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Mechanisms of Coastal Cliff Retreat and Hazard Zone Delineation in Soft Flysch Deposits

Vicki G. Moon and Terry Healy

Department of Earth Sciences
University of Waikato
Private Bag 3105
Hamilton, New Zealand

ABSTRACT

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The Auckland coastline is characterised by steep cliffs formed in soft, alternating sandstone and siltstone flysch deposits. A rapidly increasing population of the greater Auckland urban area has meant increased development on the cliff tops, with little consideration given to cliff erosion rates.

Three mechanisms of cliff instability are recognised: gravitational collapse of jointed sandstone blocks following removal of underlying frittered siltstone beds; planar failure along steeply dipping fault planes; and planar failure along gently dipping bedding planes. Erosion of overlying soils to maintain a stable slope configuration occurs contemporaneously with erosion of the rock faces.

Average rates of cliff retreat of 2–6 m per century derived from dated structures are attributed to the joint block fall mechanism. Planar failure along fault planes produces up to 5 m of retreat in a single event, and one event may be expected in any ten year period along an individual fault plane. A limit of approximately 10 m maximum retreat exists for this mechanism. Bedding plane failures are very large, but infrequent events.

Planar failures along fault planes represent the greatest risk to structures over their expected lifetime. Thus, a *minimum* coastal hazard zone width of 16 m can be derived, based on 10 m of erosion through fault plane failure, and 6 m set-back due to the stable angle of the overlying soils. Allowing for a factor of safety increases this hazard zone to 23 m.

More refined hazard zones may be derived by mapping of cliff-face structure, hence recognising zones of increased susceptibility to failure. Risks may also be minimised by careful stormwater disposal, which presently often discharges directly onto unstable cliff faces, increasing the instability.

ADDITIONAL INDEX WORDS: *Coastal management, slope failure, control erosion, hazard zone, shore erosion.*

INTRODUCTION

Steep, exposed coastal cliffs characteristically occur around the margin of the Waitemata Harbour of Auckland, New Zealand. Recent population pressure has seen a large increase in the demand for cliff-top property, with land along the cliffs being considered the prime residential real estate, due primarily to the panoramic views. This has led to a marked increase in the number and value of houses built close to the cliff faces, a trend which is likely to continue for some time in many areas. Likewise, the increasing population has resulted in a much greater use of the base of the cliffs as a recreational area.

Despite this increase in the use of the cliff region, little recognition has been given to the risks associated with cliff instability. Over most of the area, development has proceeded apace, with any study of the likely instability restricted to private engineering reports on individual site developments. In some areas, a narrow (3–4 m wide) strip along the cliff top has been appropriated by the local authority and used for public walkways with minimal development, but this is regarded more as a means of allowing public access than as a serious attempt to allow for a coastal hazard zone.

The following discussion systematically outlines the mechanisms by which cliff retreat in such flysch deposits occurs, and uses this as a basis to assess the risks associated with the use of the cliff-adjacent land, and hence to derive a coastal hazard zone for this type of coastal landform.

COASTAL ENVIRONMENT

The Waitemata Harbour and inner Hauraki Gulf area is shown in Figure 1. The coastal type area is an indented lee shore, sheltered from the prevailing westerly and southerly waves (PICKRILL and MITCHELL, 1979). Further protection from swell waves of Pacific origin is provided by

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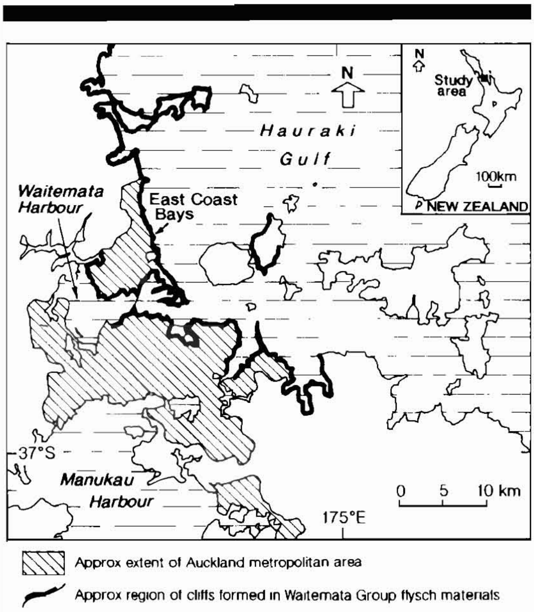


Figure 1. The approximate extent of cliffs formed in the flysch deposits of the Waitemata Harbour and inner Hauraki Gulf adjacent to metropolitan Auckland.

the nearby islands, giving a maximum fetch of approximately 25 km. Hence, most of the waves acting on the cliffs are locally wind generated, resulting in a moderate to low energy wave environment. Tidal regime is mesotidal with an approximately 2 m tidal range.

SHORELINE MORPHOLOGY

The cliffs throughout the Auckland area are consistently very steep (Figure 2), forming angles of 10–20° to the vertical in general. Their height varies from approximately 5 m to 30 m with a typical cliff height of approximately 20 m.

At the top of the cliffs, it is common to see an area across which the slope decreases markedly behind the edge (Figure 2). This "bevelled zone" is believed to be the result of past failure of the overburden, and thus represents an equilibrium configuration in which the weathered regolith has established a stable angle. In the context used here, "bevelled" applies to the area of low slope developed specifically in the weathered soil material at the top of the cliff; however, the terminology is consistent with that used by YOUNG (1972) and TRENHAILE (1987).

At the base of the cliffs, characteristically, a shore platform has developed. Most commonly, this consists of a broad, gently sloping surface extending some 40–200 m seaward of the cliff face (Figure 3a), and ending abruptly in a bioeroded sea cliff (HEALY, 1967a,b, 1968). In some sectors of the shoreline, a higher level platform exists some 2–3 m above the low tide level (Figure 3b). Such high tide platforms, or high level benches (HEALY, 1967a), are generally only a few metres wide.

Numerous beaches occur along the shoreline. Usually, these front re-entrant valleys are truncated by the shoreline. Alternatively, many of the beaches are simply a thin veneer of sand over a pre-existing shore platform.

