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WORKSHOP
GLOBAL GEOPHYSICAL INFORMATICS WITH APPLICATIONS TO RESEARCH IN EARTHQUAKE PREDICTIONS AND REDUCTION OF SEISMIC RISK

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FOCAL DEPTHS AND SOURCE PARAMETERS OF THE ROCKY MOUNTAIN HOUSE EARTHQAKE SWARM FROM DIGITAL DATA AT EDMONTON

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Seismic records at Edmonton (EDM) and Suffield (SES) between January 1976 and February 1980 show 220 events with magnitudes less than 4 originating near Rocky Mountain House. Many of these events show well defined S_n, S_g, and P_g phases and a small variation in the difference of $S_g - S_n$ and $S_g - P_g$. Analysis of the theoretical travel times using a structure determined for central Alberta yields an average focal depth of 20 ± 5 km and an average epicentral distance of 175 ± 5 km southwest of Edmonton for 40 of these events. Because S_n was not clear on the remainder, it was not possible to get focal depths for all the events.

Seismic moments of 80 events with local magnitudes from 1.6 to 3.5 were found to be in the range of $6.6 \pm 2 \times 10^{18}$ to $7.9 \pm 2 \times 10^{20}$ dyn·cm ($6.6 \pm 2 \times 10^{11}$ to $7.9 \pm 2 \times 10^{15}$ N·cm). A relationship between local magnitude and seismic moment was $\log(M_0) = 1.3M_L + 16.6$. This is similar to that determined for California. Source radii, where they could be determined, were 500 ± 50 m and stress drops were 0.75 ± 0.75 bar (75 ± 75 kPa).

The energy release of 263 events recorded at EDM from the Rocky Mountain House area was 5.6×10^{17} erg (5.6×10^{10} J). The b value for this earthquake swarm was 0.8, similar to that observed in other parts of western Canada.

The depths of focus, the low stress drops, and the statistical similarity to other natural earthquake sequences suggest that at least part of the swarm is of a natural origin.

Les enregistrements sismiques à Edmonton (EDM) et à Suffield (SES) effectués entre janvier 1976 et février 1980 signalent 220 évènements de magnitudes inférieures à 4 et dont l'origine se situe près de Rocky Mountain House. Plusieurs de ces évènements sismiques présentent des phases bien définies de S_n, S_g et P_g et une faible variation de la différence $S_g - S_n$ et $S_g - P_g$. L'analyse de la durée des temps théoriques de propagation fondée sur une structure définie pour le centre de l'Alberta fournit pour 40 de ces évènements une profondeur moyenne de foyer à 20 ± 5 km et une distance moyenne de l'épicentre à 175 ± 5 km au sud-ouest d'Edmonton. Vu que S_n n'est pas défini clairement pour les autres évènements, il est impossible de déterminer leurs profondeurs de foyer.

Les moments sismiques pour 80 évènements de magnitudes locales variant de 1,6 à 3,5 sont compris entre $6{,}6 \pm 2 \times 10^{18}$ et $7{,}9 \pm 2 \times 10^{20}$ dyn·cm ($6{,}6 \pm 2 \times 10^{11}$ à $7{,}9 \pm 2 \times 10^{15}$ N·cm). La relation de la magnitude locale avec le moment sismique était $\log(M_0) = 1{,}3M_L \pm 16{,}6$. Cette relation est analogue à celle trouvée en Californie. Le rayon de la source, quand il fut possible de le déterminer, était entre 500 ± 50 m et les chutes de contrainte étaient de 0,75 ± 0,75 bar (75 ± 75 kPa).

L'énergie libérée par les 263 évènements enregistrés à EDM provenant de la région de Rocky Mountain House était $5{,}6 \times 10^{17}$ erg ($5{,}6 \times 10^{19}$ J). La valeur b de ce groupe de tremblement de terre était 0,8, semblable à celle observée pour d'autres régions de l'ouest du Canada.

Les profondeurs de foyer, les faibles chutes de contrainte et les données statistiques analogues à d'autres séquences de tremblements de terre naturels indiquent qu'au moins une partie du groupe étudié ici est d'origine naturelle.

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Introduction

Since 1976 the digital seismic station at Edmonton (EDM) has been detecting unusually numerous small seismic events apparently originating near Rocky Mountain House. This is a discussion of these data, augmented by some observations from other analog stations in western Canada. Regional epicentre locations for these events are near 52.26°N and 115.13°W (Fig. 1).

As a consequence of the fact that no destructive earthquakes have occurred, or appear likely to occur, in Alberta and Saskatchewan, the distribution of seismic stations in space and time in this area is not well suited to the study of small activity with local magnitudes of less than three (Milne *et al.* 1978). The best review of western Canadian seismicity (Milne *et al.* 1978) is a general survey. The activity near Bengough (Horner *et al.* 1973) has been well documented, but is not relevant to the Foothills of the Rocky Mountains. Significant earthquakes have occurred west of Rocky Mountain House (Rogers and Ellis 1979; Rogers *et al.* 1980) and north of Rocky Mountain House in Willmore Provincial Park (Nyland and Rebollar, in preparation), but these events are a considerable distance from the subject of this paper, an apparently localized swarm of seismic activity near Rocky Mountain House.

