Mass Movements in Great Britain

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Chapter 7

Mass-movement sites in Cretaceous strata

R.G. Cooper

INTRODUCTION

Considering their large outcrop (i.e. 4% greater in area than that of the Carboniferous strata) the Cretaceous strata of Britain have a relatively low density of landslides (38% of the number in Carboniferous strata). Of these, 58% are recorded in the national landslide survey as of unspecified type (Jones and Lee, 1994). Of those specified, 34% are complex and 19% are rockfalls. The Chalk, which is by far the largest formation considered areally (71% of the total Cretaceous outcrop) has a lower number of landslides (less than half) than each of the Upper Greensand and Gault, Upper Greensand, Weald Clay and Hastings Beds, each of which has less than one tenth of the area of the Chalk (Jones and Lee, 1994). Of these other formations, the Upper Greensand and Gault lead, with 31% of the landslips in the British Cretaceous outcrop. Unfortunately the survey does not provide statistics for these two formations separately. However, while 58% of the 273 slides identified in the Upper Greensand and Gault were of unspecified type, 49% were described as complex, and 23% were multiple rotational.

Two sites in Cretaceous strata have been selected (Figure 7.1). The first, **Folkestone Warren**, is one of the most intensively studied landslides in Great Britain. It is described in international reviews of landsliding and mass-movement processes (e.g. Zaruba and Mencl, 1969; Selby, 1982) and may therefore be claimed as a site of international significance for its mass-movement interest. The second site, **Stutfall Castle**, has two principal points of interest: it is on an abandoned marine cliff, and represents the types and sequence of mass movements characteristic of the degradation of such a cliff after removal of (marine) cliff-foot erosion. Secondly, it illustrates the way in which geotechnical understanding can, in certain circumstances, be enhanced by archaeological investigation.

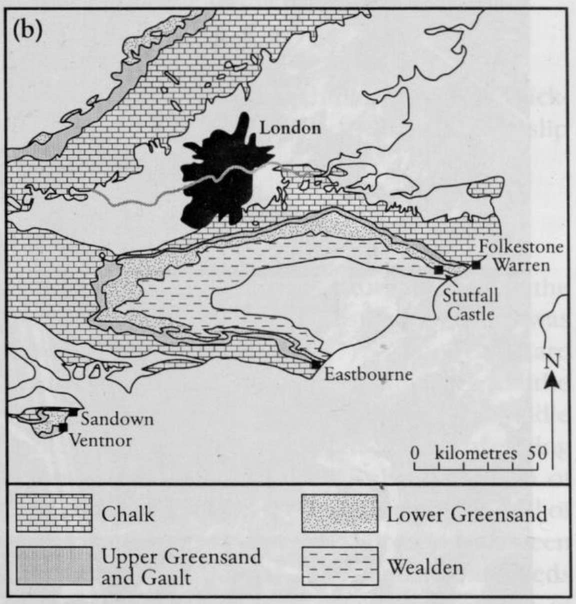


Figure 7.1 ◄(a) Areas of Cretaceous strata (shaded) and the locations of the GCR sites described in the present chapter. ▲(b) The Cretaceous strata of southern England showing the locations of the GCR sites described – **Folkestone Warren** and **Stutfall Castle**. After Hutchinson *et al.* (1980).

FOLKESTONE WARREN, KENT (TR 243 375–TR 268 385)

Introduction

(a) General

The principal reason for the detail in which Folkestone Warren has been studied has been that since 1844 the 3 km-long unstable area has been traversed by the main railway line from Folkestone to Dover (Figure 7.2). The photographs of the twisted railway tracks, one with a train on them, taken after the great landslip of December 1915, have become justly famous (Figure 7.3). Detailed investigations have been carried out by the Southern Railway and later by British Railways, and more recently by workers from a number of academic institutions and civil engineering consultancies.

The area of landslides is backed by the 'High Cliff', a 30 m-high Chalk cliff standing at about 55°, consisting of a succession of broad, irregularly spaced buttresses. In plan (Figures 7.4 and 7.5), the rear scarp of the Folkestone Warren has a generally arcuate form, concave to seaward, with the degree of concavity increasing towards the west. It is made up chiefly of three en-echelon sets of essentially joint-controlled faces (Hutchinson *et al.*, 1980), which trend at 67°,

Figure 7.2 The landslide complex at Folkestone Warren, Kent, showing the engineering structures. (Photo: Copyright reserved Cambridge University Collection of Air Photographs.)

Figure 7.3 The central part of Folkestone Warren, viewed eastwards from a point about 200 m east of the Warren Halt, shortly after the 1915 slip. (Photo: British Railways, Southern Region.)

47° and 109° (Figure 7.4). These give the rear scarp in detail a rather saw-toothed appearance. Its general form reflects the tendency of the faces at 47° to become increasingly predominant over those at 67° towards the west. The upper part of the cliff is vegetated, but much of its foot was freshened by the movements of 1915. The 300–400 m-wide area of slips between the High Cliff and the coastline is known as the 'Undercliff'. The multiple form of the Warren slips is best seen at the western end, where the rear scarps of slips of the 1930s and 1940s parallel both the associated rear scarp of the 1915 slip and the High Cliff above the ancient slide block of 'Little Switzerland' (Figure 7.6). Elsewhere the slide blocks are generally obscured by a mantle of Chalk debris derived from numerous falls from the High Cliff. There is a break in the talus at the foot of the High Cliff. This was produced by the 1915 slip. It can still be traced except where masked by later Chalk falls. The seaward edge of the Undercliff consists of sea cliffs up to 15 m high, exhibiting a great thickness of Chalk fall debris overlying the slip deposits.

(b) Stratigraphy

The arrangement of the undisturbed strata in the immediate vicinity of Folkestone Warren was described by Osman (1917). The cliffs are formed by the truncation of the scarp of the North Downs by the sea, and consist of Middle and Lower Chalk (Upper Cretaceous), overlying Gault Clay and the Folkestone Beds division of the Lower Greensand (Lower Cretaceous). All of the strata dip at about 1° in a direction between north-east and NNE. The Folkestone Beds formation is about 18 m thick and consists of coarse-grained yellowish greensands with bands of calcareous and glauconitic sandstone (Gallois, 1965). The junction between the Folkestone Beds and the overlying Gault Clay is at the top of the 'Sulphur Band', a bed of phosphatic nodules (Smart *et al.*, 1966).

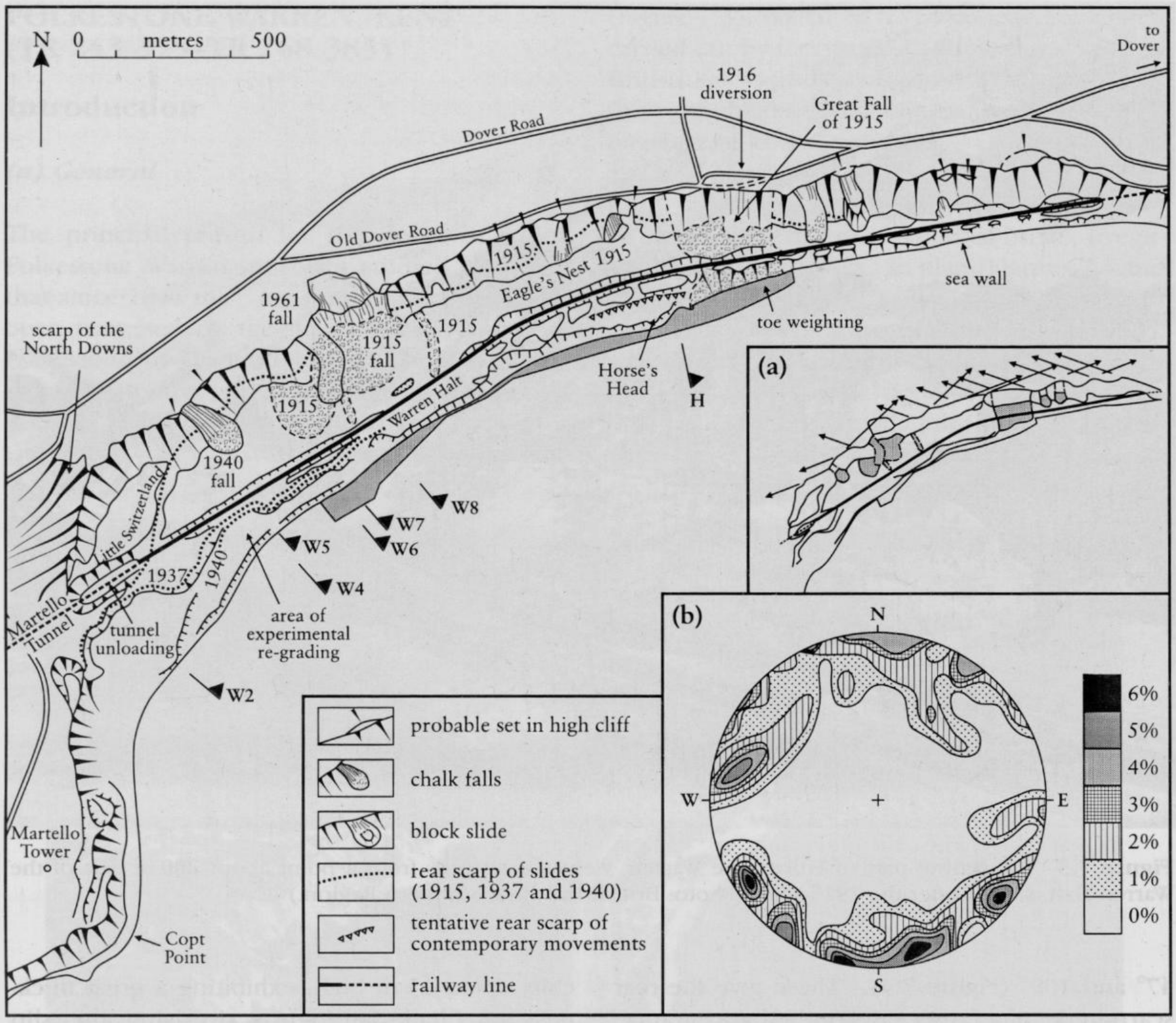


Figure 7.4 Geomorphological map of Folkestone Warren showing dated rockfalls on the near scarp and traces of some of the larger rotational slips. Position of cross-sections W2, W4–8 and G are shown (see Figure 7.5 for the postion of cross-sections W3–4, W6 and W8–9). Inset (a) shows the main scarp trend directions and inset (b) the predominant joint directions. After Hutchinson (1969).

The Gault Clay consists of hard, over-consolidated, fissured and jointed clays. It is between 40 m and 50 m thick. Thirteen lithological subdivisions have been recognized within it (Jukes-Brown, 1900). These possess large variations in physical properties, for example in the liquid limit – in the upper part of the Gault Clay, liquid limit values lie between 80% and 120%, falling to 60%–70% in the middle part, and rising again to 90%–110% in the lower part, with a

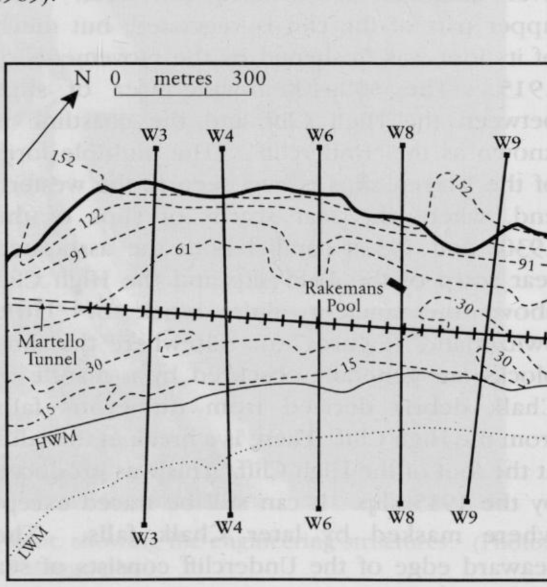


Figure 7.5▸ Part of the 2nd edition Ordnance Survey maps of 1899, showing the contours in the western part of Folkestone Warren and locations of cross-sections W3–4, W6 and W8–9. After Hutchinson (1969).

Figure 7.6 Folkestone Warren viewed from the west, with 'Little Switzerland' in the foreground. (Photo: R.G. Cooper.)

rapid fall to about 70% immediately above the Sulphur Band (Toms, 1953; Muir Wood, 1955a).

The lowest bed of the Lower Chalk is the Chloritic Marl, a relatively impermeable formation about 27 m thick, which can have as much as 50% argillaceous and arenaceous matter (Gallois, 1965). The overlying Grey Chalk, 24 m thick, becomes more pure, massive and blocky upwards into the overlying well-jointed White Chalk, 18 m thick and itself overlain by a 2 m-thick layer of *plenus* marls. The superincumbent Middle Chalk has at its base the Melbourn Rock, a band of nodular, gritty, yellowish-white Chalk about 12 m thick. This passes upward into a fine white Chalk in massive, well-jointed and highly pervious beds. At Folkestone Warren these beds extend up to the level of the presumed Pliocene or Plio–Pleistocene platform (Jones, D.K.C. 1980) which defines the top of the cliffs. Detailed descriptions of the Middle Chalk at Horse's Head, the Gault Clay west of Horse's Head, and Glauconitic Marl and Chalk Marl on the foreshore east of Horse's Head, are provided by Gale (1987).

(c) History of landsliding

The first recorded landslide at Folkestone Warren took place in 1716. Since then failures have recurred at frequent intervals. As noted by Muir Wood (1955b), slipping has been more frequent in the western part of the Warren. Since 1844 there have been 11 known deep-seated landslides. All have taken place within the December to March period, which is when groundwater levels are highest. Moreover, the three largest known failures have occurred in the first third of this period. This suggests that the landslides were brought about by particularly high seasonal peaks of groundwater pressure in the slipped masses. In the case of the 1915 slip at least, there is evidence that it took place, as might be expected, near the time of low tide.

Studies at The Roughs (Brunsden *et al.*, 1996a), 12 km to the west of Folkestone Warren, confirm this. Geotechnical data from investigations at the site have been plotted on every cell of a slope map derived from a digital terrain model (DTM) of the site, to produce a factor of safety map of the Hythe Beds escarpment. Combining this with 5% and 8% perturbations of the current climatic trend extrapolated to the years 2030 and 2050 respectively, shows that many currently dominant areas of the escarpment are vulnerable to a small rise in water levels. Statistical analysis of archive data confirms this conclusion. Since the Weald Clay escarpment is adjacent to Folkestone Warren (the next feature along the coast to the east), it is likely that this conclusion applies there too.