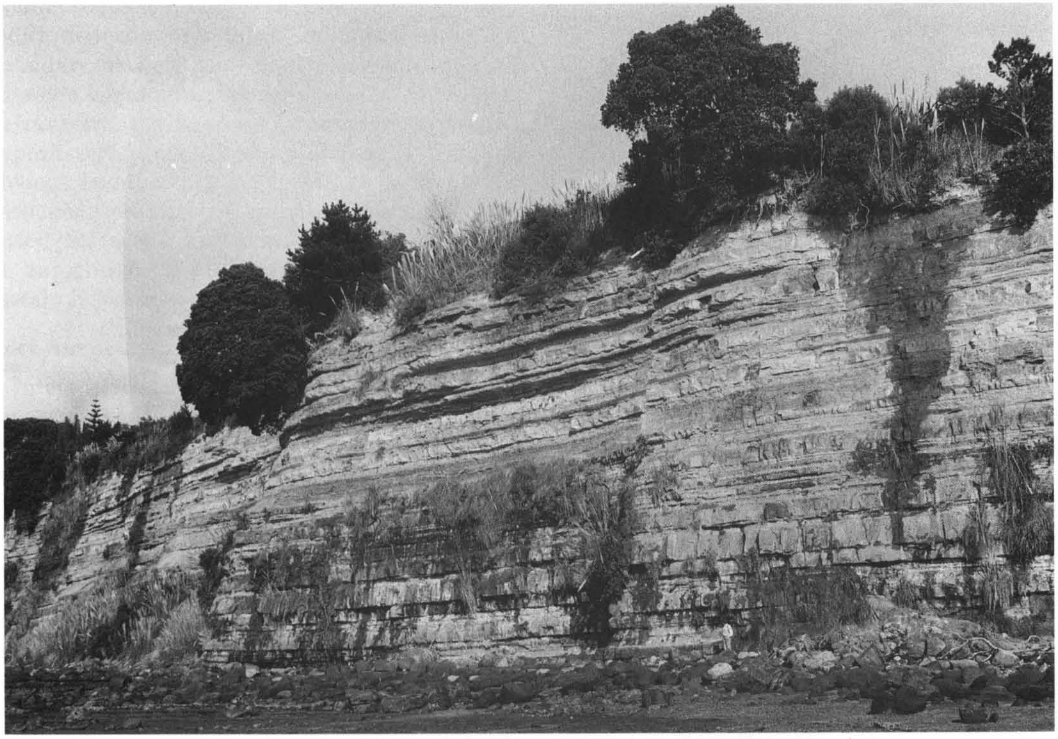
MATERIALS

Geology

The cliffs are composed of Waitemata Group rocks and comprise Miocene turbidite deposits which were laid down in a shallow, enclosed basin (BALLANCE, 1974). The sediment was largely derived from erosion of an andesitic volcanic arc immediately to the west of the basin. As well as the turbidity currents, there were occasional incursions of large mudflows into the basin, resulting in andesitic breccia beds known as Parnell Grit.

The flysch materials are composed of three identifiable units. The primary unit consists of poorly sorted sandstone beds of variable thickness ranging from approximately 0.05 m to 1.50 m. In local areas, this variability is much less, the thickness of individual beds tending to be quite constant over large distances. Intercalated with the sandstones are thinner siltstone beds which have thicknesses varying from 0.01 m to 1.00 m. The Parnell Grit deposits tend to be thick (> 10 m), homogeneous units. The muddy matrix is dominant with up to 50% of angular volcanic fragments of 5–10 mm diameter. The volcaniclastic component of the rocks is high, resulting in a significant proportion of active, swelling clays (smectites and illite) in all of the rocks (DE LA MARE, 1992).

As a general rule, the flysch deposits are essentially flat-lying, with an average dip of approximately 3–5° to the east. However, penecontemporaneous intraformational slumping was common in the depositional basin, resulting in extensive folding of the beds around many parts of the Auckland coastline (HEALY, 1967b). In these areas, the dip of the deposits is variable, reaching 90° in many instances.



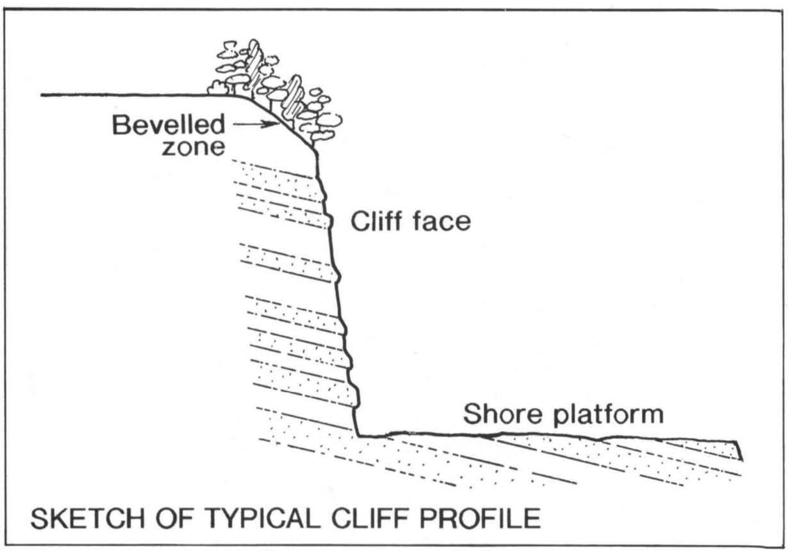


Figure 2. Steep cliffs up to 30 m in height are typical of the region. Alternating sandstone and siltstone layers with gentle dips to seaward make up the bulk of the rock faces. At the top of the profile, there is commonly a "bevelled zone" of lower slope representing the stable angle of the weathered soils.

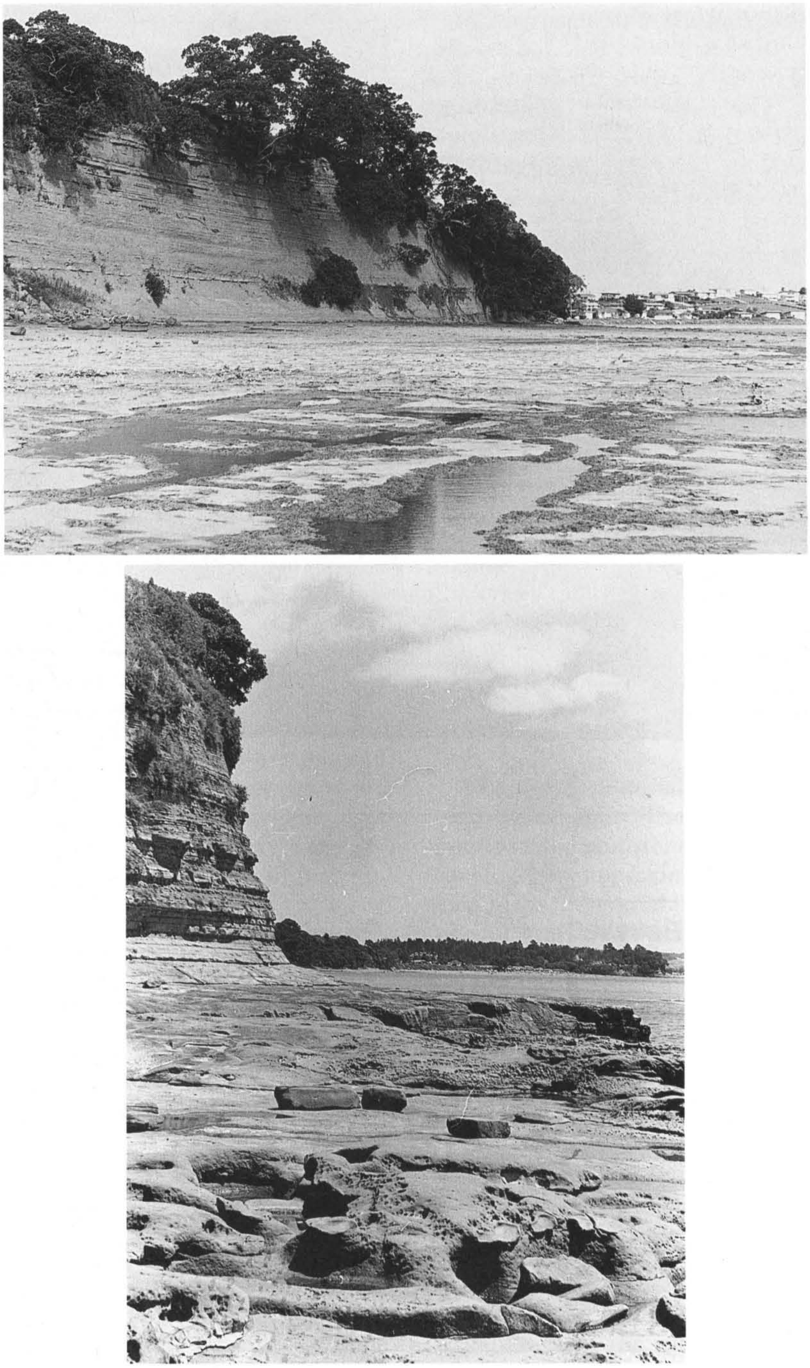


Figure 3. Shore platforms eroded into the flysch deposits: (Top, A) The common intertidal platform, which usually forms a 40–200 m wide shelf. (Bottom, B) The high-tide platform which is typically a narrow bench 2–3 m above the low-tide platform level and planed to the high tide level.