All of the activity discussed here is small. Much of it was not routinely archived anywhere except in Edmonton. We rely on $S-P$ times at Edmonton, some locations based on regional stations, and pronounced similarities in wave form to isolate events to the Rocky Mountain House region. We then examine the digital records maintained at Edmonton to identify S_n and P_n phases and analyse these. These S_n and P_n phases are not

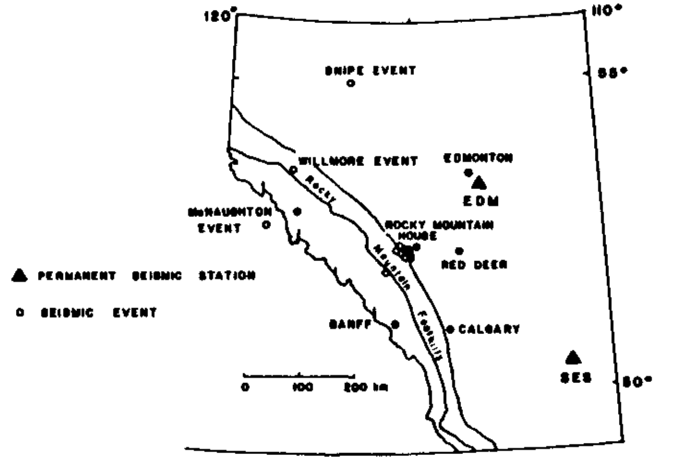

FIG. 1. Location of the Rocky Mountain House earthquake swarm. These events were published by the Earth Physics Branch, Department of Energy, Mines and Resources Canada, in its monthly bulletin.

usually evident on analog records but are very clear on the digital records at EDM. The availability of digital seismic data makes the routine analysis of spectral content of the S phases feasible. Although the results depend on assumptions about models of processes at the focus, these spectra suggest constraints on the nature of conditions in the source region of the swarm.

EDM is fortunate in being separated from Rocky Mountain House by a relatively simple, well determined, velocity structure. The station at Suffield (SES), 350 km from the epicentral area, is also separated from Rocky Mountain House by a similar simple structure and also sees some of this activity. It can be used to help in locations. With this in mind we made occasional use of the stations at Fort St. James (FSJ), 650 km from the epicentral area, and Penticton (PNT), 460 km from the epicentral area. Obviously records from other areas (such as Mica Creek array, 200 km from the epicentral area) provide additional information, but accurate interpretation of such data requires the analysis of propagation effects through rather complex structures and is beyond the scope of this paper.

It is perfectly true that better location of this activity would be desirable. Unfortunately, most of it is too small to locate well without a local network, which is not available. Use of data from other nearby stations such as Mica leads to only marginal improvement in location. We demonstrate that interesting things can be said about the source depth of the activity from an analysis of S_n and P_n recorded in the digital data archives at EDM.

We suggest in what follows that the seismic swarm has several similarities to natural activity. If it is in fact a natural, as opposed to an induced, phenomenon, an explanation of its cause is desirable. This cause must lie in the stress states in the region. Recent evidence (Bell and Gough 1979; Gough and Bell 1981) shows that significant compressive stress exists perpendicular to the Rocky Mountains. The small earthquakes that occur along the Canadian Rockies could be evidence of release of the associated internal energy.

This stress could be due to regional isostatic adjustment of

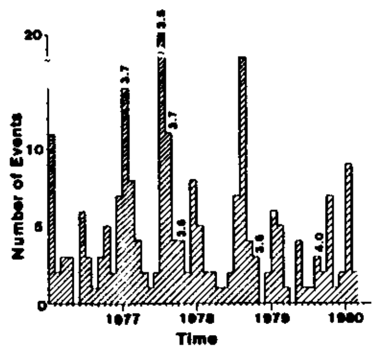

FIG. 2. Histogram of local seismic activity ($S-P$ times between 19.5 and 22.5 s) detected at EDM since 1976.

the Rocky Mountains and the foreland basin. Beaumont (1981) considered a model of lithosphere flexure in this area under laterally migrating loads and predicted the bend of the basement and consequently a deep root of the Rocky Mountains. Monger and Price (1979), compiling the work of several authors, suggested a Mohorovičić discontinuity (Moho) depth of 50 km. Therefore, this possible flexure of the crust could create enough stresses to generate earthquake activity in the basement of the Rocky Mountains. It is also possible that the stress reflects the stress induced by motion of the North American Plate (Nyland 1981). The directions derived by Gough and Bell are consistent with simple models of intraplate stress (Solomon *et al.* 1980).

Data acquisition

An average of five events per month, apparently originating in the Foothills of the Rocky Mountains (Fig. 2), has been detected on analog records at EDM since 1976. There were random bursts of up to 25 such events. This is unusually high activity compared with the local activity detected at EDM from 1964 to 1976.

