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Imaging of Subsurface Faults using Refraction Migration with Fault Flooding

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ABSTRACT

We propose a novel method for imaging shallow faults by migration of transmitted refraction arrivals. The assumption is that there is a significant velocity contrast across the fault boundary that is underlain by a refracting interface. This procedure, denoted as refraction migration with fault flooding, largely overcomes the difficulty in imaging shallow faults with seismic surveys. Numerical results successfully validate this method on three synthetic examples and two field-data sets. The first field-data set is next to the Gulf of Aqaba and the second example is from a seismic profile recorded in Arizona. The faults detected by refraction migration in the Gulf of Aqaba data were in agreement with those indicated in a P-velocity tomogram. However, a new fault is detected at the end of the migration image that is not clearly seen in the traveltimes tomogram. This result is similar to that for the Arizona data where the refraction image showed faults consistent with those seen in the P-velocity tomogram, except it also detected an antithetic fault at the end of the line. This fault cannot be clearly seen in the traveltimes tomogram due to the limited ray coverage.

INTRODUCTION

In this paper we focus on seismic refractions to delineate and image the shallow subsurface faults. Figure 1 shows several types of waves recorded in a seismic survey. The seismic reflections are typically used to estimate the subsurface reflectivity distribution from the

depths of 5 m to 50 m in engineering surveys and 0.25-1.0 km to 10 km in exploration surveys (Yilmaz, 2001). Surface waves, on the other hand, probe no deeper than about 1 km for most surveys, and can only be used to estimate the S-wave velocity to that depth. Unfortunately, reflectivity estimates are typically not possible at depths from 0 to 5 m because near-surface scattering and environmental noise interfere with the shallow-reflection signals at short source-receiver offsets. However, the first-arrival traveltimes of refractions (see Figure 1) can be picked and inverted for the shallow P-wave velocity distribution. The resulting velocity image is known as the P-velocity tomogram, but typically is too smooth to unambiguously indicate the locations of near-surface faults with sharp velocity contrasts.

An alternative imaging method is refraction migration (Aldridge and Oldenburg, 1992; Zhang, 2006), but it cannot image faults because it is restricted to imaging the refraction interfaces, not the fault boundaries. For example, the head-wave raypath abruptly changes its angle of propagation across the vertical fault, and is characterized by a sharp inflection in the slope of the observed refractions. Therefore, the refraction ray does not propagate along the fault surface so the subsurface fault plane will not be imaged by standard refraction migration.

To remedy this problem, we propose a novel method for detecting near-surface faults: migration of transmitted refraction arrivals for shallow-fault detection. We denote this method refraction migration with fault flooding where the materials across one side of the fault are flooded to the other side as if there was no fault. The key identifier of a near-surface fault in the data is that the trace position of the inflection (fault location) does not change with source position even if the source is on the opposite side of the fault plane. In the fault-flooded refraction migration image, the fault boundaries will be automatically imaged in the final migrated section so the interpretation of the faults will not strongly depend on the interpreter's judgment.

This paper is organized into four sections. Following the introduction, we present the theory for refraction migration with fault flooding. There are several possible implementations, but the one used in this paper is presented as a workflow. The next section validates our refraction method with tests on both synthetic data and field data from two seismic

surveys. All tests demonstrate the effectiveness in detecting fault locations and estimating dip angles in the final refraction image. The final section is a summary of the benefits and limitations of refraction migration with fault flooding.

THEORY

Refraction migration is used to delineate the location of refracting interfaces (Hill, 1987; Aldridge and Oldenburg, 1992; Zhang, 2006; and many others). The diffraction-stack migration equation for refractions is given by

$$m(\mathbf{x}) = \sum_{\omega} \sum_r \sum_s D(\mathbf{g}, \mathbf{s}, \omega) e^{-i\omega(\tau_{sx} + \tau_{xg})}, \quad (1)$$

