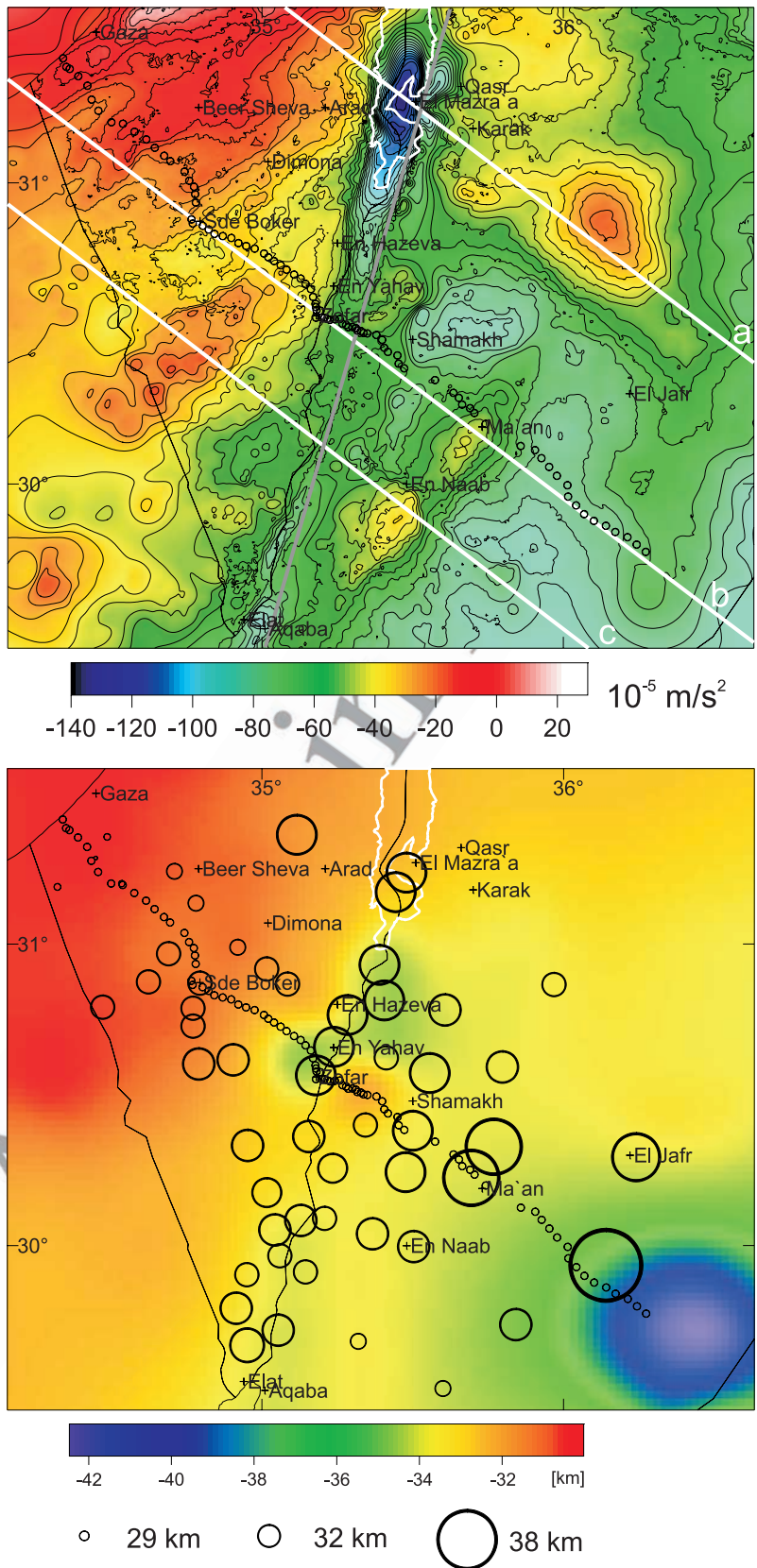


Figure 12. Bouguer anomaly and the gravity-derived Moho depth in the vicinity of the DST. (top) Measured (complete) Bouguer anomaly. The three white lines, subparallel to the seismic profile, mark the positions of the profile view cross sections shown in Figure 13. The WRR profile is indicated by small black circles. (bottom) Moho depth beneath the 3-D gravity modeling area. Also shown is the Moho depth (open circles) from the RFM [Mohsen *et al.*, 2005]. Modified from Götze *et al.* [2006] with kind permission of Springer Science and Business Media.

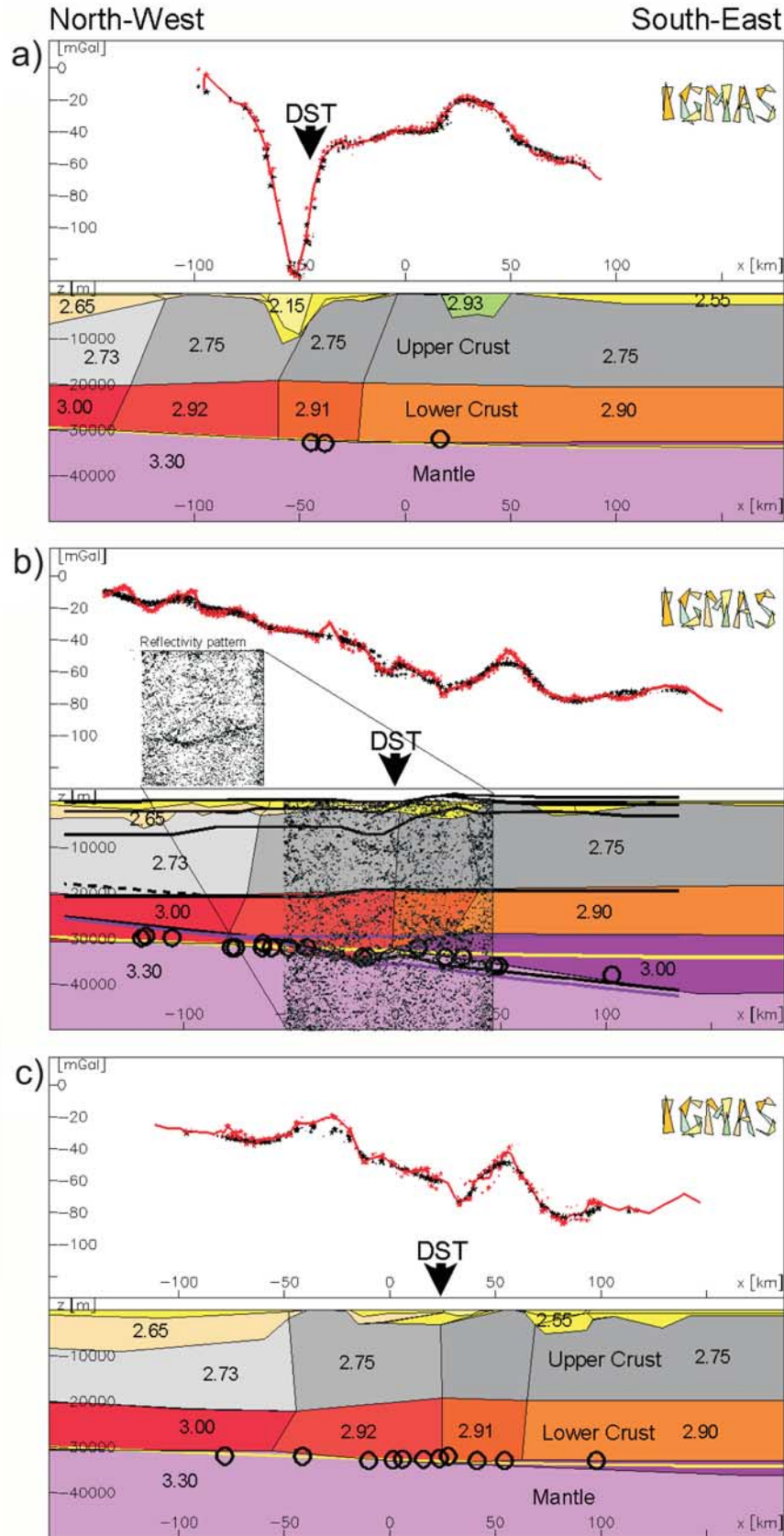


Figure 13

618 rocks, and class 4 shows typical values of clastic or
 619 consolidated sediment at sufficient depth to have undergone
 620 compaction. Class 3 shows moderate seismic velocities,
 621 similar to class 4, but extremely low resistivity, lower even
 622 than that found in the surface sediments. The most plausible
 623 explanation for this class is the presence of a few percent of
 624 hypersaline fluid [Ritter et al., 2003; Maercklin et al., 2005;
 625 Bedrosian et al., 2007]. This is consistent with the obser-
 626 vation of saline brines within the Zofar 20 well a few
 627 kilometers west of the study area. Class 3 terminates along a
 628 vertical plane ~ 500 m east of the surface trace of the AF.
 629 For a more detailed discussion of this buried fault (fossil AF
 630 (FAF)), see Maercklin et al. [2004, 2005] and Kesten et al.
 631 [2008]. The change in lithology across the AF, from classes
 632 3, 4, and 5 in the west to class 2 in the east, suggests that the
 633 fault is, at most, a few hundred meters wide, at a depth of
 634 0.5–3.0 km below the surface. The vertical uplift of the
 635 basement is at least 1.3 km, which is in accordance with
 636 gravity modeling [Tašárová et al., 2006].

637 [22] A 3-D image of the electrical conductivity across and
 638 along the AF (Figure 16) reveals the continuation of the
 639 upper crustal brine zone toward the north. The AF seems to
 640 act primarily as an impermeable barrier to cross-fault fluid
 641 transport. However, the conductivity images do not reveal a
 642 pronounced fault zone conductor associated with the dam-
 643 age zone of the fault as it has been observed, e.g., at the
 644 SAF [Unsworth et al., 1999, 2000; Bedrosian et al., 2002;
 645 Becken et al., 2008] and the West Fault in Chile [Hoffmann-
 646 Rothe et al., 2004].

647 [23] A 3-D image of seismic contrasts along a 7 km long
 648 stretch of the central AF between 1 and 4 km depth (Figure 17)
 649 is derived from arrays of controlled seismic sources and
 650 receivers, respectively (controlled source arrays (CSA) I
 651 experiment; for details, see Maercklin et al. [2004]). The
 652 well-resolved 7 km wide area (from ~ 3 to 10 km in the y
 653 direction in Figure 17) begins ~ 1 km north of line 1 and
 654 ends ~ 1 km south of line 3 (Figure 2f). Using a 3-D seismic
 655 migration method based on beam forming and coherency
 656 analysis of P -to- P scattered waves, a subvertical reflector
 657 (red color, FAF in Figure 17) offset roughly 1 km toward the
 658 east of the surface (active) AF can be imaged. The dashed
 659 line gives the downward projection of the AF (Figure 17a).
 660 Resolution tests show that the accuracy of the location of
 661 the reflector (FAF) is ~ 200 m in the E–W direction.
 662 Because of the configuration of source and receiver arrays
 663 the extent of this scatterer could not be resolved above 1 km
 664 and below 4 km depth and also not farther toward the south,
 665 thus not directly encompassing line 1 of the CSA I exper-
 666 iment and the NVR profile (Figure 2f). Integration of the
 667 results of this independent data set based on scattered

seismic waves with the results of Ritter et al. [2003], 668
 Maercklin et al. [2005], Ryberg et al. [2007], and Bedrosian 669
 et al. [2007] suggests that the reflector imaged here is the 670
 northward extension of the western boundary of the high- 671
 velocity sliver of crystalline basement (Figures 14b and 15c, 672
 respectively). 673

[24] A detailed local 3-D density model covers an area of 674
 $\sim 30 \times 30$ km down to a depth of ~ 5 km, partly shown in 675
 Figure 18. That model was computed from a newly com- 676
 piled Bouguer gravity anomaly database corrected for 677
 effects of regional structures such as the Moho [Götze et 678
 al., 2006]. The 3-D structural image of the upper crust 679
 reveals that the basement is vertically offset across the AF 680
 (Figures 18b–18d). It also shows an abrupt change in the 681
 physical parameters of the two lithological blocks, i.e., of 682
 the sediments in the west and the sliver of crystalline 683
 basement in the east. Analysis of the calculated gravity 684
 gradients (not shown) furthermore suggests that the AF 685
 could be offset at depth as shown in Figure 17. It should 686
 also be noted that such a shift in the location of the active 687
 transform was suggested by Joffe and Garfunkel [1987] to 688
 be the result of a shift of $\leq 5^\circ$ in the local direction of plate 689
 motion which increased transtension some 5 Ma ago or 690
 somewhat earlier, whereas ten Brink et al. [1999a] sug- 691
 gested continuous small variations in plate motion as the 692
 cause of this shift. Note also that Rotstein et al. [1992] 693
 found evidence for a similar 2.5 km shift of the active fault 694
 trace in the area south of the Sea of Galilee. 695

[25] The 3-D image presented so far is also supported by 696
 the combination and interpretations of geology and seismic 697
 reflection studies (Figure 19) with multispectral (ASTER) 698
 satellite images (Figure 20) [Kesten et al., 2008]. Such a 699
 combination allows us to analyze geologic structures in 700
 space and time since reflection seismics image deep faults 701
 possibly inactive at present, whereas satellite images reveal 702
 neotectonic activity in shallow young sediments. 703

[26] Starting in the west, we will now discuss distribution 704
 of slip along the central part of the NVR profile, i.e., of 705
 faults in the Arava/Araba Valley and its immediate vicinity. 706
 West of the AF, between profile kilometers 41 and 52, 707
 strong sedimentary reflections and indications of faults are 708
 visible in the NVR data (Figure 19c). Here, the Barak Fault 709
 and the ZF (Figures 19a and 20) are clearly imaged, as 710
 previously by Bartov et al. [1998] and Frieslander [2000]. 711
 Whereas the vertical offset at the Barak Fault seems to 712
 be minor, the Zofar Fault shows a vertical separation of 713
 ~ 500 m and has been interpreted as a major fault in the 714
 northern Arava/Araba Valley, where it is the western border 715
 fault of the southern Dead Sea Basin [Bartov et al., 1998; 716
 Frieslander, 2000]. On the basis of the general tectonic 717

Figure 13. (a–c) Three parallel vertical cross sections through the 3-D density model (for location of the profiles in Figures 13a–13c, see Figure 12). The distance between the cross sections is ~ 60 km. The density of the various geologic units is given in Mg/m^3 . The black circles indicate the Moho depth from the RFM (Figure 11a) [Mohsen et al., 2005]. The black overlay is the NVR line drawing (Figure 11c), and the black lines indicate the interfaces of the seismic refraction model (Figure 10a). At the top of each section, the line of black dots indicates the modeled gravity, and the red line gives the measured gravity along the vertical cross section. Modified from Götze et al. [2006] with kind permission of Springer Science and Business Media.

