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Contemporary Plate Motion and Crustal Deformation

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The measurement of relative plate velocities during the past few years is a signal accomplishment in earth science, leading to refinement of the precepts of steady motion of plate interiors and cyclic deformation along plate margins. Regrettably, deformation premonitory to an earthquake has yet to be detected with confidence. Delineation of the spatial and temporal buildup of strain between earthquakes, however, has put limits on models of the earthquake cycle. A diverse set of fault structures and crustal rheologic conditions can explain the pattern of surface strain accumulation and release along strike-slip faults. In contrast, the geometry of thrusts and normal faults, revealed by earthquake deformation, has been found to differ markedly from expectations. Geodetic observations have proved vital to monitor the ascent of magma through the Earth's crust and to predict volcanic eruptions at the Earth's surface. Episodic vertical and horizontal deformation in southern California remains a subject of dispute; if anything, it is less certain than it once seemed.

Plate Motions Inferred From Terrestrial And Space Geodesy

Perhaps the most important advance in geodesy during the past half century was achieved during the past 4 years: direct measurement of global plate motion. Ironically, this is not the first time such a claim has been made. "We begin the demonstration of our theory with the detection of present-day drift of the continents by repeated astronomical positioning," wrote *Alfred Wegener* in the 1929 edition of *The Origin of Continents and Oceans* (p. 23), "because only recently this method furnished the first real proof of the present-day displacement of Greenland—predicted by drift theory—and because it also constitutes a good quantitative corroboration." From repeated longitude measurements Wegener found that Greenland had drifted 1,610±285 m from Europe between 1823 and 1907, a rate of 19±3 m/yr. This result agreed with longitude observations performed elsewhere on Greenland between 1860 and 1921. Wegener carefully considered and ultimately dismissed the possibility that gradual improvement in measurement precision, rather than drift of Greenland, accounted for the longitude changes: "This accumulation of similar results which do not stand in opposition to any others makes it highly improbable that it is all just a matter of unfortunate combination of extreme errors of observation." Today Wegener stands at once vindicated and discredited: Greenland is drifting away from Europe, but at a velocity now measured to be 19±10 mm/yr, one-thousandth Wegener's proffered rate. We also stand forewarned that our systematic errors may, too, turn out to be improbably large.

Relative Plate Motion

The relative motions of as many as five plates have been observed by four independent techniques during the past decade. The methods include terrestrial laser ranging using the Geodolite, satellite laser ranging (SLR), satellite radio positioning with the Global Positioning System (GPS), and Very Long Baseline Interferometry (VLBI) on extragalactic radio sources. Agreement among these measurement systems is generally good and steadily improving [for Geodolite versus GPS, see *Prescott and Svarc,* 1986; for GPS versus VLBI, see *Kroger and Davidson,* 1986; for VLBI versus SLR, see *Kolenkiewicz et al.,* 1985, and *Lyzenga et al.,* 1986]. Even more encouraging, however, is that the observed plate velocities accord well with averaged velocities over the past 1–2 m.y. inferred from magnetic anomaly profiles, earthquake slip vectors, and fault azimuths by *Minster and Jordan* [1978]. Thus, at a precision of about 15 mm/yr, plate motion measured sufficiently far from plate boundaries appears steady over periods of 1–10^6 yr.

Both the steady velocities of plates measured since 1981 and the agreement between current and geologic plate rates argue that jerky plate motion is confined to plate boundaries; no internal deformation of plate interiors has been reliably measured. *Tapley et al.* [1985] found that changes in four intercontinental baselines measured by SLR from Australia to the North American and Pacific plates coincided within 3 mm/yr of the average rates and over the past 2 m.y. *Christodoulidis et al.* [1985] reported that the relative motion of six pairs of plates measured since 1979 by SLR have a correlation of 0.61 with respect to the velocities predicted by *Minster and Jordan* [1978]. Recent SLR results were reviewed by *Cohen and Smith* [1985]. *Herring et al.* [1986] found that the North American and European plates have been separating at 19±10 mm/yr as measured by VLBI since 1981, a rate indistinguishable from the 17±3 mm/yr averaged over the past 1 m.y., but also not significantly different from zero. The more rapid motion of the Pacific plate relative to North America and Eurasia was also measured by VLBI; those rates of 50–100 mm/yr agree with *Minster and Jordan's* [1978] values with a mean discrepancy of 17 mm/yr [*Carter and Robertson,* 1986].

