Geodynamic Problems

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Introduction. This article is a transcription of a talk given at the end of the conference, with the intentions of summarizing some leading themes thereof and supplementing the report of the conference with my impressions of other geodynamic problems to whose solution geodesy could contribute. The emphasis of the article is on our understanding of the solid earth; suggestions of what measurements should be undertaken are based on estimates of instrumental feasibilities in Kaula et al. [1978], a committee report which was rather widely reviewed.

The organization of the conference report is by observational technique. A problem-oriented approach should take a different cut: the most obvious is a combination of spatial and temporal spectra, or length and time scales. Figure 1 is an attempt at such a cut; the discussion is succeeding sections is based thereon: (1) earth evolution & mantle convection; (2) lithosphere-asthenosphere-surface load interaction; (3) glacier-ocean-solid earth interaction; (4) solid earth interactions with the sun, moon, core, oceans, and atmosphere; (5) zones of strain accumulation; and (6) earthquakes.

1. Earth Evolution and Mantle Convection

The greatest events in earth history--in the sense of the amount of mass and energy involved--were formation (probably by planetesimal infall), possibly a major impact great enough to knock off the protolunar material, and separation of the core. From isotopic constraints and dynamical plausibility, all these events happened more than 4.4×10^9 years ago. Since then, the most important process has been solid state convection in the mantle. The major long term questions are: the energy sources for this convection and the geodynamo: the division between primordial and radiogenic, etc.; the manner of crustal separation and ocean & atmosphere outgassing; the degree of inhomogeneity, thermal as well as compositional, of the lower and upper mantle; the extent to which there has been material as well as heat transfer among parts of the mantle; and the nature of fluctuations about a general trend of declining activity. Although past history is significantly constrained by petrological and isotopic data as well as by thermomechanical reasoning, our ideas about mantle convection must be largely based on its present state and recent history: in particular, the last ∿200 My, for which there exist seafloor spreading and other evidences to infer the major surface motions. This restriction raises the question of whether current conditions are atypical. For example, present heat flow could be misleading if the mantle convective system can undergo fluctuations in heat transfer of a factor of two or more, as do some numerical experiments in convection. Certainly the present is atypical in the extent to which the land is broken up into several continents.

Major constraints on mantle convection are the properties of the lithosphere, which is part of the flow system. The plate velocities, oceanic topography, and plate margin locations--particularly the subducted slabs--are all significant boundary conditions. Because of the strong temperature dependence of the rheology, the lithosphere is a remarkably thick boundary layer, and screens much of the properties of mantle convection from observation. Geodetic data contribute to inferring these convective properties in two ways: (1) the variations in the gravity field indicate the amount of density inhomogeneity; and (2) the rates of uplift and sinking in response to the glacier-ocean mass transfer, discussed below, indicate the effective viscosity of the mantle. Both of these properties relate to stress, the distribution of which determines the flow.

The magnitudes of density anomalies $\Delta\rho$ inferable from gravity anomalies Δg depends on their characteristic vertical length scale L, beyond which they are in effect compensated. This length scale L_D is probably much less than the mantle thickness M. As is well known, the visible density anomalies constituted by the topography must be largely compensated in the uppermost 100 km. If this is true within the lithosphere, then it must be all the more so in the more plastic interior. Hence we can write, roughly,

$$\Delta g \propto 2\pi G \Delta\rho L_D \qquad (1)$$

Because the rest of the earth attracts the density anomaly stresses are set up, which will also be proportionate to the length scale L_D:

$$\sigma \propto g \Delta\rho L_D \propto g \Delta g / G 2\pi \qquad (2)$$

Viscosity is related to stress through the strain rate, which can be inferred from the plate velocities v:

$$\sigma \propto \nu \dot{\varepsilon} \propto \nu v / L_V \qquad (3)$$

If we assume the two length scales L_V and L_D to be the same and take Δg to be the typical value for $1^o \times 1^o$ square, say 25 mgal, use 10^{22} for ν and 5 cm/yr for v, then L is ∿300 km and σ is ∿60 bars.

