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Application of Focal-Time Analysis for Improved Induced Seismicity Depth Control: A Case Study from the Montney Formation, British Columbia, Canada

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Right Running Head: Focal-time Analysis & Montney Formation

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ABSTRACT

Characterization of induced seismicity and associated microseismicity is an important challenge for enhanced oil recovery and development of tight hydrocarbon reservoirs. In particular, accurately correlating hypocenters of induced events to stratigraphic layers plays an important role in understanding the mechanisms of fault activation. Existing methods for estimating focal depth, however, are prone to a high degree of uncertainty. A comprehensive analysis of inferred focal depths is applied to induced events that occurred during completions of horizontal wells targeting the Montney Formation in British Columbia, Canada. Our workflow includes a probabilistic, nonlinear global-search algorithm (NonLinLoc), a hierarchical clustering algorithm for relative relocation (GrowClust) and depth refinement using the recently developed focal-time method. The focal-time method leverages stratigraphic correlations between P-P and P-S reflections to eliminate the need for an explicit velocity model developed specifically for hypocenter depth estimation. We show that this approach is robust in the presence of noisy picks and location errors from epicenters obtained using a global-search algorithm, but it is limited to areas where multicomponent 3-D seismic data are available. A novel

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method to determine statics corrections is developed here, to ensure that both passive seismic observations and 3-D seismic data share a common datum in areas of moderate to high topography. Our results highlight the importance of transverse faults, which appear to provide permeable pathways for activation of other faults at distances of up to 2 km from hydraulic fracturing operations.

INTRODUCTION

Fluid injection is one of the main mechanisms for triggering induced seismicity in the subsurface (Ellsworth, 2013; Eaton, 2018). With increasing development of unconventional reservoirs during the past decade, there has been a surge in induced seismicity linked to hydraulic fracturing (e.g., Holland, 2013; Skoumal et al., 2015; Bao and Eaton, 2016; Lopez-Comino et al., 2018; Schultz et al., 2018; Galloway et al., 2018; Meng et al., 2019). The Triassic Montney Formation, located in northeastern British Columbia and northwestern Alberta, is one of the most prospective resource plays in North America, with estimated recoverable resources of 449 Tcf natural gas, 14.5 billion barrels of natural-gas liquids, and 1.1 billion barrels of oil (NEB 2013). As unconventional

oil and gas development in this play has progressed, significant cases of induced seismicity linked to hydraulic fracturing (HF) have been documented (BC Oil and Gas Commission, 2014; Mahani et al., 2017). Detailed characterization of induced seismicity is therefore important for responsible development of this resource play.

Induced seismicity can be recognized by spatial and temporal correlation between the triggering factors and earthquake occurrence (Holland, 2013; Atkinson et al., 2016). Seismic events observed during HF operations are normally in the range of local magnitude $M_L < 0$, herein referred to as microseismicity. In contrast, rare events associated with the activation of faults during HF are referred to as anomalous induced events. Reliable depth location of anomalous induced events and associated microseismicity can provide important clues regarding the mechanism of fault activation, which in turn can be used to understand and mitigate seismic hazards.

Most methods for determining hypocenters (locations) of induced events rely on fitting P- and S-wave arrival time picks using a calibrated velocity model (Lomax et al., 2001; Trugman and Shearer, 2017; Meng et al., 2019). However, in cases where surface

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seismograph networks are used to determine event hypocenters, the uncertainty region for focal-depth solutions is often highly elongate in the depth direction and is sensitive to the presence of sharp velocity contrasts (Lomax et al., 2009; Jones et al., 2014). In addition, challenges for building velocity models include lack of sufficient data, particularly for models that include anisotropy and 3-D heterogeneity (Zhang et al., 2015). These characteristics mean that uncertainty in focal depth is often significantly greater than uncertainty in the epicenter location (i.e., horizontal coordinates of an event), even if a well calibrated velocity model is available.

