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Fluid Mechanics of Tropical Cyclones

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Abstract. Typhoons in the northwest Pacific and hurricanes in the northeast Atlantic are particular instances of a global phenomenon with frequently disastrous consequences known as the Tropical Cyclone (TC). This is an intense cyclone, generated over a tropical ocean with kinetic energy 10^{18} J or more, which extends over several hundred kilometres and yet is above all characterized by its calm central region: "the eye of the storm". In a TC (not, of course, to be confused with such completely different phenomena as tornadoes) both the energy input and its dissipation mainly occur within that boundary layer between air and ocean which, at high TC wind speeds of 50–60 m/s, comprises essentially "a third fluid": ocean spray. Afterwards, as a TC reaches land, disastrous effects of several different kinds may occur, and this paper outlines how fluid mechanics contributes towards worldwide struggles to reduce the human impact of TC disasters.

1. Introduction

In grateful reciprocity towards many foregoing papers in which old friends of the author have epitomized scientific efforts of long standing, this paper describes his own abiding interest in the fluid mechanics of Tropical Cyclones. The scientific term Tropical Cyclone describes (Anthes, 1982; WMO, 1995) a single well-defined natural phenomenon, which in the northwest Pacific is called a "typhoon" (literally, "great wind") while in the northwest Atlantic it is known by the name "hurricane" (of Caribbean origin); the latter name being used also in the northeast Pacific, whereas the scientific expression Tropical Cyclone (or, more simply, just cyclone) is preferred in tropical areas of the southwest Pacific and of the Indian Ocean.

In meteorology, of course (Holton, 1979), the word cyclone on its own is used to denote any centre of low pressure towards which, by the well-known Coriolis effect, surface winds spiral inwards "cyclonically"; that is, anticlockwise in the Northern, but clockwise in the Southern, Hemisphere. In addition to satisfying this general requirement for a cyclone, however, the Tropical Cyclone proper possesses an altogether special feature—clearly exhibited (Figure 1) in satellite imagery—which is absent from other cyclones. This is the famous "eye of the storm": a calm region of circular shape which is often nearly free of clouds, although it is surrounded by a wall of extremely dense cloud (the "eyewall") where wind speeds are as high as 50 m/s or more.

The fierce one-eyed monster of Greek myth was called Cyclops. Now a Tropical Cyclone, which its formidable threats to human life and property, is undoubtedly fierce; it has moreover a single circular eye (actually, the word Cyclops in Greek signifies "circular eye"); and, above all, its huge diameter of several hundreds of kilometres makes it truly a monster among meteorological phenomena. So three special features of a Tropical Cyclone—its immense size, its one circular eye, its terrifying ferocity—are summed up in the alliterative phrase "Cyclone as Cyclops".

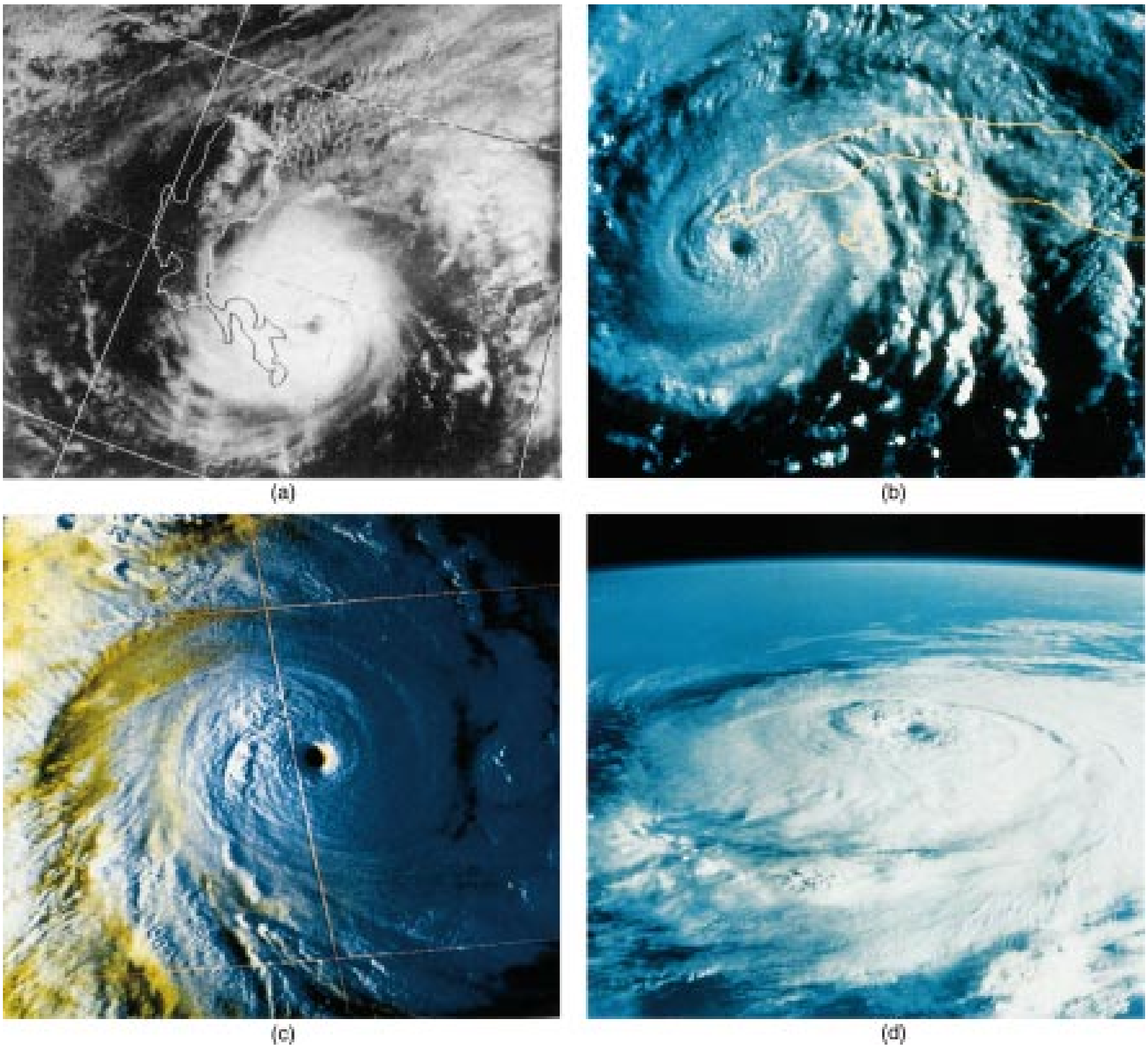

Figure 1. Views from meteorological satellites of (a) a Typhoon near the Philippines, (b) a Hurricane near Cuba, (c) a TC in the Southern Hemisphere (with clockwise rotation); alongside (d) a slant view of a hurricane from the spacecraft *Discovery*.

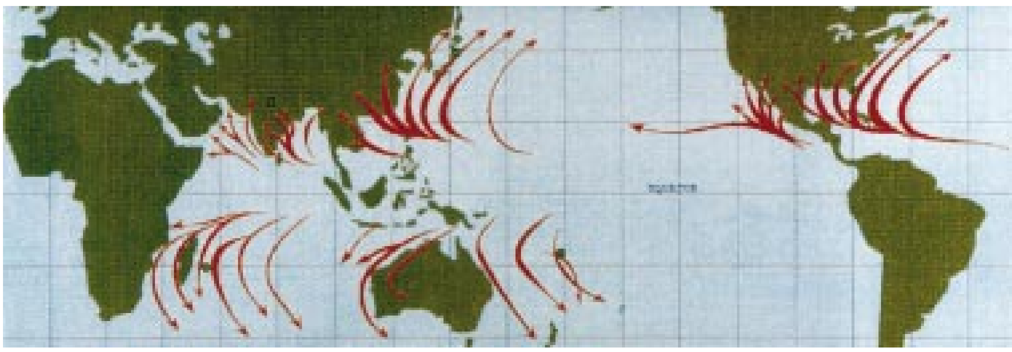

Figure 3. Schematic diagram for the world's oceans of (drastically smoothed) "frequent tracks" followed by TCs.