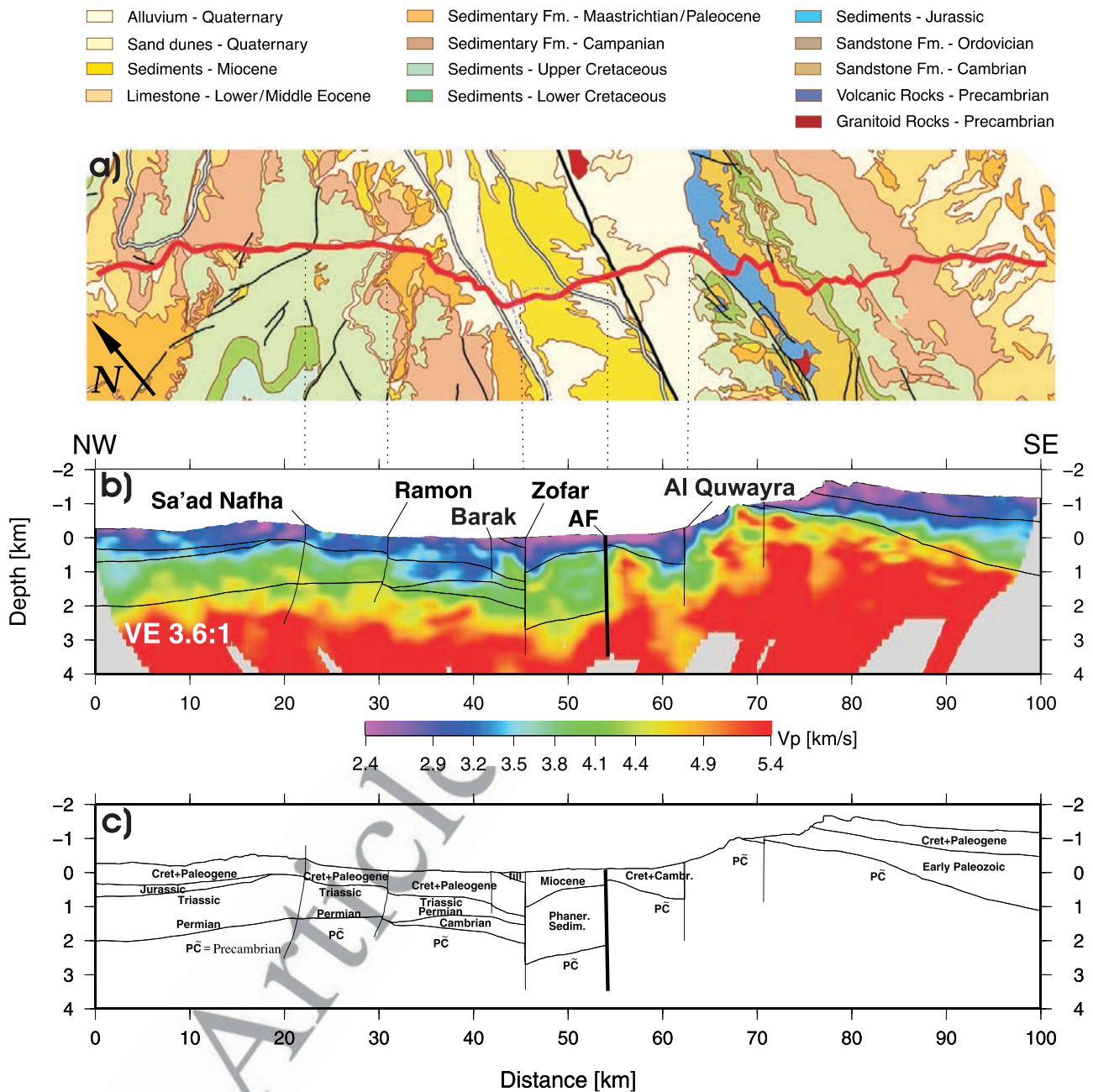


Figure 14. Geology, P velocity cross section, and geologic interpretation along the 100 km long NVR profile (white line in Figure 2b) centered on the AF. (a) Simplified geological map along the NVR profile (modified after *Sneh et al.* [1998] and *Bartov et al.* [1998]). The red line is the slightly smoothed reflection line of the NVR experiment. Six dominant faults (Sa'ad Nafha, Ramon, Barak, Zofar, AF, and Al Quwayra) are indicated by black lines (see also Figures 14b and 14c). Here the nomenclature of *Frieslander* [2000] and *Calvo and Bartov* [2001] is used. (b) Tomographic P velocity model along the NVR profile to a depth of 4 km using vibroseis and shot data [*Ryberg et al.*, 2007]. The vertical exaggeration is 3.6. The velocity model is characterized by strong horizontal gradients (e.g., at 20, 45, 55, and 62 km model distance, corresponding to the Sa'ad Nafha, Zofar, AF, and Al Quwayra faults, respectively). Several near-surface sedimentary basins (blue; low-velocity regions) bordered by faults can be seen. (c) Geologic cross section interpreted from the NVR profile and well data, to a depth of 4 km [after *DESERT Group*, 2004]. The vertical exaggeration is 3.6 as in Figure 14b. For more details of this area, see Figure 19 and *Kesten et al.* [2008]. Modified from *Ryberg et al.* [2007].

718 setting, a sinistral strike-slip component has been assumed,
 719 even if the total amount of lateral displacement is unknown
 720 (Z. Garfunkel, personal communication, 2008), but left-
 721 lateral strike-slip creep episodes along the ZF were indicated

by InSAR data [*Finzi*, 2004]. An antithetic fault, east of the 722
 Zofar Fault at profile km 47 called “Eastern Fault” by 723
Frieslander [2000], shows no surface expression and hardly 724
 any vertical displacement but is linked to the Zofar Fault at 725

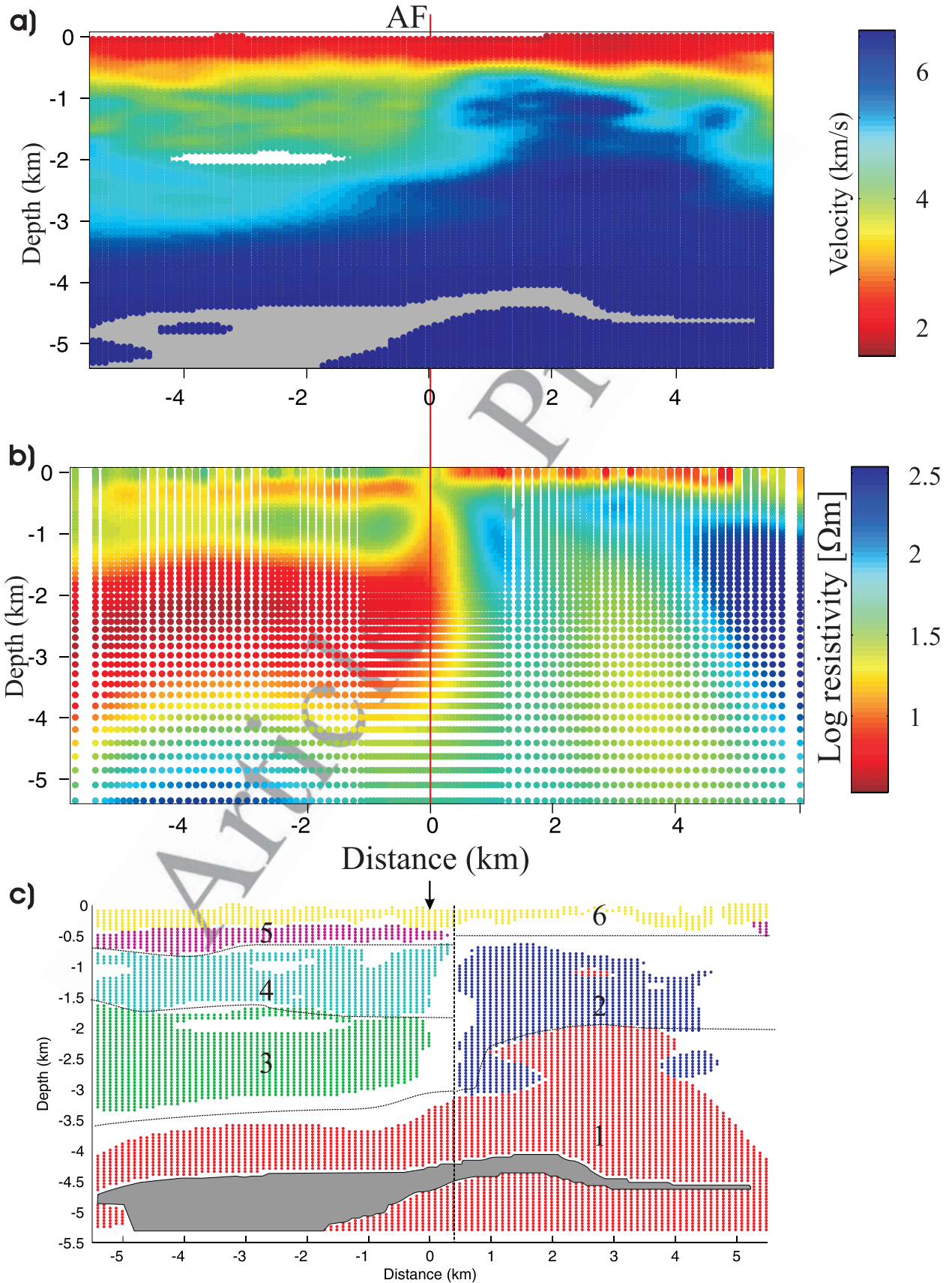


Figure 15

726 ~2.5 km depth (Figure 19c). The fault at profile kilometer 49
 727 does not seem to cut through the uppermost sediments and is
 728 not recognized at the surface. Thus, we assume it has not
 729 been active recently. There is, however, a marked contrast in
 730 reflectivity across this fault, indicative of previous strike-slip
 731 movement. The fault's near-vertical geometry corroborates
 732 this interpretation. A subsurface fault (Western Fault in
 733 Figures 2c and 19) is revealed ~4 km west of the AF in
 734 the industry reflection seismic profile VWJ-9 (Figure 19b)
 735 and the NVR profile (Figure 19c). Whereas seismic data in
 736 Figure 19b show flower structures typical for strike-slip
 737 faults, the satellite image does not reveal the Western Fault
 738 in post-Miocene sediments. Taking into account the age
 739 estimates from Table 2 and the fact that these sediments
 740 overlie the fault trace, the period of inactivity of the "old"
 741 strike-slip fault lies between 2 and 7 Ma. For a more detailed
 742 discussion, see *Kesten et al.* [2008]. The lack of coherent
 743 sedimentary reflections in the direct vicinity of the AF might
 744 result from intense brittle deformation of the rocks close to
 745 the fault but could also be caused by the absorption of high
 746 frequencies in this area covered by sand dunes and alluvium
 747 (Figure 19a). There is, however, a slight change in the
 748 character of (diffuse) reflectivity across the subsurface
 749 continuation of the AF (Figure 19c). Farther to the north
 750 (Figure 19b), there is strong indirect evidence of the AF in
 751 the seismic profile VWJ-9. Here the AF lies between a zone
 752 with strong sedimentary reflections to the west and a purely
 753 diffusely reflective upper crust toward the east. The AF itself
 754 is imaged as a nonreflective zone of ~800 m width at the
 755 surface, probably consisting of three branches (Figure 19b).
 756 The central one was also delineated by fault zone guided
 757 waves [*Haberland et al.*, 2003] and high-resolution reflection
 758 seismics and tomography to be discussed in section 2.4.
 759 Farther to the south of the study area, similar subparallel
 760 fault segments and flower structures related to the DST are
 761 also known from the Evrona playa site [e.g., *Shtivelman et al.*,
 762 1998]. About 1 km east of the AF, another, now buried
 763 fault called FAF (Figures 2c and 2f) is detected, in agreement
 764 with the FAF scatterers in Figure 17 and the joint interpreta-
 765 tion of seismic waves with magnetotellurics [*Ritter et al.*,
 766 2003; *Maercklin et al.*, 2005; *Bedrosian et al.*, 2007]; see
 767 also Figure 14 for the vicinity of profile km 57. The AQF
 768 (Figures 19 and 20) [see also *Rabb'a*, 1991] is the most
 769 prominent fault in SW Jordan. This N–S running fault is
 770 more than 100 km long and extends into Saudi Arabia. In an
 771 outcrop of the AQF on the NVR profile the Upper Creta-
 772 ceous sediments dip nearly vertically and show intense,

subhorizontal slickensides, indicating sinistral movement. 773
Abu Taimah [1988] proposed a left-lateral displacement of 774
 8 km for the AQF based on offset biotite muscovite aplite 775
 granites around latitude 30.3°N and on an 8 km long rhomb- 776
 shaped graben in this region. This argument seems to be 777
 supported by the geological map by *Bender* [1974] and 778
Ibrahim [1991]. *Barjous* [1988], on the other hand, sug- 779
 gested an even larger horizontal slip in the range of ~40 km. 780
 His assumption is mainly founded on the southernmost 781
 outcrops of Precambrian rhyolites that occur ~40 km farther 782
 to the NE of the AQF. This higher value is also supported by 783
 stratigraphic observations (for details, see *Kesten et al.* 784
 [2008]). The distribution of slip on the faults in the vicinity 785
 of the AF, especially the AQF, led *Kesten et al.* [2008] to 786
 propose that the AF is clearly the main active fault segment 787
 of the southern DST. Despite this dominant role we postulate 788
 that the AF has accommodated only a limited part (at least 789
 15 km and up to 60 km) of the overall 105 km of sinistral 790
 plate motion since the Miocene. 791

