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USGS/NEIS
(Inside Back Cover)
DISPERSION OF SHORT PERIOD RAYLEIGH WAVES WITHIN
THE OZARK UPLIFT AND ILLINOIS BASIN

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ABSTRACT

We use data recorded by four arrays of portable instruments to investigate the
propagation of short period (0.2 ≤ T ≤ 2.0 sec) surface waves within the Ozark
Uplift and Illinois Basin. At the regional scale, we construct group velocity dispersion
curves for five suites of propagation paths, and invert them for shear velocity structure. The
best model in each case consists of a single layer above a halfspace, and we can correlate
the model units with geologic formations. The upper layer in the two Ozark
Uplift models represents Ordovician and Cambrian carbonate strata, the halfspace
corresponds to the Precambrian crystalline basement. Differences between our new
models and an earlier one reflect the different parts of uplift that were sampled, and
show the thickening of the Paleozoic section away from the uplift core. In the Illinois
Basin models, the upper layer represents Pennsylvanian age clastics and the halfspace
represents older Paleozoic carbonates. This interpretation is substantiated by a velocity
log from a nearby deep well. Again, differences between our new models and earlier
efforts result from different passband data that sample different parts of the basin.

We also extract interstation phase velocities from array data recorded at the
western edge of the Illinois Basin, over Mississippian age outcrop. By comparing this
local dispersion curve with one calculated from the appropriate regional scale model,
we conclude that the local structure can also be modeled as a single layer above a
halfspace. Local shear velocities are, however, 30% faster than the regional average,
which reflects the absence of the slow Pennsylvanian clastic strata around the basin’s
periphery.

INTRODUCTION

To help delineate the shallow velocity structure of the
midcontinent, we have investigated the propagation of short
period surface waves at four sites on the flanks of the Ozark
Uplift and Illinois Basin. Our primary goal was to measure
interstation Rg phase velocities as wavefronts traversed small
aperture, portable seismograph arrays deployed at these
sites. With such information for these and other locations,
the regional variation of depth to crystalline basement could
be mapped easily and economically. The recorded data,
however, provided more reliable measurements of source-
receiver group velocity, with which we have extended the
work of McEvilly and Stauder (1965). Their dispersion stu-
dies of short period data allowed them to model the average
basement depth over the targeted geologic provinces, but the
few propagation paths available to them were comprehensive
in neither bandwidth nor geographic coverage. Our data sup-
plement their dispersion curves in the period range
0.2 ≤ T ≤ 2.0 sec, especially for paths within the center of
the Illinois Basin where they had no observations for T ≤ 3.0
sec. Previous investigations using Rayleigh waves in this
range have outlined velocity versus depth profiles over the
Cincinnati Arch (Herrmann, 1969) in Tennessee (Sodbinow
and Bollinger, 1978), New England (Kafka and Dollin,
1985), and eastern Scotland (Macbeth and Burton, 1986) for
depths down to 2 km.

INSTRUMENTATION

At each study site, we deployed five Sprengnether
MEQ-800 analog recorders coupled with Mark Products L-4C
vertical component seismometers. Although readily avail-
able, the field packages were not optimum for these experi-
ments. They had a passband from 0.5 to 35 Hz and a peak
magnification of about 10,000 at 10 Hz with a -24 dB per
octave slope at the low frequency limit. Shake table cali-
bration of the instruments before deployment revealed no
significant differences among their amplitude or phase
responses.
In most cases, the seismometers were buried in about 50 cm of topsoil, about 100 m from local roads. Upon installation, the sites were located on USGS 7.5 min topographic sheets, and so even though they were not surveyed, their positions are known to within 25 m. Site location errors therefore contributed about 3% of the total error in the interstation phase velocities, where the path lengths varied from 0.8 to 3.0 km, but were negligible for the group velocity measurements, where path lengths averaged 100 km.

Relative timing errors were also negligible, despite recording the data from each station separately, instead of on a common medium with a single clock. Hoping to circumvent this expected problem, we adjusted each instrument’s internal clock to WWV twice during each recording day, but did not discern any drift over this interval.

