INTERMEDIATE PERIOD RESPONSE OF WATER LEVELS IN WELLS TO CRUSTAL STRAIN:
SENSITIVITY AND NOISE LEVEL

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Abstract. The response of water levels in wells to earth tides indicates that wells can be used to detect small crustal strain. Vertical groundwater flow between the well intake and the water table can significantly attenuate this sensitivity. The attenuation of strain sensitivity as a function of frequency can be inferred from the response of water wells to atmospheric loading. For the wells examined in this study, significant attenuation due to water table drainage can occur when strain accumulates gradually over periods of days to weeks. Despite the presence of attenuation the wells are still sensitive strain meters over this range in period. At a frequency of 2.5 cycles/day, the noise level of the water level records examined is -130 dB relative to 1 strain^2/Hz. This noise level is about the same as that reported for dilatometers and wire strain meters, but is at least 10 dB higher than that reported for laser strain meters. In the frequency band of 0.025 to 2.5 cycles/day, the noise level of the water level records examined increases roughly 25 dB-per-decade decrease in frequency. Some of this noise is due to the influence of atmospheric loading. When the effects of atmospheric loading are removed from the record, the noise level is reduced to roughly 20 dB-per-decade for frequencies above 0.08 cycles/day, a rate typical of high-quality strain meters. For periods slightly less than a month, the wells have a lower noise level in terms of areal strain than that of the best geodetic distance measurements.

Introduction

Water level fluctuations in response to earth tides have long been observed in wells. Despite the apparent sensitivity of wells to strain and a well-established theory that quantitatively describes the interaction between pore pressure and elastic rock deformation [Biot, 1941] the use of water wells to monitor tectonic strain has been limited in extent [e.g., Sterling and Smets, 1971; Johnson et al., 1973, 1974; Roeloffs and Rudnicki, 1984; Lippincott et al., 1985]. The lack of use of water wells as strain meters is partly traceable to the often enigmatic coseismic response of water levels in wells to earthquakes [e.g., Bower and Heaton, 1978]. However, even if the coseismic response of water levels in wells could be generally explained, there would be some obvious problems with using water wells as strain meters. These problems can be separated into two broad classes: those that are related strictly to the bulk material properties of the rock near the well; those that are due primarily to groundwater flow.

The principal problem related to bulk material properties is that the sensitivity of a well to strain is highly dependent on the porosity and elastic properties of the rock or sediment that surround the well intake. A well open to rock or sediment that is highly compressible and porous cannot be expected to be highly sensitive to tectonic strain. This point is discussed in detail elsewhere (S. Rojstaczer and D. C. Agnew, The static response of the water level in an open well to areally extensive deformation under confined conditions, submitted to Journal of Geophysical Research, 1988, hereinafter referred to as submitted manuscript).

Many other problems with using water wells as strain meters are due primarily to the influence of groundwater flow. One problem related to groundwater flow is that the water level in a well responds to hydrology as well as to strain. For example, a year of above average rainfall will likely cause a long-term rise in water level. The influence of rainfall can be viewed as noise placed on any strain signal that might be present in the water level record.

Groundwater flow can cause another significant problem inherent in the use of water wells as strain meters; it may reduce strain sensitivity and thus restrict the usefulness of water wells to a narrow frequency band. There are at least three potential ways for groundwater flow to reduce strain sensitivity (see Figure 1). The first source is the limited hydraulic communication between the well and the saturated rock; if the well is to be a gage of pore pressure, changes in pore pressure must be accompanied by flow into or out of the borehole. For strains above some limiting frequency (that depends on the borehole geometry and the material properties of the saturated rock), groundwater flow will be insufficient to allow the full pore pressure signal to be seen in the well.

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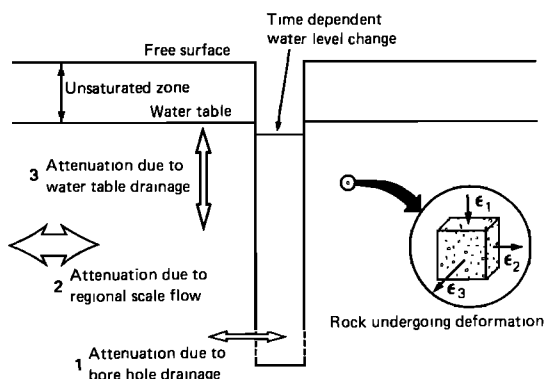
Fig. 1. Cross section of well responding to
deformation and principal sources of attenuation
of well response.

The second source of sensitivity reduction due
to groundwater flow is large-scale horizontal
transient flow from or to the monitored region of
strain-induced pore pressure change. For strain
of tectonic origin, the wavelengths of the
pressure disturbance are likely of the order of
kilometers or greater, so that this problem is
probably significant only at very low
frequencies.

The third source of sensitivity deterioration
is vertical flow from or to the water table.
Since the water table is particularly insensitive
to crustal strain [Bredehoeft, 1967], any
hydraulic connection between the water table and
the zone of saturated rock monitored can cause
significant attenuation of the strain-induced
pore pressure signal. Water wells are typically
in communication with rocks that are less than
200 m below the water table, and this source will
likely be significant at frequencies higher than
those of the second source.

The problem of noise in the water level signal
has not been quantitatively examined, but
problems of strain sensitivity related to
groundwater flow have been studied previously in
some detail. Sensitivity attenuation caused by
fluid flow into and out of the well has been
extensively examined [e.g., Cooper et al., 1965;
Bodvarrson, 1970; Johnson, 1973; Gieske, 1986]
and is discussed elsewhere [Rojstaczer, 1988].
The influence of the water table on attenuation
has been examined by Johnson [1973], A. G.
Johnson and A. Nur (unpublished manuscript,
1978), and Bower and Heaton [1978]; these studies
theoretically examined this influence assuming
that the unsaturated zone above the water table
did not significantly influence water table
response.

This study examines the influence of vertical
groundwater flow to the water table on the
sensitivity of water wells as strain meters and
the noise levels of water wells in comparison
with those of other strain meters. The influence
of vertical groundwater flow on attenuation of
sensitivity is examined by: extending the

theoretical work of Johnson [1973] and A. G.
Johnson and A. Nur (unpublished manuscript, 1978)
to include the influence of the unsaturated zone
above the water table on atmospheric pressure
induced strain; applying these theoretical
results to the response of two wells to
atmospheric loading in order to examine how
vertical groundwater flow influences strain
sensitivity as a function of frequency. The
comparative performance of water wells as strain
meters is assessed by calibrating two wells to
known strains and examining noise levels in the
frequency band of 0.025 to 2.5 cycles/day.

Static-Confined Sensitivity to Strain and Equations Governing Groundwater Flow

Before we examine the influence of water table
drainage on water level response, it is useful to
briefly examine (1) the response when the depth
interval tapped by the well is hydraulically
isolated from the water table and the influence
of fluid flow between the well intake and the
rock formation is negligible; (2) the equations
governing groundwater flow under conditions of
areally extensive deformation. I focus on the
response of water level to two types of
deformation: imposed areal strain and imposed
atmospheric loading. In a later section the
response of wells to atmospheric loading will be
used to infer how water table drainage influences
the response of water levels to areal strain.
The relations contained in this section can be
readily obtained from the Biot [1941] theory of
poroelasticity, and details on the derivations
can be found elsewhere (S. Rojstaczer and D. C.
Agnew, submitted manuscript).

When the influences of water table drainage
and borehole storage are negligible, the response
of the water level in the well reflects the
undrained response of the rock formation. In
this paper, the response to deformation under
these conditions is called the static-confined
response. The static-confined response
(extension taken to be positive) to areal strain,
A'_s, is given by [Van der Kamp and Gale, 1983]

$$A'_s = \frac{w}{\epsilon_a} = \frac{3}{\rho g \beta}\left[\frac{B(1-2\nu)}{2\alpha B(1-2\nu)-3(1-\nu)}\right] \qquad (1)$$

where w is the water level in the well, ϵ_a is the
areal strain (sum of the principal horizontal
strains), β is the rock matrix compressibility, ρ
is the fluid density, g is gravitational
acceleration, ν is Poisson's ratio, B is
"Skempton's coefficient" which relates mean
stress to pore pressure under undrained
conditions [Skempton, 1954; Rice and Cleary,
1976]:

$$B = \frac{\beta - \beta_u}{(\beta - \beta_u) + \phi(\beta_f - \beta_u)} \qquad (2)$$

and α is the fraction of rock strain taken up by

the pore space under drained conditions [Biot and Willis, 1957; Nur and Byerlee, 1971]:

$$\alpha = 1 - \beta_u/\beta \qquad (3)$$

It should be noted that β_f and β_u are the fluid and rock grain compressibility, respectively, and ϕ is the porosity.

