I.O.S.

AN OUTLINE REVIEW OF MEDIUM FREQUENCY WAVES

J M HUTHNANCE

REPORT NO. 81

1979

NATURAL ENVIRONMENT RESEARCH COUNCIL

INSTITUTE OF OCEANOGRAPHIC SCIENCES

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CONTENTS

SUMMARY

Motions with frequencies in the range 10^{-4} to 10^{-2} Hz appear to be dominated by edge waves constrained by local topography. Some localities, notably partially enclosed sea areas, may be especially responsive at particular frequencies. Travelling pressure and wind disturbances are known to be effective in generating edge waves, and surf beat is a well-established phenomenon at higher frequencies in the range. Internal waves and tidal stream instabilities merit further investigation as sources of medium-frequency motion.

1. INTRODUCTION

Large structures (eg some schemes for wave-energy extraction) moored in seas of depth 100m or more have natural oscillation periods of a few minutes. Although these seem most likely to be excited by varying radiation stress due to swell (see 4c below), any motion of such periods is of interest. This is particularly so around northern Scotland, the principal region envisaged for wave-energy extraction, where the presence of oscillations with periods of some minutes is already well-known in tide gauge records (see 4a below).

The 'medium frequency' range 10^{-4} to 10^{-2} Hz (order of magnitude limits only) includes such 'local seiche' phenomena and many others, including surf beats, edge waves, tsunamis, internal waves and tidal stream instabilities. Wind waves or swell, at higher frequencies, and tides, at lower frequencies, are usually more energetic but are excluded here. The medium frequency grouping is convenient, since for such motions on continental shelves the sea is effectively shallow and the earth's rotation is unimportant (see 2b below). Amplitudes at many sites are low, typically 1cm, and there are consequent instrumentation problems (Munk, 1962). This seems to have resulted in a certain lack of interest (as compared with wind waves or tides), except in the particular field of tsunamis. However, some understanding has been gained since the previous reviews by Munk (1962) and Tucker (1963).

2. SHALLOW WATER AND EDGE WAVES

(a) Archetype.

A progressive plane wave form $\zeta(x-ct)$ advances with speed $c=(gh)^{\frac{1}{2}}$ in water of uniform depth h, provided that $\zeta \ll h \ll$ wavelength and that friction and the earth's rotation are negligible. The forward velocity is $u = c\zeta/h$, uniform in depth and in constant relation to the elevation ζ everywhere beneath the progressive wave. There is no transverse velocity: $v=0$. The pressure is hydrostatic.

(b) Modifications.

Relaxation of the conditions in (a) modifies the progressive wave form. We note the main individual effects of the following on a wave of period T (frequency $\sigma = 2\pi/T$) in 100m of water:

(i) rotation (Coriolis parameter f; eg inertial period 15 hr) causes transverse velocities v in a plane wave ($v/u = O(f/\sigma) > 10\%$ if $T > 1\frac{1}{2}$ hr) and increases the phase speed c (by a fraction $O(f^2/2\sigma^2) > 10\%$ if $T > 7$ hr);

(ii) bottom friction (eg $\rho C U u$ where U(tide) = 20cm/sec, $C=3\times10^{-3}$) causes wave decay (the rate CU/2h means a 10% reduction in 9 hours) and a bottom boundary layer (thickness $O(\kappa u_* /\sigma = 0.4\, UC^{\frac{1}{2}} T/2\pi) > 10$m if $T > 4$hr);

(iii) if $h \not\ll$ wavelength, the velocity and perturbation pressure decrease with depth together (the bottom pressure perturbation $\sim \rho g \zeta/\cosh(\sigma^2 h/g)^{\frac{1}{2}}$, which is 10% below the hydrostatic value $\rho g \zeta$ if $T < 45$ sec) and there is a smaller reduction in wave speed;

(iv) finite elevation amplitude a removes the arbitrariness in the wave form ζ.

A progressive wave train (cnoidal waves) is 'peaky' with broadened troughs. However, the wave speed given by $c^2 = g(h+a)$ is increased by 10% only for $a > 21$m, and the forward velocity is still essentially $u = \zeta(g/h)^{\frac{1}{2}}$, uniform in depth.

The parameters in this section have been chosen for relevance on the Scottish continental shelf. In the medium frequency band we can generally expect effects of rotation, bottom friction and finite amplitude to be small.

