that variability was not affected by flow regulation through the first dam. Thus we assumed that the slope constant in Eq. 1 also holds for the earlier record. Then, deviations of the annual maximum river height (H') may be converted to annual volume discharge deviations (V') by setting

$$V' = 20.8\,H' \qquad (2)$$

If V' is normalized through division by some reference value $\overline{V}$, the significance of variations is brought out. We have chosen $\overline{V} = 91$ km^3, the average discharge for the last century. Even if this value varied somewhat over the centuries, the record becomes readily comparable to that of rivers such as the Colorado (V = 18 km^3/year), the Columbia, and other rivers with widely varying mean flow. However, according to Jarvis (*4*, p. 1022), "... one can conclude that the flood levels about 3000 B.C. (which are recorded on the Palermo stone) and the later data concerning Nile floods given by the Greek and Roman authors agree well with those of today."

A major problem is to find values of $\overline{H}$ over the centuries from which the deviations H' are to be computed. The river bed has risen at an average rate of 10 to 15 cm per century as a result of sedimentation. A first attempt to remove this trend linearly fails. The rise of the river bed was not nearly linear; there were only small rises and even reductions during the early centuries of data and then rapid increases thereafter. Therefore, factors other than sedimentation affect the 1300-year trend. We considered that the best procedure would be to plot the "mass curve" showing the changes of slope by 5-year means (*4*), 10- and 100-year means (*6*), and 25-year means. From these changes, the onset and termination of short-period climate fluctuations were determined; only such cases where the duration of an "episode" was at least 50 years and where the accumulation $\Sigma V'/\overline{V}$ exceeded ± 1 at the extreme value were considered. Altogether, eight such episodes were found (Fig. 1 and Table 1); another maximum has been hypothesized to have occurred near 1600 (*6*). Most remarkable is the fact that, after several small early episodes, fluctuations on the 100-year scale died out completely for over 250 years; this period is coincident with the well-known minimum of storminess of the "little climatic optimum" in the North Atlantic (*6*). At the end of this quiescent period the Nile fluctuations resumed, at first weakly and then strongly after 1300, presumably coupled with strengthening and expansion of the belt of upper west winds (*6*). Herewith interaction between extratropical and tropical regions, well known on very short time bases, is clearly suggested as a major control of the 100-year episodes. Butzer *et al.* (*7*) have found evidence of discharge variations on the time scale of centuries during earlier historical periods.

A second point of interest is that the episodes consist of a combination of high-flow and low-flow (or vice versa) years. One could think of a long-term constant basic flow upon which an occasional period of high-flow or of low-flow years is superimposed. In that case the mass curve would make a single jog and then return to its previous slope. But this pattern was not found. There are always two jogs, so that we are confronted with wavelike oscillations in time, some of them solitary events and some a sequence of events.

Table 1 shows characteristic numbers for all eight episodes. Their duration could be eliminated by the introduction of nondimensional time. There is, however, no obvious way to reduce the other variables so that a single episode model would emerge. Rather, an irregular spectrum of episode properties is brought out. The change in the available water supply across the episode peak (column 4) lies predominantly in the range 10 to 15 percent. However, in the last two events nearly double that change occurred; hence the significance for water supply is much larger. The mean slope of the curves in Fig. 1 may be a measure of the episode severity (column 5); it is well correlated with the preceding column, since the cases were all fairly symmetrical during growth and decay stages. The last column is an attempt to define an impact measure. We calculated these values by multiplying percent volume change by the duration. On the basis of this index, only the last three episodes can be rated as having had major impact on the economy.