[27] In summary, integrating the results of these studies 792
 leads to the structural image of the crust in the vicinity of the 793
 AF given in Figure 3b. Only the integration and cross 794
 validation of several seismic methods (2-D and 3-D tomog- 795
 raphy, scattering mapping, and reflection seismics from 796
 controlled-source experiments), 2-D and 3-D MT experi- 797
 ments, 3-D gravity models, the analysis of multispectral 798
 satellite images, geology, and novel mathematical approaches 799
 (joint classification) allow the imaging of the 3-D structure 800
 under the AF on crustal to kilometer scale (Figure 3b). On 801
 the basis of these detailed studies of faults in the vicinity of 802
 the Arava/Araba Valley we suggest that at the beginning of 803
 transform motion deformation occurred in a ~20–30 km 804
 wide belt, possibly with the reactivation of older approxi- 805
 mately north to south striking structures. Later, deformation 806
 became concentrated in the region of the present-day Arava/ 807
 Araba Valley. Until ~5 Ma ago, there might have been 808
 other, now inactive faults in the vicinity of the present-day 809
 AF that took up lateral motion (FAF, Figures 2c and 3b). 810
 Together with a rearrangement of plates ~5 Ma ago, the 811
 main fault trace shifted then from the FAF to the position of 812
 today's AF. 813

2.4. Meter to Microscopic Scale 814

[28] In a high-resolution small-scale seismic experiment 815
 (CSA II, Figure 2f and *Haberland et al.* [2006]) the shallow 816
 structure of the (active) AF is analyzed down to a maximum 817
 depth of a few hundred meters. The experiment consists of 818

Figure 15. Lithological cross section derived from joint interpretation of magnetotelluric and seismic models along the seismic NVR profile (white line in Figure 2c) and the 10 km long magnetotelluric pilot profile (dark blue dots in Figure 2d). (a) *P* velocity model [*Ryberg et al.*, 2007]. The seismic model was calculated on a mesh of 10,000 cells. The gray and white areas denote regions where velocities cannot be constrained. (b) Resistivity model to 5 km depth [*Ritter et al.*, 2003]. The model was calculated on a mesh of 5136 cells. No vertical exaggeration is applied. The red line marks the surface trace of the AF. (c) Spatial distribution of lithological classes derived from an automatic class selection. Colors correspond to the classes enumerated in Table 1. The arrow marks the surface trace of the AF, and the thin black lines are our preferred structural interpretation. Gray region indicates missing model data. Modified from *Bedrosian et al.* [2007], copyright Blackwell Publishing.

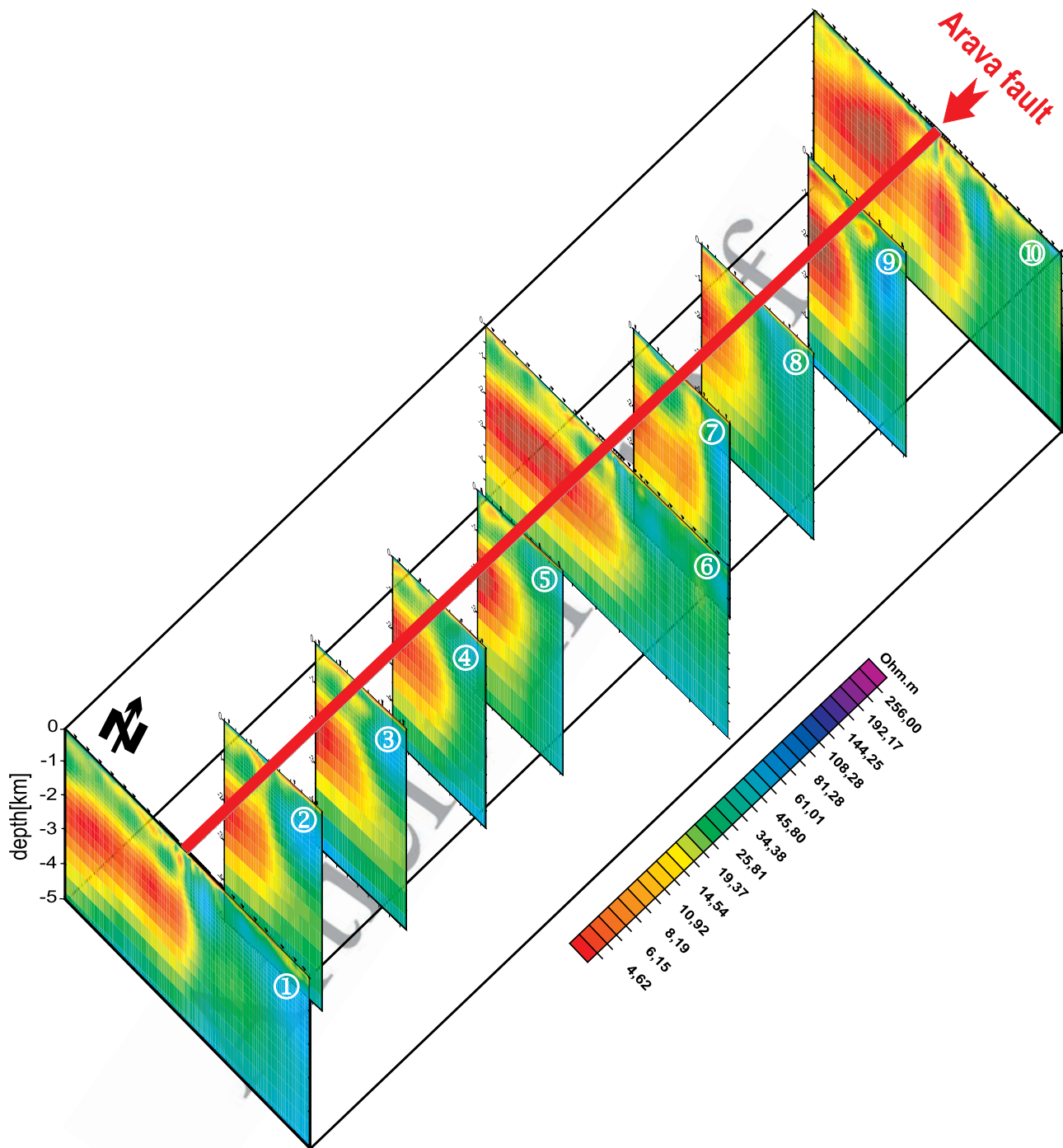


Figure 16. Three-dimensional electrical resistivity image around the AF (red line at top) in a $10 \times 10 \times 5$ km box. The resistivity is given in Ω m, and red and yellow indicate good conductors. The resistivity distribution is derived from 2-D inversions of the three long profiles (10 km length, profiles 1, 6, and 10) and seven short profiles (5 km length, profiles 2, 3, 4, 5, 7, 8, and 9). Profile 1 is also shown in Figure 15b. The locations of the ten magnetotelluric profiles are given by dark blue dots in Figure 2d. The main anomaly is a good conductor from 1.5 to 4 km depth west of the AF, showing some variation in the north-south direction. This anomaly is most likely caused by saline fluids in a sedimentary layer, with the AF acting as a seal for cross-fault fluid flow toward the east [Ritter *et al.*, 2003; Maercklin *et al.*, 2005; Bedrosian *et al.*, 2007]. From south to north a shallowing of this brine layer can be observed. No vertical exaggeration. Modified from Weckmann *et al.* [2003].

819 eight subparallel 1 km long seismic lines. The combination
 820 of first break tomography and reflection seismic images
 821 (Figures 21a and 21b) shows a subvertical (main) fault
 822 separating two blocks with different seismic (physical)

properties, positive flower structures, and, based on their
 823 seismic velocity, indications for different sedimentary layers
 824 on the two sides of the main fault. Often, the superficial
 825 sedimentary layers are bent upward close to the AF, 826

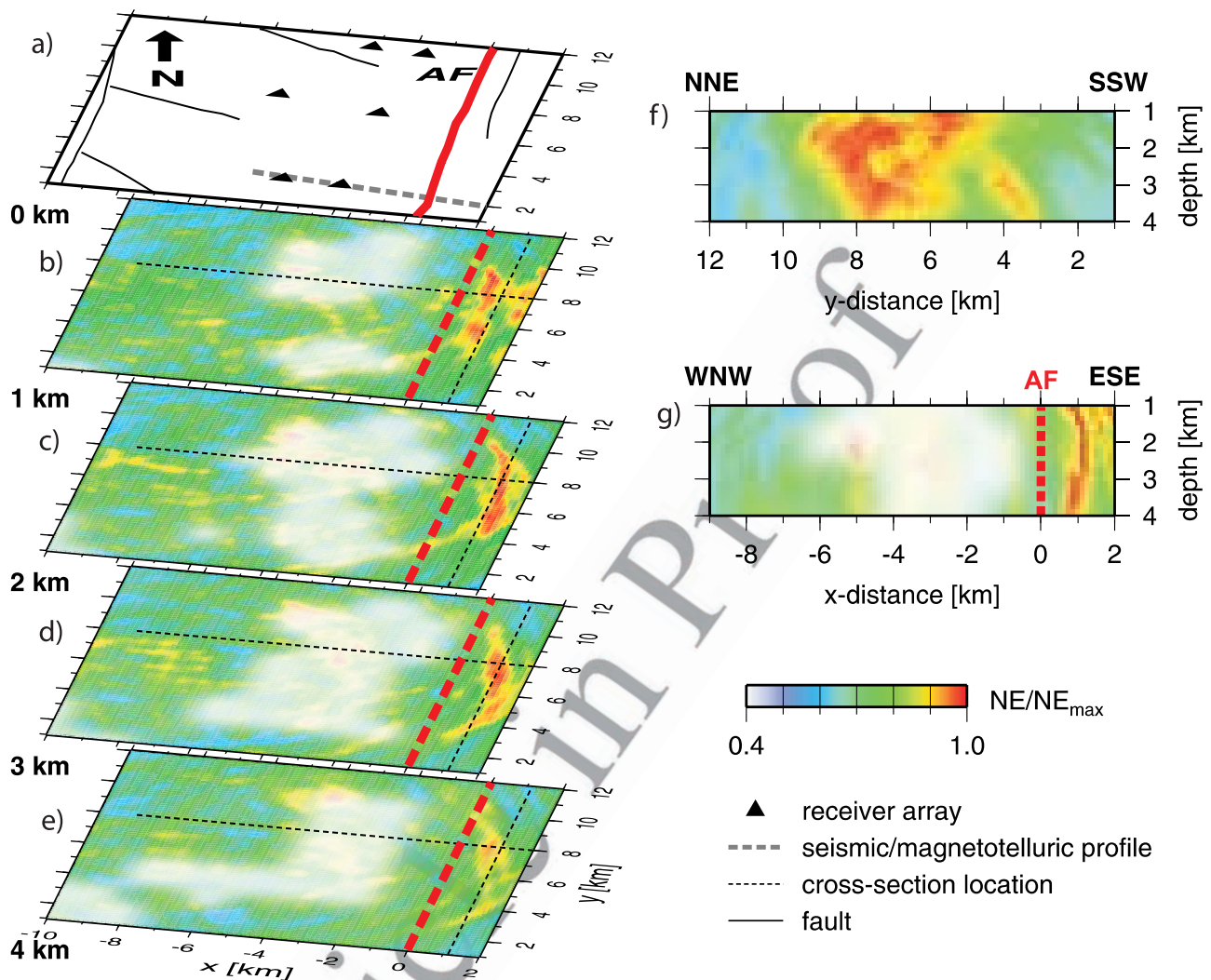


Figure 17. Color coded 3-D scatter image of the AF area in the central Arava/Araba valley. The model is derived from scattered seismic waves in the CSA I experiment. See Figure 2c: 3×10 km long profiles (thin black lines in Figure 2c); 9 miniarrays with 10 seismometers each (black triangles in Figure 2c); and 7×5 , 4×3 , and 3×2 shots (white stars in Figure 2c). Zones of strong scattering are in red. At these zones, high semblance values NE are observed; that is, these are the sources of secondary, scattered waves from the 53 shots located with the 9 miniarrays and the 3 profiles. For details of the method and the processing, see Maercklin *et al.* [2004]. Areas with poor resolution are whitened out. (a) A zoom of Figure 2c, with fault traces in black (except the AF, shown as a red line) and miniarrays (black triangles). (b–e) The four color-coded horizontal depth slices show the distribution of scatterers from 1 to 4 km depth, respectively. The downward projection of the AF is indicated by the dashed red line in each of the four depth slices. Note the 1 km offset of the scatterers (red) toward the east relative to the AF (dashed red line). Two vertical cross sections through the imaged volume at the two locations indicated by thin dashed black lines in the depth slices: (f) cut along the fossil AF (FAF) at $x = 1.125$ km running NNE to SSW and (g) cut across FAF at $y = 8$ km running WNW to ESE. The location of the AF at 0 km is indicated as a dashed red line. Modified from Maercklin *et al.* [2004], copyright Blackwell Publishing.

Figure 18. Selected cross sections of the 3-D density model of the central Arava/Araba Valley (see Figure 2e). (a) Study area and the residual gravity field in 10^{-5} m/s^2 . The local survey gravity stations are marked by small black dots (green dots in Figure 2e). The 3-D density model consists of 13 cross sections (thin black oblique lines) with 1–3 km of separation. The position of the five seismic velocity cross sections from the 3-D seismic tomography model [Maercklin, 2004; Maercklin *et al.*, 2005] used to constrain the gravity modeling are marked by thick black lines (L1, L12, L2, L23, and L3). The AF (thick black line) and the ZF (thick dashed black line) are also indicated. Cross sections along (b) L3, (c) L2, and (d) L1, corresponding to lines 3, 2, and 1, respectively, in Figures 2c and 2f, of the 3-D gravity model with the corresponding density values in Mg/m^3 . Above each density cross section the observed (red) and modeled (black) gravity anomalies in 10^{-5} m/s^2 are shown. The dashed lines are the velocity isolines of 3.5 km/s (top line), 4 km/s (middle line), and 4.5 km/s (bottom line) from the tomography model (P velocity cross sections) of Maercklin *et al.* [2005], shown below each panel. Distance and depth are in kilometers. Modified from Tašárová *et al.* [2006].