It is always over a tropical ocean (Atlantic, Pacific, or Indian) that a Tropical Cyclone is formed, but human beings experience its ferocity when it approaches land. Then great destruction may result, either from the direct force of extreme winds on manmade structures, or else from coastal inundation by a storm surge; as can occur when nearshore water, acted on by such winds, piles up against a coastline. The worst recorded storm surge, produced by a 1970 Tropical Cyclone in the Bay of Bengal, caused 300,000 deaths in Bangladesh (Frank and Husain, 1971).

Even a city as far from the equator as 35° N, such as Kyoto, is by no means immune from the threat of extreme Tropical Cyclone winds. For example, on 29 September 1994, "Supertyphoon Orchid" passed close to Kyoto (ESCAP/WMO, 1995) with some seriously damaging consequences; especially to structures which, having been built before the introduction of modern building codes for wind-hazard resistance, had not been adapted later to take those into account.

A map (Figure 2) showing the track followed by the eye of supertyphoon Orchid recalls the severe problems which forecasters face in predicting such a track. The track of a Tropical Cyclone (TC) commonly involves just as many sudden changes of direction as were observed for Orchid. Accordingly, any approach to TC track prediction by a simple process of extrapolation of the path followed in the last 24 or 48 hours (meteorologists call this a "persistence" approach to forecasting) can achieve only relatively poor reliability. Here is one of the reasons why fluid mechanics needs to play an important role in TC forecasting.

In another map, designed to give a summary view of the TC as a worldwide phenomenon, the highly irregular character of individual TC tracks is altogether suppressed as a result of a sort of statistical "smoothing". This map (Figure 3) shows "frequent tracks" (drastically smoothed) of TCs commonly appearing in the Caribbean, in the Pacific, around Australasia, and in the Indian Ocean. Very large numbers of islands, and extremely extensive coastal regions of continents, are seen from this map to be seriously threatened by TC hazards. Formidable hazards may arise moreover from TCs with clockwise rotation (in the Southern Hemisphere, of course); one of those, for example, having destroyed three-quarters of Darwin, Australia in 1974.

The map reconfirms also that all TCs are formed over oceanic regions (even though the hazards they generate arise only later, when they approach land). Their study, then, involves not simply fluid mechanics but specifically the mechanics of (at least) two interacting fluids: ocean and atmosphere.

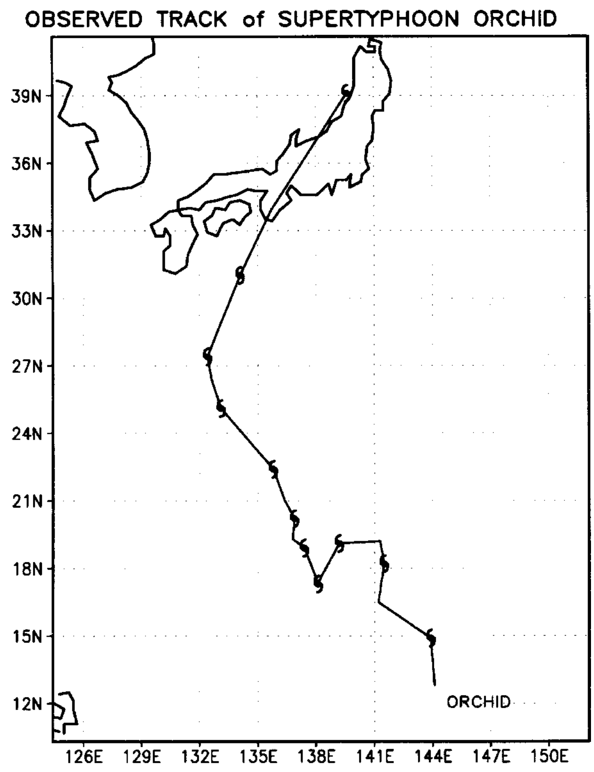

Figure 2. On this observed track of supertyphoon Orchid, the symbols show the eye's position at 0000Z on 20 September 1994 and at 24-hourly intervals thereafter.

Thus TCs are global hazards, crying out for international collaboration on a firm foundation of studies which span different disciplines (meteorology, oceanography, etc.). For example, such collaboration was powerfully achieved in these three international symposia:

(i) "Intense Atmospheric Vortices" held in Reading, England (Bengtsson and Lighthill, 1982) by two member Unions of the International Council of Scientific Unions (ICSU); being devoted not only to TCs but also to tornadoes—even though those completely different phenomena are formed over land and are smaller in horizontal scale by two orders of magnitude;

(ii) "Tropical Cyclone Disasters" held in Beijing, China (Lighthill *et al*., 1993) by ICSU as a whole jointly with the World Meteorological Organization (WMO), with strong global participation (from five continents) being attained alongside excellent cooperation between meteorologists and oceanographers; and

(iii) "Global Climate Change and Tropical Cyclones" held in Huatulco, Mexico (Lighthill *et al*., 1994) by ICSU and WMO on important questions of whether expected directions of global climate change may or may not modify the frequency or intensity of tropical cyclones in the twenty-first century.

Conclusions from all these symposia are referred to later in this paper. First, however, it is necessary to return to the theme of "Cyclone as Cyclops", and ask the key questions: "Why is it that intense Tropical Cyclones possess a circular eye?"

2. Wet-Air Thermodynamics and a Disappearing Spiral

In every cyclone, of course, the surface winds spiral inward cyclonically; yet in TCs, on the other hand, this spiral motion's inward component performs an extraordinary "disappearing trick" at the eyewall: that circular wall of extremely dense cloud (of the type known to meteorologists as "convective") which surrounds an eye often nearly free of cloud. What causes the inward component of the spiral motion to disappear there?

The answer is somewhat astonishing: in the eyewall, fast upward motions are able to lift the air right up to the base of the stratosphere (situated at about 15 km altitude), where—as Figure 1(d) shows—it then spirals outward in a broadly anticyclonic (although not usually very symmetrical) motion. Thus, the surface spiral "disappears" because fast upward motions lift surface air to stratospheric altitudes.

Yet, under ordinary circumstances, air simply cannot rise like this! This is because rising air, as its pressure drops, expands and therefore cools; actually (see below) by 1 °C per 100 m through energy lost in the work of expansion. Now, in any stable atmosphere, the surrounding air is not so cold; in other words, the atmosphere's temperature drop with height is less than 1 °C per 100 m. The rising air, then, being colder than its surroundings, necessarily falls back.

The simple thermodynamics of air rising "under ordinary circumstances" yields such a conclusion from the easily derived equation (1). Because the pressure p drops with height z at a rate ρg, equal to the weight of air per unit volume, it follows that, during any height increase *dz*,

$$-\,g\;dz = \frac{dp}{\rho} = c_p\;dT \tag{1}$$

for a perfect gas in an adiabatic change. Here, c_p is the specific heat of air at constant pressure (about 1000 J/kg per 1 °C), so that heat input at constant pressure is $c_p\;dT$; on the other hand, heat input at constant temperature equals the work of expansion $p\;dV$ if specific volume V increases by dV, where by Boyle's law $p\;dV = -V\;dp = -\rho^{-1}\;dp$; and these two heat inputs must cancel under adiabatic conditions (of no exchange of heat with the surroundings). Equation (1) gives the rate of temperature drop with height as

$$-\,\frac{dT}{dz} = \frac{g}{c_p} = \frac{10\,\mathrm{m/s^2}}{1000\,\mathrm{J/kg\ per\ 1\,{}^\circ C}} = 1\,{}^\circ\mathrm{C\ per\ 100\,m}, \tag{2}$$

just as stated in the previous paragraph.

2.1. Wet-Air Thermodynamics

By contrast, any corresponding conclusions from wet-air thermodynamics are altogether different. Wet air here means air that is 100% humid; in other words, saturated with water vapour. Now, whereas rising air that is wet (in this sense) does still cool, nevertheless this cooling causes some condensation of water vapour into rain drops; in which process latent heat is released, so that the degree of cooling is less.

In wet-air thermodynamics the equation governing rising air is changed from (1) to

$$-\,g\;dz = \frac{dp}{\rho} = c_p\;dT + L\;dq, \tag{3}$$

where the latent heat L means the excess energy per unit mass of water vapour over water in condensed form, while q is the concentration (by mass) of water vapour. The crucial fact when wet air rises is that q continues to take its saturated value q_{s} which is a steeply increasing function of temperature; this, of course, is why cooling air experiences that reduction in vapour concentration q which necessarily implies some condensation.