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Recent Crustal Uplift in Yellowstone National Park

Abstract. *Comparison of precise leveling measurements made in 1923 with those made in 1975, 1976, and 1977 reveals that the 600,000-year-old Yellowstone caldera is being uplifted relative to its surroundings. Maximum relative uplift since 1923 is in excess of 700 millimeters—about 14 millimeters vertically per year. The most likely cause of this rapid and unusually large surface deformation is a recent influx of molten or partially molten material to a location within the crust beneath Yellowstone National Park.*

Yellowstone National Park is an important area of Pleistocene intracontinental volcanism; three similar cycles of intense silicic volcanism have occurred there in the past 2 million years (*1, 2*). The latest cycle began about 1.2 million years ago with the generation of two adjacent ring-fracture zones in central Yellowstone above two magma chambers that were probably connected at deeper levels with a larger magma body. Growth of the ring-fracture zones and minor rhyolitic volcanism eventually led to an explosive eruption of rhyolitic pumice and ash (1000 km^3) 600,000 years ago. Immediately after this eruption the roofs overlying the two magma chambers collapsed to form the elliptical Yellowstone caldera. A resurgent dome in the northeastern part of the caldera developed shortly after formation of the caldera; another resurgent dome in the southwestern part, near Old Faithful, appears to be the result of renewed magmatic activity beginning 150,000 years ago. Rhyolite flows as young as 70,000 years are associated with this recent activity.

In this report we present evidence for recent uplift of the Yellowstone caldera in excess of 700 mm (about 14 mm/year). This rapid and unusually large vertical movement is of considerable interest because geophysical studies indicate that a hot upper crustal body, which may be at least partially molten, still underlies the

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caldera (3–8). Moreover, the possibility of a volcanic eruption in Yellowstone in the future cannot be ignored (2).

The basis for vertical crustal movement studies in Yellowstone was established in 1923 when level lines of second-order precision were observed throughout the park by the U.S. Coast and Geodetic Survey (9). Nearly all of the 1923 level lines were reobserved to first-order precision by the U.S. Geological Survey (Topographic Division) in cooperation with the University of Utah during the summers of 1975, 1976, and 1977 (10). Assuming that the benchmarks did not move relative to one another between 1975 and 1977, a 52-year interval (1923 to 1975) has been defined for which the following quantities may be computed for each benchmark relative to an arbitrarily chosen reference level: (i) the change in height (Δh) and (ii) the average relative vertical velocity (V).

A contour map of Δh and V computed relative to benchmark K12 is shown in Fig. 1. The principal feature of this map is an elongated uplift that approximately coincides with the Yellowstone caldera. Benchmarks with Δh greater than 300 mm (V = 5.8 mm/year) are confined to the caldera, and benchmarks located along the line between Fishing Bridge and West Thumb indicate that the zone of largest Δh (greater than 500 mm) extends along a northeast trend parallel to the caldera axis. The maximum Δh is 726 mm (V = 14.0 mm/year) and occurs at a benchmark 3 km north of Fishing Bridge near the margin of one of the resurgent domes. These positive Δh values are among the largest that have been discovered in a continental setting far from a plate boundary (except in landslides and man-made disturbances). The nature of the uplift northeast and southwest of the caldera rim is unknown because of the absence of leveling data for those areas.

The comparison of repeated leveling measurements has become an important technique for detecting vertical surface movements, but its use demands careful error analysis. The accuracy of a Δh value depends on the accuracy of the height difference measured in each of the two constituent levelings. For the levelings in 1923 and 1975, 1976, and 1977, the random error of a measured height difference was assumed to be normally distributed with mean zero and standard deviation equal to $\alpha\,(L/N)^{1/2}$, where L is the distance between the two points connected by leveling, α is a monotonically increasing function of L that describes the quality of the leveling, and N is the number of repeated measurements made of the height difference during the leveling. The assumption of distance-dependent α is an attempt to include undefined systematic errors that can accumulate with distance in the random error (11). Since this statistical model cannot represent systematic errors that accumulate with height, we checked the Δh data in profile form for a correlation with topography. For example, the profile across the caldera between benchmarks Q13 and B14 (Fig. 2a) shows that although large Δh values occur in areas of high elevation, there is no consistent correlation between Δh and the topography. We were also unable to find a straightforward correlation between Δh and topography along other profiles. This evidence does not rule out the possibility of systematic errors accumulating with height in the 1923 and 1975–1977 levelings, but it does indicate that such errors are probably not a significant component of the Δh data.