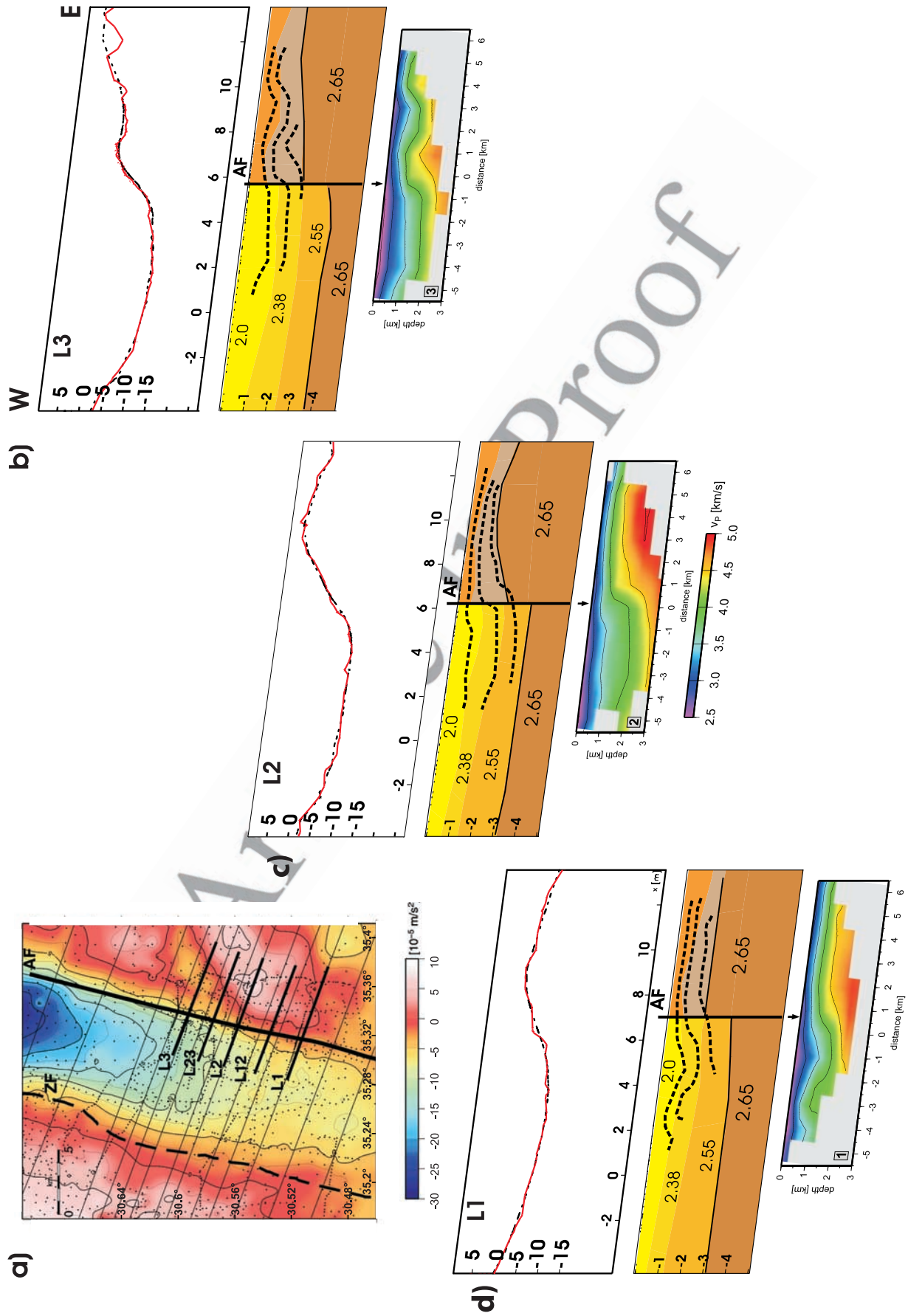


Figure 18

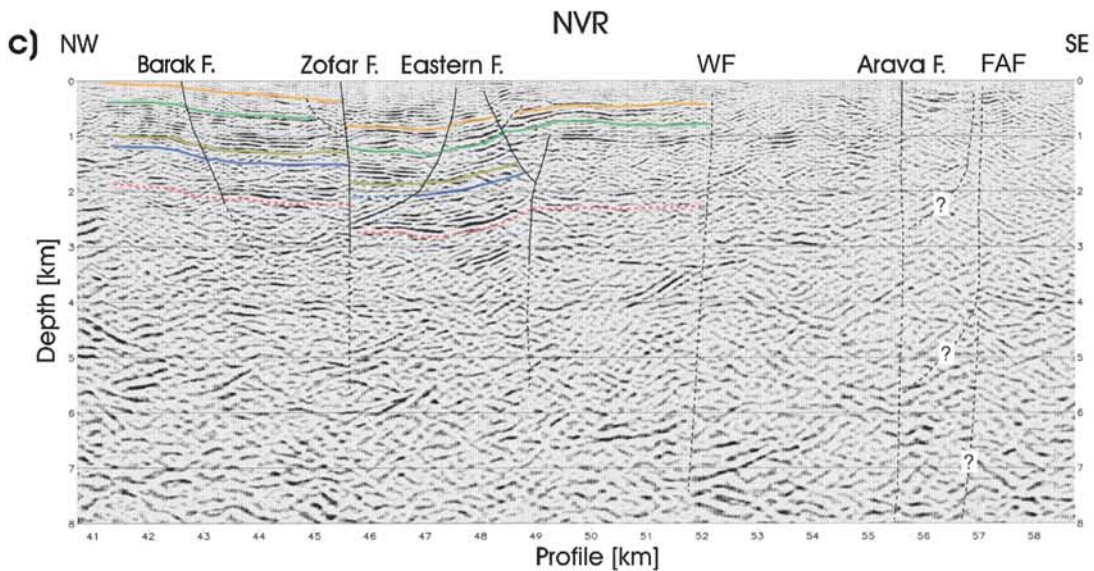
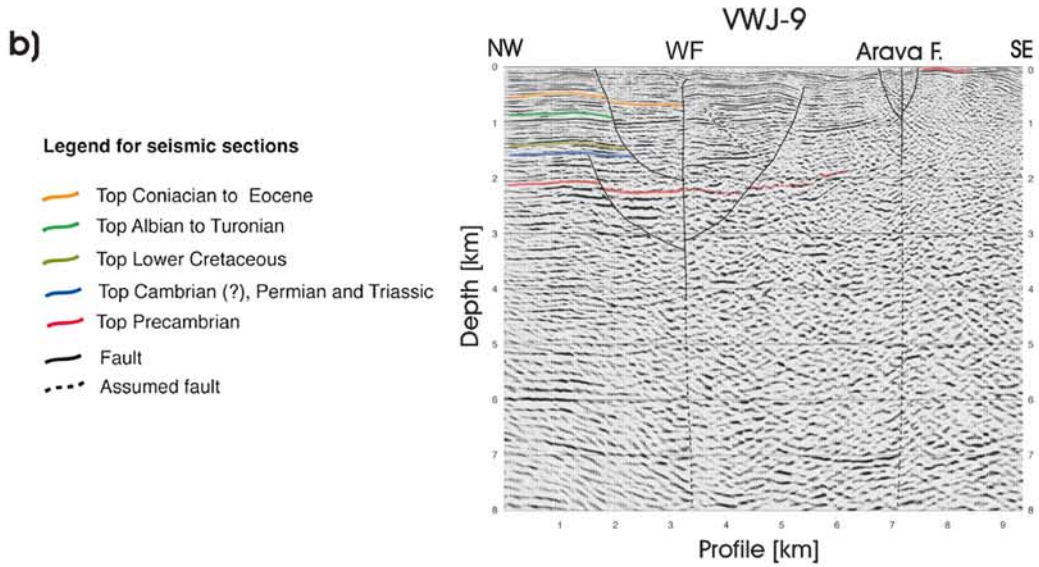
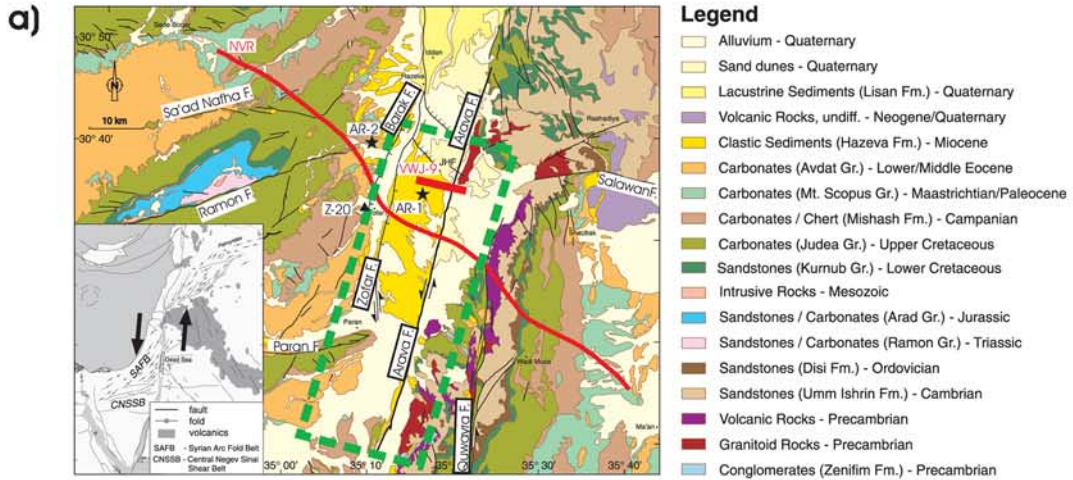


Figure 19

827 indicating that this section of the fault (at shallow depths) is
 828 characterized by a transpressional regime [see, e.g., *Harding*
 829 *and Lowell*, 1979; *Lowell*, 1985]. To the north, these
 830 structures correspond to the transpressional elements at
 831 the surface (pressure ridges, see Figure 1c) caused by the
 832 local restraining eastward bend of the AF near Jebel Humrat
 833 Fidan (Figure 20a at 30.62°N). A 100–300 m wide hetero-
 834 geneous zone of deformed and displaced material (Figures 21b
 835 and 21c) is detected which, however, is not characterized
 836 by low seismic velocities at a larger scale. Note that the
 837 shallow part of the DST in the southern Arava/Araba Valley
 838 is also several hundred meters wide [*ten Brink et al.*, 2007;
 839 *Shtivelman et al.*, 1998]. At depth below ~400 m, geophys-
 840 ical images indicate a blocked cross-fault structure (Figures 21d
 841 and 21e). The fault cores are not wider than ~10 m, in
 842 agreement with the study of fault zone guided waves from
 843 *Haberland et al.* [2003] indicated by black triangles in
 844 profiles P6, P7, and P8 (Figures 21c–21e). For a layout
 845 of that experiment, see Figures 2f and 2g. Such a narrow
 846 fault core is consistent with the up to 60 km displacement
 847 postulated for the AF [*Kesten et al.*, 2008]. Similar, wide
 848 (~1 km) flower structures related to the DST had been
 849 revealed within the sediments of the Evrona playa basin just
 850 north of the Gulf of Aqaba/Elat by high-resolution seismic
 851 studies [*Shtivelman et al.*, 1998]. In trenching analysis,
 852 *Niemi et al.* [2001] also found subparallel fault strands,
 853 pressure ridges, and comparable narrow fault zones at the
 854 AF zone ~50 km north of our study area. At a segment
 855 ~200 km farther to the north, the main faults of the DST are
 856 observed as wide zones of deformation rather than as
 857 distinct fault planes [*Rotstein et al.*, 1992].

858 [29] Geological and geochemical studies of carbonate
 859 fault rocks at the AF [*Janssen et al.*, 2004, 2005, 2007a,
 860 2007b] document mesoscale to microscale faulting and
 861 veining in limestone at two locations (areas I and II in
 862 Figure 22) representing faulting at depths of 2–5 km and up
 863 to 3 km, respectively. The role of fluids in faulting defor-
 864 mation in the AF is locally quite different from that of other
 865 major fault zones, like, e.g., the SAF. At the AF, the small
 866 amount of veins and the lack of alteration and dissolution
 867 processes in limestone suggest reduced fluid-rock interac-
 868 tions and limited fluid flow within the fault. Note that on the
 869 kilometer to meter scale the AF acts as a barrier for fluids
 870 (see, e.g., Figure 15). Hydrothermal reactions (cementation
 871 and dissolution) did not affect the strength of the fault zone,
 872 indicating that the AF is a strong fault near the surface down
 873 to a few kilometers in the upper crust. Other segments of the
 874 AF show some iron mineralization, which might be related

to an older fault system, known as the Central Negev-Sinai 875
 shear zone (CNSSB in Figure 19a (inset)) [see also *Frieslander*, 876
 2000] or the Syrian Arc Fold Belt System (SAFB in Figure 877
 19a (inset)) [see, e.g., *Bartov*, 1974; *Sneh et al.*, 1998; 878
Shamir et al., 2005]. In area I, where pressure ridges expose 879
 the exhumed fault and samples could be taken (red dia- 880
 monds in Figure 2e), calcite mineralization reveals an open 881
 fluid system with fluids originating from two sources. 882
 Stable isotopes ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) and trace elements indicate 883
 predominant infiltration of descending meteoric water, possi- 884
 bly supplied from the high eastern escarpment in Jordan 885
 [*Janssen et al.*, 2005]. This source is indicated by the good 886
 shallow electrical conductor east of the AF interpreted as a 887
 clay layer and possibly acting as caprock for meteoric water 888
 (near-surface red color east of the AF in Figure 22 (bottom); 889
Ritter et al. [2003] and class 6 in Figure 15). This good 890
 conductor is also confirmed in the Short Offset Transient 891
 Electromagnetics experiments (blue profiles in Figure 2f; S. 892
 Helwig (personal communications, 2008)), showing that the 893
 subsurface groundwater flow is blocked at the AF. In area II 894
 (Figure 22, top right), geochemical data indicate only local 895
 (small-scale) fluid redistribution along profile B. These 896
 fluids were derived from the adjacent limestone under 897
 nearly closed system conditions [*Janssen et al.*, 2004]. 898