(c) Edge waves

Although the shallow water approximation appears to be valid by (b.iii) above, depth variations play a crucial role. In 100m of water, the wave speed is $(gh)^{\frac{1}{2}} \triangleq 1.88$ kilometres per minute. That is, wavelengths in the medium frequency band correspond to most coastal and continental shelf length scales.

Considerable amplifications may occur locally in consequence. For the simple step-shelf model in figure 1, Proudman (1925) found that a plane wave, normally incident from the deep water with amplitude a, induces elevations at the coast as great as $2a/\Delta^{\frac{1}{2}}$ when the shelf width L is near the quarter wavelength 'organ-pipe resonance' condition, ie $L \triangleq (g\Delta h_0)^{\frac{1}{2}} T/4$. A finite length of such shelf, ie. an embayment (figure 2), may be even more responsive. When the relative length ℓ is moderate ($\ell\Delta < 0.1$ say), the corresponding amplification factors have maxima $4/\pi\ell\Delta$ for the organ-pipe mode and $\pi/\ell\Delta^2(1+\ell^2/4)$ for the 'sloshing mode' with opposite ends of the length ℓL of shelf in antiphase (Huthnance, 1980). The factor Δ^{-2} may cause the latter to be particularly large, but the response is correspondingly highly tuned and susceptible to frictional damping. Naturally, tides form the best documented examples. M_2 tides in the Bristol channel and M_3 on the broad Brazilian shelf west of Rio de Janeiro illustrate the organ-pipe mode, and M_2 on the Argentinian shelf appears to be a sloshing mode modified by the earth's rotation. However, one may expect higher frequencies associated with smaller-scale topography to be more common, with less frictional energy loss per cycle.

Any profile of depth which increases offshore and is uniform alongshore supports a set of trapped 'edge waves' (Huthnance, 1975). The n'th mode (n = 0,1,2,..) has n offshore nodes and a dispersion relation $\sigma = \sigma_n(k)$ of frequency as an increasing (usually) function of longshore wavenumber k; σ_n increases with n. Propagation is the same in either sense along the shelf if the earth's rotation is negligible. There is a minimum 'cut-off' frequency and wavenumber where the wave-speed $|\sigma/k| = (g h_0)^{\frac{1}{2}}$, provided the depth increases to a finite maximum h_0 offshore. The edge waves are in $|\sigma/k| \ll (g h_0)^{\frac{1}{2}}$; in $|\sigma/k| > (g h_0)^{\frac{1}{2}}$ trapping is impossible but there is a continuum of waves exchanging energy with the sea far offshore. Figure 3 illustrates these features for the profile $h = h_0(1-\exp(-ax))$ (Ball, 1967), which is a useful model for a coastal beach merging into a level continental shelf. A submarine ridge (Buchwald, 1968) and an island (Summerfield, 1972) support similar sets of waves, although those for an island slowly lose energy by radiation to the distant sea. Edge waves may be regarded as trapped by shoreward refraction in the deeper water offshore, the discrete forms being selected by a requirement of reinforcement when the refracted wave returns to the coast (Shen, Meyer and Keller, 1968).

3. OBSERVED SPECTRA

Munk, Snodgrass and Tucker (1959) surveyed much of the data then available, and concluded that various sites have spectral peaks at frequencies characteristic of the site, although the overall energy level varies. For example, tsunami response is apparently site dependent more than tsunami dependent. The characteristic frequencies vary from $O(10^{-2}\text{Hz})$ at Guadalupe, which has a narrow steep shelf, to $O(10^{-4}\text{ Hz})$ at Mar del Plata on the broad, shallow

Argentine shelf (Inman, Munk and Balay, 1961). The latter site is particularly energetic with elevation amplitudes of 5-10cm.

The Southern California shelf has relatively little medium frequency energy, with typical elevations of only 1 cm. However, Munk, Snodgrass and Carrier (1956) have found that energy levels at Scripps and Oceanside, 38 km apart, vary together, although the waves are not coherent between the two sites. From the offshore amplitude variation they interpreted the 2-5 min period motion as surf beat (see 4c below) and the 10-30 min period motion as edge waves. By synthesising co-spectra from pairs of records with various longshore separations, Munk, Snodgrass and Gilbert (1964) later found that over 90% of the energy with frequencies below 10^{-2} Hz fell along edge wave dispersion curves $\sigma = \sigma_n(k)$ (illustrated in figure 3c). Hence almost all the energy was trapped on the shelf rather than in exchange with the deep sea. At Hell's mouth in North Wales (4°35'W, 52°50'N), Huntley (1976) found energy peaks in the range 3×10^{-3} to 3×10^{-2} Hz around the low wavenumber cutoff frequencies of the modes (see figure 3c).

