Landsliding:

Most of the ⊕'s surface is not covered by streams. Streams are merely the conveyer belt of sediment to the oceans.

How do soil & rock get into streams?

- overland flow
- slope wash
- rill & gully incision

} exposed soil, e.g. plowed fields

- rock and soil creep
- solifluction
- rockfalls & avalanches
- landslides — particularly in rugged mountain ranges with rapid erosion rates (Taiwan, Papua New Guinea, Himalaya, Southern Alps of New Zealand) this is the most important erosion agent

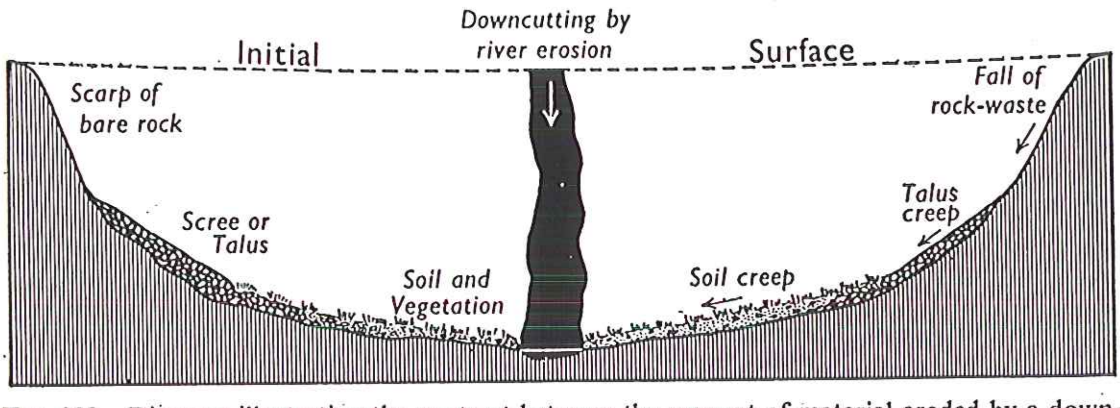
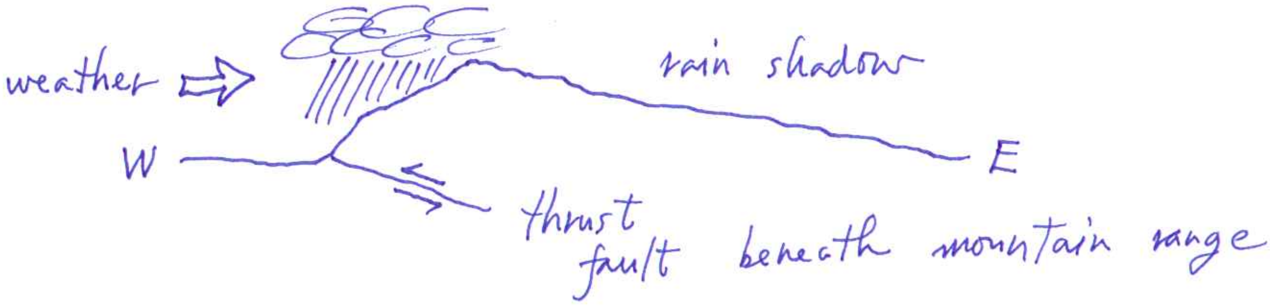
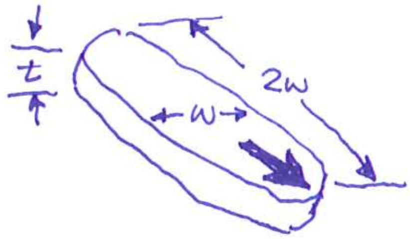


FIG. 323 Diagram illustrating the contrast between the amount of material eroded by a downcutting stream and that supplied to it by the processes that wear back the valley sides

Southern Alps of New Zealand – a case study



Typical landslide scar elliptical in shape:



area $A = \frac{\pi}{2} w^2 = 1.6\, w^2$

thickness $t = \underbrace{(0.04 \pm 0.02)}_{\text{call this } \varepsilon} \sqrt{A}$