In an open well, the static-confined response to atmospheric loading, σ_a, is

$$E'_B = \frac{w\rho g}{\sigma_a} = 1 - \frac{2B(1+\nu)}{3-(1-2\nu)\alpha B} \qquad (4)$$

The parameter, E'_B, is called the static-confined barometric efficiency (S. Rojstaczer and D. C. Agnew, submitted manuscript). Equation (4) differs slightly from the derivation of E'_B that can be obtained from Van der Kamp and Gale [1983]. The essential difference is that (4) assumes that the atmospheric load is placed upon an elastic half-space. Under this assumption, atmospheric loading produces areal strain, ϵ_a, equal to the vertical strain, ϵ_z, at typical well depths [Farrell, 1972]. Van der Kamp and Gale [1983] assume that surface loading induces vertical strain only (a traditional assumption in hydrology [e.g., Jacob, 1940]). The appropriateness of either assumption ($\epsilon_a = \epsilon_z$ or $\epsilon_a = 0$) is discussed elsewhere (S. Rojstaczer and D. C. Agnew, submitted manuscript).

The static-confined areal strain sensitivity, A'_s, and barometric efficiency, E'_B, influence the degree that groundwater flow is caused by areally extensive deformation. At typical well depths, groundwater flow in response to laterally extensive areal strain such as that imposed by earth tides and broad-scale tectonic deformation is governed by [Van der Kamp and Gale, 1983]

$$D' \frac{\partial^2 p}{\partial z^2} = \frac{\partial p}{\partial t} + \rho g A'_s \frac{\partial \epsilon_a}{\partial t} \qquad (5)$$

where p is pore pressure, D' is a hydraulic diffusivity for imposed areal strain under conditions of plane stress [Van der Kamp and Gale, 1983]:

$$D' = \frac{k}{\mu}\left\{\rho g \alpha \beta [1 - \frac{2\alpha(1-2\nu)}{3(1-\nu)}] + \phi(\beta_f - \beta_u)\right\}^{-1} \qquad (6)$$

and k and μ are the permeability of the formation and the viscosity of the pore fluid, respectively.

In response to areally extensive surface loading (such as that produced by atmospheric loading), groundwater flow at the near surface of a half-space is governed by (S. Rojstaczer and D. C. Agnew, submitted manuscript)

$$D \frac{\partial^2 p}{\partial z^2} = \frac{\partial p}{\partial t} + (1-E'_B) \frac{\partial \sigma_a}{\partial t} \qquad (7)$$

where D is a hydraulic diffusivity under

conditions of areally extensive imposed surface stress:

$$D = \frac{k}{\mu}\left\{\rho g \alpha \beta [1 - \frac{\alpha(1-2\nu)}{3}] + \phi(\beta_f - \beta_u)\right\}^{-1} \qquad (8)$$

Equations (5) and (7) indicate that pore pressure response to imposed areal strain and atmospheric loading are both described by diffusion equations that contain a source term. The source term in the case of imposed areal strain has (owing to the hydraulic diffusivity, D', being slightly larger than the hydraulic diffusivity, D, for equal-bulk material properties) a weaker influence on time dependent changes in pore pressure. The diffusion equations given above are an accurate description of pore pressure response as long as we can assume that in the frequency range of interest there is no lateral pore pressure dissipation.

Influence of Water Table Drainage on the Sensitivity of Water Levels in Wells to Areally Extensive Strain

Figure 2 shows idealized descriptions of water level response to two types of deformation: an imposed areal strain and an imposed atmospheric load. In both descriptions I neglect the influence of borehole drainage and assume that the water level in the well is a direct reflection of pore pressure in the rock
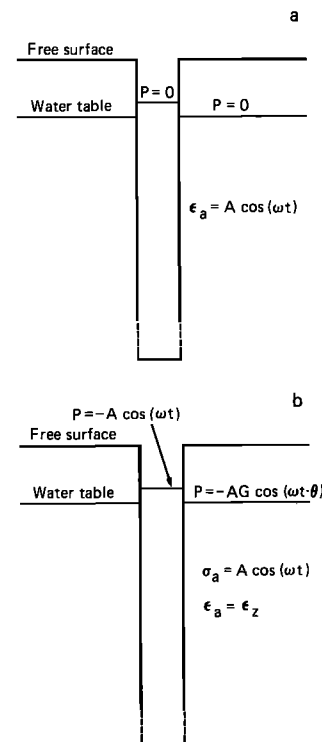

Fig. 2. (a) Idealized model of response of well to areally extensive periodic strain; (b) Idealized model of response of well to areally extensive periodic atmospheric loading.

formation. For the areal strain problem, the
rock is subject to a periodic areal strain
$A \cos(\omega t)$. The water table is at zero pressure
and the well is an accurate gage of pore
pressure.

For the atmospheric loading problem the rock
is subject to a periodic vertical stress
$A \cos(\omega t)$, and areal strain, ϵ_a, equals the
vertical strain ϵ_z. The water table is subject
to a pressure $-AG \cos(\omega t + \theta)$, where G and θ
account for the attenuation and phase advance of
the atmospheric pressure owing to diffusion of
air through the unsaturated zone above the water
table. The water in the well is subject to a
periodic pressure $-A \cos(\omega t)$ (i.e., the well is
open to the atmosphere), and the relation between
water level, w, and pore pressure is [Rojstaczer,
1988]

$$w = [A \cos(\omega t) + p]/\rho g \qquad (9)$$

Since fluid flow is assumed to be one
dimensional, both problems outlined above are
readily solved analytically.

Frequency Response of a Well to Imposed Areal Strain

The response of a water well to imposed
periodic areal strain is given by (Appendix A)

$$w = [AC \exp(-\sqrt{\overline{Q'}}) \cos(\omega t - \sqrt{\overline{Q'}}) - AC \cos(\omega t)]/\rho g \qquad (10)$$

where A is the amplitude of the areal strain and
Q' is a dimensionless frequency:

$$Q' = z^2 \omega / 2D' \qquad (11)$$

The result given by (10) is qualitatively similar
to the frequency dependent response of water
wells to strain given by Johnson [1973] and A. G.
Johnson and A. Nur (unpublished manuscript,
1978). The major differences between their
analogous solution and that given here are due to
their approximation of the water table as a
spherically shaped boundary. Equation (11)
indicates that the most important parameter
governing sensitivity as a function of frequency
is the depth from the water table, z. Hydraulic
diffusivity is also an important factor, with low
diffusivity favoring low attenuation at a given
frequency.

The gain and phase (phase advance is positive)
of the sensitivity to areal strain are plotted as
a function of the dimensionless frequency, Q', in
Figure 3 under the assumption that the
static-confined sensitivity of the well to areal
strain, A_s, is 0.05 cm water level drop per areal
nanostrain. Attenuation and phase shift begin to
significantly deviate from the static-confined
response when the dimensionless frequency
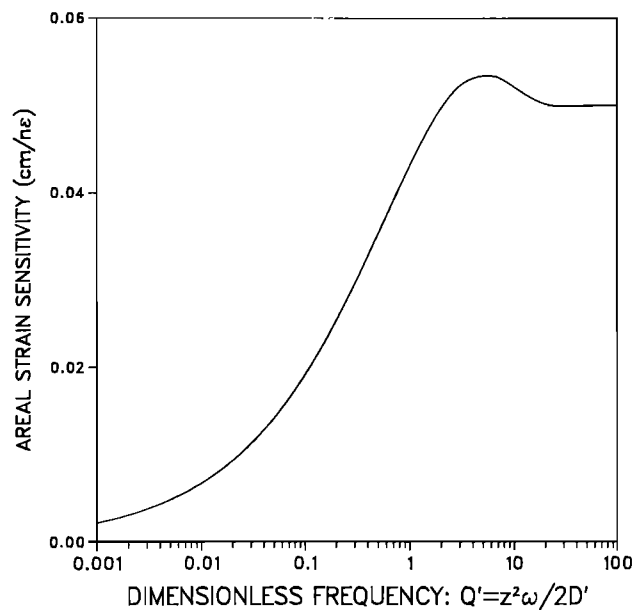decreases to a value of 10. Between a

DIMENSIONLESS FREQUENCY: $Q'=z^2\omega/2D'$

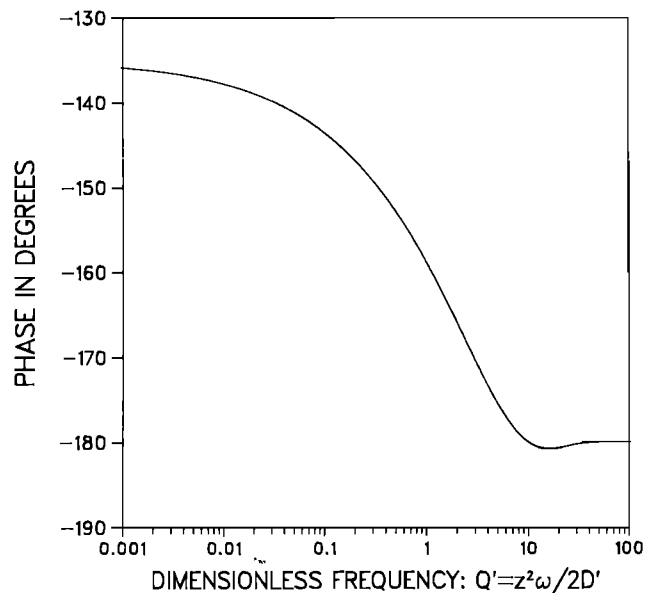Fig. 3a

DIMENSIONLESS FREQUENCY: $Q'=z^2\omega/2D'$

Fig. 3b

Fig. 3. Response of well to areally extensive
periodic strain in terms of (a) areal strain
sensitivity (centimeters water level drop per
areal nanostrain) and (b) phase.
Static-confined areal dilatational efficiency
is 0.05 cm/nϵ.

dimensionless frequency of 1 and 10, the strain
signal is slightly amplified relative to
static-confined conditions; near-complete
attenuation of the strain signal takes place for
strain with dimensionless frequency 0.001 or
less.