In this case study, we use a recently developed data-driven technique called the focal-time method (Poulin et al., 2019), which partly overcomes these uncertainties and has potential to reduce depth scatter and other velocity-model artifacts. Rather than inverting for event depths using a known (or assumed) P- and S-wave velocity model, the focal-time method uses stratigraphic correlations between P-P and P-S reflections derived from three-component (3-C) seismic data. This method exploits the property that P-P and P-S reflections from surface sources, as well as waves radiated from

microseismic events, propagate through the same medium. The velocity model, used to process the seismic data, enters into the focal-time method indirectly, through a lookup-table approach that relates the P-wave time to depth. In our implementation of the focal-time method, we make use of event epicenters obtained using the NonLinLoc method, a direct global search algorithm (Lomax et al., 2001; Lomax, 2005), with further refinement based on the GrowClust method, a cross-correlation based approach for hypocenter relocation (Trugman and Shearer, 2017). Our study is located in areas of moderate to high topography, and we introduce a novel approach to determine statics corrections to ensure that both passive seismic observations and the P-P and P-S datasets share a common datum. Joint interpretation of the relocated seismicity distribution with 3-D seismic data highlights a series of transverse faults that appear to provide permeable pathways for hydraulic communication between the treatment wells and other faults.

DATASETS

This study uses a variety of datasets within our area of interest (AOI), including 3-D P-P and PP-PS seismic surveys, continuous monitoring with an array of seismograph

stations, a regional 2-D seismic line, and six wells with an extensive set of logs including dipole-sonic, density, and gamma logs. Figure 1 outlines the AOI, along with locations of wells, the monitoring array and seismic datasets. Two treatment wells (A and B) are located within the area of 3-D PP-PS seismic coverage (covering ~40 km²). We note that PP-PS datasets are much less common than conventional P-wave land 3-D datasets; however, at least in some areas, a great deal of converted-wave data have been acquired, but not processed. In the case of the present study, only the vertical-component data were processed initially; the unprocessed horizontal-component data are available and were processed as part of this case study.

Seismicity was monitored using a local array consisting of 27 seismograph stations, of which 24 were located within the PP-PS seismic coverage. These stations were buried to a depth of approximately 1 m and were comprised of a mix of five broadband seismometers and twenty-two 4.5 Hz three-component geophones. The stations were deployed to monitor two horizontal wells that were completed using hydraulic fracturing over a period of about two weeks. Commercial processing of the

data from the monitoring array yielded a catalogue of more than 1700 events. The events were processed by picking phase arrivals using the short-time-average/long-time average (STA/LTA) method (Earle and Shearer, 1994), which were manually reviewed to eliminate bad picks. Continuous data were further analyzed to detect other signals that were missed by the STA/LTA triggering algorithm, such as due to the occurrence of closely spaced event times. During initial processing, events were located using a grid-search method, based on an initial velocity model, and then were relocated using 3-D velocity model derived from a joint inversion of the 3-D isotropic velocity model, event location and station delays. Ideally, calibration sources with known locations, such as perforation shots, are used to validate the velocity model (Warpinski et al., 2015). For the dataset considered here, no perforation shots with reliable S-wave picks were observed. Moreover, calculated event depths were found to be highly sensitive to small changes in the velocity model, providing a strong motivation to obtain independent focal-depth estimates using a different approach.

For events observed during this treatment program, the maximum moment magnitude (M_W) was 1.7. In addition, the estimated Gutenberg-Richter b -value was 1.16 ± 0.4 (Figure 2), obtained using a maximum-likelihood method with 95% confidence limits (Aki, 1965). The b -value corresponds to the slope of the semilogarithmic magnitude-frequency distribution and is a measure of the relative abundance of low- to high-magnitude events (Eaton, 2018). The estimated b -value of ~ 1 is lower than expected for operationally induced microseismicity during hydraulic fracturing ($b \sim 2$) but is typical of b -values observed for active fault systems (e.g., Eaton and Maghsoudi, 2015).

METHODS

NonLinLoc

NonLinLoc is a probabilistic, direct global-search nonlinear package based on method of Lomax et al. (2001) and Lomax (2005). This method determines the epicenter location based on a probability density function (PDF), which is obtained by global sampling algorithms. It uses two likelihood functions (or minimum misfit) to build the PDF;

the first function includes the least-squares, L_2 norm and the second function uses the equal differential-time (Lomax, 2005). For sampling, the Oct-tree sampling method is incorporated, which is based on recursive subdivision and sampled to create a cascade structure in a way that the spatial density of samples follows the target PDF values (Lomax, 2005). The inputs for this method are the phase arrival times, velocity model, and reference station locations. In this case study, a 1-D velocity model was used to calculate hypocenter locations, obtained using P- and S-wave velocity logs from the nearest well.