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Pacific–North American Plate Boundary

The geologic and geodetic displacement along the San Andreas fault zone has been subjected to closest scrutiny. During the past 2 m.y., 56±3 mm/yr of transform motion has been accommodated along the San Andreas and other right-lateral strike slip and oblique faults that span California from the coastline to the Nevada border, although work in progress [*Demets et al.*, 1986] suggests that the plate rate may be less than 50 mm/yr. *Sieh and Jahns* [1984] found that the central San Andreas fault has slipped 30–40 mm/yr during the past 15,000 yr, a finding in accord with the 33±1 mm/yr rate *Thatcher* [1979] resolved from the past century of triangulation. *Lyzenga and Golombek* [1986] analyzed VLBI measurements of a 150 × 300-km triangle spanning the southern San Andreas fault during the period 1980–84 and found a displacement rate of 24±4 mm/yr. Thus about half of the plate-boundary motion is accommodated by the San Andreas fault. *Minster and Jordan* [1987] use these data in concert with other continental VLBI results to explain the discrepancy in velocity and direction between the plate motions and the San Andreas slip rate. They find that 13±5 mm/yr of right-lateral slip must occur largely west of the California coastline along such offshore faults as the Hosgri. In addition, they resolve 9±3 mm/yr of compression perpendicular to the San Andreas fault, which gives rise to thrust faults and folds in the central California Coast Ranges.

A single 880-km baseline coincident with the California VLBI measurements has been observed by satellite laser ranging since 1972 [referred to as the San Andreas Fault Experiment, or SAFE]. These SLR observations have undergone considerable instrument and orbit development, during which the calculated motion decreased dramatically from original values that were reported to be twice the geologic plate rate [*Smith*, 1980]. *Christodoulidis et al.* [1985] calculated a rate of 61±25 mm/yr for 1981–83, but data taken since 1983 now suggest a rate of about 20±5 mm/yr [D.E. Smith, pers. comm.], commensurate with the VLBI result.

Earthquake Deformation

The discipline of precise measurement of strain accumulation and release was born in this country on April 18, 1906. In his brilliant analysis of the movement of triangulation stations along the San Andreas fault that preceded and accompanied the great San Francisco earthquake, *Harry Fielding Reid* [1910] offered a geodetic strategy for earthquake prediction. "To measure the growth of strains, we should build a line of piers, say a kilometer apart, at right angles to the fault," wrote Reid in 1910 (p. 31). "A careful determination from time to time, of the directions of the lines joining successive piers, their differences of level, and the exact distance between them, would reveal any strains which might be developing along the region the line of piers crosses." Reid reasoned that when the shear strain accumulated to an amount equal to what was released during the previous earthquake release, a strong shock would follow. "Measures of the class described would be extremely useful, not only for the purpose of prediction, but also to reveal the nature of the earth-movements taking place, and thus lead to a better understanding of the causes of earthquakes." Thousands of reference points in the western U.S. are now surveyed annually to detect the buildup of crustal strain, and much has been learned, as Reid forecast, from these measurements about the cycle of deformation of which an earthquake is a part. These insights have led to a new picture of fault geometry and slip patterns at depth and to useful estimates of earthquake repeat times. Unambiguous detection of premonitory deformation nevertheless continues to elude us. We have succeeded only in establishing that accelerated fault slip before earthquakes is either uncommon, restricted to isolated patches, or has a magnitude less than a few percent of the eventual earthquake displacement.