The incidence of Chalk falls appears to be little influenced by the seasonal variations in groundwater level. This no doubt reflects the fact that the falls involve chiefly the body of Chalk situated above the highest position of the groundwater table in the slipped masses and adjacent Chalk. The boreholes of Trenter and Warren (1996) reveal three en-echelon slips that commence north of the railway, trend eastwards and swing south-east to the beach, surfacing on the foreshore at the Warren Point, Horse's Head and Warren East End locations. Measurements made at the toes of each of these slips show that in each case the rate of movement increased eastward to a maximum where the slip crossed the foreshore. On the adjoining landward side, movements were markedly smaller.

There has also been sliding or falling of large masses of Chalk from the High Cliff at the rear of Folkestone Warren. Chalk falls from the rear of the Warren are commonly preceded by slight, chiefly downward, movements known as 'sets'. These affect the Chalk behind the High Cliff for distances of probably up to 20 m, and may also involve the underlying Gault Clay (Toms, 1953). A subsidence of as much as 1.5 m has been recorded, but movement is usually much less than this. It is noteworthy that the great majority of the three dozen recorded failures have consisted of renewals of movement in the slipped masses that form the Undercliff and have in no case involved a general recession of the rear scarp of the landslips. Thus, although locally scarred by Chalk falls, or slightly shifted by sets, the High Cliff is a feature of considerable age.

The largest slip about which detailed information has been collected took place in 1915. Other notable movements occurred in 1937 and 1940 (see Tables 7.1 and 7.2). The 1937 landslip (Toms, 1946; Hutchinson, 1969) was more than 900 m wide, and affected the whole of the slope seawards of the railway line (Figure 7.7). Upheaval of the foreshore took place. Seaward movement varied from about 27 m in the western part of the slip to zero at Warren Halt. The 1940 landslip (Toms, 1953; Muir Wood, 1955b; Hutchinson, 1969) took place in about 6 ha of the Warren, with a length along the coast approaching 700 m. Movements were slight and gradual, beginning in 1940 and continuing episodically to 1947, amounting to 1.5 m horizontally at most. Its essential features are shown on two cross-sections (Figure 7.8).

Since 1936–1940, apart from a slight renewal of movement of the 1940 slip in 1947, no major movements have occurred. The improved stability of the Warren since 1915 is probably the cumulative result of coast protection measures, drainage works in the slipped masses and the extensive weighting of the toe described by Viner-Brady (1955).

Description

(a) Hydrology

Investigations in 1948–1950 left little doubt that most of the groundwater in the slipped masses derives from the aquifer provided immediately inland by the Chalk. Using data from boreholes in the Warren, Muir Wood (1955b) has drawn

Table 7.1 Folkestone Warren: summary of the average values of ϕ_r' (°), σ_n' and *s* in the Gault Clay at failure in the 1940, 1937 and 1915 landslips. The original pre-metric data have been used. After Hutchinson (1969).

Landslip	ϕ_r' (°)		σ_n' (pounds per square foot)		*s* (pounds per square foot)	
	Max *u*	Min *u*	Max *u*	Min *u*	Max *u*	Min *u*
1940	15.1	14.0	4510	4950	1215	1235
1937	16.3	14.0	8340	9740	2440	2430
1915	16.6	13.9	13 170	15 620	3925	3865

σ_n' average effective normal stress on slip-surface in Gault Clay determined graphically using computed values of internal forces
u porewater pressure acting on slip-surface
s average shear-strength, $\sigma_n'\tan\phi_r'$, along slip-surface in Gault Clay

Table 7.2 Results of stability analyses. After Hutchinson (1969).

	Cross section	Pore pressure assumption on slip surface	Value of ϕ_r' Gault Clay required for F = 1.0		Remarks
			Janbu (°)	Morgenstern and Price (°)	
1915 landslip	W4 lower profile	Maximum Minimum	10.2 8.3		Slip surface entirely in Gault Clay
	W4 upper profile	Maximum Minimum	10.3 8.5		Slip surface entirely in Gault Clay
	W4 average profile	Maximum Minimum	10.25 8.4	9.7 7.7	Slip surface entirely in Gault Clay
	W6	Maximum Minimum	16.55 14.3	16.3 13.8	ϕ_r' for small length of slip surface in Chalk Marl at rear of slip = 20°
	W8	Maximum Minimum	23.85 20.1	22.2 18.7	ϕ_r' for small length of slip surface in Lower Chalk at rear of slip = 23°
	Weighted average of sections W4 (average), W6 and W8	Maximum Minimum	17.6 14.8	16.6 13.9	
1937 landslip	W1	Maximum Minimum	17.4 14.9		ϕ_r' for small length of slip surface in Lower Chalk at rear of slip = 23°
	W2	Maximum Minimum	18.3 15.9	16.2 14.0	ϕ_r' for small length of slip surface in Lower Chalk at rear of slip = 23°
1940 landslip	W5	Maximum Minimum	13.9 13.0	12.5 11.6	ϕ_r' for small length of slip surface in Chalk Marl at rear of slip = 20°
	W7	Maximum Minimum	19.2 17.6	17.6 16.4	ϕ_r' for small length of slip surface in Chalk Marl at rear of slip = 20°
	Average of (W5 + W7)/2	Maximum Minimum	16.55 15.3	15.05 14.0	ϕ_r' for small length of slip surface in Chalk Marl at rear of slip = 20°

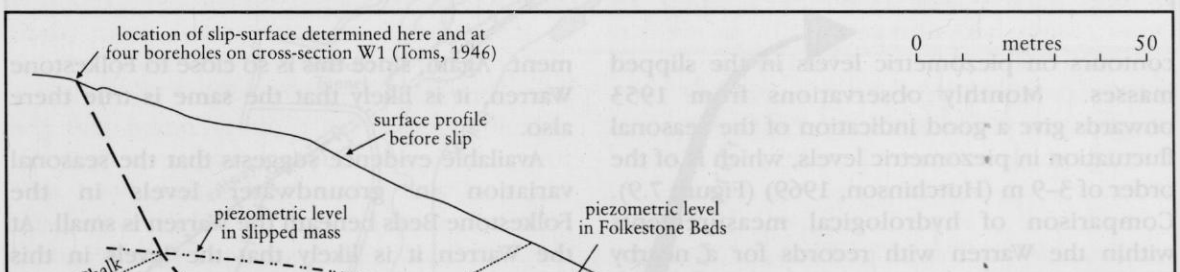


Figure 7.7 Section through the 1937 landslide of Folkestone Warren, transformed/transferred from cross-section W2. After Hutchinson (1969).

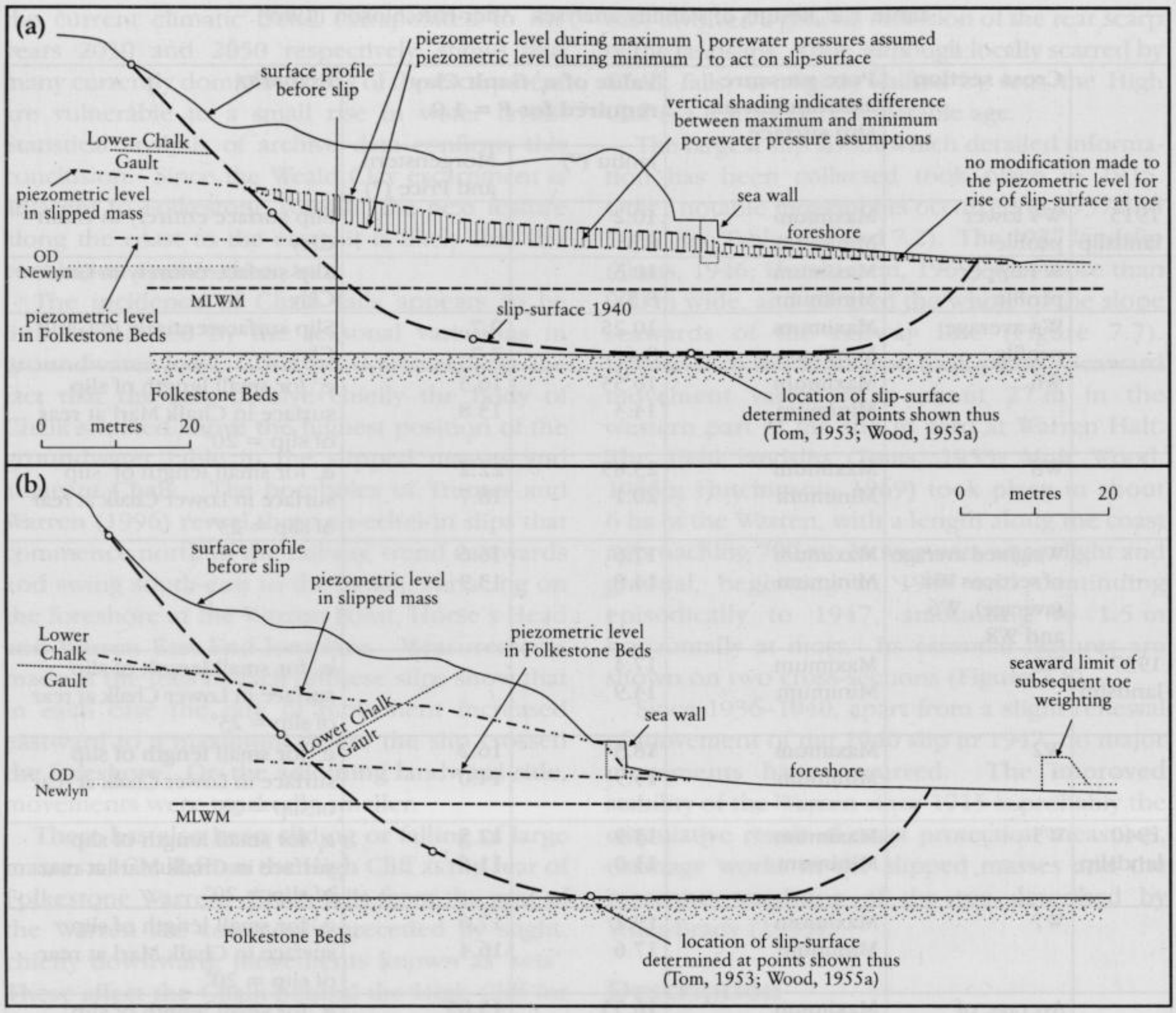


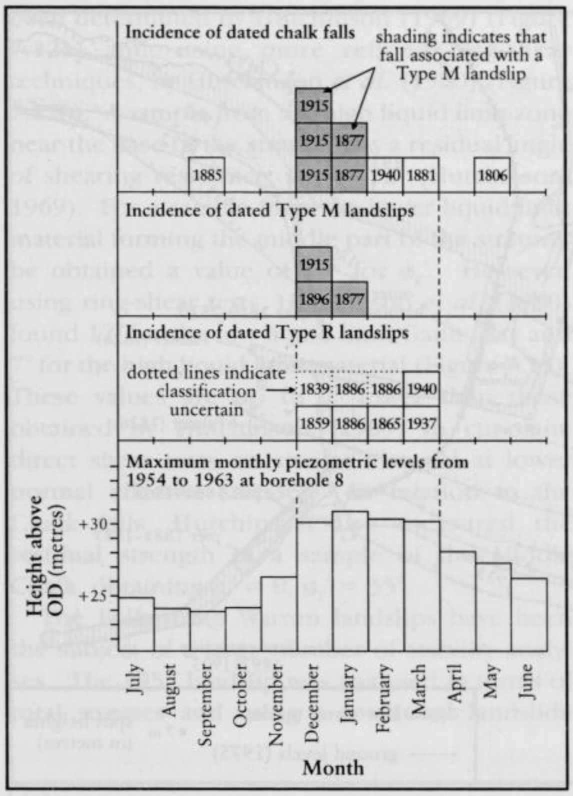
Figure 7.8 (a) The 1940 landslide at Folkestone Warren based on cross-section W5. (b) The 1940 landslide based on cross-section W7. After Hutchinson (1969).

contours on piezometric levels in the slipped masses. Monthly observations from 1953 onwards give a good indication of the seasonal fluctuation in piezometric levels, which is of the order of 3–9 m (Hutchinson, 1969) (Figure 7.9). Comparison of hydrological measurements within the Warren with records for a nearby Chalk well shows that an intimate connection exists between the groundwater bodies in the slipped masses and the adjacent in-situ Chalk (Muir Wood, 1955b).

At The Roughs, analysis using the 5% and 8% perturbations of the current climatic trend mentioned (Brunsden *et al.*, 1996a), showed that only a small rise in water levels would bring groundwater to ground surface level over large areas of the Hythe Beds escarpment. Again, since this is so close to Folkestone Warren, it is likely that the same is true there also.

Available evidence suggests that the seasonal variation in groundwater levels in the Folkestone Beds beneath the Warren is small. At the Warren it is likely that the levels in this confined aquifer fluctuate slightly in response to, but lag somewhat behind, the tidal variations in sea level (Hutchinson, 1969).