The foregoing estimate is, at best, an order-of-magnitude, and is likely to be deceptive when applied to a regime which must be organized at some level, as is convection. But the order-of-magnitude of 100-500 km for the length scale is what arises from other estimates, such as those based on Rayleigh number considerations. Hence it is desirable that the gravity be known to a resolution not worse than 100 km to infer mantle

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convective patterns therefrom.

To infer whether there are density inhomogeneities at depth require consideration of, firstly, the spectral distribution of gravity and secondly the correlation of gravity with surface topography. The leading feature of the gravity spectrum is the "$10^{-5}/\ell^2$" law: a fairly sharp drop off in the rms magnitude of the normalized potential coefficients $\bar{C}_{\ell m}$, $\bar{S}_{\ell m}$ with spherical harmonic degree ℓ. The principal properties of the cross-correlation between gravity and topography are (1) negligible correlation for the long wavelengths $\ell \lesssim 5$ and (2) moderately positive correlation for the long wavelengths $\ell \gtrsim 6$ [Kaula, 1977]. Studies which assume that the density spectrum $\Delta\rho_\ell$ is "white": i.e., a comparable amount of variability in each degree conclude that the steep drop off in the gravity spectrum requires inhomogeneities at depth [Kaula, 1977; Lambeck, 1976]. The equally plausible assumption that the length scale L in eqs. 1-3 also has a spectrum L_ℓ such that L_ℓ varies inversely with ℓ (i.e., directly with horizontal wavelength) suggest that $\Delta\rho_\ell$ may vary directly with ℓ, because of the stresses entailed; requiring density anomalies at depth all the more. Analyses based on numerical models of convection get a flatter spectrum than $10^{-5}/\ell^2$, with even a reversal in slope at the longest wavelengths when the bottom boundary conditions are free [McKenzie, 1977].

For the present, it seems prudent to assume that there are density anomalies throughout the mantle comparable in magnitude to those near the surface. Hence any study of gravity should employ harmonic analysis, to separate shallow and deep effects.

2. Lithosphere-Asthenosphere-Surface Load Interaction

Moving to shorter wavelengths and more recent phenomena makes it fruitful to explain associated topographic and gravitational features by specific physical models, as discussed by Watts, Roufosse, and Turcotte in these proceedings. The common theme of all these studies is to treat the lithosphere as an elastic layer over a fluid half space. The leading property then becomes the flexural rigidity [Walcott, 1970]:

$$D = \frac{\mu(\mu+\lambda)T^3}{3(2\mu+\lambda)} \approx \frac{2\mu T^3}{9} \qquad (4)$$

in which λ,μ are the elastic moduli and T is the effective thickness of the lithosphere. In general, the flexural rigidity is less than what is calculated from seismic values for the elastic moduli λ,μ and the Rayleigh-wave lid thickness for T, as should be expected. Typical results are $\sim 10^{30}$ dyne-cm (compared to $\sim 10^{32}$ dyne-cm from seismic data), and an inverse correlation with both lithospheric age (since creation at the ocean rise) and duration of loading.

More dynamic problems utilizing gravity & topography are those associated with subduction zones, where the plate velocity and thus the asthenospheric viscosity may be of significance. Thus, for example, the topographic and gravitational high oceanward from the trench have been explained as due to the elastic bending of the lithosphere [Watts & Talwani, 1974]. Probably the most elaborate models leading to prediction not only of gravity anomalies and topography but also to rate-of-topographic rise are the thermomechanical finite element calculations of Himalaya and Zagros underthrusting by Bird [Bird et al., 1975; Bird, 1978].