It is useful to examine the theoretical
response of a well that taps rock with reasonable
hydraulic and mechanical properties. Given a
well tapping rock with a vertical permeability of

10^{-14} m² (the average permeability of nonargillaceous materials inferred for the crust [Brace, 1980, 1984]), a matrix compressibility of 1 x 10^{-10} Pa^{-1}, a fluid compressibility of 4 x 10^{-10} Pa^{-1}, a solids compressibility of 2 x 10^{-11} Pa^{-1}, a porosity of 0.10, and a well depth (relative to the water table) of 100 m, the dimensionless frequency will have a value greater than 10 for strain with frequency greater than 3 cycles/day. This result indicates that for reasonable material properties and geometries, vertical groundwater flow to the water table can have a strong influence on strain sensitivity in the frequency range of practical interest.

Frequency Response of a Well to Atmospheric Loading

The response of a well to atmospheric loading is considerably different than that given above. These differences are due primarily to the diffusion of the load through the air phase of the unsaturated zone [Weeks, 1979]. Water level response in an open well to periodic fluctuations in atmospheric loading is given by (Appendix B)

$$w = [(-M+\gamma)A \ exp(-\sqrt{Q}) \ cos(\omega t-\sqrt{Q}) - (\gamma-1)A \ cos(\omega t) - NA \ exp(-\sqrt{Q}) \ sin(\omega t-\sqrt{Q})]/\rho g \quad (12)$$

where M and N are

$$M = \frac{2 \ cosh(\sqrt{R}) \ cos(\sqrt{R})}{cosh(2\sqrt{R}) + cos(2\sqrt{R})} \quad (13a)$$

$$N = \frac{2 \ sinh(\sqrt{R}) \ sin(\sqrt{R})}{cosh(2\sqrt{R}) + cos(2\sqrt{R})} \quad (13b)$$

Q and R are dimensionless frequencies referenced to fluid diffusivity, D, and air diffusivity, D_a, respectively:

$$Q = z^2\omega/2D \quad (14a)$$

$$R = L^2\omega/2D_a \quad (14b)$$

and L is the depth from the Earth's surface to the water table. The gain (barometric efficiency, E_B) and phase of the response of a water well to atmospheric pressure fluctuations are shown in Figure 4 for a well with a static-confined barometric efficiency, E_B', of 0.5.

The response is plotted as a function of two dimensionless parameters: dimensionless frequency, Q, and the ratio of dimensionless frequencies R and Q. The dimensionless ratio, R/Q, is a measure of the time taken for atmospheric pressure changes to reach the water table versus the time taken for water table drainage to significantly influence water well response.

For values of R/Q less than 0.0001 the water table is fully influenced by atmospheric pressure changes at dimensionless frequency, Q, and the
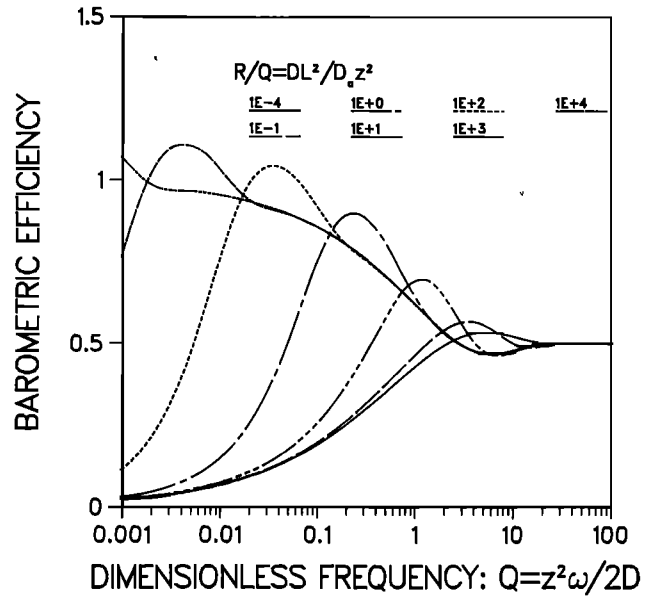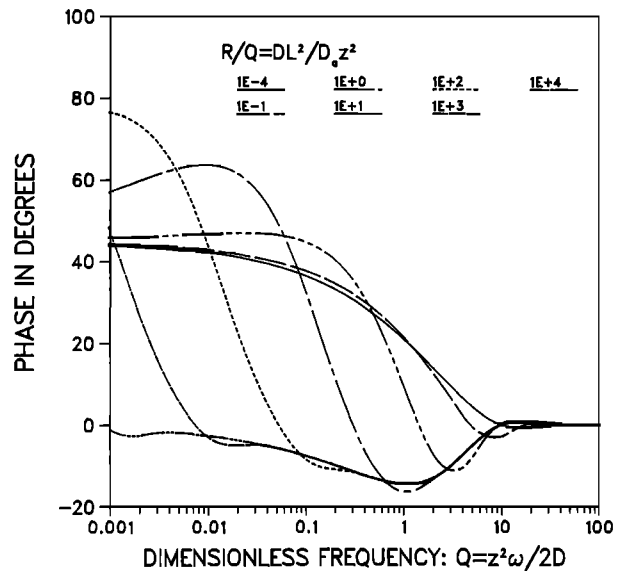

Fig. 4a

Fig. 4b

Fig. 4. (a) Barometric efficiency and (b) phase of response to atmospheric loading as a function of frequency and R/Q. Static-confined barometric efficiency is 0.5.

response of the water well to atmospheric loading is functionally identical (except for a phase shift of 180°) to the response to strain which is applied directly to the solid phase alone (equation (10)). Under these conditions the attenuation of sensitivity to tectonic or earth tide induced strain can be inferred from the air pressure response once an allowance is made for the slightly different hydraulic diffusivity that governs fluid flow. For values of R/Q greater than 0.1 there is significant attenuation and phase shift of the atmospheric pressure signal at the water table; under these conditions we can

expect that over a frequency band whose width and
location are a strong function of R/Q, the water
well response will be significantly amplified
relative to the static-confined response. When
R/Q is greater than 10, we can expect a water
well to respond significantly to atmospheric
loading in a frequency band where response to
tectonic or earth tide induced strain is strongly
attenuated. Hence the response of a water well
to atmospheric loading, when R/Q is significantly
greater than 0, is not indicative of how the
sensitivity to tidal or tectonic strain
attenuates with frequency.

Although it is not possible, on the basis of
the response to atmospheric loading, to directly
infer how the sensitivity of water wells will
attenuate in response to imposed areal strain
when R/Q is 0.1 or greater, the atmospheric
pressure response gives us an indirect means to
determine how water levels in wells respond to
tectonic strain. The material and fluid flow
properties of the rock and/or sediment which
control well response can be determined by
fitting the atmospheric pressure response of a
well to the theoretical response (equation (12)
or (B6)). These elastic and fluid flow
properties can then be used in conjunction with
the theoretical response to imposed areal strain
(equation (10) or (A3)) to infer how sensitivity
to tectonic strain attenuates with frequency.

Overview of Wells to Be Examined

Figure 5 shows the water level record for the
two wells examined here. These wells were chosen
because: they are located in areas which have
been the focus of many crustal deformation
studies over the past decade; and they exhibited
the largest amplitude response to the strain
induced by earth tides of any of the wells
monitored in these areas over the time shown.
The hydrograph labeled LKT is the water level
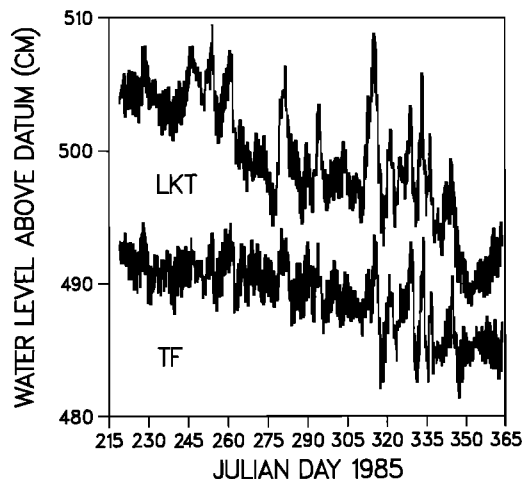record from a well in the Long Valley caldera

Fig. 5. Water level record at LKT and TF during
the second half of 1985.

near Mammoth Lakes, California; the hydrograph
labeled TF is the record from a well near the San
Andreas fault at Parkfield, California.

