THERMOELASTIC STRAIN IN A HALF-SPACE COVERED BY UNCONSOLIDATED MATERIAL

BY YEHUDA BEN-ZION AND PETER LEARY

ABSTRACT

An algorithm to predict crustal thermoelastic strain from observed local atmospheric temperature is given and applied to a 24-month crustal strain record of one test strainmeter site located near Bouquet Reservoir in southern California. We use a crustal model that consists of an elastically decoupled layer overlying a uniform elastic half-space, and a thermal source that is given by a stationary temperature wave whose wavelength is related to local topography and/or lateral material heterogeneity. The decoupled layer delays, attenuates, and low-pass filters the source temperature field. The thermoelastic strain in the underlying half-space, resulting from the temperature variations at the base of the decoupled layer, is calculated using the Berger (1975) solution for thermoelastic strain in a uniform half-space. Applying our model to the test data, we obtain a good fit between predicted and observed strains if we filter the surface thermal signal through a 63-cm-thick decoupled layer. Much of the remaining strain variations clearly correspond to other environmental sources (reservoir loading and rainfall). Our analysis suggests that the horizontal thermoelastic strain is inversely proportional to the wavelength of local topography and/or lateral material heterogeneity. Thus, the horizontal thermoelastic strain will be greater in areas of local topography and/or lateral material heterogeneity and smaller in more homogeneous and flat areas. An upper layer of loose material, natural or artificial, acts as a thermoelastic strain insulator. Burial of strainmeters in places where such a layer exists can reduce the thermoelastic strain noise considerably even for shallow strainmeter emplacements.

INTRODUCTION

Geophysical strainmeters, being close to the surface, are greatly affected by surface and atmospheric phenomena. Several responses to the shallow crust strain noise problem have been made. Goulty and Gilman (1978) situated their strainmeters near enough to large shallow creep displacement sources in central California to overcome the environmental strain noise. The geophysical group at the University of California San Diego, working at Pinon Flat between the San Jacinto and San Andreas faults in southern California, has lengthened the strainmeter baseline to 750 m, placed their instruments in an arid site of low local topographic relief, and installed supportive strain and tilt sensors to provide compensation for the environmental effects (e.g., Wyatt and Berger, 1980; Wyatt, 1982). The University of Southern California has installed strainmeters in six abandoned tunnel sites (mines, water diversion adits, and aqueduct tunnels) in the San Gabriel Mountains of southern California (Leary and Henyey, 1982). At three sites, Bouquet (BQ), Dalton (DT), and Jackson (JK), the tunnel conditions were sufficiently stable to permit continued operation of the strainmeters (Figure 1, Table 1). The largest nontectonic effect in the data recorded by these strainmeters is the thermoelastic strain induced by variations of the atmospheric temperature. Elimination of the large thermoelastic strain signal will permit a better understanding of the remaining noise sources. Proper compensation for the remaining noise should result in a strain instrument suited to the detection of short-term tectonic strain signals. In this paper, we model
the near-surface thermoelastic strain component in the observed strain data in order to devise a filter for its removal.

Berger (1975) gives a solution for the thermoelastic strain induced in an infinite homogeneous flat half-space by a traveling thermal wave. The solution contains two terms: an equivalent body force term due to thermal expansion at a given point, and an equivalent surface traction term due to elastic coupling of thermal expansion at distant points. For depths greater than the thermal boundary layers, only the latter term is of importance. The removal of thermally induced strain by direct

<table>
<thead>
<tr>
<th>Site</th>
<th>Location</th>
<th>Name</th>
<th>Length (m)</th>
<th>Orientation</th>
<th>Site Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bouquet Reservoir</td>
<td>N34°34.7’</td>
<td>BQ1</td>
<td>27</td>
<td>E10°N</td>
<td>Abandoned gold mine 1 km north of Bouquet Reservoir; 10 m overburden; perpetually wet with seasonal variations in standing water.</td>
</tr>
<tr>
<td></td>
<td>W118°22.8’</td>
<td>BQ2</td>
<td>20</td>
<td>E10°N</td>
<td></td>
</tr>
<tr>
<td>Sierra Pelona Valley,</td>
<td>N34°30.9’</td>
<td>JK1</td>
<td>30</td>
<td>N45°W</td>
<td>Abandoned gold mine in detached granite outcrop in sedimentary basin; average 25 m overburden; dry throughout year.</td>
</tr>
<tr>
<td>Soledad Canyon</td>
<td>W118°17.6’</td>
<td>JK2</td>
<td>30</td>
<td>N45°E</td>
<td></td>
</tr>
<tr>
<td>Big Dalton Canyon,</td>
<td>N34°10.2’</td>
<td>DT1</td>
<td>30</td>
<td>20°W</td>
<td>Bypass tunnel for Big Dalton Dam; tunnel entrance in face of 100 m near-vertical wall; DT1 normal to wall face at depth 10–40 m; moist, with standing water during wet months.</td>
</tr>
<tr>
<td>Glendora, California</td>
<td>W117°48.4’</td>
<td>DT2</td>
<td>60</td>
<td>N60°E</td>
<td></td>
</tr>
</tbody>
</table>