Slide volume $V = \varepsilon A^{3/2}$

$$= (0.04 \pm 0.02)\, A^{3/2}$$

Cumulative frequency distribution of landslides:
Exhibits a power-law distribution analogous to Gutenberg–Richter earthquake distribution

$$\boxed{\log_{10} N(\text{with area} \leq A) = a - bA}$$

N ← # per km^2 per year

$a = 5.4 \cdot 10^{-5}\ \mathrm{km}^{-2}\,\mathrm{yr}^{-1}$

$b \approx 1$

Can find landslide erosion rate by adding up volumes of all slides — analogous to finding total energy released by all earthquakes.

area range (km^2)	# per km^2 per year in this range	total # in 2670 km^2 study area per year	total slide volume per km^2 per year from slides in this range [# per km^2 per year $\times \varepsilon A_{\text{middle of range}}^{3/2}$]
$1 \le A \le 1/4$	$3a$	0.5	$15/8\ \varepsilon a \equiv V_{biggies}$
$1/4 \le A \le 1/16$	$12a$	1.8	$\frac{1}{2} V_{biggies}$
$1/16 \le A \le 1/64$	$48a$	7.3	$\frac{1}{4} V_{biggies}$
$1/64 \le A \le 1/256$	$132a$	29	$1/8\ V_{biggies}$
$1/256 \le A \le 1/1024$	$768a$	117	$1/16\ V_{biggies}$
			$\vdots$

The total volume of slide material per km^2 per yr is

$$\dot{e}_{landsliding} = V_{total} = \left(1+\frac{1}{2}+\frac{1}{4}+\frac{1}{8}+\cdots\right) V_{biggies} = 2V_{biggies}$$

$$= \frac{15}{4}\varepsilon a = \frac{15}{4}(0.04 \pm 0.02)(5.4 \cdot 10^{-5})$$

$$= (8 \pm 4) \cdot 10^{-6}\ km/yr$$

$$\boxed{\dot{e}_{landsliding} = 8 \pm 4\ km/Myr}$$

From stream gauging data we find

$$\boxed{5\ km/Myr \le \dot{e}_{streamloads} \le 11\ km/Myr}$$

Conclusion — the rapid erosion rate in the Southern Alps of NZ is dominated by landslide-derived material.

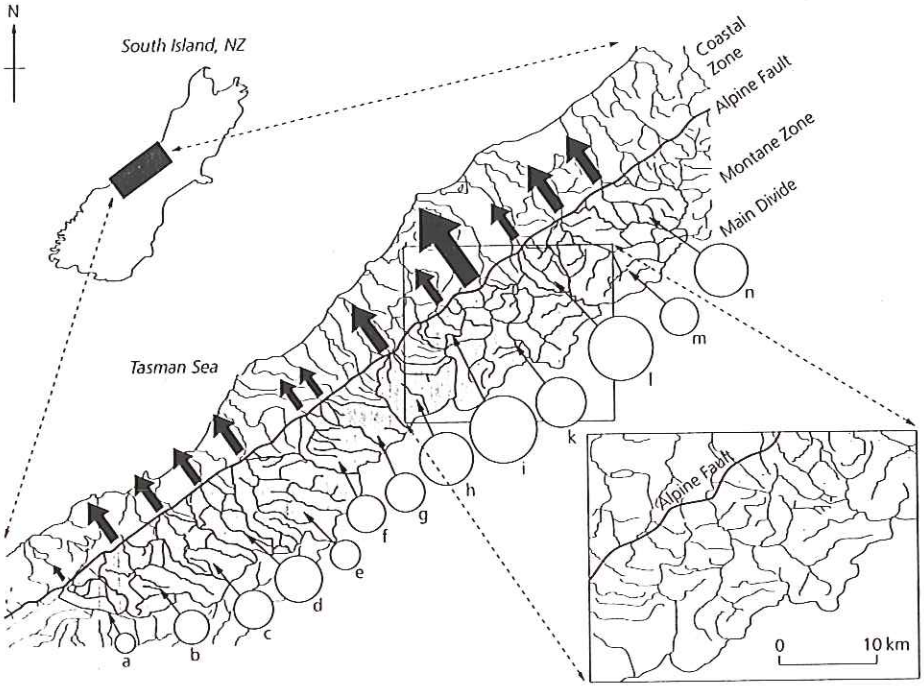


Fig. 3.18 The landslide efflux of mountainous hillslopes, Southern Alps, New Zealand, derived from the same dataset as in Fig. 3.17. Denudation rates of each of 13 catchments draining the western side of the Southern Alps are shown proportional to the areas of circles and vary from 1.8 mm y^{-1} in Moeraki catchment (a) to 18.1 mm y^{-1} in Waitangitaona (j) and Poerua (l) catchments. The sediment discharges are proportional to the size of arrows leaving the catchments, ranging from 1.2×10^5 $m^3 y^{-1}$ in Moeraki catchment (a) to 5.1×10^6 $m^3 y^{-1}$ in Whataroa catchment (k). After Hovius *et al.* 1997 [24].