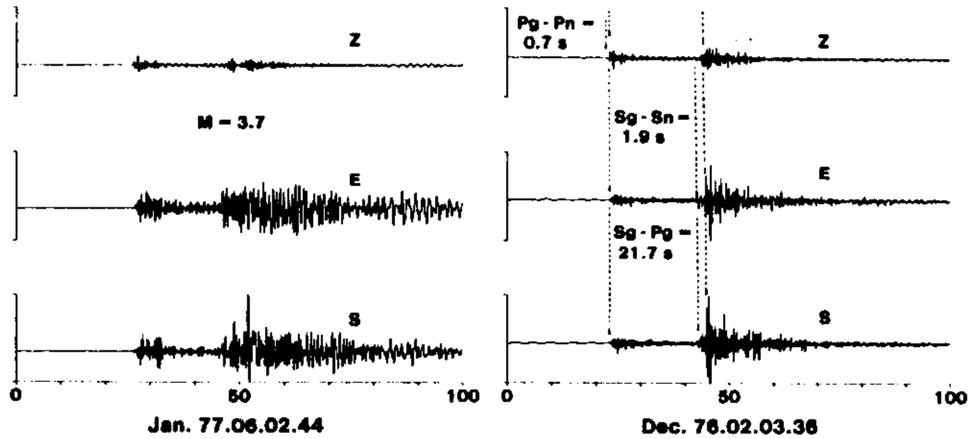

FIG. 3. Events recorded at Edmonton from the Rocky Mountain House earthquake swarm (left, complex event; right, simple event).

Some of the events of the swarm are simple, with well defined P_g, S_n, and S_g phases. However, some of the large events (Fig. 3) are complicated, showing similar amplitudes of P and S waves. Usually, P_n is barely above the noise level. Neither the complex nor the simple events appear on the long-period records. This suggests relatively deep foci.

From the travel-time curves we know approximately the distance and depth of the Rocky Mountain House activity. The approximate location was calculated using a few $S-P$ times at SES. A typical value of $S-P$ at SES is 34.7 s, which corresponds to an epicentral distance of 295 km using an average velocity of V_p of 6.5 km/s.

Origin times for this kind of activity are quite uncertain, but this does not mean that differential travel times are uncertain. These differences between observed phases can provide valuable information on the events being recorded. In the analysis that follows we use differential times only.

From January 1976 to February 1980, it was possible to read $S_g - S_n$ and $S_g - P_g$ in 41 events out of 220. Reading errors of P_g and P_n are of the order of 0.05 s when the onset is clear and 0.1 s or greater for S_g and S_n when a clear phase was recorded. The average time difference for $S_g - S_n$ was 1.9 ± 0.2 s and for $S_g - P_g$ was 21.7 ± 0.7 s. The average $P_g - P_n$ difference of 0.6 ± 0.1 s was based on 6 of the 41 events. This group of events shows remarkably consistent differential travel times, and as a consequence we feel justified in treating the data as a group. The average of the differential times should be representative of the group and it is this representative value we analyse.

Analysis of refracted phases

These average differential travel times have implications for the average depth of seismicity at Rocky Mountain House as observed at EDM. In order to estimate source depth we assume a structure derived from seismic refraction profiles, calculate travel-time curves for P_g, S_g, P_n, and S_n as a function of source depth, and use the average $S_g - S_n$, $P_g - P_n$, and $S_g - P_g$ times to constrain the depths. Because $S_g - S_n$ and $S_g - P_g$ were tightly constrained, we can place limits on the depths from which the seismicity originates if our structure is correct. We have tested the entire range of possible plains structures and conclude that no reasonable structure permits shallow source depths for the portion of the Rocky Mountain House seismicity that has observable S_n at Edmonton.

This conclusion is important. It seems highly unlikely that seismicity at depths exceeding 10 km is triggered by the relatively small changes associated with hydrocarbon recovery operations on the Strachan gas field, which is near the swarm. The events studied at EDM are most likely not induced. This does not preclude the possibility that other activity exists and is related to hydrocarbon recovery. The possibility that the basement structures near the Foothills are seismically active, albeit at a low level, also has interesting implications for the geodynamics of the North American Plate.

Several observational constraints can be placed on the problem. The first, and most important, is the requirement that S_n must arrive before S_g. This must hold no matter what velocity model is used and no matter what source depth is used. The second derives from the spread in $S_g - S_n$. Although it was 0.2 s in our data, we accepted solutions in which deviations of up to 0.35 s appeared. A third constraint, which is also crucial, is that P_n arrive before P_g. Here again we accept the pessimistic view that a 0.35 s deviation from an observed average is acceptable.

Given these constraints, we must solve an inverse problem that has no unique solution. We determine the range of velocity models, source depths, and epicentral distances that fit our observations. In order to explore the problem we first derived what appeared to be a reasonable velocity model for the problem and then searched for a source depth and a distance to the activity that satisfied our average of $S_g - S_n$ and $S_g - P_g$.