This all suggests that local topography is important in determining the energy characteristics of medium frequency motion via the edge wave dispersion relations.

4. SOURCES OF MEDIUM FREQUENCY WAVES

(a) Seiches

Harbour oscillations are an obvious example of local geometry controlling medium frequency motions. Cape Town (Tucker, 1963) and Port Kembla (Clarke, 1974) are well known for their energetic seiching: ships sometimes have to put out to sea from the latter. A review is given by Miles (1974).

Harbours are an extreme case in that the (usually) narrow entrance implies fine tuning, determined by the harbour geometry, and large responses at the selected frequencies. However, section 2c indicated the possibility of quite large responses at comparatively open locations. This certainly appears to be realised at many places around the British Isles in the form of a 'local seiche' evident in their tide gauge records. The oscillations at Lerwick, with a period of about 28 min, have typical amplitudes of a few centimetres, but up to 40 cm amplitude has been recorded (Rossiter, 1971). They have been interpreted by Cartwright and Young (1974) as a standing combination of edge waves along the east side of Shetland; oscillations of similar period occur further north at Balta sound in Shetland. George (1977) has studied oscillations at Plymouth, where amplitudes exceeded 1m following a line squall on 3rd July 1946. At Newlyn, Cornwall, there are energy peaks at wave periods of about 12 minutes and 30-40 minutes (Darbyshire, 1958). The latter's amplitude reaches 20 cm a few times each year, following passage of an atmospheric front with a particular northward velocity component. An oscillation of most of Mount's bay is probably responsible.

Similarly, Swansea tide gauge shows oscillations, with a period of 30-50 minutes and amplitudes up to 20 cm, an hour or so after sudden changes of wind (A. Wilding, personal communication). Tide gauge records held at Bidston (S. Brown, personal communication) show persistent oscillations with periods of about 25 minutes at Stornoway and Ullapool and 12 minutes at Wick and Buckie. Table 1 shows typical and maximum ranges over a one month winter period 21st February to 20th March 1975, chosen only for easy availability of the records. Fair Isle (1°38'W, 59°32'N) shared neither the Wick nor Lerwick oscillations during that period, although its occasional oscillations of about 5 minutes period also had a maximum range of about 13cm on 21st-22nd February.

It appears that seiches with amplitudes of 10-20 cms are quite widespread under appropriate but reasonably common meteorological conditions. Those with periods of 10 minutes or more can represent oscillations extending for some tens of kilometres, e.g. across the Minch (Stornoway - Ullapool) or Moray Firth (Wick - Buckie).

(b) Atmospheric forcing.

The seiches above almost certainly result from forcing by atmospheric pressure and wind stress fluctuations (local geometry then selecting those frequencies to which it responds best). The quasi-equilibrium 'inverse barometer' response is a sea surface rise of about 1 cm beneath an atmospheric pressure low of 1 mb. However, the dynamical response to a plane wave pressure disturbance $-F(x-Ut)$ travelling with speed U is a sea surface elevation $F(x-Ut)/(1-U^2/gh)$ (Proudman, 1953). This is large for disturbance speeds U nearly matching the shallow water wave speed $(gh)^{\frac{1}{2}}$.

More recent calculations are essentially refinements of the same idea. Hurricane Carol, travelling along the US east coast on 26th August 1954, was followed by sea level changes of 60 cm amplitude and 6 hr period (Munk et.al, 1956). Greenspan (1956) modelled this as an edge wave response (on a shelf of slope α) to a cyclone with a Gaussian form of pressure perturbation (½-pressure radius a) moving with speed U along the shelf. The parameter $\alpha g a/U^2$ relates the cyclone radius to the offshore length scale of the zero order edge wave (the main response). Amplifications (ie maximum sea level elevation, cm/maximum pressure perturbation, mb) of approximately

5,3,2,1 result if $\alpha g a/U^2$ = 2.4, 3, 3.6, 4.2 approximately. Cyclones appear generally to be too large in horizontal extent for maximum effect. Buchwald and de Szoeke (1973) calculated the edge wave wake following a pressure front travelling along a step-shelf (figure 1). When Δ = 0.05, the amplification represented by the amplitude of mode zero is 2.1, 2.8, 8, ∞ for $U/(g\Delta h_0)^{\frac{1}{2}}$ = 2,3,4, $\sqrt{20}$ (the last corresponding to $U = (gh_0)^{\frac{1}{2}}$ and resonance by the Proudman result).