Bedding features are apparent with incomplete Bouma sequences visible in most individual turbidite deposits. Within an individual unit, the boundary between the sandstone and overlying siltstone is gradational over several centimetres. In contrast, the boundaries between successive deposits are generally sharp and marked by an obvious bedding plane. Numerous fault planes with small offsets (< 3 m) slice through the exposed cliff faces. Two general strike directions can be recognised: approximately E and approximately NW.

Rock Strength

Intact strength measurements of the major sandstone and siltstone rock units comprising the cliffs show that the rocks are very weak. The sandstones are all very low strength rocks in the classification of DEERE and MILLER (1966) with unconfined compressive strengths of < 7 MPa (DE LA MARE, 1992); although not readily determined due to intense fracturing of the rocks, the siltstones are believed to have unconfined compressive strengths of < 1 MPa. The Parnell Grit is somewhat stronger with an unconfined compressive strength of approximately 12 MPa (DE LA MARE, 1992)—a "weak" rock in the classification of DEERE and MILLER (1966). All of these rocks thus fall into the category of "soft" rocks—materials transitional between rocks and soils with uniaxial compressive strengths approximately in the range 0.5–25 MPa (JOHNSTON, 1993).

However, the rock mass characteristics of the three units are quite distinctive. The sandstone beds are dissected by sets of regularly spaced joints. Typically, they have spacings of 0.5–1.0 m (DE LA MARE, 1992), and occur in two intersecting conjugate sets though the sets differ between individual sandstone beds. The joints are almost always continuous through the thickness of the bed, and hence, in conjunction with the bedding planes, break the sandstones into regularly shaped blocks. In contrast, the Parnell Grit is relatively unjointed, with widely spaced vertical joints running through the deposit, but usually only in one orientation. The units thus appear as massive bodies of homogeneous rock.

The siltstone beds have a frittered and spalled appearance in outcrop—the entire surface of the material is broken into small, curved fragments less than 15 mm in length. This is the result of the high proportion of swelling clays, which undergo a considerable volume change on wetting and drying, together with stress release following uplift of the materials. Repeated cycles of wetting and drying cause complete fragmentation of the rock (slaking), a common occurrence in mudrocks (BELL and PETTINGA, 1988; DICK and SHAKOOR, 1992). Such fragmentation probably only extends a short distance into the rock body, being the distance to which moisture content changes caused by surface water penetration and evaporation-driven capillary action are effective.

Table 1. *Index properties for two main soil types (after McLeod, 1988).*

Soil On	Clay Content (%)	Plastic Limit (%)	Liquid Limit (%)	Plasticity Index (%)	Activity
Flysch	51	50	75	25	0.49
Parnell Grit	34	47	67	20	0.59

Soils

Two broad soil types can be identified along the cliff tops, corresponding to soils formed on the flysch material and soils formed on the Parnell Grit. The soil formed on the flysch deposits is a very firm, yellow, clayey soil with a well-developed prismatic structure. Kaolinite is volumetrically the dominant clay mineral type with significant secondary amounts of smectite and vermiculite (MCLEOD, 1988). On the Parnell Grit, the exposed soil has a gritty texture; it is a firm, silty clay, with a blocky structure and numerous thin (< 0.01 m) ironpan lenses in the profile. Kaolinite and smectite clays are both abundant in this soil (MCLEOD, 1988).

Index data for these two soil types are presented in Table 1: all values are from a standard depth of 2 m and represent averages of several measurements (MCLEOD, 1988). The Parnell Grit soil contains a considerably lower clay percentage than the soil formed on the flysch deposits, accounting for its much grittier texture and more open structure in the field. However, the Atterberg limits for the two soils are very similar, reflecting the similar overall composition of the materials. Both soils are of normal activity as expected for materials dominated by kaolinite. The activity of the Parnell Grit soil is slightly higher, indicating a somewhat greater capacity for shrinking and swelling in response to moisture content variations, than that of the flysch soils.

MECHANISMS OF CLIFF RETREAT

TRENHAILE (1987) sees mass movement as being a dominant process in the evolution and maintenance of the form of marine cliffs, where mass movement is defined as the movement of material downslope in response to gravitational stress (HANSEN, 1984; TRENHAILE, 1987). Thus, the erosion of coastal cliffs depends not only on the processes of cliff attack, but also upon the composition and structure of the cliffs, which influence the nature of the mass movement (KOMAR and SHIH, 1993).

It is commonly recognised that in hard, unweathered rock, the stable height and angle of a slope are determined by the discontinuities in the rock mass, rather than by the intact strength of the rock (TERZAGHI, 1962). Types of failure commonly encountered in discontinuous rock masses include free falls or topples, and slides (HOEK and BRAY, 1977; ZARUBA and MENCL, 1982; WHALLEY, 1984; TRENHAILE, 1987). Falls involve the detachment and fall of surficial material from steep rock faces; they may occur following the removal of underlying support for a jointed block (notches cut by wave action, for example (TRENHAILE, 1987)) or by toppling, which involves rotation of blocks of rock about a fixed base (HOEK and BRAY, 1977; YOUNG and YOUNG, 1992). In contrast, sliding failures involve the sliding of blocks of material along discontinuity planes, and may include planar failures along dipping discontinuity planes and wedge failures along intersecting discontinuity planes (HOEK and BRAY, 1977). Planar failure occurs when a discontinuity surface strikes parallel ($\pm$ 20°) to the strike of the face, whereas the more general wedge failures occur when the plane of the discontinuities strikes across the plane of the face, and sliding occurs along the line of intersection of two such discontinuities (HOEK and BRAY, 1977). Conditions for planar and wedge failures require that the dip of the failure plane (or line of intersection) is greater than the angle of sliding friction of discontinuity surfaces, yet less than the slope angle of the face (the plane "daylights" from the slope) (TERZAGHI, 1962; HOEK and BRAY, 1977).

TERZAGHI (1962) presents an analysis of the stable angles of slopes on jointed rock masses, as summarised for the coastal environment by YOUNG (1972) and TRENHAILE (1987). TERZAGHI (1962) considers the specific case of stratified sedimentary rocks with continuous bedding planes and cross-joints which are essentially perpendicular to the bedding planes (the common situation in Auckland). When the bedding planes are horizontal, no sliding (planar failure) can occur, and the theoretical stable slope angle is vertical; this is reflected by the typically steep Auckland cliffs. As the bedding angle increases, the critical slope angle becomes a function of the bedding plane orientation and the spacing of the cross-joints. This simple situation may be complicated by the presence of faults within the rocks, and water along discontinuity surfaces.