**Fig. 1.** Location map for carbon fiber strainmeter array in San Gabriel Mountains of southern California.
application of the Berger (1975) solution is not, however, in general possible. Two phenomena appear to require modification of the half-space approximation.

1. Owing to weathering and depositional processes, as recognized by Berger (1975), the upper part of the crust is commonly characterized by a zone of incompetent and loose material. The presence of a surface layer may be expected to alter the thermoelastic response of an underlying uniform half-space to a surface temperature field. In particular, the elastically weak surface layer will be a poor transmitter of thermally induced strain. In the limit of a truly unconsolidated surface layer, thermal strain in the elastic half-space will be generated only by the delayed, attenuated, and low-pass-filtered thermal pulse which has traveled through the unconsolidated surface layer. Accordingly, we generalize the half-space model to a composite model that consists of a half-space covered by an elastically decoupled unconsolidated material. In our model, the expression for the thermoelastic strain is still Berger's (1975) solution for the uniform half-space; however, the source field for the strain may now be given by the temperature variations at the boundary between the half-space and the decoupled layer.

2. The actual crustal surface has topography and lateral material heterogeneities. These heterogeneities play important roles in the thermoelastic strain problem. Harrison and Herbst (1977) discuss coupling between the body force term and topography to allow horizontal thermoelastic strain in the presence of topography. Their solution, however, does not contain the Berger (1975) traction term and thus, while it may be valid for the near-surface area, it does not provide for a significant thermoelastic strain below the shallow penetration depth of the thermal pulse. In our opinion, the important effect of topography and lateral material heterogeneities is to create the local spatial variations in the temperature field and in the response of the rock. This is a longer wavelength effect which penetrates deeper than does the thermal pulse treated by Harrison and Herbst (1977). We incorporate this effect in the composite model by using a thermoelastic strain source defined as a stationary thermal wave whose wavelength is related to topography and lateral material heterogeneities in a given area.

Using the composite model, we are able to predict with good accuracy the waveform and amplitude of the thermoelastic component of the observed BQ strain record from the observed local atmospheric temperature time series. The residual crustal strain record bears a strong resemblance to the water level fluctuations of a nearby reservoir. The two parameters of the composite model for the BQ strain site, the thickness of the equivalent upper layer (~65 cm), and the wavelength of the source field (~4 km ~ local topography), are sufficiently plausible that our model appears to have a physical basis. The qualitative properties of a second (DT) strain record tend to support the physical validity of our simple thermoelastic model for the near-surface crust.

The Strain Data

The University of Southern California tunnel-sited strain instruments are modifications by R. Bilham of the original King and Bilham (1976) Cambridge wire strainmeter. A light wire (or in our case bundle of molecular carbon fibers) of high extensional modulus is strung between two piers firmly attached to the ground. In a tunnel, the exposed bedrock renders pier construction a straightforward and reliable procedure. For the six University of Southern California instruments, the distance between the piers varies between 17 and 60 m. The fiber is fixed directly to one pier. On the second pier, a weighted horizontal lever arm is used to place the
fibervunder tension. As the distance between the piers expands or contracts, the weighted lever arm rises or falls, a motion that can be detected with a precision of 1 micrometer. Because the lever arm operates against the fiber with a mechanical advantage of 20:1 to 25:1, extensional sensitivity is on the order of $2 \times 10^{-9}$ strain.

Siting the instruments in tunnels conveniently isolates the strainmeter from direct environmental effects. In particular, the air temperature in the tunnels remain stable to annual fluctuations of $\pm 0.01^\circ C$ at DT and $\pm 0.05^\circ C$ at BQ. Thus, direct thermal effects are limited to a few per cent of the 1 $\mu$ strain environmental signal. The instruments, however, remain sensitive to surface environment effects transmitted elastically by the host rock to the strainmeter cavity.

