Unusual Marine Erosion in San Diego County from a Single Storm

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Observations of wave-induced geological damage were made along the San Diego coastline to a depth of 25 m, following the storm of 17–18 January, 1988. Massive damage to limestone reefs occurred, including the shearing of individual sections with in-water weights of over 20 tonnes at the remarkable depth of at least 22 m. Large sections of the walls of a submarine canyon were broken off at a depth exceeding 20 m. The drag and inertial forces from the waves in this storm are shown to be about twice those in the largest previous storms of the century, and it appears to be a 200 year event. In addition to the kelp mortality reported in Seymour *et al.*, 1989, there was extensive mortality among encrusting algal and animal communities. The apparent age of the mature successional communities in the deeper sites supports the engineering estimates of the rarity of this event. The movement of cobbles and boulders at depths almost twice as great as the previously assumed limits on effective sediment transport may require coastal engineers to revise cross-shore transport models.

Introduction

The storm of January 17–18 1988, which so devastated the Point Loma and La Jolla kelp forests (Seymour *et al.*, 1989), also resulted in massive effects on geological structures along the coastline within our study areas off Point Loma, San Diego County, California. These unusual effects extended from the high water mark to the remarkable depths of at least 25 m. A wind-driven storm surge of approximately 0·4 m, which is significant by West Coast standards, was superimposed on an astronomical high tide of 2·4 m (Cayan *et al.*, 1988). Coupled with the increased runup associated with the very large waves, this resulted in substantial flooding of low-lying coastal areas, overwash fans of sand and cobbles and the deposition of thousands of tons of kelp wrack. The waves, which exceeded 10 m in significant height offshore and 6 m in the surf zone, caused extensive erosion of sand from exposed beaches. The sand scour levels around the Scripps Pier pilings were lower than ever recorded.

Kelp forest observations

We have been studying selected parts of the Point Loma kelp forest since 1970 (Figure 1). Five study areas, chosen to encompass a range of physical conditions, are each visited
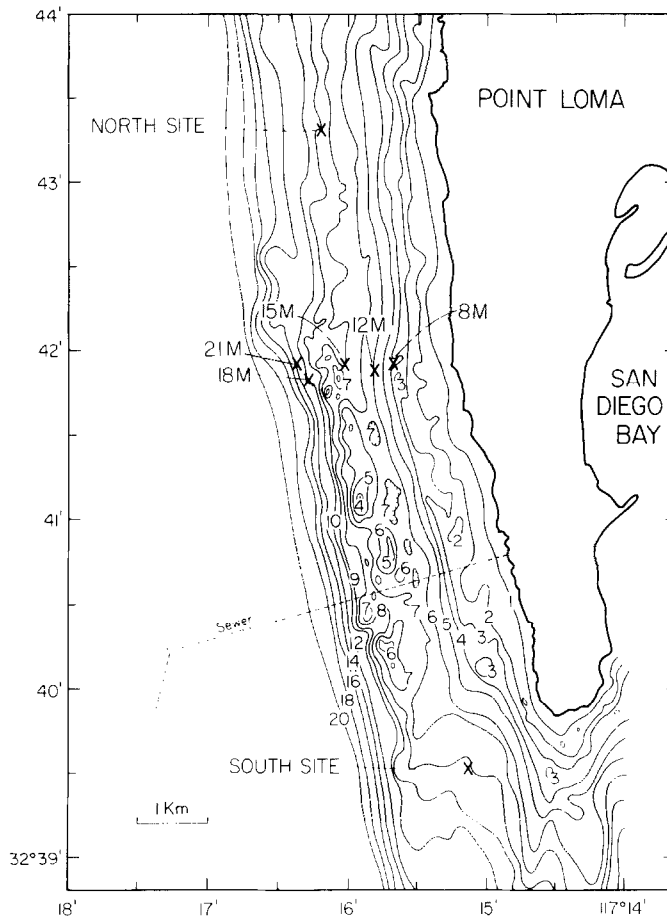
Figure 1. The Point Loma kelp forest is located on a broad shelf west of San Diego, California. Study sites described in the text are marked. Contour lines are in fathoms.

several times a month. Algal and animal populations are monitored regularly, and a census was completed during the month before the storm. During the nearly two decades of this study, there have been several storm periods which caused considerable mortality to *Macrocystis* populations (summarized in Seymour *et al.*, 1969). In spite of severe battering, the shorter, more robust understory kelps usually survived these storm (Dayton and Tegner, 1984) and only twice has there been sufficient energy at the kelp holdfast depths to move large rocks.