Seismic Deformation Cycle

If all strain energy that accumulated between earthquakes were released during shocks of equal magnitude, no permanent deformation would result, and if the rate of accumulation were constant throughout the buildup, knowledge of only the coseismic strain drop and its interseismic rate measured at any point in the cycle would be sufficient to forecast earthquakes. *Savage* [1983*a*] employed this rationale to estimate earthquake repeat times along the southern San Andreas fault from a decade of Geodolite observations. The measured shear strain there was 0.3 ppm/yr [*Savage*, 1983*a*; *Snay et al.*, 1983]. Assuming a typical strain drop of 50 ppm for the next earthquake on the southern San Andreas fault, Savage obtained a repeat time of 160 yrs, in accord with *Sieh's* [1984] and *Weldon and Sieh's* [1985] measured prehistoric repeat times of 150–200 yr. *Thatcher* [1984*a*] found the 1906 earthquake strain drop on the northern San Andreas to have been about 130 ppm, however, which would lead to a repeat time of 430 years. *Thatcher* [1983, 1984*a*] used triangulation surveys along the San Andreas fault to demonstrate that large earthquakes are followed by rapid postseismic deformation, after which strain diffuses outward from the fault at a slower and nearly steady rate. This pattern can be explained by a partially or fully faulted lithosphere coupled to a viscoelastic asthenosphere or intracrustal layer. Immediately after the earthquake the deepest portion of the fault creeps, giving rise to the postseismic transient. Between earthquakes, the asthenosphere flows to relieve traction imposed at the base of the lithosphere by the coseismic slip. These interseismic processes accelerate the accumulation of strain near the fault shortly after the earthquake; as a result, the strain buildup is no longer constant but varies as a function of time and proximity to the fault. Repeat times are overestimated if these processes are neglected, with the effect more pronounced for underthrust earthquakes.

Because dip-slip earthquakes drop and lift parts of the crust, gravitational forces are exerted and the seismic and interseismic displacements are not equal. This imbalance results in the growth of geologic structures, such as basins and ranges. At convergent plate margins, where great earthquakes typically repeat at 100-yr periods, the seismic and interseismic deformation have been measured for a complete cycle landward of the trench. *Thatcher* [1984*a*, 1984*b*] and *Thatcher and Fujita* [1984] used leveling, tide gauge records, triangulation, and chained distance measurements in Japan to study the Nankai and Sagami Trough subduction boundary. There the coastline slowly drops down between earthquakes, but it rises an amount greater than the interseismic subsidence during each major shock [*Kato,* 1983*a*, 1983*b*], resulting in marine terraces formed by upraised former shorelines [*Lajoie,* 1986]. Models that incorporate the interaction of elastic and viscous layers and the effect of buoyant restoring forces upon the earthquake cycle have been advanced by *Li and Rice* [1983*a*, 1983*b*], *Savage* [1983*b*], *Thatcher and Rundle* [1984], and *Savage and Gu* [1985*a*]. *Reilinger and Kadinsky-Cade* [1985] and *Reilinger* [1986] have shown that these models can also explain the postseismic deformation observed for large dip-slip earthquakes in plate interiors.

Some seismic gaps—segments of plate boundaries that have not suffered a large historical or recent earthquake–are currently accumulating strain at rates consistent with the approach of a large shock. Of two closely monitored sites where the Pacific plate subducts beneath Alaska, the Yakataga gap is accumulating strain at a rapid rate (0.3 ppm/yr; *Savage and Lisowski* [1986*a*]), whereas the gap centered on the Shumagin Islands is not accumulating measurable strain [*Savage and Lisowski,* 1986*b*]. This leaves an impending Shumagin gap-filling earthquake in doubt despite recent short-term fluctuations of tilt measured by releveling and a tidal-gauge network [*Beavan et al.,* 1983, 1984]. Subduction along the Shumagin gap may be aseismic [*Savage et al.,* 1986*a*], or the rate of strain accumulation may have slowed late in the earthquake cycle. A large gap-filling earthquake apparently struck the Shumagin Islands in 1788, an occurrence which is hard to reconcile with aseismic subduction. The eastern margin of the Basin and Range province of the western U.S. produced a series of large prehistoric earthquakes along the Wasatch fault in Utah [*Schwartz and Coppersmith,* 1984]. Here again, the measured rate of extension across the fault zone (0.02–0.05 ppm/yr) [*Snay et al.,* 1984; *Savage et al.,* 1985], is too small to be confidently resolved above noise.