The water level records at LKT and TF over the
second half of 1985 both show a long-term decline
in water level; this low-frequency decline is at
least partially due to the relative lack of
precipitation in central California over the
winter of 1984-1985. The power spectra of the
water level records are shown in Figure 6. Both
wells have considerable power in the semidiurnal
and diurnal frequency bands owing to earth tides
and atmospheric loading. Power levels increase
with decreasing frequency.

Figure 6 also shows the power spectra of
atmospheric pressure at TF and LKT over the
second half of 1985 (pressure is in terms of
equivalent water level). Comparison of the
atmospheric pressure power spectra with the water
level power spectra at TF and LKT indicates many
similarities. There is substantial power in the
semidiurnal and diurnal band and a strong
increase in power with decreasing frequency.

The depth and near-hole lateral permeabilities
of these two wells are shown in Table 1. For
both wells, the well intake is isolated from the
near surface to a depth in excess of 100 m. The
depth to water is about 20 m at both LKT and TF;
if near-hydrostatic conditions prevail at these
wells, then this depth indicates the depth to the
water table at both sites. The lateral
permeabilities of the rock in direct hydraulic
communication with these wells were inferred from
a slug test [Kipp, 1985] and pumpage data from
LKT and TF, respectively. These permeabilities
are high in relation to the range of
permeabilities that have been measured or
inferred for the crust [Brace, 1980, 1984]
indicating that hydraulic communication between
the well and rock is relatively good. The
borehole diameter is roughly 15 cm at both wells.

At each site, atmospheric pressure and water
level were measured with silicon strain bridge
transducers four times per hour. The water level
and atmospheric pressure records over the time
interval shown in Figure 5 contain 2% and 5% gaps
at LKT and TF, respectively; these gaps range
from 1/2 hour to a few days in length. The
method by which these gaps in the time series
were filled is described in Appendix C. The
tidal areal strain time series at each site was
determined from the body tide with no corrections
made for ocean loading, topographic, or geologic
effects. The results of Beaumont and Berger
[1975], Berger and Beaumont [1976] and Evans et
al. [1979] suggest that the amplitude of actual
tidal strain agrees with those determined from
the homogeneous earth tide with an error of about
±50%.

Inferred Strain Sensitivity of Wells

The relation of water level to crustal strain
induced by atmospheric loading and earth tides at
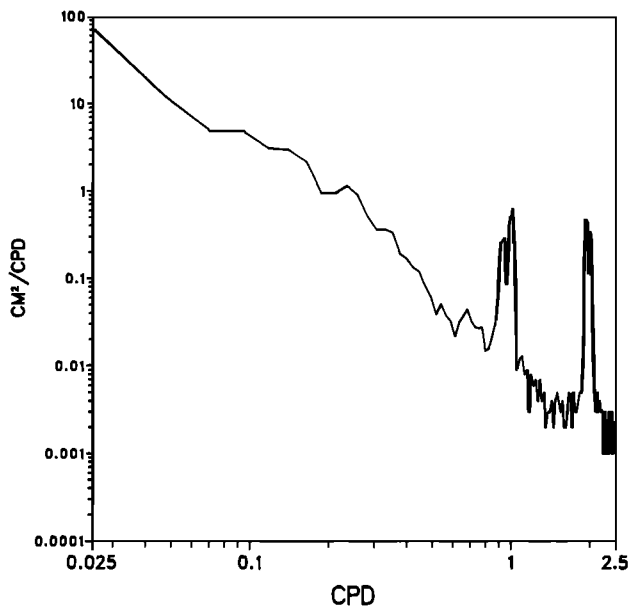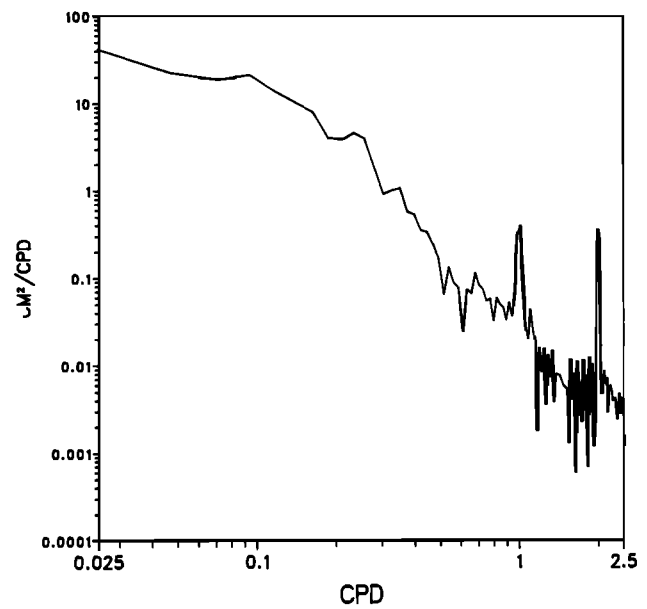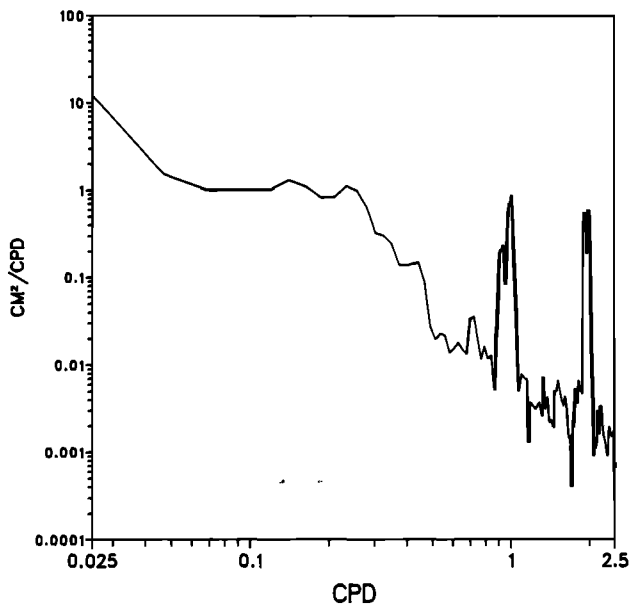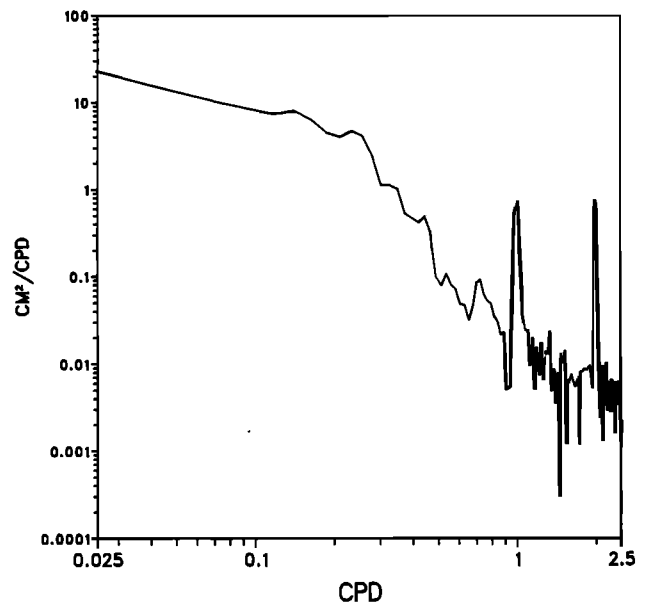the two wells was determined from cross-spectral

Fig. 6. Power spectral densities for (a) water level and (b) atmospheric pressure at LKT; power spectral densities for (c) water level and (d) atmospheric pressure at TF. The abbreviation "CPD" denotes cycles per day.

estimation (Appendix D). Table 2 indicates the response of these wells to the M_2 and O_1 earth tides in terms of their areal strain sensitivity (centimeters/nanostrain) and phase. Their response to the M_2 tide in terms of areal strain sensitivity is identical; for both sites the areal strain sensitivity with regard to the M_2 component of the tidal potential is 0.034 cm/nϵ. The areal strain sensitivity of the O_1 tide is slightly lower, particularly at TF. At TF this slightly greater difference in amplitude may be due partly to water table drainage which (as is shown below) significantly attenuates strain sensitivity in the frequency band of interest

(0.025 to 2.5 cycles/day). Phase differences at LKT and TF between the O_1 and M_2 tidal constituents are slight. The phase shift of both constituents as well as the small difference in areal strain sensitivity between the M_2 and O_1 constituents at LKT are likely due to elastic anisotropy, local inhomogeneities, or ocean loading effects. At TF the phase shift of both constituents may be due to a combination of water table drainage, elastic inhomogeneity or anisotropy, and ocean loading influences.