A major plate tectonic problem to which it is hoped future geodetic data will apply is the variation of plate velocities about their values inferred from remanent magnetic striations. The plate velocities appear to be quite steady for periods of about 10^7 years, after which there is an adjustment of the velocity pattern, apparently due to change in physical circumstances at plate margins [Larson & Pitman, 1972]. This dependence on margin conditions is consistent with plate tectonic models which attempt to account for the velocity patterns as arising from plausible combinations of forces on the margins and on the lithosphere-asthenosphere interface, dependent on the extent to which the plate is oceanic or continental. These studies conclude that the plates move mainly as a consequence of push at the rise and pull at the trenches; the asthenosphere acts as a minor drag on the lithospheric plates [Forsyth & Uyeda, 1975; Richardson et al., 1976]. Since single events, like the 1960 Chile earthquake, may account for several decades' motion over a major segment of plate margin, it is plausible that some of this jerkiness is transmitted to the overall plate motion. This post-seismic stress propagation may be geodetically detectable over 1000's of kilometers.

Geodetic techniques may also detect variations about steady rates in vertical motion arising from tectonic causes both close to and remote from plate margins. Estimates of vertical motion rates from either geologic observations or thermomechanical modeling suggest that vertical motion rates greater than $\sim$1 mm/yr should be exceptional. While the inherent sluggishness of the solid earth makes faster rates implausible and, at present, inexplicable, our understanding is imperfect enough that such rates cannot be ruled out a priori. It is therefore important to determine relative uplift rates more reliably between sites for which the geologic setting is well known.

3. Glacier-Ocean-Solid Earth Interaction

The dominant effect on the rate of uplift or sinking, and on the height relative to sea level, of the solid earth in the time scale of $\sim$300 to 30,000 years is the 3×10^{22} gram mass transfer from continental glaciers to the ocean surface which mostly took place 18,000 to 8,000 years ago. Because the distribution of the glacial unloading is relatively well known [Paterson, 1972; Peltier & Andrews, 1976] and of the ocean loading very precisely known, measurements of the response of the solid earth are the most significant data on its rheology.

The principal data type pertaining to this regime is carbon-14 dating of ancient shorelines, which gives points in the historical record over the last few 1000 years. The principal interpretations are that the characteristic decay

time $\tau(\lambda)$ varies directly with wavelength λ for $\lambda \gtrsim 1000$ km and inversely for $\lambda \lesssim 1000$ km. The former appears consistent with a mantle of 10^{22} poise; the latter, with an elastic lithosphere $\sim$100 km thick. Current vertical rates are 10 mm/yr near the center of the formerly glaciated area, but $\sim$1 mm/yr away from it.

Since most of the adjustment to the glacier-ocean load transfer appears to be completed, geodetic leveling measurements are to some extent supplementary. However, they are a valuable supplement, because the C^{14} shoreline data give rather incomplete coverage of the peripheral bulge area which is critical to inferring elastic and non-linear rheological effects.

The post-glacial response is one of the two or three principal evidences that the mantle indeed flows. However, there are some cautions as to the application of the inferred viscosity to the longer term convective problem [Kaula, 1979]. Foremost among these is the possibility that the post-glacial effect is transient. While it may seem strange to call an effect on a $\sim 10^3$ year timescale 'transient', it is true that steady state rheology has been attained in the laboratory only at strains appreciably greater than the 10^{-4} to 10^{-3} characteristic of post-glacial load transfer [Goetze & Brace, 1972; Weertman, 1978]. The present level-of-detail of the data warrants at most linear viscoelastic models [Clark et al., 1978]. To go beyond these models and infer non-linear viscosity or transient effects may require an amount of detail which only geodetic data can supply.

4. Solid Earth Interations with the Sun, Moon, Core, Oceans, and Atmosphere

Although tidal friction can be said to apply to the entire history of the earth, most of these exogenic effects lie well within a $\lesssim$300 year timescale in their characteristic periods. The common thread of the exogenic disturbances is that they are all quite small in their effects at the earth's surface, and hence are observable geodetically only if they affect things which can be measured very precisely, such as pole position or tide height.