Fault Geometry

Interseismic and coseismic strain measurements have afforded new insights into the mechanics of transform faults. *Thatcher* [1975] showed that slip extended no deeper than 15 km and possibly only to 5 km during the 1906 earthquake; today, seismicity extends no deeper than 12–15 km along most of the San Andreas fault. Transform faults have been successfully modeled as vertical dislocations either partially or fully locked between earthquakes from the surface to a transition depth and freely slipping at the long-term slip rate below that depth. Competing models with distributed deformation below a rheological transition [*Thatcher,* 1983] are also compatible with the observations, but the specifics of even the simplest single-fault geometry have proved difficult to delimit. *Prescott and Yu* [1986] demonstrate that shear strain distributed across the San Andreas and associated faults in the northern San Francisco Bay region can be modeled with transition depths from 6 to 30 km. *Segall and Harris* [1986] found that transition depths of 14–22 km fit the dense network of observations along the Parkfield segment of the San Andreas. *King and Savage* [1983; 1984] favored locking depths of 15 km on the southern San Andreas and San Jacinto faults, but they emphasized that extreme values of 5 and 20 km cannot be excluded. *Matsu'ura et al.* [1986] find the transition to occur at a depth of 10±5 km along the central segment of the San Andreas fault, south of San Francisco Bay. Because a given profile of surface strain across a fault can be fit by varying the slip rate with the transition depth, neither parameter can be determined independently. This is a practical limitation of geodetic data for determination of either the long-term fault slip rate or the depth of the locked zone of a fault. In principal, it can be overcome by measuring the strain profile normal to the fault over an aperture many times wider than the fault depth. In practice, however, such surveys are limited by the competing strain fields of adjacent faults and by the precision of the measurements.

The deformation associated with dip-slip earthquakes has revealed important surprises about the subsurface geometry of thrusts and normal faults. The 1983 $M =$ 6.7 Coalinga earthquake struck on a thrust fault with no surface trace; in fact, analysis of repeated leveling surveys show that the fault slip did not penetrate the uppermost 5 km of the crust [*Stein and King,* 1984]. The larger 1977 $M = 7.4$ Caucete, Argentina, earthquake took place on a fault that did not slip within the upper 15 km of the crust [*Kadinsky-Cade et al.,* 1985]. In the past, thrust faults that do not offset surface deposits were not classed as active. At Coalinga the fault is masked by a young surface fold in layered sedimentary rocks. The amplitude of this fold grew 0.75 m during the earthquake, revealing that at least some folds grow in jumps during earthquakes, rather than through progressive ductile deformation as is generally believed. Thus identification of past and future earthquake sources in regions of crustal contraction must include a search for active folds.

Investigation of the geodetic record for the three largest historical normal-faulting earthquakes in the Basin and Range province, the 1954 $M = 7.2$ Fairview Peak, Nevada, 1959 $M = 7.3$ Hebgen Lake, Montana, and 1983 $M = 7.0$ Borah Peak, Idaho, shock indicates that these earthquakes struck on planar faults dipping 45°–60° and extending to depths of 10–20 km [*Snay et al.,* 1985; *Stein and Barrientos,* 1985; *Ward and Barrientos,* 1986; *Barrientos et al.*, 1987]. This result contrasts with previous interpretations that the Basin and Range is extending by slip on shallow, gently-dipping faults that are reactivated thrust faults from a former episode of contraction.

Precursory Deformation

Near-field measurement of preseismic deformation shortly before an earthquake is a nearly impossible task; too many candidate faults exist for them to be frequently monitored. The 1984 $M = 6.2$ Morgan Hill, California, earthquake [*Bakun et al.*, 1984], however, struck within a Geodolite ground-based laser ranging net, with a precision of 0.2 ppm. *Prescott et al.* [1984] measured lines within 5 km of the epicenter 2 weeks, 8 days, and 1 day before the main shock. No preseismic deformation was detected, indicating that no preseismic slip greater than 100 mm occurred on a fault that sustained 425 mm of coseismic slip. Near-field recordings of dilatational strain (resolution of 10^{-4} ppm) and shear strain (resolution of 10^{-3} ppm) measured in borehole instruments by *Johnston et al.* [1987] 30–50 km from the $M = 6.7$ 1983 Coalinga, $M = 6.0$ 1984 Kettleman Hills, $M = 6.2$ 1984 Morgan Hill, 1984 $M = 5.8$ Round Valley earthquakes in California and the 1978 $M = 7.0$ Izu earthquakes in Japan fail to reveal earthquake precursors. Any precursory slip within several hours to days must have been less than a few percent of the coseismic slip during these moderate to large earthquakes to explain the absence of preseismic strain changes. An alternate explanation is that precursory fault slip is restricted to very strong or very weak patches of the fault, an idea supported by seismic evidence that ruptures start and stop at bends or offsets of faults [*King and Nabelek*, 1985].