Preliminary processing of the raw strain records prepared the data for the thermoelastic strain analysis. Small gaps were filled by interpolation to form a continuous data base, and the data were detrended to a level baseline. The removed trend is believed to be the result of drift of the carbon fiber or other mechanical components in the instrument since: (i) it is registered as $\sim 10^{-6}$ strain/yr of apparent contraction, while the geodetically estimated long-term regional tectonic strain rates are a factor of 10 lower and in a sense of extension (e.g., Savage et al., 1981); (ii) values of the detrended component and similar near-linear terms of other years (Figure 2) fall on a curve which suggests a decay in the drift rate with time; (iii) the presence and properties of this component are largely independent of site and instrument orientation. In any case, the component that was removed by detrending the data, whether entirely instrumental or partially tectonic or environmental (as from local groundwater trend) in origin, has little impact on the present discussion in view of the cyclic untrended nature of the thermal driving source to be modeled here. In the following sections of the paper, we refer to the corrected (continuous detrended) data set as “observed strain.”

**Thermoelastic Strain**

Consider a surface temperature field (Figure 3) given locally by a stationary wave

$$T(x, 0, t) = \sum_{\omega} T_{\omega} \cos(kx)e^{i(\omega t - \phi)}$$

(1)
where $\omega$ is the angular frequency, $T_\omega = (A_\omega^2 + B_\omega^2)^{1/2}$, $A_\omega$ and $B_\omega$ being the sine and cosine coefficients of the pertinent $\omega$, respectively, $k = 2\pi/\lambda$, $\lambda$ being the wavelength of the surface stationary temperature wave, and $\phi = \arctan(A_\omega/B_\omega)$. The solution of the heat conduction equation with (1) as the surface boundary condition, provides the two-dimensional temperature profile within the elastic half-space

$$T(x, y, t) = \sum_\omega T_\omega \cos(kx)e^{-\gamma t}e^{i(\omega t - \phi)}$$  \hspace{1cm} (2)

where $\gamma = k[1 + i\omega/\kappa k^2]^{1/2}$, $\text{Re}(\gamma) > 0$, and $\kappa$ denotes the thermal diffusivity. From (2), we see that the crustal temperature variations are delayed, attenuated, and low-pass filtered with depth. The boundaries of the thermal layers may be defined to be at $y = 2\pi/\text{Re}(\gamma)$. Using this definition and characteristic thermal diffusivity of many rocks, $\kappa = 10^{-2}$ cm$^2$/sec (Clark, 1966), we get thermal boundary layers of 20, and 1 m for the annual and diurnal periods, respectively.

Thermoelastic strain at depth $y = y'$ can be thought of as the superposition of (A) strain resulting directly from temperature variations at $y = y'$ and (B) traction from the thermally induced strain at all shallower depths ($y < y'$) that are elastically coupled to $y = y'$.

Taking Berger's solution for the horizontal plane strain $\varepsilon_{xx}$ with (2) as a source, we get

$$\varepsilon_{xx} = \sum_\omega \left( \frac{1 + \sigma}{1 - \sigma} \right) \left( \frac{k}{\gamma + k} \right) \cdot \left\{ \left[ 2(1 - \sigma) + \frac{k}{\gamma - k} - ky \right] e^{-ky} - \frac{k}{\gamma - k} e^{-\gamma y} \right\} \cdot [\beta T_\omega \cos(kx)e^{i(\omega t - \phi)}]$$  \hspace{1cm} (3)

where $\sigma$ is Poisson's ratio and $\beta$ the coefficient of linear thermal expansion. Solution
(3) contains an equivalent thermal body force term falling off with depth as $e^{-Re(y)}$, and an equivalent thermal surface traction term falling as $e^{-ky}$. These terms correspond, respectively, to the body force term and traction term of (A) and (B) above. We note that the body force is negligibly small below the thermal boundary layers. However, if the thermal layers are elastically coupled to the rock below, thermoelastic strains generated within these layers persist through surface traction to a depth that is on the order of the surface temperature wavelength.