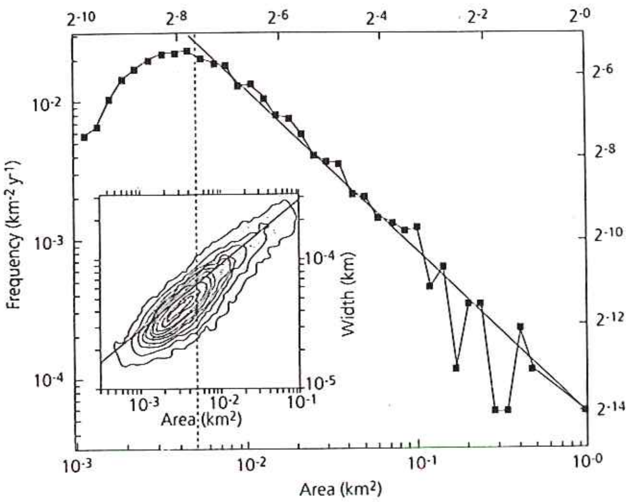


Fig. 3.17 The power law magnitude–frequency distribution of landslides in the Southern Alps (New Zealand), derived from mapping 4984 landslide events east of the Alpine fault which have occurred over the last 60 years. The area to the left of the vertical dashed line is below mapping resolution and should be ignored. The inset graph shows the relationship between the landslide area and the width of the best-fit ellipse, giving information on the plan-view aspect of the landslide scars. After Hovius *et al.* (1997) [24].

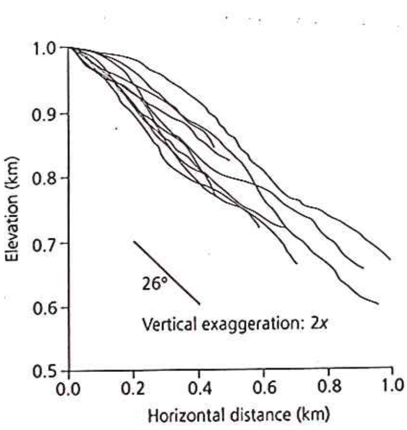
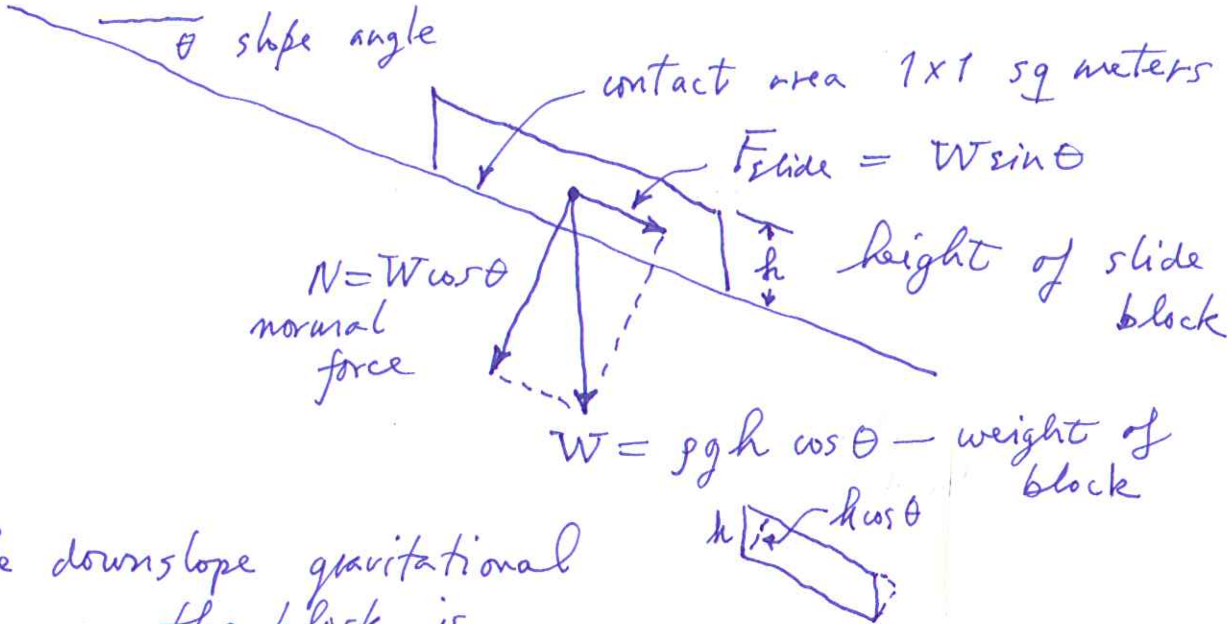