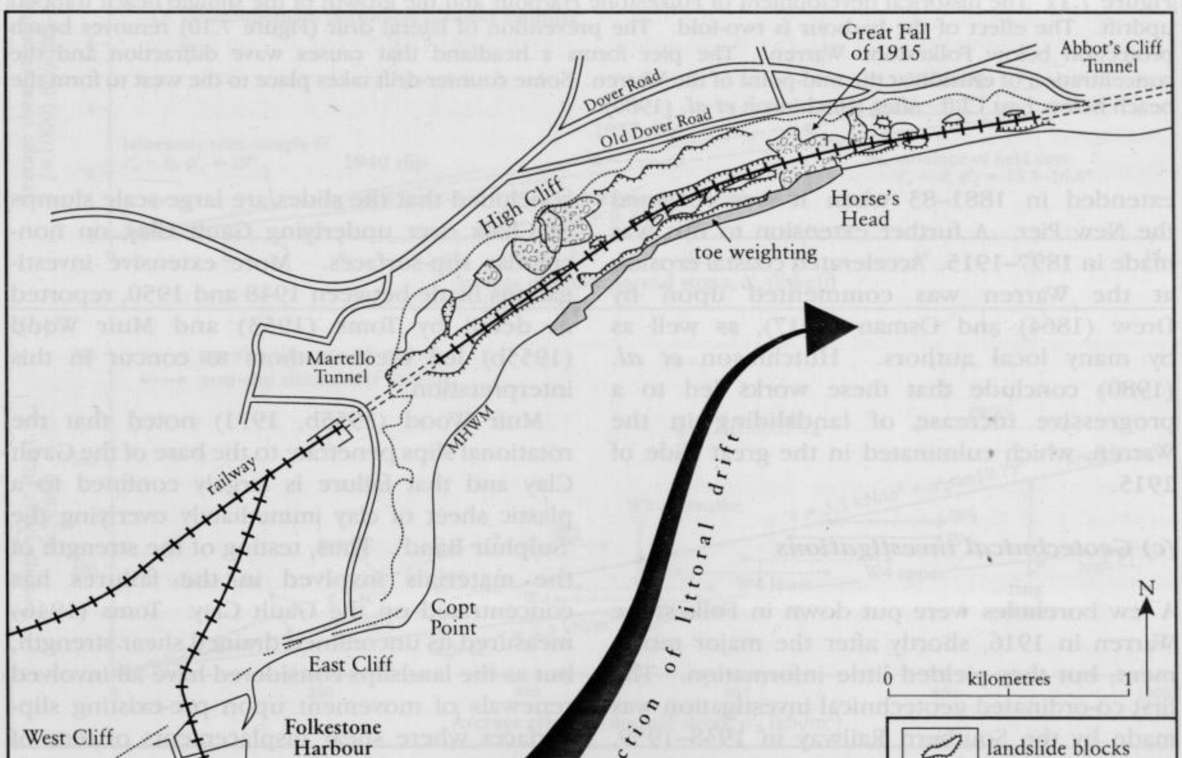
Figure 7.10▸ The influence of dominant wave energy and littoral drift on the stability of the landslide complex at Folkestone Warren. The asymmetrical, zeta-bay shape is a typical setting for such large landslides on the south coast of Britain. After Bromhead (1986), from Jones and Lee (1994).



◄**Figure 7.9** The relationship between the incidence of landslides and seasonal variation in piezometric level in the slipped masses. After Hutchinson (1969).

(b) Coastal factors

Hutchinson *et al.* (1980) have discussed the predominant eastward littoral drift on this part of the coast (Figure 7.10), and the way that interference with this drift in the Folkestone area tends to deplete the beaches downdrift and hence stimulate landsliding in Folkestone Warren. Major interference dates from the construction of the masonry harbour at Folkestone in the first decade of the 19th century. Construction began in about 1807. The West Pier was completed by 1810 and the Haven by 1820. However, by 1830 the harbour mouth was completely choked by sand and shingle. From 1842 onwards, successive pier extensions were carried out in order to produce shingle-free, deep-water berths (Figure 7.11). In 1861–1863 the Promenade Pier was built, initially of timber but stone-cladded and



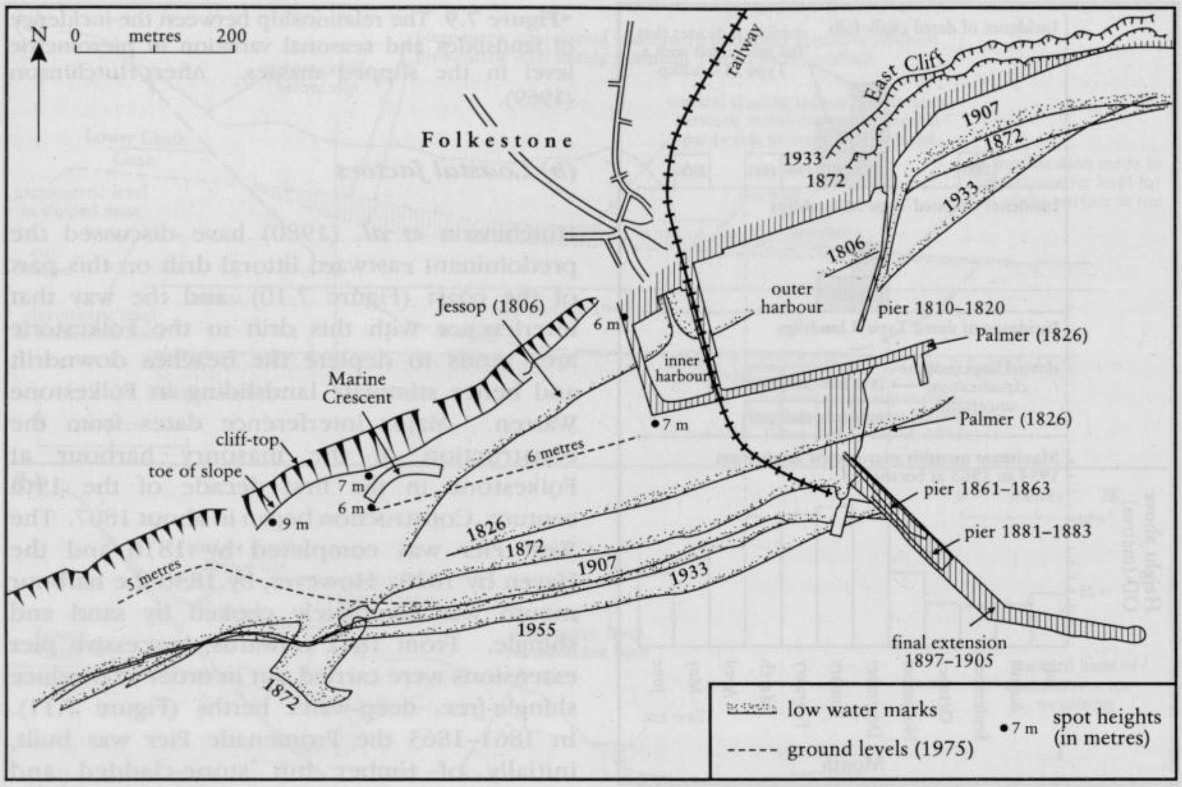


Figure 7.11 The historical development of Folkestone Harbour and the growth of the shingle beach trapped updrift. The effect of the harbour is two-fold. The prevention of lateral drift (Figure 7.10) removes beach protection below Folkestone Warren. The pier forms a headland that causes wave diffraction and the concentration of erosion at the mid-point of the Warren. Some counter-drift takes place to the west to form the beach below East Cliff. After Hutchinson *et al.* (1980).

extended in 1881–83 when it was renamed the New Pier. A further extension to this was made in 1897–1915. Accelerated coastal erosion at the Warren was commented upon by Drew (1864) and Osman (1917), as well as by many local authors. Hutchinson *et al.* (1980) conclude that these works led to a progressive increase of landsliding in the Warren, which culminated in the great slide of 1915.

(c) Geotechnical investigations

A few boreholes were put down in Folkestone Warren in 1916, shortly after the major movement, but they yielded little information. The first co-ordinated geotechnical investigation was made by the Southern Railway in 1938–1939, following the landslip movements in 1936 and 1937. The investigation was concentrated towards the western end of the unstable area, and was described by Toms (1946), who concluded that the slides are large-scale slumps of Chalk over underlying Gault Clay, on non-circular slip-surfaces. More extensive investigations made between 1948 and 1950, reported in detail by Toms (1953) and Muir Wood (1955b) led these authors to concur in this interpretation.

Muir Wood (1955b, 1971) noted that the rotational slips penetrate to the base of the Gault Clay and that failure is largely confined to a plastic sheet of clay immediately overlying the 'Sulphur Band'. Thus, testing of the strength of the materials involved in the failures has concentrated on the Gault Clay. Toms (1946) measured its unconfined drained shear strength, but as the landslips considered have all involved renewals of movement upon pre-existing slip-surfaces where shear displacements of tens of metres have taken place, the shear strength mobilized at failure can be taken to be residual. The residual strength of the Gault Clay, taking samples of both high and low liquid limit, has been determined by Hutchinson (1969) (Figure 7.12a) and, using more refined ring-shear techniques, by Hutchinson *et al.* (1980) (Figure 7.12b). A sample from the high liquid limit zone near the base of the stratum has a residual angle of shearing resistance, ϕ_r', of 12° (Hutchinson, 1969). For a sample from the lower liquid limit material forming the middle part of the stratum, he obtained a value of 19° for ϕ_r'. However, using ring-shear tests, Hutchinson *et al.* (1980) found 12° for the low liquid limit Gault Clay and 7° for the high liquid limit material (Figure 7.13). These values are up to 7° lower than those obtained by Hutchinson (1969) in cut-plane direct shear tests on similar material at lower normal effective stresses. In relation to the Chalk falls, Hutchinson also measured the residual strength of a sample of the Middle Chalk, obtaining $c_r' = 0$, $\phi_r' = 35°$.

The Folkestone Warren landslips have been the subject of a large number of stability analyses. The 1937 landslip was analysed in terms of total stresses and using a rotational landslide failure model, by Toms (1946) and Skempton (1946). Similar analysis of the 1915 and 1940 landslips was carried out by Muir Wood (1955b). The 1915, 1937 and 1940 landslips have been analysed in terms of effective stresses by Hutchinson (1969) and Hutchinson *et al.* (1980), an approach more appropriate for long-term problems. In contrast to the earlier work of Toms (1946, 1953) they employed a non-circular, multiple failure model. Hutchinson (1969) found that the average strengths mobilized on the non-circular failure surfaces in the Gault Clay approximate to the residual, and are bounded by the envelopes $c_r' = 0$, $\phi_r' = 13.9°$ and $c_r' = 0$, $\phi_r' = 16.6°$. However, Hutchinson *et al.* (1980) found that in the range of average normal effective stress levels in the Warren (about 200–800 kN m^{-3}), the values of ϕ_r' indicated as likely by stability analyses are more probably in the range 7.5° to 15°. Clearly, these values tend to be higher than those derived from ring-shear tests. Trenter and Warren (1996) made residual effective shear

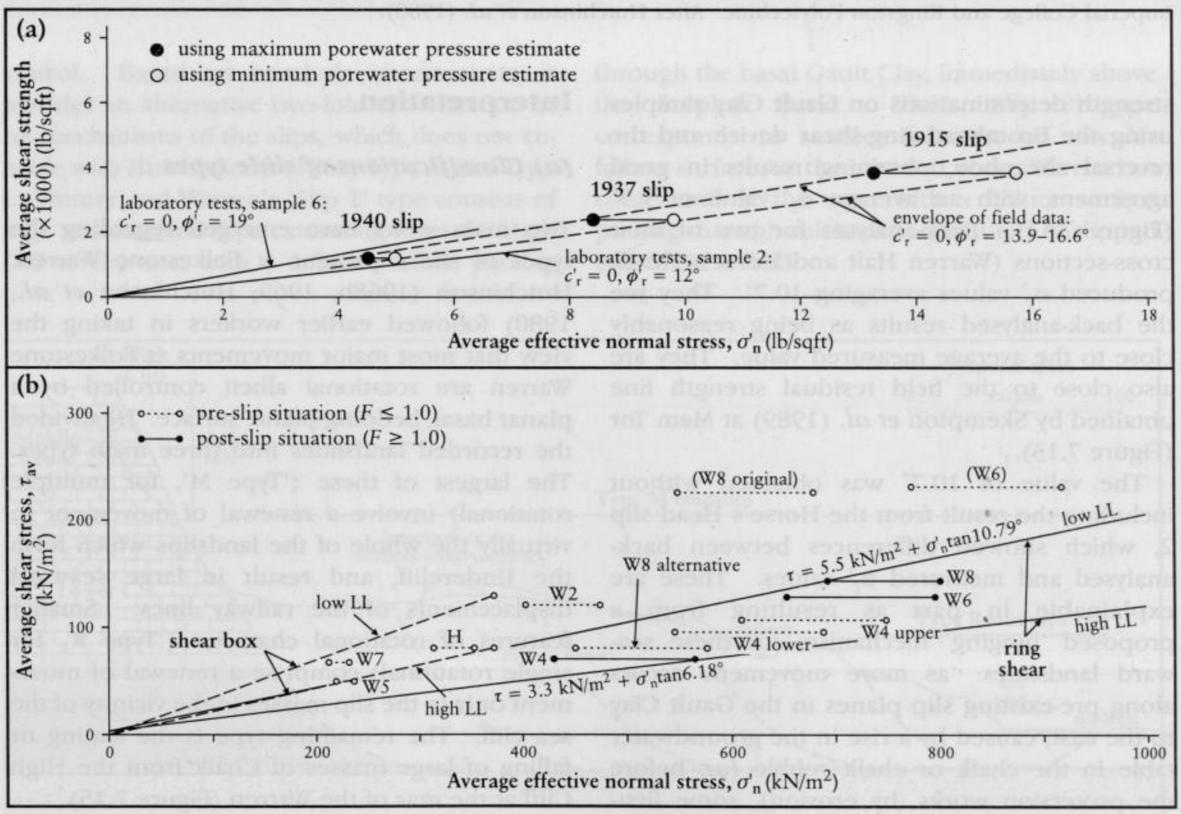


Figure 7.12 (a) Plot of average shear-strength mobilized in the Gault Clay at failure against average effective normal stress for the 1915, 1937 and 1940 Folkestone Warren landslips (after Hutchinson, 1969). (b) Comparison of residual strengths in the Gault Clay derived from back-analyses with the corresponding envelopes obtained in the laboratory (after Hutchinson *et al.*, 1980).