Three primary mechanisms of cliff retreat are recognised in the flysch deposits of the Auckland region:

(1) fall of jointed blocks (referred to as "joint block fall")
(2) planar failure along fault planes ("fault plane failure")
(3) planar failure along bedding planes ("bedding plane failure")

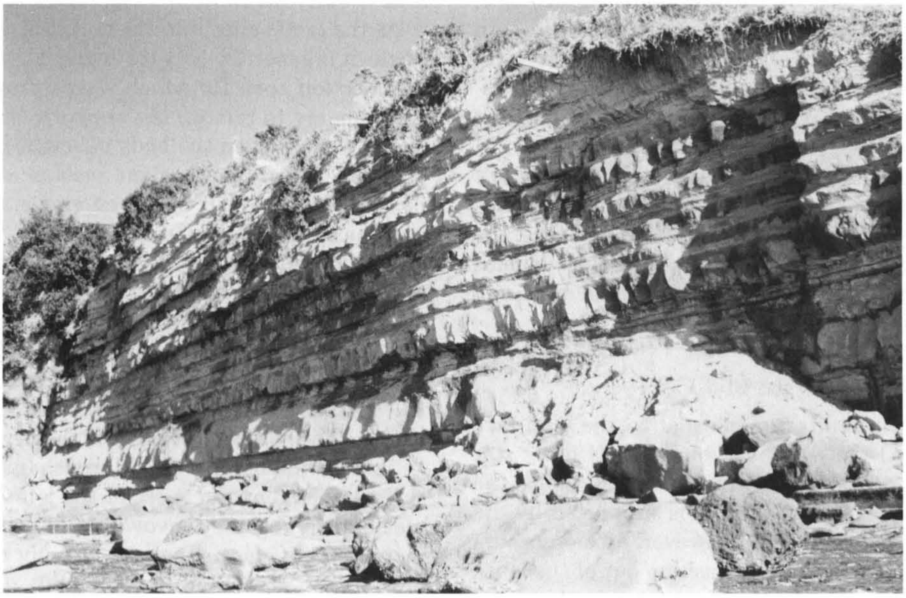
In addition, account must be taken of the failure of folded strata and of the soils overlying the rock faces. These five processes will be considered in turn.

Joint Block Fall

Collapse of jointed sandstone blocks is the most common form of instability of these cliffs. Figure 4 shows a typical stretch of coastal cliffs which is actively retreating by this mechanism. The actual failure itself is the result of simple gravitational collapse of the regularly shaped joint blocks described previously. Initially, the sandstone beds are supported by the underlying siltstone. However, the tendency of the siltstone to fragment by slaking results in loose, frittered material which is easily removed by surface water, waves, and even wind (reports of blowing silt are common from householders on the cliff tops). Once the sandstone blocks are undercut, there is very little support for them, as the joint planes tend to be continuous and open. The blocks can, therefore, fall readily from the face. This process is exacerbated by water flowing down the joint planes, as shown by wet surfaces and extensive ironstaining on newly exposed joint faces.

Fault Plane Failure

Fault planes provide weak zones on the cliff face along which instability can occur. Generally, the faults in this area have a narrow (< 0.1 m wide) zone of gouge material lining the fault plane. As



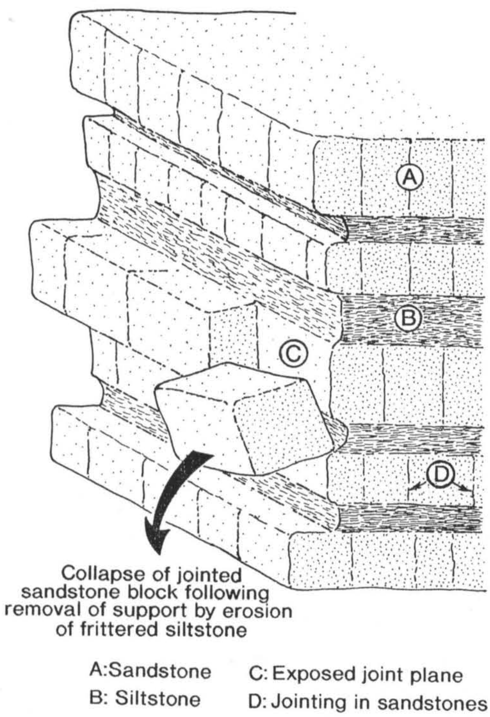


Figure 4. Stretch of cliffs actively eroding by falling joint blocks. Note the tendency for the siltstone beds to be preferentially eroded leaving the sandstone layers protruding. Continuous vertical joints through the sandstones break this material into regular blocks. Loss of support for these blocks results in simple gravitational collapse. The diagram shows the simplest case of a single displaced block; commonly a number of blocks will fall together, removing the support for a sizeable body of material above.

it is wide enough to separate the rock surfaces on either side of the fault plane, the shear strength along the fault becomes equal to the shear strength of the gouge. However, this crushed rock has minimal frictional strength and, hence, little resistance to failure. Fault planes, therefore, are regions of very low shear strength.

Two approximately perpendicular sets of fault planes occur in these cliffs: both are steeply dipping, one runs almost normal to the cliff face, the other sub-parallel to the face. The mechanisms for failure along these two directions are quite different, and treated separately.

Faults Normal to the Cliff Face

When the faults run normal to the cliff face (Figure 5A) the situation is inherently quite stable, with no real capacity for sliding along the fault plane as the weak zone is supported by the rock mass on each side. However, erosion occurs by means of preferential washing out of the loose material along the fault plane with some localised joint block fall along the walls of the fault as support for the sandstone blocks is removed. This results in a narrow gully developing along the fault plane (Figure 6), similar to the erosion of weak, vertically dipping rocks described by TRENHAILE (1987). Unfortunately, associated soil failure above such natural gullies tends to produce a topographic feature which is seen as a prime site for disposal of stormwater. This enhances the gully erosion significantly, as it is mainly flowing surface water which is the cause of removal of the fault gouge initially.

Faults Sub-Parallel to the Cliff Face

When the fault planes run approximately parallel to the cliff face (Figure 5B), the situation is much less stable. In particular, the cliffs are susceptible to planar failure along the fault plane. TERZAGHI (1962) describes such a situation where the fault intersects the cliff face at an acute angle, and shows that the wedge-shaped body of rock between the slope and the fault may drop out. As many of the faults in the Auckland area are nearly vertical, they do not daylight from the slope, so the cliffs are buttressed by the rock at the base of the cliff. Unfortunately, in the coastal environment, waves provide a ready mechanism for transporting away loose material at the base of the slope, thus removing support for the overlying rock mass. As this support is lost, planar failure of the material seaward of the fault plane becomes a possibility, as shown in Figure 7.

As the faults slice into the cliff face at an acute angle, it is generally only the material in the acute intersection zone for which waves provide sufficient energy to remove the support. Observation suggests that when the body of rock between the cliff face and the fault plane reaches a thickness of approximately 10 m, it remains stable as its own mass provides sufficient overburden stress to maintain enough friction along the faulted contact. Thus, the planar failure is generally limited to a length of 30–40 m along the cliff face at any one time. Over time, however, as the cliffs steadily retreat by joint block fall (as described above), the instability progresses along the line of the cliffs, exposing the fault plane as the new cliff face.