Statistical estimates for α were obtained from discrepancies between the forward and backward levelings of a section between benchmarks ($0 < L < 2$ km) and from circuit closures where $L \doteq 125$ km (12). Values of α at intermediate distances were computed by straight-line interpolation between the statistically estimated endpoints (Fig. 2b). For small L we obtained $\alpha(1923)$ = 3.6 mm/km$^{1/2}$ and $\alpha(1975\text{–}1977)$ = 1.3 mm/km$^{1/2}$; these are typical α values for the order and date of the leveling work considered here. The 1975–1977 circuit closures are all within a few millimeters of the random error predicted by $\alpha(1975\text{–}1977)$ = 1.3 mm/km$^{1/2}$, L = 125 km, and N = 2, indicating that $\alpha(1975\text{–}1977)$ is a constant and that systematic error accumulating with distance is probably not a factor in the leveling data for those years. In contrast with this result, the 1923 circuit closures are four times the expected value for $\alpha(1923)$ = 3.6 mm/km$^{1/2}$, L = 125 km, and N = 1, suggest-

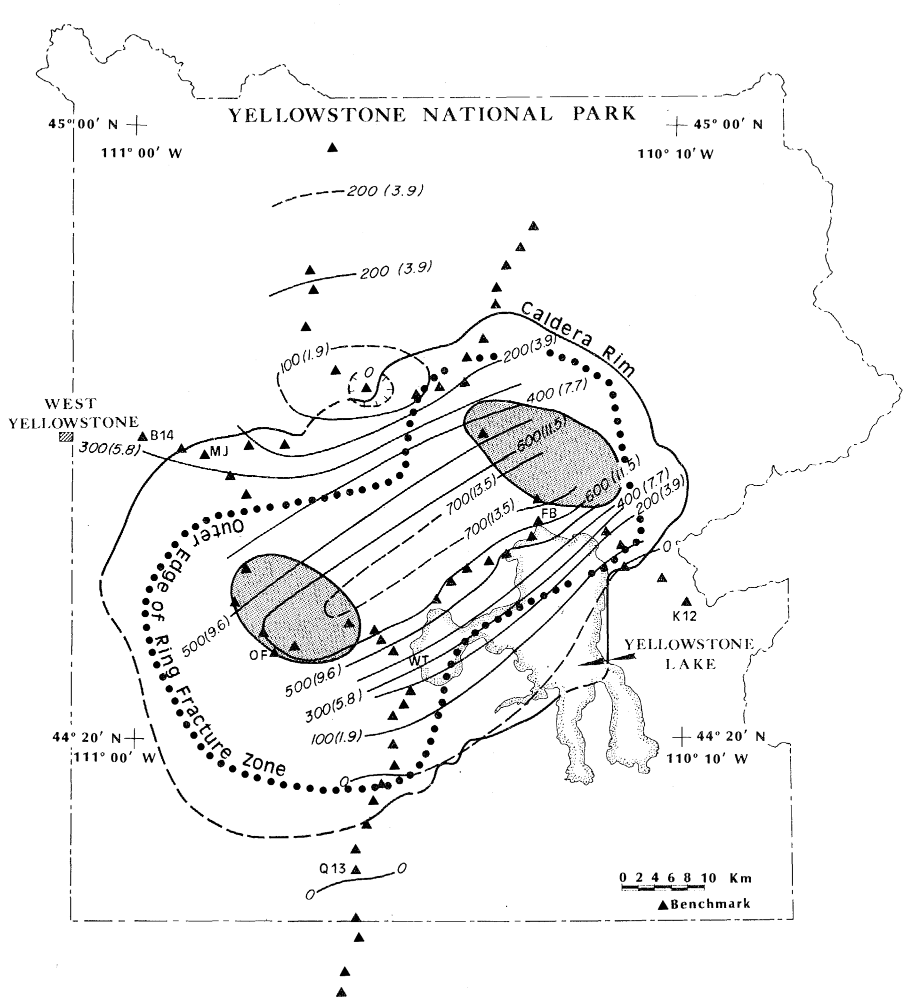

Fig. 1. Contour map of Δh for the time interval 1923 to 1975. Reference point is benchmark K12. Contour interval is 100 mm; corresponding values of V are given in parentheses in millimeters per year. Shaded areas are resurgent domes. Abbreviations: *FB*, Fishing Bridge; *MJ*, Madison Junction; *OF*, Old Faithful; and *WT*, West Thumb.