GrowClust

GrowClust is a program that relocates earthquake hypocenters using differential times obtained by waveform cross-correlation (Trugman and Shearer, 2017). The method uses a common set of seismic stations along with hierarchical clustering to simultaneously group and relocate events. The inputs to this program are (1) phase arrival times, (2) cross-correlation values for P and S body-wave phases, (3) a velocity model for computing predicted differential travel times, and (4) initial event locations. Differential

times help to relocate the events, whereas the cross-correlation values are utilized to weigh the data by waveform similarity (Trugman and Shearer, 2017). The program is computationally efficient and numerically stable. In addition to determining relocated event locations, GrowClust resamples the input data for bootstrap uncertainty analysis.

Focal-time analysis

Focal-time analysis is a data-driven approach to estimate the focal depth for induced and microseismic events. This algorithm does not provide a solution for epicenter locations, which need to be obtained using another method. The focal-time technique requires P- and S-wave arrival time picks, 3-D multicomponent surface seismic data to correlate P-P and P-S reflection times and one or more well ties for calculating the time-to-depth relationship (Poulin et al., 2019). In essence, the focal-time method is based on the kinematic principle that waves radiated from a point source in the subsurface are characterized by an S-P time difference at zero offset that is equal to the time difference between zero-offset P-S and P-P reflections from the same location (Figure 3). Based on this principle, the focal-time method combines regression analysis of S and P time picks

of microseismic events with a lookup-table approach to determine an equivalent P-wave time, which is then converted to depth based on independent constraints such as a seismic well tie (Poulin et al., 2019). Unlike NonLinLoc and GrowClust, the focal-time method does not require an explicit velocity model for hypocenter depth location, although this information is indirectly incorporated into the algorithm during processing of the P-P and P-S seismic volumes.

The workflow used in our implementation of the focal-time method is shown in Figure 4. The microseismic/induced events and seismic data are analyzed in parallel streams, which are merged to derive the focal time for each event. The horizons for several geological markers are interpreted and registered using P-S synthetic seismograms from density and dipole sonic logs. The P-P and P-S horizon registration is an essential step in the focal-time workflow since the difference between P-P and P-S horizons are linked to the differential time between P and S phases in the microseismicity or induced events via the Garotta equation (Garotta and Grange, 1987):

$$\frac{V_P}{V_S} = 2 \left(\frac{t_{PS}}{t_{PP}} \right) - 1, \quad (1)$$

where V_P and V_S are the average P- and S-wave velocity from the surface to a picked horizon. The difference between P-S horizon t_{PS} time and P-P seismic horizon t_{PP} time is given by (Poulin et al., 2019)

$$\Delta t_S \cong t_{PS} - t_{PP} = \frac{V_P - V_S}{V_P V_S} z, \quad (2)$$

where z is the depth of the interface.

The main concept of focal-time analysis is that, for a given microseismic event with phase arrival time picks t_P and t_S , the value of Δt_S from equation 2 is equal to $t_S - t_P$ at zero offset,

$$t_S - t_P|_{x=0} \cong t_{PS} - t_{PP}, \quad (3)$$

where x denotes the offset distance from the event epicenter to the recording stations. In practice, the offset distance is based on an independently determined epicenter location (the surface location directly above the hypocenter). The intercept is then obtained by linear regression in $t-x^2$ space, in which the moveout for both phases is approximately

linear (Poulin et al., 2019). One of the strengths of this approach is that it uses the $t_S - t_P$ times and so does not depend on the (uncertain) event origin times.

Re-datuming

The focal-time method requires datum adjustments to be applied so that the microseismic/induced events are referenced to the same elevation datum as the seismic data. Figure 5 is a schematic illustration of the datum difference between the reference datum of seismic survey and the induced events. Our goal is to re-datum all observations to the reference datum. These static corrections are especially important in the case of our AOI, due to the presence of moderate surface topography ($\pm 200\text{m}$) in this area.