Permissive evidence for deformation or accelerated fault creep before an earthquake is widespread but equivocal: During the 6 months preceding the Coalinga earthquake, for example, wire creepmeters on the adjacent Parkfield segment of the San Andreas fault 40 km from Coalinga measured creep that was faster than their long-term averages [*Mavko et al.*, 1985]. Before the most recent (1966) earthquake on this segment of the San Andreas, new cracks were noted along the fault and a pipe crossing the fault apparently ruptured. Seasonal expansion and contraction of the surface materials could, however, plausibly explain both the 1966 and 1983 changes. *Segall and Harris* [1986] inverted the Geodolite line-length changes along this segment and showed that the deficit in slip since 1966 is now equal to the coseismic slip measured in 1966. This geodetic evidence fulfills Reid's prescription for an earthquake prediction; it also complements a forecast predicated on the seismic history of the Parkfield segment of the fault, which has ruptured in a $M = 5^3/_4$ earthquake every 22±5 yr since 1857 [*Bakun and Lindh*, 1985].

Recent Aseismic Deformation in Southern California?

Despite concerted efforts both to measure the variations of strain accumulation and vertical deformation in the vicinity of the big bend of the San Andreas fault and to model the errors intrinsic to these measurements, considerable doubt still attends reports of episodic deformation in southern California. Simply put, the observed strain increment in 1979 [*Savage and Gu*, 1985*b*] and the uplift at Palmdale [*Castle et al.*, 1984] are close enough to measurement uncertainty to be suspect.

Aseismic uplift of southern California during the period 1953–73 has been a subject of continuing examination since it was first reported by *Castle et al.* [1976]. Systematic leveling errors caused by atmospheric refraction of the line of sight were examined in detail during the past 4 years. Most of these studies found the errors to be significant [*Holdahl*, 1983; *Jackson et al.*, 1983; *Shaw and Smietana*, 1983; *Strange*, 1981; *Heroux et al.*, 1985; *Stein et al.*, 1986]. A field test of atmospheric refraction showed that the error can be modeled and removed from surveys made since 1930; when this is done the aseismic uplift at Palmdale during 1955–65 shrinks to 56±16 mm [*Stein et al.*, 1986]. Others contend that the experiment was inadequate [*Castle et al.*, 1983*a*, *Craymer and Vaníček*, 1986] or that the errors are small and do not merit correction [*Castle et al.*, 1983*b*, 1984, 1985; *Burford and Gilmore*, 1984]. An equally large systematic error in leveling conducted with the Zeiss Ni1 automatic level instrument used during 1972–80 has been studied in the laboratory [*Whalen*, 1984] and empirical [*Strange*, 1985; *Holdahl et al.*, 1986] methods. Under most circumstances, the error can be removed, but the laboratory investigations report the error to be twice as large as analysis of the field records would indicate. Magnetic errors require further study; they may affect some of the southern California leveling results presented by *Jachens et al.* [1983].

Strain changes measured by southern California Geodolite networks showed that an anomalous contraction measured perpendicular to the San Andreas fault since 1973 was abruptly released during mid-1979. This change was immediately preceded by large line-length changes on the VLBI triangle that circumscribes the Palmdale Geodolite net, and it was accompanied by changes in gravity and elevation [*Jachens et al.*, 1983]. These VLBI results were never published; analyses of the VLBI measurements made since 1980 show no such excursions [*Lyzenga and Golombeck*, 1986]. *Jackson et al.* [1983] argued that the Geodolite observations were contaminated by a large zero or offset error and systematic temperature-dependent errors. *Savage and Prescott* [1983] replied that these errors are too small to account for the strain excursions. *Savage and Gu* [1985], however, acknowledge that since 1980 the principal strain changes measured with the Geodolite have been more erratic than they were before 1979, casting renewed doubt on the earlier Geodolite anomaly. Statistically significant correlations between dilatation, gravity, elevation, and the local magnetic field provide some independent corroboration of the strain changes [*Jachens et al.*, 1983; *Johnston*, 1987], but the signal/noise ratio for the individual observations is quite small. A 2-color geodimeter installed at Palmdale has been recording measurements weekly to quarterly since 1980. Excursions in strain over periods as short as 2 weeks have been reported, but these anomalies suffer from similar uncertainty [*Langbein et al.*, 1982]. Unlike the principal strains, the rate of shear strain accumulation has been very steady since 1972 [*Savage et al.*, 1986*b*]. Because shear is less subject to scale or temperature errors, the changes in dilatation remain questionable.