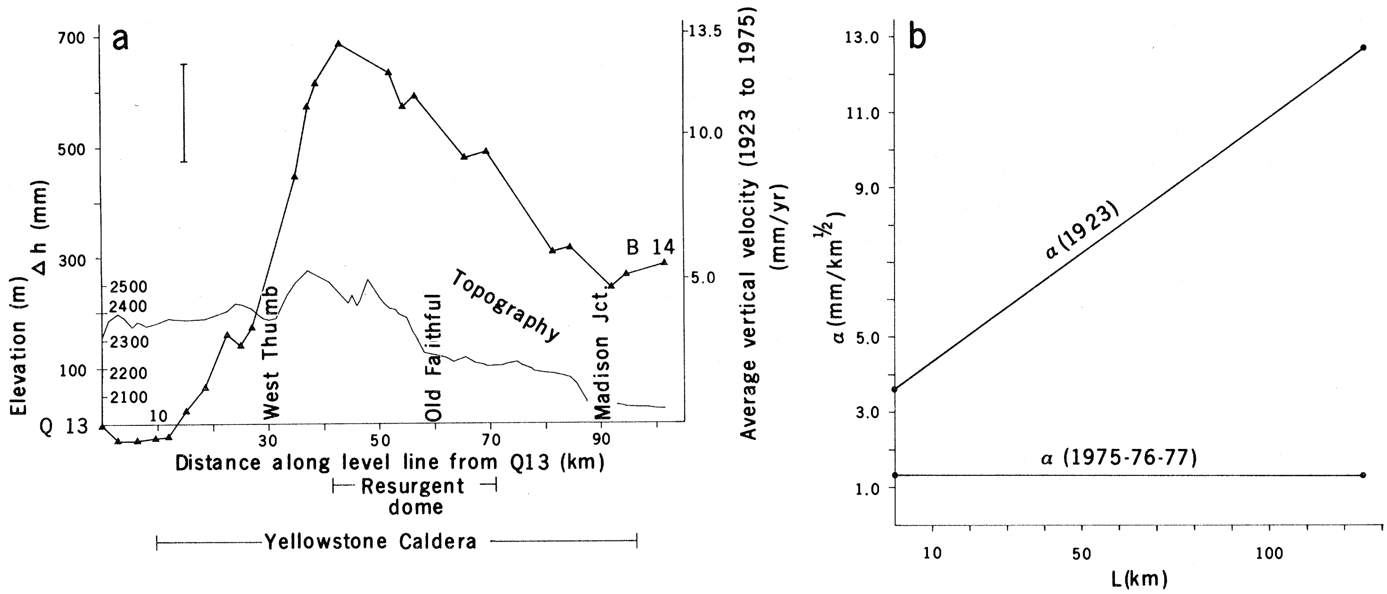

Fig. 2. (a) Profile of Δh, V, and topography for time interval 1923 to 1975. Bar length represents twice the average of the standard deviations of Δh and V profile values. (b) Plot of estimated α values.

ing that systematic error did accumulate with distance in 1923. The 95 percent confidence interval for α(1923) at 125 km is unavoidably large because of the small sample (only two circuits), but 12.7 mm/km$^{1/2}$ is the best point estimate and has been used as a means of including the undefined 1923 distance-dependent systematic error in the random error. We investigated the possibility that the large 1923 circuit closures are the result of a failure to make orthometric corrections to the observed height difference data. The maximum circuit closure resulting from a failure to make orthometric corrections during otherwise errorless leveling around a 125-km circuit in Yellowstone was estimated to be approximately 12 mm. The 1923 circuit closures are an order of magnitude larger than 12 mm and therefore must be due to some other source of systematic error.