We conclude that edge waves may respond strongly to individual travelling pressure disturbances. Phillips' (1957) theory of wind-wave generation by travelling turbulent pressure fluctuations gives the amplification (in the same sense as above) represented by the root-mean-square sea surface elevation response as

π (duration of forcing in wave periods)$^{\frac{1}{2}}/2^{\frac{1}{4}}$.

Although this formula is derived for short waves (wavelength $\ll$ water depth) it appears to hold without essential change for shallow water waves with frequency O(10^{-2}Hz), and again predicts fairly large amplifications.

No analogous calculations appear to have been carried out for travelling wind stress fluctuations. However, a numerical model has simulated oscillations of 50 cms amplitude along the Dutch coast following a cold front moving at 33 knots ($\simeq (gh)^{\frac{1}{2}}$) and its associated change in direction of 30 knot winds from 200° to 270° (Timmerman, 1971).

(c) Swell.

Munk (1949) and Tucker (1950) recorded sea floor pressure variations ('surf beat') a few hundred metres offshore. Periods were a few minutes and amplitudes near 10% of the swell amplitude - a <u>linear</u> relation. Correlation with the envleope (ie wave groups) of the shoreward propagating swell was best if the latter was lagged by the return travel time $\int (gh)^{-\frac{1}{2}} ds$ to the shore, and indicated a lower mean level ('set-down') at wave group maxima. Set-down is a theoretical result of Longuet-Higgins and Stewart (1962), but is there predicted to be proportional to the <u>square</u> of the swell amplitude, mean surface slopes balancing variations of the (quadratic) mean wave-momentum flux. Gallagher (1971) proposed a mechanism for the generation of edge waves as the set-down and mass and momentum flux vary through wave groups; his mechanism has been verified in recent laboratory experiments (Bowen and Guza, 1978).

Varying wave-momentum flux and set up will directly exert a changing force (and probably the largest) on any obstacle such as a wave power device which affects the progressive swell. For normally incident, reflected and transmitted wave amplitudes a_I, a_R and a_T respectively, Longuet-Higgins (1977) found the force per unit length to be $\frac{1}{4}\rho g(a_I^2+a_R^2-a_T^2)\beta$, where $\beta \equiv 1 + 2kh/\sinh 2kh$ lies between 1 and 2 according to the swell wavenumber k. If a_I= 1m and $a_R^2 = a_T^2$, this is ¼ to ½ tonnes weight per metre.

Instability of the swell may also cause lower frequency motion. A wave train in <u>deep</u> water, with amplitude a and wavenumber k in $0 < ak <$ 0.346, is known to be unstable to subharmonics of sufficiently low frequency (Longuet-Higgins, 1978). Furthermore, swell approaching a beach may form a resonant triad with two edge waves. For normal incidence, Guza and Davis (1974) found that edge waves at half the swell frequency were preferred, corresponding with observations by Huntley and Bowen (1973) on Slapton beach, Devon. They recorded edge waves of 10 sec period while the incident wind waves had a 5 sec period. Obviously, such frequencies are above our range, but Guza and Bowen (1975) extended the analysis to oblique incidence. Then one of the edge waves has less than half the swell frequency.

(d) Tsunami

Tsunamis from the Lisbon earthquake of 1755 caused oscillations of 1m amplitude at Plymouth (George, 1977). However, they are so rare in British waters that we simply refer to the large body of literature arising from their more frequent and sometimes devastating occurrence in the Pacific (e.g. Murty, 1978). Nevertheless, an 'afterglow' which can remain for days after the first arrival and has a site-dependent energy spectrum (Munk, 1962) suggests that some energy becomes trapped in local edge wave forms as already discussed for meteorological forcing.