In addition to the massive kelp mortality, the storm of 17–18 January 1988 resulted in dramatic physical changes. Understandably, these disturbances were most pronounced in shallow water, but they extended seaward to depths in which such physical perturbations had not previously been observed in this area. At our 8 and 12 m sites, the bottom had the appearance of being sandblasted. In addition to the kelps, between 50–85$^{\circ}_{\circ}$ of the nearly 100% cover of encrusting coralline algae had been removed, exposing bare, smooth rock. The substratum in these areas is a well-cemented conglomerate. Many of the embedded rocks and cobbles had been ripped from the bedrock leaving scattered depressions and cavities. A thin ledge, about 15–20 cm thick, which faced into the wave-driven bottom
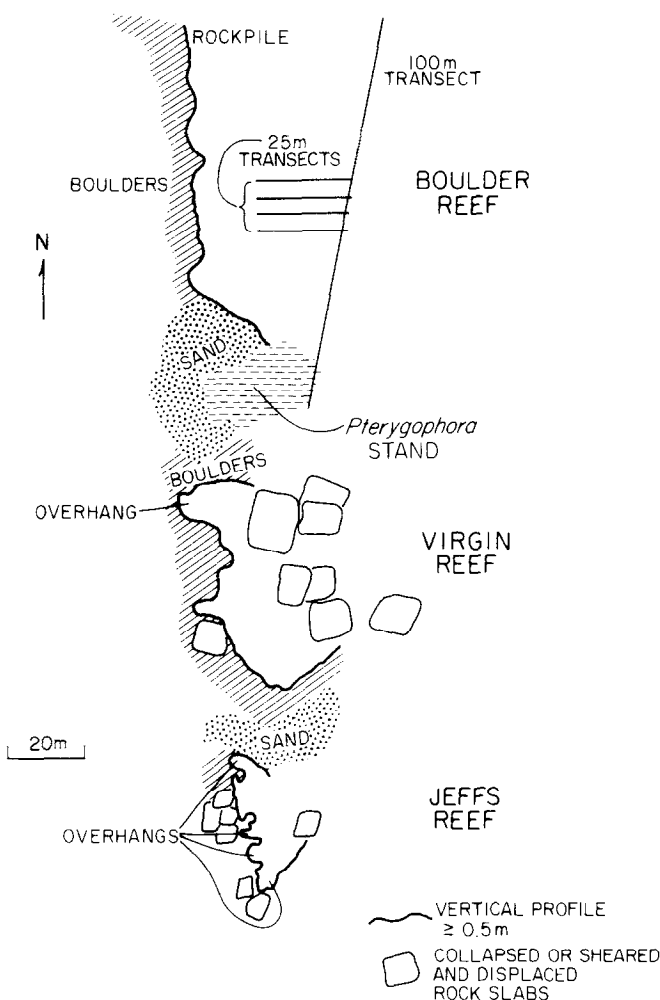
Figure 2. Diagram of the reefs at the 15 m site illustrating large rock slabs displaced during the storm. The *Pterygophora* stand is a large patch of understory algae.

surge, was split from the bottom along a bedding plane. Many smaller pieces of bedrock were ripped away, exposing boring and settling clams, sipunculids and snails which had been smashed or abraded by the surface erosion. The same fate befell chitons, small limpets and urchins whose smashed remains were visible in cracks and crevices. We had previously marked the 12 m site with a string of six concrete-filled scuba tanks joined with a heavy chain. After the storm, two of these heavy tanks and pieces of the chain were found inshore of the 8 m site—at least 0·5 km from their original location. The remaining tanks—and an expensive thermograph—have not yet been located. The large experimental quadrants had been marked with 12 mm leaded line attached to the bottom with heavy railroad spikes. Following the storm, most of the lines were gone. Many of the spikes were bent, some as much as 90° or were gone—leaving holes in the bottom that marked their former location.