which is similar to that for Kirchhoff migration of reflections except the frequency-domain data $D(\mathbf{g}, \mathbf{s}, \omega)$ consist of windowed refraction arrivals. Here, $e^{-i\omega(\tau_{sx} + \tau_{xg})}$ is the migration kernel for a source at $\mathbf{s}$ and a geophone at $\mathbf{g}$ and τ_{xg} is the time of a refraction signal to propagate from the trial image point at $\mathbf{x}$ to the receiver at $\mathbf{g}$. Likewise, τ_{sx} is the refraction propagation time from the source at $\mathbf{s}$ to the trial image point at $\mathbf{x}$. For example, if the point on the actual refraction ray path is denoted by $\mathbf{x}_o \in B_r$, where B_r is the set of points along the raypath in Figure 2a, then the normalized refraction can be described by $D(\mathbf{g}, \mathbf{s}, \omega) = e^{i\omega(\tau_{sx_o} + \tau_{x_o g})}$ in the frequency domain. Setting the trial image point $\mathbf{x}$ to be at $\mathbf{x}_o$ in the migration equation 1 will cancel out the phase term in $D(\mathbf{g}, \mathbf{s}, \omega)$ and lead to coherent summation over all frequencies for all trial image points along the refraction raypath. For a sufficiently dense distribution of sources and receivers only the migration image $m(\mathbf{x})$ along the first-Fresnel zone of the refraction raypath B_r in Figure 2a will emerge with strong intensity.

Refraction migration is restricted to imaging the refraction interfaces, not imaging the fault boundaries. For example, the head-wave raypath (denoted as the actual ray in Figure 2b) abruptly changes its angle of propagation across the vertical fault, and is characterized by a sharp *inflection* in the slope of the observed refractions. Therefore, the transmitted refraction ray does not propagate along the fault plane so the fault will not be

imaged by refraction migration. The key identifier of a near-surface fault is that the trace position of the inflection at **A** does not change with source position even if the source is on the opposite side of the fault.

A field data example is shown in Figure 3 for different shot positions where the inflection point (jagged portion of yellow line) is roughly at the same geophone position $X \approx 75\text{ m}$ for different shot positions. This suggests the possibility of a fault plane just below the inflection point.

To image the fault boundary in Figure 2b, we propose a fault-flooding scheme that floods one side of the fault with the same material as on the other side. In this way the red virtual events in Figure 2b can be generated that have no change in their moveout velocity across the fault. The migration algorithm then images the intersection point $\mathbf{x}_o$ of the virtual and actual raypaths in Figure 2b as long as their arrival times agree with one another at $\mathbf{x}_o$.

The next section describes the algorithmic details of fault-flooding migration, which is then followed by numerical results.

Workflow for Fault-Flood Migration of Refractions

We present the algorithm for migrating virtual and recorded refractions to the blue fault boundary in Figure 2b. For convenience, the canonical velocity model is illustrated in Figure 2b where the shots are in the leftside medium with velocity v_1 and the rightside medium has the slowest velocity v_2 . The deepest layer has the fastest velocity denoted by v_3 .

1. The refraction traveltimes are picked and inverted to get the refraction tomogram and an estimate of $v(x, z)$. The refraction events are identified whose $x - t$ moveout curve sharply bends at a common surface location for shot gathers on one side of the fault. This sharp bend might be accompanied by a sharp change in the traveltimes moveout of the surface waves and the presence of backscattered diffractions. The observed $D(\mathbf{g}, \mathbf{s}, \omega)_{obs}$ and virtual $D(\mathbf{g}, \mathbf{s}, \omega)_{virt}$ refraction data on the surface can be

mathematically represented by

$$D(\mathbf{g}, \mathbf{s}, \omega)_{obs} = e^{i\omega(\tau_{sx_o}^1 + \tau_{x_o g}^2)}; D(\mathbf{g}', \mathbf{s}, \omega)_{virt} = e^{i\omega(\tau_{sx_o}^1 + \tau_{x_o g'}^1)}, \quad (2)$$

where $\mathbf{g}'$ is the virtual geophone location associated with the observed data recorded at $\mathbf{g}$, and the superscripts in the traveltimes indicate the traveltime in the medium with velocity v_1 or v_2 . The location $\mathbf{x}_o$ indicates the intersection point of the observed and virtual refraction rays at the fault. If the leftside of the fault material is homogeneous, then the virtual red refractions in Figure 2b are created by a straight-line continuation of the moveout curve from the leftside to the rightside of the fault.