The crust is nearly 45 km thick in much of southern Alberta (Clowes and Kanasewich 1970; Chandra and Cumming 1972). Ganley and Cumming (1974) found a possible Moho from

TABLE 1. The velocity structure (Richards and Walker 1959)

Layer thickness (km)	V_p (km/s)	V_s (km/s)
2.0	3.6	1.81
1.5	6.1	3.07
28.5	6.2	3.51
13.0	7.2	4.15
Half space	8.2	4.73

TABLE 2. A modified version of the velocity structure

Layer thickness (km)	V_p (km/s)	V_s (km/s)
1.3	2.68	1.35
1.3	4.59	2.31
29.4	6.20	3.57
13.0	7.20	4.15
Half space	8.20	3.75

TABLE 3. A selection of predicted differential travel times for one plains velocity model

Depth (km)	ED (km)	$P_c - P_n$	$S_c - S_n$	$S_c - P_c$
5	170	−1.06	−1.71	20.90
5	185	−0.47	−0.68	22.68
10	170	−0.51	−0.75	20.92
10	185	0.07	0.27	22.70
15	170	0.06	0.23	20.95
15	185	0.64	1.25	22.73
20	170	0.65	1.27	21.00
20	185	1.23	2.28	22.78

reflection records at 35 km near Edmonton. However, those profiles are too far from the Rocky Mountain House activity to be useful. The closest reversed seismic refraction structure to the earthquake swarm is that determined by Richards and Walker (1959) from an approximately north–south refraction profile near 113.5°W and between 50.8 and 51.9°N (Table 1).

At a distance corresponding to that between Rocky Mountain House and Edmonton, the $S_g - P_g$ time does not depend on the depth of focus. It is thus possible to adjust this depth to satisfy observations of refracted S (S_n) that we obtained from digital data archives at Edmonton. It was, however, not always possible to match refracted P (P_n) as well. This is not surprising, for P_n is not well observed even on the records we used. We found that the seismicity was associated with depths in excess of 10 km.

Obviously such depth calculations are model dependent. One way to explore the degree of this dependency is to determine the effect of small changes in the model. Another plausible structure (Table 2) also yields the result that the average depth of the activity is greater than 10 km. The predicted values of the time differences for various phases (Table 3) for this model are such that the average of the observed values indicates depths of the order of 20 km and epicentral distances in the range 170–180 km. Again $P_r - P_n$ does not fit well. On the travel-time plots for events at various depths with a dipping Moho and a flat Riel discontinuity, we note that only once when the source depth exceeds 15 km does S_n arrive before S_g. We cannot distinguish the effect of Moho dip on the travel times; hence we compared our data (Fig. 4) with the travel times predicted for a flat Moho.

TABLE 4. The bounds on the search for feasible models of the velocity structure

165 km	≤	Distance	≤	180 km
0.1 km	≤	Depth	≤	40 km
5.0 km/s	≤	α_1	≤	6.5 km/s
30.0 km	≤	Layer 1	≤	35.0 km
7.0 km/s	≤	α_2	≤	7.6 km/s
10.0 km	≤	Layer 2	≤	20.0 km
7.9 km/s	≤	α half space	≤	8.3 km/s

TABLE 5. The only satisfactory models in a suite of 1000 reasonable ones

H	V_p	V_s	H	V_p	V_s
34.14	6.25	3.45	30.83	6.25	3.45
14.10	7.30	4.01	10.80	7.50	4.15
Half space	8.10	4.48	Half space	8.23	4.55

The data at EDM do not contradict the existence of shallow activity at Rocky Mountain House. Only 41 of the 220 observed events in the digital system showed positive $S_n - S_g$ times. Small $S_n - S_g$ or negative values of this phase difference would lead to masking of S_n in the S_g wave train. Those events for which S_n cannot be picked can be explained as occurring at depths of less than 10 km. The sharpness of the observed distribution of $S_n - S_g$ suggests a concentration of activity at 20 ± 5 km depth.

In this kind of calculation there is a trade off between computational effort and results. At this stage we have a strong suggestion, but certainly no proof, that the average depth of a significant fraction of the seismicity is in excess of 20 km. In order to explore this further, we simplified the characteristics of the velocity model and defined a priori reasonable bounds beyond which the solution of our inverse problem could not lie. We admit that seven parameters are unknown to some degree. These are the average epicentral distance of the activity, the average depth of the activity, the P-wave velocities of two layers and a half space, and the layer thicknesses. We assumed that Poisson's ratio was 0.28 in order to relate S-wave velocities to P-wave velocities.

It is possible with a computer search algorithm (ZXSRCH from the International Mathematical Statistical Libraries (IMSL 1979)) to determine N possible heptads of numbers contained within a seven-dimensional rectangle such that the N heptads are evenly distributed in the possible range (Table 4) of the solutions. For no particularly good reason, other than the desire to limit computation, we set N to 1000 and let the computer find those heptads that define solutions to the inverse problem; only two models fit (Table 5). They both require focal depths in excess of 10 km.

Obviously a computer search of this type does not prove rigorously that pathological solutions do not exist between the points tested. We do, however, have now a reasonable basis for the statement that at least some of the seismicity at Rocky Mountain House originates at depths that suggest natural rather than induced causes. We also state that this depth can be determined if $S_n - S_g$ can be observed.