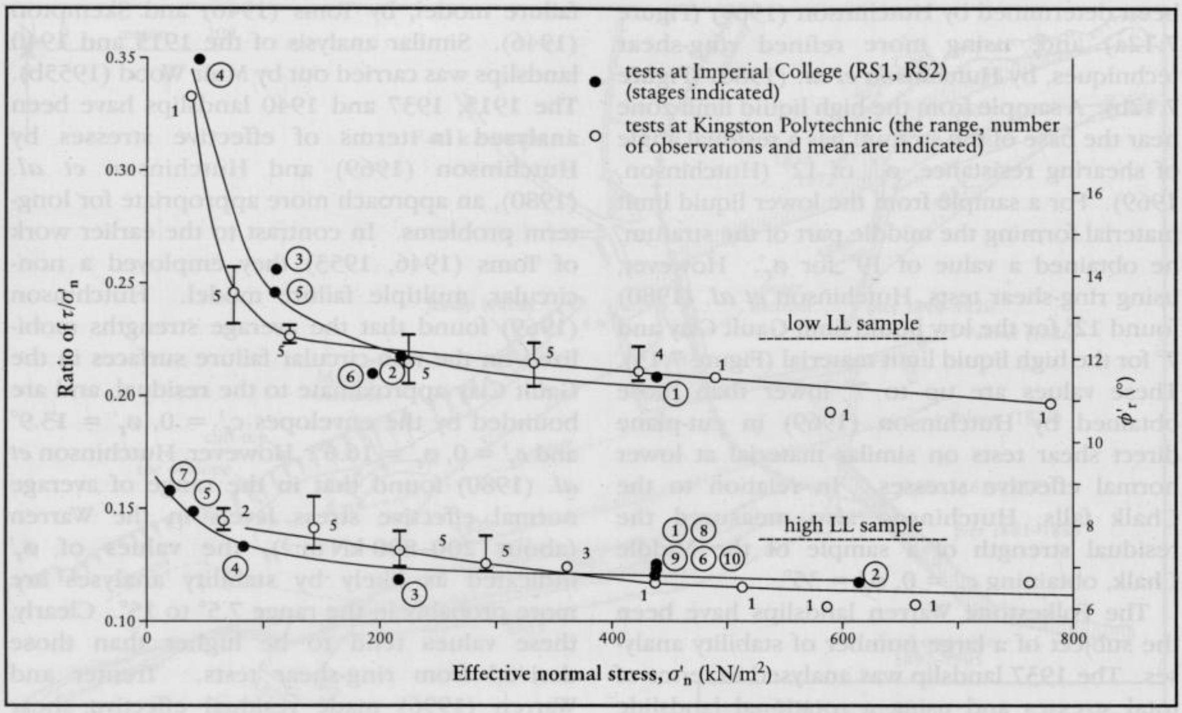


Figure 7.13 Summary of ring-shear test results, showing the comparison between the measurements of Imperial College and Kingston Polytechnic. After Hutchinson *et al.* (1980).

strength determinations on Gault Clay samples using the Bromhead ring-shear device and the reversal shear-box, obtaining results in good agreement, with an average $ø_r'$ value of 9.5° (Figure 7.14). Back-analyses for two of their cross-sections (Warren Halt and Horse's Head) produced $ø_r'$ values averaging 10.7°. They see the back-analysed results as being reasonably close to the average measured value. They are also close to the field residual strength line obtained by Skempton *et al.* (1989) at Mam Tor (Figure 7.13).

The value of 10.7° was obtained without including the result from the Horse's Head slip 2, which showed differences between back-analysed and measured $ø_r'$ values. These are explainable in part as resulting from a proposed 'hinging' mechanism for these seaward landslides: 'as more movement occurs along pre-existing slip planes in the Gault Clay to the east, caused by a rise in the groundwater table in the chalk or chalk rubble (or, before the protection works, by erosion), some first-time movement will be provoked about the hinge further west' (Trenter and Warren, 1996, p. 618).

Interpretation

(a) Classifications of slide types

Two main views have emerged regarding the types of slides present at Folkestone Warren. Hutchinson (1968b, 1969; Hutchinson *et al.*, 1980) followed earlier workers in taking the view that most major movements at Folkestone Warren are rotational albeit controlled by a planar basal, bedding-plane, surface. He divided the recorded landslides into three main types. The largest of these ('Type M', for multiple rotational) involve a renewal of movement in virtually the whole of the landslips which form the Undercliff, and result in large seaward displacements of the railway lines. Smaller features of rotational character ('Type R', for single rotational) comprise a renewal of movement only in the slip masses in the vicinity of the sea cliff. The remaining type is the sliding or falling of large masses of Chalk from the High Cliff at the rear of the Warren (Figure 7.15).

The second and more recent view is that of Trenter and Warren (1996) who gave more emphasis to the effect of the planar bedding

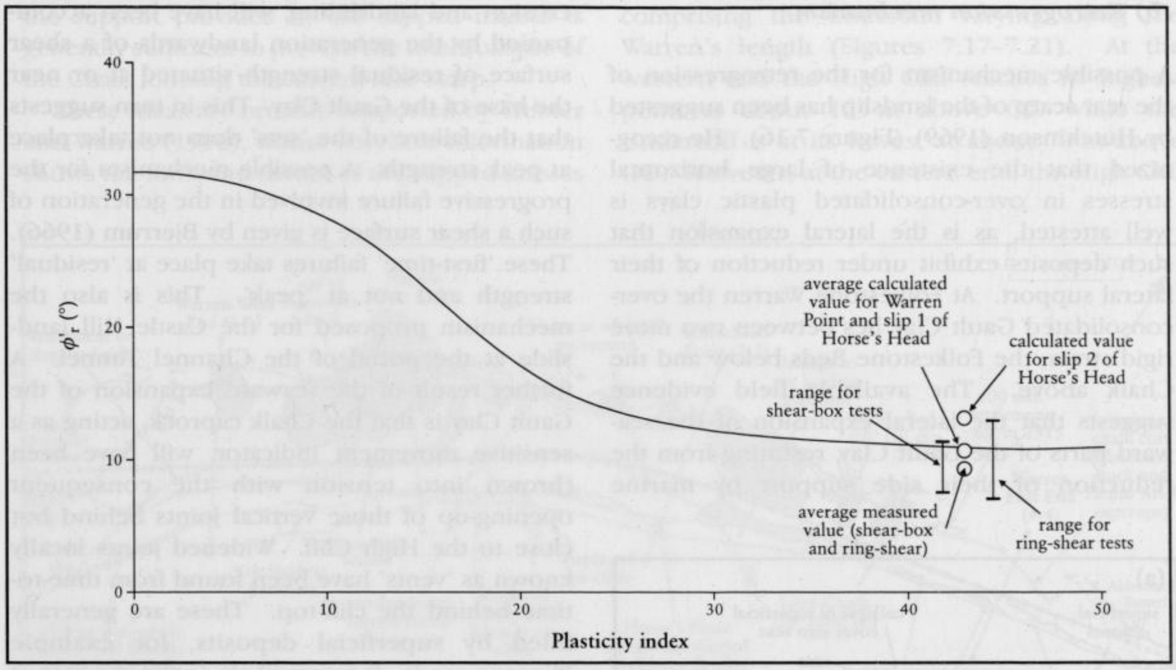


Figure 7.14 Correlation between residual strength and plastic index. From Skempton *et al.* (1989) with data from Trenter and Warren (1996).

control. Based on borehole observations, it provides an alternative two-fold classification of the mechanisms of the slips, which does not co-incide with Hutchinson's (1969) Type M and Type R. Trenter and Warren's 'Slip 1' type consists of large translational slips extending from the High Cliff to the sea, with a failure surface passing through the basal Gault Clay, immediately above the Sulphur Band. Their 'Slip 2' type corresponds to smaller features, on circular failure surfaces at the east end of the Warren but compound at the west with a failure surface passing through substantial quantities of slipped and broken chalk.

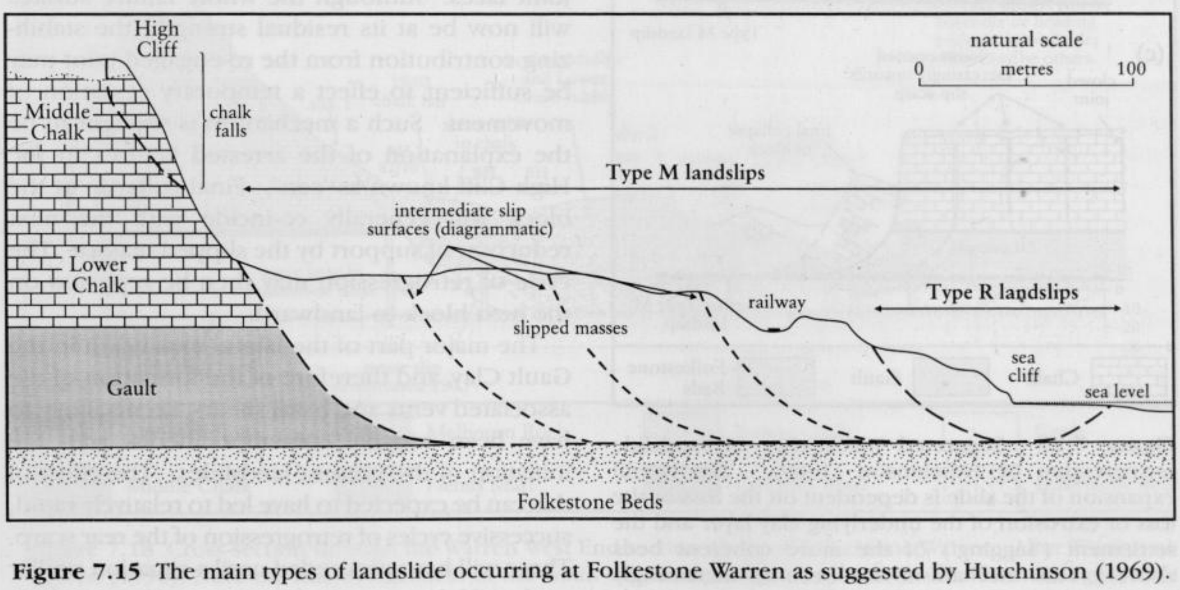


Figure 7.15 The main types of landslide occurring at Folkestone Warren as suggested by Hutchinson (1969).

(b) Retrogression mechanism

A possible mechanism for the retrogression of the rear scarp of the landslip has been suggested by Hutchinson (1969) (Figure 7.16). He recognized that the existence of large horizontal stresses in over-consolidated plastic clays is well attested, as is the lateral expansion that such deposits exhibit under reduction of their lateral support. At Folkestone Warren the over-consolidated Gault Clay lies between two more rigid strata, the Folkestone Beds below and the Chalk above. The available field evidence suggests that the lateral expansion of the seaward parts of the Gault Clay, resulting from the reduction of their side support by marine erosion and landsliding, will have been accompanied by the generation landwards of a shear surface of residual strength situated at or near the base of the Gault Clay. This in turn suggests that the failure of the 'sets' does not take place at peak strength. A possible mechanism for the progressive failure involved in the generation of such a shear surface is given by Bjerrum (1966). These 'first-time' failures take place at 'residual' strength and not at 'peak'. This is also the mechanism proposed for the Castle Hill landslide at the portal of the Channel Tunnel. A further result of the seaward expansion of the Gault Clay is that the Chalk caprock, acting as a sensitive movement indicator, will have been thrown into tension with the consequent opening-up of those vertical joints behind but close to the High Cliff. Widened joints locally known as 'vents' have been found from time-to-time behind the cliff-top. These are generally filled by superficial deposits, for example Pliocene sands. It seems likely, in the absence of anticlinal flexure and/or cambering, that they were initiated by lateral extension of the hill.

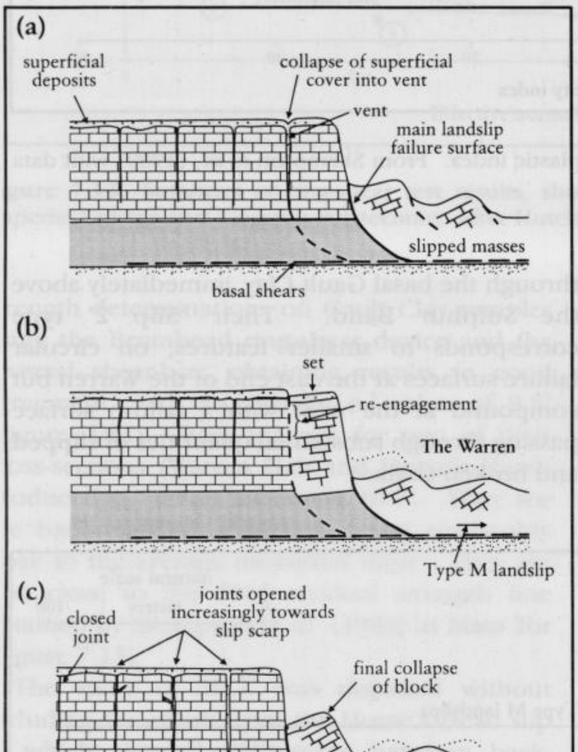


Figure 7.16 Suggested mechanism of landslide retrogression by Hutchinson (1969). The lateral expansion of the slide is dependent on the loss or the loss or extrusion of the underlying clay layer and the settlement ('sagging') of the more coherent beds above. This mechanism has become increasingly important in recent work.

The forces resisting the collapse of the Chalk mass isolated between the cliff-face and the most seaward vent will have been greatly diminished by these movements. If insufficient support is provided by the slipped masses to seaward, the block will fail and begin to subside. With increasing subsidence the curvature of the failure surface produces a back-tilt of the failing block (the apparent rotational failure) which will bring about a re-engagement of part of the irregular joint faces. Although the whole failure surface will now be at its residual strength, the stabilizing contribution from the re-engaged joint may be sufficient to effect a temporary cessation of movement. Such a mechanism is thought to be the explanation of the arrested failures in the High Cliff known as 'sets'. Final collapse of the block will generally co-incide with the next reduction of support by the slipped masses. The cycle of retrogression may then be repeated on the next block to landward.

The major part of the lateral expansion in the Gault Clay, and therefore of the formation of the associated vents and basal shears, seems likely to have been roughly contemporaneous with the initiation of the present landslides. The effect of this can be expected to have led to relatively rapid, successive cycles of retrogression of the rear scarp. These will have proceeded until a situation similar to that of the present day was reached in which the support provided by the slipped masses is generally sufficient to prevent the total collapse of the Chalk forming the current rear scarp.