The site at 15 m depth had been characterized by a large, sometimes undercut limestone ledge, locally called 'Virgin Reef' (Figure 2). This reef had approximately 1–2·5 m of

vertical relief. Some 100 m to the north was the 'Boulder Patch', an area littered with boulders from 0·5 to 3 m in diameter. About 75 m south of Virgin Reef, the limestone ridge elevated to form Jeff's Reef—a sloping, undercut reef about 3 m above the bottom. Post storm, the 'Boulder Patch' was massively disturbed with over 50% of the rocks turned and/or moved as far as 4 m (Plate 1). At least 28 m of the total 35 m length of Virgin Reef suffered having the top meter of rock ripped away. The average volume of the pieces removed from the reef was about 2 m³. The largest displaced rock that we observed was $2·5 \times 1·8 \times 0·75$ m. It was inverted and about 10 m inshore of the reef. Jeff's Reef lost similar sized rocks and one piece, representing about 25% of its upper surface, came to rest some 15 m inshore. A rock about $2 \times 1·5 \times 1$ m had been moved from elsewhere to abut Jeff's Reef.

Of all our marker buoys, only one at a nearby 15 m depth site survived the storm. It was a subsurface buoy with several floats shackled to an anchor in a large reef system. The large flotation volume had always resulted in a very rigid system that resisted being pulled aside by an attached boat even in strong winds and currents. After the storm, however, the buoy mooring line was abraded about 7 m above the bottom, suggesting that it had been pulled under sufficiently to contact a small reef about 7 m from the anchor point!

Just offshore of the 18 m site was a 20 m² rock reef extending about 2·5 m above the local substratum. This outcrop was reduced to rubble during the storm and much of it was projected shoreward, leaving nearly complete kelp mortality in its wake. At this depth, the surrounding kelps had about 25% survivorship, indicating a significant direct effect of these wave-driven projectiles. Finally, the central 21 m site was also heavily damaged in localized patches. The tops of reefs had been stripped away only in a few locations, but there were windows of gravel and cobbles about a meter thick piled against some of the ledges, with heavy damage along their trajectory. In this area, the kelps were either gone or heavily abraded. In many areas, the encrusted coralline algae was completely eroded away and there were many dead animals. Other areas at this remarkable depth understandably had very little damage to either the substrate or the biota. The presence of ridges or reefs seemed to have offered protection from damage.

We had two other sites at 17 m depth at the northern and southern ends of the kelp forest which were also damaged. It was interesting to note that the damage was considerably less at the southern site. We were not able to explain definitively the reasons why the southern end was spared relative to the extensive damage at the center and especially at the northern end. However, we suspect that a reduction in the availability of cobbles or coarse sand at the southern end may have reduced abrasion and impact damage. The wave directions through the storm could be not defined with sufficient accuracy to deduce whether there was a significant defocussing of the waves because of bathymetric differences. The northern 17 m site suffered a surprising amount of damage, given its great depth. This area lost almost all of its *Macrocystis*, most of the understory, and again, the crusting algae was severely sandblasted. There was overturn of many ledges, including one section at least 25 m long which was only about 15 cm above the bottom and had no overhang. Several colonies of *Diopatra* worms were removed, as were chitons, small snails and clams and various echinoderms.