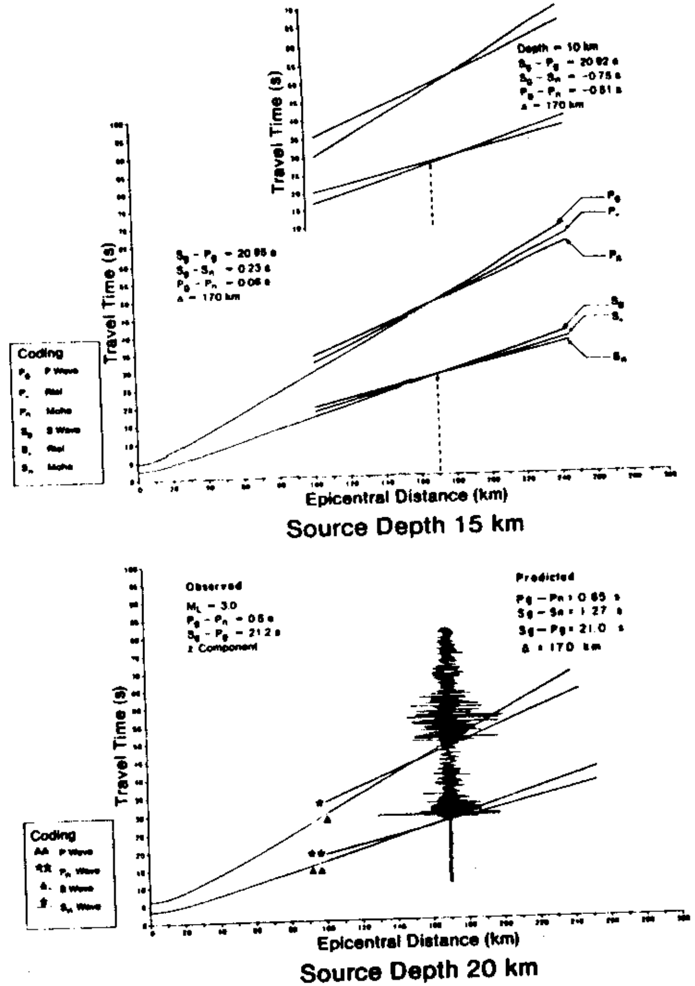

FIG. 4. Theoretical travel-time curves for the Alberta model at different depths. The refracted waves in the Riel discontinuity (P_r and S_r) are not apparent in our travel-time curves. The best match of P_g, S_n, and S_g with the theoretical travel-time curves was for a source depth of 20 km.

There is also shallow seismicity at Rocky Mountain House. One week of observations with a Sprengnether DR100 recording system (Rebollar *et al.* 1982) found mainly shallow activity, but one event showed an $S-P$ time of 2.1 s. Unfortunately, this event cannot be located for the Earth Physics Branch portable analog stations (R. Wetmiller, personal communication, 1981), and the Edmonton digital station was not operating. We are continuing a program of recording at Rocky Mountain House and hope soon to acquire data for several swarm events recorded both at Rocky Mountain House and EDM. This would provide final confirmation of deep seismicity at Rocky Mountain House and raise the very interesting problem of its cause.

A definitive proof would consist of a simultaneous observation at EDM and at Rocky Mountain House of a deep event. We suggest this may be difficult. In 5 years there were 48 events on which S_n could be picked at EDM. Although we are continuing to observe, it may take some time to establish the cause for seismic activity in the basement structures of the Rocky Mountain Foothills.

Spectral analysis

Digital data provide other information on source conditions, for spectral analysis of the body waves becomes feasible without tedious digitization on doubtful records. These digital data