These ideas are broadly supported by Trenter and Warren (1996), whose borehole information shows the form and nature of the slipped masses comprising the Undercliff varying along the Warren's length (Figures 7.17–7.21). At the western end the High Cliff reaches its highest point at about 165 m above OD while the Undercliff is at its lowest at about 70 m above OD. However, at the eastern end, the High Cliff

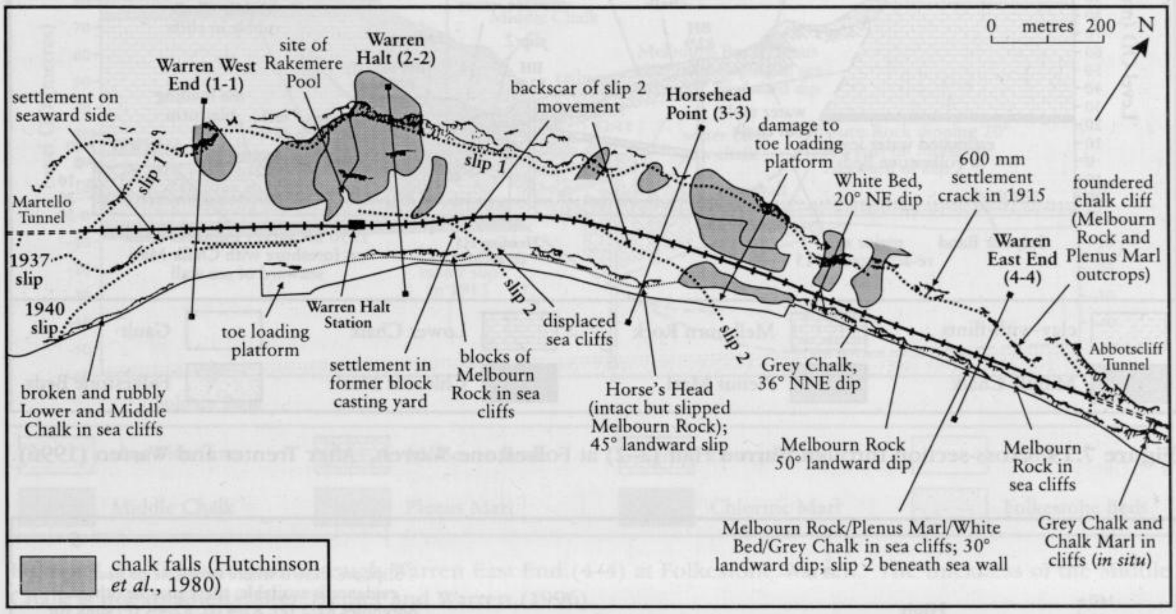


Figure 7.17 Site plan showing the location of boreholes and the cross-sections described by Trenter and Warren (1996).

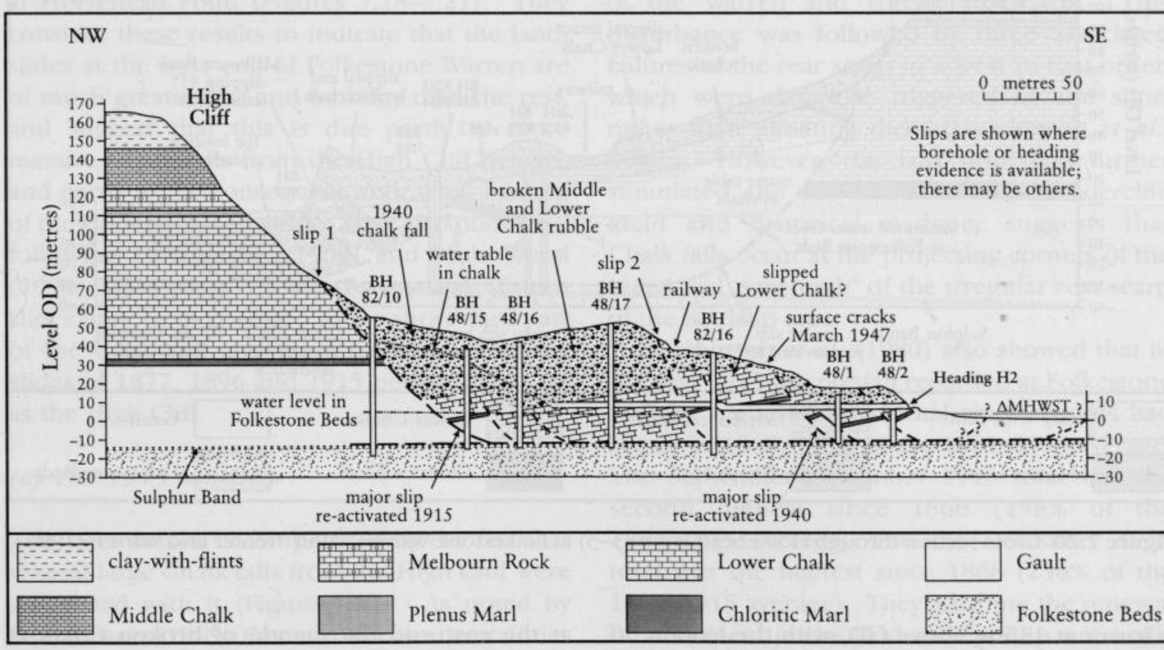


Figure 7.18 Cross-section through the Warren West End section (1–1) of Folkestone Warren. After Trenter and Warren (1996). Note the loss of thickness of the Gault Clay and Lower Chalk strata. The former suggests clay extrusion. The latter suggests that the original failure took place from a cliff that sloped towards the sea.

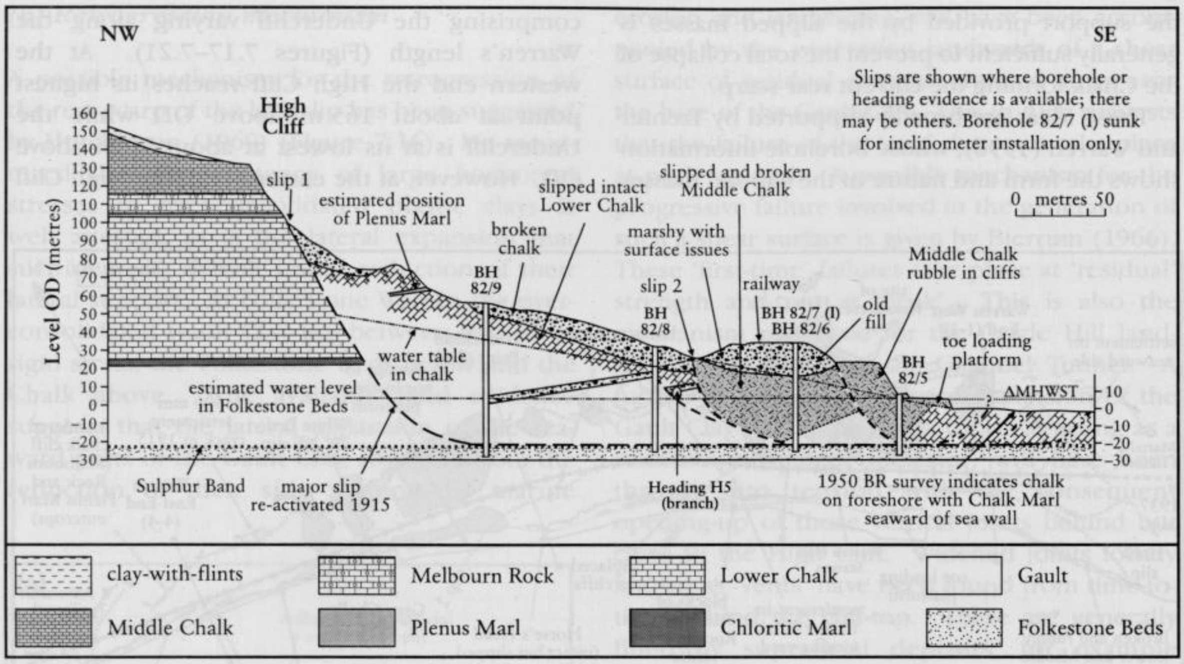


Figure 7.19 Cross-section through Warren Halt (2–2) at Folkestone Warren. After Trenter and Warren (1996).

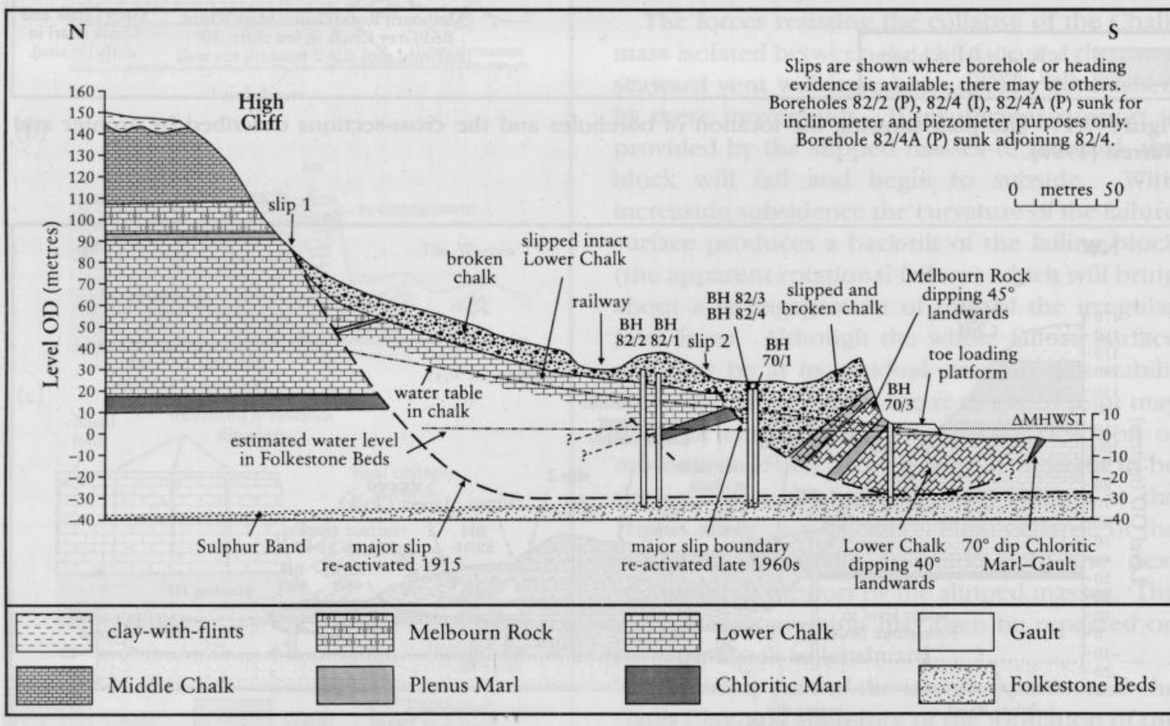


Figure 7.20 Cross-section through Horsehead Point (3–3) at Folkestone Warren. After Trenter and Warren (1996).

is lower at 135 m above OD, with the Undercliff at its highest, 90 m above OD. Trenter and Warren's (1996) boreholes show up to 45 m thickness of broken Chalk at the west end, while at the east end the mantle of broken Chalk is only 10–15 m thick. The boreholes also indicate larger amounts of slipped but intact Lower and Middle Chalk at the east end, the prime example

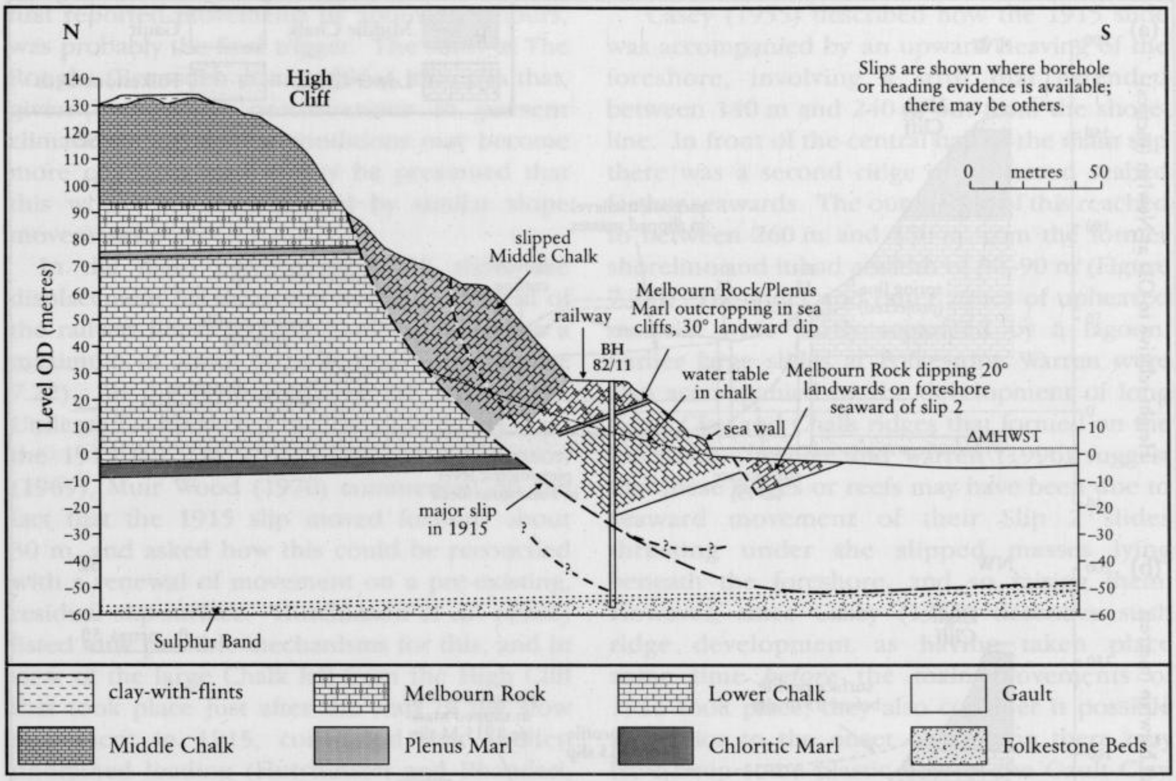


Figure 7.21 Cross-section through Warren East End (4–4) at Folkestone Warren. The thickness of the Middle Chalk is unexplained. After Trenter and Warren (1996).

being the Horse's Head, a prominent small hill, at Horsehead Point (Figures 7.18–7.21). They consider these results to indicate that the landslides at the west end of Folkestone Warren are of much greater size and mobility than the rest, and suggest that this is due partly to more massive Chalk falls from the High Cliff behind, and partly to the consequent undrained loading of the Undercliff landslides. They also point out, following Hutchinson (1969) and Muir Wood (1994), that most of the Folkestone Warren slides have been confined to the shoreward part of the Undercliff and the foreshore. Only the slides of 1877, 1896 and 1915 penetrated as far as the High Cliff.