2.5. Modeling and Interpretation 899

[30] Finite element 2.5-D thermomechanical modeling of 900
 the DST on plate tectonic to kilometer scale is used by 901
Sobolev et al. [2005] to study the dynamics of this conti- 902
 nental transform boundary between the Arabian and African 903
 plates. 2.5-D means calculation of all three components of 904
 the displacement (velocity) vectors under assumption of no 905
 changes in material properties, temperature, and velocity 906
 along the strike of the DST. The results of geological, 907
 geophysical, geothermal, and petrophysical observations, 908
 mostly from the DESERT project reported here, are used 909
 to constrain initial and boundary conditions and to choose 910
 the thermal and rheological parameters. The preferred 911
 model combines plate-scale transtension (strongly dominated 912
 by a strike-slip deformation component of 105 km) with 913
 thinning of the mantle lithosphere of the Arabian Shield at 914
 about 5–10 Ma and has a relatively weak crust [*Sobolev et* 915
al., 2005]. Figure 23 shows the setup of the thermomechan- 916
 ical model. In the initially cold lithosphere expected at the 917
 DST, shear deformation localizes in a 20–40 km wide zone 918
 (Figure 23) where the temperature-controlled mantle 919
 strength is minimal. The largest strain rates and finite strain 920
 are concentrated in the 5 km wide core of this zone. The 921

Figure 19. Geological map of the central Arava/Araba Valley and near-surface structure derived from seismic reflection data. Faults: BF, ZF, EF, WF, AF, and FAF. (a) Geological map (compiled after *Sneh et al.* [1998], *Bender* [1968], and *Frieslander* [2000]). The NVR and the VWJ-9 common depth point lines are given in red (white lines in Figure 2c). The green dashed box indicates the area of Figure 20a. Information on the stratigraphy can be found in Table 2. The inset shows information on some major tectonic elements in the Dead Sea region [after *Sneh et al.*, 1998]. Z-20, Zofar-20 well; AR-1 and AR-2, outcrops of Arava/Araba Formation; JHF, Jebel Humrat Fidan; CNSSB, Central Negev-Sinai shear zone; SAFB, Syrian Arc Fold Belt System. (b) Depth migrated shallow seismic reflection profile VWJ-9 [*Kesten*, 2004]. (c) Upper central part of the depth migrated NVR profile, where only the 18 km range in the green box in Figure 19a is shown (profile km 40.5–58.5). Modified from *Kesten et al.* [2008] with kind permission of Springer Science and Business Media.

922 resulting mechanically weak decoupling zone (20–40 km
923 wide with the 5 km wide weakest core), controlled by shear
924 heating and temperature and strain rate dependence of the
925 viscosity, extends subvertically through the entire litho-

sphere. That is in general agreement with seismological, 926
seismic, and magnetotelluric observations (Figures 6–9 927
and 3a), although important details remain unexplained 928
(see discussion later in this section). Modeling furthermore 929

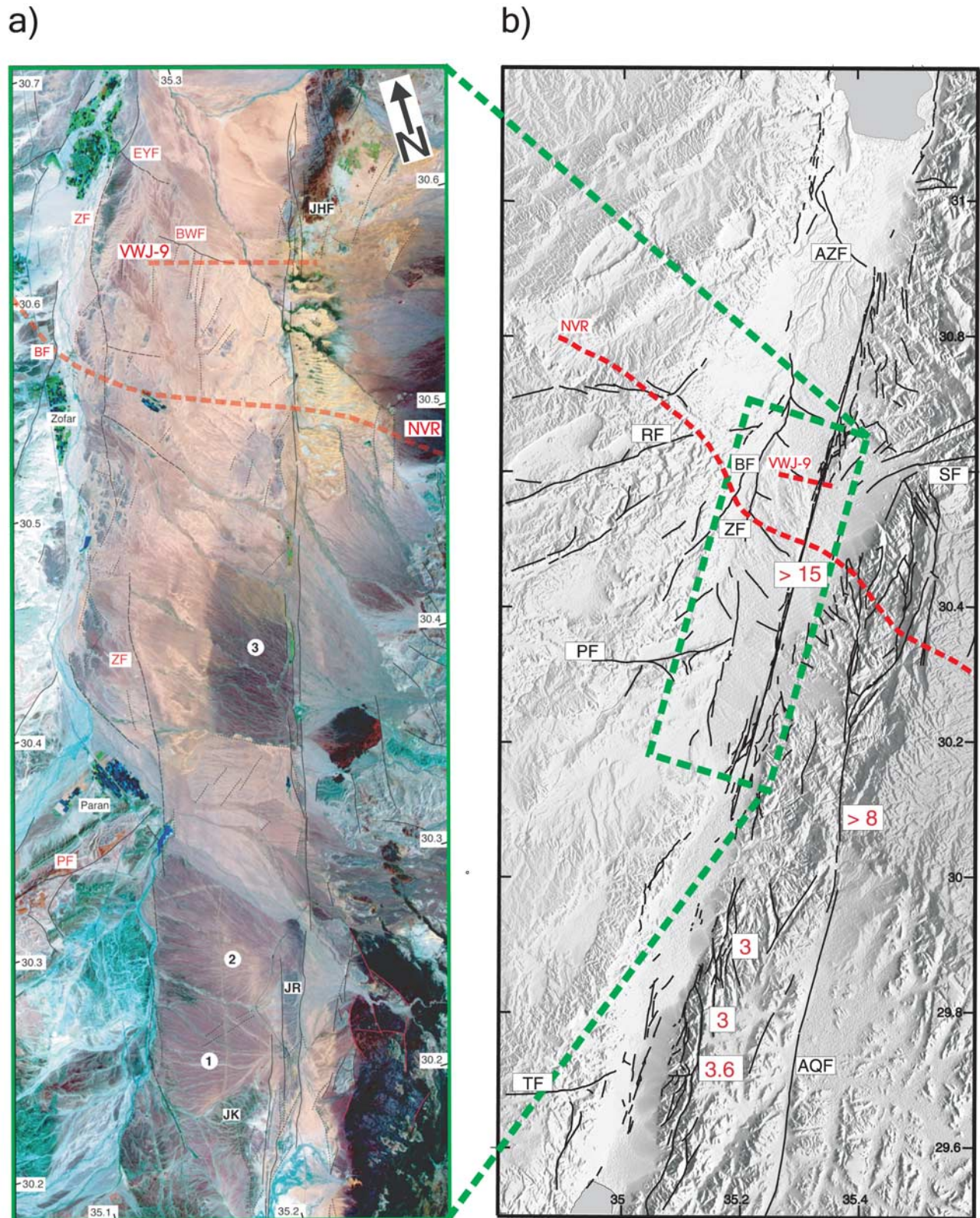


Figure 20

930 suggests that the location of the AF has been controlled by
 931 the minimum in lithospheric strength possibly associated
 932 with the margin of the Arabian Shield lithosphere and/or by
 933 the increased crustal thickness toward the east, also visible
 934 in seismic data [Koulakov and Sobolev, 2006; DESERT
 935 Group, 2004; Mechie et al., 2005]. In the crust, one or two
 936 major faults take up most of the transform displacement, but
 937 a few kilometers of displacement occur also at several minor
 938 faults. These modeling results are consistent with geological
 939 and geophysical observations of several faults in the crust
 940 (Figures 14–20, 3a, and 3b) and the lithosphere structure
 941 imaged along the DESERT profile. The modeling also
 942 shows that less than 3 km of transform-perpendicular
 943 extension occur (Figure 24), suggesting that the AF segment
 944 of the DST is a dominantly strike-slip plate boundary. Note
 945 also the area of strong fault-parallel deformation in the lower
 946 crust (especially in the weak crust model (Figures 24d–
 947 24f)), which possibly is responsible for the LCR in Figure 11.
 948 We would furthermore like to point out that the hypothesis of
 949 the AF as a strong fault [Janssen et al., 2004, 2005, 2007a,
 950 2007b] holds only for the near-surface region but not for the
 951 crustal scale discussed here. Other geological, geophysical,
 952 and geodetic observations, like the slight asymmetric topog-
 953 raphy of the Moho, are also well reproduced. The uplift
 954 of the Arabian Shield adjacent to the DST requires young
 955 (<10 Ma) thinning of the lithosphere at of the plate boundary.
 956 Such lithospheric thinning is consistent with seismological
 957 observations and the high temperatures derived from mantle
 958 xenoliths in Neogene–Quaternary basalts [Sobolev et al., 2005].
 959 [31] An interesting question is the nature of the westward
 960 shift of the zone of higher seismic anisotropy and of the MT
 961 conductor relative to the surface trace of the DST, i.e., the
 962 AF. If both seismic anisotropy and conductor are associated
 963 with the locus of the shear deformation in the lower crust
 964 and lithospheric mantle as we suggest, then the present-day
 965 shear zones in the upper crust (AF) and in the deeper crust
 966 and mantle lithosphere (marked by the MT conductor and
 967 zone of seismic anisotropy) appear to be mutually shifted.
 968 Thermomechanical modeling [Sobolev et al., 2005] demon-
 969 strates that such a shift is indeed possible (Figures 23c and
 970 23f) because of heterogeneity of the lithospheric strength.
 971 However, the model suggests shift of the mantle shear zone
 972 to the east rather than to the west of the AF (Figure 23f).
 973 The most likely reason for such contradiction is that the
 974 model does not include the Dead Sea pull-apart basin which
 975 is located <100 km north of the DESERT line. If the mantle

deformation zone is placed right beneath the Dead Sea basin 976
 and continues to the south parallel to the strike of the DST, 977
 as it is suggested by the numerical model of a pull-apart 978
 basin [Petrunin and Sobolev, 2006], then it must cross the 979
 DESERT line indeed west of the AF. Another possibility is 980
 that the AF at the DESERT line is a relatively young 981
 feature, and most of the strike-slip displacement is taken 982
 by other faults west of it, located above the zone of strongly 983
 anisotropic mantle. This idea is in line with seismic obser- 984
 vation suggesting that the uppermost section of the AF is 985
 very narrow (10 m) but apparently contradicts geological 986
 observations [Bartov et al., 1998]. 987

[32] Finally, we would like to discuss how the DST 988
 modifies strength of the lithosphere, which it cuts through, 989
 on the basis of thermomechanical model and multidisciplinary 990
 observations. Various seismic observations presented in 991
 sections 2.2 and 2.3 indicate reduced seismic velocities and 992
 increased seismic anisotropy in the crust beneath the DST. 993
 This is likely due to the high crack density and indicates 994
 mechanical weakening of the crust, suggesting that the 995
 hypothesis of the AF as a strong fault [Janssen et al., 996
 2004, 2005, 2007a, 2007b] holds only for the near-surface 997
 region but not for the crustal scale discussed here. The 998
 thermomechanical modeling [Sobolev et al., 2005] also 999
 suggests that major faults in the upper crust of the DST 1000
 must be significantly weaker than the bulk of the crust. The 1001
 model also infers that the zone of mechanical weakening 1002
 continues deeper below the brittle-ductile transition and 1003
 crosses the entire lithosphere (see distribution of viscosity 1004
 in Figures 23d and 23g). There are at least two reasons why 1005
 viscosity drops in the DST shear zone. One is the strain rate 1006
 dependency of the viscosity related to the dislocation creep 1007
 in lithospheric rocks, which leads to the reduction of the 1008
 viscosity in the zone of the highest strain rate. Another 1009
 reason is shear heating, which increases temperature in the 1010
 shear zone. As a result of a number of processes also 1011
 including fluid flow (recall MT conductor in the middle- 1012
 deep crust), the lithospheric-scale shear zone appears to be a 1013
 self-weakening body crossing the entire lithosphere. 1014

3. DISCUSSION AND COMPARISON OF THE DEAD 1016 SEA TRANSFORM AND THE SAN ANDREAS FAULT 1017

[33] The DST has a relatively slow present-day plate 1018
 motion of ~ 5 mm/a with a total displacement of 105 km 1019
 in 20 Ma (up to 60 km of it along the AF). This puts the 1020

Figure 20. Satellite image and fault map of the Arava/Araba Valley. (a) Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) scene of the central Arava/Araba Valley taken on 6 April 2001. Three types of lineaments can be distinguished: (1) Solid lines represent lineaments that are clearly recognized as faults (by displaced geological units, offset alluvial fans, or clear “doglegs” of streams); (2) dashed lines are lineaments that were identified as faults in other, mainly shallow seismic studies [e.g., Frieslander, 2000]; and (3) dotted lines are lineaments whose origin could not be clarified for lack of geological or geophysical information. EYF, En Yahav Fault; BWF, Buweirida Fault; PF, Paran Fault; JR, Jebel Er Risha; JK, Jebel El Khureij. Numbers 1–3 indicate alluvial fans. (b) Fault map of the southern DST, derived from the interpretation of ASTER satellite images [Kesten, 2004; Kesten et al., 2008] over shaded relief map. The red numbers indicate the minimum amount of left-lateral strike-slip displacement along the respective faults in kilometers. AZF, Amaziahu Fault; RF, Ramon Fault; SF, Salawan Fault; TF, Themed Fault. Here the nomenclature of Frieslander [2000] and Calvo and Bartov [2001] is used. Modified from Kesten et al. [2008] with kind permission of Springer Science and Business Media.

t1.1 **TABLE 1. Resistivity (ρ), Seismic Velocity (V_p), and Thickness (t) of Classes Denoted in Figure 15c^a**

t1.2	Class	ρ (Ω m)	V_p (km/s)	t (km)	Age	Lithology Type	Stratigraphic Unit
t1.3	1	16–92	5.5 ± 0.3	1.5+	<i>l</i> PC	ark, cgl, volc	Zenifim formation (W); Aqaba complex: Ghuwayr volcanic, Araba complex (undifferentiated) (E)
t1.4	2	26–173	4.8 ± 0.4	3+	<i>l</i> PC	cgl, qtz porph	Fidan granite, Araba complex: Ahaymir volcanic (E)
t1.5	3	2–11	4.4 ± 0.5	1.6	C-K	sst, lst, dolm, mar, clst	Yam Suf, Negev, Ramon, Kurnub, Judea groups (W) \approx 1.7 km
t1.6	4	6–39	3.9 ± 0.3	1.0	<i>l</i> K-T	chk, cht, lst, mar	Mount Scopus, Avedat groups (W) (equivalent to Belqa group), 700 m
t1.7	5	10–43	3.0 ± 0.3	0.5	Tm	sst, clst, cgl	Hazeva group (W) \approx 500 m
t1.8	6	11–39	2.2 ± 0.4	0.2	Tp-Qp	alluv: cgl, sd, slt, grv	Arava formation (W, E)

^aAlso included are inferred age (*l*PC, Late Precambrian; C-K, Cambrian-Cretaceous; *l*K-T, Late Cretaceous-Tertiary; Tm, Miocene; Tp-Qp, Pliocene-Pleistocene), lithology type (ark, arkose; cgl, conglomerates; volc, volcanic rocks; qtz, quartz; porph, igneous rocks; sst, silt stone; lst, limestone; dolm, dolomite; mar, marl; clst, clastic rocks; chk, chalk; cht, chert; alluv, alluvium; sd, sand; slt, silt; grv, gravel), and stratigraphic units. Classes spanning the Arava Fault are identified by geologic unit on both the west (W) and east (E) sides of the fault. Modified from *Bedrosian et al.* [2007].