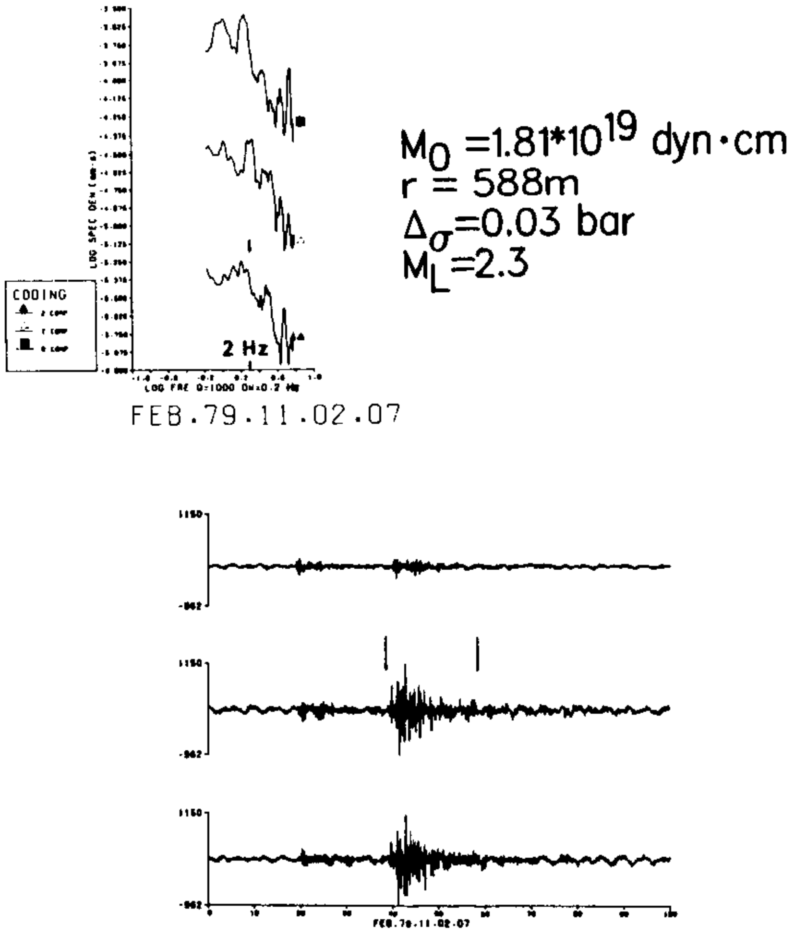

FIG. 5. Spectra of a typical event. The amplitude of radial and tangential components has been increased in order to plot them together. The spectra were smoothed with a Daniel window (DW) of 0.2 Hz.

can yield source parameters. In order to extract good source information, a high sampling rate, a high dynamic range, and a broad band system are desirable. Unfortunately, EDM was designed to detect teleseismic activity. It has a sampling rate of 18 samples per second and a reliable band of frequencies between 0.5 and 7 Hz.

Our basic data are the digitized components of the S phases of the signals identified as originating in the Rocky Mountain House area. These were always band-pass filtered in the range 0.5–7 Hz. By suitable rotations, S_H was isolated on the transverse component and was combined with the vertical and radial components to isolate S_V.

Clowes and Kanasewich (1970), studying the attenuation in southern Alberta, found a Q of 300 in the sedimentary layers and a Q of 1500 for the basement. Therefore, we approximate the attenuation with a Q of 1000 in order to take into consideration the sediments. The signal was corrected for attenuation, instrument response, and distance. We use a sample length of 20 s in which we included S_H, S_V, and scattered waves near the station. Corner frequencies (Fig. 5) were difficult to recognize in many cases because of the narrow bandwidth.

Source parameters can be derived from these spectra (Brune 1970, 1971). As only one station was used in this study, the radiation pattern was approximated as the root-mean-square (rms) average over the focal sphere for S waves (0.63). We calculate the correction for amplification at the free surface for S_V waves using Nuttli's (1961) formulation. The correction for free surface amplification is 1.04 for the S_V vertical component and 1.75 for the S_V radial component considering an angle of incidence of 24.5°.

The source parameters, moment M_0 (dyn·cm), source radius r (km), and stress drop (bar) are related to the spectral characteristics (Brune 1970, 1971) by

$$M_0 = 4\pi\rho R\beta^3\Omega/R_{\theta\phi}A$$

$$r = 0.37/f$$

$$\Delta\sigma = 106\rho R\Omega f^3/10^6$$

where β = 3.37 km/s is the S wave velocity; ρ = 2.9 g/cm³ is the density; Ω is spectral amplitude at zero frequency; f is corner frequency; R is epicentral distance; R is radiation pattern correction; and A is free surface amplification correction.

Computations were made on the spectra of 78 events. Usually the seismic moment calculated from the S_H wave was higher than that calculated from S_V. Their average was used. The error in seismic moments is 2 SD (standard deviations). The corner frequency was identified (between 1.8 and 2.9 Hz) for 34 events. Those corner frequencies gave source radii of 500–600 m and stress drops of 0.01–1.5 bar (1–150 kPa) (Fig. 6). Obviously at this epicentral distance and with this narrow band system, we sampled only a restricted number of events with similar source parameters.

It should be noted that the stress drops reported here are at best lower bounds of the actual values of the stress drops.

Seismic moment and local magnitude

Because we deal here with a relatively large collection of data, average behaviour is probably the best representation of the spectral information. In addition to seismic moment, local magnitude is relatively easy to determine, and we have calculated local magnitudes for all events for which we obtained seismic moments. Seismic moments range from 6.39 ± 2 × 10^{18} (M_L = 1.6) to 2.21 ± 2 × 10^{20} (M_L = 3.5) dyn·cm (6.39 ± 2 × 10^{11} to 2.21 ± 2 × 10^{13} N·cm). Our data show more scatter (Fig. 7) than those on similar plots (Wyss and Brune 1969; Thatcher and Hanks 1973), but our linear trend is similar. The best fit to our data is

$$\log(M_0) = 1.3 M_L + 16.6$$

If we are to interpret the two numbers that seem to represent the observations to some degree we have to assume some model of the earthquake source. Such models are legion, but they all have some features in common. Therefore, we constructed a relation between local magnitude and seismic moment assuming that the Sato and Hirasawa (1973) circular model is representative of the source processes at Rocky Mountain House.

If we assume that all events of the swarm have the same orientation, rupture velocity, and rupture area, we can calculate the predicted spectral amplitude at 0.8 Hz (and hence the predicted local magnitude) as a function of seismic moment. Then the S-wave amplitude at frequency 0.8 Hz will have the form (Sato and Hirasawa 1973)

$$A(\omega) = M_0R_{\theta\phi}B(\omega, v, \zeta, \phi, \tau)4\pi\rho r\beta^3$$

where β is S-wave velocity; v is rupture velocity; ζ is rise time; r is source radius; $R_{\theta\phi}$ is radiation pattern; ϕ is azimuth; M_0 is seismic moment; B is source function; and ω is angular frequency.