Consider now an upper layer of loose material covering the half-space as illustrated in Figure 3. Any combination of the following processes is expected to occur through that layer and at the boundary between it and the half-space below: variations with depth of the thermal diffusivity coefficient; partial decoupling of surface traction; and full decoupling of the surface traction. All of these processes have the effect of further delaying, attenuating, and low-pass filtering the temperature variations and the resultant thermoelastic strain. Thus, without distinguishing between these possible mechanisms, we represent the total effect of the upper coverage by introducing an “equivalent upper layer” that is fully decoupled from the half-space below. The composite model consists of this environment of a half-space covered by the “equivalent upper layer.” This representation is chosen for its simplicity, since it enables us to keep the half-space solution for the strain. The effects of the “equivalent upper layer” are confined only to changes in the temperature field that acts as a source for the thermoelastic strain. The half-space solution (3) then gives the thermoelastic strain in the competent rock as a response to the altered temperature source field at the base of the unconsolidated layer. The thickness of the “equivalent upper layer” is found, as we shall show, by fitting the delay and frequency content of the predicted thermoelastic strain to the observed data. The magnitude of the predicted thermoelastic strain can then be fitted by using either the wavenumber of the surface temperature variations or the position of the strainmeter in the temperature field as a free scaling parameter. In the following analysis of BQ strain data, we used the wavenumber of the surface temperature field as the free scaling parameter. The best-fitting wavenumber found in our strain prediction procedure corresponds closely to the dominant wavenumber of the local topography at BQ.

**Analysis and Results**

To examine the validity of the composite model and to demonstrate its application, we use strain data recorded by our strainmeter station at BQ. The two instruments at BQ are parallel and closely redundant, giving us confidence in the strain record. The tunnel which houses the BQ strainmeters is located in a hilly terrain covered by a layer of dirt and dense vegetation. The overburden of rock column above the strainmeters at that site is about 10 m. In Figure 4, we show 2 yr of BQ2 strain data, along with daily minimum and maximum atmospheric temperature time series recorded at the weather station in Bouquet Reservoir. It is seen that the thermally induced strain signal dominates the strain record.

In order to best fix a surface location for the strainmeter, we observe that the general location of minimum temperature variations is likely to be at Bouquet Reservoir where the crustal surface is covered with water, and that the general location of maximum temperature variations is likely to be at the south-facing slope of the local topography where the strainmeter site is situated. We thus set in (3), $x = 0$. The thickness of the “equivalent upper layer” and the wavenumber of the spatial variations of surface temperature remain free parameters that are used to
fit the phase delay, frequency content, and magnitude of the thermoelastic prediction to the observed strain. We start by assuming a thickness of zero for the "equivalent upper layer" and a horizontal wavelength of temperature variations that is similar to the scale length of local topography and heterogeneity in the considered environment. For BQ, we take $\lambda = 3$ km, compatible with $\lambda/4 = 750$ m $\approx$ the separation distance between the reservoir and the strainmeter site. Substituting in (3) the following values

\[
\begin{align*}
\kappa &= 8.64 \cdot 10^{-2} \text{ m}^2/\text{day} \\
\sigma &= \frac{1}{3} \\
\beta &= 10^{-5} \text{C}^{-1} \\
\kappa &= 2\pi/3 \cdot 10^3 \text{ m}^{-1} \\
y &= 10 \text{ m} \\
x &= 0
\end{align*}
\]

and using the temperature time series shown in Figure 4, we generate the predicted thermoelastic strain for the half-space model. When the half-space model prediction is compared with the observed strain (Figure 5), two discrepancies appear

1. the observed strain is delayed with respect to the predicted strain;
2. the predicted strain contains high frequencies that do not appear in the observed strain.

For the case we are studying, where the strainmeter is below the thermal boundary layers of the short-period components of the temperature field, these two discrepancies cannot be accounted for by varying parameters in the framework of the half-space model. This is understood, since at these depths the contribution of the delayed, attenuated, and low-pass-filtered body force term is negligible, leaving the waveform of the thermoelastic strain to be established by the traction term which transmits strains from within the thermal boundary layers down to the strainmeter.
position. If, for example, we increase $\kappa$ and/or decrease $\gamma$, so much as to effectively model the strainmeter position in the short-period thermal boundary layers, a change in the waveform of the model thermoelastic strain can be obtained. However, such operations will advance the phase and increase the high-frequency content of the predicted thermoelastic strain, further increasing the discrepancies between predicted and observed strains. If, on the other hand, we decrease $\kappa$ and/or increase $\gamma$, so as to effectively model the strainmeter position further away from the thermal source, the contribution of the body force term will become more negligible while the traction term will continue to transmit the same waveform from the thermal boundary layers (wherever they are) down to the strainmeter position (wherever it is). To illustrate the second case, we generated a second prediction for the half-space model by decreasing $\kappa$ and increasing $\gamma$, three orders of magnitude each. Normalizing this second prediction to the first and plotting both half-space model predictions on the same figure, the two curves appeared indistinguishable on the plot. Thus, in order to account for the discrepancies in phase and frequency content between half-space model prediction and observed strain, it is necessary to alter the thermal signal that drives the half-space. This we do by introducing the elastically decoupled "equivalent upper layer."