Volcanic Deformation

Magmatic Injection

In 1980, an intense seismic swarm punctuated by four moderate earthquakes ($M_L > 6$) began beneath Long Valley caldera in California. The caldera had formed when the volcanic edifice collapsed into its partially voided magma chamber during a spectacular eruption 700,000 yrs ago, when 600 km^3 of magma was expelled and ash was spewed as far as Kansas, 1500 km away. The caldera has since produced smaller peripheral eruptions every 500–1000 yr. *Savage and Clark* [1982] showed that the spate of earthquakes was accompanied by extension and domal uplift of the caldera floor, deformation which they attributed to injection of 0.1 km^3 of magma into the crust at a depth of about 10 km and to strike-slip faulting adjacent to the caldera. The caldera abruptly uplifted 0.25 m between 1975 and 1980 after at least 60 years of stability [*Castle et al.*, 1984]. A succeeding seismic swarm in 1982–1983 was monitored in great detail by increased horizontal and vertical geodetic coverage. *Savage and Lisowski* [1984] and *Savage and Cockerham* 1984] explained the renewed activity by intrusion of 0.03 km^3 of magma into an inclined tabular body, or sill, at a depth of 8 km, accompanied by strike-slip displacement on a vertical fracture extending from the sill.

Denlinger and Riley [1984] and *Denlinger et al.* [1985] analyzed a concentrated electronic distance measurement network and suggested that the 1980–83 deformation was produced by opening of a horizontal sill accompanied by movement on a separate strike-slip fault, but *Rundle and Whitcomb* [184] and *Whitcomb and Rundle* [1985] argued that two magma cupolas at depths of 5 and 8 km, or fingers extending upward from a larger chamber, fit the geodetic and repeat gravity observations better than a sill. As with attempts to determine the transition depth of strike-slip faults, interpretations of the caldera surface inflation are nonunique, the depth of the source varying with the volume of the injection. In light of changes in the surface hydrothermal system, and results of seismic imaging studies, however, it is now inescapable that magma has been injected into the upper crust beneath some portions of the caldera. Whether this process will culminate in an eruption or a large earthquake is unknown. A network of tiltmeters, dilatometers, lake level gauges, and a 2-color laser geodimeter have been installed to provide real-time indices of the state of deformation. Since 1983 the rate of extension and elevation change has slowed [*Linker et al.*, 1986], but two more $M_L > 6$ earthquakes have struck outside of the caldera [*Gross and Savage*, 1986; *Gross and Savage*, 1987].

Rabaul caldera in Papua New Guinea, and Campi Flegrei ("Fiery Fields") in Pozzuoli Italy, also awoke during the past several years, heralded by dramatic increases in seismicity and inflation but, as yet, no eruptions. Raubul, which last erupted in 1937, and Campi Flegrei, which erupted in 1538, both exhibit deformation compatible with very shallow sites of magma injection. *McKee et al.* [1985] suggest that inflation observed at Raubul during 1983–84 can be explained by two sources at 1 and 3 km depths. The rate of extension has subsided since 1984. At Pozzuoli, a pulse of uplift during 1982–84 totaling 160 cm may have resulted from sources of inflation at a depth of 3–5 km [*Decker*, 1986]. The deformation rate has also slowed during the past year at Pozzuoli. Prehistoric eruption of both of these calderas led to the formation of deep coastal harbors; as a result the areas are densely populated, and must be closely monitored for earthquakes and deformation.

Eruption Monitoring

Recent eruptions at Mt. St. Helens Volcano in Washington, and Mauna Loa and Kilauea Volcanoes in Hawaii have provided convincing evidence that these violent events can be predicted, if the magma-ascent process is understood and vigilant geodetic surveillance of the volcanic edifice is maintained. The 18 May 1980 eruption of Mt. St. Helens was preceded by bulging of the north flank at an astonishing rate of 2 m/d as well as by a sequence of summit earthquakes. *Swanson et al.* [1983, 1985] and *Dzurisin et al.* [1983] document thirteen minor dome-building eruptions of Mt. St. Helens through 1982 that were predicted by precursory seismicity, deformation, and gas emission. Measuring the daily movement of the congealed plug of magma within the summit crater proved a reliable but dangerous technique to gauge deeper magma movement.