The standard deviation of Δh is given by the square root of the sum of the squares of the standard deviations for the two constituent measured height differences. The mean of the Δh standard deviations for the 60 1923 benchmarks recovered for use in this study is 74.7 mm and the maximum is 121.9 mm; corresponding mean and maximum standard deviations for V are 1.4 and 2.3 mm/year, respectively. We therefore feel that the Δh and V values for most of the recovered 1923 benchmarks are statistically significant.

As mentioned above, the uplift at Yellowstone occurs in close association with the Yellowstone caldera, a known center of extensive Pleistocene volcanism. This coincidence immediately suggests that the uplift is the result of magmatic processes operating in the crust beneath the caldera. An increase in pressure acting against the boundary-enclosing magma (or partially molten material) is a basic mechanism for uplift in volcanic areas. The increase in pressure could result from an influx of molten or partially molten material from deeper levels, or from intense vesiculation. If a significant change in temperature accompanies either process, then thermal expansion or contraction may become an important effect.

Uplift mechanisms that are not related to magmatic processes but could produce the vertical movements in Yellowstone include horizontal compressive stress associated with active tectonics, dilatancy, and glacioisostatic rebound. The predominance of Quaternary normal faults (*13*) and normal earthquake fault-plane solutions (*7*) for the Yellowstone area is an indication that the direction of regional maximum compressive stress is vertical. This evidence argues against interpreting the uplift at Yellowstone as a result of tectonic horizontal compressive stress. If dilatancy is responsible for the uplift, then it must be developed to the stage at which expansion of a crustal body has occurred. A dilatant mass 5 km thick beneath the caldera must undergo a volumetric strain ($\Delta v/v$, where v is volume) of about 0.0001 to be consistent with an average uplift of 400 mm (*14*). If we may apply the results of laboratory experiments, this volumetric strain would require an unreasonably large stress difference (> 5 kbar) for granite at a confining pressure of 1 kbar (*15*), assuming that dilatancy occurs in situ in competent rock by the opening of microfractures. Thus either dilatancy is not responsible for the uplift, or dilatancy in situ may be occurring by a process that requires a much lower stress difference for the onset of volume increase.

A prominent terrace was formed not more that 9000 years ago around Yellowstone Lake at a level 18 to 20 m above the present shoreline (*16*). The fact that the terrace maintains a nearly constant elevation around the lake is of critical importance because this means that the entire terrace has experienced the same net vertical displacement. The contemporary pattern of average relative vertical velocities in Yellowstone thus cannot have been maintained for more than a few hundred years; otherwise the portion of the terrace located near Fishing Bridge would now be measurably tilted up with respect to another portion of the same terrace about 12 km to the southeast. This information provides a strong argument against any interpretation of the Yellowstone uplift which is based on long-term maintenance of the modern pattern of relative vertical velocities such as glacioisostatic rebound.

The extremely high convective heat loss in Yellowstone National Park (4.02×10^{16} cal/year) can be explained by the crystallization and cooling of 0.1 km^3 of rhyolite magma annually from 900° to 500°C (*17*). If we adopt a simplified cooling history for a rhyolite magma body to explain the heat loss at Yellowstone, then in the 600,000 years since formation of the caldera, a layer of solid granite 26 km thick should have formed beneath the caldera. This conclusion is inconsistent with anomalously low seismic velocities deduced from relative teleseismic P-wave delays, which suggest a partial melt in the upper crust (*5*, *6*). One way to resolve this problem is to postulate a sporadic input of molten or partially molten material from the magmatic source. Such inputs might occasionally produce a temporary swelling of the surface followed by stress relaxation. On this basis it seems reasonable to interpret the contemporary uplift at Yellowstone as one phase in a series of deformation episodes centered on the caldera and related to an ongoing process of intrusion. This hypothesis does not conflict with the youth of the uplift or the fact that there has been no eruption in Yellowstone for 70,000 years (*2*). It has been suggested that intrusion of magma may be responsible for contemporary doming (3 to 5 mm/year) of an 8000-km^2 area just northwest of Yellowstone (*18*), but whether there is any relationship between this doming and the uplift at Yellowstone is an unsettled question.