Because rotation and tides are precisely measurable and mathematically modelable, they have an intellectual appeal which may result in their receiving more attention that may be warranted by the criterion of illuminating causes. As discussed by Lambeck in this volume, matters which are still very much in doubt are: the relative magnitudes of atmospheric, earthquake, and aseismic creep effects on the free polar wobble; the amount of non-tidal rotational acceleration; the manner of dissipation of both polar wobble and tides; and the nature of core-mantle coupling. While for all these problems an increase in accuracy of measurement of their rotational & tidal effects should be of some help, it also can be said for all of them that enhanced insight will require either or both of appreciable improvement in non-geodetic modeling or significant additional non-geodetic observations.

Thus to infer whether the atmosphere makes a major contribution to the excitation of the long term variations in LOD and free polar wobble already observable requires a better understanding of both seasonal and decade-scale changes: such questions as the year-to-year variations in atmospheric mass distribution, water and energy transfers between the ocean & atmosphere, and the influence of the solid earth on the atmosphere, most likely through volcanism [Lambeck & Cazenave, 1976, 1977; Wilson & Haubrich, 1976]. All of these phenomena have broad spectra, and hence require observations capable of measuring changes on a much finer scale than global. Furthermore, some of the most feasible observations--such as satellite photography of cloud motions--require significant supplemental observations and modeling to be applied. The recent revisions of the magnitude-moment relationships for great earthquakes [Kanamori, 1977] revive seismicity as a possible source of wobble excitation, but there has not yet been a convincing modeling of the effect of any one earthquake on the pole path.

The non-dissipatory part of the response of the earth to tidal and rotational effects depends on bulk properties, and hence is more amenable to mathematical attack. Significant advances have been made in modeling the effects of the ocean [Dahlen, 1976] and the core [Smith, 1977]. On the other hand, the dissipatory part of the response is probably associated more with the liquid-solid interfaces: ocean-crust and core-mantle. Consequently, the dissipation is more difficult to model. The mechanisms of both the polar wobble damping and tidal friction are still not understood. However, some constraints can be placed on both these processes, even if details of how the dissipation occurs are unknown. Thus the tidal friction at present must be anomalously high, because the present 1/Q extrapolated into the past brings the moon close to the earth much too recently. But this high tidal dissipation is not surprising: by paleontological and other indications, the present continental configuration is remarkably broken up, so that a greater-than-usual portion of tidal energy is transferred from the second to the higher harmonics of the tide, eventually to be dissipated, willy-nilly. An interesting theme to pursue might be the extent to which the need to maintain a small tidal 1/Q constrains the tectonic style in the past. Some geological discussions suggest that there is a real dearth of subduction zone associations from Proterozoic, 0.6 to 2.5×10^9 years ago.

In summary, measurements of pole position and LOD appear to be in the Hubble situation: "We do not know <u>what</u> the future will want, but we do know they will want it accurate". To which we can add "continuous": as new techniques are phased in, there should be some years' overlap with the older measurements.

5. Zones of Strain Accumulation

The somewhat arbitrary distinction from topic #2, lithospheric-asthenospheric interaction, is that we are concerned here about endogenic motions in the solid earth at rates sufficient both to be measured and to test tectonophysical models. These criteria narrow attention to plate margins

(using a broad definition of the term) and very recent times. The equally arbitrary distinction from topic #6, earthquakes, is that we are concerned here about the entire process of strain change in a region, rather than just the phenomena associated with particular abrupt releases of strain. The possibility of gradual, non-earthquake release of strain makes the title 'strain accumulation' somewhat incomplete in its implications.