DATA AND VELOCITY ANALYSES

The data consisted of short period, fundamental mode Rayleigh waves, generated by quarry and surface mine explosions throughout the study area. We located the sources using P arrivals recorded by a telemetered regional-scale network (Stauder et al., 1976), and the program FASTHYPO (Herrmann, 1979). Computed source location errors were typically ~ 3.5 km N-S and E-W, and these contributed about 86% of the total uncertainty in the group velocities. Errors in the origin times averaged 0.49 sec and contributed about 13% of the group velocity uncertainty. The epicenter and origin time errors contributed little, however, to the scatter of the observed interstation phase velocities.

As seen in Figure 1, the waveforms were generally simple and well dispersed. They were uncontaminated by beats, which indicated freedom from multipath interference, but several unused waveforms suggested the presence of an overtone, and some displayed an Airy phase. Because of the generally low amplitudes, the presence of high frequency noise, and the difficulty of removing curvature from the pen-written traces, we chose not to digitize the analog records and to analyze the data in the time domain. We therefore calculated source-receiver group velocities in the classical manner, by dividing the propagation path length by the travel time. We did not, however, divide the various paths into “pure-province” segments (e.g. Kafka and Dollin, 1985; Macbeth and Burton, 1986) because the province boundaries are not as sharp as shown schematically in Figures 2 and 5, and because paths to the arrays FVM, DON, and to some extent ELC and CSI sample similar structures. Estimates of the segment lengths would therefore have been somewhat arbitrary.

For phase velocity analyses we calculated local, interstation values using the arrival times of waveform peaks, troughs, and zero crossings, correlated among the stations of each array, in a least-squares technique that assumes planar

--- 10 sec ---

Fig 1. Typical seismograms recorded by portable instrument arrays. Recording speed was 300 mm/min; gains varied between 78 and 96 dB depending on field conditions. Note curvature of traces for large amplitude signals.
Short Period Group Velocity Dispersion

Fig 2. Setting of Ozark Uplift experiments in Gauss-Krüger (UTM) projection. Uplift delineated by outcrop extent of Ordovician sediments (Bayer, 1983). Closed triangles and large fonts represent portable seismograph arrays in this investigation and stations used by McEvilly and Stauder (1965); closed circles source locations; open circles and small fonts wells providing geologic control for velocity models.

wavefronts with a constant propagation azimuth (Aki, 1961). We could not, however, read these times with sufficient precision to achieve consistently accurate results. Only for the RLC array could we establish an unambiguous dispersion curve, presented below. The values from the FVM, DON, and CSI arrays were extremely scattered, and could not be interpreted with confidence. We suspect that the station separations at these arrays were too small; wavefronts arrived nearly simultaneously at all of the stations. Future investigations of interstation phase velocities at these and other sites will require array apertures of at least 5 km.

GROUP VELOCITY STUDIES

Ozark Uplift

Figure 2 illustrates the setting and geometry of the experiments in the Ozark Uplift. The shaded area depicts the outcrop extent of Ordovician strata, which are the dominant surface units within the uplift (Bayer, 1983). Smaller outcrops of Cambrian strata surround the exposed crystalline uplift core. The propagation paths to the arrays of portable instruments at FVM and DON have an average length about 80 km and back-azimuths between 0° and 90°. The paths to FVM might have been divided into two families with azimuths ~90° apart, but the group velocities obtained from these families at any period were within one standard devia-

tion of each other. We therefore concluded that azimuthal effects were minimal, and we averaged the suite of dispersion curves to the FVM array and the suite to the DON array to form mean curves for each. For comparison, we have also shown the two propagation paths to station ROL from which McEvilly and Stauder (1965) constructed model OZARK. These paths are considerably longer (178 and 255 km) and cross the center of the uplift. They too have widely separated back-azimuths (72° and 129°), but no azimuth-dependent variation of group velocity was reported for these, either.