The thickness of the equivalent upper layer that is needed to model the phase and frequency content of the observed strain can be found by measuring the delay between components of the same period of observed strain and half-space model prediction. For the annual periods, we find that the observed strain is delayed 23 days with respect to the half-space model prediction. Using the approximation of $\gamma$ for real conditions where $\omega/\kappa \gg k^2$

$$\gamma \approx (1 + i)(\omega/2\kappa)^{1/2} \quad (4)$$
the delay is given by

$$\Delta t = \frac{1}{2} \gamma_b \sqrt{\tau / \pi \cdot k}$$

(5a)

and \( \gamma_b \), the thickness of the "equivalent upper layer," is given by

$$\gamma_b = 2 \cdot \Delta t \sqrt{\pi k / \tau}$$

(5b)

where \( \tau \) in (5a) and (5b) denotes the annual period. Substituting in (5b) the thermal diffusivity coefficient of soil, \( k = 2.16 \cdot 10^{-2} \text{ m}^2/\text{day} \), we find for the BQ strainmeter site that

$$\gamma_b = 2 \cdot 23 \cdot \sqrt{\pi} \cdot 2.16 \cdot 10^{-2} / 366 = 63 \text{ cm}.$$ 

We continue, accordingly, by evaluating the atmospherically produced temperature variations at a depth of 63 cm by use of equation (2), and then taking the temperature variations at that depth to be the input source for (3). The resultant predicted thermoelastic strain of the composite model is shown in Figure 6. The overall phase and frequency content agreement of predicted and observed strains is improved over the fit of the half-space model shown in Figure 5.

The magnitude of the thermoelastic strain is a function of a number of parameters: elastic and thermal moduli; burial depth of the strainmeter; and most importantly the wavelength of the temperature field and the position of the strainmeter in that field. Thus, scaling the predicted thermoelastic strain waveform to the observed strain can be done in various ways. The simplest scaling procedure involves only one free parameter. Depending on our relative ability to estimate the above parameters at a given site, the free parameter should probably be either \( k \) or \( x \). In this paper, analyzing BQ strain data, we use \( k \) as the free scaling parameter. In other cases, using \( x \) may be more appropriate. Scaling the thermoelastic strain to the...
observed strain by use of $k$ is done as follows: since, by assumption of our model, $k$ is independent of the frequency, we fix $\omega$. Having the amplitude of the temperature variations in the decoupling boundary for that frequency, we can generate, using equation (3), the curve $\varepsilon_{2\pi} |_{\omega} = f(k)$. Having the magnitude of the observed strain for that frequency, we find $k$ from the curve $\varepsilon_{2\pi} |_{\omega} = f(k)$. For the annual period, $T_0$ at 63 cm is 5.63°C, $\varepsilon_{2\pi}$ of the observed strain is $4.11 \times 10^{-7}$ strain, and from Figure 7, giving $\varepsilon_{2\pi} |_{\omega=(2\pi/366)} = f(k)$, we find the best-fitting wavelength at BQ to be $\lambda = 3517$ m. To get the final model result, we repeat the composite model prediction using $k = 2\pi/3517$ m$^{-1}$.

To be sure, the recorded strain contains other sources, and it is not expected that the thermoelastic prediction will be identical to the observed strain. The residual strain, left after removal of the thermoelastic prediction from the observed data, is shown in Figure 8 along with rain data and changes in the height of the water in Bouquet Reservoir. It is seen that most of the remaining strain variations are related in time to these other two sources.

**DISCUSSION**

We represent a real thermoelastic environment by a half-space covered with a decoupled "equivalent upper layer" which alters the source temperature field and whose thickness is found by fitting the waveform of the predicted thermoelastic strain to the observed crustal strain. A qualitative conclusion about the relative thickness of the "equivalent upper layer" can be made by examining the frequency content of the recorded strain data, while the exact thickness may be found by comparing the phase delay of components of the same period of observed and half-space model-predicted strains. At sites where the uppermost material is loose, the