The response of LKT to atmospheric loading is shown in Figure 7. The phase is generally flat and near $0°$ for frequencies greater than 0.02

TABLE 1. Description of Wells Examined

Well Name	Location in California	Open Interval, m	Local Permeability, m^2
LKT	Long Valley	152-296	2×10^{-11}
TF	Parkfield	152-177	2×10^{-14}

Local permeability refers to the permeability near the borehole.

cycles/day. The admittance or barometric efficiency begins to show some attenuation below a frequency of 0.05 cycles/day. This attenuation may not reflect any influence of groundwater flow to the water table; rather it may be the result of some error in the estimate of barometric efficiency at low frequencies. For this study, use of error estimates for the admittance determined from cross-spectral estimation [e.g., Bendat and Piersol, 1986] yield the following 95% confidence interval for the barometric efficiency, $E_{B.95}$:

$$E_{B.95} = E_B \pm 1.2\sqrt{(1-\Gamma^2)}E_B \qquad (15)$$

where Γ is the coherence. Figure 8 shows the coherence squared, as a function of frequency for LKT. At frequencies less than 0.05 cycles/day, the coherence squared is substantially less than 1, and the error bounds on barometric efficiency are greater than ±0.27. A flat barometric efficiency of 0.48 (the mean value of the barometric efficiency for frequencies greater than 0.05 cycles/day) over the entire frequency range examined is thus not inconsistent with the admittance values and is consistent with the general frequency insensitivity of the phase. Alternatively, the attenuation that begins to occur at frequencies less than 0.05 cycles/day is a real physical phenomenon, presumably due to the influence of groundwater flow to the water table.

The fit to the data, based on the assumption that there is some water table drainage in the observed frequency band, is also shown in Figure

TABLE 2. Tidal Response of Wells Examined

Well Name	O_1 Gain	O_1 Phase	M_2 Gain	M_2 Phase
LKT	0.031	-166	0.034	-171
TF	0.030	-189	0.034	-188

Gain expressed in terms of centimeters of water level drop per areal nanostrain. Phase expressed in degrees.
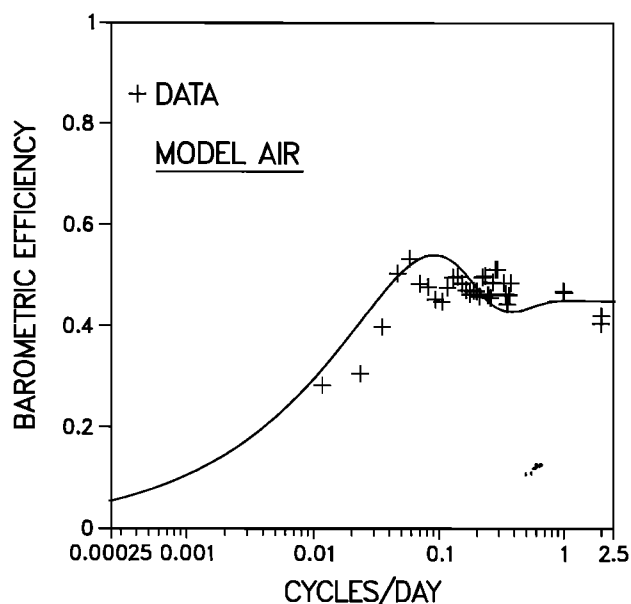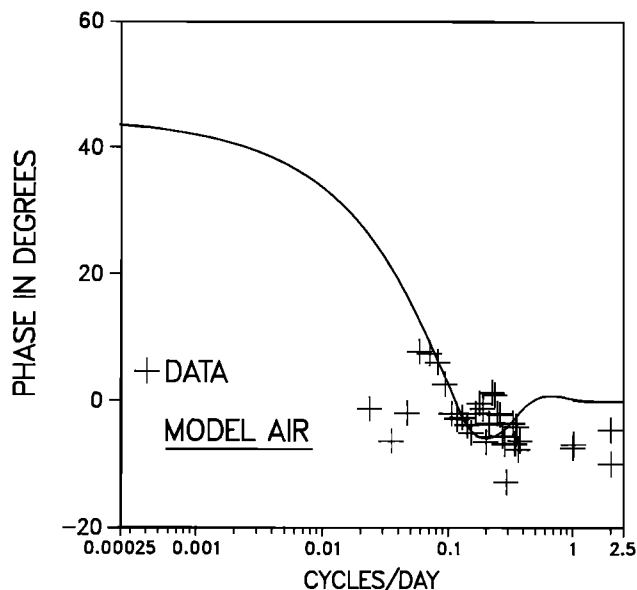

Fig. 7a

Fig. 7b

Fig. 7. Response of LKT to atmospheric loading in terms of (a) barometric efficiency and (b) phase. Fit to data is solid line denoted as "MODEL AIR."

7. The best fit to the data is achieved with a static-confined barometric efficiency, E_B', of 0.45, a value for dimensionless frequency Q of 4.8ω, and a value for dimensionless frequency R of 0.94ω, where frequency ω is in units of radians/day. This value of R yields a pneumatic diffusivity of 2×10^{-3} m^2/s. Since pneumatic diffusivities under field conditions are not commonly measured, it is difficult to assess whether or not the pneumatic diffusivity inferred here is reasonable. It is a factor of 30 lower than a pneumatic diffusivity inferred by Weeks
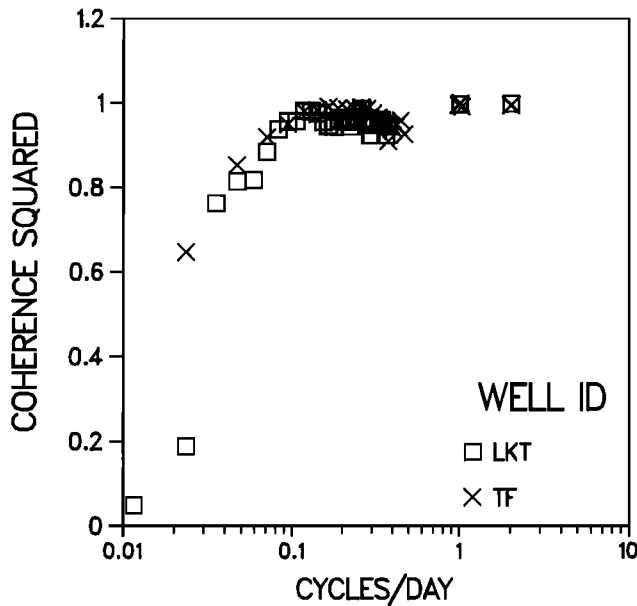
Fig. 8. Multiple coherence squared for relation
of water level to atmospheric loading and earth
tides at LKT and TF.

[1979] from the response of a well in Lubbock,
Texas, to atmospheric loading.

The value of Q' consistent with the
atmospheric loading response is 4.1ω. This value
for Q' indicates that the observed M_2 and O_1
tidal responses are not influenced by water table
drainage. The value for the static-confined
barometric efficiency given by this fit combined
with the static-confined areal strain sensitivity
estimated from the M_2 tidal response (shown in
Table 2) allow us to calculate (S. Rojstaczer and
D. C. Agnew, submitted manuscript) a
compressibility for the rock in communication
with the well. Assuming a Poisson's ratio of
0.25, the drained rock matrix compressibility is
1.5×10^{-10} Pa^{-1}. Given this compressibility and
assuming that the depth to water is indicative of
the depth to the water table, it is also possible
to estimate the vertical permeability of the zone
between the water table and the depth of the well
intake. This permeability is 7×10^{-16} m^2, which
is much less than the lateral permeability given
in Table 1. The difference indicates either that
moderately low permeability layers exist above
the monitored zone or that there is considerable
anisotropy in the permeability of the rock and
sediment above the zone monitored. It should be
noted that this inferred permeability is a
maximum permeability based upon the assumption
that the attenuation indicated at low frequencies
is a real phenomenon.

The elastic and fluid flow properties
determined from the response of LKT to
atmospheric loading are used to infer how this
well responds to crustal strain (Figure 9; note
that we assume that the attenuated response to
atmospheric loading reflects a real physical

phenomenon). In terms of attenuation of
sensitivity, it is not appreciably different than
the atmospheric loading response. This
correspondence suggests that the water table is
in good pneumatic communication with the Earth's
surface. The inferred sensitivity remains
moderately high when frequencies are less than
0.001 cycles/day.