Fig. 3.21 Hillslope profiles from Santa Cruz Mountains. The long straight segments indicate that diffusion is overwhelmed by landsliding. The mean slope is about 26°. After Anderson (1994) [25].

Mechanical analysis of a landslide:



The downslope gravitational force on the block is

$$F_{slide} = W \sin\theta$$

The block will slide if this exceeds the resistive force (note: since we are considering a $1 \times 1\ m^2$ block, all forces are per unit area)

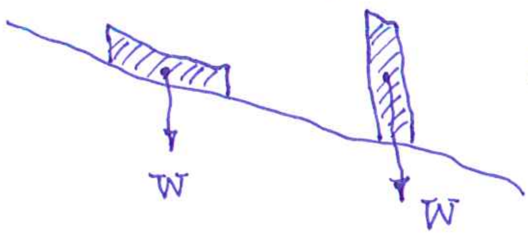
$$F_{resist} = C + \mu N = C + \mu W \cos\theta$$

cohesive force

friction — proportional to normal force

μ = coefficient of friction

Note — friction is independent of contact area!



same frictional force

The slide condition is: $F_{slide} = F_{resist}$

$$W \sin\theta = C + \mu W \cos\theta$$

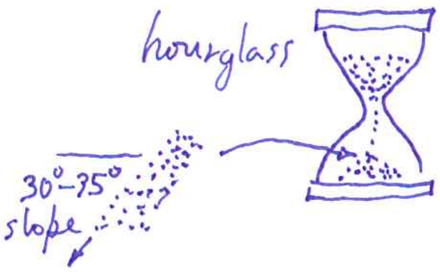
$$\mu = \tan\theta + \frac{C}{W\cos\theta}$$

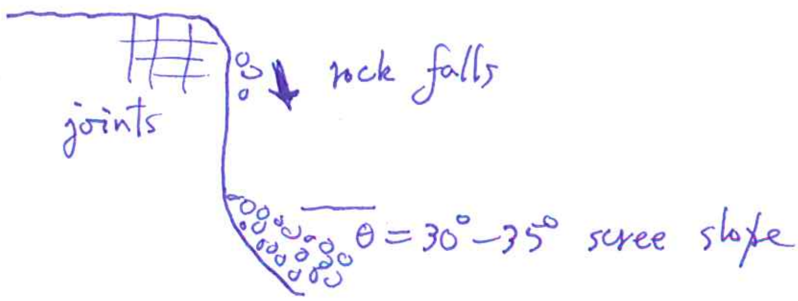
$$\boxed{\mu = \tan\theta + \frac{C}{\rho g h \cos^2\theta}}$$

The cohesion C and coefficient of friction μ are material properties.

Example: dry sand $\mu = 0.6 - 0.7$
$C = 0$ MPa

$$\boxed{\theta = \arctan\mu = 30^\circ - 35^\circ \quad : \text{angle of repose}}$$





μ	θ	
0.85	40°	← Byerlee "maximum" friction — ind. of rock type
0.6	31°	← nominal value
0.4	22°	clay-rich rocks
0.2	11°	

Cohesion highly variable:

Brunswick shale $C \sim 5$ MPa

Stockton sandstone $C \sim 15$ MPa

Rocky Hill diabase $C \sim 100$ MPa

Clay-rich rocks, e.g. fault zones $C \lesssim 0.2$ MPa

Example — Frank slide, Alberta, Canada

$$\theta \simeq 50^\circ \quad \rho = 2700 \text{ kg/m}^3, \quad h \approx 150 \text{ m}$$