![Figure 7](image-url)
thickness of the "equivalent upper layer" and the reduction in the thermoelastic noise, especially in the high-frequency components, will be substantial. In contrast, areas with a competent solid surface rock will be characterized by a thin "equivalent upper layer" or by a half-space only. A preliminary examination of DT1 strain record shown in Figure 9 shows strain signals at periods in the range of days to weeks that are not observed at BQ site (compare with Figure 5). Indeed, these two strain sites are clearly located in different environments. BQ tunnel is located in a hill that is covered by a layer of soil and dense bush, while DT tunnel is situated in bare rock with only sparse patches of soil and vegetation. At BQ, our model suggests that the surface temperature field is low-pass filtered by the unconsolidated ground cover before it generates strains that are recorded by the strainmeter while at DT, with no such cover layer, the strainmeter is subjected to short-period strain noise. The comparison of DT and BQ strain data also shows that the overall strain level in DT is lower than in BQ. We attribute the overall magnitude difference to the factor $f(k)\cos(kx)$ in (3). The topography is similar at both sites; however, BQ strainmeter is located in a south-facing slope and at a distance of about 750 m from Bouquet Reservoir, while DT strainmeter is located near the bottom of a canyon and at a distance of about 150 m from Dalton Reservoir. We suggest that the different positions of the strainmeters in their respective local source fields are responsible for the different strain magnitudes at these two sites.

The sources for the horizontal thermoelastic strain are spatial variations in the applied temperature and in the response of the rock. These variations are created locally by topography and lateral material heterogeneity. Thus, although modeling the environment by a flat and laterally homogeneous media, we expect that the characteristic length of the "source" of the thermoelastic strain will be associated with the characteristic length of topography and lateral material heterogeneity in the considered area. The dependency of the "source" wavelength on the local...
topography is supported by BQ strain data. Figure 10 shows the amplitude spectra of four BQ topography cross-sections of the following directions: N-S; N45E-S45W; E-W; and E45S-W45N. Each cross-section is a 10 km line, centered at BQ strain-meter site. It is seen that the dominant wavelength in BQ local topography is 3.3 ± 1.0 km, comparable with the 3.5 km wavelength found by the fitting process to produce the observed strain magnitude. The prominent 10 km wavelength in the E-W amplitude spectrum comes from the fact that the total cross-section length is 10 km and is actually a representative of a much longer wavelength. Like the topography, or general distribution of heterogeneities, a realistic stationary source field is expected to be more complicated than the pure cosine wave we are using and to contain a spectrum of wavelengths, each contributing to the thermoelastic strain. However, if we assume a uniform thickness for the “equivalent upper layer” and the same time function for all existing wavelengths, each “source” wavelength will generate the same waveform, and the contributions from all the wavelengths will be in phase. In such case, the whole wavelength domain can be substituted by a single “effective” wavelength that can produce the same result, reducing this case back to our simple model. Alternatively, our model may be considered as a one component in the wavelength domain of a more complicated model, where each wavelength is treated separately by the model to produce its waveform and then all the waveforms are summed up to give the final result.

The magnitude of the thermoelastic strain is inversely proportional to the “source” wavelength (see Figure 7), and so the thermoelastic strain will be small in places where the wavelengths of topography and lateral material heterogeneity are large. In the limit of a flat, homogeneous medium, the horizontal thermoelastic strain vanishes. Our model, and the comparison of BQ and DT strain records, also suggest that in a given area the magnitude of the thermoelastic strain varies with position of the strainmeter in the source field.
The roles of the topography and the upper coverage could be better defined by monitoring the three-dimensional temperature distribution in the area surrounding the strainmeter station. In the future, we plan to install several thermistors in shallow holes near the BQ strainmeter site. The residual $1.1 \times 10^{-6}$ strain, obtained by removing the thermoelastic prediction from the 2-yr-long strain record, corresponds to an increase in the ability to detect tectonic signals using these instruments. Successful modeling of the remaining surface effects will enhance that ability further.

![Amplitude spectra](image)

**Fig. 10.** Amplitude spectra of N-S, N45E-S45W, E-W, and E45S-W45N topography cross-sections at BQ.

**CONCLUSIONS**

1. The solution derived by Berger (1975) for an infinite homogeneous, flat, half-space can be applied in an iterative procedure described by our composite model to give the horizontal thermoelastic strain at depth.

2. The horizontal thermoelastic strain will be greater in areas of local topography and/or lateral material heterogeneity (e.g., water/rock boundaries) and smaller in more homogeneous and flat areas.

3. In a given area, the magnitude of the thermoelastic strain varies with position of the strainmeter in the source field.

4. Burial of instruments at shallow depths in places where the upper coverage is thick and well developed will greatly reduce the thermoelastic noise, especially in the high-frequency components.

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