Very roughly indeed (see below for precise details), the effect of the added term $L\;dq$ appearing in (3), with q rising so steeply as a function of T, is to double the coefficient of dT on the right-hand side from its simple value c_p occurring in (1). For rising wet air, then, the resulting rate of temperature drop with height,

$$\Gamma = -\frac{dT}{dz}, \tag{4}$$

assumes very roughly half the former value (2), becoming about $\frac{1}{2}$ °C per 100 m. Accordingly, if the surrounding fluid's rate of temperature drop with height exceeds this (lying then, somewhere between $\frac{1}{2}$ °C per 100 m and its maximum possible value of 1 °C per 100 m for a stable atmosphere), rising wet air will always remain warmer than its surroundings and therefore can continue to rise. Such a capability, nonetheless, exists only for wet air in the strict sense of 100% relative humidity; by contrast, (1) is satisfied by air "under ordinary circumstances"; that is, with relative humidity less than 100%.

It is necessary, then, to interpret a TC's eyewall as that location where the very long spiral path pursued by winds directly over the surface of the ocean has finally caused them to become saturated with water vapour (100% humidity). Then the resulting wet air is indeed able to rise if, in the sense sketched above, the surrounding atmosphere is "not too stable".

Far more precision can be injected into the above arguments if (3) is used with the density ρ given by the perfect-gas law as p/RT while q takes its saturated value

$$q_{\mathrm{s}} = 0\cdot 62\frac{p_{\mathrm{v}}(T)}{p}. \tag{5}$$

Here, 0·62 is just the water/air molecular-weight ratio, while the vapour pressure $p_{\mathrm{v}}(T)$ may be plotted (Figure 4) as a function of temperature in two alternative curves, representing partial pressure of water vapour (i) over liquid water and (ii) over ice. In the lower part of the eyewall, of course, we are concerned with vapour condensing to liquid water. Higher up, however, vapour condenses to ice crystals; which, indeed, are found to be abundant in the upper air over a TC. Transition from one curve to the other takes place somewhere between 0 °C and -10 °C, depending on the amount of supercooling of condensed water.

Values of Γ, the rate (4) of temperature drop with height for wet air, were obtained from (3) and (5) by List (1951); and are here plotted (Figure 5) against temperature for a range of different pressures, with both $p_{\mathrm{v}}(T)$ and L given their values for vapour over liquid water when $T > 0$ °C while values for vapour over ice are used when $T < -10$ °C. Dotted lines indicate possible transitions between the two sets of curves.

Equations (3) and (5) also specify the ratio dp/dT as a function of p and T in the form of a differential equation which can be solved given a boundary condition at (say) the base of the eyewall. When typical base values of p = 950 mb and T = 30 °C are used to derive this solution, the corresponding values of Γ are as shown by the broken line. (They are a little less than the very rough approximation $\frac{1}{2}$ °C per 100 m in the lower part of the eyewall, and a little more in the upper part.) Air in the eyewall can rise when rates of temperature drop in the surrounding atmosphere exceed these Γ values.

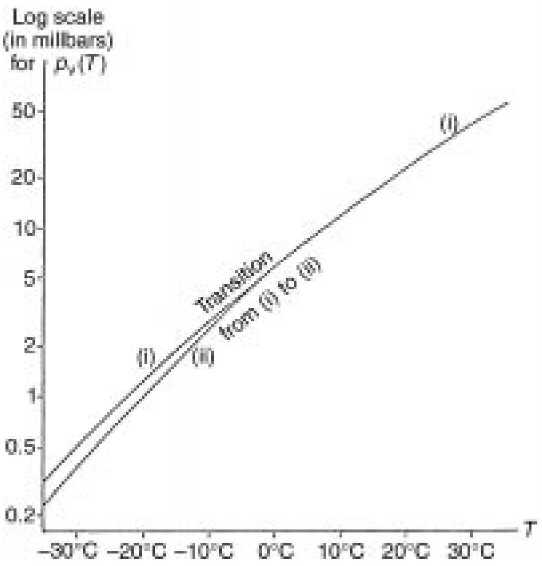

Figure 4. The partial pressure $p_{\mathrm{v}}(T)$ of water papour (i) over liquid water and (ii) over ice.

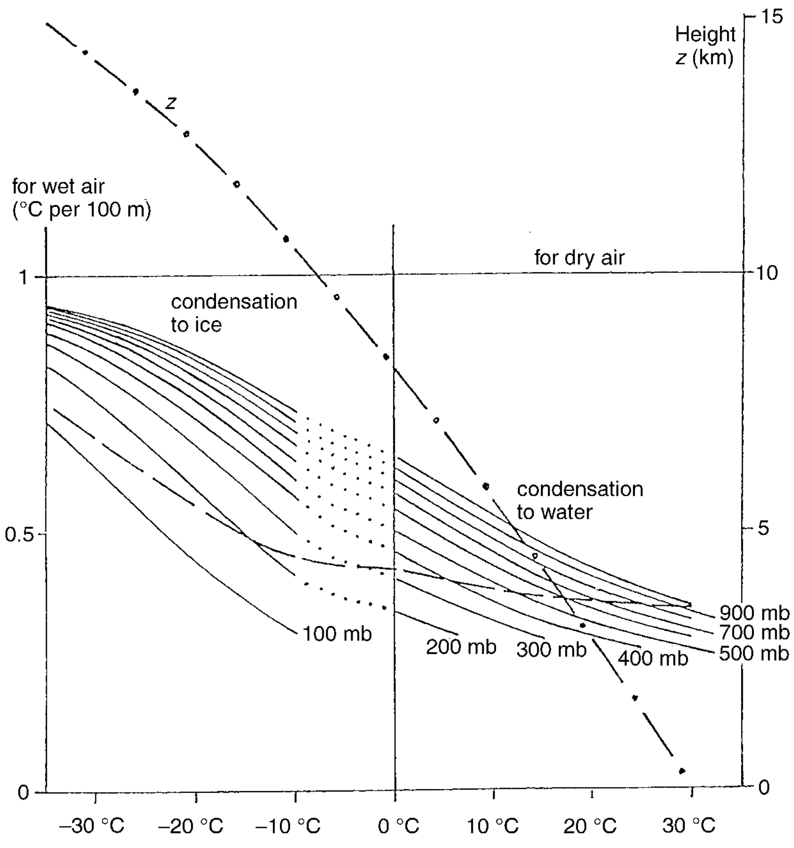

Figure 5. Solid curves show Γ, the rate of temperature drop (4) with height for adiabatically rising wet air, computed (List, 1951) as a function of temperature T in °C and pressure p in millibar. Dotted curves suggest possible transitions between values for vapour condensing (i) into liquid water and (ii) into ice. The broken-line curve plots values of Γ in an eyewall where wet air is assumed to rise adiabatically from a base at 30 °C and 950 mb, while the dash-dotted line gives corresponding values of the height z (again assuming a strictly adiabatic process).

The corresponding heights z at which the Γ values would be found if wet air rose (as assumed) in a strictly adiabatic process are readily deduced from the broken-line values of Γ by numerically integrating (4) from a base value of $T = 30$ °C at $z = 0$. This derives the height z as the function of T shown in the dash-dotted line.

Real processes within the rising fluid of the eyewall include, of course, some departures from strictly adiabatic wet-air behaviour; the most important one arising, perhaps, from mixing caused by entrainment of ambient fluid. Under these circumstances the value of T reached at an altitude of 15 km typical of the base of the stratosphere will not be nearly so far above a typical ambient temperature of -70 °C as the value derived (-35 °C) on the assumption of a strictly adiabatic process. Nonetheless the adiabatic curves for wet air give an impressive indication of those lifting forces that power a TC.