As discussed by Savage in this volume, the rate-of-strain as measured near active plate margins in California, Japan, and New Zealand appears to be on the order of 0.3×10^{-6}/yr. This strain accumulation is distributed across a zone ∿200 km wide. Such a magnitude is roughly consistent with the prediction from plate tectonics,

$$\dot{\varepsilon} \approx \frac{v}{W} \qquad (5)$$

where $\dot{\varepsilon}$ is strain rate, v is the relative velocity of the plates and W is the width of the strain accumulation zone. However, this generalization is based on relatively few surveys of differing temporal intervals in regions of differing tectonic context and seismicity. Factors aside from relative plate velocities which may affect the pattern of rate-of-strain include: (1) the regional fault configuration; (2) the depth to an 'asthenosphere' or other place which can undergo plastic deformation; and (3) the recent seismic history of the region. Thus, for example, the San Andreas fault appears to have at least three distinctly different regional behaviors. In northern California, the strain evolution appears to be still strongly influenced by the 1906 earthquake: since then, the seismicity has been much lower than in the previous ∿50 years, and the strain accumulation appears to be much more about the Hayward & Calaveras faults [Thatcher, 1975ab]. In the Hollister region, there are sporadic motions at ∿ monthly intervals associated with very small earthquakes, suggestive of slip on a shallow plastic zone [Huggett et al., 1977, Johnston et al., 1977]. In southern California, where the San Andreas bends appreciably and there are subsidiary faults, in six years of precise measurements there has been a steady north-south contraction of 0.3×10^{-6}/year [Savage et al., 1978].

The San Andreas, being a transcurrent fault, is relatively simple among zones of strain accumulation. However, it shows significant variations on a length scale of ∿300 km and on a temporal scale of ∿100 years for major events, as indicated by the sedimentary record in the south [Sieh, 1978] as well as by the historical record in the north [Thatcher, 1975ab]. A consequence of these irregularities is that the strain accumulation pattern in a region may be quite different from that suggested by simple plate tectonic considerations, as is the case now in southern California. Hence, while more geodetic measurements are manifestly desirable, it can be expected that a purely empirical approach will not solve the problem. There must be a development of models as well as elaboration of the data. Other improvements which should be forthcoming are the combination of horizontal and vertical data in analysis and better integration of geodetic measurements in the geological context.

6. Earthquakes

As emphasized in the preceding section, earthquakes are not isolated phenomena. However, they do have a distinctive character, and there is nothing like a ≥7.5 magnitude earthquake to test a model of a strain accumulation zone. From a purely scientific point-of-view, it is a pity they do not happen more often. We are concerned here, firstly, about when, where, & why earthquakes occur in a strain accumulation region and, secondly, for a seismic event the pattern of precursory, coseismic, and subsequent activity and its causes.

It has been evident for sometime that earthquakes are most likely to occur at gaps in strain accumulation zones: places where earthquakes have not recently occurred and hence where presumably the greatest stresses have accumulated [e.g., Kelleher et al., 1973]. However, the stress redistributed by earthquakes in adjacent regions can vary significantly; some stress can apparently be released aseismically, or, in complex fault zones, by motion on adjacent faults; and tectonic configurations vary appreciably in their ability to withstand stress. Hence in an earthquake-prone region the recurrence interval can vary over almost an order-of-magnitude. For example, at Pallett Creek, in the south-central San Andreas fault, the interval has varied from 50 to 300 years in the last 1400 years [Sieh, 1978].

So far, geodetic measurements have contributed only slightly to the depiction of motions precursory to a sizeable earthquake. The 1964 Niigata earthquake (7.5) was preceded by rises on the order of 10 cm during the previous 60 years, with an acceleration in the last 5 years [Kisslinger, 1974]. The 1971 San Fernando shock (6.4) was located about 30 km from the center of an uplift of 20 cm over 10 years, with some migration of this center toward the shock a couple of years before the event [Castle et al., 1975]. The 1973 Point Mugu quake (6.0) was preceded by uplifting and then downwarping of about 4 cm during the previous 13 years [Castle et al., 1977]. However, these changes are not remarkably bigger than those inferred from leveling data in relatively non-seismic regions [Brown, 1976], and more recent experience with the Palmdale bulge has indicated that leveling must be treated with care and used in combination of other data [Kumar & Strange, 1978]. This data should include both gravity measurements and horizontal geodetic survey. However, the former can be appreciably affected by non-tectonic processes such as ground water movement [Lambert & Beaumont, 1977], while the latter have so far been only sketchily related to a shock [Thatcher, 1974].