(e) Ice calving

Floating 'elastic' glaciers, and icebergs calving from them, oscillate vertically with periods of a few minutes (Reeh and Engelund, 1971); $2\pi(89D/11g)^{\frac{1}{2}}$ for an iceberg of total depth D and density 89% of the seawater density. Oscillations arise as an iceberg is released on calving from a position below (by an amount d, say) its equilibrium floating level. Much of the potential energy release $\frac{1}{2}(0.11)\rho g d^2$ per unit area is eventually radiated away as water waves of the same period which are large nearby ('kaneling').

Greenland is the main source of such wave energy in the North Atlantic. An order of magnitude calculation based on an annual ice volume V = ('rainfall') (area of Greenland) = O(10cm x 2 x10^6 km^2 = 200 km^3) gives a rate of energy release $\frac{1}{2}$ $0.11\rho g d^2 V/D$ per annum, or 700 kilowatts if d = 10m, D = 500m. Distribution along 10^4 km of coastline as a shoreward progressing wave in 100m of water (to represent the situation near coasts other than Greenland) gives an amplitude of 0.66 mm. This is small enough to be neglected despite the rough estimation, particularly since the approach in the calculation is the basis of an upper bound on the wave energy.

(f) Internal waves

In the ocean, internal waves may have frequencies between f, the inertial frequency, and the maximum Brunt-Väisälä frequency $N = (-g/\rho \; d\rho/dz)^{\frac{1}{2}} = O(10^{-3}\text{Hz})$ in a thermocline. Garrett and Munk (1972) roughly estimate the typical internal wave energy density as 0.382 joules/cm^2 distributed through the ocean depth. A rough fraction $1-(2/\pi)\sec^{-1}\sigma/f \triangleq 12\%$ is in frequencies exceeding 10^{-4} Hz if f = 1 cycle/15 hrs (Garrett and Munk, 1975). In other words, medium frequency wave energy in the ocean is O(500 joules/m^2). This represents current amplitudes of 3 - 5 cm/sec in the more energetic surface layers. It seems reasonable to expect these currents to extend onto the shelf whether or not the shelf sea is stratified. They have certainly been observed in the stratified shelf water of N.W. Africa, as internal waves (Gordon, 1978). Since barotropic currents of this size in 100m of water correspond to surface elevations of 10 - 16 cms, the possibility seems to warrant closer examination.

(g) Tidal stream instability

We have already mentioned large oscillations of tidal frequency associated with coastal features (section 2c). Smaller scale features such as abound off northern Scotland may also be associated with tide races, recirculating eddy patterns and tidal streams resembling free jets with slower-moving water on either side. Yell Sound (1°15'W, 60°30'N) in Shetland is a good example (Blackman, Graff and Vassie, 1979). The northward tidal stream, having emerged from a constricted channel, is jet-like with a speed of at least 1½ m/sec, but may show non-tidal fluctuations with amplitudes up to 1 m/sec and time scales of an hour or less, particularly at spring tides. Such apparent instabilities (detached eddies?) are suggestive of river flow between piers of a bridge. This example is probably unusual only in being well-documented.

5. CONCLUSIONS

Frequencies of 10^{-2} to 10^{-4} Hz correspond to wavelengths of 3 to 300 km in 100 m of water, typical of British shelf seas. Since these scales match coastal and shelf features, local topography strongly influences medium frequency motion. Even along an open straight coast, increasing depth offshore enables energy to be trapped as edge waves. Longshore variations and partial enclosure of sea areas may further assist energy trapping. A wealth of observations bears this out, and indicates energy levels varying widely from place to place. 'Background' levels in the absence of storms are O(1-10 cm) elevation amplitude, according to position.

Atmospheric pressure and wind stress fluctuations are effective in generating edge waves, particularly when travelling near to the edge wave speed. This is an average value of $(gh)^{\frac{1}{2}}$ over the spatial extent of the edge wave, and is therefore readily matched by atmospheric fronts over shallower shelf waters. Swell propagating shorewards with varying amplitude generates motions at the 'beat' frequency, and sometimes also at other frequencies at the top end of the medium frequency range owing to instabilities. We have essentially discounted tsunami and ice calving contributions in British waters. However, deep ocean internal wave motions are energetic enough to make a significant contribution if their energy extends onto the shelf. Tidal current fluctuations may be important locally.

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Table 1

Medium frequency oscillations recorded on tide gauges from 21/2/75 to 20/3/75.