1021 DST in marked contrast to other major plate-bounding
 1022 transform fault systems such as the NAF zone (presently
 1023 20 to 30 mm/a; 80 km in \sim 5 Ma) located amidst an orogenic
 1024 belt [see, e.g., *Sengör, 1979; Barka, 1992; McClusky et al.,*
 1025 *2000; Hubert-Ferrari et al., 2002*] and the dominantly
 1026 transpressional SAF system (50 mm/a distributed among
 1027 numerous faults from offshore to Colorado plateau; 300 km
 1028 since 5–6 Ma) which originated from a complicated inter-
 1029 action between oceanic subplates and continental rocks [see,
 1030 e.g., *Nicholson et al., 1994; Holbrook et al., 1996; Henstock*
 1031 *et al., 1997; Fuis, 1998; Atwater and Stock, 1998; Powell,*
 1032 *1993; Oskin et al., 2001; Mooney et al., 2007; Fuis et al.,*
 1033 *2008*]. A comparison of the characteristics of the DST and
 1034 the SAF is given in Table 3. In contrast to the SAF the recent
 1035 seismicity at the DST is moderate. As at the SAF, earth-
 1036 quakes are clustered in time at the DST, with large earth-
 1037 quakes occurring in a long cycle of several hundred to
 1038 thousand years [*Marco et al., 1996; Amit et al., 2002; Marco*
 1039 *et al., 2005*], whereas the recurrence intervals at the SAF
 1040 are typically 100–400 years [*Fumal et al., 2002*] depending
 1041 on the segment. *Lyakhovskiy et al.* [2001] find that where
 1042 the rate of healing is large compared to the rate of loading,
 1043 the system exhibits short memory, and fault geometry
 1044 evolves along several seismic cycles. The low rate of loading
 1045 on the AF is compatible with such a system and with the
 1046 distribution of the total slip across several strands, each
 1047 active at a different time (see Figures 3, 9, 14, 15, 17, 19, 20,
 1048 23, and 24).
 1049 [34] In the subcrustal lithosphere the DST is a narrow 5–
 1050 20 km wide anisotropic zone of (most likely) fault-parallel

mineral alignment, suggesting subhorizontal fault-parallel
 mantle flow within a zone with distinct mineral alignment
 that extends through the entire lithosphere (Figures 6 and 9).
 The SAF also cuts through the crust [*Holbrook et al., 1996;*
Henstock et al., 1997; Unsworth et al., 2000; Silver, 1996;
Mooney et al., 2007; Becken et al., 2008] and even the
 lithosphere [*Fuis et al., 2008*], but estimates of the width of
 the fault-related seismically anisotropic zone below the crust
 range from 30 to 150 km [*Silver, 1996; Savage, 1999*]. This
 difference between the DST and the SAF could possibly
 indicate that the width of the decoupling zone scales with the
 total strain accumulated along the fault. One of the most
 likely main driving forces for the differences in strain and
 decoupling zones between the SAF and the DST is the
 difference in mantle thermal structure. While the DST
 originated in thick, cold lithosphere, which only recently
 may have been thinned to \sim 80 km [*Sobolev et al., 2005*],
 the SAF has been generated in thin lithosphere underplated
 by hot mantle penetrating the opening slab window [e.g.,
Dickinson and Snyder, 1979; Furlong et al., 1989; ten Brink
et al., 1999b; Fuis et al., 2008; Wilson et al., 2005].
 [35] Both the DST and the SAF systems show a strong
 asymmetry in subhorizontal lower crustal reflectors (LCR in
 Figure 11) and a deep reaching narrow deformation zone of
 about 5 km width in the middle and lower crust (Figures 7
 and 9 for the DST and *Holbrook et al.* [1996], *Silver* [1996],
Henstock et al. [1997], *Mooney et al.* [2007], *Becken et al.*
 [2008], and *Fuis et al.* [2008] for the SAF). Whereas two
 large subvertical crustal conductivity zones are found in the
 Arava/Araba Valley and its vicinity (Figure 9), one SW

Figure 21. Tomographic and reflection seismic images of the top few hundred meters along a \sim 8 km long N–S segment of the (active) AF derived from the CSA II experiment (eight short white lines in Figure 2f, P1–P10, station spacing 5 m, and shot spacing of 20 m). (a) Tomographic inversion of the eight seismic lines. P wave velocities are in color (see scale at the bottom), and unresolved regions are masked (white). Depth range down to \sim 200 m. (b) Migrated reflection seismic sections down to \sim 500 m depth of all eight lines. Positive and negative amplitudes are shown as blue and red, respectively. (c) Migrated reflection seismic images (black wiggles) overlaid on the tomography (color) together with geological interpretation. Black thick lines indicate inferred faults, and dashed lines indicate less well constrained faults. M, S, and C indicate the main fault, secondary/flanking faults, and the sedimentary cover east and west, respectively. Inverted black triangles indicate the position of FZGW observations from *Haberland et al.* [2003]. The shots and receivers producing and recording FZGW are indicated in Figure 2g by red triangles and stars, respectively. (d) Summary of interpreted sections down to 500 m depth along a \sim 8 km long N–S segment of the AF based on Figures 21a–21c. The main fault is indicated by the red area. (e) Reconstruction of the fault structure based on seismic results. A few subparallel faults form the AF system in the study area; to the north the pressure ridge structure dominates. Modified from *Haberland et al.* [2006], copyright 2006, Elsevier.

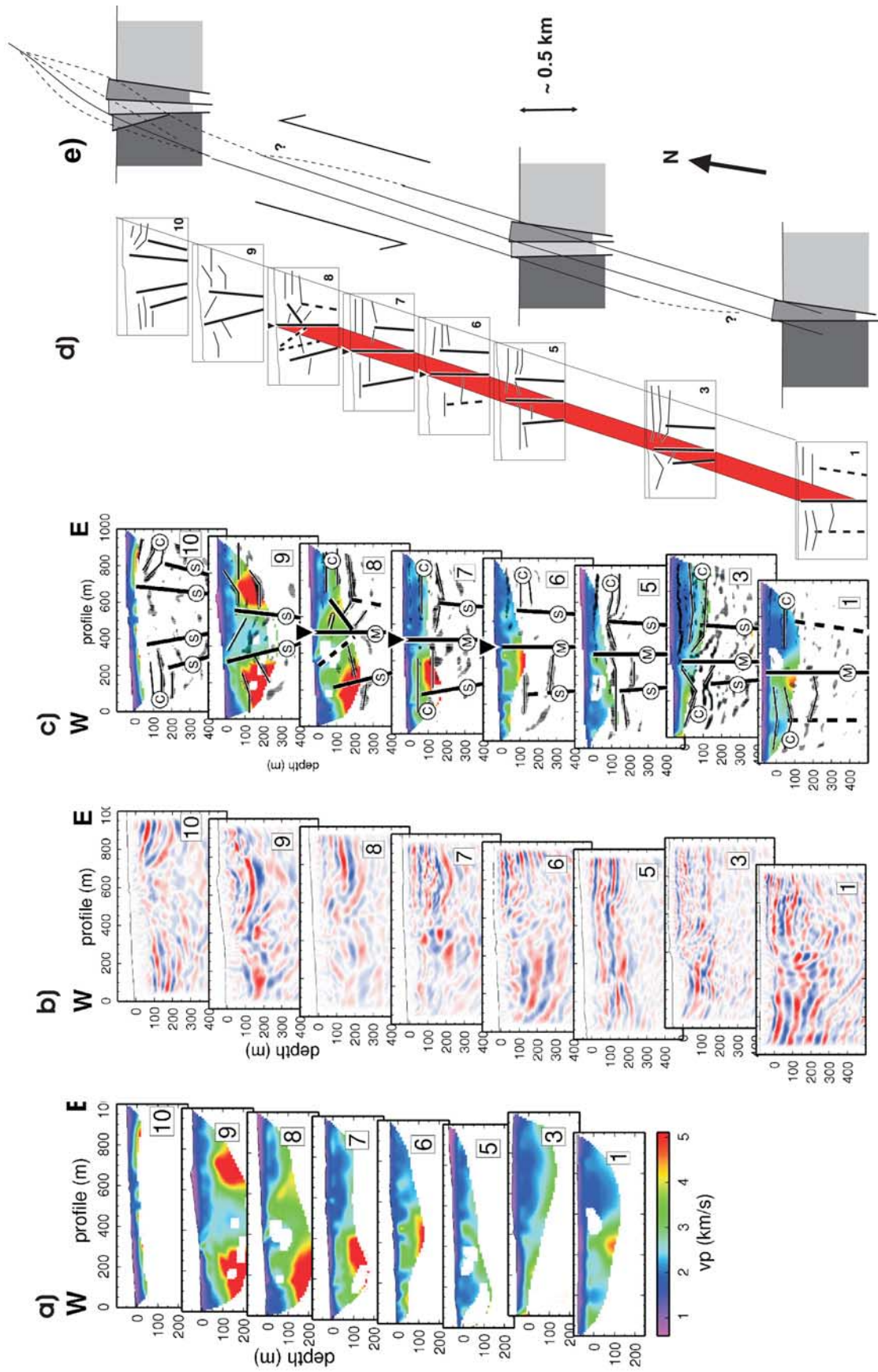


Figure 21

1081 tilted crustal conductivity zone is found in the vicinity of the
 1082 SAF at Parkfield [Becken et al., 2008, Figure 14]. Such
 1083 anomalies could be due to ascending fluids; see Janssen et
 1084 al. [2005, 2007a] for the AF and Wiersberg and Erzinger
 1085 [2007] for the SAF. At the SAF this conductive zone widens
 1086 in the lower crust and seems connected to a broad conduc-

tivity anomaly in the upper mantle, but as stressed by 1087
 1088 Becken et al. [2008], the upper crustal branch of the inferred
 1089 fluid conduit is located NE of the seismically defined SAF,
 1090 suggesting that the SAF itself does not act as a major fluid
 1091 pathway, at least not in the area near Parkfield. At the AF,
 1092 only reduced rock interactions and limited fluid flow were

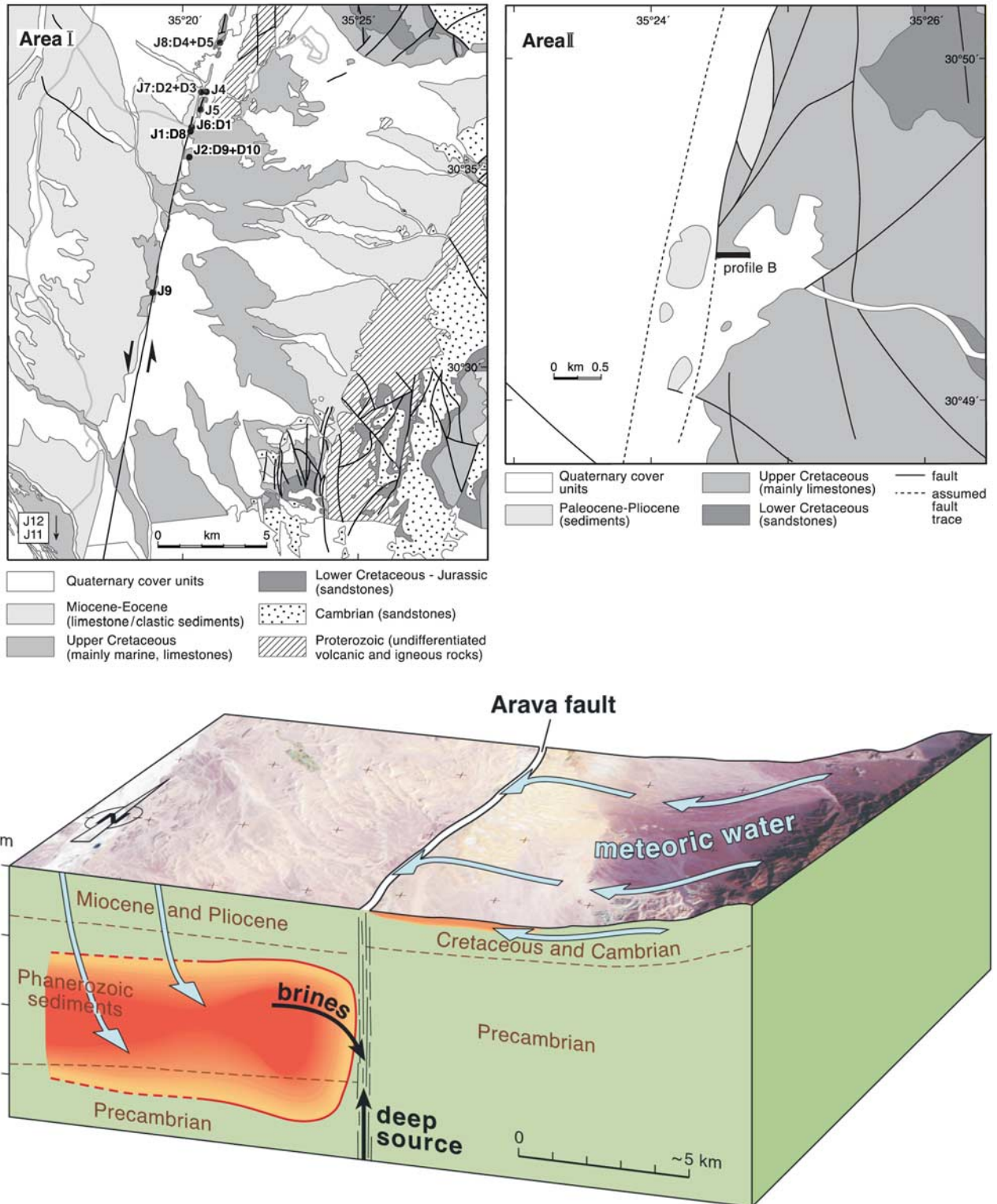


Figure 22

1093 observed [Janssen et al., 2005]. This leads us to conclude
 1094 that neither the SAF (at least at Parkfield) nor the AF act as
 1095 important fluid conduits, despite their strong signal in
 1096 seismological and MT studies. We would also like to point
 1097 out that the bright spot under the San Gabriel Mountains
 1098 [Ryberg and Fuis, 1998] implies large fluid contents. As
 1099 this feature is connected geometrically to the SAF in crust
 1100 and mantle the SAF is interpreted as a fluid conduit there
 1101 [Fuis et al., 2001].