In the P-S seismic processing stream, seismic horizons (in time) are required to re-datum from floating datum (the datum of three-component seismic volume) to the reference datum. It is typically difficult to identify P-S first breaks to constrain a replacement velocity for PS static solution. Thus, the P-P replacement velocity was used with residual statics applied to improve the imaging (Cary and Eaton, 1993). In our

workflow, we applied a time adjustment as follows. As illustrated in Figure 5, the $t_{PP-floating}$ (P-P time from floating datum to reference datum) was used to calculate the $t_{PS-floating}$ (P-S time from floating datum to reference datum) based on

$$\begin{aligned} t_{PS-floating} &= \frac{t_{PP-floating} + t_{SS-floating}}{2} \\ &= \frac{\left(1 + \frac{V_P}{V_S}\right)}{2} t_{PP-floating} \end{aligned} \quad (4)$$

In this equation, $t_{SS-floating}$ is calculated using

$$t_{SS-floating} = \left(\frac{V_P}{V_S}\right) t_{PP-floating} \quad (5)$$

We derived time adjustment based on the difference of $t_{PS-floating}$ and $t_{PP-floating}$ using

$$t_{adj} = t_{PS-floating} - t_{PP-floating} = \left(\frac{\left(\frac{V_P}{V_S} - 1\right)}{2}\right) t_{PP-floating} \quad (6)$$

where this adjustment time is applied for every PS and PP horizon. For instance, in the Montney Horizon (MNTN), the adjusted time difference of t_S and t_P , $(t_S - t_P)_{MNTN}$, can be obtained as

$$(t_S - t_P)_{MNTN} = MNTN_{PS} - MNTN_{PP} + t_{adj} \quad (7)$$

For the induced/microseismic processing stream, in order to have the same reference datum as seismic data, we used P-statics from the seismic data to obtain the adjustments needed for the P phase times, assuming vertical ray paths in the near surface. For S-wave times, we used S-statics calculated using

$$S_{statics} = P_{statics} \times \left(\frac{V_P}{V_S}\right) + PS_{residual} \quad (8)$$

In above equations, V_P/V_S ratio is the average V_P/V_S of PP-PS seismic data. In addition, it was found that applying a P-S residual static correction ($PS_{residual}$) is a necessary step to obtain robust results. This correction was obtained by correlating two distinct horizons (horizons that are clearly discernible and can be accurately picked) in receiver stacks, both with and without residual statics corrections. Figure 6 shows examples of time picks for two interpreted seismic horizons- derived from receiver stacks. More than 85% of the data required residual-statics adjustments in the range of ± 10 ms. Larger residual-statics corrections, as illustrated in Figure 6, were needed for about 15% of the AOI due to elevation fluctuations.

In the next step, a set of lookup tables is constructed that relate the $t_{PS}-t_{PP}$ time difference to the t_{PP} time in seismic workflow. To have a smooth interpolation table at each location, we used spline interpolation between picked horizons. In addition, we extrapolated the lookup tables, as needed, to account for events and stations that fall just outside the area of P-S data coverage. A 1-D lookup table was computed corresponding to the x - and y - position of each common midpoint bin center throughout the AOI. Taken together, this method provides a data-driven approach to incorporate 3-D subsurface heterogeneity that implicitly incorporates velocity heterogeneity and seismic anisotropy (Poulin et al., 2019).

Turning next to the microseismic/induced workflow, the next step is to calculate the zero-offset time difference (Δt). For this purpose, a simple $\ell - x^2$ regression method is used, where x represents the epicentral distance for each sensor. Since the P- and S-wave moveout for induced events is approximately hyperbolic, a $\ell - x^2$ method is expected to result in an approximately linear relationship for the respective time picks. This method also provides the associated uncertainty, based on linear regression

analysis. We reduced uncertainty by excluding a small number of outliers in either the P- or S-wave time picks based on the random sample consensus (RANSAC) algorithm (Meer et al., 1991). Finally, the two separate streams are merged, to obtain an estimate of the focal time for each event. In this step, the lookup table result is used to map the re-datumed Δt values into the corresponding t_{PP} values, and then into depth based on time-depth relationships from 3-D seismic well ties (Poulin et al., 2019).