An individual event by this mechanism is considerably larger than those of the joint block fall mechanism and may involve up to some tens of metres of cliff face. However, usually the events are somewhat smaller, and the cliffs degrade by a series of successive failures on the fault plane. Such a failure is the most common form of large-scale failure of the faces. Again, failure of the soils above the unstable zone produces a ready area for stormwater disposal, and the problem is significantly exacerbated by the weakening and removal of fault gouge material by flowing water.

Bedding Plane Failure

As the predominant dip of the bedding planes is seaward (out of the cliff face), the capacity exists for large-scale planar failure along these planes. Such a mechanism is believed to cause the largest scale failure of these cliffs, but geomorphic evidence suggests that these events are relatively uncommon.

The mechanism for planar failure along the bedding planes is shown in Figure 8: large volumes of material lying above an unstable bedding plane seem to be rafted seaward out of the face. Fortunately, the dip angle of the bedding is generally very low, so the frictional resistance across the bedding planes is sufficient to maintain stability under normal circumstances. However, water entering these bedding plane surfaces would evidently reduce the stability immensely. Water along joint planes has the capacity to develop a cleft water pressure (TERZAGHI, 1962; WHALLEY, 1984) which effectively separates the two surfaces, and, hence, reduces the frictional resistance between them. The overlying material will thus fail when

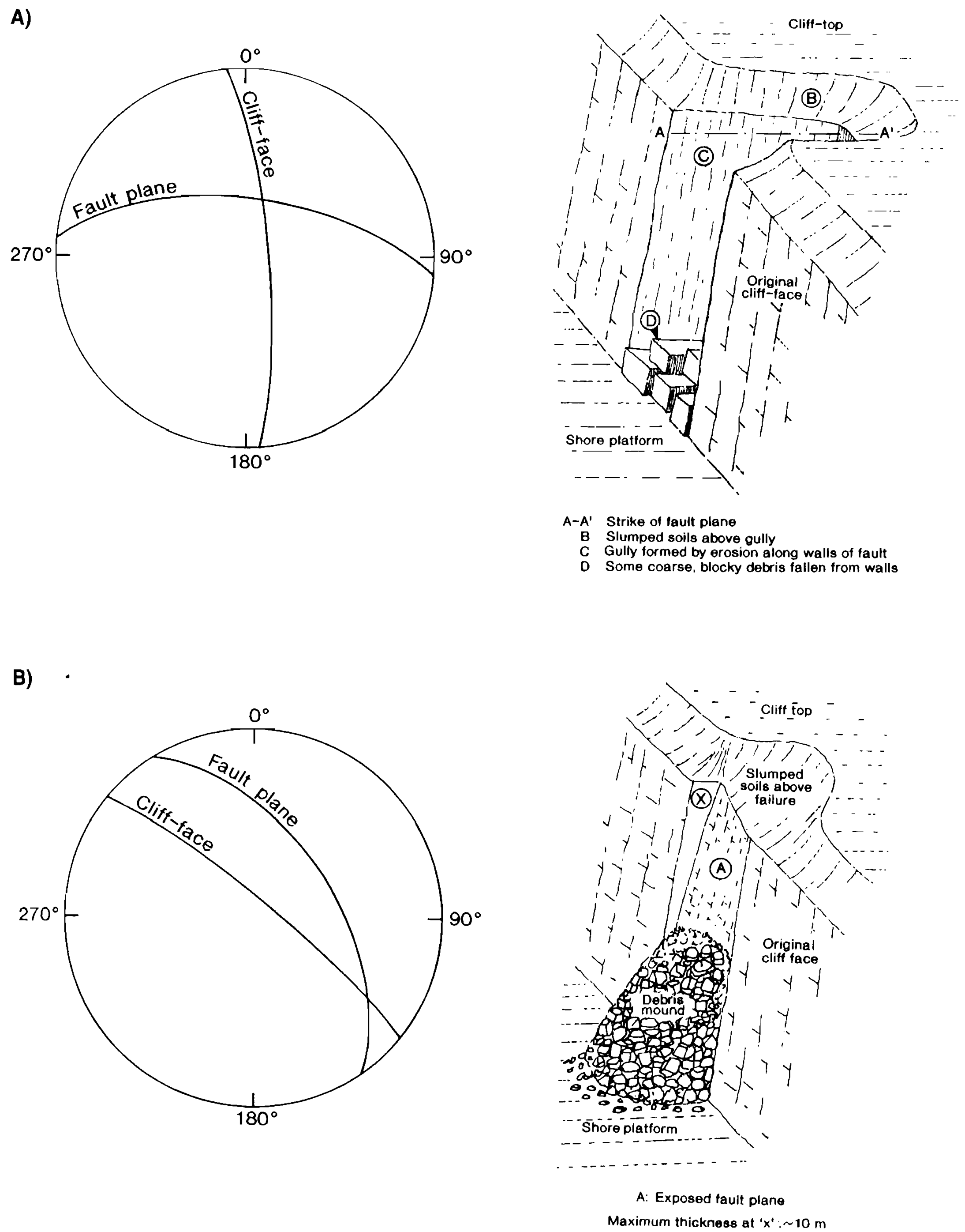


Figure 5. (A) Stereonet analysis showing the fault planes striking normal to the strike of the cliffs. In this situation a narrow gully is incised along the fault plane. (B) Stereonet analysis of fault planes striking sub-parallel to the cliff face. In this case, the capacity exists for planar failure of the material in the acute intersection zone between the fault and the cliff face.

Figure 6. Gully development along a small fault striking normal to the cliff face. Note the recent extension of the stormwater discharge pipe to the base of the cliffs—in 1983 this discharge occurred at the cliff top, a significant cause of enhanced erosion along this plane.

Figure 7. Planar failure along fault planes striking sub-parallel to the cliff face. The recent mass movement has been along the fault in the foreground, but similar movements have occurred along parallel faults slightly seaward of this one in the past. These are now forming the cliff face over short lengths of the cliff, but each one has only eroded to a depth of approximately 10 m before the mass of material seaward of the fault plane is sufficient to maintain stability.

the effective friction angle is less than the dip of the beds.

Large failures of this type have demonstrably occurred in the East Coast Bays area in at least three instances. The largest one (Figure 9) has clearly occurred since the rise in sea level following the last glacial, as the stream discharging from the slumped area has not incised through the present shore platform. This failure covers a basal failure plane area of some 500 m^2, and the total area directly affected by the slump is some 900 m^2. Instability of the slopes around this failure is still very evident, with cracking and slumping of the soils, disruption to house foundations, and movement of survey pegs being recorded.

Due to the strong influence of cleft water pressures on such failures, the control of water entering the strata near the cliff face is essential, and this does not apply to just the immediate area of the cliffs, but for some distance back. For some individual dwelling developments, drainage measures have been undertaken to ensure that water along bedding surfaces can drain freely and, hence, not develop cleft water pressures. The insertion of such drains has resulted in the escape of considerable quantities of water from the face, and has been notable for a long relaxation time for the flow.

Failure of Folded Strata

Folded strata fail by the same mechanisms as described above, but the significance of each mechanism depends upon the inclination of the individual beds. Hence, where the bedding planes are tilted so as to be dipping steeply seaward, the propensity for bedding plane failure becomes very great—the amount of material affected by an individual event is probably quite small, however.