Bedding plane slip in a jointed limestone

Cohesion needed to prevent sliding

$$C_{critical} = \rho g h \cos^2\theta \, (\tan\theta - \mu)$$

μ	$C_{critical}$
0.6	1 MPa
0.85	0.5 MPa

A posteriori measured cohesion $C \sim 0.2$ MPa

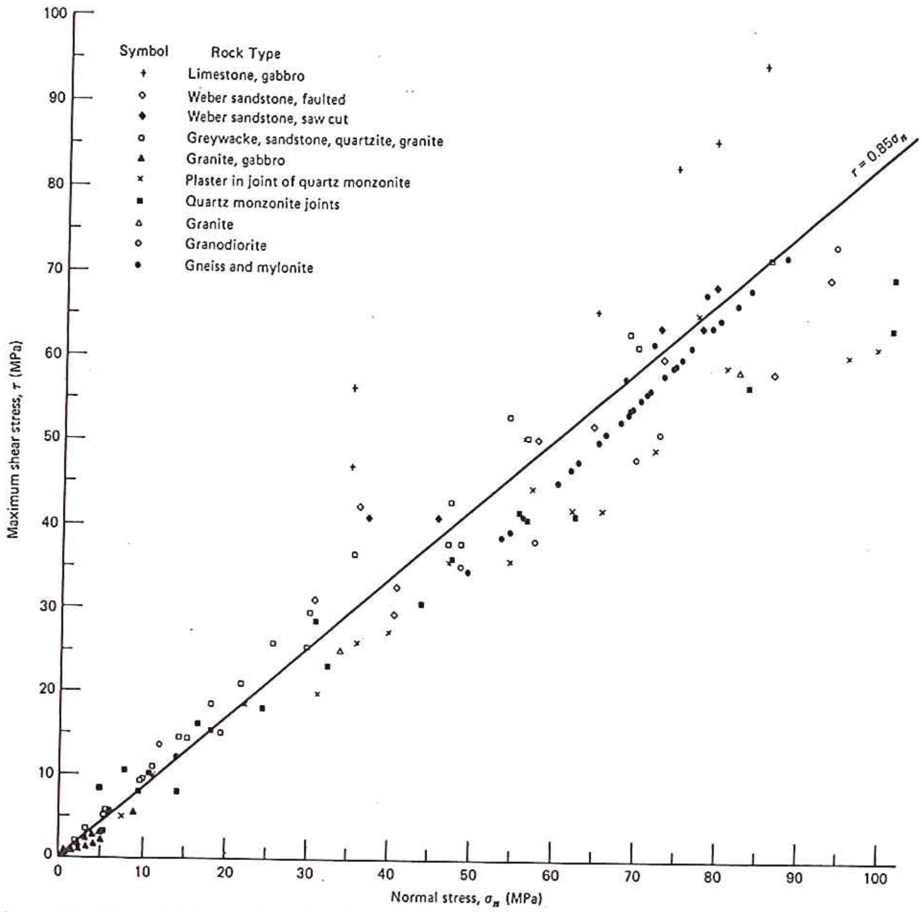


Figure 8-6 Maximum shear stress to initiate sliding as a function of normal stress for a variety of rock types. The linear fit defines a maximum coefficient of static friction max f_s equal to 0.85. Data from J. D. Byerlee, Friction of rocks, in *Experimental Studies of Rock Friction with Application to Earthquake Prediction*, ed. J. F. Evernden, pp. 55–77, U.S. Geological Survey, Menlo Park, Calif., 1977.

FRANK ROCKSLIDE, ALBERTA, CANADA

TABLE I

Summary of shear strength parameters

$\phi = \arctan \mu$

Type of sample	Peak		Ultimate	
	ϕ	c(kN/m^2)	ϕ	c(kN/m^2)
Bedding plane	51.7	262	32.3	55
Flexural-slip surface	28.0	221	15.6	124
Joints	32.0	172	14.0	83
Diamond-saw cut	29.0	0	29.0	0
Surface lapped with 45/80 grit	—	—	37.2	34.5