VALIDATION USING SYNTHETIC DATA

Methods for estimating hypocenter locations are subject to errors and uncertainties, arising from factors such as inaccuracy of the time picks or a lack of calibration of the velocity model. The focal-time method was proposed and tested by Poulin et al. (2019), who demonstrated that this method could reduce uncertainty in estimation of focal depth arising from velocity-model uncertainty, including 3-D heterogeneity and anisotropy. Here, we present additional validation of this approach, based on our implementation of the focal-time method in which epicenters obtained using a global-search algorithm are used to determine the offset distance for regression

analysis as part of the focal-time method. Figure 7a shows a simple 2-D velocity model that we used for synthetic testing. The model contains dipping, isotropic layers with depth-varying V_P/V_S ratio. Known locations for four test events are used to quantify location uncertainty related to use of an inaccurate velocity model. In this analysis, it is assumed that a P-wave time-depth relationship is available at the wellbore; in practice, this can be achieved using a well tie to seismic data. Several scenarios are considered: (1) a stratified (1-D) velocity model is used to determine initial hypocenter locations, based on lateral extrapolation of layering that exists at the well; (2) the velocity model used to locate events is perturbed, relative to the true velocity model, by scaling the layer velocities by $\pm 5\%$. These approximations are reasonable; for example, the use of a 1-D velocity model to compute hypocenter locations is common for studies of both induced and natural seismicity (e.g. Michelini et al., 1998; Musumeci et al., 2003; Yukutake et al., 2010; Eaton et al., 2018). In addition, in cases where calibration shots are available, velocity corrections of 5% (or more) are applied to log-derived seismic velocities to compute hypocenter locations (e.g., Eyre et al., 2019). These corrections may reflect velocity-

dispersion or anisotropic effects that cause well-log and seismic velocities to differ (Eaton, 2018).

One of the first steps in focal-time analysis is registration of P-P and P-S reflections to construct a lookup table. Figure 7b compares the zero-offset P-P and P-S (converted wave) traveltimes for the four interfaces based on the velocity model in Figure 7a. To match the horizons approximately, the P-S time has been divided by a factor of 1.5, the appropriate scaling for a V_P/V_S ratio of two (e.g., Weir et al., 2018). Once the P-P and P-S horizons are correlated, this is used to create a table of P-S minus P-P time, versus P-P reflection time. This table is interpolated (and extrapolated) linearly to provide a set of lookup tables that are regularly sampled, covering the spatial extent and time range of interest. In this 2-D synthetic example, the spatial extent is linear along the profile, but in the case study discussed below, the lookup tables cover a 2-D area of interest. Similarly, the time-depth relationship at the wellbore is formulated to allow P-P time to be converted into depth. In practice, lookup table for P-S minus P-P times can extend deeper than the well control, depending on the presence of correlatable deeper reflections, whereas the

time-depth relationship must be extrapolated to handle events below the available well control. Here, a linear extrapolation is used.

For each of the four test locations, a series of synthetic tests was implemented based on: (1) the actual 2-D velocity model with dipping layers; (2) a 1-D horizontally layered velocity model, extrapolated away from the well, and (3) $\pm 5\%$ adjustments to the 1-D and 2-D velocity models. This procedure provided six estimated hypocenter locations for each test event, for a total of 24 synthetic test cases. For each of these simulations, aleatory uncertainty was incorporated by generating 100 realizations by adding uniform random errors to the synthetic times of up to ± 10 ms for P-wave picks and ± 20 ms for S-wave picks. The initial hypocenter locations were obtained using a least-squares global search on a 10m grid, where the optimal hypocenter location was chosen based on the ℓ_2 norm applied to the summed residuals, based on calculated P and S times. The resulting location uncertainties are summarized in Figure 8. The results of this analysis indicate that:

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1. If the correct velocity model is used, the least-squares global search (LSGS) algorithm obtains the correct hypocenter locations (as expected), with small aleatory scatter due to noisy picks.
2. If a 1-D velocity model is used in the presence of dip, the x -locations based on LSGS method are systematically displaced in an up-dip direction.
3. The x -locations of the calculated hypocenters based on LSGS method are relatively insensitive to the accuracy of the velocity model (i.e., whether it is too high or too low). This assumes low-to-moderate lateral heterogeneity.
4. Conversely, for the LSGS method the depth (z) errors are strongly affected by whether the velocity model is erroneously high or low. On the other hand, depth errors are relatively insensitive to inclusion of dip in the velocity model.