2.2. The TC Viewed as a Heat Engine

Emanuel (1986, 1991) has pointed out how wet-air thermodynamics allows us to view the TC as a heat engine where the working fluid is just that mix of dry air with water in all its forms (vapour, droplets, ice crystals) which appears in the atmosphere. From this standpoint all the heat intake occurs over the ocean, and essentially consists of latent heat of evaporation transferred during the long spiral path pursued by winds before they reach saturation. Somewhat remarkably, all this heat intake occurs at a practically constant temperature (that of the ocean surface). This is because most of the cooling of air which could be expected to result from provision of latent heat is cancelled by the vigorous processes of radiative and turbulent heat transfer which take place at the interface between ocean and atmosphere. After that the heat engine's nearly adiabatic work-output phase is concentrated in the eyewall; while, finally, the heat-loss phases takes place at a nearly stratospheric temperature.

Essentially, the heat-engine cycle so described is close to a Carnot cycle: one with heat-intake and heat-loss phases occurring at different constant temperatures, and separated by an adiabatic work-output phase. Here, the large difference between the heat-intake temperature T_1 and the heat-loss temperature T_0 suggests a substantial value for the "thermal efficiency" η: that proportion of the heat intake which appears in the output of mechanical work. Thus, the classical efficiency value

$$\eta = \frac{T_1 - T_0}{T_1} \tag{6}$$

(with temperature in kelvins) for an ideal Carnot cycle could take values of order one-third if T_1 were about 300 K (a typical sea-surface temperature) and T_0 about 200 K (a typical stratospheric temperature).

Whatever a realistic value for η may be, that proportion η of the heat intake at the ocean surface which generates mechanical energy—above all in the form of extreme winds—is required in a TC to balance all the frictional dissipation of energy in these winds occurring near the ocean surface itself. Study of this important balance explains (Emanuel, 1986, 1991) why the TC is a tropical phenomenon: the heat intake per unit mass of air depends critically on the concentration q_s (by mass) of water vapour under saturated conditions, which increases steeply with temperature (Figure 4), while there is no such dependence on temperature in the dissipation rate per unit mass. (However, see also Section 3.3 below.)

3. Air and Ocean in Strong Interaction

A TC's energetics, then, depend critically upon extremely strong interactions between atmosphere and ocean; "extremely" strong in the sense that the interactions take place at extreme wind speeds. They include

(i) that transfer of water vapour from ocean to atmosphere which is necessary to allow 100% humidity to be reached—so that air in the eyewall can rise to great heights; along with
(ii) such heat transfer from ocean to nearby air as is required to keep their temperatures equal; yet opposed by
(iii) a transfer of momentum from air to ocean associated with its frictional resistance to surface winds.

Thus, while in any TC the ocean gives energy to the atmosphere, nonetheless the atmosphere in turn "gives back" momentum to the ocean—with some potentially disastrous results of just two main kinds, each whipped up by extreme surface winds. On the deep ocean these tend to produce intense surface waves (see Section 3.1) while in shallow water (see Section 4.3) they may generate those powerful bulk motions which are known as storm surges, and which can gravely threaten coastal populations.

3.1. Exceptional Ocean Waves

Ocean waves of exceptional height may of course be of immediate local importance, as severe hazards to shipping—while being capable also, "at long range", of augmenting the flooding threat to a coast. Now a well-known requirement for a high wind to generate intense waves is for it to act over a sufficiently long distance: the so-called "fetch" needed for wave amplitudes to reach large values. Similarly, any exceptional wave heights may be attainable only when extreme winds are applied over a long distance—so how is it that the relatively localized extreme winds around a TC's eyewall can act over a long enough fetch?

An answer to this question is suggested by the fact that the energy in a group of ocean wave travels at the group velocity (half the velocity of their crests). Accordingly, extreme TC winds can continue to energize one and the same wave group over a long distance or fetch provided that the travel velocity of the TC as a whole coincides with the group velocity of the waves (WAMDI, 1988). For a typical TC travel velocity around 10 m/s this condition identifies waves of length around $\lambda = 250\,\mathrm{m}$ as those which might continue to be excited; and, certainly, an extended fetch would be required for such long waves to build up to near maximum heights.

A schematic diagram of a travelling TC shows (Figure 6) which part generates waves moving in the direction of the travel velocity. Energy in a wave group there generated, at wavelengths satisfying the above condition, must travel at this velocity. Therefore, these waves can continue to be amplified by extreme winds, so that they are able to reach very great heights; as with the wave thus produced with height 32 m (from crest to trough) which was observed in the early hours of 11 September 1995 by the crew of the celebrated liner *Queen Elizabeth II*.

It goes without saying, of course, that TC winds generate waves with a most diverse range of lengths—not just those very big wavelengths which have been stressed up to now (because they allow such exceptional wave heights to be attained). Moreover, many far shorter waves are quite important; for example, as helping to determine those roughness lengths which affect values of transfer coefficients for momentum, heat, and water vapour. In addition, the continual breaking of these waves can be described as creating between air and ocean enormous volumes of "a third fluid"—spray—which will be reemphasized later (Section 3.3) as a principal feature of intense interactions between air and ocean.

3.2. Conditions for TC Formation and Intensification

Before that, however, those conditions under which such strong interactions have been observed to produce TC formation and intensification must be enumerated. In an important series of papers (see pp. 3–20 of

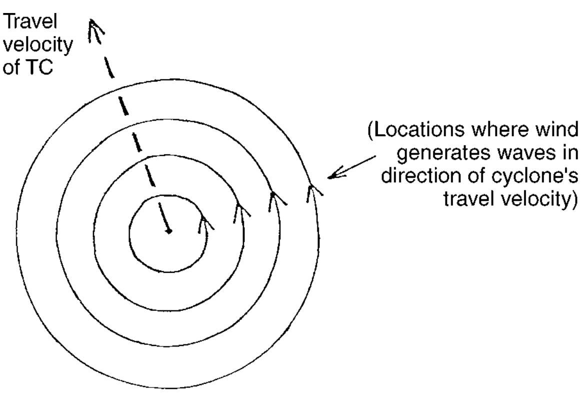

Figure 6. Schematic diagram of a travelling TC, showing which part generates ocean waves moving in the direction of the travel velocity.

Bengtsson and Lighthill (1982) and pp. 116–135 of Lighthill *et al.* (1993)) Gray, after summarizing and analyzing very fully the observational record, concluded that TC initiation—often called "cyclogenesis"—requires all six of the following conditions to be satisfied (here, Gray's well-accepted list of six necessary conditions is accompanied with brief bracketed comments that relate each condition to physical discussions in Section 2 above):

(i) latitudes, whether north or south of the equator, must be a least 5° (for Coriolis effect to yield a cyclonic spiralling of surface winds);
(ii) rates of temperature drop with height in surrounding air must exceed the value Γ appropriate to wet air rising adiabatically (for ascent of air in the eyewall to be adequately powered);
(iii) temperatures T at the ocean surface must be at least 26 °C (for the saturated water-vapour concentration (5) to provide sufficient latent-heat input to cyclonically spiralling winds);
(iv) vertical shears, i.e., gradients of wind with height in surrounding air, must be relatively small (for disruption of the TC flow structure's axisymmetry and vertical coherence to be avoided);
(v) relative humidities in the middle troposphere must be sufficiently high (for prevention of any possible "drying out" of the eyewall upflow by entrained air); and, finally,
(vi) some rather substantial amounts of cyclonic vorticity must previously be present at low altitude (for that process which is described in Section 2.2 as the Carnot cycle heat engine to be able to start).

It is noteworthy that all six conditions are shown by Gray from the observational record to be necessary for TC formation and intensification.

In particular, condition (vi) refutes any idea that a TC arises from some sort of "instability to small-amplitude disturbances". On the contrary, only a pre-existing atmospheric disturbance of substantial amplitude can, by means of strong interaction between air and ocean, be amplified into a TC. Among examples of how this can happen, two may be briefly mentioned.

In the western Pacific, a commonly observed phenomenon is the "monsoon trough": a long line of low pressure formed in the aftermath of a monsoon. At the eastern end of a monsoon trough, a TC is often formed through strong interaction of local cyclonic vorticity with the tropical ocean.

Over the Atlantic, on the other hand, the meteorology of the Sahara acts to generate waves with continually growing amplitude which steadily travel westward in the upper atmosphere. Then, after sufficient amplification of the westward-travelling waves, a "crest" may break away to form a "cut-off low" with strong cyclonic rotation which may then penetrate down to the sea surface and "ignite the Carnot engine". For further discussion of the fluid mechanics of TC initiation, see Section 4.1.