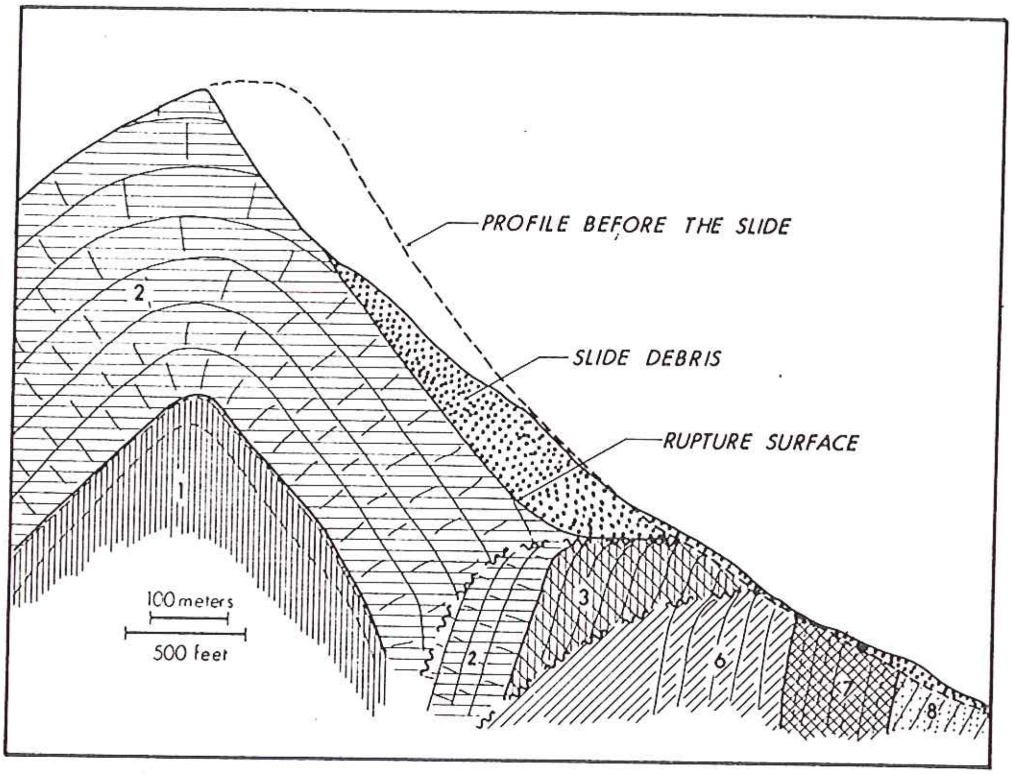


Fig. 8. Cross-section through Turtle Mountain along line C—C' shown in Fig. 5. For legend, see Fig. 5.