The response of TF to atmospheric loading is
shown in Figure 10. This response has a
signature which indicates that the influence of
groundwater flow to the water table is present
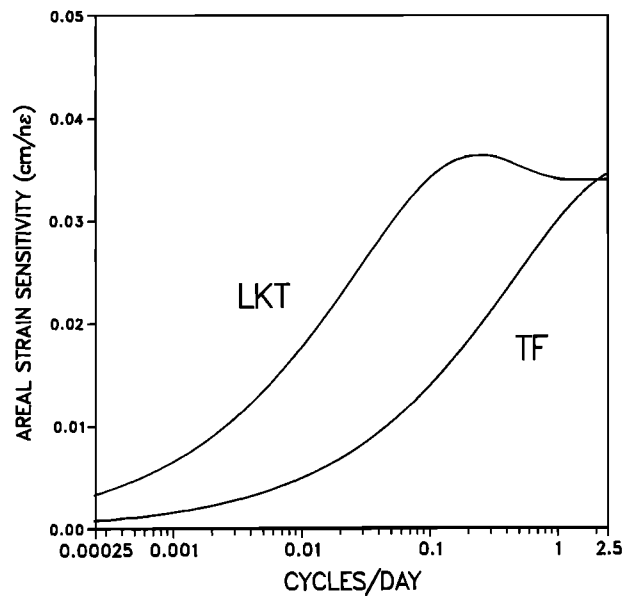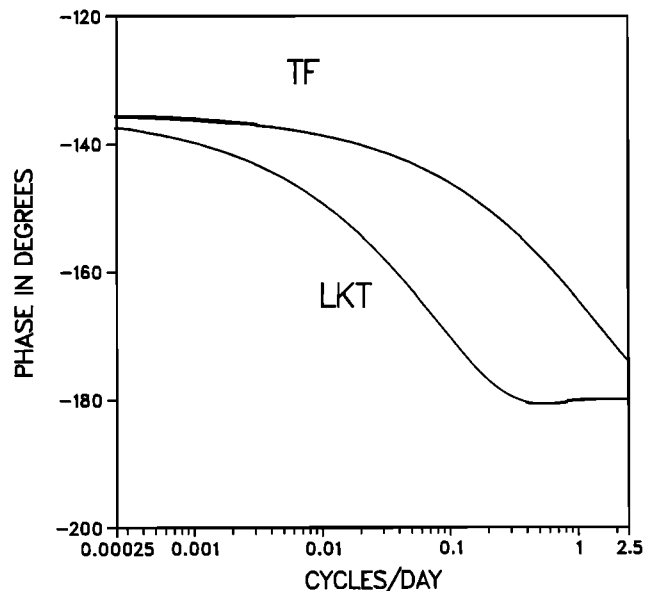throughout the observed frequency band. The

Fig. 9a

Fig. 9b

Fig. 9. Inferred strain response of LKT and TF
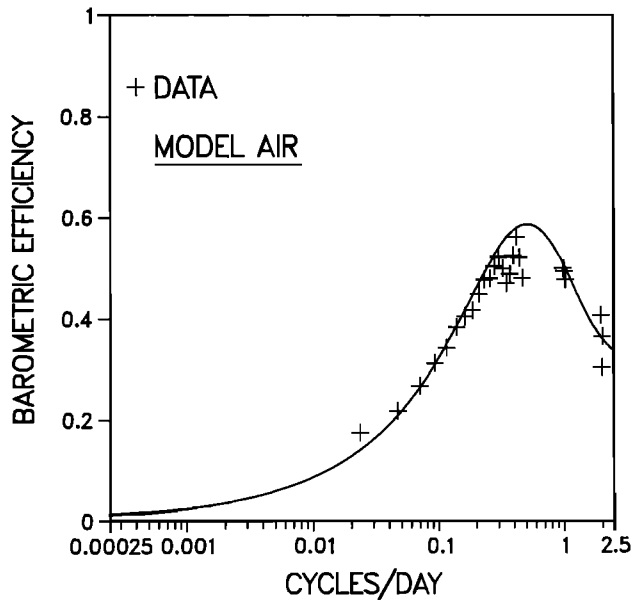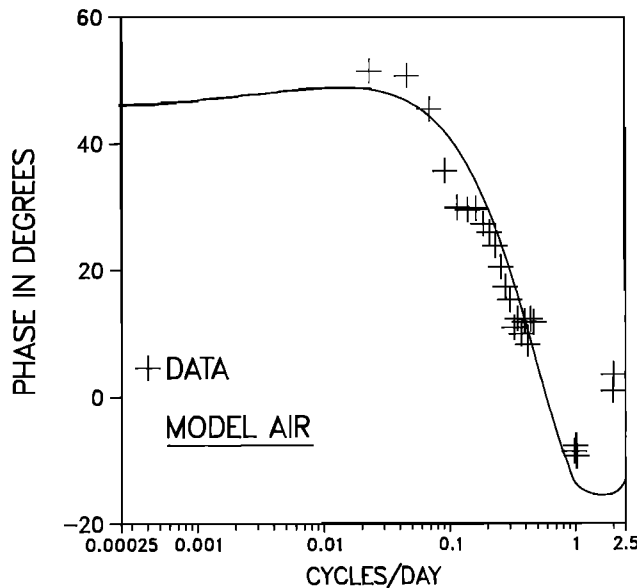in terms of (a) areal strain sensitivity and (b)
phase.

Fig. 10a

Fig. 10b

Fig. 10. Response of TF to atmospheric loading in terms of (a) barometric efficiency and (b) phase. Fit to data is solid line denoted as "MODEL AIR."

relative lack of ambiguity in this signal is evident in the coherence squared for the transfer function (Figure 8), which is near 1 for frequencies in excess of 0.06 cycles/day. The response to atmospheric pressure has a maximum at about 0.5 cycles/day. It might be suspected that decreasing sensitivity with increasing frequency above the frequency of 0.5 cycles/day would be due (at least partially) to limited groundwater flow into and out of the borehole; this source of attenuation is unlikely because it is inconsistent with the near-flat areal strain sensitivity for the semidiurnal and diurnal tidal

constituents shown in Table 2 and is also inconsistent with the phase of the atmospheric loading response. Rather, this decreasing sensitivity at higher frequencies is likely due to air pressure diffusion effects which amplify the atmospheric loading response in the frequency band of 0.2 to 2 cycles/day. As Figure 4 indicates, such a response would be theoretically possible if the dimensionless numbers, R and Q, were of the same magnitude.

The best fit to the data is given with a static-confined barometric efficiency of 0.37 and a value for both Q and R of 0.34ω, where frequency is in terms of radians/day. The fit to the atmospheric loading response indicates that water table drainage also influences areal strain response at frequencies less than about 5 cycles/day. The value of R yields a pneumatic diffusivity of 5×10^{-3} m^2/s. The value for Q' which is consistent with the atmospheric loading response is 0.27ω. Fitting the strain response to the observed M_2 areal strain sensitivity given in Table 2 yields a static-confined areal strain sensitivity of 0.033 cm/nϵ. These values of static-confined barometric efficiency and areal strain sensitivity indicate that the drained rock matrix compressibility for the rock in direct communication with the well bore is 1.9×10^{-10} Pa^{-1} (S. Rojstaczer and D. C. Agnew, submitted manuscript). Given this compressibility, the vertical permeability above the well intake is estimated to be 1×10^{-14} m^2, a value that is only slightly less than the lateral permeability inferred from pumping data. This correspondence indicates that the material above and within the zone monitored is largely homogeneous and isotropic with respect to permeability.

The inferred response to tectonic strain is shown in Figure 9. In terms of attenuation of sensitivity it is considerably different than the atmospheric loading response. The inferred response to strain approaches static-confined conditions at a frequency of 2.5 cycles/day, but the static-confined response is never fully observed in the frequency band analyzed. Below a frequency of 0.001 cycles/day, where strain sensitivity is less than 1/10 of the inferred static-confined response, strain sensitivity asymptotically approaches zero.

Strain Spectra of Wells

The raw and filtered (atmospheric loading effects removed) strain spectra for LKT and TF are shown in Figure 11. These strain spectra were derived by (1) obtaining power spectral densities for the water level records in terms of centimeters^2 per cycle/day; (2) converting these spectra into units of strain^2/hertz through the use of the estimated static-confined areal strain sensitivities for the wells (0.034 cm/nϵ for LKT and 0.033 cm/nϵ for TF); (3) adjusting the strain
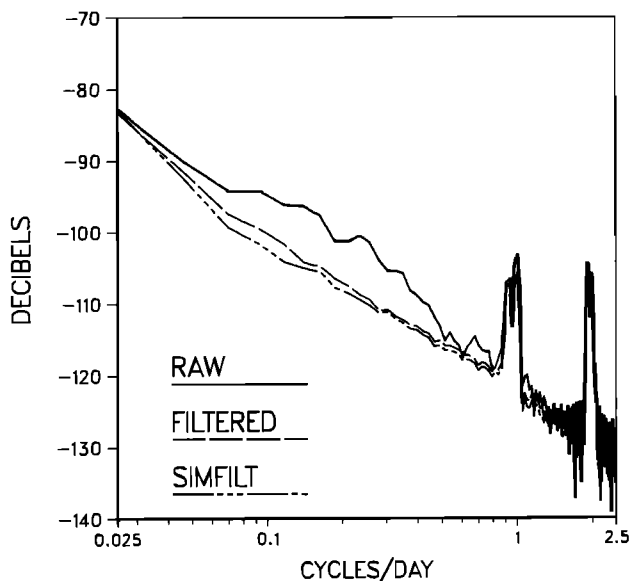
Fig. 11a
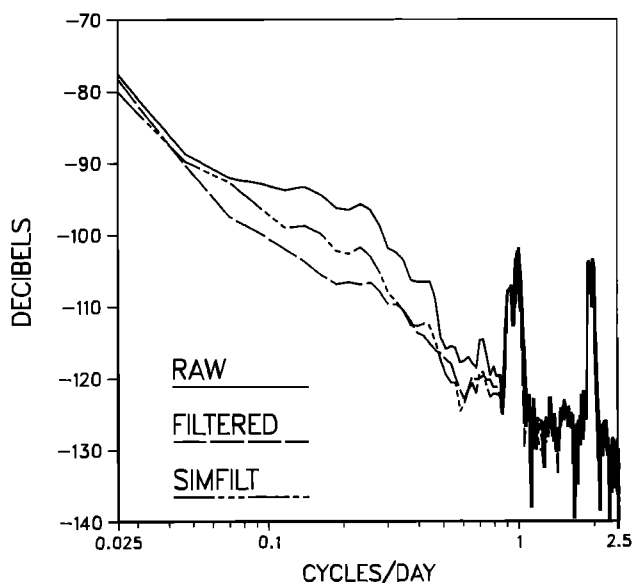