Focal time was determined using the method described above, using the epicenter location derived from LSGS analysis. For all of the scenarios tested (apart from use of the exact 2-D velocity model, where the LSGS method yields the correct location within

uncertainty), errors in focal depths calculated using the focal-time method are significantly lower, in general, than those obtained using the LSGS method. Moreover, unlike the LSGS locations, the focal-time depth errors do not bear any systematic relationship to inaccurate scaling of the velocity model used to compute hypocenters. This lack of sensitivity arises because the focal-time method is referenced to the P-P and P-S seismic data, based on the simple principle that the seismic reflections and the waves generated by point sources propagate in the same medium. In this analysis, most of the focal-time depth error arises solely from the use of an approximate time-depth function based on a single well tie. The extrapolation of this time-depth function becomes increasingly inaccurate away from the well. Indeed, intermediate calculation of P-P times (focal times) are more accurate than the focal depth, since they are not affected by an uncertain time-depth relationship. In the absence of comprehensive velocity-model calibration, we observe that co-rendering with 3-D seismic data (in P-P time) of focal-time hypocenters yields more robust stratigraphic inferences than is possible with hypocenters obtained using a conventional least-squares approach.

RESULTS

NonLinLoc

The NonLinLoc method was applied using a 1-D velocity model that was created using well log data for one of wells in the area. Smoothing was applied to the velocity model to obtain a more stable solution for hypocenter locations, since the presence of sharp velocity contrasts can lead to artifacts in which events are tightly grouped along interfaces (Lomax et al., 2009). Figure 9 shows the velocity model used for NonLinLoc method. Phase time picks and locations of stations were two other inputs to the NonLinLoc program. Figure 10 shows the NonLinLoc results in map view and cross-section view; events and perforation locations are colored by time of occurrence.

In map view, the events fall into three main clusters. The largest cluster is concentrated along the two treatment wells; in addition, two seismicity clusters are evident, one nearly 2 km west of the nearest treatment well, and another nearly 1.5 km east of the nearest treatment well. For the cluster located west of the treatment wells, the

apparent sharp ~north-south edge is not an artifact of a boundary of the model domain, because the domain extends farther west. In cross section, the majority of the events are located below the treatment wells, but a significant fraction of the events (including the highest magnitude event) are located above the wells. The event locations exhibit a high degree of scatter, which renders it difficult to jointly interpret the results using 3-D seismic data or other geological information. In particular, the focal depths are highly scattered, reflecting an overall higher depth uncertainty depth compared with epicentral parameters (x - and y - locations). This scatter is exacerbated by challenges in accurately accounting for near-surface velocity variations at each of the seismograph stations, in the presence of moderate surface topography.

GrowClust

To reduce the scatter in event locations, we applied the GrowClust program to relocate the microseismic and anomalous induced events. The GrowClust algorithm is based on differential times and cross-correlation of waveforms observed at a set of common seismic stations. An example of the cross-correlation procedure for two example

events is shown in Figure 11. The velocity model for GrowClust is the same as the model used NonLinLoc program (Figure 9). Results of relocated events are shown in Figure 12.

The application of GrowClust reveals a number of notable features, especially in map view. The cluster of events located west of the treatment wells is now expressed as a series of NW-SE trending lineaments, while the cluster of events located east of the wells exhibits an apparent WNE-ESE linear fabric. In cross section, there is still a high degree of depth scatter, although a smaller fraction of events (and none of the larger events) are located above the treatment wells.