The parameters of this relation are of varying uncertainty. Clearly the rupture velocity is hard to define a priori other than by bounds. To say that the rupture velocity must be in the interval of 0.65 ± 0.35 times the shear-wave velocity seems reasonable. Similar bounds on the shear-wave velocity are 3.53 ± 0.1 km/s. Those corner frequencies we have obtained suggest that the source radius will be 0.6 ± 0.2 km. Lacking any orientation information, we took an average value for the radiation pattern correction of 0.6 and an azimuth of 90°. The source function B depends on the model and the angular frequency at which we do the calculation (by definition at 0.8 Hz). We set the rise time to be the ratio of the rupture size to the rupture velocity.

If we assume a rupture velocity of 0.99β, an S-wave velocity of 3.52 km/s, a density of 2.9 gm/cm³, and a source dimension 2r of 1 km we get

$$\log(M_0) = M_L + 17.6$$

This relationship seems to fit our data. The intercept (17.6) of the relationship was modelled for a wide variety of rupture

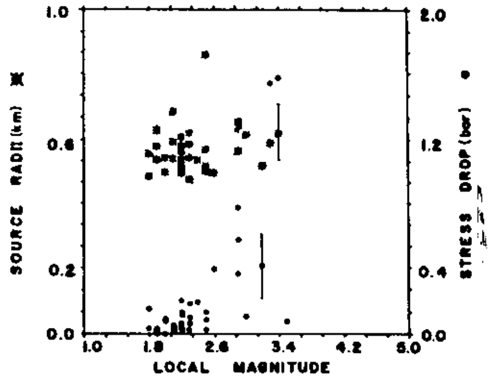

FIG. 6. Source radii and stress drops derived from S_H spectra plotted against local magnitude. Error bars were calculated according to the propagation errors of corner frequency and seismic moment.

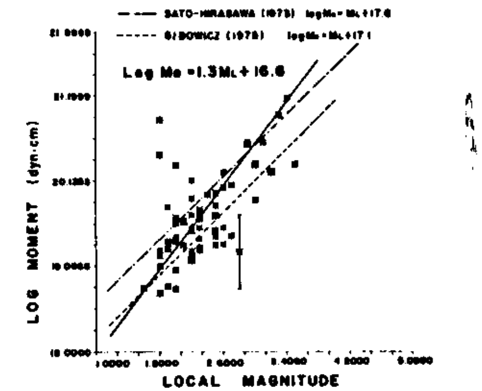

FIG. 7. Comparison of theoretical relationship between M_0 and M_L for a circular fault (Sato and Hirasawa 1973 model) and Gibowicz (1975) relationships with our experimental relationship.

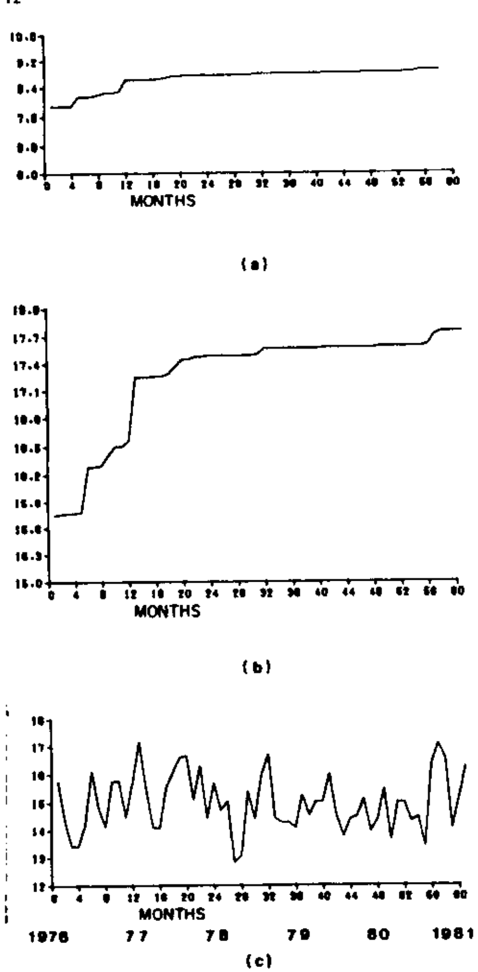

FIG. 8. Strain release (*a*), cumulative energy release (*b*), and energy release (*c*) from the Rocky Mountain House earthquake swarm calculated at EDM, 1976–1980.

velocities, shear-wave velocities, densities, source dimensions, radiation pattern, and azimuths, using again the subroutine ZXSRCH (IMSL 1979). The best fit was found to be log $(M_0) = M_L + 17.1$, assuming a shear and rupture velocity of the same magnitude (3.5 km/s), a density of 2.8 gm/cm^3, a source dimension of 100 m, a radiation pattern of 0.5, and an azimuth of 71°. Therefore, this search for the best fit tends to give smaller source dimensions than those observed in our spectral analysis. Gibowicz (1975) calculated a similar theoretical relation using the Randall (1973) graphical relation given by

$$\log (M_0) = M_L + 17.16$$

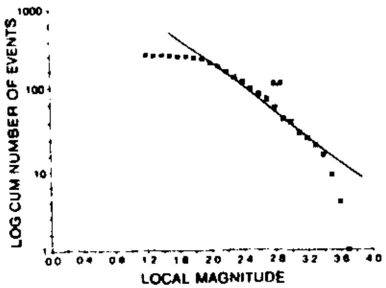

FIG. 9. Evaluation of the *b* value for the Rocky Mountain House earthquake swarm.

for events with source dimensions less than 500 m. This is essentially the same relationship that we found.