2. Correlate the recorded and virtual traces to give the correlated data, which in the frequency domain takes the form

$$\begin{aligned} \Phi(\mathbf{g}, \mathbf{g}', \mathbf{s}, \omega) &= D(\mathbf{g}, \mathbf{s}, \omega)_{obs} D(\mathbf{g}', \mathbf{s}, \omega)_{virt}^* \\ &= e^{i\omega(\tau_{x_o g}^2 - \tau_{x_o g'}^1)}. \end{aligned} \quad (3)$$

3. The migration kernel that cancels the phase of the frequency-domain correlogram in equation 3 trace at $\mathbf{x}_o$ is $e^{-i\omega(\tau_{x_o g}^2 - \tau_{x_o g'}^1)}$. Therefore, the migration equation for imaging the fault is given by

$$m(\mathbf{x}) = \sum_{\omega} \sum_{\mathbf{s}} \sum_{\mathbf{g}} \sum_{\mathbf{g}'} \Phi(\mathbf{g}, \mathbf{g}', \mathbf{s}, \omega) e^{-i\omega(\tau_{x_o g}^2 - \tau_{x_o g'}^1)}. \quad (4)$$

Equation 4 is equivalent to interferometrically detecting the location of a point source (Schuster, 2009) at $\mathbf{x}_o$, except now the source is at the location of the fault boundary. This migration procedure has poor resolution properties perpendicular to the recording plane, but this problem can be overcome by using a wavepath-like migration algorithm described in the next step.

4. Wavepath migration (Sun and Schuster, 2001) can be implemented by using a local

slant stack to estimate the slope of the recorded refractions at $\mathbf{g}$ to construct the *observed* refraction ray that terminates at $\mathbf{g}$. The *observed* refraction arrival is then backprojected along the *observed* refraction ray with the v_2 velocity to give the *observed* backprojected data $O(\mathbf{x}, \mathbf{g}, \mathbf{s}, \omega)$. The virtual arrival at $\mathbf{g}'$ is also backprojected along its ray to give $V(\mathbf{x}, \mathbf{g}', \mathbf{s}, \omega)$ using the v_1 velocity. The intersection of these two rays at $\mathbf{x}$ defines a possible location of the fault boundary. If the corresponding arrival times for both the backprojected observed and virtual rays at the intersection point $\mathbf{x}$ also agree within $1/4$ of a period then the migration energy $O(\mathbf{x}, \mathbf{g}, \mathbf{s}, \omega) * V(\mathbf{x}, \mathbf{g}', \mathbf{s}, \omega)$ at $\mathbf{x}$ is retained. Otherwise it is not used. Therefore, the formula for the final migration image is given by

$$m(\mathbf{x}) = \sum_{\omega} \sum_{\mathbf{g}} \sum_{\mathbf{g}'} \sum_{\mathbf{s}} T(\mathbf{x}, \mathbf{s}, \mathbf{g}, \mathbf{g}', \omega) V(\mathbf{x}, \mathbf{g}', \mathbf{s}, \omega) O(\mathbf{x}, \mathbf{g}, \mathbf{s}, \omega)^*, \quad (5)$$

where the threshold function $T(\mathbf{x}, \mathbf{s}, \mathbf{g}, \mathbf{g}', \omega) = 1$ if the two backprojected rays intersect at $\mathbf{x}$ and the backprojected virtual and observed traveltimes agree to within $1/4$ of a period. Otherwise $T(\mathbf{x}, \mathbf{s}, \mathbf{g}, \mathbf{g}', \omega) = 0$. This procedure is closely related to imaging salt flank boundaries with transmitted PS waves (Sheley and Schuster, 2000).

NUMERICAL RESULTS

Both simulations and field data are now used to validate the efficacy of imaging fault boundaries with the fault-flooding migration scheme.

Data for all synthetic models are calculated using finite-difference solutions to the 2D acoustic wave equation. In the synthetic models we computed 150 common shot gathers (CSG) with 150 receivers per shot gather, where both the receivers and shots are spaced at 30 m intervals.

Vertical Fault Model

The first synthetic example represents a vertical fault model (Figure 4a), where Figure 5a shows an example of the computed common shot gathers (CSG). A synthetic data set is generated using the same survey geometry for the non-faulted (flooded) velocity model (Figure 4b) in order to get the virtual data depicted in Figure 5b. In other words, one side of the fault is flooded with the same material as the other side.