3.3. Observing Air–Sea Interaction at Extreme Wind Speeds

In the meantime, this section has amply shown how air–sea interaction at extreme wind speeds can exert many different influences on a TC's formation and development, as well as on its possibly hazardous consequences; so that great importance must be attached to direct observations of the detailed nature of that interaction. It has above all been the gifted and courageous crews of Russian research ships who have contributed such direct observations by deliberate navigation through typhoons—and those from ICSU who planned the "Tropical Cyclone Disasters" Symposium (Beijing, 1992) were pleased to be able to draw the attention of the world's meteorologists to these Russian achievements. In particular, the excellent symposium paper given by V.D. Pudov (see pp. 367–376 of Lighthill *et al.* (1993)—along with other papers referred to therein) summarized that valuable research programme.

Here, special attention is drawn to Dr. Pudov's 1988 observations, made with great care near typhoons "Tess" (8830) and "Skip" (8831), of a steep increase with surface wind speed in the concentration of water droplets ("spray"), found alongside a closely parallel fall in air temperature below that of the ocean surface (Figure 7). These observations suggest a slight correction to the thermodynamic picture painted in Section 2. Thus, if

(a) much of the vapour transfer to surface winds came from spray droplets, then
(b) cooling from the corresponding latent heat transfer might not be fully made up by heat transfer from the ocean surface, so that

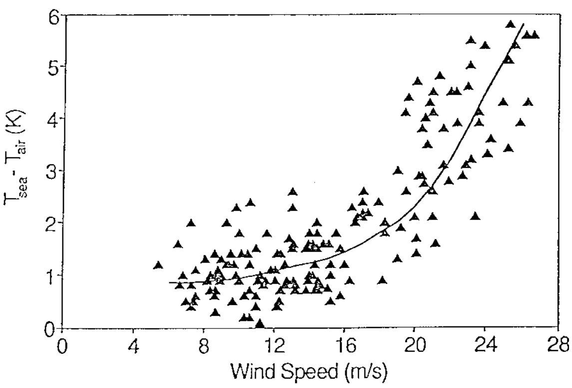

Figure 7. The 1988 measurements by Pudov, demonstrating for increasing wind speed an increasing difference (in °C) between temperatures at the sea surface and in neighbouring air—as ascribed to air cooling by spray evaporation in a paper (Fairall *et al.*, 1994) from which this replot of Pudov's data is taken.

(c) air temperature (as observed) would reach an equilibrium value below that of the ocean surface; and, in consequence,
(d) the average temperature of saturated air around the base of the eyewall would be less than the sea-surface temperature.

From the thermodynamic viewpoint, the importance of such a correction (d) to the heat-intake temperature T_1 lies, of course, not in the rather modest resulting drop in the Carnot efficiency (6), but in the much more significant reduction in the overall latent heat intake per unit mass of air associated with the very steep dependence (Figure 4) of saturated water-vapour concentration q_s on the heat-intake temperature T_1.

Fairall *et al.* (1994), in a careful study of Pudov's data, have given a theoretical analysis in support of the above interpretation. Breaking waves generate spray in two ways: surface bursting of bubbles of entrapped air makes smaller droplets (radii up to 60 μm), while spume formation at whitecaps makes larger droplets (radii over 40 μm). An increase in the fraction of ocean surface covered by whitecaps is viewed as the main influence of increasing wind speed on spray formation; nonetheless these authors recognize that, at wind speeds above the value (around 40 m/s) for which that whitecap fraction approaches unity, spray formation may yet continue to increase "in an unknown manner" and so they discourage application of their theory at greater wind speeds. Very briefly, their predictions at 40 m/s are that the mass density of spray should reach only 0.008 kg/m^{-3} (less than 1% of the air density), and yet that vapour transfer from spray to air should exceed direct transfer from the ocean surface by an order of magnitude. Thus, hypothesis (a) wins from this analysis quite strong support—which, moreover, would not be qualitatively altered even if the rate of generation of spray droplets had been overestimated by up to a factor of 2 as suggested by Katsaros and de Leeuw (1994).

At relatively high, although not "extreme", wind speeds, the major international programme HEXOS (humidity exchange over the sea) gained much valuable information (Smith *et al.*, 1996) on water vapour transfer rates. In particular, as wind speeds increased to 18 m/s, North Sea measurements on the fixed platform Meetpost Noordwijk found no statistically significant increase in the standard vapour transfer coefficient. In this range of wind speeds, on the other hand, changes are absent also in Pudov's above-noted data. For the TC environment, then, there remains considerable confidence in hypothesis (a) and its suggested implications (b), (c), and (d).

Future observations of air–ocean interaction at extreme wind speeds will depend increasingly on satellite imagery. Already, a large amount of information on sea-surface roughness, including directional spectra of waves, is obtained from the earth resources satellite *ERS*-1 of the European Space Agency by means of a C-band radar (at 5.3 GHz) which measures Bragg back-scatter from the rough ocean surface. Such "scatterometer" data (Quilfen *et al.*, 1994) are concerned, not with "exceptional" waves developed over a long fetch (Section 3.1), but with the general statistical distribution of wave roughness (as a function of length and orientation) associated with local winds. Accordingly, they are often used to suggest an approximate distribution of surface wind vectors. These data may increase in value if, as proposed by the Institute Français de recherche pour l'exploitation de la mer (IFREMER), current scatterometer measurements at 25 km resolution can be made routinely available.

3.4. Global Climate Change and TCs

Yet another problem closely related to strong interactions between air and ocean is the question, posed at an ICSU/WMO symposium held from 22 November to 1 December 1993, of whether expected directions of global climate change are liable to produce significant effects on the frequency or on the intensity of TCs. Very briefly, the report on the symposium (Lighthill *et al.*, 1994) concluded that, even after an expected doubling of CO_2 in the atmosphere, those effects were likely to be small relative to the usual large year-by-year variations.

Indeed, models of climate change following such a doubling suggest a mean surface temperature rise in the tropical oceans of only 1 °C (because ocean temperatures respond less than land temperatures, and tropical oceans less than oceans in general). Admittedly, in relation to the six conditions of Section 3.2, a rise of 1 °C helps to satisfy condition (iii), which might be thought to make TC formation and intensification significantly more likely. Nonetheless, all consideration of other influences, including those of conditions (ii), (iv), (v), and (vi), tends to work the other way and diminish expectations of a rise in TC frequency.

As to any effect on TC intensities, the thermodynamic argument of Section 2.2 as it stands does indicate a positive influence of mean sea-surface temperature. This, however, is where considerations about spray droplets outlined in Section 3.3 may yet again tend to diminish any such influence. Indeed, if any increase in TC intensity produces an enhancement of effects (a)–(d), then this increase may readily become self-limiting because saturated air rising from the base of the eyewall may fall short, to an increasing extent, of gaining that full latent-heat intake (for powering the TC) which would be associated with the sea-surface temperature.

Next, the report gives detailed consideration to TC statistics, stressing not only the large year-to-year variation which has already been mentioned but also the absence of any discernible trends in recent decades. Finally, it studies scatter diagrams in which each TC is represented by a single point related to when and where it reached maximum intensity, that intensity being plotted against the actual sea-surface temperature found there during the month in question. The absence of significant correlation when the data are plotted in this way is seen as confirming the report’s broadly negative conclusions about climate-change influences on TC statistics.

4. Fluid Mechanics and TC Disasters

Tropical Cyclone disasters can of course arise quite soon after TC formation, when a small oceanic island may suffer the impact of TC winds. Far more frequently, however, major disasters occur when a TC reaches a substantial land mass, where extreme winds may cause severe damage either directly by their action on manmade structures or indirectly through storm-surge flooding; and where, in addition, river valleys can be massively flooded if the TC’s huge water content is released over their catchment areas.

Scientific studies of all types of TC disaster—and of their possible mitigation—are founded primarily on fluid mechanics. Moreover, many different aspects of the mechanics of what the late A.E. Gill (1982) called “the earth’s fluid envelope” (atmosphere, ocean, rivers, lakes, groundwater) contribute to such studies, in ways which are briefly outlined in this Section 4.