Energy release and b *value*

We used Gutenberg and Richter's (1942, 1956) energy–magnitude empirical relationship in order to calculate the seismic energy release of the Rocky Mountain House earthquake swarm; this is given by $\log_{10} E_s = 9.4 + 2.1m_l - 0.024m_l^2$, where E_s is seismic energy and M_l is local magnitude.

This empirical relationship depends on the theoretical study of seismic radiation at short epicentral distances; therefore, it can give a reasonable estimate of the seismic energy for small earthquakes from the Rocky Mountain House earthquake swarm.

An average of 10^{11} erg/month (10^4 J/month) was released during 58 months since 1976. The contribution of small earthquakes to the total seismic energy release is negligible (Fig. 8). The total seismic energy released during this period was 5.6×10^{17} erg (5.6×10^{10} J), equivalent to a single earthquake of magnitude (M_s) 3.9. The total seismic energy release of possible deep events, i.e., earthquakes that show clear S_n refracted phases, was 5.49×10^{17} erg (5.49×10^{10} J). This means that 98% of the total seismic energy release was released by possible deep events (depths greater than 10 km). The cumulative seismic strain release (Fig. 8*a*) yields a maximum strain release of 7.5×10^8 erg$^{1/2}$ (75 J$^{1/2}$).

The strain release as a function of time (Fig. 8*b*) shows a maximum release of strain energy during the first year of the swarm (1976). After that the strain is accumulated and represented by the flat part of the plot. However, whether the strain energy is released in small earthquakes or is accumulated cannot be answered from this short period of observations.

Using 242 events we calculated the cumulative number of events versus local magnitude (Fig. 9) or the Gutenberg and Richter (1956) frequency–magnitude relation given by $\log_{10} N = a - bm_l$, where N is the number of earthquakes for unit time, a and b are constants, and m_l is local magnitude. We used some of the magnitudes calculated by the Earth Physics Branch (EPB), Department of Energy, Mines and Resources Canada, and reported in their bulletin. For events not reported by EPB, we calculated the local magnitude according to Richter (1958).

There are three main factors that limit the accuracy of the evaluation of the *b* value. The magnitudes are uncertain, small events are undetected, and there are few events because of the short time of observations (Milne *et al.* 1978). Uncertainties in the evaluation of magnitudes are difficult to estimate if the

magnitudes are calculated at a single station. The Edmonton station detects events from the Rocky Mountain House area with magnitudes greater than 1.2, as can be seen from the frequency–magnitude relationship. However, some events in the range from 1 to 2 are undetected. The *b* value for the Rocky Mountain House earthquake swarm is 0.8 and lies in the range of values (between 0.6 and 1.5) found by Everden (1970) using worldwide seismic data, and is similar to the values found by Milne *et al.* (1978) in western Canada (those values range from 0.65 to 0.82).

Conclusions

Events detected at Edmonton from the Rocky Mountain House earthquake swarm appear to have a source depth greater than 10 km. This conclusion is based on the match of refracted phases with theoretical travel-time curves for a model of central Alberta. A portable digital seismic station was operated during August of 1980 at 52.26°N and 115.13°W, approximately 155 km southwest of EDM. Events detected during this period of time indicate mainly shallow activity (less than 6 km). Nevertheless, there is one event with an $S-P$ time greater than 2 s, indicating a distant or deep event. Therefore, two kinds of activity may exist: deep activity possibly associated with thrust faults of the basement of the Foothills of the Rocky Mountains, and shallow activity possibly induced by secondary and tertiary recovery methods in the Strachan gas field.

Seismic moments calculated for events with a local magnitude of 1.6–3.5 vary from $6.39 \pm 2 \times 10^{18}$ to $2.21 \pm 2 \times 10^{20}$ dyn·cm ($6.39 \pm 2 \times 10^{11}$ to $2.21 \pm 2 \times 10^{13}$ N·cm). A relation between seismic moment and local magnitude was found to be log $(M_0) = 1.3M_L + 16.6$ for magnitudes from 1.6 to 3.5. This relationship suggests average properties of the source that, in terms of the Sato and Hirasawa (1973) model, give small source dimensions, approximately 100 m, and near-sonic rupture velocities.

The total seismic energy release for a period of 58 months was 5.6×10^{17} erg (5.6×10^{10} J); 98% of this energy was probably released by deep events (events with clear refracted phases at Edmonton). The *b* value of this earthquake swarm was 0.8, similar to that observed in other parts of western Canada.

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