Then, the refractions are isolated to get the observed (Figure 5c) and virtual refraction arrivals (Figure 5d). These observed and virtual refraction traces are correlated with one another to get the correlograms for all shot positions shown in Figure 5e. These correlograms for all shot positions are migrated by equation 5 with the dual image conditions of ray intersection and arrival-time coincidence. This gives the final migration image shown in Figure 5f, where most of the fault plane is well imaged.

Tilted Fault Model

The second synthetic example represents a vertical fault model (Figure 6a), where Figure 7a shows an example of the computed common shot gathers (CSG). The shots and receivers are located on the surface, and synthetic data are generated for the non-faulted velocity model (Figure 6b) to get the virtual data in Figure 7b.

Then the refractions are isolated to get the observed (Figure 7c) and the virtual refraction arrivals (Figure 7d). The windowed virtual and observed traces are cross correlated with one another to get the correlograms in Figure 7e. Equation 5 is used to migrate ¹ the correlograms to get the final migration image of the fault in Figure 7e. It is obvious that the image provides the correct dip angle and location of the fault. However, the fault image extends below the actual refractor interface because the crosscorrelation imaging condition has poor vertical resolution (Schuster, 2009). This problem can be mitigated if the refraction tomogram is used to outline the refracting interface.

¹In practice, the migration velocity is obtained from the P-velocity tomogram.

Vertical Fault Model with Topography

The third synthetic example represents a vertical fault model (Figure 8a), where Figure 9a shows an example of the computed common shot gathers (CSG). In this case the sources and receivers are located on the irregular topographic surface shown in Figure 8, which might affect the accuracy in locating the fault by migration. Here the elevation difference between the highest and lowest source-receiver locations is 100 m.

To test the influence of topography on the migration image, a synthetic data set is generated for the non-faulted velocity model in Figure 8b. The refraction migration method is then used to get the virtual data set in Figure 9b using the same coordinates for the sources and receivers. Then, from the observed and the virtual data sets, the refractions are isolated to get the observed (Figure 9c) and virtual refraction arrivals (Figure 9d). The refraction traces are correlated with one another to get the correlograms for all shot positions shown in Figure 9e. These correlograms are then migrated by equation 5 with the dual-image conditions of ray intersection and arrival-time coincidence. This gives the final migration image shown in Figure 9f, where the correct fault location is misplaced due to the effect of topography as shown with the red arrows, where the correct location is indicated by the black dashed line.

To avoid mispositioning the fault in the migration image, elevation statics are used to redatum the data to a horizontal datum at the highest elevation. This will eliminate the false kink in the refractions due to topography. The virtual data are then created and migrated to get the migration image. This procedure is demonstrated in Figure 10a, where the air is filled with the same materials on the left side of the fault and all the sources and receivers are placed on the surface, then a synthetic data set is generated for the velocity model in Figure 10a using a finite-difference solutions (see Figure 11a) using the same parameters. The refraction migration method is then used to get the virtual data set in Figure 11b using the same coordinates for the sources and receivers. Then, from the observed and the virtual data sets, the refractions are isolated to get the observed (Figure 11c) and virtual refraction arrivals (Figure 11d). The refraction traces are correlated with one another to get

the correlograms for all shot positions shown in Figure 11e. These correlograms are then migrated by equation 5 with the dual-image conditions of ray intersection and arrival-time coincidence. This gives the final migration image shown in Figure 11f, where the correct fault location is indicated by the red arrows.

Gulf of Aqaba Field Data

The fault-flood migration algorithm is applied to field data collected near the Gulf of Aqaba. Figure 12 shows the location of the 2D survey where a total of 120 common shot gathers were recorded. Each shot gather has 120 traces with shot and receiver intervals of 2.5 meters along a straight recording line that is 297.5 m long. Traces were recorded using a 1 ms sampling interval for a total recording time of 0.3 seconds. A 90 kg accelerated weight drop was used as the seismic source, with 10 to 15 stacks at each shot location.

To generate the virtual data, the observed CSG's were modified to resemble the refraction arrivals of the non-faulted model by flattening the kinks associated with the normal faults. This was done by shifting the kinked areas to be in line with the slope of the other first arrivals, as depicted in Figures 13a and 13b. Muting was also performed in order to isolate the refraction arrivals in Figures 13c and 13d. The observed and virtual traces were then correlated with one another to give the correlograms shown in Figure 14.