In Figure 3 we present the mean group velocity dispersion curves determined for the FVM and DON arrays. These curves agree with each other to within one standard error for periods \( T \leq 0.85 \) sec. In this range, our observations also agree reasonably well with values computed from model OZARK. For example, at \( T = 0.73 \) sec the three curves agree to within 4%. At longer periods they diverge slightly; at \( T = 1.5 \) sec the differences between them are as large as 11%. To interpret these dispersion curves, we conducted an series of inversions using a variety of starting velocity models, and have superimposed the fitted curves from our preferred models on the observations. The inverse fits for both the

Fig 3. Mean source-receiver group velocity dispersion observed at FVM and DON arrays with preferred inverse fits superimposed. Error bars represent one standard error of the mean, scaled to the 95% confidence interval via Student’s \( t \) statistic. Error estimates incorporate both observational error and scatter of individual event curves. Solid curve represents group velocity dispersion determined by McEvilly and Stauder (1965).
FVM and DON curves are good; they easily lie within one standard error of the experimental results.

The velocity models FVM and DON are illustrated in Figure 4. Here we have graphed both the shear and compressional wave velocities, the latter having been computed from the inverted $\beta$ vs. $z$ profile with the assumption that Poisson’s ratio $\sigma$ equals 0.25. Three features of the models stand out. First is that both consist of a single layer overlying a halfspace. This is the simplest of many layering schemes tested. Although multilayer models fit the data well, additional layers were not necessary to obtain an adequate match.

![Velocity profile diagram](image)

**Fig 4.** Velocity models for Ozark Uplift's eastern flank, inverted from observed dispersion curves in previous figure. Model DON is thin line, model FVM is thick line. Model parameters listed in Table 1. Compressional velocities calculated from shear velocity profile with the assumption that Poisson’s ratio was 0.25.

Furthermore, with this layering scheme the inverted velocities were resolved exactly. The second feature is that the velocities in the models are similar, both above and below the interface. The difference between the layer values is about 4%; the difference between the halfspace values is less than 1%. Moreover, the layer values are similar to that found in model OZARK (Table 1), but the halfspace values are about 11% lower. Birch (1942) listed $2.47 \leq \beta \leq 2.99$ km/sec for “limestones” and $2.82 \leq \beta \leq 3.52$ km/sec for “granites” at the 80% confidence limits. Thus, it is reasonable to interpret models FVM and DON as representing the Ordovician and Cambrian carbonate sequence and the underlying crystalline Precambrian basement. The third, and most apparent, feature of the two inverse models is that the upper layers in both are thicker than found in OZARK, and differ from each other by 200 m. We believe the differences among these three models reflect variation of the lithologic composition and thickness of the Paleozoic section in the Ozark Uplift.

Single-layer velocity models such as OZARK, FVM and DON can be characterized by four dimensionless parameters: the shear velocity contrast, $\beta_1 / \beta$, the density contrast, $\rho_2 / \rho_1$, and Poisson’s ratio in each medium, $\sigma_1$ and $\sigma_2$ (Mooney and Bolt, 1966). Of these, the shear velocity contrast exerts the greatest control over the shape of the dispersion curve. The stronger the contrast, the steeper the curve at the inflection point between its short period minimum and longer period asymptote. For the three models, we have velocity ratios of 1.40, 1.29, and 1.24 respectively. If we assume a reasonable range for this ratio, say from 2.55 (e.g. weakly consolidated sandstone over granite) to 1.0, the three models offer a 10% variation, which largely explains the similarity of the observed dispersion curves.

Next in importance is the value of Poisson’s ratio in the layer, $\sigma_1$. The less rigid the material, the larger the value, and again the steeper the curve at its inflection point. As mentioned above, these models share a value of 0.25, a commonly used average. We note, however, that weakly consolidated material, as found in the upper 10 m of the “weathered layer”, may have a value as high as 0.45. The remaining two parameters, $\rho_2 / \rho_1$, and $\sigma_2$, have significantly less influence on the shape and position of the dispersion curve in a velocity-period field.

The slight differences among the dispersion curves and among the velocity models are understood by examining the propagation paths used in each study (Figure 2). Recall that model OZARK is based on the two long paths to station ROL. These traverse the uplift’s center, where the Ordovician and Cambrian strata that overlie the basement are thinnest. In contrast, the shorter paths used in this study approach the FVM and DON arrays from the east and northeast, over the flanks of the uplift where the sedimentary section should be thicker. It thus seems clear that the interface in models FVM and DON represents the top of the Precambrian basement.