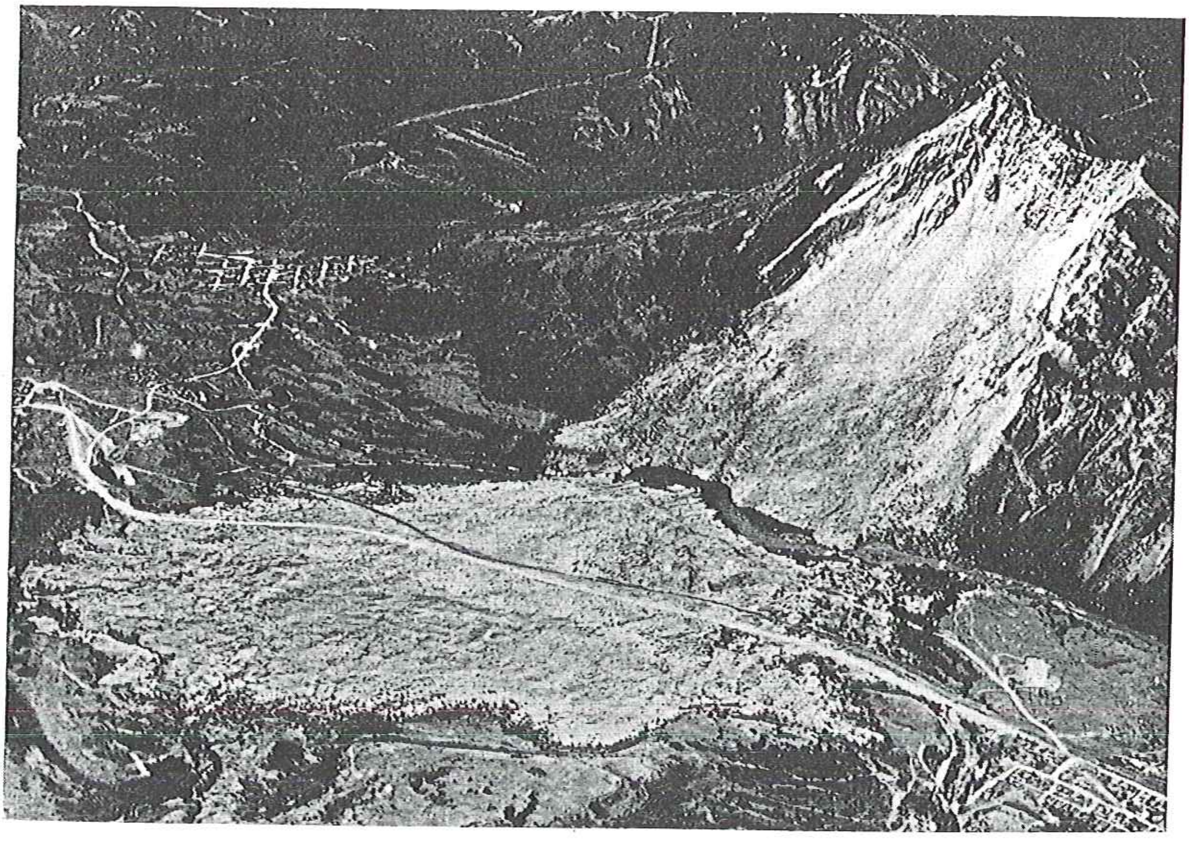


Fig. 9. Oblique aerial view of Frank slide from the northeast.

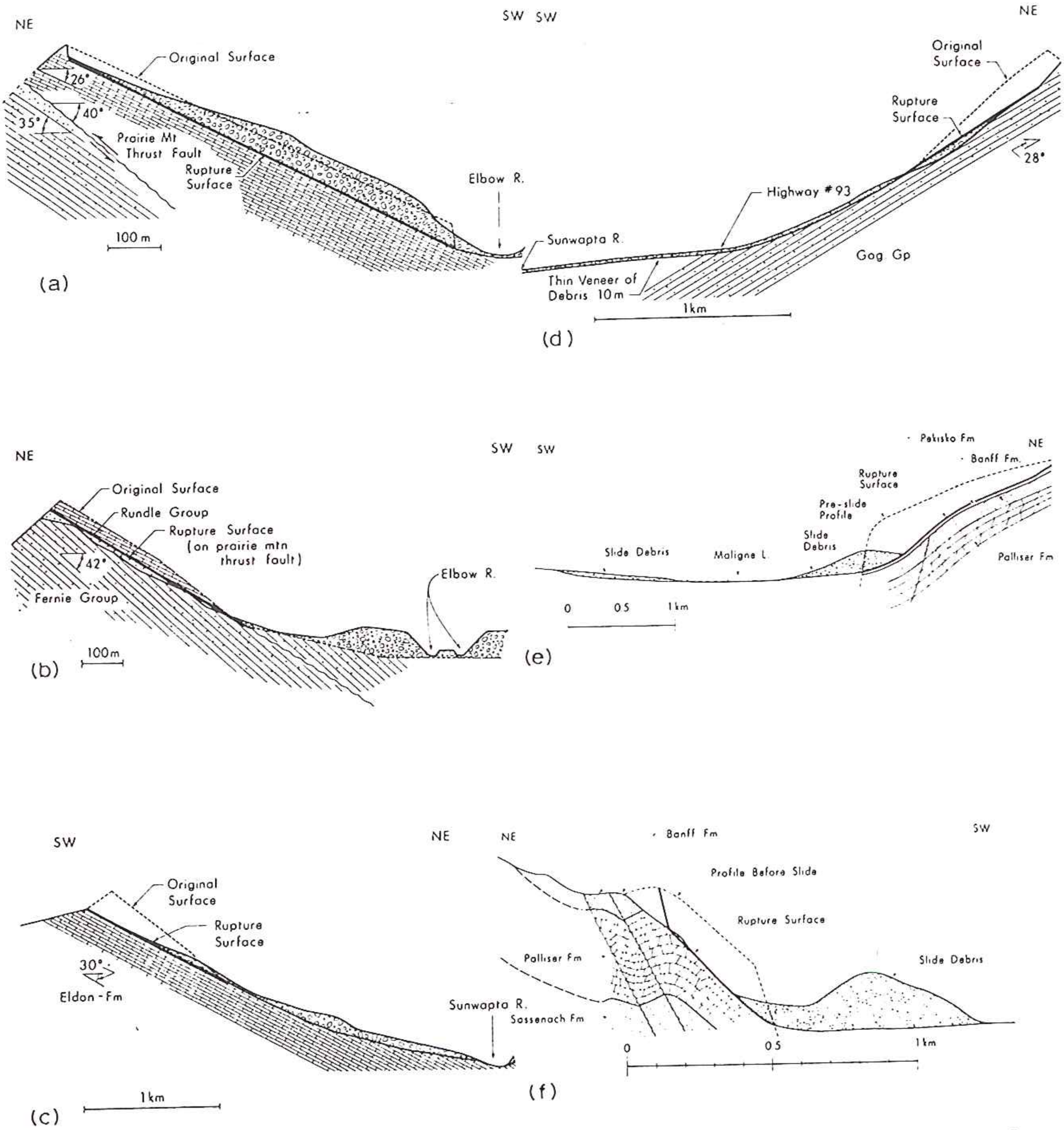
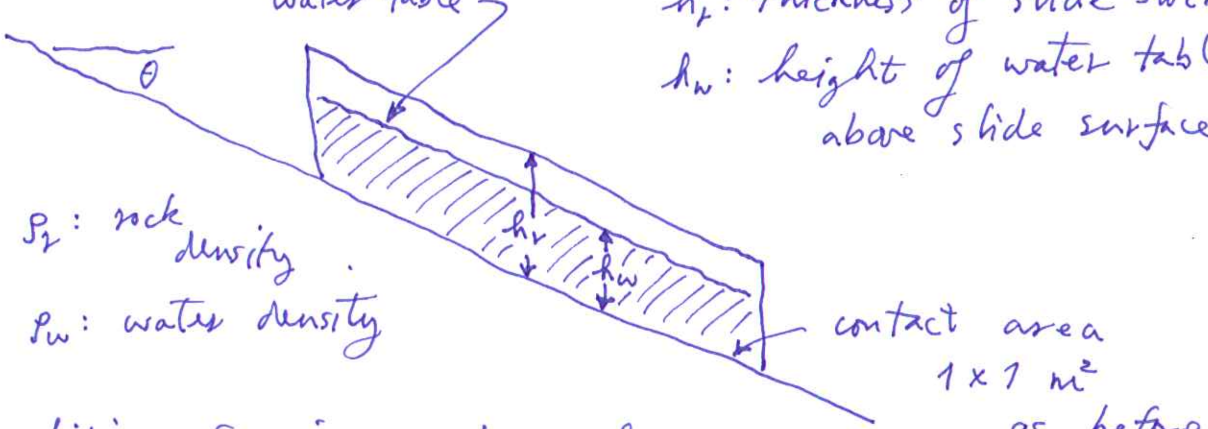