After migration, the image representing the possible locations of the faults in the Aqaba region is depicted in Figure 15c. The migration image of the fault agrees with the dips and locations of the faults indicated in the P-wave velocity tomogram (Figure 15a) with its ray-path density (Figure 15b). In this example, the wavepath migration was not used and all correlograms were simply migrated by a Kirchhoff migration method.

Arizona Field Data

In March 2008, University of Utah researchers (Liu, 2009) carried out a 2-D high resolution seismic survey perpendicular to the Washington fault scarp near the Arizona-Utah border. Figure 16 shows the seismic survey site. The 2-D seismic data were collected using 96

vertical-component geophones spaced 1 m apart for a total line length of 95 m. Seismic sources, using a 7.2 kg sledgehammer striking a small metal plate, were initiated at every second geophone and stacked five times for each shot position to improve the signal-to-noise ratio of each record. The traces were recorded by a 120-channel Bison data recorder.

In this case the elevation difference between the highest and lowest receiver location is 8 m as shown in Figure 17. To correct for topography, we apply elevations statics to the data to avoid the false kinks in the traces due to the topography. To apply the statics corrections we flooded the air with the first-layer velocity of 300 m/s and then, by using Snell's law, the time shift is computed for the raypath from each source to receiver as shown in Figure 18. This shift is applied to the recorded data, which shows that the observed kinks are due to the existing faults and not due to the topographic change.

After elevation corrections, the virtual data were generated in the same way by straightening the kinks in the observed data as in the Gulf of Aqaba dataset. This was done by shifting the kinked areas to be in line with the slope of the other first arrivals, as depicted in Figures 19a and 19b. Muting was also performed in order to isolate the refraction arrivals in Figures 19c and 19d. Then we undid the statics corrections on both the observed and the virtual datasets and applied muting again to isolate the refraction arrivals as in Figure 20. These muted traces were correlated with one another to give the correlograms shown in Figure 21.

After migration, the image representing the possible locations of the faults in the Arizona region is depicted in Figure 22c. The location of the faults in the migration image agree with the P-wave velocity tomogram (Figure 22a), and suggest similar dip angles. The normal fault indicated at the offset of about 60 meters is also seen in a nearby trench log (Hanafy et al., 2015). In this case an antithetic fault on the left corner of the refraction image is indicated in Figure 22c, which is not visible on the P-wave velocity tomogram due to the limited-offset coverage as shown in the ray-path density diagram (Figure 22b). Our simulations show that the limited data coverage also affects the migration image at the two ends of the survey, but not nearly as much as in the velocity tomogram.

CONCLUSIONS

A novel algorithm is presented for imaging a near-surface fault boundary by migrating recorded and virtual refractions to their points of intersection. The virtual refractions are created by continuing the incident refractions across the fault after flooding it with the velocity of the incident medium. The key assumptions are that one side of the fault has a much faster velocity than the other side, and there is an underlying refractor interface. Tests with synthetic data and recorded refraction data validate the efficacy of this method.

This method is valid when the following evidence indicates a fault at the near surface:

1. A sharp change in the slope of refraction arrivals occurs at the same geophone location in CSGs with different shot positions on either side of the fault.
2. A sharp change in the moveout of surface wave arrivals is visible in the shot gather at the inflection point of the refractions.
3. Backscattered diffractions originate at the inflection point in the recorded surface waves.

The limitations of this method are the following:

1. A strike-slip fault is not likely to create a large enough velocity contrast at the near surface to generate a strong change in the moveout of refractions.
2. The procedure for windowing around the refraction events and creating virtual arrivals on one side of the faults requires user intervention.
3. The recording line should be perpendicular to the fault scarp otherwise 3D data must be collected to image the fault plane.
4. Elevation corrections should be applied before using this technique, otherwise the correct location of the fault might be misplaced.

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LIST OF FIGURES

1. a) Schematic diagram showing different types of waves after generating seismic waves using an impulsive source. b) Seismic section after trace by trace normalization showing the different types of recorded waves (Pasasa et al., 1998).