<table>
<thead>
<tr>
<th>Model</th>
<th>$H$ (km)</th>
<th>$\alpha$ (km/sec)</th>
<th>$\beta$ (km/sec)</th>
<th>$\rho$ (g/m/sec$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>OZARK</td>
<td>0.65</td>
<td>4.30</td>
<td>2.50</td>
<td>2.67</td>
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<tr>
<td></td>
<td></td>
<td>6.10</td>
<td>5.00</td>
<td>2.67</td>
</tr>
<tr>
<td>FVM</td>
<td>0.70</td>
<td>4.18</td>
<td>2.43</td>
<td>2.34</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5.40</td>
<td>3.12</td>
<td>2.38</td>
</tr>
<tr>
<td>DON</td>
<td>0.90</td>
<td>4.33</td>
<td>2.50</td>
<td>2.37</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5.38</td>
<td>3.11</td>
<td>2.38</td>
</tr>
</tbody>
</table>

The available borehole data, although incomplete, generally support this interpretation. Wells SF1 and SF2 (Figure 2) recovered Precambrian felsic rock at about 230 m (Kisvarsanyi, 1975). Neither is shown on Buschbach’s (1985) map of the top of the crystalline basement, but the depths of penetration are in concert with the local NE dip of the basement.
Short Period Group Velocity Dispersion

surface. Because the propagation paths to the FVM array came from this direction, we take the layer thickness in model FVM as being the minimum value consistent with both the observed group dispersion and published estimates of the Precambrian basement depth. Supporting model DON is an interpreted driller’s log from the DC1 well (Figure 2). It lists the thicknesses of the encountered formations (Groshkopf, 1955), which are consistent with the generalized Missouri stratigraphic column (e.g., Kisvarsanyi, 1976). Although the well did not penetrate older, deeper strata (maximum depth was 552.4 m), it is reasonable to extrapolate their thicknesses from known values elsewhere, because no major unconformities are known to exist in the area. Assuming the strata to be continuous to the northeast, in the direction of the propagation paths, the summed thickness of the Paleozoic section is between 554 and 989 m. These values agree reasonably well with the range 610 to 910 m estimated from Buschbach’s map, and bracket the depth of 900 m required by model DON.

Illinois Basin

The shaded outline of the Illinois Basin in Figure 5 generally follows the outcrop pattern of Mississippian strata (Bayer, 1983). In this province, the investigated propagation paths vary in length and azimuth more than in the Ozark Uplift. Nevertheless, we find that they may be grouped into only three sets. Most of the paths to the CSI array are confined to the relatively shallow western and southern shelves of the basin and, with the exception of the two paths with back-azimuth = 62°, are aligned roughly along 110°. They yield group velocities that are sufficiently alike that we consider them as forming one family. The two paths that extend into the basin interior yield group velocities significantly different from those obtained from the paths along the shelves, and so are considered as a second family. The third set consists of the three paths to the ELC array. They sample the same area as the CSI(shelf) suite, but at a different back-azimuth (0°), and provide a means to check the reproducibility of the results across the western shelf.

McEvilly and Stauder (1965) based their average model BASIN almost exclusively on data from the paths to FLO and SLM. These paths cross the shallowest parts of the basin, but provided abundant group velocities for periods $T \leq 5$ sec. On the other hand, the path from the earthquake epicenter in southeast Missouri to station BLO yielded only three group velocities, all for $T \geq 3.0$ sec. Thus, our group velocities from the paths in the basin’s interior are an important supplement to the existing data base.

The group velocities extracted from the three suites of propagation paths form distinct dispersion curves (Figure 6). With standard errors that average 0.04 km/sec, none of the curves intersect. It is interesting though, that the curves observed at CSI and ELC over the basin’s western shelf have nearly the same shape; translation parallel to the period axis allows the curves to coincide. This isomorphism implies that the curves sample similar velocity structures. The dispersion curve constructed for the two propagation paths that extend from CSI into the basin interior exhibit a smaller slope than the shelf curves, which implies a milder vertical gradient of shear velocity in the basin center than along its periphery. The best inverse fits to these dispersion curves are graphed, again agreeing with the experimental results to within one standard error.