Fig. 14. Sections of major rockslides of the Canadian Rocky Mountains (after Cruden, 1976): (a) Beaver Flats north, (b) Beaver Flats south, (c) Mt. Kitchener, (d) Jonas Creek north, (e) Maligne Lake, (f) Medicine Lake. See Table II for data. Courtesy of the National Research Council of Canada.

Analysis of a wet landslide (Hubbert – Rubey effect)



h_r: thickness of slide block
h_w: height of water table above slide surface

ρ_r: rock density
ρ_w: water density

The sliding force is unchanged:

$$F_{slide} = \underbrace{\rho_r g h_r \cos\theta}_{W} \sin\theta$$

But the effective normal force is reduced by the buoyancy force exerted on the rock by the water:

$$N_{effective} = (\rho_r h_r - \rho_w h_w) g \cos^2\theta = W_{effective} \cos\theta$$

$$W_{effective} = (\rho_r h_r - \rho_w h_w) g \cos\theta$$

The resistive force is $F_{resist} = C + \mu N_{effective}$

Ignoring cohesion ($C \equiv 0$) for simplicity we find:

$$\boxed{\tan\theta = \mu \left(1 - \frac{\rho_w h_w}{\rho_r h_r}\right) = \mu_{effective}}$$

μ	θ_{dry}	$\theta_{saturated}$ ($h_w = h_r$)
0.85	40°	27°
0.6	31°	20°
0.4	22°	13°
0.2	11°	7°

assuming $\rho_r = 2500 \frac{kg}{m^3}$