2. Common shot gathers for the a) two-layer and b) two-layer-fault models. The red virtual refractions in b) are obtained by creating the red arrivals (with $1/v_1$ slope) to the right of the inflection point at **A**. Here, $v_2 < v_1 < v_3$. The position $\mathbf{x}_o$ must be at the fault if the backprojected virtual and actual rays intersect at $\mathbf{x}_o$ and their arrival times at $\mathbf{x}_o$ agree with one another.
3. Common shot gathers for different shot positions. The yellow curve depicts the move-out pattern of refractions for a shot at the 110 m position, and is overlaid on the shot gathers with 115 m and 120 m shot locations. The inflection point is at $X \approx 75$ m, which suggests the possible surface-projection location of a hidden near-surface fault. These shot gathers were recorded near Olduvai gorge and are only used as an illustration of an inflection point which hardly changes with shot position.
4. Velocity models for a) the faulted and b) the fault-flooded models.
5. Synthetic data for a) the faulted and b) fault-flooded models. c) and d) are the same data after windowing about the refractions. e) Correlograms of observed and virtual data. f) Final migration image of fault plane.
6. Velocity models for the hidden a) tilted fault and b) the fault-flooded models.
7. Synthetic data for the a) hidden fault and b) fault-flooded models. c) and d) are the same except only the windowed refraction arrivals are shown. e) Correlograms computed from the observed and virtual data. f) Final migration image of the buried fault plane.
8. Velocity models for the a) vertical fault and b) the fault-flooded models with topography (elevation differences).
9. Synthetic data for the a) vertical fault and b) fault-flooded models with topography (elevation differences), where the true kink due to the fault is highlighted with a black circle while the false kink due to topography is highlighted by a red circle. c) and d) are the same except only the windowed refraction arrivals are shown. e) Correlograms computed from the observed and virtual data. f) Final migration image of the fault

plane misplaced due to the effect of topography as shown by the red arrows, where the true location of the fault is shown by a black dashed line.

10. Velocity models for the a) vertical fault and b) the fault-flooded models after filling the air with the same materials on the left side of the fault.
11. Synthetic data for the a) vertical fault and b) fault-flooded models after placing all the sources and receivers on the surface. c) and d) are the same except only the windowed refraction arrivals are shown. e) Correlograms computed from the observed and virtual data. f) Final migration image of the fault plane is placed at the correct location as indicated by the red arrows.
12. Map showing the Gulf of Aqaba with a zoomed picture showing the survey line.
13. CSG with kinks a) highlighted and b) removed. c) and d) only contain the windowed refractions.
14. Correlograms for the shot at $X = 20\text{ m}$.
15. a) P-wave traveltime tomogram obtained by inverting the first-arrival traveltimes using a dynamic smoothing steepest descent technique (Hanafy et al., 2014), the red circle shows the low velocity colluvial wedge associated with the fault. b) Ray-path density diagram c) Final migration image for the Gulf of Aqaba data showing the same fault locations as the red dashed lines.
16. The map of the Washington fault showing the survey site with a zoomed picture of the survey line (Liu, 2009).
17. Elevation values at each source-receiver location.
18. Time shift required for statics corrections for each source-receiver location.
19. CSG after applying the statics corrections with kinks a) highlighted and b) removed. c) and d) only contain the windowed refractions.
20. CSG a) with kinks and b) removed. c) and d) only contain the windowed refractions after undoing the static correction (i.e. the left hand side has kinks due to both effects

from the topography and the faults as well, while the right hand side (fault-flooded) has only kinks due to the effect of topography only).

21. Correlograms for a shot at $X = 95\text{ m}$.
22. a) P-wave traveltimes tomogram (constructed by inverting the first arrival traveltimes using a dynamic smoothing steepest descent technique) with the possible fault locations as black dashed lines. b) Ray-path density diagram. c) Final migration image of the Arizona data showing the faults at the same locations. An antithetic fault on the eastern corner of the image is shown which is not visible on the P-wave traveltimes tomogram due to the limited offset coverage.

Figure 1: a) Schematic diagram showing different types of waves after generating seismic waves using an impulsive source. b) Seismic section after trace by trace normalization showing the different types of recorded waves (Pasasa et al., 1998).

Figure 2: Common shot gathers for the a) two-layer and b) two-layer-fault models. The red virtual refractions in b) are obtained by creating the red arrivals (with $1/v_1$ slope) to the right of the inflection point at **A**. Here, $v_2 < v_1 < v_3$. The position x_0 must be at the fault if the backprojected virtual and actual rays intersect at x_0 and their arrival times at x_0 agree with one another.