1102 [36] In the uppermost crust the AF occurs as a barrier to
 1103 fluid flow (Figures 9, 15, 16, and 3b) and not as a single,
 1104 wide damage zone, a characteristic element of large, brittle
 1105 fault zone structures [Chester and Logan, 1986; Scholz,
 1106 1987, 2000, 2002]. The SAF has multiple strands and is at
 1107 ~ 3 km depth a barrier to fluids, at least at SAFOD
 1108 [Unsworth et al., 1999, 2000; Bedrosian et al., 2002; Becken
 1109 et al., 2008]. The West Fault in Chile [Hoffmann-Rothe et
 1110 al., 2004] also shows a pronounced fault zone conductor in
 1111 the top 2–3 km. The central segment of the SAF near
 1112 Parkfield is a location in transition between locked and
 1113 creeping. Here the zone of high conductivity within the
 1114 upper 2–3 km is attributed to fluids within a highly fractured
 1115 damage zone [Unsworth et al., 2000; Ritter et al., 2005a].
 1116 The depth extent of the corresponding seismic low-velocity
 1117 zone is ~ 3 km, the base of which coincides with a cluster of
 1118 small earthquakes. The width of this seismic low-velocity
 1119 zone inferred from fault zone guided waves [Li et al., 1990,
 1120 1997; Mooney et al., 2007] is 100–700 m. Note also that at
 1121 the NAF, Ben-Zion and Sammis [2003] found similar fault
 1122 zone widths of ~ 100 m. Possible reasons why no shallow,
 1123 single, wide damage zone is observed at the AF could
 1124 include the lower slip or reduced seismic activity along the
 1125 AF in the last few hundred years. At the AF, strain may have
 1126 been localized for a considerable time span along a narrow,
 1127 meter-scale damage zone, with a sustained strength differ-
 1128 ence between the shear plane and the surrounding host rock.
 1129 As a consequence, the existence or nonexistence of high
 1130 electrical conductivity in the central part of large-scale
 1131 strike-slip fault zones may be an indicator for the degree
 1132 of strain localization during faulting [see also Ritter et al.,
 1133 2005a].

1134 [37] In the top kilometer a network of (subparallel)
 1135 individual faults characterized by narrow fault cores/
 1136 damage zones is observed at the AF (Figures 21 and 3c).
 1137 A narrow fault zone width of the main strand, between

5 and 20 m, was found by analysis of fault zone guided 1138
 waves [Haberland et al., 2003]. These narrow faults (black 1139
 lines in Figure 21c) then form a broad heterogeneous zone 1140
 of deformed and displaced material, which, however, is not 1141
 characterized by low seismic velocities at a larger scale. On 1142
 the other hand, throughgoing, subvertical low-velocity 1143
 zones with a typical width of 100–300 m have been found 1144
 at large shear zones such as the SAF and the NAF [see, e.g., 1145
 Li et al., 1997, 1998; Ben-Zion and Sammis, 2003; Lewis et 1146
 al., 2005]. A comparison of the few-meters-wide main 1147
 strand of the AF with the thickness of gouge in laboratory 1148
 experiments is consistent with corresponding total slip 1149
 along the AF of 60 km [Kesten et al., 2008]. The apparent 1150
 distribution of deformation across several fault strands and 1151
 the concentration of the deformation to individual narrow 1152
 fault zones might be related to the low loading to healing 1153
 ratio at the AF. The size of the damage zone is determined 1154
 by a competition between localization and delocalization 1155
 processes and thus depends strongly on the segment of the 1156
 fault studied. The SAF zone does not heal completely on the 1157
 time scale of the seismic cycle, and ruptures tend to repeat 1158
 on the same smooth trend [Stirling et al., 1996] and in very 1159
 narrow zones [Rockwell and Ben-Zion, 2007]. 1160

[38] Structural and fluid properties of large fault systems 1161
 vary in time and space [Evans and Chester, 1995; Caine et 1162
 al., 1996; Evans et al., 1997; Hoffmann-Rothe et al., 2004]. 1163
 The internal structures of the Nojima fault zone (Japan) and 1164
 the SAF are very similar, exhibiting a continuous meter- 1165
 thick fault core containing foliated and nonfoliated ultra- 1166
 cataclastites and alteration minerals surrounded by a wider 1167
 zone of damaged host rocks [Chester et al., 1993; Ohtani et 1168
 al., 2000]. Such a narrow fault core, composed of chemi- 1169
 cally altered rocks, was only observed in one outcrop at the 1170
 DST, the Serghaya fault section in Syria, ~ 500 km north of 1171
 the study area [Janssen et al., 2007a, 2007b], but not on the 1172
 AF. In fluid properties, however, some similarities between 1173
 the DST, the SAF, and the Nojima fault zone systems exist. 1174
 Both the DST and SAF show significant variations in the 1175
 intensity of fluid-rock interaction depending on fault seg- 1176
 ment. In all three fault systems, fluids originated from a 1177
 variety of sources under different flow conditions, and 1178
 geochemical results show that the fluids are predominantly 1179
 of meteoric origin and migrated downward at shallow to 1180
 moderate depths [Kharaka et al., 1999; Lin et al., 2001; Pili 1181
 et al., 2002; Janssen et al., 2004, 2005, 2007a, 2007b]. 1182

Figure 22. Detailed geological maps of two locations at the AF and sketch of fluid flow at the AF. (top left) Geological map of the Fidan region (area I) slightly modified from Rabb'a [1991] with locations of mesostructural and microstructural analysis indicated by black dots with labels (J, location number; D, sample number from Janssen et al. [2004]). These locations are also indicated by red diamonds in Figure 2e. (top right) Geological map of the Fifa region (area II in Figure 2b, slightly modified from Tarawneh [1992]). (bottom) Sketch of the fluid movement at the AF (area I, Figure 22 (top left)) illustrating fluid infiltration from different sources. Geological units are from Figure 14c [DESERT Group, 2004]. Red and yellow areas are regions of high electrical conductivity, indicative of (saline) fluids [Ritter et al., 2003; Maercklin et al., 2005; Bedrosian et al., 2007; Weckmann et al., 2003] (see also Figures 15 and 16). Meteoric water (the red near-surface conductor east of the AF) is supplied from the high eastern escarpment in Jordan. Rare earth element and Sr isotopes analysis also suggest minor involvement of fluids from a deep source that ascend through the AF. Modified from Janssen et al. [2004] (copyright 2004 by the University of Chicago) and Janssen et al. [2005] (with kind permission of Springer Science and Business Media).

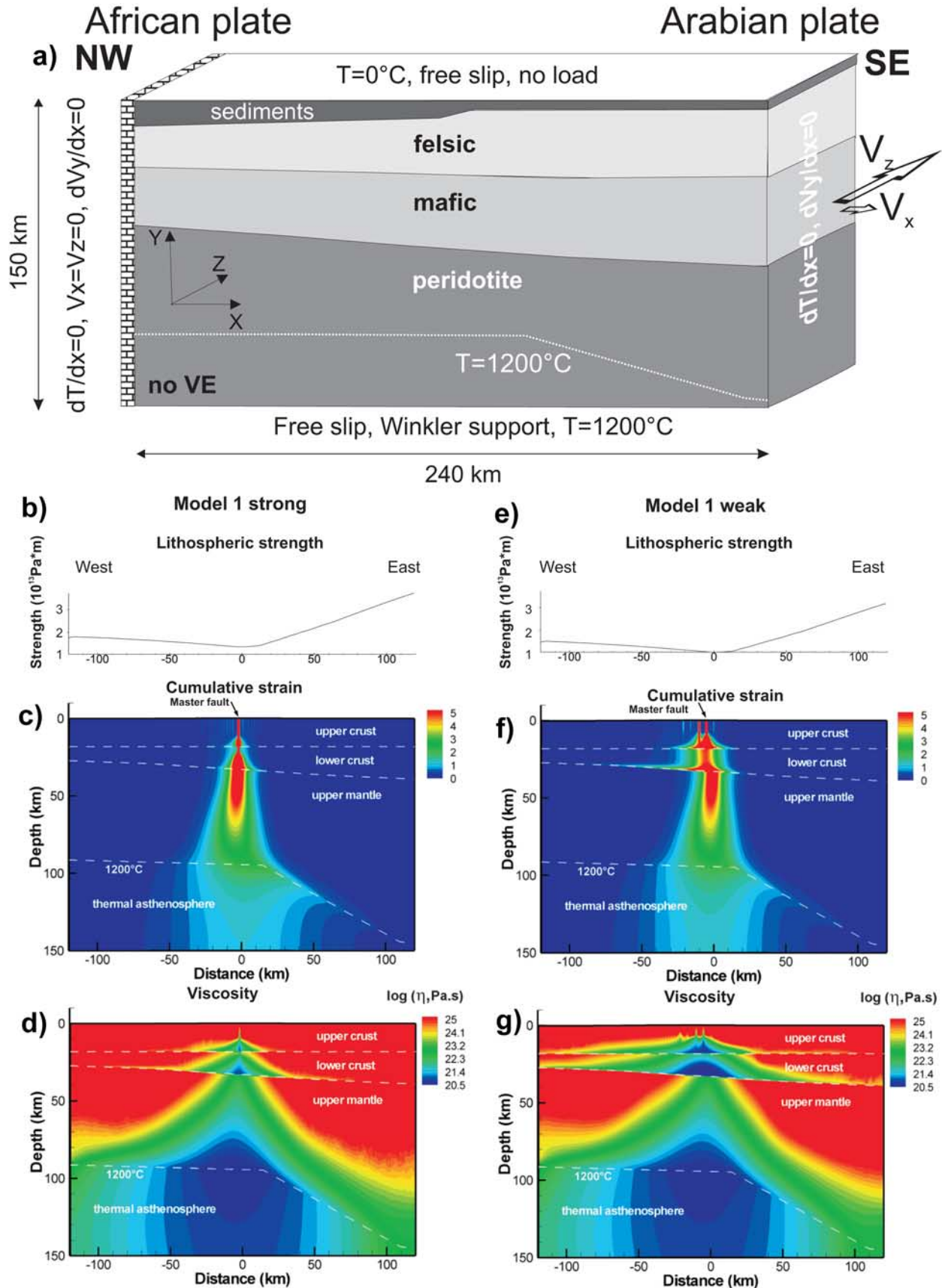


Figure 23

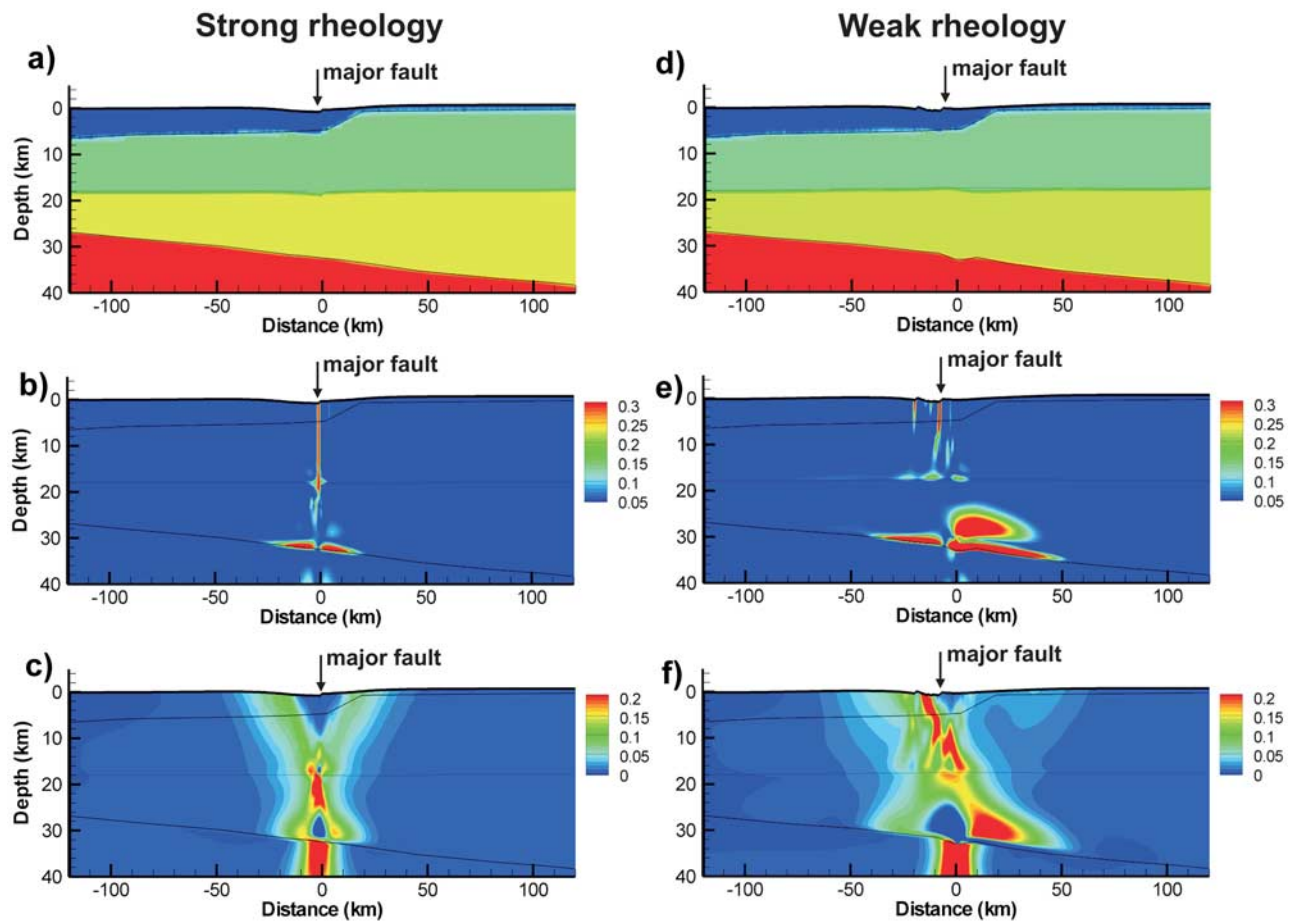


Figure 24. Distribution of crustal structure, shear strain, and extension at $t = 17$ Ma in the strong and weak crust models shown in Figure 23. Additionally, the lithosphere was thinned at $t = 12$ Ma; that is, the mantle lithosphere was then replaced by the asthenosphere with a temperature of 1200°C , and a few kilometers of transform-perpendicular (east-west) extension were added. For details, see *Sobolev et al.* [2005]. (a–c) Model with strong crust. (d–f) Model with weak crust. Figures 24a and 24d show crustal structure. Note the occurrence of Moho flexure in the weak-crust model (Figure 24d). Figures 24b and 24e show distributions of absolute values of shear strain $|e_{12}|$ (horizontal shear at horizontal plane or vertical shear at vertical plane). Note intensive shear deformation in the lower crust in the weak crust (Figure 24e) and its absence in the strong crust (Figure 24b). Figures 24c and 24f show distributions of transform-perpendicular extension (e_{11} component of the finite strain tensor). Note asymmetric deformation in the weak crust (Figure 24f). The vertical exaggeration is 2. Modified from *Sobolev et al.* [2005], copyright 2005, Elsevier.