Fig. 11b

Fig. 11.   Power spectral densities at (a) LKT and (b) TF normalized to 1 strain^2/Hz.   "RAW" curve is water level record without atmospheric loading effects removed.   "FILTERED" curve is water level record with atmospheric loading effects removed with a frequency dependent transfer function. "SIMFILT" curve is water level record with atmospheric loading effects removed with a filter of length one.

spectra, when necessary, to account for any unambiguous frequency dependent changes in areal strain sensitivity inferred from the response to atmospheric loading; (4) normalizing the spectra relative to 1 strain^2/Hz and converting to a decibel scale.   The strain spectra were determined with no detrending of the data. Analysis of the computation of the strain spectra indicated that including the trends caused only

slight changes in the strain spectra (generally less than 1 dB) for frequencies greater than 0.05 cycles/day.

The spectral estimates TF were adjusted upward to account for the observed decreasing sensitivity with period by essentially adding a slope of 8 dB-per-decade decrease in frequency to the strain spectra for frequencies less than 1 cycle/day; since the observed attenuation at LKT is slight and ambiguous in the frequency band examined here, no adjustments in the strain spectra were made.   The raw strain spectra at LKT and TF are very similar.   At 2.5 cycles/day the noise levels for both wells are at about -130 dB relative to 1 strain^2/Hz.   This noise level is at least 10 dB higher than that reported for laser strain meters [Berger and Levine, 1974; Beavan and Goulty, 1977; Agnew, 1986] but is comparable to that reported for dilatometers [Johnston et al., 1986] and wire strain meters [Bilham et al., 1974].   At higher frequencies, noise levels increase at a rate of roughly 25 dB-per-decade decrease in frequency, a rate that is slightly high compared to at least one dilatometer [Johnston et al., 1986].   The M_2 tide at both wells is roughly 20 dB above the background noise level.

Considerable reduction in noise level can be achieved by removing the effects of atmospheric loading on water level response.   These effects were removed in two ways.   In the first method the fit of the observed frequency response shown in Figures 7 and 10 was assumed to be the transfer function between water level and atmospheric pressure.   This transfer function was then multiplied with the Fourier transform of the atmospheric pressure record, and the resultant frequency response was then inverted into the time domain and subtracted from the water level time series.   In the second method a single coefficient between the water level record and the atmospheric pressure record was found by linear regression of the two time series (the coefficient was 0.48 for LKT and 0.35 for TF); the atmospheric pressure record was then simply multiplied by this coefficient, and the resultant time series was then subtracted from the water level record.

At LKT the strain spectra for the time series obtained with regression and the frequency dependent transfer function are very similar in the frequency range of interest.   In the frequency band of 0.05 to 0.8 cycles/day, both strain spectra have noise levels as much as 7 dB lower than that obtained from the unfiltered water level record.   The regression filter is slightly better at lowering the noise level of the water level record, suggesting that the inferred frequency dependent transfer function overestimates the effects of water table drainage on water level response to atmospheric loading.

At TF there is substantial difference between the strain spectra determined by filtering with the regression coefficient and the frequency

dependent transfer function. The difference is
due to the considerable attenuation and phase
shift of the atmospheric pressure signal in the
water level record. Use of the frequency
dependent transfer function reduces noise levels
as much as 12 dB in the frequency band of 0.05 to
0.8 cycles/day. The filtered strain spectra do
not have a significantly lower noise level than
the unfiltered strain spectrum for frequencies
greater than 0.9 cycles/day.

   Both filtered records at LKT and the record
filtered with a frequency dependent transfer
function at TF yield strain spectra whose slopes
increase at a rate of roughly 20 dB-per-decade in
frequency in the frequency band of 0.08 to 2.5
cycles per day. This is comparable to the rate
of increase seen in this frequency band for the
strain spectrum reported for a dilatometer
[Johnston et al., 1986]. Presumably, one could
also remove the effects of atmospheric loading on
dilatometers [Suyehiro, 1982] and significantly
reduce the noise level of the instrumentation.
For instruments that measure horizontal strain
only, however, the reduction in noise level
achieved by removing atmospheric loading effects
may be slight [Bilham et al., 1974]. The
similarity between the noise levels in these
wells and those of dilatometers and wire strain
meters suggests that either the effects of
precipitation on the water level record are small
for frequencies greater than 0.08 cycles/day or
are similar to those of other strain meters.

   Following the approach of Agnew [1987], I
compare the strain response shown in Figure 11
with standard electronic distance measurement
response by transforming standard error estimates
for distance measurements into strain spectra.
The most accurate electronic distance measurement
currently available is obtained with a two-color
laser device [Linker et al., 1986]. The standard
error for a two-color distance measurement taken
over a baseline of 10 km is 1.2×10^{-7}. If
distance measurements 10 km in length were made
to determine areal strain with a two-color
geodimeter once a day, the resultant strain
spectrum would be (in the absence of a tectonic
signal) a flat -83 dB relative to 1 strain^2/Hz.
The raw and corrected strain spectra for both
wells are well below this noise level until
frequencies are less than 0.04 cycles/day. This
comparison indicates that these wells outperform
the ability of electronic distance measurements
to measure areal strain for deformation that
takes place over days to weeks.

Conclusions

   The results given here suggest that
theoretical models which describe the response of
water wells to atmospheric loading and areal
strain can be used, in conjunction with
cross-spectral estimation, to yield some valuable
information on the influence that vertical
groundwater flow to the water table has on strain

sensitivity. When the water table is weakly
isolated from atmospheric pressure changes, the
attenuation of the atmospheric loading response
of water wells is qualitatively similar to the
attenuation of the earth tide and tectonic strain
response. Well sensitivity in both cases
gradually attenuates with decreasing frequency
owing to water table drainage, and the frequency
at which significant attenuation begins to take
place is a strong function of well geometry and
rock material properties. For wells open to rock
in excess of 100 m below the water table,
attenuation of sensitivity will be slight at
periods of days to weeks if the rock above the
open interval has a vertical permeability of
1×10^{-16} m^2 or less and a compressibility of
1×10^{-10} Pa^{-1}.

   The wells examined here are most useful as
strain meters over a limited frequency band.
Both have a lower noise level than standard
distance measurement techniques for strain events
with periods of days to weeks; at longer periods,
electronic distance measurements can have
superior performance. At TF, increasing noise
with increasing period appears to be the result
of decreasing strain sensitivity in the presence
of a 15 to 20 dB-per-decade increase in the water
level power spectrum. At LKT, the increasing
noise appears to be largely due simply to a 20 to
25 dB-per-decade increase in the water level
spectrum.

   The noise levels of these wells in the
frequency band of 0.08 to 2.5 cycles/day compare
well with dilatometers and wire strain meters.
Laser strain meters offer comparatively better
performance. While water wells may be limited in
their ability to accurately detect teleseismic
strain, the relative quality of these wells as
strain meters in the frequency band analyzed here
suggests that wells can be of considerable value
in continuously monitoring intermediate period
strain. In areas where conventional means of
continuously monitoring strain may be difficult
to install or maintain, water wells may provide a
unique way of monitoring deformation. In areas
where strain monitoring networks exist, it is
possible that water wells can augment the ability
of the network to reliably detect aseismic
crustal deformation.