Figure 3: Common shot gathers for different shot positions. The yellow curve depicts the moveout pattern of refractions for a shot at the 110 m position, and is overlaid on the shot gathers with 115 m and 120 m shot locations. The inflection point is at $X \approx 75 \text{ m}$, which suggests the possible surface-projection location of a hidden near-surface fault. These shot gathers were recorded near Olduvai gorge as an illustration of an inflection point which hardly changes with shot position.

Figure 4: Velocity models for a) the faulted and b) the fault-flooded models.

Figure 5: Synthetic data for a) the faulted and b) fault-flooded models. c) and d) are the same data after windowing about the refractions. e) Correlograms of observed and virtual data. f) Final migration image of fault plane.

Figure 6: Velocity models for the hidden a) tilted fault and b) the fault-flooded models.

Figure 7: Synthetic data for the a) hidden fault and b) fault-flooded models. c) and d) are the same except only the windowed refraction arrivals are shown. e) Correlograms computed from the observed and virtual data. f) Final migration image of the buried fault plane.

Figure 8: Velocity models for the a) vertical fault and b) the fault-flooded models with topography (elevation differences).

Figure 9: Synthetic data for the a) vertical fault and b) fault-flooded models with topography (elevation differences), where the true kink due to the fault is highlighted with a black circle while the false kink due to topography is highlighted by a red circle. c) and d) are the same except only the windowed refraction arrivals are shown. e) Correlograms computed from the observed and virtual data. f) Final migration image of the fault plane misplaced due to the effect of topography as shown by the red arrows, where the true location of the fault is shown by a black dashed line.

Figure 10: Velocity models for the a) vertical fault and b) the fault-flooded models after filling the air with the same materials on the left side of the fault.

Figure 11: Synthetic data for the a) vertical fault and b) fault-flooded models after placing all the sources and receivers on the surface. c) and d) are the same except only the windowed refraction arrivals are shown. e) Correlograms computed from the observed and virtual data. f) Final migration image of the fault plane is placed at the correct location as indicated by the red arrows.

Figure 12: Map showing the Gulf of Aqaba with a zoomed picture showing the survey line.

Figure 13: CSG with kinks a) highlighted and b) removed. c) and d) only contain the windowed refractions.

Figure 14: Correlograms for the shot at $X = 20$ m.

Figure 15: a) P-wave traveltime tomogram obtained by inverting the first-arrival traveltimes using a dynamic smoothing steepest descent technique (Hanafy et al., 2014), the red circle shows the low velocity colluvial wedge associated with the fault. b) Ray-path density diagram. c) Final migration image for the Gulf of Aqaba data showing the same fault locations as the red dashed lines.

Figure 16: The map of the Washington fault showing the survey site with a zoomed picture of the survey line (Liu, 2009).

Figure 17: Elevation values at each source-receiver location.

Figure 18: Time shift required for statics corrections for each source-receiver location.

Figure 19: CSG after applying the statics corrections with kinks a) highlighted and b) removed. c) and d) only contain the windowed refractions.

Figure 20: CSG a) with kinks and b) removed. c) and d) only contain the windowed refractions after undoing the static correction (i.e. the left hand side has kinks due to both effects from the topography and the faults as well, while the right hand side (fault-flooded) has only kinks due to the effect of topography only).

Figure 21: Correlograms for a shot at $X = 95$ m.

Figure 22: a) P-wave traveltimes tomogram (constructed by inverting the first arrival traveltimes using a dynamic smoothing steepest descent technique) with the possible fault locations as black dashed lines. b) Ray-path density diagram. c) Final migration image of the Arizona data showing the faults at the same locations. An antithetic fault on the eastern corner of the image is shown which is not visible on the P-wave traveltimes tomogram due to the limited offset coverage.

Highlights

A novel algorithm is presented for imaging a near-surface fault boundary by migrating recorded and virtual refractions to their points of intersection. The virtual refractions are created by continuing the incident refractions across the fault after flooding it with the velocity of the incident medium. The key assumptions are that one side of the fault has a much faster velocity than the other side, and there is an underlying refractor interface. Tests with synthetic data and recorded refraction data validate the efficacy of this method.

Keywords:

Seismic; Tomography; Refraction Migration; Fault Imaging; Fault-Flooding.