Submarine canyon observations

Between Point La Jolla and the pier at the Scripps Institution of Oceanography there is sharply incised submarine canyon, called La Jolla Canyon, that extends from very deep
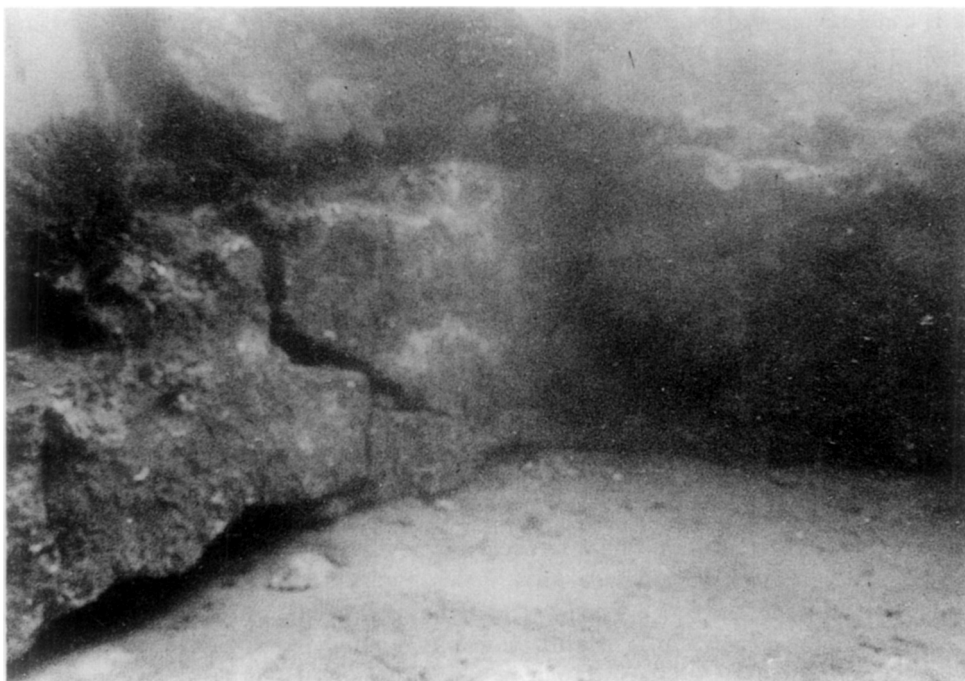
Plate 1. Underwater photograph of part of the top of Virgin Reef after the storm. A slab has been removed from the foreground and cracks are visible in the remainder of the reef. The vertical face visible in the center of the photograph is about 0·5 m high.

water into a depth of about 20 m. It was eroded by turbidity currents formed from beach sand suspensions over geological time scales (see Seymour, 1986). It is no longer active as a sand sink—this function having been superceded by a newer canyon about 1 km north. Some 80% of the eastern edge of the rim at the shoreward end of the canyon was dislodged and tumbled into the canyon. The ejected pieces were about 2–3 m wide and 1–2 m thick. The only undamaged spots were in areas protected by large outcroppings in the canyon wall.

Physical causes of the observed damage

It is clear from the observations of massive fractures of rock structures that had survived for very long periods of time, and which are known to have survived the onslaughts of many intense storms in the past decade alone, that some extraordinary forces came into play during the 17–18 January 1988 storm. To understand these forces, it is necessary to consider how water flows load objects that resist their motions (Sleath 1984; Morison *et al.*, 1950; Wiegel 1964). The most common load, and the one that we tend to understand best from our own experience, is the friction force—usually called drag. This load is caused by the velocity of the fluid over and around the resisting object. It is higher for objects that offer a lot of resistance (large, rough and not streamlined). The drag force on a

body can be expressed approximately as (Sarpkaya & Isaacson, 1981; Grace & Casciano, 1969)

$$F_d = 0.5 \, \rho \, C_d A \, v^2$$

where: F_d is the drag force; ρ is the water density; C_d is a coefficient of drag, approximately unity; A is the frontal area of the body; and v is the water velocity.

Let us consider the effects of a given maximum flow speed on rocks of various sizes. The drag force is proportional to the frontal area, the cross-sectional area opposing the flow. Simplifying to spherical (or, at least, geometrically similar) rocks, this means that the maximum dislodging force will be proportional to the square of some characteristic diameter. The resistance of the rock to being dragged with the flow comes from friction against the bottom, which is approximately proportional to the submerged weight of the rock. This says that the rock resists motion with a force that is proportional to the diameter cubed. With the dislodging force increasing only as the square of the diameter while the resisting forces increases as the cube, it is clear that for a given speed of flow, there is some limiting diameter of rock which is not going to be moved.

There is a second load, however, which only occurs with flows that are changing rapidly with time. These loads do not tend to occur in rivers or in winds, which are relatively steady, but do occur in wave-driven flows, which change direction completely every few seconds. These loads are called inertial or acceleration loads. They result from the inertia of the resisting body, which is a measure of its unwillingness to change the magnitude or direction of its present speed (which may be zero). This inertial force can be expressed as (Carstens, 1952)

$$F_i = (1 + C_m) \, \rho \, A \, L \, a$$

where: F_i is the inertial force; C_m is a coefficient of added mass, approximately unity; L is a characteristic length of the body; and a is the water acceleration.