A much better geodetic description of strain accumulation prior to a major earthquake can be expected in the next decade or two. The San Andreas has been anomalously quiet since the intersification of instrumentation programs. More

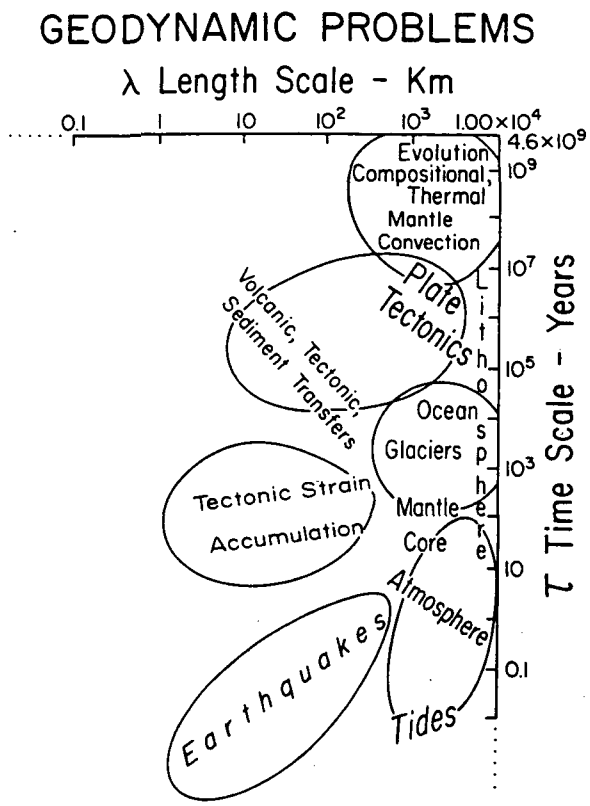

Fig. 1. Length scale versus time scale breakdown of geodynamic problems.

careful attention to leveling observation, computation, and adjustment should also payoff. Point measurements--strain, tilt, and gravity--appear to have greater inherent difficulties of sorting out regional from local effects [Johnston et al., 1977; McHugh & Johnston, 1977; Lambert & Beaumont, 1977]. For a while at least, seismicity itself will remain the leading precursory data type: see, e.g., Shimazaki [1978] and Ishida & Kanamori [1978] for studies on two different scales.

Postseismic geodetic measurements [e.g., Brown et al., 1977] are important because they indicate some aspects of the stress redistribution precursory to subsequent creep and seismicity in the zone, and because they are affected by the same tectonic context as the earthquake itself. In fact, so far earthquake models have been more successful in explaining post-seismic adjustment than the precursory buildup and the earthquake itself [e.g., Nur & Mavko, 1974; Anderson, 1975; Rundle & Jackson, 1977; Savage & Prescott, 1978; Walcott, 1978]. The situation appears to be analogous to that discussed in connection with rotation: phenomena dependent on bulk properties, such as viscoelastic post-seismic adjustment, are easier to model than those dependent on interface properties, such as earthquake occurrence. Geodetic measurements also constrain the bulk response more than the interface effects, for which more stress- and energy-sensitive indicators (primarily seismic data, but also heat flow, radon, etc.) will continue to be needed, as well as experimental rheology [Dieterich, 1978].

Conclusions

It is many years since Love [1911] wrote his book of similar title, but progress seems to be accelerating in recent years in both measurement and modeling of geodynamic phenomena. Rather than attempt to recapitulate the main themes, I close with a personal wish list, in priority:

1. Intensify geodetic measurements in strain accumulation zones within ∿100 km of the best-studied faults. The data, as well as modeling considerations, indicate that this is where most of the action occurs, and that it is quite observable with current techniques at major plate boundaries [Savage & Prescott, 1978; Walcott, 1978]. This is not to say that a comprehensive insight does not require widespread measurements entailing space techniques; the globe is one big coupled system [Anderson, 1975]. Rather, given limited resources, the most likely payoff, scientifically as well as practically, is close to the big faults: scientifically, because meaningful models must deal with stress, which in turn entails strain & strain rate, taking detailed measurements to describe. The rocks know each other mainly through their nearest neighbors.