(c) The 1915 landslip

The 1915 failure is complicated by the fact that several large Chalk falls from the High Cliff were associated with it (Figure 7.22). As noted by Hutchinson (1969) this introduces uncertainties into stability analyses. An unusually large number of eyewitness accounts of the slide were collected. These tend to show that the renewal of movement commenced at the western end of the Warren and spread eastwards. This disturbance was followed by three associated failures of the rear scarp in a west to east order, which were therefore triggered by the slips, rather than initiating them (Hutchinson *et al.*, 1980). However, the falls doubtless further stimulated the movements of the Undercliff. Field and historical evidence suggests that Chalk falls occur at the projecting corners of the individual 'saw-teeth' of the irregular rear scarp of the landslip area.

Hutchinson *et al.* (1980) also showed that in the year 1915 the rainfall recorded at Folkestone had the highest annual total since records had begun in 1868, 138% of the 1881–1915 average. The September–December 1915 total was the second highest since 1868 (158% of the 1881–1915 average), and the December 1915 total was the highest since 1868 (256% of the 1881–1915 average). They attribute the renewal of movement in the Undercliff to this unusually high rainfall, and to the erosion of the bay as intensified by pier construction. The low tide (two days before Spring tides) that preceded the

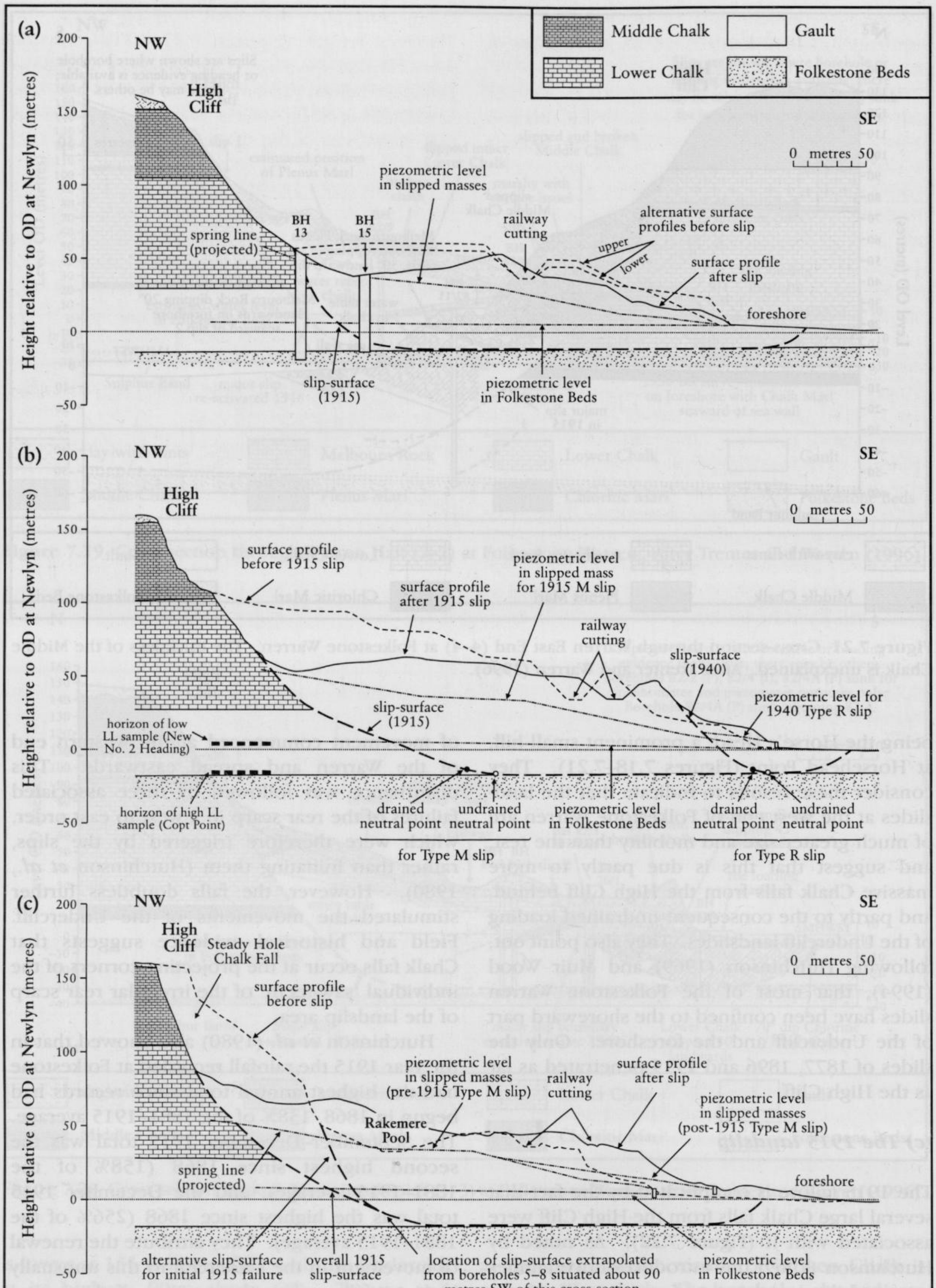


Figure 7.22 Cross-sections through Folkestone Warren. (a) 1915 landslide section W4; (b) section W6; (c) 1915, section W8. After Hutchinson (1969).

first reported movements by about 1.25 hours, was probably the final trigger. The study at The Roughs (Brunsden *et al.*, 1996a) indicates that, given only small perturbations to present climatic trends, similar conditions may become more common, and it may be presumed that this would be accompanied by similar slope movements.

In the main part of the 1915 slide, the displacement (as measured by the movement of the railway lines) was 10–20 m, increasing to a maximum of about 50 m at one point (Figure 7.22). In the central part of the Warren, the Undercliff is blanketed by flow slide debris from the 1915 fall. In a discussion of Hutchinson (1969), Muir Wood (1970) commented on the fact that the 1915 slip moved forward about 30 m, and asked how this could be reconciled with a renewal of movement on a pre-existing, residual slip-surface. Hutchinson *et al.* (1980) listed nine possible mechanisms for this, and in view of the large Chalk fall from the High Cliff that took place just after the start of the slow movement in 1915, concluded that sudden undrained loading (Hutchinson and Bhandari, 1971) of the rear of the slide by this rockfall, was the most likely explanation.

Casey (1955) described how the 1915 slide was accompanied by an upward heaving of the foreshore, involving a strip that extended between 140 m and 240 m out from the shoreline. In front of the central half of the main slip there was a second ridge of upheaved seabed farther seawards. The outer edge of this reached to between 260 m and 350 m from the former shoreline and it had a width of 80–90 m (Figure 7.23). The inner and outer zones of upheaved material were partly separated by a lagoon. Earlier large slides at Folkestone Warren were also accompanied by the development of long Gault Clay and Chalk ridges that formed on the foreshore. Trenter and Warren (1996) suggest that these ridges or reefs may have been due to seaward movement of their Slip 2 slides thrusting under the slipped masses lying beneath the foreshore, and so raising them. However, since Casey (1955) describes such ridge development as having taken place some time *before* the main movements of 1916 took place, they also consider it possible that prior to the onset of slipping there may have been some plastic flow of the Gault Clay, with the failed Gault Clay erupting at the foreshore.

Figure 7.23 View to the east of the sea cliff and heaved foreshore of the Warren, taken just after the 1915 slip. The partly eroded 'cape' of the debris of the Great Fall can be seen just beyond the Horse's Head. (Photo: British Railways, Southern Region.)

(d) Contemporary movements

Trenter and Warren's (1996) triangulation and observations within drainage headings, supplemented by piezometer readings, have shown that an area in the vicinity of the 'Horse's Head', which is the most prominent sea cliff in the Warren, in the central part of the Undercliff, has been moving seawards since at least 1938. The area of greatest movement co-incides with the location of the debris tongue of the 1915 fall, and is also the area in which coastal erosion was at a maximum until the extension of the present sea wall across it. Piezometer results suggest that at least in the vicinity of the seaward edge of the Warren and possibly elsewhere, the slipped masses consist of blocks of Gault Clay of various sizes among Chalk blocks and debris which in general have become unloaded as the slips have developed (Hutchinson *et al.*, 1980). The piezometric pressures within the Chalk and the smaller masses of Gault Clay had equalized with the long-term groundwater conditions, but negative excess porewater-pressures still existed within the larger blocks of slipped Gault Clay (Figure 7.24). As these swelled back to equilibrium the factor of safety on any slip-surfaces traversing these blocks steadily decreased.

(e) Age

It is unlikely that Folkestone Warren was initiated before the virtual completion of the Flandrian transgression. An immature soil profile situated

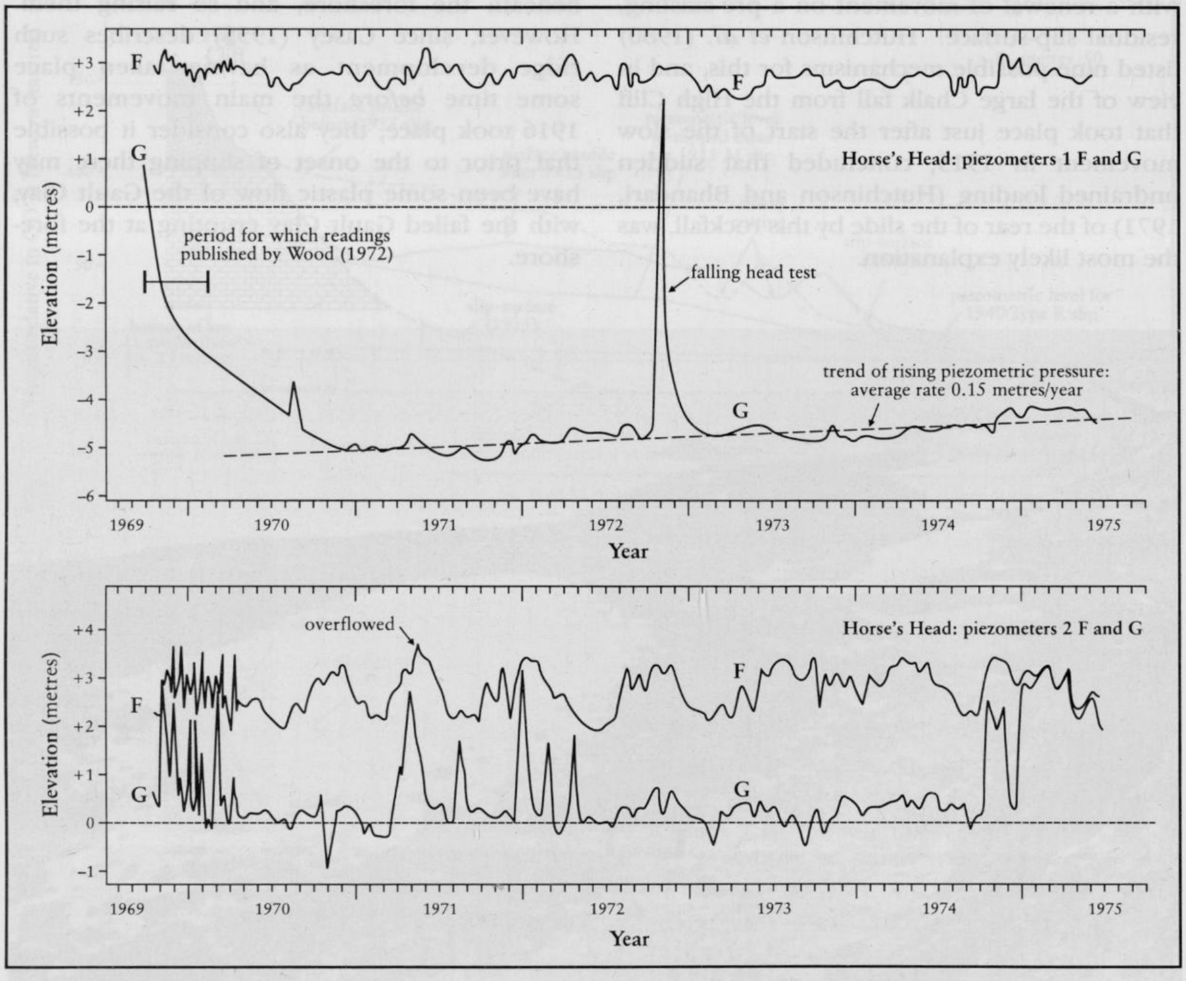


Figure 7.24 Record of piezometric levels in boreholes 1 and 2 in the Horse's Head area from installation in 1969 until readings ceased in 1975. The letters F, G refer respectively to piezometers with their tips in the Folkestone Beds and the slipped Gault Clay. After Hutchinson *et al.* (1980).

towards the base of Chalk debris overlying a bluff of steeply back-tilted Middle Chalk, at the Horse's Head, has been dated by its contained fauna to Atlantic or Sub-Boreal age. As these sea cliffs are the oldest remaining part of the present Warren landslides, and the soil layer is also strongly back-tilted, it can be inferred that the main slipping movements affecting it have occurred since its formation, suggesting a Sub-Boreal date (5500–2500 BC) for the initiation of the present landslides (Hutchinson, 1968b). The weathered and vegetated appearance of those parts of the present High Cliff that have not suffered relatively recent Chalk falls suggests that its age of at least two centuries indicated by the historical records is on the low side.

Conclusions

At least before the construction of the sea wall and toe weighting, Folkestone Warren was in a state of dynamic equilibrium under the combined influences of its topography, hydrology, geology and its exposure to marine attack. This is reflected chiefly in the widening of the Undercliff towards the west, which compensates for the gradual reduction in passive resistance at the toe as the level of the Gault Clay rises.

Folkestone Warren is a mass-movement site of great importance, particularly due to the great detail in which, and timescale over which, it has been studied. Other large-scale Chalk slips on the south coast, for example the Undercliff at Ventnor (Isle of Wight), and White Nothe in Dorset, seem to be broadly similar in nature. The results of the studies on Folkestone Warren are quoted in many studies of mass movements still further afield. The site is fundamental in the development of understanding of both translational and rotational slips, and the relationships between them.

The early recognition of the clay-extrusion model is of fundamental importance and has since become important in the interpretation of the major slides at Castle Hill on the Channel Tunnel portal, and is being increasingly recognized as a fundamental failure mechanism. New research on this topic is in progress in many countries.