In general, because this second force is neither steady nor common in our experience, it is more difficult to envision. Consider a wave moving through the water. Ahead of it some distance, the water is either still or even moving in the wrong direction relative to the speed observed locally. To have the form of the wave propagate forward, it must accelerate (or decelerate) an appropriate mass of water ahead of it. Now assume a mass of water which is displaced by a rock or other fixed object. If the rock were not there, we could calculate the force that the water surrounding this volume would have to exert on the water within in order to overcome its inertia. Unless the rock is so huge that it disrupts the whole wave flow, then we can say pictorially that the surrounding water does not know that the rock is there and attempts to accelerate the volume as though it contains water. The force on the rock is then the force that would have been exerted on the water it displaced. This force is proportional to the displaced water volume—that is, to the cube of the rock diameter. Fluid mechanics specialists are also concerned with an apparent or added inertia associated with water that tends to stay attached to the rock. However, because this discussion only concerns relative sizes, this linear multiplicative factor need not be considered here. The resistance of the rock to being accelerated we have already defined as its inertia, which is proportional to its mass, or to its diameter cubed, for a given fixed density. Thus, we have the result that the inertial load on the object and its ability to resist that load are both proportional to the volume of the object. Size does not matter with this force. A grain of quartz sand and a cobble that is geometrically similar and formed principally of the same material are equally likely to be dislodged by the inertial force of an oscillatory flow. There is another feature of these forces that is difficult to comprehend. The inertial forces are in

quadrature with the drag forces—that is, one is maxium when the other is zero. If the drag forces greatly exceed the inertial forces, as is most often the case under waves, then sand may move under drag loads and cobbles remain stationary (Hallermeier, 1980). As will be seen below, it is possible under some conditions to have inertial loads grow faster and dominate over drag loads. In these cases, every free object of a given density that is small relative to the length of a wave will move the same distance under inertial loading, because the accelerations will be identical.

In the above discussion, it has been assumed that the resisting objects are free to roll, as with sand grains or cobbles. Consider an object fixed to the ocean bed, such as a rock ledge. To dislodge such an object, it is in general necessary to break (shear) it along a plane roughly parallel to the flow. The force to do this depends upon the strength of the rock and the area sheared or fractured (which is roughly proportional to the square of a characteristic dimension). For geometrically similar rock ledges of the same material strength, then, the ratio of drag force to fracture resistance is therefore constant for a given speed, regardless of ledge size. However, to fracture a ledge with an inertial load, the bigger the ledge the easier this is to accomplish. The load goes up by the cube and the resistance by the square of the characteristic dimension. Because of internal defects and nonhomegeneities, shear can occur on skewed planes, so that this argument merely suggests the trend. There are further subtleties associated with how shape changes affect drag and inertial loads that tend to favour breaking off ledges by inertial forces, but they will not be discussed here.

Both types of loads on resisting objects are proportional to the height of the wave. The differences between them among storms (or within storms as wave characteristics change) is caused by the wave period. Flow speeds, which generate drag, are proportional to the inverse of the period, or the frequency. Flow accelerations, which generate the inertial loads, are proportional to the frequency squared. Therefore, in two storms with the same characteristic wave height and different characteristic periods, the one with the shorter period will yield drag and inertial loads which are higher—but the ratio between inertial and drag loads will be increased compared to the storm with the longer period. There is a complementary effect in which both velocities and accelerations are decreased with depth faster under waves with short periods. However, in shallow water, particularly in the surf zone, this effect is less important than the increase in flow, even at the bottom. Taking the ratio of the inertial to the drag forces by dividing the first equation given above into the second gives

$$\frac{F_i}{F_d} \simeq \frac{4aL}{v^2}$$

Noting from the table below that for major storms of interest in this area, using the units shown,

$$v^2 \simeq 5a$$

which yields the result that inertial (acceleration) forces will exceed drag forces whenever L, the characteristic length of the body in the direction of flow, is greater than about 1·25 m.

Considering the same three winter seasons (1982–83, 1985–86, 1987–88) discussed in Seymour *et al* (1989), the approximate maximum values of the velocity squared (proportional to drag force) and the acceleration (proportional to inertial force) near the bottom can be tabulated for a nominal depth of 18 m.