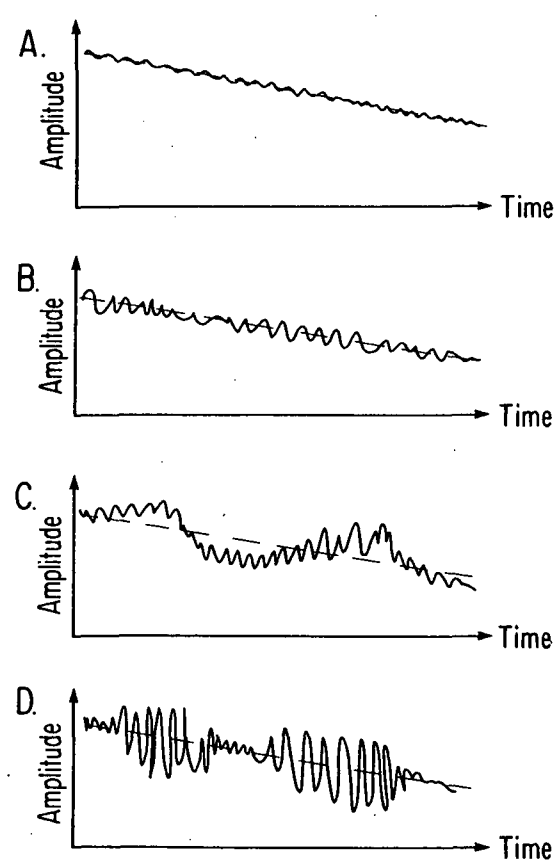

Fig. 2. Schematic representation of the possible forms of short term oscillations about long term geodynamic trends.

2. Develop better devices for geodetic measurements over distances less than ∿30 km. 'Better' means not only more accurate, but more rapid & responsive. Some that look promising are:

a. Three-wavelength ranging system good to within 5×10^{-8}, such as that described by Levine in this volume;

b. mini-interferometers, such as those described by Shapiro in this volume;

c. three-wavelength angle measuring devices, to obtain level differences more rapidly.

Since important tectonic phenomena are inherently inaccessible to direct observation and must be inferred from surface measurements, there is no distinct point of diminishing returns in the accuracy of measurements, while we are still far from saturation in spacing and frequency.

3. Tidy-up the vertical networks. There have been too many suspicious leveling results, such as the now-here, now-gone discrepancy from tidal bench marks between San Diego and San Francisco [Douglas, 1978]. Leveling has the inherent accuracy to tell us a lot which should be realized. While it is indeed possible that short term variations may have appreciably higher rates than the long term trends (see Figure 2), there are limits to what is physically plausible, particularly over distances more than some 10's of kilometers.

4. Measure gravity globally to 100 km resolution. As discussed in section 2, such detail is needed to see through the lithosphere to infer the underlying mantle convective action. The SEASAT altimeter will obtain it for most of the oceans, but satellite-to-satellite tracking systems, such as described by Fischell in this volume, appear to be the way over the land.

5. Develop portable 3 μgal (30nm/sec) gravity meters. Despite the difficulties with local effects, a comparably accurate gravity meter seems to be a necessary concommittant to leveling.

The earth is rather big and has been around a long time. While geodesy has gained a global range spatially, it is still limited temporally to the last finger-snap of geologic time. However, there are important, as well as entertaining, processes on the range of time scales within geodetic reach. The task of geodesy might be said to measure the short-term fuzz on the long-term trends: for each phenomenon, to determine which sort of pattern--A, B, C, or D in Figure 2--prevails.

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