4.1. Fluid Mechanics and TC Formation

Even though knowledge of Gray’s six necessary conditions for TC formation (see Section 3.2) is of very great value, nonetheless some severe difficulties still stand in the way of any attempt to forecast just when and where a TC will appear. Admittedly, it can be argued that obstacles to forecasting TC initiation present no huge embarrassment; after all, satellite pictures soon pinpoint a TC, and for very many purposes (see Section 4.2) a good forecast of future movements, along with any intensity changes, for a TC already observed by satellite may be quite sufficient.

Most TC forecasters, on the other hand, who also take into account the above-noted dangers to oceanic islands, feel strongly committed to watch out for likely TC originators. These include, of course, those monsoon troughs and westward-moving waves which were mentioned in Section 3.2. Recently, moreover, a simple idea from fluid mechanics—potential vorticity—has begun (Hoskins *et al.*, 1985) to prove useful in this context. Potential vorticity for the atmosphere was defined by Ertel (1942), and it turns out that anomalously large cyclonic values of potential vorticity can be viewed as "having the potential" to produce strong cyclonic rotation; which, after gaining energy from air–sea interaction as described in Section 3.2, may initiate a TC.

From the fluid-mechanics standpoint, potential vorticity may be interpreted as follows. The vertical component of the atmosphere's absolute vorticity may be written $f + \zeta$, where the Coriolis expression

$$f = 2\Omega \sin \theta \tag{7}$$

(with θ as latitude) gives the contribution from Earth's rotation at angular velocity Ω, while ζ represents the vertical vorticity component of the winds themselves (air motions relative to the rotating Earth). If now h stands for the vertical spacing between two nearby surfaces of constant entropy S—these are surfaces tending to move with the fluid—then the absolute vorticity $f + \zeta$ tends to respond to any vertical stretching of fluid elements by a variation in direct proportion to h. This implies a tendency for the quantity

$$(f + \zeta)\frac{\partial S}{\partial z}, \tag{8}$$

defined as potential vorticity, to remain constant for each particle of fluid.

The stress laid by modern meteorologists on distributions of potential vorticity recalls that earlier revolution in low-speed aerodynamics which arose from Prandtl's committed recognition of how a vorticity field both uniquely determines the flow around an aircraft and can display its key features. Indeed, the likeness runs deep: an analogous "inversion theorem" exists (see pp. 143–156 of Lighthill *et al.* (1993)) for the meteorological potential-vorticity field; which, furthermore, may help to indicate key cyclonic developments. Forecasters are able to utilize such indications because modern programs (see Section 5.2) which apply computational fluid dynamics to weather prediction are able to give printouts of potential-vorticity distributions.

4.2. Fluid Mechanics of TC Tracks

In the general context of TC disaster prediction on the other hand the primary need is to forecast, for a TC already observed by satellite, its future "track". This means the path pursued by the eye of the storm.

An associated need, second only to the tracking requirement, may be to forecast any changes in TC intensity. Admittedly, while a TC travels over a tropical ocean, there exists an approximate balance between energy gain from the Carnot cycle and energy loss by friction at the surface. Nonetheless, this balance can only be approximate; accordingly, questions about intensity changes need to be put even over a tropical ocean. Furthermore, rates of energy gain (but not of loss) diminish over colder ocean areas, where the TC therefore weakens—while over land there remains no energy supply at all (and no physical basis for the presence of an eyewall) so that weakening is even more rapid. Yet these considerations tend perhaps to strengthen the key demand for track prediction, just because the gravest threats of disaster may be centred around where a TC first reaches land.

The track which a TC follows is often extremely complicated (Figure 2); primarily, because it is influenced by the TC's interactions which the entire surrounding field of atmospheric motions. Such far-ranging influences can be fully taken into account only in a global numerical model (Section 5.2) of those motions. Nevertheless, a preliminary analysis by the methods of theoretical mechanics may help both in interpreting the outputs of numerical models and in suggesting (see Section 6.2) how a model might be further improved.

The fluid mechanics of TC tracks views them as influenced to a similar extent both by the general ambient-flow pattern (including both first and second horizontal derivatives of ambient flow velocities) and by the well known "beta effect". Here β stands for the northward gradient in the Coriolis parameter (7); evidently, horizontal gradients in ambient values of the absolute vorticity $f + \zeta$ are fully specified by this mix of data.

Those influences, taken together, can perturb the TC's inherently axisymmetric structure, in a manner highlighted by the paper of Chan and Williams (1987). Also, studies by Smith (see pp. 267–279 of Lighthill *et al.* (1993)) have still further elucidated the nature of this perturbation; yet here the contribution of Dr. Chan is emphasized as a rather striking example of good theoretical mechanics which, later, would lead him directly to propose valuable practical improvements to TC track forecasting (see Section 6.2).

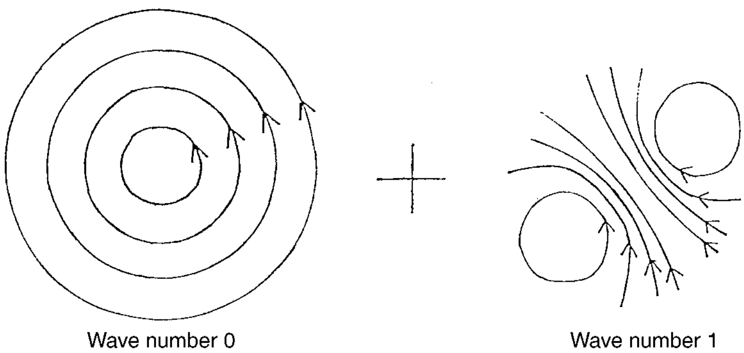

Figure 8. For a TC wind field, a much better approximation than a purely axisymmetrical pattern of azimuthal winds is a linear combination of such a component (with wave number 0) and another component (with wave number 1) which tends to convect the whole TC pattern.

The departure from axisymmetry may be described in terms of Fourier components having different wave numbers with respect to an azimuthal angle; then a purely axisymmetric TC pattern is represented by a component of wave number 0. A key discovery by the authors cited is that a component of wave number 1 represents (Figure 8) by far the biggest perturbation to that axisymmetric motion. Furthermore, the form of this component near the TC's centre is such that it tends to convect the TC pattern. Here, then, is at least one of the mechanisms underlying the nature of TC tracks—a subject returned to in Sections 5 and 6.

4.3. Wind Forcing of Shelf Waters

Strong winds in general (including TC winds) may force shallow waters over a continental shelf into large "bulk motions", nearly uniform with respect to depth except in a bottom boundary layer. Such a response is called barotropic.

The especially extensive shelf waters off Bangladesh—where depths of order 10 m extend over almost 100 km—can massively respond to TC winds. During the years since the 1970 storm surge (mentioned in Section 1 as the worst ever), another megadisaster hit Bangladesh on 29 April 1991, when the combination (Katsura *et al.*, 1992) of storm-surge flooding and extreme winds led to 138,000 deaths. This was a case when, although TC warnings had been widely disseminated from 26 April, some of the human action called for in response was tragically delayed until intense winds began to make such action difficult. Here it may be noted that many TC forecasters acknowledge a special problem in winning public confidence in storm-surge warnings; which at present suffer from enhanced false-alarm rates because of extreme sensitivity to TC track-forecast timing (briefly, storm surges are much more damaging if a TC's arrival coincides with high tide).

Wind forcing of shelf waters is important too in a non-TC context. For example, the Netherlands and Britain need to be always on the alert against a possibly disastrous storm-surge response of the North Sea. This is an area where increasingly reliable numerical models (barotropic, with bottom friction and an appropriate horizontal eddy viscosity) have been developed to assist decision-making: a model developed (Flather *et al.*, 1991) at the Proudman Oceanographic Laboratory near Liverpool determines when the Thames Barrier must be raised to protect London, while one developed (Verboom *et al.*, 1992) at the Delft Hydraulics Laboratory (with Rijkswaterstaat) has a key role for Dutch coastal defences. Both laboratories have collaborated (Mastenbroek *et al.*, 1993; Wu *et al.*, 1994) in developing a new, highly promising wave-tide-surge coupled model (which allows, e.g., for the influence of wind waves on augmenting both surface friction coefficient and bottom dissipation).