TABLE 1. Approximate maximum values of the velocity squared and acceleration near the bottom (nominal depth 18 m)

Winter season	Maximum velocity squared $(m^2 s^{-2})$	Maximum acceleration $(m s^{-2})$
1982–83	5·34	1·07
1985–86	3·42	0·98
1987–88	10·11	1·91

From these data it is clear that both the drag and the inertial forces during the January 1988 storm were about a factor of two higher than those seen in the 1982–83 winter season, which had produced a number of very intense and destructive storms, including the worst seen in Southern California during this century until that time.

Discussion

While the massive kelp mortality (Seymour *et al.*, 1989) was unusual, the severe geological damage that we observed was perhaps even more unusual. In our experience, any major storm along this coast can be expected to move sand and cobbles in water depths less than 18 m and several have sheared pieces of overhanging ledges and moved large boulders. For example, twice in the previous 17 years we have observed small sections of Virgin Reef broken loose. In the January 1988 storm, fully 80% of the top of a very hard limestone outcropping was sheared and displaced. Massive pieces in the range of 12 m^3 (weighing more than 20 tonnes in water) were transported more than 10 m from their original location, often being inverted in the process, with no indication that they had touched the intervening bottom during their trajectories. Thus, the displacing forces must, in addition to shearing the rock segment from its base, lift and project it. Virgin Reef stood 2 m above the bottom. Perhaps even more impressive was the lifting and shearing of ledges with only 15–20 cm of elevation and no overhang, existing at depths of at least 22 m. Only exceptional velocities and accelerations could be expected to disrupt these low profile, streamlined structures.

The destruction of the lip at the top of La Jolla Canyon is remarkable because the great depths of these submarine canyons relative to the adjacent flanks cause a predictable (and observed) substantial reduction in the wave heights locally.

Observation and hindcasting of major storms over long time periods allow coastal engineers to make estimates of the apparent return interval for storms of given intensities. These return intervals are not intended to suggest a periodicity, but imply that over very long periods (say thousands of years) there would be a number of storms of this intensity given by the predicted average interval between storms. Using relationships from Seymour *et al.* (1984), which is based upon a study of the first 83 years of this century, suggests a return interval for the January 1988 storm of about 200 years. It is interesting to compare this estimate with circumstantial biological evidence about the probable rarity of such an intense storm. At depths of 15 m or more, it appears likely that the reefs which were destroyed had communities at such an advanced successional state that some of the deeper ones must have been secure for more than 100 years. There is little hard evidence to support this statement. However common biological sense and some studies (see Warme & Marshall, 1969; Warme *et al.*, 1971) would suggest that the advanced condition of the

boring communities indicates a great age. For example, it probably takes pholad clams many years to become well established in these hard substrates, yet in some of the deeper reefs the pholad stage is largely gone, to be replaced by settling clams, and clams which have finally eroded the rocks enough that the holes go all the way through. These holes have been colonized by colonial anemones, tunicates and sponges, many of which themselves are very slow growing and could be expected to take decades to grow as large as they were prior to the storm. The biological evidence appears to be consistent with the engineering estimates for the probable interval between disruptions of this magnitude. This suggests that these observations may be unique—the only occurrence of such destruction along this coast since man was able to reach such depths. The geological evidence has another significance to coastal engineers. For those concerned with the transport of sediment in the cross-shore direction by waves and currents, there is a conceptual depth at which there is no longer any net transport. At this depth, profiles of the bottom contour taken normal to the shoreline should always pass through a fixed elevation. Shoreward of this depth, the profiles would be expected to change seasonally and in response to major storms as sand moves in the cross-shore direction. Seaward of this point, the tractive forces of the waves are assumed to be too diminished by depth to effectively transport (as opposed to simply displacing on occasion) any significant amount of sand. In Southern California, this depth has been assumed, based upon not very precise profile surveys, to be in the range of 10–12 m. The observations here of cobbles, gravel and boulders weighing many tonnes being displaced at depths far exceeding 20 m suggests that coastal engineers must seriously reconsider their definitions of the depths at which extinction of cross-shore transport can be expected.

Acknowledgements

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