Whereas model BASIN required two layers over a halfspace, we find that for the bandwidth of our data, single layer models will suffice. The parameters of the three inverse models are listed in Table 2 and are illustrated in Figure 7. Two features of the models are again readily apparent. First is the thickness of the upper layer varies as expected for the geographic distribution of the suites of propagation paths. In models CSI(shelf) and ELC, derived from

![Fig. 5. Setting of Illinois Basin experiments in Gauss-Krüger (UTM) projection. Basin delineated by outcrop extent of Paleozoic (Pennsylvanian and Mississippian) sediments (Bayer, 1983). Closed triangles and large fonts represent portable seismograph arrays in this investigation and stations used by McEvilly and Stauder (1965); closed circles source locations; open circles and small fonts wells providing geologic control for velocity models.](image-url)
paths that sampled the basin periphery, where the sedimentary section is thin, $H = 0.6$ km. In model CSI(center), derived from the two paths that sample the basin interior, where the section is thicker, $H = 0.9$ km. The second feature is the narrow range spanned by the shear and compressional velocities. In the upper layer, $1.81 \leq \beta \leq 1.96$ km/sec, only an 8% difference. In the halfspace, this range is narrower still, $2.72 \leq \beta \leq 2.86$ km/sec, or 5%. Moreover, models CSI(shelf) and ELC have identical velocity contrast: $\beta \beta_1 = 1.50$, which helps explain the isomorphism of the dispersion curves observed over the basin's western shelf. The contrast is slightly less for model CSI(center), $\beta_2 \beta_1 = 1.42$.

A velocity log obtained from the well TC1 (Figure 5) strongly supports these inverse models. It equals the model compressional velocities within 3%, which helps justify, a posteriori, our assumption of 0.25 for Poisson's ratio in the computation of $\alpha_i$ from $\beta_i$. Only in the upper 100 m of the section is the log appreciably lower (~ 15%) than the model velocities. Note especially the excellent match between the log and model CSI(center). The log's sharp increase at $z = 0.9$ km agrees exactly with the depth of the model interface. Sexton et al. (1986) identified this increase as the top of the St. Genevieve Limestone, the uppermost of several hard, relatively fast Mississippian carbonates. Above the carbonates, the section consists of Pennsylvanian clastic strata. The range of velocities for sandstone-shale mixtures (Press, 1966) easily brackets the velocities we calculate for the upper layers.

The use of realistic rock densities, computed from Nafe and Drake’s (1957) velocity-density relationship, lends our models further credence. Of course rock density is extremely variable and not diagnostic, but our comparatively low values are consistent with shales, sandstones, and most limestones. The fixed value of 2.67 gm/cm$^3$ in models OZARK and BASIN is more appropriate for granite. Because the shallow strata of the Ozark Uplift and Illinois Basin are known to consist of the former, we are confident that our models better reflect the near-surface structure in these provinces.

**PHASE VELOCITY STUDIES**

As mentioned above, the interstation phase velocity curve determined at the ELC array is the best defined of
those from the four study sites (Figure 8). The scatter of the observations is comparable to that obtained by Sodbinow and Bollinger (1978) with a similar time-domain technique from data in the same passband. The scatter has two possible causes: lateral velocity heterogeneity beneath the array that would refract and disrupt the wavefronts, and the finite reading precision of the arrival times. As velocity heterogeneity is probably minimal over the characteristic dimension of the array (~3 km), we attribute most of the scatter to the reading precision. It is clear, however, that the velocities are mildly dispersed about an average value of 3 km/sec.

![Image of ELC Array Station Geometry and Composite Phase Dispersion](image)

Fig. 8. Left, Station configuration for ELC array; see Fig. 5 for array location. Right, Phase velocities observed at ELC array with average regional dispersion curve and suite of dispersion curves computed for local velocity models. Model parameters listed in Table 3.