For TC forcing around US shores, NOAA operates an excellent model pioneered by Jelesnianski (1965). Fine models have been developed also in China and India (see pp. 423–451 of Lighthill *et al.*, (1993))—as well as at Kyoto University's Disaster Prevention Research Institute with its highly effective model (Katsura *et al.*, 1992) for TC-generated surges: one of those (see also Flather, 1994) that stood up well to the test of being applied retrospectively to the April 1991 disaster in Bangladesh. Yet again, however, it is remarked that future operational use even of outstanding models in the Bay of Bengal may need to await improved accuracy in TC track forecasts.

Some workers in this field regard CFD alone as a useful guide, but others believe that theoretical fluid mechanics can make a contribution; above all, in deciding whether or not shelf waters exhibit a resonant type of response (as opposed to one that is mainly determined by a balance between wind forcing and dissipative effects). In a joint paper (pp. 410–422 of Lighthill *et al.* (1993)), Lighthill and Johns did identify a resonant response on any rather long continental shelf—taking the east coast of India as an example. Very briefly, the "transient forcing" (Lighthill, 1996) produced by a TC crossing the shelf at an acute angle may generate a short "group" (in the wave-theory sense) of shelf waves, having a phase velocity equal to the along-shore component of TC travel velocity, provided that the wave-number so determined matches the TC's dimensions.

Actually, coastal responses to wind forcing have long been observed, both in the North Sea (Gordon and Huthnance, 1987) and around Australia (Fandry *et al.*, 1984), to include shelf waves. These, however, are Kelvin waves; that is, they propagate along a shelf "anticyclonically relative to the coast" with a radian frequency ω which is less than the Coriolis parameter f.

The general theory of shelf waves, on the other hand, shows (Huthnance, 1975) that cyclonic (as well as anticyclonic) propagation is possible when $\omega > f$. For the Bay of Bengal, Lighthill and Johns found that the relevant frequencies have $\omega \gg f$ (making Coriolis effect negligible) and established that a TC is able to generate a group of shelf waves which propagates relative to the east coast of India "cyclonically" (that is, in the northeast direction).

There is one highly simplified case, that of a shelf with uniform bottom slope, when such shelf waves become the famous "edge waves" discovered by Stokes, and it is a pleasure to acknowledge that the possibility of these being excited by a hurricane had been suggested 40 years ago by Greenspan (1956). Moreover the present author outlined a nonlinear theory of edge waves, and applied it to the propagation of a group of such waves, in a 1992 lecture; however, while preparing that lecture for publication, he completed a literature search which revealed that, in extending Stokes's edge-water theory to a nonlinear analysis, he had been preceded by his old friend and erstwhile pupil Whitham (1976)!

4.4. Kinematics of River-Valley Floods

Furthermore, in relation to those other flood disasters which may arise in river valleys—often from extreme precipitation—Lighthill and Whitham (1955) collaborated long ago to emphasize how the speed, with which a flood wave travels down the river, is determined above all by fluid kinematics. Essentially, at each place along the river, any local level of the water surface determines simultaneously both the cross-sectional area of water A and the downstream volume-flow rate Q (the latter arising from a local balance between gravity forcing and frictional dissipation); thus permitting for each place the construction of a plot of Q against A, with a slope

$$C = \frac{\partial Q}{\partial A} \qquad \text{for } x \text{ constant.} \tag{9}$$

Here, x stands for the distance down the river.

Simple kinematics now shows that C represents the local speed with which the flood wave propagates downstream from where new runoff into the river is occurring. This is because (9) gives

$$\frac{\partial Q}{\partial t} = C\frac{\partial A}{\partial t} = -C\frac{\partial Q}{\partial x}, \tag{10}$$

where the last two expressions are identical by the equation of continuity (law of conservation of volume). Equation (10), confirming that values of the flow rate Q travel down river at speed C, is yet another useful insight from theoretical fluid mechanics.

5. Forecasts and Warnings

The UN World Conference on Natural Disaster Reduction (Yokohama, 23–27 May 1994) issued a strong Yokohama message (Elo, 1994), stressing above all the contribution that preparedness can make to natural disaster reduction of all kinds. Such preparedness includes proper use of (and provision of)

(i) hazard resistant structures, and
(ii) forecasts and warnings;

which were the titles of two of the conference's Technical Sessions. Both (i) and (ii) are specially important for reducing the human impact of TC disasters—and indeed both involve fluid mechanics!—but there is space here for only a brief summary of the TC-related recommendations (Lighthill, 1994) from the Technical Session on Hazard Resistant Structures.

Half of these recommendations are concerned with structures able to resist extreme winds. Happily, local building codes in countries subject to TC threats do offer good advice, so our key recommendation was necessarily that BUILDERS COMPLY WITH CODES. (Usually, such compliance demands only rather small additional costs, yet it is vital that these should not be avoided.) For example, a new roof needs to be of "hipped" construction (alternatively, an existing gable roof must have a parapet added) and to be anchored, e.g., by "hurricane straps"; masonry walls need bracing; metal sheeting must have proper thicknesses and fastenings; etc.

Again, for construction to resist strong storm surges, just one main recommendation suffices: build at least 10 m above mean sea level. For example, on threatened Bangladesh islands, low artificial hills are being created in the impressive "Multipurpose Cyclone Shelter Project". It is called multipurpose because the structure built on such a hill acts primarily as either a school or a health centre—or as whatever else the islanders themselves feel is most needed—and secondarily as a shelter to house the local population safely (with their livestock crowded nearby) after receipt of a storm-surge warning.

5.1. From Forecasts to Effective Warnings

Of course any such reference to how an effective warning may be of vital importance for disaster mitigation reminds us that two problems of closely comparable difficulty are faced by forecasting experts:

(a) to apply good science (including good fluid mechanics) to making timely and reliable forecasts of patterns of wind, wave, surge, flood; and
(b) on the basis of such forecasts, to disseminate useful warnings that will convincingly advise people under threat about actions needed for their protection.

The essential processes which link (a) to (b) represent a key part of TC disaster reduction and one that is taken particularly seriously by the World Meteorological Organization. In this paper about TC fluid mechanics on the other hand, proper stress having been laid on (b), all of what follows is devoted to (a).

5.2. Advances in Global Weather Forecasting

For weather forecasting in general, the big improvements achieved during the past two decades came above all (Manabe, 1985) from combining better fluid mechanics with better atmospheric physics in a well-constructed numerical model of global atmospheric processes. Such a model for numerical weather prediction (NWP), then, uses refined CFD methods into which the physics of air/water mixtures, and also of radiative heat transfer, have been comprehensively injected.

Here, before describing just one of the many good global NWP models that are in operational use, it may perhaps be necessary to address a question on whether those big improvements actually resulted from meteorological satellites. However the answer to this question is "no", because satellites do "now-casting": they tell us about the weather at this moment, not about the future. Improvements in forecasting, which is concerned of course with weather in the future, have come almost entirely from NWP developments.

Admittedly, initial conditions are highly important for any CFD model, and for NWP in particular. Every NWP model, indeed, starts from initial data (comprehensively "smoothed" in a certain sense to ensure compatibility with the model) which are internationally derived for the global atmosphere each day at 0000Z and 1200Z (midnight and noon at longitude zero). Such data are obtained

(i) by weather stations over land from regularly released radiosonde ballons (telemetering data from all altitudes); and also
(ii) from local weather radars; and, especially,
(iii) from geostationary and polar-orbiting satellites; as well as
(iv) from ships and (above all) aircraft;

these keen users (iv) of global forecasts being happy to feed valuable data—alongside data from (i) and (iii)—into the WMO's massive World Weather Watch system.

One of today's many good NWP models is that used (Davies and Hunt, 1995; Cullen, 1993) by the UK Meteorological Office (UKMO), both to give global advice to aircraft and shipping and other industries, and to issue public forecasts and warnings. It is this model that (partly for a reason which emerges in Section 6.2) is described below in brief summary.

The UKMO model uses a CFD grid with 19 different levels in the vertical and with horizontal spacings of five-sixths degrees in latitude and five-fourths degrees in longitude. The numerical scheme is "conservative" (in the technical sense that quantities which physically should be conserved remain conserved even in the finite-difference representation) and includes parametrizations for

(a) type of boundary layer (dependent on Richardson number),
(b) radiative characteristics of land surfaces (as affected by soil type, vegetation, snow depth, etc.),
(c) radiative characteristics of atmosphere (as affected by greenhouse gases),
(d) both large-scale cloud cover and "convective" clouds,
(e) air–topography interaction (with correct localization of level for loss of air momentum due to gravity-wave drag), and of course
(f) air–ocean interaction; as well as
(g) horizontal and vertical eddy diffusion.