STUTFALL CASTLE, LYMPNE, KENT (TR 117 345)

Introduction

Romney Marsh in Kent is backed by an abandoned marine cliff (Figure 7.25) (May, 2003). The cliff has been degrading by a process involving several types of mass movement since isolation from wave action by the growth of Romney Marsh. For example, The Roughs, a site in the Hythe Beds between Hythe and West Hythe, has 17 separate landslide scars, with lobate features in the accumulation zone down-slope. Stutfall Castle, at Lympne, is the remains of a Roman fort built in AD 280, which has broken up under the influence of these mass movements. The bastions which constitute the remaining fragments of the fort have moved relative to the crest and foot of the slope, and relative to each other, throwing light on the nature and effects of mass movements on the slope of the degraded cliff over the 1600+ years since the fort was abandoned by its last occupants.

Abandoned cliffs undergo successive changes following the cessation of erosion at their foot. Studies aimed at describing and explaining these changes have concentrated on locations where a cliff has been successively abandoned along its length, either by the growth of a coastal spit along a coastline (e.g. Savigear, 1952) or by the migration of a river meander (Brunsden and Kesel, 1973). The 'ergodic hypothesis' (substitution of space for time) is used to elucidate the probable sequence of events at any particular point along the cliff-line. At Stutfall Castle a different approach has been adopted: the abandoned cliff has been subject to investigation by a combined team of geotechnical engineers and archaeologists (Hutchinson *et al.*, 1985), in the expectation that the archaeology of the fort would throw light on the later stages of land-sliding and slope development (and that geotechnical investigation would illuminate some of the archaeological problems of the site). Previous excavations of the fort had been carried out by Roach Smith (1850, 1852), and by Cunliffe (1980a).

The crest of the slope is capped by the Hythe Beds, which form a plateau to the north. These are underlain thinly by Atherfield Clay, but most of the cliff slope is formed in Weald Clay, which

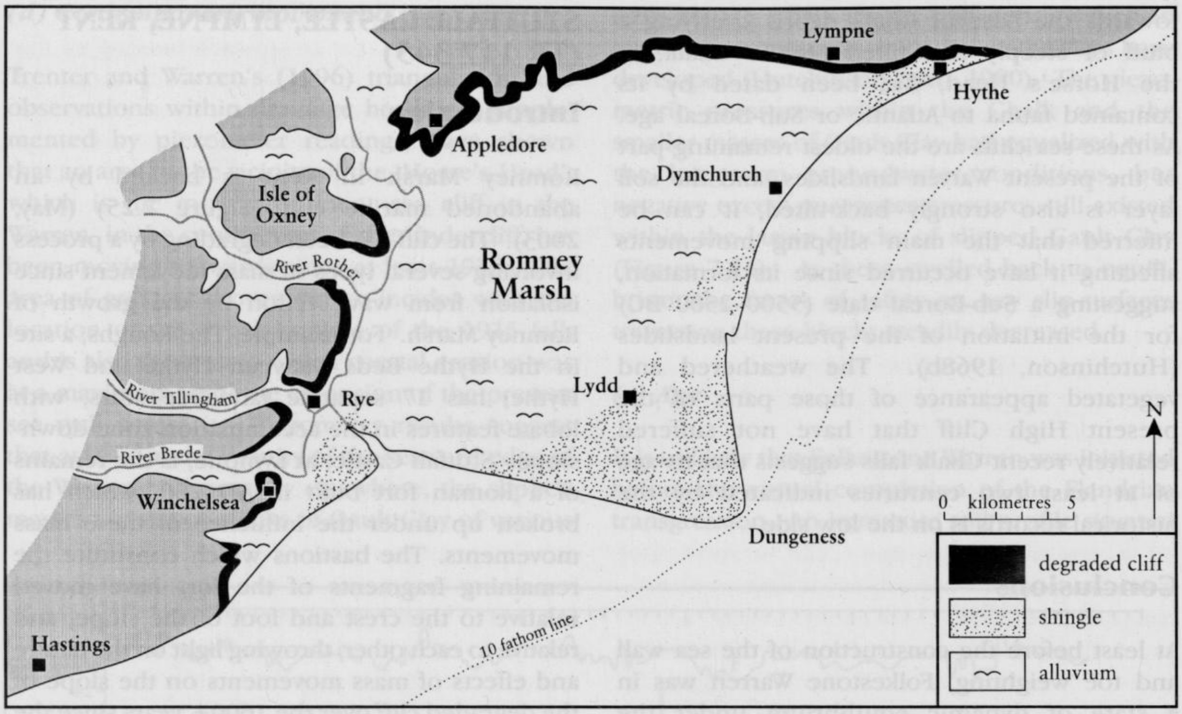


Figure 7.25 The former sea cliff which was abandoned as a result of the formation of Romney Marsh behind a major barrier of sand and shingle. Based on Jones, DKC (1981) and Jones and Lee (1994).

is estimated to extend to about 70 m below the slope foot (Smart *et al.*, 1966). The regional dip is 1°–2° to the NNE or north-east.

Description

The slope at Lympne is part of an extensive abandoned marine cliff, protected from marine erosion by Romney Marsh and its associated shingle spits. The cliff extends about 8 km from Hythe, 3 km to the east of Lympne, continuously to Appledore, 18 km to the west, and then discontinuously southwards to the coast 5 km to the east of Hastings (Figure 7.25). It is of fairly constant height, at about 90–100 m above OD. There is a tendency for the average steepness and signs of instability to increase from west to east, which may result from geological changes, but suggests that active marine erosion ceased more recently in the eastern parts. For example, The Roughs, close to the eastern end of the abandoned cliff, shows signs of current instability (Brunsden *et al.*, 1996a).

At Lympne the abandoned cliff has a height about 100 m above the marsh at its foot. The present slope profile comprises two main elements: at its head is a 15 m-high scarp in the Hythe Beds, standing at 35°, below which a 550 m-long, slightly irregular slope, predominantly in Weald Clay, extends down to the marsh at about 9°. The 9° slope can be divided into an upper degradation zone and a lower accumulation zone. The degradation zone possesses marked cross-slope undulations, characteristic of landsliding. The accumulation zone is smoother. The fact that the average slope angles of the degradation and accumulation zones are virtually identical indicates that the slope has developed to a condition of long-term stability so that, under present conditions of climate and vegetation, further landsliding is unlikely (Hutchinson *et al.*, 1985). This is in contrast with The Roughs, 1 km to the east along the Hythe Beds escarpment, which is still showing signs of active degradation and has not yet reached the angle of ultimate stability (Brunsden *et al.*, 1996a).

In the accumulation zone a sheet of shallow landslide debris and hillwash obscures the traces of the earlier, more deep-seated landslides which dislocated the fort. Three significant features may be identified (see Figure 7.26), using the combined evidence of the slope morphology and the remains of the fort:

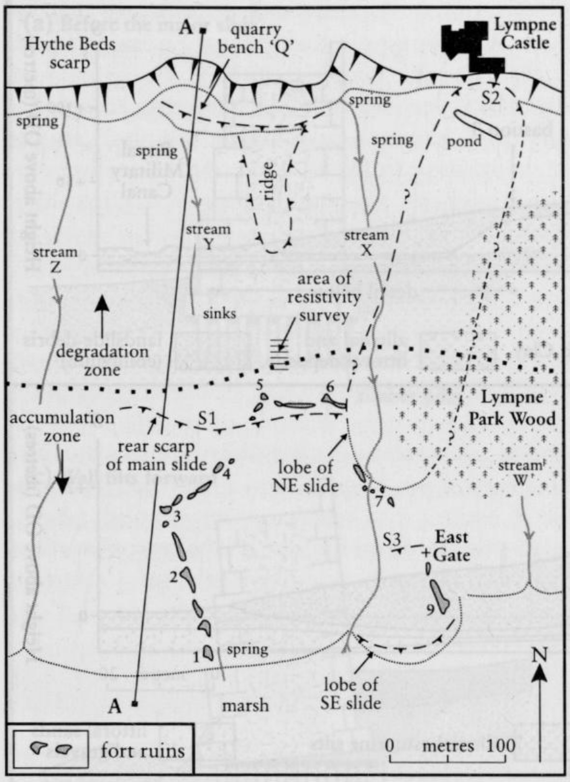


Figure 7.26 Geomorphology of the Stutfall Castle GCR site. The ruins of the fort are stippled. Numbers 1–7 and 9 are Bastion Numbers. After Hutchinson *et al.* (1985).

1. A scarp, S1, which runs beneath the north wall of the fort, cuts across the north-west wall and can be inferred to have transected the north-east wall. This represents the rear scarp of the slide termed the 'main landslide' (Hutchinson *et al.*, 1985), which chiefly disrupted the fort.
2. A lobe that projects well into the marsh between stream X and point K. This is associated with the toe of a 'south-east slide' which caused additional displacement and damage in much of the east wall of the fort and the eastern part of its south wall.
3. A pronounced lobe which encroaches on the north-east corner of the fort, forming the toe of a 'north-east slide'. This appears to cover the eastward continuation of scarp S1 and is therefore inferred to be the later event.

The toe of the slope is fronted by the extensive post-glacial, littoral and alluvial accumulations of Romney Marsh.

Descending the cliff slope at intervals along the slope of 100–150 m, is a succession of streams originating from the spring line near the crest. One of these, 'stream X' (Figure 7.26), runs through the eastern part of the fort ruins.

Interpretation

By analogy with other degraded cliffs (e.g. Hutchinson and Gostelow, 1976) it was expected that the area immediately below the crest of the escarpment would be occupied by the remains of rotational slips, consisting of back-tilted masses of Hythe Beds over Atherfield Clay, and possibly also Weald Clay. Geotechnical investigations (including boreholes) showed, however, that instead the area is occupied by a fairly level bench, Q (Figure 7.26), which a 2.5 m-deep pit showed to consist of 2.2 m of loose loam, with flecks of charcoal, over Hythe Beds debris. Hutchinson *et al.* (1985) concluded that this bench was produced in the course of quarrying of the slipped, and possibly also the in-situ, Hythe Beds in that vicinity for building materials. As Stutfall Castle is constructed predominantly of Hythe Beds material, for which this bench is the nearest accessible source, they consider it likely that the quarrying dates from the construction of the Roman fort.

From the quarry bench to about halfway down the main slope, the ground is mantled by a thin sheet of landslip debris, varying between 1.5 m and 3.5 m in thickness, which shows evidence of part successive-rotational, part translational slipping. This sheet thickens to around 8 m or 9 m from about 30 m downslope of the line of the north wall to just below Bastion 3.

In the main part of the accumulation zone, the landslide debris thickens further to nearly 20 m where it buries a former sea cliff, cut into the in-situ Weald Clay during the last phase of strong marine erosion. The sub-surface investigations define the position and form of this cliff and also show the extent to which the associated beach and alluvial deposits have subsequently been over-ridden by landsliding (Figure 7.27). The crest of the former sea cliff is situated just downslope of the present position of Bastion 3: its foot is a further 35 m downslope. Since erosion ceased there, the landslide debris has advanced approximately 130 m seawards of that point. The buried cliff is fronted by a shore platform, also formed in the Weald Clay. This is sub-planar and declines slightly towards the

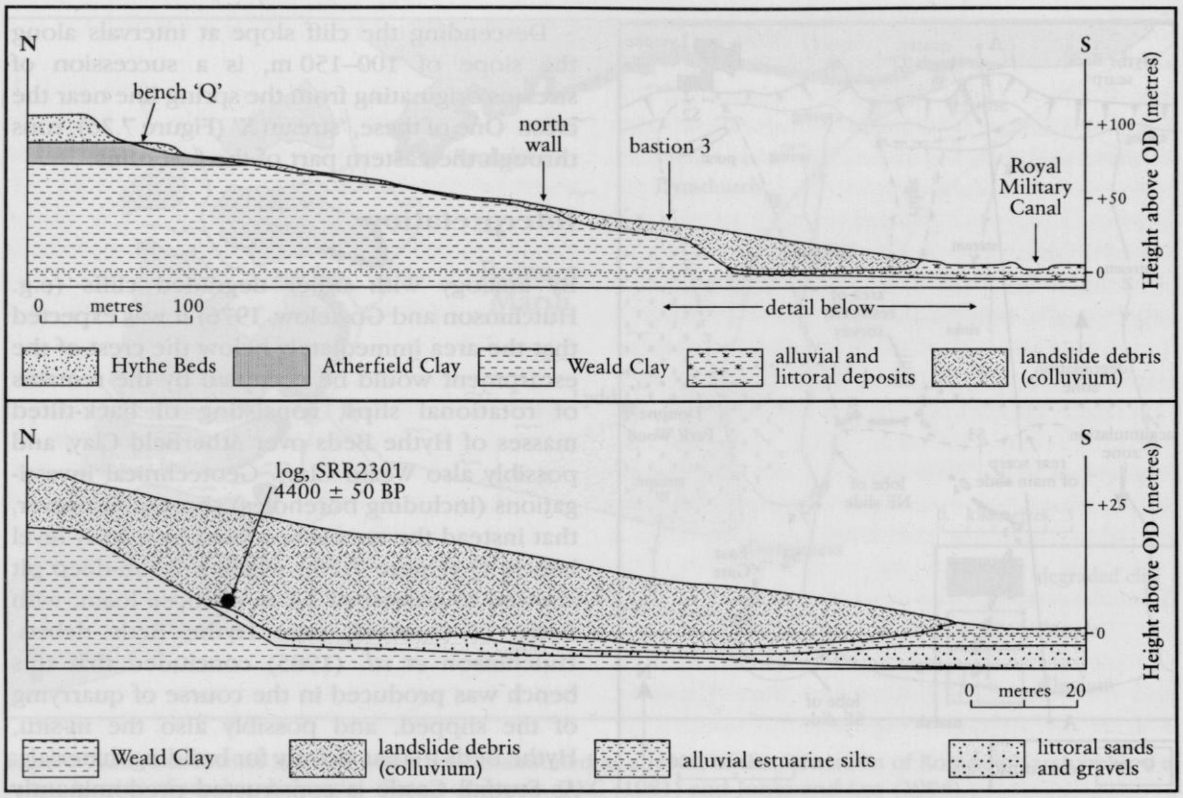


Figure 7.27 Cross-sections through the abandoned cliff at Stutfall Castle. Top – complete section; Bottom – detail of the base of the slope. After Hutchinson *et al.* (1985).

south. At its contact with the buried cliff it has an elevation of 1.0 m below OD. Overlying the shore platform is an irregular spread of alluvial silts, and littoral sands and gravels, often shelly, up to 3.8 m thick (Figure 7.27).