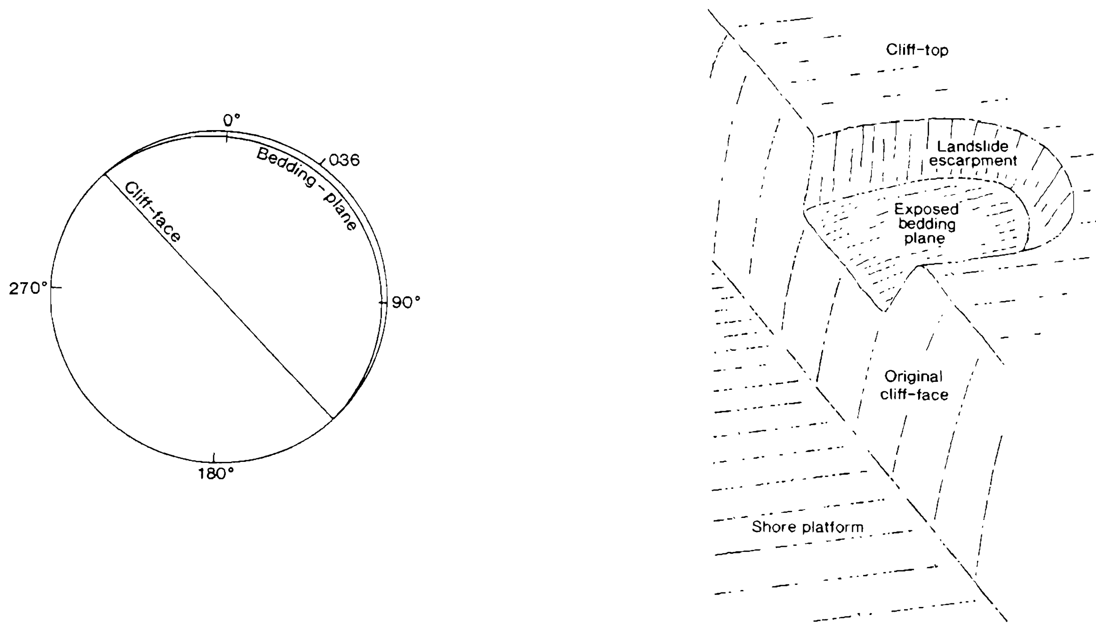


Figure 8. Stereonet analysis of the bedding plane failure mechanism at Churchill Reserve in the East Coast Bays area (see Figure 9). A very shallow seaward dip on the beds is normally sufficient to maintain stability, but loss of frictional resistance along this plane due to high cleft water pressures may result in a very large mass of material becoming unstable.

In contrast, when the beds strike normal to the cliff face, which often occurs with the steeply dipping strata, the tendency exists for failure similar to the type described for faults normal to the face, as described by TRENHAILE (1987). In this case, erosion of the siltstone beds results in sandstone layers which stand proud from the face. These fail by joint block fall in the same way as described for horizontal strata.

Failure of Overlying Soils

Most of the immediate hazard to dwellings on the cliff face comes not from failure of the rocks below, but from instability of the soils in which they are founded. Any rock failure which results in cliff retreat will lead to oversteepening of the soils, and, hence, must ultimately be matched by failure of the soils to maintain a stable angle. This stable angle is reflected by the "bevelled zone" noted in the cliff morphology; where such a zone is absent, it can be assumed that the soils are in a potentially unstable situation.

Due to the different clay proportions and structure of the various soil types, the mechanisms of failure in the soils developed on the flysch and Parnell Grit appear quite different. The flysch soils characteristically show rotational type failures, with deep, curved failure planes. In contrast, the soils derived from the Parnell Grit appear to form translational failures, which have steep, straight failure planes, but only extend to a shallow depth and hence involve only small volumes of material.

Geomorphically, the Parnell Grit soils appear far more prone to mass movement than those formed on the flysch material. This is attributed to the more open structure and slightly higher activity of the Parnell Grit soils. Individual failures are comparatively small, but occur very often, resulting in continual retreat of the upper cliff zones. Extensive revetment wall construction has been necessary in one area of Parnell Grit soils in order to attempt to maintain stability on these steep failures (Figure 10).

The other type of soil movement is slow, continual creep of the soil body. This is a particular problem in the flysch soils which contain a very high clay content. Such creep is a problem for all of the region, but particularly on the cliffs where the lateral support for the soils is lost.

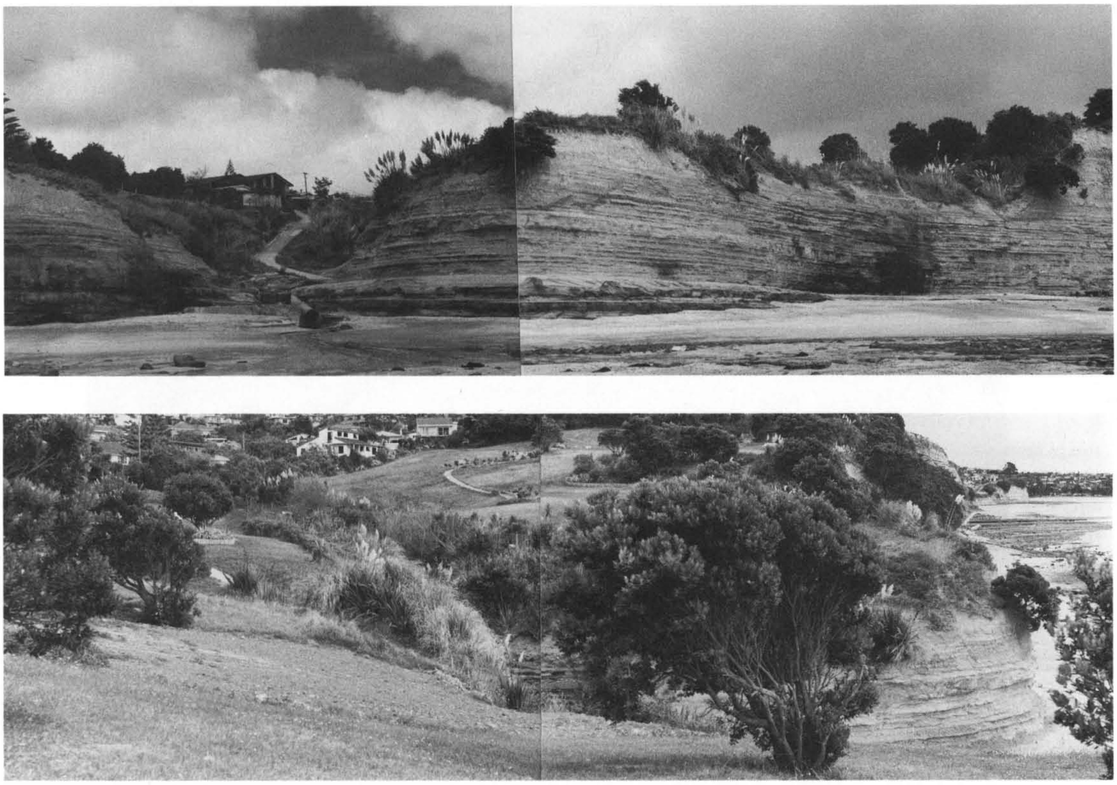


Figure 9. The Churchill Reserve, East Coast Bays, landslide. (Top, A) The sudden lowering of the cliff height at the right of the photograph represents the boundary of the landslide; the boundary on the left is beyond the edge of the photograph. Above the lowered patch of cliffs a thickness of some 10–15 m of material has been rafted off along a gently dipping bedding plane. The stream, which has been piped for a short distance, has not incised a course through the shore platform, suggesting that this landslide occurred after the post-glacial sea level rise. (Bottom, B) The grassed area follows approximately the base of the failure and is surrounded by a typical scarp face on which the houses are built. Evidence of continued movement exists in this area, particularly the hummocky nature of the grassed area, disruption to house foundations, and movement of survey pegs.