Location	Longitude (W)	Latitude (N)	Range, cm typical	Range, cm maximum	Date of maximum
Stornoway	6°23'	58°12'	5	20	6/3
Ullapool	5°10'	57°54'	3	13	6/3
Wick	3°6'	58°26'	10	30	22/2
Buckie	2°58'	57°40'	8	17	21/2, 24/2

CAPTIONS

Figure 1. Step-shelf profile

Figure 2. Embayment plan

Figure 3. (a) Depth profile $h = h_0(1 - exp(-ax))$

(b) Elevation forms for edge wave modes $n = 0, 1, 2$ when $k/a = \pi$.

(c) Dispersion curves for edge wave modes $n = 0, 1, 2$

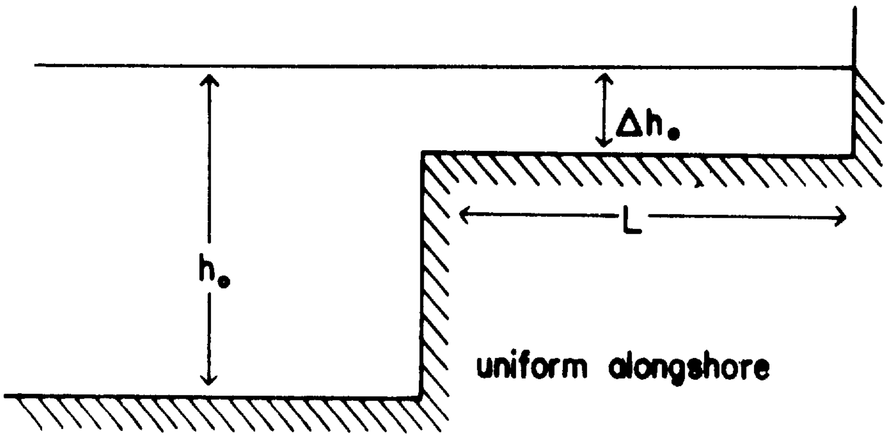

Figure 1

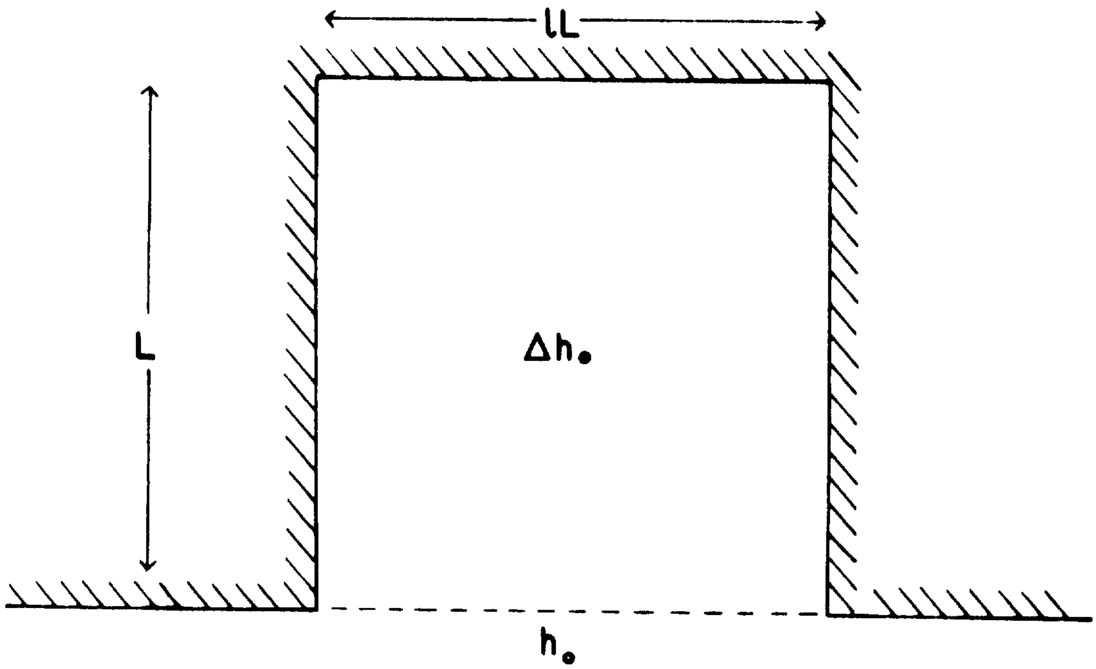

Figure 2

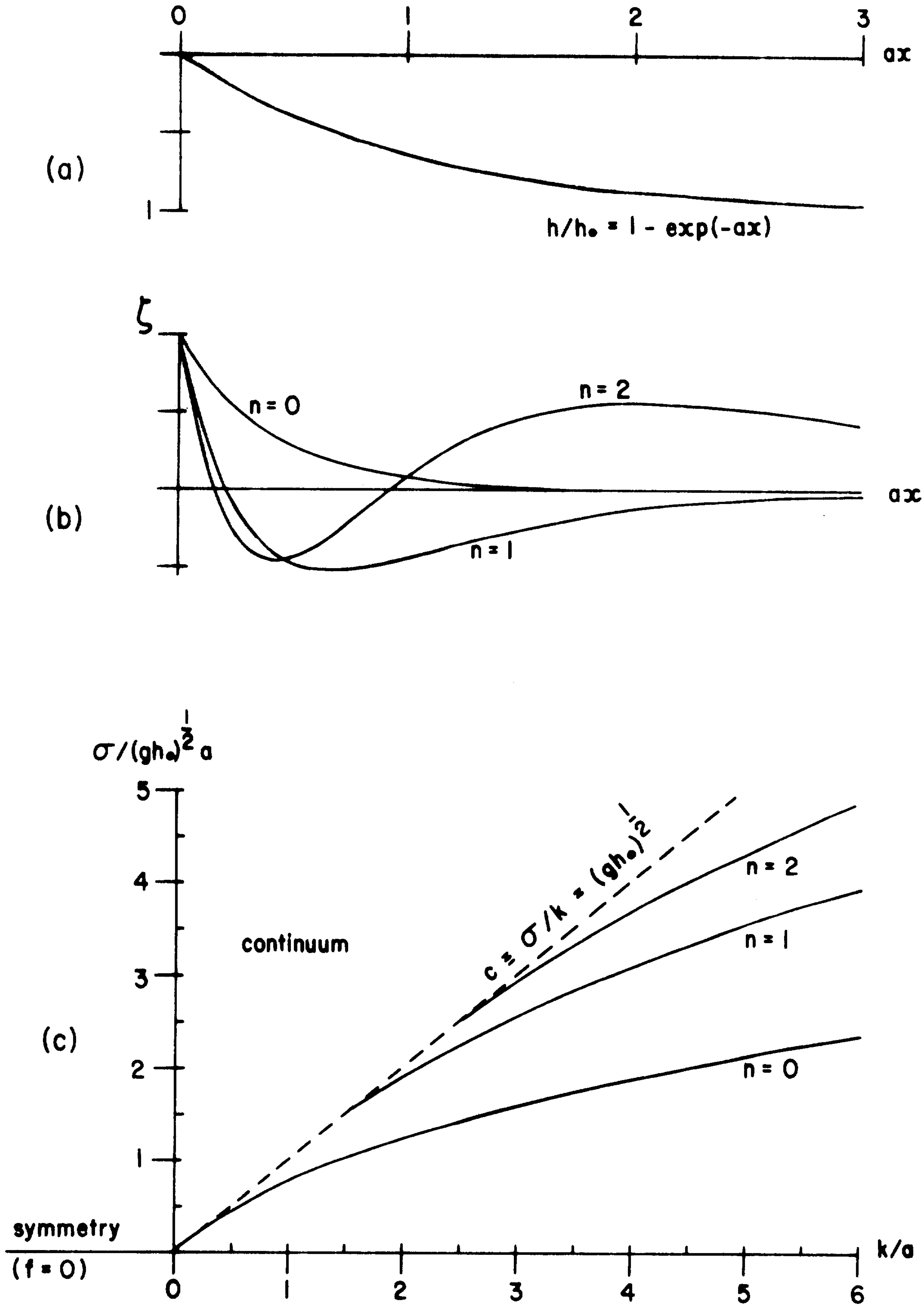
0
1
2
3
ax
(a)
1
h/h₀ = 1 - exp(-ax)
ζ
n = 0
n = 2
ax
(b)
n = 1
σ/(gh₀)^½ a
5
4
3
2
1
continuum
c = σ/k = (gh₀)^½
n = 2
n = 1
n = 0
(c)
symmetry
(f = 0)
k/a
0
1
2
3
4
5
6

Figure 3