Using geodetic and seismic precursors of the 1975 summit eruption of Mauna Loa in Hawaii to guide interpretation of data gathered during 1981–83, *Decker et al.* [1983] forecast that an eruption of Mauna Loa would occur during the succeeding 2 years. Accelerated line-length changes across the volcano summit accompanied by shallow seismicity increases led these investigators to suspect magma movement at a depth of about 3 km

[*Klein*, 1984]. The eruption followed in 1984 [see *Lockwood et al.*, 1985]. South of Mauna Loa, a $M = 7.2$ earthquake and associated eruption of Kilauea volcano in 1975 displaced the entire volcanic edifice south of the summit 8 m seaward and 3.5 m downward along a near-horizontal slip surface at a probable depth of 6–10 km [*Lipman et al.*, 1985]. This eruption was preceded by large and abrupt changes in line length on the flank of the volcano and by inflation of the summit [see also *Davis*, 1986, and *Dvorak et al.*, 1986].

The observation that these volcanoes of different ages and histories all possess magma chambers at depths of 3–5 km has led *Decker* [1986] to argue that the chamber must migrate upward as the volcano grows and evolves. If this were not the case, older volcanoes would increase the distance to their magma source as they built up successive layers of volcanic deposits. These upper crustal magma chambers must in turn have conduits to larger and deeper reservoirs, because the penultimate eruptions expel more magma than the shallow chambers can hold.

WHAT'S AHEAD

Space Geodesy

Space and satellite-based geodesy is developing rapidly; within the next few years, repeated Very Long Baseline Interferometry (VLBI) and Global Positioning System (GPS) measurements will begin to flood the literature. Portable VLBI receivers make possible an expanded program of measurements [see, for example, *Davidson and Trask*, 1985; *Herrring*, 1986; *Kolenkowicz et al.*, 1985; *Kroger et al.*, 1986]. The ability to measure baselines as long as 10,000 km insures that VLBI will remain vital for the study of plate motion. The Global Positioning System promises to become the most widely used geodetic tool for baselines less than about 500 km in length, in part because of its portability and low cost [see *Goad*, 1985, for the current status of GPS development]. *Bock et al.* [1985] measured a 35-station network with a precision of 1 ppm. Atmospheric corrections using water vapor radiometers should allow this precision to be maintained during inclement weather [*Ware et al.*, 1985]. Currently, 4–45 km baselines are being measured simultaneously with GPS and the laser Geodolite; GPS displays an accuracy of 0.5–1.0 ppm relative to the laser Geodolite, which has a demonstrated accuracy of 0.2 ppm [*Prescott and Svarc*, 1986]. Planned satellite launches will strengthen the current constellation, and improved satellite tracking by GPS receivers co-located with VLBI radio antennae offers the prospect of increased precision during the next few years. The deformation of the sea floor accompanying great subduction earthquakes, hidden from us by the ocean, may also soon be measured from associated changes in the gravity field by the Geopotential Research Mission satellites [*Wagner and McAdoo*, 1986].

The Need for Deformation Metrology

A few perennial problems have hampered deformation monitoring in the U.S. Instruments have too often been placed in field operation before they were proven to be reliable, drift free, and accurate. New generations of instruments or new methods for conducting measurements have also succeeded older ones without parallel running of both systems to assess their systematic differences. Finally, in an attempt to monitor as many tectonically active regions as possible, too few instruments have been dispersed too widely. These practices have been more costly to our understanding of active faults rather than active volcanoes, because volcanic regions tend to deform and generate earthquakes at a much higher rate than do tectonic provinces. The establishment of the Piñon Flat Observatory in California [see, for example, *Wyatt et al.*, 1984; *Agnew*, 1986] has been vital for redundant testing and intercomparison of virtually all short- and long-baseline deformation measurement systems at a stable site on crystalline bedrock. Efforts to monitor the approach of the next earthquake on the San Andreas fault along the Parkfield segment and to watch for signs of a potential eruption at Long Valley caldera have also created two concentrated instrument deployments. Evaluation of a dozen competing and ancillary measurements will improve our confidence in some instruments and remove our faith in others.

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