RATES OF CLIFF RETREAT

Some attempts have been made to estimate rates of cliff retreat around the Auckland coastline, the most recent being that of Brodnax (1991). Mean erosion rates derived from aerial photograph analysis suggest retreat rates of 11–18 m per century (Brodnax, 1991). However, this is in conflict with quoted values for measurements of the actual retreat around clearly dated anthropogenic structures, which give values of 2–6 m per century (Brodnax, 1991). This latter range is more in keeping with the observations of the authors; the discrepancy between the two sets of values may in part be due to the short time span covered by the aerial photographs, and in part to the method employed which assumes an 8–10 m per century erosion rate for areas where there is no discernible change in the aerial photographs. Average rates of 2–6 m per century will be assumed for this paper.

However, average rates take little account of spatial variations in cliff retreat rates due to mass movement. No rigorous attempt has been made to estimate these, so the values suggested here are estimates based on observations over a 10–15 year time frame. Although not rigourous, they represent a best estimate of the likely rates of cliff retreat. Each of the mechanisms outlined above will be considered in turn.

Joint Block Fall

Being the most common mechanism of cliff retreat, joint block fall is probably best represented

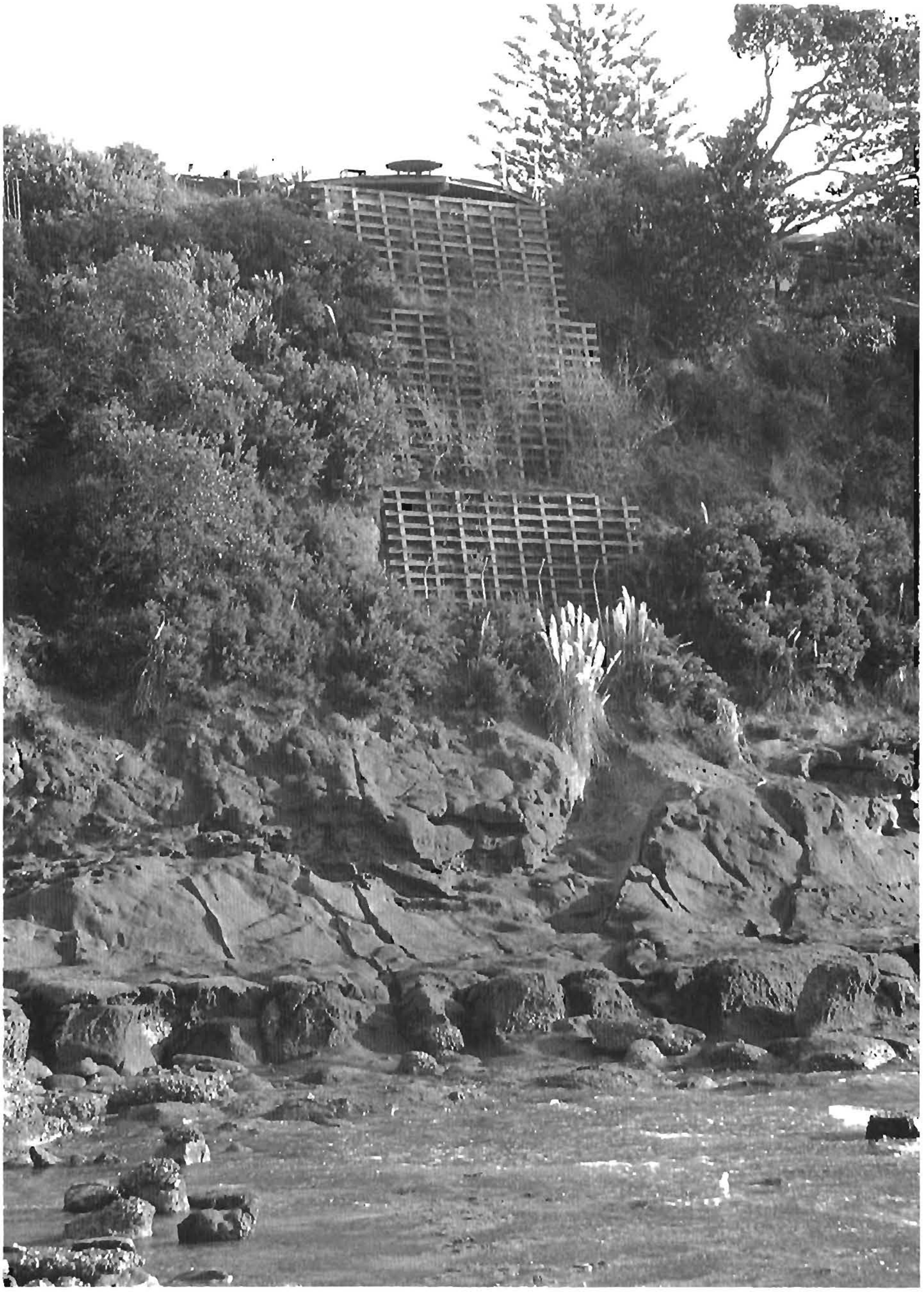


Figure 10. Translational failures in the Parnell Grit soils result in very steep (60–70°), unstable faces which undergo frequent, shallow failures. Extensive revetment wall construction has occurred in this area in an attempt to stabilise such failures.

by the average retreat rates for the area (*viz.*, in the order of 2–6 m per century), and is certainly the mechanism responsible for the measured rates around the structures considered by BRODNAX (1991). For most of the cliffs, rates of 2–4 m per century are probably applicable, with higher rates of retreat in localised areas where a very high proportion of siltstone in the profile hastens retreat by this mechanism. These areas are often readily recognised by the lack of a "bevelled zone" in the overburden, which indicates rapid retreat of the underlying cliffs.

Fault Plane Failure

Faults Normal to the Cliff Face

Fault planes aligned normal to the cliff face erode a gully into the face along the strike of the fault. The width of this gully and the rate at which it extends back into the cliff are strongly dependent upon the nature of (a) the fault itself, and (b) the strata surrounding the fault plane. However, generally this erosion is not great, suggesting rates only slightly higher than the average for the cliffs, in the order of 6–10 m per century.

Faults Sub-Parallel to the Cliff Face

Where the faults strike approximately parallel to the cliff faces, the erosion tends to occur at rates much greater than the average. Again, the actual retreat rates depend upon a number of properties of the individual fault, particularly the inclination of the fault plane, its strike in comparison to that of the cliffs, and the frictional resistance along the fault plane. However, in general, a retreat of at least 3–5 m may reasonably be expected in any single mass movement event, and observations suggest that it is reasonable to expect an event in any 10 year period.

Rates of 30–50 m per century could thus be derived for these areas. However, this is unreasonable, as once the failure has occurred the instability progresses along the strike of the fault, and, hence, along the cliff face until a maximum depth of approximately 10 m is reached. Thus, individual events resulting in 5 m of retreat may be expected within any 10 year period with maximum retreat distances of 10 m along any single fault plane.