### Table 3: ELC Array Local Velocity Models

<table>
<thead>
<tr>
<th>Model</th>
<th>(H) (km)</th>
<th>(\alpha) (km/sec)</th>
<th>(\beta) (km/sec)</th>
<th>(\rho) (g/m³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ELC04</td>
<td>0.4</td>
<td>4.84</td>
<td>2.79</td>
<td>2.46</td>
</tr>
<tr>
<td></td>
<td>---</td>
<td>6.10</td>
<td>3.50</td>
<td>2.67</td>
</tr>
<tr>
<td>ELC06</td>
<td>0.6</td>
<td>4.84</td>
<td>2.79</td>
<td>2.46</td>
</tr>
<tr>
<td></td>
<td>---</td>
<td>6.10</td>
<td>3.50</td>
<td>2.67</td>
</tr>
<tr>
<td>ELC08</td>
<td>0.8</td>
<td>4.84</td>
<td>2.79</td>
<td>2.46</td>
</tr>
<tr>
<td></td>
<td>---</td>
<td>6.10</td>
<td>3.50</td>
<td>2.67</td>
</tr>
</tbody>
</table>

We disagree with Sodbinow and Bollinger (1978) that such high phase velocities must reflect higher mode propagation. Shear velocities in the sedimentary section need not be greater than 3.5 km/sec to explain high values for the fundamental mode. If we assume that the velocity structure beneath the ELC array can again be modeled as a single layer above a halfspace, the 30% elevation of the local phase velocities above the average regional dispersion curve requires only that \(\beta_1 \geq 2.5\) km/sec. Moreover, the average slope of the local curve is less than that of the regional curve. From this we infer that the local shear velocity contrast, \(\beta_2/\beta_1\), is less than the regional value of 1.50. With these constraints it is possible to construct a suite of geologically plausible models that yield appropriate phase velocities.

We have superimposed theoretical dispersion curves from three such models over the local observations. We based these models (Table 3) on information provided by the local geological setting and the regional average structure, established above. From Bayer (1983) we know that the ELC array sat atop Mississippian strata, and from the group velocity inversions and the velocity log we know that the Mississippian and underlying Paleozoic section have a shear velocity about 2.8 km/sec. We take this value for \(\beta_1\). Below the Paleozoic section, the Precambrian granite has a shear velocity of 3.5 km/sec (McEvilly and Stauder, 1965), which we take for \(\beta_2\). Therefore, the local shear velocity contrast is 1.25. We then vary the layer thickness \(H\) to obtain the illustrated curves. The middle curve, with \(H = 0.6\) km, appears to fit the best, but it is difficult to confirm this value, as only chip samples are available from the nearby well OFI (Figure 5). Nevertheless, a thickness of 600 m is consistent with the expected thinning of the Paleozoic section toward the basin's periphery.

### CONCLUSIONS

This effort has been directed toward a preliminary exploration of the variation of the shallow shear wave structure in the Ozark Uplift and Illinois Basin, using two techniques that offer complementary views: regional and local. Group velocity techniques work well, but yield an average velocity model for relatively long paths. They can, however, be combined easily with observations at other stations to obtain a tomographic view of the shear wave structure. Phase velocity techniques yield a velocity model specific to the location of the recording array but are sensitive to the design of the array and to the timing of the measurements.

The analysis of short period surface waves can define the shear wave structure at shallow depths, an important constraint for models of ground motion in earthquake hazard studies. The problem of limited depth resolution can be addressed by using a wider signal bandwidth in the analysis, which is now possible given digital recording and newer broadband sensors. The next step is to extend and refine the velocity models spatially, and to focus on the equally important parameter, Q.

### ACKNOWLEDGMENTS

We would like to thank Denis Reidy for help during the field program, Michael Hamburger for providing a crucial BLO seismogram, and Ken Taylor for relocating a troublesome epicenter. This research began while M.T.W. and D.R.R. were at St. Louis University, and was funded in part by the U.S. Department of the Interior, Geological Survey under Contract 14-08-0001-G1090.

### REFERENCES


Bayer, K. (1983). Generalized structural lithologic and physiographic provinces in the fold and thrust belts of the United States (exclusive of Alaska and Hawaii), USGS, 1:2,500,000, 2 sheets.


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