Focal-time analysis

Next, we applied focal-time analyses to refine the hypocenter depth calculations. In the first step of this procedure, we imported re-located event epicenters derived from the GrowClust method. We then calculated $t_S - t_P$ for the microseismicity and anomalous induced events, as discussed in the methodology section. Figure 3 shows an example of a $t_S - t_P$ calculation from a linear regression for zero-offset determination, where a random

sample consensus (RANSAC) algorithm (Meer et al., 1991) was applied to remove outliers in the data. The 95% confidence limit for P and S intercept errors was applied based on the RANSAC program. The zero offset ($t_S - t_P$) for induced events is equivalent to $t_{PS} - t_{PP}$ from seismic data, as shown in Figure 13. The intercept error is then used to determine the P-P time and depth based on lookup tables, as illustrated in Figure 13. Based on linear regression analysis, the average uncertainty value (95% confidence interval) for the calculated focal depths is 91 m. The magnitude of this uncertainty depends on the number of events and number and distribution of stations. When the number of events and station increase, the error in the focal-time analyses will decrease.

Figure 14 shows the results from application of the focal-time method to events within the AOI that met our quality control measures based on acceptance criteria (at least three picks and regression coefficient $r^2 > 95\%$ for both P- and S-waves, for a given event). The use of these acceptance criteria helps to eliminate weak events with very low signal-to-noise ratio. The map view of event epicenters is thus very similar to the results from GrowClust, except that only 1400 events are plotted in Figure 14 compared with

1700 events in Figure 12. There is, however, a conspicuous difference in terms of the depth distribution, since all of the events that met the criteria for focal-time analysis plot near, or below, the depth of the horizontal wells. We caution that for events with shallowest depth, the relatively small scatter in depths that is evident in Figure 13 reflects analytical uncertainties in the regression analysis ($\sim \pm 60\text{m}$) but does not capture all of the inherent uncertainties in the method, which are difficult to quantify. Nevertheless, this is a robust result, because after applying statics corrections (as described above) all of the valid events were characterized by a $t_S - t_P$ intercept difference greater than or equal to the $t_{PS} - t_{PP}$ time difference corresponding with the top of the Montney Formation. As elaborated below, this improved spatial delineation of seismicity is important in terms of both hydrocarbon recovery as well as hazard characterization.

DISCUSSION

Velocity models play an important role in hypocenter location methods. To illustrate the sensitivity of hypocenter locations to the velocity model, we generated an adjusted velocity model, by blocking the P- and S-wave velocities into 15 layers (compared with

300 layers in the original version) and then re-smoothing the models using a running-average filter (Figure 15). As shown in Figure 16, when event hypocenters are recomputed, using the same procedure as described previously but with the new velocity model, it leads to a substantial shift in event locations in terms of both event epicenters and focal depths. We acknowledge that, in practice, such velocity model adjustments would be subject to calibration tests; however, the aforementioned lack of reliably observed perforation shots presents a challenge for calibration. Rather, this example serves as a reminder of the high sensitivity of event locations to changes in the velocity model, which in this case consists of a simple adjustment to the smoothing used to construct the model. In this sense, the focal-time method provides an advantage, inasmuch as the computed focal depths rely primarily on horizon correlations derived from 3-D multicomponent seismic data rather than a specified velocity model.

Figure 17 shows an updated version of Figure 14, in which only induced events with $M_w \geq 0$ are illustrated. This shows that event depths estimated using the focal-time method (Figure 14) are concentrated at, or below, the level of horizontal wells A and B

within the Montney Formation, indicating that positive magnitude events are mainly located at the deeper locations. Although this depth distribution differs from typical operationally induced microseismicity, where the microseismic event distribution provides a proxy for hydraulic fracture (e.g., Boroumand and Eaton, 2015), this result is consistent with other induced-seismicity studies in the Montney trend in which traditional methods are used to obtain event locations (Davies et al., 2013; Mahani and Kao, 2018; Yu et al., 2019). It should be noted that the magnitude of completeness (minimum reliable earthquake magnitude) of our dataset is approximately -0.5 (Figure 2), which is considerably larger than the magnitudes of typical microseismicity during hydraulic fracturing (Eaton, 2018). This suggests that operationally induced microseismicity was generally too weak to be detected by the induced seismicity monitoring network used in this study. This result is also in good agreement with studies summarized by Rogers (2018), who found that virtually all of the large induced seismic events reported in the Montney play are associated with faulting within the underlying Belloy and Debolt Formations. Moreover, a recent numerical investigation of induced seismicity in the Montney play, which considered the geomechanical properties of the Montney and

adjacent layers, found that larger fault slip displacements occur in these underlying formations, which are characterized by significantly higher stiffness (Amini and Eberhardt, 2018). This stratigraphic/geomechanical relationship is also consistent with focal depths of induced events within the Duvernay play in Alberta, where some studies show that events are vertically displaced from the injection level, with the largest events in the massive Wabamun carbonate (Poulin et al., 2019; Eyre et al., 2019).