The model is run twice daily, with initial data derived (see above) at 0000Z and 1200Z, in each case leading to global forecasts for regularly spaced instants up to 6 days ahead.

This, like the world's many other excellent models, has achieved big measurable improvements in forecasting accuracy. For all of them, on the other hand, there may be a need to ask what limits, if any, exist on the possible extent of further improvements.

5.3. Weather Predictability in General

The past decade has brought increasingly precise knowledge about "predictability horizons" for systems like the global weather that, in a now well-defined sense, exhibit "chaotic" properties. Furthermore, the existence of several good NWP models has made possible the quantitative determination of such horizons; in other words, of effective limits on how far ahead prediction is possible.

Every NWP model, indeed, allows "predictability experiments"—in which the model is run with a range of different sets of initial conditions varying only slightly among themselves. This generates what is called (Palmer *et al.*, 1992) an "ensemble" of different forecasts; which, as the time t increases, diverge more and more from one another. After a certain time, forecast differences resulting from quite small shifts in initial conditions are seen to have reached levels where no confidence can be placed in any of the forecasts. Moreover, the limits on predictability identified in this way are increasingly identified as "real" limits for the actual global atmosphere, primarily because NWP models have become so good.

In the general theory of "deterministic chaos", such sensitivity to initial conditions is described (Argyris *et al.*, 1994) by Lyapunov exponents, of which the largest characterizes an exponential growth with time t in solution differences resulting from shifts in initial conditions. Similarly, refined studies of NWP models have determined such exponents; furthermore, they did so in a way that is wholly consistent with what is known about the real atmosphere's sensitivity to small disturbances (e.g., the "baroclinic instability" phenomenon arising over mid-latitude oceans).

Both for models and for the real atmosphere, the existence of an exponential growth in solution differences explains first why limits on predictability, though by no means precisely defined, may nonetheless possess useful approximate values. A second, perhaps even more vital, inference is that—given the good models now available—any attempt to move still closer towards a predictability horizon may above all demand YET MORE ACCURATE INITIAL CONDITIONS.

All these remarks apply with undiminished force when NWP is used for TC forecasting. Admittedly, unusual difficulties in numerical representation may then arise, e.g., from steepness of gradients near an eyewall, and forecasting improvements are being vigorously sought in two ways (notably, by members of the Geophysical Fluid Dynamics Laboratory at Princeton—see Kurihara *et al.* (1990)):

(i) with specialized local refinements to the model, and
(ii) with more accurate initial conditions.

Yet a relatively short paper can hardly include both; so, for the reasons just indicated, the last section is focussed on the big improvements which (ii) on its own may be able to make.

6. Improved Initial Data for TC Forecasting

Improved initial data for TC forecasting may be of two kinds: either specially measured (the best kind, as described in Section 6.1) or specially estimated (as argued in Section 6.2 to constitute "a good second-best").

6.1. Possible Futures for Measured Initial Data

Measured initial data for TC forecasting need imperatively to be three-dimensional in character; see, for example, Section 3.2 on the subject of how important for TC behaviour are questions of how winds, temperatures, humidities, etc., vary initially with height. This essential need for three-dimensional data is recalled here to emphasize the point that the required data cannot come just from two-dimensional satellite imagery. Yet neither can other regular sources of initial data (Section 5.1) help: the oceanic areas where TCs appear tend to lack radiosonde stations (and even weather radars); while aircraft, in general, avoid TCs!

An honourable exception to this last generalization is offered in the eastern USA by NOAA's Aircraft Operations Center, from which manned aircraft fly regularly into Caribbean hurricanes. Since 1982, moreover, such aircraft have deployed (see p. 6 of AMS (1995)) a most valuable instrument for deriving three-dimensional initial data; namely, the OMEGA Dropsonde, able to telemeter to the aircraft variations in physical quantities from any level below it. Use of such initial data has been shown to produce a statistically significant reduction in forecast errors.

In all other tropical regions, however, very high costs have continued to offer an overwhelming obstacle to TC reconnaissance with manned aircraft. It was against this background that the joint ICSU/WMO symposium (Beijing, 1992) adopted as a major goal the identification of a new and satisfactorily cost-effective means of acquiring three-dimensional initial data for TC forecasting. Then extensive debate during the symposium produced full agreement on what has been described (see pp. 583–586 of Lighthill *et al.* (1993)) as the meeting's "chief conclusion"; namely, that participants should give strong support to the vigorously proposed development of the small computer-piloted TC reconnaissance aircraft Aerosonde.

An impressive feature of the piston-engined Aerosonde, with its 3 m wing span and 12–15 kg all-up weight, is its combination of low cost with high technology—that is, with an advanced computer as its pilot, with an advanced GPS navigation system (yielding wind vectors by comparison of aircraft motions relative to ground and to air), and with meteorological instrumentation as advanced as in modern radiosondes. Low cost, of course, makes tolerable the occasional loss of an aircraft, although it should be emphasized that the 24 Aerosonde flights during 1995 involved only one such loss (from engine failure). In another of these flights, lasting over 8 hours, the aircraft monitored conditions at altitudes 0–4 km, flying into a heavy thunderstorm (which had an intense gust front) yet returning safely to base. Future developments of Aerosonde by Sencon Environmental Systems, in association with Australia's Bureau of Meteorology, are aimed at a doubling of endurance by 1997, followed by introduction of a supercharged powerplant which will progressively extend the aircraft's ceiling to 16 km.

6.2. Fruitful Uses of Estimated Initial Data

In the meantime, ingeniously estimated initial data have been shown, where measured data were not available, to constitute "a good second-best". Yet another happy result of the Beijing symposium was increased cooperation of TC experts from Asia with those from other continents—including a much valued visit by the able Hong Kong meteorologist Dr. J.C.L. Chan to the UKMO.

There he worked with UKMO's gifted NWP specialists on developing a major improvement to existing schemes for applying "bogus" (or, more correctly, estimated) inital data. This improvement uses the fluid-mechanical models outlined in Section 4.2, including those of Chan and Williams (1987), to yield appropriate three-dimensional interpretations of satellite imagery; winds at four heights, and at four radii from the storm centre, being estimated (Heming *et al.* 1995) as the vector sum of

(i) an expected axisymmetrical (that is, azimuthal) wind field and
(ii) the TC's observed travel velocity.

Here, the wind-field component with wave number 1 can appropriately be approximated by its central value (ii) in the region where initial data are most needed, and the blend of this addition with good choices for the four radii has generated a major improvement over previous "bogusing" methods.

Indeed, the new initial-data scheme was assessed experimentally in detailed trials, lasting from 25 August to 12 September 1994, on that UKMO model which was briefly described in Section 5.2; and the striking successes observed (including, for example, a reduction from 201 km to 123 km in mean TC tracking error—as averaged over fortyfive 24-hour forecasts) led to a decision to introduce the scheme operationally from 25 October 1994. Furthermore, its good success has since been maintained; for example, a recent independent assessment of 10 different NWP models for all the Atlantic hurricanes in 1995 gave the UKMO model lowest mean tracking errors for 1-day forecasts (143 km), and for 2-day forecasts (225 km), and for 3-day forecasts (335 km).

6.3. Concluding Remarks

Reference was made in Section 6.2 to just one example among many fruitful "post-Beijing" collaborations, after accounts in Section 6.1 of that promising Aerosonde development which had also been stimulated by discussion in Beijing, and in Section 3.3 of some advances in understanding air-sea interaction at extreme wind speeds that were influenced by an unusual interdisciplinary blend of contributions to the ICSU/WMO "Tropical Cyclone Disasters" symposium. Thus collaboration on TC disasters between ICSU and WMO has continued fruitfully to stress the key contributions of

(a) oceanography in TC science; of
(b) improved initial conditions in TC forecasting; and of
(c) growing links between TC experts from Asia and from other continents.

Moreover, at the WMO/ICSU International Workshop on Tropical Cyclones (Hainan, China, April 1998), all of these points will be emphasized still further—in the ever exciting context of TC fluid mechanics.

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