The absence of periglacial solifluction features on the Lympne slope, in contrast to their presence on the geologically analogous escarpment at Sevenoaks, Kent (Skempton and Weeks, 1976) indicates that marine erosion ceased at its foot after the Younger Dryas period of the Late-glacial (10 850–10 050 BP). A radiocarbon date of 4400 ± 50 years BP on a piece of waterlogged wood recovered from the landslip debris just above the foot of the buried cliff (Figure 7.27), suggests that the cliff was abandoned around 4000 BP. Since then the slope has developed chiefly through the burial of the cliff and the re-establishment of the accumulation zone by the supply of debris from the degradation zone upslope. This process must have been relatively far advanced when the Romans chose the site for the fort.

The previous archaeological excavations had not exposed the foundations of the fort. Hutchinson *et al.* (1985) used a tracked excavator to dig eight trenches, each about a metre wide and with one exception they were taken down into the weathered surface of the Weald Clay, 3–4 m below ground level. The northern walls of the fort enabled the investigators to establish the relationship of the Roman foundations to the landslide debris and the in-situ strata. Thirty timber piles were exposed in the floor of one of these trenches, in an area of about 0.9 × 5.0 m. Most of the piles were about 3 m in length. The northernmost of the piles, and those nearest to it, were close to vertical, but those to the south leant successively more and more downslope so that in the southern part of the trench, they occupied sub-horizontal positions. It was evident that the piles, each sharpened to a point at the lower end, had originally been driven and emplaced vertically. The increasing forward tilt of the piles towards the southern end of the trench was produced by the sliding downhill over them of the wall which they had supported (Figure 7.28). The wall slid downslope for a distance of about 5.5 m, tilting forward in the process, and leaving its piles

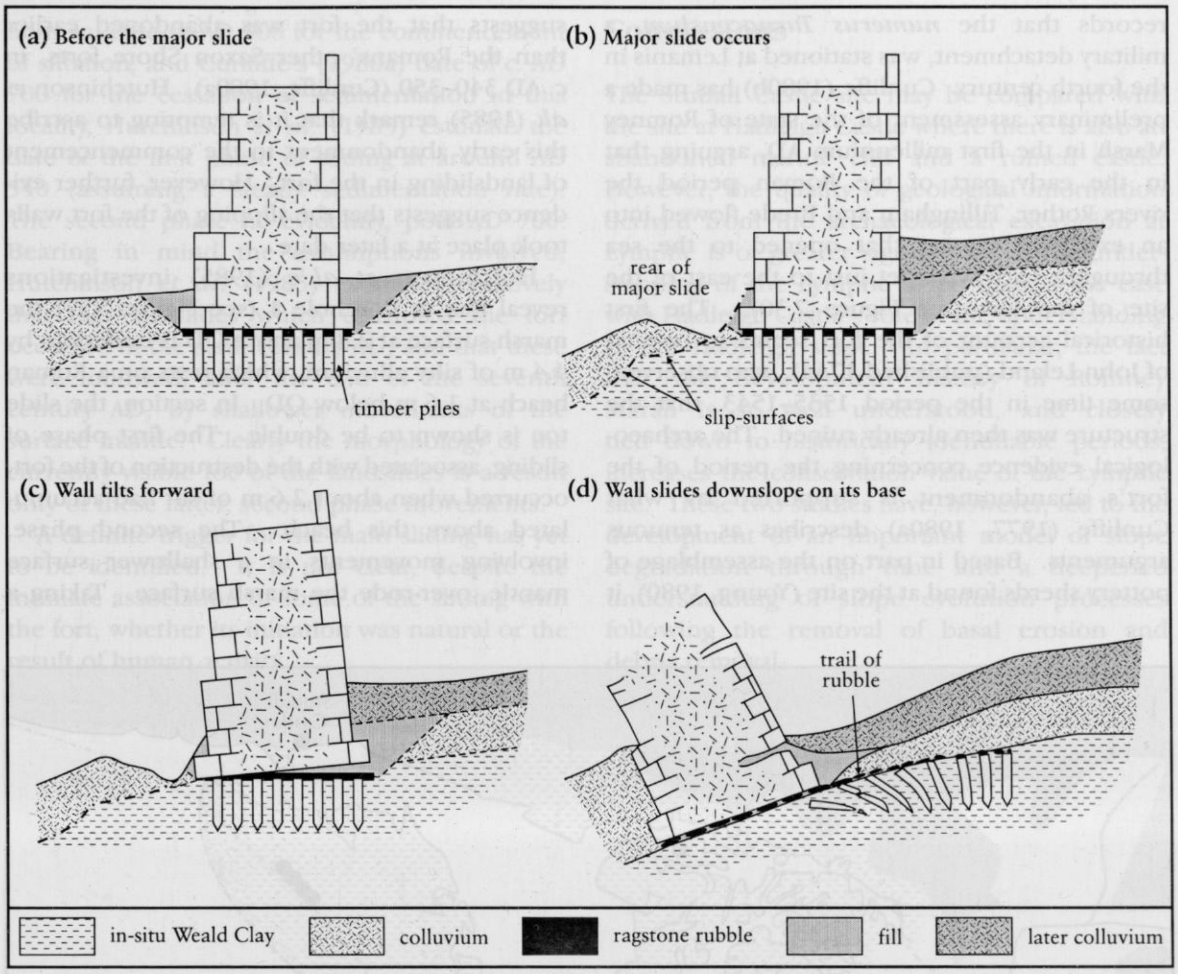
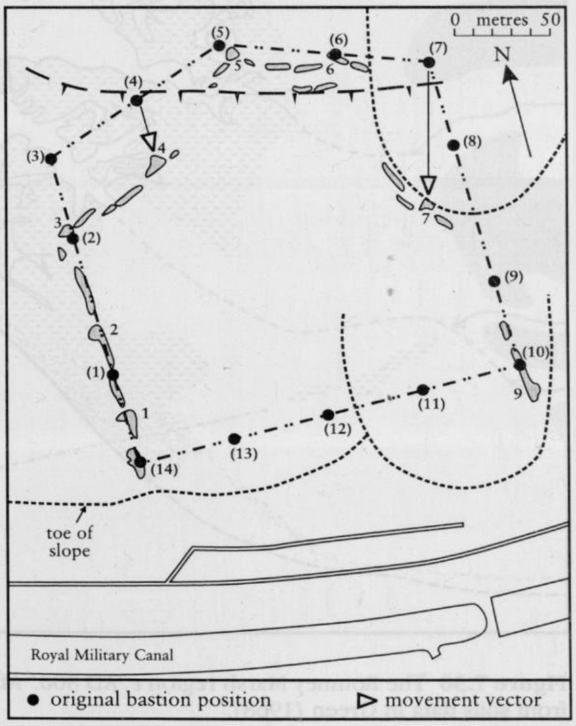


Figure 7.28▲ The inferred mode of failure of the north wall of the fort. After Hutchinson *et al.* (1985).



behind. Similar observations and measurements at other trenches enabled the original positions of the bastions and walls of the fort to be identified, and so the original shape and size of the fort was reconstructed (Figure 7.29).

The date (or dates) of the landslides which disrupted the fort are not easily established, but clearly they post-date the construction of the fort. According to Cunliffe (1980a) two classical texts refer to the fort, enabling it to be identified as *Portus Lemanis*: the Antonine Itinerary, compiled in the early third century, mentions Portus Lemanis, while the *Notitia Dignitarum*

Figure 7.29▸ Reconstructed plan of the Roman Fort (Stutfall Castle) showing the original shape and absolute position of the central parts of the northern walls and the inferred outline of the remainder. After Hutchinson *et al.* (1985).

records that the *numerus Turnacensium*, a military detachment, was stationed at Lemanis in the fourth century. Cunliffe (1980b) has made a preliminary assessment of the state of Romney Marsh in the first millennium AD, arguing that in the early part of the Roman period the rivers Rother, Tillingham and Brede flowed into an extensive estuary that opened to the sea through a narrow outlet just to the east of the site of the fort (see Figure 7.30). The first historical account of the fort seems to be that of John Leland (published 1744), who observed, some time in the period 1535–1543, that the structure was then already ruined. The archaeological evidence concerning the period of the fort's abandonment is supported by what Cunliffe (1977, 1980a) describes as tenuous arguments. Based in part on the assemblage of pottery sherds found at the site (Young, 1980), it suggests that the fort was abandoned earlier than the Romans' other Saxon Shore forts, in *c.* AD 340–350 (Cunliffe, 1980a). Hutchinson *et al.* (1985) remark that it is tempting to ascribe this early abandonment to the commencement of landsliding in the fort. However, further evidence suggests that the slipping of the fort walls took place at a later date.

Hutchinson *et al.*'s (1985) investigations reveal that at Borehole 1 (see Figure 7.27) the marsh surface at 2.8 m above OD is underlain by 4.4 m of silty alluvium, which rests on a Roman beach at 1.6 m below OD. In section, the slide toe is shown to be double. The first phase of sliding, associated with the destruction of the fort, occurred when about 2.6 m of silt had accumulated above this beach. The second phase, involving movements in a shallower surface mantle, over-rode the marsh surface. Taking a

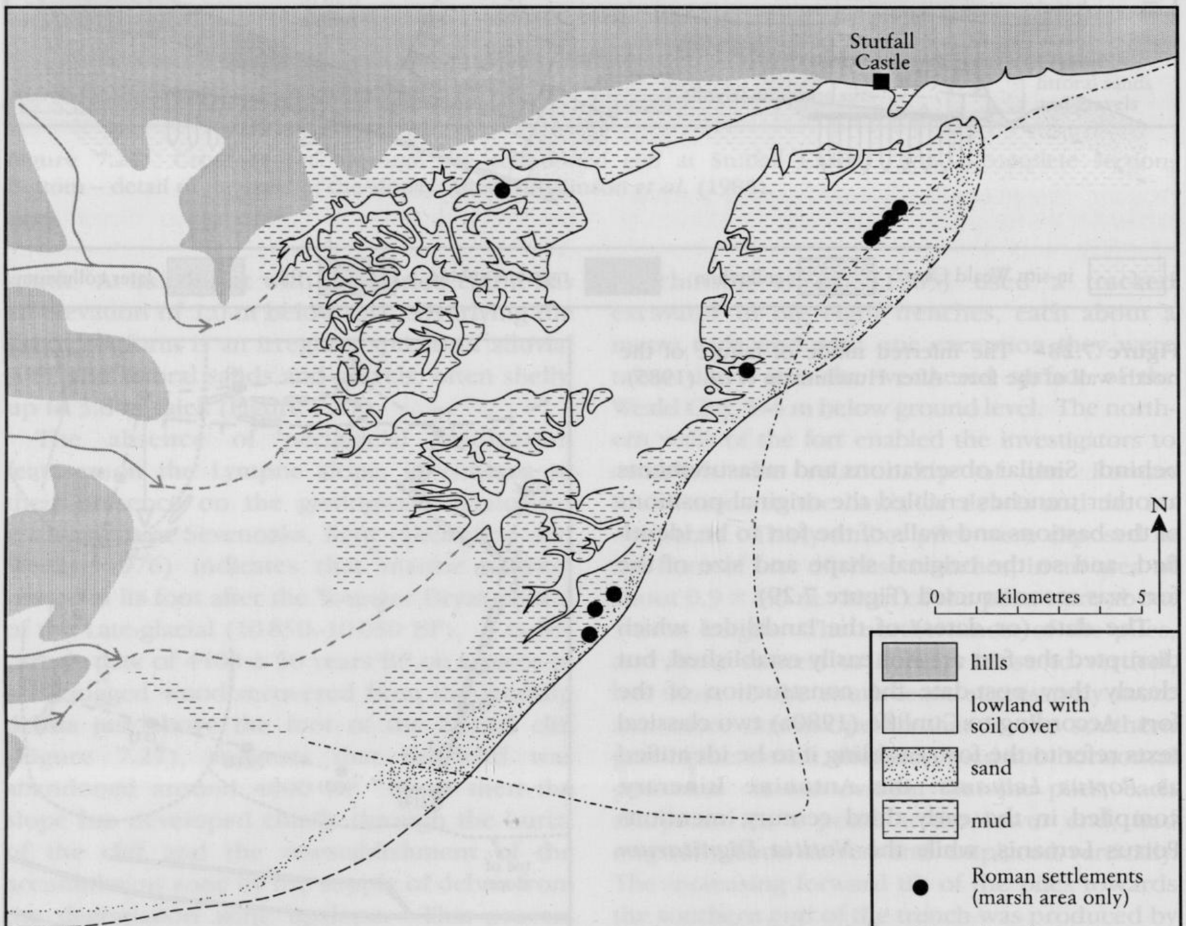


Figure 7.30 The Romney Marsh region *c.* AD 300. After Cunliffe (1980a); details of creeks inferred by Cunliffe from soils data in Green (1968).

nominal date of AD 300 for the commencement of siltation, and Cunliffe's (1980a) date of *c*. AD 700 for the cessation of sedimentation in this locality, Hutchinson *et al*. (1985) estimate the date of the first phase of sliding at around AD 540 (assuming a steady sedimentation rate). The second phase is, evidently, post AD 700. Bearing in mind the assumptions involved, Hutchinson *et al*. (1985) estimate tentatively that the landslides which disrupted the fort occurred in the sixth century AD, and that these were followed, after the end of the seventh century AD, by shallower movements of the surface mantle. Clearly the morphology of the currently visible toe of the landslides is a result only of these latter, second phase movements.

A definite trigger for the main sliding has yet to be identified. It is not clear, despite the intimate association of some of the sliding with the fort, whether its initiation was natural or the result of human activity.

Conclusions

The Stutfall Castle site may be compared with the site at Hadleigh Castle where there is also an abandoned marine cliff and a ruined castle. However, the quality of geological information derived from the archaeological excavation at Lympne is of greater significance in the understanding of the Lympne slope than is the case with Hadleigh Castle's role in the understanding of the Hadleigh slope. In particular, the fact that the development history of Romney Marsh is so well understood, and closely tied down to historically identifiable periods, increases the conservation value of the Lympne site. These two studies have, however, led to the development of an important model of slope degradation through time and a deepened understanding of slope evolution processes following the removal of basal erosion and debris removal.