Appendix A: Solution to the Response of a Well
to Periodic Areal Strain

   The solution to the response of a well to
periodic areal strain is obtained by solving for
the pore pressure response to periodic areal
strain. I assume that the well is an accurate
gage of the average pore pressure of the
saturated rock with which it is in hydraulic
communication and also assume that the open
interval of the well is very small relative to
the change in pore pressure with depth. Pore
pressure response to a periodic areal strain,
$A \cos(\omega t)$, is governed by the equation:

$$D' \frac{\partial^2 p}{\partial z^2} - \frac{\partial p}{\partial t} - \rho g A_s' A\omega \sin(\omega t) \quad (A1)$$

Equation (A1) was obtained from (5) by substituting $A \cos(\omega t)$ for ϵ_a. This equation must be solved subject to the boundary conditions:

$$p(0,t) - 0 \quad (A2a)$$

$$p(\infty,t) - -\rho g A_s' A\cos(\omega t) \quad (A2b)$$

where $z=0$ is taken to be the water table. No initial condition is imposed because we seek the periodic steady state solution. This problem is easily solved by employing complex notation. The solution in terms of gain and phase is

$$\text{Gain} - |P/\rho g A_s' A| - \sqrt{(J^2 + K^2)} \quad (A3a)$$

$$\text{Phase} - \text{Tan}^{-1}(-K/J) \quad (A3b)$$

where J and K are

$$J = \exp(-\sqrt{Q'}) \cos(\sqrt{Q'}) - 1 \quad (A4a)$$

$$K = \exp(-\sqrt{Q'}) \sin(\sqrt{Q'}) \quad (A4b)$$

Since the well is assumed to be an accurate gauge of pore pressure in the frequency band of interest, water level changes are related to pore pressure changes by $w-p/\rho g$, and the solution in terms of water level change per change in strain can be obtained from (A3) by multiplying the gain by A_s'. The solution in the real domain is given in (10).

Appendix B: Solution to the Response of a Well to Periodic Atmospheric Loading

The solution for the response of a well to periodic atmospheric loading is obtained, as in Appendix A, by solving for the pore pressure response. In Appendix A it was assumed that the water table ($z=0$) was always at zero pressure; this is not the case for periodic atmospheric loading, and before pressure at depth can be known, the water table pressure must be determined. The water table response to periodic atmospheric loading is determined through the use of a diffusion equation for flow of air through unsaturated porous materials [Buckingham, 1904; Weeks, 1979]:

$$D_a \frac{\partial^2 p_a}{\partial z^2} - \frac{\partial p_a}{\partial t} \quad (B1)$$

subject to the following boundary conditions:

$$p_a(-T,t) - A \cos(\omega t) \quad (B2a)$$

$$p_a(T,t) - A \cos(\omega t) \quad (B2b)$$

where p_a is the air pressure. The boundary $-T$ is

taken to be the Earth's surface; the zone from depth 0 to depth T is simply an artifice to assure that at the water table, $z=0$, there is no air flux. As in Appendix A, we seek the periodic steady state solution. The air pressure at the water table ($z=0$) is readily solved by employing complex notation:

$$p_a - MA \cos(\omega t) + NA \sin(\omega t) \quad (B3)$$

where M and N are given in equation 13. The pore pressure at the water table is $-p_a$.

With the pressure at the water table known, the solution to pore pressure response to a periodic atmospheric pressure fluctuation, $A \cos(\omega t)$, is obtained from the following modified version of (7):

$$D \frac{\partial^2 p}{\partial z^2} - \frac{\partial p}{\partial t} - (1-E_B')A\omega \sin(\omega t) \quad (B4)$$

The appropriate boundary conditions are

$$p(0,t) = -MA \cos(\omega t) - NA \sin(\omega t) \quad (B5a)$$

$$p(\infty,t) - -(1-E_B')A \cos(\omega t) \quad (B5b)$$

The solution in terms of gain and phase is given by

$$\text{Gain} - |w\rho g/A| - \sqrt{(U^2 + V^2)} \quad (B6a)$$

$$\text{Phase} - \text{Tan}^{-1}(-V/U) \quad (B6b)$$

where U and V are

$$U = (-M+\gamma)\exp(-\sqrt{Q}) \cos(\sqrt{Q}) + N \exp(-\sqrt{Q}) \sin(\sqrt{Q}) - (\gamma-1) \quad (B7a)$$

$$V - (-M+\gamma)\exp(-\sqrt{Q}) \sin(\sqrt{Q}) - N \exp(-\sqrt{Q}) \cos(\sqrt{Q}) \quad (B7b)$$

The response in terms of water level within the well can be obtained (as in (9)) by adding $A \cos(\omega t)$ to the real part of the solution and dividing by ρg. The solution in the real domain in terms of water level is given in (12).

Appendix C: Method by Which Gaps Were Filled in the Water Level and Atmospheric Pressure Record

Gaps in the water level and atmospheric pressure record over the time period examined were filled by an iterative process. The gaps were originally filled by linear interpolation. A zero-phase high-pass filter [Otnes and Enochson, 1978] with a cutoff frequency of 10^{-8} Hz was then applied to the water level and atmospheric pressure data to remove any long-term trends. The autocovariance with a maximum length of 40 days was then calculated for each time series. Gaps were then filled with a symmetric linear filter of length 120 with weights determined from the structure of the

autocovariance and the distance between the interpolated point and the nearest data point. The spacing between each value used in the filter was eight data points (2 hours), and weights were calculated by solving the following system of linear equations:

$$
\begin{bmatrix}
\sigma^2 & C^1\!\uparrow^2 & .. & C^1\!\uparrow^{60} & 1 \\
C^2\!\uparrow^1 & \sigma^2 & .. & C^2\!\uparrow^{60} & 1 \\
.. & .. & .. & .. & 1 \\
C^{60}\!\uparrow^1 & .. & .. & \sigma^2 & 1 \\
1 & 1 & 1 & 1 & 0
\end{bmatrix}
\begin{bmatrix}
w_1 \\ w_2 \\ .. \\ w_{60} \\ \mu
\end{bmatrix}
=
\begin{bmatrix}
C^1\!\uparrow' \\ C^2\!\uparrow' \\ .. \\ C^{60}\!\uparrow' \\ 1
\end{bmatrix}
\qquad (C1)
$$

where σ^2 is the variance, μ is the mean, $Ci\uparrow j$ is the autocovariance between the ith data value and the jth data value used in the interpolation, w_i is the weight of the ith data point in the filter, and $Ci\uparrow'$ denotes the autocovariance for the length between the interpolated point and the ith nearest data value. This linear system of equations provides an estimate of the interpolated value for which the estimation variance is at a minimum [Journel and Huijbregts, 1978]. The estimation variance σ^2_e is defined as

$$
\sigma^2_e = \sigma^2 + \mu - \Sigma w_i Ci\uparrow' \qquad (C2)
$$

Once the gaps were interpolated using the system of equations given in (C1), the autocovariance was recalculated, and the system of equations was solved again. This process was repeated until convergence was achieved. The residual time series (the original time series with linearly interpolated gaps minus the high pass filtered time series with linearly interpolated gaps) was then added to the new time series to preserve the long-term trend.

Appendix D: Method by Which the Transfer Functions of Water Level to Earth Tides and Atmospheric Pressure Were Determined

The relation of water level to earth tides and atmospheric pressure at the wells in the frequency domain was obtained by solving the following complex system of equations for every frequency [e.g., Bendat and Piersol, 1986]:

$$
\begin{bmatrix}
BB & BT \\
TB & TT
\end{bmatrix}
\begin{bmatrix}
HB \\
HT
\end{bmatrix}
-
\begin{bmatrix}
BW \\
TW
\end{bmatrix}
\qquad (D1)
$$

where BB and TT denote the power spectra of the atmospheric pressure and earth tides, respectively; BT denotes the cross spectrum between atmospheric pressure and earth tides; TB denotes the complex conjugate of the cross spectrum between atmospheric pressure and earth tides; BW and TW denote the cross spectra between atmospheric pressure and water level and earth tides and water level, respectively; and HB and HT denote the transfer function between water level and atmospheric pressure and water level and earth tides, respectively.

The power spectra and cross spectra were obtained for the time series of interest by using the Blackman-Tukey procedure after removing the mean and long-term trend from the time series. This procedure is computationally inefficient because it requires that the autocorrelation or cross-correlation function be calculated in the real domain in order to obtain spectral quantities; it was used because the frequencies of interest were low relative to the length of the data set. The spectral estimates obtained from the Blackman-Tukey procedure were smoothed using a Hanning window.

At TF the long-term trend was removed with a zero-phase high-pass filter [Otnes and Enochson, 1978] with a cutoff frequency of 10^{-8} Hz. Spectral estimates were then obtained using a maximum correlation length of 20 days. At LKT the time series for water level, atmospheric pressure, and earth tides were lengthened to include the first half of 1986, and the data were decimated to two samples per hour; the time series was lengthened to see whether any attenuation of sensitivity could be unambiguously identified at low frequencies. The long-term trend was removed with a zero-phase high-pass filter with a cutoff frequency of 5×10^{-9} Hz. Spectral estimates were obtained using a maximum correlation length of 40 days.

The barometric and areal strain sensitivity as a function of frequency are simply the admittance of the transfer functions HB and HT, respectively. The multiple coherence squared, Γ^2, of the transfer functions as a function of frequency was obtained from the following spectral estimates [e.g., Otnes and Enochson, 1978]:

$$
\Gamma^2 = \frac{(WB \times TT - WT \times TB)BW + (WT \times BB - WB \times BT)TW}{(TT \times BB - TB \times BT)WW} \qquad (D2)
$$

where WB and WT denote the complex conjugates of BW and TW, respectively, and WW is the water level power spectrum.

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