Interpretation of 3-D seismic data within the AOI reveals a number of faults adjacent to the treatment wells (Figure 18). Offset reflectors show that these steeply dipping faults extend through underlying layers up to at least the lowermost Montney Formation. One set of faults strikes NNW-SSE, parallel to the Late Cretaceous – Paleocene Rocky Mountains (situated west of the AOI). These faults, including two larger faults F1 and F2 as well as smaller faults (e.g. F3), are interpreted to be part of a buried contractional fault system contemporaneous with the Rocky Mountains. Other faults, including F4-F5 (Figure 18), transect the NNW-SSE faults; this second set of faults has uncertain age and is herein referred to as transverse faults. By overlaying re-located

seismicity (Figure 14) onto the mapped faults, some correspondence in orientation is observed. In particular, a conspicuous linear event cluster west of wells A and B is close to, and sub-parallel with, fault F1 mapped from 3-D seismic data. Another linear event cluster with a WNW-SSE trend, located mainly east of the treatment wells, appears to be parallel to transverse fault F4.

One perplexing question for understanding the distribution of induced seismicity associated with this treatment program is the mechanism for activation of fault structures, such as fault F2, located at distances of up to approximately 2 km from the HF operations. Although it is well established that elevated pore pressure can activate a fault, this distance significantly exceeds the expected reach of hydraulic fractures, typically a few hundred meters (Yu et al., 2014). Similarly, the low matrix permeability of the Montney siltstone ($10^{-5} - 10^{-4}$ mD) (Riazi et al., 2017) means that poroelastic diffusion through the Formation over this distance would require timescales much longer than the duration of the completion program (Atkinson et al., 2016). On the other hand, the pattern of seismicity found in this study, coupled with interpretation of 3-D seismic data, suggests

that transverse faults could play an important role in the transmission of elevated pore pressure (or stress perturbation) from the treatment well to offset faults. Indeed, one distinct possibility is that transverse faults, and/or associated fracturing within damage zones near these faults, could provide permeable pathways for communication of elevated pore pressure from the treatment well to the activated faults. Further research is required in order to assess this model.

CONCLUSIONS

This paper provides a case study of induced seismicity associated with hydraulic fracturing within the Montney Formation in northwestern Canada. We implemented the recently developed focal-time method to aid in constraining the depth of induced and microseismic events. For determining epicenter locations of induced events, we utilized the NonLinLoc direct global search method and then refined event locations using GrowClust. Both NonLinLoc and GrowClust require an explicit velocity model, which is a source of uncertainty for the depth locations of induced events. The focal-time method requires co-located, multicomponent surface seismic data to map the time difference

between correlated P-P and P-S horizons, which provides the basis for generating lookup tables to convert P- and S-wave time picks into depth. Synthetic tests conducted here show that this approach is robust, even in the presence of uncertain velocity models for the initial hypocenter location.

Caution is needed to ensure that both induced seismic and reflection seismic are referenced to the same datum. We have developed a new procedure to re-datum time picks of induced seismic events. Our results show that the focal-time method yields a depth-distribution of induced events that exhibits less scatter than other methods. Microseismic event locations are at or below the treatment depth, consistent with other field observations from the Montney Formation. Joint interpretation of faults mapped using 3-D seismic data, together with the distribution of induced events, suggests that WNW-trending transverse faults, which intersect the main NNW-trending reverse faults in this area, play an important role in the distribution of induced seismicity at distances of up to 2 km from the treatment wells.

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DATA AND MATERIALS AVAILABILITY

Data associated with this research are confidential and cannot be released.