1183 Besides meteoric fluids, infiltration of brines is observed
 1184 in all three faults. Upward fluid migration is found at the
 1185 SAF (mantle fluids [*Kennedy et al.*, 1997; *Wiersberg and*
 1186 *Erzinger*, 2007]) and to a minor extent at the DST
 1187 (hydrothermal fluids from crystalline basement in area II
 1188 [*Janssen et al.*, 2005]). However, as pointed out previously
 1189 in this section, the geochemical and geophysical studies of

Wiersberg and Erzinger [2007] and *Becken et al.* [2008] 1190
 1191 indicate that mantle fluids seem to migrate through the
 1192 northeastern fault block of the SAF on the North American
 1193 Plate, while the seismically defined SAF is not very
 1194 permeable in the vertical direction. In addition to fluid
 1195 migration under open system conditions, local fluid redistri-
 1196 bution under closed conditions is found at the DST and

Figure 23. Setup of thermomechanical models with two pure strike slip models and strong and weak crust, respectively. Lithospheric strength, cumulative finite strain, and viscosity are shown. (a) General model setup with boundary conditions and lithospheric structure. Results for pure strike-slip models of cold and thick lithosphere with (b–d) a strong crust and (e–g) a weak crust. Figures 23b and 23e show lithospheric strength prior to deformation ($t = 0$ Ma) on a W–E profile. Figures 23c and 23f show the distribution of the cumulative finite strain (square root of the second invariant of the finite strain tensor) at $t = 17$ Ma, corresponding to present conditions. Thin white lines indicate major lithospheric boundaries from *DESERT Group* [2004] and *Mechie et al.* [2005]. Figures 23d and 23g present the distribution of viscosity at $t = 17$ Ma. No vertical exaggeration. Modified from *Sobolev et al.* [2005], copyright 2005, Elsevier.

t2.1 **TABLE 2. Miocene to Recent Stratigraphic Units in Northern Arava/Araba^a**

t2.2	Age	Stratigraphic Unit	Absolute Age (Based on)
t2.3	Pleistocene to recent	Zehiha formation/Lisan formation, alluvium, sand dunes, playa deposits	1.5–0.5 Ma (basalt flows, Jordan)
t2.4	Pliocene	Arava formation/Mazar formation	3.7–1.7 Ma (basalt flows, Jordan), 6 Ma (En Yahav dike, Israel, and basalt flows, Jordan)
t2.5	Miocene	Hazeva formation/Dana conglomerate	19–21 Ma (dolerite dikes, Karak)
t2.6	Eocene	Avedat group/Um Rijam Chert-Limestone formation	

t2.7 ^aAfter *Sneh et al.* [1998] and *Avni et al.* [2001]. Age dating after *Steinitz and Bartov* [1991].

1197 the SAF [*Evans and Chester*, 1995], but it has to be
1198 pointed out again that the SAF (at Parkfield) and the AF
1199 do not act as important conduits for deep fluids.

1200 [39] The principal difference in the geodynamic evolution
1201 of the DST and SAF is that the SAF is evolving within a
1202 thin lithosphere and in a system with a highly variable
1203 thermal state. During the northward migration of the Men-
1204 docino triple junction along the Pacific–North America
1205 plate boundary, the slab being subducted beneath it was
1206 replaced by hot asthenospheric material in a slab window or
1207 slab gap [e.g., *Dickinson and Snyder*, 1979; *Furlong et al.*,
1208 1989; *Wilson et al.*, 2005]. The transform deformation along
1209 the northern part of the plate boundary thus developed
1210 simultaneously with thermal reequilibration of the litho-
1211 sphere. Along the southernmost part of the transform, active
1212 rifting with high heat flow is occurring. Along the central
1213 SAF, older rocks are offset and heat flow is moderate
1214 [*Lachenbruch and Sass*, 1980] (see also Table 3). Another
1215 key difference between the DST and SAF systems is that the
1216 North American lithosphere must have been affected by the
1217 passage of the Mendocino triple junction below it [*Goes*
1218 *et al.*, 1997; *Furlong and Govers*, 1999]. This makes the

geodynamic situation at the SAF strongly three-dimensional, 1219
since right-lateral shear is accommodated at the SAF, but 1220
compressional deformation occurs over a much wider belt, 1221
basically across the entire Coast and Transverse Range 1222
province. In contrast to the SAF, at the AF, strike-parallel 1223
changes in lithospheric structure can be ignored in a first- 1224
order approximation [*Sobolev et al.*, 2005]. 1225

4. CONCLUSIONS AND OUTLOOK 1226

[40] On the basis of the multidisciplinary approach pre- 1227
sented here, we find that the main characteristics identified 1228
for the Arava/Araba Fault (AF) segment of the DST are as 1229
follows: (1) a narrow, subvertical zone cutting through the 1230
entire crust, extending into the lithosphere, where the fault 1231
zone width ranges between 20 and 30 km; (2) a Moho depth 1232
that smoothly increases from 26 to 39 km from NE to SW; 1233
(3) a subhorizontal lower crustal reflector east of the AF; 1234
(4) the existence of several large faults in the upper crust in 1235
the vicinity of the AF, none of which has a large zone of 1236
decreased seismic velocities/high conductivity typical for 1237
damage zones; (5) that the AF acts as a barrier to fluids and 1238

t3.1 **TABLE 3. Comparison of Several Characteristics of the Dead Sea Transform and the San Andreas Fault^a**

t3.2	Characteristic	DST	SAF
t3.3	Age of onset	≈20 Ma	28 Ma
t3.4	Total offset of main FZ	105 km	480 km (1000 km if distributed shear east to Colorado Plateau allowed)
t3.5	Offset in last 5–6 Ma	≈60 km	≈300 km
t3.6	Present displacement rate	≈5 mm/a	35 mm/a in northern/central CA and a few mm/a convergence, 30 mm/a in southern CA and 10 mm/a convergence
t3.7	Current thickness of lithosphere	≈80 km	≈30 km in northern CA (crust only), 35+ km in southern CA
t3.8	Quake size	≈M7.5	M8
t3.9	Quake frequency	100–1000 years	100–400 years
t3.10	Number of faults currently active	1–2	3–4 depending on segment
t3.11	Number of inactive faults in FZ	3–4	3–4+
t3.12	Thickness of fault core, active fault	5–20 m	100+ m
t3.13	Fluid involvement/barriers and fault manifestation at depths	0–1 km: shallow fluid layers separated by barrier 1–4 km: west of AF, brine body stops at fossil AF	3 km: gas and fluid barrier at SAFOD 30+ km: interpreted fluid channel east of SAF at SAFOD, dips steeply west
t3.14		10–50+ km: vertical FZ ≈ 10 km west of AF	15–20 km: subhorizontal fluid-rich layer connecting to SAF (southern CA)
t3.15		5–30 km: AQF dipping east	
t3.16	Heat flow	low to moderate	high at north and south ends, moderate in center

t3.18 ^aDST, Dead Sea Transform; SAF, San Andreas Fault; FZ, fault zone; AF, Arava/Araba Fault; AQF, Al Quwayra Fault; SAFOD, San Andreas Fault Observatory at depth. See also Figure 3.

1239 abrupt changes in lithology across the AF occur down to a
 1240 depth of 4 km; (6) ongoing tectonic activity in shallow
 1241 sediments in at least two AF strands (fossil and active AF);
 1242 (7) a damage zone of individual faults with widths of only
 1243 5–20 m; and (8) a mainly meteoric origin of fluids in the
 1244 AF. As pointed out in more detail at the end of each section,
 1245 these eight features (see Figure 3 for a representation at
 1246 different scales) could not have been identified reliably if
 1247 only one (geophysical) method had been used. Only an
 1248 interdisciplinary approach integrating findings from seis-
 1249 mology, seismics, electromagnetics, gravity, geothermics,
 1250 petrology, geochemistry, and field mapping based on sur-
 1251 face geology, multispectral satellite images, and remote
 1252 sensing gives sufficient independent confirmation, which
 1253 then is tested by thermomechanical modeling. Considering
 1254 the dynamics of the DST we find that deformation began in
 1255 a 20–40 km wide zone, which later became concentrated in
 1256 one or two major faults in today's Arava/Araba Valley. Until
 1257 ~5 Ma ago, other, now inactive fault traces in the vicinity of
 1258 the present-day AF took up lateral motion. Then, together
 1259 with a rearrangement of plates, the main fault trace shifted
 1260 to its present position. The AF is the main active fault of the
 1261 DST system, but it has only accommodated up to 60 km of
 1262 the overall plate motion. The AF is a system of almost pure
 1263 strike-slip faulting and the shear deformation is controlled
 1264 by the location of minimum mantle strength, which pro-
 1265 duced the subvertically mechanically weak decoupling zone
 1266 extending through the entire lithosphere.

1267 [41] A comparison of the AF with the SAF and other
 1268 large faults shows that the width of the AF is significantly
 1269 smaller than that of other major faults, most likely because
 1270 of less total slip on the AF (up to 60 km). The narrow
 1271 damage zone at the AF could be the result of a faulting
 1272 mechanism where strain is extremely localized. Prominent
 1273 similarities between the DST and the SAF, on the other
 1274 hand, are that both have an asymmetry in subhorizontal
 1275 lower crustal reflectors and deep reaching deformation
 1276 zones and show flower structures in transpressional regimes
 1277 at local scale. Such features are most likely fundamental
 1278 characteristics of large transform plate boundaries.

1279 [42] Large transform faults represent varying structure
 1280 and dynamics in both time and space. However, a number
 1281 of common features can be detected. The dominant mech-
 1282 anisms for the development and occurrence of large fault
 1283 systems are as follows: (1) the large-scale forces acting on
 1284 the plates and their transfer into the contact area between the
 1285 plates, where the faults then develop; (2) the previous
 1286 history and the geochemical, thermal, and petrological
 1287 fabrics inherited during geological time in the contact area
 1288 of the plates; and (3) the local hydrological and geological
 1289 setting controlling the strength of the fault(s) as they
 1290 develop in time. The dependence of the development of
 1291 faults on such a large range of spatial and temporal scales,
 1292 their specific inherited structures, and the limited number of
 1293 large faults studied in detail up to now makes it nontrivial to
 1294 isolate fundamental and characteristic features of large
 1295 transform faults. The only way to gain a better understand-
 1296 ing of the controlling forces and settings that determine how

1297 faults come into being, evolve, and become inactive is the
 1298 multidisciplinary study of more active and fossil faults. Of
 1299 special interest for future studies is also how more compli-
 1300 cated 3-D structures like deep sedimentary basins/pull-apart
 1301 basins (e.g., the Dead Sea Basin) evolve along large faults.
 1302 This is the topic of an ongoing study.

[43] **ACKNOWLEDGMENTS.** The work presented here by
 the Dead Sea Rift Transect (DESERT) Group was supported by
 Deutsche Forschungsgemeinschaft (DFG), the GeoForschungs-
 Zentrum Potsdam (GFZ), and the Alexander von Humboldt Foun-
 dation. N. Kukowski and G. Fuis provided constructive reviews of
 an earlier version of this manuscript, though their views do not
 necessarily agree with those presented here. We thank the editor M.
 Manga, the reviewers W. Mooney and T. Stern, and an anonymous
 reviewer for comments on an earlier version of this manuscript and
 A. Siebert for her help in drafting Figures 1, 2, 4, 11, 14, 17, 19, 20,
 and 21, most of them created with the GMT software [Wessel and
 Smith, 1998]. We thank the National Ministry of Infrastructure of
 Israel, the Natural Resources Authority of Jordan, and the An-Najah
 National University in Nablus, Palestine, for their support during the
 field work. We thank our contractors the Geophysical Institute of
 Israel, the Site Group (Jordan), and the Chemical and Mining
 Industries (Jordan) for their excellent work under difficult logistic
 conditions. The instruments for the field work were provided by the
 Geophysical Instrument Pool of the GFZ. The experiments were
 funded by the DFG, the GFZ, and the Minerva Dead Sea Research
 Centre. For more information on DESERT and the Dead Sea
 Integrated Research Project (DESIRE), its successor, see [http://](http://www.gfz-potsdam.de/portal/-?%3Bpart=CmsPart&docId=2434369)
www.gfz-potsdam.de/portal/-?%3Bpart=CmsPart&docId=2266356 (DESIRE).

[44] The Editor responsible for this paper was Michael Manga.
 He thanks Walter Mooney and Timothy Stern as technical
 reviewers and one anonymous cross-disciplinary reviewer.

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