There is a possibility that the uplift at Yellowstone represents a new magmatic insurgence heralding the start of a fourth volcanic cycle. At this time we cannot distinguish this possibility from one in which only late third-cycle volcanism occurs or from one in which the intrusion of mobile material simply deforms the surface without eruption. If a new eruption were to occur, it would typically be preceded by such phenomena as increased numbers of earthquake swarms, increased hydrothermal activity, and further deformation of the surface.

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Dolphin Lung Collapse and Intramuscular Circulation During Free Diving: Evidence from Nitrogen Washout

Abstract. *Intramuscular nitrogen tensions in* Tursiops truncatus *after a schedule of repetitive ocean dives suggest a lung collapse depth of about 70 meters and suggest that intramuscular circulation is maintained during unrestrained diving in the open ocean. Therefore, the bottle-nosed dolphin is not protected by lung collapse from the decompression hazards of dives to depths shallower than 70 meters.*

Ocean diving studies with a trained dolphin have shown that dolphins may be protected from decompression sickness by alveolar collapse during dives deeper than about 100 m (*1*). Recent studies have shown that dolphins dive more frequently to depths shallower than 100 m (*2*). Therefore, we decided to investigate tissue nitrogen accumulation after a long series of such shallow dives.

A Medspect MS-8 medical mass spectrometer (Searle Scientific Research Instruments Corp., Baltimore) (*3*) was used in this study to monitor intramuscular nitrogen washout after diving. This spectrometer measures the minute amounts of gases that continuously diffuse through a Teflon membrane enclosing a vacuum chamber at the end of an implantable catheter. The gases then enter the mass spectrometer in quantities proportional to their partial pressures.

Two bottle-nosed dolphins named Blue and Brown were trained to dive repetitively for about 1 hour in the open ocean to a depth of 100 m, as depicted in Fig. 1. Descent, ascent, and surface interval times were recorded for each of the 25 dives made by Blue and the 23 dives by Brown. Mean dive time was about $1^1/_2$ minutes, with about 1-minute surface intervals. During surface intervals, the ventilation rate of the dolphins was about ten breaths per minute. After completion of a dive series, the catheter from the spectrometer was inserted, trans-

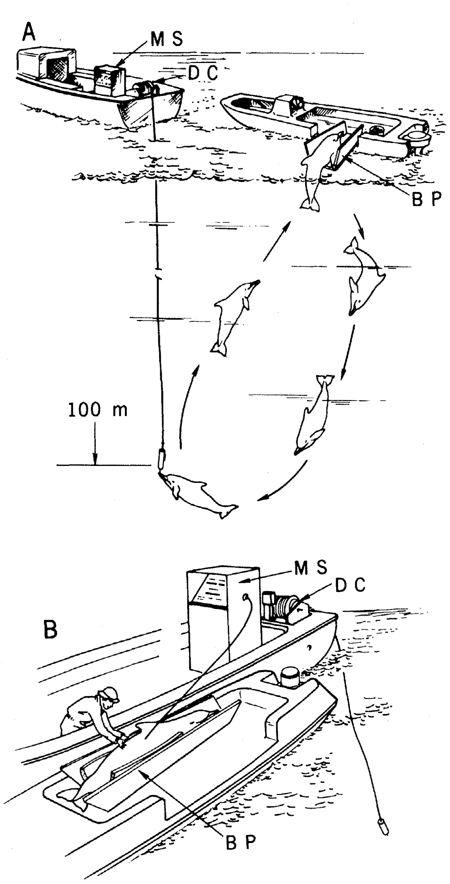

Fig. 1. (A) Each dolphin made multiple dives to a depth of 100 m to press a switch at the end of the dive cable (*DC*). After each dive it returned to the boat for a food reward during an interval of 45 seconds, after which it was commanded to dive again. After completion of the last dive, the dolphin jumped onto the beaching pad (*BP*). (B) The small boat containing the dolphin on the beaching pad was brought alongside the larger craft and the mass spectrometer (*MS*) probe was inserted. The animals remained in this position while the N_2 washout was measured.