Bedding Plane Failure

As failure along bedding planes is uncommon, its significance in terms of retreat rates is difficult to assess. The only known example for which an indication of the time frame involved can be inferred simply implies that failures of the order of 200 m may be expected within 10,000 years.

Failure of Folded Strata

Retreat rates for folded strata depend very strongly on the exact nature of the geological structure. Generalisations cannot readily be drawn, but it is probably reasonable in most instances to assume that the rates of cliff retreat are significantly greater than for the joint block fall mechanism operating on horizontal beds, but less than for the planar failures along fault planes.

Failure of Overlying Soils

Rates of retreat of the soils are influenced by a variety of factors: moisture content, vegetation patterns, and land-use practices in particular. If the soils are assumed to be in a stable configuration initially, then any retreat of the cliffs will be reflected in a subsequent retreat of the soils to maintain stability. Thus, the rates suggested for the cliffs can be taken as minimum retreat rates for the soils. However, changing land-use patterns may well result in a brief (decades to century order) period of comparatively rapid soil retreat to establish a new equilibrium. This is illustrated by numerous examples of soil failure induced by channelling of household stormwater drainage into the soils near the cliff top.

Again, little account can be taken of the spatial variations in soil retreat rates due to mass movement; and, indeed, simple rates of retreat may not be as pertinent as the frequency of mass movement events in terms of hazard analysis. Although the Parnell Grit deposits are characterised by low erosion rates compared with the flysch rocks, the translational failures in the soils, although individually very small, are frequent events. As such they pose a significant threat to structures on the cliff top. In contrast, the rotational failures occurring in the soils formed on the flysch rocks are larger events and occur much less frequently. The overall threat to structures over a short time frame is thus considerably less.

RISKS ASSOCIATED WITH CLIFF FAILURE

Each of the mechanisms of cliff retreat discussed above poses a risk (as defined by VARNES, 1984) to the users of the cliff area. However, each mechanism represents a different level of risk based on the likely magnitude and frequency of the event.

Joint block fall is of little relevance to homes on the cliff top over a relatively short time span. As most homes are naturally built some metres from the face, collapse of jointed blocks provides little immediate threat. Localised areas of more rapid retreat may become a significant problem within one lifetime.

Likewise, the failure along bedding planes probably represents only a minimal risk, in that the events are very rare, with individual events likely to be separated by a time span of centuries. Although the property value concerned for the area involved may be considerable, the risk to human life is relatively small, as the mechanism is such that some forewarning of an impending failure is likely.

Failure along fault planes probably represents the most significant cause of risk to overlying property though the risk does not directly arise from the failure events along the fault plane. Rather, it is the failure of the soils following oversteepening by the cliff collapse which causes most concern to homeowners. Above an actively eroding fault plane, the soil profile must continually respond to the cliff retreat by successive failures to maintain a stable angle. This mechanism has resulted in the undermining of foundations, generally of decks and swimming pools, but also of houses themselves on occasion. Recognition of areas where this mechanism is operative is largely based on the recognition of the fault planes in the cliff below. As there are very many such faults in the Auckland region, the area affected by this hazard is considerable, and the frequency of events is such that the risk becomes significant.

Likewise, failure of the soils above the Parnell Grit is of major concern to the homeowners in these regions, but they are restricted areally, so the overall hazard is quite small. In contrast, soil creep affects the entire region, but the damage done is small and progressive, and does not represent a major concern in general. However, with much development having occurred in a comparatively short time span (15–20 years), the effects of soil creep in the future may become very apparent quite suddenly as many dwellings are affected at one time.

Joint block fall poses the major hazard to people using the base of the cliffs due to the frequency with which blocks are dislodged. Despite this, sensible precautions mean that the risk from this form of failure is minimal.

DELINEATION OF HAZARD ZONES

Two approaches may be considered when attempting to define a coastal hazard zone for this area: either consider each mechanism in turn and define a specific hazard zone associated with each, or take a general zone for the entire area based upon the mechanisms which represent the greatest risk over an appropriate time frame. Although the first approach is most rigourous and will produce the most appropriate hazard zones, it requires detailed mapping of the cliff geology, soils, and geomorphology. As such maps do not exist for most of the Auckland region, a more practical approach is provided by the latter method.

As a convenient human time frame, 100 years will be used following the recommendation of HEALY (1981) and GIBB and ABURN (1986), as this is likely to equal or exceed the life expectancy of most cliff-top structures. Over this time frame, failure along fault planes poses the greatest risk, for which we have retreats of 5 m expected in any 10 year period with a likely maximum retreat at any single point of about 10 m. Obviously, a 10 m retreat may be expected within the 100 year time frame, so this is a sensible initial hazard zone width.

As the instability progresses along the cliff face, the fault plane becomes the new cliff. Joint block fall then becomes the mechanism for eroding into this face until sufficient thickness of rock is removed to make the next fault plane unstable. In general, the time frame required for this is probably greater than 100 years for the bulk of the cliff region.

However, it is the soils which must ultimately respond to the cliff retreat. As discussed previously, they will do this at a minimum rate equal to the rate of rock retreat and at a possibly higher rate as land-use practices vary. Hence, any hazard zone must be taken as a set-back from the point at which the soils rest at their stable angle rather than the cliff face. This, as a general rule for the flysch soils, is approximately 6 m back from the rock face.

Hence, a sensible **minimum** coastal hazard zone is of the order of 16 m from the cliff face. Obviously, this may be too small in areas of complex structure such as numerous intersecting faults or folded strata, but provides a sensible average for the region in general. Note that although this zone was derived using a 100 year time span, due to the episodic nature of the failures the same zone would be derived for a much shorter time, say 50 years.

GIBB and ABURN (1986) suggest the addition of a "safety factor" (F) to the hazard zone width which is equal to two-thirds of the width derived from a 100 year retreat rate. For this analysis, taking 10 m as the 100 year retreat, this method would involve adding a further 7 m to the hazard zone width, giving a total hazard zone of 23 m. This approach should take sufficient account of most failure mechanisms (except for bedding plane failures) in even the most complexly folded strata.

CONCLUSION—MITIGATION OF HAZARDS

(1) Cliffs carved within soft flysch deposits such as occur around the Auckland coasts are dynamic features undergoing continual erosion and will thus always pose a hazard to people using the cliff regions for either housing or recreation.

(2) Recognition of the mechanisms of cliff retreat is essential for assessing the magnitude of the risk at any one site, and the extent of the hazard. Thus, mitigation of the risks posed is probably best viewed as avoidance of areas where the risk becomes too great. For this reason, a **minimum** coastal hazard zone (where the development of permanent structures for dwelling or commercial use is restricted) of 16 m is recommended.

(3) Unfortunately, it is probably too late for such avoidance practises along most of the cliffs as development has already proceeded. Under these circumstances, it must be recognised that the geological makeup of the cliffs is such that erosion will continue, and little real influence can be brought to bear on this.

(4) The one factor which is common to the various mechanisms of cliff retreat is that they are all exacerbated by water, which either induces increased cleft water pressures along joint and bedding planes, washes soft material out of fault planes, or causes subsurface piping along the soil-rock interface. Control of surface water is, therefore, of paramount significance for any attempt to slow or control the cliff erosion. This aspect has not been recognised historically, and zones of slumped soil (which frequently represent an underlying instability in the rocks) are used for stormwater disposal as they provide a low point in the topography. Changing this practice is one practical method which can be readily applied to increase overall cliff stability.

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